Downstream coarsening in headwater channels

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[1] Field data from four mountain drainage basins in western Washington document systematic downstream coarsening of median bed surface grain size (D_{50}) and a subsequent shift to downstream fining at a drainage area of about 10 km^2. Analyses of network-wide patterns of unit stream power derived from both channel surveys and digital elevation models reveal maximum unit stream power that in all four study areas roughly corresponds with both the grain size maxima and an inflection in the drainage area-slope relation thought to represent the transition from debris flow-dominated channels to fluvially dominated channels. Our results support the hypothesis that basin-wide trends in D_{50} are hydraulically controlled by systematic variations in unit stream power in addition to lag deposits forced by mass-wasting processes. The similar relations found in our four study areas suggest that the tendency for downstream coarsening may be ubiquitous in headwater reaches of mountain drainage basins where debris flow processes set the channel gradient. *INDEX TERMS:* 1824 Hydrology: Geomorphology (1625); 1815 Hydrology: Erosion and sedimentation; 1848 Hydrology: Networks; 1860 Hydrology: Runoff and streamflow; *KEYWORDS:* downstream coarsening, downstream fining, debris flow, unit stream power, headwater channel, lag deposit

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1. Introduction

[2] Numerous studies have addressed the relative importance of various mechanisms for the gradual downstream fining observed in many alluvial rivers. Previous work has attributed downstream fining to selective transport of finer grains [*Paola et al.*, 1992, *Ferguson et al.*, 1996; *Wilcock*, 1997; *Gasparini et al.*, 1999; *Hoey and Bluck*, 1999; *Montgomery et al.*, 1999], particle abrasion [*Krumbein*, 1941; *Schumm and Stevens*, 1973; *Parker*, 1991a, 1991b; *Kodama*, 1994], in situ clast weathering during storage [*Bradley*, 1970; *Heller et al.*, 2001], and the spatial distribution of sources for resistant lithologies [*Pizzuto*, 1995]. In addition, downstream trends in grain size can be disrupted by mixing of discrete sediment populations at tributary junctions [*Miller*, 1958; *Knighton*, 1980; *Rice*, 1998] or by continuous mixing along mountain headwater channels that are ‘‘coupled’’ to adjacent hillslopes [*Grant and Swanson*, 1995; *Rice and Church*, 1996; *Church*, 2002]. Although considerable work has focused on downstream fining, only a few researchers have explored downstream variations in grain size in headwater channels in mountain drainage basins [e.g., *Miller*, 1958; *Benda*, 1990; *Grimm et al.*, 1995; *Lambert et al.*, 1996].

[3] In steep mountainous terrain, episodic disturbance by debris flows dominates the transport and routing of sediment from low-order headwater channels to higher-order alluvial channels [*Dietrich and Dunne*, 1978; *Benda*, 1990; *Lancaster et al.*, 2001]. Hence headwater channels provide an important sediment-transport link between sediment production on hillslopes and delivery to downstream channel reaches [*Caine and Swanson*, 1989; *Benda and Dunne*, 1997]. A number of workers [e.g., *Seidl and Dietrich*, 1992; *Montgomery and Foufoula-Georgiou*, 1993; *Stock and Dietrich*, 2003] have identified the topographic signature of the transition from debris flow-dominated channels to fluvially dominated channels as an inflection in the relationship between drainage area and channel slope (Figure 1a). Under steady-state conditions, channel slope (S) has been observed to vary empirically as a function of drainage area (A):

$$S = kA^{-\theta} \qquad (1)$$

where k and θ are empirical coefficients that represent profile steepness and concavity, respectively [e.g., *Hack*, 1957; *Howard et al.*, 1994; *Sklar and Dietrich*, 1998]. The exponent θ generally ranges from less than 0.3 in steep headwater channels dominated by mass wasting to greater than unity in some alluvial channels [*Seidl and Dietrich*, 1992; *Montgomery and Foufoula-Georgiou*, 1993; *Montgomery*, 2001; *Stock and Dietrich*, 2003].

[4] Variation in θ can influence the rate of energy expenditure per unit area of the channel bed, which is defined in terms of unit stream power:

$$\omega = \frac{\rho g Q S}{W} \qquad (2)$$

where ρg is the unit weight of water, Q is discharge, and W is channel width. Bankfull discharge may be assumed to vary with drainage area A via

$$Q = eA^d \qquad (3)$$

where e and d are determined empirically. The coefficient d typically ranges from 0.7 for semi-arid regions to 1.0 for

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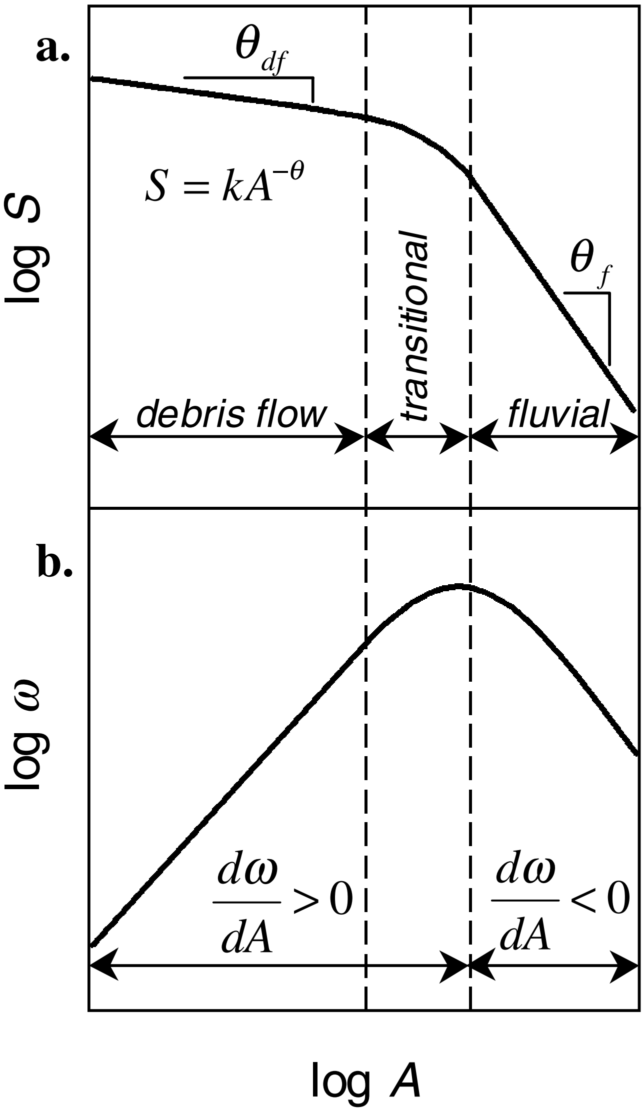

Figure 1. Schematic illustration of the transition in concavity index between a debris flow process domain (θ_{df}) and fluvial process domain (θ_f) and the hypothesized influence on the downstream rate of change in unit stream power predicted by equation (8).

humid landscapes that drain small catchments, such as our field sites [*Dunne and Leopold*, 1978; *Leopold*, 1994; *Rice*, 1998; *Whiting et al.*, 1999]. Combining equations (2) and (3) allows unit stream power to be recast in terms of parameters that are readily measured either in the field or from digital elevation models (DEMs):

$$\omega = \frac{\rho g e A^d S}{W} \tag{4}$$

Classic hydraulic geometry relations hold that channel width varies as a power function of downstream changes in discharge (which can be converted to drainage area based on the empirical relation discussed above):

$$W = cA^b \tag{5}$$

where c and b are empirical constants. Numerous studies report b to be about 0.5 for alluvial channels [e.g., *Leopold and Maddock*, 1953; *Ibbitt*, 1997; *Knighton*, 1998], and the limited data available indicate values of 0.3 to 0.6 for bedrock channels [*Snyder et al.*, 2000; *Montgomery and Gran*, 2001]. Finally, equations (4) and (5) may be combined and simplified into a form readily extracted from DEMs:

$$\omega = \rho g(e/c)A^{(d-b)}S \tag{6}$$

where the coefficient $\rho g(e/c)$ is retained to render equations (2) and (6) dimensionally equivalent. On the basis of the ranges of typical values for d (0.7 to 1.0) and b (0.3 to 0.6), $d - b$ is expected to range from 0.1 to 0.7. Combining equations (1) and (6) allows unit stream power to be expressed as a function of drainage area:

$$\omega = c_2 A^{(d-b-\theta)} \tag{7}$$

where the constants in equations (1) and (6) have been folded into c_2. The variation in unit stream power with drainage area can be evaluated from the first derivative of equation (7):

$$\frac{d\omega}{dA} = c_2(d - b - \theta)A^{(d-b-\theta-1)} \tag{8}$$

Equation (8) predicts a constant value of ω when $\theta = d - b$, a downstream increase in ω for $\theta < d - b$, and a downstream decrease in ω for $\theta > d - b$ (Figure 1b).

[5] Although previous studies of river networks have demonstrated a downstream variation in unit stream power [*Magilligan*, 1992; *Lecce*, 1997; *Knighton*, 1999], few studies have explicitly coupled systematic downstream patterns of unit stream power in mountain channel systems with either longitudinal variations in surface grain size or the downstream sorting and delivery of sediment by debris flows (*Whiting et al.* [1999] provide a notable exception). Previous field studies of mountain channels suggested that bed surface grain size increased downstream and reached a maximum at the transition from debris flow-dominated channels to fluvially dominated channels [*Montgomery and Buffington*, 1993; *Lambert et al.*, 1996]. Identification of the fundamental transition from debris flow to fluvial processes is important for understanding the relationship between short-term bed surface organization and longer-term controls on channel slope, as well as controls on the disturbance ecology of mountain streams, the routing of sediment through mountain channel networks, and assumptions inherent in many landscape evolution models. Here we present field data and results of DEM analyses from four mountain drainage basins in western Washington that document systematic downstream bed surface coarsening and evaluate the correspondence of these trends to downstream variations in unit stream power.

2. Study Areas

[6] Channel surveys were conducted in two study areas in the northern Cascades (Boulder River and Finney Creek) and two on the Olympic Peninsula (South Fork Hoh River and Barnes Creek), all located in western Washington

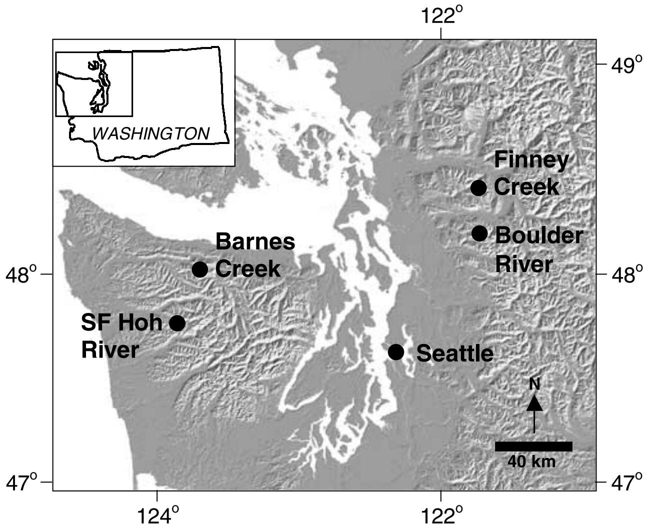

Figure 2. Shaded relief map of western Washington showing locations of study areas.

(Figure 2). Both northern Cascade study areas share a similar lithology and climate history but are distinguished by contrasting degrees of land management; the Boulder River watershed is a pristine wilderness area, whereas the Finney Creek basin has been heavily logged. Although the South Fork Hoh River originates in Olympic National Park, some lower portions outside the park were logged extensively in the 1980s, whereas Barnes Creek is located entirely within Olympic National Park.

[7] The Boulder River drains 63 km^2 and is a tributary to the North Fork Stillaguamish, which flows into Puget Sound. Elevations in the Boulder River basin range from 90 m at the basin outlet to 2090 m on Three Fingers Peak. Permanent snowfields and small alpine glaciers occupy north-facing cirques located below the highest peaks. Finney Creek drains 139 km^2 and is a tributary to the Skagit River, which also flows into Puget Sound. Elevations in the Finney Creek basin are lower than those in the Boulder and range from 40 m at the confluence with the Skagit River to 1550 m on Finney Peak. Both basins are underlain by Mesozoic marine metasedimentary and metavolcanic rocks [*Dragovich et al.*, 2002]. Late Pleistocene advances of the Cordilleran ice sheet deposited exotic glacial till and outwash gravels in both of these basins [*Crandall*, 1965]. In addition, alpine glaciers scoured high divides and delivered autochthonous glacial sediments to proximal, upper-basin locations.

[8] The South Fork Hoh River originates in Olympic National Park and drains 129 km^2 from the western flank of the Olympic Mountains (2428 m) to the confluence with the main stem Hoh River (130 m), which flows west to the Pacific Ocean. Alpine glaciers in the core of the Olympic Mountains provide continuous summer snowmelt to the South Fork. Lithology of the South Fork Hoh River basin is dominated by Tertiary sediments of the Western Olympic Lithic Assemblage [*Tabor and Cady*, 1978]. Upper reaches of the basin are confined within a bedrock gorge, whereas the lower two thirds of the basin have been glacially sculpted into a broad, U-shaped valley filled with stratified outwash and moraine deposits. Elevated river terraces and local exposures of bedrock within the unconfined valley floor provide evidence of ongoing tectonic uplift and commensurate fluvial incision.

[9] Barnes Creek drains 41 km^2 from Happy Lake Ridge (1667 m) to its outlet at Lake Crescent (176 m), a 21-km^2 freshwater lake on the northern Olympic Peninsula. This region lies at the southern extent of Pleistocene glaciation; hence the glacial legacy evident in the other study areas is not as pronounced in Barnes Creek. Elevation of the Cordilleran Ice Sheet was considerably lower in the vicinity of the northern Olympics than it was in the northern Cascades. Ice reached a maximum elevation of 600 m near Lake Crescent and advanced up the lower 4.5 km of Barnes Creek [*Dragovich et al.*, 2002]. In addition, alpine glaciers incised cirques into the upper 150 m of relief in the headwater reaches and modified the upper 1.5 km of the Barnes Creek longitudinal profile. In contrast to the other study basins, Barnes Creek is largely confined by both steep, forested slopes and (locally) deep bedrock gorges incised into Tertiary basalt and volcaniclastic sediments of the Crescent Formation [*Tabor and Cady*, 1978]. Base level for Barnes Creek is thought to have risen by 24 m during blockage of ancestral Lake Crescent by a series of large landslides sometime following retreat of the Cordilleran ice sheet about 13,000 year ago [*Logan and Schuster*, 1991]. Subsequent aggradation formed a prograding fan halfway across Lake Crescent and reduced the channel gradient in the lower 3 to 4 km of Barnes Creek.

3. Methods

[10] We combined recent field data collected during 2001 and 2002 field seasons with results from prior surveys conducted since 1992 [*Montgomery and Buffington*, 1997] in the Boulder River, Finney Creek, and South Fork Hoh River study areas. To acquire data from a broad range of drainage basin sizes, sampling in these three basins was performed in both main-stem and tributary channels, as well as in channels from adjacent basins. Although macroscopic basin properties (i.e., lithology, climate, topography, land use history) were consistent within each collective study area, reach-scale attributes (e.g., channel morphology and disturbance history) varied considerably between sample locations in different tributaries. Therefore data were collected in a fourth basin (Barnes Creek) in 2002 by sampling systematically along a single longitudinal profile. We chose this sampling protocol to control for the variability of conditions encountered in previous field studies due to sampling of a suite of channels from throughout a channel network and consequently at various stages of recovery from prior disturbances.

[11] In all four study areas, channel reaches of 10–20 channel widths in length were selected systematically to minimize the influence on channel morphology by woody debris, tributary junctions, or direct, proximal anthropogenic disturbances related to timber harvest (e.g., roads and culverts). Likewise, tributary reaches located on the floodplain of the South Fork Hoh River were excluded from this data set because their slopes are controlled by the significantly larger main stem of the South Fork. In confined valley reaches that lacked floodplains, flow indicators such as vegetation patterns, stain lines, and the height of snagged debris were used to approximate the flow depth

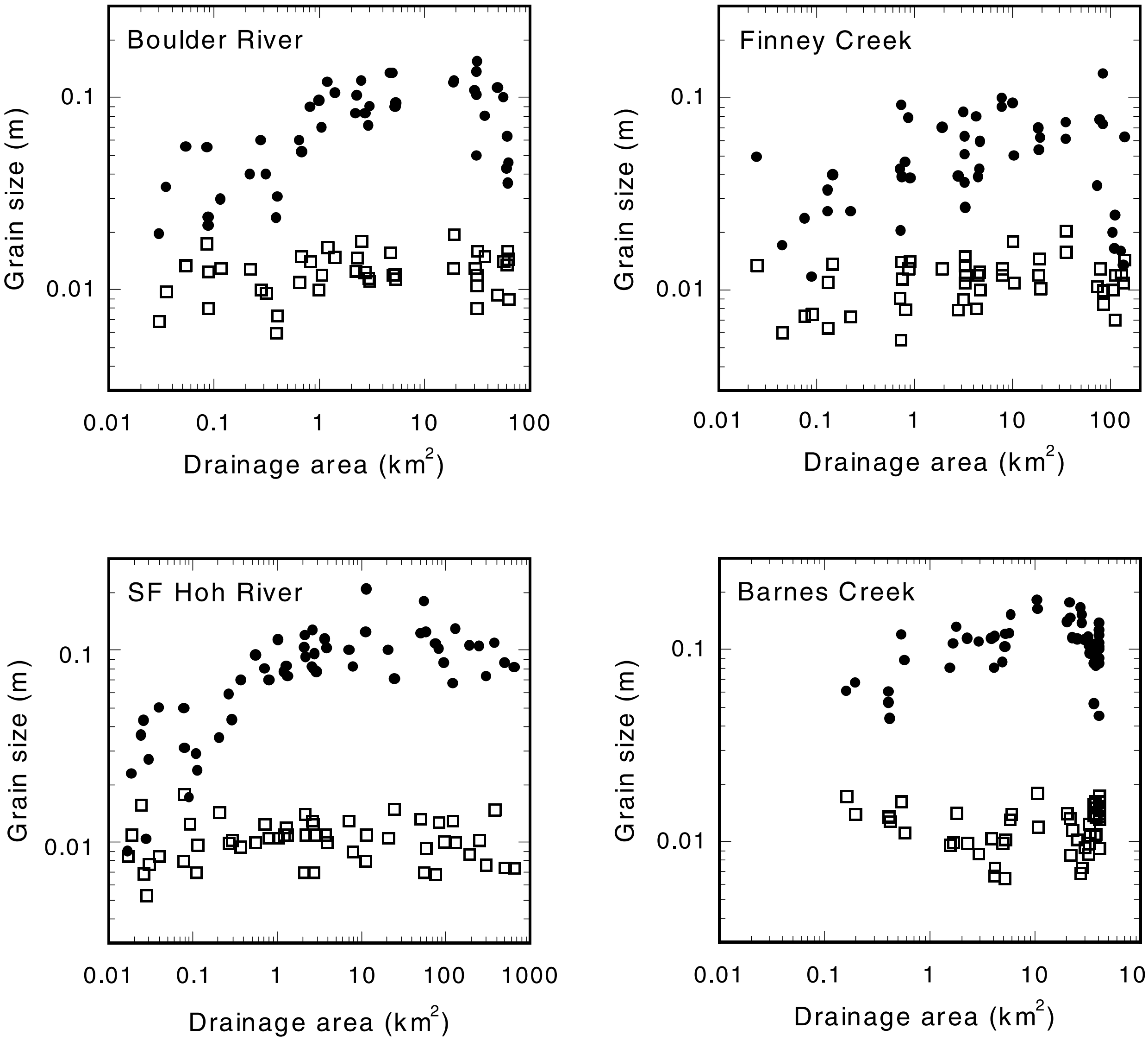

Figure 3. Plots of drainage area versus grain size (D_{50}) for field sites showing the transition from downstream coarsening to downstream fining. Symbols represent surface D_{50} (solid circles) and subsurface D_{50} (open boxes). Note the different horizontal scales.

corresponding to the recurrence interval of bank-full flows in floodplain channels (i.e., the seasonally high flow). Channel width and depth were approximated from one to three cross sections surveyed at each reach using a tape, stadia rod, and hand level or engineer's level. Slope was calculated by surveying the bed between two geomorphically similar channel units (e.g., steps or riffles) located at upstream and downstream ends of each reach. Drainage areas were calculated from DEMs using locations mapped onto U.S. Geological Survey 1:24,000 scale topographic maps. Colluvial channels were distinguished in the field from distinct types of alluvial channels by bed morphology and a valley fill composed of unsorted, matrix-supported colluvium overlain by a thin substrate of alluvium [*Montgomery and Buffington*, 1997]. Channel width was regressed against drainage area for data from each study area to compare hydraulic geometry relations between debris flow-dominated and fluvially dominated channels. For these purposes, process domains were defined from observable breaks in log binned average drainage area-slope relations derived from DEMs. The level of significance of hydraulic geometry relations was evaluated using the two-sample t statistic.

[12] The grain size distribution of sediment was characterized at each reach using the pebble count method of *Wolman* [1954] for bed surface sediment and the method of *Buffington* [1996] for subsurface sediment. Reach average values of bed surface sediment were obtained by randomly selecting at least 100 surface grains from both exposed bars and submerged locations with the tip of an index finger. Bed surface sampling occurred across the entire width of the channel (if feasible) but avoided pools. The medial axis of each clast was then measured and recorded, and the grain was discarded to prevent resampling. Subsurface pebble counts were conducted on exposed bars at water level by first removing surface material to a depth of the largest grain size and then exposing approximately 1 m^2 of subsurface material. The substrate was then loosened and mixed with a

hand shovel to a depth of 15 to 30 cm, and sampling then proceeded following the same methodology employed for surface sampling. Grain size data were truncated at 4 mm and are reported as median grain size (D_{50}) for both surface and subsurface data.

[13] We used simplified forms of equations (4) and (6), along with the assumption that discharge varies linearly with drainage area (i.e., $d = 1$) to define transport capacity in terms of two indices of unit stream power. The first index (AS/w_{bf}) was calculated at each sample location using both local reach slope and bank-full width (w_{bf}) surveyed in the field. A second index of unit stream power $[(1/c)A^{(1-b)}S]$ was predicted throughout the entire channel system from 10-m grid size DEMs using basin-specific c and b coefficients derived from relations between drainage area and channel width.

4. Results

[14] Our analysis focuses on general relations between drainage area and three morphological characteristics of headwater channels: surface and subsurface grain size, channel width, and unit stream power. We also compare field data sampled randomly from different tributaries with data sampled systematically along a single longitudinal profile. Finally, we use results of DEM analyses to complement generalization inferred from limited field data.

4.1. Downstream Coarsening

[15] A systematic increase in surface D_{50} with drainage area is found in basins smaller than ~1 to 10 km^2 (Figure 3). In contrast, subsurface D_{50} shows no appreciable dependence on drainage area. Deviation of surface D_{50} from subsurface D_{50} allows downstream coarsening to be expressed in terms of a downstream increase in armoring through the ratio of surface D_{50} to subsurface D_{50} (here defined as D^*_{50}). Values of D^*_{50} range from approximately 3 in drainage basins <0.1 km^2 to more than 25 in basins >10 km^2. Variance in surface D_{50} for a given drainage area is greatest in channels draining basins <0.5 km^2 and throughout the industrialized Finney Creek basin. Downstream coarsening of surface D_{50} shifts abruptly to rapid downstream fining both where our study channels flow across broad floodplains immediately upstream of the confluences with larger rivers (i.e., Boulder River and Finney Creek) and upstream of Lake Crescent at the outlet of Barnes Creek. In contrast, downstream coarsening along the South Fork Hoh River yields to a more gradual rate of downstream fining over a range of drainage basin areas that span 10 km^2 to at least 650 km^2 (the largest drainage area sampled). *Heller et al.* [2001] attributed the relatively low rate of downstream fining observed in the main stem of the Hoh River to the continuous resupply of glacial debris stored in floodplain deposits, despite the fact that these weathered grains abrade rapidly in tumbling mill experiments.

[16] Combined field data from pristine basins further emphasize the positive dependence of surface D_{50} and D^*_{50} on drainage area (Figure 4). Power-law regression of combined surface D_{50} and D^*_{50} data for pristine basins shows that drainage area accounts for, respectively, 71% and 75% of the variance in grain size for basins <10 km^2. Overall, measurements of surface D_{50} in our four study areas range from 9.0 to 210 mm, whereas subsurface D_{50} is considerably finer and ranges from 5.3 to 21 mm.

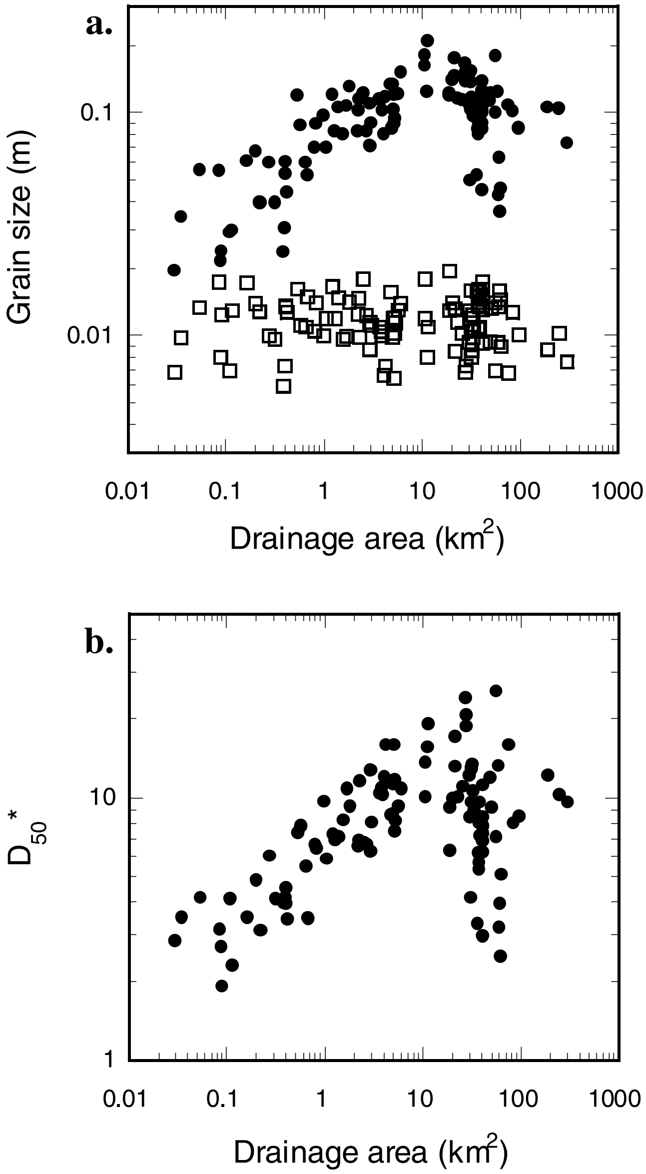

Figure 4. Composite plots of grain size (D_{50}) and armoring (D^*_{50}) versus drainage area for pristine watersheds. (a) Downstream coarsening of surface D_{50} occurs in channels draining basins less than ~10 km^2, whereas rapid downstream fining occurs where our basins drain into large glaciated valleys (see text for further explanation). The downstream coarsening regression of surface D_{50} for basins <10 km^2 follows the relation $D_{50} = 0.074A^{0.30}$ ($r^2 = 0.71$). Symbols are the same as in Figure 3. (b) Downstream trends in D^*_{50} parallel those of surface D_{50} and exhibit a similar relation with drainage area for basins <12 km^2: $D^*_{50} = 6.5A^{0.29}$ ($r^2 = 0.75$).

4.2. Drainage Area-Slope Analysis

[17] Log bin averaged drainage area-slope relations derived from 10-m grid size DEMs of our channel networks exhibit relations of the general form described by *Montgomery and Foufoula-Georgiou* [1993] and found in mountain channel networks by previous workers

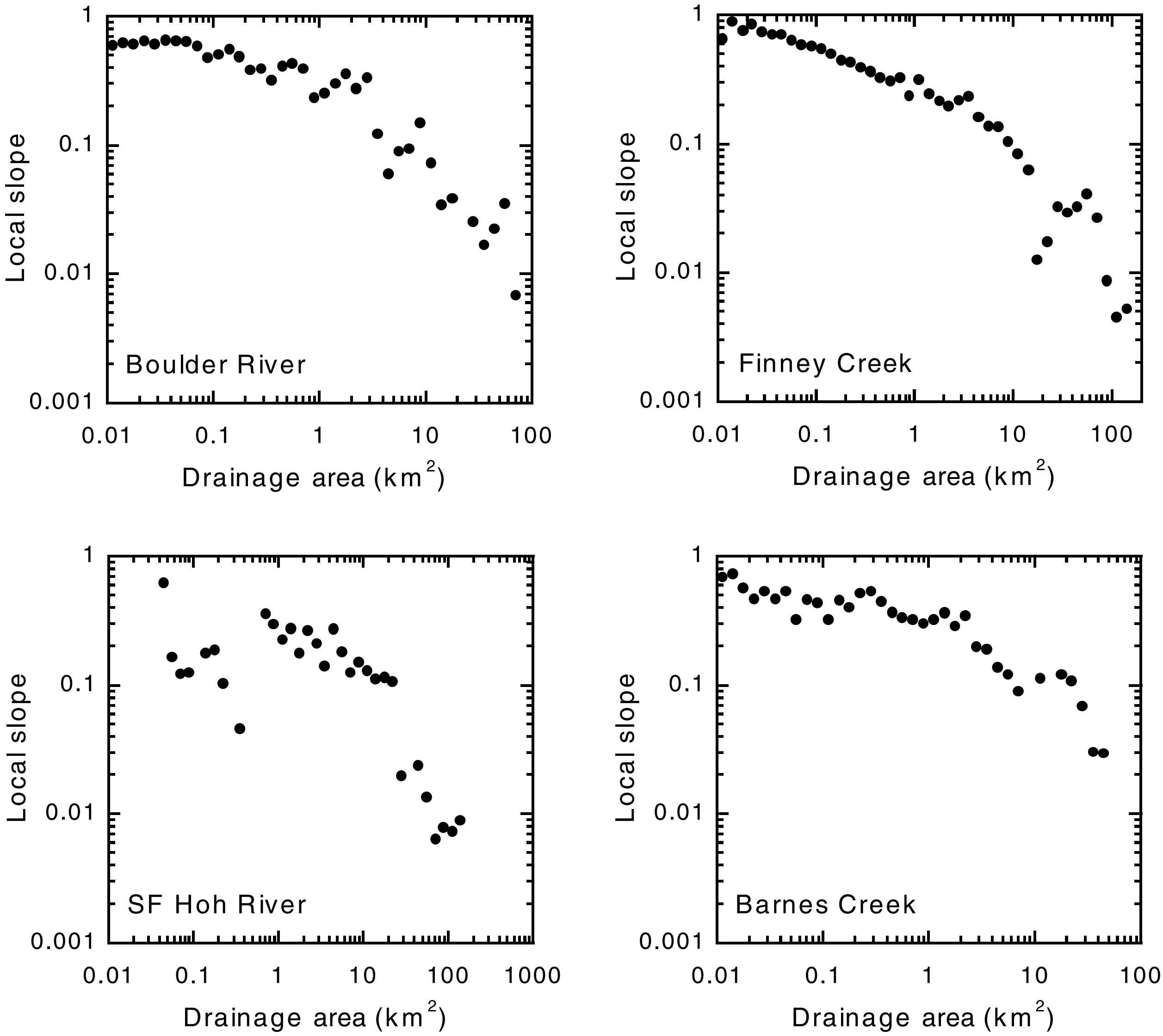

Figure 5. Log bin averaged local slope derived from 10-m grid size DEMs versus drainage area for the four study areas.

[e.g., *Whipple and Tucker*, 1999; *Snyder et al.*, 2000; *Montgomery*, 2001; *Stock and Dietrich*, 2003]. Drainage area-slope relations define two discrete domains in the Boulder River and Barnes Creek basins but are more transitional in the Finney Creek basin, whereas there is no clear relation in the South Fork Hoh River basin (Figure 5). With the exception of the South Fork Hoh River, inflections in drainage area-slope scaling between debris flow- and fluvial domains occur at a drainage area of approximately 3 km^2, which correspond with log binned-average local slopes ranging from 0.1 to 0.3. This transitional domain includes most of the cascade and some of the step-pool channel types surveyed in the field.

4.3. Channel Width

[18] Channel width increases with drainage area according to equation (5) for both debris flow-dominated and fluvially dominated reaches surveyed in our study areas (Figure 6). With the exception of Barnes Creek, the conventional value of $b = 0.5$ falls within the 95% confidence interval of b values calculated for combined field data (Table 1). In addition, differences in both c and b values for debris flow- and fluvially dominated reaches are significant ($p < 0.001$) at the 95% confidence level for the Finney Creek data, whereas only b values are significantly different for the Barnes Creek data. In the Finney Creek study area, debris flow-dominated channels are significantly wider than fluvially dominated channels of comparable drainage area. Smaller b values for debris flow reaches indicate that the rate of downstream increase in channel width is less than that of fluvially dominated segments in Finney Creek. In contrast, fluvially dominated channels are significantly wider than debris flow-dominated channels along the Barnes Creek profile. The few bedrock-floored reaches surveyed in our study areas plot below the trend of the regressions; hence they are narrower than most debris flow- and fluvially dominated channels.

4.4. Unit Stream Power Indices

[19] Both the field-based index of unit stream power (AS/w_{bf}) and channel type vary systematically with drainage area in the Finney Creek, South Fork Hoh River, and

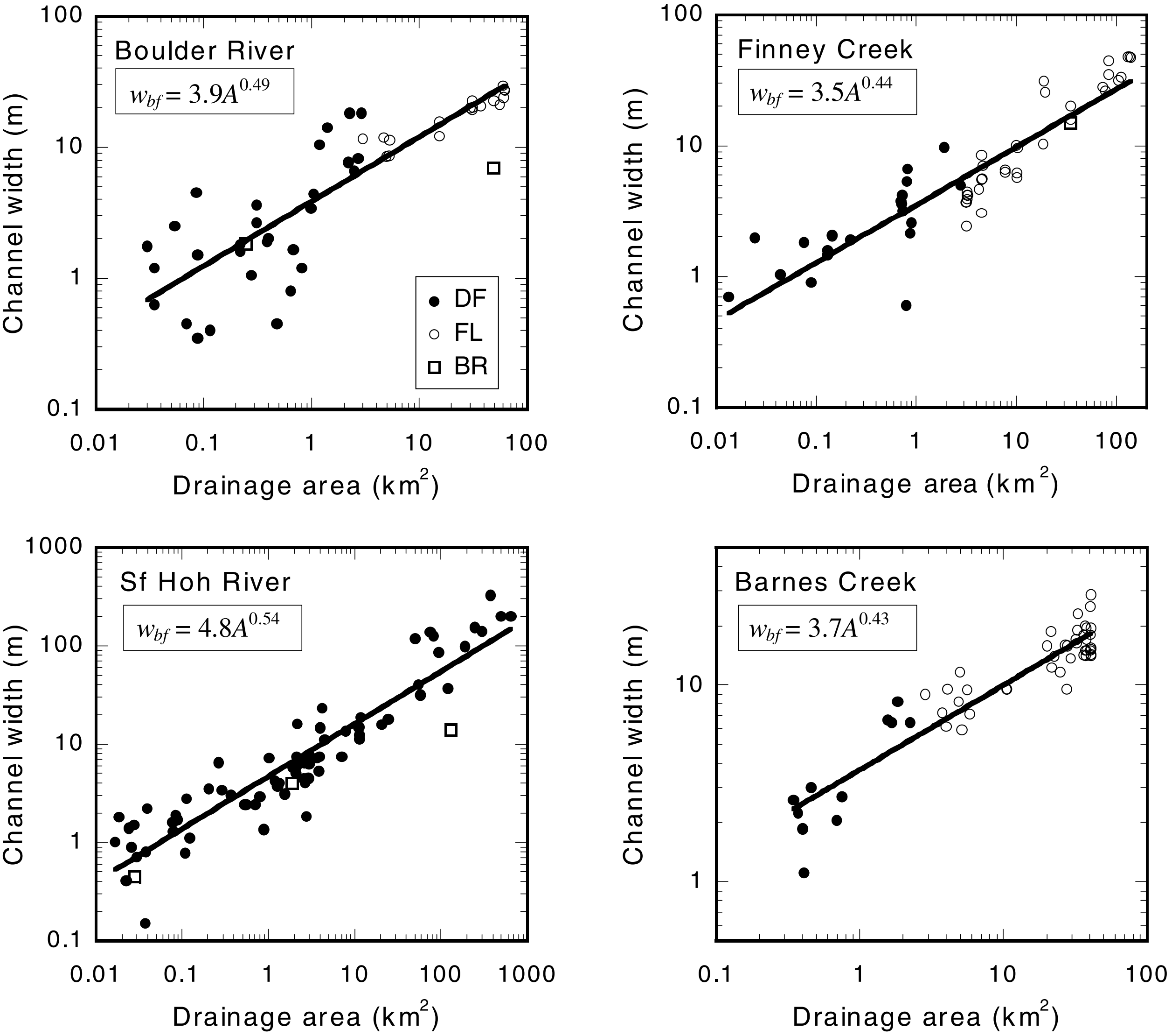

Figure 6. Plots of relations between channel width and drainage area for debris flow (DF), fluvial (FL), and bedrock (BR) channels in the Boulder River, Finney Creek, and Barnes Creek study areas and combined channel types in the South Fork Hoh River study area. Power function regressions are shown for combined (debris flow and fluvial) relations. Bedrock channels are shown for comparison and plot below the regression trend for the combined data.

Barnes Creek study areas (Figure 7). The lowest values of unit stream power occur in both colluvial and pool-riffle channels located at the smallest and greatest drainage areas of our channel networks, whereas the highest values are found in cascade and step-pool channels located within mid-basin reaches. the overall pattern of unit stream power observed the three study areas is consistent with data from previous field studies [e.g., *Magilligan*, 1992; *Lecce*, 1997; *Knighton*, 1999], which found unit stream power increased downstream in headwater channels before reversing to a decreasing trend in lower-gradient alluvial reaches. The weakest patterns occur in the three study areas where data are spatially aggregated (Boulder River, in particular), whereas the most systematic relation between drainage area and the field-based index of unit stream power occurs in Barnes Creek, where data were collected along a single longitudinal profile.

[20] The field-based index of unit stream power measured along the Barnes Creek profile parallels downstream trends in both surface D_{50} and D^*_{50} (see Figures 4 and 7). Not surprisingly, there are weak, but significant log linear relationships between unit stream power index and both surface D_{50} (r^2 = 0.49) and D^*_{50} (r^2 = 0.36) along the Barnes Creek Profile (Figure 8). The degree of correlation varies among the other study areas. For instance, surface D_{50} is uncorrelated with unit stream power index in both the Boulder River (r^2 = 0.01) and Finney Creek (r^2 = 0.16) basins but exhibits a weak correlation (r^2 = 0.27) in the South Fork Hoh River basin. Although there is considerable variability in surface D_{50} and D^*_{50} at any particular value of unit stream power, the upper envelope of grain size data tends to increase with unit stream power.

[21] DEM-based indices of unit stream power [$(1/c) A^{(1-b)}S$] were predicted from 10-m grid size DEMs and

Table 1. Summary of Hydraulic Geometry Relations and Basin Parameters

Study Area/Domain[a]	Hydraulic Geometry Relations ($w = cA^b$) c[b]	b	R^2	n	Basin Parameters θ	R^2	$(d - b - \theta)$[c]
Boulder River							
Debris flow	4.1 ± 1.7	0.54 ± 0.23	0.46	29	0.09	0.69	0.37
Fluvial	6.0 ± 1.3	0.35 ± 0.07	0.88	18	0.73	0.73	−0.08
Combined	3.9 ± 0.8	0.49 ± 0.08	0.76	47			
Finney Creek							
Debris flow	3.6 ± 1.2	0.35 ± 0.18	0.48	20	0.26[d]	0.93	0.38
Fluvial	1.9 ± 0.4	0.66 ± 0.08	0.90	33	0.93	0.80	−0.59
Combined	3.5 ± 0.5	0.44 ± 0.06	0.83	53			
South Fork Hoh River							
Combined	4.8 ± 0.7	0.54 ± 0.05	0.87	71			
Barnes Creek							
Debris flow	4.0 ± 0.8	0.78 ± 0.26	0.82	12	0.14	0.63	0.08
Fluvial	4.4 ± 1.1	0.37 ± 0.08	0.69	40	1.0[d]	0.97	−0.43
Combined	3.7 ± 0.5	0.43 ± 0.05	0.87	52			

[a]Transition between debris flow and fluvial domains defined at 3 km^2 based on drainage area-slope inflection from DEMs except for the South Fork Hoh River, where there is no clear transition.

[b]Values for c are reported in m and for drainage area are in km^2.

[c]Coefficient in equation (8) calculated with an assumed d value of 1.0.

[d]Concavity calculated in basins <1 km^2 for the debris flow domain in Finney Creek and in basins between 3.0 and 10 km^2 for the fluvial domain in Barnes Creek.

equation (6) using basin-specific c and b values derived from combined hydraulic geometry relations. Plots of log binned-averaged unit stream power index versus drainage area show downstream trends of increasing unit stream power in headwater channels followed by a decrease in unit stream power in fluvial channel types (Figure 9). These trends are consistent with the downstream variation of unit stream power exhibited by field data in the Finney Creek, South Fork Hoh River, and Barnes Creek study areas (see Figures 7 and 9) and predicted by equation (8) using an assumed d value of 1.0 and appropriate b values for each basin (Table 1). In general, the highest values of unit stream power indices derived from DEMs occur in mid-basin reaches for drainage areas ranging from 1 to 10 km^2, which correspond to the transition in DEM-derived drainage area-slope scaling (Figure 5). Minimum values of unit stream power are predicted to occur in both extreme headwater reaches at the tips of the channel networks and along low-gradient, alluvial channels; however, relatively high values are also predicted in main-stem channels at knickpoints.

5. Discussion

[22] We posit that to some degree the lack of correlation between bed surface grain size and unit stream power in the Boulder River, Finney Creek, and South Fork Hoh River study areas results from the spatial aggregation of field data collected from various tributaries with different disturbance histories. In the Barnes Creek study area, where sampling along a continuous profile may have controlled for the spatial variance in physical parameters, the correlation between unit stream power and the upper envelope of D_{50}^* suggests a dynamic adjustment of the bed surface to the local rate of energy expenditure. Previous studies have related bed surface armoring to transport capacity in excess of sediment supply [e.g., *Dietrich et al.*, 1989; *Lisle and Madej*, 1992; *Buffington and Montgomery*, 1999b]. These prior studies ascribed variability beneath the upper bound in grain size data to bed surface fining in response to an increase in the sediment supply rate. The close association of the upper envelope of both surface D_{50} and D_{50}^* data with unit stream power in the Barnes Creek study area leads us to hypothesize that the systematic pattern of downstream coarsening reflects a basin-scale trend in the magnitude of transport capacity relative to sediment supply and that the observed transition from downstream coarsening to downstream fining signals a reversal in the downstream rate of change of this ratio.

[23] We attribute at least two factors to the systematic, order-of-magnitude shift between the inflection in drainage area-slope relations derived from DEMs (3 km^2) and the range in unit stream power maxima (1–10 km^2). As predicted by equation (8) and illustrated graphically in Figure 1, the downstream rate of change in unit stream power is determined by the sign of $(d - b - \theta)$, which, for our field areas, remains positive until the transition to the fluvial domain is complete. The correspondence between maxima in both the field derived index of unit stream power and surface D_{50} suggests that the characteristic length scale of this transition ranges between about 1 and 10 km^2. This recognizable pattern in bed surface grain size can provide an additional means of identifying the fundamental transition from debris flow to fluvially dominated channels.

[24] Unlike alluvial channels, where slope is an additional degree of freedom that may adjust in response to feedbacks between transport capacity and sediment supply, our findings point to grain size as a key dependent variable above this transition in steep headwater channels. Our field investigation and DEM analyses suggest channel beds above the drainage area-slope inflection are shaped by persistent, short-term fluvial processes that organize the bed surface upon a slope set by longer-term debris flow processes. We further hypothesize that the mismatch between the area-slope inflection and the shift from downstream coarsening to downstream fining, concomitant with depositional slopes of most debris flows, may also be forced by the supply of immobile boulders from debris flows and other mass-wasting events. These lag deposits may act in concert with hydraulic forcing in hillslope-coupled, transitional reaches to extend the range of downstream coarsening and delay the onset of downstream fining to well beyond the area-slope inflection, within uncoupled floodplain channels.

[25] Previous studies have shown that mechanical sorting by debris flows tends to focus the largest clasts to the flow front. On the basis of flume experiments, *Parsons et al.* [2001] attributed the development of a course snout to stripping of fines and water from the flow front by a relatively dry, rough bed. They also found evidence of vertical sorting by kinetic sieving, which caused larger clasts to become entrained by higher surface velocities and preferentially transported to the debris flow front. *Suwa* [1988] described a similar mechanism whereby coarser grains commenced motion before finer grains because of reduced pivoting angles and greater protrusion into higher velocity flows. Scale dependence on both fluid drag and inertia caused coarser grains to accelerate and attain higher

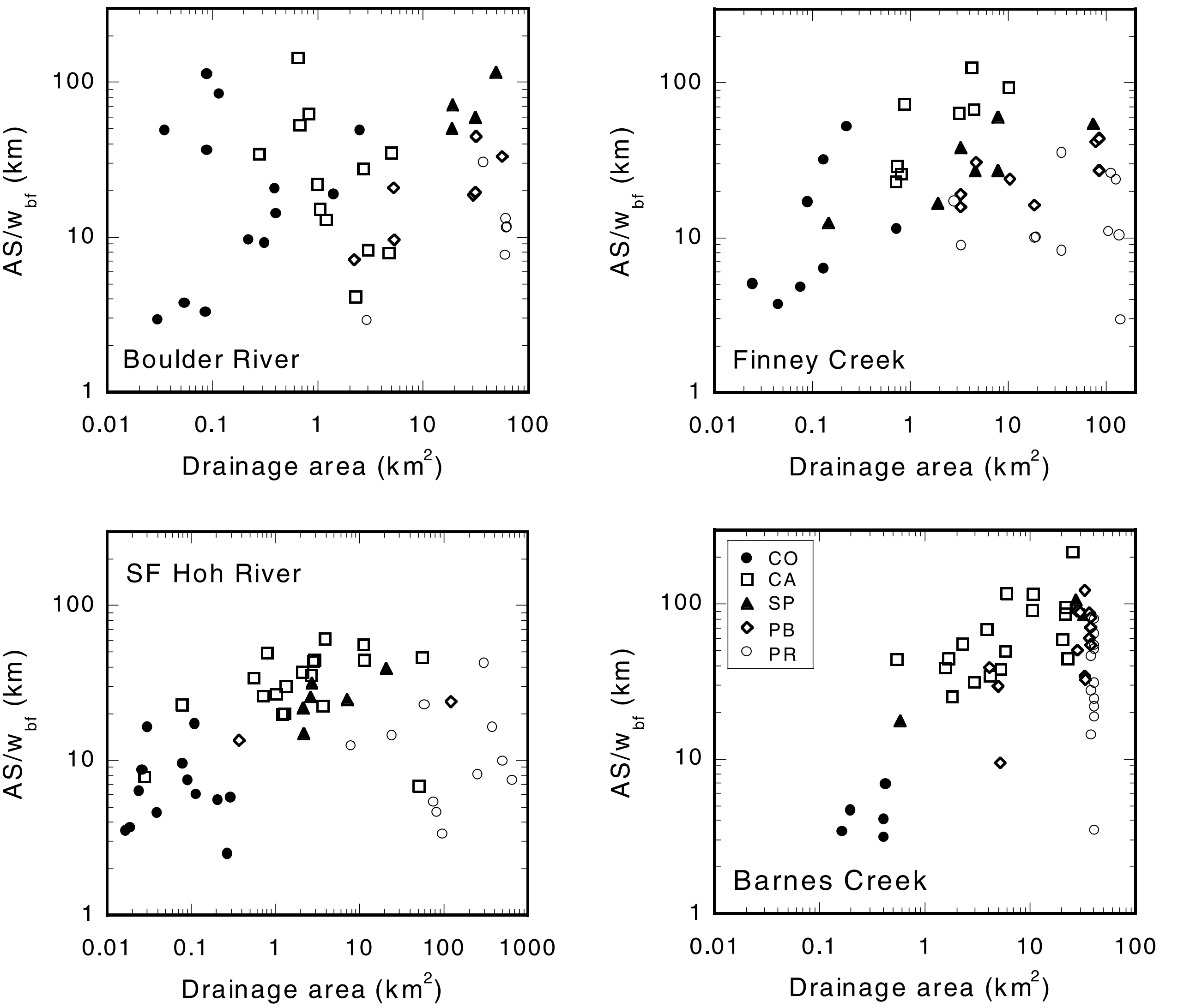

Figure 7. Plots of drainage area versus the index of unit stream power (AS/w_{bf}) by channel type according to the classification of *Montgomery and Buffington* [1997]: colluvial (CO), cascade (CA), step pool (SP), plane bed (PB), and pool riffle (PR).

terminal velocities than smaller grains, which *Suwa* [1988] posited maintained a coarse snout. Field observations in Barnes Creek indicate that partially reworked debris flow deposits at tributary junctions are frequently associated with the highest values of both surface D_{50} and D^*_{50}. In addition, large boulders, such as those in debris flow deposits, commonly form cascade and step-pool features and exhibit fluted or faceted surfaces: erosional features that indicate these clasts are mobile only during infrequent flood events and are abrading in place. Hence the sorting and delivery of coarse sediment by debris flows, a direct result of the strong hillslope-channel coupling in our study areas, may provide a secondary control on observed patterns of downstream coarsening and locally force relatively high values of surface D_{50} that do not necessarily represent the dynamic response of the bed surface to fluvial processes. Further examination of nonfluvial processes in mountain channels is necessary to isolate the effects of hydraulic sorting and supply caliber on downstream trends in bed surface grain size.

[26] The degree of armoring in our channels is considerably greater than that measured in previous grain size studies of gravel-bed channels, possibly due in part to our small sample volume, as the largest clast found on the surface was often absent in the subsurface sample. Unfortunately, the large caliber of bed surface material encountered in most of our channels prohibits adherence to the sampling protocol proposed by *Church et al.* [1987] and adopted by other workers for gravel-bed channels, whereby the mass of the largest clast is $\leq$1% of the total sample mass. In many instances, this would require field processing nearly 20 tons of subsurface material. However, our sampling methods were consistent with the *Buffington* [1996] method, which found that homogenized subsurface pebble counts approximated bulk sieve distributions fairly well. It is possible that the volume recommended by *Buffington* [1996] (i.e., 1 m^2 to the depth of a shovel blade) is inappropriate for subsurface sampling in steep headwater channels, where the sediment population includes boulders that are $\sim$1 m in diameter. We also acknowledge that

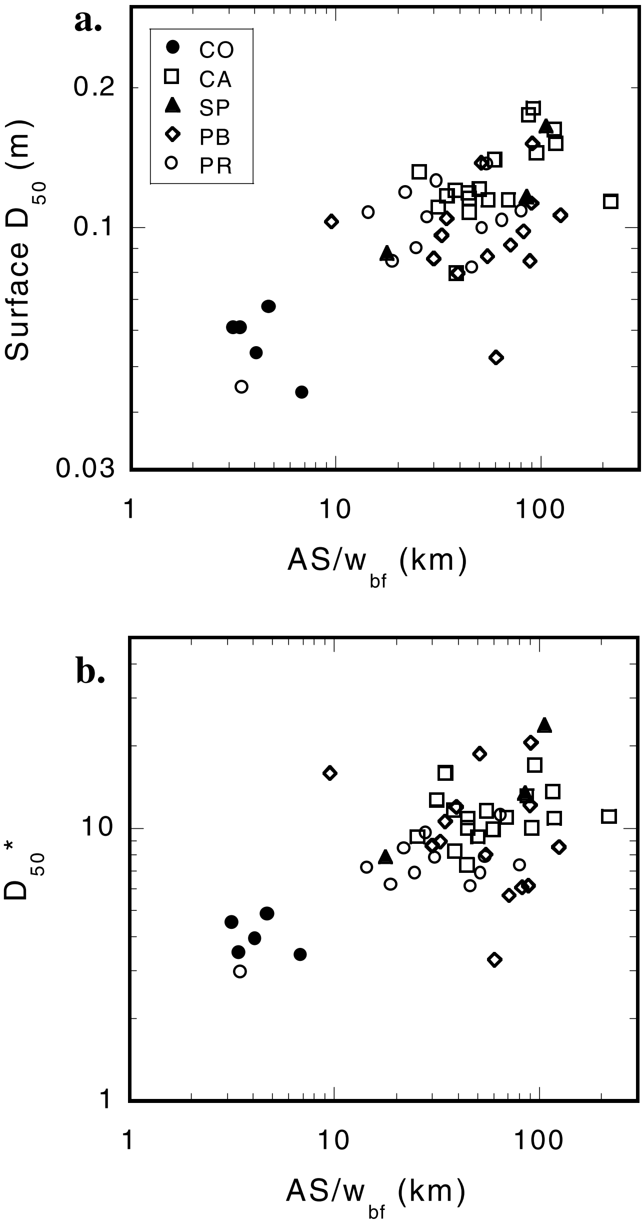

Figure 8. Plots showing relations between the index of unit stream power (AS/w_{bf}) and grain size for the Barnes Creek study area according to channel type. (a) Surface D_{50} increases with unit stream power index according to the log linear relation $D_{50} = 0.045(AS/w_{bf})^{0.23}$ ($r^2 = 0.49$). (b) Relations between unit stream power and armoring (D_{50}^*) parallel those of surface D_{50} : $D_{50}^* = 3.2(AS/w_{bf})^{0.23}$ ($r^2 = 0.36$). More importantly, the upper envelope of grain size data in both plots increases with unit stream power.

removal of suspendable fines from the subsurface grain size distribution, as done by *Buffington and Montgomery* [1999a], could reduce D_{50}^* toward more traditional values. However, bed shear stress, which is necessary for the calculation of suspendable fines, was difficult to infer at all of our sample locations because of high form roughness. In addition, flow-resistance equations that rely on limited velocity measurements and the assumption of a logarithmic velocity profile [e.g., *Wiberg and Smith*, 1991] may not apply to our steep, rough channels [*Byrd et al.*, 2000] and would likely provide an inaccurate estimate of the effective shear stress at the bed. Such complications illustrate shortcomings of conventional models in steep, rough channels and further highlight some of the fundamental differences between headwater channels and their lower-gradient alluvial counterparts.

[27] In general, results of our field studies do not allow discrimination between debris flow-dominated and fluvially dominated channels based solely on drainage area-width relations, except for channels in the heavily logged Finney Creek study area, which have experienced widespread disturbance by recent debris flows. In these reaches, channel widening by debris flow scour may have reduced unit stream power, which, in combination with increased sediment supply in downstream reaches, may have contributed to the relatively low values of D_{50}^* found in this basin. *Montgomery and Gran* [2001] also found that recent scouring of channels by debris flows skewed drainage area-width relations toward wider channels (increase in *c* value) with a concomitant decrease in the *b* value. Field observations, as well as previous studies of landslide frequency and associated sedimentation in the Finney Creek study area, indicate that headwater channels in this basin have been recently impacted by debris flows [*Parks*, 1992; *Cooper*, 1994; *Paulson*, 1997]. The few bedrock reaches surveyed in our study areas plot below the mean for all channel types (Figure 6), a trend that indicates these bedrock channels are narrower than most debris flow- and fluvially dominated channels draining catchments of the same size. This finding is consistent with those of *Montgomery and Gran* [2001] who found lower *c* values for bedrock channels relative to alluvial channels. Although *b* values are more variable in both debris flow-dominated and small bedrock channels, both studies show that the channel-width scaling of combined data for these channels types generally follows the classic hydraulic geometry relations for alluvial channels (i.e., $b \approx 0.5$). On the basis of these results, the systematic, basin-scale variation in unit stream power found in our study areas appears to be controlled largely by changes in profile concavity and to a lesser extent by changes in channel-width scaling. The overall correspondence between the channel-width scaling of debris flow-dominated, bedrock, and alluvial channels supports the use of a constant *b* value of 0.5 for the entire channel network in these small mountain drainage basins. We speculate that bed surface organization (and hence changes in both flow resistance and mean velocity) is more responsive to variations in channel gradient than are adjustments to channel width, which may account for the nearly continuous width function measured across process domains.

[28] Our study areas retain features inherited from previous climate and geomorphic regimes that disrupt the otherwise systematic trends in unit stream power. For instance, discontinuities in drainage area-slope relations and anomalous peaks in unit stream power extracted from DEMs of the Boulder River and Finney Creek basins (Figures 5 and 9) correspond with the hanging-valley knickpoints that flank the North Fork Stillaguamish and Skagit River valleys, respectively. The discontinuity in drainage area-slope scaling at 0.5 km^2 in the South Fork Hoh study area (Figure 5) corresponds with bedrock tributaries incised into glacially

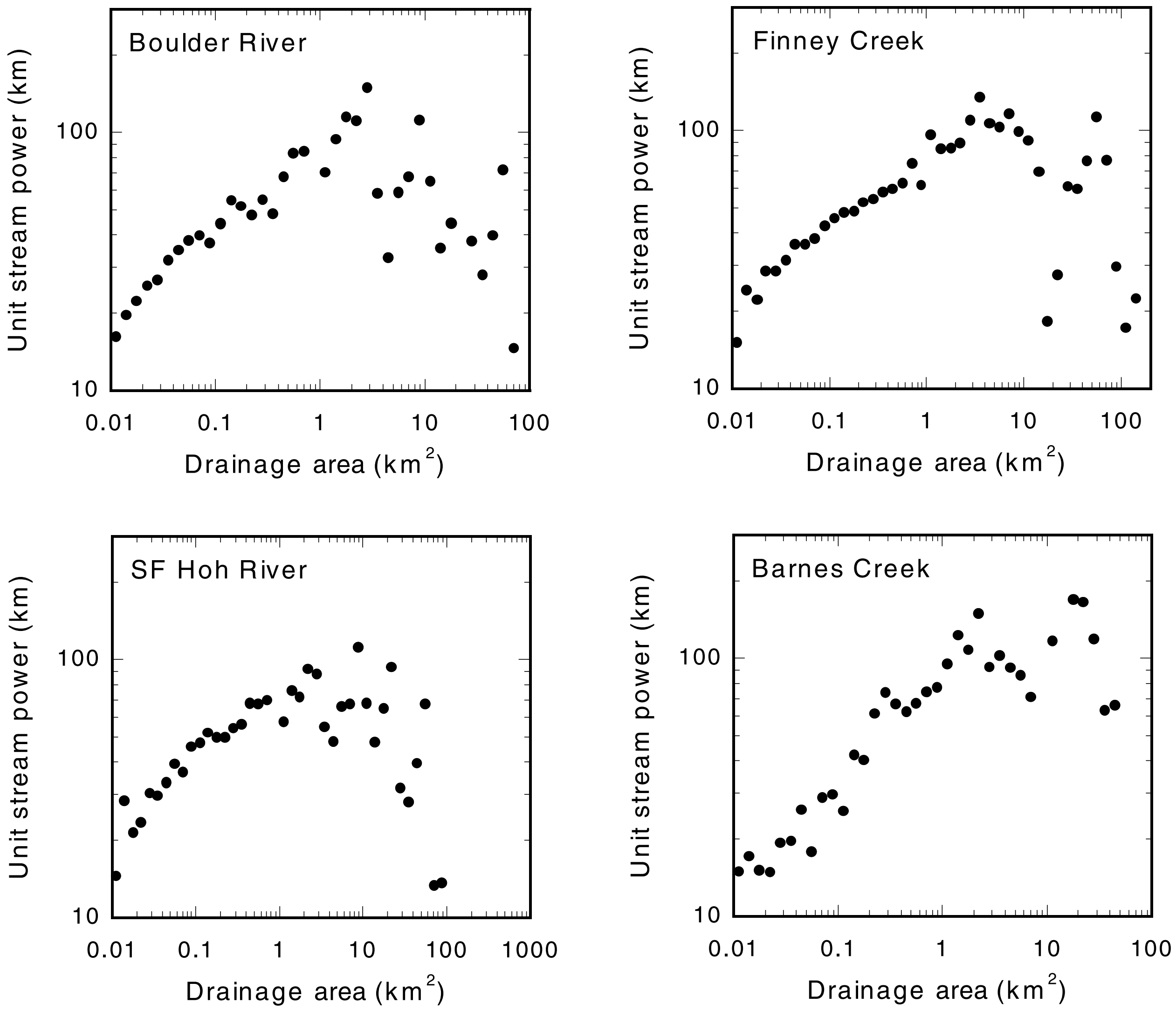

Figure 9. Log bin averaged index of unit stream power $[(1/c)A^{(1-b)}S]$ derived from DEMs versus drainage area for the four study areas.

oversteepened valley walls along the main stem of the South Fork. Reductions in unit stream power and surface D_{50} related to aggradation following base level rise at the outlet of Barnes Creek provide another example of how the adjustment of channel networks to external forcing may be reflected in basin-scale variations in both unit stream power and grain size. In addition, aggregation of drainage area-slope relations from various tributaries within our basins may have diffused the inflection in basin-scale relations and contributed to the scatter apparent in unit stream power-grain size relations. In contrast, surface D_{50} is well correlated with unit stream power index along a continuous stream profile from Barnes Creek. Even though portions of these basins were glaciated, they exhibit strong drainage area-slope relations typical of both hillslope and fluvial processes.

[29] Observations from our field studies can be generalized into basin-scale trends in both surface and subsurface D_{50} (Figure 10). Results of our sediment sampling reveal no systematic downstream trend in subsurface D_{50}. Overlap of surface and subsurface populations occurs in low-order, colluvial channels described by *Montgomery and Buffington* [1997] as exhibiting weak or ephemeral flow. Downstream coarsening occurs in debris flow-dominated channels and shifts to the conventional pattern of downstream fining in alluvial valley segments. Maximum surface D_{50} occurs coincident with maximum unit stream power in cascade and step-pool reaches, channel types that we consider transitional between these two process domains. We posit that the bed material of these hillslope-coupled, transitional reaches is supplied by mass wasting events, which in the case of cascade and perhaps some step-pool channels, forms lag deposits that exhibit mobility thresholds out of phase with the annual hydraulic regime. We further hypothesize that local variability in surface textures reflects the sampling of channels that are at various stages of recovery from debris flow disturbance.

[30] Our conceptual model implies that downstream variations in grain size convey hydraulic information about the state of mountain channels (e.g., roughness, energy expenditure, disturbance history, channel morphology, and sediment supply) and have fundamental implications for

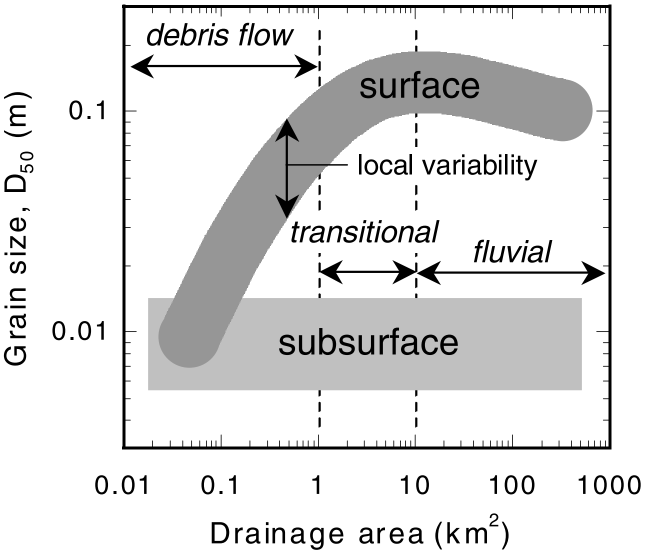

Figure 10. Schematic illustration showing relations between process domains and systematic trends in surface and subsurface grain sizes as inferred from field studies of mountain drainage basins. Results of our sediment sampling reveal no systematic downstream trend in subsurface D_{50}. Overlap of surface and subsurface populations occurs in low-order channels exhibiting weak or ephemeral flow. Downstream coarsening occurs in debris flow-dominated channels and shifts to the conventional pattern of downstream fining in fluvial valley segments. Maximum surface D_{50} occurs coincident with maximum unit stream power in transitional reaches between these two process domains. In addition, local variability in surface grain size may be forced by debris flow deposits.

models of sediment transport, landscape evolution, and riverine ecosystems that are often applied over a wide range of both spatial and temporal scales. The heterogeneity of specific channel properties (e.g., drainage area-slope relations, unit stream power, grain size, and roughness) measured at the spatial scale of the transition in process domains (1–10 km^2 for our basins) runs counter to principles underlying the optimal channel network concept that form the basis for a number of theoretical models of river-system evolution. Our results parallel findings of previous studies on the upper slope limit of the bedrock incision law [e.g., *Sklar and Dietrich*, 1998; *Whipple and Tucker*, 1999; *Snyder et al.*, 2000; *Stock and Dietrich*, 2003]. However, these models may not apply to the majority of the drainage network length in mountain drainage basins, since first- and second-order channels account for nearly 70% of the total channel length in mountain drainage basins [*Shreve*, 1967]. For the same reason, downstream coarsening would appear to characterize longitudinal trends in grain size for more of the total channel network length in mountain rivers than the traditionally recognized pattern of downstream fining.

6. Conclusions

[31] We present results from both field and DEM analyses of four mountain drainage basins in western Washington that describe the character of downstream coarsening, a largely unexplored but potentially widespread pattern of grain size variation in headwater channels that runs counter to conventional patterns of downstream fining. Our results build upon previous studies to show that headwater channels are fundamentally different from alluvial channels in many respects (e.g., drainage area-slope relations, increasing unit stream power, hillslope-channel coupling, and downstream coarsening) but share similar elements of downstream scaling of channel width. Downstream coarsening is found in basins up to 10 km^2 but is most prominent below a drainage area of ~1 km^2, which corresponds to the inflection in drainage area-slope relations thought to represent the transition from debris flow-dominated headwater channels to fluvially dominated channels. We assert that the shift from downstream coarsening to downstream fining can be thought of as an additional morphological indicator of this fundamental transition. The correspondence between grain size, unit stream power, and drainage area-slope relations, which is illustrated best in the study area where sampling occurred along a single profile, suggests that downstream coarsening is controlled primarily by the systematic increase in unit stream power, with second-order controls from the temporary accumulation of lag deposits forced by mass wasting events. We suspect that downstream coarsening is common in landscapes that exhibit a drainage area-slope inflection (i.e., those that are dominated by debris flows) and therefore that downstream coarsening may be a ubiquitous characteristic, rather than a curious exception, in headwater channels of mountain drainage basins.

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