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Development of channel organization and roughness following sediment pulses in single-thread, gravel bed rivers

Mary Ann Madej

Abstract. Large, episodic inputs of coarse sediment (sediment pulses) in forested, mountain streams may result in changes in the size and arrangement of bed forms and in channel roughness. A conceptual model of channel organization delineates trajectories of response to sediment pulses for many types of gravel bed channels. Channels exhibited self-organizing behavior to various degrees based on channel gradient, presence of large in-channel wood or other forcing elements, the size of the sediment pulse, and the number of bed-mobilizing flows since disturbance. Typical channel changes following a sediment pulse were initial decreases in water depth, in variability of bed elevations, and in the regularity of bed form spacing. Trajectories of change subsequently showed increased average water depth, more variable and complex bed topography, and increased uniformity of bed form spacing. Bed form spacing in streams with abundant forcing elements developed at a shorter spatial scale (two to five channel widths) than in streams without such forcing mechanisms (five to 10 channel widths). Channel roughness increased as bed forms developed.

1. Introduction

Watershed disturbances can increase the rate and magnitude of sediment supply to a channel. Sediment pulses (defined in this paper as an episodic input of sediment that is several times greater than the average annual sediment input to a channel) may be caused by natural events, such as volcanic eruptions, fire, and floods, or by human activities, such as surface mining, timber harvest, road failures, dam breaks, etc. Associated mechanisms of increased sediment supply (sediment pulses) in a channel network, such as mass movements, gullying, surface erosion, or release of stored sediments, are stochastic processes that can generate regions of high bed load transport and bed material storage in channels [*Benda and Dunne*, 1997a]. Many studies have addressed the magnitude and movement of sediment pulses in natural channels [*Gilbert*, 1917; *Pickup et al*., 1983; *Nicholas et al*., 1995; *Madej and Ozaki*, 1996; *Benda and Dunne*, 1997b; *Lisle et al*., 1997], although these studies have not quantified the development and arrangement of bed forms following the generation of a sediment pulse.

Channel bed forms cover a range of scales from boulder clusters (microform) to unit bars (macroform) to sedimentation zones (megaforms) [*Church and Jones*, 1982]. Although the evolution of bed forms as a type of channel organization has been well studied over a range of hydraulic conditions in sand-bedded rivers [*Jain and Kennedy*, 1974], the development of regular bed forms with characteristic length scales in gravel bed rivers is not well known. *Montgomery and Buffington* [1997] broadly define bed form patterns in various types of channel reaches to be either “multi-layered, laterally oscillatory, vertically oscillatory, featureless, random, irregular or variable.” A rigorous analysis of bed form variability across a range of channel gradients is lacking, however, as is the concept of changes in channel structure, organization, and roughness values following disturbances. An understanding of bed form development at several scales is important because bed forms provide resistance to flow and consequently influence flow properties.

A study of changes in channel morphology accompanying high supplies of coarse sediment to forested mountain streams must consider three aspects: the material composing the channel bed and banks, how that material interacts to form channel structures, and how those channel structures are organized. Material forming the channel boundary can be fine or coarse sediment, small or large wood, or bedrock outcrops and can be considered the building blocks of the channel and its bed forms. Common metrics of channel material are particle size and distribution; bed and bank composition and heterogeneity; and presence and size of large wood. The interactions of channel material with flow give rise to channel structure. In this paper, I consider channel structure as being analogous to the architecture of an individual building, in which channel material is arranged to form higher-order features or structures. Metrics used to describe structure range over several scales and include bed imbrication and sorting and the size and type of channel features, such as boulder clusters, pools, bars, steps, and riffles. At a higher order still, channel organization develops. By extension of the architectural analogy, channel organization is comparable to the arrangement of individual buildings in a community and refers to the spatial arrangement or pattern of channel units or other bed features. This spatial distribution can be random, regular, or clustered; the metrics used in quantifying channel organization in the channel are the spacing, frequency, variability, and regularity of structures, and they are measured by pattern analyses and various spatial autocorrelation techniques.

The present study focuses on changes in channel structure, organization, and roughness accompanying high supplies of coarse sediment to forested, mountain streams, based on boundary conditions, channel gradient, the presence of forcing

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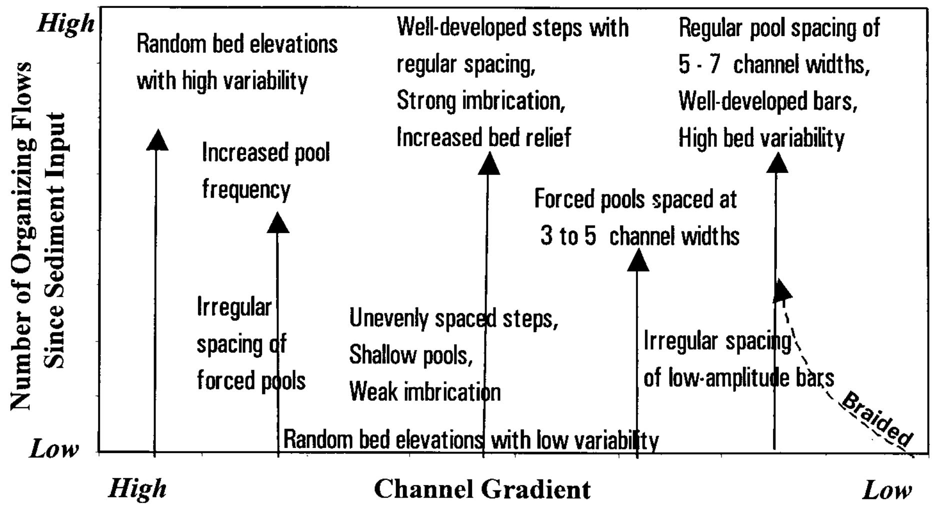

Figure 1. Conceptual model of the development of channel bed structure and organization following sediment pulses in single-thread, coarse-grained rivers. The schematic shows expected trajectories of channel morphologic change in streams across a range of channel gradients.

elements, and the number and magnitude of flows capable of mobilizing bed material (called "organizing flows"). Consequently, as a channel processes a sediment pulse, the signal of the pulse may be manifested through a change in structure, organization, and roughness in the channel, although not necessarily by movement of a discrete sediment wave.

2. Conceptual Model

The purpose of this paper is to propose a conceptual model of a developmental sequence of channel changes following sediment pulses and to evaluate the model using field studies across a range of channel gradients. I hypothesize first that spatial scales of channel structure and organization, as well as the time required to develop channel organization, differ in different parts of the channel network, based on channel gradient and occurrence of organizing flows. Second as the channel becomes organized into distinct bed forms, channel roughness should increase.

Figure 1 depicts the conceptual model of the development of bed organization following a sediment pulse. The x axis represents different channel gradients ranging from steep, headwater streams to low-gradient, high-order gravel bed rivers. The arrows in Figure 1 represent trajectories of change leading from a poorly organized state immediately following a sediment pulse to a more strongly organized channel. Channel bed organization is defined by variability in bed topography and the degree of regularity of bed form spacing. The degree to which a channel is organized is dependent upon the number of organizing events since disturbance (the y axis). An organizing flow (defined as a discharge greater than the critical discharge required for bed load transport, $Q > Q_{cr}$) is considered a necessary condition for channel organization because bed particles must be mobilized in order to be rearranged and for associated bed forms to be altered. For example, if a rockfall or earth flow delivers a large volume of boulders to a cobble bed stream, the coarse particles will be deposited in the channel, but flows may not be competent to reorganize the boulders. In contrast, a large volume of more easily mobilized sediment entering a channel will be reorganized to various degrees, depending on the presence of in-channel wood or other forcing elements and the frequency of organizing flows.

The frequency of organizing flows varies in different parts of the channel network because of different dominant bed particle sizes, channel gradients, and stream power. Other factors also influence bed mobility, such as the degree of armoring, imbrication, bed packing, grain protrusion, boulder clusters, and bed relief. In addition, the magnitude of an organizing flow should not be considered a constant through time at a given channel location. As the channel processes a sediment pulse, the channel bed commonly becomes coarser; thus the magnitude of an organizing flow increases through time, and the frequency of such flows decreases.

Channel response to sediment pulses varies in different parts of the channel network. In steep, low-order channels recently subjected to debris flows, bed variability is initially low, or absent if scoured to bedrock, but the bed develops more complexity through time as additional wood and sediment enter the channel. In these streams the trajectory of change (Figure 1) leads to increased variability of the channel bed, with a few pools but with no regular spacing or frequency of bed forms. Flows great enough to mobilize the wood and dominant sediment size (organizing events) in such channels are rare.

In mountain streams with gradients of ~3–10%, steps composed of wood or boulders are common. Fine sediment following a pulse may initially bury the previous stepped structure, but it can be transported through these channels quickly. In this case, channel structure and organization will be reestablished through exhumation of wood, boulders, and other structural elements. However, if the underlying step-forming clasts and wood are disturbed, as will be examined in this investigation, the development of steps with regular spacing, height, and length characteristics will be slow, because the frequency of organizing flows is low. Initially, steps will be weakly imbricated, pools will not be well developed, and the steps will not display a regular spatial scale or pattern for a given channel gradient. Through time, bed material is scoured from the pools, pools become deeper, and step-forming clasts become reorganized into more regularly spaced, stable, strongly imbricated steps with distinctive height and length characteristics. Examples of well-developed step structure are described by *Whittaker and Jaeggi* [1982], *Grant et al*. [1990], *Abrahams et al*. [1995], *Wohl et al*. [1997], and *Chin* [1999].

In lower-gradient, higher-order streams, pool-riffle channel morphology is common. Bar and pool topography generated by local flow convergence and divergence may be either freely formed by cross-stream flow and sediment transport or forced by channel bends and obstructions [*Lisle*, 1986; *Montgomery and Buffington*, 1997]. A sediment pulse in such channels can result in a more planar bed as pools fill, roughness elements are buried, and the channel aggrades. Subsequently, the channel evolves from a random, low-variability channel bed to one with forced bed forms caused by scour and deposition around obstructions and finally to a well-organized, highly heterogenous channel bed with regularly spaced bed forms. Depending on the abundance of forcing elements, a channel may not evolve to the end-member of a channel formed only through alternating convergent and divergent flow patterns (a self-formed channel). In a single-thread pool-riffle channel, if the magnitude of the sediment pulse is great enough, a braided planform may emerge temporarily (dashed line in Figure 1), which has no characteristic length scale [*Sapozhnikov and Foufoula-Georgiou*, 1999]. As the channel processes the sediment pulse, it may eventually revert to a single-thread, pool-riffle system. The development of organization at any stage can be interrupted by the input of additional sediment or wood.

3. Previous Studies of Channel Organization

Many recent studies have addressed the phenomenon of spatial self-organization in physical systems [*Hallet*, 1990]. Certain types of stream channels, such as those displaying pool-riffle or meander morphology, have long been recognized to have a regular pattern. The spatial scale of this organization has usually been studied at a channel unit or reach scale. *Nelson* [1990] considered alternate bar formation to be a natural self-patterning phenomenon and showed that bar wavelengths increased following perturbations of a flat channel bed until eventually the bed forms stabilized. Both horizontal and vertical self-organization have been identified in braided rivers, which evolve toward a critical state [*Sapozhnikov and Foufoula-Georgiou*, 1999]. The present study will consider only vertical and downstream organization in river profiles (development of bed forms) and not horizontal organization (meander development) because many steep mountain streams have limited channel migration zones and meander development. In addition, the degree of vertical organization can be compared across many channel types.

Studies of bed form spacing have covered a range of conditions. Regular spacing of pools and riffles, commonly at five to seven channel widths, has been documented by many researchers [*Richards*, 1976; *Keller and Melhorn*, 1978; *Milne*, 1982]. Pools are not necessarily regularly spaced in streams with a high number of forced pools, such as those with high wood loading [*Montgomery et al*., 1995]. In Alaskan streams with high wood loading, autocorrelation analyses showed no significant regularity in stream depth [*Robison and Beschta*, 1989]. *Madej* [1999] used spatial autocorrelation coefficients to quantify the gradual development of regularity in a pool-riffle channel following the input of high sediment loads.

In rivers too steep to display a pool-riffle morphology, different types of channel organization have been described. Even in a gravel bed stream without well-defined bars, a coarse channel bed can develop stone cells [*Church et al*., 1998], stone lines [*Laronne and Carson*, 1976], boulder clusters [*Brayshaw*, 1984], or transverse ribs [*Koster*, 1978]. On a microscale, *Robert* [1991] used spatial autocorrelation to document bed roughness due to skin friction in coarse-grained channels. *Furbish et al*. [1998] suggested that dominant wavelengths of alternate and midchannel bars in steep, rough channels can be discerned. *Kaufmann* [1987] showed weak regularity in thalweg depths in streams subjected to debris torrents.

In channels dominated by steps and pools, morphologic characteristics and spacing of steps have been quantified by several researchers [*Grant et al*., 1990; *Abrahams et al*., 1995; *Wohl et al*., 1997; *Chin*, 1999]. *Grant et al*. [1990] described the pattern and origin of step-pool topography and found channel units were associated with distinct bed slopes and sequences. *Billi et al*. [1998] studied various levels of bed organization in step-pool, pool-riffle, and mixed reaches in steep mountain streams following a large flood and attempted to define the recurrence interval of significant organizational events.

Changes in channel morphology following large sediment inputs have been demonstrated in several regions. *Lisle* [1982] showed a decrease in pool depths following a large flood and associated channel aggradation. *Madej and Ozaki* [1996] quantified the decreases in both pool depth and frequency associated with a sediment pulse. A debris flow in a third-order mountain stream resulted in a reach with short, disordered channel units and decreased channel complexity [*Lamberti et al*., 1991]. *Beschta* [1984] documented changes in channel morphology due to increased sediment loads in both New Zealand and Oregon. Following a dam break flood, *Pitlick* [1993] documented aggradation that completely filled the channel downstream until subsequent flows eroded most of the sediment. Although these studies have documented changes in sediment flux and pool characteristics, they have not specifically addressed how bed forms reorganize in response to a sediment pulse.

4. Data Sources

The conceptual model presented in Figure 1 was evaluated using surveys of channels adjusting to sediment pulses. Figure 2 shows the distribution of study sites according to drainage area and stream gradient. Study reaches were selected to represent a range of channel conditions, as well as by what surveys were available for analysis. Table 1 lists characteristics of the study reaches. Several study reaches were located in Redwood Creek and its tributary Bridge Creek in the northern Coast Ranges of California, United States of America, which have had episodes of massive landsliding associated with large floods. Lost Man Creek, a tributary of Redwood Creek and the site of a small dam removal project, was also included. *Madej* [2000] provides more detailed descriptions of these basins. To broaden the scope of study beyond the Redwood Creek basin, analyses based on results from other studies [*Lisle et al*., 1997; *Martinson et al*., 1986; *Maita*, 1991; *Kaufmann*, 1987; *Sutherland*, 1999] were used. These examples included an artificial channel undergoing a sediment pulse (a flume experiment), a stream impacted by a volcanic eruption (Smith Creek), two rivers affected by large landslides (Higashigochi and Navarro Rivers), and three streams in which a debris torrent had af-

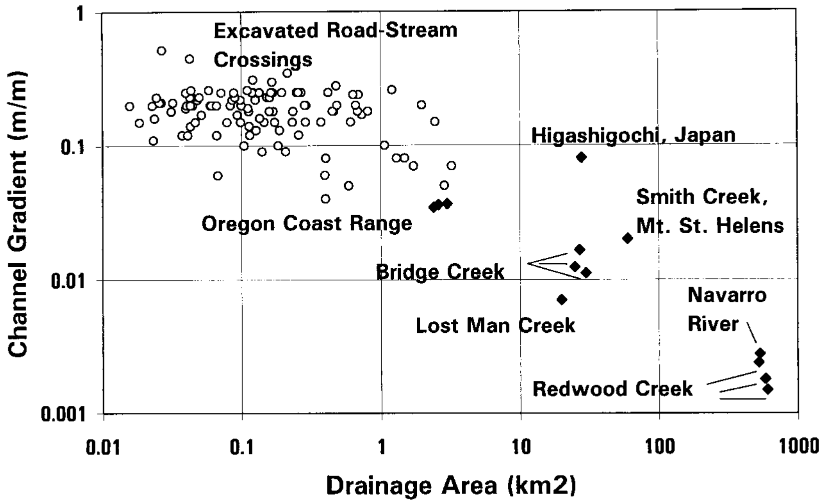

Figure 2. Distribution of study sites by drainage area and channel gradient.

fected the channel bed. All study reaches were relatively straight and had sinuosities of <1.2.

In addition to the high-order streams described above, channel development was documented in 124 steep, low-order stream reaches that were modified during a watershed restoration program. The focus of the restoration program was the removal of abandoned logging haul roads in order to restore natural drainage patterns and to reduce sediment production from the roads [*Madej*, 2001]. During road removal, road fill was excavated from stream crossings, and heavy equipment formed new channels through the former road prism. Sediment supply was high in these newly excavated channels, because of the availability of decompacted road fill. These newly formed channels ranged from 4 to 52% in slope and drained between 0.02 and 3.2 km^2.

Table 1 also lists the relative magnitude of the sediment pulse entering study reaches. The volume of sediment entering the channel was scaled as a ratio of sediment input to mean annual sediment transport through that channel reach (sediment pulse/annual sediment load). Because most streams did not have any associated sediment transport data, these are simply estimates based on regional trends and adjusted for drainage area. Although rough, these estimates are useful as indicators of the relative influence of a sediment pulse in a given channel network.

Longitudinal profile surveys formed the basis for much of the following data analysis. In Redwood, Bridge, and Lost Man Creeks, elevations of channel bed and water surface were measured using a self-leveling level and stadia rod. The spacing of survey shots averaged about 1/4 channel width, but it was not constant because all major breaks in slope were surveyed. Similar types of survey data were obtained for the Navarro River. Details of the method are given by *Madej* [1999]. The abundance of forcing elements (large woody debris, boulders, and bedrock outcrops) was noted qualitatively during the surveys. In the Higashigochi River and Smith Creek, longitudinal

Table 1. Channel Characteristics of Study Reaches

Study Reach	Drainage Area, km^2	Bank-Full Width, m	Channel Gradient, %	D_{50}, mm	Size of Sediment Pulse/ Annual Sediment Load
Little Lost Man Creek	9.0	8	2.50	50	no pulse (control)
Lost Man Creek	32	17	0.70	30	3
Upper Bridge Creek	25	23	1.24	30	3
Bridge Creek Canyon	27	12	1.66	60	3
Lower Bridge Creek	30	15	1.12	32	3
Redwood Creek at Weir Creek	520	60	0.24	22	25
Redwood Creek at Bond Creek	590	70	0.18	18	25
Redwood Creek at Elam Creek	600	110	0.15	15	25
Excavated road-stream crossings	0.1–3.2	1–2	4–52	~200	10
Higashigochi River	28	80	8.37	?	4
Smith Creek	30	70	2.21	?	10^4
Navarro River	530	60	0.28	22–32	0.3
Gwynn Creek	2.1	3.3	3.25	?	?
Cape Creek	2.9	4.2	3.65	?	?
Little Cummins Creek	2.6	3.4	3.61	?	?
Flume	NA[a]	1	1.00	0.57	10[b]

[a]NA is not available.
[b]Rate of input of sediment pulse was about 10 times the background bed load transport rate.

profile data were reconstructed from published survey plots, and the spacing of the survey shots is not known.

Data for analysis were also obtained from a flume experiment [*Lisle et al*., 1997]. The flume was 1 m wide and 160 m long, with a slope of 0.01 and a bed composed of a poorly sorted mixture of sand and fine gravel. First, the flume was run until a series of migrating alternate bars formed. Next, sediment was introduced over a section of flume 60–80 m downstream of the flume entrance. The flume was run at steady water discharge until the sediment accumulation dispersed and seemed to disappear. Bed elevations were measured at 0.5-m intervals down three longitudinal profiles, one located over the channel's centerline and the other two located half the distance to either bank. From these data I constructed thalweg profiles for four runs: one before the sediment input, two during the dispersal of the sediment wedge, and one at the conclusion of the flume run. The thalweg profiles covered the distance 80–140 m downstream of the flume entrance (downstream of the sediment introduction). Points in the longitudinal profile were linearly interpolated to form a data set with thalweg elevations spaced 0.3 m apart.

From the various sets of longitudinal survey data, distributions of residual water depths (the depth of water below the elevation of the downstream riffle crest [*Lisle*, 1987]) were calculated. Standard statistics (mean, median, and standard deviation) were calculated for the populations of residual water depths to document changes in channel structure. The standard deviation of water depths was used as an indicator of the variability of bed topography and was normalized by bankfull width in comparisons across channel types. In channels that did not display well-defined pools and riffles (Higashigochi River and Smith Creek), residual water depths could not be defined. In these cases a linear regression was fit through the bed profile, and the absolute values of residuals from the regression line were used as an indication of bed variability. The greater the standard deviation of the residual values, the greater is the variability of bed topography.

Channel organization was analyzed through spatial autocorrelation techniques that are described more fully by *Madej* [1999]. A spatial autocorrelation coefficient, Moran's I [*Legendre and Fortin*, 1989, Appendix A], documented the presence of regularity in the bed profiles, based on surveys of bed elevations. Two scales of spatial autocorrelation were studied. Short lag distance correlation represents the tendency of points close to one another in the channel bed to have similar elevations (i.e., neighboring points on a riffle crest). This scale of correlation is called "patch size" in landscape ecology, but here I refer to it as "bed relief patch size" to distinguish it from geomorphic "patches" defined by areas of uniform surface bed material. At longer lag distances, significant positive or negative spatial autocorrelation represents pairs of points that are more similar or more dissimilar from each other, respectively, than expected. A positive Moran's I represents regular spacing of channel units, such as between pool and pool or riffle and riffle. I refer to this scale of autocorrelation as "topographic regularity." In this paper, the correlation distance is expressed in terms of channel width to compare regularity across channels of different sizes. The strength of topographic regularity is indicated by the magnitude of the value of Moran's I.

A different approach was used in evaluating channel bed patterns in newly excavated channels at restored road crossings. Excavated stream crossings were surveyed with a clinometer to measure stream gradient, and the number and type of steps, if present, were inventoried. Steps were defined as a distinct break in slope greater than 0.3 m high with a flatter tread upstream (to distinguish them from cascades, which are more continuous runs of turbulent water). A total of 124 crossings, with a combined channel length of 2800 m, were used to analyze channel development.

To analyze changes in channel roughness values through time, U.S. Geological Survey water discharge records from five gauging stations in the Redwood Creek basin were used. Four stations were located on reaches of Redwood Creek that had aggraded in the past, and records from a gauging station in an undisturbed, unlogged tributary, Little Lost Man Creek, were used as a control. Gauging stations were purposely constructed in straight reaches to avoid complicated hydraulics due to bends, so roughness due to channel curvature is negligible. In addition, an analysis of sequential aerial photographs from these sites confirmed that other factors which could affect channel roughness, such as changes in channel planform, in-channel wood, or stream bank vegetation, were not important at the gauging stations.

Channel roughness was calculated using the Manning's equation:

$$n = \frac{(d^{2/3})(S^{1/2})}{v}, \tag{1}$$

where n is the roughness coefficient Manning's n, d is mean flow depth in meters, as an approximation of hydraulic radius, S is the water surface slope (m/m), and v is mean velocity (m/s). Values for these variables were obtained from gauging station records (U.S. Geological Survey (USGS) Form 9-207). Mean depth, not specifically listed on the USGS form, was calculated as "area/width." Water surface slope was assumed to be equivalent to the gradient of the channel bed surveyed at low flow. Channel bed profiles (surveyed at the gauging stations at low flow) showed no significant change in channel grade during the study period. Water surface slope may change with discharge, so this value is only an approximation of the true energy gradient at the gauging station. Because slope was used as a constant for all discharges at a given site, absolute values of roughness may be slightly off, but patterns in changing n values should still be valid.

Before examining time trends in roughness, roughness values must be adjusted for the influence of discharge. Extreme summer low flows, when water depth and bed particle size are equivalent, were not used in the analysis. To test whether trends in time were significant, I used the method described by *Helsel and Hirsch* [1992, p. 335]. A multiple regression analysis was constructed using both time and discharge as variables in the model:

$$\ln(\text{roughness}) = B_0 + B_1(\text{time}) + B_2[(\ln(\text{discharge})] + e. \tag{2}$$

The t statistic for B_1 tests for significant changes with time.

The size distribution of particles on the streambed at the gauging stations was determined through pebble counts [*Wolman*, 1954]. In this study, pebble counts consisted of a random selection of at least 100 particles from a riffle crest near each gauging station. The intermediate axis of each pebble was measured and tallied using Wentworth size classes. Cumulative size distribution curves were constructed, from which D_{50} (the median particle size) and D_{84} (the particle size that is coarser than 84% of the bed material) were calculated.

Table 2. Results of Thalweg Profile Analyses

Case Study	Study Reach	Time of Survey	Median Residual Water Depth, m	Standard Deviation, m/CW[a]	Bed Relief Patch, m/CW	Topographic Regularity, m/CW	Moran's I[b]
2	Higashigochi	before flood	NA[c]	0.0120	NA	NA	NA
		after flood, August 1982a	NA	0.0065	NA	NA	NA
		after flood, August 1982b	NA	0.0110	NA	NA	NA
		after flood, May 1983	NA	0.0140	NA	NA	NA
	Smith Creek	1983	NA	0.0033	NA	NA	NA
		1984	NA	0.0040	NA	NA	NA
		1995	NA	0.0064	0.2	6.3	0.11
3	Upper Bridge Creek	1986	0.08	0.0100	0.4	none	NA
		1995	0.08	0.0104	0.4	3	0.15
		1997	0.01	0.0052	0.2	none	NA
		1999	0.08	0.0104	0.2	none	NA
	Bridge Creek Canyon	1995	0.08	0.0125	0.5	none	NA
		1999	0.05	0.0175	0.8	none	NA
	Lower Bridge Creek	1986	0.01	0.0100	0.5	none	NA
		1995	0.03	0.0093	0.3	none	NA
		1997	0.00	0.0060	0.3	none	NA
		1999	0.01	0.0120	0.3	4.1	0.10
	Gwynn Creek	recent torrent	0.01[d]	0.0150	2	10	?
	Cape Creek	12-year old torrent	0.20[d]	0.0240	1	5	?
	Little Cummins Creek	120-year old torrent	0.17[d]	0.0240	1	5	?
4	Lost Man Creek	1 year postdam	0.08	0.0080	0.7	none	NA
		6 years postdam	0.19	0.0140	0.5	10.2	0.20
		control reach	0.23	0.0180	0.6	4.5	0.24
5	Navarro River	1995	0.34	0.006	1.5	8.6	0.36
		1996	0.75	0.013	1.2	8.9	0.50
		1997	0.71	0.013	1.2	9.3	0.58
	Redwood Creek at Weir Creek	1977	0	0.002	NA	NA	NA
		1983	0.36	0.006	1.1	4.3	0.09
		1986	0.46	0.009	0.8	3.6	0.09
		1995	0.51	0.010	0.6	3.3	0.09
		1997	0.32	0.007	0.5	2.9	0.08
	Redwood Creek at Bond Creek	1977	0	0.002	NA	NA	NA
		1983	0.16	0.004	0.6	5.3	0.24
		1995	0.24	0.006	0.5	8.0	0.18
		1997	0.22	0.004	0.3	8.2	0.07
	Redwood Creek at Elam Creek	1977	0	0.002	NA	NA	NA
		1983	0.06	0.002	0.6	3.4	0.17
		1986	0.14	0.003	0.5	2.6	0.07
		1995	0.31	0.004	0.4	2.5	0.18
		1997	0.17	0.003	0.3	2.4	0.11
6	flume	before sediment pulse	0.010	0.006	0.9	8.0	0.29
		during sediment pulse a	0.009	0.005	0.9	none	NA
		during sediment pulse b	0.009	0.004	0.9	none	NA
		after dispersion of sediment pulse	0.011	0.006	1.2	8.0	0.19

[a] CW is bank-full channel width.
[b] Moran's I is defined in Appendix A.
[c] NA is not available.
[d] Value indicates mean residual water depth.

There are several limitations to this study of channel organization. First, many of the surveys used in this analysis were conducted for reasons other than documenting responses to sediment pulses, and the surveys were not necessarily located in the best stream reaches to analyze such responses. The spatial resolution of the surveys used (spacing of survey shots about 1/4 channel width apart) was useful in detecting intermediate-sized bed forms, but the surveys cannot detect small-scale structures and organization. Likewise, because the survey transects were generally 20–30 channel widths long, longer-scale organization, if present, was not detectable by these surveys. Surveys were not available across the full range of channel gradients or for streams with wide valleys, survey resolution was unknown for some streams, and only sediment pulses that could be easily mobilized by a stream were studied.

5. Results

Trajectories of changes in channel structure and organization following sediment pulses (Figure 1) were evaluated by measuring four aspects of channel bed topography: median residual water depth, the standard deviation of distributions of water depths, bed relief patch size, and topographic regularity (Table 2). Bed forms in steep channels dominated by step morphology were analyzed using the number, spacing, and height of steps. The following discussion describes responses to sediment pulses in channels ranging in channel gradient from >20% to 0.15%.

5.1. Case 1: Development of Steps in Excavated Road-Stream Crossings

The first case involves the development of steps and pools in artificially manipulated channels. Since 1978, hundreds of kilometers of abandoned roads have been removed in Redwood National and State Parks, California. A major part of the road removal process is to excavate culverts and road fill from road-stream crossings using heavy equipment. When culverts and the overlying fill material are removed, the equipment operators form a smooth channel bed connecting the channel upstream of the old road bed to the channel downstream of the road prism. Although this newly excavated channel bed is in the same general location as the channel bed before the road was constructed, it is not necessarily in the same alignment or at the same elevation as the original streambed. The newly excavated channels can be considered to have undergone a sediment pulse because a large supply of poorly sorted road fill material that remained in the channel bottom and formed the new channel banks was suddenly available for reworking by the stream.

Following excavation, channels adjusted by varying degrees of incision, head cutting, bank erosion, and transport of sediment [*Madej*, 2001]. As a stream incised a freshly excavated crossing, fine-grained road material was eroded and transported, leaving coarser material in the channel bed. Depending on the particle size and stream gradient, these coarse particles became organized into steps. In some locations, previously buried boulders, tree roots, and wood were exhumed and formed steps. Nevertheless, in most cases the new channel was not simply an exhumation of the preroad channel because of the degree of disturbance in these channels during the original road construction and later in the road removal process.

Photographs taken immediately after the excavations provide evidence that most new channels were smooth and sloped evenly through the remaining road fill material. Moderate floods (3- to 5-year recurrence intervals) occurred in 1983 and 1986, and a 12-year flood occurred in 1997. By the time the channels were mapped in 1997 and 1998, many steps had developed; 38% were formed dominantly by wood, and 62% were formed by cobbles or boulders. The average step height (0.5–0.6 m) was similar for both wood and boulder steps and was about twice the size of the dominant bed particle.

The frequency of steps (number of steps per 30 m of channel length) was computed for each excavated crossing. The initial step frequency was assumed to be negligible, based on an analysis of photographs taken immediately after excavation, and measured step frequencies of <2 steps/30 m were considered to be within probable detection levels. Any step frequency of ≥2 steps/30 m was considered significant development of bed variation following stream-crossing excavation. According to this definition, 62% of the excavated crossings showed significant step development through time, and in these crossings mean step frequency was 5.3 (±2.6) steps per 30 m of channel length.

The relationship of step spacing and channel gradient was compared to that of another step-pool stream, Rio Cordon [*Billi et al.*, 1998] (Figure 3). Although many of the excavated crossings fell within the range of Rio Cordon, many others had steps spaced significantly farther apart. The fact that there were fewer, more widely spaced steps in excavated crossings than in a natural step-pool stream is consistent with the idea that the frequency of organizing flows in step-pool channels is low and that the organization of the channel bed into regularly spaced steps following road restoration will probably take several decades. At the time of the inventory in 1997 and 1998, steps were still randomly distributed with no characteristic step length, but channel beds showed increased bed variability since excavation, thus fitting the general conceptual model displayed in Figure 1. Eventually, after a period with several organizing flows, steps may become more regularly spaced, as the channel adjusts hydraulically to the size and type of bed material available in the channel.

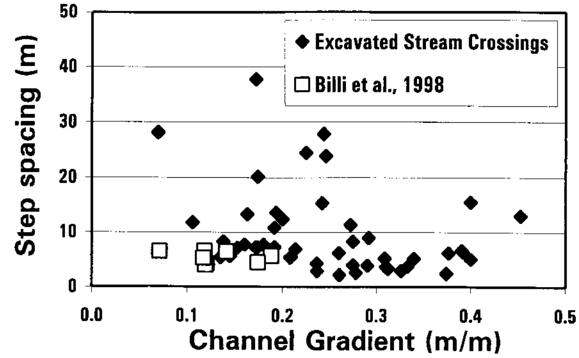

Figure 3. Graph of step spacing against channel gradient in the Rio Cordon and in excavated road-stream crossings.

The development of steps is an important component of energy dissipation in these excavated crossings. In the 124 crossings the average percent of total elevation drop due to steps (which is assumed to be proportional to the percent of energy dissipated) was 35%. Although 62% of the channels in the present study showed step development and thus higher roughness values than in freshly excavated channels, they have not yet attained the roughness values typical of well-developed step-pool channels [*Wohl et al.*, 1997]. Plunge pools downstream of the steps in the present study were shallow (<0.1 m) and may scour deeper with time, leading to further energy loss due to turbulence and hydraulic jumps over steps.

5.2. Case 2: Sediment Pulses in Steep (2–8% Grade) Channels

Few studies document channel changes in steep channels following sediment pulses. In a steep gravel bed river in Japan, the Higashigochi, *Maita* [1991] surveyed several longitudinal profiles following a pulse of sediment caused by landslides. This river has the highest stream power (based on "drainage area times channel gradient") of the sites considered in this paper. In August 1982 a flood accompanied by extensive landsliding raised the streambed from 3 to 8 m, and this sediment was subsequently eroded from the reach. By May 1983 the channel bed had returned to its former elevation. Almost no woody debris was observed in the surveyed reach, which was steep (average gradient is ~8%) and did not have a pool-riffle morphology. Initially, roughness elements were primarily boulders within the coarse gravel bed, and many of these were buried following the sediment pulse. Table 2 lists the standard deviations of the regression residuals based on the longitudinal profile surveys, which are an indication of bed variability. Immediately following the sediment pulse, the standard deviation decreased and subsequently increased through time (Table 2).

The first survey of August 1982 represents the peak of the deposition, whereas the second survey of August 1982 followed the recessional flow when the river had partially eroded the flood deposits [*Maita*, 1991]. By May 1983 the river had eroded down to its preflood level, and the standard deviation of residuals had increased to slightly above the preflood level. These results support the conceptual model presented earlier of high sediment loads initially leading to decreased heterogeneity, with an increase in bed variation through time. The resolution of these surveys was too coarse to examine trends in spatial autocorrelation.

Another example of response in steep channels following a sediment pulse is from Mount St. Helens, Washington, which erupted in 1980 and generated a lateral blast and extensive lahars in the Lewis River drainage basin. The newly deposited bed material was very poorly sorted and ranged from small bits of pumice to boulders. A tributary of the Lewis River, Smith Creek, was surveyed several times following the eruption, and the channel exhibited up to 10 m of aggradation [*Martinson et al*., 1986; U.S. Geological Survey, unpublished data, 1996]. Although there were many downed trees in the channel following the eruption, the influence of woody debris on Smith Creek morphology was dwarfed by the volume of volcanic deposits that obliterated the preeruption channel. Even though surveys were conducted during a relatively low-flow period (1982–1995) when most floods exhibited a $<$5-year recurrence interval, these flows can be considered organizing flows because the fine-grained volcanic deposits were readily mobilized. Surveys were analyzed in a similar manner to the Higashigochi River data set (Table 2).

The general trend in Smith Creek is of increasing bed variability through time. Cross-sectional surveys [*Martinson et al*., 1986] show a concomitant decrease in mean bed elevation during this period as sediment was transported out of the reach. The resolution of the early surveys was too coarse to examine trends in spatial organization, but weak regularity had appeared by 1995. More detailed surveys could certainly shed more light on the development of channel organization in this system, which is responding to an extreme increase in sediment load from the eruption. The flood of record ($>$40-year return interval) occurred in 1996, and the USGS is currently researching the effects of recent high flows on channel morphology.

5.3. Case 3: Channel Reorganization Following Debris Torrents

In 1997 a debris torrent that originated from a road fill failure deposited about 15,000 m^3 of sediment and wood in Bridge Creek, a 30-km^2 tributary of Redwood Creek. Profile surveys had been conducted previously (in 1986 and 1995) in three reaches of Bridge Creek downstream of the torrent site. These surveys were repeated in 1997 and 1999 to document channel changes following this sediment pulse. Formation and destruction of debris jams downstream of the debris torrent site have caused secondary effects in sediment routing in this stream.

Table 2 reports the results of the surveys. The upstream reach of Bridge Creek, which was directly impacted by the debris torrent deposition, responded immediately with decreases in water depth and channel bed variation. The channel down cut through the debris torrent deposits rapidly (mean channel bed elevation had dropped $\sim$0.5 m by 1999, following the increase of 0.7 m in 1997). Although median residual water depth and standard deviation recovered quickly (by 1999 the values approached the predebris torrent levels), the pattern of regularly spaced bars that was present in 1995 was obliterated in 1997 and had not reappeared by 1999. Large woody debris in this channel, much of it deposited by the debris torrent, strongly affects channel morphology in this reach and contributes to the irregular bed form spacing by causing many forced pools.

Farther downstream, a canyon reach, which is narrow, steeper, and bouldery, showed little response to the release of sediment from erosion of the debris torrent deposit upstream and the release of sediment from a debris jam that broke upstream of the canyon. No bed form regularity was apparent either before or after the debris torrent. This is not surprising because the reach was characterized by boulders and bedrock outcrops, as well as large woody debris, both before and after the debris torrent. The possibility of exceeding the critical discharge to move such boulders to rearrange the bed is negligible.

The downstream reach of Bridge Creek, also with high woody debris loading, did not have well-developed alternate bars. This reach responded to the debris torrent in 1997 by both a decrease in water depth and bed variation. The lack of significant short and long lag distance autocorrelation in this reach in 1997 indicates that the channel bed topography was random and not organized into any regular bed forms. By 1999, water depth and channel variability had increased, bed relief patches had returned, and weak topographic regularity was apparent at a spacing of four channel widths. A newly formed debris jam upstream of this reach which trapped some sediment from the debris torrent may have assisted in the recovery of channel variation in this reach through the metering of sediment supply and transport. In-channel wood plays an important role in this stream, and as wood from the debris torrent is redistributed during future flows, channel morphology will likely continue to change.

Results from Bridge Creek are consistent with those from another study involving channel surveys in streams affected by debris torrents. *Kaufmann* [1987] substituted space for time and studied morphologic characteristics in three channels subjected to debris torrents of various dates. Channel beds were surveyed every meter in a 100-m reach in three channels: Gwynn, Cape, and Little Cummins Creeks, Oregon. Kaufmann showed that channel complexity increased with time because of the creation of transverse bars, glides, riffles, pools, and side channels. In Gwynn Creek, which had a torrent 2–3 years prior to the study, mean depth and standard deviation of bed elevations were low (Table 2), pools were infrequent, and roughness values were low. Water depth, standard deviation of water depths, pool frequency, and roughness values were significantly higher in both Cape and Little Cummins Creeks, which had not experienced torrents for $\sim$12 and $\sim$120 years, respectively. In terms of spatial autocorrelation, Kaufmann found short lag distance autocorrelation (bed relief patches) about two channel widths long in the newly disturbed channel (Gwynn Creek), which decreased to one channel width with time (Cape and Little Cummins Creeks). He also reported weak regularity in the profiles at ten channel widths in the more recently disturbed channel and at five channel widths in channels with older debris torrents.

Hogan et al. [1998] report similar trends in streams in British Columbia, where debris torrents are an important mechanism leading to woody debris jams in channels. Here they recognized that initial response to jam formation was a loss of stream channel complexity, less variable water depths, and more prevalent riffles. In some cases, channels became braided. In more recently disturbed channels the distance between stable riffle-pools was greater than in channels with older disturbances. Through time, pools became more extensive, a single-thread channel developed, previously buried large wood was exhumed, and channel morphology became more complex.

5.4. Case 4: Reorganization of a Channel Bed Following Dam Removal

In 1989 a 3-m-high dam was removed from Lost Man Creek, a 32-km^2 tributary of Redwood Creek, and roughly three fourths of the 4000 m^3 of sediment stored upstream of the dam was also removed. Heavy equipment excavated a new straight channel through the remaining sand and gravel in the bed at an elevation ~2 m lower than the channel bed that existed when the dam was in place. The remaining accumulation of loose, easily mobilized sediment upstream of the old dam site can be considered a sediment pulse because it represents a large sediment supply suddenly available for reworking by the stream. Longitudinal profile surveys were conducted at the dam site in 1990 and 1996 to document channel development in this area. In addition, a 1990 survey of Lost Man Creek upstream of the dam-influenced reach was used as a control to evaluate what the channel condition may have been without the influence of a dam.

Following dam removal and one high flow season (1990), Lost Man Creek at the dam site exhibited low values of median residual water depth and standard deviation when compared to the control reach located upstream of the dam influence (Table 2). No regularly spaced bed forms had appeared. By 1996 the depth and variability of the channel bed had increased. Bed relief patch size had decreased, and topographic regularity had formed at a spacing of 10.2 channel widths. Because of previous logging activity along this stream as well as the dam construction, the supply of in-channel wood is low, and so the influence of large wood on channel morphology is also low in this reach. In contrast, the control reach had deeper water and higher bed variability and showed strong topographic regularity spaced 4.5 channel widths apart. This suggests that the dam-affected reach will evolve to a more complex channel with shorter wavelength features as it becomes better organized through time.

5.5. Case 5: Reorganization in Low-Gradient, High-Order Streams Following Landslides

Three reaches of Redwood Creek were studied intensely from 1977 (immediately following a large sediment input from a flood and associated landsliding) to 1997 [*Madej*, 1999]. Results from these longitudinal profile surveys are reported in Table 2. In general, water depth and bed variability increased with time, and bed relief patch size decreased with time since the 1975 flood and associated sediment input. In January 1997 a flood renewed aggradation in the three reaches, causing a decrease in water depth and variability measured in the 1997 surveys. Although the three reaches of channel display alternate bars, bar spacing is not completely regular. Topographic regularity is less than the five-to-eight channel width spacing frequently reported in the literature. This probably reflects the influence of other forcing mechanisms in the channel, such as channel bends, bedrock outcrops, boulder deposits, large wood, etc., that are present in these reaches. Forcing elements are most common in the upstream reach, Redwood Creek near Weir Creek, which showed the weakest regularity.

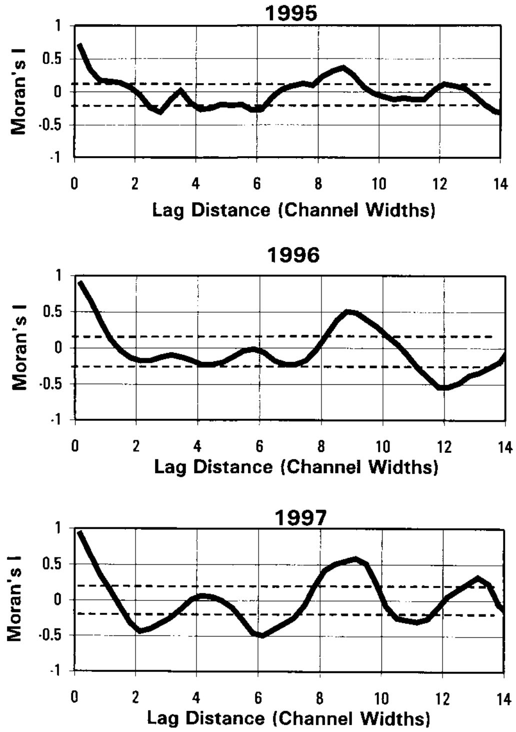

Figure 4. Correlograms based on Moran's *I* spatial autocorrelation coefficient for the Navarro River. The horizontal dashed lines represent the 95% confidence intervals for Moran's *I*. Points plotted above the line show lag distance with significant positive correlation.

In the Navarro River basin a single landslide contributed about 60,000 m^3 of sediment to the channel in 1995 and dammed the river. Subsequent surveys [*Sutherland*, 1999] showed several meters of downcutting in the channel at the landslide site. My analysis of those surveys shows results consistent with the other rivers in this study (Table 2). Median water depth and standard deviation increased with time since the sediment pulse. Although the spacing of topographic regularity hardly changed (Figure 4), the strength of the regularity increased (Table 2). The fact that the alternate bar structure was not eliminated after the sediment pulse may be due to the relatively small size of the pulse compared to the total annual sediment load of the Navarro River (Table 1) or because much of the landslide material abraded rapidly and was transported as suspended sediment [*Sutherland*, 1999].

5.6. Case 6: Sediment Pulse Under Controlled Channel Conditions (Flume Experiment)

The previous cases of channel organization involved forcing mechanisms of channel development due to the presence of wood, bedrock outcrops, boulders, etc. to various degrees. To examine bed organization without such factors, data from a flume experiment with an introduced sediment pulse [*Lisle et al.*, 1997] were analyzed. Median residual water depth and standard deviation initially decreased after the sediment input and then increased through time to their original values (Table 2). Bed relief patch size remained about the same through the run. Topographic regularity was spaced at eight channel widths

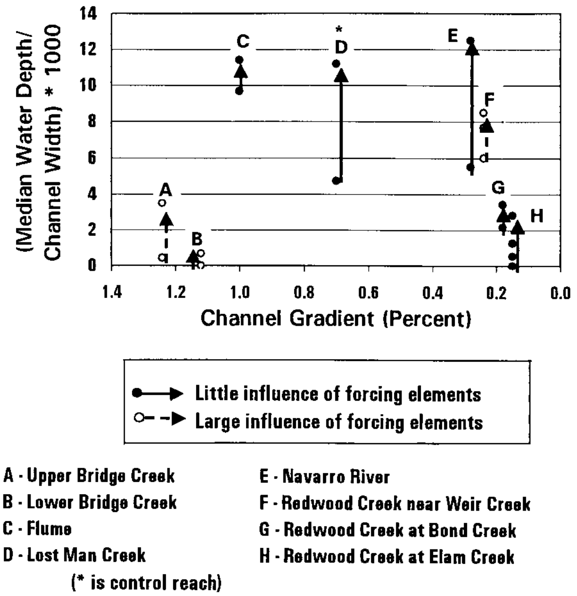

Figure 5. Trends in median residual water depth, scaled by channel width, with increasing time since sediment pulse.

before the sediment pulse but was destroyed during the pulse. Visually, the experimenters noted a change in bar spacing and migration of bar fronts that corresponded to these results (T. E. Lisle, personal communication, 1999). The experimenters stopped the flume run when it seemed that the sediment pulse had disappeared. At this point the bed had reorganized into distinct periodic bed forms, and topographic regularity reappeared at a spacing of 8 m.

In this case study the trajectory of stream channel evolution did not go through a phase of primary scale organization due to forced bars and pools (Figure 1), because the flume was devoid of such forcing mechanisms. In the absence of wood or bank irregularities the channel reestablished its original structure and organization. The processing of the sediment pulse in the flume was detectable through changes in channel organization, although not necessarily through tracking the movement of a distinct wave down the flume.

5.7. Trends in Channel Structure and Organization

The results of the case studies presented in section 5.1–5.6 support the concept of increasing channel structure and organization with time since disturbance. Figure 5 summarizes the trends in median residual water depth in streams with various channel gradients. In all study reaches in which a median water depth could be calculated, initial water depths were low following a sediment pulse, and depth increased with time following disturbance, whether or not forcing elements were abundant.

Figure 6 depicts trends in the variability of bed topography through time, scaled by channel width. In all streams, both with and without abundant forcing elements, variability of bed topography increased with time since a sediment pulse. This variability shows increasing development of complexity in a longitudinal direction. Most likely, complexity was also increasing in a cross-channel direction as well, but a three-dimensional view of channel structure and organization awaits further work. Bed variability also increased in the excavated stream crossings, but those results are not shown on this graph because a different survey procedure was used.

In the three Redwood Creek reaches, bed variability increased through time, but the increase was less in downstream, wider reaches (J and K in Figure 6) than a narrower, steeper reach (I). This may be because continued channel aggradation in the downstream reach indicates that the channel is still processing high sediment loads, and topographic variability may not have yet reached the stage of development of the upstream reach. Alternatively, the variability of bed topography may not increase in a downstream direction as quickly as channel width increases, and so the normalized value of variability is smaller in downstream reaches.

Figure 7 summarizes another indicator of channel organization, that of the bed relief patch size, scaled by channel width. If the distribution of bed elevations is random (the bed elevations of adjacent surveyed points are not related), the value of the index is about zero. Bridge Creek reaches that had a recent debris torrent and high wood loading (B and C in Figure 8) had short bed relief patches. In contrast, most channels responded to a sediment input by a smoothing of the channel (filling in the pools). In these cases, there were initially long stretches of channel bed with similar bed elevations, with correspondingly high values of bed relief patch size. As the channel developed more pools and complex bed topography, the similarity between adjacent points diminished, and the lag distance of significant autocorrelation decreased with time. It should be noted that a limitation in this type of study using channel surveys from many sources is the resolution of the surveys themselves. It is difficult to compare values of bed relief patch size without knowing the level of detail of the surveys because the spacing between survey shots determines the scale of feature and autocorrelation that can be detected. For this reason, profiles from the Higashigochi River and Smith Creek were not used in spatial autocorrelation analyses.

A fourth index used to describe channel structure and organization is topographic regularity, defined by regularly spaced features such as steps, pools, and riffles (Figure 8). Upper

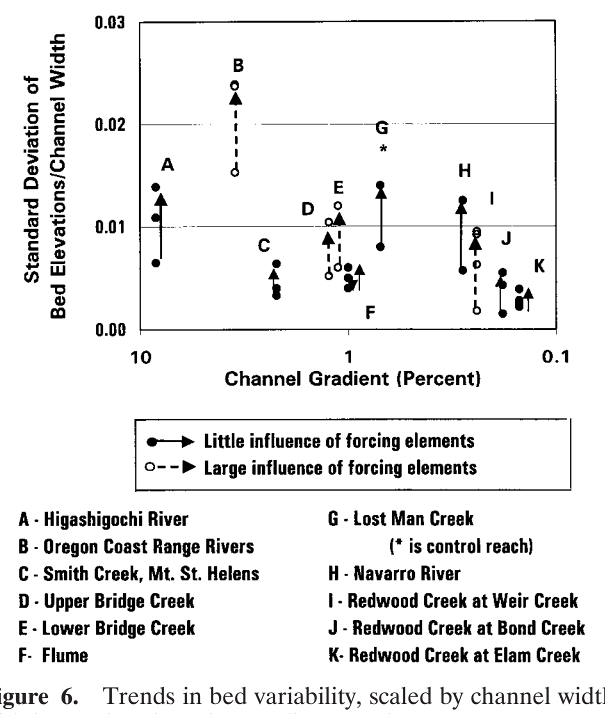

Figure 6. Trends in bed variability, scaled by channel width, with increasing time since sediment pulse.

Bridge Creek (B) showed no development of regularity 2 years after a debris torrent, perhaps because of the high wood loading in this channel, although water depth and variability had increased within 2 years. This observation suggests that the development of channel structure occurs more quickly than channel organization, which is consistent with the conceptual model introduced earlier. In lower Bridge Creek, the flume, and dam removal case studies (C, D, and E, respectively) the channel bed initially showed no regularity (a value of zero), but regularity of bed forms developed through time. Regularity at a scale of longer than seven channel widths only developed in channels without many forcing elements, consistent with the conceptual model introduced earlier. Topographic regularity barely changed in the Navarro River (F), but the strength of the regularity increased with time (Table 2). In this case the alternate bar forms downstream of the landslide were not destroyed by the sediment pulse; rather, fine-grained sediment from the landslide initially smoothed out the bed, and complexity redeveloped through time. In Redwood Creek at Bond Creek (H) the length scale of regularity increased, while the strength of the regularity decreased following large landslide inputs in this reach in 1997. In other Redwood Creek reaches and Oregon Coast Range rivers (A, G, and I) the length scale of regularity decreased with time as the channels became more complex.

In excavated stream crossings, not shown on Figure 8 because a different survey methodology was used, regularity in step spacing was developing in some of the channels but was still less than that reported in the literature for other step-pool channels. Crossings without wood or coarse clasts, or with boulders too coarse to transport, showed little or no step development. These results from many types of channels suggest that the degree of regularity and organization that develops in a channel depends on the time since disturbance (number of organizing flows), the size of sediment pulse, and the presence of forcing elements that can influence channel morphology.

5.8. Changes in Roughness Values Through Time

Both skin friction and form drag contribute to flow resistance in a channel. If a channel bed becomes coarser, the roughness due to boundary materials (skin friction) increases. In addition, as bed topography becomes more complex, form roughness should also increase. Increases in form roughness should have the most influence at low to moderate flows, whereas the roughness due to bed topography would be drowned out at very high flows, when mean depth of water is much greater than the vertical dimension of bed topography. The hypothesis tested here is that an increase in bed form development detected in the study reaches should manifest itself in a concomitant increase in Manning's n roughness coefficient at low to moderate flows. Water discharge measurements collected at five gauging stations along Redwood Creek were used to test this idea. Up to 25 years of water discharge measurement records (1972–1997) were examined, using the relationship defined by the Manning's equation (equation (1)).

Besides bed topography other factors, such as in-channel wood, meanders, abrupt changes in channel geometry, and hydraulic jumps, can theoretically contribute to flow resistance, but field observations show that these factors are not important at the gauging stations used in this study. The channels at the gauging stations are highly confined and have no floodplains; consequently, energy loss during overbank flow is not an important factor. Another consideration is that a sediment pulse may change the channel gradient in a reach, affecting the slope variable in the Manning's n calculation. However, to account for the observed changes in Manning's n (for example, at a discharge of 4 m^3/s at Redwood Creek near Blue Lake, n was 0.24 in 1972 and increased to 0.60 in the 1990s), channel gradient would have had to have increased by a factor of 8, from 0.0005 to 0.0040 m/m. Instead, surveys at the gauging stations from the 1970s, although limited, depict little or no change in slope during the study period.

Table 3 shows results of the analysis of Manning's n values through time. There was strong evidence that Manning's n increased through time after accounting for the effect of discharge (p values for time were all <0.09 for streams with a sediment

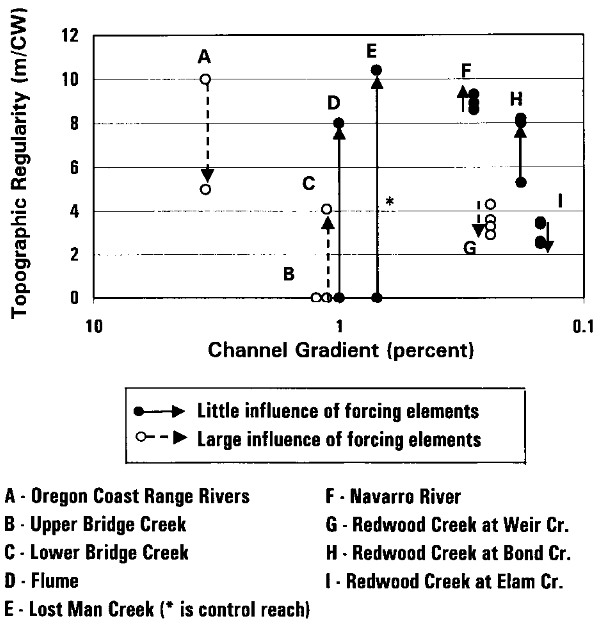

Figure 8. Trends in topographic regularity, scaled by channel width, with increasing time since sediment pulse.

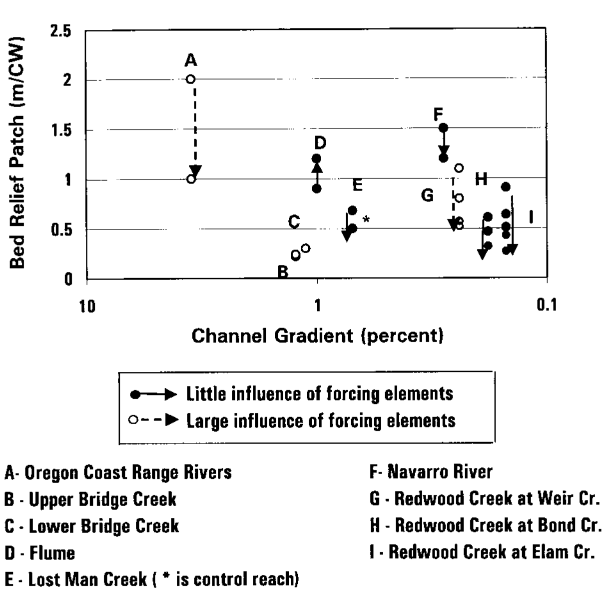

Figure 7. Trends in bed relief patch size, scaled by channel width, with increasing time since sediment pulse.

Table 3. Results of Multiple Regression Analysis Relating Channel Roughness (Manning's n) Values to Discharge and Date of Measurement

Gaging Station	Drainage Area, km^2	Channel Gradient, %	Period of Record	Measurements	Adjusted r^2	p Value for Discharge	p Value for Date
Little Lost Man Creek (control)[a]	9	2.5	1974–1988	133	0.48	0.0001[b]	0.7024
Redwood Creek near Blue Lake[c]	67.7	0.25	1972–1997	287	0.72	0.0001[b]	0.0001[b]
Redwood Creek at Panther Creek ($Q < 20$ m^3/s)[d]	389	0.28	1980–1988	54	0.83	0.0001[b]	0.0001[b]
Redwood Creek at South Park Boundary ($1 < Q < 30$ m^3/s)[e]	479	0.31	1975–1981	46	0.76	0.0001[b]	0.0001[b]
Redwood Creek at Weir Creek[f]	523	0.24	1977–1982	13	0.72	0.0004[b]	0.0922[b]

[a]Regression equation is ln Manning's $n = -0.916 - 8.19 \times 10^{-6}$ date $-$ 0.270 ln discharge.
[b]Variable is statistically significant at the 90% confidence level or better.
[c]Regression equation is ln Manning's $n = -3.57 + 9.93 \times 10^{-5}$ date $-$ 0.192 ln discharge.
[d]Regression equation is ln Manning's $n = -4.68 + 1.81 \times 10^{-4}$ date $-$ 0.383 ln discharge. Q is discharge.
[e]Regression equation is ln Manning's $n = -5.66 + 2.83 \times 10^{-4}$ date $-$ 0.277 ln discharge.
[f]Regression equation is ln Manning's $n = -5.69 + 4.26 \times 10^{-5}$ date $-$ 0.209 ln discharge.

pulse). In contrast, in Little Lost Man Creek, the control stream, the coefficient for time is not statistically significant.

The increases in flow resistance are consistent with either a coarsening of the channel bed through time or increased development of bed topography or both. To analyze the possible effects of increased particle size on roughness values, size distribution data based on pebble counts at the gauging stations were analyzed. These results show mixed trends (Figures 9a and 9b). Although pebble counts were centered on a given channel cross section each year, the total area of channel bed area sampled each year varied with different survey crews. This causes a problem in interpreting the results because, for example, a crew may have sampled a fine-grained eddy deposit one year but may not have included it in the sample area another year. Because the pebble counts were not necessarily conducted in a standardized manner, these results should be used cautiously.

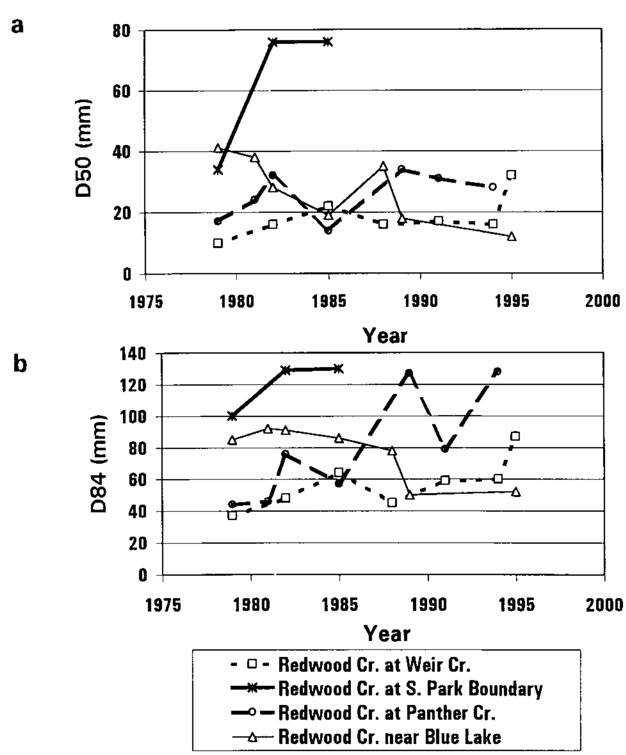

Figure 9. (a) Changes in D_{50} based on pebble counts at gauging stations. (b) Changes in D_{84} based on pebble counts at gauging stations.

Although the pebble count data have these constraints, some generalizations can be made. Redwood Creek near Blue Lake, which had a significant increase in roughness values through time, actually showed a decrease in surface bed particle size during the same period. Other factors, such as more pronounced bed relief, must be responsible for the increase in flow resistance. At the two other stations exhibiting increased roughness values through time (Redwood Creek above Panther Creek and Redwood Creek at Weir Creek), the channel bed surface did coarsen during the same period. Skin friction at the gauging stations has the vertical dimension of 30–130 mm (range of D_{84} in the channel bed). In contrast, the vertical dimension of bed topography (the depth of pools and heights of bar faces) is 1–3 m. Because the increase in flow resistance is seen at all stages, not just summer low flow, it is probably due to the development of bed topography as well as coarsening of the bed. This idea is supported by *Prestegaard* [1983], who showed that bar resistance accounts for 50–75% of total resistance in wide, low-sinuosity gravel bed streams, and *Parker and Peterson* [1980], who demonstrated the importance of bar resistance at low to moderate flows.

6. Discussion and Conclusions

The results presented in this paper support a conceptual model which proposes trajectories of channel recovery following sediment pulses in gravel bed streams across a range of channel gradients. A sediment pulse commonly diminishes channel structure, organization, and roughness. All the case studies presented in this paper followed this general pattern of decreased water depth and bed variability following a sediment pulse. Median water depth, bed variability, and development of forced pools and bars subsequently increased with the number of organizing flows. The model also suggested an increase in spatial organization of the channel bed through time. In most streams the length of bed relief patches decreased with time since a sediment pulse, indicating increased bed complexity. Topographic regularity developed at five to 10 channel widths in streams without many forcing elements and two to five channel widths in streams with forcing elements. One stream reach with high wood loading did not develop any regularity within 2 years of a debris torrent, which suggests that the time required to organize the channel into regularly spaced bed forms is greater than that needed to develop an increase in bed variability and structure.

Not all gravel bed streams are expected to perform as the conceptual model (Figure 1) proposes. For example, a low-variability bedrock-dominated stream that receives a sediment pulse may actually respond by an increase in variability with the addition of mobile material. Alternatively, if a sediment pulse causes bank erosion and associated tree fall, scour pools may increase in frequency. However, channel changes in many types of situations followed the trajectories proposed by the conceptual model. As a sediment pulse is processed by a stream system, the channel structure, organization, and roughness will change, and the manifestation of a sediment pulse may be reflected in changes in these elements. Future research will focus on examining such changes under a wider range of field conditions.

The development of regularity in many channels indicates the self-adjusting nature of channels. Fluvial processes, over time, were able to organize previously random hillslope inputs into regularly spaced bed forms. Regularity in streams with forcing mechanisms, such as large in-channel wood and bedrock outcrops, developed at a shorter spatial scale (two to five channel widths) than in streams without such forcing mechanisms (five to ten channel widths). Alternatively, if forcing elements are abundant, there may be an absence of any topographic regularity. Owing to the random distribution of bedrock outcrops, wood, and other obstructions, natural rivers will not display as regular a spacing as a artificial channel in a flume. The channel in the flume experiment exhibited self-organizing behavior that processed sediment inputs into a strong pattern of regularly spaced bed forms.

Roughness increased through time at several gaging stations in Redwood Creek during the same time period as channel bed relief was increasing. Because neither bed material size, channel gradient, nor riparian vegetation changed significantly at most sites, the development of bed forms probably plays an important role in the observed increase in flow resistance in the channel. The development of bed topography in a gravel bed river contributes to channel roughness and flow resistance and can influence bed particle mobility [*Church et al*., 1998].

Increased development of bed forms and increased variability of bed topography will influence the distribution and magnitude of secondary flows within a channel. An understanding of the spatially varied conditions of the channel bed may further the understanding of nonuniform flow. In addition, by identifying scales of longitudinal bed forms we may increase the understanding of sediment transport and alluvial sedimentation. Models of sediment routing could also benefit from a knowledge of channel organization, because how a channel processes its sediment load in one part of the channel network influences the timing and magnitude of sediment input to reaches farther downstream.

Sediment pulses are of concern to land managers because they influence sediment routing, sediment input to downstream reaches and reservoirs, and the stability of infrastructure features, such as bridge crossings and buried pipelines. The effect of sediment pulses on a channel may not necessarily be detectable as a distinct sediment wave moving downstream. Instead, this research showed that more subtle changes in channel structure and organization can also accompany sediment pulses.

An issue of concern in the Pacific Northwest regarding management of forested lands is the range of morphological diversity in natural systems. Under many ecosystem management activities, land managers attempt to replicate the range of natural variability in watershed processes. To date, variability in the magnitude and frequency of many processes has not been adequately quantified. The disturbance regime in a watershed affects the range of variability in channel conditions. The results presented here provide a basis upon which to compare variability of channel bed patterns in differently sized streams in response to sediment pulses. In addition, an understanding of the trajectories of physical recovery in disturbed stream systems will help predict the biological response to disturbances. Channel structure plays an important role in providing aquatic habitat, and the types and scales of roughness elements influence the distribution of microhabitats, benthic invertebrates, and channel complexity.

Geomorphic recovery, as proposed by *Wolman and Gerson* [1978], requires the attainment of a preexisting landform. Geomorphic work in stream channels not only entails a transport of material and a change in form but also a reorganization of bed forms. Through this research I suggest that measures of geomorphic effectiveness can include not only channel form but also the arrangement of such forms following a disturbance.

Appendix A: Definition of Spatial Autocorrelation Coefficient

Moran's I at distance class d is as follows:

$$I(d) = \frac{n \sum \sum w_{ij}(x_i - \bar{x})(x_j - \bar{x})}{W \sum (x_i - \bar{x})^2},$$

where x is residual water depth at points i and j in the channel and $\bar{x}$ is mean residual depth. All summations are for i and j varying from 1 to n, the number of data points, but exclude the cases where $i = j$. The w_{ij} take the value of 1 when the pair (i, j) pertains to the distance class d and are 0 otherwise. W is the number of pairs of points used in computing the coefficients for the given distance class. Moran's I may be positive or negative, with values usually ranging between -1 and $+1$. Moran's I compares values for pairs of points (residual water depths) at different distance classes (lag distance).

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M. A. Madej, U.S. Geological Survey Western Ecological Research Center, 1655 Heindon Road, Arcata, CA 95521. (mary_ann_madej@usgs.gov)

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