2
Channel Processes, Classification, and Response

David R. Montgomery and John M. Buffington

Overview

• This chapter discusses physical processes, classification, and response potential of channels in mountain drainage basins of the Pacific coastal ecoregion.

• A relatively simple set of physical processes leads to a wide variety of natural stream channels, the classification of which can guide recognition of functionally similar zones in mountain drainage basins. Different portions of mountain channel networks are dominated by different geomorphic processes and relationships between transport capacity (a function of discharge and boundary shear stress) and sediment supply (size and amount of material available for transport).

• Channel classifications use similarities of form and function to impose order on a continuum of natural stream types or morphologies. No single classification can satisfy all possible purposes or is likely to encompass all possible channel types.

• This chapter reviews geomorphological channel classifications and their use for systematizing channel morphology and physical processes for the purpose of assessing physical channel condition and response potential. Early classification systems tend to neglect the influence of woody debris or emphasize single scales of influences on channel morphology and processes. In contrast, a hierarchical approach to channel classification addresses different factors influencing channel properties over a range of spatial and temporal scales and is well suited for assessment of channel conditions and response potential in mountain drainage basins.

• The spatial distribution of reach types within a drainage basin influences the distribution of potential impacts and responses to disturbance. Alluvial channels with high transport capacities relative to sediment supply generally maintain their morphology while transmitting increased sediment loads; channels with lower ratios of transport capacity to sediment supply tend to exhibit greater morphologic response to increased sediment loads. Steep channels thereby act as sediment delivery conduits connecting zones of sediment production on hillslopes to more responsive lower-gradient channels.

• Consideration of channel bed morphology, confinement (the ratio of the width of the valley floor to the width of the bankfull channel), position in the channel network, and external influences (such as riparian vegetation and in-channel woody debris) can guide evaluation of channel condition and response potential in forested mountain drainage basins.

Introduction

Stream channels are important both as avenues of sediment transport that deliver eroded material from continents to the oceans and as environments for freshwater ecosystems that

River Ecology and Management
Robert Naiman & Robert Bilby, Editors

have economic and social significance (Chapter 23). Variability in sediment delivery, hydraulic discharge, and channel slope give rise to spatial and temporal variations in channel morphology and response. Over geologic time, channels respond to tectonic uplift, erosion of the landscape, and climate change. Over historical time, channels respond to changes in discharge and sediment supply from both land use and such extreme events as floods and droughts. Concern over such impacts on aquatic and riparian ecosystems, as well as human uses of fluvial systems, motivates assessment of channel change to evaluate past response to disturbance and to predict response to climate change or land use. The wide variety of channel types, adjustment of individual channels to local factors, and potential time lags between perturbation and response complicate the interpretation and prediction of changes in channel form and processes. This complexity fostered the development of classification schemes to guide identification of functionally similar portions of channel networks. This chapter discusses both the conceptual basis for understanding channel response and how channel classification can aid the study of watershed processes, assessment of channel condition, and evaluation of channel response in mountain drainage basins of the Pacific coastal ecoregion.

Channel Processes

Channels ranging in size from small ephemeral rivulets to large rivers exhibit a wide variety of morphologies, but share a number of basic processes. Over decades to centuries, channel morphology is influenced by both local and systematic downstream variation in sediment input from upslope sources (sediment supply), the ability of the channel to transmit these loads to downslope reaches (transport capacity), and the effects of vegetation on channel processes (Figure 2.1). Potential channel adjustments to altered discharge and sediment load include changes in width, depth, velocity, slope, roughness, and sediment size (Leopold and Maddock 1953).

A few key relationships describe the physics governing channel processes and illustrate controls on channel response. Conservation of energy and mass describe sediment transport and the flow of water through both the channel

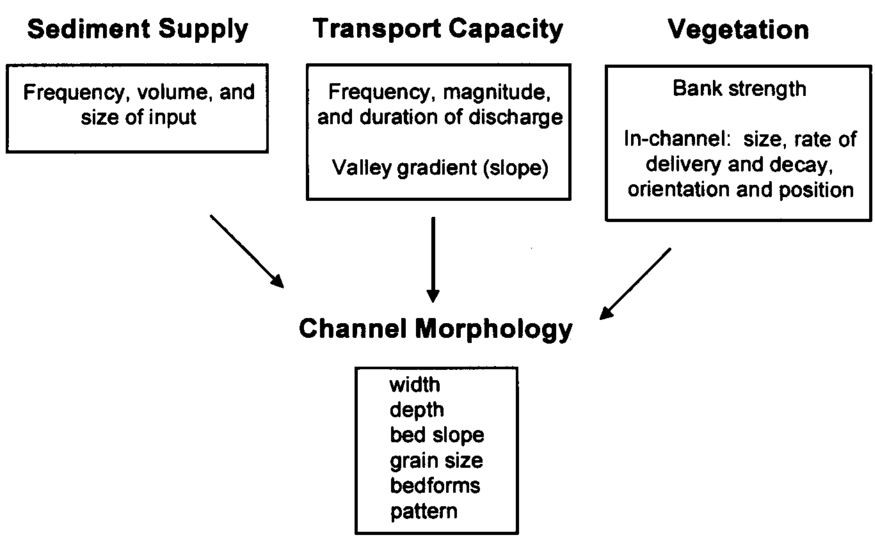


FIGURE 2.1. Decade- to century-scale influences on channel morphology include sediment supply, transport capacity, and direct and indirect effects of vegetation.

network and any point along a channel. Other relationships describe energy dissipation by channel roughness elements, the influence of boundary shear stress on sediment transport, and the geometry of the active transport zone.

Precipitation falling on a landscape moves downslope, causing erosion and maintaining channels. The frequency and magnitude of precipitation and the topographic relief onto which it falls provide the potential energy that drives these processes. Downslope movement of water converts this potential energy into kinetic energy, which is dissipated by friction and turbulence generated by the channel bed and banks. For a nonaccelerating fluid, the downstream gravitational force of the water ($mgS = \varrho ALgS$) is balanced by the shear resistance of the channel bed and banks ($\tau_0 LP$), and hence

$$\varrho ALgS = \tau_0 LP \tag{2.1}$$

where m is the mass of the water, g is gravitational acceleration, S is the water surface slope, ϱ is the fluid density, A is the channel cross-sectional area, L is the channel length, P is the wetted perimeter, and τ_0 is the total basal shear stress. Rearranging the force balance in equation (2.1) and solving for τ_0, the shear stress that flowing water exerts on the channel bed is

$$\tau_0 = \varrho gRS \tag{2.2}$$

where the hydraulic radius is $R \equiv \frac{A}{P}$. For natural channels with a width (W) much larger than mean flow depth (D), $R \approx D$ and thus $\tau_0 \approx \varrho gDS$.

The hydraulic discharge of a channel (Q) is defined as

$$Q = WDu \tag{2.3}$$

where u is the mean flow velocity, which depends on the fluid driving force and frictional resistance of the channel. Several empirical equations relate mean flow velocity to channel resistance

$$u = \frac{D^{2/3}S^{1/2}}{n} = C\sqrt{DS} = \sqrt{\frac{8gDS}{f}} \tag{2.4}$$

where n is the Manning roughness coefficient, C is the Chezy resistance factor, and f is the Darcy–Weisbach friction factor. The total roughness of a channel reflects the rate of energy dissipation and incorporates resistance offered by bed-forming particles, bedforms, and in-channel obstructions such as large woody debris (LWD). In general, the total boundary shear stress is related to the square of velocity ($\tau_0 \propto u^2$).

The basal shear stress acting on the channel bed drives sediment transport. The fraction of the total boundary shear stress available for sediment transport, defined as the effective shear stress (τ'), depends upon the amount of energy dissipation caused by in-channel roughness other than grain resistance (e.g., bedforms, LWD, channel bends). The critical shear stress (τ_c) represents the shear stress necessary to mobilize a given grain size (d_i)

$$\tau_c = \tau_c^*(\varrho_s - \varrho)gd_i \tag{2.5}$$

where ϱ_s is the sediment density and τ_c^* is the dimensionless critical shear stress (Shields 1936). Transport of bed material occurs at effective boundary shear stresses greater than or equal to the critical shear stress ($\tau' \geq \tau_c$). In gravel- and cobble-bed channels, the bankfull stage establishes channel morphology and accomplishes most sediment transport (Wolman and Miller 1960). The frequency of the bankfull discharge varies for different channels, but commonly occurs about every 1.5 years (Williams 1978).

Modes of sediment transport include both suspension of grains within the flow (suspended load) and rolling, sliding, and saltation of grains near the channel bed (bedload). Suspended load typically accounts for the majority of transported sediment, but bedload transport dominates channel morphology. Many bedload transport equations are based on the difference between applied and critical grain shear stresses

$$Q_b = k(\tau' - \tau_c)^n \tag{2.6}$$

where Q_b is the bedload transport rate (kg/s) and k and n are empirically determined values, with n typically being greater than one. The dependence of basal shear stress on flow depth, and thus discharge, indicates that a significant change in discharge directly influences sediment transport, bed stability, and scour.

Continuity requires that differences between the sediment supply (Q_s) and the bedload transport rate be accommodated by a change in the amount of sediment stored (S_s) within a reach

$$Q_s - Q_b = \Delta S_s \tag{2.7}$$

If more sediment enters a channel reach than it can transmit, then the amount of stored sediment must increase. Continuity further requires that the thickness of the active transport layer is related to the bedload transport rate (Carling 1987) as

$$D_s = \frac{Q_b}{u_b W \rho_s (1 - e)} \tag{2.8}$$

where D_s is the mean depth of scour, u_b is the average bedload velocity, and e is the bed porosity. Hydraulic discharge thereby controls the depth of scour defined by the thickness of the active transport layer.

Conceptual Models of Channel Response

The relatively simple set of channel processes outlined above results in a wide array of possible channel responses to changes in sediment supply, discharge, and external influences such as LWD flow obstructions. In response to changes in sediment supply or discharge, a channel may widen or deepen; change its slope through aggradation, degradation, or modified sinuosity; alter bedforms or particle size, thereby changing the frictional resistance of the bed; or alter the thickness of the active transport layer defined by the depth of channel scour. Drawing on both theory and empirical evidence, previous researchers developed conceptual models of channel response to changes in sediment load or discharge.

Gilbert (1917) hypothesized that the slope of an alluvial stream adjusts through erosion or deposition in order to transport the imposed load. Where the channel slope exceeds that necessary to transport the load, the channel incises and slope decreases, indicating a proportionality between bedload transport and channel slope

$$Q_b \propto S \tag{2.9}$$

With this reasoning, Gilbert (1917) anticipated and subsequently confirmed downstream channel incision in response to dam construction.

Lane (1955) later hypothesized that bedload discharge and sediment size adjust to hydraulic discharge and slope

$$Q_b d_i \propto QS \tag{2.10}$$

Based on this expression, Lane (1955) argued that corresponding changes in channel slope and sediment size accommodate changes in hydraulic discharge or bedload transport. Lane's expression provides a more complete model than Gilbert's, but neither expression accounts for the ability of a channel to change its basic geometry.

Schumm (1971) combined empirical relationships between discharge, bedload transport, and other descriptive morphological variables into general relationships for channel response that include channel geometry (i.e., W and D)

$$Q \propto \frac{WD\lambda}{S} \tag{2.11}$$

and

$$Q_b \propto \frac{W\lambda S}{Dp} \tag{2.12}$$

where λ is meander wavelength and p is sinuosity. Nunnally (1985) elaborated on Schumm's approach to include the median grain size of the bed surface (d_{50}).

Considering general relationships between sediment supply, discharge, and channel attributes (such as equations 2.2–2.8) additional factors can be incorporated into more comprehensive conceptual relationships for channel response in mountain drainage basins:

$$Q \propto \frac{WDQ_b D_s d_{50} n}{S_s S} \tag{2.13}$$

and

$$Q_s \propto \frac{WQ_b D_s S_s S}{Dd_{50} n} \tag{2.14}$$

Some variables in the conceptual relationships between discharge, sediment supply, and channel morphology summarized in equations (2.13) and (2.14) have thresholds of response, while others possess continuous response potential. Channel width and depth are related to discharge through equation (2.3). Discharge changes resulting in altered flow depth or velocity (2.3) have a direct influence on shear stress and hence both bedload transport (2.2, 2.5, 2.6) and depth of scour (2.8). Similarly, the control of discharge on shear stress directly affects bed surface grain size (2.5). In the absence of concurrent changes in sediment supply, increased discharge may cause higher bedload flux from a reach (2.2, 2.3, 2.6), resulting in depletion of sediment storage (2.7) or channel incision (2.3) and decreased slope; bank cutting and meander development caused by increased discharge can also decrease channel slope. Increased discharge resulting in higher ratios of transport capacity to sediment supply may be balanced by bed surface coarsening (Dietrich et al. 1989) and bedform development (Whittaker and Jaeggi 1982), both of which serve to increase resistance and stabilize the channel. In contrast, increased sediment supply can cause bed surface fining (Dietrich et al. 1989) and pool filling (Whittaker and Davies 1982), smoothing the channel bed and decreasing roughness. Bed surface fining caused by elevated sediment supply can result in greater bedload transport (2.5, 2.6) and consequently increased scour depths (2.8). Excessive sediment loading that exhausts the channel transport capacity can lead to bed aggradation, resulting in increased sediment storage (2.7) or decreased channel depth, which may, in turn, trigger alluvial bank cutting and channel widening if there are no concurrent changes in discharge (2.3); alternatively, aggradation may elevate channel slope.

Although conceptual response relationships, such as (2.13) and (2.14), allow prediction of the general direction of potential channel changes, specific responses often arise from some combination of altered discharge and sediment supply. Consequently, attribution of channel change to altered discharge or sediment supply often requires independent constraints on one of these factors. The predictions of (2.13) and (2.14) apply throughout channel networks, but the type and magnitude of response vary with site-specific channel processes and conditions. These relationships also illustrate a fundamental problem in predicting or reconstructing channel response: there are seven variables, but only two relationships—thus specific channel response is somewhat indeterminate. Fortunately, a great deal of accumulated experience relates to channel response.

Examples of Channel Change

An extensive literature on channel change highlights common responses and provides a large body of empirical evidence with which to develop and test conceptual models. Studies of channel change reveal a wide range of responses to changes in sediment supply and discharge. Increased sediment supply can induce channel widening and aggradation, decrease roughness through pool filling, and decrease bed sediment size. These responses are consistent with those predicted by (2.14). Increased discharge can cause channel widening, incision, and bed armoring, effects predicted by (2.13). Channel response to dam construction can involve a variety of effects due to changes in discharge and sediment supply, which covary.

Sediment Supply

Channel response to increased sediment supply depends on the ratio of transport capacity to the sediment supply. Significant aggradation, channel widening, bed fining, pool filling, or braiding occur where the amount of introduced sediment overwhelms the local transport capacity. Spatial variability in sediment supply may govern channel morphology in different portions of a drainage network.

Temporal variations in sediment supply also influence channel form. A classic study that illustrates progressive downstream aggradation and subsequent degradation in response to an episodic increase in sediment input is Gilbert's (1917) report on the effects of huge additions of hydraulic mining debris to rivers in the foothills of the Sierra Nevada of California from the early 1850s to the 1880s. Aggradation occurred sequentially throughout the downslope channel network as the mining debris was gradually transported through the system. Locally, channel aggradation approached 40m by the late 1870s (Whitney 1880). Subsequent reincision of the channels was still occurring just after the turn of the century and some channels continued to respond over one hundred years after hydraulic mining ceased (James 1991).

The 1964 floods in northern California and southern Oregon also illustrate morphologic response to increased sediment supply. Channel widths doubled at some gaging stations and channel beds aggraded as much as 4m (Kelsey 1980, Lisle 1982), except for channels with nonalluvial banks confined in narrow valleys (Lisle 1982). Kelsey (1980) estimated that a pulse of sediment originating in steep headwaters of the Van Duzen River migrated downstream at a rate of about 1km/yr. Lisle (1982) reported that pool filling decreased channel roughness and accelerated sediment transport within aggraded reaches. Both pool frequency and the mean size of bed material also decreased in response to aggradation (Kelsey 1980). Helley and LaMarche (1973) reported increased sediment storage in large gravel bars along channel margins and described evidence for a comparable response to prehistoric floods. Channel widening on the middle fork of the Willamette River, Oregon, in response to the 1964 flood reflected increased sediment delivery from hillslopes and disturbance of riparian vegetation (Lyons and Beschta 1983). Debris flows also scoured many steep channels to bedrock (Grant et al. 1984). Over twenty years later, significant flood-delivered material remained stored in low-gradient reaches where braiding continued to rework flood deposits (Sullivan et al. 1987). Such changes in sediment storage within a channel system can persist for decades, as sediment gradually mobilizes from the reaches in which it accumulated.

The South Fork Salmon River in central Idaho presents another example of impact and recovery from significant sediment inputs. Severe storms in the early 1960s following extensive logging and road construction dramatically increased sediment loads, resulting in pool filling, burial of gravels with sand, decreased bed roughness, and fining of the channel bed (Platts et al. 1989). A coincident decline in the fish population resulted in a moratorium on logging in the watershed, which reduced the sediment supply to impacted channels. Cross sections monitored over subsequent years showed progressive reincision, as pools were reexcavated and sand was transported out of spawning gravels (Megahan et al. 1980).

An important characteristic of channel response to increased sediment loads is that different portions of a drainage network may respond differently to a single disturbance. An excellent example of spatial patterns of channel response occurred as a result of a 100-year storm in the Santa Cruz Mountains that caused widespread landsliding in January, 1982 (Ellen and Wieczorek 1988). Debris flows and high discharges scoured many of the channels with gradients steeper than 10%, resulting in major sediment delivery to lower-gradient channels (Nolan and Marron 1988). Channel response in intermediate-gradient channels was variable, with significant local aggradation associated with landslide deposition (Nolan and Marron 1988). In these channels, sand filled many pools, buried riffles, and deposited in the interstices of coarse bed material (Coats et al. 1985). Substantial aggradation and overbank deposition also occurred along steep channels from routing of landslide debris (Nolan and Marron 1988). Later that winter, subsequent flows in steep- and intermediate-gradient channels scoured much of the aggraded sediment and redistributed it downslope. Pool filling and riffle burial persisted for a longer time in lower-gradient channels (Coats et al. 1985), illustrating a strong difference in the type and persistence of channel response at different locations in the drainage network.

Changes in sediment supply also influence the character of the channel bed. For example, Perkins (1989) studied the effect of landslide-supplied sediment on channel morphology in Salmon Creek in southwestern Washington. Based on considerations of the relationship between transport capacity and sediment supply, she argued that accelerated sediment delivery increased the amount of material stored in bedforms (expanding bar volumes at the expense of pool volumes) and decreased the average grain size in the reach. In her study area, elimination of landslide-supplied sediment resulted in a long-term decrease in the amount of material stored in the bed and a greater degree of bedrock control on bed morphology. Her study illustrates how channel form and sediment storage reflect the relative balance between sediment supply and transport capacity.

The size of bed surface material also responds to changes in sediment supply. In a series of flume experiments, Jackson and Beschta (1984) showed that increasing the amount of sand in a mixed sand/gravel bed increased gravel transport and scoured previously stable gravel riffles. They also showed that the median grain size (d_{50}) of the flume bed decreased with increased sediment transport. Dietrich et al. (1989) proposed that the degree of bed surface coarsening reflects the relationship between sediment supply and transport capacity. Their flume experiments showed that decreased sediment supply resulted in surface armoring and constriction of the zone of active sediment transport. Knighton (1991) reported that channel response to large inputs of fine sediment involved both a wave of aggradation and a general fining of bed material. After passage of such a wave, the channel substrate coarsened as the bed degraded toward its original condition.

Discharge

Changes in the magnitude and frequency of channel discharge may result from alteration of either the total precipitation falling in a watershed or from changes in runoff production and routing through the channel network (Chapter 3). Climate change provides the most direct precipitation-related impact on discharge in channel networks, but opportunities to monitor the influence of climate change on channels are rare. In contrast, the impact of land management on the discharge regime and morphology of stream channels is well documented. Watershed urbanization, for example, dramatically increases peak discharges because of increased impervious area, which increases the proportion of rapid surface runoff at the expense of infiltration (Leopold 1968). Channel response to these changes typically involves channel expansion through an increase in either channel width or depth. Hammer (1972) compared relationships between drainage area and channel width for urbanized and rural drainage basins in Pennsylvania and found significant channel widening in response to increased peak flows. He also found that large impervious surfaces (such as parking lots) directly connected to the channel network (via storm sewers) enhanced channel widening. Many others have reported significant channel widening and incision as a result of urbanization in both temperate (Graf 1975, Booth 1990) and tropical catchments (Whitlow and Gregory 1989).

Changes in watershed vegetation may affect the flow regime in downstream channels through changes in water yields, summer low flows, and peak flows. Paired watershed experiments indicate that forest clearance generally increases water yields (Bosch and Hewlett 1982), but in some regions, species which revegetate a cleared forest may have higher rates of evapotranspiration and thereby reduce discharges below original levels (Harr 1983). Although they may be very important biologically, changes in low-flow conditions are generally unimportant for morphological channel response. In contrast, increases in peak runoff caused by road construction (Jones and Grant 1996) or rain on snow events in clear cut areas (Harr 1986) may significantly affect channels because of the possible change in either the frequency or magnitude of the channel-forming discharge. Channel responses to high peak flows during rain on snow events include bank erosion, channel incision, and mobilization of both bedload and LWD (Harr 1981). These effects are similar to those occurring from natural large discharge events, but a change in their frequency could impact biological systems and reach-level sediment transport.

Dams

Dam construction changes both the discharge regime and sediment supply of downstream channels, resulting in channel incision, constriction or widening, and changes in channel substrate. Many studies document channel incision and bed surface coarsening immediately downstream of dams in response to sediment impoundment (Gilbert 1917, Williams and Wolman 1984). Tributary channels also may incise in response to mainstream channel incision through upstream knickpoint propagation from their confluence. Decreased discharge below a dam may cause narrowing of the active channel width (Leopold and Maddock 1953). Tributary sediment inputs downstream of dams can cause channel aggradation (Allen et al. 1989) and accumulation of fine sediment (Wilcock et al. 1996) because of decreased transport capacity resulting from dam construction. Channel-spanning log jams can also act like dams, causing upstream fining and downstream coarsening (Rice 1994).

Collectively, the case studies presented in this section illustrate that (2.13) and (2.14) provide a reasonable conceptual framework for examining channel response. However, differences in channel form and function affect the probability of specific responses to a given perturbation. Channel classification can aid interpretation and assessment of response potential by grouping functionally similar physical environments.

Geomorphological Channel Classification

Channel classifications use similarities of form and function to impose order on a continuum of natural stream types or morphologies. A voluminous literature on channel classification attests to the wide variety of stream morphologies. Each of the channel classifications in common use has advantages and disadvantages in geological, engineering, and ecological applications (Kondolf 1995), and no single classification can satisfy all possible purposes or likely encompass all possible channel types. This chapter reviews geomorphological channel classifications and their use for systematizing channel morphology and physical processes, and for assessing physical channel condition and response potential.

Past Classifications

Early geomorphological delineations of different channel types focused on broad criteria (Powell 1875, Gilbert 1877), but recent classifications include more detailed consideration of channel pattern, bed material or mobility, sediment transport mechanisms, position within the channel network, and various combinations of slope and valley characteristics. Most geomorphological classifications are designed for large floodplain rivers, although Schumm's (1977) general delineation of erosion, transport, and deposition zones provides a conceptual framework within which to couple channel type and channel response potential throughout mountain drainage basins.

Stream order. The concept of stream order proposed by Horton (1945), and later modified by Strahler (1957), remains the most widely used channel classification. In Strahler's system, the channel segment from the head of the channel network to the first confluence constitutes a 1st-order channel. Second-order channels lie downslope of the intersection of two 1st-order channels, and so on down through the channel network. Stream order correlates with channel length and drainage area, thereby providing an indication of relative channel size and position within a channel network. Although channel ordering is a useful conceptual and organizational tool, comparisons between channel networks can prove misleading because the order assigned to a channel segment depends on the criteria used to determine where 1st-order channels begin. Representations of the extent of channels in a given watershed vary on maps of different scales, and basin topology influences the size of channels classified as a particular order. Moreover, channel networks defined from blue lines on maps, the curvature of topographic contours, or a critical gradient or drainage area can differ substantially from the network identifiable in the field (Morisawa 1957). Aside from the tautology that higher-order channels tend to be larger, there is no inherent association of channel morphology and process with stream order.

Channel bed. The nature of the channel bed provides the basis for perhaps the most fundamental geomorphological classification of stream channels. Gilbert (1914) recognized that bedrock channels have a greater transport capacity than the sediment supply, whereas alluvial channels have a transport capacity less than, or equal to, the sediment supply. Gilbert further recognized that different portions of a channel network may be composed of different channel types and patterns. Henderson (1963) later recognized two alluvial channel types based on grain size and sediment mobility characteristics: "live bed" channels that are actively mobile at most stages, and "threshold" channels that exhibit significant mobility only during high flows. These different styles of bed movement are strongly correlated with grain size, and thus with the common distinction of sand- and gravel-bed channels.

Channel patterns and processes. Several channel classifications broadly characterize general differences in channel patterns and processes. Leopold and Wolman (1957) quantitatively differentiated straight, meandering, and braided channel patterns based on relationships between slope and discharge. Schumm (1977) classified alluvial channels based on dominant modes of sediment transport (i.e., suspended, mixed, or bedload transport) and recognized three geomorphic zones within a watershed: degrading headwater channels that are the primary source of sediment and water inputs, stable mid-network channels with roughly balanced inputs and outputs, and aggrading channels low in the network characterized by extensive depositional floodplains or deltas. Mollard (1973), and later Church (1992), classified floodplain rivers into a continuum of channel patterns related to differences in discharge, slope, sediment supply, and channel stability. Kellerhals et al. (1976) used an extensive list of descriptive features to characterize large alluvial rivers in terms of stream pattern, frequency of islands, bar type, and lateral channel migration. They further emphasized that channel form and processes depend on surficial geology, the frequency and magnitude of sediment and water inputs, the relationship of a channel to its floodplain and valley walls, and the history of geologic, climatic, and anthropogenic disturbance. Church and Jones (1982) subsequently presented a classification of bar types and patterns that explicitly relates channel morphology to gradient and the volume and size of sediment supply. These fundamental distinctions can guide general interpretation of channel condition and response potential in mountain drainage basins.

A more detailed channel reach classification developed by Rosgen (1994) recognizes 7 major and 42 minor channel types based on channel pattern, entrenchment, width-to-depth ratio, sinuosity, slope, and bed material size. Although Rosgen (1994) demonstrated that his major channel types exhibit distinct roughness coefficients and hydraulic geometry relationships, the classification is not process based; a lack of any explanation of the rationale underlying Rosgen's assessment of response potential for each minor channel type emphasizes this shortcoming. In contrast, Whiting and Bradley (1993) presented a process-based classification for headwater channels that associates channel morphology with the potential for debris flow impacts, channel substrate size, and processes and rates of fluvial sediment transport. Paustian et al. (1992) provided an example of a valley-scale classification emphasizing region-specific associations with channel morphology and processes.

The classifications discussed above serve a variety of purposes, but generally are incomplete for comprehensive channel assessments in forested mountain drainage basins due to either neglect of the influence of woody debris or an emphasis on a single scale of influence on channel morphology and processes. A hierarchical channel classification approach addresses these issues and is well suited for assessment of channel conditions and response potential in mountain drainage basins.

Hierarchical Channel Classification

A hierarchical approach to channel classification addresses different factors influencing channel properties over a range of spatial and temporal scales scales (Figure 2.2) (Frissell et al. 1986). A hierarchy of spatial scales that reflects differences in processes and controls on channel morphology includes geomorphic province, watershed, valley segment, channel reach, and channel unit (Table 2.1, Figure 2.3). Each level of this spatial hierarchy provides a framework for comparing channels at increasingly finer spatial scales.

Geomorphic Provinces

Geomorphic provinces consist of regions with similar land forms that reflect comparable hydrologic, erosional, and tectonic processes over areas greater than 1,000 km^2 (Table 2.1, Figure 2.3). Major physiographic, climatic, and geological features bound geomorphic provinces and impose broad controls on channel processes. Watersheds within a geomorphic province tend to share roughly similar relief, climate, and lithologic assemblages. General controls on channel processes and morphology are reasonably similar for most large watersheds within a geomorphic province; thus, their channels are potentially comparable in terms of relationships between drainage area, discharge, sediment supply, and substrate size. Environmental conditions or the legacy of climate history may impose similar general external constraints on channels within a geomorphic province. Channels in the Olympic Mountains geomorphic province, for example, have an abundant supply of extraordinarily large logs that profoundly influence channel morphology and dynamics. Although geomorphic provinces identify broad areas likely to host comparable watersheds, the concept remains too general for predicting specific channel attributes or responses. Geomorphic provinces do, however, provide a general context for investigating and interpreting channel processes, and therefore channel response.

Watersheds

Watersheds, or drainage basins, define natural systems for routing sediment and runoff into

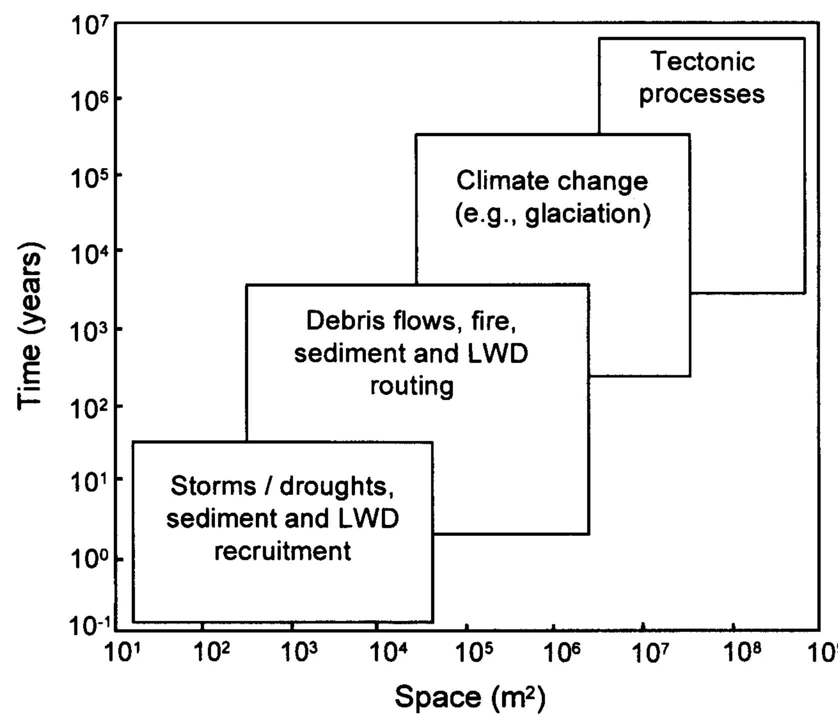


FIGURE 2.2. Range of spatial and temporal influences on stream channels.

and through channel networks (Table 2.1, Figure 2.3). While the term watershed classically refers to a drainage divide, contemporary usage also considers the drainage area upslope of any point along a channel network as a watershed. Although the appropriate scale of watershed-level classification ultimately is site specific, drainage basins of 50 to 500 km^2 provide practical units for examining the influence of watershed processes on channel morphology and disturbance regimes (Montgomery et al. 1995a). Although watershed-level classification differentiates major drainage basins within geomorphic provinces, large rivers can traverse several provinces. Classifying watersheds based on similar geologic and climatic history, lithology, and land use may highlight areas with similar controls on channel processes and identifies river systems as either well or ill suited for comparison. However, watershed level classification of channel networks neglects fundamental differences in sediment production and transport processes of finer-scale valley morphologies.

TABLE 2.1. Hierarchical levels of channel classification and associated spatial scales.

Classification level	Spatial scale
Geomorphic province	1,000 km^2
Watershed	50–500 km^2
Valley segment	10^2–10^4 m
Colluvial valleys	
Bedrock valleys	
Alluvial valleys	
Channel reaches	10^1–10^3 m
Colluvial reaches	
Bedrock reaches	
Free-formed alluvial reaches	
Cascade reaches	
Step-pool reaches	
Plane-bed reaches	
Pool-riffle reaches	
Dune-ripple reaches	
Forced alluvial reaches	
Forced step-pool	
Forced pool-riffle	
Channel units	10^0–10^1 m
Pools	
Bars	
Shallows	

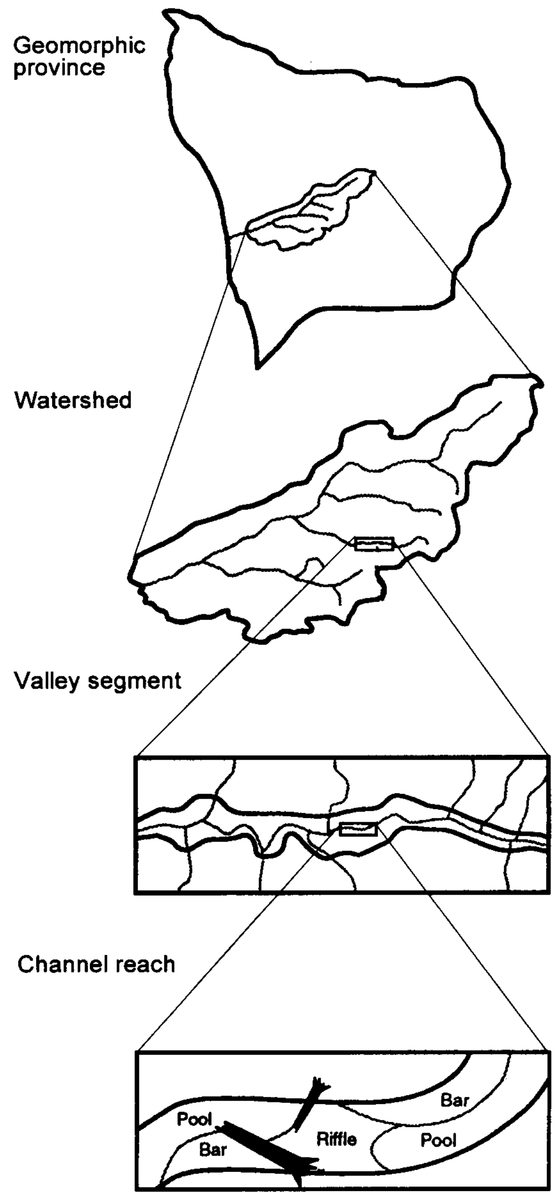


FIGURE 2.3. Geomorphic province, watershed, valley segment, channel reach, and channel unit (e.g., pools, bars, riffles) scales of classification illustrated for the Olympic Peninsula, Washington.

Valley Segments

Valley segments define portions of the drainage network exhibiting similar valley-scale morphologies and governing geomorphic processes at a scale of 10^2 to 10^4 m (Table 2.1, Figure 2.3).

In mountain drainage basins, valley segments are classified into colluvial, bedrock, and alluvial valley types based on valley fill, sediment transport processes, channel transport capacity, and sediment supply. A fourth type, estuarine valleys, are important links between terrestrial and marine environments. These divisions are similar to the valley segment classification developed by Frissell and Liss (1986) for the Oregon Coast Range. More elaborate region-specific valley segment classifications may help link channel classification with resource assessments (Paustian et al. 1992).

Colluvial valleys. Shallow and ephemeral fluvial transport in headwater valleys is relatively ineffective at transporting sediment delivered from surrounding hillslopes, resulting in accumulation of colluvial valley fills. Colluvial valley bottoms lacking evidence of a well-defined channel indicate insufficient hydraulic erosion to initiate and maintain a channel; these unchanneled valleys (regionally referred to as hollows, swales, or headwalls) extend upslope of the smallest channels in many soil-mantled environments. In steep landscapes, unchanneled valleys gradually accumulate colluvial soils transported from surrounding hillslopes and periodically deliver the stored sediment to downstream channels via debris flows. Hillslope sediment transport processes subsequently refill excavated hollows, resulting in a cycle of long-term accumulation punctuated by periodic catastrophic erosion (Dietrich and Dunne 1978). This cycle of hollow accumulation and erosion can take thousands of years (Reneau et al. 1990).

Channeled colluvial valleys downslope of hollows indicate the emergence of fluvial transport. Nevertheless, the influence of fluvial processes on colluvial valley form and incision is often secondary to transport by periodic debris flows in steep landscapes. In contrast, the maintenance of colluvial valleys in low-gradient landscapes requires incision by streams; extensive networks of colluvial valleys in low-gradient landscapes are likely a result of long-term climate change.

Bedrock valleys. Bedrock valleys typically are confined and lack significant valley fill. Narrow valley bottoms result in relatively straight channels, although deeply incised bedrock meanders may occur. Channel floors in bedrock valleys consist of either exposed bedrock or thin, patchy accumulations of alluvium. Insignificant sediment storage in bedrock valley segments indicates downstream transport of virtually all the material delivered to the channel, implying that transport capacity exceeds sediment supply over the long term.

Alluvial valleys. Channels in alluvial valleys transport and sort sediment loads supplied from upslope, but lack the transport capacity to routinely scour the valley to bedrock. Channels in alluvial valley segments may support either narrow or wide floodplains. Thick alluvial deposits in unconfined valley segments imply a long-term excess of sediment supply relative to transport capacity. Both the specific channel morphology and degree of confinement reflect the local channel slope, sediment supply, and hydraulic discharge.

Valley segment morphology is useful for distinguishing dominant sediment transport processes (fluvial versus mass wasting), inferring general long-term sediment flux characteristics (transport- versus supply-limited), and providing insight into the spatial linkages that govern watershed response to disturbance. Alluvial channels, however, exhibit a variety of morphologies, some of which appear functionally similar at the valley segment level, but which respond differently to similar perturbations in sediment load and discharge. Consequently, channel reach morphology often proves more useful than valley segment morphology for understanding channel processes and response potential.

Channel Reaches

Channel reaches exhibit similar bedforms over stretches of stream that are many channel widths in length (Table 2.1, Figure 2.3). In mountain drainage basins, channel reaches are classified into colluvial, bedrock, and alluvial reach types. These general reach types are briefly described below (for further detail see Montgomery and Buffington 1997).

Colluvial Reaches

Colluvial reaches typically occupy headwater portions of a channel network (Figure 2.4a) and occur where drainage areas are large enough to sustain a channel for the local ground slope (Montgomery and Dietrich 1988). Soil creep, tree throw, burrowing by animals, and small-scale slope instability introduce sediment into colluvial reaches. Intermittent flow reworks some portion of the accumulated material, but does not govern deposition, sorting, or transport of most valley fill because of low shear stresses (Benda 1990). Large grains, woody debris, bedrock steps, and in-channel vegetation reduce the energy available for sediment transport. Ephemeral, low discharges in colluvial reaches result in a poorly sorted bed with finer grain sizes than in downstream alluvial reaches. Episodic transport by debris flows accounts for most of the sediment transport in steep headwater channels (Swanson et al.1982).

Bedrock Reaches

Bedrock reaches exhibit little, if any, alluvial bed material or valley fill, and are generally confined by valley walls and lack floodplains (Figure 2.4b). Bedrock reaches occur on steeper slopes than alluvial reaches with similar drainage areas (Montgomery et al. 1996), an observation supporting Gilbert's (1914) hypothesis that bedrock reaches lack an alluvial bed due to a higher transport capacity than sediment supply. Steep headwater channels in mountain drainage basins may alternate through time between bedrock and colluvial morphologies in response to periodic scour by debris flows (Benda 1990). In general, bedrock reaches in low-gradient portions of a watershed reflect a high transport capacity relative to sediment supply, whereas those in steep debris-flow-prone channels also reflect recent debris flow scour.

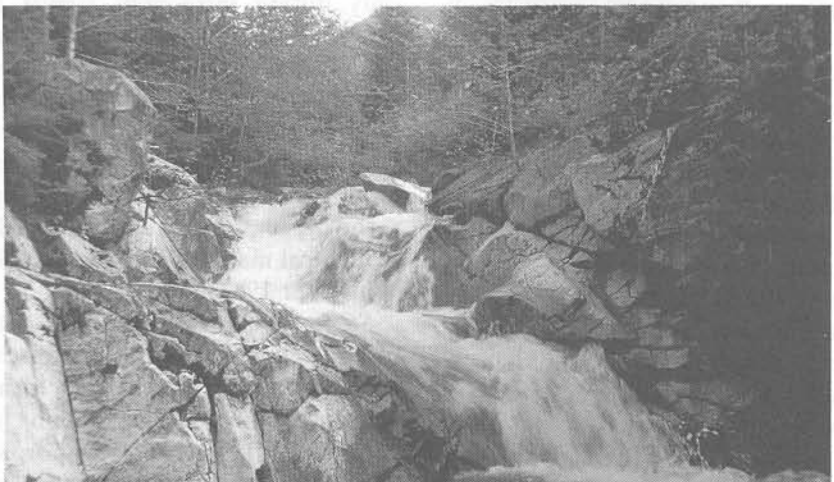
Free-Formed Alluvial Reaches

Alluvial reaches exhibit a wide variety of bed morphologies and roughness configurations that vary with slope and position within the channel network. Montgomery and Buffington (1997) suggest that the ratio of transport capacity to sediment supply controls the roughness configurations that shape alluvial reach morphology, which can be categorized into five free-formed alluvial channel reach types: cascade, step-pool, plane-bed, pool-riffle, and dune-ripple. Transitional morphologies also occur, as this classification imposes order on a natural continuum.

Cascade reaches. Cascade reaches occur on steep slopes with high rates of energy dissipation and are characterized by longitudinally and laterally disorganized bed material, typically consisting of cobbles and boulders confined by valley walls (Figure 2.4c). Flow in cascade reaches follows a tortuous convergent and divergent path over and around individual large clasts; tumbling flow over these grains and turbulence associated with jet-and-wake flow around grains dissipates much of the mechanical energy of the flow (Peterson and Mohanty 1960, Grant et al. 1990). Large particle size relative to flow depth make the largest bed-forming material of cascade reaches mobile only during infrequent events (i.e., >25 yr, Grant et al. 1990). In contrast, rapid transport of the smaller bedload material over the more stable bed-forming clasts occurs during flows of moderate recurrence interval. Bedload transport studies demonstrate that steep, alluvial, mountain streams are typically supply limited, receiving seasonal or stochastic sediment inputs from local mass wasting events (Griffiths 1980, Whittaker 1987).

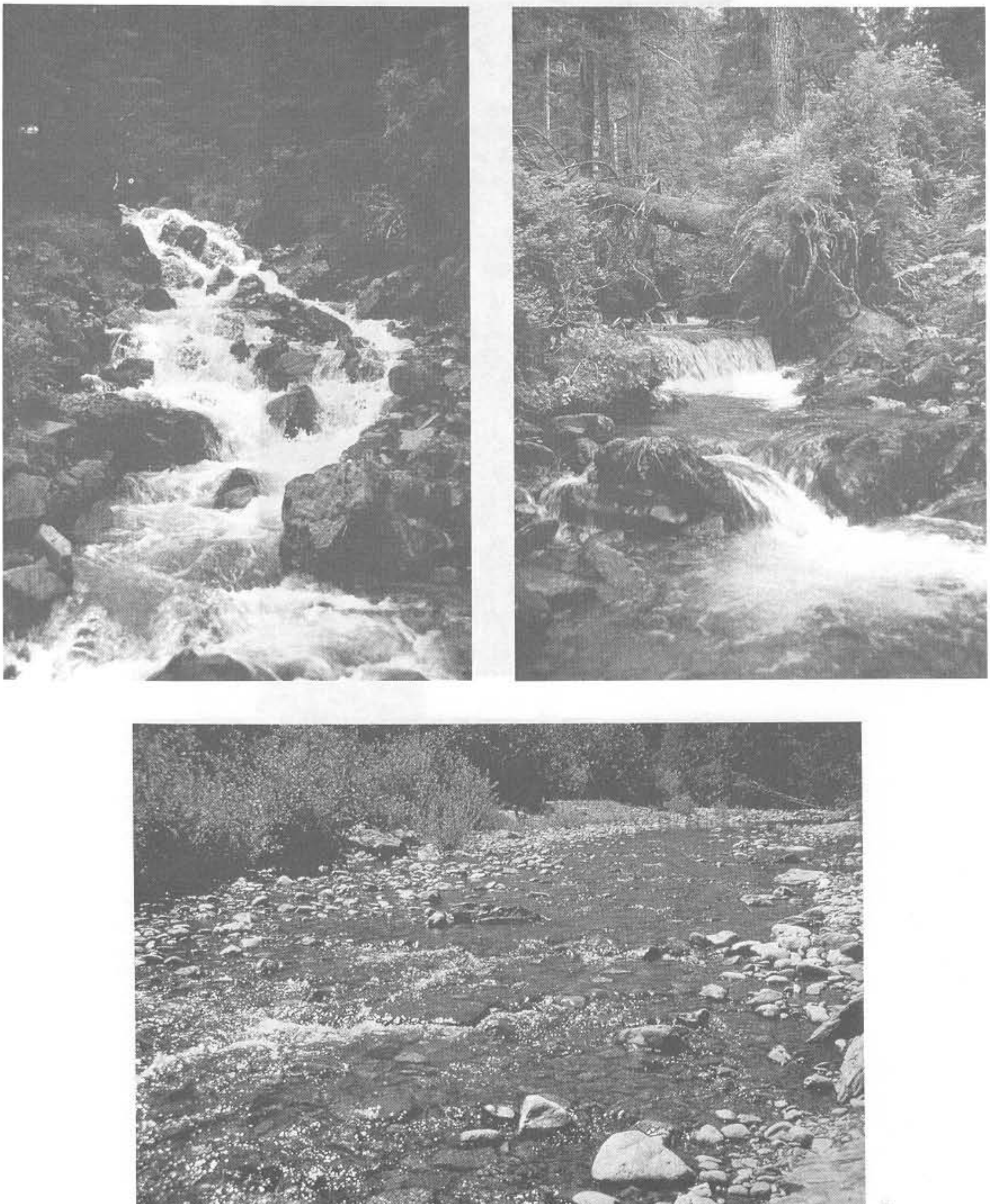
Step-pool reaches. Step-pool reaches consist of large clasts organized into discrete channel-spanning accumulations that form a series of steps separating pools containing finer material (Figure 2.4d). The stepped morphology of the bed results in alternating turbulent flow over steps and tranquil flow in pools. Channel-spanning steps provide much of the elevation drop and roughness in step-pool reaches (Whittaker and Jaeggi 1982). Step-forming clasts may be viewed as a congested zone of large grains that causes increased local flow resistance and further accumulation of large particles (Church and Jones 1982), or as

a



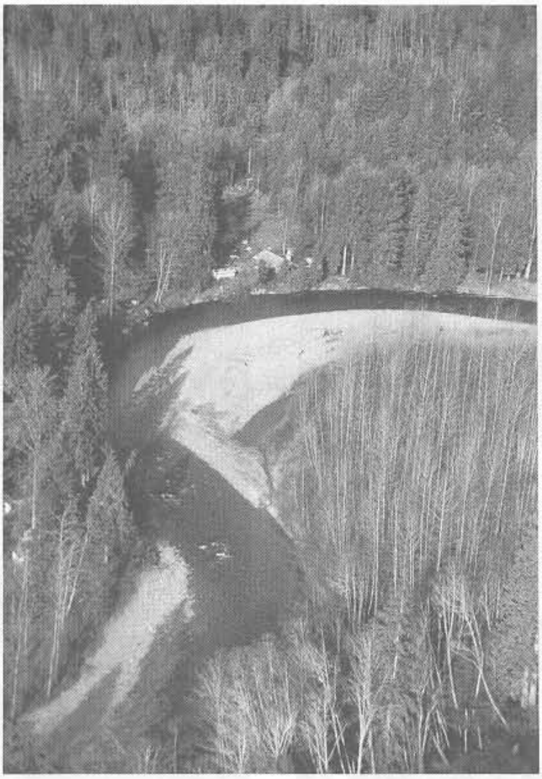
b

FIGURE 2.4. Photographs of reach-level channel types: (a) colluvial; (b) bedrock; (c) cascade; (d) step-pool; (e) plane-bed; (f) pool-riffle; and (g) dune-ripple.



c d e

FIGURE 2.4. *Continued.*



f

g

FIGURE 2.4. *Continued.*

macroscale antidunes (Whittaker and Jaeggi 1982). Step-pool morphologies develop during infrequent flood events and are associated with supply-limited conditions, steep gradients, coarse bed materials, and confined channels (Chin 1989). Like cascade reaches, large bed-forming material mobilizes infrequently (Grant et al. 1990), while finer pool-filling material is transported annually as bedload (Schmidt and Ergenzinger 1992). Step-pool reaches often receive episodic slugs of sediment input that travel downstream as bedload waves (Whittaker 1987).

Plane-bed reaches. Plane-bed reaches are characterized by a relatively featureless gravel/cobble bed (Figure 2.4e) that encompasses channel units (described later) often termed glides, riffles, and rapids (Bisson et al. 1982). Plane-bed reaches may be either unconfined or confined by valley walls and are distinguished from cascade reaches by the absence of tumbling flow and smaller relative roughness (ratio of the largest grain size to bankfull flow depth). Plane-bed reaches lack sufficient lateral flow convergence to develop pool-riffle morphology (discussed in next section) because of lower width-to-depth ratios and greater relative roughnesses, which may decompose lateral flow into smaller circulation cells (Ikeda 1977 and 1983). Bed surfaces are often armored in plane-bed reaches and are calculated to have a near-bankfull threshold for general mobility (Buffington 1995). Lack of depositional features, such as barforms, and typically armored bed surfaces demonstrate some supply-limited characteristics of plane-bed reaches (Dietrich et al. 1989). However, studies of armored gravel-bed channels show a general correlation of bedload transport rate and discharge during armor-breaching events (Jackson and Beschta 1982), indicating that sediment transport is not limited by supply once the bed is mobilized. Hence, plane-bed reaches represent a transition between supply- and transport-limited morphologies.

Pool-riffle reaches. Pool-riffle reaches are typically unconfined by valley walls and consist of a laterally oscillating sequence of bars, pools, and riffles (Figure 2.4f) resulting from oscillating cross-channel flow that causes flow convergence and scour on alternating banks of the channel. Concordant downstream flow divergence on the opposite side of the channel results in local sediment accumulation in discrete bars. Bedform and grain roughness provide the primary flow resistance in pool-riffle reaches. Bedforms in many pool-riffle reaches are relatively stable morphologic features, even though the material forming the bed is transported annually. Alluvial bar development requires a large width-to-depth ratio and small grains easily mobilized and aggraded by the flow (Church and Jones 1982). Pool-riffle reaches are commonly armored, exhibiting a near bank-full threshold for general bed surface mobility (Parker 1979, see also review by Buffington 1995) and a mixture of supply- and transport-limited characteristics similar to plane-bed reaches. Although the presence of depositional barforms in pool-riffle reaches suggests that they are generally more transport-limited than plane-bed reaches, the transport-limited character of both of these morphologies contrasts with the more supply-limited character of step-pool and cascade reaches.

Dune-ripple reaches. Dune-ripple reaches are unconfined, low-gradient, sand-bedded channels (Figure 2.4g). They exhibit a succession of mobile bedforms with increasing flow depth and velocity that proceeds as lower-regime plane bed, ripples, sand waves, dunes, upper-regime plane bed, and finally antidunes (Gilbert 1914, Simons et al. 1965). The primary flow resistance is provided by bedforms (Kennedy 1975), several scales of which may coexist; ripples and small dunes that climb over larger dunes as they all move down the channel. Sediment transport in dune-ripple reaches occurs at most stages, and strongly depends on discharge; as such, these reaches are transport-limited.

Forced Alluvial Reaches

External flow obstructions, such as LWD and bedrock outcrops, force local flow convergence, divergence, and sediment impoundment that respectively form pools, bars, and steps (Figure 2.5). The morphologic impact of LWD, in particular, depends on the amount, size,

FIGURE 2.5. A pool-riffle reach morphology forced by large woody debris (LWD).

orientation, and position of debris, as well as channel size (Bilby and Ward 1989, Montgomery et al. 1995b, Chapter 13, this volume) and rates of debris recruitment, transport, and decay (Bryant 1980, Murhphy and Koski 1989). In small channels, LWD is generally stable over years to decades, and individual logs can dominate channel morphology by anchoring pool and bar forms. Logs oriented transversely and located low in the flow may form steps that create local plunge pools and hydraulic jumps, buttress significant amounts of sediment, and dissipate energy otherwise available for sediment transport (Keller and Swanson 1979, Marston 1982). Single logs oriented obliquely can result in scour pools and proximal sediment storage by both upstream buttressing and downstream deposition in low-energy zones. However, in large rivers where individual pieces are mobile, debris jam formation is necessary for LWD to significantly influence channel morphology (Abbe and Montgomery 1996). When accumulated in jams, LWD can influence channel pattern and floodplain processes in large forest channels by armoring banks, developing pools, bars, and side channels, and forcing bank cutting and channel avulsions (Bryant 1980, Nakamura and Swanson 1993, Abbe and Montgomery 1996).

Flow obstructions can force specific channel morphologies on steeper slopes than is typical of analogous free-formed alluvial morphologies. In particular, LWD may force pool-riffle formation in otherwise plane-bed or bedrock reaches (Montgomery et al. 1995b, 1996). Consequently, plane-bed reaches are rare in undisturbed forested environments where LWD dominates formation of pools and bars. LWD may also force step-pool morphologies in otherwise cascade or bedrock reaches. It is important to recognize forced morphologies as distinct reach types because the interpretation of whether such obstructions govern bed morphology is crucial for understanding channel response.

Channel Units

Channel units are morphologically distinct areas that extend up to several channel widths in length and are spatially embedded within a channel reach (Figure 2.3); they are the morphologic building blocks of a reach. Channel units are classified as various types of pools, bars, and shallows (i.e., riffles, rapids, and cascades) (Bisson et al. 1982, Sullivan 1986, Church 1992). Distinctions among these units focus on topographic form, organization and areal density of clasts, local slope, flow depth and velocity, and to some extent, grain size. Different channel units have characteristic velocities and depths and provide specific habitat characteristics associated with different patterns of fish use (Chapter 9). In practice, however, definitions of these channel unit morphologies tend to overlap and vary with discharge; and channel unit classification by different observers often yields inconsistent results.

Although there is a general association of specific channel unit morphologies with reach type, prediction of channel unit properties is complicated by site-specific controls, particularly in forest environments. For example, the size, location, and orientation of individual flow obstructions control the distribution of specific pool types and their dimensions (Lisle 1986, Cherry and Beschta 1989). Furthermore, similar channel unit morphologies will have different response potentials depending on reach type and associated physical processes and conditions. For example, plunge pools in a LWD-forced pool-riffle reach may exhibit prolonged and extensive pool filling in response to increased sediment supply, while similar pools in a step-pool reach may be less responsive due to a potentially greater ratio of transport capacity to sediment supply. While the channel unit scale is biologically relevant, interpretation of the abundance, characteristics, and response potential of channel units depends upon the context imposed by reach-level channel types.

Channel Disturbance and Response Potential

Response to land use or environmental change varies for different channel types. Alluvial channels, in particular, exhibit a wide variety of potential responses. The predictive ability of conceptual models of channel response can be dramatically improved by considering the influence of reach-level channel type. Reach morphologies are associated with physical processes and environments that limit the range and magnitude of possible channel responses to changes in hydraulic discharge and sediment supply. Reach-specific response potential is further affected by external influences, such as channel confinement, riparian vegetation, and in-channel LWD. The impact of both isolated and cumulative watershed disturbance(s) on a particular reach also depends on the location of the reach within the drainage basin and the sequence of upstream reach types.

Reach-Level Response

Differences in reach morphology and physical processes result in different potential responses to similar changes in discharge or sediment supply. The specific response of a particular channel reach also reflects the intensity of disturbance or the magnitude of change, as well as the condition of the reach. While there are many possible response scenarios depending on site-specific factors, Table 2.2 illustrates typical patterns of potential response for Pacific Northwest channel reaches; direction of response is indicated by relationships (2.13) and (2.14). Changes in sediment storage dominate the response of colluvial reaches to altered sediment supply because of transport-limited conditions and low fluvial transport capacities; depending on the degree of valley fill, increased discharge can significantly change reach morphology. In contrast, bedrock, cascade, and step-pool reaches are resilient to most discharge or sediment supply perturbations because of high transport capacities and

TABLE 2.2. Interpreted reach-level channel response potential to moderate changes in sediment supply and discharge (+ = likely to change; p = possible to change; − = unlikely to change).

	Reach level morphology	Width	Depth	Roughness	Scour depth	Grain size	Slope	Sediment storage
Response	dune-ripple	+	+	+	+	−	+	+
	pool-riffle	+	+	+	+	+	+	+
	plane-bed	p	+	p	+	−	+	p
Transport	step-pool	−	p	p	p	p	p	p
	cascade	−	−	p	−	p	−	−
	bedrock	−	−	−	−	−	−	−
Source	colluvial	p	p	−	p	p	−	+

Modified from Montgomery and Buffington, 1997.

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

Segment-Level Response

Position within the network and differences between ratios of transport capacity to sediment supply allow aggregation of channel reaches into source, transport, and response segments. Source segments are headwater colluvial channels that act as transport-limited, sediment storage sites subject to intermittent debris flow scour. Transport segments are composed of morphologically resilient, supply-limited reaches (bedrock, cascade, and step-pool) that rapidly convey increased sediment inputs (Table 2.2). Response segments consist of lower-gradient, more transport-limited reaches (plane-bed, pool-riffle, and dune-ripple) in which significant morphologic adjustment occurs in response to increased sediment supply (Table 2.2).

The spatial distribution of source, transport, and response segments generally reflects the distribution of potential impacts and recovery times. Distribution of these segment types define watershed-scale patterns of sensitivity to altered discharge and sediment supply (Figure 2.6). Sediment delivered to transport segments rapidly propagates to downstream response segments, where sediment accumulates. Consequently, locations in the channel network where transport segments flow into response segments indicate places particularly susceptible to impacts from accelerated sediment supply. In this regard, the general classification of source, transport, and response segments identifies areas most susceptible to local increases in upstream sediment inputs. Because response segments are sensitive to increases in sediment supply, they are excellent sites for monitoring the effects of upstream actions and can be considered locations of critical importance in watershed monitoring programs. The relationship between channel classification and response potential provides an understanding of the linkage between upstream sediment inputs and downstream response.

External Influences

External influences on channel response include factors such as confinement, riparian vegetation, and in-channel LWD. Specific effects of these factors vary both with channel type and position within the network.

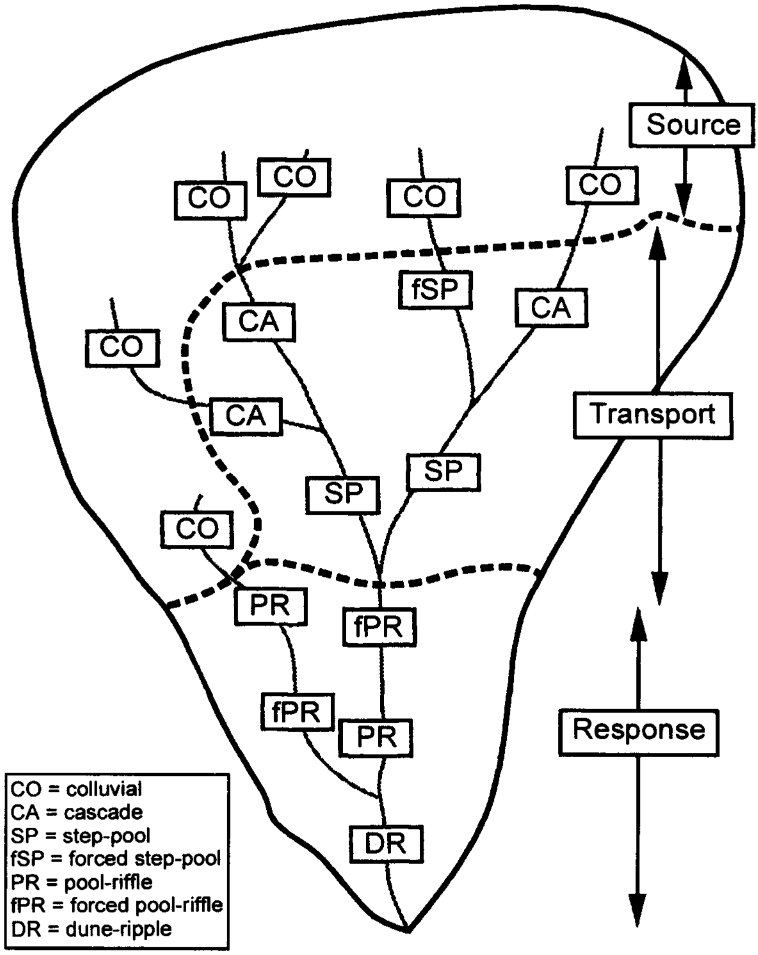


FIGURE 2.6. The spatial distribution of reach types within a drainage basin influences the distribution of potential impacts and responses to disturbance.

Confinement

Channel confinement strongly influences channel response. Channel migration and avulsion, for example, typically are rare in confined channels. The geometry of the channel above the bankfull stage also controls the response of the channel bed to high-discharge flow. Unconfined channels possess extensive floodplains across which over-bank flows spread, which limits the effect of peak discharges on channel morphology. In contrast, confined channels efficiently translate high flows into increased basal shear stress.

The degree of channel confinement may be influenced by either the long-term sediment balance of the channel or by external influences. Unconfined channels may reflect long-term alluvial aggradation and flood plain development where sediment supply exceeds transport capacity. Alternatively, unconfined channels may be controlled by tectonic boundary conditions (as in the case of alluvial fans at the base of block-faulted mountains), or reflect an inherited morphology (as in the case of underfit channels or u-shaped glacial valleys).

Isolation of unconfined channels from their floodplains can entail dramatic consequences for many biological systems. Prevention of overbank flows by dikes, or other flood control measures, may trigger channel entrenchment.

Flow diversions or regulation that prevent or decrease the frequency of floodplain inundation change both side-channel and floodplain processes. Abandonment of side channels and ponds may eliminate important aquatic habitat. Prevention of over-bank flows also stops sediment and nutrient delivery to floodplain soils, which may affect both floodplain-dwelling organisms and the long-term productivity of agricultural land.

Riparian Vegetation and Large Woody Debris

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

LWD provides significant control on the formation and physical characteristics of pools, bars, and steps, thereby influencing channel type and the potential for change in sediment storage and bedform roughness in response to altered sediment supply, discharge, or LWD loading. LWD may also decrease the potential for channel widening by armoring stream banks; alternatively it may aid bank erosion by directing flow and scour toward channel margins. Furthermore, bed surface textures and their response potential are strongly controlled by hydraulic roughness resulting from in-channel LWD and debris-forced bedforms (Buffington 1995).

Changes in amount, size, and decay rate of LWD may also affect channel processes and morphology. Alteration of channel margin vegetation may change both the age and species of wood entering the fluvial system (Murphy and Koski 1989). In small channels where LWD provides significant sediment storage, decreased supply of LWD accelerates sediment transport (Smith et al. 1993a). Channels in which LWD provides a dominant control on pool formation and sediment storage (e.g., forced pool-riffle or forced step-pool channels) are particularly sensitive to changes in the size, species, and amount of recruited LWD (Figure 2.7). Removal of LWD from forced pool-riffle reaches may alter the size and location of pools (Smith et al. 1993b) and lead to either a pool-riffle or plane-bed morphology depending on channel slope (Montgomery et al. 1995b). Similarly, loss of LWD may transform a forced step-pool reach into a step-pool or cascade reach, depending on channel slope and discharge. Where transport capacities are in extreme excess of sediment supply, forced alluvial reaches may become bedrock reaches following LWD removal (Montgomery et al. 1996).

Debris Flow Disturbance

Debris flows are primary agents of channel disturbance in mountain drainage basins. Debris flows tend to be pulsed disturbances, the effects of which vary with slope and position in the channel network. Passage of a debris flow can scour steep channels to bedrock. Deposition in lower-gradient channels typically results in local aggradation and can even obliterate the channel as a morphological feature. Recovery from debris flow impacts also differs for steep and low-gradient channels. Steep, high-energy channels (bedrock, step-pool, and cascade reaches) recover quickly from sediment deposition because of high transport capacities. In contrast, lower-gradient channels (plane-bed and pool-riffle reaches) typically take longer to recover from debris flow deposition because of their lower transport capacity. The morphology of mountain channels prone to debris flows thus reflects the time since debris flow scour, as well as position within the fluvial system. Although channel gradient generally determines the type of debris flow impacts, channel network architecture influences debris flow routing. Assessment of potential impacts of debris flow involves differentiating areas of potential debris flow initiation, scour, and deposition (Benda and Cundy 1990).

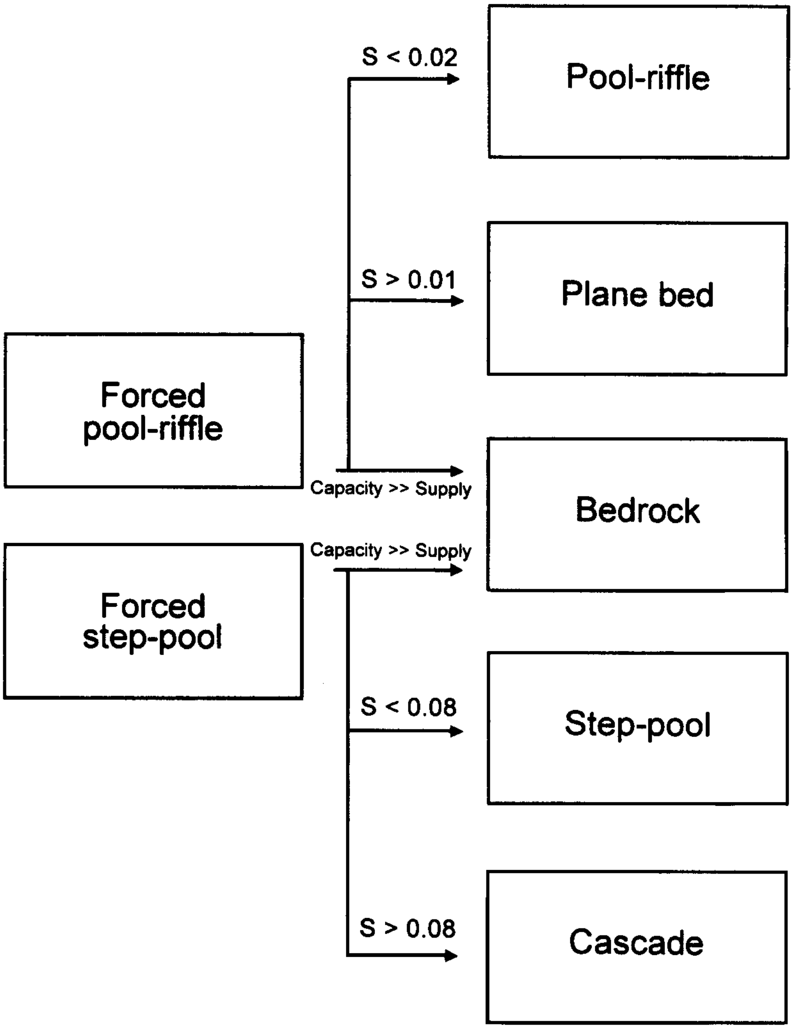


FIGURE 2.7. Potential morphologic response to LWD removal in channel reaches with forced alluvial morphology. Response depends on both channel slope (S) and the relative magnitudes of transport capacity and sediment supply.

Channel slope and tributary junction angles exert important controls on debris flow routing. Debris flows originating at the heads of long straight channels tend to scour long channel segments, and deliver sediment to downslope alluvial channels (Grant et al. 1984, Benda and Dunne 1987). Such events also may scour the base of adjacent hillslopes, hollows, and tributary channels, activating smaller failures that contribute to the sediment load imposed upon downslope channels. Debris flows originating in obliquely oriented tributaries tend to deposit at channel confluences (Grant et al. 1984, Benda and Cundy 1990). Subsequent events large enough to scour the accumulated material in the main channel can have catastrophic impacts on downstream alluvial channels. Debris flow deposition occurs when the channel slope declines to the extent that the yield strength of the flowing debris is sufficient to resist further transportation and deformation. This gradient is commonly between 5% and 10% for the range of water contents typical of debris flows (Takahashi et al. 1981, Benda and Cundy 1990), but incorporation of LWD in the leading edge of a debris flow may result in deposition on steeper slopes.

Massive inputs of sediment, such as from extensive synchronous landsliding within a basin, can set up a sediment wave that pulses through downstream channels (Figure 2.8). Passage of such waves involves local aggradation accompanied by fining of the bed. If large enough, passage of a sediment wave can change channel type; bedrock channels may become alluvial (Perkins 1989) or pool-riffle channels may become braided. Debris flow inputs can set up oscillations in channel morphology, the frequency of which varies with position in the network (Benda 1994). In historical old-growth forests of the Pacific Northwest, large stable log jams likely damped propagation of sediment waves through channel networks. Today the availability of such sediment capacitors is low in many watersheds due to stream cleaning and salvage operations, as well as harvesting large logs, from riparian forests; these large logs would otherwise serve to stabilize wood jams (Abbe and Montgomery 1996, Montgomery et al. 1996). Hence, large-scale sediment waves may be more important in today's industrial forests than in primeval forests. Channel morphology generally recovers after passage of a sediment wave at a rate that depends upon slope, confinement, sediment supply, and position in the network.

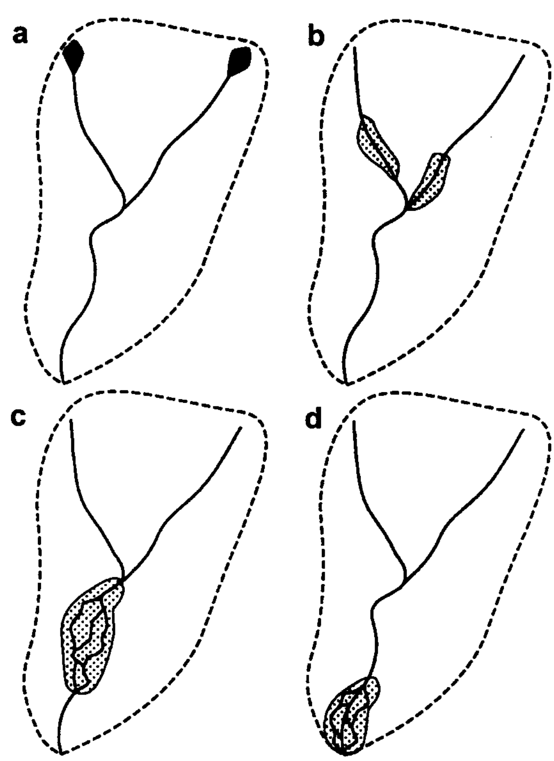


FIGURE 2.8. Sediment wave propagation through a channel network: (a) landsliding originating at the head of the channel network rapidly propagates (b) through steep headwater channels to lower-gradient reaches (c) where deposition may trigger channel braiding, which eventually (d) recovers to a single thread morphology.

Applications for Ecosystem Analysis

While the channel types discussed earlier are readily identified in the field, a method for classifying channel reaches from topographic maps or aerial photographs is useful for watershed-scale analysis and rapid assessment of reach types and response potential in mountain drainage basins. A classification that does not require visual assessment is particularly useful for designing channel assessment plans and interpreting conditions across entire watersheds. Such a classification can be based on channel gradient and confinement, two key features readily estimated from topographic maps and digital elevation models (DEMs).

Empirical association of channel type with different reach slopes provides a method for predicting reach type from topographic maps or DEMs. Frequency distributions of surveyed reach slopes for Pacific Northwest channels illustrate associations between reach type and slope gradient: pool-riffle reaches are most common on slopes less than 1%; plane-bed reaches are most common on slopes of 2% to 4%; step-pool channels are common on slopes of 4% to 8%; cascade morphologies dominate on slopes of 8% to 20%; and channels steeper than 20% typically have colluvial bed morphologies (Figure 2.9). The slope ranges for each reach type overlap, and bedrock reaches

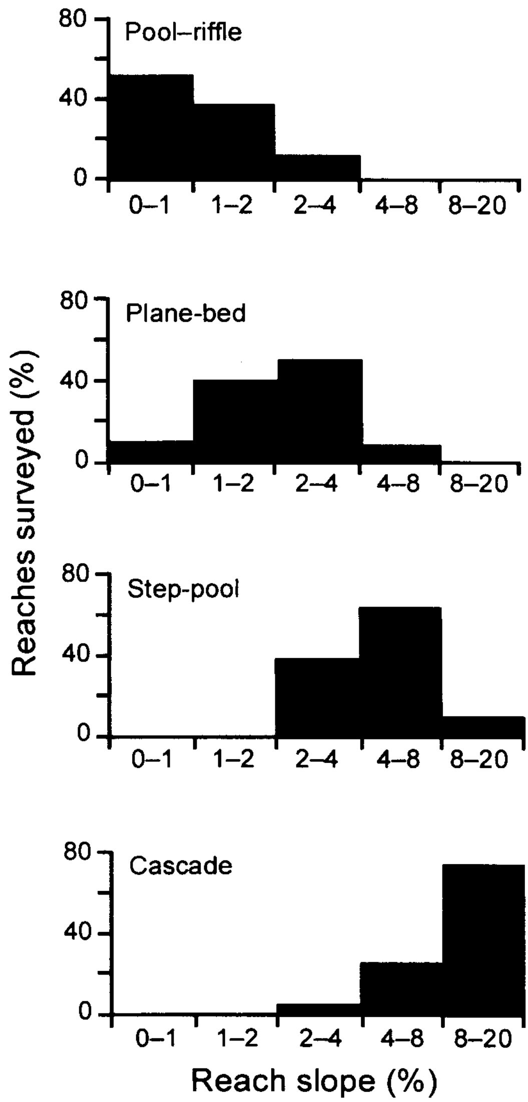


FIGURE 2.9. Frequency distributions of reach slopes for pool-riffle (n = 48), plane-bed (n = 57), step-pool (n = 51), and cascade (n = 78) morphologies surveyed in Oregon, Washington, and southeast Alaska.

are not associated with any particular slope range. Moreover, the LWD loading that controls forced reach morphologies cannot be predicted from maps or DEMs. Although predictions of channel type based on reach-average slope can provide a reasonable stratification of channel networks in mountain drainage basins of the Pacific Northwest, field observations are necessary to verify reach type, condition, and response potential.

Three simple confinement classes (totally confined, moderately confined, and unconfined) can be estimated from aerial photographs or valley widths portrayed on 7.5′ topographic maps. Although a gradient/confinement index provides a map-based stratification of channel networks useful in watershed analyses (Figure 2.10), channel confinement can be difficult to estimate from topographic maps because channel width is poorly expressed; thus, predicted slope and confinement may prove inaccurate in the field. Nevertheless, explaining discrepancies between expected conditions and those observed in the field can provide insight into local channel processes or disturbance history.

In mountain drainage basins, channel gradient and confinement provide a rough guide for identifying channels with different frequencies and magnitudes of ecologically relevant disturbance processes. Identification of headwater colluvial channels, confined alluvial channels, and unconfined alluvial channels can quickly characterize spatial patterns in the types, magnitudes, and frequencies of geomorphological disturbances that influence ecological

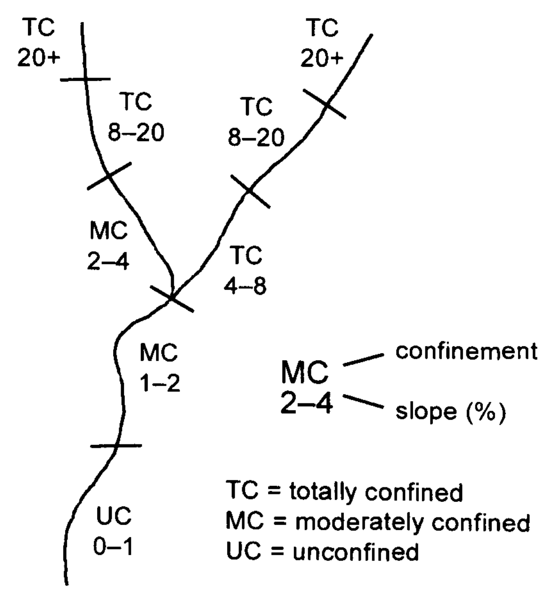


FIGURE 2.10. Application of a gradient/confinement index to channel classification.

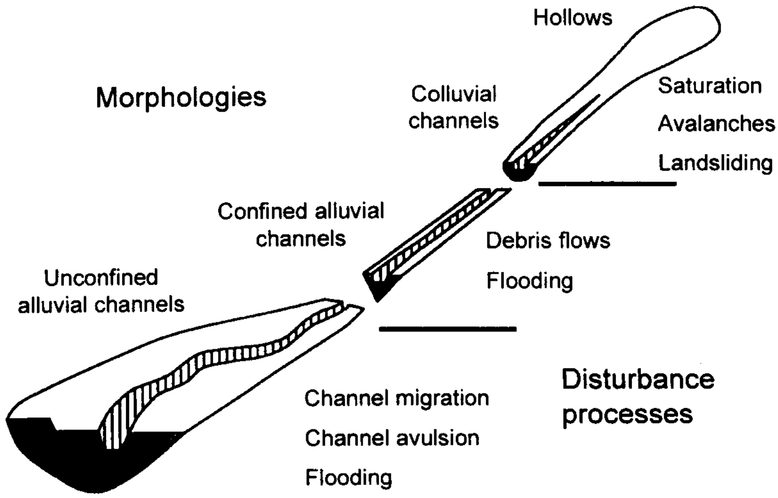


FIGURE 2.11. Differences in disturbance processes among colluvial, confined alluvial, and unconfined alluvial channels in mountain drainage basins.

organization and variability (Figure 2.11). More specific differences in disturbance processes can be evaluated based on consideration of bed morphology and disturbance history. The ecological significance of the environmental characteristics and variability associated with specific process domains depends upon the organism(s) of interest.

Channel classification cannot substitute for focused observation and clear thinking about channel processes. Channels are complex systems that need to be interpreted within their local and historical context. Classification simply provides one of a variety of tools that can be applied to particular problems—it is not a panacea. Classifications that highlight specific aspects of the linkages between channel networks and watershed processes are likely to be most useful, but careless application of any channel classification may prove misleading; no classification can substitute for an alert, intelligent, well-trained observer. Nonetheless, it is difficult to fully understand a channel reach without reference to the context defined by its bed morphology, confinement, position in the network, and disturbance history. Consideration of these factors within a spatial hierarchy can further guide interpretations of field observations and evaluation of channel conditions.

Acknowledgments. Preparation of this chapter was supported by the Washington State Timber-Fish-Wildlife agreement through grant TFW FY95-156 and by US Forest Service cooperative research agreements 93-0441 and 94-0617. The authors thank Leslie Reid and Tom Lisle for their critiques of a draft manuscript.

Literature Cited

Abbe, T.B., and D.R. Montgomery. 1996. Large woody debris jams, channel hydraulics and habitat formation in large rivers. Regulated Rivers: Research & Management **12**:201–221.

Allen, P.M., R. Hobbs, and N.D. Maier. 1989. Downstream impacts of a dam on a bedrock fluvial system. Bulletin of the Association of Engineering Geologists **26**:165–189.

Arcement, G.J., and V.R. Schneider. 1989. Guide for selecting Manning's roughness coefficients for natural channels and flood plains. United States Geological Survey Water Supply Paper 2339. US Government Printing Office, Washington, DC, USA.

Benda, L.E. 1994. Stochastic geomorphology in a humid mountain landscape. Unpublished Ph.D. dissertation. University of Washington, Seattle, Washington, USA.

Benda, L. 1990. The influence of debris flows on channels and valley floors in the Oregon Coast Range. USA Earth Surface Processes and Landforms **15**:457–466.

Benda, L., and T.W. Cundy. 1990. Predicting deposition of debris flows in mountain channels. Canadian Geotechnical Journal **27**:409–417.

Benda, L., and T. Dunne. 1987. Sediment routing by debris flows. Pages 213–223 *in* R.L. Beschta, R. Blinn, G.E. Grant, G. Ice, and F.J. Swanson, eds. Erosion and Sedimentation in the Pacific Rim. IAHS Publication 165, International Association of Hydrological Sciences, Wallingford, UK.

Bilby, R.E., and J.W. Ward. 1989. Changes in characteristics and function of woody debris with increasing size of streams in western Washington. Transactions of the American Fisheries Society **118**:368–378.

Bisson, P.A., J.L. Nielson, R.A. Palmason, and L.E. Gore. 1982. A system of naming habitat types in small streams, with examples of habitat utilization by salmonids during low stream flow. Pages 62–73 *in* N.B. Armantrout, ed. Acquisition and utilization of aquatic habitat information. American Fisheries Society, Portland, Oregon, USA.

Booth, D.B. 1990. Stream channel incision following drainage basin urbanization. Water Resources Bulletin **26**:407–417.

Bosch, J.M., and J.D. Hewlett. 1982. A review of catchment experiments to determine the effect of vegetation changes on water yield and evapotranspiration. Journal of Hydrology **55**:2–23.

Bryant, M.D. 1980. Evolution of large, organic debris after timber harvest: Maybeso creek, 1949 to 1978. USDA Forest Service General Technical Report PNW-101. Pacific Northwest Forest and Range Experiment Station, Portland, Oregon, USA.

Buffington, J.M. 1995. Effects of hydraulic roughness and sediment supply on surface textures of gravel-bedded rivers. Masters thesis. University of Washington, Seattle, Washington, USA.

Carling, P.A. 1987. Bed stability in gravel streams, with reference to stream regulation and ecology. Pages 321–347 *in* K.S. Richards, ed. River channels: Environment and process. Blackwell, Oxford, UK.

Cherry, J., and R.L. Beschta. 1989. Coarse woody debris and channel morphology: A flume study. Water Resources Bulletin **25**:1031–1036.

Chin, A. 1989. Step pools in stream channels. Progress in Physical Geography **13**:391–407.

Church, M. 1992. Channel morphology and typology. Pages 126–143 *in* P. Calow and G.E. Petts, eds. The rivers handbook. Blackwell, Oxford, UK.

Church, M., and D. Jones. 1982. Channel bars in gravel-bed rivers. Pages 291–338 *in* R.D. Hey, J.C. Bathurst, and C.R. Thorne, eds. Gravel-bed rivers: Fluvial processes, engineering and management. John Wiley & Sons, Chichester, UK.

Coats, R., L. Collins, J. Florsheim, and D. Kaufman. 1985. Channel change, sediment transport, and fish habitat in a coastal stream: Effects of an extreme event. Environmental Management **9**:35–48.

Dietrich, W.E., and T. Dunne. 1978. Sediment budget for a small catchment in mountainous terrain. Zeitschrift für Geomorphologie, Supplementband **29**:191–206.

Dietrich, W.E., J.W. Kirchner, H. Ikeda, and F. Iseya. 1989. Sediment supply and the development of the coarse surface layer in gravel-bedded rivers. Nature **340**:215–217.

Ellen, S.D., and G.F. Wieczorek. 1988. Landslides, floods, and marine effects of the storm of January 3–5, 1982, in the San Francisco Bay Region, California. United States Geological Survey Professional Paper 1434. US Government Printing Office, Washington, DC, USA.

Frissell, C.A., and W.J. Liss. 1986. Classification of stream habitat and watershed systems in south coastal Oregon and an assessment of land use impacts. Progress report prepared for the Oregon Department of Fish and Wildlife, Oak Creek Laboratory of Biology, Corvallis, Oregon, USA.

Frissell, C.A., W.J. Liss, C.E. Warren, and M.D. Hurley. 1986. A hierarchical framework for stream habitat classification: Viewing streams in a watershed context. Environmental Management **10**:199–214.

Gilbert, G.K. 1877. Report on the geology of the Henry Mountains. United States Geological and Geographical Survey, Rocky Mountain Region. US Government Printing Office, Washington, DC, USA.

——. 1914. The transportation of debris by running water. United States Geological Survey Professional Paper 86. US Government Printing Office, Washington, DC, USA.

——. 1917. Hydraulic-mining débris in the Sierra Nevada. United States Geological Survey Professional Paper 105. US Government Printing Office, Washington, DC, USA.

Graf, W.L. 1975. The impact of suburbanization on fluvial geomorphology. Water Resources Research **11**:690–692.

Grant, G.E., M.J. Crozier, and F.J. Swanson. 1984. An approach to evaluating off-site effects of timber harvest activities on channel morphology. Proceedings, Symposium on the Effects of Forest Land Use on Erosion and Slope Stability. Environment and Policy Institute, East–West

Center, University of Hawaii, Honolulu, Hawaii, USA.

Grant, G.E., F.J. Swanson, and M.G. Wolman. 1990. Pattern and origin of stepped-bed morphology in high-gradient streams, Western Cascades, Oregon. Geological Society of America Bulletin **102**:340–352.

Griffiths, G.A. 1980. Stochastic estimation of bed load yield in pool-and-riffle mountain streams. Water Resources Research **16**:931–937.

Hammer, T.R. 1972. Stream channel enlargement due to urbanization. Water Resources Research **8**:1530–1540.

Harr, R.D. 1981. Some characteristics and consequences of snowmelt during rainfall in western Oregon. Journal of Hydrology **53**:277–304.

——. 1983. Potential for augmenting water yield through forest practices in western Washington and western Oregon. Water Resources Bulletin **19**:383–393.

——. 1986. Effects of clearcutting on rain-on-snow runoff in western Oregon: A new look at old studies. Water Resources Research **22**:1095–2000.

Helley, E.J., and V.C. LaMarche, Jr. 1973. Historic flood information for northern California streams from geologic and botanical evidence. United States Geological Survey Professional Paper 485-E. US Government Printing Office, Washington, DC, USA.

Henderson, F.M. 1963. Stability of alluvial channels. Transactions of the American Society of Civil Engineers **128**:657–686.

Horton, R.E. 1945. Erosional development of streams and their drainage basins; Hydrophysical approach to quantitative morphology. Geological Society of America Bulletin **56**:275–370.

Ikeda, H. 1983. Experiments on bedload transport, bed forms, and sedimentary structures using fine gravel in the 4-meter-wide flume. Environmental Research Center Papers, Environmental Research Center, University of Tsukuba, Ibaraki, Japan.

Ikeda, S. 1977. On the origin of bars in the meandering channels. Bulletin of the Environmental Research Center, University of Tsukuba **1**:17–31.

Jackson, W.L., and R.L. Beschta. 1982. A model of two-phase bedload transport in an Oregon Coast Range stream. Earth Surface Processes and Landforms **7**:517–527.

Jackson, W.L., and R.L. Beschta. 1984. Influences of increased sand delivery on the morphology of sand and gravel channels. Water Resources Bulletin **20**:527–533.

James, L.A. 1991. Incision and morphologic evolution of an alluvial channel recovering from hydraulic mining sediment. Geological Society of America Bulletin **103**:723–736.

Jones, J.A., and G.E. Grant. 1996. Peak flow responses to clear-cutting and roads in small and large basins, western Cascades, Oregon. Water Resources Research **32**:959–974.

Keller, E.A., and F.J. Swanson. 1979. Effects of large organic material on channel form and fluvial processes. Earth Surface Processes **4**:361–380.

Kellerhals, R., M. Church, and D.I. Bray. 1976. Classification and analysis of river processes. Journal of the Hydraulics Division, American Society of Civil Engineers **102**:813–829.

Kelsey, H.M. 1980. A sediment budget and an analysis of geomorphic process in the Van Duzen River basin, north coastal California, 1941–1975. Geological Society of America Bulletin **91**:1119–1216.

Kennedy, J.F. 1975. Hydraulic relations for alluvial streams. Pages 114–154 *in* V. Vanoni, ed. Sedimentation engineering. Manual 54, American Society of Civil Engineers, New York, New York, USA.

Knighton, A.D. 1991. Channel bed adjustment along mine-affected rivers of northeast Tasmania. Geomorphology **4**:205–219.

Kondolf, G.M. 1995. Geomorphological stream channel classification in aquatic habitat restoration: Uses and limitations. Aquatic Conservation: Marine and Freshwater Ecosystems **5**:127–141.

Lane, E.W. 1955. The importance of fluvial morphology in hydraulic engineering. Proceedings of the American Society of Civil Engineers **81**:745–761.

Leopold, L.B. 1968. Hydrology for urban land planning: A guidebook on the hydrologic effects of urban land use. United States Geological Survey Circular 554. US Government Printing Office, Washington, DC, USA.

Leopold, L.B., and T. Maddock, Jr. 1953. The hydraulic geometry of stream channels and some physiographic implications. United States Geological Survey Professional Paper 252. US Government Printing Office, Washington, DC, USA.

Leopold, L.B., and M.G. Wolman. 1957. River channel patterns: Braided, meandering and straight. United States Geological Survey Professional Paper 282-B. US Government Printing Office, Washington, DC, USA.

Lisle, T.E. 1982. Effects of aggradation and degradation on riffle-pool morphology in natural gravel channels, northwestern California. Water Resources Research **18**:1643–1651.

——. 1986. Stabilization of a gravel channel by large streamside obstructions and bedrock bends, Jacoby Creek, northwestern California. Geological Society of America Bulletin **97**:999–1011.

Lyons, J.K., and R.L. Beschta. 1983. Land use, floods and channel changes: Upper Middle Fork Willamette River, Oregon (1936–1980). Water Resources Research **19**:463–471.

Marston, R.A. 1982. The geomorphic significance of log steps in forest streams. Annals of the American Association of Geographers **72**:99–108.

Megahan, W., W.S. Platts, and B. Kulesza. 1980. Riverbed improves over time: South Fork Salmon. Proceedings of the Symposium on Watershed Management, American Society of Civil Engineers, New York, New York, USA.

Mollard, J.D. 1973. Air photo interpretation of fluvial features. Proceedings of the Canadian Hydrology Symposium. National Research Council, Ottawa, Ontario, Canada.

Montgomery, D.R., and J.M. Buffington. 1997. Channel reach morphology in mountain drainage basins. Geological Society of America Bulletin **109**:596–611.

Montgomery, D.R., T.B. Abbe, N.P. Peterson, J.M. Buffington, K.M. Schmidt, and J.D. Stock. 1996. Distribution of bedrock and alluvial channels in forested mountain drainage basins. Nature **381**:587–589.

Montgomery, D.R., J.M. Buffington, R.D. Smith, K.M. Schmidt, and G. Pess. 1995b. Pool spacing in forest channels. Water Resources Research **31**:1097–1105.

Montgomery, D.R., and W.E. Dietrich. 1988. Where do channels begin? Nature **336**:232–234.

Montgomery, D.R., G.E. Grant, and K. Sullivan. 1995a. Watershed analysis as a framework for implementing ecosystem management. Water Resources Bulletin **31**:369–386.

Morisawa, M.E. 1957. Accuracy of determination of stream lengths from topographic maps. Transactions, American Geophysical Union **38**:86–88.

Murphy, M.L., and K.V. Koski. 1989. Input and depletion of woody debris in Alaska streams and implications for streamside management. North American Journal of Fisheries Management **9**:427–436.

Nakamura, F., and F.J. Swanson. 1993. Effects of coarse woody debris on morphology and sediment storage of a mountain stream system in western Oregon. Earth Surface Processes and Landforms **18**:43–61.

Nolan, K.M., and D.C. Marron. 1988. Stream channel response to the storm. Pages 245–264 *in* S.D. Ellen and G.F. Wieczorek, eds. Landslides, floods, and marine effects of the storm of January 3–5, 1982, in the San Francisco Bay Region, California. United States Geological Survey Professional Paper 1434. US Government Printing Office, Washington, DC, USA.

Nunnally, N.R. 1985. Application of fluvial relationships to planning and design of channel modifications. Environmental Management **9**:417–426.

Parker, G. 1979. Hydraulic geometry of active gravel rivers. Journal of the Hydraulics Division, American Society of Civil Engineers **105**:1185–1201.

Paustian S.J., K. Anderson, D. Blanchet, S. Brady, M. Cropley, J. Edgington, et al. 1992. A channel type users guide for the Tongass National Forest, southeast Alaska. USDA Forest Service Technical Paper R10-TP-26. Alaska Region R10.

Perkins, S.J. 1989. Interactions of landslide-supplied sediment with channel morphology in forested watersheds. Unpublished Masters thesis. University of Washington, Seattle, Washington, USA.

Peterson, D.F., and P.K. Mohanty. 1960. Flume studies of flow in steep, rough channels. Journal of the Hydraulics Division, American Society of Civil Engineers **86**:55–76.

Platts, W.S., R.J. Torquedada, M.L. McHenry, and C.K. Graham. 1989. Changes in Salmon spawning and rearing habitat from increased delivery of fine sediment to the South Fork Salmon River, Idaho. Transactions of the American Fisheries Society **118**:274–283.

Powell, J.W. 1875. Exploration of the Colorado River of the west and its tributaries. Government Printing Office, Washington, DC, USA.

Reneau, S.L., W.E. Dietrich, D.J. Donahue, A.J.T. Jull, and M. Rubin. 1990. Late Quaternary history of colluvial deposition and erosion in hollows, central California Coast Ranges. Geological Society of America Bulletin **102**:969–982.

Rice, S. 1994. Towards a model of changes in bed material texture at the drainage basin scale. Pages 159–172 *in* M.J. Kirkby, ed. Process models and theoretical geomorphology. John Wiley & Sons, Chichester, UK.

Rosgen, D.L. 1994. A classification of natural rivers. Catena **22**:169–199.

Schmidt, K.H., and P. Ergenzinger. 1992. Bedload entrainment, travel lengths, step lengths, rest periods—studied with passive (iron, magnetic) and active (radio) tracer techniques. Earth Surface Processes and Landforms **17**:147–165.

Schumm, S.A. 1971. Fluvial Geomorphology: Channel adjustment and river metamorphosis. Pages 5-1–5-22 *in* H.W. Shen, ed. River mechanics.

Volume 1. H.W. Shen, Fort Collins, Colorado, USA.

——. 1977. The fluvial system. John Wiley & Sons, New York, New York, USA.

Shaler, N.S. 1891. The origin and nature of soils. Pages 213–345 *in* J.W. Powell, ed. United States Geological Survey Twelfth Annual Report. US Government Printing Office, Washington, DC, USA.

Shields, A. 1936. Anwendung der Änlichkeits-mechanik und der Turbulenzforschung auf die Geschiebebewegung. Translated by W.P. Ott and J.C. van Uchelen. California Institute of Technology, Pasadena, California, USA.

Simons, D.B., E.V. Richardson, and C.F. Nordin. 1965. Sedimentary structures generated by flow in alluvial channels. Pages 34–52 *in* G.V. Middleton, ed. Primary sedimentary structures and their hydrodynamic interpretation. Society of Economic Paleontologists and Mineralogists, Tulsa, Oklahoma, USA.

Smith, D.G. 1976. Effect of vegetation on lateral migration of anastomosed channels of a glacier meltwater river. Geological Society of America Bulletin **87**:857–860.

Smith, R.D., R.C. Sidle, and P.E. Porter. 1993a. Effects on bedload transport of experimental removal of woody debris from a forest gravel-bed stream. Earth Surface Processes and Landforms **18**:455–468.

Smith, R.D., R.C. Sidle, P.E. Porter, and J.R. Noel. 1993b. Effects of experimental removal of woody debris on the channel morphology of a forest, gravel-bed stream. Journal of Hydrology **152**:153–178.

Strahler, A.N. 1957. Quantitative analysis of watershed geomorphology. Transactions, American Geophysical Union **38**:913–920.

Sullivan, K. 1986. Hydraulics and fish habitat in relation to channel morphology. Unpublished Ph.D. dissertation. Johns Hopkins University, Baltimore, Maryland, USA.

Sullivan, K., T.E. Lisle, C.A. Dolloff, G.E. Grant, and L.M. Reid. 1987. Stream Channels: The link between forests and fishes. Pages 39–97 *in* E.O. Salo and T.W. Cundy, eds. Streamside management: Forestry and fishery interactions. Institute of Forest Resources Contribution Number 57, University of Washington, Seattle, Washington, USA.

Swanson, F.J., R.L. Fredriksen, and F.M. McCorison. 1982. Material transfer in a western Oregon forested watershed. Pages 223–266 *in* R.L. Edmonds, ed. Analysis of coniferous forest ecosystems in the western United States. Hutchinson Ross, Stroudsburg, Pennsylvania, USA.

Takahashi, T., K. Ashida, and K. Sawai. 1981. Delineation of debris flow hazard areas. Pages 589–603 *in* T.R.H. Davies and A.J. Pearce, eds. Erosion and sediment transport in Pacific Rim steeplands. IAHS Publication 132, International Association of Hydrological Sciences, Wallingford, UK.

Whiting, P.J., and J.B. Bradley. 1993. A process-based classification for headwater streams. Earth Surface Processes and Landforms **18**:603–612.

Whitlow, J.R., and K.J. Gregory. 1989. Changes in urban stream channels in Zimbabwe. Regulated Rivers: Research and Management **4**:27–42.

Whitney, J.D. 1880. The Auriferous Gravels of the Sierra Nevada. Memoir 6. Harvard University Museum of Comparitive Zoology, Cambridge, Massachusetts, USA.

Whittaker, J.G. 1987. Sediment transport in step-pool streams. Pages 545–579 *in* C.R. Thorne, J.C. Bathurst, and R.D. Hey, eds. Sediment transport in gravel-bed rivers. John Wiley & Sons, Chichester, UK.

Whittaker, J.G., and T.R.H. Davies. 1982. Erosion and sediment transport processes in step-pool torrents. Pages 99–104 *in* D.E. Walling, ed. Recent developments in the explanation and prediction of erosion and sediment yield. IAHS Publication Number 137, International Association of Hydrological Sciences, Wallingford, UK.

Whittaker, J.G., and M.N.R. Jaeggi. 1982. Origin of step-pool systems in mountain streams. Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers **108**:758–773.

Wilcock, P.R., G.M. Kondolf, W.V. Matthews, and A.F. Barta. 1996. Specification of sediment maintenance flows for a large gravel-bed river. Water Resources Research **32**:2911–2921.

Williams, G.P. 1978. Bank-full discharge of rivers. Water Resources Research **14**:1141–1154.

Williams, G.P., and M.G. Wolman. 1984. Downstream effects of dams on alluvial rivers. United States Geological Survey Professional Paper 1286. US Government Printing Office, Washington, DC, USA.

Wolman, M.G., and J.P. Miller. 1960. Magnitude and frequency of forces in geomorphic processes. Journal of Geology **68**:54–74.