Channel-reach morphology in mountain drainage basins

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ABSTRACT

A classification of channel-reach morphology in mountain drainage basins synthesizes stream morphologies into seven distinct reach types: colluvial, bedrock, and five alluvial channel types (cascade, step pool, plane bed, pool riffle, and dune ripple). Coupling reach-level channel processes with the spatial arrangement of reach morphologies, their links to hillslope processes, and external forcing by confinement, riparian vegetation, and woody debris defines a process-based framework within which to assess channel condition and response potential in mountain drainage basins. Field investigations demonstrate characteristic slope, grain size, shear stress, and roughness ranges for different reach types, observations consistent with our hypothesis that alluvial channel morphologies reflect specific roughness configurations adjusted to the relative magnitudes of sediment supply and transport capacity. Steep alluvial channels (cascade and step pool) have high ratios of transport capacity to sediment supply and are resilient to changes in discharge and sediment supply, whereas low-gradient alluvial channels (pool riffle and dune ripple) have lower transport capacity to supply ratios and thus exhibit significant and prolonged response to changes in sediment supply and discharge. General differences in the ratio of transport capacity to supply between channel types allow aggregation of reaches into source, transport, and response segments, the spatial distribution of which provides a watershed-level conceptual model linking reach morphology and channel processes. These two scales of channel network classification define a framework within which to investigate spatial and temporal patterns of channel response in mountain drainage basins.

INTRODUCTION

Geologists and engineers have long recognized fundamental differences between mountain channels and their lowland counterparts (e.g., Surell, 1841; Dana, 1850; Shaler, 1891). In contrast to self-formed flood-plain channels, the gradient and morphology of mountain channels are tremendously variable and prone to forcing by external influences. Although mountain channels provide important aquatic habitat (e.g., Nehlsen et al., 1991; Frissell, 1993), supply sediment to estuaries and the oceans (e.g., Milliman and Syvitski, 1992), and transmit land use disturbances from headwater areas down through drainage networks (e.g., Reid, 1993), they have received relatively little study compared to lowland rivers.

Improved ability to relate morphology and processes in mountain channels would facilitate understanding and predicting their response to both human and natural disturbance. Classification schemes can organize such understanding into conceptual models that provide further insight into channel processes (e.g., Schumm, 1977). With few exceptions (e.g., Paustian et al., 1992; Whiting and Bradley, 1993), classifications of mountain channels are not process based, which compromises their use for assessing channel condition, response potential, and relations to ecological processes.

In order to provide a useful general classification of mountain channels, a typology should be applicable on more than a regional basis, yet adaptable to regional variability; otherwise proliferation of regional channel classifications could impede rather than enhance communication and understanding. Moreover, a classification should rely on aspects of channel form that reflect channel processes. Furthermore, it should encompass the whole channel network, rather than consider only channels inhabited by desirable organisms or indicator species. A process-based understanding of spatial linkages within a watershed is essential for assessment of channel condition, prediction of channel response to disturbance, and interpretation of the causes of historical channel changes.

Herein we systematize a channel classification that expands on Schumm's (1977) general delineation of erosion, transport, and deposition reaches and provides a framework for examining channel processes in mountain drainage basins. We also report a field test of the classification using data from drainage basins in Oregon and Washington and propose a genetic explanation for the distinct channel morphologies that we recognize. The tie to channel processes and morphogenesis provides a defensible theoretical and conceptual framework within which to classify channel morphology, assess channel condition, and interpret response potential. In particular, coupling of process-based channel classification with landscape-specific spatial linkages can provide insight into how disturbances propagate through drainage basins. Our classification arose from field work in mountain drainage basins where we repeatedly observed the same general sequence of channel morphologies down through the channel network. Here we draw on previous work and our own field observations to discuss these morphologies and propose a theory for the origin of distinct alluvial channel types. Although developed based on literature review and field observations in the Pacific Northwest (Montgomery and Buffington, 1993), subsequent field work confirms the relevance of the classification in other mountainous regions.

Channel-reach Morphology

A voluminous literature on channel classification attests to the wide variety of morphologies exhibited by stream channels. No single classification can satisfy all possible purposes, or encompass all possible channel types; each of the channel classifications in common use have advantages and disadvantages for use in geological, engineering, and ecological applications (see discussion in Kondolf, 1995). Although stream channels possess a continuum of characteristics identifiable at spatial scales that range from individual channel units to entire drainage basins (Frissell et al., 1986), channel reaches of at least 10 to 20 channel widths in length define a useful scale over which to relate stream morphology to channel processes, response potential, and habitat characteristics.

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TABLE 1. DIAGNOSTIC FEATURES OF EACH CHANNEL TYPE

	Dune ripple	Pool riffle	Plane bed	Step pool	Cascade	Bedrock	Colluvial
Typical bed material	Sand	Gravel	Gravel-cobble	Cobble-boulder	Boulder	Rock	Variable
Bedform pattern	Multilayered	Laterally oscillatory	Featureless	Vertically oscillatory	Random	Irregular	Variable
Dominant roughness elements	Sinuosity, bedforms (dunes, ripples, bars) grains, banks	Bedforms (bars, pools), grains, sinuosity, banks	Grains, banks	Bedforms (steps, pools), grains, banks	Grains, banks	Boundaries (bed and banks)	Grains
Dominant sediment sources	Fluvial, bank failure	Fluvial, bank failure	Fluvial, bank failure, debris flows	Fluvial, hillslope, debris flows	Fluvial, hillslope, debris flows	Fluvial, hillslope, debris flows	Hillslope, debris flows
Sediment storage elements	Overbank, bedforms	Overbank, bedforms	Overbank	Bedforms	Lee and stoss sides of flow obstructions	Pockets	Bed
Typical confinement	Unconfined	Unconfined	Variable	Confined	Confined	Confined	Confined
Typical pool spacing (channel widths)	5 to 7	5 to 7	None	1 to 4	<1	Variable	Unknown

We recognize three primary channel-reach substrates: bedrock, alluvium, and colluvium. Bedrock reaches lack a contiguous alluvial bed and reflect high transport capacities relative to sediment supply; they are typically confined by valley walls and have steep slopes. In contrast, alluvial channels exhibit a wide variety of morphologies and roughness configurations that vary with slope and position within the channel network, and may be either confined, with little to no associated flood plain, or unconfined, with a well-established flood plain. We recognize five distinct alluvial reach morphologies: cascade, step pool, plane bed, pool riffle, and dune ripple. Colluvial channels form an additional reach type that we recognize separately from alluvial channels, despite the common presence of a thin alluvial substrate. Colluvial channels typically are small headwater streams that flow over a colluvial valley fill and exhibit weak or ephemeral fluvial transport. Each of these channel types is distinguished by a distinctive channel-bed morphology, allowing rapid visual classification. Diagnostic features of each channel type are summarized in Table 1 and discussed below.

Cascade Channels

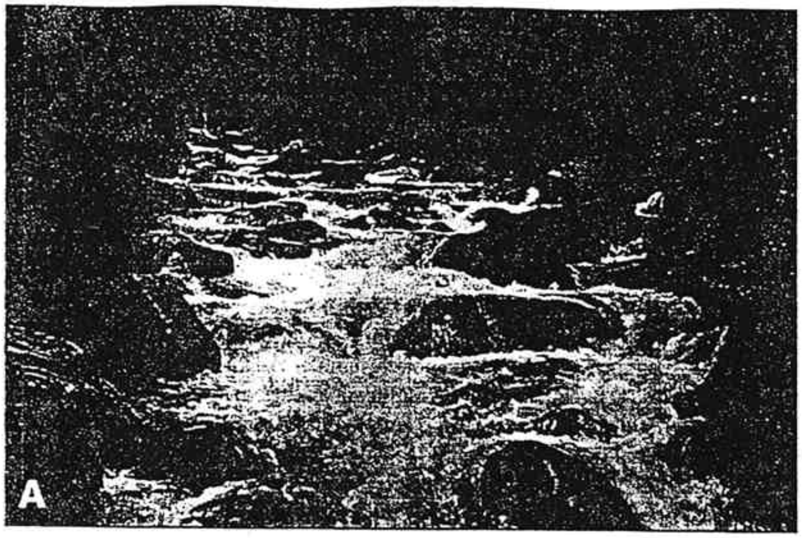
The term "cascade" connotes tumbling flow, although its specific morphologic definition varies and often is applied to both channel units and reaches (e.g., Bisson et al., 1982; Grant et al., 1990). Our delineation of cascade channels focuses on streams in which energy dissipation is dominated by continuous tumbling and jet-and-wake flow over and around individual large clasts (e.g., Peterson and Mohanty, 1960) (Fig. 1A). Cascade channels generally occur on steep slopes, are narrowly confined by valley walls, and are characterized by longitudinally and laterally disorganized bed material typically consisting of cobbles and boulders (Fig. 2A). Small, partially channel-spanning pools spaced less than a channel width apart are common in cascade channels. Tumbling flow over individual grain steps and turbulence associated with jet-and-wake flow around grains dissipates much of the mechanical energy of the flow (Fig. 3A).

Large particle size relative to flow depth makes the largest bed-forming material of cascade reaches effectively immobile during typical flows. Studies of steep-gradient channels report that large bed-forming grains typically become mobile only during infrequent (i.e., 50–100 yr) hydrologic events (Grant et al., 1990; Kondolf et al., 1991; Whittaker, 1987b). Mobilization of these larger clasts is accompanied by high sediment transport rates due to the release of finer sediment trapped under and around large grains (Sawada et al., 1983; Warburton, 1992). During lesser floods, gravel stored in low-energy sites is mobilized and travels as bedload over larger bed-forming clasts (Griffiths, 1980; Schmidt and Ergenzinger, 1992). Gravel and finer material are locally stored on stoss and lee sides of flow obstructions (i.e., large grains and large woody debris) due to physical impoundment and generation of velocity shadows. One tracer study (Kondolf et al., 1991) showed that material in such depositional sites was completely mobilized during a seven-year recurrence-interval event, whereas no tracer movement was observed during flows of less than the annual recurrence interval.

These observations suggest that there are two thresholds for sediment transport in cascade channels. During moderate recurrence-interval flows, bedload material is rapidly and efficiently transported over the more stable bed-forming clasts, which have a higher mobility threshold corresponding to more infrequent events. The lack of significant in-channel storage (Kondolf et al., 1991) and the rapid scour of depositional sites during moderately frequent high flows suggest that sediment transport is effectively supply limited in cascade channels. Bedload transport studies demonstrate that steep channels in mountain drainage basins are typically supply limited, receiving seasonal or stochastic sediment inputs (Nanson, 1974; Griffiths, 1980; Ashida et al., 1981; Whittaker, 1987). Because of this high transport capacity relative to sediment supply, cascade channels function primarily as sediment transport zones that rapidly deliver sediment to lower-gradient channels.

Step-Pool Channels

Step-pool channels are characterized by longitudinal steps formed by large clasts organized into discrete channel-spanning accumulations that separate pools containing finer material (Figs. 1B and 2B) (Ashida et al., 1976, 1981; Griffiths, 1980; Whittaker and Jaeggi, 1982; Whittaker and Davies, 1982; Whittaker, 1987a, 1987b; Chin, 1989; Grant et al., 1990). Primary flow and channel bed oscillations in step-pool reaches are vertical, rather than lateral, as in pool-riffle channels (Fig. 3B). The stepped morphology of the bed results in alternating critical to supercritical flow over steps and subcritical flow in pools (Bowman, 1977; Chin, 1989). Step-pool channels exhibit a pool spacing of roughly one to four channel widths (Bowman, 1977; Whittaker, 1987b; Chin, 1989; Grant et al., 1990), significantly less than the five to seven channel widths that typify self-formed pool-riffle channels (Leopold et al., 1964; Keller and Melhorn, 1978). Steps provide much of the elevation drop and roughness in step-pool channels (Ashida et al., 1976; Whittaker and Jaeggi, 1982; Whittaker, 1987a, 1987b; Chin, 1989). Step-pool morphology generally is associated with steep gradients, small width to depth ratios, and pronounced confinement by valley walls. Although step-forming clast sizes typically are comparable to annual high flow depths, a stepped longitudinal profile also may develop in steep sand-bedded channels (G. E. Grant, 1996, personal commun.).



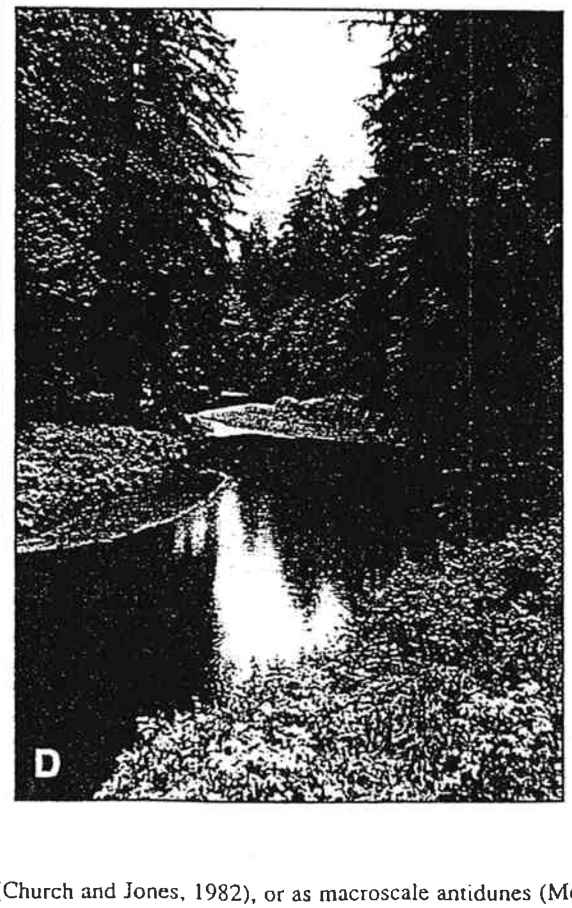
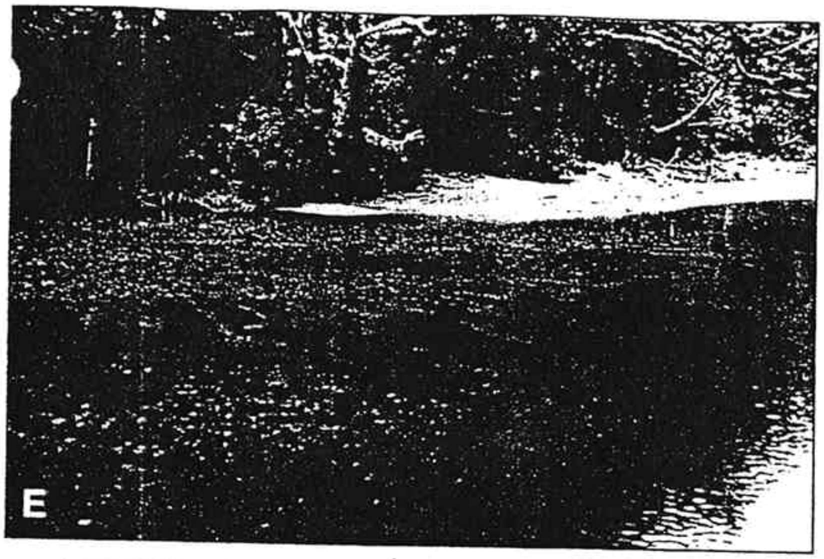


Figure 1. Alluvial channel-reach morphologies: (A) cascade; (B) step pool; (C) plane bed; (D) pool riffle; (E) dune ripple; (F) colluvial (channel in photo is 0.5 m wide); and (G) forced pool riffle.

Step-forming material may be viewed as either a kinematic wave (Langbein and Leopold, 1968), a congested zone of large grains that causes increased local flow resistance and further accumulation of large particles (Church and Jones, 1982), or as macroscale antidunes (McDonald a Banerjee, 1971; Shaw and Kellerhals, 1977; Grant and Mizuyama, 199 Step-pool sequences form through armoring processes under high d



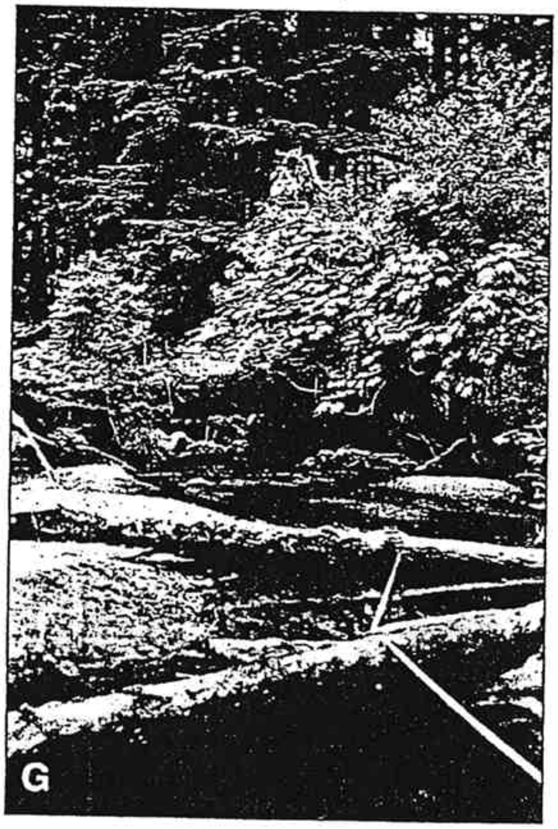
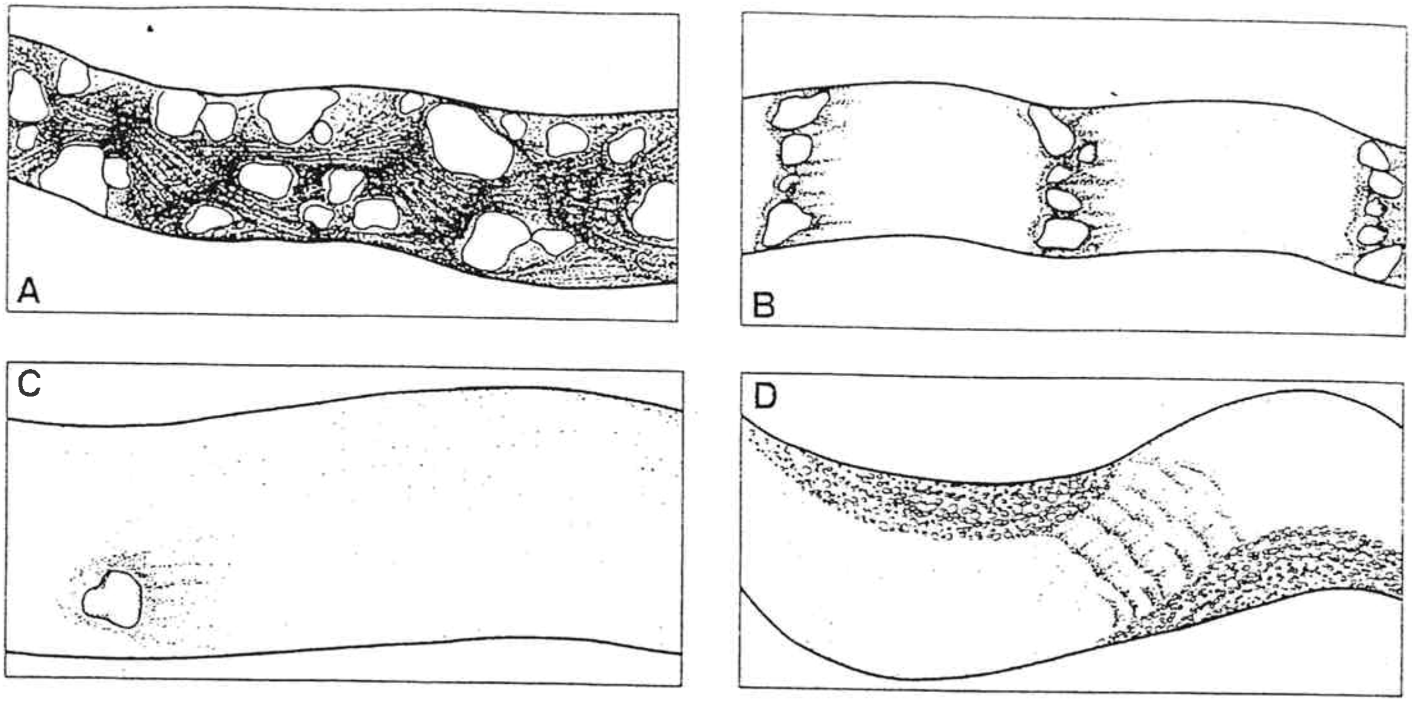


Figure 1. *(Continued—caption on facing page).*

charges and low sediment supply (Ashida et al., 1981; Whittaker and Jaeggi, 1982). Grant et al. (1990) suggested that low sediment supply and infrequent discharges capable of moving the coarsest sediment are required for development of stepped-bed morphology, and Grant and Mizuyama (1991) suggested that step-pool formation requires a heterogeneous bed mixture and near-critical flow. Furthermore, step spacing corresponds to maximum resistance, providing stability for a bed that would otherwise be mo- (Whittaker and Jaeggi, 1982; Abrahams et al., 1995).

Step-pool channels have several sediment transport thresholds. Large bed-forming material generally is mobile only during relatively infrequent hydrologic events (Whittaker, 1987a, 1987b; Grant et al., 1990), although Warburton (1992) showed that step-forming clasts in steep proglacial channels may be mobile annually. Significant movement of all grain sizes occurs during extreme floods, and step-pool morphology is reestablished during the falling limb of the hydrograph (Sawada et al., 1983; Whittaker, 1987b; Warbuton, 1992). During more frequent discharges, finer material stored in pools travels as bedload over stable bed-forming clasts (Ashida et al., 1981; Whittaker, 1987a, 1987b; Ergenzinger and Schmidt, 1990; Grant et al., 1990; Schmidt and Ergenzinger, 1992). In a series of tracer tests in a step-pool channel, Schmidt and Ergenzinger (1992) found that all of the tagged particles placed in pools mobilized during frequent, moderate discharges and were preferentially redeposited into pools. Transport of all the pool-filling material indicates that sediment transport of non–step-forming grains is supply limited. Bedload studies in step-pool channels demonstrate complex relations between discharge and sediment transport; transport rates are dependent on seasonal and stochastic sediment inputs, flow magnitude and duration, and antecedent events (Nanson, 1974; Griffiths, 1980; Ashida et al., 1981; Sawada et al., 1983; Whittaker, 1987a, 1987b; Warburton, 1992). Ashida et al. (1981), for example, observed a 10 hr lag between the hydrograph peak and onset of bedload transport for step-pool channels scoured of all pool-filling sediment during previous storms. Hydrograph peaks and bedload transport were, however, directly correlated during a subsequent storm due to the availability of sediment deposited in pools. Warburton (1992) suggested three phases of sediment transport in step-pool channels: a low-flow flushing of fines; frequent high-flow mobilization of pool-filling gravel (also noted by Sawada et al., 1983); and less-frequent higher-discharge mobilization of step-forming grains.

Although step-pool and cascade channel morphologies both reflect supply-limited transport, they are distinguished by differences in the spatial



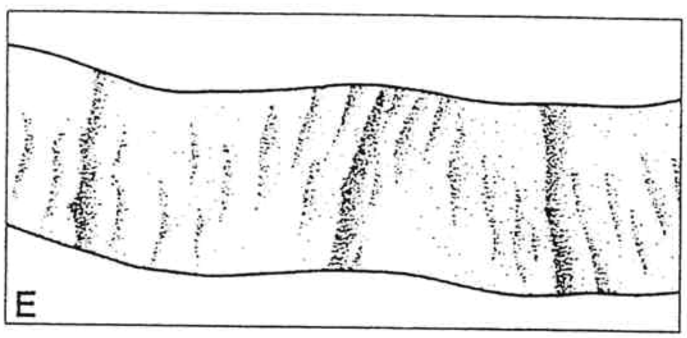


Figure 2. Schematic planform illustration of alluvial channel morphologies at low flow: (A) cascade channel showing nearly continuous, highly turbulent flow around large grains; (B) step-pool channel showing sequential highly turbulent flow over steps and more tranquil flow through intervening pools; (C) plane-bed channel showing single boulder protruding through otherwise uniform flow; (D) pool-riffle channel showing exposed bars, highly turbulent flow through riffles, and more tranquil flow through pools; and (E) dune-ripple channel showing dune and ripple forms as viewed through the flow.

density and organization of large clasts. Step-pool channels are defined by discrete channel-spanning steps less than a channel width in length that separate pools spaced every one to four channel widths. Cascade channels are defined by ubiquitous tumbling and jet-and-wake flow over a series of individual large clasts that together exceed a channel width in length, with small, irregularly placed pools spaced less than a channel width apart. The regular sequence of pools and steps in step-pool channels probably represents the emergence of a fluvially organized morphology in alluvial channels. In contrast, the disorganized large clasts of cascade channels may include lag deposits forced by nonfluvial processes (e.g., debris flows, glaciers, and rock falls).

Plane-Bed Channels

The term "plane bed" has been applied to both planar bed phases observed to form in sand-bed channels (Simons et al., 1965) and planar gravel and cobble-bed channels (Florsheim, 1985) like the coarse-grained, threshold canals described by Lane and Carlson (1953). Our use of the term refers to the latter and encompasses glide (run), riffle, and rapid morphologies described in the fisheries literature (e.g., Bisson et al., 1982). Plane-bed channels lack discrete bars, a condition that is associated with low width to depth ratios (Sukegawa, 1973; Ikeda, 1975, 1977) and large values of relative roughness (ratio of 90th percentile grain size to bankfull flow depth). Church and Jones (1982) considered bar formation unlikely at relative roughnesses of 0.3 to 1.0. Plane-bed reaches occur at moderate to high slopes in relatively straight channels that may be either unconfined or confined by valley walls. They typically are composed of sand to small boulder grain sizes, but are dominantly gravel to cobble bedded.

Plane-bed channels differ morphologically from both step-pool and pool-riffle channels in that they lack rhythmic bedforms and are characterized by long stretches of relatively featureless bed (Figs. 1C and 2C). The absence of tumbling flow and smaller relative roughness distinguish plane-bed reaches from cascade and step-pool channels (Fig. 3C). Plane-bed channels lack sufficient lateral flow convergence to develop pool-riffle morphology due to lower width to depth ratios and greater relative roughness, which may decompose lateral flow into smaller circulation cells. However, introduction of flow obstructions may force local pool and bar formation.

Plane-bed channels typically exhibit armored bed surfaces calculated to have a near-bankfull threshold for mobility, although elevated sediment loading can cause textural fining and a lower calculated mobility threshold (Buffington, 1995). Plane-bed channels with armored bed surfaces indicate a transport capacity greater than sediment supply (i.e., supply-limited conditions), whereas unarmored surfaces indicate a balance between transport capacity and sediment supply (Dietrich et al., 1989). Nevertheless, beyond

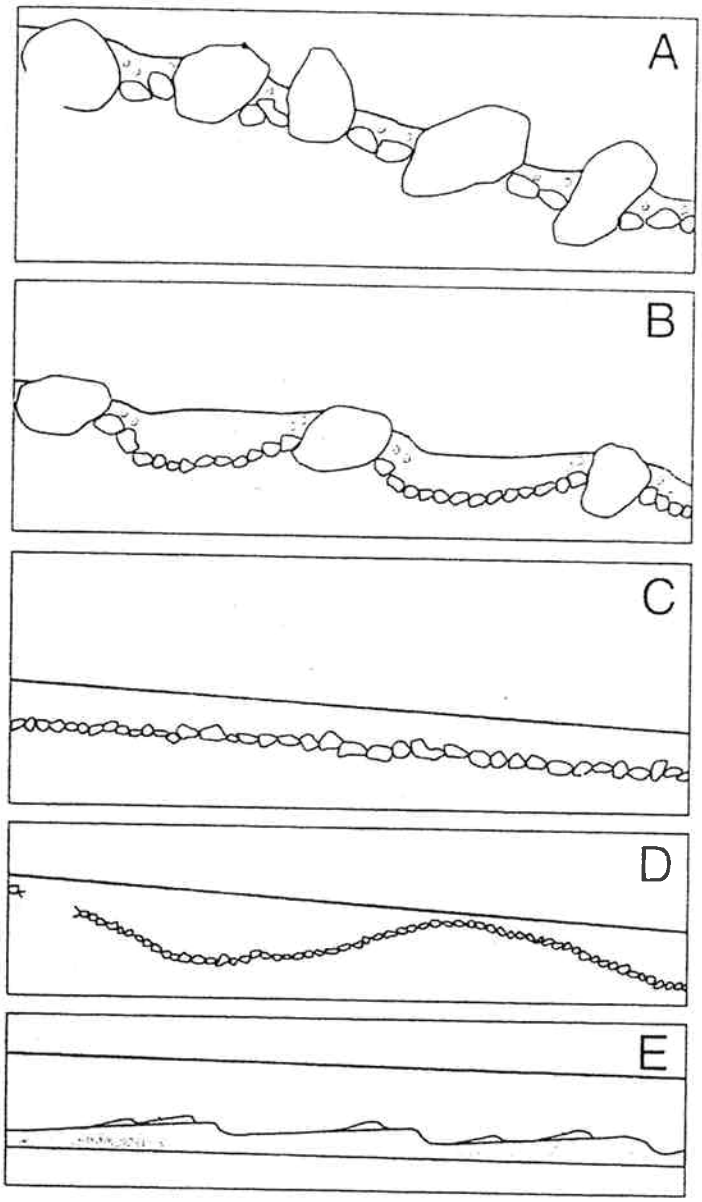


Figure 3. Schematic longitudinal profiles of alluvial channel morphologies at low flow: (A) cascade; (B) step pool; (C) plane bed; (D) pool riffle; and (E) dune ripple.

the threshold for significant bed-surface mobility, many armored gravel-bedded channels exhibit a general correspondence between bedload transport rate and discharge (e.g., Milhous, 1973; Jackson and Beschta, 1982; Sidle, 1988), implying transport-limited conditions. The above observations suggest that plane-bed channels are transitional between supply- and transport-limited morphologies.

Pool-Riffle Channels

Pool-riffle channels have an undulating bed that defines a sequence of bars, pools, and riffles (Leopold et al., 1964) (Fig. 1D). This lateral bedform oscillation distinguishes pool-riffle channels from the other channel types discussed above (Fig. 2D). Pools are topographic depressions within the channel and bars are corresponding high points (Fig. 3D); these bedforms are thus defined relative to each other (O'Neill and Abrahams, 1984). Pools are rhythmically spaced about every five to seven channel widths in self-formed, pool-riffle channels (Leopold et al., 1964; Keller and Melhorn, 1978), but channels with a high loading of large woody debris exhibit smaller pool spacing (Montgomery et al., 1995). Pool-riffle channels occur at moderate to low gradients and are generally unconfined, and have well-established flood plains. Substrate size in pool-riffle streams varies from sand to cobble, but typically is gravel sized.

Bar and pool topography generated by local flow convergence and divergence may be either freely formed by cross-stream flow and sediment transport, or forced by channel bends and obstructions (e.g., Lisle, 1986). Free-formed pool-riffle sequences initially result from internal flow perturbation that causes flow convergence and scour on alternating banks of the channel; concordant downstream flow divergence results in local sediment accumulation in discrete bars. Topographically driven convective accelerations reinforce convergent and divergent flow patterns, and thus pool-riffle morphogenesis (Dietrich and Smith, 1983; Dietrich and Whiting, 1989; Nelson and Smith, 1989). Alluvial bar development requires a sufficiently large width to depth ratio and small grain sizes that are easily mobilized and stacked by the flow (Church and Jones, 1982). Bar formation in natural channels appears to be limited to gradients ≤0.02 (Ikeda, 1977; Florsheim, 1985), although flume studies indicate that alternate bars may form at steeper gradients (Bathurst et al., 1983; Lisle et al., 1991). Bedform and grain roughness provide the primary flow resistance in free-formed pool-riffle channels.

Pool-riffle channels have heterogeneous beds that exhibit a variety of sorting and packing, commonly with a coarse surface layer and a finer subsurface (Leopold et al., 1964; Milhous, 1973). Armored gravel-bed channels typically exhibit a near-bankfull threshold for general and significant bed-surface mobility (e.g., Parker et al., 1982; Jackson and Beschta, 1982; Andrews, 1984; Carling, 1988; Buffington, 1995). Movement of surface grains releases fine sediment trapped by larger grains and exposes finer subsurface sediment to the flow, contributing to a steep rise in bedload transport with increasing shear stress (Milhous, 1973; Jackson and Beschta, 1982; Emmett, 1984). Bed movement is sporadic and discontinuous, depending on grain protrusion (Fenton and Abbott, 1977; Kirchner et al., 1990), friction angle (Kirchner et al., 1990; Buffington et al., 1992), imbrication (Komar and Li, 1986), degree of burial (Hammond et al., 1984; Buffington et al., 1992), and turbulent high-velocity sweeps of the channel bed. Very rarely is the whole bed in motion, and material eroded from one riffle commonly is deposited on a proximal downstream riffle.

Pool-riffle channels, like plane-bed channels, exhibit a mixture of supply- and transport-limited characteristics depending on the degree of bed-surface armoring and consequent mobility thresholds. Unarmored pool-riffle channels indicate a balance between transport capacity and sediment supply, while armored surfaces represent supply-limited conditions (e.g., Dietrich et al., 1989). Nevertheless, during armor-breaching events, bedload transport rates are generally correlated with discharge, demonstrating that sediment transport is not limited by supply once the bed is mobilized. Considerable fluctuations in observed transport rates, however, reflect a stochastic component of grain mobility caused by grain interactions, turbulent sweeps, and transient grain entrapment by bedforms (Jackson and Beschta, 1982; Sidle, 1988). Magnitudes of bedload transport also may vary for similar discharge events, depending on the chronology of antecedent transport events (Milhous, 1973; Reid et al., 1985; Sidle, 1988). Although both pool-riffle and plane-bed channels display a mix of supply- and transport-limited characteristics, the presence of depositional barforms in pool-riffle channels suggests that they are generally more transport limited than plane-bed channels. The transport-limited character of both of these morphologies, however, contrasts with the more supply-limited character of step-pool and cascade channels.

Dune-Ripple Channels

Dune-ripple morphology is most commonly associated with low-gradient, sand-bed channels (Figs. 1E, 2E, and 3E). A flow regime–dependent succession of mobile bedforms provides the primary hydraulic resistance in dune-ripple channels (e.g., Kennedy, 1975). However, even gravel-bed channels can exhibit a succession of multiple-scale bedforms during extreme discharges (e.g., Griffiths, 1989; Dinehart, 1992; Pitlick, 1992). The bedform configuration of dune-ripple channels depends on flow depth, velocity, bed-surface grain size, and sediment transport rate (e.g., Gilbert, 1914; Middleton and Southard, 1984), but generally follows a well-known morphologic sequence with increasing flow depth and velocity: lower-regime plane bed, ripples, sand waves, dunes, upper-regime plane bed, and antidunes (Gilbert, 1914; Simons et al., 1965; Harms et al., 1975). In channels transporting moderately to poorly sorted sediment, migrating bedload sheets composed of thin accumulations of sediment also may develop (Whiting et al., 1988). Several scales of bedforms may coexist in a dune-ripple channel; ripples, bedload sheets, and small dunes may climb over larger mobile dunes. A complete theoretical explanation for the development of such multiple-scale bedforms does not yet exist, but they are typically associated with low relative roughness. Dune-ripple channels also exhibit point bars or other bedforms forced by channel geometry. In contrast to the threshold sediment transport of plane-bed and pool-riffle streams, dune-ripple channels exhibit "live bed" transport (e.g., Henderson, 1963), in which significant sediment transport occurs at most stages. Hence, dune-ripple channels are effectively transport limited. The frequency of bed mobility and the presence of ripples and/or dunes distinguish dune-ripple channels from pool-riffle channels.

Colluvial Channels

Colluvial channels are small headwater streams at the tips of a channel network that flow over a colluvial valley fill and exhibit weak or ephemeral fluvial transport (Fig. 1F). Little research has focused on colluvial channels, even though first-order channels compose approximately half of the total length of a channel network (Montgomery, 1991). Dietrich et al. (1982) recognized that shallow flows in headwater channels have little opportunity for scour, and therefore sediment delivered from neighboring hillslopes generally accumulates to form colluvial valley fills. Benda and Dunne (1987) examined sediment in steep headwater valleys in the Oregon Coast Range and concluded that beneath a water-worked coarse surface layer, the valley fill consists of relatively unsorted colluvium delivered from surrounding hillslopes. Shallow and ephemeral flow in colluvial channels appears insufficient to mobilize all of the colluvial sediment introduced to the channel, resulting in significant storage of this material (Dietrich and Dunne, 1978; Dietrich et al., 1982; Benda, 1990). Large clasts, woody debris, bedrock steps, and in-channel vegetation further reduce the energy available for sediment transport in colluvial channels. Intermittent flow may rework some portion of the surface of the accumulated material, but it does not govern deposition, sorting, or transport of the valley fill.

Episodic transport by debris flows may account for most of the sediment transport in steep headwater channels. A sediment budget for a small basin in northern California indicated that debris flows account for more than half of the long-term sediment yield (Lehre, 1982). Swanson et al. (1982) estimated that only 20% of the total sediment yield from a first-order channel in the Cascade Range is accommodated by fluvial transport. Hence, the long-term sediment flux from low-order channels in steep terrain appears to be dominated by debris-flow processes. Differences in channel profiles support the hypothesis that different processes dominate the erosion of steep headwater channels and lower-gradient alluvial channels in the Oregon Coast Range (Seidl and Dietrich, 1992).

Dietrich and Dunne (1978) recognized that the residence time of sediment in headwater debris-flow–prone channels was on the order of hundreds of years. Kelsey (1980) also estimated that the sediment stored in first- and second-order channels is scoured by debris flows every 300 to 500 yr. Benda (1990) proposed a conceptual model for the evolution of channel morphology in steep headwater channels that involves cyclical alteration of bed morphology from gravel to boulder to bedrock in response to episodic sediment inputs. The accumulation of colluvial valley fills during periods between catastrophic scouring events indicates that transport capacity, rather than sediment supply, limits fluvial transport in colluvial channels.

Bedrock Channels

Bedrock channels lack a continuous alluvial bed. Although some alluvial material may be temporarily stored in scour holes, or behind flow obstructions, there is little, if any, valley fill. Hence, bedrock channels generally are confined by valley walls. Evidence from both anthropogenic badlands and mountain drainage basins indicates that bedrock channels are steeper than alluvial channels having similar drainage areas (Howard and Kerby, 1983; Montgomery et al., 1996). It is reasonable to adopt Gilbert's (1914) hypothesis that bedrock channels lack an alluvial bed due to high transport capacity associated with steep channel gradients and/or deep flow. Although bedrock channels in low-gradient portions of a watershed reflect a high transport capacity relative to sediment supply, those in steep portions of a watershed may also reflect recent catastrophic scouring.

Forced Morphologies

Flow obstructions can force a reach morphology that differs from the free-formed morphology for a similar sediment supply and transport capacity. In forested mountain drainage basins, for example, large woody debris may force local scour, flow divergence, and sediment impoundment that respectively form pools, bars, and steps (Fig. 1G). In an extreme example, Montgomery et al. (1996) found that log jams forced alluvial streambeds in otherwise bedrock reaches of a mountain channel network in western Washington.

Forced pool-riffle and step-pool channels are the most common obstruction-controlled morphologies in forested mountain drainage basins. A forced pool-riffle morphology is one in which most pools and bars are forced by obstructions such as large woody debris, and a forced step-pool channel is one in which large woody debris forms most of the channel-spanning steps that define the bed morphology. Forced morphologies can extend beyond the range of conditions characteristic of analogous free-formed morphologies (i.e., to steeper gradients and/or lower sediment supply). We recognize forced morphologies as distinct channel types because interpretation of whether such obstructions govern bed morphology is important for understanding channel response.

Intermediate and Other Morphologies

The channel types described above represent identifiable members along a continuum that includes several intermediate morphologies: riffle bar (pool riffle–plane bed); riffle step (plane bed–step pool); and cascade pool (step pool–cascade). Mixed alluvial and bedrock reaches exhibit subreach scale variations in alluvial cover. In our experience, however, it is simple to replicate identification of the seven basic reach types, even though they lie within a continuum of channel morphologies. Whether intermediate channel types are useful for classification purposes depends on the context of the application. Although our proposed classification does not cover all reach types in all environments (e.g., estuarine, cohesive-bed, or vegetated reaches), we have found it to be applicable in a variety of mountain environments.

TABLE 2. STUDY AREA CHARACTERISTICS

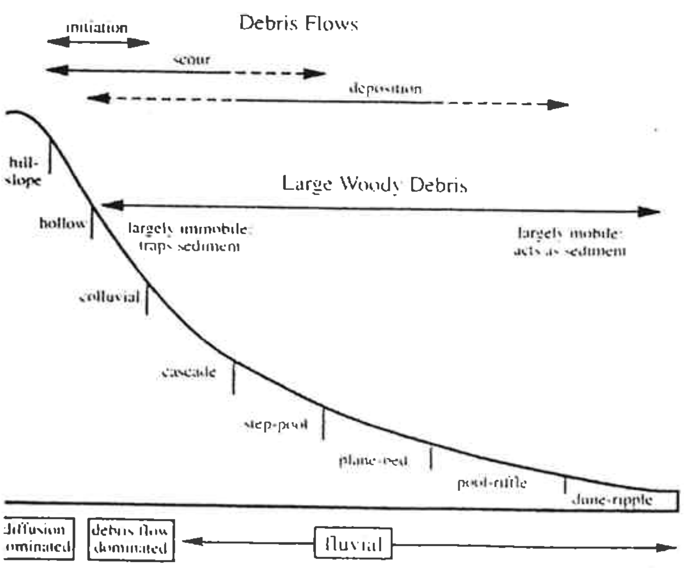
Study area	Geology	Drainage area (km^2)	Relief (m)	Land use
Finney Creek, Washington	Phyllite, greenschist, glacial sediments	128	1476	U.S. Forest Service, state forestry
Boulder River, Washington	Phyllite, glacial sediments	63	1985	U.S. Forest Service wilderness area
South Fork Hoh River, Washington	Sandstone, glacial sediments	129	>882	State forestry, national park
Deton Creek, Oregon	Sandstone	8	327	Private forestry

IELD TEST

Process differences associated with reach morphology should result in stinct physical characteristics for each reach type. Data compiled from eld studies in the Pacific Northwest reveal systematic association of chan-el types with slope, drainage area, relative roughness, and bed-surface ain size. Furthermore, these data suggest an explanation for the origin of stinct channel types.

udy Areas and Methods

Field surveys were conducted in four drainage basins in western Wash-gton and coastal Oregon: Finney Creek, Boulder River, South Fork Hoh iver, and Deton Creek (Table 2). In each study area, channel reaches)–20 channel widths in length were surveyed throughout the drainage isin. Each reach was classified into one of the above-defined channel pes. Reach slopes were surveyed using either an engineering level or a ind level and stadia rod. Topographic surveys and channel-spanning peb-e counts of 100 grains (Wolman, 1954) were conducted at representative oss sections. Reach locations were mapped onto U.S. Geological Survey 24,(scale topographic maps from which drainage areas were measured ing gital planimeter. Reach slopes were determined from topographic aps for some additional reaches where morphologies were mapped, but ope and grain-size measurements were not collected. We also included in ir analysis data collected using similar field methods in related studies in western Washington and southeast Alaska (Montgomery et al., 1995; Buffington, 1995).



Fi‥re 4. Idealized long profile from hillslopes and unchanneled ll lownslope through the channel network showing the general stribution of alluvial channel types and controls on channel ocesses in mountain drainage basins.

Results

In each study area, there is a general downstream progression of reach types that proceeds as colluvial, cascade, step pool, plane bed or forced pool riffle, and pool riffle (Fig. 4); we encountered no dune-ripple reaches in the study basins, although we observed them in neighboring areas. Bedrock reaches occur at locally steep locations throughout the channel networks, and not all of these channel types are present in each watershed. Furthermore, the specific downstream sequence of reach types observed in each drainage basin reflects local factors controlling channel slope, discharge, sediment supply, bedrock lithology, and disturbance history.

Data from alluvial, colluvial, and bedrock reaches within each study basin define distinct fields on a plot of drainage area versus reach slope (Fig. 5). These data provide further evidence that, for a given drainage area, bedrock reaches have greater slopes, and hence greater basal shear stress and stream power, than either alluvial or colluvial reaches (Howard and Kerby, 1983; Montgomery et al., 1996). Alluvial reaches occur on slopes less than about 0.2 to 0.3, and different alluvial channel types generally segregate within an inversely slope-dependent band within which pool-riffle and plane-bed channels occur at the lowest slopes, and step-pool and cascade channels occur on steeper slopes. Colluvial reaches occur at lower drainage areas and extend to steeper slopes. Data from colluvial reaches define a relation between drainage area and slope that contrasts with that of lower-gradient alluvial reaches. This general pattern holds for each of the study basins, implying consistent differences among colluvial, alluvial, and bedrock reaches in mountain drainage basins.

The different drainage area–slope relation for colluvial and alluvial channel reaches implies fundamental differences in sediment transport processes. For equilibrium channel profiles, channel slope (S) and drainage area (A) are related by

$$S = KA^{-m/n} \qquad (1)$$

where K, m, and n are empirical variables that incorporate basin geology, climate, and erosional processes (e.g., Howard et al., 1994). A log-linear regression of reach slope and drainage area data from alluvial and colluvial channels in Finney Creek yields m/n values of 0.72 ± 0.08 (R^2 = 0.72) and 0.26 ± 0.05 (R^2 = 0.58), respectively, which implies long-term differences in sediment transport processes between these channel types. This correspondence between the inflection in the drainage area–slope relation and the transition from colluvial to alluvial channels is consistent with the interpretation that scour by debris flows is the dominant incisional process in colluvial channels (Benda, 1990; Seidl and Dietrich, 1992; Montgomery and Foufoula-Georgiou, 1993).

Although slope ranges of free-form alluvial channel types overlap, they have distinct medians and quartile ranges (Fig. 6). Examination of the composite slope distributions indicates that reaches with slopes of less than 0.015 are likely to have a pool-riffle morphology; reaches with slopes of

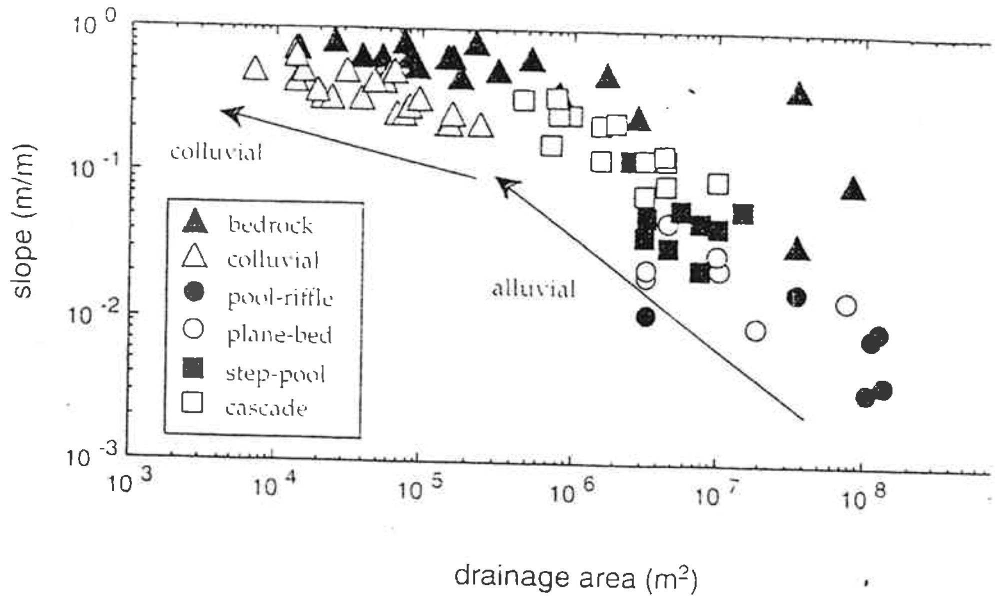


Figure 5. Drainage area versus reach slope for channels in the Finney Creek watershed, Washington.

0.015 to 0.03 typically have a plane-bed morphology; reaches with slopes of 0.03 to 0.065 are likely to have a step-pool morphology; and alluvial reaches with slopes greater than 0.065 typically have a cascade morphology. These core slope ranges define zones over which each channel type is the

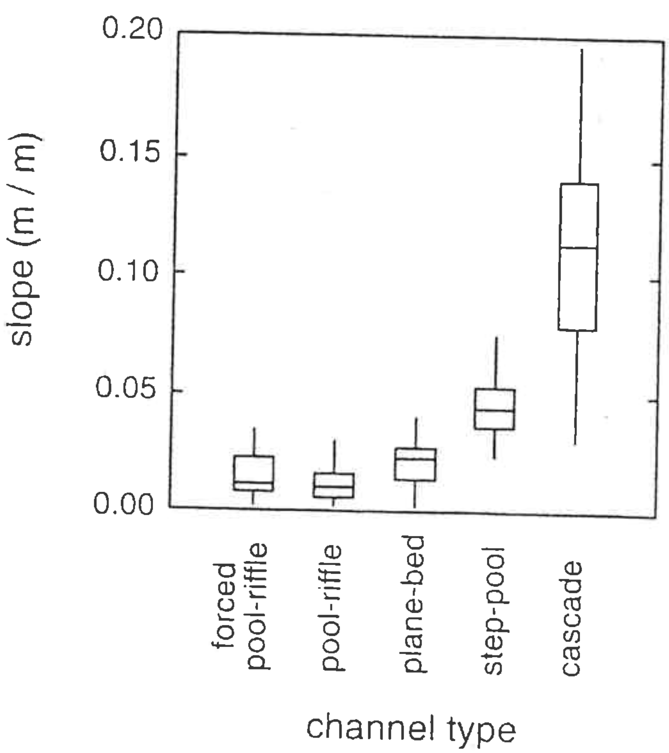


Figure 6. Composite slope distributions for channel reaches surveyed in this and related studies (Buffington, 1995; Montgomery et al., 1995); boxes represent inner and outer quartiles; vertical lines represent inner and outer tenths.

most likely to occur; however, the distributions overlap and channel ty[...] not uniquely related to reach slope. Furthermore, forced pool-riffle rea[...] span the slope ranges for pool-riffle and plane-bed reaches, indicating introduction of large woody debris can extend a forced morpholog[...] slopes where such a morphology would not be expected under low wo[...] debris loading (Montgomery et al., 1995). Nonetheless, the general se[...]gation of reach type by slope allows prediction of likely channel morph[...]ogy from topographic maps or digital elevation models.

Relative roughness (the ratio of the ninetieth percentile grain size to bankfull flow depth [d_{90}/D]) and reach slope together differentiate allu[...] reach types (Fig. 7): pool-riffle channels have relative roughness less t[...] about 0.3 and occur on slopes <0.03; plane-bed channels exhibit rela[...] roughness of roughly 0.2 to 0.8 on slopes of 0.01 to 0.04; step-pool reac[...] occur on steeper slopes and have relative roughness of 0.3 to 0.8; and the [...] of the largest clasts on the bed of steeper cascade reaches can approach th[...] of bankfull flow depth. Relative roughness and reach slope together prov[...] a reasonable stratification of channel morphology. In pool-riffle and pla[...] bed channels relative roughness increases rapidly with increasing slo[...] whereas there is little relation between relative roughness and slope [...] steeper step-pool and cascade reaches.

Composite bed-surface grain-size distributions for pebble counts fr[...] different channel types exhibit systematic coarsening from pool-rif[...] through cascade channels. For reaches in the Finney Creek waters[...] (Fig. 8), the median grain size increases from 17 mm for pool-riffle ch[...]nels to 80 mm for cascade morphologies, and d_{84} increases from 57 mm [...] 250 mm. These systematic changes in bed-surface grain-size distributi[...] indicate that progressive fining of the bed material accompanies the form[...]tion of different channel types downstream through a channel network.

The data reported above demonstrate that qualitatively defined chan[...] types exhibit quantitatively distinguishable characteristics. Our data furt[...] indicate that channel morphology is related to reach-average bankfull sh[...] stress (Fig. 9). Bedrock channels occur in reaches with the greatest sh[...] stress; cascade and step-pool reaches plot at lower values, which in turn [...] greater than those for plane-bed and pool-riffle channels. Hence, it appe[...] that, in part, local flow hydraulics influence the general distribution of ch[...]nel types in a watershed.

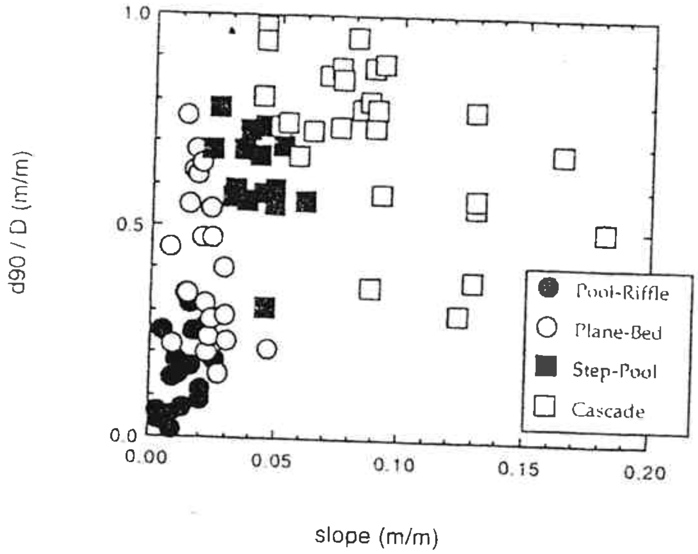


Figure 7. Composite plot of relative roughness (d_{90}/D) versus field surveyed reach slope for data from alluvial reaches in our study areas.

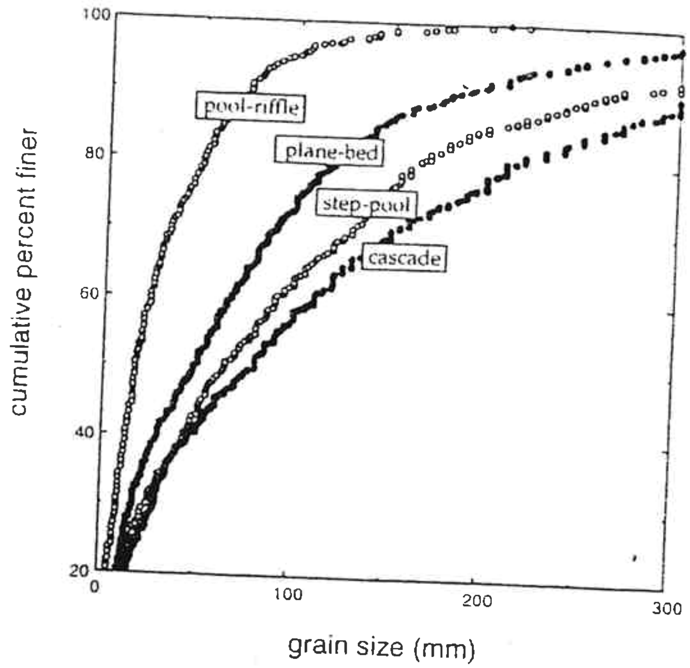


Figure 8. Aggregated cumulative grain-size distributions for alluvial channels of reaches with different bed morphologies in the Finney Creek watershed.

ORIGIN OF REACH-LEVEL MORPHOLOGIES

The typical downstream sequence of channel morphologies (Fig. 4) is accompanied by a progressive decrease in valley-wall confinement, which in stream-formed valleys may reflect opposing downstream trends of sediment supply (Q_s) and transport capacity (Q_c). Transport capacity is defined here as a function of the total boundary shear stress and is distinguished from the effective transport capacity (Q_c'), which is a function of the effective shear stress available for sediment transport after correction for shear stress dissipation caused by hydraulic roughness elements. Transport capacity generally decreases downstream due to the slope decreasing faster than the depth increases, whereas total sediment supply generally increases with drainage area, even though sediment yield per unit area often decreases (Fig. 10). This combination may result in long-term patterns of downstream deposition and development of wide flood plains and unconfined valleys. Insignificant sediment storage in a valley segment indicates that virtually all of the material delivered to the channel is transported downstream. In contrast, thick alluvial valley-fill deposits imply either a long-term excess of sediment supply over transport capacity, or an inherited valley fill.

These general patterns and our field observations discussed above lead us to propose that distinctive channel morphologies reflect the relative magnitude of transport capacity to sediment supply, which may be expressed as the ratio $q_r = Q_c/Q_s$. Colluvial channels are transport limited ($q_r << 1$), as in-

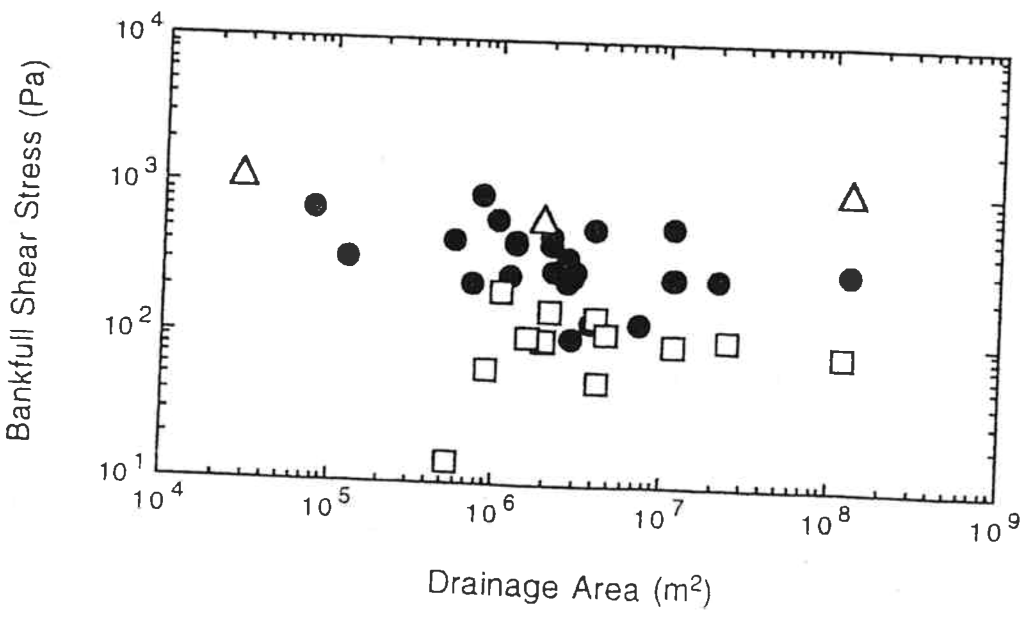


Figure 9. Plot of drainage area versus reach. Average shear stress for bedrock (triangles), cascade and step-pool (circles), and plane-bed and pool-riffle (squares) channel morphologies are from the South Fork Hoh River study area.

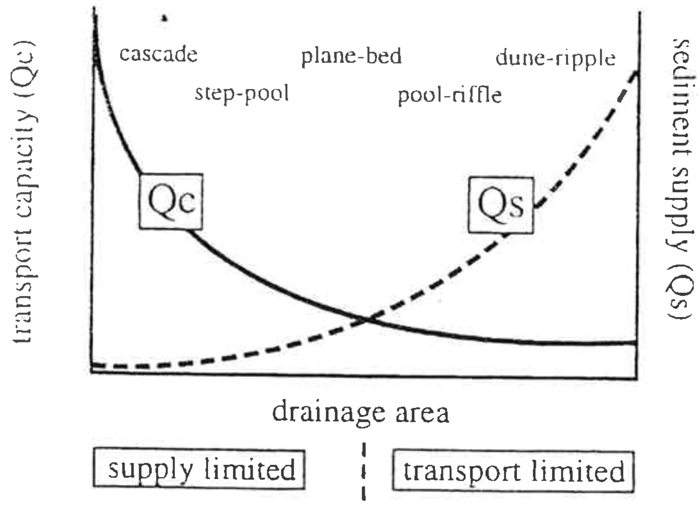


Figure 10. Schematic illustration of generalized relative trends in sediment supply (Q_s) and transport capacity (Q_c) in mountain drainage basins.

dicated by the accumulation of colluvium within valley bottoms. In contrast, the lack of an alluvial bed indicates that bedrock channels are supply limited ($q_r >> 1$). For a given drainage area (and thus Q_s), bedrock reaches have greater slopes and shear stresses (Figs. 5 and 9), implying that they have higher transport capacities and thus greater q_r values than other channel types. Alluvial channels, however, probably represent a broad range of q_r: steep alluvial channels (cascade and step-pool) have higher shear stresses (Fig. 9) and thus higher Q_c and q_r values for a given drainage area and sediment supply; the lower-gradient plane-bed and pool-riffle channels are transitional between $q_r > 1$ and $q_r = 1$, depending on the degree of armoring (e.g., Dietrich et al., 1989) and the frequency of bed-surface mobility; and the live-bed mobility of dune-ripple channels indicates that $q_r \leq 1$. The variety of alluvial channel morphologies probably reflects a broad spectrum of q_r expressed through fining and organization of the bedload (Fig. 11), which leads to formation of distinct alluvial bed morphologies that represent the stable bed form for the imposed q_r. This hypothesized relation between q_r and stable channel morphologies in mountain drainage basins provides a genetic framework for explaining reach-level morphologies that elaborates on Lindley's (1919) regime concept. An alluvial channel with $q_r > 1$ will become stable when the bed morphology and consequent hydraulic roughness produce an effective transport capacity that matches the sediment supply ($Q_c' = Q_s$).

Different channel types are stabilized by different roughness configurations that provide resistance to flow. In steep channels energy is dissipated primarily by hydraulic jumps and jet-and-wake turbulence. This style of energy dissipation is pervasive in cascade channels and periodic in step-pool channels. Skin friction and local turbulence associated with moderate particle sizes are sufficient to stabilize the bed for lower shear stresses characteristic of plane-bed channels. In pool-riffle channels, skin friction and bedform drag dominate energy dissipation. Particle roughness in dune-ripple channels is small due to the low relative roughness, and bedforms govern hydraulic resistance. The importance of bank roughness varies with channel type, depending on the width to depth ratio and vegetative influences, but in steep channels bank resistance is less important compared to energy dissipation caused by tumbling flow. These different roughness configurations represent a range in q_r values that varies from high in cascade reaches to low in dune-ripple channels.

Our hypothesis that different channel types represent stable roughness configurations for different q_r values implies that there should be an association of channel type and roughness. Even though the general correlation of morphology and slope (Fig. 6) implies discrete roughness characteristics among channel types, different channel morphologies occurring on the same slope should exhibit distinct roughness. Photographs and descriptions of channel morphology from previous studies in which roughness was determined from measured velocities (Barnes, 1967; Marcus et al., 1992) allow direct assessment of the roughness associated with different channel types. For similar slopes, plane-bed channels exhibit greater roughness than pool-riffle channels, and step-pool channels, in turn, appear to have greater roughness than plane-bed channels with comparable gradients (Fig. 12). Moreover, intermediate morphology reaches plot between their defining channel types. These systematic trends in roughness for a given slope strongly support the hypothesis that reach-level channel morphology reflects a dynamic adjustment of the bed surface to the imposed shear stress and sediment supply (i.e., the specific q_r value).

CHANNEL DISTURBANCE AND RESPONSE POTENTIAL

Natural and anthropogenic disturbances that change hydrology, sediment supply, riparian vegetation, or large woody debris loading can alter channel processes and morphology. The effect that watershed disturbance has on a particular channel reach depends on hillslope and channel coupling, the sequence of upstream channel types, and site-specific channel morphology. In particular, the variety and magnitude of possible morphologic responses to

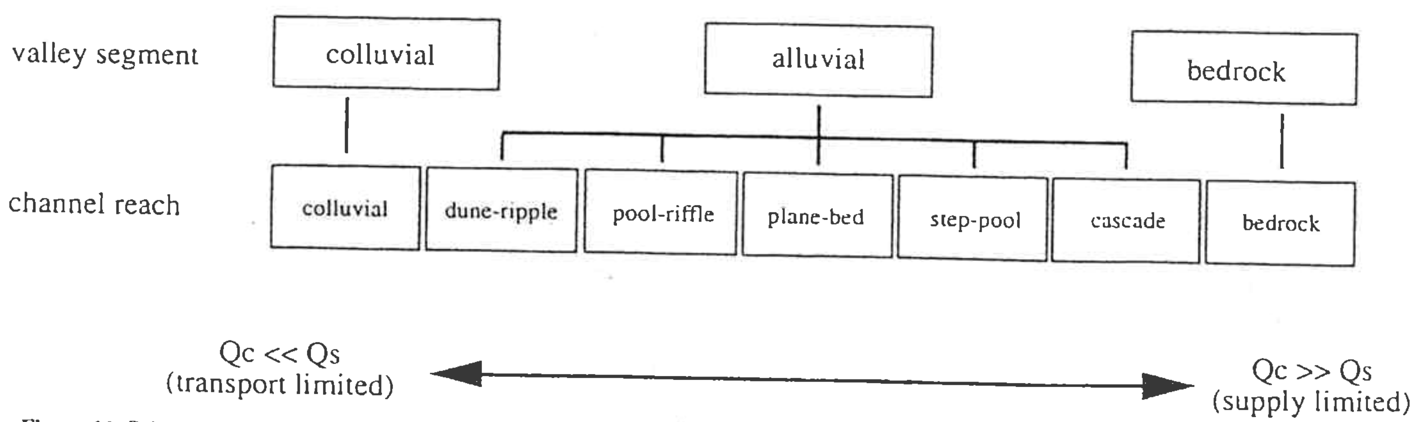


Figure 11. Schematic illustration of the transport capacities relative to sediment supply for reach-level channel types.

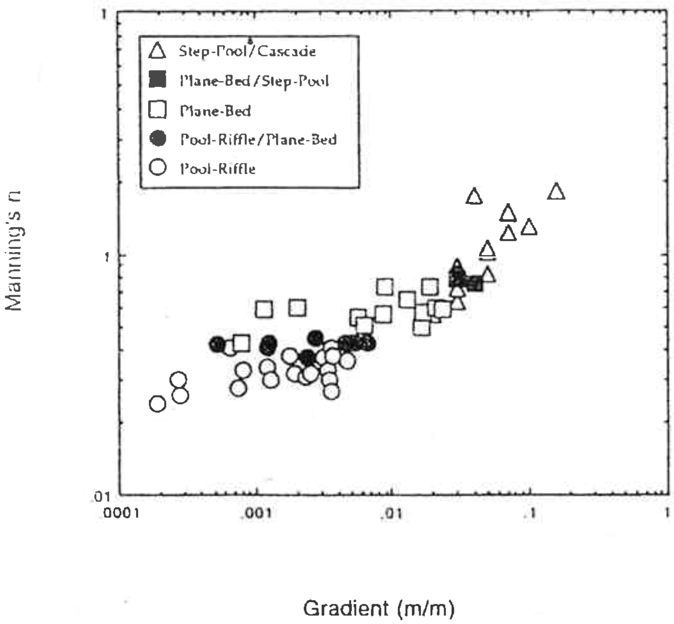


Figure 12. Plot of reach roughness coefficient (Manning's n) versus each slope for channels classified according to our system using data nd photographs in Barnes (1967) and Marcus et al. (1992). Note that hannel types interpreted to reflect greater relative transport capacity iave higher roughness over similar slopes.

. give- disturbance depend on channel type, external influences (e.g., coniner, riparian vegetation, large woody debris), and disturbance history. ogether these considerations provide an integrative approach for examinng spatial and temporal patterns of channel disturbance and response in nountain watersheds.

›patial Distribution of Channel Types

The spatial distribution of channel types and their coupling to both hill-lopes and one another can strongly influence the potential for a channel to e affected by a disturbance. In general, the degree of hillslope-channel couıling changes downstream through mountain channel networks, resulting in hanges in both the characteristics and delivery mechanisms of sediment upplied to a channel (e.g., Rice, 1994). Furthermore, the general downtream progression of channel morphologies in mountain drainage basins Fig. 4) causes an association of hillslope coupling and channel type. Headvater colluvial channels are strongly coupled to adjacent hillslopes, and net ediment transport from these weakly fluvial reaches is affected by the freuency of upslope debris flows and mass movements. Valley-wall confinenent allows direct sediment input by hillslope processes to cascade and stepool channels, which makes them prone to periodic disturbance from iillslope failures. Debris flows can dominate the disturbance frequency in ieadwater portions of the basin, scouring high-gradient channels and agrading the first downstream reach with a gradient low enough to cause depsition of the entrained material (e.g., Benda and Dunne, 1987). Consequently, the effects of debris-flow processes on channel morphology can be livided into those related to scour, transport, and deposition. Farther downtream, the coupling between hillslopes and lower-gradient channels (i.e., ıla: 1, pool-riffle, and dune-ripple) is buffered by wider valleys and depsitional flood plains, making these reaches less susceptible to direct disturance from hillslope processes. Sediment characteristics, delivery, and transport are generally dominated by fluvial processes in these lower-gradient channels, although forcing by large woody debris and impingement of channels on valley walls can have a significant influence on the local transport capacity and sediment supply (e.g., Rice, 1994).

The downstream sequence in which channel types are arranged also affects the potential for a disturbance to impact a particular reach. Position within the network and differences between q_r values allow general aggregation of channel reaches into source, transport, and response segments. In steep landscapes, source segments are transport-limited, sediment-storage sites subject to intermittent debris-flow scour (i.e., colluvial channels). Transport segments are morphologically resilient channels with a high q_r (i.e., bedrock, cascade, and step-pool channels) that rapidly convey increased sediment loads. Response segments are channels with a low q_r (i.e., plane-bed, pool-riffle, and dune-ripple) in which significant morphologic adjustment occurs in response to increased sediment supply. These distinctions build upon Schumm's (1977) concept of erosion, transport, and deposition zones within a watershed to provide a conceptual model that allows identification of reach-specific response potential throughout a channel network.

The spatial distribution of source, transport, and response segments governs the distribution of potential impacts and recovery times within a watershed. Downstream transitions from transport to response reaches define locations where impacts from increased sediment supply may be both pronounced and persistent. Transport segments rapidly deliver increased sediment loads to the first downstream reach with insufficient transport capacity to accommodate the additional load. Consequently, the "cumulative" effects of upstream increases in sediment supply may be concentrated in response segments where longer time and/or significant morphological change is required to transport the additional sediment. In this regard, reach-level classification identifies areas most sensitive to increases in upstream sediment inputs. Hence, downstream transitions from transport to response segments can provide ideal locations to monitor network response and should serve as critical components of watershed monitoring studies. Most important, the relation between channel type and response potential provides a direct link between upstream sediment inputs and downstream response. Identification of source, transport, and response segments thereby provides a context for examining connections between watershed modifications, impacts on channel morphology, and biological response.

Influence of Channel Type

Differences in confinement, transport capacity relative to sediment supply, and channel morphology influence channel response to perturbations in sediment supply and discharge. Thus, it is important to assess channel response potential in the context of reach type and location within a watershed. An understanding of reach morphologies, processes, and environments allows reach-specific prediction of the likely degree and style of response to a particular perturbation. Small to moderate changes in discharge or sediment supply can alter channel attributes (e.g., grain size, slope, and channel geometry); large changes can transform reach-level channel types. On the basis of typical reach characteristics and locations within mountainous watersheds, we assessed the relative likelihood of specific morphologic responses to moderate perturbations in discharge and sediment supply for each channel type (Table 3).

Channels with different bed morphology and confinement may have different potential responses to similar changes in discharge or sediment supply. Changes in sediment storage dominate the response of colluvial channels to altered sediment supply because of transport-limited conditions and low fluvial transport capacities (Table 3); depending on the degree of valley fill, increased discharge can significantly change channel geometry. In contrast, bedrock, cascade, and step-pool channels are resilient to most discharge or

TABLE 3. INTERPRETED REACH-LEVEL CHANNEL RESPONSE POTENTIAL TO MODERATE CHANGES IN SEDIMENT SUPPLY AND DISCHARGE

	Width	Depth	Roughness	Scour depth	Grain size	Slope	Sediment storage
Dune ripple	+	+	+	+	o	+	+
Pool riffle	+	+	+	+	+	+	+
Plane bed	p	+	p	+	+	+	p
Step pool	o	p	p	p	p	p	p
Cascade	o	o	p	o	p	o	o
Bedrock	o	o	o	o	o	o	o
Colluvial	p	p	o	p	p	o	+

Notes: +—likely, o—unlikely, p—possible

sediment-supply perturbations because of high transport capacities and generally supply-limited conditions. Many bedrock channels are insensitive to all but catastrophic changes in discharge and sediment load. Lateral confinement and large, relatively immobile, bed-forming clasts make channel incision or bank cutting unlikely responses to changes in sediment supply or discharge in most cascade and step-pool channels. Other potential responses in step-pool channels include changes in bedform frequency and geometry, grain size, and pool scour depths, whereas only limited textural response is likely in cascade channels. Lower gradient plane-bed, pool-riffle, and dune-ripple channels become progressively more responsive to altered discharge and sediment supply with decreasing q_r, smaller grain sizes, and less channel confinement. Because plane-bed channels occur in both confined and unconfined valleys, they may or may not be susceptible to channel widening or changes in valley-bottom sediment storage. Smaller, more mobile grain sizes in plane-bed and pool-riffle channels allow potentially greater response of bed-surface textures, scour depth, and slope compared to cascade and step-pool morphologies. Unconfined pool-riffle and dune-ripple channels generally have significant potential for channel geometry responses to perturbations in sediment supply and discharge. Changes in both channel and valley storage are also likely responses, as well as changes in channel roughness due to alteration of channel sinuosity and bedforms. There is less potential for textural response in dune-ripple than in pool-riffle and plane-bed channels simply because of smaller and more uniform grain sizes. At very high sediment supply, any of the above channel types may acquire a braided morphology (e.g., Mollard, 1973; Church, 1992). The general progression of alluvial channel types downstream through a channel network (Fig. 4) suggests that there is a systematic downstream increase in response potential to altered sediment supply or discharge.

The above predictions of response potential are largely conceptual, based on typical reach processes, characteristics, and locations within a drainage basin. Nevertheless, our approach provides a rational, process-based alternative to channel assessments based solely on descriptive typologic classification. For example, a channel-reach classification developed by Rosgen (1994) recognizes 7 major and 42 minor channel types primarily on the basis of bed material and slope; there is also the option of more detailed classification using entrenchment, sinuosity, width to depth ratio, and geomorphic environments. However, the classification lacks a basis in channel processes. The lack of an explanation of the rationale underlying Rosgen's (1994) assessment of response potential for each minor channel type emphasizes this shortcoming. Furthermore, Rosgen's (1994) classification combines reach morphologies that may have very different response potentials: Rosgen's (1994) C channels may include reaches with dune-ripple, pool-riffle, plane-bed, or forced pool-riffle morphologies; his B channels may include plane-bed, forced-pool riffle, and step-pool morphologies; and his A channels may include colluvial, cascade, and step-pool reaches. Although bed material and slope provide a convenient classification for many channels, the lack of a process-based methodology compromises such an approach to structuring channel assessments, predicting channel response, and investigating relations to ecological processes.

External Influences

Channel response potential also reflects external influences on channel morphology, the most prominent of which are confinement, riparian vegetation, and large woody debris loading. Valley-wall confinement limits changes in both channel width and flood-plain storage and maximizes channel response to increased discharge by limiting overbank flow. Although there is a general downstream correspondence between channel type and valley-wall confinement in many mountain watersheds, structural control and geomorphic history can force confinement in any portion of the channel network.

Riparian vegetation influences channel morphology and response potential by providing root strength that contributes to bank stability (e.g., Shaler, 1891; Gilbert, 1914), especially in relatively noncohesive alluvial deposits. The effect of root strength on channel bank stability is greatest in low gradient, unconfined reaches, where loss of bank reinforcement may result in dramatic channel widening (Smith, 1976). Riparian vegetation is also an important roughness source (e.g., Arcement and Schneider, 1989) that can mitigate the erosive action of high discharges.

Large woody debris provides significant control on the formation and physical characteristics of pools, bars, and steps (Heede, 1985; Lisle, 1986; Montgomery et al., 1995; Wood-Smith and Buffington, 1996), thereby influencing channel type and the potential for change in sediment storage and bedform roughness in response to altered sediment supply, discharge, or large woody debris loading. Woody debris may decrease the potential for channel widening by armoring stream banks; alternatively, it may aid bank erosion by directing flow and scour toward channel margins. Furthermore, bed-surface textures and their response potential are strongly controlled by hydraulic roughness resulting from in-channel wood and debris-forced bedforms (Buffington, 1995). Although large woody debris can force morphologic changes ranging from the scale of channel units to reaches, its impact depends on the amount, size, orientation, and position of debris, as well as channel size (Bilby and Ward, 1989; Montgomery et al., 1995) and rates of debris recruitment, transport, and decay (Bryant, 1980; Murphy and Koski, 1989). In general, individual pieces of wood can dominate the morphology of small channels, whereas debris jams are required to significantly influence channel morphology in larger rivers where individual pieces are mobile (Abbe and Montgomery, 1996). Thus, the relative importance of large woody debris in controlling channel morphology and response potential varies through a channel network.

Temporal Changes in Channel Morphology

The spatial pattern of channel types within a watershed provides a snapshot in time of a channel network, but history also influences the response potential of mountain channels, because past disturbance can condition channel response. Temporal variations in macroscopic channel morphology reflect (1) changes in large woody debris loading (e.g., Beschta, 1979;

Heede, 1985); (2) changes in discharge and sediment input (e.g., Hammer, 1972; Graf, 1975; Megahan et al., 1980; Coats et al., 1985); and (3) routing of ment waves through the channel network (e.g., Gilbert, 1917; Kelsey, 19 Church and Jones, 1982; Madej, 1982; Reid, 1982; Beschta, 1983).

Channels in which large woody debris forces pool formation and sediment storage are particularly sensitive to altered wood loading. For example, removal of large woody debris from forced pool-riffle channels may ead to either a pool-riffle or plane-bed morphology (Montgomery et al., 995). Similarly, loss of large woody debris may transform a forced step-ool channel into a step-pool, cascade, or bedrock channel, depending on hannel slope, discharge, and availability of coarse sediment.

Changes in reach-level channel type resulting from increased sediment upply typically represent a transient response to a pulsed input, although a onger-term response may result from sustained inputs. A landslide-related ulse of sediment may result in a transient change to a morphology with a ower q_r that subsequently relaxes toward the original morphology as the erturbation subsides. Pool-riffle reaches, for example, can develop a raided morphology while transmitting a pulse of sediment and subsequently revert to a single-thread pool-riffle morphology. Channel reaches vith high q_r should recover quickly from increased sediment loading, because they are able to rapidly transport the load downslope. Reaches with a ow q_r should exhibit more persistent morphologic response to a comparable increase in sediment supply. Transient morphologic change can also result from debris-flow scour of steep-gradient channels. For example, colluial and cascade channels that are scoured to bedrock by a debris flow may lowly revert to their predisturbance morphologies.

The spatial pattern of channel types provides a template against which to ssess channel response potential, but the disturbance history of a channel etwork also is important for understanding both current conditions and reponse potential. Reach-level channel morphology provides a general indiati differences in response potential, but specific responses depend on ne ure, magnitude, and persistence of disturbance, as well as on local onditions, including riparian vegetation, in-channel large woody debris, ank materials, and the history of catastrophic events. Furthermore, concurent multiple perturbations can cause opposing or constructive response, depending on both channel type and the direction and magnitude of change. ence, assessment of either present channel conditions or the potential for iture impacts in mountain drainage basins should consider both disturance history and the influences of channel morphology, position in the netork, and local external constraints.

ONCLUSIONS

Systematic variations in bed morphology in mountain drainage basins rovide the basis for a classification of channel-reach morphology that reects channel-forming processes, serves to illustrate process linkages within e channel network, and allows prediction of general channel response potial. The underlying hypothesis that alluvial bed morphology reflects a able roughness configuration for the imposed sediment supply and transort capacity implies a fundamental link between channel processes and rm. The association of reach types and ratios of transport capacity to sediment supply combined with identification of external influences and the spal coupling of reaches with hillslopes and other channel types provides a nceptual framework within which to investigate channel processes, assess annel conditions, and examine spatially distributed responses to watershed sturbance. Integration of this approach into region-specific landform and lley segment classifications would provide a common language to studies fl processes and response to disturbance. This classification, however, n al for all purposes; characterization of river planforms, for example, useful for classifying flood-plain rivers. The development of specific restoration designs requires further information on reach-specific characteristics. Our classification simply characterizes aspects of reach-level channel morphology useful for assessing channel condition and potential response to natural and anthropogenic disturbance in mountain drainage basins.

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