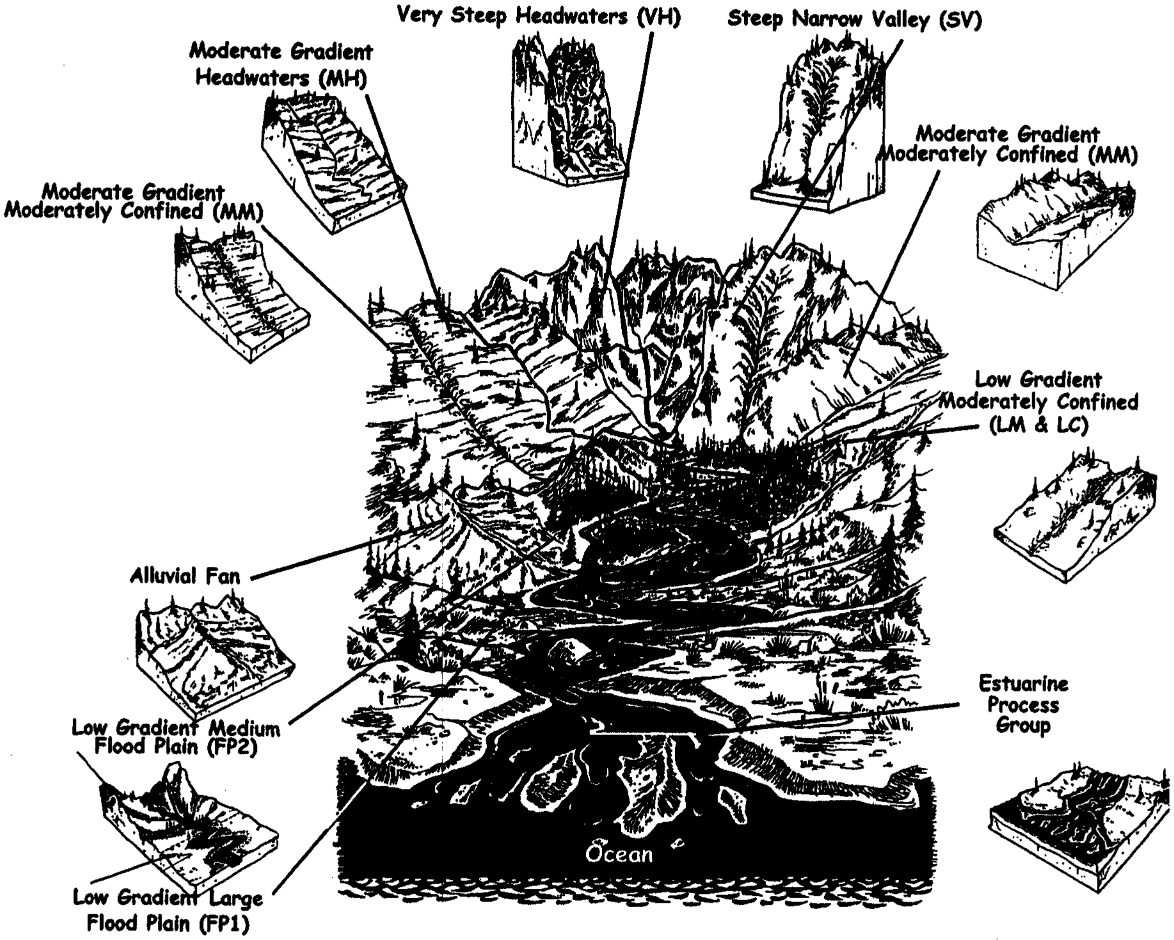

Figure 5. Typical distribution of CHTs in a mountainous watershed.

provides a framework and mechanism for evaluating (1) basin-wide stream channel conditions, (2) the influences of land management activities on specific stream reaches, and (3) potential restoration opportunities.

> **See Historical Conditions and Channel Modification components to map human impacts to the stream channels.**

The Historical Conditions Assessment component provides a methodology for obtaining and interpreting historical material to identify where potential impacts may have occurred, and to describe what the watershed may have looked like before human activities. The Channel Modification Assessment component systematically determines how the channels in the watershed have been modified by various human uses. Often, it is hard to separate impacts resulting from current, ongoing land and water management activities from historical or ***legacy activities***. For that reason, knowledge of past use can provide a context to help us understand the current condition of a watershed.

Schumm (1960) presented cogent arguments to suggest that channel shape, as defined by *W/D* ratio is determined primarily by the nature of the sediment in the channel perimeter. Where perimeters have a high percentage of silt and clay (particles <0.074 mm), channels tend to be narrow and deep. In contrast, wide, shallow channels seem to be characteristic of rivers having coarse-grained perimeters. Schumm's data, collected from semiarid and arid climate streams, show that *W/D* ratio is related to the percent silt-clay by the equation

$$F = 255M^{-1.08}$$

where F is the width-depth ratio and M is the percent silt and clay in the channel perimeter. The magnitudes of the mean annual discharge or the mean annual flood do not seem to affect this relationship (Schumm 1971).

The relationship developed by Schumm is partly related to sediment transport mechanics. Bank materials are generally a reflection of the type of sediments transported through a given stream reach. In streams that are not aggrading or degrading, coarse-grained bedload is more efficiently transported in a wide, shallow channel because higher velocities (and steeper velocity gradients) are located near the channel floor (Pickup 1976; Morisawa 1985). In contrast, suspended loads are carried best in channels having lower width-depth ratios (narrow, deep channels). This relationship between the nature of the load in transport and channel shape is so striking that it prompted Schumm (1963a) to classify rivers on the basis of load types (fig. 6.23). More important, it demonstrates another viable method by which a river can be adjusted to alterations in sediment and discharge. A channel reach suddenly burdened with a different type of load than the one it previously carried may as easily alter its channel shape to accommodate the new load as change its gradient by deposition or erosion.

It is also clear that the *W/D* ratio is influenced by the erosional resistance of the channel banks. Although bank strength cannot be defined by a single parameter, it does depend on the cohesion of the bank material, which can be expressed by the silt and clay content of the channel sediments (Knighton 1987). Recall, however, that alluvial bank deposits often are vertically stratified with fine-grained, cohesive sediments overlying a coarse-grain basal layer. In such cases, bank erosion may be related more to the ability of the stream to remove the basal, noncohesive layer than the corrasion of the fine-grained alluvium (ASCE 1998). In addition, the erosional resistance of the bank may be related to mass wasting processes and have little to do with fluvial mechanics.

Factors other than the character of sediment also can affect the cross-sectional shape of channels (Miller 1990). For example, root systems of riparian vegetation drastically decrease bank erosion and therefore control the channel width (Smith 1976). In contrast, large trees that fall into a channel may increase bank erosion by diverting flow around a tree jam into an unprotected bank (Keller and Swanson 1979; Keller and Tally 1979).

CHANNEL PATTERNS

The discussion of adjusting mechanics in rivers has thus far centered on the balancing factors that function in a channel cross-section or a series of cross-sections considered in a downstream direction. Rivers also have characteristic plan-view forms that display a distinct geometric pattern over long stretches of their total length. The plan-view pattern that a river adopts is now recognized as another manifestation of channel adjustment to the prevailing conditions of discharge and load.

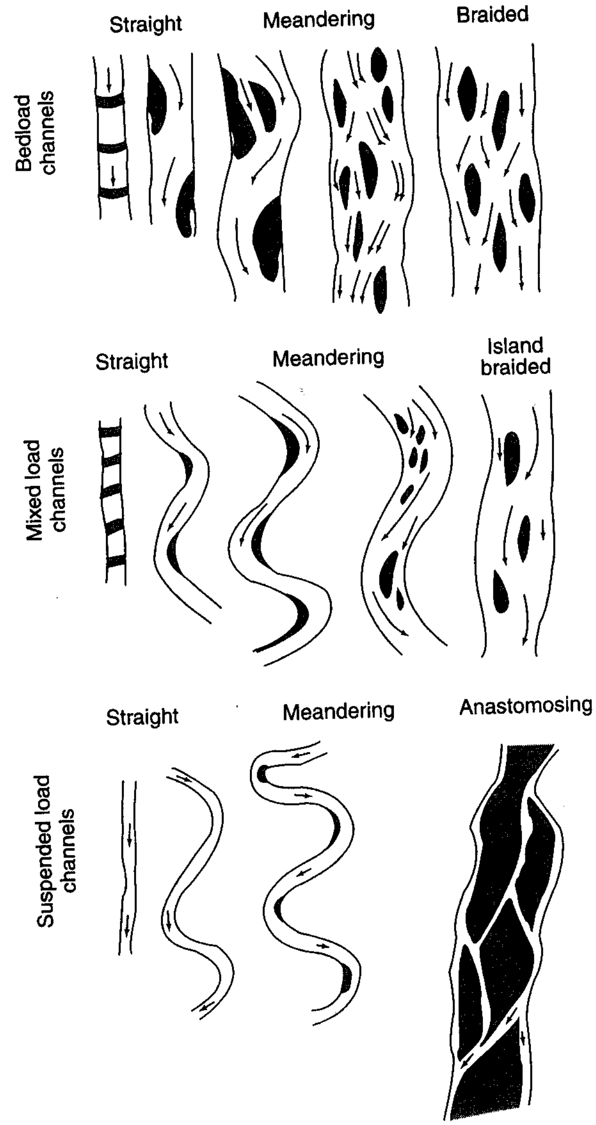

Figure 6.23
Classification of channel patterns presented by Schumm (1981, 1985).

In that sense, channel patterns are another important aspect of river mechanics.

Historically, channel patterns were typically classified as straight, meandering, or braided (Leopold and Wolman 1957). Many geomorphologists now recognize a wider range of river patterns and believe that this traditional classification scheme is insufficient in describing the types of patterns that exist in nature (Knighton 1998). Schumm (1981), for example, combined the original nomenclature (straight, meandering, and braided) with the type of load transported by the river, ultimately recognizing 14 different categories (fig. 6.23). Others have divided stream patterns into two broad groupings including single-channeled (straight and meandering) and multi-channeled (braided and anabranching) systems (Knighton 1998). This latter classification scheme emphasizes the existence of a number of patterns that are distinct from the classical braided configuration in which the flow moves through an interconnected network of channels.

Regardless of the classification system that is utilized, it is clear that the boundaries between types are arbitrary and indistinct. The distinction between straight and meandering, for example, is based on a property called **sinuosity,** which is the ratio of stream length (measured along the center of the channel) to valley length (measured along the axis of the valley) (Schumm 1963b; see Leopold and Wolman 1957 for a slightly different definition). The transition between straight and meandering streams is usually placed at a sinuosity of 1.5, but this value has no particular mechanical significance.

The braided pattern, characterized by the division of the river into more than one channel, is more easily discerned. The designation becomes vague, however, when only part of the river's total length is multi-channeled. How much of the river must consist of divided reaches to constitute a braided system is an individual decision. Another complication is that some single channeled rivers become distinctly braided in times of high flow, requiring that stage be considered when deciding on the pattern of classification. In addition, there is no reason to expect that a river will display the same pattern for its entire length. Because patterns are a function of discharge and load, minor variations in those factors downstream may easily generate different patterns, especially in very large watersheds.

Straight Channels

In spite of the fact that a broad range of river patterns is now recognized, the traditional subdivision into straight, meandering, and braided provides a useful framework for examining the processes that are responsible for creating the types of river patterns that exist. We will, therefore, utilize this traditional scheme, starting our discussion with straight channels. We will then conclude our discussion by examining the anabranching pattern, which is now considered a distinct type of plan obtained by some rivers. Most streams do not have straight banks for any significant distance, making the straight pattern a rather uncommon one. It may seem strange, then, that straight streams display many of the same features as the more common meandering rivers. As figure 6.24 illustrates, straight reaches often contain accumulations of bed material, known as *alternate bars,* that are positioned successively downriver on opposite sides of the channel. A line connecting the deepest parts of the channel, called the *thalweg,* migrates back and forth across the bottom. In rivers with a poorly sorted load, the channel floor undulates into alternating shallow zones called *riffles* and deeps called *pools* (figs. 6.22, 6.24). The pools are directly opposite the alternate bars, and the riffles are about midway between two successive pools. Clearly, a straight channel implies neither a uniform streambed nor a straight thalweg, and the spacing of bars, riffles, and pools is closely analogous to that in a meandering channel (Leopold et al. 1964).

Sequences of pools and riffles are very important manifestations of how bedforms, flow, and sediment transport are interrelated in rivers to maintain quasi-equilibrium. This is true regardless of channel pattern, indicating that the tendency to develop bars and/or pool and riffle sequences must be due to some fundamental

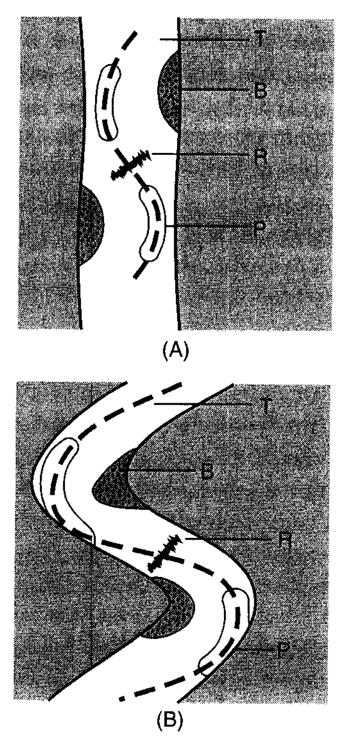

Figure 6.24
Features associated with (A) straight and (B) meandering rivers. T = thalweg, B = bar, R = riffle, P = pool.

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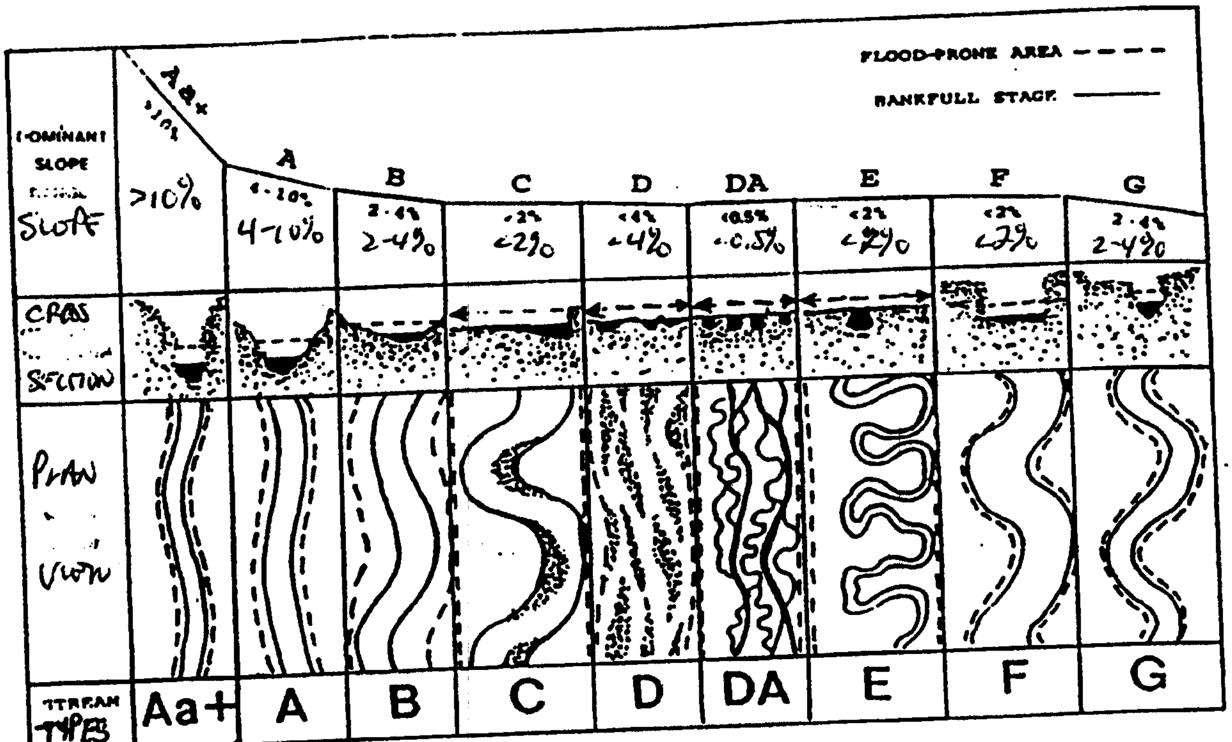

FIG. 2 ~~Fig. 1~~. Longitudinal, cross-sectional and plan views of major stream types.

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Rosgen, 1994

Table 2
Summary of delineative criteria for broad-level classification

Stream type	General description	Entrenchment ratio	W/D ratio	Sinuosity	Slope	Landform/soils/features
Aa+	Very steep, deeply entrenched, debris transport streams.	< 1.4	< 12	1.0 to 1.1	> 0.10	Very high relief. Erosional, bedrock or depositional features; debris flow potential. Deeply entrenched streams. Vertical steps with deep scour pools; waterfalls.
A	Steep, entrenched, cascading, step/pool streams. High energy/debris transport associated with depositional soils. Very stable if bedrock or boulder dominated channel.	< 1.4	< 12	1.0 to 1.2	0.04 to 0.10	High relief. Erosional or depositional and bedrock forms. Entrenched and confined streams with cascading reaches. Frequently spaced, deep pools in associated step-pool bed morphology.
B	Moderately entrenched, moderate gradient, riffle dominated channel, with infrequently spaced pools. Very stable plan and profile. Stable banks.	1.4 to 2.2	> 12	> 1.2	0.02 to 0.039	Moderate relief, colluvial deposition and/or residual soils. Moderate entrenchment and W/D ratio. Narrow, gently sloping valleys. Rapids predominate with occasional pools.
C	Low gradient, meandering, point-bar, riffle/pool, alluvial channels with broad, well defined floodplains	> 2.2	> 12	> 1.4	< 0.02	Broad valleys with terraces, in association with floodplains, alluvial soils. Slightly entrenched with well-defined meandering channel. Riffle-pool bed morphology.
D	Braided channel with longitudinal and transverse bars. Very wide channel with eroding banks.	n/a	> 40	n/a	< 0.04	Broad valleys with alluvial and colluvial fans. Glacial debris and depositional features. Active lateral adjustment, with abundance of sediment supply.
DA	Anastomosing (multiple channels) narrow and deep with expansive well vegetated floodplain and associated wetlands. Very gentle relief with highly variable sinuosities. Stable streambanks.	> 4.0	< 40	variable	< 0.005	Broad, low-gradient valleys with fine alluvium and/or lacustrine soils. Anastomosed (multiple channel) geologic control creating fine deposition with well-vegetated bars that are laterally stable with broad wetland floodplains.
E	Low gradient, meandering riffle/pool stream with low width/depth ratio and little deposition. Very efficient and stable. High meander width ratio.	> 2.2	< 12	> 1.5	< 0.02	Broad valley/meadows. Alluvial materials with floodplain. Highly sinuous with stable, well vegetated banks. Riffle-pool morphology with very low width/depth ratio.
F	Entrenched meandering riffle/pool channel on low gradients with high width/depth ratio.	< 1.4	> 12	> 1.4	< 0.02	Entrenched in highly weathered material. Gentle gradients, with a high W/D ratio. Meandering, laterally unstable with high bank-erosion rates. Riffle-pool morphology.
G	Entrenched "gulley" step/pool and low width/depth ratio on moderate gradients.	< 1.4	< 12	> 1.2	0.02 to 0.039	Gulley, step-pool morphology with moderate slopes and low W/D ratio. Narrow valleys, or deeply incised in alluvial or colluvial materials; i.e., fans or deltas. Unstable, with grade control problems and high bank erosion rates.

Montgomery & Buffington, 1997

TABLE 1. DIAGNOSTIC FEATURES OF EACH CHANNEL TYPE

	Dune ripple	Pool riffle	Plane bed	Step pool	Cascade	Bedrock	Colluvial
Typical bed material	Sand	Gravel	Gravel-cobble	Cobble-boulder	Boulder	Rock	Variable
Bedform pattern	Multilayered	Laterally oscillatory	Featureless	Vertically oscillatory	Random	Irregular	Variable
Dominant roughness elements	Sinuosity, bedforms (dunes, ripples, bars) grains, banks	Bedforms (bars, pools), grains, sinuosity, banks	Grains, banks	Bedforms (steps, pools), grains, banks	Grains, banks	Boundaries (bed and banks)	Grains
Dominant sediment sources	Fluvial, bank failure	Fluvial, bank failure	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	Overbank	Bedforms	Lee and stoss sides of flow obstructions	Pockets	Bed
Typical confinement	Unconfined	Unconfined	Variable	Confined	Confined	Confined	Confined
Typical pool spacing (channel widths)	5 to 7	5 to 7	None	1 to 4	<1	Variable	Unknown

We recognize three primary channel-reach substrates: bedrock, alluvium, and colluvium. Bedrock reaches lack a contiguous alluvial bed and reflect high transport capacities relative to sediment supply; they are typically confined by valley walls and have steep slopes. In contrast, alluvial channels exhibit a wide variety of morphologies and roughness configurations that vary with slope and position within the channel network, and may be either confined, with little to no associated flood plain, or unconfined, with a well-established flood plain. We recognize five distinct alluvial reach morphologies: cascade, step pool, plane bed, pool riffle, and dune ripple. Colluvial channels form an additional reach type that we recognize separately from alluvial channels, despite the common presence of a thin alluvial substrate. Colluvial channels typically are small headwater streams that flow over a colluvial valley fill and exhibit weak or ephemeral fluvial transport. Each of these channel types is distinguished by a distinctive channel-bed morphology, allowing rapid visual classification. Diagnostic features of each channel type are summarized in Table 1 and discussed below.

Cascade Channels

The term "cascade" connotes tumbling flow, although its specific morphologic definition varies and often is applied to both channel units and reaches (e.g., Bisson et al., 1982; Grant et al., 1990). Our delineation of cascade channels focuses on streams in which energy dissipation is dominated by continuous tumbling and jet-and-wake flow over and around individual large clasts (e.g., Peterson and Mohanty, 1960) (Fig. 1A). Cascade channels generally occur on steep slopes, are narrowly confined by valley walls, and are characterized by longitudinally and laterally disorganized bed material typically consisting of cobbles and boulders (Fig. 2A). Small, partially channel-spanning pools spaced less than a channel width apart are common in cascade channels. Tumbling flow over individual grain steps and turbulence associated with jet-and-wake flow around grains dissipates much of the mechanical energy of the flow (Fig. 3A).

Large particle size relative to flow depth makes the largest bed-forming material of cascade reaches effectively immobile during typical flows. Studies of steep-gradient channels report that large bed-forming grains typically become mobile only during infrequent (i.e., 50–100 yr) hydrologic events (Grant et al., 1990; Kondolf et al., 1991; Whittaker, 1987b). Mobilization of these larger clasts is accompanied by high sediment transport rates due to the release of finer sediment trapped under and around large grains (Sawada et al., 1983; Warburton, 1992). During lesser floods, gravel stored in low energy sites is mobilized and travels as bedload over larger bed-forming clasts (Griffiths, 1980; Schmidt and Ergenzinger, 1992). Gravel and finer material are locally stored on stoss and lee sides of flow obstructions (i.e., large grains and large woody debris) due to physical impoundment and generation of velocity shadows. One tracer study (Kondolf et al., 1991) showed that material in such depositional sites was completely mobilized during a seven-year recurrence-interval event, whereas no tracer movement was observed during flows of less than the annual recurrence interval.

These observations suggest that there are two thresholds for sediment transport in cascade channels. During moderate recurrence-interval flows, bedload material is rapidly and efficiently transported over the more stable bed-forming clasts, which have a higher mobility threshold corresponding to more infrequent events. The lack of significant in-channel storage (Kondolf et al., 1991) and the rapid scour of depositional sites during moderately frequent high flows suggest that sediment transport is effectively supply limited in cascade channels. Bedload transport studies demonstrate that steep channels in mountain drainage basins are typically supply limited, receiving seasonal or stochastic sediment inputs (Nanson, 1974; Griffiths, 1980; Ashida et al., 1981; Whittaker, 1987). Because of this high transport capacity relative to sediment supply, cascade channels function primarily as sediment transport zones that rapidly deliver sediment to lower-gradient channels.

Step-Pool Channels

Step-pool channels are characterized by longitudinal steps formed by large clasts organized into discrete channel-spanning accumulations that separate pools containing finer material (Figs. 1B and 2B) (Ashida et al., 1976, 1981; Griffiths, 1980; Whittaker and Jaeggi, 1982; Whittaker and Davies, 1982; Whittaker, 1987a, 1987b; Chin, 1989; Grant et al., 1990). Primary flow and channel bed oscillations in step-pool reaches are vertical, rather than lateral, as in pool-riffle channels (Fig. 3B). The stepped morphology of the bed results in alternating critical to supercritical flow over steps and subcritical flow in pools (Bowman, 1977; Chin, 1989). Step-pool channels exhibit a pool spacing of roughly one to four channel widths (Bowman, 1977; Whittaker, 1987b; Chin, 1989; Grant et al., 1990), significantly less than the five to seven channel widths that typify self-formed 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, 1987b; Chin, 1989). Step-pool morphology generally is associated with steep gradients, small width to depth ratios, and pronounced confinement by valley walls. Although step-forming clast sizes typically are comparable to annual high flow depths, a stepped longitudinal profile also may develop in steep sand-bedded channels (G. E. Grant, 1996, personal commun.).

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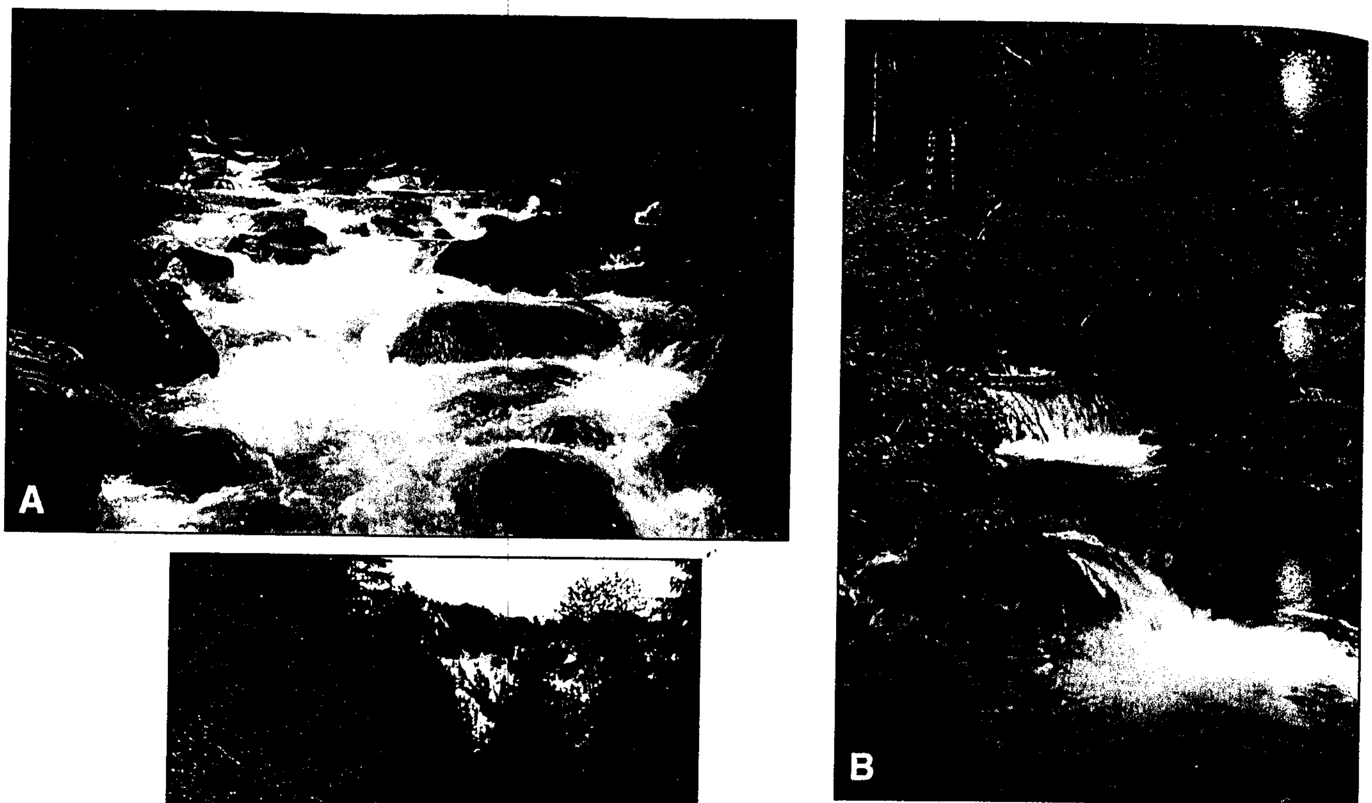

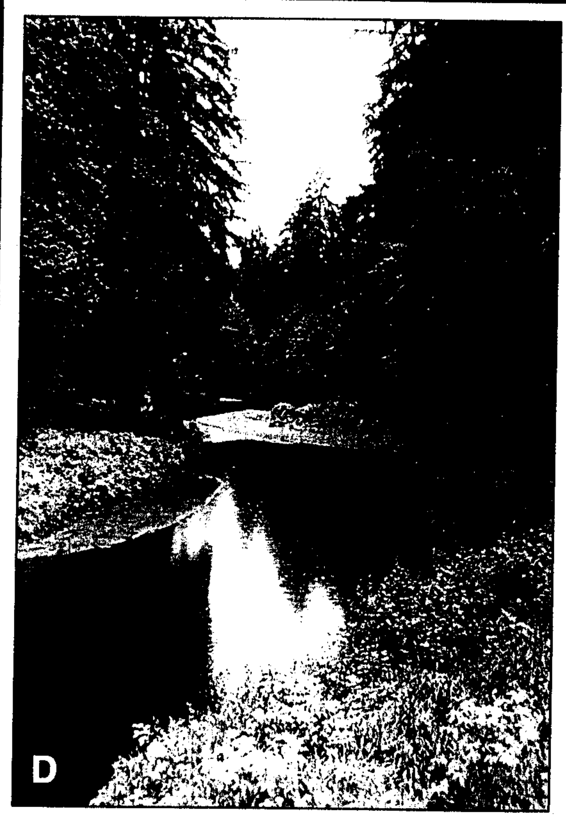

Figure 1. Alluvial channel-reach morphologies: (A) cascade; (B) step pool; (C) plane bed; (D) pool riffle; (E) dune ripple; (F) colluvial (channel in photo is 0.5 m wide); and (G) forced pool riffle.

Step-forming material may be viewed as either a kinematic wave (Langbein and Leopold, 1968), a congested zone of large grains that causes increased local flow resistance and further accumulation of large particles (Church and Jones, 1982), or as macroscale antidunes (McDonald and Banerjee, 1971; Shaw and Kellerhals, 1977; Grant and Mizuyama, 1991). Step-pool sequences form through armoring processes under high dis-

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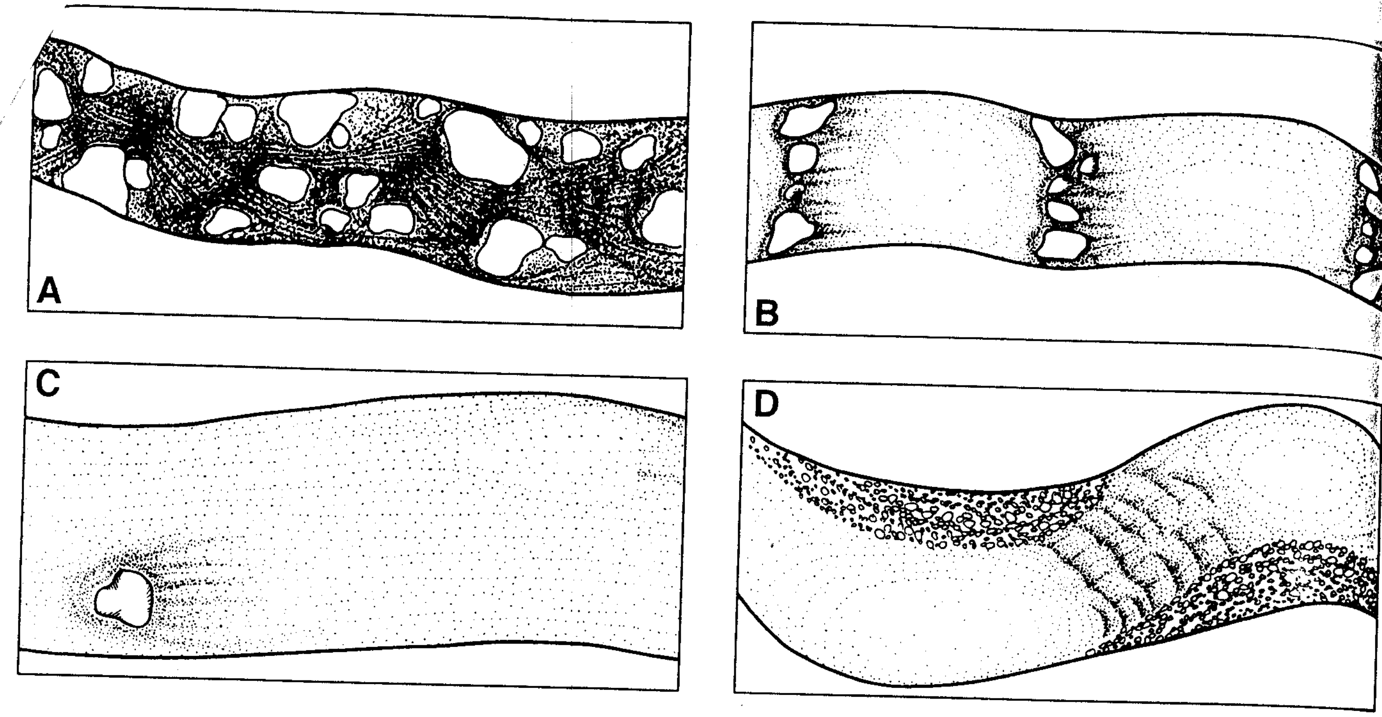

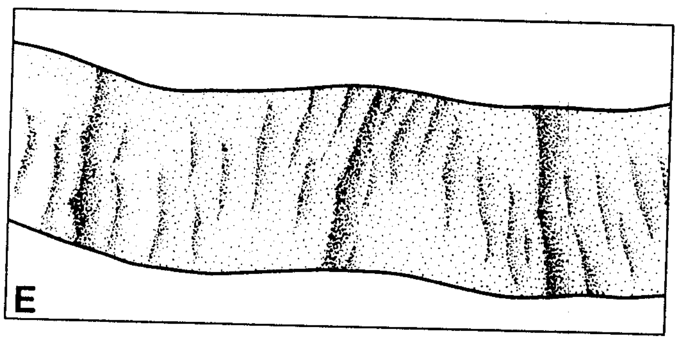

Figure 2. Schematic planform illustration of alluvial channel morphologies at low flow: (A) cascade channel showing nearly continuous, highly turbulent flow around large grains; (B) step-pool channel showing sequential highly turbulent flow over steps and more tranquil flow through intervening pools; (C) plane-bed channel showing single boulder protruding through otherwise uniform flow; (D) pool-riffle channel showing exposed bars, highly turbulent flow through riffles, and more tranquil flow through pools; and (E) dune-ripple channel showing dune and ripple forms as viewed through the flow.

density and organization of large clasts. Step-pool channels are defined by discrete channel-spanning steps less than a channel width in length that separate pools spaced every one to four channel widths. Cascade channels are defined by ubiquitous tumbling and jet-and-wake flow over a series of individual large clasts that together exceed a channel width in length, with small, irregularly placed pools spaced less than a channel width apart. The regular sequence of pools and steps in step-pool channels probably represents the emergence of a fluvially organized morphology in alluvial channels. In contrast, the disorganized large clasts of cascade channels may include lag deposits forced by nonfluvial processes (e.g., debris flows, glaciers, and rock falls).

Plane-Bed Channels

The term "plane bed" has been applied to both planar bed phases observed to form in sand-bed channels (Simons et al., 1965) and planar gravel and cobble-bed channels (Florsheim, 1985) like the coarse-grained, threshold canals described by Lane and Carlson (1953). Our use of the term refers to the latter and encompasses glide (run), riffle, and rapid morphologies described in the fisheries literature (e.g., Bisson et al., 1982). Plane-bed channels lack discrete bars, a condition that is associated with low width to depth ratios (Sukegawa, 1973; Ikeda, 1975, 1977) and large values of relative roughness (ratio of 90th percentile grain size to bankfull flow depth). Church and Jones (1982) considered bar formation unlikely at relative roughnesses of 0.3 to 1.0. Plane-bed reaches occur at moderate to high slopes in relatively straight channels that may be either unconfined or confined by valley walls. They typically are composed of sand to small boulder grain sizes, but are dominantly gravel to cobble bedded.

Plane-bed channels differ morphologically from both step-pool and pool-riffle channels in that they lack rhythmic bedforms and are characterized by long stretches of relatively featureless bed (Figs. 1C and 2C). The absence of tumbling flow and smaller relative roughness distinguish plane-bed reaches from cascade and step-pool channels (Fig. 3C). Plane-bed channels lack sufficient lateral flow convergence to develop pool-riffle morphology due to lower width to depth ratios and greater relative roughness, which may decompose lateral flow into smaller circulation cells. However, introduction of flow obstructions may force local pool and bar formation.

Plane-bed channels typically exhibit armored bed surfaces calculated to have a near-bankfull threshold for mobility, although elevated sediment loading can cause textural fining and a lower calculated mobility threshold (Buffington, 1995). Plane-bed channels with armored bed surfaces indicate a transport capacity greater than sediment supply (i.e., supply-limited conditions), whereas unarmored surfaces indicate a balance between transport capacity and sediment supply (Dietrich et al., 1989). Nevertheless, beyond

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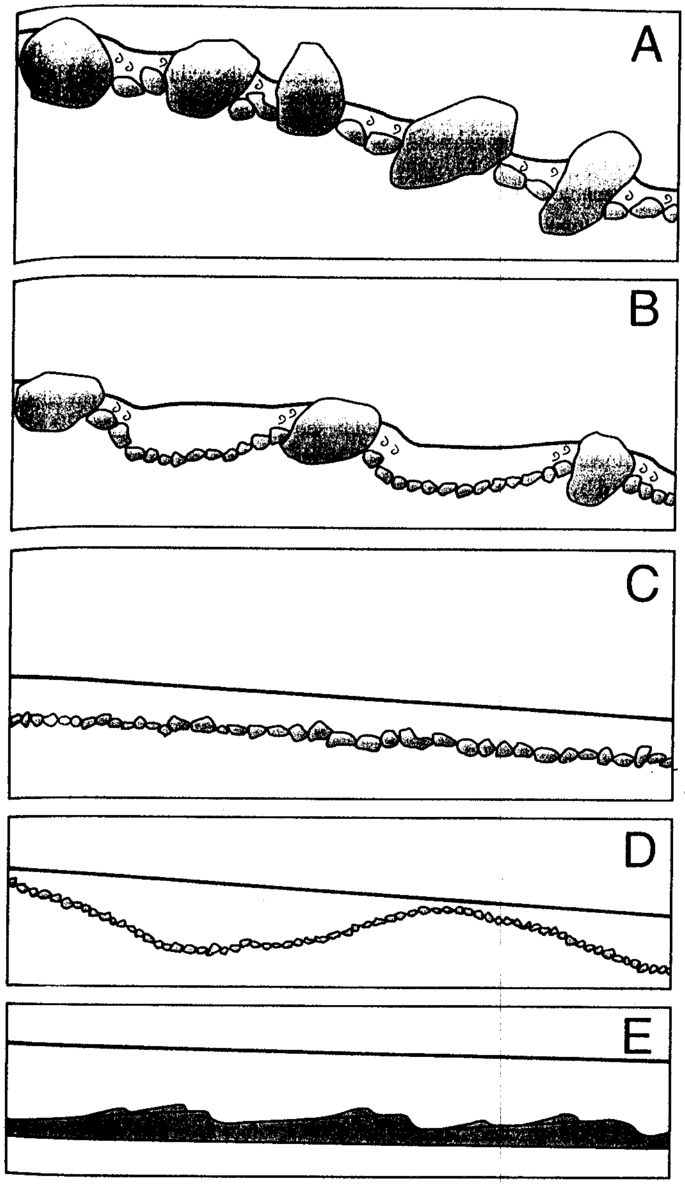

Figure 3. Schematic longitudinal profiles of alluvial channel morphologies at low flow: (A) cascade; (B) step pool; (C) plane bed; (D) pool riffle; and (E) dune ripple.

the threshold for significant bed-surface mobility, many armored gravel-bedded channels exhibit a general correspondence between bedload transport rate and discharge (e.g., Milhous, 1973; Jackson and Beschta, 1982; Sidle, 1988), implying transport-limited conditions. The above observations suggest that plane-bed channels are transitional between supply- and transport-limited morphologies.

Pool-Riffle Channels

Pool-riffle channels have an undulating bed that defines a sequence of bars, pools, and riffles (Leopold et al., 1964) (Fig. 1D). This lateral bedform oscillation distinguishes pool-riffle channels from the other channel types discussed above (Fig. 2D). Pools are topographic depressions within the channel and bars are corresponding high points (Fig. 3D); these bedforms are thus defined relative to each other (O'Neill and Abrahams, 1984). Pools are rhythmically spaced about every five to seven channel widths in self-formed, pool-riffle channels (Leopold et al., 1964; Keller and Mellhorn, 1978), but channels with a high loading of large woody debris exhibit smaller pool spacing (Montgomery et al., 1995). Pool-riffle channels occur at moderate to low gradients and are generally unconfined, and have well-established flood plains. Substrate size in pool-riffle streams varies from sand to cobble, but typically is gravel sized.

Bar and pool topography generated by local flow convergence and divergence may be either freely formed by cross-stream flow and sediment transport, or forced by channel bends and obstructions (e.g., Lisle, 1986). Free-formed pool-riffle sequences initially result from internal flow perturbation that causes flow convergence and scour on alternating banks of the channel; concordant downstream flow divergence results in local sediment accumulation in discrete bars. Topographically driven convective accelerations reinforce convergent and divergent flow patterns, and thus pool-riffle morphogenesis (Dietrich and Smith, 1983; Dietrich and Whiting, 1989; Nelson and Smith, 1989). Alluvial bar development requires a sufficiently large width to depth ratio and small grain sizes that are easily mobilized and stacked by the flow (Church and Jones, 1982). Bar formation in natural channels appears to be limited to gradients ≤0.02 (Ikeda, 1977; Florsheim, 1985), although flume studies indicate that alternate bars may form at steeper gradients (Bathurst et al., 1983; Lisle et al., 1991). Bedform and grain roughness provide the primary flow resistance in free-formed pool-riffle channels.

Pool-riffle channels 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). Armored gravel-bed channels typically exhibit a near-bankfull threshold for general and significant bed-surface mobility (e.g., Parker et al., 1982; Jackson and Beschta, 1982; Andrews, 1984; Carling, 1988; Buffington, 1995). Movement of surface grains releases fine sediment trapped by larger grains and exposes 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). Bed movement is sporadic and discontinuous, depending on grain protrusion (Fenton and Abbott, 1977; Kirchner et al., 1990), friction angle (Kirchner et al., 1990; Buffington et al., 1992), imbrication (Komar and Li, 1986), degree of burial (Hammond et al., 1984; Buffington et al., 1992), and turbulent 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.

Pool-riffle channels, like plane-bed channels, exhibit a mixture of supply- and transport-limited characteristics depending on the degree of bed-surface armoring and consequent mobility thresholds. Unarmored pool-riffle channels indicate a balance between transport capacity and sediment supply, while armored surfaces represent supply-limited conditions (e.g., Dietrich et al., 1989). Nevertheless, during armor-breaching events, bedload transport rates are generally correlated with discharge, demonstrating that sediment transport is not limited by supply once the bed is mobilized. Considerable fluctuations in observed transport rates, however, reflect a stochastic component of grain mobility caused by grain interactions, turbulent sweeps, and transient grain entrapment by bedforms (Jackson and Beschta, 1982; Sidle, 1988). Magnitudes of bedload transport also may vary for similar discharge events, depending on the chronology of antecedent transport events (Milhous, 1973; Reid et al., 1985; Sidle, 1988). Although both pool-riffle and plane-bed channels display a mix of supply- and transport-limited characteristics, the presence of depositional barforms in pool-riffle channels suggests that they are generally more transport limited than plane-bed channels. The transport-limited character of both of these morphologies, however, contrasts with the more supply-limited character of step-pool and cascade channels.

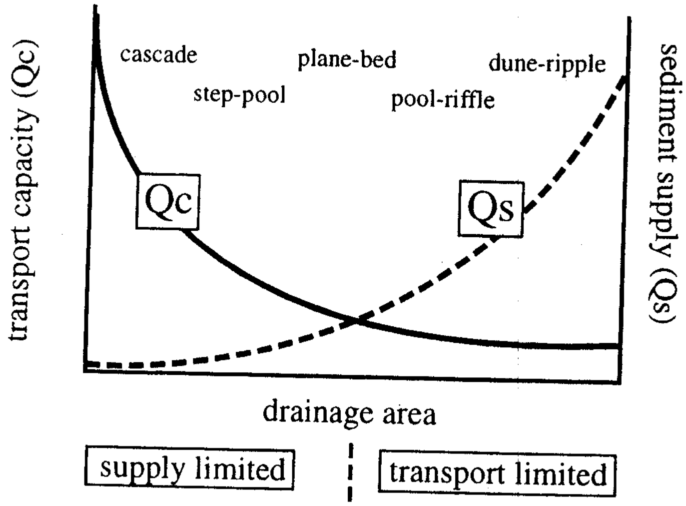

Figure 10. Schematic illustration of generalized relative trends in sediment supply (Q_s) and transport capacity (Q_c) in mountain drainage basins.

dicated by the accumulation of colluvium within valley bottoms. In contrast, the lack of an alluvial bed indicates that bedrock channels are supply limited ($q_r >> 1$). For a given drainage area (and thus Q_s), bedrock reaches have greater slopes and shear stresses (Figs. 5 and 9), implying that they have higher transport capacities and thus greater q_r values than other channel types. Alluvial channels, however, probably represent a broad range of q_r: steep alluvial channels (cascade and step-pool) have higher shear stresses (Fig. 9) and thus higher Q_c and q_r values for a given drainage area and sediment supply; the lower-gradient plane-bed and pool-riffle channels are transitional between $q_r > 1$ and $q_r \approx 1$, depending on the degree of armoring (e.g., Dietrich et al., 1989) and the frequency of bed-surface mobility; and the live-bed mobility of dune-ripple channels indicates that $q_r \leq 1$. The variety of alluvial channel morphologies probably reflects a broad spectrum of q_r expressed through fining and organization of the bedload (Fig. 11), which leads to formation of distinct alluvial bed morphologies that represent the stable bed form for the imposed q_r. This hypothesized relation between q_r and stable channel morphologies in mountain drainage basins provides a genetic framework for explaining reach-level morphologies that elaborates on Lindley's (1919) regime concept. An alluvial channel with $q_r > 1$ will become stable when the bed morphology and consequent hydraulic roughness produce an effective transport capacity that matches the sediment sup ($Q_c' \approx Q_s$).

Different channel types are stabilized by different roughness configu tions that provide resistance to flow. In steep channels energy is dissipa primarily by hydraulic jumps and jet-and-wake turbulence. This style of ergy dissipation is pervasive in cascade channels and periodic in step-p channels. Skin friction and local turbulence associated with moderate pa cle sizes are sufficient to stabilize the bed for lower shear stresses char teristic of plane-bed channels. In pool-riffle channels, skin friction and b form drag dominate energy dissipation. Particle roughness in dune-rip channels is small due to the low relative roughness, and bedforms gov hydraulic resistance. The importance of bank roughness varies with cha nel type, depending on the width to depth ratio and vegetative influenc but in steep channels bank resistance is less important compared to ener dissipation caused by tumbling flow. These different roughness configu tions represent a range in q_r values that varies from high in cascade reach to low in dune-ripple channels.

Our hypothesis that different channel types represent stable roughne configurations for different q_r values implies that there should be an asso ation of channel type and roughness. Even though the general correlation morphology and slope (Fig. 6) implies discrete roughness characteristi among channel types, different channel morphologies occurring on th same slope should exhibit distinct roughness. Photographs and descriptio of channel morphology from previous studies in which roughness was d termined from measured velocities (Barnes, 1967; Marcus et al., 1992) a low direct assessment of the roughness associated with different chann types. For similar slopes, plane-bed channels exhibit greater roughness th pool-riffle channels, and step-pool channels, in turn, appear to have great roughness than plane-bed channels with comparable gradients (Fig. 12 Moreover, intermediate morphology reaches plot between their defini channel types. These systematic trends in roughness for a given slop strongly support the hypothesis that reach-level channel morphology r flects a dynamic adjustment of the bed surface to the imposed shear stre and sediment supply (i.e., the specific q_r value).

CHANNEL DISTURBANCE AND RESPONSE POTENTIAL

Natural and anthropogenic disturbances that change hydrology, sedime supply, riparian vegetation, or large woody debris loading can alter channe processes and morphology. The effect that watershed disturbance has on a particular channel reach depends on hillslope and channel coupling, the se quence of upstream channel types, and site-specific channel morphology. In particular, the variety and magnitude of possible morphologic responses to

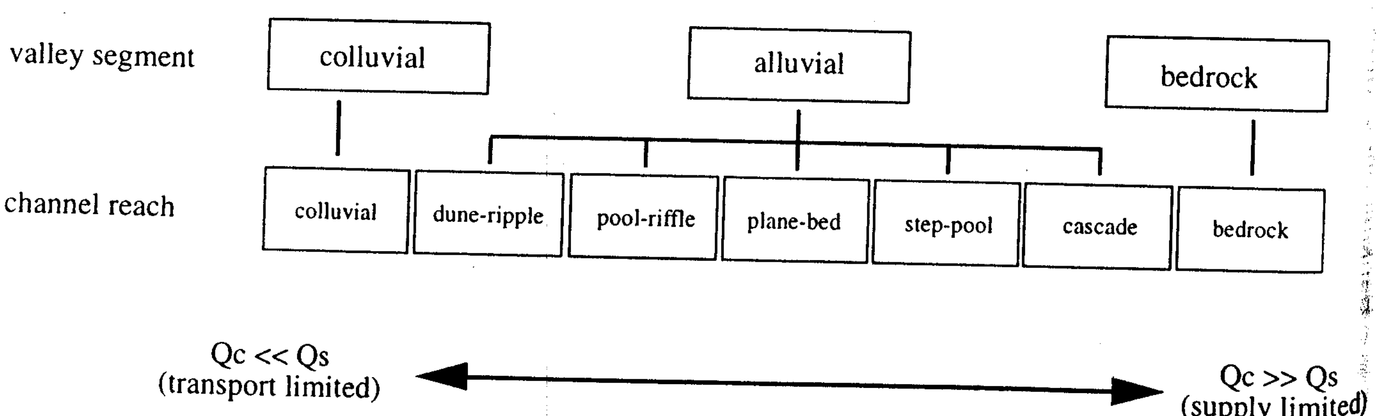

Figure 11. Schematic illustration of the transport capacities relative to sediment supply for reach-level channel types.

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FLUVIAL LANDFORMS

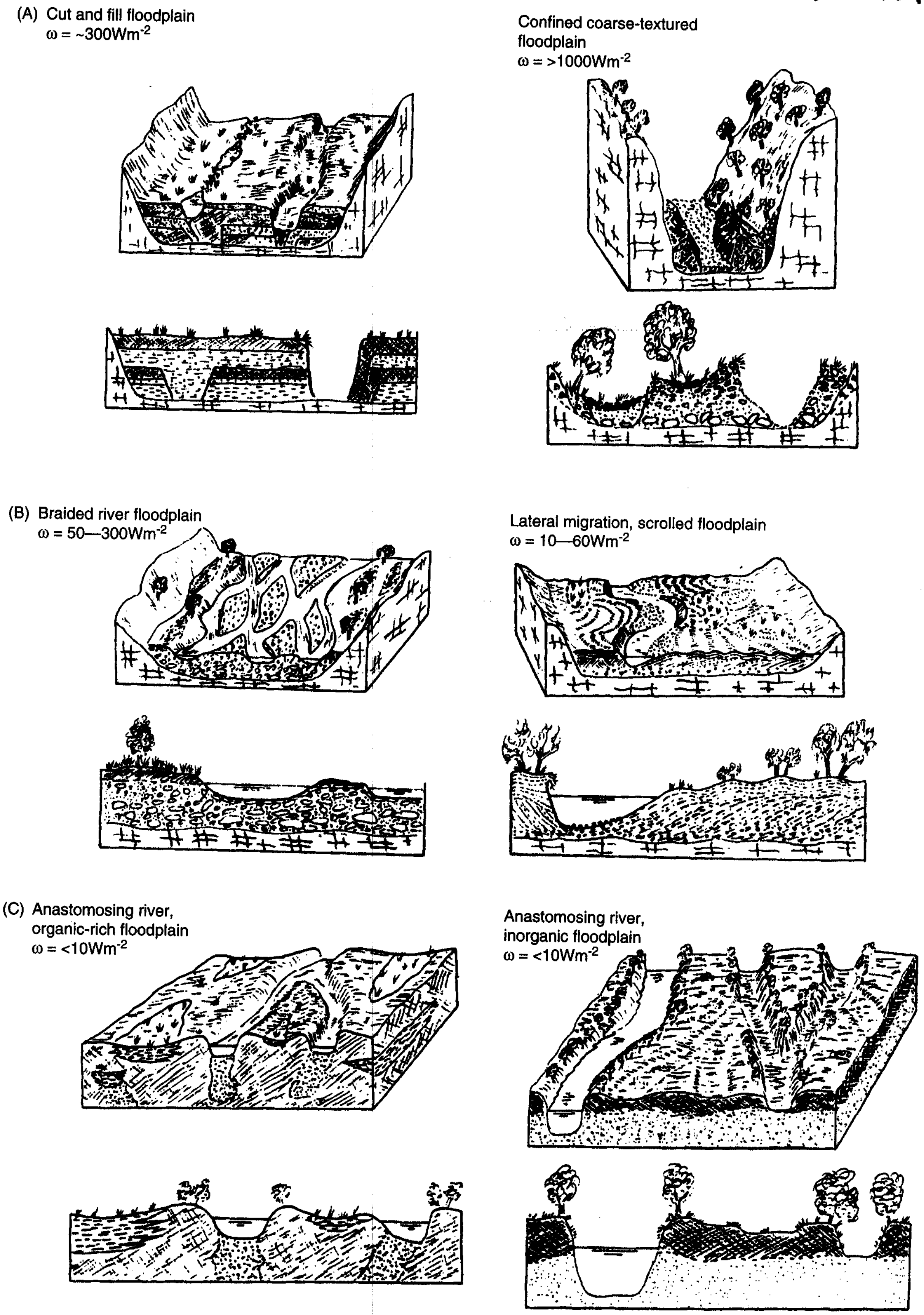

Figure 7.1
Examples of floodplain types in the classification of genetic floodplains. (A) High energy, noncohesive; (B) Medium energy, noncohesive; (C) High energy, noncohesive. (After Nanson and Croke 1992.)

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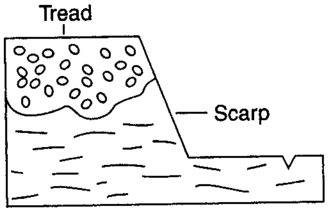
by simultaneous processes of lateral migration and overbank flooding. The deposits of lateral accretion are spread in a rather even sheet across the valley bottom, whereas overbank deposits accumulate over the entire floodplain surface away from the channel. The floodplain acts as a storage area for sediment that cannot be transported directly from the basin when it is eroded. In arid climates, floodplain development is more complex and may not manifest a process-form equilibrium between river flow and the feature. In many cases overbank flows are rare or absent. Braided rivers tend to shift laterally without the undercut banks and point bar phenomena so common in humid-climate meandering patterns. Abandoned channels and islands occupy much of the valley bottom, and these are also subject to periodic shifts of position.

Floodplains are usually considered to be associated with stable rivers, but there is no overriding reason why they cannot be present when a channel is undergoing long-term aggradation or degradation. In fact, the observed frequency of overbank flooding can continue during valley filling if the channel floor and the floodplain surface are raised at the same rate. Once the thickness of the valley deposits exceeds the limits of a reasonable scouring depth, however, the sediment below that depth can no longer be considered part of the active floodplain. In a degrading channel, the floodplain becomes a terrace when channel incision prevents the river from inundating the surface on a regular and frequent basis.

Fluvial Terraces

Terraces are abandoned floodplains that were formed when the river flowed at a higher level than at present. Topographically, a terrace consists of two parts: a **tread,** which is the flat surface representing the level of the former floodplain, and the **scarp,** which is the steep slope connecting the tread to any surface standing lower in the valley (fig. 7.10). The surface of a terrace is no longer inundated as frequently as a normal, active floodplain. It is important to remember, however, that floodplain surfaces elevated by vertical accretion also have a low frequency of inundation. Thus, it may be problematic to define a floodplain (and by contrast, a terrace) on the basis of the recurrence interval of overbank flow. Most geomorphologists would recognize a surface constructed by continuous vertical accretion as a floodplain; that is, the surface and the river are still hydrologically related even though overbank flow rarely occurs. For our purposes then, the distinguishing factor in terrace formation is that the river channel floor associated with the abandoned floodplain surface (tread) must have previously been at a higher level. In fact, the presence of a terrace demands an episode of downcutting (channel entrenchment), and indicates that some significant change must have occurred between the conditions that prevailed during development of the tread and those that produced the scarp. Usually the downcutting phase begins as a response to climatic or tectonic changes, but these are not always necessary. The tread surface normally is under-



(A)

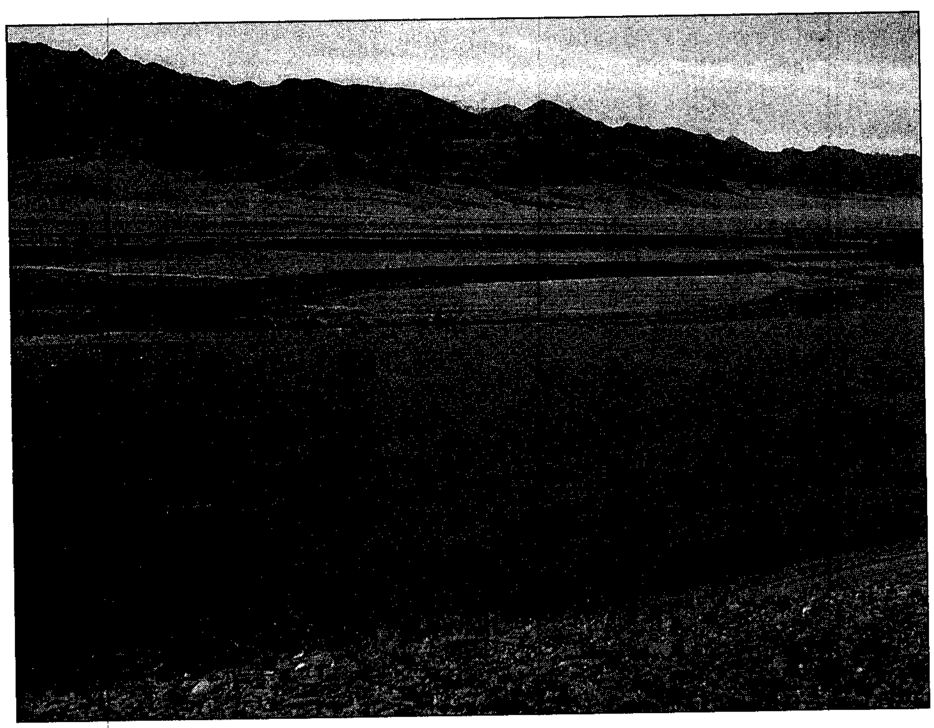


Figure 7.10
(A) Parts of a fluvial terrace.
(B) Terraces along the Madison River upstream from Ennis, Mont.

(B)

lain by alluvium of variable thickness, but these deposits are not a true part of the terrace. To avoid confusion, it is better to limit the term to the topographic form and refer to the deposits as fill, alluvium, gravel, and so on.

Types and Classification

In general, terraces are broadly categorized as erosional or depositional. **Erosional terraces** are those in which the tread has been formed primarily by lateral erosion. If the lateral erosion truncates bedrock, the terms *bench, strath,* or *rock-cut terrace* are commonly used. If the erosion crosses unconsolidated debris, the terms *fill-cut* or *fillstrath* may be used. **Depositional terraces,** the second major grouping, are terraces where the tread represents the surface of a valley fill. Figure 7.11 illustrates both types.

Erosional terraces, especially rock-cut types, are identifiable by the following, rather distinct, properties (Mackin 1937): (1) they are capped by a uniformly thin layer of alluvium in which the total thickness is controlled by the scouring depth of the river involved, and (2) the surface cut on the bedrock or older alluvium is a flat mirror image of the surface on top of the capping alluvium. In contrast, the alluvium beneath the tread of depositional terraces varies in thickness and commonly exceeds any reasonable scouring depth of the associated river. Although the tread surface may be flat, the surface beneath the fill can be very irregular.

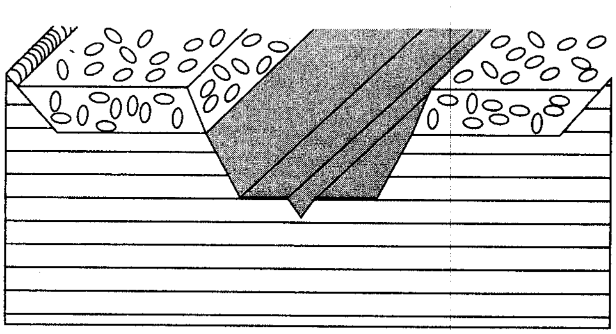
The classification of terraces as erosional or depositional is clearly a genetic distinction, and must be supported by evidence of the formative processes. Bull (1990) refines this approach by suggesting that major erosional terraces (straths) are the fundamental tectonic stream-terrace landform; that is, major straths are tectonically controlled. In contrast, terrace treads resulting from significant depositional events are the fundamental climatic stream-terrace landform; they are related to climatically controlled aggradation.

Another classification scheme is based on the topographic relationship between terrace levels within a given valley, as illustrated in figure 7.12. In this method, terrace treads that stand at the same elevation on both sides of the valley are called **paired** (matched) **terraces** and presumably are the same age. If the levels are staggered across the valley, they are said to be **unpaired** (unmatched) **terraces.** Most investigators interpret unpaired terraces as erosional types formed by a stream simultaneously cutting laterally and downcutting very slowly. Levels across the valley, therefore, are not exactly equivalent in age but differ by the amount of time needed for the river to traverse the valley bottom.

The Origin of Terraces

Depositional Terraces The development of a depositional terrace always requires a period of valley filling and subsequent entrenchment into or adjacent to the fill. This cyclic pattern is necessary because the alluvium at the tread surface takes its form from purely depositional processes. The tread, in fact, represents the highest level attained by the valley floor as it rose during aggradation. The initial entrenchment that forms the terrace scarp is primarily vertical, and so the tread surface is virtually unaffected by subsequent lateral erosion at a lower level (see fig. 7.11).

Valley filling occurs when, over an extended period, the amount of sediment produced in a basin exceeds the amount that the river system can carry away. Prolonged aggradation is usually triggered by (1) glacial outwash, (2) climate change, or (3) changes in base level, slope, or load caused by rising sea level, rising local or regional base level, or an influx of coarse load because of uplift in source areas. Where tectonics are ruled out, the balance between load and discharge is determined primarily by climatic processes, although it may be driven by glaciation and may be complexly interrelated with sea level changes. Although entrenchment has been considered


Erosional (strath) terrace
(A)

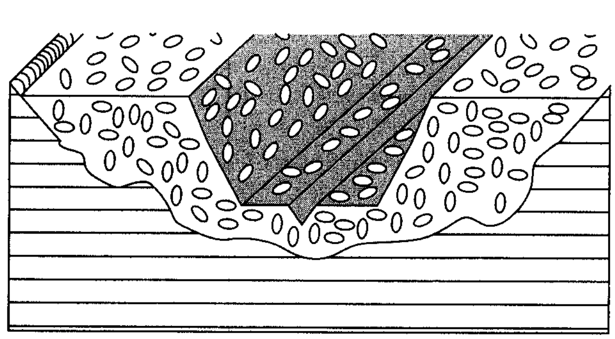

Depositional (fill) terrace
(B)

Figure 7.11
(A) Erosional (strath) terrace. Thin alluvial cover with truncation of underlying bedrock along smooth, even surface.
(B) Depositional (fill) terrace. Terrace scarp underlain by alluvium that is highest level of fill deposited in valley. Note thickness of alluvium and irregular bedrock surface beneath the fill.

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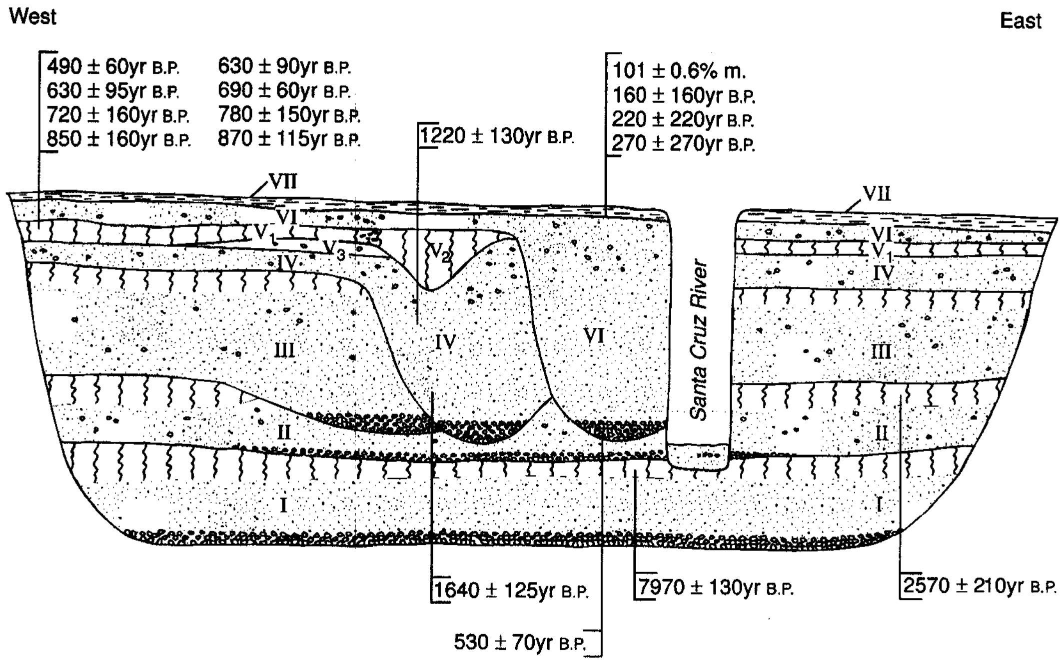


Figure 7.16
Generalized composite geologic cross-section of the northern portion of the San Xavier reach of the Santa Cruz River. Not to scale. Note that the level of entrenchment prior to fill IV and VI is the same as previous channel bottoms.
(Waters 1988)

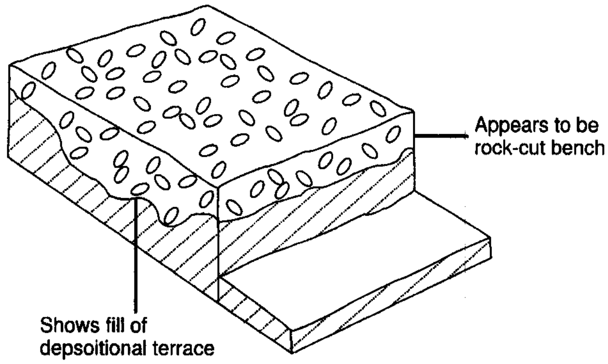


Figure 7.17
Difficulty in interpreting terrace origin from field data. If downcutting exposes only part of fill, terrace may appear to be rock-cut. Data across the terrace are needed to determine true thickness of the fill.

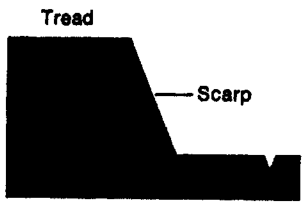
Piedmont Environment: Fans and Pediments

The topography of almost every region reflects an adjustment between dominant surficial processes and lithology. When the rocks have diverse resistances, geomorphic processes tend to maximize the relief between regions of greatest and least resistance. Nowhere is this more apparent than in areas where mountains and plains adjoin, especially where the climate is arid or the region has undergone recent tectonism. Aridity buffers the smoothing effects of vegetation; vertical tectonic activity accentuates relief by bringing more resistant basement rocks toward the surface, where they become the cores of topographic mountains.

The sloping surface that connects the mountain to the level of adjacent plains is the **piedmont.** It extends from the mountain front to a floodplain or playa, either of which can mark the base level for geomorphic processes that function on the piedmont surface (fig. 7.18 A & B). Piedmonts consist of a number of geomorphic landforms, but most commonly they are composed of eroded bedrock plains called **pediments** and depositional features called **alluvial fans.** The relative percentage of the total piedmont area occupied by either of these features probably depends on a unique combination of local geomorphic variables.

Alluvial Fans

Alluvial fans have been investigated most extensively and in great detail in regions of arid or semiarid climate. This does not mean that fans are absent in other climatic zones. On the contrary, humid-climate fans and deposits have been examined in such diverse settings as humid-glacial (Boothroyd and Ashley 1975), humid-periglacial (Ryder 1971a, 1971b; Wasson 1977), humid-tropical (Mukerji 1976; Wescott and Ethridge 1980), and humid-temperate



(A)

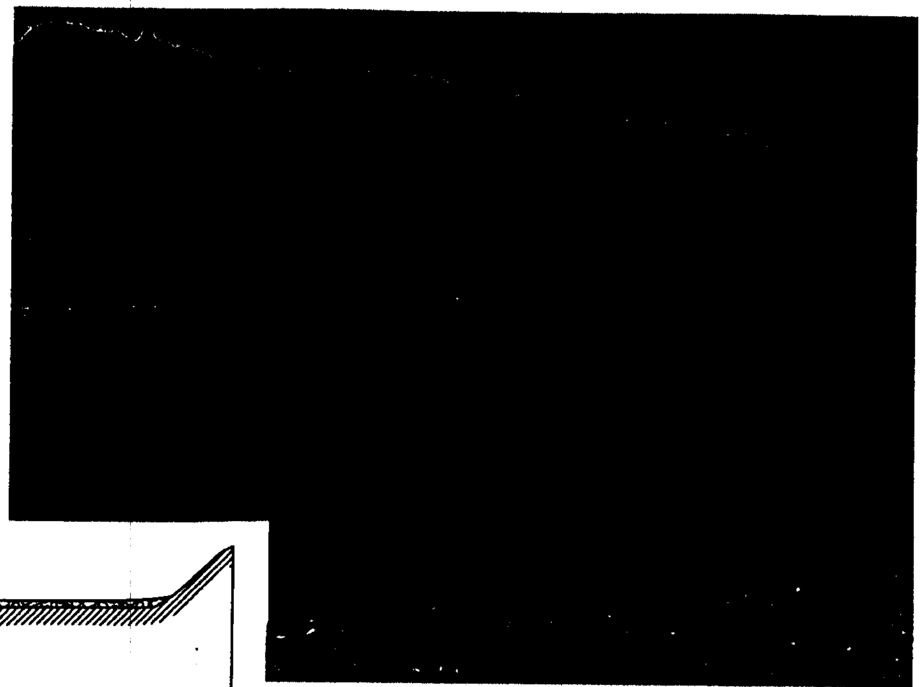


FIGURE 7.10
(A) Parts of a fluvial terrace.
(B) Terraces along the Madison River upstream from Ennis, Mont.

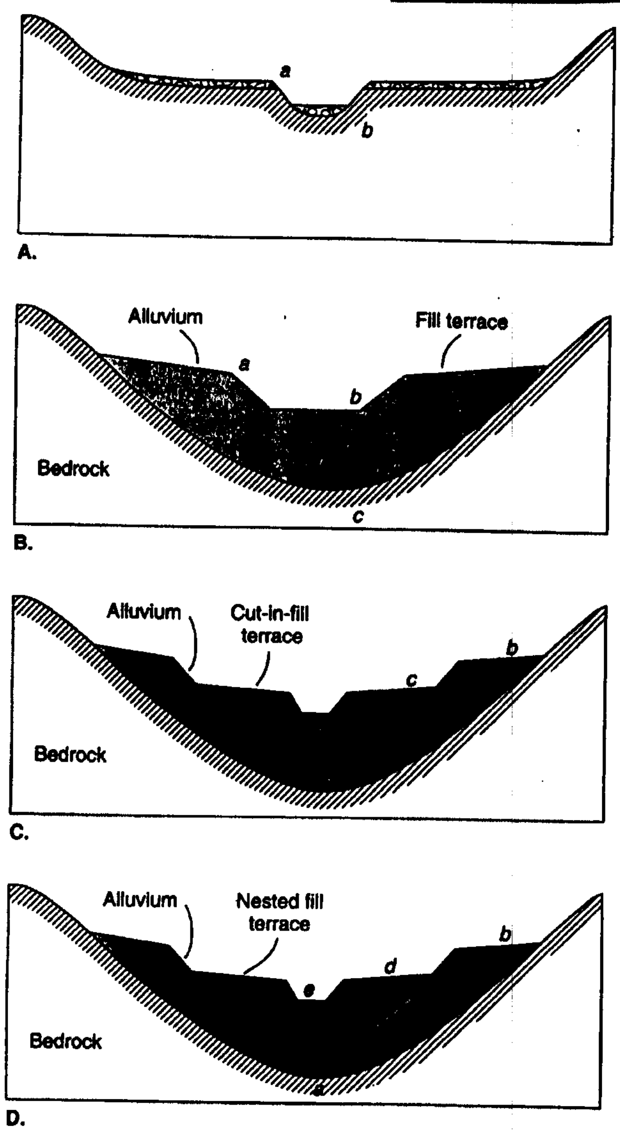


FIGURE 6–47
Types of stream terraces: (A) bedrock (cut) terrace, (B) fill terrace, (C) cut-in-fill terrace, (D) nested fill terrace.

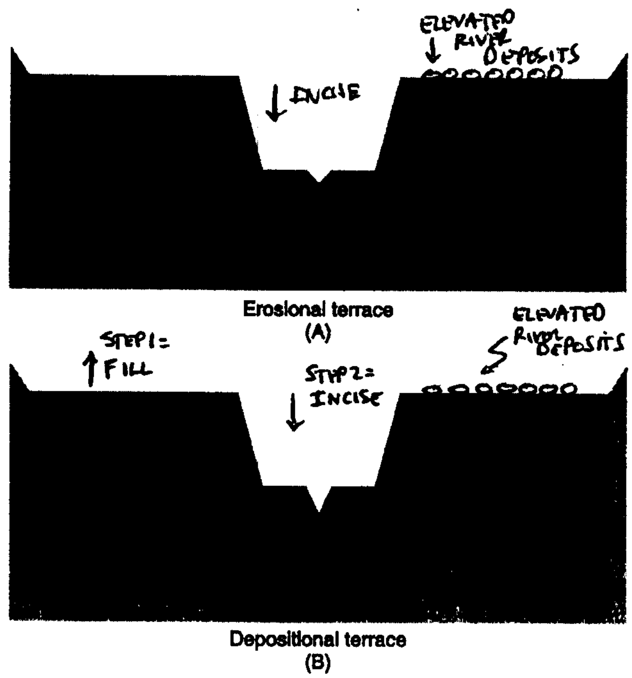


FIGURE 7.11
(A) Erosional terrace. Thin alluvial cover with truncation of underlying bedrock along smooth, even surface.
(B) Depositional terrace. Terrace scarp underlain by alluvium that is highest level of fill deposited in valley. Note thickness of alluvium and irregular bedrock surface beneath the fill.

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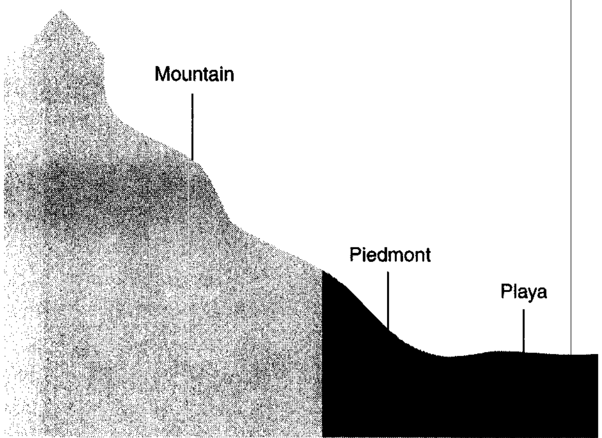

Figure 7.18A
Physiographic components of a mountain-basin geomorphic system showing the position of the piedmont zone, which contains pediments and alluvial fans.

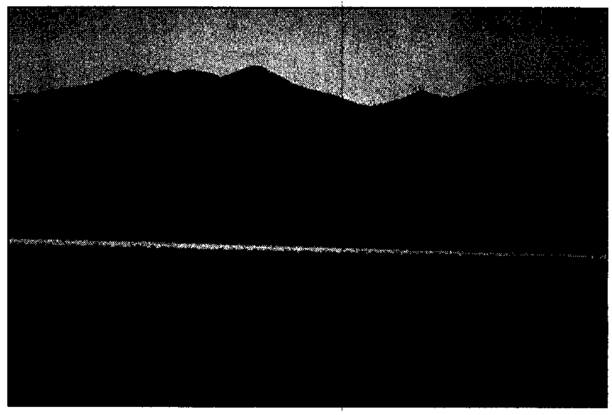

Figure 7.18B
Photo of mountain front environment in an arid basin of central Nevada: Mountain is in rear and playa is white area on basin floor. Piedmont zone in center.
(Photo by D. Germanoski)

TABLE 7.3 Generalized Characteristics of Alluvial Fans Formed in Different Environments

Parameter	Arid fans	Humid-glacial Fans	Humid-tropical Fans	Virginia Humid-temperate Fans
Fan morphology				
Plan view	Broad, fanlike, symmetrical	Broad, fanlike, symmetrical	Broad, fanlike, symmetrical	Broad, fanlike to elongated
Axial profile	Segmented (20–100m/km)	Smooth (1–20m/km)	Smooth	Segmented (40–100m/km)
Thickness	Up to 100s m	Up to 100s m	Up to 100s m	5 to 20 m
Area	Small	Very large	Large	Small
Depositional processes				
Major processes	Debris flow Braided stream Sheet flood Sieve flood	Braided stream	Braided stream Debris flow	Debris flow (avalanche)
Return interval	1–50 yrs, discrete events	0–few days, seasonally constant	Seasonally constant to discrete	3000–6000 yrs, discrete events
Fan area activated	10–50%	80–100%	30–70%	10–70%
Triggering processes	Heavy rain Snow melt	Meltwater Outwash	Heavy rain Monsoon	Heavy rain Hurricane
Discharge	Flashy	Seasonal	Seasonal	Flashy

After Kochel and Johnson (1984). Used with permission of the Canadian Society of Petroleum Geologists.

(Hack and Goodlett 1960; Williams and Guy 1973; Kochel and Johnson 1984; Wells and Harvey 1987). Fans developed in every climatic setting are linked together by a similar plan-view geometry, but other aspects of morphology and depositional processes may vary considerably (table 7.3). More recent and excellent reviews of fan processes and deposits can be found in Blair and McPherson (1994a, b) and Harvey (1997).

Alluvial fans are one end of an erosional-depositional system, linked by a river, in which rock debris is transferred from one portion of a watershed to another. Fans are largest and most well developed where erosion takes place in a mountain and the river builds the fan into an adjacent basin. Deposits tend to be fan-shaped in plan view and are best described morphologically as a segment of a cone radiating away from a single point source (fig. 7.19). Adjacent fans often merge at their lateral extremities; the individual cone shape is lost, and a rather nondescript deposit is formed covering the entire piedmont. The coalesced

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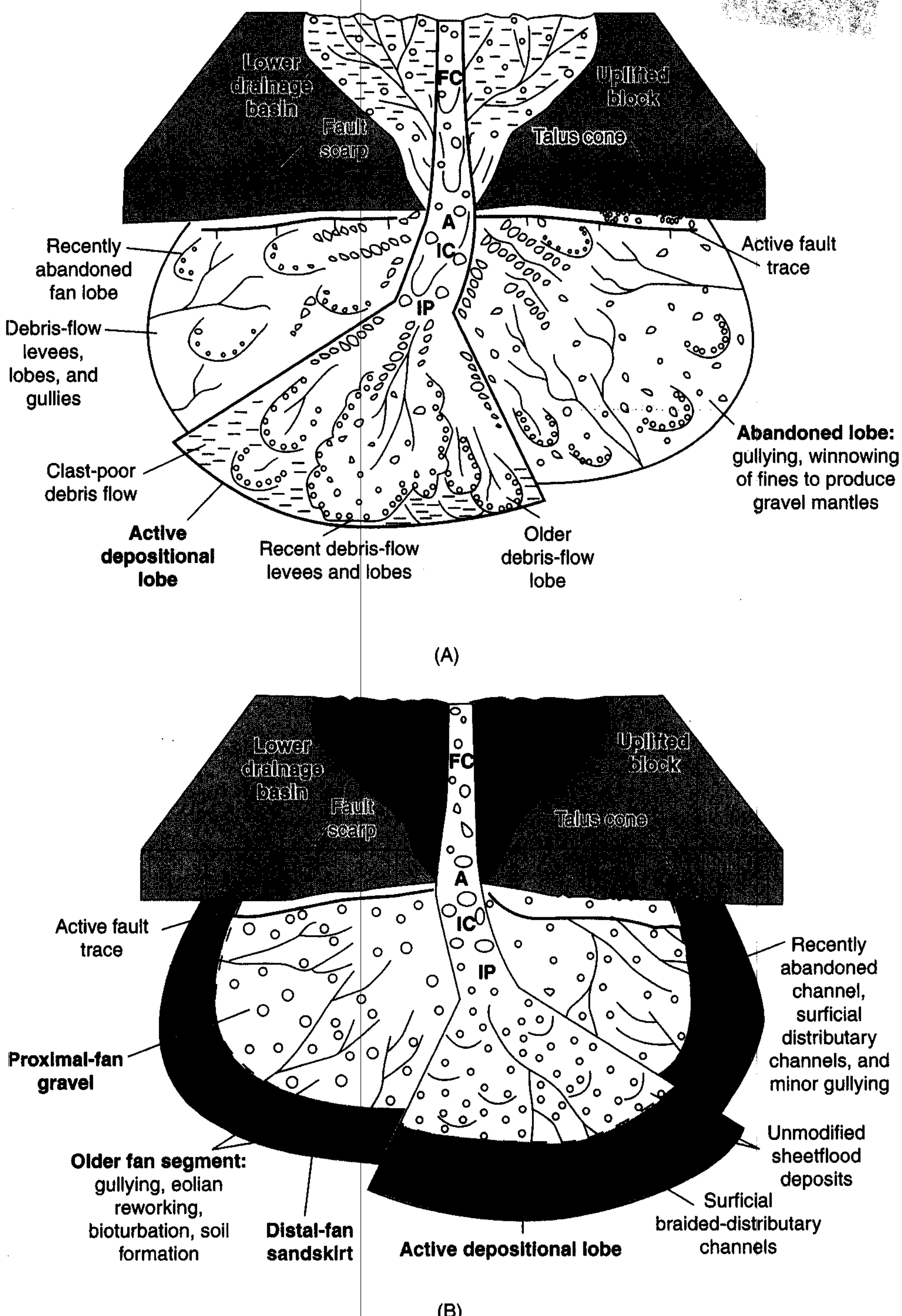

Figure 7.20
Schematic diagrams showing the lower drainage basin, and primary and secondary depositional features of alluvial fans, including those dominated by (A) debris-flow processes and (B) sheetflood processes. Abbreviations: A = fan apex; FC = drainage-basin feeder channel; IC = incised channel on fan; IP = fan intersection point.

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Size and Shape Pediments vary in size from less than 1 square kilometer to hundreds of square kilometers, probably depending on fundamental geomorphic controls. Shape is also variable; pronounced irregularities occur when the rocks cut by the pediment surface have wide differences in resistance to erosion. Generally they tend to be fan-shaped in plan view, narrowing toward the mountain front and widening downslope. Across the pediment, the shape can be either convex or concave.

Surface Topography Contrary to lay opinion, pediments are not monotonous, smooth, flat surfaces but are dissected by incised stream channels and dotted with residual bedrock knobs, called **inselbergs,** that stand above the general level of the pediment itself. Inselbergs have been investigated repeatedly with regard to pedimentation (e.g., Twidale 1962, 1978; Kesel 1973, 1977; Twidale and Bourne 1975). In some cases, these residual hills might be the last unconsumed vestiges of a landscape that has been totally pedimented. More likely, however, most inselbergs represent areas of rock that are more resistant to weathering and erosion (Kesel 1977; Twidale 1978).

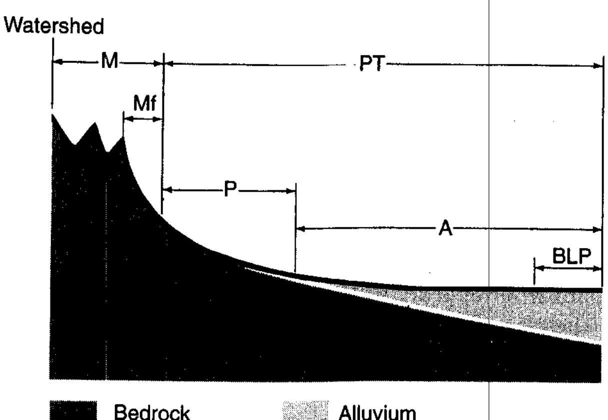

Figure 7.27
Landforms in the mountain-basin geomorphic system: M = mountain area, Mf = mountain front, P = pediment, PT = piedmont plain, A = alluvial plain, BLP = base-level plain.

The frequency and size of both incised valleys and inselbergs seem to increase toward the mountain front, sometimes giving the topography the aspect of gently rolling hills and valleys. Some of the channels and other depressions may be filled with alluvium up to 3 m thick (Cooke and Warren 1973), giving the false impression that the bedrock surface is smooth and perfectly planed.

The Piedmont Angle The upper boundary of the pediment is usually marked by an abrupt change from the steep slopes of the mountain front to the low declivity of the pediment surface. In plan view the boundary is usually linear, but embayments into major valleys of the mountain front can give the trace a rounded or crenulate appearance. The angle formed by the junction of the two surfaces is the **piedmont angle** (fig. 7.28). Its development and maintenance have traditionally been cited as evidence in theoretical models of pediment origin.

In detail, the piedmont angle can take the form of a narrow zone of intense curvature rather than a distinct angle. Twidale (1967) reports mountain front slopes of 22° changing to pediment gradients of 3°over a transition zone 100 m wide. Both the magnitude of the piedmont angle and the sharpness of the angular relationship are probably related to structural or lithologic control (Denny 1967; Twidale 1967; Cooke and Reeves 1972). For example, the piedmont angle is most distinct when formed in granitic and gneissic rocks, perhaps because these rocks tend to break down into bimodal debris consisting of sand and boulders. The assumption is that the piedmont angle represents an adjustment of the two slopes to the size of debris they are required to transport; the boulder size of the mountain front is presumably controlled by joint spacing. Piedmont angles seem to be less well-defined in other rock types, but exceptions do occur (Selby 1982). Other factors such as joint density and subsoil weathering may affect the character of the piedmont angle (Mabbutt 1966; Twidale 1967, 1978).

Slope The longitudinal profile of almost all pediments is slightly concave-up, although local convexities do occur. Overall longitudinal convexities have been suggested as a theoretical possibility if the suballuvial bench is also considered (Lawson 1915), but available observation and geophysical data (Langford-Smith and

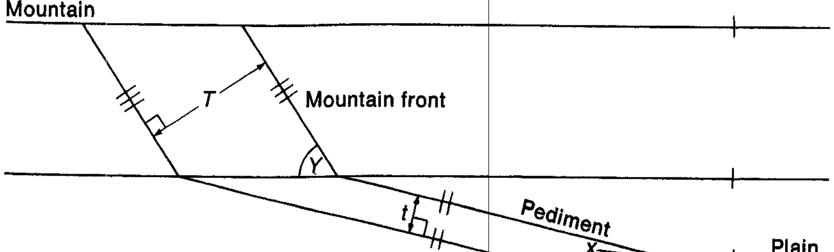

Figure 7.28
Retreat of pediment surface parallel with expanding plain. Piedmont angle at *Y* is maintained. Ratio of hillslope erosion (*T*) to pediment erosion (*t*) must be constant, and the rectilinear profile is perpetuated.
(Ruxton and Berry 1961)

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River Incision, erosion, knickpoints

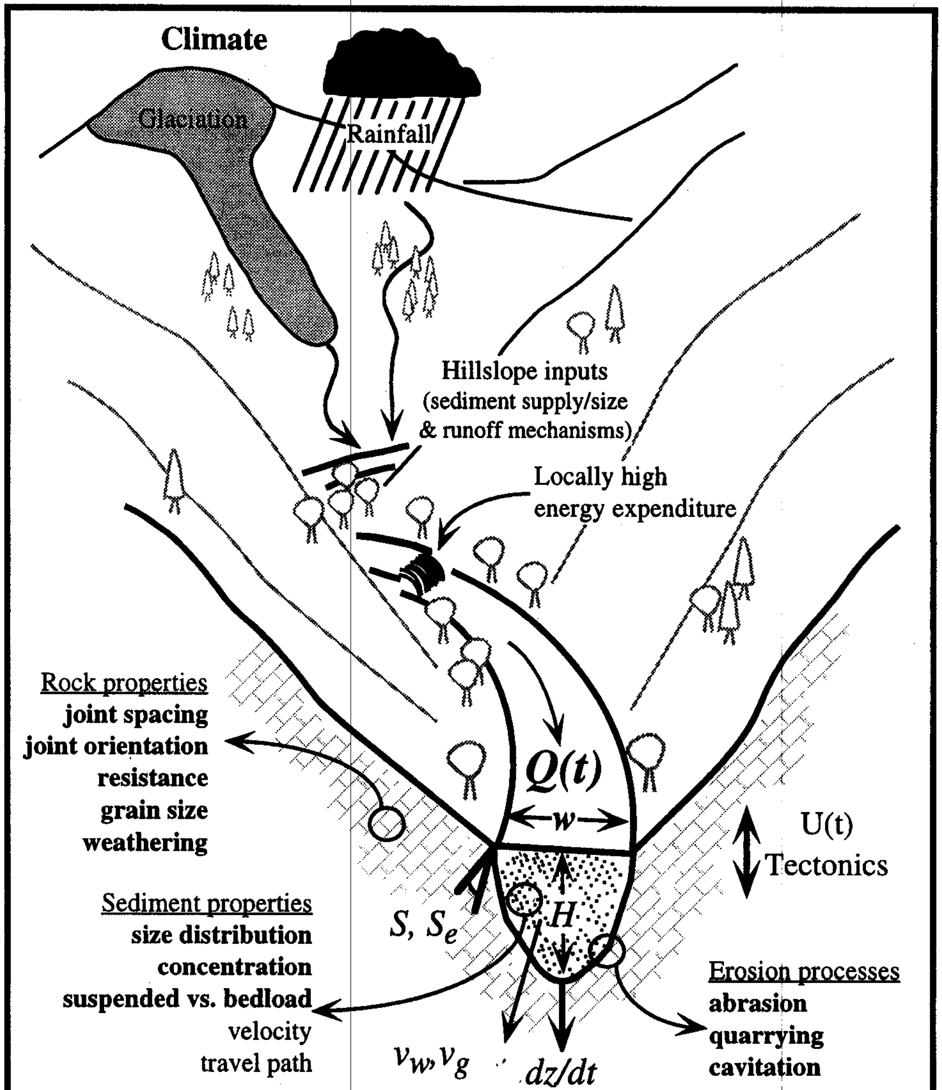

Figure 1. A schematic of a bedrock channel system, showing important variables that act to set the erosion rate, *dz/dt*, of the channel and terms used in the text. Channel variables are: $Q(t)$: discharge as a function of time; w: channel width; H: flow depth; v_w: water velocity; v_g: sediment velocity; A_D: drainage area; and S and S_e: channel and energy slope, respectively. Variables that are at least partially subsumed within the parameter, K, found in all reach-scale erosion rules, are shown in bold.

2.1. *Abrasion*

2.1.1. *Theory*. Rock erosion by abrasion is accomplished by removal of material from a rock surface through forcible impact by entrained sediment. The rate at which material is removed depends upon the kinetic energy flux to the surface, delivered by impacting grains, and the "susceptibility" of the rock surface to abrasion [e.g., *Anderson*, 1986; *Foley*, 1980a]. The impacts produce fractures within minerals, dislodge individual grains, or break off flakes from the rock surface. Experimental studies of abrasion by aeolian sediment transport reveal that the mass of material removed is roughly proportional to the to the kinetic energy delivered by the impact [e.g., *Greeley and Iversen*, 1985; *Suzuki and Takahashi*, 1981]. At the grain scale, the grain velocity, v_g, diameter, D, and density, ρ_g, set its kinetic energy, and the delivery of this kinetic energy increases as grain impact angle, a, relative to the bed, increases toward vertical (90 degrees). The "susceptibility", S_a, relates kinetic energy delivery to mass of rock material removed, and is dependent primarily on the density, hardness, and fracture-mechanical properties of the

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Table 1. Published long-term average rates of bedrock channel incision (after Schumm and Chorley, 1983, Table 3-5, and Wohl et al., 1994b, Table 1).

Rate (cm/kyr)	Lithology	Location	Drainage Area (km^2)	Climate, tectonics	Time range of incision	Source
9	granite, andesite	Sierra Nevada, California, USA	35 000	arid, uplift	Pliocene-Quaternary	Huber, 1981
30	sedimentary	Colorado, USA	11 800	semiarid, uplift	Miocene-Quaternary	Larson et al., 1975
7	metamorphic	Colorado, USA	----	semiarid, uplift	Pliocene-Quaternary	Scott, 1975
45-130	sedimentary	Nahal Zin, Israel	1 540	hyperarid, uplift	Quaternary	Goldberg, 1976 Schwarcz et al., 1979 Yair et al., 1982
10	sedimentary	Nahal Paran, Israel	3 600	hyperarid, uplift	Quaternary	Wohl et al., 1994b
30	basalt, limestone	Utah, USA	9 900	semiarid, uplift	Quaternary	Hamblin et al., 1981
9.5	sedimentary	Arizona, USA	68 500	semiarid, uplift	Quaternary	Rice, 1980
23-25	basalt	Jalisco, w. Mexico	----	arid, uplift	Pliocene-Quaternary	Righter, 1997
15	suggested average rate of bedrock channel incision in the middle latitudes					Pitty, 1971
25-47	sedimentary	Utah, USA	115 000	semiarid, uplift	Quaternary	Harden and Coleman, 1989
1 000	igneous and metamorphic	Pakistan	260 000	semiarid, uplift	Quaternary	Leland et al., 1995
70-180	sedimentary	n. California, USA	655	Mediterranean, uplift	Holocene	Merritts et al., 1994
<25	sedimentary, intrusive igneous	Central California, USA	10-20	Mediterranean, uplift	Quaternary	Rosenbloom and Anderson, 1994
0.5-8	basalt	Kauai, USA	0.1-90	seasonal tropical to semiarid, uplift	Pliocene-Quaternary	Seidl et al., 1994
*40-100	basalt	Kauai, USA	0.1-90	seasonal tropical to semiarid, uplift	Pliocene-Quaternary	Seidl et al., 1997
50-690	sedimentary	Montana, USA	1 420	humid temperate, uplift	Quaternary	Foley, 1980b
≤ 1000	mudstone	s. Japan	0.15-0.4	humid temperate, uplift	Quaternary	Mizutani, 1996
5.7	limestone	New Guinea	0.02	humid tropical, uplift	Quaternary	Chappell, 1974
*$\leq 1.57 \times 10^5$	sedimentary	Ontario, Canada	686 000	humid temperate, passive	Quaternary	Tinkler et al., 1994
0.5-3	basalt, metamorphic	southeastern Australia	20-400	humid temperate, passive	Miocene-Quaternary	Young and McDougall, 1993
300	basalt, sedimentary	Svalbard, Norway	----	subpolar, passive	Quaternary	Büdel, 1982
2.7	Carbonates, Crystalline rocks	Virginia, USA	----	humid temperate, passive	Quaternary	Granger et al., 1997

* knickpoint migration upstream

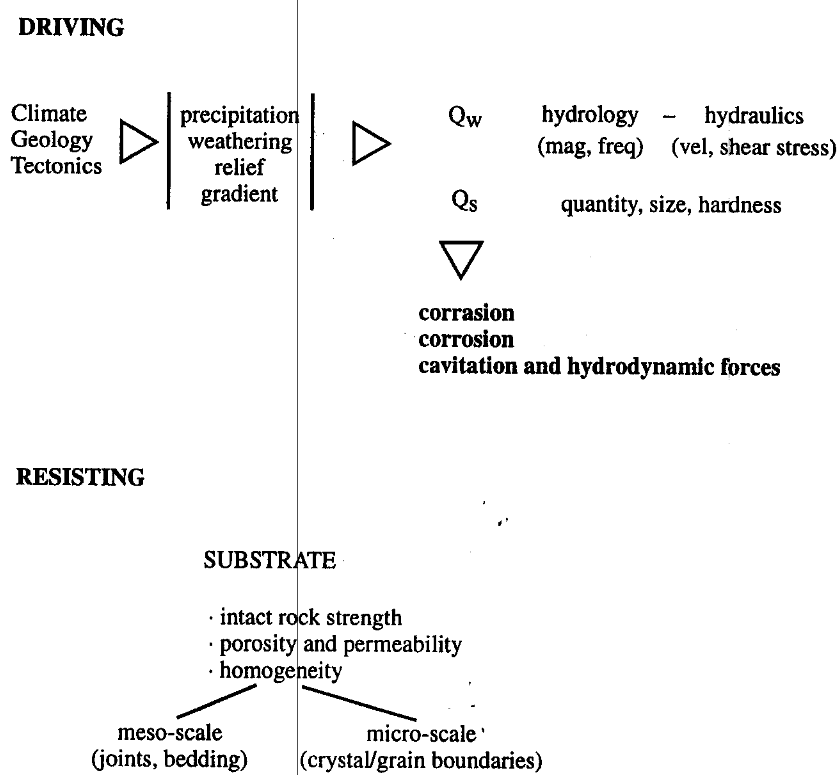

Figure 11. Variables influencing driving and resisting forces acting along bedrock channels.

CHANNEL GEOMETRY AS A REFLECTION OF EROSION

The controls on channel boundary (ir)regularity, width/depth ratio, planform, and gradient come under the two basic categories of substrate and hydraulics. The specific characteristics of these controls influence channel geometry at various spatial scales (e.g. micro scale variability in permeability to macro-scale regional joint patterns) and temporal scales (e.g. flood abrasion to low-flow corrosion). Meso-scale erosive processes and the resulting channel form appear to be dominated by hydraulic controls in some settings [e.g. *Wohl,* 1993, 1994], whereas substrate variability dominates channel form in other examples [e.g. *Miller,* 1991]. In most cases, we know too little about the forces acting along bedrock channels to precisely delineate the influences of individual controls.

Because most bedrock channel boundaries are very stable over the timespan of individual studies, most estimates of rates of bedrock channel incision are based on long-term averages (Table 1, chpt 1). These rates have a very large range (0.5-1000 cm kyr^{-1}) as a result of differences in basin area, discharge, gradient, substrate resistance, and tectonic regime. The few direct measurements of short-term incision rates are also quite variable (Table 2, chpt. 1) for the same reasons as the long-term rates, as well as the increased importance of recent flow history. Numerous investigators have suggested that high magnitude, low frequency floods are particularly important in shaping bedrock channels because of the high boundary resistance of these channels. If the majority of channel incision occurs during frequent floods, then short-term measurements of incision will be largely influenced by the (non)occurrence of such floods during the measurement period. Direct measurements of bedrock channel erosion during individual floods as well as during intervals between floods would be extremely useful.

CONCLUSIONS

At the most basic level, bedrock channel morphology reflects an adjustment between driving and resisting forces, which are controlled by the variables in Figure 11. Our ability to predict channel morphology as a function of controlling variables will thus be governed by our ability to quantify these driving and resisting forces. Both types of forces are dynamic in bedrock channels, varying both across space and as a function of time in response to rate of weathering, magnitude, frequency, and duration of flows, sediment supply to the channel, and so forth. The prevailing assumption has been that substrate dominates bedrock channel morphology. However, the description of morphological changes that seem to be independent of

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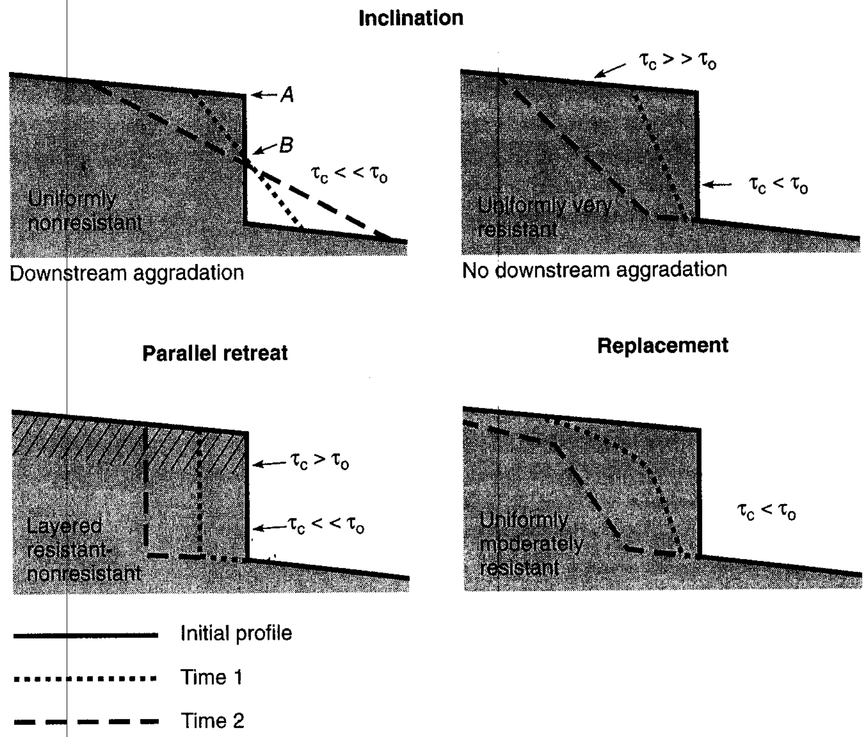

Figure 6.42
Models of knickpoint evolution for various types of bed material. τ_c is critical bottom shear stress needed to initiate erosion. τ_o is actual bottom shear stress. The knickpoint lip, shown at *A*, is the break in slope where the channel becomes oversteepened. The knickpoint face at *B* extends from the lip to the base of the knickpoint.
(Gardner 1983)

of the geologic and hydrologic setting. Miller (1991), for example, found that the distribution, shape, and headward migration of bedrock knickpoints along streams in southern Indiana are highly dependent on the stratigraphic and structural characteristics of the rock into which they are cut.

Gardner (1983) suggests that knickpoint evolution can occur in any of three different modes, depending on the balance between the shear resistance of the bed material and the shear stress produced in the river flow (fig. 6.42). In *knickpoint inclination* a uniform change in slope of the knickpoint face occurs, the details of which depend on the material resistance and whether the river can transport the sediment away from the front of the knickpoint. Where the shear stress needed to erode this debris (τ_c) is less than the actual shear stress in the river (τ_o), aggradation will occur and the fill becomes part of the readjusted gradient. *Parallel retreat* is characterized by retreat of the near-vertical knickpoint face without a change in its inclination. This mode of migration is produced best if layering exists in the parent material such that a more resistant zone overlies a less resistant zone. The third type of migration, *knickpoint replacement,* occurs when erosion of the bottom takes place upstream from the knickpoint lip as well as along the face. It results in a knickpoint profile consisting of two distinct zones in which the original slope has been modified.

Adjustment of Shape and Pattern

In the ideal case, any prediction of the adjustments in channel shape would include an understanding of the possible direction of change in the hydraulic variables (width, depth, sinuosity, etc.), the magnitude of the changes, and the rates of alterations in response to a threshold crossing event. Given our current understanding of fluvial processes, it appears that the best we can do is to decipher the potential directions in which the form variables will adjust. The most extensively utilized model for this purpose has been presented by Schumm in his concept of *river metamorphosis.* In a series of papers, Schumm (1965, 1968, 1969) pieced together many of the empirical equations we have examined into a comprehensive, though qualitative, model of possible river adjustments to altered hydrology and load. Four equations are produced when changes in both sediment load and discharge are considered. They are expressed as follows (Schumm 1969):

$$Q_w^+Q_t^+ = w^+ L^+ F^+ P^- S^\pm d^\pm$$

$$Q_w^-Q_t^- = w^- L^- F^- P^+ S^\pm d^\pm$$

$$Q_w^+Q_t^- = d^+ P^+ S^- F^- w^\pm L^\pm$$

$$Q_w^+Q_t^- = S^+ F^+ d^- P^- w^\pm L^\pm$$

In these equations Q_t is the percentage of the total load transported as bed material load (sand-sized or larger), and Q_w can be either the mean annual discharge or the mean annual flood. The other variables are width (w), depth (d), slope (S), meander wavelength (L), width-depth ratio (F), and sinuosity (P). The plus or minus exponents indicate whether the dimensions of the variables are increasing or decreasing.

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SOIL BASICS

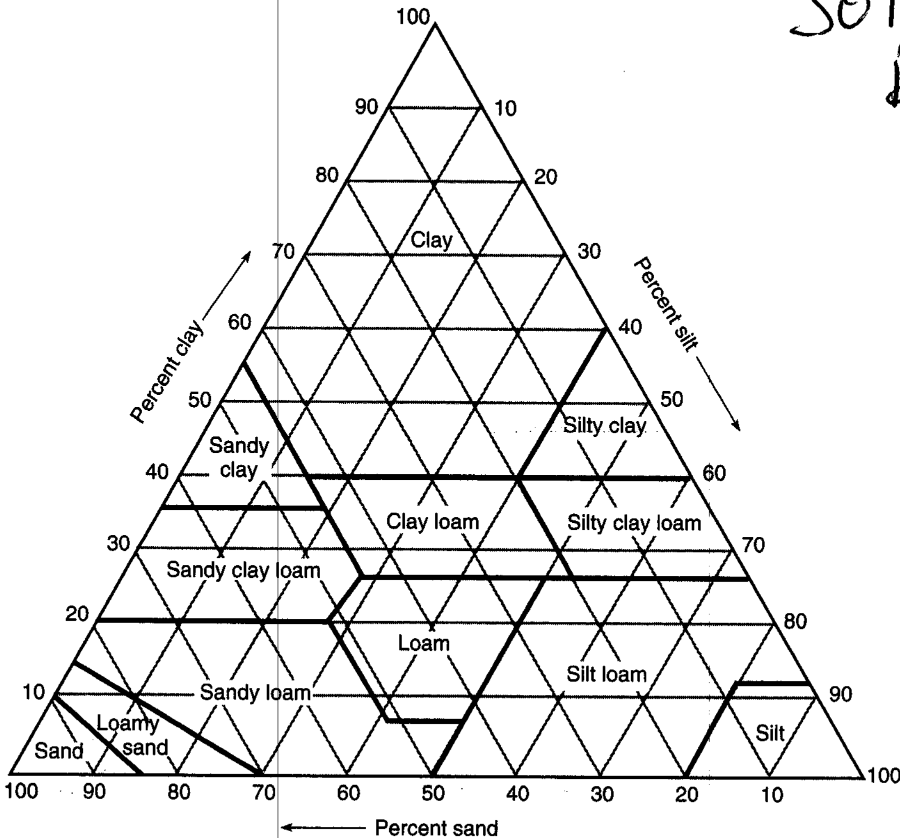
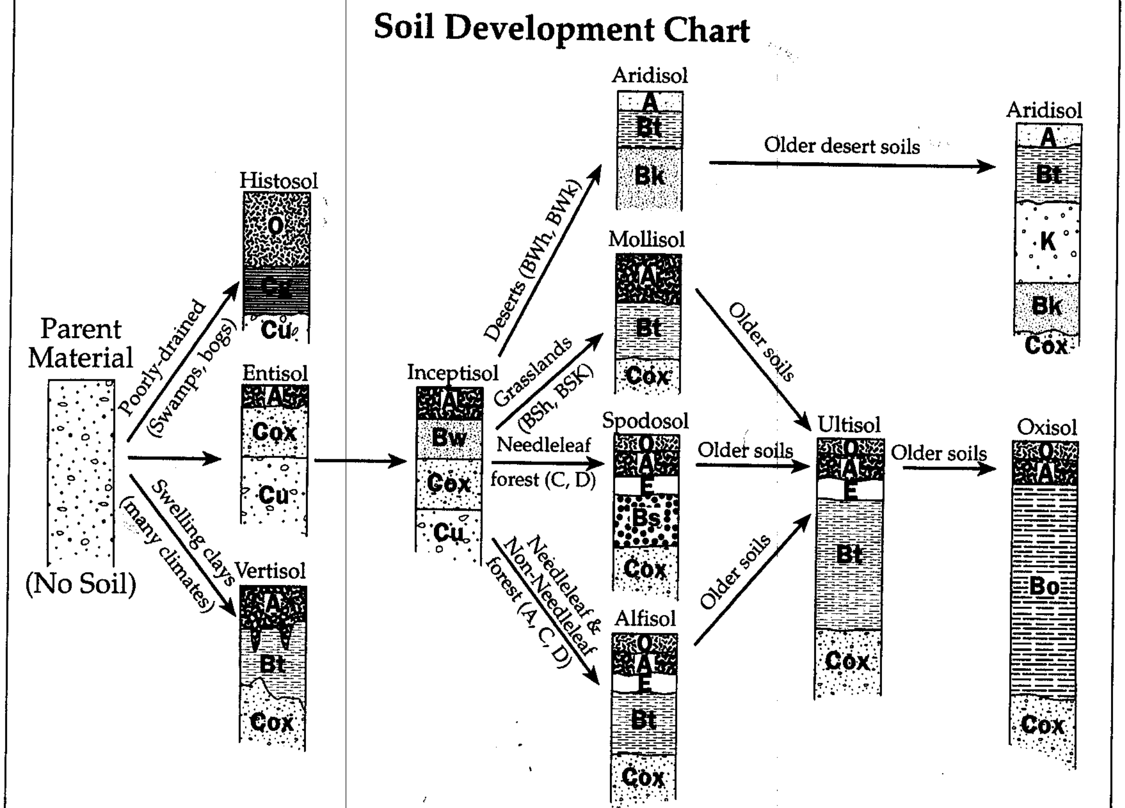


Figure 3.12
Percentages of clay (< 0.002 mm), silt (0.002–0.05 mm), and sand (0.05–2.0 mm) in basic soil textural classes as defined by the U.S. Department of Agriculture.
(Soil Survey Staff, 1951)

ganic content are typically black to dark brown, whereas the presence of ferric iron is indicated by yellow-brown to red colors. Light gray to white colors are associated with a concentration of SiO_2 or $CaCO_3$. It should be recognized, however, that small amounts of a pigmentor can cause rather intense discoloration, and so color alone may be a poor index of the total quantity of the pigmenting substance. *Texture* is simply the relative proportions of different particle sizes in a soil horizon, and is somewhat analogous to the property of sorting as used by geologists. However, it is based only on the particles that are less than 2 mm in diameter (fig. 3.12). *Structure* in soils is a unique characteristic in that it designates the shape developed when individual particles cluster together into aggregates called *peds* (fig. 3.13). In clay-rich soils, the openings between peds may play an extremely important geomorphic role by providing the primary avenues for downward percolation through an otherwise impermeable soil. *Organic matter* in soils consists mainly of dead leaves, branches, and the like, called *litter,* and the amorphous residue, called *humus,* that develops when litter is decomposed. Litter may form at mean annual temperatures as low as freezing, but its optimum production occurs at about 25° to 30° C and decreases rapidly above those levels. Microorganisms that convert litter to humus begin to function at temperatures slightly above freezing (5° C), but the optimum temperature for their life activities may be as high as 40° C (fig. 3.14). It is significant that at temperatures between 0° and 25° C, humus is produced in abundance, but above 25° C little if any humus is accumulated. Humus has an important effect on soil formation because it includes chelators that promote the leaching of iron and aluminum, and increases water absorption. In addition, the development of humus releases CO_2 in high concentrations, leading to unusual amounts of carbonic acid within the humic zone and an associated lowering of the pH.

The total quantity of water that can be held in a soil is the *available water capacity* (AWC). By combining this parameter with the bulk density (dry weight of soil/unit volume), an estimate of the depth of wetting can be made (see Birkeland 1999). Such information is significant in that it relates to many soil properties, especially those affected by the redistribution of material during the downward percolation of water.



O Horizon:	Surface horizon dominated by organic matter (e.g., leaves) in various stages of decay
A Horizon:	Surface horizon or beneath O horizon dominated by mineral matter, but with sufficient humus to darken color
E Horizon:	A light-gray subsurface horizon that has been leached of pigments by organic acids produced by the decay of needleleaf litter
Bw Horizon:	A young B horizon that has been slightly reddened by oxidation, but is not yet clay-enriched
Bs Horizon:	Illuvial accumulation of amorphous organic matter – sesquioxide (Al_2O_3; Fe_2O_3) complexes
Bk Horizon:	A light-colored B horizon due to the coating of mineral grains by carbonates
Bt Horizon:	A 'mature' B horizon that is reddened by oxidation as well as clay-enriched
Bo Horizon:	A deep red, highly weathered and leached B horizon rich in residual sesquioxides; usually restricted to very old soils in tropical climates
Cox Horizon:	An oxidized horizon beneath a B or A horizon
K Horizon:	Similar to a Bk horizon, but so greatly enriched in carbonates that the horizon is white
Cg Horizon:	A 'gleyed' subsurface horizon that shows patchy zones of blue, gray, and green colors from reduced conditions; may be local areas of oxidized material
Cu Horizon:	Unweathered parent material

Figure 2.7 Soil orders arranged into a development scheme, with greater development (and age, in most cases) to the right. Parent material controls the first step in the development, and time and bioclimate beyond that. Letters in parentheses depict climate classification of Köppen. *(Courtesy of Dennis Netoff, 1997.)*

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Table 1.1 Soil-Horizon Nomenclature

Description of master horizon horizon and subhorizons
O horizon—Surface accumulations of mainly organic material; may or may not be, or has been, saturated with water. Subdivided on the degree of decomposition as measured by the fiber content after the material is rubbed between the fingers.
Oi horizon—Least decomposed organic materials; rubbed fiber content is greater than 40% by volume.
Oe horizon—Intermediate degree of decomposition; rubbed fiber content is between 17 and 40% by volume.
Oa horizon—Most decomposed organic materials; rubbed fiber content is less than 17% by volume.
A horizon—Accumulation of humified organic matter mixed with mineral fraction; the latter is dominant. Occurs at the surface or below an O horizon; Ap is used for those horizons disturbed by cultivation.
E horizon—Usually underlies an O or A horizon, and can be used for eluvial horizons within or between parts of the B horizon (e.g., common above fragipan, x). Characterized by less organic matter and/or fewer sesquioxides (compounds of iron and aluminum) and/or less clay than the underlying horizon. Many are marked by a concentration of sand and silt. Horizon is light colored due mainly to the color of the primary mineral grains because secondary coatings on the grains are absent; relative to the underlying horizon, color value will be higher or chroma will be lower.
B horizon—Underlies an O, A, or E horizon, and shows little or no evidence of the original sediment or rock structure. Several kinds of B horizons are recognized, some based on the kinds of materials illuviated into them, others on residual concentrations of materials. Subdivisions are:
Bh horizon—Illuvial accumulation of amorphous organic matter–sesquioxide complexes that either coat grains or form sufficient coatings and pore fillings to cement the horizon.
Bhs horizon—Illuvial accumulation of amorphous organic matter–sesquioxide complexes, and sesquioxide component is significant; both color value and chroma are three or less.
Bk horizon—Illuvial accumulation of alkaline earth carbonates, mainly calcium carbonate; the properties do not meet those for the K horizon.
B*l* horizon—Illuvial concentrations primarily of silt (Forman and Miller, 1984). Used when silt cap development reaches stages 5 and 6.
Bo horizon—Residual concentration of sesquioxides, the more soluble materials having been removed.
Bq horizon—Accumulation of secondary silica.
Bs horizon—Illuvial accumulation of amorphous organic matter–sesquioxide complexes if both color value and chroma are greater than three.
Bt horizon—Accumulation of silicate clay that has either formed in situ or is illuvial (clay translocated either within the horizon or into the horizon); hence it will have more clay than the assumed parent material and/or the overlying horizon. Illuvial clay can be recognized as grain coatings, bridges between grains, coatings on ped or grain surfaces or in pores, or thin, single or multiple near-horizontal discrete accumulation layers of pedogenic origin (clay bands or lamellae). In places, subsequent pedogenesis can destroy evidence of illuviation. Although Soil Survey Division Staff (1993) does not include this, clay accumulation that lacks evidence for illuviation is included (could have been formed in situ, for example).
Bw horizon—Development of color (redder hue or higher chroma relative to C) or structure, or both, with little or no apparent illuvial accumulation of material.
By horizon—Accumulation of secondary gypsum.
Bz horizon—Accumulation of salts more soluble than gypsum.
K horizon. A subsurface horizon so impregnated with carbonate that its morphology is determined by the carbonate (Gile and others, 1965). Authigenic carbonate coats or engulfs nearly all primary grains in a continuous medium. The uppermost part of a strongly developed horizon is laminated, brecciated, and/or pisolithic (Machette, 1985). The cemented horizon corresponds to some caliches and calcretes.
C horizon—A subsurface horizon, excluding R, like or unlike material from which the soil formed, or is presumed to have formed. Lacks properties of A and B horizons, but includes materials in various stages of weathering.
Cox and Cu horizons—In many unconsolidated deposits, the C horizon consists of oxidized material overlying seemingly unweathered C. The oxidized C does not meet the requirements of the Bw horizon. In stratigraphy, it is important to differentiate between these two kinds of C horizons. Here Cox is used for oxidized C horizons and Cu for unweathered C horizons. Cu is from the nomenclature of England and Wales (Hodson, 1976). Alternatively the Cox can be termed BC or CB.
Cr horizon—In soils formed on bedrock, there commonly will be a zone of weathered rock between the soil and the underlying rock. If it can be shown that the weathered rock has formed in place, and has not been transported, it is designated Cr. Such material is the saprolite of geologists; in situ formation is demonstrated by preservation of original rock features, such as grain-to-grain texture, layering, or dikes. If such material has been moved, however, the original structural features of the rock are lost, and the transported material may be the C horizon for the overlying soil. Those Cr horizons with translocated clay, as shown by clay films, are termed Crt.
R horizon—Consolidated bedrock underlying soil.

(continued)

tographs, so one can become familiar with the classification criteria. However, to avoid becoming overwhelmed, one should read Buol and others (1997). This readable text covers all the orders, chapter by chapter, and each is subdivided into the central concept and geographic distribution of the order, the setting, pedogenic processes, uses, and classification. The reader should be aware, however, that the classification has gone through many revisions, so consult Soil Conservation Service (1994) for the latest (watch for still more recent keys).

The new classification has much appeal for geomorphologists and ecologists, at least down to the suborder level. This is because soil profile development is included in the classification, as well as base saturation, amount of organic matter, and properties indicative of relative wetness and dryness. Most geomorphological soil studies deal with climatic and time factors, and the above properties will later be shown to be related to both.

Eleven orders are now recognized in the new classification; these are subdivided into 55 suborders, and the latter into over 200 great groups. One should be impressed with the magnitude of this task as it begins at the lowest level of classification, the approximately 12,000 soil series in the United States. The orders are basically differentiated by a particular horizon or horizon combinations that occur in the soil profile. These usually can be recognized in the field without recourse to laboratory analysis. One criticism of Soil Taxonomy, from a geomorphological point of view, is the overemphasis on classification at the order level by surface horizon (e.g., Mollisol). In contrast, horizons beneath the A horizon commonly are more important to geomorphologists. Classification into suborders requires an increasingly quantitative knowledge of soil properties and soil-moisture and soil-temperature regimes. However, it is often not necessary to take these measurements because, with experience, soil classification can be estimated from properties recognizable in the field (e.g., thin Av horizon with a calcic horizon at depth points toward aridic moisture regime).

To classify a soil at the order and suborder level, one must be able to identify the diagnostic horizons as well as the soil-moisture and -temperature regimes. The reader will find that some of these definitions are exceedingly complicated, but in time they will make some sense. Diagnostic horizons are so named because they are essential to classify the soil. Epipedons are the surface diagnostic horizons. The diagnostic horizons are somewhat similar to the field-designated soil horizons, although in places they can encompass several different field-designated soil horizons. In an extreme example of the latter, the mollic epipedon can include both the A and B horizons, as long as mollic properties are obtained. For some diagnostic horizons, the criteria are so complex that one has to read the defining criteria in detail using both the field and laboratory data (e.g., spodic horizon). Only the main discriminating criteria are given here (Table 2.1); Soil

Table 2.1 Common Diagnostic Horizons Used in Soil Taxonomy

Diagnostic Horizon	Defining Criteria	Probable Field Horizon Equivalent
Epipedons		
Mollic epipedon	Must be 10 cm thick if on bedrock, otherwise a minimum of 18 or 25 cm thick depending on subhorizon properties and thicknesses; color value darker than 3.5 (moist) and 5.5 (dry); chroma less than 3.5 (moist); organic carbon content at least 0.6%; structure developed and horizon not both massive and hard; base saturation ≥50%	A, A + E + B, A + B
Umbric epipedon	Meets all criteria for mollic epipedon, except base saturation <50%	A
Ochric epipedon	Epipedon that does not meet requirements of either mollic or umbric epipedons	A
Histic epipedon	Complex thickness requirements, but >20 cm thick; >12% organic carbon, with some adjustment for percent clay; saturated with water for 30 consecutive days or more per year, or artificially drained	O
Subsurface Horizons		
Albic horizon	Light colored with few to no coatings on grains—light color is that of grains; if color value (dry) is 7 or more, or color value (moist) is 6 or more,	E

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(continued)

Table 2.1 (continued)

Diagnostic Horizon	Defining Criteria	Probable Field Horizon Equivalent
	chroma is 3 or less; if color value (dry) is 5 or 6, or color value (moist) is 4 or 5, chroma is closer to 2 than to 3	
Argillic horizon	Complex thickness requirements, but at least 7.5 or 15 cm thick depending on texture and thickness of overlying horizons; must have these greater amounts of clay relative to overlying eluvial horizon(s) or underlying parent material: (a) if the latter horizons have <15% clay, argillic horizon must have a 3% absolute increase (10 vs. 13%); (b) if the latter horizons have 15 to 40% clay, the ratio of clay in argillic horizon relative to them must be 1.2 or more, and (c) if the latter horizons have >40% clay, the argillic horizon must have an 8% absolute increase (42 vs. 50%); in most cases, evidence for translocated clay should be present (clay as bridges between grains or clay films in pores or on ped faces)	Bt
Kandic horizon	Minimum thickness is either 15 or 30 cm; complex clay increase requirements relative to overlying eluvial horizon(s) or underlying parent material: if the latter horizons have <20% clay, the kandic horizon must have a 4% absolute increase; if the latter horizons have 20–40% clay, the kandic horizon must have at least 20% more clay; if the latter horizons have >40% clay, the kandic horizon must have at least 8% absolute increase; complex depth-texture relations; CEC <16 meq/100 g clay	Bt
Natric horizon	In addition to properties of argillic horizon: prismatic or columnar structure; 15% or more exchangeable sodium; exchangeable magnesium and sodium exceed exchangeable calcium and exchange acidity	Btn
Spodic horizon	Minimum thickness is 2.5 cm, and contains >85% spodic materials: the latter are amorphous materials composed of organic matter and Al, with or without Fe; usually beneath an albic or Ej horizon	Bh, Bs, Bhs
Cambic horizon	Base usually at least 25 cm deep; stronger chroma or redder hue relative to underlying horizon; soil structure or absence of rock or sediment structure; weatherable minerals present; carbonates removed if originally present; no cementation or brittle consistence	Bw
Oxic horizon	At least 30 cm thick; >15% clay and sandy loam or finer; cation exchange capacity ≤16 meq/100g soil; few weatherable minerals	Bo
Calcic horizon	At least 15 cm thick; 15% $CaCO_3$; relative to underlying horizon, has at least 5% more $CaCO_3$, or at least 5% by volume secondary carbonate	Bk
Petrocalcic horizon	Horizon continuously cemented with $CaCO_3$	Km
Gypsic horizon	At least 15 cm thick; at least 5% more gypsum than underlying horizon; product of thickness(cm) times content (%) is 150 or more	By
Petrogypsic horizon	Strongly cemented gypsic horizon, commonly with greater than 60% gypsum	Bym
Salic horizon	At least 15 cm thick; at least 2% salts more soluble than gypsum; product of thickness (cm) times content (%) is 60 or more	Bz
Duripan	Silica cementation is strong enough that fragments do not slack in water	Bqm
Fragipan	Horizon of high bulk density relative to overlying horizons; formed in loamy material; although seemingly cemented with a brittle appearance, slacks in water; slowly permeable to water, so usually mottled; very coarse prismatic structure, usually with some bleached faces	Bx, Cx

Taken from Soil Conservation Service (1994).

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high K-feldspar content (ca. 70%) in the felsic rocks and high Ca-plagioclase (ca. 60%) in the basic rocks. Finally, the MgO and Fe oxides increase in the more mafic rocks because of the increase in mafic minerals (hornblende, pyroxenes, olivine). The ultrabasic rock dunite (mainly olivine) is included, as it is an extreme composition, and, as such, the direction of pedogenesis can be much different than in adjacent rocks of more average composition.

Average sedimentary rock compositions are quite variable (Table 7.2). Sandstone compositions reflect the source area rocks, and their high SiO_2 content is due to high quartz content. The cementing agent will vary, and common ones are carbonate and silica. Shale composition is dominated by that of the included clay minerals, with illite being common to many. Finally, limestone composition is dominantly that of CaO in $CaCO_3$ (CO_2 content usually not given). The few included mineral grains explain the rest of the chemistry.

Molar SiO_2/Al_2O_3 ratios (Table 7.2) give an idea of the changes that would have to take place during soil formation to produce the characteristic clay minerals from rocks of different composition (compare with Table 4.2).

■ Susceptibility of Rocks to Chemical Weathering

Rocks weather and erode at different rates, as can be seen in the variations in topographic relief that accompany variations in rock type. For rocks from widely spaced localities, however, factors other than rock type might influence the weathering variations observed. Nahon (1991) reviews data for several kinds of rocks in a variety of climatic settings. To compare rocks of differing lithology under similar conditions of weathering, however, it is best to study a sedimentary deposit, such as bouldery till or outwash, which includes a variety of rock types. Rocks from the same depth below the surface should have weathered under similar conditions.

Goldich's stability series for minerals (Table 7.1) can be used to predict igneous rock stability in the weathering environment. Rocks with a high content of more weatherable minerals should weather more rapidly than rocks with a high content of minerals resistant to weathering. To make a valid comparison, however, the rocks should be similar in crystallinity and grain size. For igneous rocks, therefore, resistance to weathering should be

$$\text{rhyolite} > \text{granite} > \text{basalt} > \text{gabbro}$$

Clay production should follow these trends, and it seems to (Barshad, 1958).

Grain size has an effect on the rate of weathering, for it is observed that coarser-grained igneous rocks commonly weather more rapidly than finer-grained rocks (Smith, 1962). This is readily seen in many tills in the Cordilleran Region. In the Sierra Nevada, for example, till of probable O-isotope stage 6 age has the

Table 7.2 Chemical Composition of Selected Igneous and Sedimentary Rocks

Oxide	Granite	Granodiorite	Diorite	Basalt	Ultrabasic rock	Average shale	Average sandstone	Average limestone
SiO_2	71.3	66.1	57.5	49.2	38.3	63.3	78	5.2
TiO_2	0.3	0.5	1	1.8	0.1	0.8		
Al_2O_3	14.3	15.7	16.7	15.7	1.8	17.2	7.2	0.8
Fe_2O_3	1.2	1.4	2.5	3.8	3.6	0.8	1.7	
FeO	1.6	2.7	4.9	7.1	9.4	5.5	1.5	0.5
MgO	0.7	1.7	3.7	6.7	37.9	3	1.2	8
CaO	1.8	3.8	6.6	9.5	1	3.5	3.2	43
Na_2O	3.7	3.8	3.5	2.9	0.2	1.5	1.2	0.1
K_2O	4.1	2.7	1.8	1.1	0.1	3.6	1.3	0.3
Molar SiO_2/Al_2O_3	8.5	7.3	6	5.5	32	6.5	18.6	9

Data from Garrels and Mackenzie (1971, Table 9.1), Best (1982, Appendix D), and Boggs (1991, Table 7.7).

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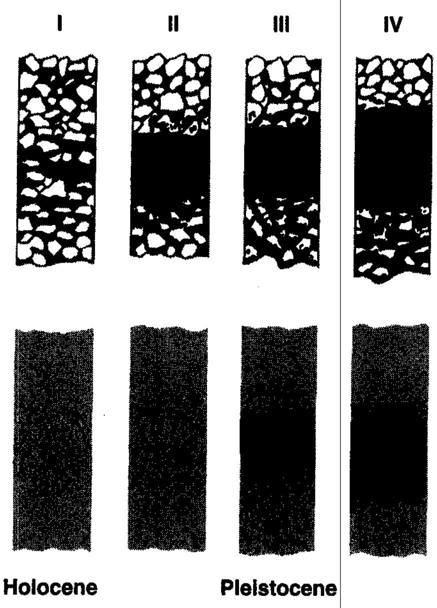


Stage	Gravelly sequence	Nongravelly sequence
I	Thin, partial or complete carbonate coatings.	Carbonate filaments and/or faint coatings on grains.
II	Carbonate coatings are thicker and there are some fillings in interstices.	Carbonate nodules separated by low-carbonate material.
III	Carbonate occurs essentially throughout the horizon, which is *plugged* in the last part of the stage.	
IV	Upper part of horizon nearly pure $CaCO_3$ exhibiting weakly platy structure; remainder of horizon plugged with carbonate.	
V	Laminar layering and strongly expressed platy structure; incipient brecciation and pisolith formation.	
VI	Brecciation and recementation of $CaCO_3$ layers are common.	

Figure 3.15
Schematic diagram of the stages of carbonate horizon formation in gravelly and nongravelly parent materials. Carbonate accumulations are indicated in black for clarity. Morphologic descriptions include stage additions by Machette (1985).
(Modified from Gile 1975)

with inconsistencies, Dokuchaiev's scheme had a great influence on American pedologists. His groups were defined in part by climate and vegetation, and until recently these genetic criteria were accepted as the basis of all soil classifications used in the United States. In fact, many Russian terms are still integrated in American soil nomenclature.

In the United States C. F. Marbut was the leading force in efforts to systematize soils. Marbut's classification, developed in the 1920s and 1930s, was based on characteristics that are present only in "mature" soils. The system, therefore, had no place for soils that were not fully developed. A more refined classification (Baldwin et al. 1938; Thorp and Smith 1949), designed to rectify many of the problems inherent in Marbut's system became the most extensively used soil classification in the United States until a fundamentally new system was introduced. In 1960 the S.C.S. completely revised the descriptive nomenclature and the classification of soils (Soil Survey Staff 1960). This revision is not simply a formulation of new class names (although that occurred) but represents a fundamental departure from the philosophical basis of earlier classifications. Until the new system was devised, all classifications were essentially genetic in scope; that is, the major soil classes were based on climatic and vegetal factors. Although the subdivisions were linked to observable aspects of the profile, these were not explicitly defined, and soil scientists inescapably allowed their knowledge of climate and vegetation to influence decisions about placing a soil in a particular group. In many cases, pedologists were classifying the genetic factors and not the tangible resulting

Available water capacity can be calculated if the moisture content at an upper limit (*field capacity*) and a lower limit (*permanent wilting point*) are known. Field capacity is determined by allowing a saturated sample to drain by gravity for at least 48 hours, by which time the remaining water content is held by adhesion to mineral and organic particles. After field capacity is reached, water can still be taken from the soil by plants until the tensional stresses holding the water in place become too great for the plants to break. At that point, the vegetation wilts. The water remaining in the soil is defined as the permanent wilting point. Both field capacity and permanent wilting point are expressed as a weight percentage according to the following equation:

$$P_w = \frac{W_s - W_d}{W_d} \times 100$$

where P_w is moisture percentage, W_s is total soil weight, and W_d is weight of soil after drying at 105° C. The available water capacity is simply the difference between the moisture content at field capacity and that at the permanent wilting point.

Soil Horizon Nomenclature and Description

Assuming that the vertical arrangement of the properties described above are distinct enough to identify a soil horizon, the pertinent consideration then becomes what nomenclature should be used to convey that information. In the United States, two systems of soil nomenclature are now in use. One is outlined in the *Soil Survey Manual* and is used in field descriptions of soil profiles (Soil Survey Division Staff 1993). The second is designed for the systematic classification of soils, and is based on the definition of diagnostic horizons, which, in many cases, can only be delineated following detailed laboratory analyses. The soil classification system used in the United States will be discussed in the next section. We will concentrate here on the nomenclature used to describe soils in the field.

Three kinds of symbols are used to denote horizons and layers in a soil profile. Capital letters, as shown in table 3.5, designate master horizons. Lowercase letters are used as suffixes to indicate specific characteristics of layers in the master horizon (table 3.6), and numbers are used as suffixes to connote vertical subdivision within a horizon or layer. In addition, numbers are prefixed to the master horizon designations to indicate a significant change in particle size or mineralogy within the soil. These signify a difference in the material from which the horizons have formed. In 1975 the Soil Conservation Service (S.C.S.) used Roman numerals as the prefix but have since changed to Arabic numerals (Soil Survey Division Staff 1993) (Note that the S.C.S. is now referred to as the National Resources Conservation Service). The number 1 is never used because it is implied to represent the material in the surface zones. Therefore, if no changes occur downward into the profile, prefix numbers are not needed.

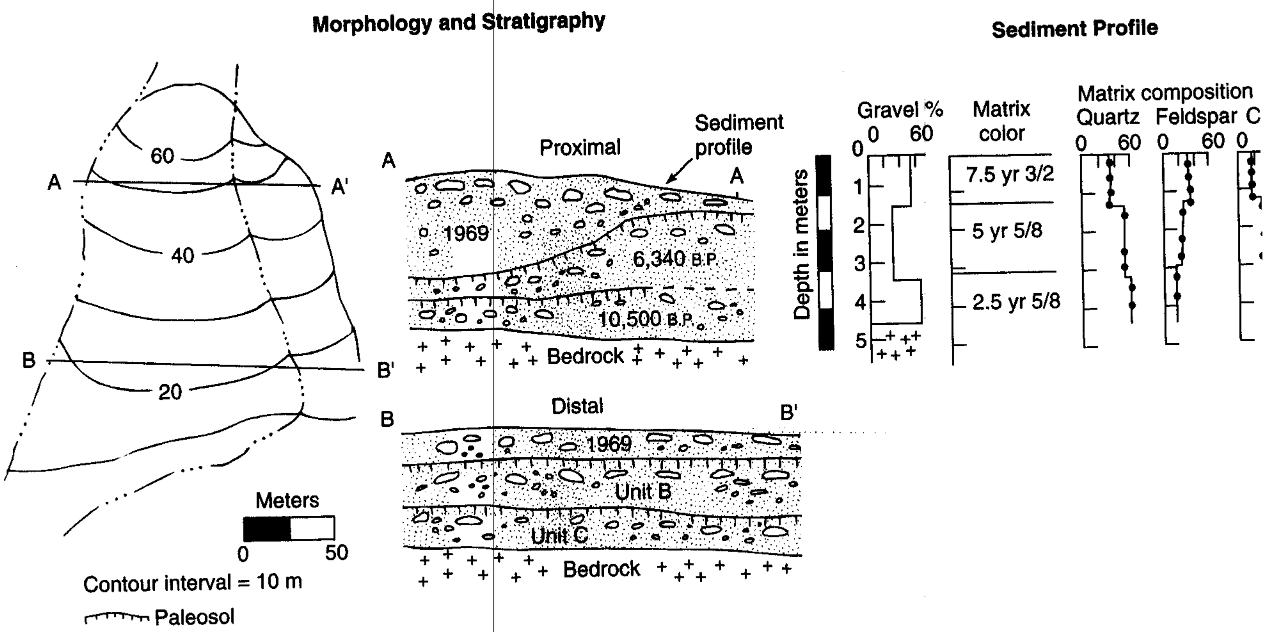
TABLE 3.5 Nomenclature of Soil Horizons

Horizon[a]	Characteristics
O	Upper layers dominated by organic material above mineral soil horizons. Must have > 30% organic content if mineral fraction contains > 50% clay minerals, or > 20% organics if no clay minerals.
A	Mineral horizons formed at the surface or below an O horizon. Contains humic organic material mixed with mineral fraction. Properties may result from cultivation or other similar disturbances.
E	Mineral horizons in which main characteristic is loss of silicate clay, iron, or aluminum, leaving a concentration of sand and silt particles of resistant minerals.
B	Dominated by obliteration of original rock structure and by illuvial concentration of various materials including clay minerals, carbonates, sesquioxides of iron and aluminum. Often has distinct color and soil structure.
C	Horizons, excluding hard bedrock, that are less affected by pedogenesis and lack properties of O, A, E, B horizons. Material may be either like or unlike that from which the solum presumably formed.
R	Hard bedrock underlying a soil.

Adapted from the Soil Survey Staff, 1960, 1975, 1981.

[a]Horizons can be divided into subhorizons by adding Arabic numbers.

The master horizons are designated by the capital letters O, A, E, B, C, and R (table 3.5). Horizons at the ground surface are called either O or A depending on the nature and amount of organic constituents they contain. The O horizon is dominated by undecomposed or partially decomposed materials, such as leaves, needles, lichens, or fungi. Mineral fragments represent only a small fraction (generally less than 50 percent) of the horizon by weight. In contrast, the A horizon is dominated by mineral grains and is normally considered to be the thin, dark-colored surface layer where decomposed organic matter is concentrated and where clays and mobile components are continuously leached downward, or *eluviated.* The E horizon underlies the O or A horizon. It is characterized by intense leaching that removes Fe^{+3} or organic coatings from the mineral grains, a process that usually imparts a bleached gray color to the horizon. The C and R horizons exist at the base of the profile. The C horizon is usually thought of as the underlying, unconsolidated parent materials that have been unmodified, or only very slightly modified, by soil-forming processes. The R horizon is simply consolidated bedrock beneath the soil.



(A)

(B)

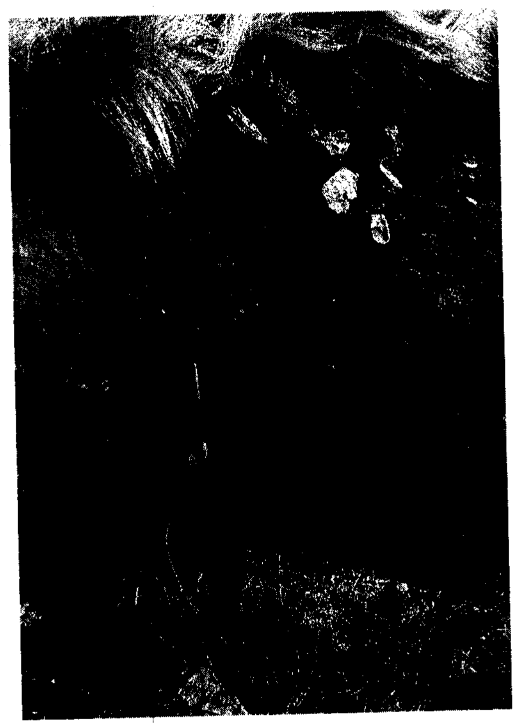


Figure 4.47
Stratified debris fan sediments in Nelson County, Va. (A) Radiocarbon-dated organic debris shows that events like the Camille flows have occurred there at least three times in the last 11,000 years. (B) Stratified floodplain sediments in Davis Creek, Va. The top layer above the glove was deposited by the Camille flood in 1969. The two lower units have paleosols developed on them and erosional upper boundaries.
(From R. C. Kochel, Holocene Debris Flows in Central Virginia, in "Debris Flows/Avalanches: Process, Recognition, Mitigation," *Reviews in Engineering Geology*, vol. 7, p. 148; 1987.)
(B): (R. Craig Kochel)

remote regions (Michaels 1985). As population continues to expand into more remote regions of the Appalachians, damage like that caused by the Camille storm in 1969 may become increasingly common. Detailed histories of debris flow frequency and magnitude have also been reconstructed by studying debris flow-induced damage in tree rings. Hupp (1984) used a dendrogeomorphic approach at Mount Shasta, California, to elucidate a 300-year record of debris flow activity. Bowers et al. (1997) were able to determine debris flow history in the Grand Canyon using ecological aspects of plant colonization on debris flow surfaces.

Progress is being made in identifying threshold intensities required to destabilize slopes in selected regions of rainfall and snowmelt (Campbell 1975; Caine 1980; Church and Miles 1987; Wieczorek 1987; Takabatake et al. 1998). Recent studies have clarified the relationships between hillslope fires and subsequent debris flows (Parrett 1985) and indentified a wide range of mechanisms that may ultimately produce debris flows