Sediment and Wood Routing in Steep Headwater Streams: An Overview of Geomorphic Processes and their Topographic Signatures

Christine May

Abstract: Headwater streams in steep terrain pose a significant challenge for the development of best management practices (BMP) in forested watersheds. There is an incomplete understanding of the processes that govern the input, storage, and transport of sediment and wood, and these processes differ geographically. At the upstream extent of the channel network, headwater streams represent a transition from hillslope to channel processes. At the downstream extent, many of these channels transition from mass wasting to fluvial process dominance. Large-scale sediment routing processes typically consist of debris flows, earth flows, and/or gully erosion. In the interval between episodic transport events, headwater streams can interrupt the delivery of sediment from hillslopes to larger river systems by storing large volumes of sediment and wood. Forest management guidelines typically identify headwater streams as occurring upstream of the distribution of fish; however, a topographically based designation of headwater streams could prove useful for identifying which erosional processes are dominant in a particular area. Broad-scale terrain analysis based on unique topographic signatures can be used to identify the spatial domain of different geomorphic landforms that govern headwater stream processes, how they may be affected by forest management, and to infer the type and severity of downstream disturbance. FOR. SCI. 53(2):119–130.

Keywords: headwater streams, watershed processes, erosion, sediment, large woody debris

HEADWATER STREAMS represent an important transition in physical processes, morphologic characteristics, ecological communities, and management practices. At the upstream extent of the channel network, headwater streams represent a transition from hillslope to channel processes. From a geomorphic perspective the channel head is the first expression of convergent transport of water and sediment. At the downstream extent, many headwater streams in steep terrain transition from mass wasting to fluvial process dominance. In some catchments this transition is abrupt, and mass wasting events terminate in discrete fans or massive log jams. In contrast, mainstem river channels in other areas are steep enough to maintain transport and mass wasting events can transition into sediment-laden "debris floods" and reorganize long distances of channel downstream. Within headwater systems, channel and valley morphology is strongly influenced by roughness elements such as boulders, bedrock outcrops, wood, and small organic debris accumulations, which impede sediment transport and increase the storage capacity of the system.

Morphologically, headwater streams differ from larger alluvial channels in many respects. Headwater streams are typically dominated by bed material and valley fill composed of unsorted, matrix-supported colluvial substrate (Montgomery and Buffington 1997), exhibit an increase in wood abundance that varies inversely with stream size (Bilby and Ward 1989, Hassan et al. 2005a), display increasing stream power and downstream coarsening (Brummer and Montgomery 2003), and do not exhibit a corresponding increase in channel slope with a decrease in drainage area (Stock and Dietrich 2003). Headwater streams often coincide with alluvial channels in the scaling of channel width with drainage area (Brummer and Montgomery 2003), dependence on wood for sediment storage in steep reaches (May and Gresswell 2003a), and armoring of the surface layer (Brummer and Montgomery 2003).

Because of the many unique attributes of headwater streams, their abundance across the landscape, their sensitivity to disturbance, and downstream impacts on fish habitat and water quality, they have become an important consideration for watershed management. Small streams (first- and second-order; Strahler 1964) are the most expansive portion of the channel network and can occupy 60 to 80% of the cumulative channel length in mountainous terrain (Schumm 1956, Shreve 1969). Thus, they are difficult to avoid when planning silvicultural treatments and constructing forest roads. In addition to their abundance across the landscape, the degree of impact on headwater streams from forest management practices differs from larger rivers. For example, the percentage of a large watershed harvested is typically much lower than the total percentage of a small watershed harvested, and there are more small stream crossings per mile. Recognition of headwater streams as important sources of sediment and wood to downstream resources has led to calls for management actions to protect and preserve headwater stream functions (e.g., Independent Multidisciplinary Science Team 1999, Prichard 1998,

Christine May, PhD, Department of Geology and Environmental Science, James Madison University, Memorial Hall, 7100, Harrisonburg, VA 22807—Phone: (540) 568-3339; Fax: (540) 568-8058; maycl@jmu.edu.

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USDA 1994). Concomitantly, geomorphic models are increasingly being used to support public policy and natural resources management (Wilcock et al. 2003). This interest in headwater streams has recently prompted the publication of several articles that summarize headwater stream processes (Gomi et al. 2002), geomorphic characteristics (Benda et al. 2005), sediment transport and channel morphology (Hassan et al. 2005b), wood dynamics (Hassan et al. 2005a), and suspended sediment fluxes (Gomi et al. 2005).

Forest management practices in headwater streams differ from larger streams and have variable guidelines for riparian forest buffers, green tree retention, and road crossings. From a management perspective, headwater streams are often designated on the presence or absence of fish. Although the transition from fish to nonfish baring streams represents an important ecological change at the downstream extent of headwater streams, it rarely corresponds to the position where important physical processes change. At the upstream extent, the boundary of headwater streams has also been difficult to delineate in a way that is both consistent and practical. Channel heads represent a shift in sediment transport processes and are typically associated with the initiation of erosion. The critical contributing area, or threshold that defines the minimum drainage area required to initiate a channel, varies by initiation process. These initiation processes include overland flow, seepage, piping, and landsliding (Montgomery and Foufoula-Georgiou 1993). Models for channel initiation by overland flow and shallow landsliding predict inverse relationships between critical contributing area and local slope; however, the exact topographic position of channel heads can be difficult to identify on low-resolution topographic data (Montgomery and Foufoula-Georgiou 1993).

Terrain-based analysis using digital topographic data is an important tool for delineating and understanding headwater streams because it provides a consistent, accurate, and low-cost assessment that can be used over a broad spatial. In this article a synthesis of recent literature is presented that draws primarily from examples in the Pacific Northwest. From this information a framework is proposed for designating steep headwater streams based on topographic signatures that identify the spatial variation in dominant erosional processes. Specific objectives of the article to (1) identify and review large-scale geomorphic processes that route material through headwater streams; (2) describe terrain-based approaches for identifying geomorphic process domains; and (3) discuss how river profiles can be interpreted and used to infer the spatial extent of headwater stream processes and their downstream consequences.

Geomorphic Processes in Headwater Streams

The erosion and transport of materials through headwater streams is primarily a function of the amount and intensity of precipitation, the texture and depth of soil, the steepness of the slope, and the type and density of vegetation. Therefore, climate, geology, uplift rates, and silvicultural practices provide important controls on processes and rates of sediment delivery to stream channels. Many erosional processes occur in specific landscape positions, which express unique topographic signatures. These topographic signatures can be used to identify process domains, which can be defined as spatially identifiable areas characterized by distinct suites of geomorphic processes (Montgomery 1999). Delineating process domains is particularly useful for forest management because it provides a way of systematically identifying structurally and functionally similar areas. Three dominant processes for the large-scale routing of sediment and wood in headwater streams are (1) earth flows, (2) gully erosion, and (3) debris flows. Some headwater streams could have been formed under different climatic conditions and therefore represent "relic" channels that undergo no modern-day mass wasting processes; however, little is known about their current state or spatial domain.

Earth Flows

Earth flows are large, deep-seated landslides that have complex forms of movement, including block gliding, slumping, and viscous flowing. Headwater streams typically bound either side of an earth flow or form an axial channel that drains the centerline. A large valley bottom channel often flows at the base of the feature and receives a direct sediment feed from the toe of the earth flow and from channels draining the earth flow itself. Rills and gullies often form on active earth flow surfaces but the drainage pattern may be constantly changing due to continued surficial movement (Kelsey 1978).

Along with complex forms of movement, earth flows also exhibit a variety of failure rates. Some are dormant for long periods, others experience prolonged and slow movement during the rainy season, while others have large pulses of movement that occur episodically during extremely large storm events (Kelsey 1978, Swanston and Swanson 1976). In some areas, debris flows can preferentially initiate from the toes or margins of earth flows (Reid et al. 2003). Root reinforcement that is lost by vegetation removal may have limited effects on earth flow movement because the failure plane is typically below the rooting depth (Swanston and Swanson 1976). However, forest management activities may accelerate the rate of movement if surface or subsurface hydrologic characteristics of the site are altered. Although forest management effects have not been well documented, there is some evidence to suggest that the loss of canopy interception and evapotranspiration can increase the soil water content and lead to elevated pore pressures, which reduces shear resistance and increases slippage rates (Swanston 1981, Miller and Sias 1998). Road drainages can also divert additional water onto the site, and further accelerate movement (Swanston and Swanson 1976). Because movement rates respond to increases in pore pressure, a substantial increase in the quantity of water delivered to an active earth flow site can increase the rate of movement and sediment delivery to downstream areas.

In addition to the direct effects of precipitation, earth flow movement is also effected by changes in the stress distribution within the soil mass. During flood events, the toe of the earth flow can protrude into a mainstem river and experience substantial erosion (Kelsey 1978). By eroding the toe of the earth flow the distribution of mass above the failure plane changes and the balance of forces is altered, which may lead to pulses of abnormally large displacement during flood events (Kelsey 1978, Miller and Sias 1998, Hungr et al. 2001). Road maintenance that carves away the toe of an earthflow can have a similar destabilizing effect (Swanston and Swanson 1976).

Earth flows can affect downstream areas by impinging on mainstem river channels, reducing valley width, and diverting channels around deposits. In some cases earth flows can completely block mainstem river channels and produce landslide-dammed lakes. Long-term impingement or blockage can cause backwater alluviation and sediment trapping, which forces the development of broad flat areas upstream of the earth flow and creates large-scale steps in the river profile (Korup 2005). Additionally, mainstem rivers below an active earth flow are often armored with large blocks, which prevents channel incision and results in unusually steep stream reaches adjacent to the toe (Kelsey 1978).

Because earth flows typically form in thick, cohesive soils (Hungr et al. 2001), these landforms and the headwater streams that drain them can deliver large amounts of fine sediment to downstream areas. Of particular concern are clay rich and highly sheared materials from the slip surface of earth flows. In the Cascade Mountains of Oregon earth flows are the primary source of expandable clays such as smectite (Ambers 2001, Hulse et al. 2002). These very fine particles, which have low settling velocities, are the cause of elevated and persistent turbidity levels in downstream areas. Extended periods of elevated turbidity levels caused by these fine clays can have detrimental effects on downstream habitat and water supply reservoirs (Ambers 2001, Hulse et al. 2002).

Recognizable terrain features, such as head scarps with distinct bowl-shaped depressions, flat benches, and hummocky terrain create a unique topographic signature for earth flows. This topographic signature provides a way of identifying areas prone to earth flow activity over a broad spatial area using digital elevation model (DEM)-based analysis. For example, McKean and Roering (2004) used high-resolution topographic data to distinguish earth flows by contrasting their roughness and surface texture. In the Oregon Coast Range, Roering et al. (2005) used a terrain-based algorithm developed from the relationship between hillslope curvature and gradient to distinguish large, deep-seated landslides from shallow landslide and debris flow terrain. Areas prone to deep-seated landsliding exhibited lower drainage density, gradient, and hillslope curvature. Based on this unique topographic signature the authors were able to develop an automated algorithm that allows for the identification and mapping of deep-seated landslides over a broad spatial area. They observed that the portion of the landscape affected by deep-seated landslides ranged from 5 to 25%, which varied systematically with geologic controls based on sedimentary facies and bedrock structure (Roering et al. 2005).

Gully Erosion

Rill and gully erosion is caused by excess surface runoff when rainfall exceeds infiltration. Gully erosion is most common when vegetation is sparse and soils are inherently impermeable, as in many arid or semi-arid regions. However, gullies can form in humid forested terrain, especially in episodic events following high-severity wildfire. If a water-repellent soil layer forms during a severe wildfire, infiltration is reduced and surface runoff increases (Letey 2001). Vegetation disturbance also reduces canopy interception and surface roughness, which can further increase runoff and accelerate gully erosion. This increase in surface runoff can lead to gully formation into previously ungullied hillslopes or further incision into existing gullied channels.

Numerous studies have documented the progressive entrainment or "bulking" of sediment from numerous rills and incising gullies in upslope areas that have recently experienced severe wildfire (Johnson 1984, Wells 1987, Wohl and Pearthree 1991, Meyer and Wells 1997, Cannon et al. 1998, 2001a,b, Meyer et al. 2001). On these disturbed sites, extensive hillslope erosion is often observed following high-intensity rainfall events, indicating that a significant proportion of material is derived from the uplands (Cannon et al. 2001a,b). This sediment-laden flow is then routed through the channel network, where it continues to entrain sediment and can lead to debris flows or hyperconcentrated flows. Some landscapes experience a two-phase erosional response to fire, where "runoff-induced" sediment pulses (i.e., rills and gully erosion) occur before the re-growth of herbaceous vegetation and development of a litter layer, and before water-repellent soil conditions are ameliorated (typically within ~5 years following a severe fire; Meyer et al. 2001). After this time period, infiltration rates typically increase to the extent that "saturation-induced" slope failures, such as shallow landsliding, can occur (typically 5–20 years post-fire; Meyer et al. 2001). Concomitant with this time period is the loss of root strength that the previous forest stand provided and the lag before major root systems from the regenerating forest have become established (Ziemer 1981). Although fire suppression and land use can affect the frequency and magnitude of fire-related sedimentation, large-scale climatic conditions can be major drivers of these events (Meyer and Pierce 2003).

Drainage density of the channel network in gullied terrain is highly variable in both space and time, as the channel network expands and contracts with episodes of gully development and infilling. Under some soil and rainfall conditions gullies can extend nearly to the ridge, and thus require a very small contributing area on previously ungullied hillslopes. Modeling by Istanbulluoglu et al. (2004) suggests that drainage density is inversely proportional to root cohesion, resulting in an increased density of gullied channels due to a reduction in forest cover. Gullies can also have high temporal variation, ranging from transient features to very persistent. Once established, gullies can become chronic sources of fine sediment for many years (Nistor and Church 2005). Fine sediment fluxes and turbidity levels to downstream areas are of particular concern because gullies tend to form in cohesive, clay rich soils.

Large wood and tree roots can be particularly important for halting the headwater migration of gullies and for trapping sediment in gullied channels (Dewey et al. 2002). Timber harvest can affect gully erosion by removing sources of in-stream wood, causing a loss of root strength, and reducing canopy interception. Concentrated runoff on bare ground and from forest roads can also lead the development of gullies on previously ungullied hillslopes and/or the transport of sediment from adjacent timber harvest sites through riparian leave areas and into larger stream channels (Rivenbark and Jackson 2004). Many headwater streams are still affected by the legacy of past forest management practices that removed in-stream wood and caused wide-spread ground disturbance, including severe compaction.

Because of the high spatial and temporal variability in gully development, there is no topographic algorithm available for broad scale identification of areas prone to this type of erosion. Based on the concept that channels initiate with smaller drainage areas on steeper slopes (Montgomery and Dietrich 1988, Montgomery and Foufoula-Georgiou 1993), Cannon et al. (2001b) empirically defined the relationship between slope and contributing area in different lithologies for debris flow generation by progressive sediment entrainment from rills and gullies. More complex, probabilistic approaches for channel initiation that depend on slope, catchment area, and the probability distributions of median grain size, surface roughness, and excess rainfall rate have also been developed (Istanbulluoglu et al. 2002).

Debris Flows

Debris flows are the most well documented, intensively researched mass transport process in steep terrain. This process is widely recognized as one of the dominant sediment routing mechanisms in steep mountainous areas (e.g., Dietrich and Dunne 1978, Benda and Dunne 1997a, Eaton et al. 2003), and can pose a significant hazard to downstream resources and infrastructure. In an extensive investigation of steep mountainous terrain, Stock and Dietrich (2003) used an international data set to estimate that >80% of channel networks in large unglaciated catchments can be composed of channels susceptible to debris flows.

Debris flows are channelized mass movements that rapidly mobilize material stored in headwater streams. Hydrologic triggers for debris flow initiation include localized groundwater inflow, prolonged moderate-intensity rainfall, and high-intensity rainfall (Reid et al. 1988). Debris flows can be initiated in a variety of topographic positions (Johnson 1984, Costa 1984), including small, shallow landslides in bedrock hollows (e.g., Dietrich and Dunne 1978, Reneau and Dietrich 1987), landslides on planar hillslopes (e.g., May 1998, Robison et al. 1999), along the margins or toes of active earthflows (e.g., Swanson and Swanston 1977, de la Fuente and Haessig 1993, Reid et al. 2003), and from the progressive bulking of sediment entrained from surface erosion combined with in-channel sources (e.g., Cannon et al. 1998, Meyer et al. 2001). However, the most common and well studied process for debris flow initiation in the Pacific Northwest is shallow landsliding in bedrock hollows (Montgomery et al. 2000). Bedrock hollows are topographic depressions expressed as unchanneled valleys where downslope soil and water transport converges and sediment accumulates (Dietrich and Dunne 1978). Soil accumulates in these topographic hollows for centuries and is episodically evacuated by shallow landsliding. Shallow landsliding commonly occurs in steep soil mantled landscapes and results in deeply dissected terrain with a high density of headwater streams whose valleys are carved by debris flows.

Shallow landslides typically form in thin and poorly cohesive soils. Tree roots increase slope stability of shallow soils by the cohesion provided by root fiber reinforcement. After vegetation removal, it takes several years for the root system of the previous forest stand to decay (Ziemer 1981). A minima in rooting strength occurs after the decay of old roots and before roots from the regenerating forest become established (Ziemer 1981). Removal of the forest canopy can also affect slope stability by altering the timing and intensity of precipitation reaching the ground surface (Keim and Skaugset 2003). Forest management activities have accelerated the frequency of shallow landsliding in many areas (e.g., Swanson et al. 1981, Sidle et al. 1985, de la Fuente and Haessig 1993, Jakob 2000, Montgomery et al. 2000, Brardinoni et al. 2002, Guthrie 2002). In some cases, the indirect legacies of forest disturbance can last for decades if the species composition, age class, or spacing of trees on unstable slopes is effected. These small-scale spatial patterns of root strength, which are not apparent in broad-scale assessments of stand age; makes some sites more susceptible to landsliding due to patches of low root cohesion (Schmidt et al. 2001, Roering et al. 2003). Roads, especially those with large sidecast and fill volumes, can also lead to an increased volume and accelerated rate of landsliding (Swanson et al. 1981, Montgomery et al. 1998, Wemple et al. 2001, May 2002).

In the interval between debris flows, headwater streams can accumulate large volumes of sediment and wood. During this interval, the export of fine sediment is influenced by the proximity of timber harvest to the stream, the detachment and routing of material through the road network, and the accumulation of smallwood in the channel (Gomi et al. 2005). The export of coarse sediment is affected by the presence of large wood and landslide deposits that overwhelm the transport capacity of the channel (May and Gresswell 2003a, Lancaster et al. 2003). In the absence of wood, steep headwater streams may lack the ability to store sediment and become chronic sources of sediment to downstream areas. When an adequate supply of wood is present, headwater streams develop a stepped profile that increases the storage capacity of the channel (Gomi et al. 2001, Lancaster et al. 2003, May and Gresswell 2003a). In addition to storing sediment, wood also plays an important role in debris flow dynamics. Velocity reduction due to the entrainment and transport of wood in the runout path extracts momentum and has the potential to decrease the travel distance of debris flows (Lancaster et al. 2003). Wood that is transported by debris flows can also provide an important input of wood to larger, mainstem channels (May 2002, May and Gresswell 2003b, Reeves et al. 2003). However, landslide and debris flow sources of wood may increase in their relative contribution to the overall wood loading in basins intensively managed for timber production because of accelerated mass wasting, removal of streamside forests, and the legacy of wood removal from channels (May and Gresswell 2003b, Montgomery et al. 2003).

In addition to routing wood, debris flows are one of the dominant processes for routing sediment through headwater streams and can be an important source for coarse sediment to downstream areas. Sediment delivered to larger rivers is typically stored in fans when there is adequate accommodation space in valley bottoms (Meyer et al. 2001, Benda et al. 2003, May and Gresswell 2004). In narrow valleys debris flows usually form massive log jams that effectively dam the channel and force deposition of a large wedge of sediment upstream (Hogan et al. 1998, Lancaster et al. 2003, May and Lee 2004). In steep and tightly confined river canyons, debris flows may continue to travel down mainstem rivers and cause substantial reorganization of the channel for long distances downstream (Benda 1985, Cenderelli and Kite 1998, Miller and Benda 2000, Miller et al. 2003).

Terrain-Based Analyses

An understanding of geomorphic process domains and their unique topographic signatures allows for their identification and mapping across broad spatial areas. Process-based mapping and modeling using digital topographic data (primarily 10-m resolution DEMs) is particularly useful for identifying areas that may be sensitive to different types of impacts from forest management activities and for hazard assessments. Some landforms, particularly topographic hollows and earth flows, occupy a very small proportion of the total area of a catchment but they can mobilize the majority of sediment that is delivered to the channel network and thus require careful identification and mapping. Terrain analysis is also useful for developing and testing hypotheses and for improving our understanding of spatial variation in landscape processes.

Several tools are currently available that facilitate terrain analysis of headwater stream processes. The algorithm developed by Roering et al. (2005) in the Oregon Coast Range, based on the relationship between hillslope curvature and gradient, can be used to differentiate between large, deep-seated landslides and debris flow terrain associated with shallow landsliding. Within debris flow prone areas of the landscape several other tools are available. Sites prone to shallow landsliding from bedrock hollows can be identified by topographic convergence which concentrates water and sediment (Dietrich and Dunne 1978). Because of the surface expression of hollows, topography can be used to locate and map areas susceptible to landsliding based on local slope and hillslope convergence (Montgomery and Dietrich 1994, Dietrich et al. 1995, Montgomery et al. 1998).

Downstream of landslide source areas, digital topographic data have been used in a variety of ways to predict debris flow runout. Simple empirical models based on channel slope, network geometry, and valley confinement are commonly used in the Pacific Northwest (Benda and Cundy 1990, Fannin and Wise 2001). Linked modeling of topographic attributes for the initiation, transport, and deposition zones of debris flows has also been developed and tested in western Oregon (Hofmeister and Miller 2003). More sophisticated modeling approaches are also available (Benda and Dunne 1997a,b, Gabet and Dunne 2003, Lancaster et al. 2003, Istanbulluoglu et al. 2004); however, these models are difficult to parameterize and validate.

One of the biggest knowledge gaps for terrain-based mapping of process domains is the inability to predict areas prone to gully erosion. The extent of gullied channels is also difficult to identify because 10-m resolution DEMs are too coarse to distinguish these narrow and incised channels. Airborne laser swath mapping (also known as LIDAR), which produces high-resolution topographic data (e.g., 2-m resolution DEMs with <10 cm vertical accuracy), is not readily available in most areas; however, data from Casper Creek experimental forest in northern California provides promising results for mapping the extent of gullied channels under dense forest cover (T. Lisle, pers. comm., US Forest Service, Pacific Southwest Research Station, 2005). As with all terrain-based mapping, predictions of the type and extent of headwater stream processes are highly dependant on the resolution of the topographic data, and predictions should be verified with local field data.

Another limitation of terrain-based mapping is the lack of information on headwater streams that may not be affected by mass wasting or gully erosion. Although their occurrence is rare in steep terrain, some headwater streams may have been formed under different climatic conditions and therefore do not undergo any modern-day mass transport processes. Research is needed to identify the conditions that formed these channels, describe their current morphology, and predict their spatial extent. Similarly, basins with low relief may not be governed by mass wasting or episodic gully incision.

Topographic Indexes in Debris Flow Terrain

A process-based approach to river profile analysis, based on the relationship between drainage area (A) and channel slope (S), has long been recognized by geomorphologists (Hack 1957). This relationship takes the form of a power function, where the coefficient (K_S) represents the "steepness index" and the exponent (Θ) represents the "concavity index."

$$S = K_S A^{-\Theta}.$$

The steepness index (K_S) characterizes the overall relief of the river profile. K_S values reported in this article have been normalized to a representative drainage area (Sklar and Dietrich 1998) of 1 km^2. Therefore, the reported values represent the characteristic slope for a stream reach that drains a 1 km^2 catchment. The concavity index (Θ) represents the shape of the profile and characterizes how abruptly or gradually the transition is from steep headwater streams to larger lowland rivers (Figure 1). High values of the concavity index represent channels that have strongly concave profiles, where steep channels abruptly grade to low-gradient river valleys. Low values represent river profiles that are poorly concave and have a very gradual transition from steep to low-gradient areas. In these river systems,

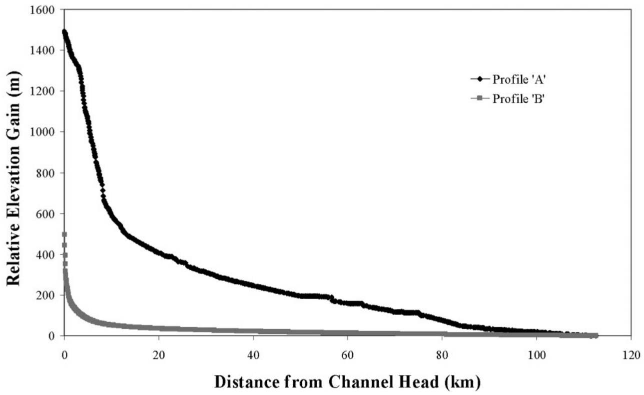

Figure 1. Examples of a river profile with high steepness and low concavity (A), and a profile with low steepness and high concavity (B).

headwater streams extend further downstream and low gradient stream reaches are less abundant.

The characteristic form of the area-slope relation displays a distinct curve in log-log space, above which there is no corresponding increase in slope with a decrease in drainage area (Montgomery and Foufoula-Georgiou 1993) (Figure 2). Stock and Dietrich (2003) provide compelling evidence that this scaling break represents a shift from fluvial process dominance in larger channels to debris flow dominance in steep headwater channels. Debris flows rarely travel down channels $<$ 3 to 10% slope, which closely corresponds to the scaling break observed in slope-area data (Stock and Dietrich 2003). The distinction between fluvial and debris flow valley incision is not simply an academic debate about a small portion of the landscape (Dietrich et al. 2003). Channels and valleys formed by debris flow incision are both extensive in length ($>$80% of large steepland basins) and comprise large fractions of the mainstem valley relief (25–100%) in steep mountainous terrain (Stock and Dietrich 2003).

Because unglaciated headwater valleys that are carved by debris flows exhibit this unique topographic signature, the inflection in the area-slope plot provides a way to objectively define the spatial extent of headwater streams in steep terrain. Channel networks derived from 10-m resolution DEMs are available in many areas of the Pacific Northwest and can be used to calculate the steepness and concavity indexes, which can be measured directly by regression analysis. It should be noted that slope-area data have considerable scatter and estimates of channel slope derived from moderate-resolution topographic data can be somewhat imprecise; however, high-resolution topographic data are rarely available for large areas (Finlayson and Montgomery 2003).

Values of the steepness and concavity index can vary both locally and regionally. Spatial variation has been attributed to differential uplift rates (Snyder et al. 2000, Kirby and Whipple 2001, Kirby et al. 2003, Kobor and Roering 2004), precipitation gradients (Roe et al. 2002), and variation in rock strength and sediment supply (Sklar and Dietrich 1998, 2001). To assess regional differences in river profiles we calculated the steepness and concavity values for nine river basins in three different geographic regions (Figure 3). Data to develop these plots used DEM-derived stream layers created by Miller (2003). Because the regression of slope-area data is only intended to represent the fluvial process domain, it was necessary to isolate that portion of the data. Headwater stream reaches that exceeded 10% slope were assumed to reside in the debris flow process domain and were omitted from the analysis; however, this criterion results in a conservative estimate of debris flow channels compared to the 3 to 10% range observed by Stock and Dietrich (2003).

Although variation occurs within each region, Figure 3 reveals large-scale variation among regions. Streams in the Oregon Coast Range have the lowest steepness values and are strongly concave. In a detailed analysis within the sandstone lithology of this region, Kobor and Roering (2004) report that steepness values ranged from 0.015 to 0.075 and concavity values ranged from 0.35 to 1.05. In our analysis, Knowles Creek and Wassan Creek in the central Oregon Coast Range are underlain by marine sandstone, while the steeper and less concave river profile of the Tillamook River is underlain primarily by volcanic rocks. Differences in channel gradient, fish habitat, and the distribution and abundance of salmonid fishes in sandstone and basalt drainages (Hicks and Hall 2003) may largely be attributable to variation the steepness and concavity of the river profile.

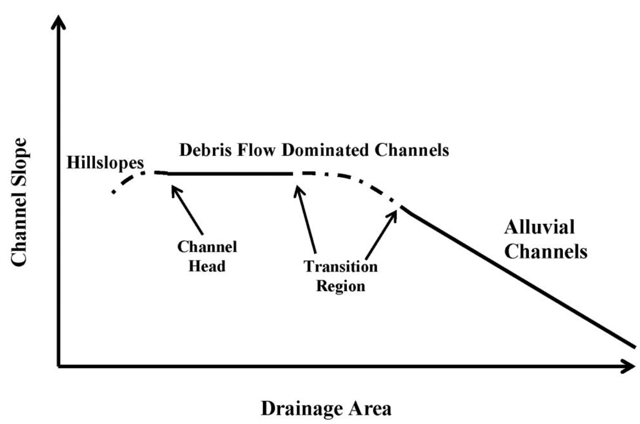

Figure 2. Schematic illustration of the topographic relationship between drainage area and channel slope (modified from Montgomery and Foufoula-Georgiou 1993). Exact slope-area values for each transition will vary based on the steepness and concavity of the profile.

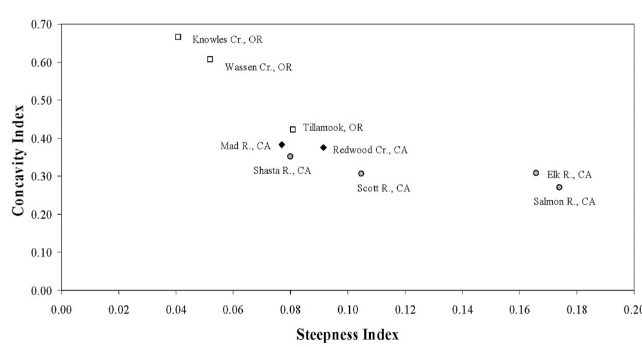

Figure 3. Regional differences in steepness (K_S) and concavity (Θ) values from river profile analysis of selected river basins in California and Oregon. Open squares represent rivers in the Oregon Coast Range, black diamonds represent the California Coast Range, and shaded circles represent rivers in the Klamath Mountains of northern California and southern Oregon.

Streams in the northern California Coast Range are underlain by schists and extensively sheared and altered sandstones and siltstones, and display intermediate values of steepness and concavity. Streams in the Klamath Mountains of northern California had the steepest and least concave profiles. This region has a diverse geology setting, ranging from easily erodable areas of decomposed granite to areas of highly resistant metamorphic rocks.

Understanding the spatial variation in steepness and concavity indexes is important because river profiles provide topographic controls on the spatial extent of headwater stream processes, the runout potential of debris flows, and their downstream consequences. For example, the ratio of steepness (K_S) to concavity (Θ) explains 53% of the variation in the abundance of headwater streams observed in 22 river basins (average drainage area = 340 km^2) in northern California and southern Oregon (Figure 4).

The shape of the river profile can also be used to assess the runout potential of debris flows and how it varies across the landscape. Debris flows rarely travel down channels that are <3 to 10% slope; however, debris flows can combine with flood flows in mainstem river channels to produce "hyperconcentrated" or "debris floods" (Costa 1984, Benda 1985, Hungr et al. 2001). In steep basins with poorly concave profiles, the transition from debris flows to debris floods is more likely because mainstem river channels are often steep enough to continue transporting the mass flow, especially in tightly confined river canyons. For example, California's Salmon River represents an extreme end-member of high steepness (0.17) and low concavity (0.27) in Figure 3. Based on this slope-area relationship the characteristic drainage area where the channel reaches a consistent slope of <6.5% (the critical channel slope identified for debris flow deposition by Benda and Cundy (1990)) is 36 km^2. In contrast, Knowles Creek has very low steepness (0.04) and high concavity (0.66), indicating that the majority of relief in the river profile occurs in steep headwater streams that abruptly transition into low-gradient mainstem rivers. Based on this slope-area relationship the characteristic drainage area where the channel reaches a consistent slope of <6.5% is only 0.5 km^2 in Knowles Creek. In

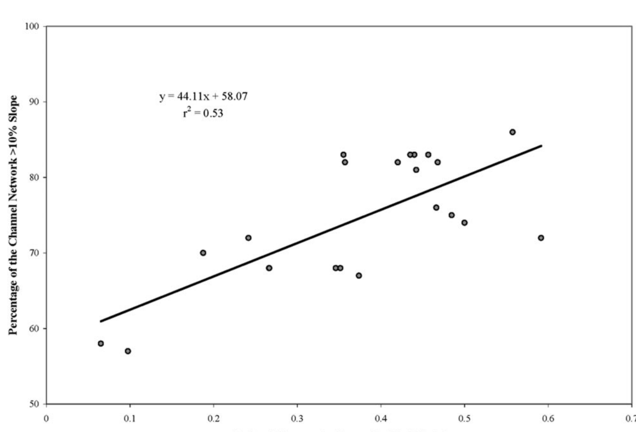

Figure 4. Abundance of headwater streams (defined as channels > 10% slope) based on the ratio of steepness (K_S) to concavity (Θ) values from river profile analysis of 22 catchments in northern California and southern Oregon (average basin size = 340 km^2).

mainstem rivers, such as the Salmon River, with high steepness and low concavity, debris flow runout can be maintained through much more of the channel network and slope of the mainstem river provides a first-order control on debris flow deposition.

The downstream impacts of long runout debris flows and debris floods can reorganize long distances of mainstem river channels, topple or bury riparian vegetation across the entire width of the valley floor, and mantle the streambed with coarse particles that may be difficult to mobilize by fluvial transport. For example, Miller and Benda (2000) documented the burial of riparian vegetation and a coarse bed inset between terraces cut by numerous shallow side channels following the passage of debris flow sediments that traveled down an alluvial river channel. Large volumes of wood delivered by debris flows have also been found to play a particularly important role in the downstream disturbance. These large batches of wood travel in a congested mode of transport, where logs move together as a single mass and occupy a large portion of the channel and/or valley floor (Braudrick et al. 1997) and can topple riparian forests for long distances downstream (Johnson et al. 2000). In contrast, in basins that have low steepness value and/or strongly concave profiles, the transition from channels that are scoured by debris flows to alluvial channels is abrupt and debris flows typically end in discrete deposits. Debris flow deposition in basins with low-gradient mainstem channels is primarily governed by tributary junctions (Benda and Cundy 1990, May and Gresswell 2004), and these deposits typically take the form of massive log jams in channels or fans at confluences (Figure 5A). Where these discrete deposits form the disturbance is patchy (Benda et al. 2004), and undisturbed areas can act as refuges. In contrast, long runout debris flows and debris floods can impact tens of kilometers of channel, resulting in a more homogenous disturbance pattern (Figure 5B). This pattern was observed in extensive mapping of debris flows in the Klamath mountains of northern California (Mondry 2004, J. de la Fuente, pers. comm., USDA, Klamath National Forest, 2005). Because the disturbance is widespread, there is limited refugia for aquatic organisms. In the few studies that have investigated the direct effects of debris floods that travel through fish-bearing streams, local extirpations of salmonids have been observed (Lamberti et al. 1991, Roghair et al. 2002, B. Harvey, pers. comm., US Forest Service, Pacific Southwest Research Station, 2005).

Knowledge Gaps

The availability of high-resolution topographic data and its use in watershed assessment was identified by Dunne (1998) as one of the critical data requirements for predicting erosion and sedimentation in mountainous terrain. Small-scale features, such as bedrock hollows and gullies, are particularly difficult to identify in mature forests and on steep slopes. Laser swath mapping, also known as LIDAR, produces high-resolution topographic data even beneath forest cover. As this technology becomes more widely available and less expensive, it will likely have greater utility for increasing the accuracy of terrain analysis in the future

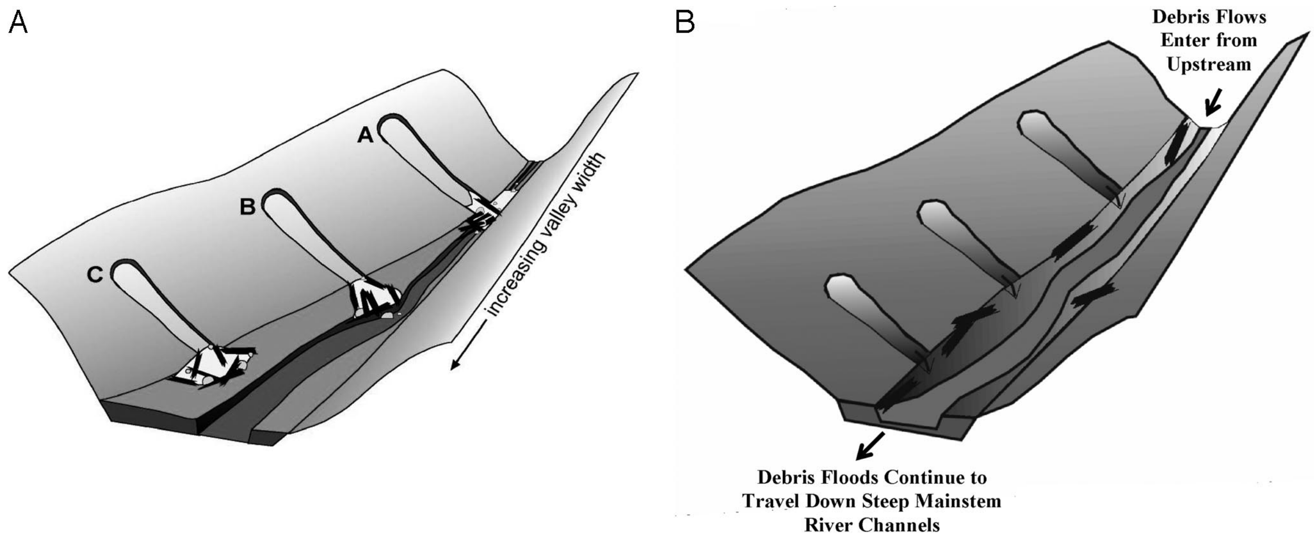

Figure 5. Patchy disturbance pattern formed by discrete debris flow deposits at tributary junctions for basins with low steepness and high concavity (A). Homogenous disturbance pattern formed by long runout debris flows and debris floods in basins with high steepness and low concavity (B).

(Carter et al. 2001). For example, LIDAR can be used to map gulley incision and will improve the predictive capabilities of other landforms such as bedrock hollows. High-resolution topography also shows great promise toward discovering new topographic signatures that will facilitate the mapping of other geomorphic landforms.

A preponderance of the literature on headwater stream processes and topographic signatures originates in the Pacific Northwest. Most small, steep streams in this region are formed and maintained by mass wasting. Gully erosion is less common and often associated with fire. A broader understanding of headwater processes in other regions is needed, especially research targeted at understanding the morphology, spatial extent, and context of channels that may not undergo any modern-day mass transport process. Similarly, areas of low relief may not be governed by mass wasting or gully erosion, and topographic signatures may not be applicable. Although topographic signatures have not been identified for all processes, nor are they applicable in all areas, they are useful for identifying major sediment sources that can be problematic for forest management.

At the present time there is limited ability to infer patterns and processes of wood delivery to streams from terrain-based mapping. Probabilistic and physically based models have been developed that provide insight into patterns of wood abundance in streams over long time periods (Benda and Dunne 1997a, b, Lancaster et al. 2003); however, these complex models are difficult to parameterize and validate. Future efforts may be capable of predicting wood delivery to stream channels with LIDAR, which has the capability of mapping the ground surface and the tree canopy.

Conclusions

Because of the numerous ways that physical and ecological processes in headwater streams vary longitudinally in the channel network and across the landscape, there is a compelling need to identify and manage headwater streams based on different geomorphic process domains (Montgomery 1999). Sediment yield from headwater streams can be extremely episodic (i.e., low-frequency, high-magnitude events), driven by debris flows, episodic gulley incision, and pulses of earth flow deformation (e.g., Kirchner et al. 2001, Eaton et al. 2003, Meyer and Pierce 2003, Istanbulluoglu et al. 2004, Ferrier et al. 2005). Therefore, identifying the location of large-scale erosional processes is critical for developing aquatic conservation strategies and planning forest management activities.

This article presents a synthesis of recent advances in identifying and mapping geomorphic landforms using digital terrain analysis, with particular emphasis on applications that can assist forest management. Delineating process domains is particularly useful because it provides a systematic way of identifying structurally and functionally similar areas. In steep mountainous terrain with a thin soil mantle, the spatial extent of headwater streams can be defined by an inflection in the relationship between drainage area and channel slope, which represents the boundary between debris flow and alluvial channels (Montgomery and Foufoula-Georgiou 1993, Stock and Dietrich 2003). Topographic indexes of steepness and concavity derived from the area-slope relationship can be used to infer the severity of debris flow disturbance to downstream areas.

In mountainous terrain with thick cohesive soils where earthflows form, the unique topographic signature of hillslope curvature and gradient provides a means for broad-scale mapping (Roering et al. 2005). Areas prone to episodic gully erosion cannot be predicted solely from topographic signatures and may require a more sophisticated, probabilistic modeling approach (e.g., Istanbulluoglu et al. 2002). At the present time, these probabilistic models are difficult to parameterize and thus have not been applied. Relic channels that formed under different climatic conditions may occur in some areas; however, there is no literature to suggest these channels are a common occurrence. Furthermore, there is overwhelming evidence that the vast majority of headwater streams in steep terrain are affected by episodic transport events (e.g., Dietrich and Dunne 1978, Wohl and Pearthree 1991, Benda and Dunne 1997a, Kirchner et al. 2001, Eaton et al. 2003, Stock and Dietrich 2003, Ferrier et al. 2005, Roering et al. 2005).

DEMs have particular utility for identifying sediment source areas from erosional processes that may be accelerated by forest management activity. The mapping of sediment source areas facilitates silvicultural planning for leave areas, identifies the intersection of roads with unstable slopes, and downstream conditions such as tributary junctions that affect the magnitude and frequency of sediment and wood delivery (Dunne 1998, Benda et al. 2004). By identifying dominant geomorphic processes, monitoring plans can be established that target specific impacts. For example, fine sediment and elevated turbidity levels can be anticipated downstream of gullies and active earth flows. In contrast, the mapping of debris flow paths can identify headwater streams that can contribute both coarse sediment and wood to downstream areas (Benda and Dunne 1997a, May and Gresswell 2003b, Reeves et al. 2003).

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