Influence of mass wasting on bed-surface armoring, lag formation, and sediment storage in mountain drainage basins of western Washington State

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This is to certify that I have examined this copy of a doctoral dissertation by

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Abstract

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This dissertation uses field studies and analyses of digital topography in western Washington State to investigate 1) hydraulic controls on the spatial variation in bed-surface armoring, 2) grain-size controls on the formation of resistant “lag” deposits, and 3) grain-size and lithologic controls on the dispersion rate and residence time of sediment in mountain channels. Field and topographic analyses document systematic downstream coarsening of the median bed-surface grain size in channels with a drainage area < 1 km^2 and a subsequent shift to the conventional pattern of downstream fining at a drainage area of about 10 km^2. The grain-size maximum corresponds with the maximum unit stream power and the inflection in the drainage area-slope relation thought to represent the transition from debris flow-dominated channels to fluvially dominated channels. The results suggest that basin-wide trends in bed-surface armoring are hydraulically controlled by systematic variations in unit stream power and should therefore be common in headwater channels where debris-flow processes set the channel gradient. Results of a four-year case study of channel recovery from blockage by a landslide dam indicate that, in addition to hydraulic controls, trends in bed-surface armoring are forced by the temporary accumulation of lag deposits arising from the reworking of mass-wasting deposits. Results indicate that the caliber of supplied sediment relative to the flow competence of the receiving channel can influence the rate of sediment-pulse dispersion and the residence time of sediment storage. Measurements of sediment volumes stored in recent and ancient landslide dams in western Washington indicate that the coarsest deposits retain > 80% of initial debris after 10^3 years, exhibit long-term incision rates that are within one to two orders of magnitude of regional bedrock-lowering rates, and create backwater lakes that provide sediment "capacitance" to mountain drainage basins. Results of this research have practical implications for understanding the residence time of sediment storage in mountain river systems, the disruption of sediment routing, the enhancement of fluvial relief through the shielding of bedrock surfaces, and the timescale of channel recovery following natural and anthropogenic disturbances.

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DEDICATION

To Zoe

CHAPTER 1: Introduction

Most confined mountain channels display a heterogeneous mixture of bed-surface particles regularly mobilized as bedload around a disorganized arrangement of apparently immobile boulders (Figure 1.1). Stream channels of this type are said to be "coupled" to adjacent hillslopes in that their morphology directly reflects fluvial sorting of non-fluvial sediment delivered by debris flows and landslides (Dietrich and Dunne, 1978; Caine and Swanson, 1989; Benda, 1990). Because the largest particles delivered to a channel may not be regularly mobilized by the stream, they accumulate as a "lag" deposit and influence channel hydraulics, sediment-transport rates, and residence time of sediment storage in mountain drainage basins. Although considerable effort has been devoted to studying the dispersion of sediment "pulses" or "waves" supplying particle sizes readily transported by typical flows (e.g., Gilbert, 1917; Lisle, 1982; Roberts and Church, 1987; Perkins, 1989; Madej and Ozaki, 1996; Lisle et al., 2001; Sutherland et al., 2002; Cui et al., 2003a, 2003b), the rates and mechanisms by which mountain channels assimilate coarse sediment inputs are poorly understood. For instance, how do mountain channels — while actively incising mountain landscapes — process sediment mixtures containing a significant fraction of coarse particles too large to be transported as bedload by typical high flows? How do lithology and network topology control the spatial distribution of lag deposits resulting from coarse sediment inputs? Do lag deposits influence dispersion rates of sediment inputs and residence time of sediment storage in mountain drainage basins? This dissertation addresses these questions through studies of bed-surface armoring, lag formation, and sediment storage in mountain drainage basins of western Washington State (Figure 1.2).

Early investigations of mountain channels recognized fundamental differences between coarse-grained, high-gradient headwater channels and low-gradient channels formed in finer alluvium (Gilbert, 1877; Miller, 1958; Leopold et al., 1964). Schumm (1977) provided an early framework for examining geomorphic processes in mountain channels by dividing the channel network into zones of erosion, transport, and deposition. Montgomery and Buffington (1997) linked the general downstream progression of channel-reach morphologies observed in mountain drainage basins to disturbance from debris flow scour and deposition. Despite an awareness of fundamental differences between headwater and alluvial reaches, predictive models developed from studies of low-gradient alluvial channels are continually applied to high-gradient, coarse-grained channels. For example, it has been widely observed that the sediment size in alluvial rivers decreases downstream in concert with declining channel gradient and the capacity to move sediment of a particular size (e.g., Paola et al., 1992, Ferguson et al., 1996; Wilcock, 1997; Gasparini et al., 1999; Hoey and Bluck, 1999). Due in part to the recruitment of coarse sediment from mass wasting, this trend has been posited by some to extend into the headwaters of mountain channel networks (Church, 2002). However, studies of river networks (e.g., Magilligan, 1992; Seidl and Dietrich, 1992; Montgomery and Foufoula-Georgiou, 1993; Lecce, 1997; Knighton, 1999; Stock and Dietrich, 2003) suggest that the transport capacity of headwater channels initially increases with drainage area, which would suggest that downstream fining may not apply well in the headwaters of mountain channels.

Landscape conditions unique to headwater channels therefore raise interesting questions regarding application of traditional models to the entire channel network. Does the conventional model of downstream fining extend throughout the entire channel network? Is the grain-size trend (if any) in headwater channels hydraulically controlled by variations in transport capacity, or is it instead determined by caliber and quantity of sediment supplied to the channel by mass wasting? Understanding the spatial distribution of sediment character in mountain channel systems requires an integrated study of systematic downstream patterns in transport capacity, longitudinal variations in surface grain size, and downstream sorting and delivery of sediment by mass-wasting processes.

The heterogeneous mixture of mobile sediment and immobile (lag) boulders observed in many mountain streams brings into question the application of surface-based, bedload-transport formulae to these channels unless the immobile particles can be segregated from the mobile fraction. Nonetheless, such formulae are the underpinnings of recent models of sediment-pulse dispersion that aim to predict how a channel responds to large-scale disturbances (e.g., Lisle et al., 2001; Sutherland et al., 2002; Cui et al., 2003b). These models explicitly neglect the effects of immobile structures found on the beds of many hillslope-coupled channels. Because 70 – 90% of the total network channel length in mountain drainage basins is composed of channels dominated by mass wasting (Shreve, 1967; Stock and Dietrich, 2003), the current suite of surface-based models may not apply to the majority of mountain channel networks.

This dissertation uses field studies and topographic analyses of digital topography to investigate hydraulic controls on spatial variation in bed-surface armoring in headwater channels of Washington State. This research also documents lag formation during a four-year field study of channel degradation following blockage of a mountain stream by a bedrock landslide. Finally, field surveys of recent and ancient valley-spanning landslide deposits are used to examine the influence of lithology and particle-size distribution of sediment inputs on the dispersion rate and residence time of sediment storage in mountain channels.

Description of the study area

The Cascade Range and Olympic Mountains of western Washington State provide a natural laboratory to study the influence of hillslope processes on bed-surface armoring, lag formation, and sediment storage in mountain river systems (Figure 1.2). Both mountain ranges have been tectonically active since at least the middle Miocene and are considered to be in erosional steady state (Brandon et al., 1998; Reiners et al., 2003). Excluding Cascade volcanoes, summits in both ranges do not exceed 2.7 km. Long-term uplift rates of 0.1 – 1.0 mm yr^{-1} promote frequent landsliding throughout both mountain ranges (Montgomery and Brandon, 2002; Reiners et al., 2003). Annual rainfall west of the Cascade Range is comparatively high relative to inland areas of Washington State and ranges from less than 1 m to over 5 m at the highest elevations (NRCS, 1999). Runoff from snowmelt and regionally high precipitation assures a high drainage density and abundance of suitable study sites. High orographic precipitation on windward sides of both ranges also promotes rapid fluvial response of mountain channel networks to disturbances from landsliding.

The North Cascades and Olympic Mountains include a diverse array of geologic rock types, which allows the investigation of lithologic controls on the particle-size composition of sediment delivered to mountain channels by mass wasting. Diverse bedrock found throughout the region also allows investigation of lithologic controls on dispersion rates and timescales of sediment storage. Rocks of the Cascade Range include pre-Cenozoic metamorphic and intrusive igneous rocks and less-resistant Tertiary volcanic and sedimentary rocks (Dragovich et al., 2002). The highly sheared and fractured marine sediments exposed in the core of the Olympic Mountains contrast with the resistant Crescent Formation basalt encircling the range on three sides (Dragovich et al., 2002).

The Cascades and Olympics are ideal settings to study the reestablishment of the fluvial network since the retreat of glacial ice about 10,000 – 13,000 years ago and to test whether the recent fluvial signal can be "read" through the prevailing glacial signal. The valley morphology of both mountain ranges has been sculpted by repeated advances and retreats of the Cordilleran Ice Sheet and alpine glaciers. The Puget Lobe of the Cordilleran Ice Sheet penetrated most major river valleys in the North Cascades and blocked the valley outlets on the north and east sides of the Olympics. Alpine glaciers flowing west from the core of the Olympics scoured most westward-draining valleys and reached the Pacific Ocean at their maximum extent. Pleistocene glaciers deposited vast amounts of outwash sediment in the valley bottoms of both mountain ranges (Crandall, 1965; Booth et al., 2004).

The study area is located in the only coastal temperate rainforest of North America. Prior to arrival of settlers in the late 1800s, low and middle elevations within the region were covered with a dense coniferous forest, with deciduous trees restricted to frequently disturbed areas (Pojar and MacKinnon, 1994). Most upland areas of the region are now managed as a vast industrial forest, with remnants of the ancient old-growth forest restricted mainly to national parks and wilderness areas located in mountain interiors. Field sites for this study were located in both old-growth and industrial forests. Although this study did not focus directly on interactions between fluvial processes and woody debris, field observations during the course of this research are consistent with findings from prior studies of wood in the region (e.g. Abbe, 2000; Brummer et al., 2006), which found the morphologic influence of wood on mountain channels to be profound.

The abundance of natural landslide dams in the North Cascades and Olympic Mountains presents an unusual opportunity to study lag deposits at various stages of development by trading space for time. Ages of many ancient landslide dams in the study area are well constrained through radiocarbon dating of submerged forests (Schuster et al., 1992; Logan et al., 1998; Pringle et al., 1998). In addition, the sediment trap formed by a landslide-dammed lake sets up the supply-limited conditions necessary to perform a controlled experiment of lag formation. In the absence of an upstream sediment flux, mobile and immobile components of a supply distribution can be segregated by comparing particle-size distributions both of the supply and the bed surface composing the lag. Finally, mechanisms governing the decay of landslide-induced sediment pulses can be examined by comparing degradation rates among landslide dams of various size, age, lithology, and particle-size distribution.

Overview of dissertation

This dissertation is organized into three main chapters. Chapter 2, entitled "Downstream coarsening in headwater channels" describes 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. Field data from four mountain drainage basins in western Washington document systematic downstream coarsening of the median bed-surface grain size (D_{50}) below a drainage area of ~1 km^2 and a subsequent shift to downstream fining at a drainage area of about 10 km^2. Field surveys and analyses of digital topography were used to derive network-wide patterns of unit stream power (represented by *AS/w*, where *A* is drainage area, *S* is channel slope, and *w* is channel width). Results of the analysis 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. The results support the hypothesis that basin-wide trends in D_{50} are hydraulically controlled by systematic variations in unit stream power, with second-order controls from the temporary accumulation of lag deposits forced by mass wasting. Similar relations found in the 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.

Chapter 3, entitled “The influence of coarse lag formation on the mechanics of sediment pulse dispersion” presents results from a four-year field study documenting lag formation during the degradation of a landslide that temporarily dammed Squire Creek in the North Cascades of western Washington. Rapid bed-surface armoring of the new channel by selective removal of the finest sediment during normal flows formed a relatively immobile lag deposit within two years. A functional relationship between total shear stress in the channel and probability that a particle of a particular size would be removed from the sediment pulse was derived from a tracer study and a comparison of supply and bed grain-size distributions. The relationship showed that covering of 20% of the bed by lag boulders with < 5% probability of entrainment was sufficient to retard vertical incision and to force considerable channel widening during a high-magnitude flood with an 8- to 152-year recurrence discharge on locally gauged streams. Results of the field study at Squire Creek indicate that the grain-size distribution of sediment supply relative to the flow competence of the receiving channel can influence the mechanism and rate of sediment-pulse dispersion and the potential for long-term sediment storage. Numerical models of sediment-pulse evolution that do not explicitly incorporate the influence of lag formation may substantially overestimate long-term dispersion rates. Shielding of sediment reservoirs by a persistent lag deposit has implications for the function of coarse-grained debris dams as large sediment capacitors in mountain channels confined by steep valley slopes prone to landsliding.

Chapter 4, entitled "Landslide dams and sediment capacitance in mountain river systems" uses measurements of sediment volumes stored in recent and ancient landslide dams in the Olympic Mountains and Cascade Range of Washington State to examine lithologic and grain-size controls on the dispersion of sediment pulses and the potential for long term sediment storage. Results of field surveys and analyses of digital topography indicate that the ratio of the supply caliber to the total flow energy expended across a landslide dam controls whether an episodic sediment input behaves as a point source of sediment input or as a long-term sediment storage element. The latter may provide sediment capacitance and ecological diversity to mountain river systems.

Chapter 4 also introduces a particle-size-based classification for landslide dams to predict the relative tendency for rapid sediment dispersion or sediment capacitance based on the particle D_{90} of the degrading bed surface and on the local unit stream power (represented by $A^{0.5}S$). Measurements of sediment volumes exported from ancient landslides by fluvial incision indicate that more than 80% of debris in the coarsest deposits is retained in storage after 10^3 years. In contrast, fine-grained landslides lost more than 80% of their initial volume available for channel reworking within a decade. The long-term average incision rates for the coarsest landslide dams examined in this study were within one to two orders of magnitude of regional bedrock-lowering rates. The long residence time of sediment impoundments created by coarse-grained landslides provides potential storage volumes that are up to three times the volume of the landslide dam from which they are formed. Based on regional erosion rates, the time to charge sediment capacitors examined in this study ranges from 10^2 – 10^3 years for the largest capacitors to < 10 years for the smaller impoundments formed in large rivers. Results of this study indicate that bedrock shielding by coarse lag deposits may temporarily retard bedrock river incision, with implications for models of landscape evolution and models of the development of "fluvial" relief in mountain drainage basins.

Chapters 2, 3, and 4 all illustrate fundamental differences between alluvial channels and headwater channels with respect to drainage area-slope relations, increasing unit stream power, downstream coarsening, hillslope-channel coupling, the mismatch between stream competence and the caliber and quantity of supplied sediment, and the effects of armoring by lag deposits on sediment-dispersion rates and timescales of sediment storage. The shift from downstream coarsening to downstream fining described in Chapter 2 can be used as a morphological indicator of the fundamental transition from debris-flow dominated and hillslope-coupled channels to fluvially dominated channels. Downstream of the transition from debris-flow scour to debris-flow deposition, bed shielding by lag deposits can provide a self-regulating mechanism by which mountain channels modulate sediment flux from coarse-grained sediment inputs. This research shows that the rate constants characterizing degradation of sediment pulses vary over time. This research also points to the need for bedload transport equations adapted to the hydraulically rough and heterogeneous sediment mixtures characteristic of hillslope-coupled mountain channels. Research into the role of lag deposits has practical implications for the residence time of sediment storage in mountain river systems, the disruption of sediment routing schemes employed in sediment budgets, the enhancement of fluvial relief through the shielding of bedrock surfaces, and the generation and maintenance of ecological diversity in mountain drainage basins. Deciphering the complex geomorphic processes in headwater channels is key to understanding the response and recovery of mountain drainage basins to natural and anthropogenic disturbances.

Figure 1.1: Typical morphology of a mountain stream in western Washington. The heterogeneous mixture of bed-surface sediment displayed on the bed surface includes sediment transported regularly as bedload around disorganized lag boulders delivered to the channel by periodic mass wasting.

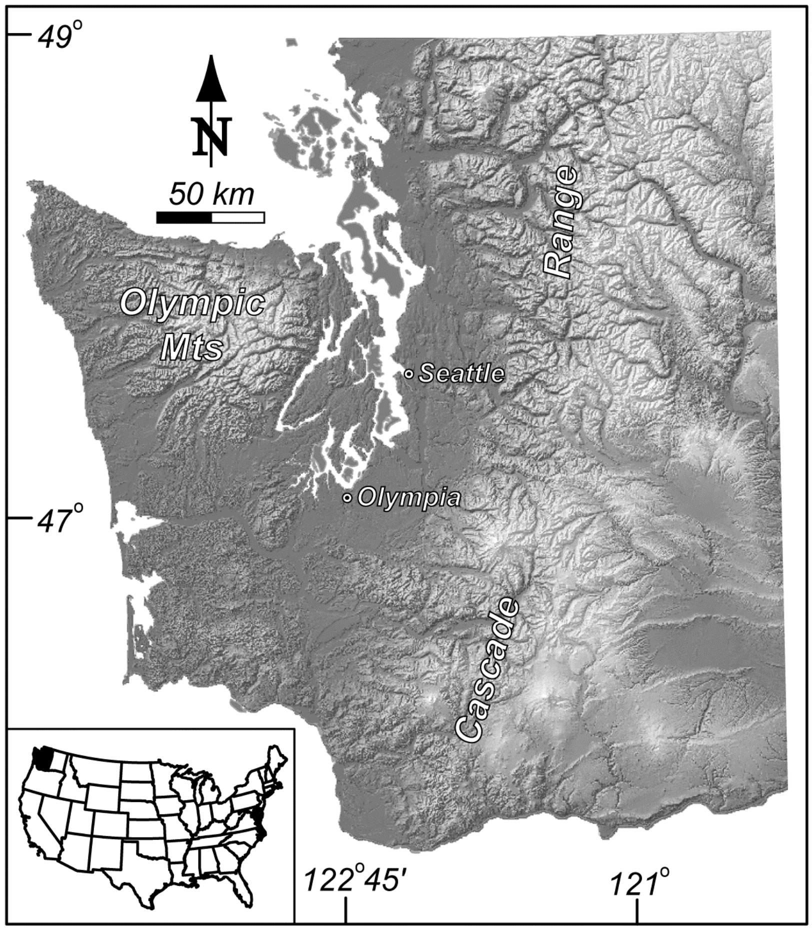

Figure 1.2: Shaded-relief map of western Washington.

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CHAPTER 2: Downstream coarsening in headwater channels

Summary

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. The 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 the 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.

Introduction

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).

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 2.1). Under steady-state conditions, channel slope (S) has been observed to vary empirically as a function of drainage area (A):

$$S = kA^{-\theta} \qquad \text{Eq. 2.1,}$$

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).

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:

$$\omega = \frac{\rho g Q S}{W} \qquad \text{Eq. 2.2,}$$

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

$$Q = eA^d \qquad \text{Eq. 2.3},$$

where e and d are determined empirically. The coefficient d typically ranges from 0.7 for semi-arid regions to 1.0 for humid landscapes that drain small catchments, such as the field sites in this study (Dunne and Leopold, 1978; Leopold, 1994; Rice, 1998; Whiting et al., 1999). Combining Equations 2.2 and 2.3 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):

$$\omega = \frac{\rho g e A^d S}{W} \qquad \text{Eq. 2.4.}$$

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):

$$W = cA^b \qquad \text{Eq. 2.5},$$

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 2.4 and 2.5 may be combined and simplified into a form readily extracted from DEMs:

$$\omega = \rho g (e/c) A^{(d-b)} S \qquad \text{Eq. 2.6},$$

where the coefficient $\rho g(e/c)$ is retained to render Equations 2.2 and 2.6 dimensionally equivalent. Based on 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 2.1 and 2.6 allows unit stream power to be expressed as a function of drainage area:

$$\omega = c_2 A^{(d-b-\theta)} \qquad \text{Eq. 2.7,}$$

where the constants in Equations 2.1 and 2.6 have been folded into c_2. The variation in unit stream power with drainage area can be evaluated from the first derivative of Equation 2.7:

$$\frac{d\omega}{dA} = c_2 (d - b - \theta) A^{(d-b-\theta-1)} \qquad \text{Eq. 2.8.}$$

Equation 2.8 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 2.1).

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. Field data and results of

DEM analyses from four mountain drainage basins in western Washington that document systematic downstream bed-surface coarsening are presented in this chapter to evaluate the correspondence of these trends to downstream variations in unit stream power.

Study areas

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 2.2). 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.

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 of (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.

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 has 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.

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.

Methods

Recent field data collected during 2001 and 2002 field seasons are combined 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. This sampling protocol was chosen 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.

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 corresponding to the recurrence interval of bankfull 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.

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.

Simplified forms of Equations 2.4 and 2.6, along with the assumption that discharge varies linearly with drainage area (i.e., $d = 1$), were used 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 bankfull 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.

Results

The 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. Field data sampled randomly from different tributaries were compared with data sampled systematically along a single longitudinal profile. Finally, results of DEM analyses were used to complement generalizations inferred from limited field data.

Downstream coarsening

A systematic increase in surface D_{50} with drainage area is found in basins smaller than ~1 to 10 km^2 (Figure 2.3). 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 the 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.

Combined field data from pristine basins further emphasize the positive dependence of surface D_{50} and D_{50}* on drainage area (Figure 2.4). 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 the 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.

Drainage area-slope analysis

Log-bin averaged drainage area-slope relations derived from 10-m-grid-size DEMs of the channel networks in this study exhibit relations of the general form described by Montgomery and Foufoula-Georgiou (1993) and found in mountain channel networks by previous workers (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 2.5). 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.

Channel width

Channel width increases with drainage area according to Equation 2.5 for both debris flow-dominated and fluvially dominated reaches surveyed in the four study areas (Figure 2.6). 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 2.1). 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 the study areas plot below the trend of the regressions; hence, they are narrower than most debris flow- and fluvially dominated channels.

Unit stream power indices

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 Barnes Creek study areas (Figure 2.7). The lowest values of unit stream power occur in both colluvial and pool-riffle channels located at the smallest and greatest drainage areas of the drainage networks analyzed in this study, 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.

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}* (cf. Figures 2.4 and

2.7). Not surprisingly, there are weak, but significant log-linear relationships between unit stream power index and both surface D_{50} (r^2 = 0.71) and D_{50}* (r^2 = 0.75) along the Barnes Creek Profile (Figure 2.8). The degree of correlation varies amongst 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.

DEM-based indices of unit stream power [$(1/c)A^{(1-b)}S$] were predicted from 10-m grid size DEMs and Equation 2.6 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 2.9). 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 (cf. Figures 2.7 and 2.9) and predicted by Equation 2.8 using an assumed *d* value of 1.0 and appropriate *b* values for each basin (Table 2.1). 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 2.5). 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 nickpoints.

Discussion

It is posited 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.

At least two factors are attributed 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 2.8 and illustrated graphically in Figure 2.1, the downstream rate of change in unit stream power is determined by the sign of ($d - b - \theta$), which, for the 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.

Unlike alluvial channels, where slope is an additional degree of freedom that may adjust in response to feedbacks between transport capacity and sediment supply, the findings of this study point to grain size as a key dependent variable above this transition in steep headwater channels. Results of the field investigations and DEM analyses suggest that 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. It is further hypothesized 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.

Previous studies have shown that mechanical sorting by debris flows tends to focus the largest clasts to the flow front. Based on 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 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 the 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.

The degree of armoring at the study sites is considerably greater than that measured in previous grain-size studies of gravel-bed channels, possibly due in part to the 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 the 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 ≤ 1% of the total sample mass. In many instances, this would require field processing nearly 20 tons of subsurface material. However, the 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 ~ 1 m in diameter. Furthermore, 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 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 steep, hydraulically rough channels (Byrd et al., 2000), such as found at the field sites, and would likely provide an inaccurate estimate of the effective shear stress applied to 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.

In general, results of the 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 the study plot below the mean for all channel types (Figure 2.6), 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., $b \approx 0.5$). Based on these results, the systematic, basin-scale variation in unit stream power found in the 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. It is hypothesized 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.

Study sites 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 2.5 and 2.9) correspond with the hanging-valley nickpoints 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 2.5) corresponds with bedrock tributaries incised into glacially 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 the study 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.

Observations from the field studies can be generalized into basin-scale trends in both surface and subsurface D_{50} (Figure 2.10). Results of 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 considered to be transitional between these two process domains. It is posited 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. It is further hypothesized that local variability in surface textures reflects the sampling of channels that are at various stages of recovery from debris-flow disturbance.

The 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 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 the study 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. Results of this investigation 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.

Conclusions

Results from both field and DEM analyses of four mountain drainage basins in western Washington are presented 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. Results of this study 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. It is asserted 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. It is suspected 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.

Table 2.1: Summary of hydraulic geometry relations and basin parameters.

Study Area/Domain‡	Hydraulic Geometry Relations ($w = cA^b$) c†	b	R^2	n	Basin Parameters θ	R^2	$(d\text{-}b\text{-}\theta)$¶
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§	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§	0.97	-0.43
Combined	3.7 +/- 0.5	0.43 +/- 0.05	0.87	52			

† c values reported in m, drainage area in km^2
‡ 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.
¶ Coefficient in Equation 2.8 calculated with an assumed d value of 1.0.
§ 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.

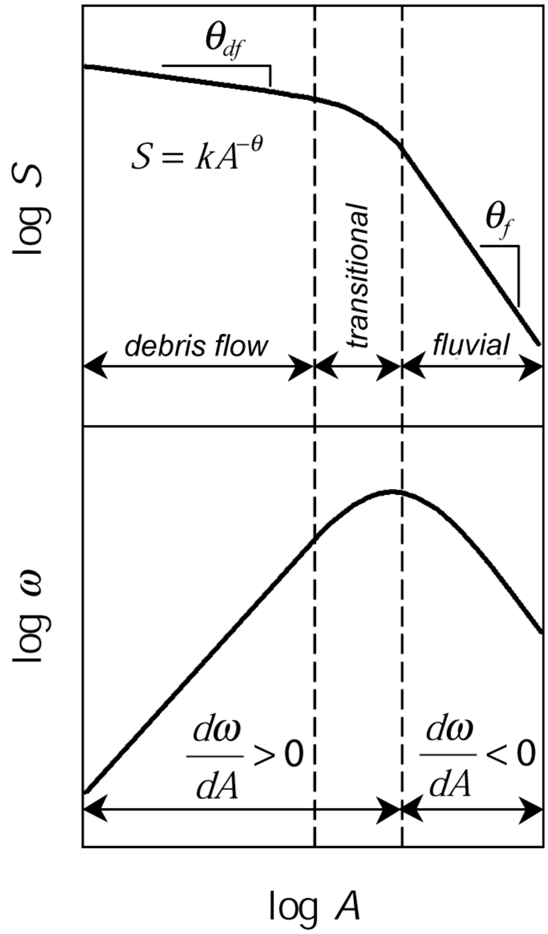


Figure 2.1: Schematic illustration of the influence of drainage areas-slope relations on the rate of change in unit stream power predicted by Equation 2.8. The transition in concavity index between a debris-flow process domain (θ_{df}) and fluvial process domain (θ_f) should correspond with the peak in unit stream power.

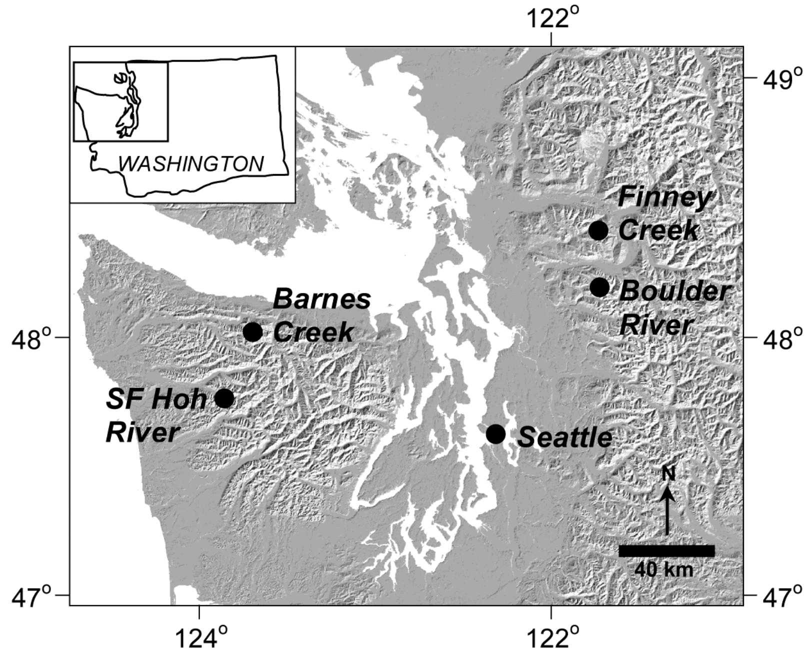


Figure 2.2: Shaded-relief map of western Washington showing locations of study areas.

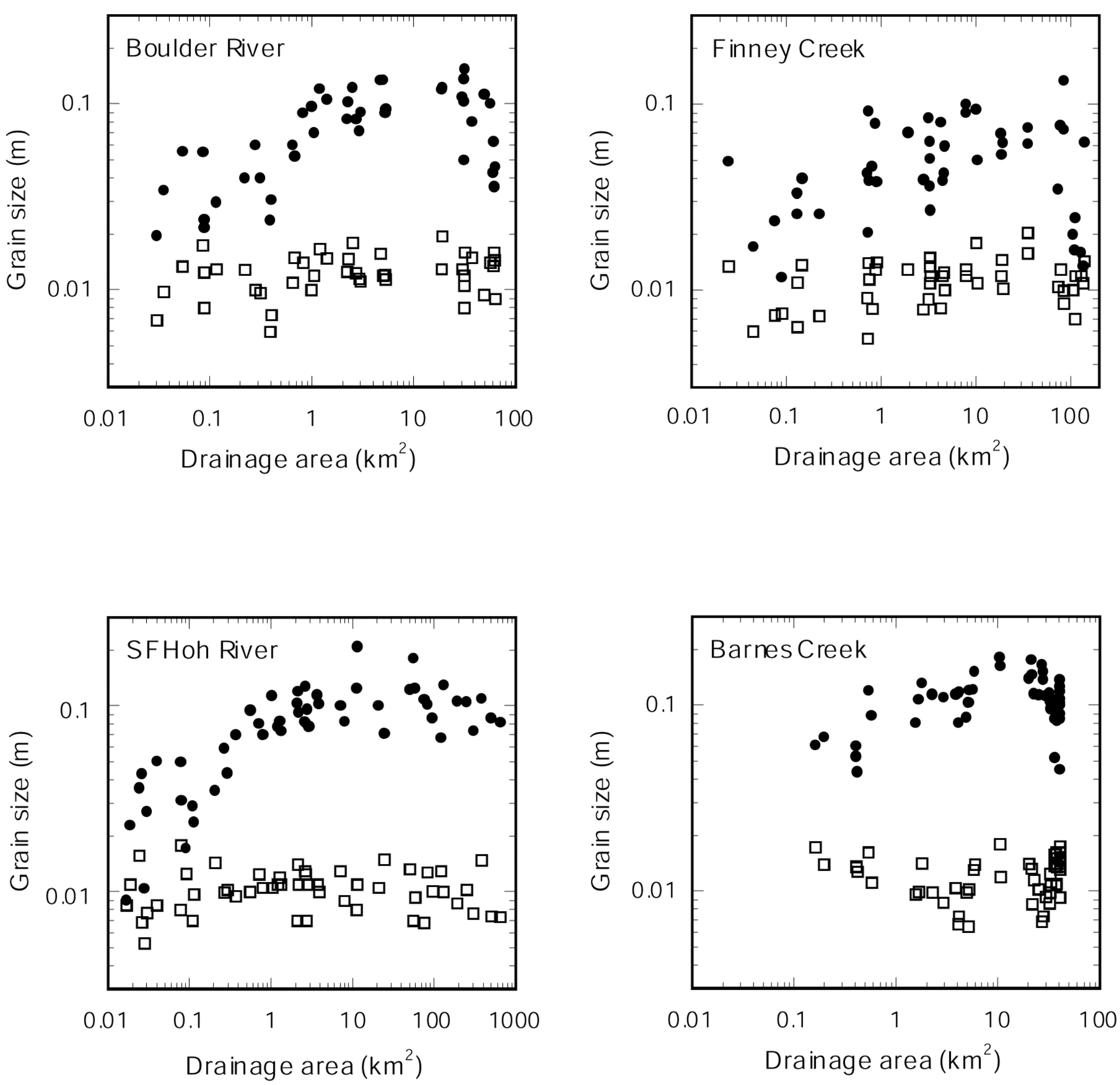
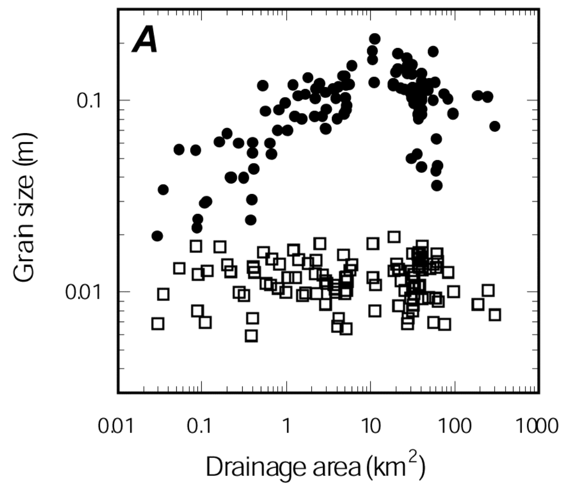


Figure 2.3: Plots of grain size (D_{50}) versus drainage area for the four field sites. Plots show the transition from downstream coarsening to downstream fining. Symbols represent surface D_{50} (solid circles) and subsurface D_{50} (open boxes). Note the different horizontal scales.



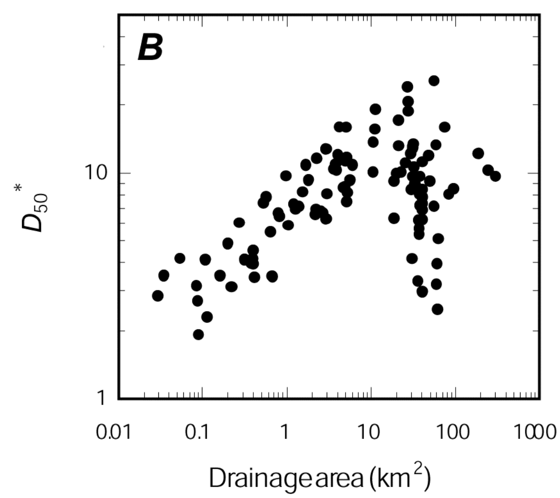


Figure 2.4: Composite plots of grain size (D_{50}) and armoring (D_{50}*) versus drainage area for pristine watersheds. A) Downstream coarsening of surface D_{50} occurs in channels draining basins less than ~ 10 km^2, whereas rapid downstream fining occurs where basins drain into large glaciated valleys (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 2.3. 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: $D_{50}* = 6.5A^{0.29}$ ($r^2 = 0.75$).

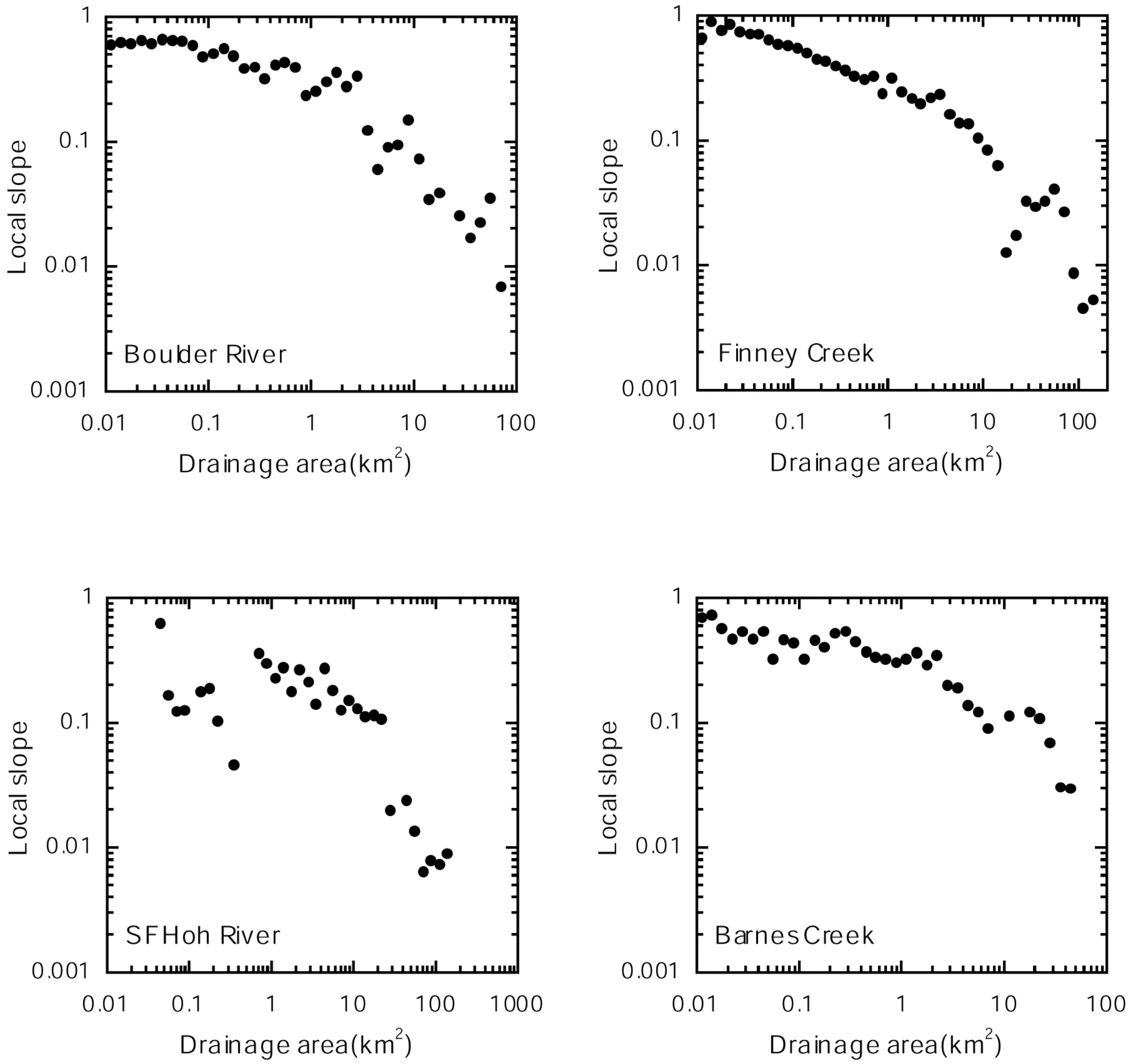


Figure 2.5: Log-bin averaged local slope derived from 10-m-grid-size DEMs versus drainage area for the four study areas.

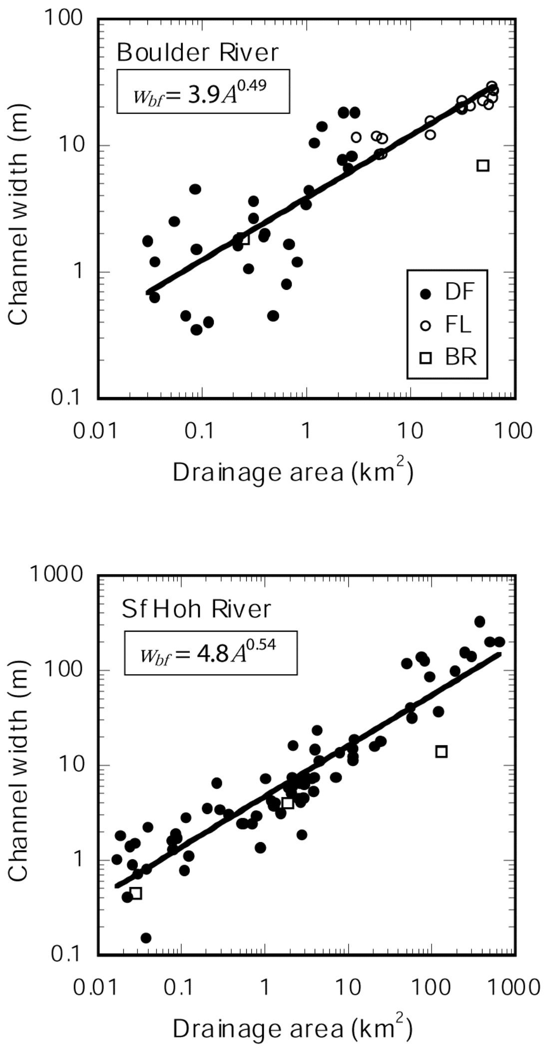


Figure 2.6: Plots of relations between channel width and drainage area. Relations show debris flow (DF), fluvial (FL), and bedrock (BR) channels in the Boulder River, Finney Creek, and Barnes Creek study areas and combined channel types in the South Fork Hoh River study area. Power function regressions 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.

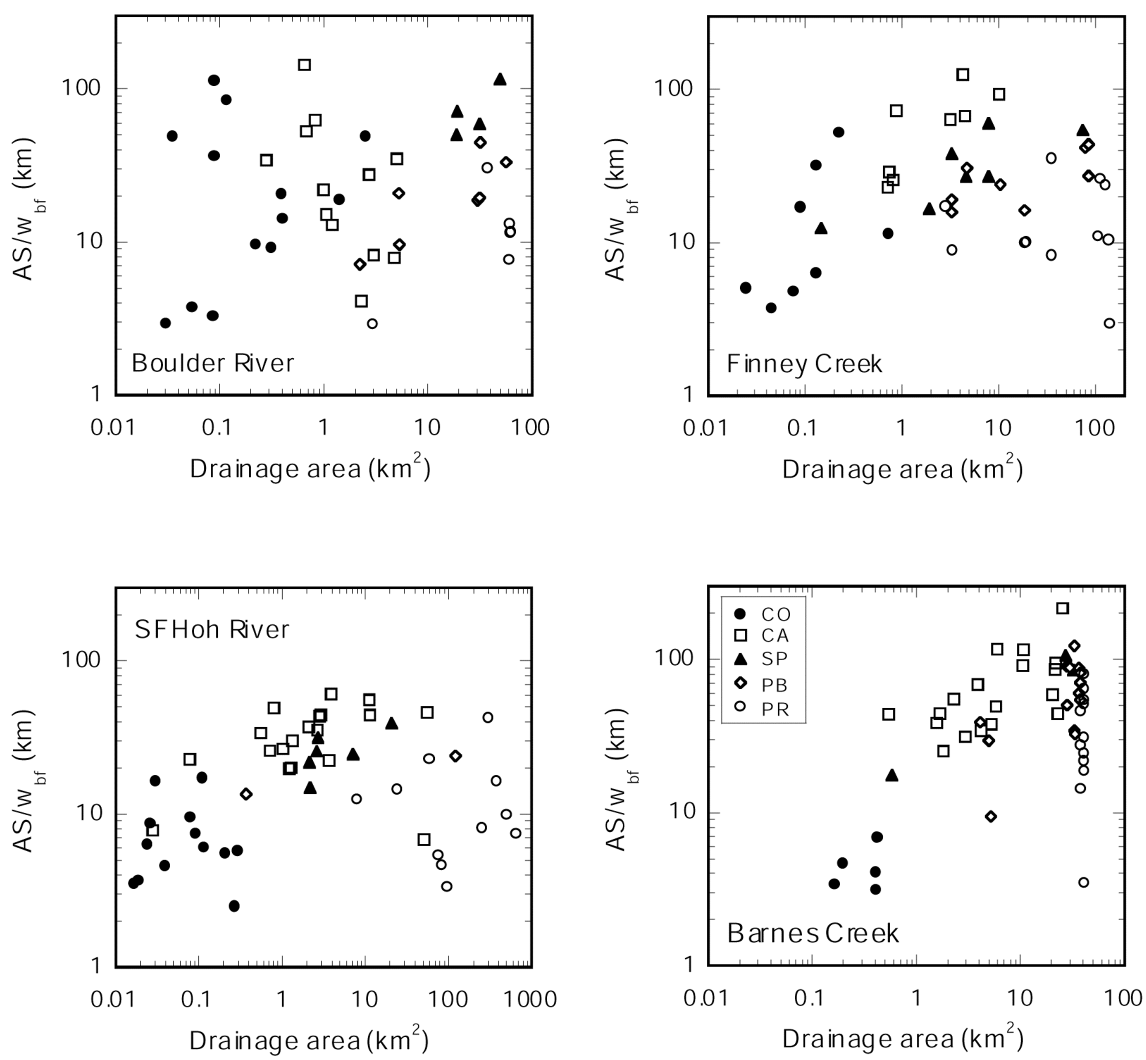


Figure 2.7: Plots of the index of unit stream power (AS/w_{bf}) versus drainage area by channel type. Channel classification follows Montgomery and Buffington (1997): colluvial (CO), cascade (CA), step pool (SP), plane bed (PB), and pool riffle (PR).

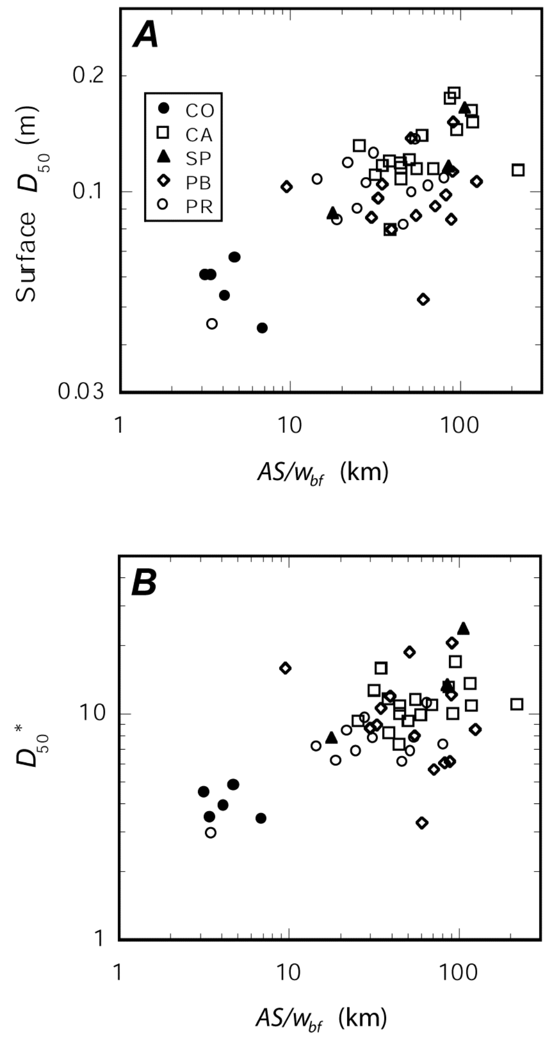


Figure 2.8: Plots of relations between the index of unit stream power (AS/w_{bf}) and grain size for the Barnes Creek study area according to channel type. A) Surface D_{50} increases with unit stream power index according to the log-linear relation $D_{50} = 0.045(AS/w_{bf})^{0.23}$ ($r^2 = 0.49$). 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.

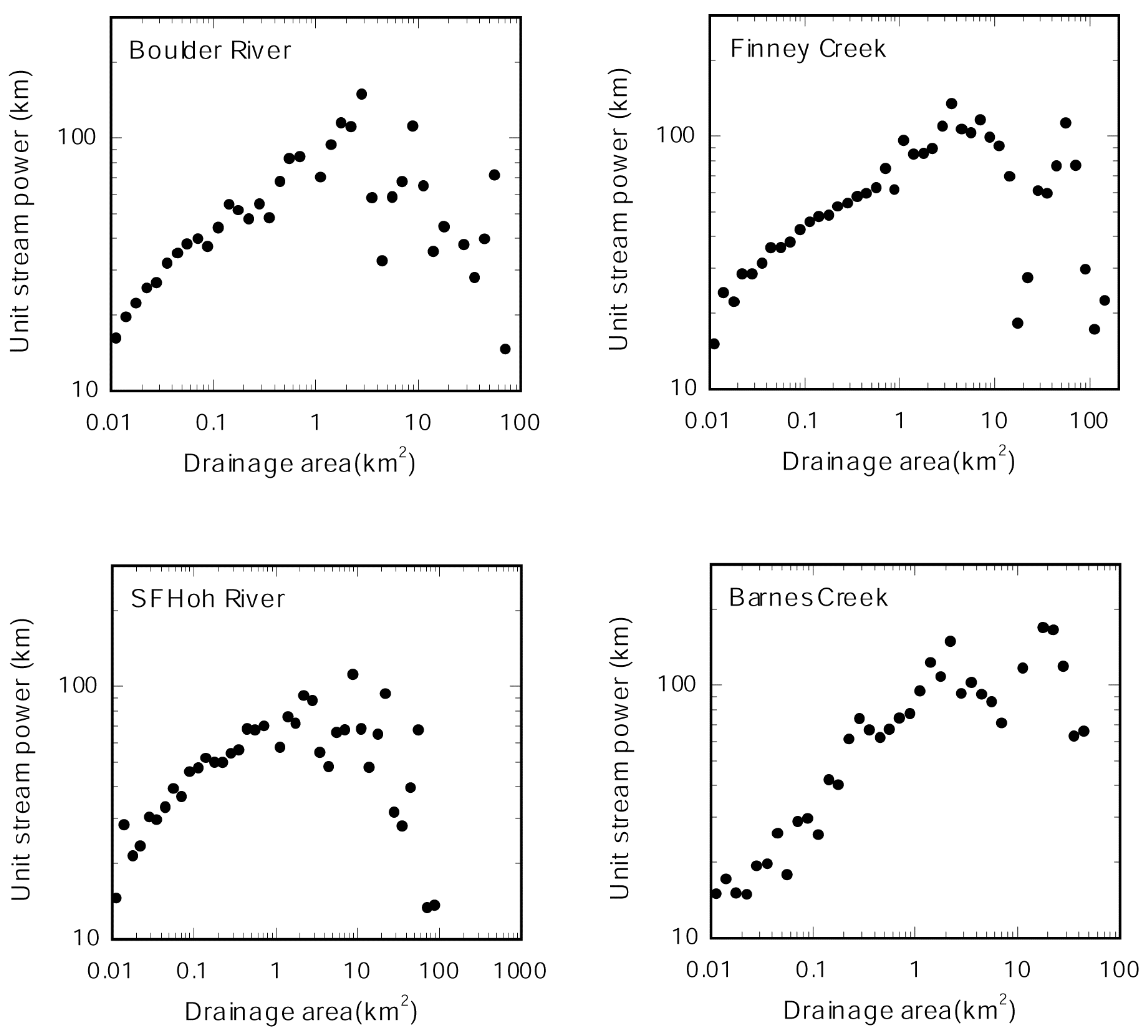


Figure 2.9: Log-bin-averaged index of unit stream power [$(1/c)A^{(1-b)}S$] derived from DEMs versus drainage area.

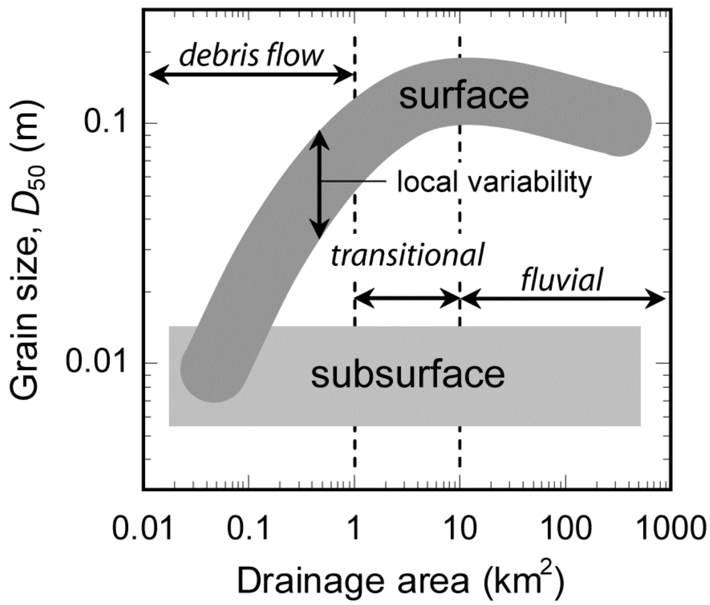


Figure 2.10: Schematic illustration showing relations between process domains and systematic trends in surface and subsurface grain sizes. Results of sediment sampling reveal no systematic downstream trend in subsurface D_{50}. Overlap of surface and subsurface populations occurs in low-order channels 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.

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CHAPTER 3: The influence of coarse lag formation on the mechanics of sediment pulse dispersion in a mountain stream

Summary

Mountain channels closely coupled to landslide-prone hillslopes often exhibit bed surface grain sizes coarser than transportable by annual high flows. Coarse particles within poorly sorted sediment delivered to channels by mass-wasting processes may not be readily transported as bedload and can consequently form lag deposits that influence the morphology, hydraulics, and sediment storage within mountain channel networks. A tracer study and comparison of supply and bed grain-size distributions from a valley-spanning landslide in the North Cascades of Washington state were used to derive relations between shear stress and the probability of particle entrainment and erosion from the sediment pulse. Rapid bed-surface armoring formed a relatively immobile lag deposit within two years. Covering of 20% of the bed by lag boulders with <5% probability of entrainment was sufficient to retard vertical incision and force considerable channel widening during a flood with an 8- to 152-year recurrence discharge on locally gauged streams. Results of this study imply that numerical models of sediment-pulse evolution that do not explicitly incorporate the influence of lag formation may substantially overestimate long-term dispersion rates. The grain-size distribution and lithology of a sediment input relative to the flow competence of the receiving channel are important factors influencing the rates and mechanisms of sediment pulse dispersion and the sediment capacitance provided by coarse-grained sediment pulses in mountain drainage basins.

Introduction

Sediment supply from mass wasting dominates the storage and routing of sediment in confined, steep-gradient mountain channels (Dietrich and Dunne, 1978; Caine and Swanson, 1989; Benda, 1990). Episodic sediment inputs characteristic of mountain streams can create pulses that translate downstream as a coherent wave or disperse in place. General controls on sediment-wave behavior are important for understanding the geomorphology and ecology of mountain streams. Recent theoretical and experimental studies suggest that sediment pulses in rivers with high subcritical flood Froude numbers will decay in place by dispersion (Cui et al., 2003a, 2003b), whereas downstream translation is favored in sand-bedded channels with relatively low Froude numbers and when the introduced sediment is finer than the ambient bed material (Lisle et al., 1997; Wohl and Cenderelli, 2000; Cui et al., 2003a, 2003b). Sutherland et al. (2002) applied the numerical model developed by Cui et al. (2003a, 2003b) to the degradation of a landslide that entered the Navarro River in northern California and found that the model results agreed well with field measurements documenting the dispersive evolution of the river bed to near its original profile within four years. However, as Lisle and Church (2002) note, the bedload transport function in the model does not account for bed forms or cellular particle structures that might form from the reworking of coarse-grained sediment. This chapter investigates the evolution of a coarse-grained sediment pulse resulting from an instantaneous input of poorly sorted sediment containing a significant fraction of coarse particles too large to be transported as bedload by typical high flows. Although coarse-grained sediment pulses triggered by landslides are common in tectonically active regions, and their morphologic effects have received considerable attention (e.g., Costa and Schuster, 1991; Schuster et al., 1992; Pringle et al., 1998; Korup, 2004, 2005; Ouimet and Whipple, 2004), the rates and mechanisms by which mountain channels assimilate coarse-grained sediment pulses are poorly understood.

The morphology of alluvial channels in mountain drainage basins varies systematically with gradient to produce a stable roughness configuration that balances transport capacity with sediment supply (Montgomery and Buffington, 1997). Adjustments to bed-surface texture can occur across a broad range in channel gradient (Kinerson, 1990; Lisle, 1995; Brummer and Montgomery, 2003). However, variations in form roughness and slope arising from lateral migration and adjustments in sinuosity are the primary regulators of sediment transport capacity in most low-gradient floodplain channels (Schumm, 1963; Leopold et al., 1964). In contrast, valley confinement in steep-gradient channels (slopes greater than about 0.03) favors the adjustments of bed-surface texture through armoring and the organization of the coarsest particles into characteristic, energy-dissipating structures. For channel gradients ranging from about 0.03 – 0.10, the coarsest particles are commonly organized into a sequence of channel-spanning steps and pools that promote bed stabilization through the maximization of resistance to flow (Whittaker and Jaeggi, 1982; Abrahams et al., 1995). Step spacing has also been found to correspond with antidunes and the theoretical wavelength of standing waves, under which large clasts come to rest during transient conditions of critical flow (Whittaker and Jaeggi, 1982; Grant, 1997; Chartrand and Whiting, 2000). However, large particles exhumed during the fluvial reworking of coarse-grained sediment pulses and exhibiting mobility thresholds that exceed typical high flows can dominate the bed-surface texture of confined channels. For example, Zimmerman and Church (2001) found support for neither the flow-resistance nor the antidune hypothesis for step-pool formation. They attributed the highly variable step wavelength observed in their study reaches to the semi-random placement of stable boulders supplied by mass wasting. The authors found that steps destroyed by floods reformed around stationary keystones. Likewise, in their laboratory flume, Curran and Wilcock (2005) observed a random spacing of steps and found no support for grain deposition beneath water-surface waves. Montgomery and Buffington (1997) concluded that the disorganized arrangement of coarse particles in some cascade-type channels represents stable lag deposits forced by mass wasting. It follows that the size distribution of sediment supplied to mountain channels by mass-wasting processes should influence bed stability and the configuration of energy dissipating structures and thereby potentially moderate the dispersion of coarse grained sediment pulses.

The self-stabilizing tendency of alluvial channels imposes a threshold for bed mobilization that varies with position in the channel network. Bed stabilization by surface armoring of poorly-sorted sediment is well documented in low-gradient channels under conditions of excess sediment transport capacity (Dietrich et al., 1989; Church et al., 1998; Buffington and Montgomery, 1999; Lisle and Church, 2002). Decades of work on gravel-bed channels has established the generality of a threshold for armor mobility corresponding to discharges at or below bankfull flow, which typically occurs roughly every 1.5 – 2 years based on flood-frequency analyses (Wolman and Miller, 1960; Andrews, 1980; Whiting et al., 1999). Higher up in the channel network, studies of step-pool formation following channel disturbance document rapid bed stabilization by typical flows and a progressive increase in the threshold for step reorganization corresponding to flows with a recurrence interval of about 20-50 years (Grant et al., 1990; Lenzi, 2001; Madej, 2001; Kasai et al., 2004; Chin and Wohl, 2005). The disorganized arrangement of lag boulders in many cascade channels suggests a stable morphology between bed-resetting debris flows, which have recurrence intervals of 10^2 – 10^3 years depending on the relative stream order (Benda, 1990). The overlap in channel gradient for the occurrence of cascade and step-pool morphologies may reflect the variability in supply caliber with respect to the largest particle that can be transported by typical high flows (i.e., the flow competence of the stream). Hence, lag deposits represent the stable end member of the continuum of mobility thresholds observed in many alluvial channels.

Although the high threshold for bed mobility in steep mountain channels is widely recognized, the mechanisms of armor formation in steep channels remain poorly understood. Previous work in debris flow-dominated channels shows a stream power dependence on armoring even for steep (lag-prone) channels (Brummer and Montgomery, 2003). Lag development in steep-gradient channels implies that armoring may also be forced by the selective transport of sediment supplied by mass wasting. Lisle and Church (2002) examined sediment transport-storage relations using flume experiments, numerical simulations, and field examples of degrading channels. They found a rapid decline in transport rate and depth of incision in response to armoring and surface structure development. The effect was greatest in poorly sorted material. Selective transport should also be more pronounced and lead to lag formation where the coarse tail of the supply distribution exceeds the competence of typical high flows. Lag development that shields coarse-grained sediment pulses from incision can form sediment storage reservoirs in headwater channels and moderate the dispersion of sediment inputs within mountain channel networks.

Lag formation in steep-gradient channels may have particular importance for the understanding of how channels process coarse-grained sediment pulses, as field studies to date have focused on sediment pulses derived from source rocks that supply particle distributions readily transported by typical flows (Gilbert, 1917; Roberts and Church, 1987; Perkins, 1989; Madej and Ozaki, 1996; Wohl and Cenderelli, 2000, Sutherland et al., 2002; Cui et al., 2003a, 2003b). Because a boulder lag may inhibit the erosion of the mass-wasting deposit from which it evolves, an assessment of the mobility of large particles is necessary to understand the dispersion rate and residence time of coarse-grained sediment pulses in mountain channel networks. This investigation uses the case study of a coarse-grained bedrock landslide and valley-spanning debris dam in the North Cascades of Washington to investigate lag formation and attendant effects of sediment capacitance in a mountain stream.

Study area and landslide description

A massive landslide in the Squire Creek basin on 25 February 2002 created a sediment pulse that temporarily blocked the creek. The study area is located 3.5 km southwest of the town of Darrington in the North Cascades of Washington (Figure 3.1). Elevations within the basin range from about 130 m at the confluence with the North Fork Stillaguamish River to 2090 m on Three Fingers Peak. Most of the basin is underlain by granitic rocks of the Squire Creek stock. Fault-bounded, ultramafic rocks of the Darrington-Devils Mountain Fault Zone (Tabor et al., 2002) form the ridge north of Jumbo Mountain, where the landslide originated. Recessional outwash deposits form terraces in the lower reach of Squire Creek and within the study area.

The basin receives an average of 2000-3400 mm of annual precipitation, which falls primarily between October and March as rain (Daly and Taylor, 1998). Snow accumulation occurs throughout the basin but typically melts by late summer even at the highest elevations. Squire Creek has been gauged approximately 8.8 km downstream of the landslide at the state route 520 bridge by the Washington State Department of Ecology since May 2004. Stream flow is dominated by storm runoff and responds within hours to precipitation. The majority of the basin is vegetated with old-growth, conifer forest, although northern portions of the valley bottom and the base of eastern slopes were logged as recently as the 1980s. These areas are densely vegetated with a second-growth, mixed deciduous and conifer forest.

At the landslide deposit, Squire Creek drains a basin area of 26.0 km^2. Adjacent reaches of Squire Creek unaffected by the landslide have a mean gradient of 0.08 and average bankfull width of 23 m. The study reach is a step-pool channel (Montgomery and Buffington, 1997) containing large boulders, some with diameters exceeding the bank height by several meters. Except for a short channel reach approximately 700 m below the study area, bedrock is not exposed in the valley bottom, and the channel is formed entirely within sediment derived from glacial outwash, mass wasting, and recent fluvial processes. Aside from the recent landslide, extensive field reconnaissance and examination of aerial photographs found no evidence of significant historical (i.e., < 100 years) slope failures in the basin.

The Squire Creek landslide occurred after a series of moist Pacific frontal systems delivered over 200 mm of precipitation in a week (Figure 3.2), as recorded at the Darrington Ranger Station (elev. 167 m, mean annual precipitation 2020 mm). The failure began as a deep-seated rock avalanche from the ridge north of Jumbo Mountain and evolved into two debris-flow tracks as it entrained soil and woody debris during the 1500-m descent to the valley bottom (Figure 3.3). An estimated 125,000 m^3 of rock, soil, and wood were deposited along a 400-m reach of the valley floor (elev. 360 m) and temporarily dammed Squire Creek. Video recorded from a U.S. Forest Service helicopter approximately two hours after the event shows formation of a backwater lake and a multi-thread channel traversing the landslide dam (Figure 3.3).

Methods

Post-landslide modification of bed-surface grain size was measured using grid-by-number particle counts (Wolman, 1954). Particle counts were conducted at nine locations on the undisturbed surface of the landslide deposit and combined into a composite grain-size distribution (942 point counts). Reach-average particle counts (>100 counts each) were conducted on the channel bed through the landslide deposit at 11 locations (Figure 3.4). Particle counts were conducted across the entire high-flow channel width and along a reach length roughly equivalent to one channel width. Fluvial incision through the landslide deposit was documented by repeated field surveys of the channel profile and cross sections. Field work was conducted during summer 2003, 2004, and 2005, whereas photographs compiled from field reconnaissance by others document early channel evolution before 2003. A longitudinal profile of the channel thalweg was constructed in 2003 by surveying every step and pool feature along ~1 km of the creek. The survey was conducted with a tape, hand level, and stadia rod. In addition, nine cross sections were established and surveyed in summer 2003 between rebar end markers spaced along the creek segment impacted by the landslide. The profile and cross sections were resurveyed in summer 2004 and 2005. Both the perimeter and surface topography of the landslide deposit were surveyed in greater detail in 2004 using a total station theodolite. The original landslide surface above the channel was inferred from photographs taken the day of the landslide and by projecting the surface topography across the top of the channel. The initial volume of the landslide deposit within the valley bottom was calculated from the surface area of the landslide and the average depth of the deposit estimated from the height of exposed channel banks and topographic relief along the landslide perimeter. The volume of material eroded between surveys was calculated by interpolating between cross sections.

Lag formation was documented by comparing transport capacity with the grain-size distribution of sediment supplied by the landslide. Empirical relationships for the maximum mobile grain size transported by winter flows (D_{max}) was derived using two independent approaches: (1) a tracer study of angular clasts derived from the landslide dam and (2) a probabilistic model of sediment entrainment derived from the difference between grain size distributions of the sediment input (composite of nine pebble counts on the landslide surface) and the creek bed (the 11 in-channel pebble counts of the lag deposit).

Hydrologic data for Squire Creek were compiled from 24-hour historical precipitation data recorded at the Darrington Ranger Station. Gaps in the Darrington record were filled using daily precipitation from the Concrete Fish Station, located 33 km north of Darrington. The Concrete data were adjusted using a linear regression of historical daily precipitation from both sites. Because flow data were not available prior to May 2004, the 24-hour precipitation from the nearest recording station in Darrington was used as a proxy for relative discharge.

Tracer study

The relative stability of bed-surface material supplied by the landslide was evaluated by tracking angular particles downstream from the landslide for up to 2 km. Particles presumed to have been supplied from the landslide were readily identified by an unweathered appearance, angular shape, and metagabbroic lithology unique to the landslide source area. Stationary particles abrading in place were identified by fluted, faceted, and potholed surfaces sculpted by impacts from fine sediment suspended in flow separation eddies (Whipple et al., 2000; Springer et al., 2005). In contrast, angular clasts transported downstream from the landslide deposit were considered mobile. Measurements at each sample location included the intermediate diameters of all angular metagabbro clasts, as well as clasts on the bed surface that exhibited evidence of long-term stability and in-place abrasion (i.e., fluting and potholing); all angular and fluted clasts larger than ~20 cm were sampled. Local channel slope (S) was calculated from surveyed elevations of the channel bed at upstream and downstream ends of each sample reach. The channel depth (h) corresponding to seasonal high flow was also measured. Because most of the sample reaches lacked a true floodplain, high-water indicators such as vegetation patterns, stain lines, and the height of snagged debris were used to approximate the flow depth corresponding to seasonally high flow in floodplain channels. Total reach-average shear stress ($\tau=\rho ghS$) was used as a proxy for tractive forces acting on the bed, where ρ is the density of water and g is gravitational acceleration. Form drag from steps and large boulders was not quantified but could reasonably account for a large fraction of the total shear stress estimated from the depth-slope product. A discriminant-function analysis was used to derive the best-fit relation between groups of stable and mobile sediment sizes as a function of total shear stress.

Probabilistic model

A modification of the sediment-trend model first described by McLaren and Bowles (1985) was used to derive empirical relationships between total shear stress, grain size, and the probability of sediment entrainment and transport from the landslide deposit (Figure 3.5). The McLaren and Bowles (M-B) model relates the grain-size distribution of the source [$G(s)$] to the distribution of the removed or eroded sediment [$R(s)$] by introducing a transfer function [$t(s)$]:

$$t(s_i) = \frac{R(s_i)}{G(s_i)} \qquad \text{Eqn. 3.1,}$$

where $G(s_i)$ and $R(s_i)$ are the proportion of sediment in the i^{th} interval of the grain-size distribution (s), and $t(s_i)$ is the transfer coefficient for sediment in the i^{th} grain-size interval. The transfer coefficient is also the probability that a particle in the i^{th} grain-size interval will be removed from $G(s_i)$. For a poorly sorted $G(s)$ containing clasts exceeding the flow competence, the M-B model predicts that the transported sediment will become finer than the source, and the remaining sediment distribution will coarsen and form a lag deposit [$F(s)$]. A general solution for the transfer function can be derived from field measurements of the source and bed-surface distributions without the need to characterize the distribution of the eroded sediment, where the sampling of coarse bedload is impractical due to the size of large boulders transported by infrequent, high-intensity storms.

In a closed system, conservation of mass requires that the grain-size distribution of removed sediment equals the difference between distributions of the source and lag such that

$$R(s) = G(s) - F(s) \qquad \text{Eqn. 3.2.}$$

Combining Equations 3.1 and 3.2 yields a solution for the transfer function in terms of the source and lag distributions, which could be readily measured:

$$t(s) = 1 - \frac{F(s)}{G(s)} \qquad \text{Eqn. 3.3.}$$

The M-B model has been used primarily to infer transport direction of fine sediments in coastal environments based on the relationships between grain-size distributions of sequential deposits (McLaren and Bowles, 1985). Sutherland (1987) conceptualized a similar model for gravel-bed channels to describe the changes that occur in sediment distributions as bed-surface armoring progresses. Sutherland (1987) combined eroded and armor distributions to posit a probability of removal for each particle size, which is equivalent to the transfer function in the M-B model. The comparison of residual (lag) deposits to their source size distribution represents the first application of such a model to the quantitative evaluation of bed-surface armoring from lag development in mountain channels.

Grain-size distributions were developed from normalized particle counts of the landslide deposit [$g(s)$] and the n=11 particle counts of the bed surface [$f_n(s)$]. Particle counts for the probabilistic model were conducted in summer 2003. Grain size was characterized using the logarithmic ψ scale, where ψ and the traditional ϕ scale are related to the median grain diameter (D) in mm by

$$\psi = -\phi = \log_2(D) \qquad \text{Eqn. 3.4.}$$

The median diameter of grains between 2ψ and 12ψ were measured with a graduated ruler or stadia rod to the nearest 0.5ψ interval.

Because only particles coarser than 2ψ (4 mm) were sampled, the volume fraction of sediment finer than 2ψ and the volume of the voids in the landslide deposit had to be accounted for. The combined volume occupied by voids and sediment $<2\psi$ within the landslide (p) was calculated from the weight of coarse sediment sampled from a known volume of landslide material. The *in-situ* volume was measured from the dimensions of an excavation on the surface of the landslide. Sediment $<2\psi$ was sieved in the field and discarded. The remaining sediment, which was coarser than 2ψ, was air dried on tarps and weighed in the field. The combined volume occupied by voids and sediment $<2\psi$ is given by the constant p calculated from

$$p = 1 - \frac{M_c}{\rho_s V_t} \qquad \text{Eqn. 3.5,}$$

where M_c is the mass of sediment $>2\psi$ (167 kg), ρ_s is the assumed sediment density (2700 kg/m^3), and V_t is the total volume of the excavation (0.095 m^3). Solving Equation 3.5 for p yields a volume fraction of the landslide occupied by both voids and sediment $<2\psi$ of 0.35.

Because the composition of the lag deposit should vary with incision depth, it is necessary to know the number of grains, by size interval, within the initial source volume above a unit area of the bed from which each lag deposit was derived. Likewise, it is also desirable to know the number of grains per unit area of the bed surface and the total number of grains in each size interval per unit volume of the supply. this quantity was calculated by converting the grid-by-number particle count of the source distribution to synthetic volume-by-number distributions [$G_n(s)$] for each of the 11 sample locations:

$$G_n(s) = g(s)k_1(1-p)h_n\left(1/D^3\right) \qquad \text{Eqn. 3.6,}$$

where h_n is the incision depth at the n^{th} sample location, D_i is the grain diameter in mm of each size interval, and k_1 is a scaling coefficient. that normalizes $G_n(s)$ by the between the grid-by-number and volume-by-number distributions such that

$$k_1 = \frac{\sum_{i=1}^{21} G_n(s_i)D_i^3}{(1-p)h_n} \qquad \text{Eqn. 3.7.}$$

The coefficient k_1 normalizes $G_n(s)$ by the unit volume of eroded sediment (denominator in Equation 3.7). Equations 3.6 and 3.7 were solved simultaneously to obtain a value of 1.91 for k_1 and solution for $G_n(s)$.

The total number of grains in each size class per unit area of the channel bed was calculated by converting the grid-by-number particle counts of the bed surface to synthetic area-by-number grain-size distributions [$F_n(s)$] for each sample location:

$$F_n(s) = f_n(s)k_2\left(1/D^2\right) \qquad \text{Eqn. 3.8,}$$

where k_2 is a second scaling coefficient (1.27), similar to k_1, which normalizes $F_n(s)$ by the total area of the bed surface covered by all of the grains in the area-by-number distribution:

$$k_2 = \sum_{i=1}^{21} F_n(s_i)D^2 \qquad \text{Eqn. 3.9.}$$

In contrast to the volume-by-number conversion, voids and fine sediment need not be accounted for in the surface calculations because an area-by-number sampling surface will always be intersected by the projection of a grain within the surface layer.

Transfer coefficients were derived for the 11 sample locations by combining Equation 3.3 with the synthetic grain-size distributions calculated for $F_n(s)$ and $G_n(s)$ in Equations 3.6 and 3.8, respectively:

$$t_n(s) = 1 - F_n(s) / G_n(s) \qquad \text{Eqn. 3.10.}$$

The coarse tails of the transfer functions were truncated at the grain-size interval in which $F_n(s) > G_n(s)$ resulted in a negative transfer coefficient. Negative transfer coefficients occur when the number of grains in a unit area of the bed within a particular size interval exceeds the number of grains in the original source volume above that unit area. Although negative transfer coefficients are not possible in a closed system, they occurred in the coarse tails of the distribution because of the small sample population within the coarsest grain-size intervals.

Relations between the form of each transfer function and corresponding shear stress were used to develop an empirical model for the probability of sediment entrainment from the landslide deposit as a function of grain size and total shear stress. An analytical solution derived from results of the field sampling (described later) was chosen that minimized residuals between the model and field-based transfer coefficients.

Results

The longitudinal profile of Squire Creek exhibits a distinct convexity along the channel reach affected by the landslide (Figure 3.6). Exposure of a buried forest floor and soil on a relic landslide diamicton beneath the recent landslide deposit indicates that the profile convexity is the result of at least two landslide events. The profile convexity increases the channel gradient at the downstream inflection and reverses the channel gradient at the upstream inflection. Cross sections constructed from field surveys show that the maximum vertical incision occurred directly above the downstream inflection (cross section 2, Figure 3.7), where the initial channel gradient through the landslide was steepest. During field surveys, the bed surface of nearly the entire study reach was observed to be organized into a coherent step-pool morphology, with most steps spanning the entire channel width (Figure 3.8). The largest step was over 4 m high and formed a deep plunge pool above cross section 2 (at station 365, where stationing is in m and increases upstream) and near the toe of the landslide deposit. The deposition of intermediate-size material eroded from the landslide formed a boulder bar immediately downstream of the landslide toe between stations 200 and 320. Additional material eroded from the landslide during winter 2003-04 aggraded the channel farther downstream and forced a channel avulsion at station 230.

A reversal in profile gradient at the upstream inflection formed a backwater lake, which began aggrading at the lake inlet with sand and gravel supplied from upstream. Field observations indicate that at moderate discharge, a portion of the flow overtops the shallow banks at the downstream end of the lake (station 720) and flows through the forest and around the western perimeter of the landslide before rejoining the mainstem at station 600. The bifurcation of flow and shallow channel depth at the lake outlet apparently limits vertical incision and the potential for reworking the bed along this upper reach.

Channel incision and lag formation

A time series of cumulative channel incision and sediment export at the landslide was constructed using field observations, channel surveys, and hydrologic data (Figure 3.9). Channel incision and the formation of an incipient lag occurred rapidly within the first two months following the landslide. Photographs of the channel taken 25 April 2002 by the U.S. Forest Service (Ron Hausinger, personal communication, 2004) document approximately 4 – 6 m of vertical incision and considerable armoring of the channel bed relative to the undisturbed surface of the landslide. During this brief period, the maximum 24-hour precipitation at Darrington was 76 mm (Figure 3.2). Photographs taken 25 days later on 20 May 2002 (Ron Hausinger, personal communication, 2004) showed no discernable bed lowering based on key boulders in the same orientation as those in the 25 April photographs. Daily precipitation between the April and May 2002 photographs did not exceed the previous maximum precipitation. Field evidence of significant bed reorganization was apparent during the first site visit approximately one year later on 30 May 2003. The earlier 24-hour precipitation maximum of 76 mm was exceeded once (91 mm, 13 March 2003) between the 20 May 2002 photographs and the 30 May 2003 site visit.

The first detailed survey of the channel profile in summer 2003 measured up to 9 m of vertical incision relative to the estimated elevation of the original landslide surface (Table 3.1). Channel incision prior to the 2003 survey eroded most of the fine sediment from the channel and coarsened the bed relative to the surface of the landslide deposit. The grain-size distribution of sediment sampled from the bed of the channel, where it had incised into the landslide, coarsened with increasing shear stress relative to the original landslide particle-size distribution (Figure 3.10). The landslide deposit exhibited a bimodal distribution, which is posited to have developed during the mixing of rockfall material with entrained colluvium and glacial sediments. The landslide material also had a geometric sorting coefficient (σ_g) of 1.44, where $\sigma_g=(\psi_{84}-\psi_{16})/2$. Results of the 2003 sediment sampling indicate that the median bed-surface grain size increased with both the local shear stress and the local incision depth (Figure 3.11). Dimensionless shear stress (Shields number) calculated for the 11 samples ranged from 0.4 – 0.8. Comparison with a typical range of 0.03 – 0.06 for incipient motion (Buffington and Montgomery, 1997) indicates energy losses of up to about 84 – 95% during the winter 2002-03 peak flow. The degree of bed armoring (surface D_{50}/supply D_{50}) ranged from 2.2 – 10.1 (mean of 5.6). Based on the interpolation of cross-sectional areas, approximately 22,000 m^3 of sediment (9% of the estimated original landslide volume within the valley bottom and available for channel reworking) was eroded from the landslide during the latter half of winter 2001-02 and through winter 2002-03.

Runoff generated by several high-intensity storms between 16-21 October 2003 resulted in significant channel adjustment at the study site and caused local flooding and property damage throughout the region. Although Squire Creek was not gauged at the time, the recurrence interval of October 2003 floods on six nearby gauged rivers ranged from 8 – 152 years. The Darrington Ranger Station recorded 417 mm of precipitation during this six-day period. The maximum 24-hour precipitation from the prior rainy season was exceeded twice during this period (141 mm and 155 mm). A resurvey of the channel thalweg and cross sections following the intense October 2003 storms documented minor vertical change in the profile, but considerable bank erosion and channel widening. The maximum vertical change in local thalweg elevation measured among the nine cross sections between the 2003 and 2004 surveys was 1.9 m and only slightly greater than the vertical precision of the survey set by the D_{90} of the bed (~1 m). However, bank retreat eroded an additional 26,000 m^3 of sediment (11% of the original landslide volume) and increased the local channel width by up to 27 m. Bank migration into the adjacent forest downstream of cross section 4 (station 419) exhumed stored sediment from the prehistoric landslide deposit. Although the channel was not resurveyed until summer 2004 when flows receded, periodic site visits and photo-documentation showed no evidence of bed reorganization between the October 2003 storms and the 2004 survey. Likewise, comparison of the 2004 and 2005 surveys showed no significant (>1 m) change in cross sections or profile. The peak 24-hour precipitation recorded between the October 2003 storms and the summer 2004 survey was 80 mm, which is less than the 2002-03 maximum, but greater than the peak 24-hour precipitation recorded in April 2002. The maximum 24-hour precipitation at Darrington between the 2004 and 2005 surveys was 118 mm on 9 December 2004, which triggered the peak discharge for the season of 126 m^3/sec on 10 December 2004.

Tracer study

The intermediate diameter of fluted and potholed particles sampled downstream of the landslide in 2003 was significantly greater than the diameter of angular particles

(Figure 3.12). The best-fit relation that discriminates between groups of stable and mobile sediment sizes is positively correlated with total shear stress (r^2=0.69). The discriminant function derived from results of the tracer study is given by

$$D = 43\tau^{0.35} \qquad \text{Eqn. 3.11,}$$

where D is the particle diameter in mm that best discriminates between stable and mobile clasts and τ is the total shear stress in Pa, as defined earlier. Grain sizes from which Equation 3.11 was derived ranged from 350 – 760 mm over a range of total shear stress of 400 – 3670 Pa.

Probabilistic model

The transfer functions calculated from Equation 3.10 display the same general form for all 11 sample locations (Figure 3.13). The asymptotic behavior toward unity for the fine and intermediate grain-size intervals indicates that all of the sediment within these size intervals was eroded during channel incision. The decay in transfer coefficients over the coarsest four to five grain-size intervals indicates a decline in the proportion of coarse sediment removed from the deposit during incision and an increase in the proportion of coarse sediment remaining to form a stable bed surface. In most cases, the intervals comprising the coarse tails of the distributions, where transfer coefficients are noticeably < 1, show the greatest variability between transfer coefficients. The variability is attributed to the small sample sizes within the coarsest size intervals. The transfer functions also become increasingly skewed toward coarser size intervals with increasing shear stress.

The slope and intercept of the discriminant function derived from the tracer study (Equation 3.11) roughly correspond to the maximum mobile grain size (D_{max}) inferred from the field-based transfer coefficients, even though the size of some angular (*i.e.*, mobile) particles sampled in the tracer study fall above the discriminant line in Figure 3.12. This may be due to the relatively low form roughness of the ambient bed (from which tracer particles were collected) in comparison to bed conditions of the highly armored channel passing through the landslide deposit. Equation 3.11 was therefore used to derive a general solution for the transfer function by incorporating the power-law regression between total shear stress and grain size into the error function (erf):

$$t(s) = \mathrm{erf}\left[c \log_2\left(\frac{a\tau^b}{D_i}\right)\right] \qquad \text{Eqn. 3.12,}$$

where c is a shape factor that scales the grain-size range over which $t(s)$ decays from near unity to zero, and D_i is the i^{th} grain-size interval (mm). The error function is defined by

$$\mathrm{erf}(x) = \frac{2}{\sqrt{\pi}}\int_0^x e^{-u^2}\,du \qquad \text{Eqn. 3.13.}$$

The coefficient and exponent derived from the discriminant analysis (Equation 3.11) were substituted for a and b in Equation 3.12, which was then solved for c by minimizing the sum of the squares of the residuals for the ensemble of field data. In doing so, the numerator in the log expression is explicitly set equal to D_{max}. The general solution for the transfer function is then given by

$$t(s) = \mathrm{erf}\left[0.61\ \log_2\left(\frac{43\tau^{0.35}}{D_i}\right)\right] \qquad \text{Eqn. 3.14.}$$

Equation 3.14 provides a satisfactory fit (composite r^2=0.84) to the ensemble of transfer functions (Figure 3.13, Table 3.1). Solving for D_i of the bed material in Equation 3.14 when $t(s)$ equals 0.95 (when 5% of D_i particles are immobile) indicates that transfer coefficients begin diverging from unity at a threshold given by

$$D_i = 9.0\tau^{0.35} \qquad \text{Eqn. 3.15.}$$

Because the transfer coefficients are undefined for negative values, $t(s)$ was set equal to zero for the condition

$$D_i > 43\tau^{0.35} \qquad \text{Eqn. 3.16.}$$

Equations 3.15 and 3.16 define the grain-size range over which the transfer function decays from near unity to zero assuming a homogeneous supply distribution. However, mechanical sorting by the landslide and topographic focusing of the landslide trajectory during transport to the valley floor may have introduced spatial variability in the supply distribution. In such cases, the general solution may diverge from the field-based transfer function. Local variability in $G_n(s)$ and $t_n(s)$ may also occur when the incision depth is less than the size of the largest particles in the supply distribution; examples of such variability are presented in Figure 3.13 for $t_1(s)$, $t_2(s)$, and $t_3(s)$, where the incision depths were limited to 1 – 2 m.

Bed shielding and step-pool formation

The fraction of the bed surface covered by relatively immobile particles in summer 2003 was evaluated using the general solution for the transfer function (Equation 3.12). Lag boulders (*D* with <5% probability of entrainment prior to the October 2003 storms) covered 8 – 30% (average of 20%) of the bed surface through the most deeply incised portion of the landslide (Table 3.1). The degree of armoring was relatively high and as much as five times greater than the armoring measured for the finer-grained sediment pulse investigated by Sutherland et al. (2002). A comparison of the median step-forming grain size with the grain size corresponding to a transfer coefficient of 0.05 indicates that all of the steps sampled were formed by lag boulders with a <5% probability of entrainment (Figure 3.14A). An additional comparison of the median step-forming grain size with the reach-average D_{90} of the bed surface indicates that most lag boulders fell within the upper 10 percentile of the bed surface grain size distribution (Figure 3.14B).

Minor vertical incision (locally <2 m) and significant channel widening of up to 27 m occurred during the storms of October 2003 (Figure 3.7). Alternating bank erosion during channel widening in October 2003 increased the thalweg sinuosity, which is illustrated in the cross sections by alternating bank erosion from the left bank (cross sections 1 – 4) to the right bank (cross sections 5 – 8) and then back to left bank (cross sections 8 – 9). The increase in sinuosity effectively reduced the channel gradient by increasing the thalweg length through the deposit (the 2004 profile was constructed along the high-flow alignment, which has a lower sinuosity).

Discussion

Higher local shear stress along the steepest reach of the new channel drove deeper incision of the landslide deposit over the negative feedback of armoring and coarsened the bed relative to the original particle size distribution. Vertical incision declined rapidly as the bed surface armored with a lag of relatively immobile particles. The decline in the value of transfer coefficients (or the probability of mobilization) at the coarse tail of the 11 grain-size distributions (Figure 3.13) indicates that lag formation was driven by the selective transport of fine material from the poorly sorted landslide deposit. The characteristic form of the transfer functions shows that all particle sizes were not mobilized equally relative to their individual fractions present within the sediment pulse. The findings of this study are consistent with those of Lisle (1995), who found that selective transport was most prevalent in steep, coarse-bedded channels, where annual bed scour did not regularly penetrate below the armor layer. In a similar study of fluvial recovery at Mount Pinatubo, Gran and Montgomery (2005) found that a seasonal decline in sediment supply during the dry season, combined with selective transport of fine material from the bed, drove vertical incision and bed armoring, which increased form drag and inhibited bed mobility. Results of this study suggest that selective transport may be widespread throughout mountain channel networks and dominate the reworking of non-fluvial sediment supplied to confined channels.

Lag formation also involved the organization of the coarsest particles into energy-dissipating steps that apparently limited scour during the October 2003 event and shielded the bed from further incision. The largest boulders provided stable nuclei for step formation. Results of this study show that a grain-size percentile coarser than D_{50} (such as D_{90}) may be the appropriate reference grain size in steep, confined mountain channels, where bed reorganization of step-pool structures is controlled by the mobility threshold of the largest step-forming clasts (Grant et al., 1990; Lenzi, 2001; Madej, 2001; Kasai et al., 2004; Chin and Wohl, 2005).

The chronology of channel incision reconstructed from photographs and channel surveys indicates that reworking of the bed by progressively larger flows during lag formation increased the threshold for further bed reorganization and forced the channel to widen during the high-magnitude flood of October 2003. The observed channel widening is interpreted to indicate a shift in the process of sediment-pulse dispersion from vertical incision to lateral erosion, a process change expected to follow bed stabilization by a relatively immobile lag. Channel widening and the increase in sinuosity caused by local variations in lateral migration further stabilized the bed. Channel widening decreased the flow depth and tractive forces acting on the bed by future flows of lesser magnitude than the October 2003 event. Likewise, the increase in sinuosity effectively reduced the channel gradient by increasing the channel length through the landslide.

Findings from this study that armor formation limits vertical incision and forces channel widening are consistent with previous studies of debris-fan erosion in the canyon rivers of the Colorado Plateau. Pizutto et al. (1999) and Webb et al. (1999) monitored the effects of the controlled flood of 1996 in the Grand Canyon and found that previous armoring of debris-fan surfaces by several small-magnitude floods inhibited significant reworking in 1996, whereas recently aggraded fans exhibited the greatest amount of reworking. Larsen et al. (2004) also reported limited reworking of previously armored debris fans by successive floods of comparable magnitude in a similar study along the Green River of Colorado and Utah. Previous studies within canyon rivers found that lateral erosion was the most significant mechanism of debris-fan dispersion (Kieffer, 1985; Hammack and Wohl, 1996; Pizutto et al., 1999; Webb et al., 1999; Larsen et al., 2004). Pizutto et al. (1999) reported that reworking of a coarse-grained debris fan during the 1996 controlled flood ended after four hours, when large boulders armored the unconsolidated banks. Similarly, armoring was also observed at the base of the banks of Squire Creek (Figure 3.8). In contrast, Hammack and Wohl (1996) suggested, on the basis of hydraulic modeling along a debris fan in the Yampa River, that erosion may continue in the portions already widened by previous, higher-magnitude flows. These authors showed that the critical shear stress for entrainment of the largest particles in the Yampa River debris fan is exceeded even during the 1-year recurrence flow and suggested that the duration of flows may be as important as flow magnitude for the dispersion of sediment pulses not dominated by the coarse tail of the grain-size distribution.

Results of the probabilistic model of particle entrainment provide insights on possible constraints for lag formation and channel widening. Results suggest that a poorly sorted sediment pulse must contain a sufficient concentration and caliber of coarse particles that are not readily mobilized during high-magnitude flows in order to form a relatively stable armor and the energy-dissipating structures capable of significantly retarding vertical incision into the deposit. This is consistent with other field studies showing the longevity of poorly sorted debris flows containing large boulders (Miller and Benda, 2000, Benda et al., 2003; Korup, 2005; Korup et al., 2006). The erosivity of the supply lithology may provide an additional control on the long-term dispersion rate of sediment pulses (Nolan and Marron, 1985; Taylor and Kite, 2006). Because armoring develops rapidly during low-magnitude flows, lag deposits can shield sediment pulses and sequester large volumes of sediment in mountain channel networks over time periods exceeding the recurrence of characteristic bed-reorganizing events.

Hence, lag deposits can attenuate the downstream delivery of sediment introduced by large-scale disturbances and should be common in confined mountain channels prone to coarse-grained sediment inputs from bedrock slopes that supply material resilient against mechanical weathering by fluvial erosion.

It is hypothesized that the relict lag boulders and profile convexities mapped elsewhere along Squire Creek and observed throughout similar mountain drainage basins in the Pacific Northwest record bed armoring and profile stabilization that may play a more significant role in retarding the decay of coarse-grained sediment pulses than previously recognized. In particular, the persistence of such deposits has implications for how coarse-grained debris dams function as large sediment "capacitors" in mountain channels by moderating sediment flux through the charging and discharging of sediment reservoirs. Results of the case study presented herein fits the two-phase model of degrading, gravel-bed channels proposed by Lisle and Church (2002) and provides a field-based measure of the time scale for phase I, which is 1 – 2 years. Results show that channel widening immediately after profile stabilization by a persistent armor layer may prolong the initial period of high sediment export. Consequently, numerical models of sediment-pulse evolution that do not explicitly incorporate the linkages between lag formation and lateral erosion may underestimate dispersion during channel widening by these end-member, coarse-grained sediment pulses but later overestimate dispersion once a lag forms. Hence, the grain-size distribution and lithology of a sediment input relative to the flow competence of the receiving channel appear to be critical factors influencing the rates and mechanisms of sediment pulse dispersion in mountain drainage basins.

Conclusions

Results from a four-year field study of a valley-spanning landslide document the effects of a persistent armor layer on the incision and reworking of a poorly sorted sediment pulse containing a significant fraction of particles not readily transported as bedload by high-magnitude flows. Rapid bed-surface armoring of the recovering channel was driven by selective transport during moderate-intensity flows in 2002 and 2003. Bed shielding by an incipient lag inhibited bed scour and forced the channel to widen during the October 2003 high-magnitude flood. Despite a decrease in vertical incision, widening in 2003 eroded more sediment than during all the previous reworking events combined. Lag boulders with <5% probability of entrainment and exceeding the reach-average D_{90} covered an average of 20% of the bed surface and comprised most of the step-forming clasts. The long residence time of exceptionally large particles and persistence of lag deposits have implications for how coarse-grained debris dams function as large sediment capacitors in mountain channels confined by steep valley slopes prone to landsliding.

Table 3.1: Results of the probabilistic analysis for summer 2003 field conditions.

Bed-Surface Sample	Profile Distance (m)	Incision Depth (m)	D_{50} (mm)	D_{90} (mm)	τ (Pa)	τ^{*} [a]	Model $D[t(s_i)=0.05]$ (mm)	Model D_{max} (mm)	r^2
f_1	653	1	108	321	665	0.37	405	426	0.89
f_2	635	2	77	301	762	0.59	425	447	0.80
f_3	607	1	131	331	1,049	0.48	475	500	0.70
f_4	423	4	190	538	1,276	0.40	509	535	0.96
f_5	490	4	97	352	1,281	0.79	510	536	0.76
f_6	442	4	120	350	1,329	0.66	517	543	0.66
f_7	469	5	150	502	1,380	0.55	524	550	0.91
f_8	408	6	251	912	2,275	0.54	624	656	0.95
f_9	363	7	374	1,225	2,659	0.43	766	693	0.88
f_{10}	386	9	452	1,060	4,076	0.54	659	806	0.89
f_{11}	337	5	373	1,564	4,722	0.76	807	849	0.85
								total:	0.83

[a] Shields number, $\tau^{*} = \tau / [g(\rho_s-\rho)D_{50}]$, where ρ_s and ρ are the densities of sediment (2700 kg/m^3) and water (1000 kg/m^3), respectively, and τ is the total shear stress not accounting for losses from hydraulic roughness. See text for explanation.

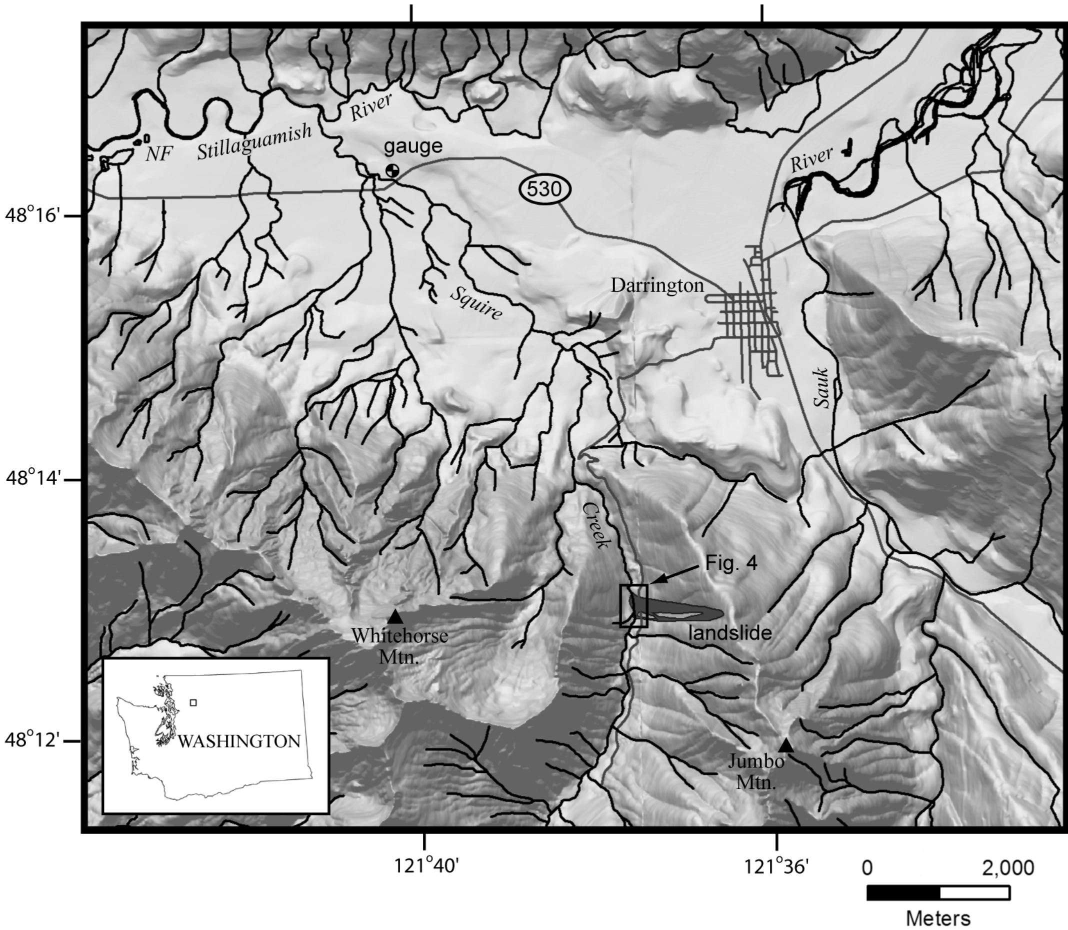


Figure 3.1: Shaded-relief map of the Squire Creek area showing the landslide, channel network, and gauging station.

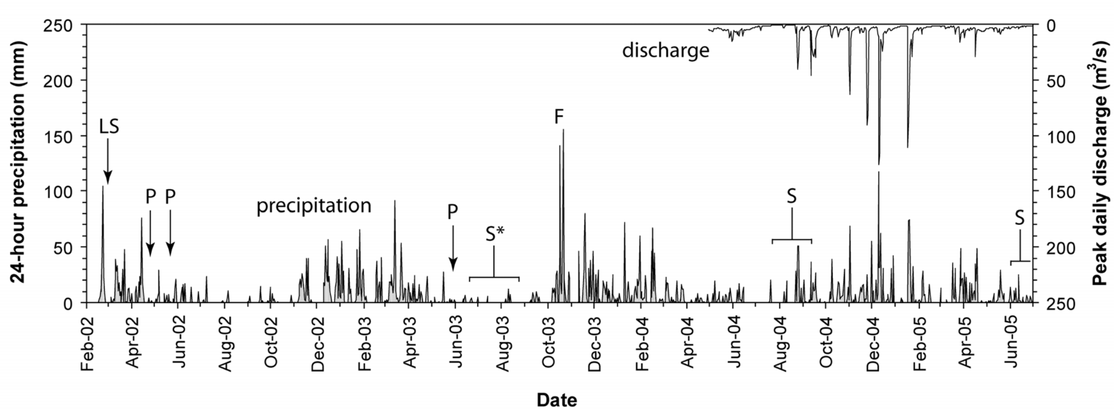


Figure 3.2: Daily precipitation at the Darrington Ranger Station and stream flow at the State Route 530 bridge. Stream flow obtained from the Washington Department of Ecology (Squire Creek at Squire Creek Park, gauge 05H070). Missing precipitation obtained from correlation with the Concrete PPL Fish Station (y=1.05x, r^2=0.48). Notation indicates dates of the landslide (LS), photo documentation (P), field surveys (S, asterisk denotes pebble counts), and the October 16-21, 2003 floods (F).

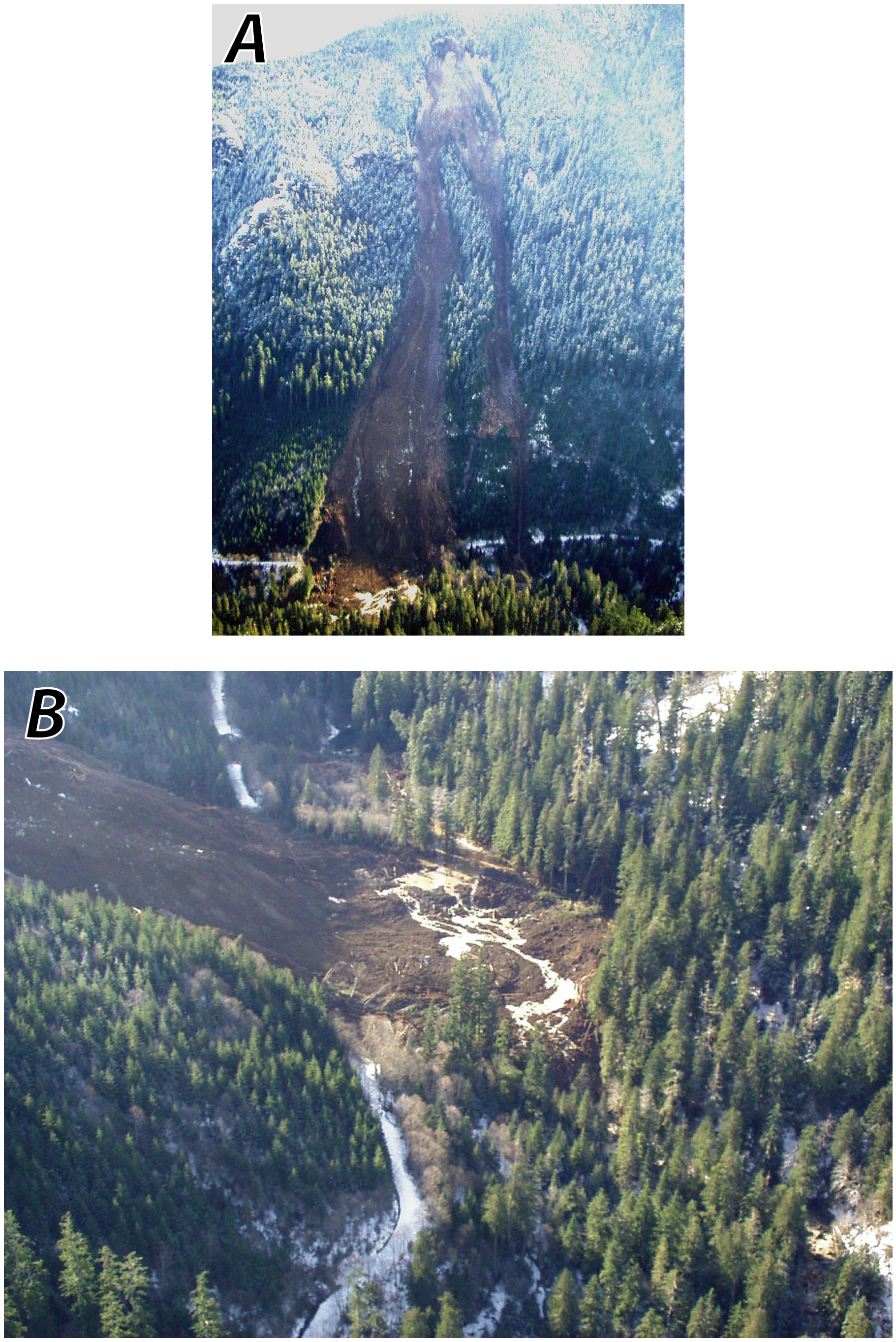


Figure 3.3: Oblique aerial photographs taken two hours after the landslide. A) The 1400-m runout. B) Water flowing over the landslide dam, view looking upstream (photo credit: Ron Hausinger, U.S. Forest Service).

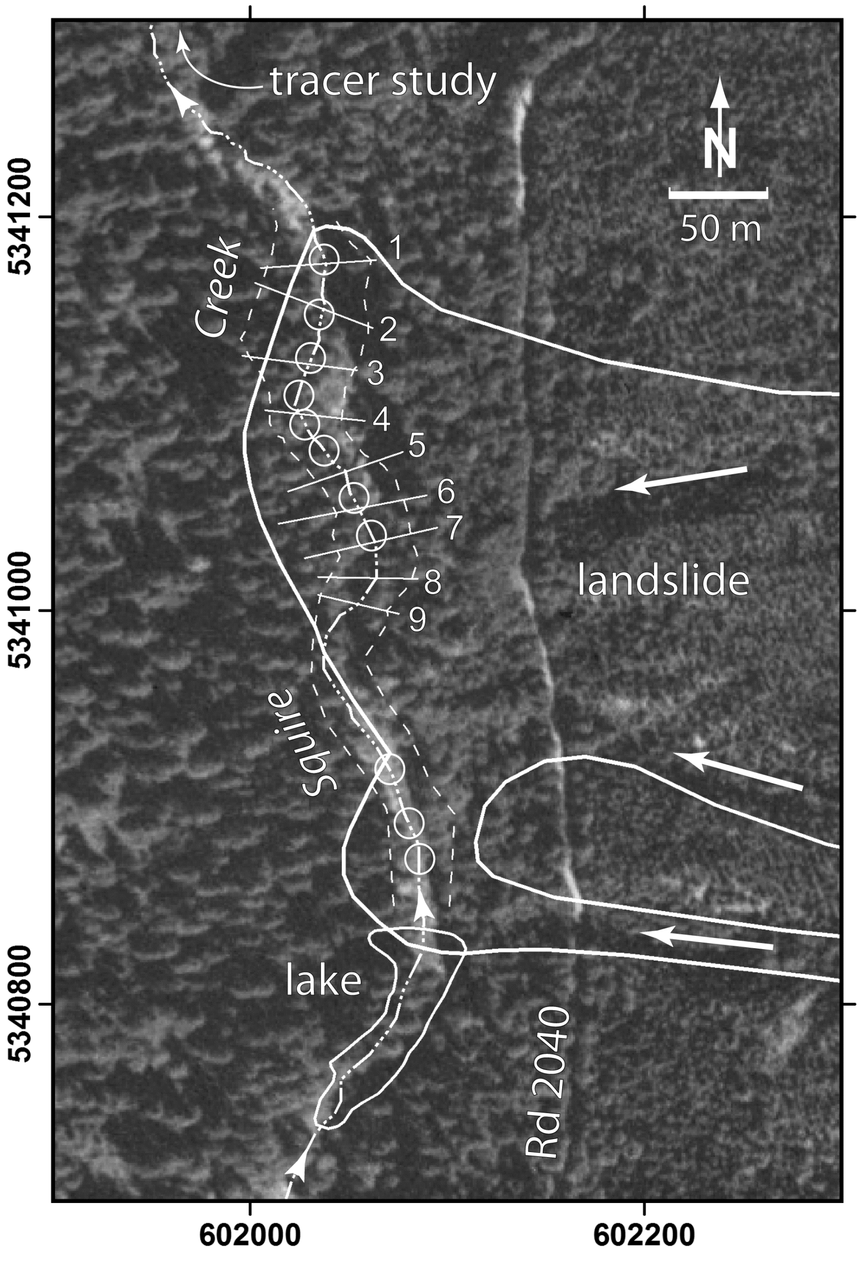


Figure 3.4: Overlay of the 1989 USGS orthophoto in the vicinity of the landslide toe. Features include the landslide perimeter and backwater lake (solid lines), thalweg, top of bank surveyed in 2004 (dashed lines), location of cross sections (1 – 9), the eleven bed-surface sampling locations for the M-B model (circles), and the tracer study. Arrows on landslide surface depict flow direction in bedrock ravines. Coordinate system is WA State Plane North, NAD83 (meters).

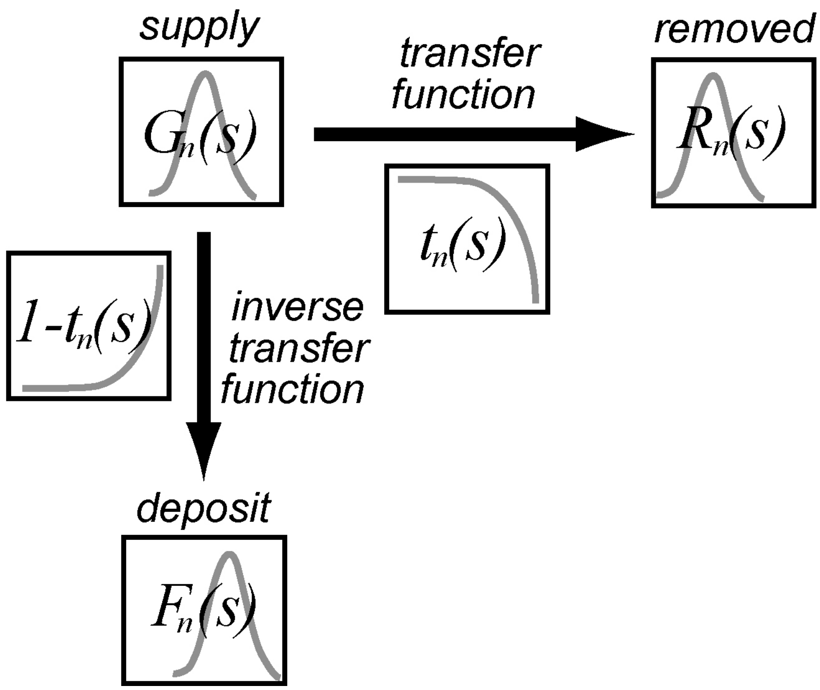


Figure 3.5: Schematic illustration of the McLaren and Bowles Model. Boxes show the generalized frequency distributions of the log-transformed grain size distribution for the supply [$G_n(s)$], removed sediment [$R_n(s)$], deposit [$F_n(s)$], and transfer function [($t_n(s)$].

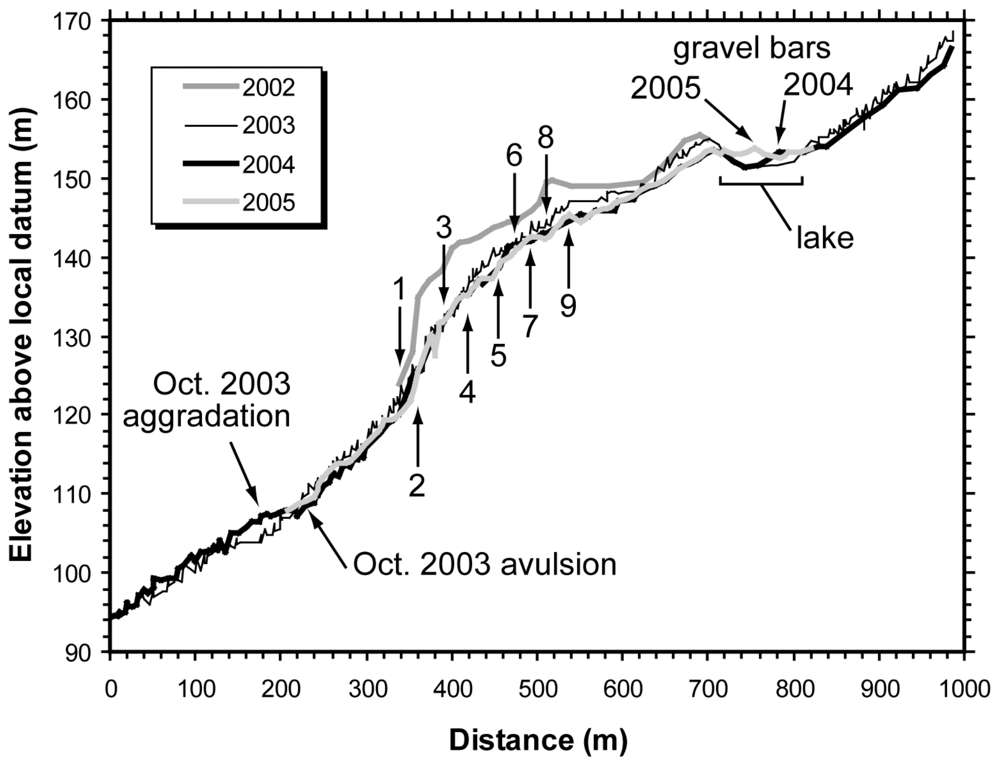


Figure 3.6: Channel profiles surveyed through the landslide deposit. Profile shows the location of cross sections (1 – 9), the landslide-dammed lake, the estimated pre-incision landslide surface in 2002, and the 2003, 2004, and 2005 surveys.

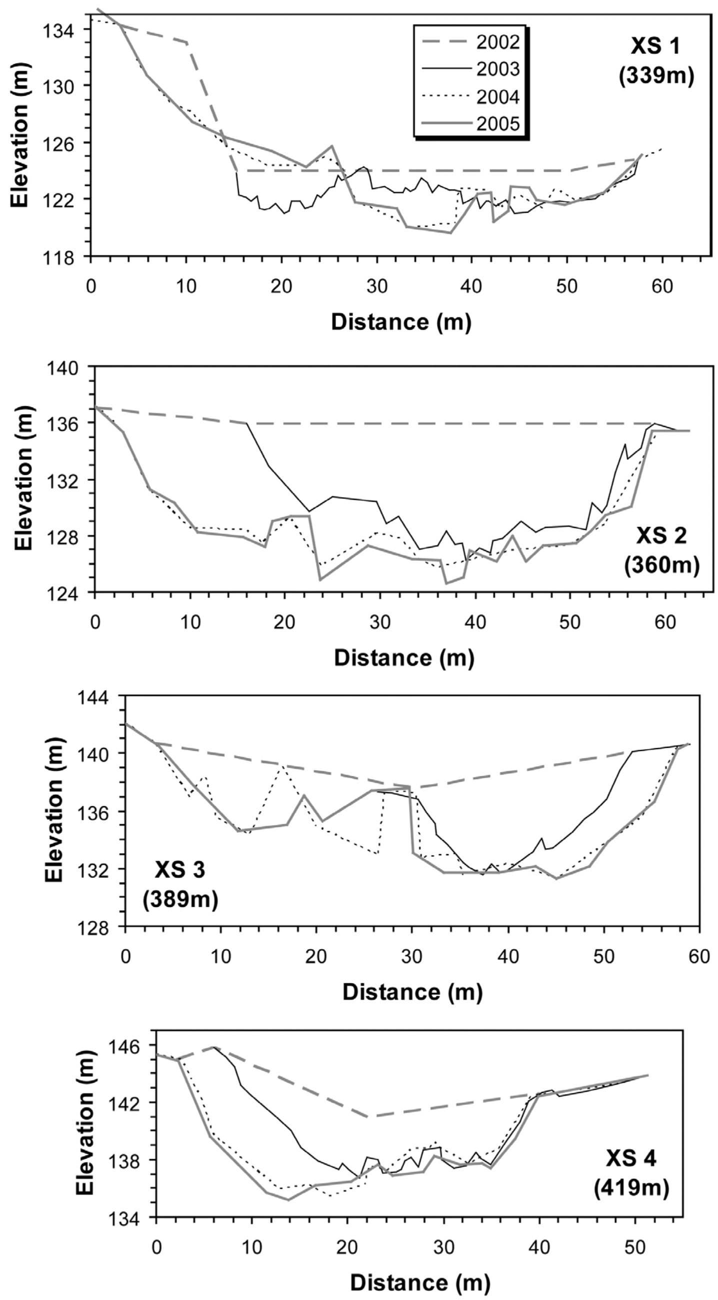


Figure 3.7A: Channel cross sections surveyed in 2003, 2004, and 2005.

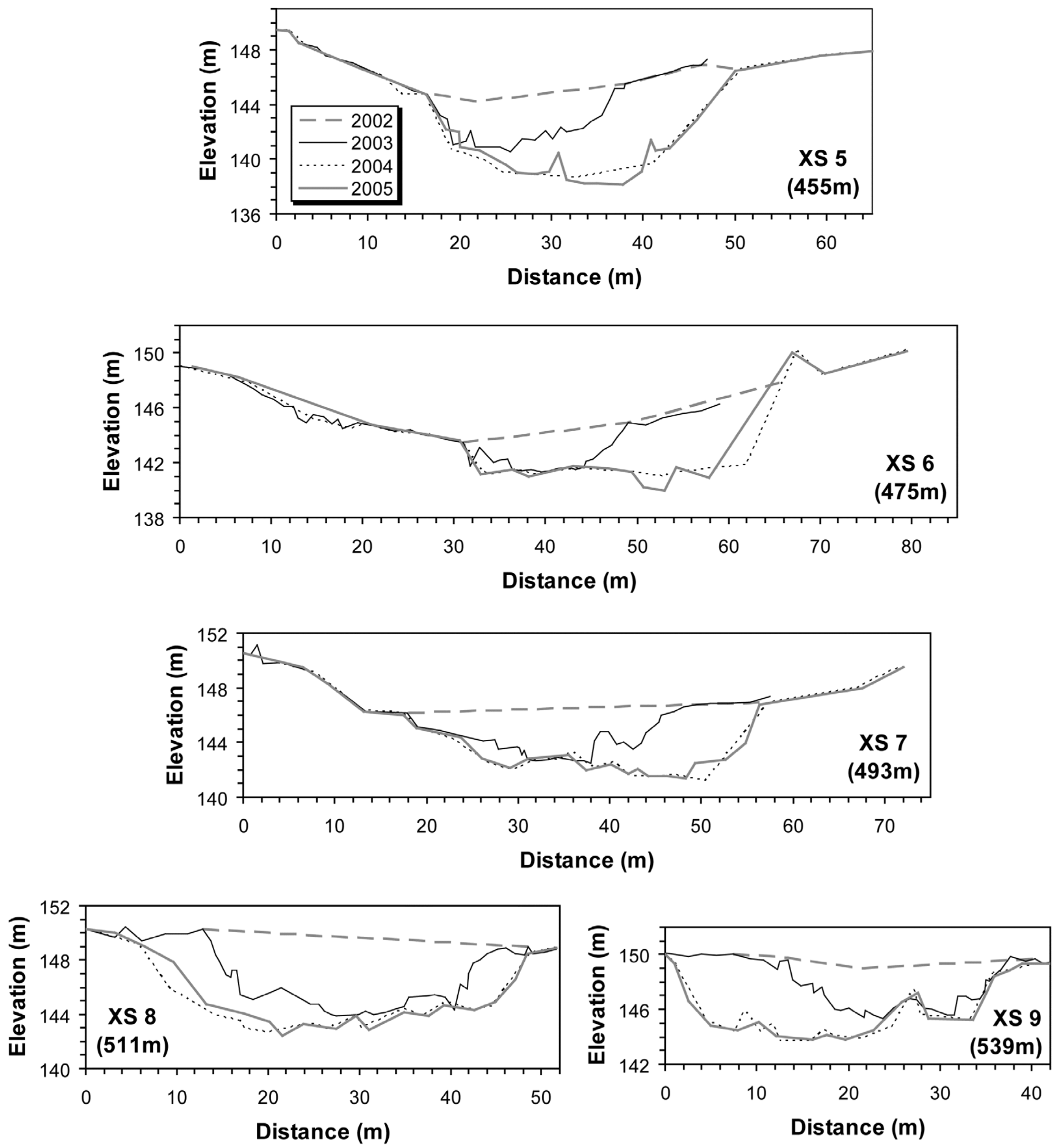


Figure 3.7B: Channel cross sections surveyed in 2003, 2004, and 2005.

Figure 3.8: Photograph of the creek bed in summer 2003 between cross sections 5 and 8. Photo shows ~6 m of vertical incision, bed-surface armoring, and incipient step-pool morphology.

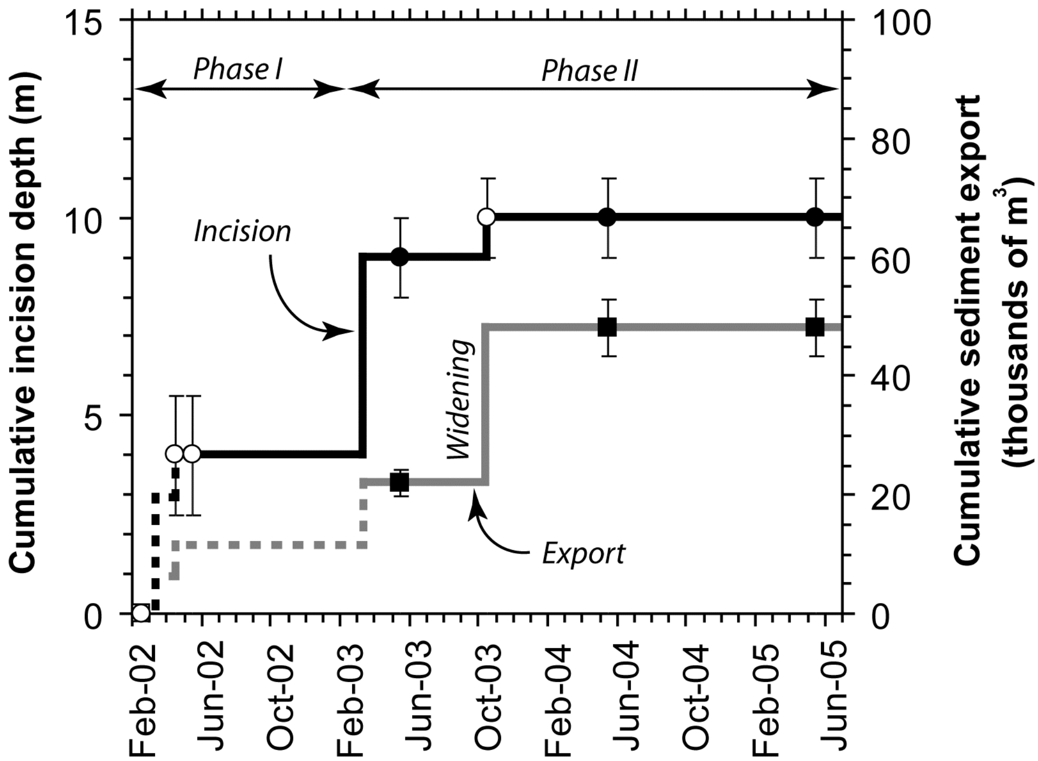


Figure 3.9: Time series of cumulative channel incision and sediment export. Trends were constructed from photo documentation (open symbols) and field surveys (solid symbols). Lines are dashed where quantities are inferred. Error bars for the surveyed incision depth correspond to the D_{90} (~1 m) or more for photo-interpreted incision, and 10% for sediment volume. Stepped increases in temporal trends correspond to storm events of increasing magnitude (Figure 2), which is supported by field observations of bed reorganization. Channel response corresponds to the two-phase model of sediment transport and storage proposed by *Lisle and Church* (2002). Rapid channel incision during winter 2002-03 corresponds to the transport-limited conditions of phase I. Bed armoring and reduction in incision rate drove channel widening and continued sediment export at the transition to supply-limited conditions of phase II.

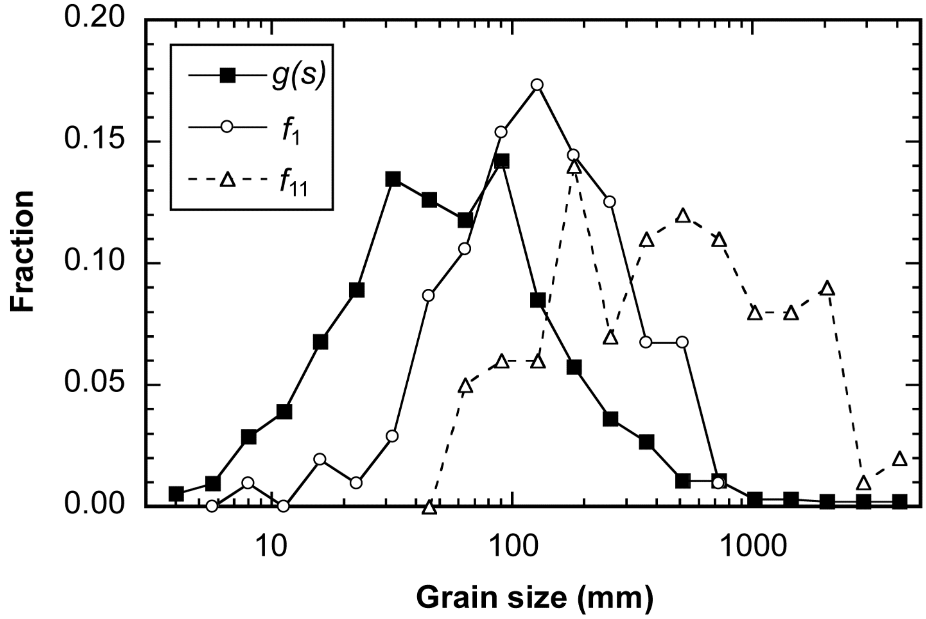
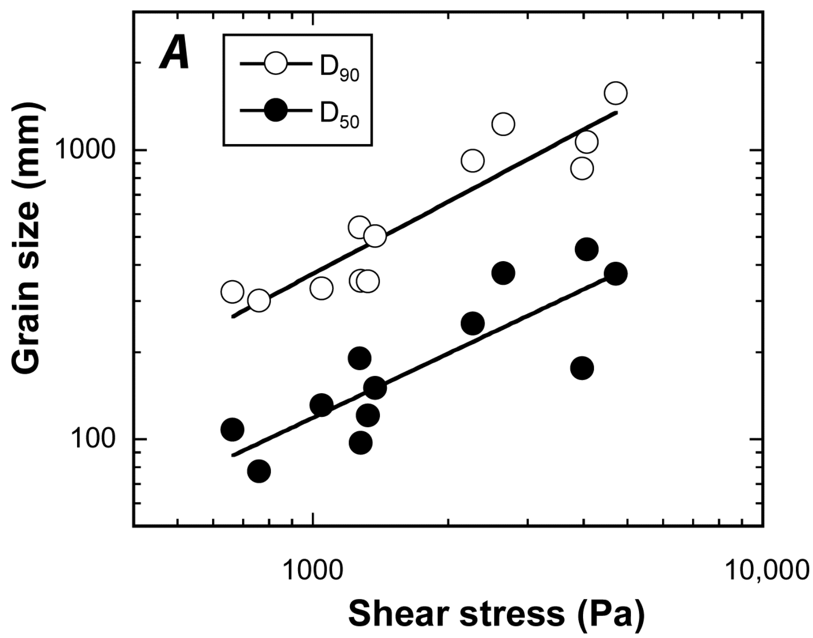


Figure 3.10: Variation in grain-size distribution of sediment supplied by the landslide [$g(s)$] and two representative sample locations from the bed surface of the channel. Bed-surface distribution coarsens relative to the supply with increasing shear stress from f_1 to f_{11} (see Table 3.1).



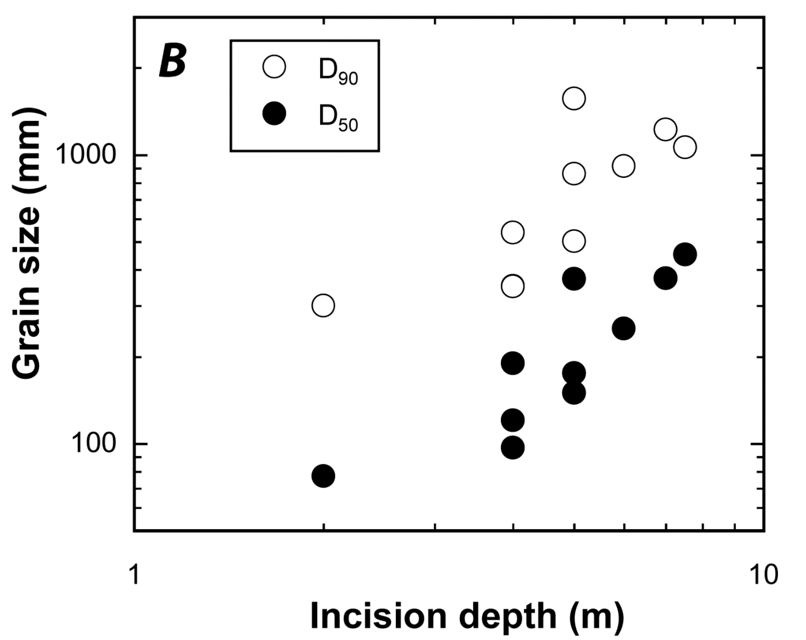


Figure 3.11: Variation in bed-surface grain size versus total shear stress and incision depth. A) total shear stress. B) incision depth.

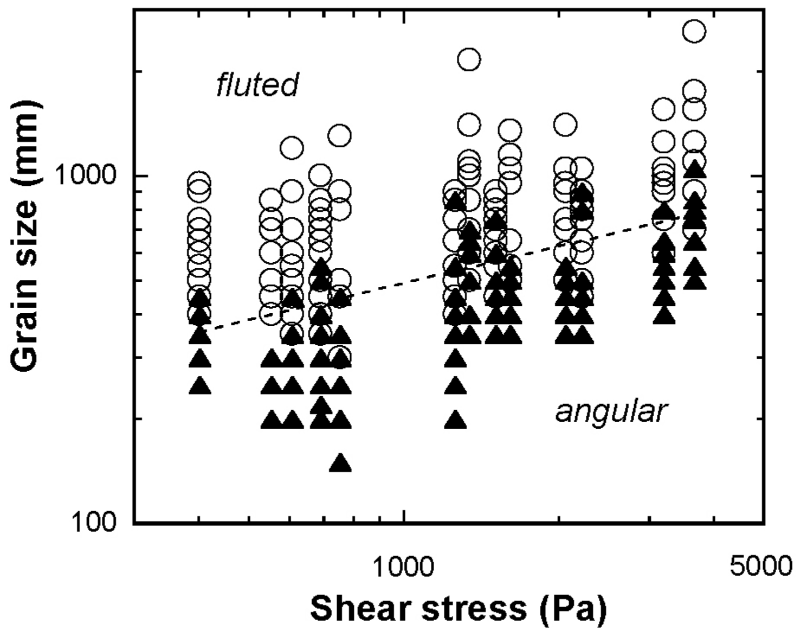


Figure 3.12: Discriminant function derived from the tracer study conducted downstream of the landslide dam. The dashed line ($D = 43\,\tau^{0.35}$, $r^2 = 0.69$) defines the relation between the grain size of fluted (stable) and angular (mobile) clasts.

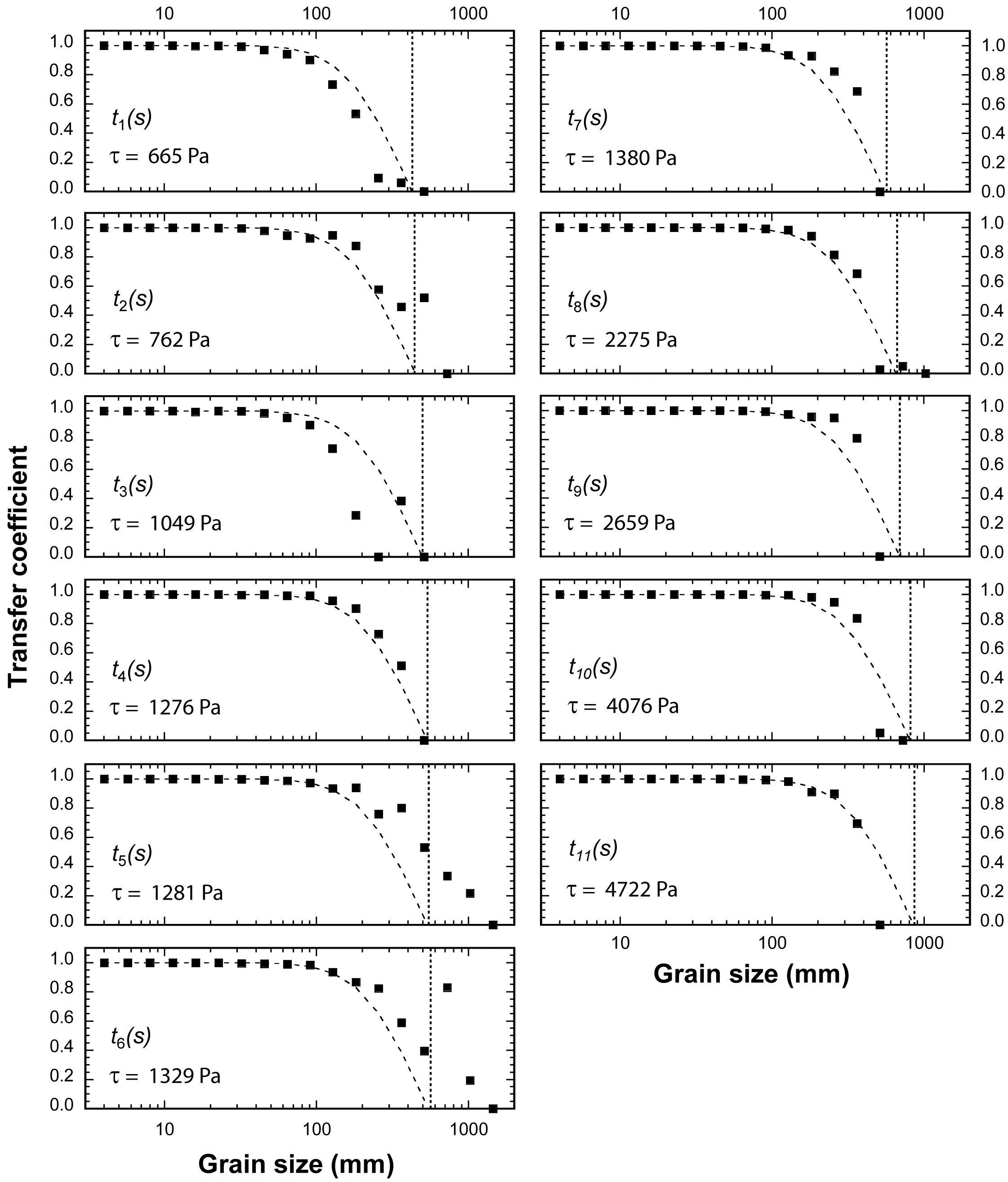


Figure 3.13: Semi-log plots of transfer coefficients versus grain size calculated from Equation 3.10 for the eleven sample locations. Dashed lines are the general solution for transfer functions (Equation 3.14), and vertical lines denote $D_{\max}$.

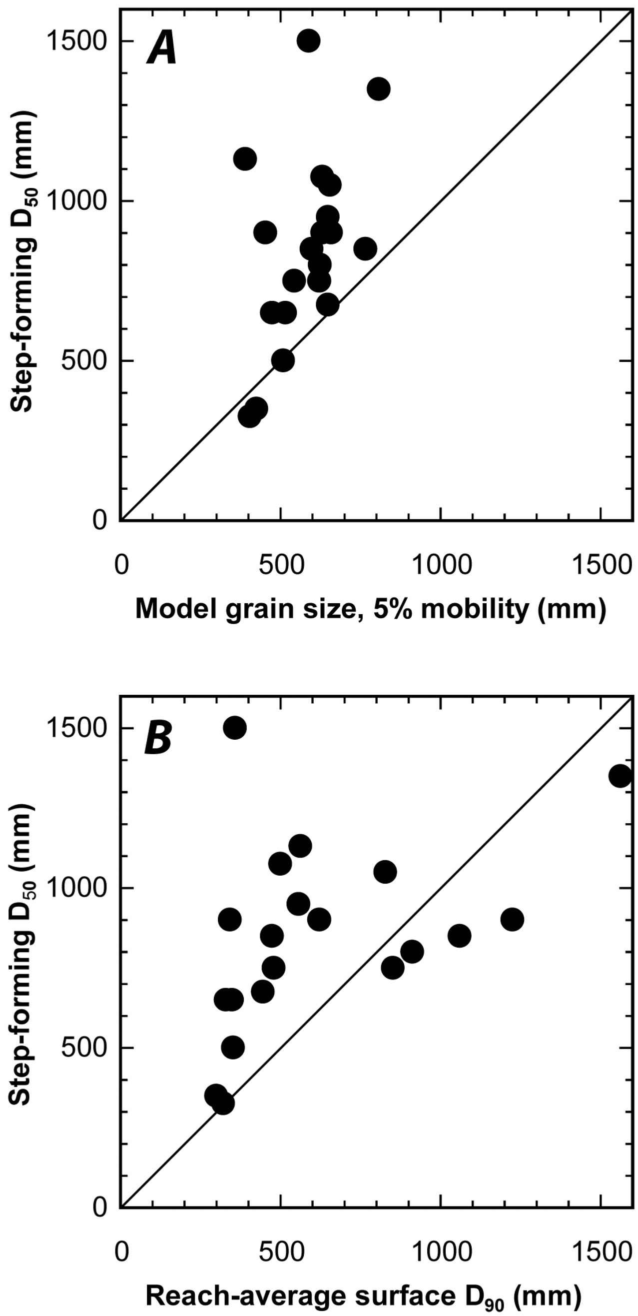


Figure 3.14: Plots of the median grain size of step-forming boulders versus the model grain size with 5% mobility and the reach-average, bed-surface D_{90}. A) the model grain size with 5% mobility. B) the reach-average, bed-surface D_{90}.

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CHAPTER 4: Landslide dams and sediment capacitance in mountain river systems

Summary

Field surveys of recent and ancient landslide dams in the Olympic Mountains and Cascade Range of Washington State indicate that the caliber of the sediment supply, in addition to the unit-width rate of energy expenditure across a landslide dam, control whether an episodic sediment input behaves as a point source of sediment input or as a long-term, sediment-storage element that may provide sediment capacitance to mountain river systems. Specifically, volumetric measurements of sediment exported from ancient landslides by fluvial incision indicate that more than 80% of debris in the coarsest deposits (defined as landslides with > 50% boulders larger than the channel depth) is retained in storage after 10^3 years. In contrast, within a decade, fine-grained landslides (defined as landslides with all sediment sizes transported by typical flows) lost more than 80% of the initial volume available for channel reworking. The long-term average incision rate for the coarsest landslide dams examined in this study is within two orders of magnitude of regional bedrock-lowering rates and may decline to within one order of magnitude after initial bed-surface armoring by an immobile lag. The long residence time of sediment impoundments created by coarse-grained landslides can provide sediment-storage volumes up to three times the volume of the landslide dam itself. Understanding the lithologic and grain-size controls on the residence time of coarse-sediment inputs has practical implications for sediment budgets and sediment routing in mountain river systems. In addition, results of this study suggest that coarse lag deposits may retard bedrock river incision and thus may have implications for models of landscape evolution and for development of "fluvial" relief in mountain drainage basins.

Introduction

The fate of landslide-induced sediment waves or pulses in rivers has been a topic of considerable interest among fluvial geomorphologists and engineers over the past decade (e.g., Lisle et al., 1997; Cui et al., 2003a, 2003b; Sutherland et al., 2002; Cao and Carling, 2003; Hoffman and Gabet, 2006). Results from numerical modeling, laboratory experimentation, and field studies in mountain rivers have demonstrated the dominance of in-place sediment dispersion over sediment-wave translation (Lisle et al., 2001; Sutherland et al., 2002; Cui et al., 2003b). Results from other studies of channel recovery (i.e., Perkins, 1989; Simon and Thorne, 1996; Lisle and Church, 2002; Brummer and Montgomery, 2006) show a reduction in the rate of profile degradation coincident with bed-surface armoring and with development of form roughness. However, relations among particle size distributions of sediment inputs, available stream power, and residence times of sediment pulses remain poorly understood. A substantial fraction of the original volume of a sediment pulse may be stored in a valley bottom if a resistant armor layer forms before degradation advances to the base of the deposit (Lisle and Church, 2002; Gran and Montgomery, 2005; Brummer and Montgomery, 2006). The long residence times of coarse-grained pulses, particularly those that form natural landslide dams, can allow trapping of additional suspended and bed load sediment in backwater impoundments. The additional storage volume may exceed the volume of sediment delivered to the channel by a sediment pulse. Knowledge of the mechanisms governing the longevity of sediment pulses is central to the understanding of how river systems assimilate coarse-grained sediment inputs while actively incising tectonically active landscapes.

Valley-spanning landslides are common in mountain river systems throughout the world (Costa and Schuster, 1988, 1991; Korup, 2002; Garde et al., 2004; Oimet and Whipple, 2004; Strom and Korup, 2006). In the North Cascades and Olympic Mountains of Washington State, frequent landsliding is promoted by moderate rock uplift rates of 0.1 – 1.0 mm yr^{-1} and Pleistocene glacial erosion (Montgomery, 2002;

Montgomery and Brandon, 2002; Reiners et al., 2003; Mitchell and Montgomery, 2006). On occasion, landslides may block valleys and create landslide dams.

Landslide dams can function as large sediment capacitors — storage elements within the channel network that periodically fill with and discharge sediment (e.g., Massong and Montgomery, 2000; Lancaster et al., 2001; Casebeer, 2004). Sediment capacitance can be provided by both the profile convexity formed by the landslide dam itself and by a backwater lake formed behind the dam. The time required to "charge" a sediment capacitor can be estimated from the landslide frequency and, in the case of a backwater lake, from the lake volume and the basin-wide erosion rate. In this study, sediment capacitance (alternatively characterized as the resistance to dispersion) is defined as the ratio of sediment-storage volume within the landslide dam to the magnitude of the potential energy lost across the storage element or landslide dam. These two parameters covary, such that the potential energy of a capacitor decreases as sediment is discharged during profile lowering by erosion. Furthermore, sediment capacitance should be a function of the material properties and geometry of the landslide dam. In this study, the volumes of sediment currently stored in backwater lakes were not measured and thus are not included in estimates of sediment capacitance.

This investigation complements previous studies documenting the rapid dispersion of fine-grained sediment pulses by examining the long residence times of coarse-grained sediment pulses. Landslide dams spanning two millennia in age are examined to test the hypothesis that sediment capacitance varies with the ratio of substrate resisting forces to hydraulic driving forces. Here, the resisting forces are expressed in terms of a reference grain size, and the driving forces are represented by the unit stream power of the degrading channel.

Background

Previous models of stream-profile adjustment incorporate a second-order term resembling the well-known diffusion equation (e.g., Begin et al., 1981; Begin, 1988; Lisle et al., 2001; Cui et al, 2003b; Doyle and Harbor, 2003):

$$\frac{dz}{dt} = K\frac{d^2z}{dx^2} \qquad \text{Eq. 4.1}$$

where z is local bed elevation, t is time, x is longitudinal distance, and K (a proportionality constant) is profile diffusivity with dimensions of unit area per time. It can be shown from equations of sediment continuity that the sediment transport rate per unit channel width, q_s, is proportional to K and channel slope (Begin et al., 1981, Lisle et al., 1997; Doyle and Harbor, 2003):

$$q_s = \rho_s K\frac{dz}{dx} \qquad \text{Eq. 4.2.}$$

Various sediment transport formulae have been incorporated into diffusion-based models such as Equation 4.1 (see Cui et al., 2005 for a review). The bedload transport rate is generally accepted to vary as a nonlinear function of the difference between effective basal shear stress, τ', and critical shear stress required to initiate grain motion, τ_c:

$$q_s = k(\tau' - \tau_c)^n \qquad \text{Eq. 4.3,}$$

where k is an empirical constant and n is an empirically derived exponent generally close to 1.5 (Gomez and Church, 1989). The effective basal shear stress (i.e. that which is available for sediment transport) is generally calculated from the difference between the total, reach-averaged basal shear stress, τ_T, and the shear stress lost to hydraulic roughness, τ_r, which is normally attributed to bedforms, boulders, woody debris, and other obstructions (e.g., Buffington and Montgomery, 1999b):

$$\tau_T = \rho ghS \qquad \text{Eq. 4.4a,}$$

$$\tau' = \tau_T - \tau_r \qquad \text{Eq. 4.4b,}$$

where ρ is fluid density, g is gravitational acceleration, h is water depth, and S is channel slope. The critical shear stress for entrainment is generally modeled after the Shields (1936) equation

$$\tau_c = \theta(\rho_s - \rho)gD_i \qquad \text{Eq. 4.5,}$$

where θ is the dimensionless Shields stress, which typically ranges from 0.03 – 0.06 in gravel-bed channels but can span a broader range of about 0.01 – 0.1 depending on bed packing and bed structures (Church et al., 1998; Oldmeadow and Church, 2005) and on D_i relative to the size of neighboring particles (Andrews and Parker, 1987; Wiberg and Smith, 1991; Buffington and Montgomery, 1997; Johnston and Andrews, 1998).

The total shear stress is related to the rate of energy expenditure per unit bed area, which is defined as unit stream power:

$$\omega = \frac{\rho g Q S}{w} = \tau_T u \qquad \text{Eq. 4.6,}$$

where Q is the mean annual flood discharge, w is channel width, and u is mean velocity. Unit stream power can be expressed in terms of drainage area, A, and slope, S, using regional hydraulic geometry relations for bankfull conditions in western Washington rivers (Montgomery and Buffington, 1997; Brummer and Montgomery, 2003) (Figure 4.1) and regional hydrologic relations between Q and A (Rice, 1998; Whiting et al., 1999; Stock et al., 2005):

$$w \propto A^{0.5} S \qquad \text{Eq. 4.7a,}$$

$$Q \propto A^{1.0} \qquad \text{Eq. 4.7b.}$$

Incorporating these relations into Equation 4.6 leads to

$$\omega \propto A^{0.5} S \qquad \text{Eq. 4.8.}$$

Expressing transport capacity in terms of unit stream power has an advantage over the use of total shear stress because A and S can be readily measured either in the field or from digital elevation models (DEMs). Without accounting for losses due to hydraulic roughness, Equation 4.8 represents the flow energy per bed area available for sediment transport and hence for profile degradation.

The fraction of shear stress lost to hydraulic roughness is strongly influenced by the arrangement and concentration of the largest particles on the bed and by the height to which they protrude into the velocity field (Wiberg and Smith, 1991; Lawrence, 1997; Buffington and Montgomery, 1999a). In cascade and step-pool channels, most of the energy losses accounted for in Equation 4.4b are due to turbulent wakes behind large particles and turbulence below steps (Whittaker and Jaeggi, 1982; Jarrett, 1984; Bathurst, 1993; Wohl and Thompson, 2000). Because the turbulent wake from a large particle can influence a bed area about twice the size of its diameter (Morris, 1950; Ferro and Giordano, 1991), the coarsest particles need not cover more than half the bed surface to exert a dominant influence on partitioning of shear stress and hence to cause a reduction in the sediment transport rate. For example, armoring of only ~20% of the bed with lag boulders after the Squire Creek landslide was sufficient to inhibit vertical incision and force channel widening (Brummer and Montgomery, 2006). Widening can be interpreted as a response to bed armoring with an immobile lag, a mechanism not considered in contemporary models of sediment-pulse dispersion. Shielding of the underlying sediment from further entrainment, followed by a sharp decline in q_s, is expected as protruding lag boulders and step-pool structures dissipate a greater proportion of the total flow energy and increase the critical shear stress necessary for bed mobilization.

Results from the field investigation at Squire Creek (Brummer and Montgomery, 2006) and from related field studies of degrading channels (e.g., Perkins, 1989; Simon and Thorne, 1996; Lisle and Church, 2002) demonstrate the strong dependence of sediment-dispersion rate on the size and concentration of the coarsest particles in a sediment pulse. For example, results of field studies at Squire Creek show that an effective lag formed when the grain size of only ~3% (by volume) of the introduced sediment pulse exceeded the maximum grain size transported by previous flows. Previous analyses of bed reorganization in step-pool channels (e.g., Grant et al., 1990; Lenzi, 2001; Madej, 2001; Kasai et al., 2004; Brummer and Montgomery, 2006) have suggested that a grain-size percentile coarser than the conventional D_{50} (such as D_{90}) may be the appropriate reference grain size for steep alluvial channels, where bed reorganization is controlled by the mobility threshold of the largest step-forming clasts. Because D_{90} (in the absence of other instream roughness such as woody debris) dominates both hydraulic roughness and critical shear stress for bed mobility, the sediment capacitance provided by landslide dams is posited to vary according to the ratio of maximum substrate size (a measure of resistance) to unit stream power as expressed by Equation 4.8 (a measure of hydraulic forcing), thereby yielding a "capacitance index" defined as $D_{90}/A^{0.5}S$. While the exponent on A of 0.5 renders the ratio dimensionless, the potential for different local exponents for Equations 4.7a and 4.7b implies that this expression is not necessarily dimensionless for all environments.

Study areas

This study uses field surveys of recent and ancient landslide deposits in the North Cascades and Olympic Mountains of Washington State, as well as data compiled from prior studies, to test the hypothesis that the decay rate and residence time of sediment stored in landslide dams vary as a function of the capacitance index. Field surveys investigated 13 valley-spanning landslides in the North Cascades and Olympic Mountains of Washington (Figure 4.2). Forests submerged by lakes formed behind the oldest landslides have provided organic material for radiocarbon dating by others

(Schuster et al., 1992; Logan et al., 1998; Pringle et al., 1998). The ages of the landslide dams investigated in this study ranged from < 1 year to approximately 2400 years before present. Data from previous field studies were used to examine three additional landslides in Washington State (e.g., Perkins, 1989; O'Connor, 2004) and one landslide in northern California (Sutherland et al., 2002). Landslide types include rock falls, rock slides, and debris slides using the classification of Cruden and Varnes (1996), which subdivides landslides on the basis of rate and style of movement, water content, and material. Basin areas range from 17 to 396 km^2. Landslides examined in the Cascade Range occurred in rock types ranging from poorly consolidated glacio-lacustrine sediments (Dragovich et al., 2003) to igneous and metamorphic rocks of the North Cascades Complex (Dragovich et al., 2002). Landslides examined in the Olympic Mountains occurred exclusively in Crescent Formation basalt (Gerstel and Lingley, 2003). The valley morphology at most of the Washington field sites was influenced by late Pleistocene advances and retreats of the Cordilleran ice sheet and alpine glaciation, or both. The locations of most of the Olympic and Cascade landslide dams examined in this study coincide with the edge of hanging valleys formed during Pleistocene glaciation.

Previous studies of landslide dams used in this analysis include the landslide dams investigated by Perkins (1989) in Salmon Creek (a tributary to the Chehalis River in southwestern Washington State), the Navarro River landslide in the Coast Ranges of northern California (Sutherland et al., 2002), and the Bonneville landslide on the lower Columbia River (O'Connor, 2004). Perkins (1989) collected grain-size data and measured the diffusivity of several small landslides and debris dams that temporarily blocked Salmon Creek in the 1980s. The Navarro River landslide, which consists of pervasively sheared sandstone and siltstone of the Franciscan Formation, created a temporary dam and backwater lake in 1995. Detailed field measurements of the landslide and of subsequent channel evolution are described by Sutherland et al. (2002). The 1805 Lewis and Clark expedition provided the first written account of the Bonneville landslide. They described the submerged forest of the Columbia River near

Stevenson, Washington and the steep channel constriction at Bonneville, later known as Cascades Rapids and now submerged by the lake created by Bonneville Dam. They posited a recent landslide from nearby hills for the origin of the lake and channel constriction at Bonneville (O'Connor, 2004). Radiocarbon dating of wood recovered from the submerged forest and the landslide deposit indicates blockage of the Columbia River around A.D. 1450 (O'Connor, 2004).

Methods

A combination of field surveys and analyses of digital topography were used to investigate relationships between particle size, unit stream power, and the residence time of sediment storage associated with valley-spanning landslides. The availability of digital topographic maps for western Washington allowed assessment of landslide areas and volumes. Field surveys performed at selected study sites characterized particle-size distributions, classified landslides, and verified digital topography.

Landslide classification

The valley-spanning landslides examined in this study were divided into three categories based a generalized particle-size distribution that could be readily discerned in the field from a representative exposure of the undisturbed deposit. The classification is defined by fine-grained and coarse-grained end members and an intermediate grain-size composition (Figure 4.3). Fine-grained deposits are defined as debris composed primarily of sand and gravel, with all sediment sizes clearly mobilized by the typical annual flow within the degrading channel developed through the landslide dam. A coarse-grained deposit is defined as debris composed of > 50% boulders (by volume) with an intermediate diameter generally > h. The largest particles observed on the beds of the ancient coarse-grained landslides examined in this study typically exhibited evidence of in-place abrasion, which indicates their resistance to frequent bed mobility. An intermediate deposit is defined as a poorly sorted mixture of fine and coarse sediment composed of > 50% sand and/or gravel and between 5% and 50% boulders with a diameter generally > *h*.

Analysis of profile diffusion

Diffusivity coefficients (K) were calculated from Equation 4.1 for selected landslides by comparing the initial (reconstructed) landslide profile with the current channel profile incised over a known period of time. Coefficients were calculated from the slope of the linear regression between the incision rate and the profile curvature. The incision rate was calculated from the incision depth and the landslide age. At Squire Creek, where detailed surveys of the channel during the first four years of profile development were conducted (Brummer and Montgomery, 2006), profile curvature was calculated from the second derivative of a polynomial fit to the profile. For the remainder of the sites, where the topographic resolution of the initial landslide surface was poorly defined, curvature was calculated from the change in slope between the two adjacent upstream and downstream survey points, and diffusivity was calculated for sites that yielded a satisfactory correlation ($r^2 > 0.5$) between incision rate and profile curvature.

Volumetric analysis

The existing sediment capacitance of landslide dams is expressed as the dimensionless volume of sediment remaining in valley storage (V/V_o), where V is the sediment volume remaining at the present time, and V_o is the initial volume of landslide debris accessible to fluvial reworking and export from the valley bottom. Sediment capacitance is also expressed as the normalized storage volume of the impoundment formed behind the landslide dam (V_s/V_o). Volumetric calculations were performed in a geographic information system (GIS) using a 10-m digital elevation model (DEM) of Washington State digitized from 7.5-minute USGS topographic quadrangles. High-resolution (2-m) DEMs generated from recent laser altimetry were available for the lower portion of the Church Mountain landslide and the North Fork Stillaguamish River valley prior to the North Fork Stillaguamish landslide. In addition, an electronic total station was used to survey stream profiles at selected sites.

The initial landslide volume was measured using one of two techniques depending on the fill geometry. The first technique was used for landslides that filled wide, unconfined valleys to a shallow depth relative to the landslide footprint. For this technique, the initial debris volume was calculated from the product of the average depth and the footprint area. Landslide depth was estimated from projections of the valley longitudinal profile and valley walls beneath the landslide. Landslide extents were delineated by analyzing a combination of historical aerial photographs, landslide polygons depicted on the 1:100,000 scale digital geologic map series for Washington State (Washington State Department of Natural Resources, 2005), morphologic expressions depicted in 10-m and 2-m (where available) DEMs, and field mapping of landslide deposits performed for this study.

A second technique was used for landslides that filled confined valleys (most of the landslides in this study). For these landslides, volumes were calculated from paired longitudinal profiles of the estimated pre-landslide thalweg and the initial landslide surface above the thalweg (Figure 4.4). Profiles were constructed from field surveys and from points sampled from the DEMs in a GIS. Because most of the landslides in this study coincide with a profile convexity, the pre-landslide profile was approximated from projections of both upstream and downstream profiles and estimated lake depths. The intersection of these projections beneath the landslide was checked against the projection of the valley walls and available lake bathymetry. Two pre-landslide profiles were generated for each site to estimate the minimum and maximum thicknesses and thus the range of initial landslide volume to be calculated later from these profiles. A profile representing the initial landslide surface (prior to fluvial incision) was constructed from the top of stream banks surveyed in the field or from points sampled from DEMs along the edge of inferred landslide terraces. A cross-sectional area representing the removed sediment was computed at each point along the profile using a bank slope of 35 degrees, which is within the range of angles measured in the field.

The cross-sectional areas also accounted for the channel width, which was either measured in the field or calculated from regional hydraulic geometry relations (Figure 4.1). The initial landslide volume was then calculated from the cross-sectional areas and profile distance between consecutive cross sections. Lake volumes were measured using the same technique. The volume of sediment removed by erosion was calculated from the difference in elevation between the inferred initial landslide surface and the field-surveyed or DEM-derived channel profile using the same technique. Relative errors in calculated volumes were generally less than ± 10% based on the upper and lower bounds used for the original landslide depth.

A normalized, long-term average sediment export rate, $(1-V/V_o)/t$, was computed using the landslide age, t, in years. Normalizing the long-term rates to V_o allows the sediment export rates to be expressed independent of the original volume and in units of yr^{-1}. The calculated export rates are considered minimum rates for landslides that have lost all of their initial sediment volume. In this case, the minimum export rate is simply the inverse of the landslide age in years. Relative errors in calculated export rates were generally less than ± 60% based on the upper and lower bounds used for the original landslide depth and ± 1σ given for calibrated landslide ages.

Particle-size and unit stream power measurements

Wolman (1954) pebble counts of at least 100 particles were used to calculate the D_{90} of both the in-situ landslide deposits and the bed surfaces of alluvial channels incised into the landslide deposits. The D_{90} of landslide debris was measured from sediment exposed in eroding stream banks and road cuts. For channels that were not wadable, D_{90} was approximated from the mean of the diameters of several step-forming boulders measured using a 7-m survey rod extended across the channel during low flow. Data reported in previous studies were analyzed to obtain D_{90} for the Squire Creek landslide (Brummer and Montgomery, 2006), Salmon Creek landslide (Perkins, 1989), Church Mountain landslide (Carpenter, 1993), Hazel landslide (Millar and Sias, 1998), Navarro River landslide (Sutherland et al., 2002), and Bonneville landslide (O’Connor, 2004). The D_{90} of the deposit was substituted for the bed-surface D_{90} for landslide dams that were not yet breached by an alluvial channel (e.g., Lena, Jefferson, Slide, and Glacier landslides). The supply-based D_{90} is considered more representative of the hydraulic resistance to incision during initial profile development. However, field observations indicate that bed-surface armoring and the development of bed structures, such as steps and particle clusters, commence rapidly. Hence, D_{90} of the bed surface is considered more representative of hydraulic roughness and is used as the reference grain size in subsequent expressions of the capacitance index.

The drainage area at each site was calculated in a GIS using a flow accumulation grid generated from the 10-m DEM, except for the Bonneville landslide, where the drainage area is reported at the U.S. Geological Survey gage 14128870 located approximately 1.5 km downstream of the landslide toe. In keeping with the definition of capacitance (i.e., the potential energy across a capacitor), the channel slope for each site was calculated from the difference in elevation between the initial dam crest (e.g., the lake outlet) and the landslide toe and the profile distance between these two points.

Results

Landslide classification and channel morphology

Most landslide dams investigated in this study were classified as either intermediate or coarse-grained (Table 4.1). The lower Church Mountain and North Fork Stillaguamish landslides, along with landslides from the previous studies used in this investigation, were classified as fine-grained landslides. Although the beds of some fine-grained landslides became armored during incision, rapid profile adjustment returned the North Fork Stillaguamish landslide and the Navarro River landslide (Sutherland et al., 2002) to their pre-landslide width and gradient within a decade. Channels incised into the coarsest landslides investigated for this study developed steep-gradient ($S > 0.02$), armored beds. The bed surfaces of these channels typically coarsened downstream over a convex profile from the lake outlet to the landslide toe (e.g., Figures 4.4 and 4.5). Well-sorted material observed on boulder bars directly downstream of the toe of the youngest landslide dams was presumably transported from the convex profile and deposited below the toe along a concave profile of declining transport capacity. Aggradation downstream of the toe is consistent with sediment pulse decay by dispersion, as observed by others (Lisle et al., 2001; Sutherland et al., 2002; Cui et al., 2003b). The channel morphology (*sensu*, Montgomery and Buffington, 1997) observed along the profile of coarse-grained landslides progressed from pool-riffle or plane-bed at the lake outlet to step-pool and cascade structures at the landslide toe. A boulder nickpoint was observed in the field upstream of the initial landslide toe at many of the coarse-grained landslides. Woody debris played a functional role in providing hydraulic roughness and sediment retention in all of the recent landslide dams investigated in this study.

Profile diffusion and incision

Except for the well-documented Squire Creek landslide, attempts to quantify the long-term, average diffusivity of most landslide dams in this study met with limited success due in part to the poor resolution of reconstructed landslide surfaces. About half of the profiles in this study yielded no or a weak correlation between incision rate and profile curvature. Within the subset of landslides in this study that did (only intermediate and coarse-grained landslides), the long-term, average diffusivity ranged from 20 – 987 $m^2 yr^{-1}$ (Table 4.2). The lowest diffusivity was measured for the relatively old Hamma Hamma landslide. The highest diffusivity (excluding Squire Creek) was measured three years after initial failure and incision of the Goodell Creek landslide, where the diffusivity is about the same order of magnitude as the diffusivity measured after the first year at Squire Creek. Diffusivities of the older, intermediate composition landslides examined in this study ranged from 28 – 126 $m^2 yr^{-1}$, one to two orders of magnitude lower than the initial rates measured at the youngest landslides. The long-term average incision rate for the coarsest landslide dams examined in this study ranged from 8 – 21 mm yr^{-1}, which is within two orders of magnitude of regional bedrock-lowering rates of 0.1 – 0.3 mm yr^{-1} for the Olympic Mountains and Cascade Range (Brandon et al., 1998; Reiners et al., 2003).

The diffusivity measured at the Squire Creek landslide (an intermediate-composition landslide) declined rapidly during the first three years of profile degradation. Squire Creek incised a maximum of 9 m into the landslide deposit by the end of the first full rainy season (Figure 4.5). Vertical incision was less than 2 m the second year, despite the higher-magnitude flows for this period. The diffusivity of the Squire Creek landslide declined from 3110 – 177 m^2 yr^{-1} during the first two years of profile evolution (Figure 4.6, Table 4.2). Channel surveys and field observations confirm that there was no further incision between the 2004 and 2005 surveys. Deviations from zero in the 2004-05 incision rates reflect the precision of survey measurements, which correspond to the diameter of the coarsest particles on the bed (~ 1 m). The three-year average diffusivity between 2002 and 2005 was 960 m^2 yr^{-1}. During the first year, the profile adjustment at Squire Creek was characterized by two distinguishable diffusion coefficients. The highest diffusivity was measured along most of the profile length, except at the toe, where the incision rate was highest but least sensitive to profile curvature. The diffusivity during the first year at Squire Creek was the same magnitude as the diffusivity measured by Perkins (1989) for the first two years of incision of two fine-grained landslides.

Nickpoints formed from a congestion of large boulders were observed at several of the study sites. At Squire Creek, several 2- to 3-m-diameter boulders formed a 4-m-high nickpoint at a distance of 15% of the profile length from the toe prior to the first survey in 2003 (Figure 4.5). A second, 5-m-high boulder nickpoint formed about 30 m downstream of the first during the high-magnitude flow between the 2003 and 2004 surveys. In four years of field surveys at Squire Creek, these nickpoints did not migrate. Two 4-m boulder nickpoints were also observed downstream of the Canyon Lake landslide. If the uppermost nickpoint formed at the toe, then it has propagated through roughly 50% of the profile length in about 130 years. Based on descriptions of the Bonneville landslide by Lewis and Clark (O’Connor, 2004), the nickpoint forming

Cascade Rapids appears to have migrated through about 70% of the landslide length in less than 350 years.

Sediment storage

The dimensionless volume of sediment remaining in storage (V/V_o) varies by landslide type and age (Table 4.1). Overall, the coarse-grained landslides (regardless of age) retained more than 80% of their initial sediment volume, which is greater than the volumes retained by the oldest intermediate and fine-grained landslides. The lowest values of V/V_o are found in the fine-grained landslides, which retained only 20% – 50% of their initial volumes after less than one decade.

Sediment impoundments forced by coarse-grained landslide dams as were as much as than three times the volume of the landslide dam behind which they were formed (Table 4.1, Figure 4.7). Lake volumes were found to be independent of landslide size and were instead determined by landslide height and the gradient, width, and confinement of the valley. For instance, landslides damming gently sloping valleys created the largest lakes investigated in this study (e.g., Bonneville, Packwood, Lena, Jefferson, Canyon, and Day landslides), whereas landslides blocking steep valleys formed relatively smaller lakes and provided less overall sediment capacitance (e.g., Squire, Goodell, Church Mt., and Sultan landslides). In addition, lakes formed by recent, fine-grained landslide dams were short-lived (e.g., North Fork Stillaguamish, Navarro, and Salmon landslides), whereas some lakes formed by coarse-grained landslides have persisted for more than a century (e.g., Bonneville, Packwood, Lena, Jefferson, Canyon, and Day landslides). Assuming that short-term erosion rates keep pace with measured, long-term erosion rates of 0.1 – 0.3 mm yr^{-1}, the time to fill the sediment impoundments examined in this study is on the order of 10^2 – 10^3 years for the largest lakes and on the order of ≤ 10 years for the smaller impoundments formed in large rivers.

Sediment export rates

The normalized, long-term average sediment export rate (1-V/V_o)/t varied by nearly four orders of magnitude among the landslides investigated in this study (Figure 4.8). Landslides plot in three distinct domains defined by both landslide type and age. The highest export rates were measured for the fine-grained landslides < 10 years old. The lowest, long-term export rates were measured for coarse-grained and intermediate landslides > 100 years old, which include landslides retaining the greatest fraction of initial sediment volume. The fine-grained, lower Church Mountain landslide plots within the group characterized by relatively low export rates because of its relatively old age; the entire volume is assumed to have been removed based on the rapid decay rates of similar fine-grained landslides. Thus, its export rate is considered a minimum. The third group of landslides is characterized by an intermediate export rate that is up to about one order of magnitude lower than the fine-grained group, but two orders of magnitude greater than the group of older, coarse and intermediate landslides. This third group consists of entirely of young (< 10 years old), intermediate and coarse-grained landslides.

Previous studies of channel recovery from intense sediment loading have used an exponential decay function to model the variation in sediment export rate with time (e.g., Simon, 1992; Simon and Thorne, 1996; Lisle and Church, 2002) Simon and Rinaldi (2006) introduced an exponential decay function of the form

$$\frac{V}{V_o} = a + (1 - a)e^{-rt} \qquad \text{Eq. 4.9,}$$

where r is an empirical coefficient that determines the rate of change of sediment-storage volume per unit time, and a is a scaling coefficient that defines the minimum value of V/V_o, where this ratio becomes asymptotic. Lisle and Church (2002) used this approach to model the exponential decay of sediment reservoirs and called r a transport-storage coefficient. The applicability of Equation 4.9 to describe the change in the dimensionless ratio of sediment volume with time was evaluated using the ensemble of fine-grained landslides ranging in age from < 1 to 2400 yrs. Figure 4.9 shows that changes in V/V_o with time for fine-grained landslides behaves according to Equation 4.9 with a transport-storage coefficient of 0.25 and a scaling coefficient of zero. This curve roughly discriminates the fine-grained landslides from the intermediate landslides. A second exponential curve (r = 0.35, a = 0.7) defines the boundary between the intermediate and coarse-grained domains. Accounting for the scaling coefficient of the second curve results in a comparable transport-storage coefficient of about 0.1, which is less than the coefficient for the fine-grained curve and indicates a relatively slower decay rate or greater sediment capacitance.

Two intermediate landslides plot within the coarse-grained domain of Figure 4.9 with anomalously high values of V/V_o. These include the Sultan River landslide, where the apparent lag-time for sediment export may reflect artificially regulated flows, and the Canyon Lake landslide, the youngest landslide in its age class. Other than the Sultan and Squire landslides (which are both < 10 years old), the Canyon Lake landslide is the only intermediate landslide with a high V/V_o ratio relative to the capacitance index (Table 1). In general, the segregation of most landslides types into three distinct domains, which are defined by various forms of Equation 4.9, supports the hypothesized dependence of sediment capacitance on particle size and is consistent with the particle-size-based landslide classification presented earlier.

Discussion

Although erosion rates are initially high for all landslide types, fine-grained landslides decayed an order of magnitude faster than the initial rate for intermediate and coarse-grained landslides. Field results suggest that the initial export rate for landslide dams with only moderate concentrations of large boulders typically declines by several orders of magnitude within a decade due to bed-surface armoring with lag boulders. At Squire Creek, about 85% of the profile adjustment occurred during the first full rainy season, although lateral erosion of the banks during high-magnitude flows between the

2003 and 2004 surveys exported as much sediment volume as did the first year. Lateral erosion during this event also exposed landslide debris stored within a profile convexity remaining from an earlier event. Based on the size of the oldest trees growing on the old landslide surface, the residence time of the sediment stored in this profile convexity is at least several hundred years. Pre-existing profile convexities and overlapping landslide deposits observed at other field sites provide evidence of repeated landsliding (e.g., Goodell, Slide, Jefferson, Bonneville, and Sultan landslides). Additionally, the coincidence of several landslide dams with hanging valleys (e.g., Packwood, Glacier, Lena, Slide, and Squire landslides) suggests that landsliding at these sites may be linked to post-glacial profile adjustment and local oversteepening of valley walls along narrow valley reaches with high profile curvature. This style of topographic control on landsliding provides a mechanism by which large sediment capacitors can be periodically "recharged." The long-term morphological significance of landslide dams should therefore depend on the profile decay rate relative to the recurrence interval of landsliding.

The rapid decline in late-stage erosion rates measured for intermediate and coarse-grained landslides signals a repartitioning of the flow energy from particle entrainment to mechanical abrasion. Field observations of pervasive polishing, fluting, and potholing of the largest bed particles suggest extensive abrasion of immobile bed particles. Field observations also document "ratcheting" open of joints in large boulders by hydraulic wedging of clasts (e.g., Hancock et al., 1998; Whipple et al., 2000; Springer et al., 2005) and downstream movement of lag boulders a distance of several times the length of their diameter afforded by the undermining of supporting clasts (e.g., Trowbridge, 1911; Gage, 1953). Excluding initially rapid incision rates, late-stage degradation rates for the intermediate and coarse-grained landslides investigated in this study may be within an order of magnitude of local bedrock erosion rates.

Field observations at the study sites suggest that nickpoint propagation may be an important mechanism of profile adjustment for intermediate and coarse-grained landslides. Nickpoints form in response to a drop in base level or an increase in channel gradient by concentrating flow energy at a single step-like discontinuity in the longitudinal profile (Wohl, 2000). Nickpoints have been observed in a variety of geologic materials but are best developed where resistant and jointed bedrock is exposed in the channel (Miller, 1991; Weissel and Seidl, 1998). Nickpoint propagation involves undercutting at the toe by a hydraulic jump and the release of sediment during bank failures triggered by profile lowering (Harsbargan and Paola, 2000). Winnowing of poorly sorted sediment released in the wake of an upstream-propagating nickpoint would leave a coarse lag on a bed of diminished gradient and thereby shield the remaining debris from incision. Therefore, focusing of energy by a migrating nickpoint may provide a mechanism of bed armoring previously unrecognized in hillslope-coupled channels.

The close association of coarse-grained landslide deposits with strong bedrock lithologies suggests that rock strength and fracture density determine grain size, which in turn determines the rate and style of sediment-pulse dispersion and the residence time of sediment storage in mountain river systems. The slow dispersion of extremely coarse, well-sorted landslide dams can also be attributed to lower flows due to leakage through porous deposits. Subsurface drainage has inhibited the development of surface flows and has lead to increased sediment capacitance at several of the study sites (e.g., Lena, Jefferson, Slide, and Glacier landslides). The thickness of landslide debris relative to the concentration of immobile boulders in the deposit should also influence sediment residence times. Shielding of landslide debris by a coarse lag before degradation to the base of the deposit increases the residence time of stored sediment and adds sediment capacitance to mountain drainage basins.

Field observations from this study and from the investigation of lag formation at Squire Creek (Brummer and Montgomery, 2006) can be synthesized into an empirical model describing possible controls on the tendency for a sediment pulse to function as a capacitor or to disperse rapidly. The plot of D_{90} versus $A^{0.5}S$ for landslides examined in this study (Figure 4.10) defines a continuum of sediment-pulse behavior ranging from long-term sediment capacitance for coarse-grained landslides to rapid sediment dispersion for fine-grained landslides. In general, bedrock landslides originating in crystalline rock types form coarse-grained pulses with the lowest long-term dispersion rates. Shallow landslides originating in weak or fractured rock types form fine-grained pulses with relatively high dispersion rates.

Data from this study suggest that grain size alone can discriminate between lag-prone deposits that sequester sediment and fine-grained deposits that disperse rapidly (horizontal, dashed line in Figure 4.10). However, results from a tracer study and analysis of lag formation during the degradation of the hydraulically rough bed at Squire Creek indicate a shear stress dependence on the maximum mobile grain size (Brummer and Montgomery, 2006). Based on results from the Squire landslide, the particle size with a 5% probability of entrainment, D_{lag}, varies with unit stream power according to the relationship $D_{lag} = 0.75(A^{0.5}S)^{0.44}$ ($r^2 = 0.87$, $n = 11$). This alternative model is represented by the dotted line in Figure 4.10, which also separates the domain of high sediment capacitance from the domain of rapid sediment dispersion. An exponent on this relationship of less than 1 indicates a stronger dependence on D_{90} than $A^{0.5}S$ for the development of a sediment capacitor. Although available data from the western Washington field sites and from the northern California site do not reveal a D_{90} dependence on $A^{0.5}S$, data from these field sites do not refute this relationship. According to this model, channels with a bed-surface D_{90} plotting above the dotted line should be hydraulically stable under conditions similar to Squire Creek. For instance, the open diamonds immediately above this line are the D_{90} particles measured at the Squire landslide and represent the smallest step-forming particle sizes measured in the channel. The diamond below the line is from the Sultan River landslide, which had not yet experienced the high flows necessary for substantial degradation and armoring. Armoring and a reduction in unit stream power following profile lowering would shift this landslide along a trajectory toward greater sediment capacitance. The square in the upper right of the plot is the Bonneville (Columbia River) landslide. Due to the dearth of data between the D_{90}-based and Squire Creek criteria, data from global landslide dams spanning a greater range in unit stream power would be required to assess the relevance of a unit stream power dependence on sediment-pulse behavior for landslide dams other than Squire Creek.

Nonetheless, the influence of bedrock lithology and relative grain size on the rate of sediment-pulse dispersion has implications for controls on the supply, storage, and routing of sediment in mountain channels, the application of sediment budgets in mountain river systems, and models of landscape evolution. For example, recent field, experimental, and numerical studies have focused on the dispersion of relatively fine-grained sediment pulses because their fluvial evolution can be readily observed in the field and can be modeled with accepted bedload-transport formulae (e.g., Lisle et al., 2001, Sutherland et al., 2002; Cui et al., 2003b). However, fine-grained landslides that disperse rapidly may predominate in regions underlain by weak lithologies (e.g., accretionary wedges or mélange belts), whereas coarse-grained landslides capable of providing substantial sediment capacitance may be more common in the crystalline core of mountain ranges. Consequently, the applicability of current models of sediment-pulse dispersion that assume mobility of all particle sizes may be limited geographically by lithology. Although an understanding of how fine-grained sediment pulses evolve has practical application to problems such as the dispersion of alluvial sediment following dam removal, the problem of how intermediate and coarse-grained pulses disperse and store sediment in mountain river systems has been largely overlooked. Results of this study find that the lag deposits left by the fluvial sorting of non-fluvial sediment delivered by intermediate and coarse-grained landslides can produce a hydraulically rough armor layer that may limit degradation to the equivalent of several surface layers (Lisle and Church, 2002; Brummer and Montgomery, 2006), thereby moderating the rate of sediment-pulse dispersion and reducing the downstream effects of large sediment inputs. In addition, the shielding of bedrock surfaces by coarse-grained deposits may retard river incision, and thereby increase the “fluvial” relief of mountain drainage basins (Lancaster and Grant, 2006).

Conclusions

Field surveys and GIS analyses of longitudinal profiles of recent and ancient landslide dams in the Olympic Mountains and Cascade Range of Washington State show that the sediment-export rate of landslide debris by mountain channels varies as a function of a representative particle size and the grain-size distribution of the landslide debris, which are largely controlled by bedrock lithology. Among the channel-spanning landslides examined in this study, channels armored with coarse pebbles or cobbles degraded rapidly and released more than 80% of their initial sediment volume to the channel network within a decade. In contrast, channel profiles that became armored with large boulders (D_{90} greater than ~ 1 m in western Washington) before degradation to the base of the landslide had the longest residence time for sediment storage and retained more than 80% of their original sediment volume after ~ 10^3 years. Additional sediment is trapped and sequestered in upstream impoundments created by landslide dams. After initial lag development, the long-term rate of profile degradation for the coarsest landslide dams may be within one order of magnitude of regional bedrock lowering rates. Coarse-grained landslides therefore function as effective sediment capacitors, sometimes providing sediment storage in a backwater lake that is up to three times the landslide volume.

The results of this study show that confined mountain channels subject to episodic inputs of coarse sediment can retain a long memory of disturbance. Large boulders that appear to be immobile during typical sediment-transporting flows can represent the legacy of selective transport and armoring of coarse-grained sediment inputs. Hence, this study supports the findings of previous studies of sediment-pulse degradation that showed the dominance of sediment-pulse evolution by dispersion rather than by wave translation, when the caliber of the sediment pulse is similar to or coarser than the maximum size transportable by the receiving channel. However, the findings of this study further suggest that the applicability of sediment-pulse dispersion models relying on bedload-transport formulae, which neglect the role of lag deposits, may be limited to mountain channels underlain by weak or highly fractured bedrock.

Table 4.1: Landslide dams examined in this study and data examined from previous studies.

Landslide	Age (yrs) [a]	Basin Area (km^2)	S [b]	$A^{0.5}S$ (km)	D_{90} [c] (m)	$D_{90}/A^{0.5}S$ [d]	V/V_o	$(1-V/V_o)/t$ (yr^{-1})	V_s/V_o [e]	Lithology
Coarse-grained landslides										
Goodell Creek	3	94.0	0.076	0.733	1.5	2.05	0.85	0.050	0.059	granite
Slide Lake	528 (53, 173)	17.2	0.050	0.207	2.18	10.5	1.0	0.0	3.6	granodiorite
Glacier Lake	704 (27, 108)	20.1	0.20	0.885	1.25	1.41	0.84	0.00023	1.4	granite/volcaniclastic
Jefferson Lake	1001 (115, 151)	19.4	0.071	0.314	1.5	4.78	1.0	0.0	2.4	basalt
Lena Lake	1115 (112, 114)	26.2	0.10	0.516	1.76	3.40	0.95	0.000047	0.26	basalt
Hamma Hamma River	1188 (104, 418)	104	0.030	0.307	1.95	6.35	0.82	0.00015	1.8	basalt
Bonneville Dam (Columbia R.)	365	619,000	0.029	22.8	10	0.438	0.85	0.00042	1.6	basalt
Intermediate landslides										
Squire Creek (2002-03)	1.5	26.0	0.10	0.512	0.82	1.60	0.82	0.12	0.048	meta-gabbro
Squire Creek (2003-04)	2	26.0	0.097	0.492	0.82	1.67	0.62	0.19		meta-gabbro
Squire Creek (2004-05)	3	26.0	0.091	0.465	0.82	1.77	0.62	0.13		meta-gabbro
Canyon Lake	134 (83, 162)	11.9	0.075	0.260	0.60	2.31	0.89	0.00082	1.6	meta-sandstone
Packwood Lake	974 (54, 186)	56.2	0.085	0.639	1.5	2.35	0.54	0.00047	0.55	intrusive/volcanic
Day Lake	1591 (155, 126)	24.6	0.036	0.179	0.75	4.22	0.55	0.00028	0.27	sheared phyllite
Upper Church Mt.	2401 (355, 188)	246	0.033	0.511	2.0	3.91	0.34	0.00027	0.35	volcaniclastic
Sultan River	1.7	223	0.10	1.54	0.7	0.454	0.96	0.023	1.1	serpentinite

Table 4.1 (continued)

Landslide	Age (yrs) [a]	Basin Area (km^2)	S [b]	$A^{0.5}S$ (km)	D_{90} [c] (m)	$D_{90}/A^{0.5}S$ [d]	V/V_o	$(1-V/V_o)/t$ (yr^{-1})	V_s/V_o [e]	Lithology
Fine-grained landslides										
Lower Church Mt.	2401 (355, 188)	246	0.013	0.209	0.3	1.43	0.00	0.00042		sandstone
N.F. Stillaguamish R.	5	396	0.027	0.545	0.01	0.0183	0.10	0.18	0.065	glaciolacustrine seds.
Navarro R. 1995 [f]	0.1	535	0.078	1.79	0.11	0.061	0.96	0.40	3.3	sheared siltstone
Navarro R. 1995-96 [f]	1	535	0.038	0.867	0.20	0.231	0.73	0.27		sheared siltstone
Navarro R. 1996-97 [f]	2	535	0.024	0.559	0.20	0.358	0.48	0.26		sheared siltstone
Navarro R. 1997-99 [f]	4	535	0.017	0.386	0.20	0.519	0.20	0.20	0.20	sheared siltstone
Salmon Creek S87a-1 [g]	1	7.5	0.093	0.254	0.19	0.734	0.52	0.48		sandstone
Salmon Creek S87a-2 [g]	2	7.5	0.093	0.254	0.19	0.734	0.38	0.31		sandstone

[a] Age in years before 2006, except for the Bonneville landslide, where age is calculated from the 1805 Lewis and Clark expedition. Calibrated ages reported as intercept age and (-1σ, $+1\sigma$). See Pringle et al. (1998) and Logan et al. (1998) for methods and radiocarbon ages.

[b] Slope calculated from the elevation drop between the initial lake outlet and landslide toe along the entire profile length of the landslide dam.

[c] Bed-surface D_{90}.

[d] The sediment capacitance index based on the bed-surface D_{90} (m) and $A^{0.5}S$ (km).

[e] The ratio of the lake volume to initial landslide volume.

[f] Sutherland et al. (2002)

[g] Perkins (1989)

Table 4.2: Diffusivity coefficients and incision rates for selected landslides.

Landslide	K ($m^2\ yr^{-1}$)	*dz/dt* (mm/yr) [a]	Age (yrs)	Landslide Type [b]
Canyon Lake	28	94	134	I
Goodell Creek	987	1300	3	C
Hamma Hamma River	20	18	1188	C
Squire Creek (2002-03)	3110	6100	1.5	I
Squire Creek (2003-04)	177	2600	1	I
Squire Creek (2002-05)	960	3300	3	I
Upper Church Mt.	126	21	2401	I
Salmon Creek, site S87a (Perkins, 1989)	1000 - 2500	600 -1600	1 - 2	F
Salmon Creek, site S500 (Perkins, 1989)	1200	600	1	F

Note: NS = not significant.
[a] Incision rate (mm/yr) measured at the point of maximum profile degradation.
[b] Abbreviations correspond to coarse-grained (C), intermediate (I), and fine-grained (F) landslides using the classification presented in this study.

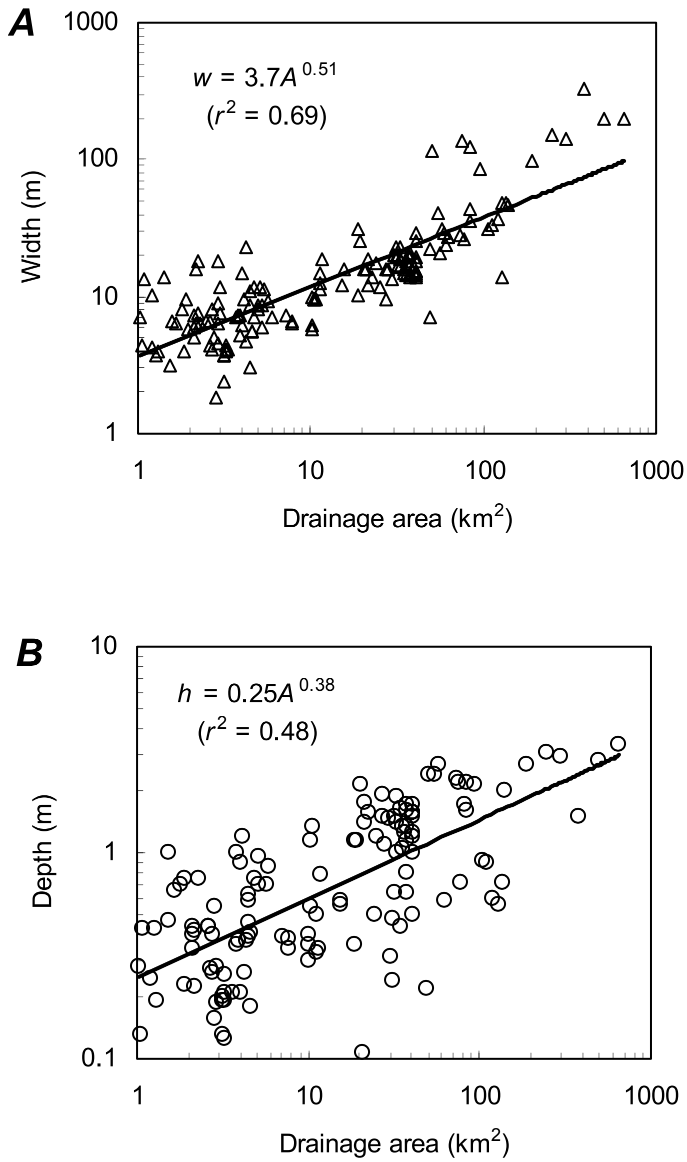

Figure 4.1: Plots of regional hydraulic geometry relations for western Washington rivers. A) bankfull width versus drainage area. B) bankfull depth versus drainage area. Both relations are for drainage basins > 1 km^2 and were developed from data collected during previous field studies (Montgomery and Buffington, 1997; Brummer and Montgomery, 2003).

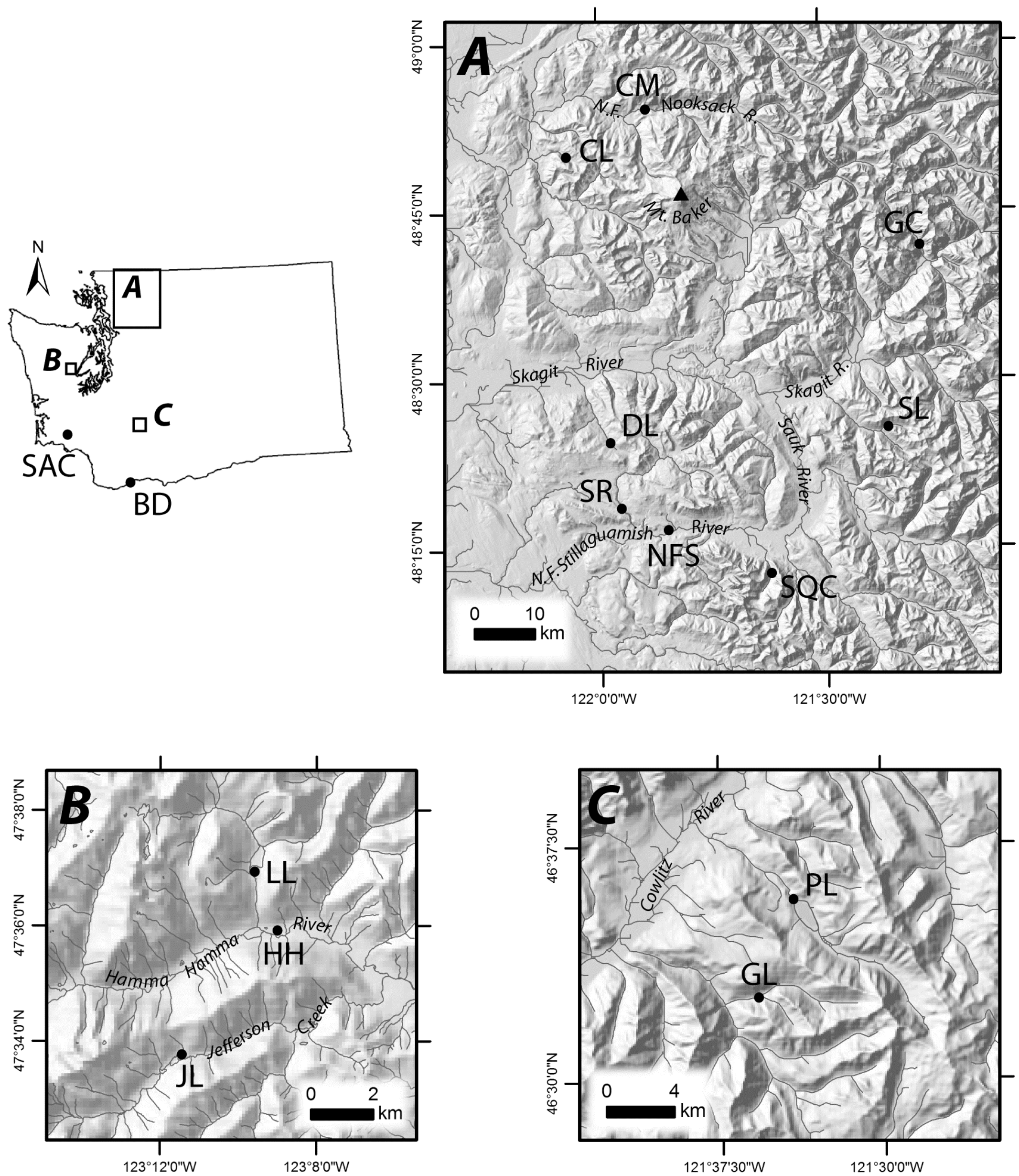

Figure 4.2: Locations of landslide dams investigated in western Washington for this study. State map shows inset maps and the locations of the Salmon Creek (SAC) landslides (Perkins, 1989) and the Bonneville Dam (BD) landslide (O'Connor, 2004). A) North Cascade landslides: Church Mountain (CM), Canyon Lake (CL), Goodell Creek (GC), Day Lake (DL), Slide Lake (SL), Sultan River (SR), North Fork Stillaguamish River (NFS), and Squire Creek (SQC). B) southeast Olympic Mountain landslides: Lena Lake (LL), Hamma Hamma River (HH), and Jefferson Lake (JL). C) southern Cascade landslides: Packwood Lake (PL) and Glacier Lake (GL).

Figure 4.3A: Field example of a fine-grained landslide dam on the N.F. Stillaguamish River. Flow is from right to left in the photograph. Aerial photos dating back to 1933 and the 1890 land survey maps show reactivation of an ancient landslide by chronic bank erosion at the outside meander bend. Reactivation of the landslide delivers large pulses of galciolacustrine sediment to the N.F. Stillaguamish River every 10 – 20 years. The most recent failure in January 2006 completely blocked the N.F. Stillaguamish River with fine sediment and woody debris. The emergency cutoff channel shown in the photo was excavated through the meander bend by the Army Corps of Engineers to alleviate flooding of riverside homes caused by the backwater impoundment. Photo credit: Dan Crowell.

Figure 4.3B: Field example of an intermediate composition landslide at Squire Creek. The landslide temporarily blocked the creek in February 2002 and created a small pond that extended 115 m upstream. The photo shows about 8 m of degradation and lag formation after four years. Note the person on the right bank for scale.

Figure 4.3C: Field example of a coarse-grained landslide at Slide Lake. The landslide dam formed about 500 year ago and traps all suspended and bed load entering the lake. Slide Lake drains through the coarse landslide dam, which has no surface outlet. The large granite boulders in the photo measure 2 – 4 m in diameter.

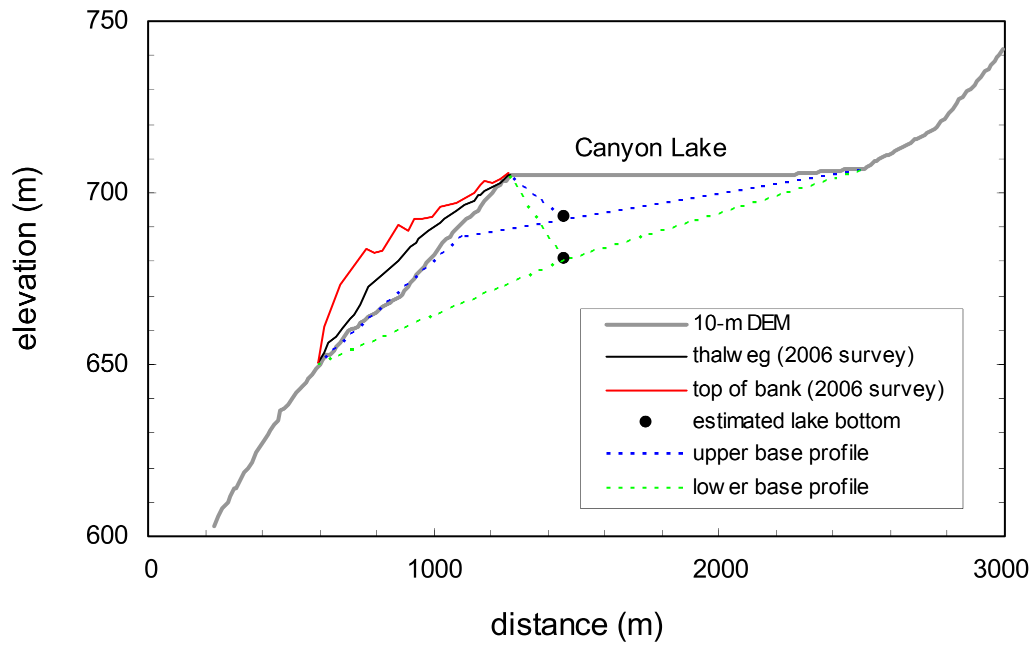

Figure 4.4: Illustration of the methods used to calculate landslide volumes. Surface profiles through the Canyon Lake landslide were generated from the 10-m-grid-size DEM and field surveys of the channel thalweg and top of bank. The solid symbols show the upper and lower estimates of the bottom of Canyon Lake based on projections of the valley beneath the lake. The minimum and maximum lake depths set the upper and lower bounds at these points for the generation of two pre-landslide profiles (colored, dashed lines used in the volumetric calculations). The location of the inflection in the upper-bound profile was fit to the projection of the surface profile extracted from the 10-m DEM. Both pre-landslide profiles tie in at the landslide toe and lake inlet.

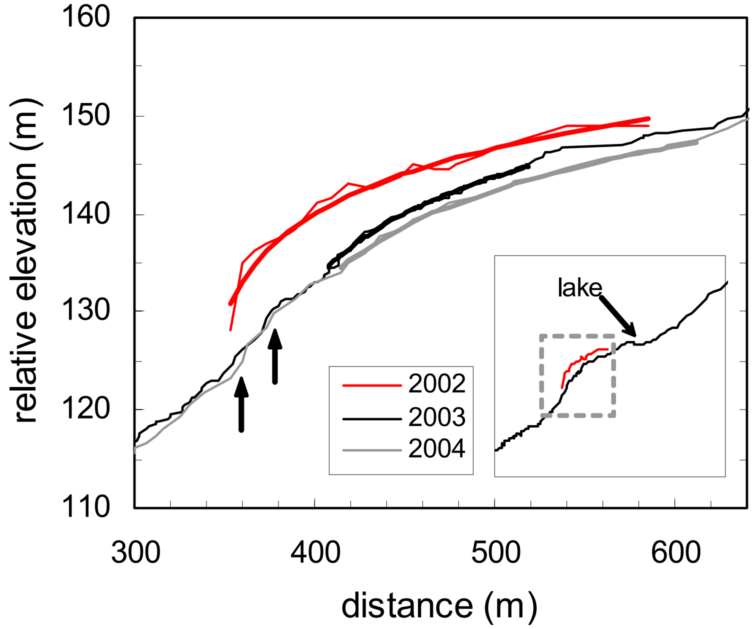

Figure 4.5: Longitudinal profile of the Squire Creek landslide. Thin lines show annual surveys. Thick lines are the regression curves used in curvature calculations. Inset shows the profile convexity in 2004. The 2005 profile (not shown for clarity purposes) is indistinguishable from the 2004 profile. Vertical arrows point to nickpoints described in the text. Note aggradation (boulder bar mentioned in text) occurring between 2003 and 2004 surveys downstream of nickpoints.

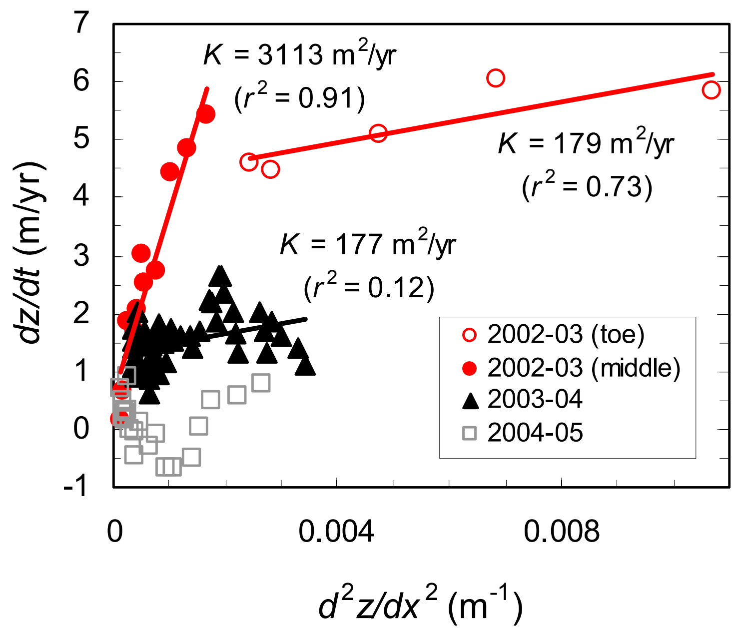

Figure 4.6: Plot of incision rate versus profile curvature for the Squire Creek landslide. The calculated diffusivity is given for selected years. The correlation between incision rate and curvature is poor after 2004 due to a reduction in degradation following lag formation in 2003. The precision of the survey is equivalent to the diameter of the largest boulders on the bed (~ 1 m).

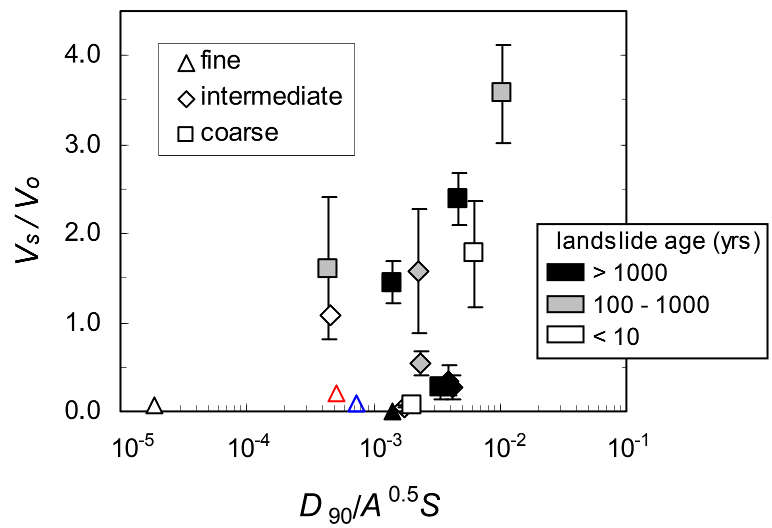

Figure 4.7: Plot of dimensionless lake volume (V_s/V_o) versus capacitance index ($D_{90}/A^{0.5}S$). Symbol shading represents landslide age (there were no landslides 10 – 100 years old). Symbol color represents the data source: black outline (this study), red outline (Sutherland et al., 2002), and blue outline (Perkins, 1989). Some of the older landslides examined in this study are not associated with a modern lake.

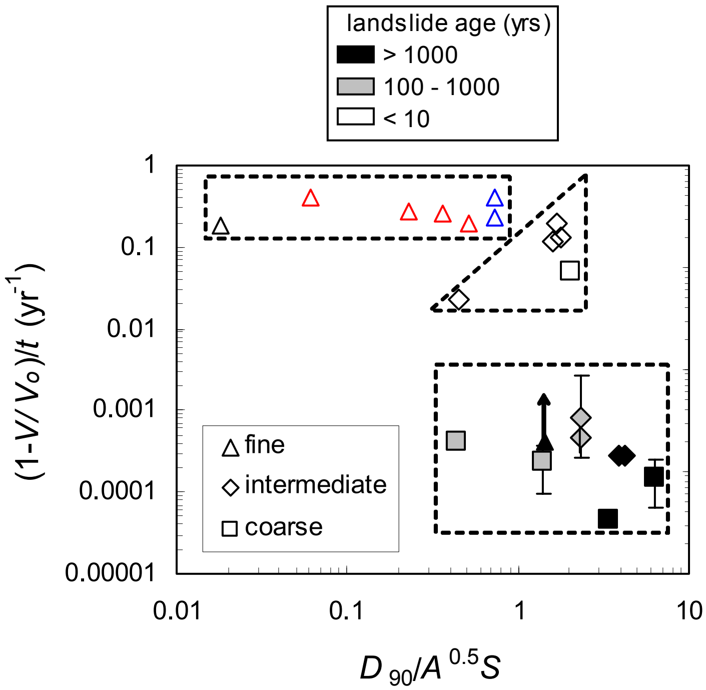

Figure 4.8: Plot of the normalized, long-term sediment export rate [(1-V/V_o)/t] versus the bed surface-based capacitance index ($D_{90}/A^{0.5}S$). Symbology is the same as in Figure 4.7. The arrow on the black triangle designates a minimum age for the lower Church Mt. landslide, where V=0. Dashed boxes define three landslide groups based on age and grain-size classification.

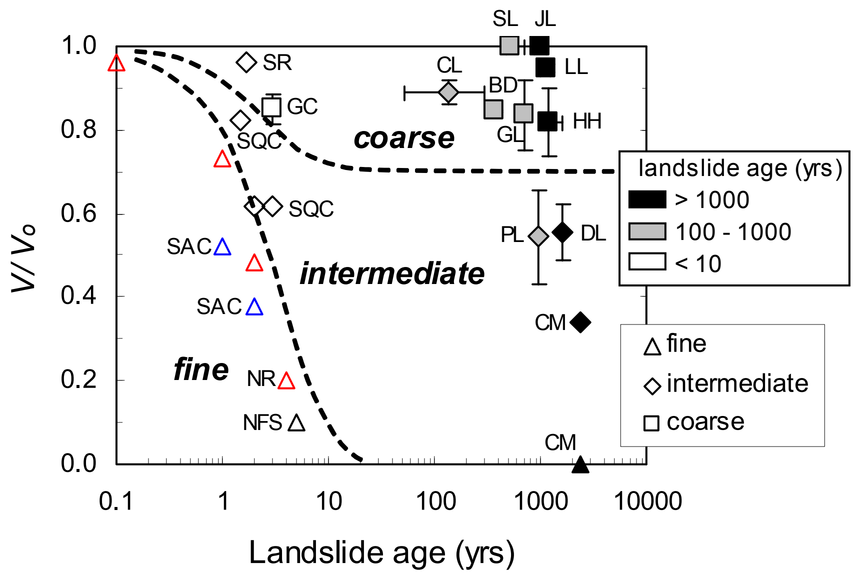

Figure 4.9: Plot of V/V_o versus landslide age. Dashed lines calculated from Equation 4.9 separate domains of landslide type and relative decay behavior. Labels identifying study sites are the same as in Figure 4.2. Three symbols shown for the Squire Creek landslide (SQC) depict the time-series of decay measured in the field. Symbol color represents the data source: black outline (this study), time-series for the Navarro River landslide (NR) in red outline (Sutherland et al., 2002), and the Salmon Creek landslides (SAC) in blue outline (Perkins, 1989).

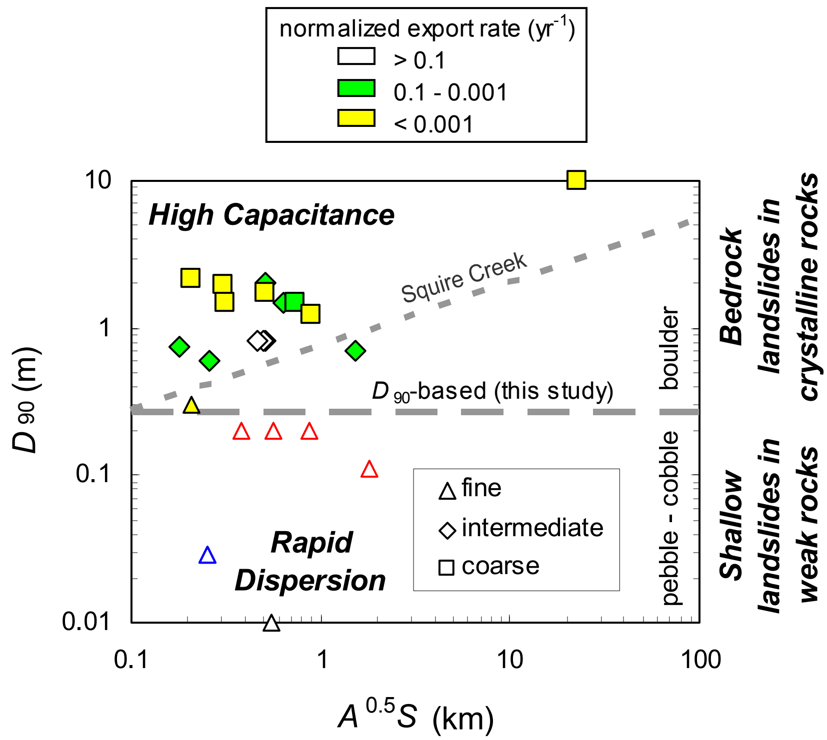

Figure 4.10: Field-based model illustrating the possible dependence of sediment capacitance and dispersion rate on bed-surface D_{90}, unit stream ($A^{0.5}S$), and bedrock lithology. Symbology is the same as Figure 4.9, except for symbol shading, which correspond to the normalized, long-term average sediment export rate, $(1-V/V_o)/t$ (see legend). The horizontal, dashed line separating regions of sediment capacitance and dispersion is a simple model based on grain size. Bedrock landslides originating in crystalline rocks form coarse-grained landslides with the greatest capacitance, whereas shallow landslides originating in weak or fractured rock types form fine-grained landslides with relatively high dispersion rates. An alternative model is represented by the dotted line, which is an empirical relationship between $A^{0.5}S$ and the particle size with a 5% probability of entrainment. This equation was derived during the degradation of the hydraulically rough bed at Squire Creek (Brummer and Montgomery, 2006) and has the form $D_{lag} = 0.75(A^{0.5}S)^{0.44}$ ($r^2 = 0.87$, $n = 11$). The dotted line represents the threshold particle size for entrainment as a function of unit stream power. Bed-surface D_{90} plotting above this line should be hydraulically stable under hydraulic conditions similar to Squire Creek.

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