TFW-SH10-93-002

CHANNEL CLASSIFICATION, PREDICTION OF CHANNEL RESPONSE, AND ASSESSMENT OF CHANNEL CONDITION

BY

David R. Montgomery and John M. Buffington

TIMBER/FISH
&WILD-LIFE

June 24, 1993

Channel Classification, Prediction of Channel Response, and Assessment of Channel Condition

David R. Montgomery and John M. Buffington
Department of Geological Sciences and Quaternary Research Center
University of Washington, Seattle, WA 98195

Report TFW-SI-110-93-002
prepared for the SHAMW committee of the
Washington State Timber/Fish/Wildlife Agreement

June 24. 1993

Channel Classification, Prediction of Channel Response,
and Assessment of Channel Condition

ABSTRACT

A process-based landscape and channel classification is proposed as a framework for assessing watershed response to natural and anthropogenic environmental change. Our proposed classification is based on a hierarchy of process-regimes at several spatial scales: i) geomorphic province, ii) watershed, iii) valley segment, iv) channel reach, and v) channel unit. The geomorphic province level identifies watersheds developed in similar materials, topography, and climates, reflecting comparable hydrologic, erosional, and tectonic processes. The watershed level distinguishes hillslopes from valleys, defining fundamental differences in transport processes within a contiguous drainage basin. Valley segment morphologies further distinguish transport processes and general relations between transport capacity and sediment supply of both channeled and unchanneled valleys. At the reach level, distinct morphologies may be identified based on sediment transport characteristics, channel roughness configuration. Channel reaches, in turn, are composed of finer-scale channel units.

Within this framework, our discussion focuses mainly on the valley segment and channel reach levels. Valley morphology and sediment transport characteristics define colluvial, alluvial, and bedrock valley segments. Unchanneled valleys (hollows) are characterized by a lack of fluvial processes, resulting in a transport-limited accumulation of colluvium that is periodically excavated by mass wasting processes. Channeled colluvial valleys am those in which fluvial sediment transport maintains a channel, but in which the transport capacity is insufficient to mobilize all of the colluvium delivered from the surrounding hillslopes. In mountain drainage basins, colluvial valleys am dominantly carved by mass wasting processes. Alluvial valleys contain predominantly alluvial fills and are characterized by fluvial transport of sediment over a variety of alluvial bed morphologies. Alluvial valley segments may be either confined or unconfined, reflecting general relations between transport capacity and sediment supply. Bedrock valley segments lack a continuous alluvial cover due to high transport capacities. Valley morphology generally reflects the relation between sediment supply and transport capacity.

At the channel reach-level of the classification, bed morphology is coupled with both the potential for debris flow impacts and the role of large woody debris loading to characterize channel processes and provide a framework within which to examine potential channel response. Colluvial and bedrock channels occupy corresponding valley segments, but we recognize six alluvial channel types: regime, braided, pool-riffle, plane bed, step-pool, and cascade. We hypothesize that observed systematic and local downstream changes in alluvial channel morphology and channel roughness correlate with changes in channel slope, sediment supply (size and amount of material available for transport) and

transport capacity (a function of the available shear stress). These differences provide the basis for interpreting the potential response of different areas of the channel network to perturbation. In general, steep alluvial channels (step-pool and cascade) tend to maintain their morphology while transmitting increased sediment loads. In contrast, low-gradient channels (regime and pool-riffle) typically respond to increased sediient loads through morphologic adjustment. In essence, steep channels effectively act as sediment delivery conduits connecting zones of sediment production on hillslopes to downslope low-gradient channels. Such distinctions allow recognition of source, transport, and response reaches. Channel morphology thus reflects the local and watershed-integrated processes influencing sediment supply and transport capacity. Evaluation of channel response potential within the context of morphologically-characteristic processes allows distinction of different response potential for different portions of a channel network.

While the proposed channel classification provides insight into potential channel response that can guide impact assessment, changes in sediment supply and transport capacity may result in either similar or opposing effects. This highlights the reality that changes in discharge and sediment supply cannot be examined in isolation; both need to be considered when assessing either watershed conditions or the potential for furture impacts. In particular, it is necessary to focus on aspects of channel morphology and dynamics that are sensitive indicators of perturbation and to consider the specific channel type and position in the channel network. A number of quantitative and qualitative approaches provide insight into evaluating watershed impacts and predicting potential responses to continuing or anticipated watershed disturbance.

"To protect your rivers, protect your mountains"
- attributed to Emperor Yu of China, 1,600 B.C.

INTRODUCTION

Addressing concerns over environmental degradation requires of strategies for assessing land management impacts on landscapes and ecosystems. Watersheds provide natural land management units because their boundaries coincide with those of natural processes. Changes in watershed processes can alter fluvial systems. At present, however, prediction of stream channel response to land use and disturbance is a weak link in watershed assessment methodologies, because channel processes am either poorly represented or viewed in isolation from the rest of the watershed. We propose a process-based classification of landscape and channel form that provides a foundation for interpreting channel morphology, assessing channel condition, and predicting reponse to natural and anthropogenic disturbance.

Stream channels integrate watershed processes. Hillslope processes generate and deliver sediment to channels; fluvial processes transport and redistribute sediment through the channel network. Analysis of channel characteristics requires a watershed context because channel response to perturbation reflects this coupling of hillslope and fluvial processes. The nature and style of fluvial sediment transport also reflects position within the channel network. Consequently, assessing channel condition and predicting channel response requires identification of functionally similar portions of the channel network. We contend that the processes governing landscape form provide the most logical context for organizing and classifying both landscapes and channel networks.

Two simple principles govern channel form and dynamics. First, conservation of mass dictates that both the water and sediment supplied to the upstream end of a channel reach must be either stored or discharged downstream. Second, the morphology and sediment transport dynamics of a channel reflect the style, magnitude, and frequency of both sediment and water input from upslope sources and the ability of the channel to transmit these loads to downslope reaches. Sediment delivery, hydraulic discharge, and channel slope vary both systematically and locally throughout a drainage network. Consequently, channel morphology, sediment transport dynamics, and response potential reflect both local conditions and the spatial context within the drainage network

Inherent in this arguement is that channel morphology is not static. Over geologic time, channels respond to tectonic uplift, subsequent erosional degradation of the landscape, and climatically-driven oscillations in discharge. Land uses such as urbanization, grazing, forestry practices, and dam construction alter channel processes in ways that affect both aquatic and riparian ecosystems, as well as human uses of fluvial systems. Concern over these impacts motivates assessing channel change to: 1) evaluate past channel response to climatic or anthropogenic disturbance, and 2) predict response to either future environmental changes or land management practices. The wide variety of channel types, the adjustment of individual channels to local factors, and potential time lags between perturbation and channel response complicate recognition and quantification of past, as well as prediction of future, changes in channel form or processes. In spite of this complexity, the intensity of contemporary land use necessitates refining a practical methodology to assess the condition and potential response of channel systems so that land managers can make informed decisions when confronted with competing interests, such as maximizing timber harvest and minimizing impacts on downstream fish populations.

This report discusses the theoretical basis for possible channel responses and reviews previous work on measuring and predicting channel change. We then synthesize previous studies of channel processes into a channel classification that illustrates how different portions of drainage basins function and respond to perturbation. This classification provides a framework for both studying watershed processes and drainage basin evolution and assessing channel condition and response potential.

THEORETICAL BASIS FOR INTERPRETING CHANNEL ADJUSTMENTS

Natural channels range in size from small ephemeral rivulets to the large rivers of the world and exhibit a wide variety of morphologies. In all channels, however, the morphology and sediment transport dynamics of a channel reflect the style, magnitude, and frequency of both sediment and water input from upslope sources, the ability of the channel to transmit these loads to downslope reaches, and the influence of vegetation on channel processes (Figure 1). Channel morphology and response potential reflect both local conditions and systematic downstream variation of the independent parameters shown in Figure 1. Changes in discharge and sediment supply result in a limited number of possible channel adjustments, which vary with channel morphology and position within the network. All of these potential adjustments may be examined in a theoretical context, allowing prediction of general channel responses to disturbance.

Potential channel adjustments to altered discharge and sediment load include changes in width, depth, velocity, slope, roughness, and sediment size (Leopold and Maddock, 1953). Equations describing the physics governing channel processes formalize controls on these possible channel adjustments. Conservation of energy and mass describe sediment transport and the flow of water through both the channel network and any point along a channel. Other equations describe the frictional dissipation of fluid energy by channel roughness elements, the relation between boundary shear stress and sediment transport, and the geometry of the active transport zone.

Precipitation over a landscape results in downslope movement of water, causing erosion and energy expenditure that forms and maintains channels. The frequency and magnitude of precipitation and the topographic relief onto which it falls provide the source of this potential energy. For the simple case of spatially-uniform rainfall, the potential energy (E_p) in a catchment is equal to the integral of the product of water mass (m), gravitational acceleration (g), and elevation (z)

$$E_p = \int m\, g\, dz \qquad (1)$$

Initially, the total energy of the system (E) consists of potential energy (mgz). Downslope movement of water converts this potential energy into kinetic energy ($mu^2/2$), pressure energy (mgD), and energy dissipated by friction and turbulence. Energy loss in this dissipative system is given by

$$AE = A(mgz) + \Delta(mu^2/2) + \Delta(mgD) \qquad (2)$$

where u and D are respectively the flow velocity and depth. Combining the bed elevation (z) and the flow depth (D) into a water surface elevation (H) and assuming that change in the downstream flow velocity is negligible (Leopold, 1953), the energy loss per unit bed area per channel length (L) can be expressed as

$$\Delta E / \Delta L \approx \Delta(D\, \rho_w\, g\, H) / \Delta L \qquad (3)$$

where ρ_w is the density of water. The amount and style of energy dissipation is a function of channel roughness (R). Thus, noting that AH/AL is the water surface slope (S), equation (3) implies that

$$R \propto DS \qquad (4)$$

In general, changes in water surface slope dominate flow depth changes (Leopold et al., 1964), so that channel roughness is primarily a function of water surface slope. Since channels tend to be steep in their headwaters and decrease in slope downstream, this implies that channel roughness generally decreases downstream.

Hydraulic discharge (Q) is fundamentally related to channel geometry. At every point within a channel network, bankfull discharge is equal to the product of the channel cross-sectional area and the velocity of the fluid, which for a rectangular channel of width W is

$$Q = W D u \tag{5}$$

The velocity of flow, in turn, depends upon the flow depth, slope, and viscosity and is inversely related to the flow resistance offered by the channel. Flow resistance may be calculated by one of several commonly used formulae (c.f., Leopold et al., 1964; Dunne and Leopold, 1978)

$$u = (D^{2/3} S^{1/2}) / n = C (DS)^{1/2} = (8gDS / f)^{1/2} \tag{6}$$

where n is the Manning resistance coefficient, C is the Chezy resistance factor, f is the Darcy-Weisbach friction factor, and D is assumed to approxmiate the hydraulic radius, an assumption most appropriate for wide channels. The total roughness of a channel reflects the rate of energy dissipation and incorporates a wide variety of factors, such as 1) resistance to flow offered by bed-forming particles, 2) bedforms, and 3) in-channel obstructions. These effects can significantly reduce the energy available for sediment transport.

The basal shear stress (τ_b) exerted on the channel bed by flowing water is the driving force for sediment transport. The average basal shear stress may be expressed as the product of fluid density, gravitational acceleration, flow depth, and water surface slope

$$\tau_b = \rho_w g D S \tag{7}$$

The fraction of the basal shear stress available for sediment transport, defined as the effective boundary shear stress (τ'), depends upon the amount of in-channel roughness and energy dissipation, as discussed above. The critical shear stress (τ_c) represents the shear stress necessary to mobilize a given gram size (d) and is expressed as

$$\tau_{cr} = \tau_* (\rho_s - \rho_w) g d \tag{8a}$$

where ρ_s is the sediment density and τ_* is a dimensionless critical shear stress (Shields, 1936). Coarse-grained channels often exhibit a threshold for significant gram mobility that is associated with a dominant discharge (Lane, 1953; Henderson, 1963; Andrews, 1984). In gravel- and cobble-bed channels the bankfull stage is the dominant discharge responsible for establishing channel morphology and accomplishing most sediment transport (Wolman and Miller, 1960, Andrews, 1980; Knighton, 1984). To a first approximation, many such streams can be viewed as bankfull threshold channels (Henderson, 1963; Li et al., 1976; Jackson and Beschta, 1982; Richards, 1982; Carling, 1987), with a threshold of bed mobility characterized by the critical shear stress of the median surface grain size (d_{50})

$$\tau_{cr50} = \tau_{*50} (\rho_s - \rho_w) \, g \, d_{50} \tag{8b}$$

Transport of bed material occurs at discharges which generate effective boundary shear stresses greater than the critical shear stress ($\tau' > \tau_c$), but the frequency of the bankfull discharge varies for different channels (Williams, 1978).

Modes of sediment transport include both suspension of grains within the flow (suspended load) and rolling, sliding, and saltating of grains near the channel bed (bedload). Suspended load typically accounts for the majority (>90%) of transported sediment (e.g, Richards, 1982), but bedload transport dominates channel morphology. Numerous expressions have been developed (Gomez and Church, 1989) describing bedload transport as a non-linear function of such inter-related hydrologic variables as shear stress, discharge, velocity, and stream power. One conceptually simple equation expresses the bedload transport rate (Q_b) as a function of the difference between the effective basal shear stress and the critical shear stress (Meyer-Peter and Müller, 1948)

$$Q_b = k \, W \, (\tau' - \tau_{cr})^{1.5} \tag{9}$$

where k is a constant. While me bedload transport rate from a reach determines the bedload supply to the next downstream teach, the bedload transport rate of any given teach is a function of both the transport capacity and the input of transportable material. The dependence of basal shear stress on flow depth, and thus discharge, indicates that a significant change in discharge directly influences sediment transport, channel bed stability, and channel scour.

The depth of scour defined by the thickness of the active transport layer is correlated with hydraulic discharge (Emmett and Leopold, 1963; Leopold et al., 1966, Hassan, 1990). However, Carling (1987) noted only a weak correlation of scour depth and discharge for a composite data set from channels with distinctly different grain sizes. This suggests that relations between discharge and depth of scour may be site specific.

Leopold and Emmett (1984) further linked the depth of scour to the bedload transport rate. Continuity requires that the thickness of the active transport layer is a function of the bedload transport rate

$$d_s = Q_b / u_b W \rho_s (1-p) \tag{10}$$

where d_s is the mean depth of scour, u_b is the average bedload velocity, and p is the bed porosity (Carling, 1987).

Potential response to altered sediment supply also is governed by sediment continuity through a channel reach. Any difference between the amount of sediment entering (Q_s) and leaving (Q_{out}) a reach is equal to the difference between the sediment supply (Q_s) and the transport rate and must be accommodated by a change in the amount of sediment stored (S_b) within the reach

$$Qs\text{-}Qout = \Delta S_b \tag{11}$$

If mom sediment is transported into a channel reach than it can transmit, then the amount of stored sediment must increase. Barform roughness reflects the amount of material stored in the channel and in turn influences sediment transport rates.

Both the discharge and sediment input to a channel reach are inherited as the output from the reach immediately upchannel. Although the sediment supply is intimately related to discharge (as discussed above), both are imposed factors to which the other variables in equations (4-11) respond (e.g., Leopold et al., 1964). In response to changes in sediment supply or discharge, a channel may: widen or deepen; change its slope through aggradation, degradation, or changes in sinuosity; alter macroscopic bedforms or the particle size of its bed and thus change the frictional resistance of the bed; or alter the depth of the active transport layer, defmed by the depth of channel scour. Each of these potential channel responses reflects the basic interplay among sediment supply, transport capacity, and sediment storage.

Conceptual Models of Channel Response

The variety of potential channel responses complicates both the prediction of specific channel response and the reconstruction of past channel changes. In spite of these complications, relations like (4-11) allow conceptual models of channel response to changes in sediment load or hydraulic discharge. Drawing on both theoretical and empirical arguments, previous workers developed conceptual models for predicting channel response to changes in discharge and sediment load supplied from upstream.

Gilbert (1917) originally hypothesized that a stream flowing over an alluvial bed is just as steep as is necessary to transport its imposed load. This hypothesis implies that where the channel slope is insufficient to transport the delivered bedload material, it is deposited and the local channel slope increases until sufficient to transmit the supplied load. Where the channel slope is greater than the minimum necessary to transport the load, the channel incises and the slope decreases. This implies that bedload discharge Q_b and channel slope S are proportional, or that

$$Q_b \propto S \qquad (12)$$

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 are adjusted to hydraulic discharge and slope through

$$Q_b\, d_{50} \propto Q\, S \qquad (13)$$

Based on this expression, Lane (1955) argued that changes in discharge or bedload transport are accommodated by corresponding changes in channel slope or sediment size. Equation (15) is more complete than Gilbert's hypothesis, but neither model accounts for the ability of a channel to change its basic geometry while preserving sediment transport rates and substrate size.

Schumm (197 1) combined empirical relations between discharge, bedload transport, and other descriptive morphological variables into general relations for channel response

$$Q \propto (W\ D\ \lambda)\ /\ S \qquad (14)$$

and

$$Q_b \propto (W\ \lambda\ S)\ /\ (D\, p) \qquad (15)$$

where λ is channel meander wavelength and p is sinuosity. Schumm (197 1) used these relations to define a series of expected channel responses to changes in sediment transport or discharge. Nunnally (198.5) later elaborated on this approach to yield a similar set of relations including d_{50}. These relations form a complex web of possible channel responses to altered conditions and include the dominant factors describing the overall geometry of a channel. However, the factors listed in (14) and (15) are only a few ways in which channels respond to perturbation. Since these relations are based on large alluvial rivers, several additional factors are needed to develop general models for channel response.

Considering the general relationships between sediment supply, discharge, and the variables in equations (4- 11) we can incorporate these additional factors into conceptual channel response models more appropriate for mountain drainage basins, as summarized below:

$$Q \quad \propto \quad Q_b \, d_s \, W \, D \, d_{50} \, / \, R \, S \, S_b \qquad (16)$$

$$Q_s \quad \propto \quad Q_b \, d_s \, W \, S \, S_b \, / \, R \, D \, d_{50} \qquad (17)$$

Some variables listed in (16) and (17), such as W or S_b, may respond only above a maximum transport capacity threshold, while others, such as d_{50}, may have a continuous response potential. When using relations such as equations (16) and (17) it is important to remember that bedload transport and sediment supply are related to discharge. While these relations allow a general prediction of the suite of possible channel changes, it is not always straightforward to reconstruct causes of channel change since a given response could reflect changes in either sediment supply, discharge, or some combination thereof. Furthermore, concurrent changes in discharge and sediment load may lead to conflicting response potential. Consequently, attributing channel change to altered discharge or sediment supply may not be possible without independent constraints on one of these factors. In order to predict channel response to future perturbation one needs to consider which of the possible disturbances and channel responses are most likely for a particular channel. Although the predictions of (16) and (17) are applicable throughout channel networks, the particular style and magnitude of response vary with channel type and pre-existing conditions.

These relations also illustrate a fundamental problem in predicting or reconstructing channel response: them am. seven variables, but only two relations! Fortunately, channel processes and morphology impose constraints on potential responses. In the next section we examine empirical studies, which combined with theoretical considerations provide the foundation for a process-based channel classification that systematizes channel morphology and response potential.

CASE STUDIES OF CHANNEL CHANGE

Reported observations of channel change highlight common channel responses and provide a large body of empirical evidence against which to test conceptual models. Evidence from a wide variety of environments supports the general predictions of relations (16) and (17). Much of this previous work focused on large, low-gradient channels, although a few workers studied smaller, high-gradient channels. The following section reviews studies that illustrate channel response to altered sediment supply, discharge, vegetation, and dam construction.

Sediment Supply

Human activities and natural processes affect the amount, distribution, and frequency of sediment transport both to and within stream channels. While all sediment transport processes are episodic over some time scale, channel response to sediment inputs depends on the ability of the channel to transport material relative to the sediment supply. In accordance with the predictions of (17), 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. Channels with a high sediment supply often exhibit a braided morphology with multiple active channelways defined by longitudinal and medial bar forms (Leopold and Wolman, 1957). Smith and Smith (1984) documented channel braiding in response to a massive increase in sediment supply along the William River in Saskatchewan Canada. They report that the channel abruptly changes from a single thread to a braided channel five times as wide and half as deep in response to a large increase in bedload as the channel traverses a sand dune field. Over this distance, other factors influencing channel width, depth, and pattern do not change appreciably. Other workers also report that channels with a high supply of coarse sediment am braided, whereas those with a restricted supply of coarse sediment are meandering (e.g., Harvey, 1991).

Temporal variations in sediment supply also influence channel form. A number of case studies illustrate progressive downstream aggradation and subsequent degradation in response to an episodic increase in sediment input. Gilbert (1917) reported the effects of huge additions of hydraulic mining debris to rivers in the foothills of the Sierra Nevada of California from the early 1850's to the 1880's. Aggradation occurred sequentially throughout the downslope channel network as the mining debris was gradually transported through the system Locally, channel aggradation approached 40 meters by the late 1870's (Whitney, 1880). Subsequent reincision of the channels was still occurring just after the turn of the cenmry (Gilbert, 1917) and some channels in the Sierra Nevada were still responding in the 1980's (James, 1989, 1990), over one hundred years after hydraulic mining ceased

This general pattern of responding to, and recovering from, increased sediment supply has been observed in many other situations. For example, Madej (1978; 1982) studied the Big Beef Creek watershed in western Washington and estimated that sediment transport rates approximately tripled in response to disturbance from certain land use changes. Disturbance associated with logging activity increased sediment delivery to channels and generated an aggradational wave that took 20 to 40 years to pass through the watershed. Channel widening also was atttibuted to increased sediment supply through comparing the existing hydraulic geometry of Big Beef Creek with surveys of channels in similar nearby watersheds.

Channel changes resulting from the 1964 floods in northern California and southern Oregon illustrate another well.-documented example of channel response to major sediment inputs. Channel widths increased by 100% at some gaging stations and channel beds aggraded as much as 4 m (Hickey, 1969; Janda et al., 1975; Kelsey, 1980; Lisle, 1981; 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 1 km/yr. Lisle (1982) reported that pool infilling decreased channel roughness and accelerated sediment transport within aggraded reaches. The mean size of bed material also decreased in response to aggradation (Nolan and Janda, 1979; Kelsy, 1980). Kelsey (1980) further noted that aggradation reduced pool spacing. 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 storm reflected increased sediment delivery from hillslopes and disturbance of riparian vegetation (Lyons and Beschta, 1983; Grant et al., 1984). Debris flows also scoured many steep channels to bedrock (Grant et al., 1984). Significant flood-delivered material is still stored in low-gradient reaches and the channel is braided in places where it is incising and reworking flood &posits (Sullivan et al., 1987). Such changes in sediment storage within a channel system may persist for decades as sediment is gradually transported from the reaches in which it accumulated.

The South Fork Salmon River in central Idaho presents another recently summarized example of impact and recovery from significant sediment inputs (Sullivan et al., 1987). Severe storms in the early 1960's following extensive logging and road construction dramatically increased sediment loads which resulted in pool filling, burial of gravels with sand, decreased bed roughness, and fining of the channel bed (e.g., 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 the subsequent thirteen years showed progressive reincision (Megahan et al., 1980), as pools were re-excavated and sand was transported out of spawning gravels (Platts et al., 1989). This reinforces the argument for a general re-establishment of original channel morphology after sediment supplies decrease.

An important characterisic 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 in 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). Scour in steep channels and aggradation in lower-gradient channels characterized channel response (Nolan and Marron, 1985; 1988). Debris flows and high discharges scoured many of the channels steeper than 6°, resulting in major sediment delivery to lower-gradient channels. Channel response in intermediate-gradient channels was variable, with significant local aggradation associated with landslide &position (Nolan and Marron, 1988). In many of these channels, sand infilled pools, buried riffles, and filled the interstices between coarse bed material (Coats et al., 1985; Nolan and Marron, 1988). Substantial aggradation and overbank deposition also occurred along steep channels that did not directly receive significant landslide debris (Nolan and Matron, 1988). Later that winter, subsequent stream flows in steep- and intermediate-gradient channels scoured much of the aggraded sediment and redistributed it downslope. In lower-gradient channels, pool filling and riffle burial persisted for a longer time (Coats et al., 1985). illustrating a strong difference in the style and persistence of channel response at different locations in the drainage network.

Changes in sediment supply also influence the character of the channel bed. Perkins (1989) studied the effect of landslide-supplied sediment on channel morphology in Salmon Creek in southwestern Washington. Based on considerations of the relation between sediment transport capacity and sediment supply she argued that accelerated sediment delivery increases the amount of material stored in bedforms (expanding bar volumes at the expense of pool volumes, for example) and decreases the average sediment size in the teach. She argued that elimination of landslide-supplied sediment results in a long-term decrease in the amount of material stored on the bed and a greater degree. of bedrock control on bed morphology. Her study provides an excellent example of how channel form and sediment storage may reflect a balance between sediment supply and transport capacity.

The size of bed surface material also responds to changes in sediment supply. Knighton (1991), for example, reported that channel response to large inputs of fiie sediment involved both a wave of aggradation and a general fining of bed material. After passage of the wave, the channel substrate coarsened as the bed degraded toward its original condition. 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 produced scour of previously stable gravel riffles. They also showed that the d_{50} of the flume bed decreased with increased sediment transport (Jackson and Beschta, 1984). Dietrich and others (1989) proposed that for gravel-bed channels the degree of bed surface coarsening reflects the relation between sediment supply and the ability of the channel to transport the imposed sediment load In a series of flume experiments they (Dietrich et al., 1989) showed that a decrease in sediment supply resulted in surface armoring and constriction of the zone of active sediment transport. They further proposed a dimensionless ratio of the sediment transport rate for surface and subsurface particles (q_*) in order to characterize the transport efficiency of a stream. Values of q_* equal unity (poorly armored) when the sediment supply rate matches the transport capacity of the channel, and decrease toward zero (well armored) as sediment supply declines relative to transport capacity. Some empirical evidence from natural channels supports this hypothesis. Kinerson (1990) reported high q_* values for channels with a high sediment supply and a low q_* for channels with low sediment supply. Lisle and Madej (in press) also report a high q_* and poorly developed surface armoring in a channel with a high sediment supply. Consequently, the character of the channel bed relative to the subsurface sediment may provide an indication of the sediment supply relative to the channel transport capacity.

Discharge

Changes in the magnitude and frequency of the discharge a channel conveys 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. Climatic change provides the most direct precipitation-related impact on discharge in channel networks. Opportunities to monitor the influence of climate change on channels are rare, but studies of dry valleys (e.g., Gregory, 1971; Gregory and Gardner, 1975) and channel initiation (Montgomery and Dietrich, 1992; in press) suggest that channel network extent expands and contracts in response to climatic forcing. The following summary of studies illustrates channel response to discharge variations in downstream portions of contemporary channel networks.

The impact of land management activities 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, in tum, increases the proportion of rapid surface runoff at the expense of infiltration (Leopold, 1968). Channel response to urbanization typically involves channel expansion through an increase in either channel width or depth as a response to increased discharge. 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. Other workers also found significant channel widening and incision as a result of urbanization in both humid (e.g., Leopold, 1973; Graf, 1975; Gregory and Park, 1976; Neller, 1988; 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, the vegetation that aggressively recolonizes a cleared forest may increase evapotranspiration (e.g., Harr et al., 1979) and may even reduce discharges below original levels (Harr, 1983). Although they may be very important biologically, changes in low-flow conditions am generally unimportant for channel morphologic response. Increases in the peak runoff caused by rain-on-snow events (Harr, 1981; 1986, Berris and Harr, 1987; Coffin and Harr, 1991) in clear cut areas may significantly affect channels due to the possible change in either the frequency or magnitude of the channel-forming discharge. Channel responses to rain-on-snow events in cleared areas include bank erosion, channel incision, and mobilization of both bedload and large, in-channel organic debris (Harr, 1981). These effects are similar to those occurring from natural large-discharge events, but a change in their frequency could impact biologic systems and the reach-level sediment transport rate.

Dams

Dam construction changes both the discharge regime and sediment supply of downstream channels leading to channel incision, constriction or widening, and changes in channel substrate. Many workers report channel incision immediately downstream of dams in response to decreased sediment supply (e.g., Gilbert, 1917; Lane, 1934, Leopold and

Maddock, 1953; Wolman, 1967; Stanley, 1972; Galay, 1983; Williams and Wolman, 1984). Tributary channels also may incise in response to mainstream channel incision (Bray and Kellerhals, 1979) through upstream knickpoint propagation from their confluence (Germanoski and Ritter, 1988). The decreased discharge below a dam may cause a narrowing of the active channel width (e.g., Leopold and Maddock, 1953; Wolman, 1967; Petts, 1977; 1979; Gregory and Park, 1974; Gregory, 1976; Park, 1977; 1981; Rahn, 1977; Williams and Wolman, 1984; Sherrard and Erskine, 1991) and aggradation of bedrock channels (e.g., Allen et al., 1989). Several workers noted, however, that some channels widen in response to dam construction, while other channels exhibit little change in width (Leopold and Maddock, 1953; Petts, 1979; Williams and Wolman; 1984). This difference in response may reflect the erodibility of channel bank material and the available size and sources of sediment. Channel incision in response to reduced sediment supply may lead to bed armoring downstream of dams (e.g., Williams and Wolman, 1984). Although a paucity of coarse sediment may limit the potential for armoring (e.g., Sherrard and Erskine, 1991), coarsening of channel bed material occurs downstream of many dams (e.g., Little and Mayer, 1976; Shen and Lu, 1983; Bradley and Smith, 1984; Kinerson, 1990). In contrast, a number of workers (e.g., Davey et al., 1987; Petts, 1988) report accumulations of fine sediment downstream of dams in gravel-bed channels, presumably due to a decrease in flows that previously flushed finer sediment. Considered together, these studies indicate that a wide variety of potential channel responses may be expected downstream of dams due to coincident alterations of both sediment supply and discharge. The specific effects likely for a given channel reflect, among other factors, the size and availability of downstream sediment and the nature of the bank-forming materials.

Vegetation

Modification of vegetation growing within, along, and near channels may induce changes in channel geometry or sediment storage and transport. For example, channel enlargement and the dissection of unchanneled valleys in most parts of the United States has been ascribed to overgrazing and removal of vegetation from valley floors (e.g., Rich, 1911; Bryan, 1928; Antevs, 1952; Brice, 1966; Daniels and Jordan, 1966; Costa, 1975; Cooke and Reeves, 1976). The increased erosion and greater drainage density resulting from channel network expansion may greatly increase sediment delivery and alter hydrograph characteristics for downstream channels. The root strength of vegetation growing along channel banks contributes to bank stability, especially in relatively uncohesive alluvial deposits. Consequently, alteration of riparian vegetation can trigger channel destabilization (e.g., Kondolf and Curry, 1986). 'Viewed in this context, vegetation growing near the channel plays an integral role in determining the channel geometry. Gregory and others (1991) discuss the biological importance of riparian corridor/stream channel interactions.

Vegetation and debris located within the channel may also strongly impact both the style and amount of sediment storage, as well as the overall channel roughness. Large woody debris (LWD), for example, provides transient storage sites for bed material (e.g., Heede, 1972; Beschta, 1979; Keller and Swanson, 1979; Mosley, 1981; Pearce and Watson, 1983; Bilby, 1984), stabilizes gravel bars (Lisle, 1986), and can provide hydraulically sheltered locations that allow fine sediment to accumulate (e.g., Zimmerman et al., 1967; Megahan, 1982). The morphology of smaller channels may respond dramatically to changes in the input, transport, and decay of LWD. For example, removal of large organic debris from small channels rapidly accelerates sediment transport (e.g., Beschta, 1979; Bilby, 1984; Smith et al., in press). Pool morphology and areal extent in some channels are strongly correlated to LWD loading (Smith and Buffington, 1991). Changes in the supply of LWD to a channel may trigger significant changes in sediment storage, channel roughness, and pool morphology.

Summary of Case Studies

The studies summarized above document a wide range of channel responses to changes in sediment supply, discharge, and vegetation. Increased sediment supply can induce channel widening and aggradation, decrease roughness through pool filling, and decrease bed sediment size. These responses am consistent with those predicted by (18). Increased discharge can cause channel widening, incision, and bed armoring, effects predicted by (17). Channel response to dam construction and vegetation alteration illustrates potential effects of covarying changes in discharge and sediment supply. Collectively, these case studies illustrate that (17) and (18) provide a reasonable conceptual framework for examining potential channel response to perturbation.

CHANNEL CLASSIFICATION SYSTEMS

A channel classification imposes order on the wide range of morphologies found in natural streams based on similarities of form or function. The need to address potential changes in fluvial systems has fueled a proliferation of channel and valley classification systems (e.g., Kellerhalls et al., 1976; Paustian et al., 1983; 1992; Rosgen, 1985; Frissell et al., 1986, Frissell and Liss, 1986, Cupp, 1989; Frissell, 1992). A classification system capable of assessing potential channel changes should differentiate streams based on both processes acting within them and their potential response to changes in either discharge or sediment loading. While there are many ways to organize a channel classification scheme, there are several essential criteria necessary for a geomorphic channel classification:

1) It should be relatively universal so that it is applicable on more than a regional basis; otherwise proliferation of regional channel classifications will impede, rather than enhance development of communication and understanding.

2) It should encompass the whole channel network. By considering only streams of biological importance, channels with desirable or sensitive attributes, or channels larger than an arbitrary size (e.g. perennial channels) one ignores the reality that each reach of a channel network inherits the products of processes acting in upslope reaches.

3) It should be process-based and rely on those aspects of channel form that reflect channel processes.

4) It should be predictive, allowing assessment of likely channel response to natural and anthropogenic environmental change.

5) It should be compatible with landscape classifications to facilitate integration with other land management goals.

Existing Ch el Classifications

A number of stream classifications have been proposed, but most am designed for large, low-gradient alluvial rivers. Mosley (1987) and Naiman and others (1991) provide recent reviews of a number of stream classifications used by geomorphologists, engineers, and ecologists. -Some of these systems have predictive power, others are purely descriptive. In our opinion, none are ideal for broad use as a foundation for either assessing channel conditions or predicting changes in channel morphology. There is a need to systematically organize the variations in channel morphology and sediment transport that characterize stream channels in environments ranging from mountain drainage basins to continental-scale river systems. Ideally, such a classification would provide a context for understanding channel processes from headwater channels to major rivers.

Systems for classifying channels can be traced as far back as Davis' (1899) conception of youthful, mature, and old landscapes. Since then a number of more detailed geomorphic classifications have been proposed for large alluvial rivers. Perhaps the first process-oriented stream classification was presented by Melton (1936) who subdivided channels into those whose floodplains are formed primarily by meandering, overbank deposition, or braiding. Leopold and Wolman (1957) presented a quantitative basis for differentiating straight, meandering, and braided channel patterns based on relationships between slope and discharge; Brice (1984) later proposed the use of channel pattern to classify streams. Schumm (1963; 1977) classified alluvial rivers on the basis of whether their beds ate stable, eroding, or aggrading and he further differentiated them through the dominance of suspended load, mixed load, or bedload sediment transport. None of these systems are appropriate for classifying steep channels in relatively small drainage basins. Consequently, they provide only a limited context for channel classification.

The most widely-used channel classification scheme is the concept of channel order first proposed by Horton (1945) and later modified by Strahler (1957). In Strahler's system, the channel segment from the tip of the channel network to the first confluence is defined as a first-order channel. Second-order channels am downslope of the intersection of two first-order channels, and so on through the channel network. Within a given channel network, stream order correlates with channel length and drainage area (Horton, 1945; Schumm, 1956). Hence stream order provides an indication of relative channel size and position within the channel network. Stream order, however, is inadequate for comparisons between channel networks because the order assigned to the same channel reach depends upon the criteria used to determine where the upstream most channels begin. A wide. varietyof methods have been used to define the extent of first-order channels: blue lines on topographic maps (e.g., Horton, 1945; Strahler, 1957), the curvature of topographic contours (Morisawa, 1957; Smart and Surkan, 1967; Howard, 1971; Abrahams, 1980; Mark, 1983), a fixed gradient (e.g., Kmmbein and Shreve, 1970; Shreve, 1974) or drainage area (e.g., Band, 1986; 1989; Zevergergen and Thome, 1987; Morris and Heerdegen, 1988), and biologic criteria (e.g., Lotspeich, 1980; Lotspeich and Platts, 1982). Most networks so defined differ substantially from the channels identifiable in the field (e.g., Morlsawa, 1957; Maxwell, 1960, Montgomery and Dietrich, 1989). Moreover, the first-order blue line channels differ on different scale maps of the same watershed (e.g., Scheidegger, 1966). Even mom problematic for using stream order to classify channels is that drainage density differs from basin to basin and thus precludes directly comparing same-order channels between different basins because they may have significantly different drainage areas, discharges, and morphologies. Hughes and Omernik (1981) provide a further discussion of the shortcomings of using stream order in channel classification schemes. While channel ordering is a useful conceptual and organizational tool for channels within a watershed, we believe it is inappropriate as a foundation for a geomorphic channel classification.

Orsborn (1990) discussed methods for segregating channels on the basis of hydrologic relationships. He found that ratios of characteristic flows can be used to group channels into hydrologic regions and subregions. Such methods provide a useful technique to identify channels that deviate from expected regional relations. They do not, however, incorporate consideration of channel morphology and are thus best used in conjuction with, rather than as, a channel classification system

Gilbert (1914) proposed the first process-based channel classification when he recognized alluvial and bedrock, or "corroding", channels. He defined alluvial channels as those in which the bed of the channel is composed of material transported by the channel. Gilbert proposed that bedrock channels reflect a transport capacity in excess of the sediment supply and that alluvial channels represent a transport capacity less than, or equal to, the sediment supply. He further recognized that different portions of a channel network may be composed of different channel types and channel patterns.

Other classification systems am defined on the basis of channel characteristics. Howard (1980; 1987), for example, also described channels as either alluvial or bedrock on the basis of whether the channel bed has a layer of active alluvium. He further subdivided alluvial streams into sand-bed and gravel-bed channels corresponding to regime and threshold channels, respectively. The bed of a regime channel is highly mobile even at low flow and both sediment transport and bedform roughness change with increasing discharge, whereas the bed of a threshold channel typically is mobile only beyond some critical discharge. This classification provides insight into the potential response of different channel types, but is too broad to be useful for distinguishing between channels within mountain drainage basins.

Frissell and others (1986) proposed a channel network classification based on spatial and temporal scales of variation in fluvial systems. They distinguished watershed, stream system, reach, pool/riffle, and microhabitat levels into which stream systems could be classified [Orsbom and Anderson (1986) proposed a similar conceptual heirarchy]. Within each of these levels, processes acting over different spatial and temporal scales determine the channel network structure. This nested heiramhy provides a powerful conceptual approach for linking channel responses at different spatial scales (i.e., influence of channel pattern changes on bed microhabitat). However, it would be difficult to either predict potential channel changes, or assess current channel conditions based upon such a heirarchical classification system. Nonetheless, such a conceptual framework is crucial for identifying the dominant factors influencing a channel network at different temporal and spatial scales. In particular, it provides a context within which to examine connections between biological and geomorphic systems [see also Vannote and others (1980) and Gregory and others (1991)].

Bisson and others (1982) defined a system for visually classifying in-channel habitat into different types of pools and riffles which they termed channel units. They (Bisson et al., 1982) reported complex associations of channel unit utilization by different fish species and recognized six types of pools and three types of riffles on the basis of position in the channel, flow characteristics, and flow-controlling features. These channel unit definitions are determined at low flow and are to some degree stage dependent, While each of these units describes a distinct variety of in-channel habitat, this classification is not useful in a predictive sense because of the association of pool types with the type and orientation of specific in-channel flow obstructions. While classification systems based on the channel unit level may be useful for interpreting the availability and quality of fish habitat, they need to be coupled to a process-based classification system in order to provide a basis for predictive channel characterization.

Several stream classification systems are applicable to the full suite of channel types present in natural channels and are presently used to classify streams and delineate similar channel types for mom detailed biologic, engineering, or geomorphic assessments. Kellerhals and others (1976) classified channels by channel pattern (e.g., straight, sinuous, meandering), the frequency of islands (e.g., occasional, frequent, braided), and bar type (e.g., side channel, mid-channel, or point). They also considered river valley features such as the relation of the channel to valley walls, land use, and surficial geology. Rosgen's (1985) stream segment classification is based on stream gradient, width to depth ratio, substrate size, channel confinement in a valley, and assessment of near-channel landform stability. Rosgen's system has been used extensively by the U.S. Forest Service, but it neither directly differentiates channels based on bed morphology (as opposed to sediment size), nor on sediment transport processes. It does, however, provide a comprehensive description of the channel that provides a solid foundation for channel assessment and restoration.

Valley segment classifications (e.g., Paustian et al., 1983; 1992; Frissell and Liss, 1986; Cupp, 1989) based on valley cross-sectional shape, valley bottom gradient, channel pattern, and channel confinement (expressed as the ratio of channel width to valley bottom width) commonly are used to stratify channels in biologic studies. Cupp's (1989) system, for example, has been used to guide extensive habitat monitoring efforts in the State of Washington (Ralph et al., 1991). Of these approaches, only Frissell and Liss' (1986) approach implies a process-based interpretation. Further information, rules, or guidelines am necessary to use these classification systems for predicting either potential channel response or assessing current channel conditions. A process-based classification should relate both channel morphology and potential channel responses to the processes governing channel form. The premise underlying this assertion is that if one can identify the types of processes operating in a given channel reach, then one may assess potential channel response(s) to specific Perturbations. A method for organizing differences in channel processes would be useful for delineating potential channel responses within a channel network. Bradley and Whiting (1991) proposed a classification system for small stream channels in mountainous terrain based on valley side slope gradients, valley to channel width ratios, and the relation between channel substrate size and boundary shear stress. Their system represents a fit attempt to generate a process-based classification applicable to streams found in mountain drainage basins.

PROPOSED LANDSCAPE AND CHANNEL CLASSIFICATION

In the following sections we develop a channel classification scheme within a broader landscape context that provides a process-based framework for geomorphic, biologic, and land management applications. Channel types are delineated based on channel morphology, sediment transport processes, and sediment flux characteristics as controlled by hydraulic discharge and sediment supply. Combined with descriptions of channel confinement, LWD loading, and potential debris flow impacts, these channel types allow prediction of potential channel response to altered sediment loads or discharge.

Channel classification encompasses a range of scales over which different factors influence channel properties. A natural division of scales that reflects differences in processes and controls on channel morphology is given by i) geomorphic province, ii) watershed, iii) valley segment, iv) channel reach, and v) channel unit scales. Channel morphology at each of these scales is related, but reflects different levels of resolution. Most importantly, each level of this spatial hierarchy provides a framework for comparing channels at finer spatial scales.

Geomorphic Province Level

Climate, geologic history, and bedrock geology influence channel morphology through large-scale controls on discharge, vegetation and the nature, amount and spatial distribution of sediment supply. Geomorphic provinces are regions of similar land forms that reflect comparable hydrologic, erosional, and tectonic processes. Geomorphic provinces thus reflect broad controls on channel processes. Typically, they are bounded by major physiographic, climatic, and geological features. They consist of watersheds of roughly similar relief and climate developed in rocks that share some geologic affinity. Thus the general controls on channel processes and morphology (Figure 1) are reasonably similar for most watersheds within a geomorphic province. Consequently, the geomorphic province level provides bounds on what channels are potentially comparable in terms of relations between drainage area and discharge, sediment supply, and substrate size. The geomorphic province level is a first-order analysis tool, useful for stratifying areas within which comparable watersheds may be found, but it is too general to be useful for predicting specific channel responses.

Watershed Level

A watershed is the drainage area upslope of any point along a channel network. The watershed defines the natural boundary for the processes generating sediment and runoff into channels and is the only logical organizational unit for channel networks. Although the scale of the watershed-level classification ultimately must be tailored to the problem at hand, for many purposes, watersheds of about 100-500 km^2 provide practical planning units. For watershed-level classification, we suggest differentiating major watersheds within geomorphic provinces. However, some large rivers may traverse several geomorphic provinces. Classifying watersheds based on similar geologic history (glaciated or unglaciated), rock type (highly fractured sedimentary rocks or competent igneous rocks), and land use (agricultural, forestry, park) can identify channel networks either well, or ill-suited for comparison. Watersheds may be subdivided into areas in which geologic materials and climate impose similar controls on channel processes.

At the most general level, a watershed can be divided into hillslopes and valleys independent of the local influences discussed above (Figure 2). Different hillslope and valley morphologies reflect fundamental differences in sediment production and transport processes. Further subdivision of valleys define higher-resolution levels of our general landscape classification.

Hillslopes

Hillslopes are the undissected portions of a landscape that are zones of sediment production and transport to valleys. Sediment flux off of hillslopes may be either transport or production (weathering) limited (e.g., Selby, 1982). On production-limited hillslopes soil transport exceeds the rate at which soil is produced and made available for transport. Soil cover is thin, if present, and slope form is controlled largely by the properties of the underlying bedrock. Thus, the soil production rate primarily controls downslope sediment flux. In contrast, soil flux on transport-limited hillslopes is less than, or equal to, the rate of bedrock weathering. Consequently, soil profiles develop and slope form is controlled by the properties of the material overlying fresh bedrock. On soil-mantled hillslopes, bedrock is converted to soil that is gradually transported downslope through the action of processes such as rainsplash, soil creep, and biologic activity. Sediment transport on such slopes may be thought of as a conveyor belt that gradually collects material as it moves downslope until the accumulated material is delivered across a channel bank. Sediment flux on transport-limited hillslopes is controlled primarily by local slope (e.g., Gilbert, 1909; Kirkby, 1971), rather than the availability of material susceptible to erosion. In general, slopes in arid environments tend to be weathering limited, whereas slopes in humid regions tend to be transport limited. However, sediment transport rates also may exceed sediment production in steep portions of humid landscapes. The presence of either thin soils and bedrock, or a well-developed soil mantle at the ground surface distinguishes production- and transport-limited hillslopes.

Valleys

Valleys am regions of the landscape that focus runoff and sediment transport through downslope topographic convergence. Networks of valleys collect and redistribute sediment delivered from adjacent hillslopes. Depending on valley position within a landscape, sediment transport and valley evolution am dominated primarily by either fluvial or mass wasting processes. Process distinctions associated with different valley morphologies are discussed further at the valley segment level of the channel classification.

Vallev Segment Level

Valley segments define portions of the valley network with similar morphologies and governing geomorphic processes. Based on the nature of the valley fill, sediment transport processes, channel transport capacity, and sediment supply, we recognize three terrestrial valley segment types: colluvial, bedrock, and alluvial (Figures 2 & 3). These divisions are similar in spirit to Frissell and Liss' (1986) valley segment classification developed for the Oregon Coast Range. Because we choose to focus on terrestrial networks we have not included estuarine valleys in our classification scheme; however, we do recognize estuarine valleys as important links between terrestrial and marine environments. More elaborate valley segment classifications (e.g., Paustian et al., 1983; 1992; Cupp, 1989) may be useful for linking channel classification with resource assessments, but they often are region specific. Furthermore, many of the features included in such systems are, in our opinion, best included in a reach level classification. The following sections describe our valley segment types.

Colluvial Valleys

Colluvial valleys may be channeled or unchanneled, but fluvial transport is relatively ineffective in transporting sediment delivered from surrounding hillslopes. In steep landscapes debris flow processes dominate colluvial valley morphology, whereas in low-gradient landscapes colluvial valleys are maintained by periodic expansion of the alluvial channel network. In general, colluvial valleys are those in which colluvial fiis accumulate and am perodically excavated.

Unchanneled Colluvial Valleys (hollows)

An unchanneled valley morphology implies that fluvial sediment transport is insufficient to initiate and maintain a channel and thus is unlikely to accomodate the sediment supply delivered from surrounding hillslopes. Unchanneled valleys (regionally referred to as hollows, swales, headwalls and a host of other terms) extend upslope of the finest scale channels in many soil-mantled environments. They are defined as topographic valleys lacking evidence of sediment transport concentrated within well-defined banks. The downslope transition from an unchanneled valley to a first-order channel represents the initiation of channel processes (Dietrich et al., 1986; Montgomery and Dietrich, 1988; 1989, 1992). Classification schemes are available for describing unchanneled valleys (see discussion in Montgomery and others (1991)), but we do not subdivide them further in our landscape classification.

Hollows function as sediment storage sites that gradually accumulate colluvial soils transported from surrounding hillslopes and periodically deliver the stored sediment to downstream channels (Dietrich and Dunne, 1978). Topographic convergence concentrates hydrologic response and elevates pore pressures along hollow axes (e.g., Dunne and Black, 1970; Anderson and Burt, 1978; Pierson, 1980; Wilson and Dietrich, 1987; Petch, 1988; Montgomery, 1991). making hollows a primary source of debris flows in steep landscapes (Figure 4). 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; Dietrich et al., 1982; 1986; Reneau et al., 1986). The depth of colluvium in a hollow is then a function of both the rate at which sediment is delivered to the hollow and the time since the deposit began accumulating (Reneau et al., 1984; 1986; 1990; Matron, 1985; Benda and Dunne, 1987). Intermittent sediment delivery from hollows to channels punctuates the more continuous supply of sediment delivered by hillslope transport across channel banks. The association of hollow failure with infrequent hydrologic events suggests that sediment flux from hollows is transport limited. On the other hand, a number of workers have argued that the colluvium in a hollow needs to exceed a minimum thickness in order to fail (Dietrich and Dunne, 1978; Okunishi and Iida, 1981, 1983; Montgomery, 1991) and that the colluvium within a hollow is not always available for transport. For the present discussion, however, we consider hollows to be a transport-limited source of episodic sediment delivery to channels.

Channeled Colluvial Valleys

Channeled colluvial valleys typically occur just downslope of hollows. However, colluvial channels may also occupy the tips of the channel network in low-gradient landscapes, or they may occur where transport capacity decreases rapidly downslope, such as where tributary channels flow across the floodplains of larger channels. Because of me nature of the valley fill, the streams within these valleys am termed colluvial channels.

Little research has focused on these small headwater channels, even though first-order channels compose approximately half of the total channel length in a channel network (Montgomery, 1991). Especially in mountainous terrain, colluvial channels are important linkages between sediment production on hillslopes and its delivery to downslope alluvial channels. Flow in colluvial channels is generally shallow and ephemeral. Consequently, basal shear stresses may be insufficient to mobilize much of the colluvial sediment introduced to the channel, resulting in storage of this material (Benda, 1990). Soil creep, tree throw, burrowing, and small-scale slope instability introduce sediment into colluvial channels from across channel banks. Intermittent flow may rework some portion of the surface of the accumulated material, but does not govern deposition, sorting, or transport of the majority of the bed-forming material. Large clasts, woody debris, bedrock steps, and in-channel vegetation may further reduce the energy available for sediment transport. Thus, fluvial processes have little influence on long-term channel form and valley development. Instead, intermittent scour by debris flows may govern valley form and incision.

Debris flows from upslope hollows might either scour or deposit material in a colluvial valley_depending on the size, velocity, and viscosity of the debris flows, as well as the effectiveness of flow obstructions. Debris flow scour may convert a colluvial valley into a bedrock valley. Unlike a hollow, fluvial processes may maintain a channel as the valley gradually refills with colluvium. Thus, sediment delivery to colluvial valleys occurs from relatively continuous delivery by hillslope processes and from intermittent mass wasting. The frequency of sediment mobilizing discharges or debris flows may thus determine the amount of sediment stored in colluvial valleys. Large volumes of sediment accumulate between infrequent scouring events, implying that sediment transport from colluvial valleys is transport-limited

We have noted that the grain size of material on colluvial channel beds is liner than found in downslope channels. This contrasts with the downstream fining typical within fluvial channels. With increasing downslope discharge, however, finer materials are selectively transported from the bed surface, resulting in bed coarsening and the dominance of fluvial processes. The maximum surface d_{50} along a channel network may indicate the transition from a dominantly fluvial to a mass wasting control on valley incision and form (Figure 5).

Bedrock Valleys

Bedrock valleys are characterized by confined fluvial channels in which a contiguous alluvial bed is ahsenr Some alluvial material **may** be temporarily stored in scour holes, or behind flow obstructions, but in general the channel bed lacks an alluvial cover and there is little, if any, valley fill. Evidence from badland landscapes suggests that bedrock channels are steeper than alluvial channels with similar drainage areas (Howard and Kerby, 1983). Although similar data for larger channel systems is not available, bedrock channels presumably **lack** an alluvial bed due to high transport capacity associated with steep channel gradients and/or deep flow. Few observations are available concerning sediment transport in bedrock channels, but it is reasonable to adopt Gilbert's (1914) hypothesis that their transport capacity exceeds the sediment supply.

Two distinct varieties of bedrock valleys reflect different processes giving rise to bedrock channel morphologies. One represents a stable fluvial morphology, while the other is temporally variable. Fluvial channels that are steep enough am bedrock floored and will not undergo significant morphological change without dramatic external change in process rates or magnitudes. Such channels often are associated with knickpoints and lithologic controls. In contrast, low-order channels in steep landscapes may alternate between bedrock and colluvial morphologies in response to periodic debris flow scour. At any one time, the proportion of bedrock channels in the debris-flow-susceptible reaches of a watershed most likely reflects the history of debris flow activity in the catchment. Thus bedrock channels in low-gradient portions of a watershed reflect a high relative transport capacity, whereas those in steep debris-flow-prone portions of a watershed may also reflect recent debris flow scour.

Alluvial Valleys

Alluvial valleys am characterized by fluvial transport of sediment over a predominantly alluvial valley fill. To varying dgrees, the channels in alluvial valleys are capable of transporting and sorting the load supplied to them from upslope channels, **but** they do not have transport capacity sufficient to scour the valley to bedrock. Alluvial channels exhibit a variety of morphologies discussed in detail at the reach level. At the valley segment level they may be either confined, with little to no associated floodplain, or unconfined, with a well established floodplain. Both the specific channel morphology and degree of confinement reflect the local channel slope, sediment supply, and hydraulic discharge.

The valley segment level is useful for distinguishing valley morphologies on the basis of the dominant sediment transport processes (fluvial versus mass wasting) and general sediment flux characteristics (transport- versus supply-limited) and providing insight into the spatial linkages that govern watershed response to disturbance. These distinctions enable first-order response predictions for major sections of the network experiencing changes in discharge or sediment supply. However, alluvial channels exhibit a variety of morphologies, some of which appear functionally similar at the valley segment level, but that may actually respond differently to similar perturbations of sediment load and discharge. In particular, the valley segment level does not allow prediction of specific response to altered discharge or sediment supply. Consequently, we believe the reach level is the most useful for understanding both morphologically significant processes and response potential.

Channel Reach Level

A reach is a length of channel that exhibits a consistent association of bedforms, or channel units and is many channel widths long. Colluvial and bedrock valley segments contain corresponding morphological teaches; namely colluvial channels and bedrock channels. However, alluvial channels have a wide variety of bed morphologies and roughness configurations. Observationally these different morphologies and roughness elements systematically vary with slope and position within the channel network. We hypothesize that such variations in channel morphology and roughness are functions of sediment supply (size and quantity) and available transport capacity (shear stress or stream power). Consequently, each bed morphology and associated roughness elements represent the stable channel configuration for a given regime of sediment supply and shear stress. Because these bed morphologies reflect process differences, one would expect associated differences in potential channel response. With this in mind, we recognize six alluvial channel reach types: cascade, step-pool, plane-bed, pool-riffle, regime, and braided channels. Transitions between morphologies may be gradual in the field, as this classification imposes order on a continuum of natural morphologies. The following discussion describes morphologies and processes characterizing each channel type; channel response is discussed later.

cascade Channels

Cascade channels are the steepest alluvial channels in our classification and arc hypothesized to reflect the highest rates of energy dissipation. Most previous studies of steep alluvial streams focused on step-pool morphologies (discussed in next section), and have either not recognized or subsumed cascade morphologies in their analyses of step-pool channels. While the term "cascade" has been used in several of these studies, it has generally been synonymous with step-pool morphologies. In contrast we recognize cascade channels as morphologically distinct from step-pool channels. Because cascade morphologies are relatively unstudied, portions of the following discussion rely heavily on knowledge of the dynamics of step-pool morphologies and alluvial channels in general.

Our delineation of cascade channels focuses on high-gradient streams in which flow is strongly three dimensional, and energy dissipation is dominated by jet-and-wake flow and hydraulic jumps around individual large clasts (Figure 6). Cascade channels are characterized by longitudinally and laterally disorganized bed material, typically consisting of cobbles and boulders. The size of the largest grains may exceed the bankfull flow depth and individual bed elements provide the primary channel roughness by creating flow obstructions and local hydraulic jumps (Grant et al., 1990). In steep landscapes, cascade channels typically exhibit supercritical flow that follows a tortuous convergent and divergent path around individual clasts. Tumbling flow over and around individual large grains dissipates much of the mechanical energy of the flow. Some cascade channels exhibit small pools that span a portion of the channel width. Because they are typically located in the upper portions of the channel network, cascade channels may be scoured by debris flows originating upslope. Cascade bed morphology, however, is dominated by fluvial sediment transport between debris-flow events.

Large particle size relative to flow depth make the largest bed-forming material of cascade channels effectively immobile during typical flows. Studies of steep-gradient channels report that large bed-forming grains only become mobile during infrequent hydrologic events (Grant et al., 1990; Kondolf et al., 1991; Whittaker, 1987a; Schmidt and Ergenzinger, 1992; Wohl, 1992). Mobilization of these larger clasts is accompanied by high sediment transport rates (Sawada et al., 1983; Warburton, 1992), due to the release of finer sediment previously trapped under and around large grains.

While the bed-forming material of cascade channels generally is boulder- and cobble-sized, limited amounts of gravel and finer material axe often deposited in low-energy zones created by flow obstructions (Kondolf et al., 1991). Grains may be trapped behind or in the wake of larger clasts and buttressed by large woody debris (LWD) or deposited in associated velocity shadows. The bed material deposited in low-energy sites in steep-gradient channels is often believed to be characteristic of the bedload (Griffiths, 1980; Schmidt and Ergenzinger, 1992). One tracer study (Kondolf et al., 1991) showed that material in such depositional sites was completely mobilized during a 7 year recurrence interval event, while no tracer movement was observed during flows of less than annual recurrence interval.

The above observations suggest that them are two thresholds for sediment transport in cascade channels. At a moderate recurrence interval threshold bedload material is rapidly and efficiently transported over the more stable bed-forming clasts; these grains may in turn be mobilized during more infrequent events. However, we hypothesize that because of steep gradients and the nature of flow through cascade channels, as well as the paucity of depositional sites (Kondolf et al., 1991), most of the finer material introduced to the channel is rapidly transported downstream. Our conception of sediment transport in cascade channels is based, in great part, on the hypothesis that continuous supercritical flow rapidly transports all but the coarsest sediment supplied to the channel. Fine sediment that remains on the bed typically consists of material trapped behind obstuctions or deposited in local low-energy environments.

The lack of significant in-channel storage and the rapid scour of depositional sites during moderately-frequent high flows suggests that sediment transport in cascade channels is supply-liited. Bedload transport studies demonstrate that steep mountainous streams can be seasonally or stochastically supply-limited (Nanson, 1974; Ashida et al., 1981). Because of this high transport capacity relative to sediment supply, we propose that cascade channels function primarily as sediment transport zones that rapidly deliver the majority of their sediment supply to downstream lower-gradient channels.

Step-pool Channels

Step-pool channels are characterized by large clasts organized into discrete channel-spanning accumulations that form a series of steps separating pools containing finer material (Figure 7) (Griffiths, 1980, Ashida et al., 1981; Whittaker and Jaeggi, 1982; Wbittaker and Davies, 1982; Whittaker, 1987a;b; Chin, 1989; Grant et al., 1990). Step-forming clast sires are typically comparable to bankfull flow depths. Primary flow and channel bed oscillations in step-pool channels are vertical, rather than lateral, as in lower-gradient pool-riffle channels. Step-pool channels tend to exhibit a pool spacing of roughly one to four channel widths (Bowman, 1977; Whittaker, 1987; Chin, 1989; Grant et al., 1990), significantly less than the five to seven channel widths that typifies pool-riffle channels (Leopold et al., 1964; Keller and Melhorn, 1978). Steps provide much of the elevation drop and roughness in step-pool channels (Ashida et al., 1976; Whittaker and Jaeggi, 1982; Whittaker, 1987a;b). Step spacing decreases with increasing channel slope (Heede, 1972; 1981; Grant et al., 1990), presumably reflecting the greater rate of potential energy dissipation imposed by steeper channel slopes.

Step-pool morphology is associated with steep gradients, coarse bed materials, and small width to depth ratios. Step-forming local accumulations of coarse materials may be viewed as a kinematic wave (Langbein and Leopold, 1968) or as a result of local congestion of large grains that causes increased local flow resistance and further accumulation of large particles (Nowell and Church, 1979; Church and Jones, 1982). Several flume studies have investigated step-pool morphogenesis as well. Whittaker and Jaeggi (1982) demonstrated that for high discharges and low sediment supply, step-pool sequences form on steep slopes (>0.075) through armoring processes. They (Whittaker and Jaeggi, 1982) further showed that step spacing corresponds to maximum flow resistance, providing stability for a bed that would have otherwise been mobile. Based on another flume study, Grant and Mizuyama (1991) suggested that step-pool formation requires a heterogeneous bed mixture and supercritical flow. From field investigations Grant and others (1990) suggested that low sediment supply and infrequent discharges capable of moving the coarse sediment are required for development of stepped-bed morphology. Ashida and others (1981) also observe that step-pool morphologies are most strongly developed in regions characterized by high discharge and low relative sediment supplies.

The style of sediment transport in step-pool channels differs from transport in other channel types. The stepped morphology of the bed results in alternating critical to supercritical flow over steps and subcritical flow in pools (c.f., Bowman, 1977; Chin, 1989). Bedload studies of steep-gradient streams (predominantly step-pool morphologies) indicate that sediment transport rates depend on both seasonal and stochastic input of material from geomorphic processes (Nanson, 1974; Griffiths, 1980, Ashida et al., 1981; Sawada et al., 1983; Whittaker, 1987a;b; Warburton, 1992). Consequently, complex relations between hydraulic discharge and bedload transport result; sediment transport reflects sediment input, flow magnitude and duration of both previous and current events (Griffiths, 1980; Ashida et al., 1981; Whittaker, 1987a; Warburton, 1992). For example, Ashida and others (1981) observed a ten hour lag between hydrograph peak and onset of bedload transport for step-pool channels that were scoured of all pool-filling sediment during previous storms. This time lag presumably represents the time required for sediment introduced upstream to reach the sampling site. Hydrograph and bedload transport were directly correlated during a subsequent storm, due to the availability of sediment deposited in storage sites during the decline of the previous storm's discharge. Temporal and pulsed variations in bedload transport can occur (perhaps due to the release of trapped finer grains by sporadic movement of larger grains) during uniform (Ergenzinger, 1988; Warburton, 1992) or decreasing (Beschta, 1981) flows.

Several thresholds for sediment transport have been documented for step-pool morphologies. Large bed-forming material is generally stable and is mobilized only during infrequent hydrologic events (Whittaker, 1987a;b; Grant et al., 1990; Schmidt and Ergenzinger, 1992). However, Warburton (1992) indicates that step-forming clasts in proglacial channels may be mobile annually due to seasonal glacial ablation. Significant movement of all grain sizes occurs during extreme floods, but step-pool morphology is re-established during the falling limb of the hydrograph (Sawada et al., 1983; Whittaker, 1987a; Warbuton, 1992).

During more typical floods of shorter recurrence interval, finer material transiently stored in pools is mobilized and travels as bedload over the large, stable, bed-forming clasts (Ashida et al., 1981; Whittaker, 1987a;b; Schmidt and Ergenzinger, 1992). In a series of tracer tests in a step-pool channel, Schmidt and Ergenzinger (1992) found that all of the tagged particles placed in pools mobilized during a series of frequent, moderate flood flows. During these events, material in transport was preferentially eroded from, and redeposited into pools. The rapid transport of all of the pool-filling material indicates that sediment transport through step-pool channels is controlled by the availability of sediment susceptible to transport. Based on a sediment transport model presented by Jackson and Beschta (1982), Warburton (1992) recently suggested three phases (thresholds) of sediment transport in step-pool channels characterized by a low-flow flushing of tines, a frequently recurring high-flow mobilization of gravel pavement and underlying fmes [also noted by Sawada et al. (1983)], and a less frequent higher-discharge mobilization of step-forming grains. However, there is some disagreement within the literature regarding models describing the different thresholds of sediment transport in step-pool channels [compare Whittaker (1987a), Ashida et al., (1981), and Warburton (1992)].

Finer material is transiently stored in pools between bedload mobilizing events (Whittaker, 1987a; Ergenzinger and Schmidt, 1990, Grant et al., 1990, Schmidt and Ergenzinger, 1992). In a flume study of step-pool dynamics Whittaker and Davies (1982) showed that flow velocity increases with pool tilling, with maximum velocity achieved when pools are completely filled. From these findings Whittaker and Davies (1982) suggest that the transport capacity of the channel increases with pool tilling. This is consistent with the hypothesis that the stepped morphology of the bed provides the primary bed roughness. Furthermore, these observations suggest that there is a negative feedback discouraging significant sediment storage in step-pool channels, as increased pool tilling enhances boundary shear velocities. Nonetheless, pool filling may be an important transient response mechanism, allowing increased sediment loads to be rapidly transported downstream.

Step-pool and cascade channel morphologies am distinguished by differences in the spatial density and organization of large clasts (Figure 8). Although these morphologies constitute an overlapping continuum in the field, we suggest distinctions based on the composition and frequency, or spacing, of zones of supercritical flow. Step-pool channels are defined by a discrete rib of channel-spanning clasts between pools that occur at a spacing of 1 to 4 channel widths. Channels with essentially continuous supercritical flow or a pool spacing of less than the channel width am cascade channels. Boundaries between these channel types are indistinct, but in our experience it is not difficult to replicate classifications using these guidelines.

Plane-bed Channels

Plane-bed channels lack well-defined bedforms and are characterized by long stretches of relatively planar channel bed that may be punctuated by occasional channel-spanning rapids (Figure 9). They are morphologically distinct from both step-pool and pool-riffle channels in that they lack rhythmic bedforms. Smaller relative roughnesses (ratio of the 90th percentile particle size to the bankfull flow depth) and the absence of both strongly three dimensional flow and significant, grain-induced, hydraulic jumps distinguish plane-bed from cascade channels. Observationally, plane-bed channels occur at gradients and relative roughnesses intermediate between pool-riffle and step-pool channels.

Plane-bed morphology encompasses channel units that have previously been termed glides, riffles, and rapids (Bisson et al., 1982; Sullivan, 1986; Grant et al., 1990), spanning a range of slopes (typically 0.01-0.03) and relative roughnesses. The flow field around particles that are large relative to the flow depth may disrupt development of channel-spanning circulation and decompose the lateral flow component into a series of smaller circulation cells. Thus, we hypothesize that plane-bed channels do not possess sufficient lateral flow convergence to cause pool development. Introduction of flow obstructions may force local pool and bar formation, but as a rule such features am not typical of plane-bed channels.

Plane-bed channels are usually armored by a bed surface layer that is coarser than the subsurface, and are threshold channels (Lane, 1953; Henderson, 1963; Li et al., 1976).

Due to their range of characteristic slopes and roughnesses they may have a range of dominant discharges. Lower-gradient and smaller-grained plane-bed channels (glides and riffles) are bankfull threshold channels, while some steep bouldery plane-bed channels (rapids) may exhibit dual-threshold bed mobility similar to step-pool and cascade channels. The lack of depositional features, such as barforms, and the presence of surface armoring, indicative of low sediment loads (Dietrich et al., 1989), demonstrate some supply-limited characteristics of plane-bed channels. However, studies of armored gravel-bedded channels (Milhous, 1973; Jackson and Beschta, 1982, Sidle, 1988) demonstrate a general correlation of bedload transport rate and discharge during bed-mobilizing flows, indicating a transport-limited nature of mobility during significant flow events. Although sediment transport in plane-bed channels is typically dependent on a dominant discharge threshold, a high sediment supply may cause surface fining, lowering the threshold for sediment transport.

Pool-riffle Channels

Pool-riffle channels have an undulating bed that defines a sequence of bars, pools, and riffles (Leopold et al., 1964) (Figure 10). This lateral bedform oscillation distinguishes pool-riffle channels from the other channel types discussed above. Pools are topographic low points within the channel and bars are the corresponding high points; these bedforms are thus defined relative to each other [see O'Neill and Abrahams (1984) for further discussion]. Although riffles am the topographic cross-over from a pool to a bar, the term riffle is also loosely applied to the entire shallow channel area (including bars) that is distinct from the pools. Pool-riffle channels are the best studied channel type and are often considered as representative of channels in general, at the expense of the other channel types outlined herein.

Bar and pool topography (Figure 8) is generated by local flow convergence and divergence that may be either freely formed by cross-stream flow and sediment transport, or forced by flow through channel bends and around in-channel obstructions that control locations of flow convergence and divergence. Lisle (1986) showed that LWD may anchor and stabilize pool and bar forms. Channels with high LWD loading exhibit particularly complex arrangements of bars, pools, and riffles.

Free-formed pool-riffle sequences result from cross-channel oscillating flow that causes flow convergence and scour on alternating banks of the channel. Downstream flow divergence results in local sediment accumulation in discrete bars. The mechanics of pool-riffle sequences are essentially the same in straight and meandering channels, with centrifugal forces becoming increasingly important with greater channel sinuosity. Flow through channel meanders causes superelevation of the water surface toward the outside of the bend, resulting in a secondary cross-channel circulation of outward surface flow and inward bottom flow (Leopold and Wolman, 1960). Although pool scour and bar maintenance have been attributed to sediment transport by secondary circulation cells resulting from channel bends (Leliavsky, 1955; Leopold and Wolman, 1960), secondary circulation may form bars and pools in straight channels (Leopold, 1982). While cross-channel circulation cells are important for bar maintenance and sediment transport (Dietrich et al., 1979), the effect of topographically-driven convective accelerations is Perhaps of mom significance in the development of convergent and divergent flow patterns and thus pool-riffle morphogenesis (Dietrich and Smith, 1983; Dietrich and Whiting, 1989; Nelson and Smith, 1989). Cross-channel and downstream convective accelerations effectively shift the high-velocity core across the riffle toward the pool and outside bank (Dietrich and Smith, 1983; Dietrich and Whiting, 1989). The resulting cross-channel shear stress may be as significant as downstream boundary shear stress (Dietrich and Smith, 1983; Dietrich and Whiting, 1989). Continued bar development reinforces the topographically-driven convective acceleration, providing a feedback mechanism through which bar development and flow oscillation result from an initial flow perturbation or deflection. Free-formed alternate bar development requires a sufficiently large width to depth ratio (W/D) and small gram sizes easily scoured by the cross-channel flow. However, streams with very large W/D may form braided, rather than alternate bars.

Pool-riffle bedforms am relatively stable morphologic features, even though the material forming the bed is transported annually. Pools are rhythmically spaced about every 5-7 channel widths in self-formed channels without significant LWD loading (Leopold et al., 1964; Keller and Mellhom, 1978). Free alternate bar formation in natural channels is limited to gradients ≤ 0.02 (Florsheim, 1985), although flume studies indicate that alternate bars may occur at steeper gradients (Bathurst et al., 1983; Lisle et al., 1991).

Significant form roughness is attributable to bedform resistance from bars in pool-riffle channels (e.g., Parker and Peterson, 1980). At low-flow conditions, pools appear as flat reaches of relatively smooth flow and riffles appear as steeper reaches of higher velocity flow. Keller (1971) proposed that as discharge increases the velocity in pools increases faster than in riffles and that at bankfull discharge the flow velocity in pools exceeds that in riffles. This velocity reversal may maintain pool-riffle sequences (Keller, 197 1) and cause deposition of finer sediments in pools on the receding limb of the hydrograph. Lisle (1979) documented such a reversal in the average bed shear stress with increasing stage for a pool-riffle type channel.

While bedforms are relatively stable features in a pool-riffle channel, bankfull events tend to cause significant local scour and fiil within a given bedform (e.g., Leopold and Maddock, 1953; Emmett and Leopold, 1963; Andrews, 1979). Campbell and Sidle (1985) reported net accumulation of coarse sediment in pools at discharges less than bankfull and excavation of pools during discharges greater than bankfull, an observation consistent with the velocity reversal hypothesis of pool-riffle sequence maintenance. Although the degree of scour and fill revealed by repeated cross-sectional profiles may be used as an indicator of changing discharge or sediment supply (Dunne and Leopold, 1978), the spatial pattern of scour and fill can be quite complex (Hassan, 1990).

Pool-riffle channels typically have heterogeneous beds that exhibit a variety of sorting and packing, commonly with a coarse surface layer and a finer subsurface (Leopold et al., 1964, Milhous, 1973). The size of the largest bedload material in pool-riffle channels is a fraction of the bankfull flow depth. Bedload transport increases non-linearly with stage (Milhous, 1973; and many others), but for armored channels the threshold for general mobility of the surface layer is associated with an approximately bankfull stage (Jackson and Beschta, 1982; Andrews, 1984). Movement of surface grains releases fine sediment trapped by larger grains and exposes the finer subsurface sediment to the flow, contributing to a steep rise in bedload transport with increasing shear stress (Milhous, 1973; Jackson and Beschta, 1982; Emmett, 1984). Observationally, the grain size composition of the subsurface sediment approximates the composition of the bedload in transport by the channel (Milhous, 1973). Bed movement at bankfull flow is sporadic and discontinuous, depending on gram protrusion (Fenton and Abbott, 1977; Kirchner et al., 1990), friction angle (e.g., Buffington et al., 1992), imbrication of grains (Komar and Li, 1986), degree of burial (Hammond et al., 1984; Buffington et al., 1992), and the occurrence of turbulence-induced high-velocity sweeps of the channel bed Very rarely is the whole bed in motion and material eroded from one riffle commonly is deposited on a proximal downstream riffle. Although one would intuitively expect small grains to travel farther than large grains, the literature presents conflicting results with respect to grain travel distances [compare Butler (1977), Laronne and Carson (1976), Sobocinski et al. (1990), and Ashworth and Ferguson (1989) with Leopold et al. (1966). Brayshaw et al. (1983). and Carling (1983; 1987)].

Bedload transport rates are generally proportional to discharge during flows that transport sediment (e.g., Milhous, 1973; Jackson and Beschta, 1982; Sidle, 1988). However, considerable fluctuations in observed transport rates reflect the stochastic component of grain mobility caused by the previously mentioned grain interactions and turbulent sweeps, transient grain entrapment by bedforms (Jackson and Beschta, 1982;

Sidle, 1988), and perhaps error introduced by the method of bedload sampling (Wilcock, 1992). Magnitudes of bedload transport may also be variable for similar discharge events, depending on season and occurrence of previous transporting events (Milhous, 1973; Reid et al., 1985; Sidle, 1988). These observations and the fact that many pool-riffle channels exhibit approximately bankfull threshold conditions for significant sediment transport (Jackson and Beschta, 1982; Andrews, 1984) suggest that pool-riffle channels are supply-limited in many regards. Nevertheless, during armor breaching events transport rates am generally correlated with discharge, indicating that during dominant discharges sediment transport is limited by transport capacity, rather than the availability of potentially mobile sediment. This is in marked contrast to sediment transport in steep channel types.

Regime Channels

Regime channels (Lindley, 1919; Lacey, 1930; Howard, 1980; Richards, 1982) exhibit a variety of mobile bedforms that provide the primary flow resistance (e.g., Kennedy, 1975) and am observationally dependent on flow depth, velocity, and bed grain sire. Regime channels typically are low-gradient, sand-bedded channels, but even gravel and boulder-bed channels may exhibit regime characteristics (e.g., Dinehart, 1992; Pitlick, 1992). In general, these channels exhibit a succession of bedforms with increasing flow velocity. In sand-bed channels, this follows the well-known sequence of planar bed, ripples, sand waves, dunes, high-energy planar bed, and finally antidunes. In channels transporting moderately- to poorly-sorted sediment, migrating bedload sheets composed of thin accumulations of sediment also may develop (Whiting et al., 1988). Several scales of bedforms may coexist; ripples, bedload sheets, and small dunes may climb over larger dunes as they all move down the channel. A complete theoretical explanation for the development of multiple-scale bedforms does not yet exist, but regime characteristics are associated with low relative roughness and low slope. In addition, regime channels may support point bars or other bedforms forced by channel geometry. Sediment transport in regime channels occurs at all stages, and is strongly discharge dependent; as such, these channels are transport-limited. Slope, frequency of bed mobility, and presence of ripples, or dunes throughout the channel bed distinguish regime channels from pool-riffle channels.

Braided Channels

A braided pattern of longitudinal and medial bars may form in both regime and threshold channels (Figure 11). Braided channels are usually wider and shallower than adjacent unbraided channels. High sediment supply and easily erodible banks favor formation of a braided channel, as frequent deposition of bars causes lateral channel shifting across the channel bed (Leopold and Wolman, 1957; Leopold et al., 1964). Bedforms am mobile and the location of the active channel in braided reaches may change rapidly. The large channel width associated with a braided channel pattern implies a shallower flow depth than for an analogous confined channel. Several workers developed empirical thresholds that define braided and meandering channel patterns on the basis of channel slope and discharge, flow characteristics, and channel geometry (e.g., Leopold and Wolman, 1957; Parker, 1976; Ikeda, 1977), but erodible banks and lack of valley confinement also are necessary for lateral channel migration to develop a braided morphology. The abundant supply of sediment and rapid response to discharge variations suggest that sediment movement in braided channels is effectively transport limited.

Channel Unit Level

Fluvial channels display a variety of morphologies within channel reaches. Channel units are morphologically distinct areas within a channel reach that are on the order of one to many channel widths in length. Common channel unit names are pools (of which as many as six types have been defined), riffles, cascades, step-pool cascades, slip-face cascades, glides, runs, and rapids (e.g., Leopold et al., 1964; Bisson et al., 1982; Sullivan, 1986; Grant et al., 1990). Distinction between these units is essentially based on organization and areal density of clasts, local slope, flow depth, flow velocity and to some extent gram size. These channel units are associated with specific habitat characteristics and thus different fish-utilization patterns (Bisson et al., 1982; Hankin and Reeves, 1988). Sullivan (1986) reported that the channel units so defined have characteristic velocities and depths, but descriptive channel unit classification provides minimal process insight into channel condition and response potential. In addition, definitions of these channel unit morphologies tend to overlap and are somewhat stage dependent; channel unit classification by different observers yield inconsistent classifications (Ralph et al., 1991).

Although striking differences in numbers and sizes of channel units have been correlated with the degree of LWD loading within a channel (Smith and Buffington, 1991), the actual distribution of specific unit types appears to be stochastic. For example, previous workers differentiated six pool types on the basis of either their mode of formation (Bisson et al., 1982) or flow characteristics (Sullivan, 1986). These specific pool types, however, are controlled by the size, location, and orientation of individual flow obstructions. No theory is available to predict these properties and thus specific pool types, based on channel type. While we recognize the biological importance of channel units, we contend that more reliable channel response predictions can be made at the reach level, and subsequently applied to specific teach inventories of channel units.

Confinement

Channel morphology and response are influenced by both the material composing channel banks and the degree of confinement by valley walls. Channel confinement may be expressed as the ratio of the width of the valley floor to the bankfull channel width. Unconfined channels may reflect tectonic boundary conditions (as in the case of alluvial fans at the base of block-faulted mountains), an inherited morphology (as in the case of underfit channels or u-shaped glacial valleys), or long-term alluvial aggradation and floodplain development where sediment supply exceeds transport capacity (Richards, 1982). Systematic downslope changes in channel confinement along a channel network generally reflect the latter case.

Steep channels in mountain drainage basins typically are confined by valley walls and shallow bedrock Insignificant sediment storage in these valley segments indicates that virtually all of the material delivered to the channel is transported downstream; the channel has a relatively high transport capacity. In contrast, thick alluvial valley fill deposits in unconfined lower-gradient channels (Figure 3) imply a long-term excess of sediment supply, reflecting the greater sediment supply (due to greater drainage area) and the lower transport capacity of gentle channel gradients. In mountainous terrain, unconfined channels should occur in reaches of the channel network where there is a local downslope decrease in sediment transport capacity, which forces local deposition of some portion of the channel load. This would occur most readily where them is an abrupt decrease in channel gradient with little increase in flow depth. Low-gradient reaches may be lithologically controlled or may reflect conditions imposed on the channel by its tectonic, geomorphic, or climatic history (Richards, 1982). In an area in which valley morphology reflects current channel morphology, channel confinement is implied from channel bed morphology; regime, pool-riffle, braided, and plane-bed channels are likely to be unconfined and step-pool and cascade channels are likely to be confined.

Vegetation

Vegetation growing within, along, and near channels influences channel morphology and processes. In particular, the root strength of riparian vegetation growing along channel banks contributes to bank stability (Gilbert, 1914), especially in relatively uncohesive alluvial deposits. The influence of riparian vegetation on channel bank stability is greatest in low-gradient, unconfined reaches where loss of bank reinforcement may result in dramatic channel widening; this is not as significant a factor in steep, confined channels. In addition to enhancing bank stability, riparian vegetation provides a source of large woody debris recruitment to channels. The composition and stand structure of riparian vegetation also reflects channel processes. Narrow bands of riparian vegetation along steep channels reflect the influence of disturbance by avalanches and debris flows. Broader riparian associations on lower-gradient floodplains reflect abandoned channels, side channels, floodplain disturbance, and channel bar recolonization. Thus, riparian vegetation both influences, and is influenced by, channel processes.

Large woody debris, on the other hand, is an external constraint to which a channel must respond. LWD forms structural elements of a channel in three ways: 1) by deflecting flow and causing local scour of pools where flow converges; 2) by forcing deposition where flow diverges; and 3) by impounding sediment. LWD can force pool and bar formation in any channel type, but the amount, size, orientation, and position of LWD determine the morphologic impact, Thus, the influence of LWD in a channel reflects rates of debris recruitment, transport, and decay (Bryant, 1980; Murphy and Koski, 1989). Additionally, the relative importance of LWD in controlling channel morphology and providing local sediment storage elements varies through a channel network.

Murphy and Koski (1989) observed that for southeast Alaskan old-growth forests, LWD loading in alluvial channels was greatest in lower-gradient (< 0.01) channels, with relatively steeper (0.01-0.03) channels presumably having higher percentages of LWD suspended above narrower bankfull widths (see also Nakamura and Swanson, 1993). They further observed that there is some tendency for depletion (transport, abrasion, and &cay) of smaller LWD to be directly correlated with channel size, while depletion rates of the largest LWD is similar at all sites (Murphy and Koski, 1989). This indicates that LWD that is introduced into smaller channels is comparatively mote stable, and thus may have a significant long-term effect on channel morphology. Furthermore, in small streams, LWD may provide the dominant control on sediment storage and bar formation, whereas larger channels exhibit a greater proportion of free bars. Some evidence suggests that woody debris smaller than about half the channel width is unstable and thus provides only very transient sediment storage sites (Bilby and Ward, 1989).

Because of the association with channel size, LWD plays an important role in the architecture of smaller channels. Transversely-oriented logs may form steps that create local hydraulic jumps, form plunge pools, and buttress significant amounts of sediment (Figure 12). Obliquely-oriented debris can result in scour pools and proximal sediment storage by both upstream buttressing and downstream deposition in low-energy zones. Flume studies indicate that LWD placed close to the bed and oriented perpendicular to the flow results in maximum scour (Cherry and Beschta, 1989); however, scour area and depth are not significantly different from obliquely-oriented LWD at similar positions within the flow. The observation that a single log can influence the formation of up to five different pools (Smith and Buffington, 1991) demonstrates the potential significance of LWD on channel morphology.

In many channel types LWD can force specific morphologies on the channel. In forested basins, for example, LWD may create steps in steep channels and dissipate energy that otherwise would be available for sediment transport (Keller and Swanson, 1979; Heede, 1981; Marsten, 1982). These organic steps may dominate the channel roughness and provide a significant proportion of sediment storage in step-pool channels. Moreover, organic steps may provide sufficient roughness to stabilize an alluvial bed in channels with a small sediment size (due, for example, to either rapid downstream fining or lack of coarse sediment). Depending on slope, discharge, and sediment load, removal of organic steps may transform a forced step-pool channel into a step-pool (Heede, 1985), cascade, or bedrock channel (Figure 13). Forced step-pool morphologies may be found on steeper slopes in channel systems where LWD is an important influence on channel morphology. Plane-bed and pool-riffle channels also respond to woody debris inputs. In particular, LWD may force pool and bar formation in channels that otherwise would be plane-bed Consequently, plane-bed channels may be rare in undisturbed forested environments where the majority of pools and bars are LWD dominated. Removal of LWD from forced pool-riffle channels may result in either pool-riffle morphology (Smith et al., in press) or metamorphosis to a plane-bed morphology (Figure 13). Pool-riffle morphology also may be forced due to imposed sinuosity or the introduction of other obstructions. Channels in which woody debris 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 (Bryant, 1980).

The influence of large woody debris on sediment storage and channel form decreases with increasing channel size. In larger rivers, sizable old growth logs are deposited on bar tops during the falling stage of flood flows (Figure 14). However, LWD may still form local scour pools in large rivers where it becomes trapped against channel banks, but such effects are transitory and LWD primarily acts as sediment. Woody debris also influences side channel development (Bryant, 1980) and bank cutting (Nakamura and Swanson, 1993) in large alluvial rivers. In many larger channels, debris jams may control channel avulsion and thus influence both channel pattern and floodplain processes.

Debris Flows

Periodic debris flow scour may dominate the morphology and disturbance frequency of steep mountain channels. Both the time since the latest debris flow and the rate of channel recovery control the morphology of channels subject to debris flow processes. Debris flows generally originate along low-order channels or in bollows steeper than 26" (50%) (e.g., Campbell, 1975). In addition to this temporal variability, the style of debris flow impacts changes downstream. Debris flows typically scour high-gradient channels and aggrade the first downstream reach with a gradient low enough to cause deposition of the entrained material. Consequently, the effects of debris flow processes on channel morphology may be divided into those related to scour and those related to deposition.

In mountainous terrain, debris flows intermittently traverse low-order channels with gradients greater than about 6° (10%) (e.g. Campbell, 1975; Ikeya, 1981; Takahashi et al., 1981; Benda and Dunne, 1987; Reneau & Dietrich, 1987a; Benda & Cundy, 1990). Channel slope and tributary junction angles are important controls on the travel distance of debris flows in mountain channels. Debris flows originating at the heads of long straight channels tend to be far travelled, scouring long channel segments, and delivering sediment to downslope alluvial channels. Debris flows originating in obliquely-oriented tributaries tend to be deposited at channel confluences and increase sediment loading in downslope channels (e.g., Grant et al., 1984; Benda and Dunne, 1987; 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. 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. The morphology of a steep mountain channel reflects the time since debris flow scour, as well as position within the fluvial system (Figure. 15).

Other effects associated with debris flow processes may influence both channel morphology and water temperature. Canopy response provides an example of how debris flows influence riparian processes, which in turn affect channel morphology. Debris flow scour and deposition typically disturb channel-marginal vegetation and expand the canopy opening over a channel (e.g., Grant et al., 1984; Grant, 1988). Following disturbance, riparian vegetation recolonizes the disturbed zone and closes the canopy opening. Riparian disturbance affects rates of LWD recruitment and may influence channel temperature by reducing channel shading. Sustained riparian disturbance also may influence rates of in-channel LWD decay by altering the type of vegetation type entering channels.

Dam break floods also scour steep alluvial channels when organic debris dams mobilize catastrophically during high discharge events (Johnson, 1991). Failure of these organic debris dams releases impounded water and sediment as a large flood wave that may propagate through downslope channels. Often it is the incorporation of woody debris into debris flow deposits that forms large organic debris jams. Thus an increase in debris flow frequency increases the probability and impact of subsequent dam break floods by increasing both the number of channel-spanning debris accumulations and the volume of impounded sediment and water.

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 angle is typically between 3" and 6" for the range of water contents typical of debris flows (e.g., Ikeya, 1981; Takahashi et al., 1981; Benda and Dunne, 1987; Benda and Cundy, 1990). Incorporation of abundant LWD in the leading edge of a debris flow may result in deposition on even steeper slopes; we have observed debris flow deposits behind LWD jams in the Olympic Peninsula and Oregon Coast Ranges in steep (>6°) channels. These deposits am subject to incorporation into subsequent debris flows. In debris-flow-prone areas, the probability of deposition is highest in the first downslope channel reach with a slope less than 6°. Lower-gradient channels (< 3") may be impacted by debris flows from tributary channels, but impacts are typically restricted to local deposition at tributary confluences where debris flow fans extend into major channels.

ORIGIN OF REACH-LEVEL MORPHOLOGES

We recognize eight distinct reach-level channel morphoiogies controlled by slope, hydraulic discharge, and sediment supply. Together with considerations of confinement, LWD loading, and the potential for debris-flow impacts, these reach-level morphologies provide a process-based classification of natural channels. Observationally, there is a downstream progression from steep headwater channels to lower-gradient channels that proceeds as colluvial, cascade, step-pool, plane-bed, pool-riffle, and regime channels (Figure 16). Bedrock channels may occur in steep reaches anywhere within the network, while braided channels am restricted to lower-gradient, unconfined reaches. Not all of these channel types are present in all watersheds and this pattern may vary downstream, reflecting factors controlling channel slope, discharge, and sediment supply. Below we elaborate our hypotheses for the origin of channel confinement and these reach-level morphologies.

Relations between drainage area, sediment supply, and transport capacity illustrate controls on the pattern of confined headwater channels and unconfined downstream channels with well-defined floodplains. The transport capacity of a channel reach is proportional to the product of channel depth and slope. Many workers (e.g., Leopold et al., 1964) report that flow depth and drainage area are positively correlated and that slope and drainage area are negatively correlated, with specific relationships characterizing river systems in different regions. In mountain drainage basins, however, slope typically decreases faster than flow depth increases. In a basin with uniform sediment production and little storage, the sediment supply of a channel reach increases with drainage area Thus, the transport capacity as defined by the depth-slope product generally decreases downstream, while the sediment supply increases. We suggest that the longitudinal arrangement of channel types and downstream transitions from confined to unconfined channels reflect both this tendency and local controls (Figure 17). This further implies that headwater channels generally are supply limited whereas lower-gradient channels are transport limited.

In free-formed alluvial channels, the total channel roughness decreases downstream and reflects the boundary shear stress, a function of the depth-slope product [see equation (4)]. We suggest that downslope and local changes in bed morphology, and thus channel type, provide this changing roughness. We further suggest that the alluvial 'bed morphology for a teach reflects the minimum roughness necessary to stabilize the channel bed for the channel-forming shear stress and the size and volume of supplied sediment. Sorting and self-organization of the supplied load generate a stable alluvial bed The degree to which the imposed sediment load is sorted and organized to produce a stable alluvial bed controls channel morphology. While them are characteristic patterns of changing morphology and roughness configuration downstream through a network (Figure 16), the roughness configuration for a given depth-slope product is not universal and depends on local sediment supply.

Different mechanisms related to channel morphology provide this roughness in different channel types. In steep channels, shear stress is dissipated dominantly by hydraulic jumps and jet-and-wake turbulence from flow over and around large bed-forming clasts. This style of energy dissipation is spatially continuous in cascade channels and intermittent in step-pool channels. Skin friction and local turbulence associated with moderate particle sizes are sufficient to stabilize the bed for the lower shear stresses characteristic of plane-bed channels. In pool-riffle channels, smaller grain sizes make both cross-channel flow oscillations and scour more effective, causing bar and pool formation, which together with grain and LWD roughness cause local flow separation and turbulence. Particle roughness in regime channels is small due to the low relative roughness, and bed roughness is controlled primarily by regime bedforms, which together with bank resistance cause large-scale eddies and turbulence. The importance of bank roughness varies with channel type, depending on width to depth ratio and vegetative influences, but in steep channels bank resistance is less important compared to energy dissipation caused by tumbling flow. These differences in channel roughness configuration control the reach-level channel morphology.

We further suggest that these distinct roughness configurations reflect the relative transport capacity of the channel, defied as the ratio of channel transport capacity (Qc) to sediment supply (Q_s). Colluvial channels are transport-limited ($Q_c < Q_s$), as indicated by the accumulation of colluvium between non-fluvial scouring events. In contrast, the lack of an alluvial bed indicates that bedrock channels are supply limited ($Q_c > Q_s$). Alluvial channels, on the other hand, maintain an alluvial bed through morphologic response that adjusts transport capacity to the sediment supply ($Q_c \geq Q_s$). In this sense, valley morphology (colluvial, alluvial, bedrock) identifies the long-term balance between sediment supply and transport capacity. Reach-level channel morphology reflects this balance over shorter temporal and spatial scales. Colluvial and bedrock channel morphologies define the end members of a continuum from transport- to supply-limited conditions (Figure 18). Steep alluvial channels (cascade and step-pool) have high relative transport capacities. The relative transport capacity of lower-gradient alluvial channels (regime, pool-riffle, plane-bed) ranges from an approximate balance for an unarmored channel bed to an excess of transport capacity for an armored bed An increase in sediment supply to an armored channel will result in fining of the bed surface until the transport rate matches the sediment supply (Dietrich et al., 1989). The variety of alluvial channel morphologies reflects bed stabilization through sorting and organization of the sediment load and thus provides an indication of relative transport capacity.

We further hypothesize that these differences in relative transport capacity cause significant differences in response potential and rates of recovery from increased sediment loads. Cascade and step-pool channels have a high relative transport capacity, and are capable of rapidly transmitting sediment smaller than the large clasts forming the stable channel bed. Channels in which bedform and particle roughness dominate energy dissipation (regime, pool-riffle, plane-bed) have a low relative transport capacity and exhibit more persistent morphologic change in response to altered sediment supply.

RESPONSE POTENTIAL

Areas of a landscape in which different processes dominate the generation, transport, and storage of sediment respond differently to changes in sediment supply or discharge. Potential impacts differ for hillslopes, hollows, and each of the channel types discussed above. Alluvial channels, in particular, exhibit a variety of potential channel responses that vary with channel processes and morphologies. Consideration of processes acting in different channel types allows assessment of the potential for specific channel responses. Differences in confinement, bed morphology, and relative transport capacity (Figure 19) suggest that different types of channels are mom or less sensitive to adjustment of channel width, depth, grain size of the bed surface, bed roughness, and sediment storage. A first step in the assessment of potential management impacts in a watershed should be to identify different channel teaches in the watershed. Potential reach-level morphological responses to persistent, moderate, perturbations are described below and summarized in Figure 20. Extreme changes in discharge and sediment loading can alter channel type. While Figure 20 lists general differences in response potential for the various channel types, exact responses depend on specific local conditions, and the effects of LWD and debris flows (see also Figure 16).

Hillslopes

Soil mantled (transport-limited) and bedrock (production-limited) hillslopes have different response potentials. Sediment flux on and from soil-mantled hillslopes is sensitive to changes in processes controlling both sediment production and transport. In general, surface erosion on undisturbed soil-mantled slopes is insensitive to hydrologic change, as runoff occurs primarily by subsurface flow. As such, sediment transport is dominated by slope-dependent processes (such as soil creep) and thus is relatively constant. The long-term supply of material available for transport, however, depends on the rate at which bedrock is converted to soil, a process that may strongly depend on biologic activity. Changes in biologic activity can also effect sediment transport and have long-term implications for soil profile development. In some areas, for example, much of the production and transport of colluvial soils on hillslopes is due to soil/bedrock mixing and net downslope soil movement from tree throw. Repeated clear-cutting effectively terminates this process, and should decrease the long-term production and delivery of sediment from hillslopes to downslope hollows and channels. Burrowing activity is another important agent of soil transport that could increase in response to climate change or land management. Ground surface disturbance also accelerates downslope movement of colluvium toward hollow axes and across channel banks. Compaction of the ground surface reduces infiltration rates and may dramatically accelerate hillslope erosion. On bedrock hillslopes with thin soils, sediment flux responds primarily to changes in processes infhtencing soil production rates. In general, however, changes in hillslope processes impact downslope morphologies and processes through changes in the frequency and magnitude of sediment input.

Hollows

Changes in land use and climate can affect both the amount of colluvium stored in hollows and their stability, thus changing the rate of sediment delivery to downslope channels. Hollows essentially have two ways to respond to altered sediment inputs or hydrologic conditions: either the rate of colluvium accumulation or the frequency of excavation may change.

The rate of soil production and hillslope sediment transport control the rate of colluvium accumulation in hollows. Direct disturbance on the surrounding hillslopes or changes in biological sediment transport processes can alter sediment delivery to hollows. Over human time scales, however, the rate of sediment influx into hollows is likely to be stable; much greater changes in sediment storage and its delivery to downslope channels results from short-term impacts on hollow stability.

Alteration of surface or subsurface hydrologic processes may influence sediment storage in hollows. Compaction of the soil surface or a reduction in the erosional resistance of the ground surface, such as often accompanies heavy grazing activity, can initiate rilling or gullying in hollows where these processes previously did not occur. These processes generally dominate the potential response of low-gradient hollows. Hollows on steep slopes am more sensitive to changes in subsurface hydrology and root strength. Changes in either the strength properties of the colluvial soil (root decay) or in the hydrologic controls on pore-pressure development within colluvial soils (canopy clearance or drainage alteration) may profoundly impact the stability of colluvium-filled hollows. Slope stability models indicate that a decrease in the root strength of the vegetation growing withii hollows reduces the piezomettic level necessary to initiate slope instability (e.g., Burroughs, 1984; Burroughs et al., 1985; Reneau and Dietrich, 1987b; Montgomery, 1991; Sidle, 1992). Consequently, changes in root strength resulting from climatically-induced vegetation change, forest clearance, intense tires, or herbicide application would tend to accelerate landsliding from hollows. Decreased evapotranspiration from forest canopy clearance and concentration of road drainage into hollows also may accelerate landsliding, due to an increase in pore-pressure response for a given rainfall. Dramatic examples of the latter occur in steep watersheds in both the Olympic Peninsula and the Oregon Coast Range, where concentration of ridgetop road drainage has destabilized many hollows (Montgomery, submitted). Management-related destabilization of hollows may dramatically accelerate sediment delivery to channels [see Montgomery (1991) and references therein].

Colluvial Channels

Colluvial channels respond to altered discharge or sediment supply primarily through changes in sediment storage. In steep landscapes, colluvial channels have only a thin alluvial mantle, typically occur in narrow, bedrock-walled valleys, and are unlikely to significantly change channel width or depth. In contrast, colluvial channels in low-gradient landscapes may occupy wide valleys with thick valley fills. Increased discharge may trigger gullying, tilling, and channel head advance that may destabilize valley fills and dramatically increase sediment delivery to downslope channels. A rise in discharge may also tend to cause both greater bed mobility and bed coarsening, resulting in an increasingly alluvial character to the channel. In steep landscapes, debris-flow scour transiently depletes valley sediment storage and may convert colluvial channels into bedrock channels until colluvium reaccumulates in the valley bottom. Conversely, sediment storage in a colluvial channel could increase from either gullying or landsliding in upslope hollows, or acceleration of colluvial transport across channel banks. Increased sediment storage may result in larger downslope deliveries of sediment during the next major channel-scouring event. The integrated nature of the channel network requires that short or long-term changes in sediment delivery to, storage in, and transport from colluvial channels affect the sediment supply of downslope channels.

Bedrock Channels

Bedrock channels are generally insensitive to short-term changes in sediment supply or discharge. Change in bedrock channel geometry generally does not occur over short time scales because bedrock channels are confined; channel width and depth will increase in response to greater discharge, not by incision, but by simple expansion of flow area. The paucity of alluvial cover in bedrock channels generally precludes significant adjustments in the bed material size, bedform roughness, or sediment storage. Only a persistent decrease in discharge and/or an increase in sediment supply sufficient to convert the channel to an alluvial morphology would significantly alter fluvial bedrock channels. Channels scoured to bedrock ‘by debris flows, on the other hand, recover their pre-scouring morphology at a rate determined by the sediment supply and fluvial transport capacity. In general, moderate changes in discharge or sediment load are transmitted downslope without significant morphological change within a bedrock reach.

Cascade Channels

The hypothesized undercapacity bedload transport rates, as well as stable bed-forming grains that are adjusted to long recurrance interval flows make cascade channels resilient to altered sediment supplies or dicharges. Cascade channels generally are floored by relatively coarse, immobile alluvium and are laterally confined by valley walls. Consequently, channel bank cutting or bed incision are unlikely responses to changes in sediment supply or discharge. However, width and depth in these confined channels can respond to increased discharges by simple flow expansion. Significant bed roughness changes are unlikely, due to the lack of distinct bedforms. With decreased discharge or large increases in sediment supply, some decrease in d_{50} may occur with deposition and/or entrapment of finer bed material behind large grains and in their wakes. However, high relative transport capacities generally imply the potential for rapid downstream transport of increased sediment loads. Because of their position within the network, cascade channels typically are subject to debris flow impacts. We hypothesize that burial of cascades by debris flow deposition is generally short-lived, as these high energy channels excavate themselves rapidly. Recovery from debris-flow scour, however, depends on the delivery of large bed-forming clasts to the channel reach.

Step-pool Channels

Adjustments to changes in flow or sediment supply in step-pool streams can be quite complex, in which a variety of possible response scenarios may be envisioned, dependent on both current and previous watershed events. Nevertheless, high relative transport capacities enable increased discharges or sediments loads to continue to be rapidly passed downstream even though morphologic response may have occurred within a step-pool reach. As with other confined channels, increased discharge may result in flow width and height expansion without bank cutting or channel incision, as bed-forming step-pool grains am essentially static. However, pool depth, and thus channel storage and depth of scour, are sensitive to changes in both sediment input and flow characteristics. For example, a decline in the frequency and duration of pool-scouring flows combined with an increase in sediment input would likely cause a larger volume of bedload material to be transiently stored in pools between mobilizing events. Pool-filling material of this sort is efficiently scoured during frequent high-flow events, but bedload travel distance depends on flood duration.

Severe increases in sediment loads will result in significant pool filing, reduction of channel roughness, and the potential for bank incision (Whittaker, 1987a). However, Whittaker and Davies (1982) suggest that pool filling is discouraged by the simultaneous reduction in step roughness, indicating the tendency for re-establishment of a stepped morphology if high sediment loads am not sustained Similarly it has been observed that channel morphology is rapidly m-established after extreme flood events (Sawada et al., 1983; Warburton, 1992). Large sustained changes in discharge may be compensated for by alteration of channel roughness by changes in step/pool spacing (see Whittaker and Jaeggi, 1982). Textural response in terms of bed armoring or fining is the more likely response to moderate changes in supply or discharge.

Plane Bed Channels

Plane-bed channels have a variety of potential responses to perturbations. They may be either confined or unconfined and may or may not be free to widen or incise with changes in discharge or sediment supply. Increased discharge can coarsen the bed, potentially causing an increase in the relative roughness and thus the turbulent energy dissipation. We have not observed sinuous plane-bed channels and so roughness changes of that sort seem unlikely. Changes in sediment loading and discharge can cause significant bed aggradation or degradation, thus changing the amount of bed storage. Increased sediment supply is expected to result in either fining of the bed surface or channel aggradation. Although plane-bed channels are characterized by a lack of bedforms, it is possible that with rapid aggradation an unconfined plane-bed channel may metamorphose into a braided morphology. Anticipated response of a plane-bed channel to altered sediment load or discharge involves changes in bed surface texture, channel geometry, or depth of scour. Additions of significant LWD to a plane-bed channel may provide sufficient flow convergence and divergence to cause pool development and transform the channel to a forced pool-riffle morphology. Conversely, a reduction in the supply of LWD to a steep-gradient (1°-3°), forced pool-riffle channel may result in conversion to a plane-bed morphology (Figure 13). Consequently, identification of channel reaches with the potential for a plane-bed morphology is an important aspect of assessing the impact of LWD loss accompanying intensive timber harvesting. Plane-bed channels are hypothesized to have lower relative transport capacities than step-pool and cascade reaches and therefore are more morphologically responsive to perturbations in discharge or sediment supply.

Pool-riffle Channels

Pool-riffle channels tend to have the widest variety of potential responses. They are generally unconfined, which allows widening in response to either increased discharge or sediment supply. Increased discharge may also cause bank cutting and meander development, potentially decreasing channel slope. Aggradation can occur due to higher sediment loads or decreased discharge. Pool filling in response to increased sediment loading reduces bedform roughness and increases sediment storage. Higher sediment loads encourage fining of the channel bed, while decreased sediment supply enhances the tendency for bed armoring. Increased sediment supply also may result in expansion of the zone of active transport within the channel. An increase in the fine sediment load can result in the development of longitudinal accumulations of fine sediment, sometimes referred to as sand stripes. Higher peak flows or more frequent sediment transporting discharges can potentially increase the depth of scour. Furthermore, higher discharge increases the difference between basal shear stress and the critical shear stress for bed mobilization, which would increase bedload transport rates, decrease sediment storage and potentially coarsen the channel bed. Less frequent channel-forming flows, on the other hand, favor pool filing and increased sediment storage. Increased sediment supply concurrent with channel widening may result in a braided channel.

The nature of the controls on pool and bar formation constrain the potential responses of pool-riffle channels. Forced pool-riffle channels are extremely sensitive to the availability of flow obstructions. Preliminary field observations indicate that forced pool-riffle channels occur in the same gradient range as plane-bed channels (.01 - .03). A decrease in the supply of LWD in a steeper-gradient, forced pool-riffle channel may result in significant morphological changes, including: pool loss, increased effective shear stress, and potential conversion to a plane-bed morphology. Even with LWD removal, lower-gradient (i.e., $S<.01$) forced pool-riffle channels may maintain a pool-riffle morphology, albeit with different pool characteristics (i.e. depth, size, and spacing) (Smith et al., submitted). In forced pool-riffle channels, LWD is the dominant roughness element and response to altered sediment supply may be limited to a narrow range of grain size response before significant bedform aggradation/degradation is initiated (Buffington and Montgomery, 1992). Due to low relative transport capacities, pool-riffle channels are morphologically responsive to changes in sediment supply and discharge. Prediction of specific response of pool-riffle channels is complicated by the large number of possible channel adjustments. Unfortunately, pool-riffle channels are both of primary concern for anadromous fish habitat and are those most likely to experience significant, persistent impacts.

Regime Channels

Regime channels am expected to have the lowest relative transport capacities, and are thus considerably sensitive to perturbations. Regime channels typically occur in low-gradient valleys and are unconfined by valley walls. Bank material usually consists of sediment previously transported by the channel. Consequently, the channel may widen in response to both increased discharge or aggradation resulting from higher sediment loads. Higher sediment loading also may result in greater floodplain storage. Regime channels typically are poorly armored with little potential for changes in bed material size. Bedform roughness is unlikely to change in response to altered sediment supply, as flow stage dominantly controls bed roughness. Changes in the discharge regime, on the other hand, am reflected in the sequence of bedfotms present in the channel. As sediment transport in regime channels does not have a distinct threshold, the rate of sediment transport increases with discharge. Increased discharge also can result in meandering, causing decreased slope and increased channel roughness. Thus the primary response potential for regime channels is changes in channel geometty or transport rates.

Braided Channels

Braided channels reflect a condition of high sediment supply relative to transport capacity. Consequently, a significant increase in discharge or decrease in sediment supply may result in conversion to a single-thread channel. Increased sediment supply may result in further widening of the active channel and/or aggradation of the bed surface. The composition of the channel bed, the number of medial and longitudinal bars, and the amount of sediment storage, also may change in response to altered sediment supply or discharge. Even so, braided channels are relatively insensitive to all but major perturbations.

Confinement

Channel confinement :is important for interpreting potential channel response. The geometry of the channel above the bankfull stage strongly controls the response of the channel bed to high-discharge flow. Unconfined channels may have extensive floodplains across which overbank flows spread At stages above bankfull, a greater increase in flow depth occurs in response to a unit increase in discharge in a confuted channel than in an unconfined channel (Figure 21). Lateral spreading of overbank flow from an unconfined channel across the floodplain effectively limits the depth of flow, and thus the basal shear stress, to about that associated with the bankfull flow depth, mitigating the effect of peak discharges on channel morphology. The geometry of a confined channel, on the other hand, translates discharge greater than bankfull into increased basal shear stress. Sediment transport rates and the depth of bed scour in unconfined channels should reflect the duration that flow is greater than that required for bed mobility. Consequently, the amount of time flow exceeds bankfull stage, rather than the peak flow magnitude, may control the response of unconfined channels to changes in discharge. In contrast, transport rates in confined channels may reflect both the full magnitude of peak discharge and the duration of flow in excess of that required for bed mobilization.

Isolation of unconfined channels from their flocxlplains can entail dramatic consequences, as connections between a channel and its floodplain are an important geomorphic component of many biologic 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 habitat elements. Prevention of overbank 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.

Vegetation

Changes in channel-margin vegetation can have dramatic impacts on channel morphology and processes. Riparian vegetation may contribute substantial cohesion to channel banks. Unconfined alluvial channels with relatively uncohesive bank-forming material are particularly susceptible to dramatic channel widening as a result of riparian vegetation clearance. Alteration of channel margin vegetation also influences the recruitment of LWD to channels, which may transiently change both the age and species of wood entering the fluvial system. This may affect the magnitude and persistence of the morphologic influence of LWD through changes in LWD sire and decay rate.

Channel response to changes in the supply of LWD depends on its role in sediment storage and pool formation. As discussed above, these effects vary systematically through a channel network. A decrease in the supply of LWD accelerates sediment transport and decreases sediment storage in small channels where woody debris provides significant sediment storage. Similarly, a decrease in the supply of LWD reduces in-channel roughness and may eliminate pools in channels where LWD controls in-channel flow convergence and divergence. For example, removal of LWD from channels in which abundant LWD maintains a pool-riffle morphology may result in either a change in the sire and location of pools or conversion to a plane-bed morphology (Figure 13). Significant morphologic change also may accompany a decreased supply of LWD to a channel in which LWD plays a major role in step formation and sediment storage (e.g., Marston, 1982). Where a supply of large, step-forming clasts is not available, a forced step-pool channel may convert to a bedrock channel following LWD removal. Hence major morphologic change may occur in small to moderate sire channels in response to changes in LWD input, transport, or decay. Changes in the amount of LWD supplied to colluvial channels marginally affect sediment storage, but may substantially affect debris flow delivery of LWD to cascade and step-pool channels. Dramatic increases in debris loading may result in more frequent dam-break floods, which may significantly alter the frequency of scour and thus morphology of downslope channels.

Changes in the amount of LWD also may impact larger channels in which LWD primarily acts as sediment. In some channels, debris jams control channel avulsions and side-channel development across floodplains. A decrease in the supply of large, jam-forming debris delivered to the main channel could lead to side channel abandonment. Channel type, sire, and position in the channel network control potential response to changes in LWD recruitement, transport, and decay. In general, field observations are required to assess potential impacts associated with LWD.

Debris Flows

The potential effect of debris flow processes reflects channel slope. and position in the channel network. Debris flow passage may mobilize bed material and scour channels to bedrock Debris flow deposition, on the other hand, may result in aggradation and obliterate the channel as a morphological feature. Recovery from debris flow impacts differs for steep and low-gradient channels. Steep, high-energy channels recover quickly from sediment deposition, recover mom slowly from bedrock scouring events, and may be greatly influenced by the availability of coarse sediment and LWD recruitment. Step-pool and cascade channels impacted by debris flow deposition recover quickly, due to the high relative transport capacities. Lower-gradient channels, on the other hand, may take significant time to recover from debris flow deposition, because of their lower relative transport capacity. Although channel gradient often correlates with the style of potential debris-flow impacts, channel network architecture also influences the routing of debris-flow impacts. Assessment of potential debris-flow impacts involves differentiating areas of potential debris-flow initiation, scour, and deposition. Digital terrain models for debris flow source areas and run out paths provide both this spatial context and assessments of relative debris flow hazard (Montgomery and Dietrich, submitted).

SOURCE, TRANSPORT, AND RESPONSE REACHES

At the most general level, network position and sediment transport characteristics of the reach-level morphologies define source, transport, and response reaches. In steep landscapes, source reaches are transport-limited, sediment storage sites subject to intermittent debris flow scour (colluvial). Transport reaches am morphologically resilient, high-gradient, supply-limited channels (bedrock, cascade, and step-pool) that rapidly convey increased sediment inputs. Response reaches are low-gradient, transport-limited channels (plane-bed, pool-riffle, regime, braided) in which significant morphologic adjustment occurs in response to increased sediment supply.

The spatial distribution of source, transport, and response reaches governs the distribution of potential impacts and recovery times. General responses for source, transport, and response reaches define patterns of sensitivity to altered discharge, sediment supply, or debris flow scour within a watershed. Downstream transitions from transport to response reaches, in particular, define locations in the channel network where impacts from increased sediment supply are both pronounced and persistent. In addition to systematic downstream changes, local impacts may occur where them is a rapid decrease in transport capacity relative to sediment supply, such as would occur at the head of local low-gradient reaches controlled by lithology or geologic structures. Upstream transport reaches rapidly deliver increased sediment loads to the first downstream reach with insufficient transport capacity to accommodate the additional load. Viewed in a landscape context, the majority of sediment delivered to transport teaches is rapidly delivered to the fit downstream response reach where sediment accumulates and is gradually transported downstream. Consequently, locations in the channel network where transport reaches flow into response teaches are particularly susceptible to impacts from accelerated sediment supply. In this regard, the reach-level classification identifies areas most susceptible to local increases in upstream sediment inputs. Thus the "cumulative" effects of upstream increases in sediment supply are magnified in a response teach where longer time and/or significant morphological change is required to transport the additional sediment. Downstream impacts will also occur, of course, but these locations provide opportunities to monitor network response and potentially may serve as a critical components of watershed monitoring studies. In many areas, these locations are those of fundamental concern for aquatic resource management because of the associated habitat values. Most importantly, this relation between channel classification and potential response provides a direct linkage between upstream sediment inputs and downstream response.

The concentration of persistent impacts at identifiable positions in the channel network has important correlations with resource vulnerability, as different species of fish utilize different portions of the channel network. In the Pacific Northwest, for example, resident trout typically occupy step-pool and cascade channels, whereas anadromous salmonids tend to spawn in pool-riffle channels. This suggests that resident trout populations are subject to different natural disturbance and recovery regimes than are anadromous species. The conceptual channel response model presented above indicates that impacts of increased sediment supply from landsliding and road construction are concentrated and persistent primarily in anadromous species habitat. Impacts on the physical habitat in steeper transport reaches are more transient, but may be dramatic nonetheless if the extent and frequency of disturbance increases significantly. This distinction of response and transport reaches provides a context for examining connections between watershed modifications, impacts on channel morphology, and biologic response. Further research on the relation between disturbance regimes and the population dynamics of different fish species should receive high priority.

The response potential for different channel morphologies highlights an important consequence of the nature of sediment transport in channel networks, and steep landscapes in particular. Channel response cannot be predicted through considering only management activities proximal to and directly impacting some portion of the channel network, such as fish-bearing waters, as is commonly done in most regulatory and management arenas. Although management regulations typically do not apply to small, steep channels because they do not harbor fisheries resources, such policies are not based on a realistic consideration of channel network processes, as increased sediment production and delivery to steep low-order channels is rapidly transmitted to downstream fish bearing waters.

APPLICATION TO CHANNEL NETWORK CLASSIFICATION

Identification of potential source, transport, and response reaches provides a first step for assessing potential channel responses. There are two approaches to mapping channel types: field inspection and prediction. Field mapping is the most reliable method for establishing the spatial distribution of channel types in a watershed, but it is both labor and time intensive. Prediction of channel type is less accurate, but easily automated Preliminary correlations of channel types and slope suggest that either topographic maps or digital elevation models may be used at a reconnaissance level to predict potential source, transport, and response reaches.

When field information is available on channel type, a simple abbreviated code may be used to delineate reach-level morphologies and illustrate their spatial linkages (Figure 22). Each channel morphology may be represented by a simple code: B = bedrock; CO = colluvium; BR = braided; CA. = cascade; SP = step-pool; PR = pool-riffle; PB = plane-bed; R = regime. Channels in which LWD provides the dominant control on bed morphology (forced step-pool and forced pool-riffle) may be designated with a subscript "f" (i.e., SP_f, PR_f).

Prelimiiary field observations in mountain drainage basins in the Pacific Northwest (Montgomery, in prep.) indicate that channel slope differentiates alluvial channel types: pool-riffle channels occur at gradients < 0.02, plane bed channels occur at gradients between 0.01 and 0.03; step-pool channels occur at gradients between 0.03 and 0.08; cascade channels occur at gradients between 0.08 and 0.30. In mountainous drainage basins in this region, colluvial channels occur at gradients > 0.20 and bedrock channels have unusually steep slopes for their drainage area (Montgomery, in prep.). These gradient divisions are not absolute, and morphologic transitions between channel types may be gradual. Drainage area and both local and regional conditions may also influence channel type. Our experience to date, however, suggests that these gradient divisions are relatively robust. For reconnaissance-level classification this can be. generalized to predict the distribution of source ($S > 0.30$), transport ($0.03 < S < 0.30$), and response ($S \sim 0.03$) reaches from digital elevation data once the channel network is delineated (see discussion in Montgomery and Foufoula-Georgiou, in press). Comparison of these slopes with general criteria for debris-flow impact discussed earlier suggests that colluvial and cascade channels am subject to scour, step-pool channels to deposition, and pool-riffle and plane-bed channels to local deposition depending upon position within the channel network.

Automated channel classification based on digital elevation data allows prediction of the spatial distribution of channel types that can guide field work. For example, channels so defined from the U.S. Geological Survey Owl Mountain 7.5' quadrangle on the Olympic Peninsula, Washington (Figure 23) indicate that source reaches occupy basin headwaters, transport reaches occupy major tributary valleys, and response reaches occupy the major alluviated valleys along the main rivers. Field mapping of channel types along the South Fork Hoh river confirms this general pattern of channel types. Moreover, in this same area land-management related impacts are concentrated and persistent at the transition from transport to response reaches where channel gradients decline rapidly as steep tributary valleys enter larger valleys. This approach provides a simple method for rapidly generating a site-specific conceptual model of general watershed processes.

ASSESSING CHANNEL CHANGE

Three general approaches axe possible for evaluating past channel changes and assessing the present state of a channel: comparison of current channel conditions with historical records of past conditions; comparison of current conditions with those in a "comparable" channel; and theoretical predictions of channel morphology (cf., Gregory, 1977). The historical approach does not have predictive power, but is the most direct and convincing method for documenting past channel changes. Such an approach is only possible, of course, when the necessary information is available. Typically, only fragmentary information exists, as concern for channel integrity usually arises only once significant change already has occurred. Comparison of channel conditions with those in undisturbed or otherwise desirable channels provides an alternative when appropriate historical information is not available. However, any such substitution of space for time involves significant uncertainties, and deciding what constitutes a desireable channel necessarily involves some judgement. To be meaningful, assessment of channel conditions using these approaches must be based on channel type. One of the most important implications of these methods is that assessment of channel conditions and prediction of channel response require fundamentally different approaches.

Assessment of past impacts requires evaluation of current conditions with respect to some expected condition. This is best accomplished through an understanding of channel processes and constraints on the potential response of different channel types. Essentially, we need to know what types of response to expect in different types of channels and how to look for evidence of past response to perturbations. Moreover, since some responses to altered discharge and sediment supply potentially offset each other, how can we distinguish the impact of variations in both sediment supply and discharge? Most importantly, we must focus assessments on those aspects of a channel that are potentially responsive to perturbations of interest (e.g., changes in sediment supply, discharge, and debris flows, or the supply of large woody debris). These aspects include channel width, depth, sediment size, surface armoring, bedform roughness, the style and amount of sediment storage, pool depths, and depth of scour. Unfortunately, however, there is no theory available at present to independently predict these channel attributes based on factors such as rainfall, drainage area, bedrock type, valley confinement, and riparian vegetation. Consequently, simply measuring any of these parameters does not assess whether a channel has changed in any of these respects. We need to have a reference frame in which to evaluate observed channel characterisitics.

An understanding of watershed processes provides a valuable framework for assessing suspected past, or potential future channel changes. While our proposed channel classification provides a framework within which to analyze channel change, there is a temporal component to assessing channel change that also must be considered. Sediment routing and storage may dramatically affect both the extent to which a disturbance will propagate through a channel network and the time lag between upslope distiurbance and downstream response. Any evidence for channel change should be evaluated against potential causal mechanisms for the watershed in question.

Historical evidence provides a powerful tool for assessing channel changes, but documentation of previous channel conditions typically is inadequate. For sites where such information is available, examination of historical maps, gaging station records, surveyed cross sections, and sequential aerial photographs may reveal whether some aspects of channel conditions changed over the period of record. Even when historical data on channel conditions are available, they typically include observations on only several of the possible channel adjustments. Changes in the annual depth of scour and bed sediment size, for example, are extremely difficult to reconstruct from most historical data sources, but may be crucial for evaluating channel condition vis-á-vis ecological concerns.

Aerial photographs provide a wealth of historical data for many watersheds. The extent of riparian canopy opening provides a disturbance indicator useful for reconstructing landslide frequency when sequential aerial photographs are available (e.g., Grant et al., 1984, Grant, 1988). Spatial patterns of riparian openings can be compared on contemporaneous coverage. Applied carefully, these techniques allow both reconstruction of the timing and spatial patterns of past events, and assessment of associations with land management.

In the situation where historical data are unavailable, assessment of past channel changes must be based on estimation of expected channel conditions. The primary quantitative opportunity for this approach lies in empirical relations that describe trends in channel geometry over regional scales. In essence, we may compare present channel conditions with those expected for a comparable "undisturbed" channel in the same region. This brings up the problem of what measurable channel attributes are sensitive to management effects, am possible to collect, and have predictive capability when extrapolated from control channels to channels of interest. A number of correlations between channel attributes may be exploited to define expected channel conditions, although defining "comparable" channels and the natural variability of channel attributes are not trivial problems.

Relationships between channel width, depth, and either drainage area, or discharge (Leopold and Maddock, 1953) are similar for channels developed on comparable lithologies in the same hydrologic region. Comparison of observed channel widths and depths with those from an "undisturbed" channel with a similar drainage area can imply gross changes in channel geometry. A similar relationship is implied in data presented by Hack (1957) for sediment size and the product of drainage area and slope. Systematic trends in bed fining or coarsening may be revealed by comparing observed sediment size with data from undisturbed channels. Similarly, comparison of the amount and size of LWD in relation to channel size (e.g., drainage area or width), as well as its influence on sediment storage and pool formation, may indicate differences between disturbed and undisturbed channels. Changes in some channel attributes, such as the annual depth of scour, am difficult to determine from this type of analysis. Such considerations do, however, provide a framework for supplementing historical data in the assessment of some channel conditions.

Sediment budgets provide another method for assessing the impact of watershed management on channel networks, but are not widely used for this purpose. A sediment budget consists of identifying and quantifying sediment sources and production rates, transport processes and rates, and storage elements and residence times (Dietrich and Dunne, 1978). Using this approach, the style and rates of sediment input and processing through a watershed may be characterized (e.g., Dietrich and Dunne, 1978; Lehre, 1981;

1982; Dietrich et al., 1982; Lehre et al., 1983; Prestegaard, 1988). This approach allows comparison of sediment transport through similar watersheds under different management conditions (e.g., Kesel et al., 1992) and thus provides insight into potential sources of observed, or inferred, channel changes and recovery times, as well as identification of areas sensitive to increased sediment loading.

Indications of the relation between sediment supply and transport capacity are another tool for inferring past changes in channel morphology and sediment transport. Two quantitative techniques are available for assessing the relation between sediment load and transport capacity from channel morphology. Dietrich and others (1989) proposed a dimensionless ratio of the sediment transport rate for the surface and subsurface bed material (q_*) as a measure of sediment supply relative to transport capacity. A poorly-armored channel with a high q_* is interpreted to have a high sediment supply. Conversely, a channel with a well-developed surface armor layer and a low q_* is interpreted to have a low sediment supply relative to transport capacity. Thus, a channel with little potential for further bed surface fining ($q_* \approx 1$) must respond to future increases in sediment supply through other morphologic adjustments (e.g., channel aggradation and pool filling). Channels with low q_* values have the potential to respond to increased loads simply by textural fining without other morphological change and therefore have a higher capacity to accomodate change. This, however, does not mean that other morphologic adjustments may not occur concurrently. Although q_* provides a quantitative assessment of the capacity for bed surface texture change in response to increased sediment loading, response to episodic increases in sediment supply may be transient and difficult to monitor.

Lisle and Hilton (1991; 1992) proposed that the average ratio of the volume of fine material overlying coarser channel bed material to total pool volume (V_*) provides a measure of the amount of sediment in transport, and thus of the sediment supply. They further showed that this index correlates with perceived sediment supply and that it may vary in response to local increases in sediment supply. While both q_* and V_* provide an assessment of the contemporary balance between sediment supply and transport capacity, neither of these approaches are directly applicable in steep channels.

A number of qualitative channel attributes indicate a high sediment supply relative to transport capacity. Aggradation, braiding, medial gravel bars, presence of sand stripes, and a wide active channel are indicative of high sediment supply (Dietrich et al., 1989). In contrast, a wide inactive zone of relatively coarse sediment on the margins of a channel implies a low sediment supply (Dietrich et al., 1989). Such observations provide qualitative insight into the relative magnitude of sediment supply and transport capacity.

It is crucial to use the above indicators of relative sediment supply within a watershed context, as some channels have a naturally high sediment supply. Identification of potential sediment sources in the watershed is necessary to assess the cause of a high sediment supply relative to transport capacity. In general, the most likely candidates are landslides, bank failures, and roads or compacted areas draining into upstream channels. One approach is to base inferences of the effect of managment on relations between q_* or V_* and quantitiative measures of management intensity, such as road density or Percent of the watershed harvested. However, these parameters (q_* and V_*) primarily reflect the chronic sediment supply of a channel. Increased sediment loading from episodic disturbance may result in other dramatic changes, that may not persist in the bed surface texture once the sediment supply is reduced Conversely, measurement of these variables during passage of an episodic disturbance may erroneously evaluate the long-term supply or discharge. Thus, it is important to recognize the context within which any of these measures are meaningful.

The size of sediment supplied to a channel may change in response to watershed disturbance. The particle size distribution of the subsurface bed material approximates that of the bedload material transported by the channel (Milhous, 1973; Parker et al., 1982). If the composition of the bedload transported by a channel changes, then the composition of the subsurface alluvium should reflect this change. Thus an increase in the proportion of fine sediment supplied to a channel will result in an increase in the fine sediment content of the subsurface sediment in portions of the channel network in which sediment of the size under consideration travels as bedload. Detection of such a change, of course, depends on knowing the pre-disturbance size distribution. In threshold alluvial channels, transport of fine sediment at discharges less than bankfull reflects either the availability of fine sediment on the bed surface or introduction of fine sediment to the channel during modest storms. The most likely source is sediment produced from runoff over unpaved mad surfaces (e.g., Reid and Dunne, 1984). Sampling programs designed to assess the impact of watershed management on the fiie sediment concentration of a channel must include subsurface, as well as bed surface particle size distributions.

Suspended load also may diffuse through the bed surface during low-flow periods when bedload is not in transit. Increased suspended loading may accelerate diffusion into the channel bed and thus the concentration of fine sediment in the subsurface sediment (Lisle, 1989). The proportion of tine material in subsurface sediment is a biologically important factor that integrates the discharge regime and sources of sediment entering the channel system (Lisle and Lewis, 1992). Many workers reported that increases in the percentage of fine channel substrate are correlated with reduced salmonid survival-toemergence [see recent reviews in Chapman (1988) and Peterson and others (1992)]. Low-gradient channels, in particular, are susceptible to increased proportions of fine sediment in the channel substrate.

Fine sediment transported over an armored bed during low flow (e.g., Jackson and Beschta, 1982) could not be derived from subsurface sediment, as the bed armor is not mobilized during low-flow events. Fine sediment in transit during low-flow events must, therefore, represent either material mobilized from the surface of the channel bed (such as fine sediment stored in pools) or sediment delivered to the channel by runoff produced during storms that do not generate armor-mobilizing discharges. The latter could come from road surface runoff, or from erosion of bare ground. Thus, the production and delivery of sediment to channels during frequent storm events may control the amount of tine sediment in transport during low-flow events and this, in turn, may control the rate of fine sediment diffusion into the channel bed.

Aggradational waves may propagate through channel networks, resulting in similar, but temporally-offset impacts in different portions of the channel network. This temporal delay in downstream impacts must be. considered in addition to the spatial distribution of potential impacts. The rate at which an aggradational wave will move through a channel system is determined by the controls on the frequency of bed material movement and the typical travel distance. This is a complex, poorly-understood problem. These effects are important, however, for examining variations in response and recovery time in natural channel networks.

Several qualitative methods have been proposed for assessing channel modifications due to altered Peak flows. Pfankuch's (1978) method relies on descriptions of factors such as bank cutting, algal staining of clasts, bed material size, and perceived bed stability to assess impacts from increased peak flows. This method essentially assumes that there is only one channel type, and that all channels respond similarly to altered peak flows. We disagree with this assumption. Metzler (1992) modified this general approach to include more channel types, but her method still is based on the assumption that the effects of altered peak flow are distinct from changes in either sediment supply or debris flow impacts. While these approaches are well-suited for either describing channel conditions or predicting the likely style of future channel response, we maintain that in many instances a watershed-level analysis is needed to ascertain whether past channel change occurred from discharge or sediment supply modifications.

Visual inventories of in-channel habitat units (Hankin and Reeves, 1988) are being used at present to assess channel conditions and provide a baseline for monitoring studies in many basins in the western United States. Unfortunately, replication of habitat classification using such methods is poor (Ralph et al., 1991). Simplification of the potential habitat types might improve replicability. More importantly, however, such detailed inventories are appropriate for assessing channel conditions (as opposed to habitat availability) only when data collection is stratified by channel type and them is some control on what defines the "expected' condition.

We contend that an anaylsis of watershed processes using all available information is necessary to assess evidence for past or the style of future channel changes. Historical data sources should be examined for direct evidence of changes in channel morphology. Comparison of existing conditions to either anticipated conditions or those in undisturbed channels should be based on channel type to allow meaningful comparison. Perhaps most importantly, channel assessments should include a watershed context incorporating spatial differences in both hillslope and channel processes. Further development of scientific methods for reconstructing and predicting channel change is strongly recommended.

WATERSHED MANAGEMENT

The effectiveness with which classification of channel processes is translated into enhanced resource quality depends on the philosophy employed to implement watershed management. Input and output management define two fundamentally distinct strategies for watershed management. Output management essentially identifies areas that have been impacted and promotes their recovery or minimizes future degradation. Input management identifies areas likely to be impacted in the future and modifies management activity to minimize impacts in sensitive areas. Both approaches are necessary for effective watershed management and restoration.

At present, the most commonly used approach is output management, which relies on the premise that a watershed can be managed to prevent or respond to channel degradation below certain threshold conditions. An inherent aspect of this management strategy is that activities which do not change channel conditions beyond a prescribed threshold are considered acceptable. Only once a channel is degraded to a sufficient state do modified management prescriptions take effect. Thus, there is a tendency over time to favor resource degradation to conditions just shy of the statutory response threshold Accurate determination of response thresholds is thus extremely important Unfortunately, a major problem with this approach is that even if we were able to identify, quantify, model, and monitor potential changes in channel systems, the biologic response to these changes may remain dauntingly complex. Moreover, discrete response thresholds do not exist in many physical and biologic systems, as incremental changes in some variable or process will result in incremental changes in associated processes. Rather, response thresholds often are management tools imposed on the natural system for the convenience of the land manager. While it is difficult to imagine ever having enough information to confidently protect biologic resources through output management, this strategy does provide an effective mechanism through which to define an acceptable limit to resource degradation. As a restoration strategy, however, a threshold-based approach favors establishing specified habitat characteristics rather than conditions mat allow natural processes to persist.

Input management relies on identifying the effects of land use disturbance on natural processes and tailoring land management operations to minimize these impacts. Input management is not simple, but is potentially mom effective in preventing long-term resource degradation. Effective implementation requires knowledge of the processes governing the system of interest. The effect of potential or existing land management practices on these processes also must be known. For the case of channel networks, the processes producing, delivering, routing, and storing sediment must be ascertained. Input management focuses on how management activities alter these processes, their rates and linkages, and adjusts the style and intensity of land management to minimize impacts to these processes. The most effective input management would preserve the magnitudes and frequency of naturally-occurring processes in the system of interest. Examples of this approach are the establishment of riparian buffer zones to provide shade and a source of woody debris for channels and the redesign of logging roads to prevent drainage accumulation and delivery of fine sediment to channels.

Input and output management provide distinct, but not mutually exclusive, strategies for watershed management. Output management is necessary to identify and restore areas that already have been impacted, whereas input management is necessary to prevent further resource degradation, especially in relatively unimpacted areas. This highlights a fundamental difference between monitoring/rehabilitation efforts and programs aimed at preventing or minimizing cumulative watershed impacts. We maintain that input management is better suited to long-term watershed management, because of the complexity of interactions between physical and biologic systems and the tenuous foundation for identifying response thresholds in natural systems. Minimizing changes in the processes that create physical habitat provides the only method for assuring that all components of that habitat will be available to biologic resources. Although operational changes generally will he necessary for implementing effective input management, it provides a solid foundation for incorporating natural resource protection into land management.

Watersheds are complex systems that integrate processes acting over different scales. The wide array of processes influencing the morphology and dynamics of natural channel systems suggests no simple solutions to problems of channel degradation, let alone the attendant biologic response. We have tried to systematize channel processes into a framework for examining the potential responses of channel systems. Our approach is not exhaustive and additional methods are needed to address channel condition, potential responses, and the impact of watershed management on both geomorphic and biologic systems. It is imperative, however, that new methods be well-founded in an understanding of channel processes and watershed dynamics.

ACKNOWLEDGEMENTS

This research was supported by the SHAMW and CMER committees of the Washington State Timber/Fish/Wildlife agreement through grant TFW FY92-010. Conversations with Kate Sullivan, Bill Dietrich, and Tom Dunne were instrumental in developing the approach presented in this paper. Anne Biklé, Kevin Schmidt, Tim Abbe, Jonathan Stock, Tom Lisle, Ellen Wohl, Lee MacDonald, Ed Rashin, Patrick Reynolds, Cindy Ricks, and George Pess provided valuable comments on drafts of this report. As is customary, we are soley responsible for any shortcomings, omissions, or errors in this work. We recognize that our review of previous work incorporates only a fraction of the immense literature on stream channels and extend our apologies to those whose work we have either overlooked or omitted for the sake of brevity.

REFERENCES CITED

Abrahams, A. D., 1980, Channel link density and ground slope: Annals of the American Association of Geographers, v. 70, p. 80-93.

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

Anderson, M. G., and Burt, T. P., 1978, The role of topography in controlling throughflow generation: Earth Surface Processes, v. 3, p. 331-344.

Andrews, E. D.. 1979, Scour and fill in a stream channel, East Fork River, Western Wyoming: U.S. Geological Survey Professional Paper 1117, 49p.

Andrew, E. D., 1984, Bed material entrainment and hydraulic geometry of gravel-bed rivers in Colorado, Geological Society of America Bulletin, v. 95, p. 371-378.

Antevs, E.. 1952, Armyo-cutting and filling: Journal of Geology, v. 60, p. 375-385,

Ashida, K., Takahashi, T., and Sawada, T.. 1976, Sediment yield and transport on a mountainous small watershed: Bulletin of the Disaster Prevention Research Institute, v. 26, p. 119-144.

Ashida, K., Takahashi, T., and Sawada, T., 1981, Processes of sediment transport in mountain stream channels: In Erosion and Sediment Transport in Pacific Rim Steeplands, IAHS publ. no. 132, p. 166-178.

Ashworth, P. J., and Ferguson, R. I., 1989. Size-selective entrainment of bed load in gravel bed streams: Water Resources Research, v. 25, p. 627-634.

Band, L. E.. 1986. Topographic partition of watersheds with digital elevation models: Water Resources Research, v. 22, p. 15-24.

Band, L. E., 1989, A terrain-based watershed information system: Hydrological Processes, v. 3, p. 151-162.

Bathurst, J. C., Graf, W. H., and Cao, H. H., 1983, Bedforms and flow resistance in steep gravel-bed channels: In Mechanics of Sediment Transport, edited by B. Mutlu Sumer and A. Muller, A. A. Balkema, Rotterdam, p. 215-221.

Begin, Z. B.. 1981, The relationship between flow-shear stress and stream pattern: Journal of Hydrology, v. 52, p. 307-319.

Benda, L., 1990, The influence of debris flows on channels and valley floors in the Oregon Coast Range, U.S.A.: Earth Surface Processes and Landforms, v. 15, p. 457-466.

Benda, L., and Cundy, T. W., 1990, Predicting deposition of debris flows in mountain channels, Canadian Geotechnical Journal, v. 27, p. 409-417, 1990.

Benda, L.. and Dunne, T., 1987. Sediment muting by debris flows: In Erosion and Sedimentation in the Pacific Rim, edited by R. L. Beschta, R. Blinn, G. E. Grant, G. Ice, and F. J. Swanson, International Association of Hydrological Sciences Publication 165, p. 213-223.

Berris, S. N., and Harr, R. D., 1987, Comparative snow accumulation and melt during rainfall in forested and clear-cut plots in the western Cascades of Oregon: Water Resources Research. v. 23, p. 135-142.

Beschta, R. L., 1979, Debris removal and its effects on sedimentation in an Oregon Coast Range Stream: Northwest Science, v. 53, p. 71-77.

Beschta, R. L., 1981, Patterns of sediment and organic-matter transport in Oregon Coast Range streams: In Erosion and Sediment Transport in Pacific Rim Steeplands, edited by T. R. H. Davies and A. J. Pearce, International Association of Hydrological Sciences Publication 132, p. 179-188.

Bilby, R. E., 1984, Removal of woody debris may affect stream channel stability: Journal of Forestry, v. 82, p. 609-613.

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

Bisson, P. A., Nielson, J. L., Palmason, R. A., and Gore, L. E., 1982, A system of naming habitat types in small streams, with examples of habitat utilization by salmonids daring low stream flow: In Acquisition and Utilization of Aquatic Habitat Information, edited by N. B. Armantrout, Western Division, American Fisheries Society, Portland, p. 62-73.

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

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

Bowman, D., 1977, Stepped-bed morphology in arid gravelly channels: Geological Society of America Bulletin, v. 88, p. 291-298.

Bradley, C., and Smith, D. G., 1984, Meandering channel response to altered flow regime: Milk River, Alberta and Montana: Water Resources Research, v. 20, p. 1913-1920.

Bradley, J. B.. and Whiting, P. J., 1991, Stream characterization and a stream classification for small streams: Interim report prepared for Washington Department of Natural Resources, pp. 20-31

Bray, D. I., and Kellerhals, R., 1979, Some Canadian examples of the response of river to man-made changes: In Adjustment of the Fluvial System, edited by D. Rhodes and G. P. Williams, Kendall/Hunt, Dubuque, Iowa, p. 351-372.

Brayshaw, A. C.. 1985, Bed microtopography and entrainment thresholds in gravel-bed rivers: Geological Society of America Bulletin, v. 96, p. 218-223.

Brayshaw, A. C., Frostick, L. E., and Reid, 1.. 1983, The hydrodynamics of particle clusters and sediment entrainment in coarse alluvial channels: Sedimentology, v. 30, p. 137-143.

Brice, J. C., 1966, Erosion and depositon in the loess mantled Great Plains, Medicine Creek drainage basin, Nebraska: U.S. Geological Survey Professional Paper 352-H. 339p.

Brice, J. C., 1984, Planform properties of meandering rivers: In River Meandering, American Society of Civil Engineers, New York, p. I- 15.

Bryan, K., 1928, Historic evidence on changes in the channel of Rio Puerco, a tributary of the Rio Grande in New Mexico: Journal of Geology, v. 36, p. 265-282.

Bryant, M. D., 1980. Evolution of large, organic debris after timber harvest: Maybeso creek, 1949 to 1978: U.S.D.A. Forest Service General Technical Report PNW-101, 30p.

Buffington, J. M., Dietrich, W. E., and Kirchner, J. W., 1992, Friction angle measurements on a naturally formed gravel streambed: Implications for critical boundary shear stress: Water Resources Research, v. 28, p. 411-425.

Buffington, J. M., and Montgomery, D. R., Effects of hydraulic roughenss and sediment supply on bed surface textures in gravel-bed streams: EOS, Transactions of the American Geophysical Union, v. 73, p. 231.

Burroughs, E. R., Jr., 1984, Landslide hazard rating for portions of the Oregon Coast Range: In Proceedings, Syposium on the Effects of Forest and Land Use on Erosion and Slope Stability, Environment and Policy Institute, East-West Center, University of Hawaii, Honolulu, p. 265-274.

Burroughs, E. R., Jr., Hammond, C. J., and Booth, G. D., 198.5, Relative stability estimation for potential debris avalanche sites using field data: In Proceedings of the International Symposium on Erosion, Debris Flow and Disaster Prevention, Tsukuba, Japan, p. 335-339.

Butler, P.R., 1977, Movement of cobbles in a gravel-bed stream during a flood season: Geological Society of America Bulletin, v. 88, p. 1072-1074.

Campbell, A. J., and Sidle, R. C., X985, Bedload transport in a pool-riffle sequence of a coastal Alaska stream: Water Resources Bulletin, v. 21, p. 579-590.

Campbell, R. H., 1975, Soil slips, debris flows and rainstorms in the Santa Monica Mountains and vicinity, southern California: U.S. Geological Survey Professional Paper 851, 51p.

Carling, P. A., 1983, Threshold of sediment transport in broad and narrow natural streams: Earth Surface Processes and Landforms, v. 8, p. 1-18.

Carling, P. A., 1987, Bed stability in gravel streams, with reference to stream regulation and ecology: In River Channels: Environment and Process, edited by K. S. Richards, Blackwell, Ltd., Oxford, p. 321-347.

Chapman, D. W.. 1988, Critical review of variables used to define effects of fines in redds of large salmonids: Transactions of the American Fisheries Society, v. 117, p. 1-21.

Cherry, J., and Beschta, R. L., 1989, Coarse woody debris and channel morphology: A flume study: Water Resources Bulletin, v. 25, no. 5. p. 1031-1036.

Chin, A., 1989, Step pools in stream channels: Progress in Physical Geography, v. 13, p. 391-407

Church, M.. and Jones, D., 1982, Channel bars in gravel-bed rivers: In Gravel-BedRivers: Fluvial Processes, Engineering and Management, edited by R. D. Hey, J. C. Bathurst, and C. R. Thorne, John Wiley & Sons, Chichester, p. 291-338.

Coats, R., Collins, L.. Florsheim, J., and Kaufman, D., 1985, Channel change, sediment transport, and fish habitat in a coastal stream: Effects of an extreme event: Environmental Management, v. 9, p. 35.48.

Coffin, B. A., and Harr, R. D., 1991, Effects of forest cover on rate of water delivery to soil during riain-on-snow: Final Report for Project SH-1 (Rain-on-Snow Field Study) submitted to Sediment, Hydrology, and Mass Wasting Steering Committee, Timber/Fish/Wildlife Agreement, State of Washington Department of Natural Resources, 106p.

Cooke, R. U., and Reeves, R. W., 1976, Arroyos and Environmental Change in the American South-West: Oxford University Press, Oxford, 213p.

Costa, J. E., 1975, Effects of agriculture on erosion and sedimentation in the Piedmont Province, Maryland: Geological Society of America Bulletin, v. 86, p. 1281-1286.

Cupp, C. E., 1989, Stream corridor classification for forested lands of Washington: Report Prepared for Washington Forest Protection Association, 44p.

Daniels, R. B., and Jordan, R. H., 1966, Physiographic history of the soils, entrenched stream systems, and gullies, Harrison County, Iowa: U.S. Department of Agriculture Technical Bulletin 1348, 116p.

Davey, G. W., Doeg, T. J., and Blyth, J. D., 1987, Change in benthic sediment in the Thomson River, Victoria, during construction of the Thomson Dam: Regulated Rivers, v. 1, p. 71-84.

Davis, W. M., 1899, The Geographical Cycle: Geographical Journal, v. 14, p. 481-504.

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

Dietrich, W. E., Dunne, T., Humphrey, N., and Reid, L., 1982, Construction of sediment budgets for drainage basins: In Sediment budgets and routing in forested drainage basins, edited by F. J. Swanson, R. J., Janda, T. Dunne, and D. N. Swanston, U.S. Department of Agriculture, Forest Service. General Technical Report PNW-141, Pacific Northwest Forest and Range Experiment Station, Portland, Oregon, p. 2-23.

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

Dietrich, W. E.. and Smith, J. D., 1983, Influence of the point bar on flow through curved channels: Water Resources Research, v. 19, p. 1173-1192.

Dietrich, W. E.. Smith, J. D., and Dunne, T., 1979, Flow and sediment transport in a sand-bedded meander: Journal of Geology, v. 87, p. 305-315.

Dietrich, W. E., and Whiting, P., 1989, Boundary shear stress and sediment transport in river meanders of sand and gravel: In River Meandering, edited by S. Ideda and G. Parker, American Geophysical Union Water Resources Monograph 12, p. l-50.

Dietrich, W. E.. Wilson, C. J., and Reneau, S. L., 1986, Hollows, colluvium, and landslides in soil-mantled landscapes: In Hillslope Processes, edited by A. D. Abrahams, Allen and Unwin, Boston, Massachusetts, p. 361-388.

Dietrich, W. E., Wilson, C. J.. Montgomery, D. R., McKean, J., in press, Analysis of Erosion Thresholds, Channel Networks and Landscape Morphology Using a Digital Terrain Model, Journal of Geology.

Dinehart, R. L., 1992, Evolution of coarse-gravel bedforms: Field measurements at flood stage: Water Resources Research, v. 28, p. 2667-2689.

Dunne. T., and Black, R. D., 1970, Partial area contributions to storm runoff in a small New England watershed: Water Resources Research. v. 6, p. 1296-1311.

Dunne, T., and Leopold, L. B., 1978, Water in Environmental Planning: W. H. Freeman and Company, New York, 818p.

Ellen, S. D., and Wieczorek, G. F., 1988, Landslides, Floods, and Marine Effects of the Storm of January 3-5, 1982, in the San Francisco Bay Region, California: U.S. Geological Survey Professional Paper 1434, 310p.

Emmett, W. W., and Leopold, L. Et., 1963, Downstream pattern of riverbed scour and fill: Proceedings of the Federal Interagency Sedimentation Conference, U.S. Department of Agriculture Research Service Miscellaneous Publication 970, p. 399-409.

Ergenzinger, P., 1988, The nature of coarse material bed load transport: In Sediment Budgets, edited by M. P. Bordas and D. E. Walling, International Association of Hydrological Sciences Publication 174, p. 207-216.

Ergenzinger, P., and Schmidt, K.-Ii., 1990, Stochastic elements of bed load transport in a step-pool mountain river: In Hydrology in Mountainous Regions. II - Artificial Reservoirs, Water and Slopes, edited by R. 0. Sinniger and M. Monbaron, International Association of Hydrological Sciences Publication 194, p. 39-46.

Fahnestock, R. K., 1963. Morphology and hydrology of a glacial stream - White River, Mount Rainier, Washington: U.S. Geological Survey Professional Paper 422-A, 70 p.

Fenton, J. D., and Abbott. J. E., 1977, Initial movement of grains in a stream bed: The effects of relative protrusion: Proceedings of the Royal Scoiety of London, part A, v. 352, p. 532-537.

Florsheim, J. L., 1985, Fluvial requirements for gravel bar formation in northwestern California: Unpublished M. S. thesis, Humboldt State University, Humboldt. California, 105 p.

Frissell, C. A., 1992, Cumulative effects of land use on salmon habitat in southwest Oregon coastal streams: unpublished Ph.d. dissertation, Oregon State University, Corvallis, Oregon, 227 p

Frissell, C. A.. Liss, W. J.. Warren. C. E., and Harley, M. D., 1986, A hierarchial framework for stream habitat classification: Viewing streams in a watershed context: Environmental Management v. 10, p. 199-214.

Frissell, C. A. and Liss, W. J., 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, 51p.

Fujita, Y., 1989, Bar and channel formation in braided streams: In River Meandering, edited by S. Ikeda and G. Parker, American Geophysical Union Water Resources Monograph 12, p. 417-462.

Galay, V. 'I'., 1983, Causes of river bed degradation: Water Resources Research, v. 19, p. 1057-1090.

Germanoski, D., and Ritter, D. F., 1988, Tributary tesponse to local base level lowering below a dam: Regulated Rivers: Research and Management, v. 2, p. 11-24.

Gilbert G. K., 1909. The convexity of hill tops: Journal of Geology, v. 17, p. 344-350.

Gilbert, G.K., 1914, The transportation of débris by running water, U.S. Geological Survey Professional Paper 86, 263 pp.

Gilbert, G. K., 1917, Hydraulic-mining débris in the Sierra Nevada: U.S. Geological Survey Professional Paper 105, 154p.

Gomez, B., and Church, M., 1989, An assessment of bed load sediment transport formulae for gravel bed rivers, Water Resources Research, v. 25, p. 1161-1186.

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

Grant, G., 1988, The RAPID technique: a new method for evaluating downstream effects of forest practices on riparian zones: U.S. Department of Agriculture, Forest Service, General Technical Report PNW-GTR-220.36~.

Grant, G. E., Crozier, M. J., and Swanson, F. J., 1984, An approach to evaluating off-site effects of timber harvest activities on channel morphology: In Proceedings, Syposium on the Effects of Forest and Land Use on Erosion and Slope Stability, Environment and Policy Institute, East-West Center, University of Hawaii, Honolulu, p. 177-186.

Grant, G. E., and Mizuyama, T., 1991, Origin of step-pool sequences in high gradient streams: A flume experiment: Proceedings of the Japan-U.S. Workshop on Snow Avalanche, Landslide, Debirs Flow Prediction and Control, p. 523-532.

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

Gregory, K. J., 1971, Drainage density changes in South West England: In Exeter Essays in Geography, edited by K. J. Gregory and W. L. D. Ravenhill, Exeter, p. 33-53.

Gregory. K. J., 1976, Drainage basin adjustments and Man: Geographica Polonica, v. 34, p. 155-173.

Gregory, K. J., 1977, The context of river channel changes: In River Channel Changes, edited by K. J. Gregory, John Wiley & Sons, Inc., Chichester, p. 2-12.

Gregory, K. J., and Gardiner, V., 1975, Drainage density and climate: Zeitschrift für Geomorphologie, v. 19, p. 287-298.

Gregory, K. J., and Park, C. C., 1974, Adjustment of river channel capacity downstream from a reservoir: Water Resources Research, v. 10, p. 870-873.

Gregory, K. J., and Park, C. C., 1976, Stream channel morhpology in northwest Yorkshire: Revue de Geomorphologie Dynamique, v. 25, p. 63-72.

Gregory, S. V., Swanson, F. J.. McKee. W. A., Cummins, K. W., 1991, An ecosystem perspective of riparian zones: Bioscience, v. 41, p. 540-551.

Griffiths, G. A., 1980, Stochastic estimation of bed load yield in pool-and-riffle mountain streams: Water Resources Research, vol. 16. no. 5, p. 931-937.

Hack, J. T., 1957, Studies of longitudinal stream profiles in Virginia and Maryland: U.S. Geological Survey Professional Paper 294-B, 97p.

Hammer, T. R., 1972, Stream channel enlargement due to urbanization: Water Resources Research, v. 8, p. 1530-1540.

Hammond, F. D. C., Heathershaw, A. D., and Langhome, D. N., 1984, A comparison between Shields' threshold criterion and the movement of loosely packed gravel in a tidal channel: Sedimentology, v. 31, p. 51-62.

Hankin, D. G., and Reeves, G. H., 1988, Estimating total fish abundance and total habitat area in small streams based on visual estimation methods: Canadian Journal of Fisheries and Aquatic Science, v. 45, p. 834-844.

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

Harr, R. D., 1983, Potential for augmenting water yield through forest practices in western Washington and western Oregon: Water Resources Bulletin, v. 19, p. 383-393.

Harr, R. D., 1986, Effects of clearcutting on rain-on-snow runoff in western Oregon: a new look at old studies: Water Resources Research, v. 22, p. 1095-2000.

Harr, R. D., Fredriksen, R. L., and Rothacher, J., 1979, Changes in streamflow following timber harvest in Southwestern Oregon: U.S. Forest Service Research Paper PNW-249, 22p.

Harvey, A. M., 1991, The influence of sediment supply on the channel morphology of upland streams: Howgill Fells, northwest England: Earth Surface Processes and Landforms, v. 16, p. 675-684.

Hassan. M. A., 1990, Scour, till, and burial depth of coarse material in gravel bed streams: Earth Surface Processes and Landforms, v. 15, p. 341-356.

Hassan, M. A., Church. M., and Schick, A. P., 1991, Distance of movement of coarse particles in gravel bed streams: Water Resources Research, v. 27, p. 503-511.

Heede. B. H., 1972, Influences of a forest on the hydraulic geometry of two mountain streams: Water Resources Bulletin, v. 8, p. 523-530.

Heede, B. H., 1981, Dynamics of selected mountain streams in the western United States of America: Zeitschrift für Geomorphologie, v. 25, p. 17-32.

Helley, E. J., and LaMarche, V. C., Jr., 1973, Historic flood information for northern California streams from geologic and botanical evidence: U.S. Geological Survey Professional Paper 485-E, 16p.

Henderson, F.M., 1963, Stability of alluvial channels, Transactions of the American Society of Civil Engineers, v. 128. Part I, p. 657-686.

Hickey, J. J., 1969, Variations in low-water streambed elevations at selected stream-gaging stations in northwestern California: U.S. Geological Survey Water Supply Paper 1879-E, 33p.

Horton, R. E., 1945, Erosional development of streams and their drainage basins; Hydmphysical approach to quantitative morphology: Geological Society of America Bulletin, v. 56, p. 275-370.

Howard, A. D., 1971. Simulation of stream networks by headward growth and branching: Geographical Analysis, v. 3, p. 29-50.

Howard, A. D., 1980, Thresholds in river regimes: In Thresholds in Geomorphology, edited by D. R. Coates and J. D. Vitek, Allen & Unwin, London, p. 227-258.

Howard, A. D., 1987, Modelling fluvial systems: Rock-, gravel-, and sand-bed channels: In River Channels: Environment and Process, edited by K. S. Richards, Blackwell, Oxford, p. 69-94.

Howard, A. D. and Kerby, G., 1983, Channel changes in badlands: Geological Society of America Bulletin, v. 94, p. 739-752.

Hughes, R. M.. and Omernik, J. M., 1981, Use and misuse of the terms watershed and stream order: American Fisheries Society Warmwater Streams Syposium, p. 320-326.

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

Ikeya, H., 1981, A method for designation for areas in danger of debris flow: In Erosion and Sediment Transport in Pacific Rim Steeplands, edited by T. R. H. Davies and A. J. Pearce, International Association of Hydrological Sciences Publication 132, p. 576-588.

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

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

James, L. A., 1989, Sustained sediment storage and transport of hydraulic gold mining sediment in the Bear River, California: Annals of the Association of American Geographers, v. 79, p. 570-592.

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

Janda, R. J., Nolan. K. M., Harden, D. R., and Colman, S. M., 1975, Watershed conditions in the drainage basin of Redwood Creek, Humboldt County. California, as of 1973: U.S. Geological Survey Open-File Report 75-568, 266p.

Johnson, A. C., 1991, Effects of landslide-dam-break floods on channel morphology: Unpublished MS. thesis, University of Washington, Seattle, Washington, 90p.

Judd, H. E., 1964, A study of bed characteristics in relation to flow in rough, high-gradient natural channels: Unpublished Ph.D. dissertation, Utah State University, Logan, Utah.

Kappesser, G.. 1992. Riffle armor stabiiity index, version 3.1: unpublished manuscript 7p.

Keller, E. A., 1971, Areal sorting of bed-load material: the hypothesis of velocity reversal: Geological Society of America Bulletin, v. 82, p. 753-756.

Keller, E. A., and Melhom, W. N., 1978, Rhythmic spacing and origin of pools and riffles: Geological Society of America Bulletin, v. 89, p. 723-730.

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

Kellerhals, R., Church, M., and Bray, D. I., 1976, Classification and analysis of river processes: Journal of the Hydraulics Division, American Soceity of Civil Engineers, v. 102, p. 813-829.

Kelsey, H. M.. 1980, A sediment budget and an analysis of geomorphic process in the Van Dozen River basin, north coastal California, 1941-1975: Geological Society of America bulletin, v. 91, p. 1119-1216.

Kennedy, J. F., 1975, Hydraulic relations for alluvial streams: In Sedimentation Engineering, edited by V. Vanoni, American Society of Civil Engineers Manual 54, p. 114-154.

Kesel, R. H., Yodis, E.G., and McCraw, D. J., 1992, An approximation of the sediment budget of the lower Mississippi River prior to major human modification: Earth Surface Processes and Landforms, v. 17, p. 711-722.

Kinerson, D., 1990, Bed surface response to sediment supply: Unpublished M.S. thesis, University of California, Berkeley, California, 420p.

Kirchner, J., Dietrich, W. E., Iseya, F., and Ikeda, H., 1990, The variability of critical boundary shear stress, friciton angle, and grain protrusion in water-worked sediments: Sedimentology, v. 37, p. 647-672.

Kirkby, M.J., 1971, Hillslope process-response models based on the continuity equation: Institute of British Geographers Special Publication No 3, p. 15-30.

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

Komar, P. D., 1987, Selective grain entrainment by a current from a bed of mixed sizes: a reanalysis: Journal of Sedimentary Petrology, v. 57, p. 203-211.

Komar, P. D., and Li., Z., 1986, Pivoting analyses of the selective entrainment of sediments by size and shape with application to gravel threshold: Sedimentology, v. 33, p. 425-436.

Kondolf, G. M., Cada, G. F., Sale, M. J., and Felando, T., 1991, Distribution and stability of potential salmonid spawning gravels in steep boulder-bed streams of the eastern Sierra Nevada: Transactions of the American Fisheries Society, v. 120, p. 177-186.

Kondolf, G. M., and Curry, R. R., 1986, Channel erosion along the Cannel River. Monterey County, California, Earth Surface Processes, v. 11, p. 307-319.

Krumbein, W. C., and Shreve, R. L., 1970, Some statistical properties of dendritic channel networks: Technical Report 13, Office of Naval Research Task 389-150, Department of Geological Sciences, Northwestern University, Evanston, Ill.

Lacey, G., 1930, Stable channels in alluvium, Proceedings of the Institution of Civil Engineers, v. 229, p. 259-292.

Lane, E.W., 1953, Design of stable channels, Proceedings of the American Society of Civil Engineers, v. 79, Paper no. 280, p. 280-1-280-31.

Lane, E. W., 1934. Retrogression of levels in riverbeds below dams: American Society of Civil Engineers Proceedings, v. 93, p. l-14.

Lane, E. W., 1955, The importance of fluvial morphology in hydraulic engineering: Proceedings of the American Society of Civil Engineers, v. 81, p. l-17.

Langbein, W. B., and Leopold, L. B.. 1968, River channel bars and dunes--theory of kinematic waves: U.S. Geological Survey Proffesional Paper, 422-L. 20 p.

Laronne, J. B.. and Carson, M. A., 1976, Interrelationships between bed morphology and bed-material transport for a small gravel-bed channel: Sedimentology, v. 23, p. 67-85.

Lehre, A. K., 1981, Sediment budget of a small California Coast Range drainage basin “ear San Francisco: In Erosion and Sediment Transport in Pacific Rim Steeplands, edited by T. R. H. Davies and A. J. Pearce, International Association of Hydrological Sciences Publication 132. p. 123-139.

Lehre, A. K., 1982, Sediment budget of a small Coast Range drainage basin in north-central California: In Sediment Budgets and Routing in Forested Drainage Basins, edited by F. J. Swanson, R. J. Janda, T. Dunne, and D. N. Swansto”, U.S. Forest Service Pacific Northwest Forest and Range Experiment Station General Technical Report PNW-141, p. 67-77.

Lehre, A. K., Collins, B. D., and Dunne, T., 1983, Post-eruption sediment budget for the North Fork Toutle River Drainage, June 1980-June 1981: Zeitschrift für Geomorphologie, Supplementband 46, p. 143-163.

Leliavsky, S., 1935, An Introdution to Fluvial Hydraulics, Constable, London, 275 p.

Leopold, L. B., 1953. Downstream change of velocity in rivers: American Journal of Science, v. 251. p. 606-624.

Leopold. L. B., 1968, Hydrology for urban land planning - a guidebook on the hydrologic effects of urban land use: U.S. Geological Survey Circular 554, 18p.

Leopold, L. B., 1973, River channel change with time: an example: Geological Society of America Bulletin, v. 84, p. 1845-1860.

Leopold, L. B., 1982. Water surface topography in river channels and implications for meander development: In Gravel-Bed Rivers, edited by R. D. Hey, J.C. Bathurst, and C. R. Thorne, John Wiley and Sons, Ltd., Chichester, p 359-388.

Leopold, L. B., and Emmett, W. W., 1984, Bedload movement and its relation to scour: In River Meandering, Proceedings of the Conference Rivers '83. edited by C. M. Elliot, ACSE, New York, p. 640-649.

Leopold, L. B., and Maddock, T. Jr., 1953, The hydraulic geometry of stream channels and some physiogmphic implications: US. Geological Survey Professional Paper 252, 56p.

Leopold. L. B., and Wolman, M. G., 1957, River channel patterns: braided, meandering and straight: U.S. Geological Survey Professional Paper 282-B, p. 39-85.

Leopold, L. B., and Wolman, M. G., 1960, River meanders: Geological Society of America Bulletin, v 71, p. 769-794.

Leopold, L. B., Wolman, M. G., and Miller, J. P., 1964, Fluvial Processes in C&morphology: W. H. Freeman and Company, 522p.

Leopold, L. B., Emmett, W. W. , and Myrick, R. M., 1966: Channel and hillslope processes in a semiarid area, New Mexico, U.S. Government Printing Office, 60 p.

Li. R.M., Simons, D.B., and Stevens, M.A.. 1976, Morphology of cobble streams in small watersheds. Journal of the Hydraulics Division, American Society of Civil Engineers, v. 102, p. 1 101- 1117.

Lindley, E. S. 1919. Regime channels. Proceedings of the Punjab Engineering Congress, v. 7, p. 63-74.

Lisle, T. E., 1979, A sorting mechanism for a riffle-pool sequence: Bulletin of the Geological Society of America, v. 90, p. 1142-I 157.

Lisle, T. E., 1981, The recovery of aggraded stream channels at gauging stations in northern California and southern Oregon: In Erosion and Sediment Transport in Pacific Rim Steeplands, edited by T. R. H. Davies and A. J. Pearce, International Association of Hydrological Sciences Publication 132, p. 189-211.

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

Lisle, T. E., 1986. Stabilization of a gravel channel by large streamside obstmctions and bedrock bends, Jacoby Creek, northwestern California: Geological Society of America Bulletin, v. 97, p. 999-1011.

Lisle, T. E., 1989, Sediment transport and resulting deposition in spawning gravels, north coastal California, Water Resources Research, v. 25, p. 1303-1319.

Lisle, T., and Hilton, S., 1991, Fine Sediment in pools: an index of how sediment is affecting a stream channel: U.S. Department of Agriculture, Forest Service, R-5 Fish Habitat Relatinship Technical Bulletin Number 6, 7p.

Lisle, T., and Hilton, S.. 1992, The volume of fine sediment in pools: An index of sediment supply in gravel-bed streams: Water Resources Bulletin, v. 28, p. 371-383.

Lisle, T. E., and Lewis, J., 1992, Effects of sediment transport on survival of salmonid embryos in a natural stream: A simulation approach, Canadian Journal of Fisheries and Aquatic Science, v. 49, p. 2337-2344.

Lisle, T. E., Ikeda, H.. and Iseya, F., 1991, Formation of stationary alternate bars in a steep channel with mixed-size sediment: A flume experiment: Earth Surface Processes and Landforms, v. 16, p. 463-469.

Lisle, T. E., and Madej, M. A., in press, Spatial variation in armouring in a channel with high sediment supply: In Dynamics of Gravel-bed Rivers, edited by P. Billi, R. Hey, and C. Thorne, John Wiley and Sons, Chichester.

Little, W. C.. and Mayer, P. G., 1976, Stability of channel beds by armoring: Journal of the Hydraulics Division, American Society of Civil Engineers, v. 102, p. 1647-1661.

Lotspeich, F. B., 1980, Watersheds as the basic ecosystem: this conceptual framework provides a basis for natural classification system: Water Resources Bulletin, v. 16, p. 581-586.

Lotspeich, F. B.. and Platts, W. S., 1982, An integrated land-aquatic classification system: North American Journal of Fisheries Management, v. 2, p. 138-149.

Lyons, J. K., and Beschta, R. L., 1983. Land use, floods and channel changes: Upper Middle Fork Willamette River, Oregon (1936-1980): Water Resources Research, v. 19, p. 463471.

Madej, M. A., 1978, Response of a stream channel to an increase in sediment load: unpublished M.S. Thesis, University of Washington, Seattle, Washington, 111p.

Madej, M. A., 1982, Sediment transport and channel changes in an aggrading stream in the Puget Lowland, Washington: In Sediment budgets and routing in forested drainage basins, edited by F. J. Swanson, R. J., Janda, T. Dunne, and D. N. Swanston, U.S. Department of Agriculture, Forest Service General Technical Report PNW-141, Pacific Northwest Forest and Range Experiment Station, Portland, Oregon, p. 97-108.

Mark, D. M.. 1983, Relations between field-surveyed channel networks and map-based geomorphometric measures, Inez, Kentucky: Annals of the Association of American Geographers, v. 73, p. 358-372.

Marron, D. C.. 1985, Colluvium in bedrock hollows on steep slopes, Redwood Creek drainage basin, northwestern California: In Soils and Geomorphology, edited by P. D. Jungerius, Catena Supplement 6, Catena Verlag, Cremlingen, Federal Republic of Germany, p. 59-68.

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

Maxwell, J. C., 1960, Quantitative geomorphology of the San Dimas Experimental Forest, California: Technical Report 19, Office of Naval Research project NR 389-042, Department of Geology, Columbia University, New York.

Megahan, W., Platts, W. S., and Kulesza, B., 1980, Riverbed improves over time: South Fork Salmon: In Proceedings of the Symposium on Watershed Management, American Society of Civil Engineers, New York, p. 380-395.

Megahan, W.F., 1982, Channel sediment storage behind obstructions in forested drainage 'basins draining the granitic bedrock of the Idaho batholith: In Sediment Budgets and Routing in Forested Drainage Basins, edited by F. J. Swanson, R. J. Janda, T. Dunne, and D. N. Swanston, US. Forest Service General Technical Report PNW-141, p. 114-121.

Melton, F. A., 1936. An empirical classification of flood-plain streams: Geographical Review, v. 26, p. 593409.

Metzler, J., 1992, Stream channel conditions assessment: A methodology to evaluate channel damage related to increased peak flow: Jones & Stokes Associates, Bellevue, Washington, 5p.

Meyer-Peter, E., and R. Müller, 1948, Formulas for bed-load transport, Second Meeting of the International Association for Hydraulic Structures Research, Stockholm, p. 39-64.

Milhous, R. T., 1973, Sediment transport in a gravel-bottom stream: Unpublished Ph.D. dissertation, Oregon State University, Corvallis, Oregon, 232 p.

Montgomery, D. R., 1991, Channel Initiation and Landscape Evolution: Unpublished Ph.D. dissertation, University of California, Berkeley, 421p.

Montgomery. D. R.. submitted, Road drainage concentration and the stability of natural landscapes: Water Resources Research.

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

Montgomery, D. R.. and Dietrich, W. E., 1989, Source areas. drainage density and channel initiation: Water Resources Research, v. 25, p. 1907-1918.

Montgomery, D. R., and Dietrich, W. E., 1992, Channel initiation and the problem of landscape scale: Science, v. 255, p. 826-830.

Montgomery, D. R., and Dietrich, W. E., in press, Landscape Dissection and Drainage Area/Slope Thresholds, in "Theory in Geomorphology" edited by M. J. Kiikby, John Wiley & Sons.

Montgomery, D. R., and Dietrich, W. E., submitted, A physically-based digital terrain model for debris flow hazard assessment: Water Resources Research.

Montgomery, D. R.. Wright, R. W., and Booth, T., 1991, Debris flow hazard mitigation for colluvium-tilled swales: Bulletin of the Association of Engineering Geologists, v. 28, p. 303-323.

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

Morris, D. G., and Heerdegen, R. G., 1988, Automatically derived catchment boundaries and channel networks and their hydrological applications: Geomorphology, v .1, p. 131-141.

Mosley, M. P.. 1981, The influence of organic debris on channel morphology and bedload transport in a New Zealand forest stream: Earth Surface Processes and Landforms, v. 6, p. 571-579.

Mosley, M. P., 1987, The classification and characterization of rivers: In River Channels: Environment and Process, edited by K. Richards, Blackwell, Oxford, p. 294-320.

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

Naiman, R. J., Lonzarich, D. G., Beechie, T.J., and Ralph. S. C., 1991, General principles of classification and the assessment of conservation potential in rivers: In River Conservation and Management, edited by P. J. Bmn, and G. E. Petts, John Wiley and Sons, Inc., Chichester. p. 93-123.

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

Nanson, G. C., 1974, Bedload and suspended-load transport in a small, steep, mountain stream: American Journal of Science, vol. 274, p. 471-486.

Weller, R. J., 1988. A comparison of channel erosion in small urban and rural catchments, Armidale, New South Wales: Earth Surface Processes and Landforms, v. 13, p. 1-7.

Nelson , J. M., and Smith, J. D., 1989, Evolution and stability of erodible channel beds: In River Meandering, edited by S. Ikeda and G. Parker, American Geophysical Union Water Resources Monograph 12, p. 321-377.

Nolan, K. M., and Janda, R. J., 1979, Recent history of the main channel of Redwood Creek, California: In A Field Trip to Observe Natural and Management Related Erosion in Franciscan Terrane of Northern California, Geological Society of America Cordilleran Section, Boulder, Colorado, pp. X-l-16.

Nolan, K. M., and Marron, D. C., 1985, Contrast in stream-channel response to major storms in two mountainous areas of California: Geology. v. 13, p. 135-138.

Nolan, K. M.. and Marron, D. C., 1988, Stream channel response to the storm: In Landslides, Floods, and Marine Effects of the Storm of January 3-5, 1982, in the San Francisco Bay Region, California, edited by S. D. Ellen and G. F. Wieczorek, U.S. Geological Survey Professional Paper 1434. p. 245-264.

Nowell, A. R. M., and Church, M.. 1979, Turbulent flow in a depth-limited boundary layer: Journal of Geophysical Research, v. 84, p. 48164824.

Nunnally, N. R., 1985, Application of fluvial relationships to planning and design of channel modifications: Environmental Management, v. 9. p. 417426.

Okunishi, K., and Iida, T., 1981, Invesitgations of topsoil horizons of mountain slopes as a basis of experimental geomorphology: Bulletin of the Disaster Prevention Research Institute, Kyoto University, v. 31, p. 131-150.

Okunishi, K., and Iida, T., 1983, Evolution of hillslopes including landslides: Transactions of the Japanese Geomorphological Union, v. 2, p. 291-299.

O'Neill, M. P. and Abrahams, A. D., 1984, Objective identificaiton of pools and riffles: Water Resources Research, v. 20, p. 921-926.

Orsbom, J. F., 1990, Quantitative modeling of the relationships among basin, channel and habitat characteristics for classification and impact assessment: Report TFW-AM3-90-010, Timber-Fish-Wildlife Agreement, State of Washington Department of Natural Resources, 45p.

Orsbom, J. F., and Anderson, J. W., 1986, Stream improvements and fish response: A bioengineering assessment: Water Resources Bulletin, v. 22, p. 381-388.

Park, C. C., 1977, Man-induced changes in stream channel capacity: In River Channel Changes, edited by K. J. Gregory, John Wiley & Sons, Chichester. p. 121-144.

Park, C. C., 1981, Man. river systems and environmental impacts: Progress in Physical Geography, v. 5, p. 1-31.

Parker, G., 1976, On the cause and characteristic scales of meandering and braiding in rivers: Journal of Fluid Mechanics, v. 76, p. 457-480.

Parker, G., Klingeman, P. C., and McLean, D. G., Bedload and size distribution in paved gravel-bed streams: Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers, v. 108, p. 544-571.

Parker, G., and Peterson, A. W., 1980. Bar resistance of gravel-bed streams: Journal of the Hydraulics Division, Proceedings of the American Society of Civil Engineers. v. 106, p. 1559-1575.

Paustian, S. J., Marion, D. A., and Kelliher, D. F, 1983, Stream channel classification using large scale aerial photography for southeast Alaska watershed management In Renewable Resources Management Symposium, American Society of Photogrammetry, Falls Church, VA, p. 670-677.

Paustian S. J., and 13 others, 1992, A channel type users guide for the Tongass National Forest, Southeast Alaska, USDA Forest Service, Alaska Region R10, Technical Paper 26, 179p.

Pearce, A. J., and Watson, A., 1983, Medium-term effects of two landsliding episodes on channel storage of sediment: Earth Surface Processes and Landforms, v. 8, p. 29-39.

Perkins, S. J., 1989, Interactions of landslide-supplied sediment with channel morphology in forested watersheds: unpublished M.S. Thesis, University of Washington, Seattle, Washington, 104p.

Petch, R. A., 1988, Soil saturation patterns in steep, convergent hillslopes under forest and pasture vegetation: Hydrological Processes, v. 2, p. 93-103.

Peterson, N. P., Hendry, A., and Quinn, T. P., 1992, Assessment of cumulative effects on salmonid habitat: Some suggested parameters and target conditions: Report prepared for the Washington Department of Natural Resources and The Coordinated Monitoring, Evaluation and Research Committee, Timber / Fiih / and Wildlife Agreement, Center for Streamside Studies, University of Washington, 75p.

Petts, G. E., 1977, Channel response to flow regulation: The case of the River Derwent, Derbyshire: In River Channel Changes, edited by K. J. Gregory, John Wiley & Sons, Chichester, p. 145-164.

Petts. G. E., 1979, Complex response of river channel morphology subsequent to reservoir construction: Progress in Physical Geography, v. 3, p. 330-362.

Petts, G. E.. 1988, Accumulation of fine sediment within substrate gravels along two regulated rivers, UK: Regulated Rivers: Research and Management, v. 2, p. 141-153.

Pfankuch, D. J., 1978, Stream reach inventory and channel stability evaluation: unpublished report, US Forest Service, Northern Region, Missoula, Montana, 26 p.

Pierson, T. C., 1980, Piezometric response to rainstorms in forested hillslope drainage depressions: Journal of Hydrology, New Zealand, v. 19, p. l-10.

Pitlick, J., 1992, Flow resistance under conditions of intense gravel transport: Water Resources Research, v. 28, p. 891-903.

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

Prestegaard, K. L., 1988, Morphological controls on sediment delivery pathways: In Sediment Budgets, edited by M. P. Bordas and D. E. Walling, International Association of Hydrological Sciences Publication 174, p. 533-540.

Rahn, P. H., 1977, Erosion below main stem dams on the Missouri River: Association of Engineering Geologists Bulletin, v. 14, p. 157-181.

Ralph, S. C., Cardoso, T., Poole, G. C., Conquest, L. L., and Naiman, R. J., 1991, Ambient monitoring project biennial progress report, 1989-91 period: Center For Streamside Studies, University of Washington, Seattle, Washington, 33p.

Reid, L. M., and Dunne, T., 1984, Sediment production form forest road surfaces: Water Resources Research, v. 20, p. 1753-1761.

Reneau, S. L., and Dietrich, W. E., 1987a, The importance of hollows in debris flow studies; examples from Marin County, County, California: In Debris Flows/Avalanches: Process, Recognition, and Mitigation, edited by J. E. Costa and G. F. Wieczorek, Geological Society of America Reviews in Engineering Geology, v. '7, p. 165-180.

Reneau, S. L., and Dietrich, W. E., 1987b, Size and location of colluvial landslides in a steep forested landscape: In Erosion and Sedimentation in the Pacific Rim, edited by R. L. Beschta, T. Blinn, G. E. Grant, G. G. Ice, and F. J. Swanson, International Association of Hydrological Sciences Publication 165, p. 39-48.

Reneau, S. L., Dietrich, W. E., Wilson, C. J., and Rogers, J. D., 1984, Colluvial deposits and associated landslides in the northern San Francisco Bay area, California, USA: In Proceedings of IVth International Symposium on Landslides, International Society for Soil Mechanics and Foundation Engineering, Toronto. Canada, p. 425-430.

Reneau, S. L., Dietrich, W. E., Dorn, R. I., Berger, C. R., and Robin, M., 1986, Geomorphic and paleoclimatic implications of latest Pleistocene radiocarbon dates from colluvium-mantled hollows, California: Geology, v. 14, p. 655-658.

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

Rich, J. L., 1911, Recent stream trenching in the semi-arid portion of southwestern New Mexico, a result of removal of vegetation cover: American Journal of Science, v. 32, p. 237-245.

Richards, K., 1982, Rivers: Form and Process in Alluvial Channels, Methuen & Co. Ltd., 361 p.

Rosgen, D. L., 1985, A stream classification system: In Riparian Ecosystems and Their Management: Reconciling Conflicting Uses, edited by R. R. Johnson, C. D. Zeibell, D. R. Patton, P. F. Ffolliott, and R. H. Hamre, U.S. Forest Service General Technical Report RM-120, p. 91-95.

Sawada, T, Ashida, K., and Takahashi, T, 1983, Relationship between channel pattern and sediment transport in a steep gravel bed river: Zeitschrift für Geomorphologie, Supplementband 46, p. 55-66.

Scheidegger, A. E., 1966, Effect of map scale on stream orders: Bulletin of the International Association of Scientific Hydrology, v. 11, p. 56-61.

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

Schumm, S. A., 1956. Evolution of drainage systems and slopes in badlands at Perth Amboy, New Jersey: Geological Society of America Bulletin, v. 67, p. 597-646.

Schumm, S. A., 1963, A tentative classification of alluvial river channels: U.S. Geological Survey Circular 477, 10p.

Schumm, S. A., 1971, Fluvial Geomorphology: Channel adjustment and river metamorphosis: In River Mechanics, Volume 1, edited by H. W. Shen, Fort Collins, Colorado, p. S-l - 5-22.

Schumm, S. A., 1977, The Fluvial System: Wiley & Sons, New York, 338p.

Schumm, S. A., and Khan, H. R., 1972, Experimental study of channel patterns: Geological Society of America Bulletin, v. 83, p. 1755-1770.

Selby, M. J., 1982. Hillslope Materials and Processes: Oxford University Press, Oxford, 264 p.

Shen, H. W., and Lu, J.-Y., 1983, Development and prediction of bed armoring: Journal of Hydraulic Engineering, v. 109, p. 611-629.

Sherrard, J. J., and Erskine, W. D., 1991, Complex response of a sand-bed stream to upstream impoundment: Regulated Rivers: Research & Management v. 6, p. 53-70.

Shields, A., 1936, Anwendung det Änlichkeitsmechanik und der Turbulenzforschung auf die Geschiebebewegung, Translated by W. P. Ott and J. C. van Uchelen, California Institute of Technology, Pasadena, California, 26p.

Shreve, R. L., 1974, Variation of mainstream length with basin area in river networks: Water Resources Research, v. 10, p. 1167-1177.

Sidle, R. C., 1988, Bed load transport regime of a small forest stream: Water Resources Research, v. 24, p. 207-218.

Sidle, R. C., 1992, A theoretical model of the effects of timber harvesting on slope stability: Water Resources Research, v. 28, p. 1897-1910.

Smart, J. S., and Surkan, A. J., 1967, The relation between mainstream length and area in drainage basins: Water Resources Research, v. 3, p. 963-974.

Smith, N. D., and Smith, D. G., 1984, William River: An outstanding example of channel widening and braiding caused by bed-load addition: Geology, v. 12, p. 78-82.

Smith, R. D., and Buffington, J. M., 1991, Effects of large woody debris on channel unit dishibution in southeast Alaska, abstract presented at Watershed 1991, U. S. Forest Service Conference, Juneau, Alaska.

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

Smith, R. D., Side, R. C., Porter, P. E., and Noel, J. R., submitted, Effects on channel morphology and aquatic habitat of experimental removal of woody debris from a forest, gravel-bed stream: Earth Surface Processes and Landforms.

Sobocinski, R. W., Cerling, T. E., and Morrison, S. J., 1990, Sediment transport in a small stream based on ^{137}Cs inventories of the bed load fraction: Water Resources Research, v. 26, no. 6, p. 1177-1187.

Stanley, J. W., 1972, Retrogression of the lower Colorado River after 1935: Transactions of the American Society of Civil Engineers, v. 116, p. 943-957.

Strahler, A. N.. 1957, Quantitative analysis of watershed geomorphology: Transactions, American Geophysical Union, v. 38, p. 913-920.

Sullivan. K., 1986, Hydraulics and fish habitat in relation to channel morphology: unpublished Ph.D. dissertation, Johns Hopkins University, Baltimore, Maryland, 430 p.

Sullivan, K., Lisle, T. E., Dolloff, C. A., Grant, G. E., and Reid, L. M., 1987, Stream Channels: The link between forests and fishes: In Streamside Management: Forestry and Fishery Interactions, edited by E. 0. Salo and T. W. Candy, College of Forest Resources, University of Washington, Seattle, Washington, p. 39-97.

Takahashi, T., Ashida, K., and Sawai, K., 1981, Delineation of debris flow hazard areas: In Erosion and Sediment Transport in Pacific Rim Steeplands, edited by T. R. H. Davies and A. J. Pearce, International Association of Hydrological Sciences Publication 132, p. 589-603.

Vannote, R. L., Minshall, G. W., Cummins, K. W., Sedell, J. R., and Cushing, C. E., 1980, The river continuum concept: Canadian Journal of Fisheries and Aquatic Sciences, v. 37, p. 130-137.

Warburton, J., 1992, Observations of bed load transport and channel bed changes in a proglacial mountain stream: Arctic and Alpine Research, v. 24, p. 195-203.

Whiting, P. J., Dietrich, W. E.. Leopold, L. B., Drake, T. G., and Shreve, R. L., 1988, Bedload sheets in heterogeneous sediment: Geology, v. 16, p. 105-108.

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

Whimey, J. D., 1880, The Auriferous Gravels of the Sierra Nevada: Harvard University Museum of Comparitive Zoology Memoir 6, 569p.

Whittaker, J. G., 1987a, Sediment transport in step-pool streams: In Sediment Transport in Gravel-bed Rivers, edited by C. R. Thorne, J. C. Bathurst, and R. D. Hey, John Wiley and Sons, Chichester, p. 545-579.

Whittaker, J. G., 1987b, Modelling bed-load transport in steep mountain streams: In Erosion and Sedimentation in the Pacific Rim, IAHS publ. no. 165, p. 319-332.

Whittaker, J. G., and Davies, T. R. H., 1982. Erosion and sediment transport processes in step-pool torrents: In Recent Developments in the Explanation and Prediction of Erosion and Sediment Yield, edited by D. E. Walling, International Association of Hydrological Sciences Publication 137, p. 99-104.

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

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

Williams, G. P., and Wolman, M. G., 1984, Downstream effects of dams on alluvial rivers: U.S. Geological Survey Professional Paper 1286, 83p.

Wilson, C. J., and Dietrich. W. E., 1987, The contribution of bedrock groundwater flow to storm runoff and high pore pressure development in hollows: In Proceedings of the International Symposium on Erosion and Sedimentation in the Pacific Rim, edited by R. L., Beschta, T. Blinn, G. E. Grant, G. G. Ice, and F. J. Swanson, International Association of Hydrological Sciences Publication 165, p. 49-59.

Wohl, E. E., 1992, Gradient irregularity in the Herbert Gorge of Northeastern Australia: Earth Surface Processes and Landforms, v. 17. p. 69-84.

Wolman, M. G., 1967, Two problems involving river channel changes and background observations: Northwestern University Studies in Geography, v. 14. p. 67-107.

Zevenbergen, L. W., and Thorne, C. R., 1987, Quantititative analysis of land surface topography: Earth Surface Processes, v. 12, p. 47-56.

Zimmerman, R. C.. Goodlet, J. C., and Comer. G. H., 1967, The influence of vegetation on channel form of small streams: In Symposium on River Morphology, International Association of Hydrological Sciences Publication 75, p. 255-275.

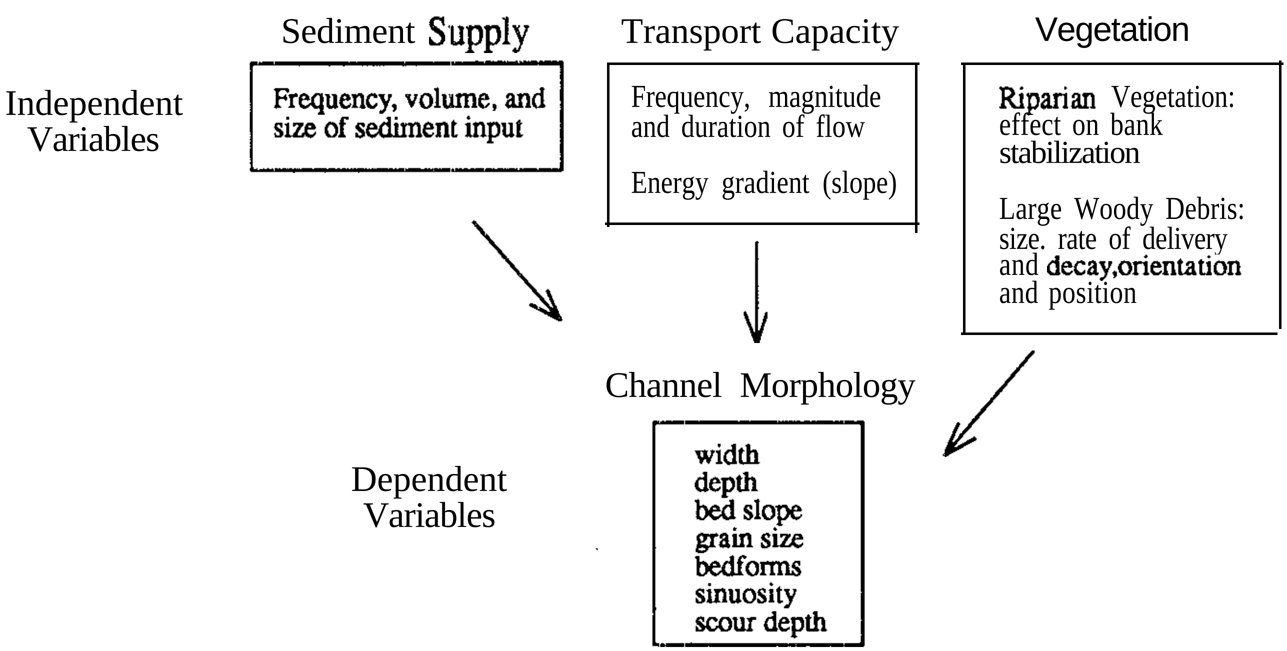

Figure 1 Illustration of the indepedent and dependent variables affecting channel morphology.

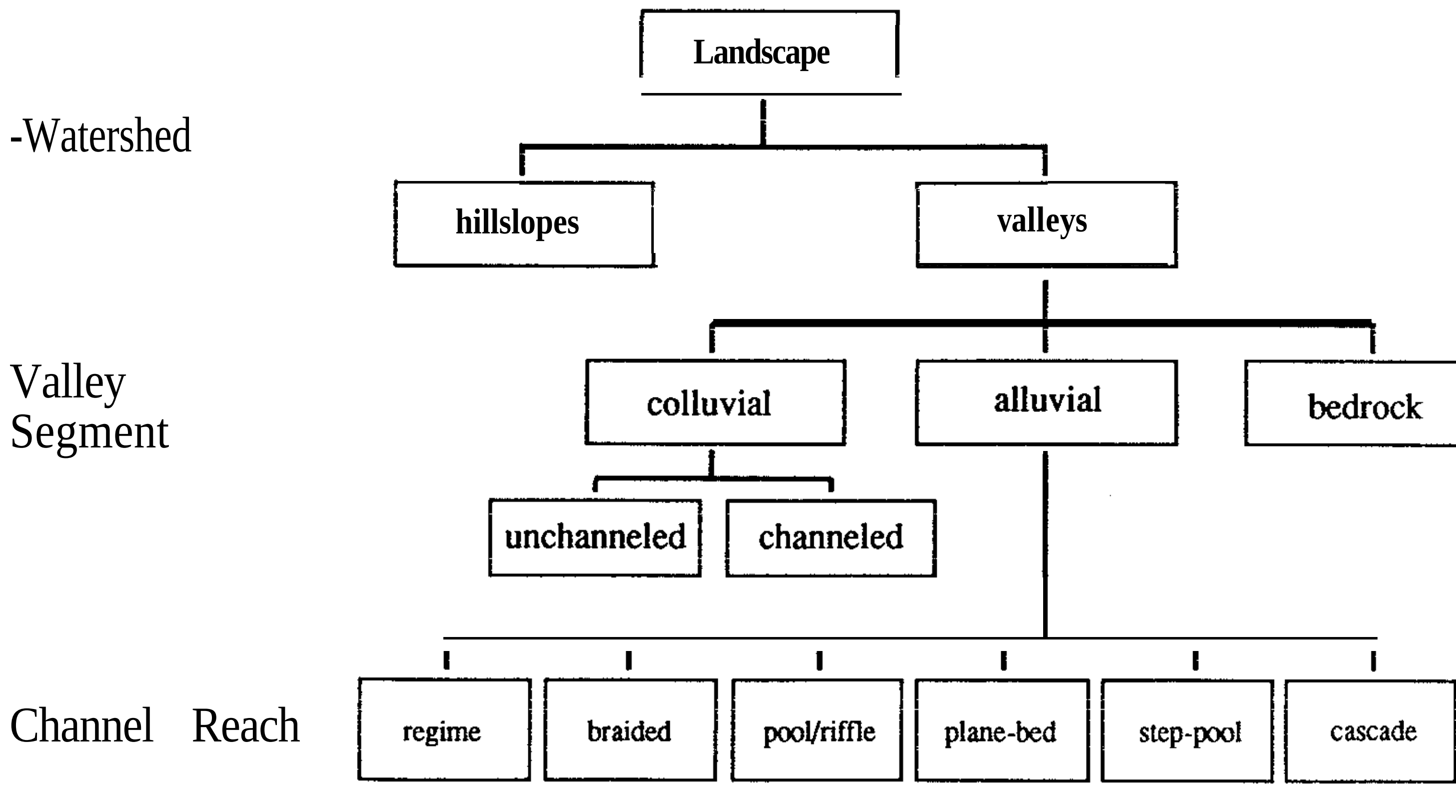

Figure 2 Landscape classification illustrating process divisions at the watershed, valley segment, and channel reach levels.

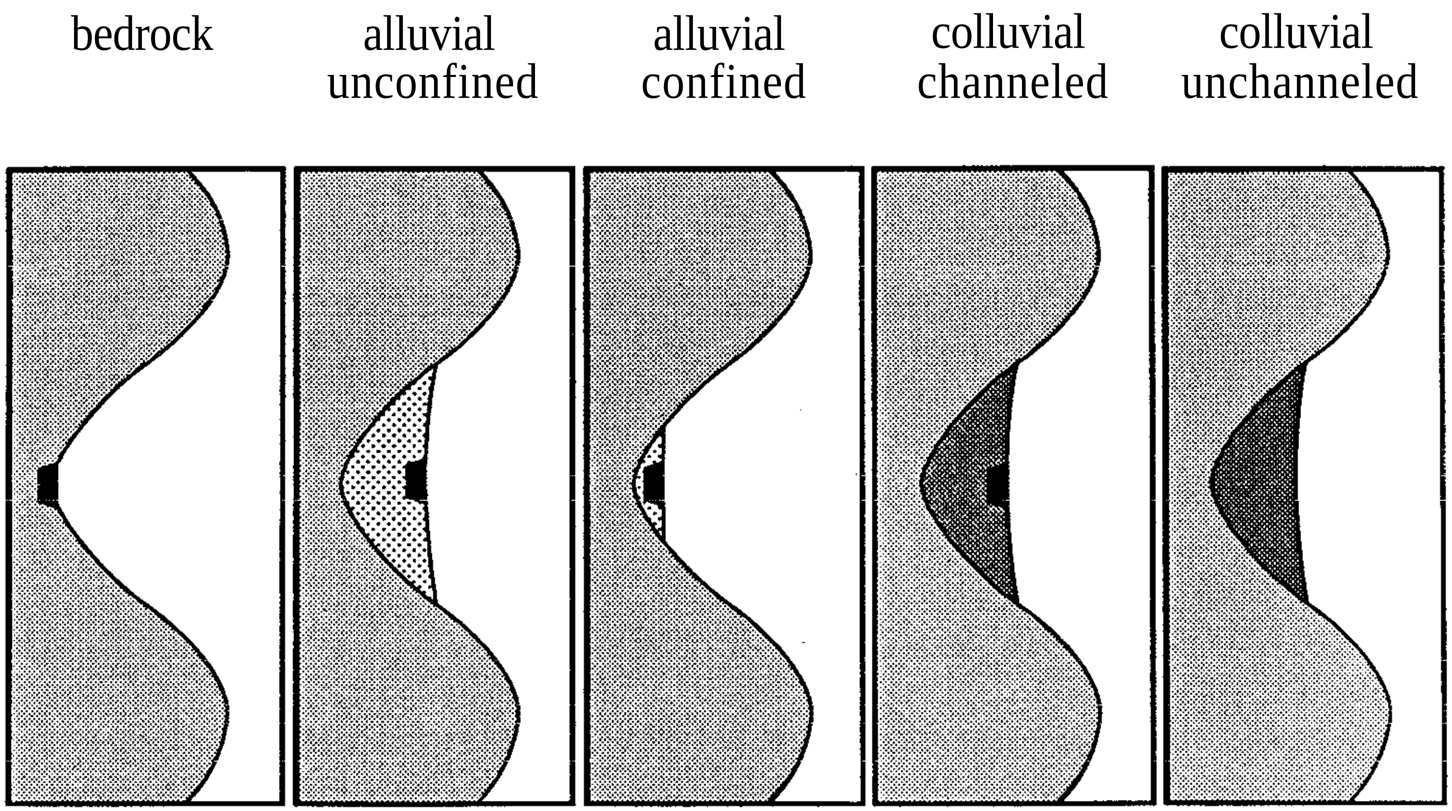

Figure 3 Illustration of valley segment morphologies.

Figure 4 Photograph of a debris flow from a hollow in the Oregon Coast Range.

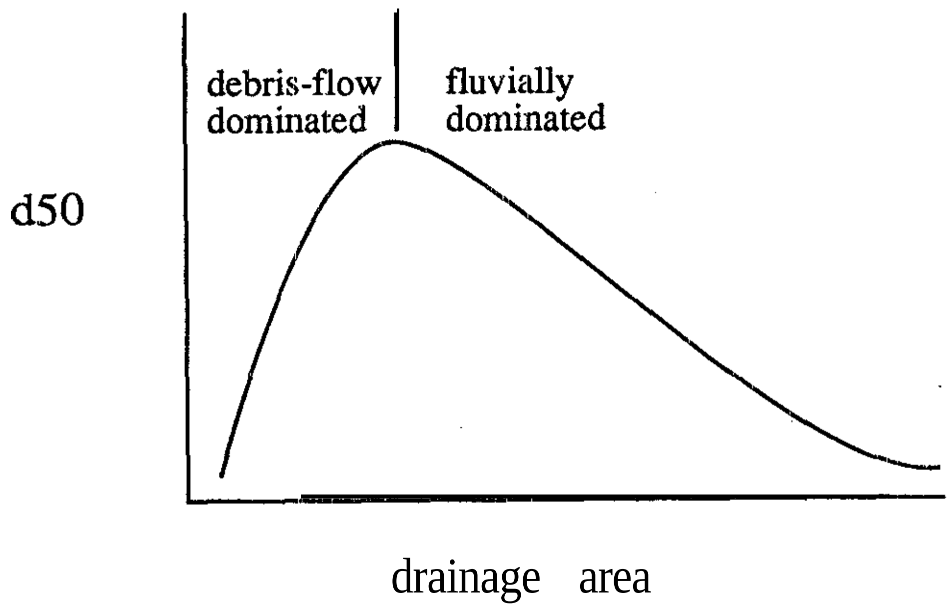

Figure 5 An inflection in the relation between median bed surface grain size (d_{50}) and drainage. area may correlate with a transition from debris flow dominated to fluvial channels.

Figure 6 Photograph

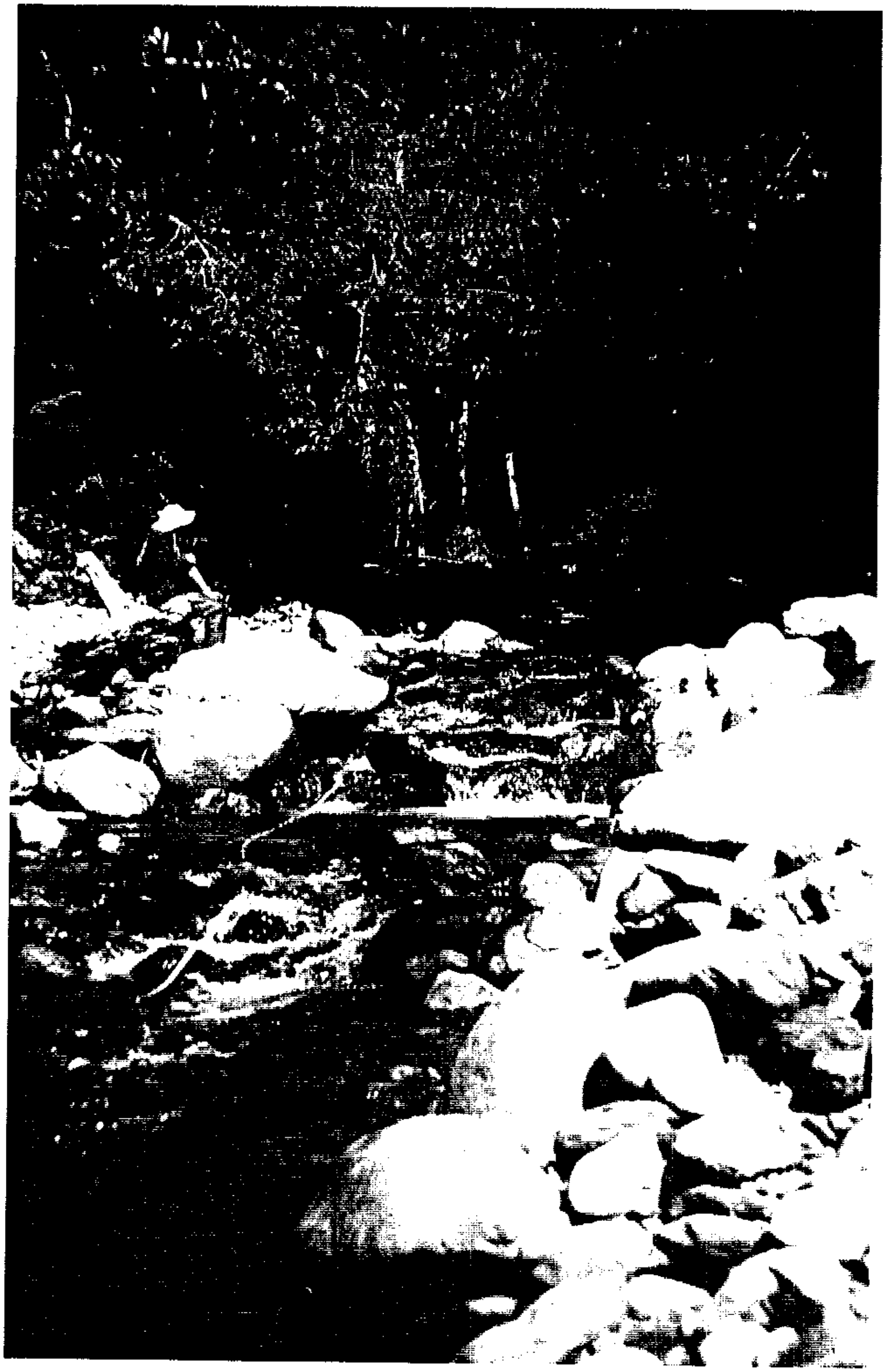

Figure 7 Photograph of a step-pool channel.

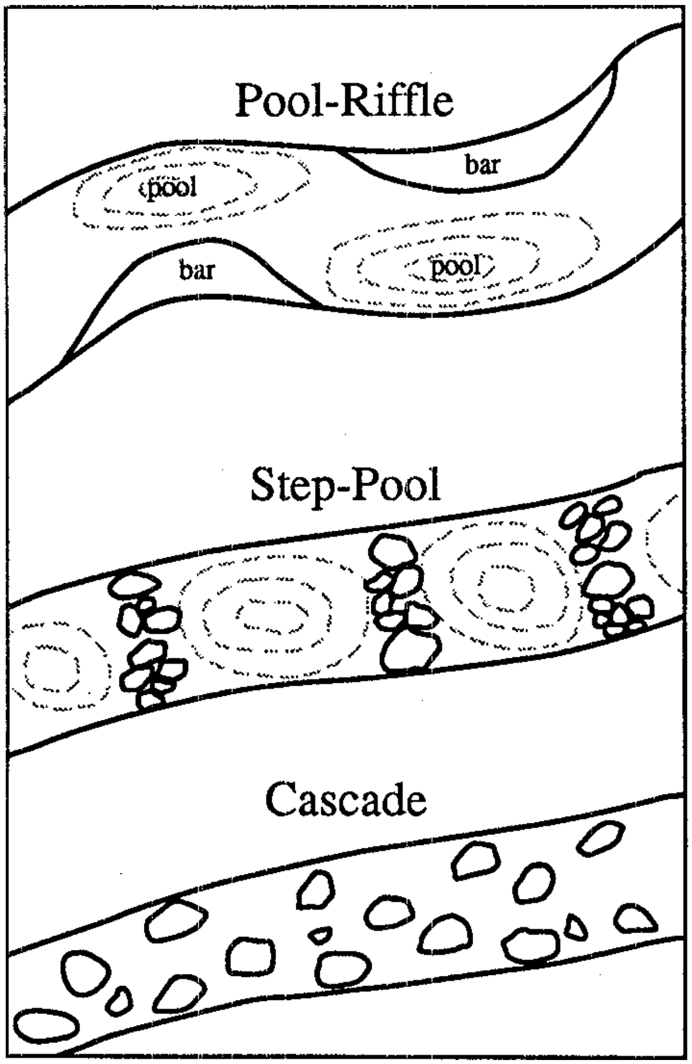

Figure 8 Illustration of differences between cascade, step-pool, and pool-riffle channel morphologies.

IFigure 9 Photograph of a plane-bed channel.

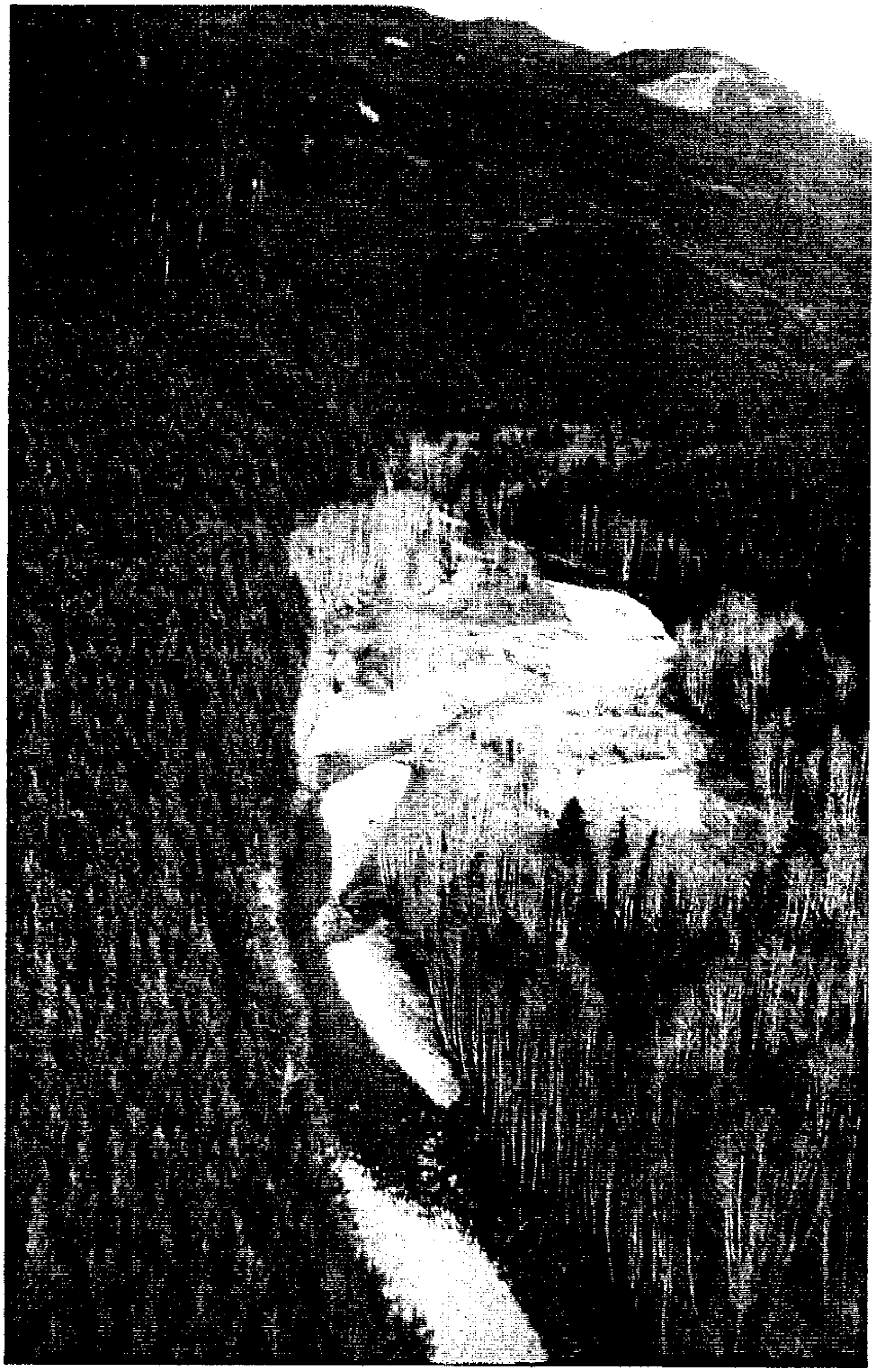

Figure 10 Photograph of a pool-riffle channel.

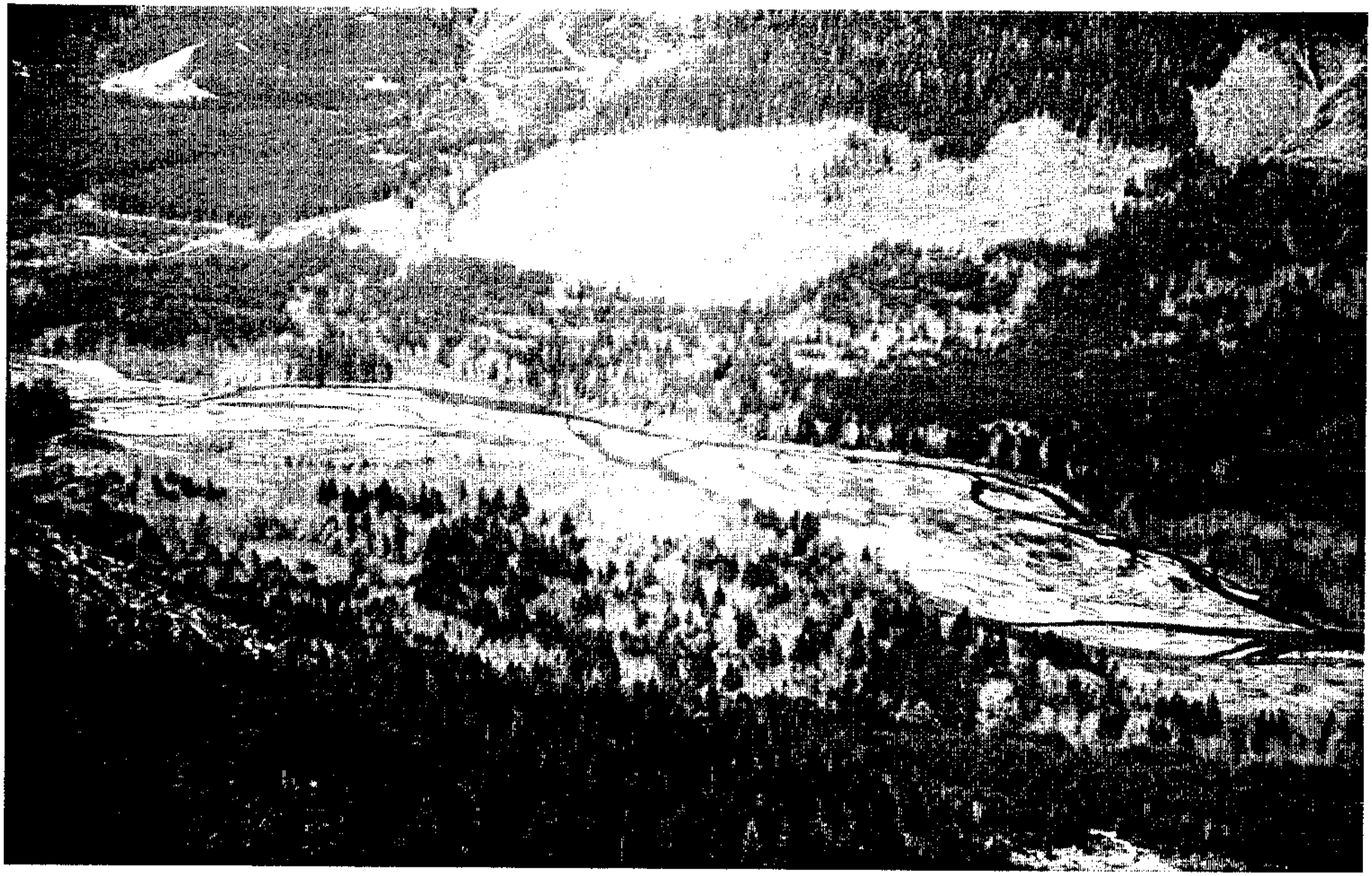

Figure 11 Photograph of a braided channel.

Figure 12 Photograph of LWD jam in small channel buttressing a significant upstream accumulation of sediment. Flow is from right to left.

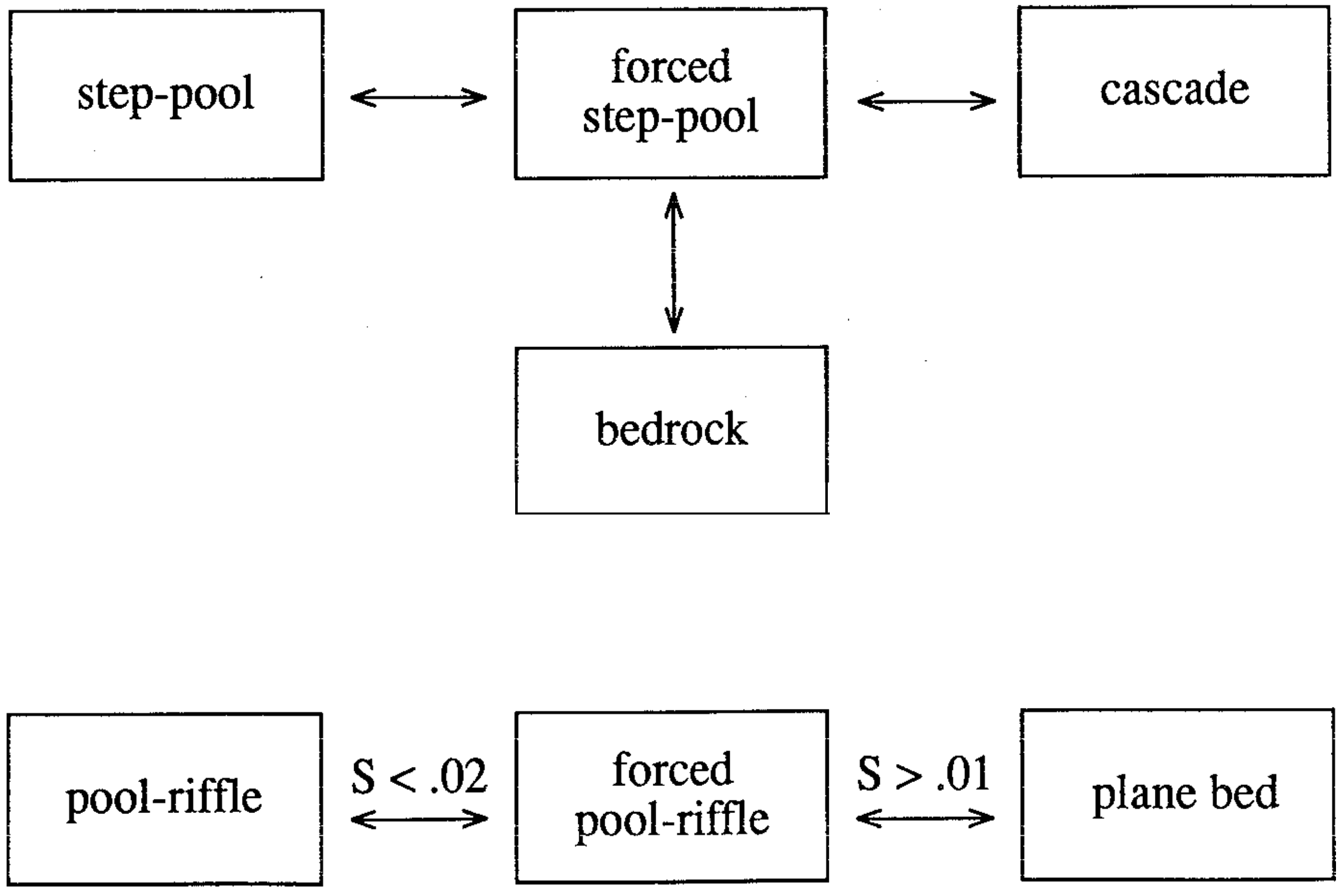

Figure 13 Potential morphologic response to woody debris removal in forced step-pool and forced pool-riffle channels.

Figure 14 Photograph of LWD on bar tops in a large, low-gradient channel.

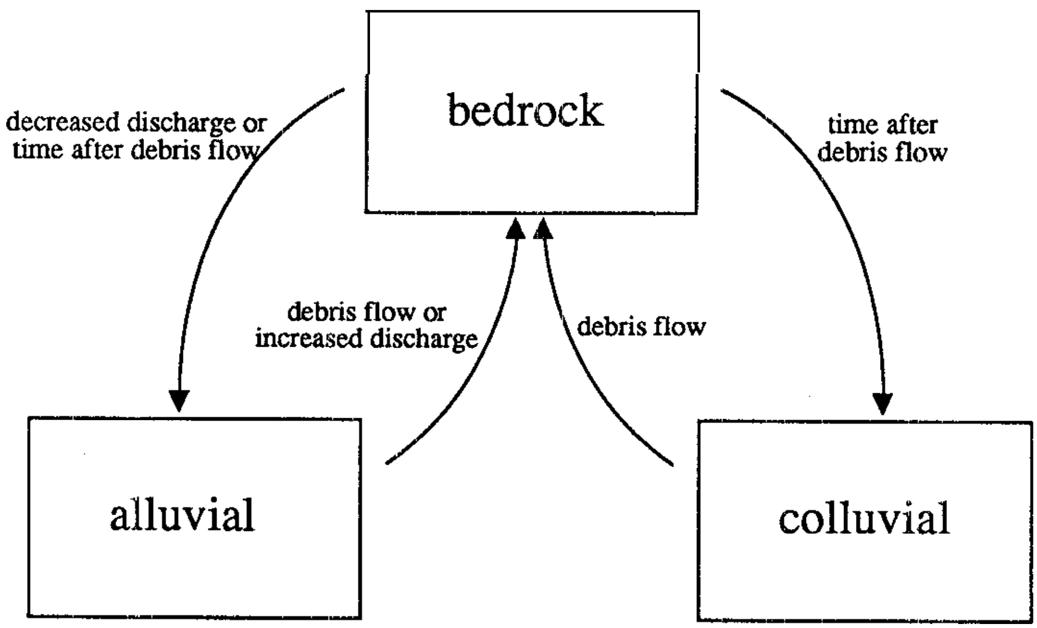

Figure 1.5 Illustration of temporal variations between channel types for steep channels subject to debris flow processes.

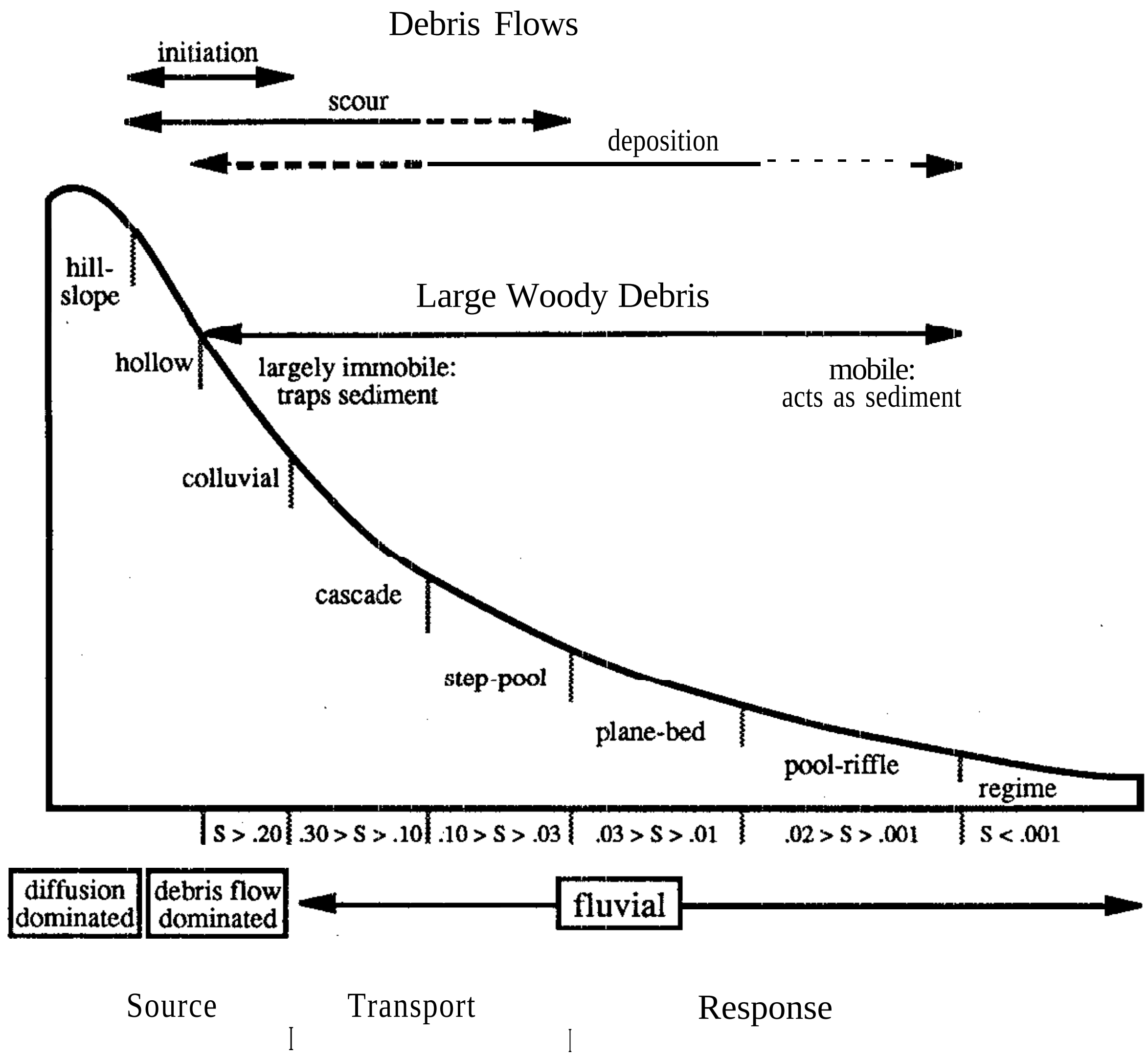

Figure 16 Illustration of idealized long profile from hill tops downslope through the channel network showing general distribution of channel types and controls on channel processes.

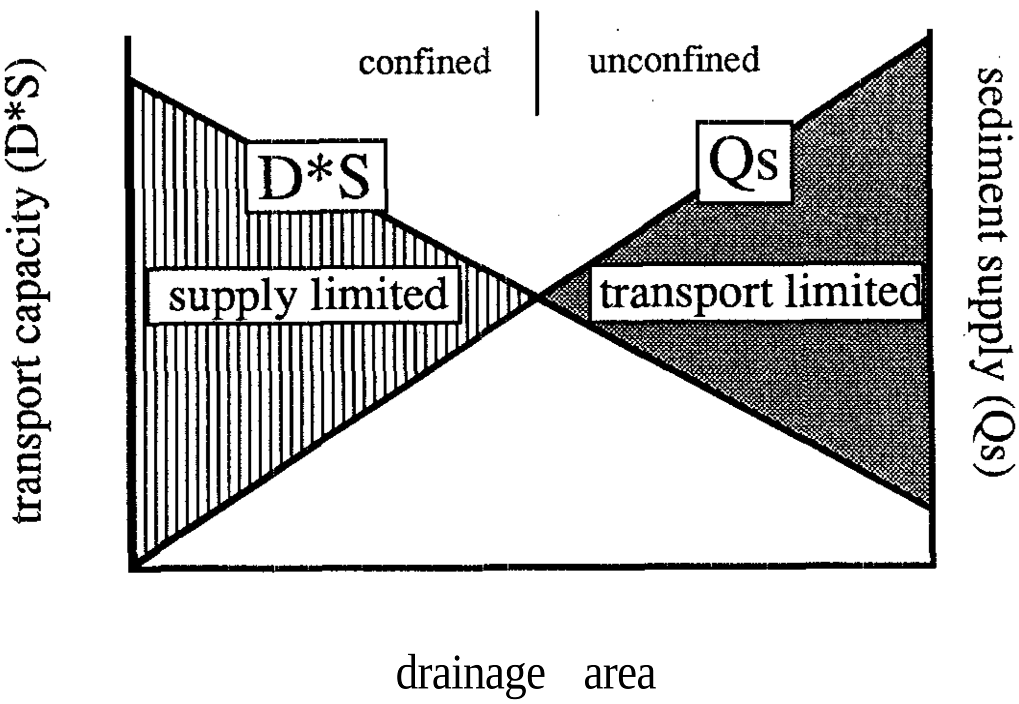

Figure 17 Illustration of general relations between transport capacity (D*S), sediment supply (Qs) and drainage area in mountain drainage basins.

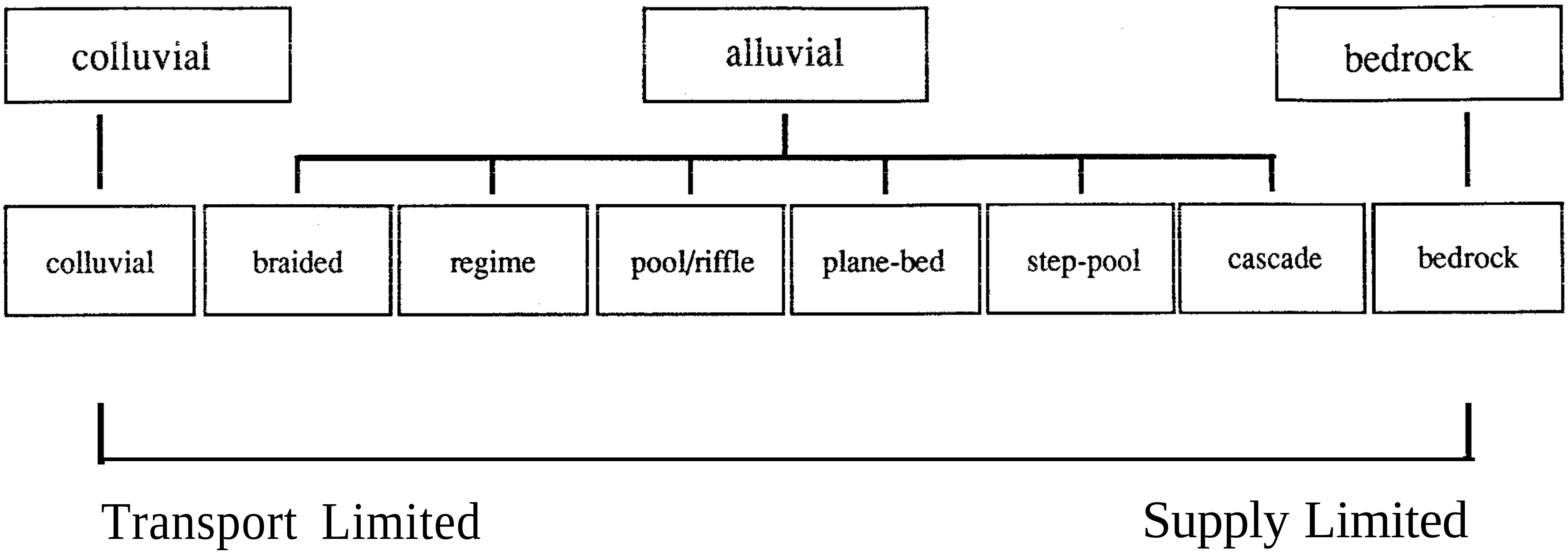

Figure 18 Schematic diagram illustrating the relative transport capacities of reach-level channel types.

	BRAIDED	REGIME	POOL-RIFFLE	PLANE-BED	STEP-POOL	CASCADE	BEDROCK	COLLUVIAL
Typical bed material	variable	sand	gravel	gravel/cobble	cobble/boulder	boulder	n/a	variable
Bedform Pattern	laterally oscillatory	multi-layered	laterally oscillatory	none	vertically oscillartory	none	-	variable
Reach Type	Response	Response	Response	Response	Transport	Transport	Transport	Source
Dominant Roughness Elements	bedforms (bars, pools)	sinuosity, bedforms (dunes, ripples, bars) banks	bedforms (bars, pools), grains, LWD, sinuosity, banks	grains, banks	bedforms (steps, pools) grains, LWD banks	grains, banks	boundaries (bed & banks)	grains, LWD
Dominant Sediment Sources	fluvial/ bank failure/ debris flows	fluvial/ bank failure/ inactive channel	fluvial/ bank failure/ inactive channel/ debris flows/	fluvial/bank failure/ debris flows	fluvial/ hillslope/ debris flows	fluvial/ hillslope/ debris flows	fluvial/hillslope debris flows	hillslope/ debris flows
Sediment Storage Elements	overbank bedforms	overbank bedforms inactive channel	overbank bedforms inactive channel	overbank inactive channel	bedforms	Lee & stoss sides of flow obstructions	-	bed
Typical Slope (m/m)	$S < 0.03$	$S < 0.001$	$0.001 < S < 0.02$	$0.01 < S < 0.03$	$0.03 < S < 0.08$	$0.08 < S < 0.30$	variable	$S > 0.20$
Typical Confinement	unconfined	unconfined	unconfined	variable	confined	confined	confined	confined
Pool Spacing (channel widths)	variable	5 to 7	5 to 7	none	1 to 4	< 1	variable	variable

19 General reach-level channel type characteristics.

	W	D	R_o	d_s	d_{50}	S	S_b
braided	x	x	x	x	x	x	x
regime	x	x	x	x	i	x	x
pool-riffle	x	x	x	x	x	x	x
plane bed	i	i	0	x	x	0	x
step-pool	i	i	i	i	i	0	i
cascade	i	i	0	0	i	0	0
bedrock	i	-	-	-	0	0	0
colluvial	i	i	0	i	i	0	x
hollow	-	-	-	-	-	-	x
hillslope	-	-	-	-	-	-	x

x likely　　0 unlikely

i possible　　- not applicable

Figure 20 Generalized channel response potential.

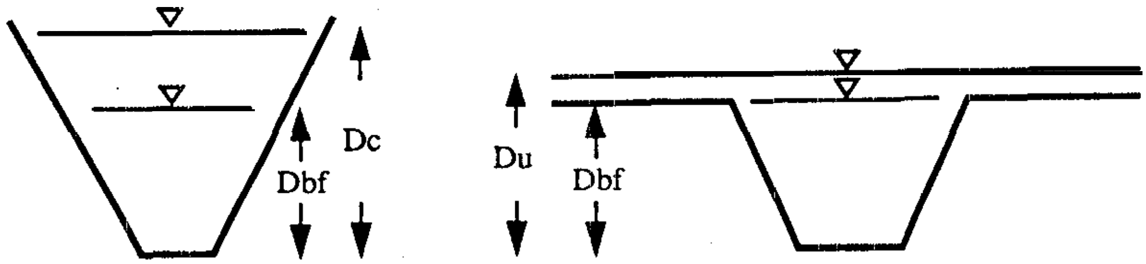

Figure 2 1 Illustration of the effect of an incremental increase in discharge above bankfull stage for confined and unconfined channels.

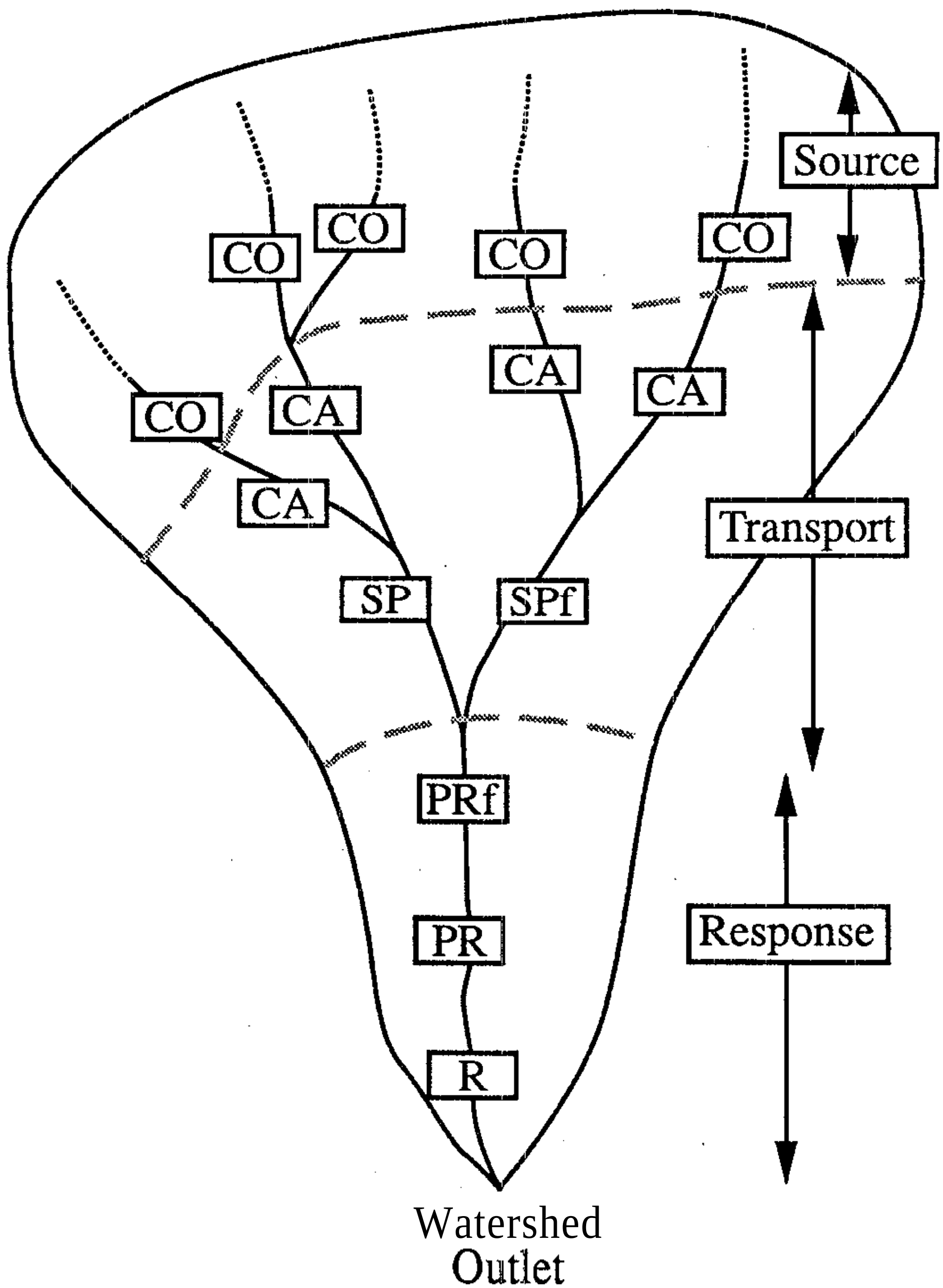

Figure 22 Watershed map illustrating application of reach-level channel classification.

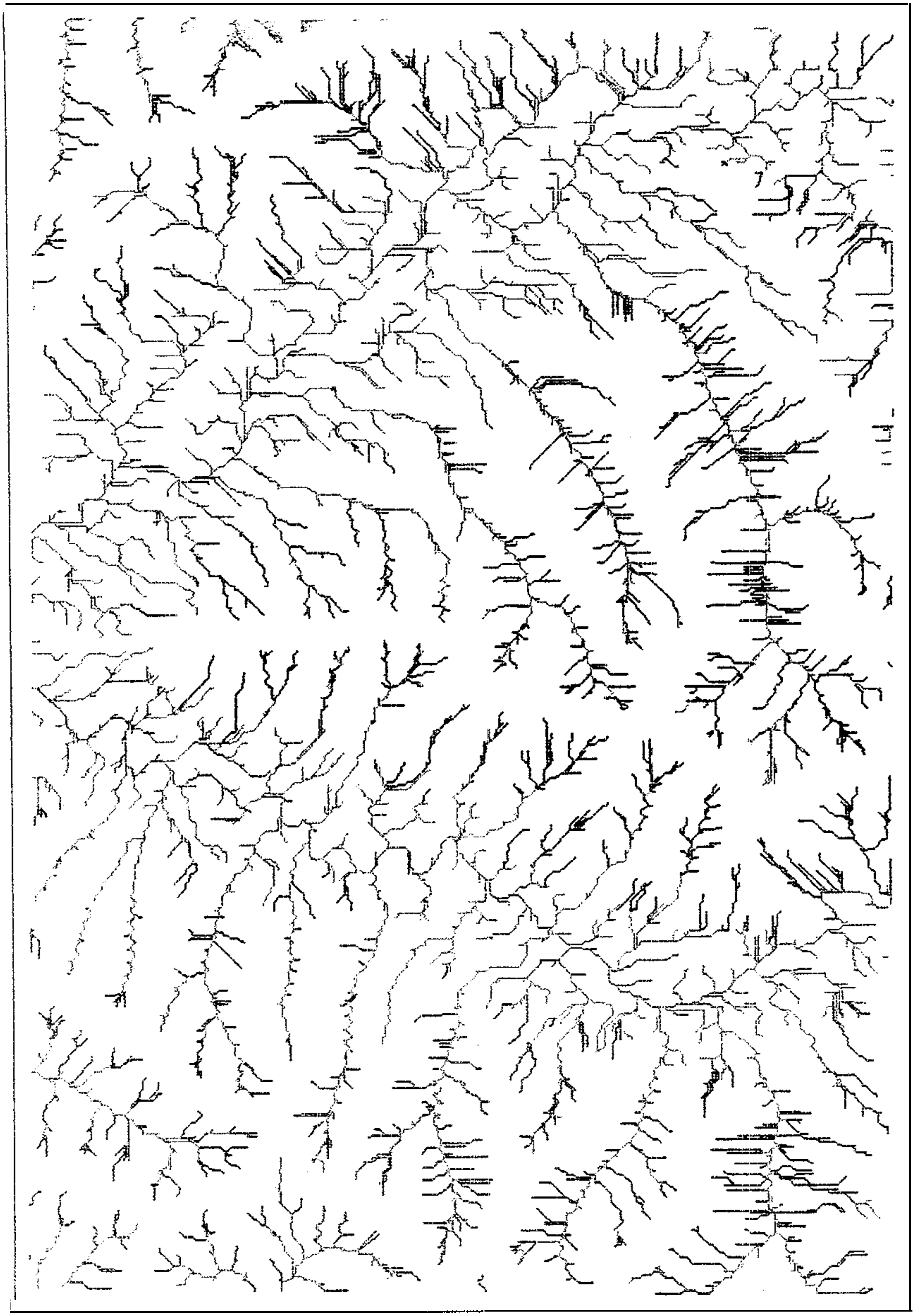

Figure 23 Map of channel types derived from U.S. Geological Survey digital elevation data for the Owl Mountain 7.5 quadrangle. Purple channels represent source reaches, green channels represent transport reaches, and red channels represent response reaches.