Influence of Headwater Streams on Downstream Reaches in Forested Areas

Lee H. MacDonald and Drew Coe

Abstract: The source areas of headwater streams typically compose 60% to 80% of a catchment. This, plus the typical increase in precipitation with elevation, means that headwater streams generate most of the streamflow in downstream areas. Headwater streams also provide other important constituents to downstream reaches, including coarse and fine sediment, large woody debris, coarse and fine organic matter, and nutrients. The relative importance of headwater streams as a source of these other constituents is highly variable because the amount and quality of each constituent can be modified by in-channel storage, dilution, biological uptake, diminution, and chemical transformations. Headwater sources of water, fine sediment, and fine particulate organic matter are more likely to be delivered to downstream reaches than coarse sediment, woody debris, nutrients, or an increase in water temperatures. The complexity and temporal variability of channel-hillslope interactions, in-channel processes, and downstream conditions makes it difficult to rigorously link upstream inputs and anthropogenic activities to the condition of downstream resources. These issues may preclude the use of adaptive management, particularly in larger basins, as adaptive management implicitly assumes that (1) downstream changes can rapidly be detected, (2) management will change rapidly in response to any adverse change, and (3) a management change will rapidly improve the affected resource. Since these assumptions may be difficult to satisfy—particularly in larger basins—the use of adaptive management must be carefully examined before it can be applied at the watershed scale. FOR. SCI. 53(2):148–168.

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HEADWATER STREAMS compose the uppermost portions of the stream network. Headwater streams typically represent from 60 to 80% of the total stream length within a catchment (Schumm 1956, Shreve 1969), and they drain 70 to 80% of the total catchment area (Sidle et al. 2000, Meyer and Wallace 2001). The small size of headwater streams means that they are particularly responsive to natural and anthropogenic disturbances such as debris flows, changes in vegetative cover, changes in sediment inputs, and changes in organic matter inputs (Benda et al. 2005, Hassan et al. 2005a, Richardson et al. 2005). The preponderance of headwater streams, when combined with their sensitivity and potential linkages to downstream resources, means that headwater streams are of increasing interest to scientists and resource managers (Whiting and Bradley 1993, Gomi et al. 2002, JAWRA 2005).

By definition, headwater streams begin where surface runoff is sufficiently concentrated to cause scour and distinct banks (Dietrich and Dunne 1993). This surface runoff may occur only during storm events or snowmelt ("ephemeral"), seasonally ("intermittent"), or continuously ("perennial"). Headwater channels can be distinguished from hillslope rills because they are relatively persistent features on the landscape and generally occur in strongly convergent areas. The channel head or the initiation point for headwater channels can migrate in response to large storm events or severe disturbance (Montgomery and Dietrich 1989, Istanbulluoglu et al. 2004), but such changes do not alter the general location of a stream within a landscape and usually do not alter stream order.

The downstream end of headwater channels is more ambiguous, but in this article the lower boundary is defined as the colluvial-alluvial transition point. When the drainage area is 1 km^2 or less, colluvial (hillslope) processes dominate channel form and in-channel fluxes. As the drainage area increases from 1 km^2 to 10 km^2, alluvial processes become an increasingly dominant control on channel morphology and fluxes (Montgomery and Foufoula-Georgiou 1993, Brummer and Montgomery 2003, Stock and Dietrich 2003). Some authors have suggested that a drainage area of 1 km^2 and a channel slope of 20 to 30% defines this colluvial-alluvial transition point (Montgomery and Foufoula-Georgiou 1993, Woods et al. 1995, May and Gresswell 2004), but variations in climate, geology, and other factors means that this transition can occur when the drainage area is as small as 0.1 to 0.3 km^2 (Benda and Dunne 1997a) or as large as 10 km^2 (Madsen 1994, Brummer and Montgomery 2003). For the purposes of this article headwater channels are defined as any channel with a drainage area of up to 10 km^2. In the Pacific Northwest this upper limit corresponds to streams that are up to about 10 m wide (Brummer and Montgomery 2003).

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Given their location in the landscape, the hydrologic, geomorphic, and biological characteristics of headwater streams result from a dynamic mix of colluvial and alluvial processes. In downstream channels the channel conditions and material fluxes are again a complex integration of the fluxes and processes from both upslope and upstream, but the hillslopes adjacent to higher-order channels generally play a much smaller role in terms of streamflow, sediment yields, nutrient fluxes, aquatic productivity, and aquatic biodiversity.

The management of headwater channels is important because they drain the vast majority of the catchment area, and they are a critical source of water, sediment, fine and coarse organic matter, and nutrients (Gomi et al. 2002). They also are the first source of aquatic life as one makes the transition from hillslopes into the channel network. From a regulatory perspective, headwater streams have been largely ignored despite their potential effect on downstream reaches. Towns and cities generally are located adjacent to larger streams and rivers, and laws such as the Clean Water Act initially focused on regulating point sources, maintaining water quality in these larger streams for domestic use, and sustaining fisheries. It was not until 1972 that the Clean Water Act explicitly addressed nonpoint pollution sources, such as hillslope erosion. Similarly, state-level forest practice regulations focused on perennial, fish-bearing streams until the late 1980s or early 1990s.

The wider recognition of the importance of headwater streams is relatively recent (Jackson et al. 2001, Gomi et al. 2002, Halwas and Church 2002, Jackson and Sturm 2002, Benda et al. 2005, Hunter et al. 2005). Relative to higher-order channels, there have been much fewer studies on the conditions and instream dynamics of headwater streams despite their relative predominance in terms of drainage area and total stream length. Although most paired-watershed studies have been conducted at the scale of headwater streams, these typically have focused on the watershed-scale changes in streamflow and sediment yields and treated the watershed as a black box. Few studies have rigorously examined the extent to which headwater streams control downstream conditions, despite the conceptual recognition of headwater-downstream linkages (e.g., Vannote et al. 1980, Benda et al. 2004b). Hence, the purpose of this article is to review existing knowledge with respect to:

1. How well are headwater streams connected to downstream areas in terms of the generation and delivery of discharge, coarse and fine sediment, coarse and fine organic matter, temperature, and nutrients?
2. To what extents do natural disturbances and anthropogenic activities alter the connectivity between hillslopes, headwater streams, and downstream areas?
3. To what extent are downstream conditions controlled by the inputs from headwater streams as compared to colluvial processes from the adjacent hillslopes and the intervening riparian and fluvial processes?

The answers to these three questions are important because they determine the extent to which downstream conditions are affected by management activities in headwater drainages. The strength of the linkages between headwater catchments and downstream conditions are important for justifying the regulation of headwater streams, and for predicting the magnitude and timing of cumulative watershed effects (CWEs).

Similarly, the magnitude and timing of the connectivity between headwater streams and downstream reaches directly affects the extent to which adaptive management can be applied in larger catchments. Adaptive management is increasingly touted as the most practical approach for resource management given the complexities and uncertainties in predicting management impacts at the watershed scale (Stednick et al. 2004). In larger watersheds the successful use of monitoring and adaptive management depends to a large extent on the strength of the linkages between upstream and downstream resources, and the responsiveness of downstream resources to a specified change in upstream management. Hence, the final objective of this article is to assess the validity of the assumptions implicit in the use of adaptive management at the watershed scale, and the relative detectability of anthropogenic effects on downstream channel conditions.

The following sections discuss the extent to which headwater streams are connected to downstream areas with respect to discharge, coarse and fine sediment, coarse and fine organic matter, temperature, and nutrients. Most of the examples are drawn from forested streams from northern California up through British Columbia, as this is where these issues are particularly controversial and much of the recent research has been conducted. The final section addresses the implicit assumptions and constraints on using adaptive management at the watershed scale.

Contribution of Headwater Streams to Downstream Runoff

The contribution of headwater streams to downstream runoff is critical for water supply as well as the transport of sediment, coarse and fine organic matter, and nutrients (Moore and Wondzell 2005). Approximately 95% of the runoff in a channel is generated on hillslopes (Knighton 1998). Since first-order channels compose the vast majority of the drainage network, it follows that headwater streams are usually the primary source of streamflow.

The delivery of water from hillslopes to channels can occur as surface flow, subsurface stormflow, or groundwater (Dunne and Leopold 1978, Novotny and Olem 1994). In headwater catchments the delivery of water to the channel and hence the magnitude and timing of peak flows are controlled primarily by hillslope processes (Dunne 1976). In a 1 km^2 basin the time to peak flow can vary from about 25 min when Horton (infiltration-excess) overland flow is the dominant runoff process, to 1 hour for saturation overland flow, and over 18 hours when subsurface stormflow is the dominant runoff process (Dunne 1976). Horton overland flow is rare in forested areas in the western United States, and water from the hillslopes is delivered to the channel network primarily by subsurface stormflow (McGlynn et al. 2004, McNamara et al. 2005) and secondarily by saturation overland flow.

The amount and timing of hillslope runoff varies as the channel network expands and contracts under different moisture conditions (Hewlett and Nutter 1970, Hunter et al. 2005), and the hillslopes become hydrologically connected or disconnected from the stream network (McNamara et al. 2005). During dry periods, hydrologic connectivity is primarily restricted to wet convergent areas and the riparian zones immediately adjacent to the channel (McGlynn et al. 2004). Subsurface stormflow is only generated under relatively wet conditions, and the delivery of this water to the stream channel is increasingly viewed as threshold-dependent. In humid forested catchments recent studies indicate that subsurface stormflow only occurs when rainfall exceeds about 20 to 55 mm under wet antecedent conditions (Weiler et al. 2005). Hillslopes with shallower soils tend to have lower thresholds (Weiler et al. 2005, Tromp-van Meerveld and McDonnell 2006), and the entire hillslope may not be hydrologically connected to the stream until the deepest soils wet up. This means that even though hillslopes compose most of the drainage area, complete hillslope-stream connectivity may only occur a couple of times per year in drier, rain-dominated forested areas, and only during spring snowmelt in snow-dominated areas (McGlynn and McDonnell 2003, Weiler et al. 2005, McNamara et al. 2005).

The variability and complexity in hillslope-stream connectivity helps explain why headwater streams can have higher peak flows per unit area and greater variability than their downstream counterparts (Figure 1) (Woods et al. 1995, Gomi et al. 2002). The higher peak flows in headwater basins can be attributed to several factors, including the general tendency for precipitation to increase with increasing elevation; the greater potential for an entire basin to be simultaneously affected by an intense rain event; the potential for the entire basin to fall within a narrow elevation band and therefore be subjected to rain or rain-on-snow rather than a mixture of rain and snow; the ability for an entire basin to be simultaneously at or near peak snowmelt rates; and the potential for runoff from different portions of a basin to be synchronized.

As basin size increases, the magnitude and timing of storm runoff are increasingly controlled by the structure and morphology of the drainage network and valley bottoms rather than hillslope-scale runoff processes (Robinson et al. 1995, Gomi et al. 2002, McGlynn et al. 2004). As a hypothetical example, the maximum channel length in a 100 km^2 basin will be on the order of 15 to 20 km. During high flows the mean water velocity should be around 1.5 m s^{-1}, so the average water molecule might spend several hours in the channel before reaching the basin outlet. Depending on the dominant runoff process, the travel time for water in the stream is comparable to the time lag between precipitation (or snowmelt) and the input of water into the stream channel. With increasing basin size the in-channel travel time increases and the hillslope runoff processes become progressively less important in defining the magnitude and timing of stormflow hydrographs.

In larger basins rainfall and snowmelt inputs will exhibit greater spatial variability, and the peak flows from the various sub-basins are more likely to be desynchronized. Larger basins generally will have more potential water storage in and on the banks and floodplains. With increasing basin size there is more potential for transmission losses, but in most mountainous environments these generally are assumed to be negligible (Table 1). Transmission losses can be substantial when a stream flows across a coarse-textured alluvial fan or alluvial plain (Herron and Wilson 1999, Woods et al. 2006), in karst terrane, and in semi-arid environments when the water table is below the deepest portion of the stream channel.

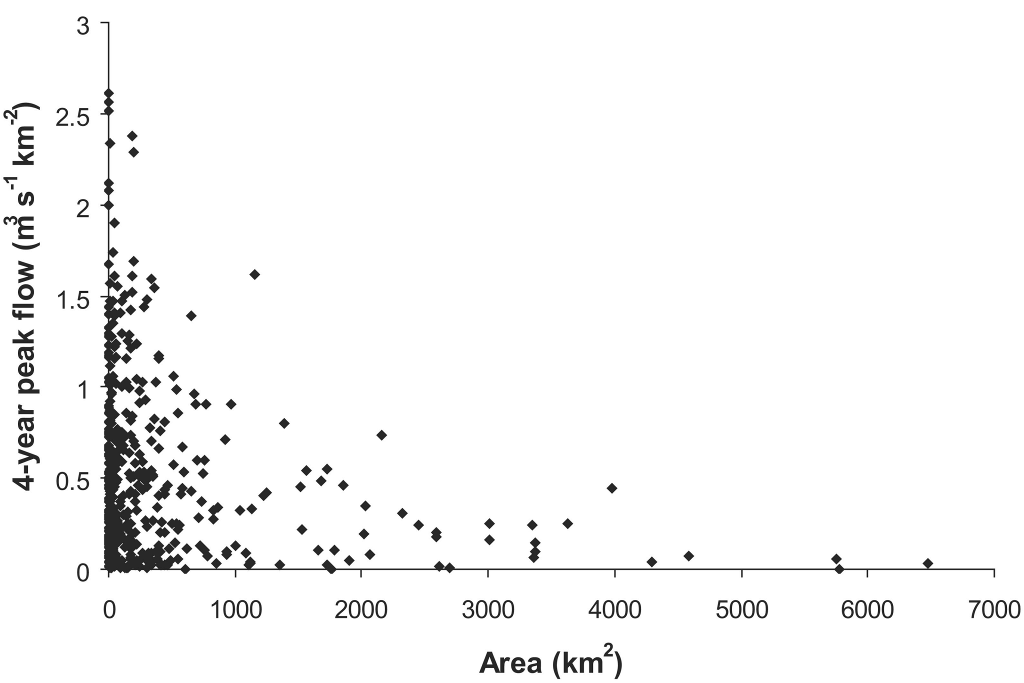

Figure 1. Unit area peak flows with a recurrence interval of four years versus catchment area for 477 gauging stations in Washington state.

Table 1. **Generalized relative likelihood of the storage, transformation, and delivery of eight different constituents from headwater streams to downstream reaches**

Constituent	Likely magnitude of Storage	Transformation	Delivery	Means of delivery
Discharge	Low	Low	High	All flows, minimal delay
Fine sediment (<2 mm)	Low to moderate	Low	Moderate to high	All flows, but predominantly high flows
Coarse sediment (>2 mm)	High	Moderate to high	Low to moderate	High flows and mass wasting events
Large woody debris	High	Low to moderate	Low	Mass wasting or extremely high flows
Coarse particulate organic matter	Moderate	High	Moderate	Primarily high flows and mass wasting
Fine particulate organic matter	Low to moderate	Moderate	Moderate to high	All flows, especially high flows
Nutrients	High	High	Low to moderate	All flows
Temperature	Low	High	Low to moderate	Low flows

Because stream channels are relatively efficient conveyers of water, particularly at higher flows, any change in runoff induced by management activities in headwater basins is likely to affect downstream runoff (Table 1). The aggregation of various management-induced changes in flow from different headwater areas can result in a cumulative watershed effect (CWE), and the magnitude of this effect depends primarily on the changes in flow resulting from each management activity, and only secondarily on how these changes in flow are transmitted downstream (MacDonald 2000).

The management-induced changes in flow can be difficult to assess in large forested watersheds for several reasons. First, paired-watershed studies have shown that the combination of forest harvest and roads can increase the size of peak flows, decrease the size of peak flows, or have no significant effect (Harr and McCorison 1979, Austin 1999, Moore and Wondzell 2005). The majority of studies conducted in small (i.e., 10–300 ha), rain-dominated catchments in the Oregon Cascades and along the Pacific Coast have shown that extensive forest harvest increases the size of the average storm peak flow by about 13 to 44% (Moore and Wondzell 2005). In snowmelt-dominated areas forest harvest generally increases the annual maximum peak flows by about 40%, but values can range up to 87% (Troendle and King 1987, King 1989, Moore and Wondzell 2005). Both field (Toman 2004) and modeling (Wigmosta and Perkins 2001) studies indicate that road runoff can increase peak stormflows in small headwater streams by up to 500%, but studies in larger basins generally have not been able to document a road-induced increase in peak stormflows. This variability means that an understanding of the underlying causal processes is needed to predict the hydrologic response of different headwater basins to a given set of management activities.

A second difficulty is whether the results from small watersheds can be extrapolated to larger basins, as most paired-watershed studies have been conducted at the headwater scale. The combined effect of forest harvest and roads on runoff is still controversial in basins larger than about 10 to 20 km^2, especially where rain-on-snow events generate the largest floods. A 1996 article, for example, claimed that forest harvest and roads increased peak flows by up to 100% in large catchments (62–559 km^2) as compared to a 50% increase in catchments smaller than 1.0 km^2 (Jones and Grant 1996). Other researchers used different statistical methods to analyze the same data set and found no significant increases in peak flows in the same large catchments (Thomas and Megahan 1998, Beschta et al. 2000). A decreasing change in the size of peak flows with increasing catchment size could be attributed to increased floodplain storage, greater spatial and temporal variability in rainfall and snowmelt, greater variability in basin characteristics (e.g., geology, soils, vegetation, and drainage network structure), and the tendency for percentage area disturbed to decrease with increasing basin size (Megahan and Hornbeck 2000, Beschta et al. 2000). In the snowmelt-dominated Rocky Mountains, the changes in flow due to harvesting 24% of a nearly 17 km^2 basin were similar to the values observed in small headwater basins (Troendle et al. 2001).

A third problem is the decline in measurement accuracy in larger basins because streamflows usually are measured in natural channels rather than with carefully engineered flumes or weirs. A lower accuracy limits our ability to detect significant change (see section on adaptive management). A final limitation is that treatments such as forest harvest do not persist over time due to vegetative regrowth. In larger basins, treatments tend to be spread out over longer time periods and regrowth will reduce both the magnitude of the changes in streamflow and the number of storms or years that can be compared (Austin 1999, Jones 2000).

We conclude that headwater streams are the dominant source of runoff. This water is generally conveyed to downstream areas (Table 1), and in-channel processes become increasingly important with increasing basin size. Management-induced changes in runoff have been repeatedly detected in small experimental watersheds, but it is much more difficult to detect the effects of forest management on runoff in larger basins due to the spatial and temporal variations in precipitation and snowmelt, transmission and storage losses, the uncertainties in quantifying the site-scale changes in runoff, measurement errors, and the difficulty of rapidly imposing a given treatment while maintaining a comparable untreated control.

Hillslope Sediment Production and Delivery

While hillslope runoff is usually delivered to headwater channels by subsurface flowpaths, sediment is transported into headwater channels and downstream reaches by surface processes (Benda and Dunne 1997a, b, Istanbulluoglu et al. 2004). The proportion of hillslope erosion that is delivered to channels varies with the transport process, proximity to the stream channel, flowpath characteristics, sediment particle size, level of disturbance, and the magnitude of runoff and erosion events (Dietrich et al. 1982, Wemple et al. 1996, Reid and Dunne 1996, Benda and Dunne 1997a, b, Croke and Mockler 2001, Sidle et al. 2004, Istanbulluoglu et al. 2004). Similarly, the ability of headwater streams to deliver sediment to downstream reaches is a function of channel type, transport process, transport capacity, and sediment particle size (Knighton 1998, Bunte and MacDonald 1999, Hassan et al. 2005a). The delivery of sediment to stream channels generally can be classified as discrete (e.g., debris flows) or relatively chronic (e.g., soil creep or the storm-by-storm delivery of sediment from roads).

Mass wasting accounts for 60% to more than 90% of long-term sediment inputs in many headwater catchments in the Pacific Northwest (Swanson et al. 1982, Benda and Dunne 1987, Raines 1991, Paulson 1997, Brardinoni et al. 2003, Benda et al. 2005). At a larger scale, mass wasting accounted for 86% of the total sediment yield for a 187 km^2 catchment in northwestern California (Raines 1991) and 44 to 98% of the total sediment yield for catchments in northwestern Washington with drainage areas of 12 to 140 km^2 (Paulson 1997).

In headwater catchments the dominant mass wasting processes are either translational slides or debris flows, and these typically originate in colluvial hollows and inner gorge landforms (Dietrich and Dunne 1978, Benda et al. 2005). The density of colluvial hollows in northern California and the Oregon Coast Range range from 22 to 100 per square kilometer (Dietrich and Dunne 1978, Montgomery and Dietrich 1989, Benda 1990), and the average cycle of infilling and failure has been estimated at 5,000–6,000 years (Dietrich et al. 1982, Benda and Dunne 1987, Reneau and Dietrich 1991). In some geologic terranes (e.g., the Franciscan formation in northwestern California), mass wasting in the form of deep-seated earthflows can be the dominant process for delivering hillslope sediment to the stream network (Kelsey 1978).

The delivery of sediment from a mass wasting event to a headwater channel depends on its location relative to the channel network and the travel distance of the landslide or debris flow. Colluvial hollows often are immediately above the heads of first-order channels, so landslides or debris flows in colluvial hollows typically have to travel only a short distance before entering the channel network, and often connect colluvial hollows to first-order channels (Montgomery and Dietrich 1989, Istanbulluoglu et al. 2004). Inner gorges are steep and parallel to the channel network, so mass failures in this landform also have a high probability for delivering water and sediment to the channel network (Paulson 1997). By combining data from two field studies (Benda and Cundy 1990, Robison et al. 1999), we found that the median travel distance for 473 landslides and debris flows in Oregon and northwestern Washington was just under 250 m, although 5% had a travel distance in excess of 1,300 m (Figure 2).

Mass wasting events deliver both fine and coarse sediment to the channel network (Reid and Dunne 1996). Shallow mass wasting processes, such as translational slides, typically remove the soil profile down to bedrock, so the particle-size distribution of this material is similar to the particle-size distribution of the soil profile (Reid and Dunne 1996). In the Oregon Coast Range, the particle-size distribution of the sediment stored in first- and second-order

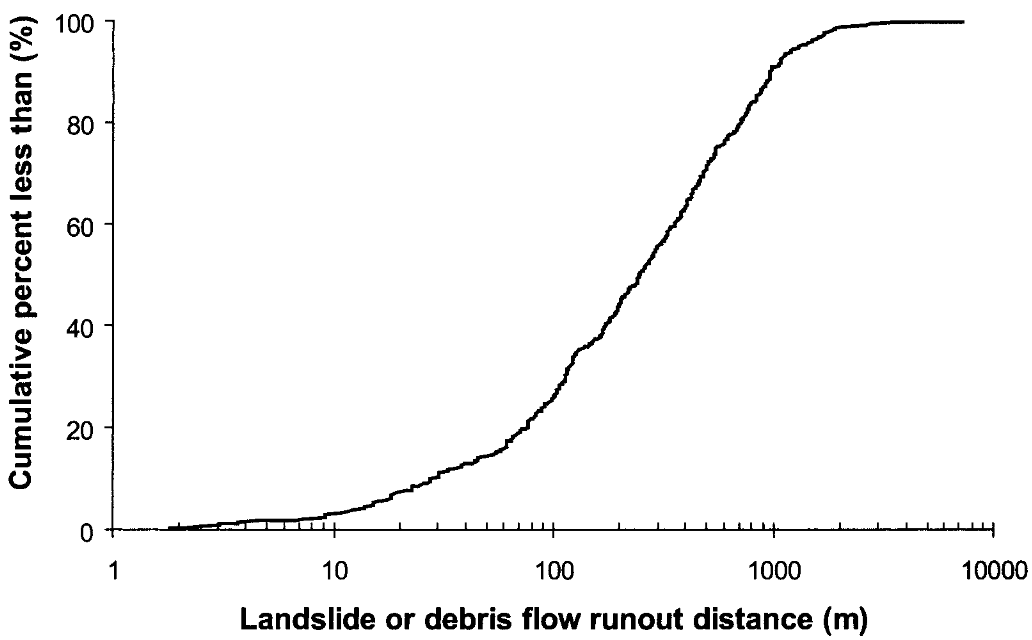

Figure 2. Cumulative frequency distribution of landslide and debris flow runout distances (n = 473) (Benda and Cundy 1990, Robison et al. 1999).

channels was almost identical to the particle-size distribution of the material stored in colluvial hollows (Benda and Dunne 1987). The larger clasts delivered into headwater streams by debris flows and deep-seated mass wasting often exceed the flow competence and remain in the channel as lag deposits (Brummer and Montgomery 2003).

Sediment inputs by shallow mass wasting generally are episodic in nature. The recurrence interval of these events decreases with distance downstream because of the rapid increase in the number of landslide-prone landforms—such as colluvial hollows—with increasing scale (Figure 3). The likelihood of disturbance (e.g., fire or timber harvest) or an exceptionally large storm event also increases with increasing catchment size (Benda and Dunne 1997a). If erosion events were completely random, annual sediment inputs and sediment yields should be progressively less variable with increasing scale. In fact, the large frontal storms that trigger mass wasting events in the Pacific Northwest tend to occur over relatively large areas, and a compilation of sediment yield data from different regions shows little change in the variability of annual sediment yields with increasing basin size (Figure 4).

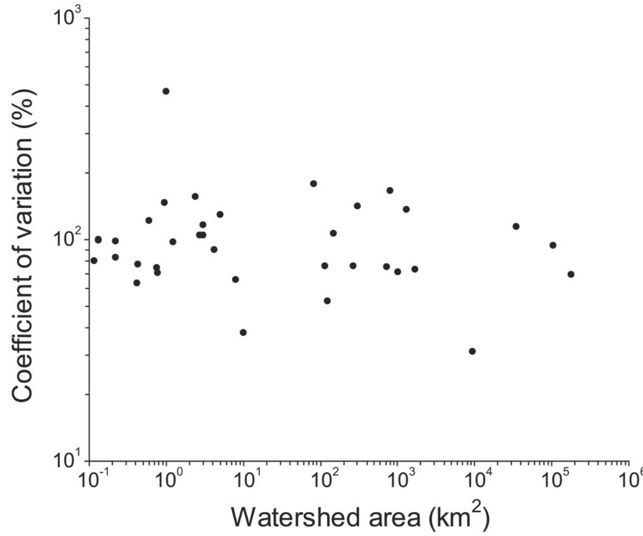

Figure 4. Coefficient of variation for annual sediment yields versus basin area (Bunte and MacDonald 1999).

Land use activities can increase both the frequency and magnitude of mass wasting and surface erosion. In forested areas, timber harvest and roads can increase the amount of sediment being produced from mass wasting by 3 to 16 times (Swanson and Dyrness 1975, Montgomery et al. 2000, Guthrie 2002, May 2002), and the amount of sediment being delivered to the channel network by 0.6 to 138 times (Guthrie 2002, Brardinoni et al. 2003, Hassan et al. 2005a). The management-induced increases in the amount of sediment being delivered to the channel network are due to an increase in travel distance as well as an increase in frequency. In the Oregon Coast Range, road-induced mass failures traveled three times as far as the mass failures in a mature forest, and the road-induced mass failures increased the amount of sediment being delivered to the channel network by nearly five times relative to mature forests (May 2002).

The chronic sources of sediment in headwater channels include soil creep, bank erosion, and surface erosion (Roberts and Church 1986, Hassan et al. 2005a). Estimated soil creep rates are 0.001 to 0.01 m yr^{-1} for the Oregon Coast Range, 0.002 m yr^{-1} for the Oregon Cascades, and 0.02 m yr^{-1} for Northern California (Dietrich and Dunne 1978,

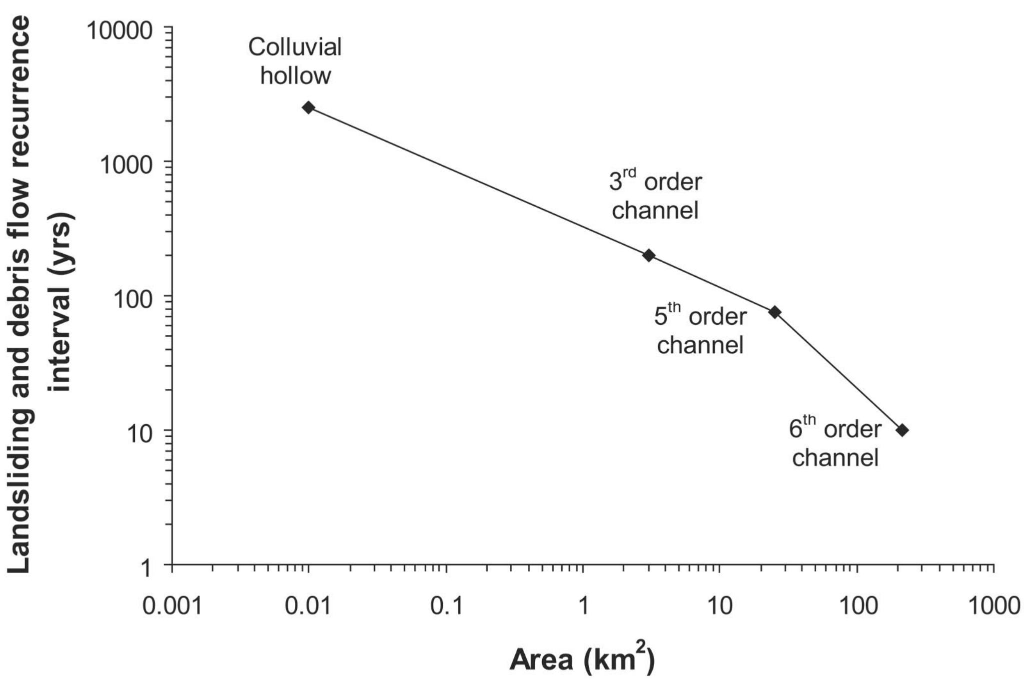

Figure 3. Predicted recurrence intervals for landslides and debris flows versus catchment area under undisturbed conditions in the North Fork of Smith Creek, Oregon. The average recurrence interval for a 3 km^2 catchment was reported as 50–100 years (Benda and Dunne 1997a), so a value of 75 years was used here. The recurrence interval for the sixth-order catchment was reported as less than a decade, and a value of 10 years was used here.

Swanson et al. 1982, Benda et al. 2005). The volume of sediment that is delivered to the stream network from soil creep depends on the creep rate, soil depth, and drainage density (Dietrich and Dunne 1978, Roberts and Church 1986).

The high infiltration rates in most undisturbed forested catchments mean that rainsplash, sheetwash, and rilling typically generate no more than a small fraction of the sediment that is delivered from hillslopes to headwater channels (Roberts and Church 1986, Hassan et al. 2005). Even if overland flow does occur, the dense vegetative cover and high surface roughness minimize overland flow velocities and sediment transport capacity (Dietrich et al. 1982, Libohova 2004).

Overland flow and the associated surface erosion processes are common on unpaved forest roads, and the resultant road surface erosion rates are commonly two or more orders of magnitude higher than the surface erosion rates in undisturbed areas (Megahan and Kidd 1972, Reid and Dunne 1984, Luce and Black 1999, MacDonald et al. 2004). Road erosion rates in the western United States have been estimated at up to 101 kg m^{-2} yr^{-1} for heavily trafficked roads in western Washington (Reid and Dunne 1984), but more typical values are from 0.2 to 2.0 kg m^{-2} yr^{-1} (MacDonald and Stednick 2003). In the Oregon Coast Range, sediment production from different road segments followed a log-normal distribution, indicating that most of the road-related sediment is being derived from a relatively few segments. Road sediment is of particular concern because it generally is fine-grained (sand-sized or smaller) (Megahan and Hornbeck 2000, Ramos-Scharron and MacDonald 2005), and this material is particularly detrimental to many aquatic organisms (Waters 1995).

There are fewer data on the delivery of sediment from roads to the stream network than on road sediment production. Studies in the western United States have reported that from 18% to 75% of the roads are hydrologically connected to the stream network (Coe 2006). A recent analysis of these connectivity data indicates that the percentage of roads connected to the stream network is directly proportional to the mean annual precipitation (Figure 5) (Coe 2006). The presence of engineered drainage structures—such as waterbars, rolling dips, or relief culverts—can decrease the proportion of roads that are connected to the stream network by about 40% (Figure 5). The small size of the particles being eroded from roads means that the hydrologically connected roads also will be delivering sediment to the stream network.

High-severity wildfires in forested basins can increase the production and delivery of sediment by several orders of magnitude (Moody and Martin 2001, Wondzell and King 2003, Benavides-Solorio and MacDonald 2005). This increase is due to the severe reduction in infiltration rates, the high overland flow velocities due to the loss of ground cover and surface roughness, and the large increase in drainage density due to the headward extension of the channel network (DeBano et al. 1998, Robichaud et al. 2000, Wondzell and King 2003, Istanbulluoglu et al. 2004,

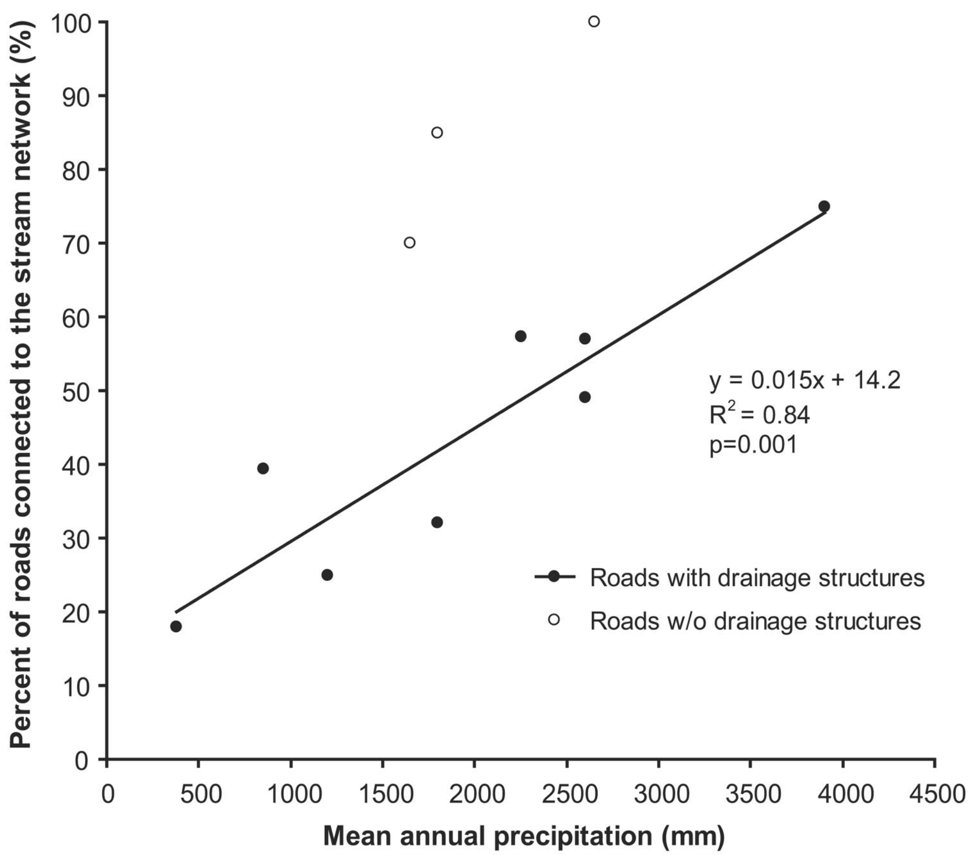

Figure 5. Percentage of roads connected to the stream channel network versus mean annual precipitation (Coe 2006).

Libohova 2004). In the Colorado Front Range most of the sediment from high-severity fires is due to channel incision (Moody and Martin 2001, Pietraszek 2006), and this is why convergent hillslopes tend to produce several times more sediment per unit area than planar hillslopes (Benavides-Solorio and MacDonald 2005). The increase in drainage density greatly increases the hillslope-stream connectivity (Libohova 2004), and most of the sediment from high-severity fires is available for transport to downstream reaches (Moody and Martin 2001, Pietraszek 2006). In contrast, postfire erosion rates after moderate- and low-severity fires are one or more orders of magnitude lower than for sites burned at high severity because the protective litter layer and soil organic matter is not completely consumed and there is correspondingly less rainsplash, soil water repellency, and overland flow (Robichaud et al. 2000, Neary et al. 2005).

Connectivity of Sediment from Headwater Channels to Downstream Reaches

The delivery of sediment from headwater channels to downstream reaches depends on the transport process, channel type, transport capacity, and sediment particle size. In steep headwater catchments, debris flows can be the dominant process for delivering sediment to first- and second-order channels as well as the predominant source of sediment (Benda and Dunne 1987, Benda 1990, May 2002, Benda et al. 2005). Debris flows typically erode the colluvium and alluvium stored along the axis of a hollow or headwater channel, and in the Oregon Coast Range they can entrain from 2 to 15 m^3 of sediment per meter of channel length (Benda 1990, May 2002). Debris flows typically deposit sediment when tributary junction angles exceed 70 degrees (Benda and Cundy 1990) and channel slopes are less than 10% (Benda et al. 2005). However, debris flows can deposit sediment in channels with gradients of up to 25%, or continue to transport sediment when the channel slope is as low as 3% (Benda et al. 2005). In general, most of the sediment from debris flows is deposited in channels that are third-order or higher (Benda and Dunne 1987, Benda 1990, May 2002, Benda et al. 2005). As noted earlier, the transport process also will affect the size of the sediment being transported from the hillslope to the channel.

The sediment transport capacity of headwater channels, and hence the delivery of sediment to downstream reaches, is limited by the high flow resistance due to large clasts, large woody debris, and vertical bedforms such as channel steps (Montgomery and Buffington 1997, Curran and Wohl 2003, Benda et al. 2005). The high roughness in headwater channels provides flow resistance, which reduces sediment entrainment and makes it more difficult to predict sediment transport rates (Hassan et al. 2005a).

The entrainment and transport of sediment in stream channels can be divided into three phases (Ashworth and Ferguson 1989, Hassan et al. 2005a), and these represent different degrees of upstream-downstream connectivity (Figure 6). In phase I, sand-sized or finer material is transported over a stable bed, and this can deliver fine sediment to downstream reaches at relatively low flows (Ashworth and Ferguson 1989). In phase II the entrainment and transport of the bed material is size-selective, and in phase III the entire bed is mobilized, resulting in a high degree of upstream-downstream connectivity. The coarse nature of the bed material in many headwater streams means that phase II and especially phase III transport are relatively infrequent, particularly in channels with strongly structured bedforms (e.g., step-pool channels) (Grant et al. 1990, Hassan et al. 2005).

These three phases of sediment transport mean that fine

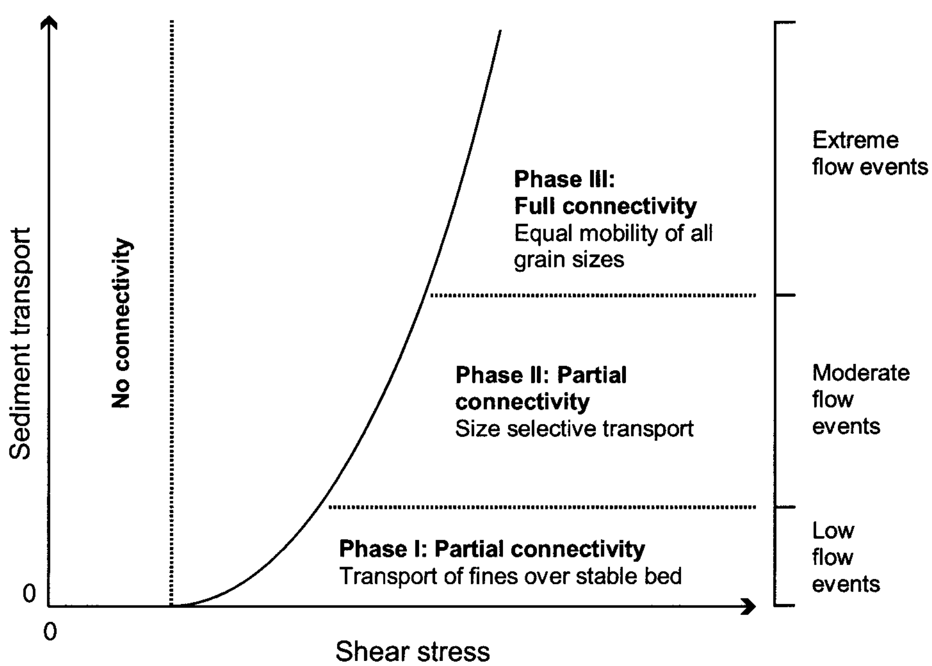

Figure 6. Schematic diagram of the different phases of bedload transport and sediment connectivity versus shear stress (adapted from Hassan et al. 2005).

sediment is more readily delivered to downstream reaches than coarser particles (Table 1). Nevertheless, large amounts of fine sediment can be stored in headwater channels in the lee of large boulders, behind large woody debris, under the surface armor layer, in the channel margins, and on floodplains and terraces (Grant et al. 1990, May and Gresswell 2003b, Benda et al. 2005). In a small ephemeral channel in western Washington, only 35% of the fine (0.063–0.5 mm) and 10% of the coarse (0.5–2.0 mm) sand particles were transported more than 95 to 125 m (Duncan et al. 1987). These low proportions were attributed to the effects of large woody debris on sediment storage and transport rates (Duncan et al. 1987).

Recent studies have documented the important role of large woody debris with respect to the storage and transport of sediment in headwater channels (Keller and Swanson 1979, Megahan 1982, Chesney 2000, May and Gresswell 2003b, Gomi and Sidle 2003). In the Pacific Northwest and Alaska, large woody debris stored an average of 0.5 m^3 of alluvial sediment per meter of channel, and the range of values was from 0 to 2.9 $m^3\ m^{-1}$ (Megahan 1982, Chesney 2000, May and Gresswell 2003b, Gomi and Sidle 2003). There may be very little sediment stored in channels that have been recently scoured by debris flows (May and Gresswell 2003b), but the amount of stored sediment will increase over time as sediment is delivered to the channel by colluvial processes and the sediment storage capacity increases with the recruitment of woody debris (May and Gresswell 2003b).

Sediment transport models developed for larger alluvial channels typically overpredict sediment transport rates in headwater streams by at least an order of magnitude (Hassan et al. 2005a). This overprediction can be attributed to the greater form roughness and sediment storage capacity in headwater streams. Alternatively, empirical models derived from tracer studies can be used to estimate the downstream travel distance and delivery of sediment. The best model developed from 50 studies explained 43% of the variability in mean annual travel distance as a function of tracer particle size and bankfull channel width:

$$\log_{10}T = 2.30 + 0.819 \log_{10}W - 0.611 \log_{10}D, \quad (1)$$

where T is the mean annual travel distance in m yr^{-1}, W is the bankfull channel width in m, and D is the particle size in mm (Bunte and MacDonald 2002). The significance of bankfull width is consistent with other analyses of tracer studies (Beechie 2001) and the cross-correlations between channel width and other variables, such as channel slope and flow depth (Knighton 1998).

Equation 1 indicates that the mean annual travel distance for a 3 m-wide headwater stream will vary from several kilometers for a particle that is only 0.05 mm in diameter to only a few tens of meters for a particle that is 64 mm in diameter (Figure 7). Both Figure 7 and our understanding of sediment transport processes indicate that there is likely to be a long lag in the delivery of larger particles to downstream reaches, and this lag increases as channel size decreases and particle size increases (Table 1). Particle abrasion will decrease travel times by decreasing particle sizes, and for weaker rock types this particle breakdown can greatly increase sediment delivery from headwater streams to downstream reaches (Benda and Dunne 1997b). The sediment derived from unpaved roads also has a relatively high likelihood of being delivered to downstream areas because it is predominantly sand-sized or smaller. Much of the sediment derived from high-severity wildfires is likely to be delivered to downstream reaches because it is largely

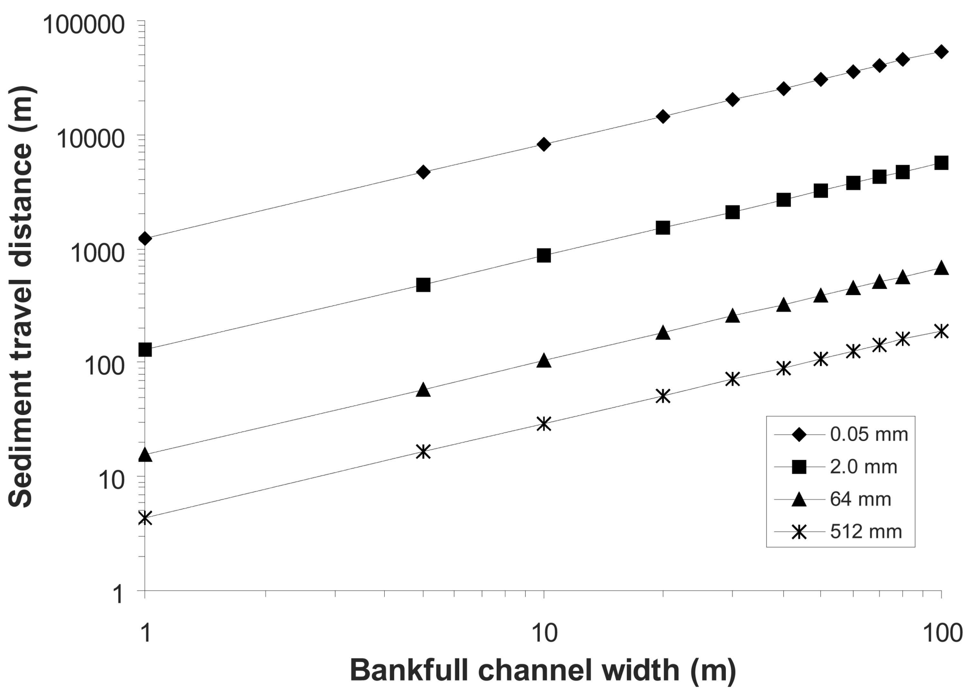

Figure 7. Predicted mean annual sediment travel distance for four different particle sizes as a function of bankfull channel width. The model explains 43% of the variability in mean annual sediment travel distance.

derived from rill and channel erosion (Moody and Martin 2001, Pietraszek 2006). High-severity wildfires also greatly increase the size of peak flows, and hence the sediment transport capacity (Helvey 1980, Istanbulluoglu et al. 2004).

Downstream Effects of Sediment Inputs

The delivery of sediment through the channel network can affect water quality, channel morphology, and aquatic organisms in downstream reaches, but the magnitude and type of these effects depend in part on whether the sediment delivery is episodic or chronic (Waters 1995, Gresswell 1999, Trombulak and Frissell 2000). Many of the documented impacts from headwater areas are due to high-magnitude, low-frequency disturbances that transport large slugs of sediment from hillslopes or headwater streams to a downstream reach. Such episodic inputs can induce debris fans, valley terrace formation, channel avulsions, coarse sediment deposits, substrate fining, increased bedload transport, and channel aggradation (Madej and Ozaki 1996, Lisle et al. 2000, Miller and Benda 2000, Sutherland et al. 2002, Benda et al. 2004a, May and Lee 2004). The effects of these episodic inputs tend to be most pronounced at the confluence of headwater streams with larger channels, as these locations are characterized by sharp changes in sediment supply, wood loading, and transport capacity. These confluence effects can include sediment deposition, a change in substrate size, and changes in channel morphology above and below the confluence (Benda et al. 2004a). Because the processes that deliver sediment depend on both the tributary and mainstem drainage area, the likelihood of a confluence effect, such as a debris fan, increases as the ratio of the tributary drainage area to the mainstem drainage area increases (Figure 8) (Benda et al. 2004a). In humid environments, debris flows typically affect the mainstem when the tributary drainage area is less than 1 km^2 and the mainstem drainage area is less than 50 km^2 (Benda et al. 2004a). The probability of a confluence effect is estimated to be greater than 50% when the ratio of tributary to mainstem drainage area exceeds 0.012 (Figure 8) (Benda et al. 2004a).

Alluvial effects occur when sediment from a tributary is transported to a mainstem channel by high flows (e.g., phase III sediment transport), and these effects can include changes in gradient, sediment deposition, a change in substrate size, and channel instability. The scale domain for alluvial effects is when tributary drainage areas are greater than 10 km^2 and mainstem drainage areas are greater than 500 km^2 (Figure 8) (Benda et al. 2004a).

Once a slug of sediment is delivered to a higher-order channel, the resulting sediment wave can move downstream via translation, dispersion, or a combination of the two (Lisle et al. 2001). Translation means that the sediment wave moves downstream as a single mass, but this is rare except when the sediment is fine-grained and the flow is tranquil (i.e., subcritical) (Lisle et al. 2001). In a purely dispersive sediment wave the upstream edge and wave apex do not travel downstream (Lisle et al. 2001). Because dispersion is the dominant process, the amplitude of sediment waves tends to decrease rapidly over time, making them difficult to detect unless the wave is relatively young or the sediment input is large relative to the channel dimensions (Lisle et al. 2001). Dispersion may be unimportant if the large clasts delivered by mass wasting or other processes exceed the flow competence of the receiving channel (Brummer and Montgomery 2006). The movement of a sediment wave can initiate a cycle of channel aggradation and degradation (Madej and Ozaki 1996, Miller and Benda 2000, Sutherland et al. 2002), while the selective transport and abrasion of material can induce downstream fining (Sutherland et al. 2002).

There is less evidence in the literature on how the low

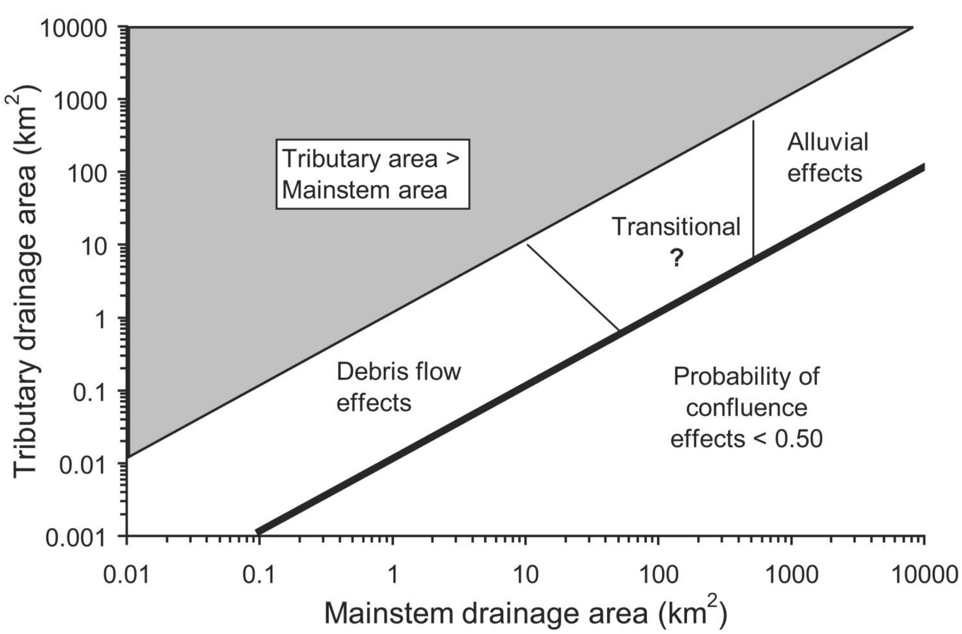

Figure 8. Process domains and the probability for confluence effects as a function of the tributary drainage area relative to the mainstem drainage area for humid regions of the Pacific Northwest (Benda et al. 2004a). The bold line represents a probability of 0.50 for a confluence effect.

magnitude, chronic delivery of sediment from hillslope sources can affect downstream reaches, as a change in sediment load often is accompanied by other changes, such as a loss of riparian cover or a change in the amount of large woody debris (Everest et al. 1987). As indicated by Table 1 and Equation 1, fine sediment is more likely to be delivered to downstream areas than coarse sediment. Numerous studies have linked an increase in fine sediment loads to changes in macroinvertebrate populations and, to a lesser extent, changes in fish habitat and fish populations (Everest et al. 1987, Waters 1995, Suttle et al. 2004). The overall trends for macroinvertebrate populations show that a shift in the substrate due to increasing fine sediment loads will tend to decrease taxa richness and abundance; decrease the abundance and richness of sensitive taxa such as Ephemeroptera, Plecoptera, and Tricoptera; and increase the number of oligochaetes and burrowing chironomids (Waters 1995).

These changes in the macroinvertebrate populations are a concern because they can directly affect the amount and type of prey available to high-value fisheries. Large increases in fine sediment loads also are a concern because of their potential adverse effect on spawning and rearing habitat (Everest et al. 1987). At low sediment loads most of the fine sediment in the channel is either on the channel margins or in the matrix of the bed material (Carling and Reader 1982, Lisle and Hilton 1999). As the fine sediment supply exceeds the storage capacity of the bed matrix, the fine particles form surficial patches on the bed surface and in pools (Lisle and Hilton 1999). In northwestern California and in Colorado, both the sediment supply and the lithology were found to affect the filling of pools with fine sediment (Schnackenberg and MacDonald 1998, Lisle and Hilton 1999).

Relatively few studies in forested areas have documented a significant relationship between downstream channel condition and watershed-scale indices of sediment supply, such as percent area harvested, road density, or equivalent roaded (or clearcut) area (McGurk and Fong 1995, MacDonald et al. 1997, Schnackenberg and MacDonald 1998, Faustini and Kaufmann 2003, Cover et al. 2006). These studies can help identify potential problems and possible causes, but they generally do not quantify the various sediment sources or provide the process-based analyses needed to guide regulators and resource managers. At the headwater scale, paired-watershed studies have shown that road construction can increase sediment loads by 100 to 200% (Brown and Krygier 1971, Rice et al. 1979), but many of these studies were done when standards for road design, construction, and maintenance were less stringent than present standards (e.g., more frequent drainage, full bench construction, limitations on wet season use, and outsloping). Recent paired watershed studies in northwestern California have demonstrated that increases in suspended sediment loads were primarily controlled by postharvest increases in the volume of streamflow during storms (Lewis et al. 2001). Although increases in suspended sediment were attributed to the watershed area occupied by roads, there was little field evidence of sediment delivery from newly constructed roads (Lewis et al. 2001).

In summary, there is a need for studies to measure directly the effects of current management activities on sediment production in headwater areas, explicitly link these sources to the channel network, evaluate sediment routing, and then document whether there is a resulting downstream physical and/or biological response. Efforts to correlate disturbance indices with downstream conditions will need to explicitly consider hillslope-channel connectivity rather than simply using watershed-scale means or totals. The identification of a downstream response may be difficult because of the temporal variability in downstream conditions and the potential effect of infrequent catastrophic erosion events or longer-term processes, such as Pleistocene glaciation or tectonic uplift, relative to contemporary hillslope erosion (Church and Slaymaker 1989, Kirchner et al. 2001, Ferrier et al. 2005).

Production and Delivery of Large Woody Debris to Headwater Channels

Large woody debris (LWD) can play an important role in modifying channel hydraulics, regulating sediment flux, and controlling channel morphology and aquatic habitat (Montgomery and Buffington 1997, Curran and Wohl 2003, Lancaster et al. 2003, May and Gresswell 2003b, MacFarlane and Wohl 2003, Hassan et al. 2005b). The connectivity of LWD from hillslopes to headwater channels depends primarily on the processes by which wood is recruited and transported.

In headwater catchments LWD is recruited through mass wasting, tree mortality, windthrow, and bank erosion (Keller and Swanson 1979, Martin and Benda 2001, Benda et al. 2002, May and Gresswell 2003a). The relative importance of each process varies with local conditions and stream size. In southeastern Alaska, bank erosion and tree mortality were the primary sources of LWD in catchments of 0.6–1.7 km^2, while mass wasting accounted for less than 3% of the total wood recruitment (Martin and Benda 2001). Bank erosion and tree mortality also were the primary sources of LWD in northwestern California (Benda et al. 2002). In contrast, landslides delivered more than half of the wood pieces to second-order streams in the Oregon Coast Range (May and Gresswell 2003a). In third-order streams in the same area, windthrow accounted for 60% of the wood pieces, while mass wasting contributed only 10% (May and Gresswell 2003a). In general, recruitment from mass wasting becomes less important and recruitment from bank erosion becomes more important with increasing drainage area (Keller and Swanson 1979, Martin and Benda 2001).

The dominant recruitment process controls which and how much of the hillslopes are connected to headwater channels and the frequency of hillslope-channel interactions. In the Oregon Coast Range, the median source distance for LWD was 40 m for a variety of mass wasting processes, 20 m for windthrow, 18 m for tree mortality, and 2 m for bank erosion (May and Gresswell 2003a). This indicates that mass wasting has a higher spatial connectivity relative to the other LWD recruitment processes, but the temporal connectivity of individual hillslopes may be much lower because landslides and debris flows are so infrequent. Individual tree mortality has a lower spatial connectivity because transport distances are less, but the temporal connectivity is much higher because the LWD inputs are more frequent and evenly spaced over time.

Delivery of LWD from Headwater Channels to Downstream Reaches

The delivery of LWD from headwater channels to downstream reaches depends on the disturbance regime and high flows in the headwater channel, as well as the number and size of LWD (Table 1) (Hassan et al. 2005b). The transport of LWD from headwater channels typically is associated with debris flows or extreme floods (May 2002, Abbe and Montgomery 2003, Benda et al. 2005, Hassan et al. 2005b), but fluvial transport in headwater channels is relatively rare because the LWD tends to be large relative to the channel dimensions and size of peak flows (Bilby and Bisson 1998, May and Gresswell 2003a). A meta-analysis of 10 studies found that the normalized loading of LWD (i.e., pieces per unit channel length divided by channel width) was greatest in channels less than 5 m wide (Hassan et al. 2005b). These results mean that the residence time of LWD in headwater channels is primarily a function of the decay rate rather than the transport rate (Keller and Swanson 1979, Lancaster et al. 2003, Hassan et al. 2005b).

Debris flows are the most important transport mechanism for LWD in first- through third-order channels (Abbe and Montgomery 2003), but fluvial transport becomes more important as channel size increases (Hassan et al. 2005b). Wood in mid-sized streams (i.e., third-order) tends to move as a congested mass (Braudrick et al. 1997), while in fourth-order and larger channels LWD is more likely to be selectively transported. The travel distance of LWD in these larger streams is a function of the ratio of wood length to both the mean channel width and the mean radius of the channel curvature (Braudrick et al. 1997).

The complex interactions between sediment and LWD affect the likelihood and magnitude of downstream effects. The entrainment of LWD can reduce debris flow velocities, runout lengths, and facilitate sediment deposition (May 2002, Lancaster et al. 2003, Bunn and Montgomery 2004). Higher LWD loadings increase the amount of sediment that can be stored in the channel as well as on floodplains and terraces (Montgomery et al. 2003). A greater sediment storage capacity can help store sediment inputs from hillslopes and release this material as somewhat smoother fluxes of fluvially transported sediment (Massong and Montgomery 2000, Lancaster et al. 2001, Bunn and Montgomery 2004). Hence, the delivery of sediment and LWD to downstream channels by debris flows can be moderated or decreased by higher wood loadings in headwater channels (Lancaster et al. 2001, May and Gresswell 2003b, Bunn and Montgomery 2004). Conversely, a reduction in LWD loading due to forest harvest, fire, or other causes can result in a more direct coupling between the fine sediment inputs into headwater reaches and the delivery of this sediment to downstream areas (Lancaster et al. 2001, May and Gresswell 2003b, Bunn and Montgomery 2004).

Inputs and Downstream Delivery of Organic Matter and Nutrients

The small size of headwater streams means that they are almost completely shaded in forested areas. The large amounts of canopy cover mean that autochthonous (i.e., inputs generated within the stream) production is usually quite low (Gomi et al. 2002, Richardson et al. 2005), and up to 90% of the organic matter inputs are derived from hillslope and riparian sources (Fisher and Likens 1973). In addition to LWD, the primary organic matter inputs include dissolved organic carbon (DOC), coarse (>1 mm) particulate organic matter (CPOM), and to a much lesser extent, fine particulate organic matter (FPOM) (Gomi et al. 2002, Richardson et al. 2005). The coarse particulate organic matter is predominantly leaves and needles, and this material can be rapidly transformed into FPOM by fungi, invertebrates, fragmentation, and abrasion (Richardson et al. 2005). Total organic matter inputs per unit stream area are high relative to downstream reaches (Richardson et al. 2005). Primary production within the stream becomes more important as drainage area and channel size increase (Bilby 1988, Richardson et al. 2005).

The delivery of organic matter from headwater streams to downstream reaches is controlled by many of the same transport processes as water and sediment (Table 1). Much of the DOC exported from headwater streams is derived from leaching of the forest litter, and the transport of DOC is directly coupled with the transport of water. FPOM typically represents the largest component of organic matter exports (Bilby and Likens 1980), as FPOM is generated by the rapid breakdown of CPOM and is more readily transported than CPOM (Table 1). FPOM and CPOM have lower specific gravities than similar-sized sediment particles, so these organic materials are more easily transported downstream.

Although FPOM and CPOM can be readily transported by streamflow, the high roughness and abundant LWD in headwater channels means that much of this organic matter can be stored in the channel, on floodplains, or on terraces (Table 1). The downstream delivery of FPOM and CPOM is primarily a function of whether the stream has access to this material rather than the transport capacity. This means that FPOM and CPOM are exported from headwater streams primarily during high flows (Table 1). Large amounts of organic matter also can be transported downstream on a much more episodic basis by debris flows or other mass wasting processes (Richardson et al. 2005).

Forest management can alter the inputs of organic matter in several ways. In the absence of buffer strips, timber harvest decreased allochthonous inputs of organic matter to small streams in western Washington by 80% (Bilby and Bisson 1992). Algal production increased by 60%, but the total organic matter inputs in the unlogged sites were still more than 50% higher than in the logged sites (Bilby and Bisson 1992). Timber harvest in riparian areas also can shift species composition from conifers to hardwoods (Richardson et al. 2005), although some states—such as Oregon and Washington—allow riparian areas to be converted from hardwoods to conifers. A change from conifers to nitrogen-fixing species such as red alder can increase in-channel productivity and decrease the total organic matter input from terrestrial sources (Piccolo and Wipfli 2002). Timber harvest also may increase the downstream delivery of organic matter by increasing mass wasting (Montgomery et al. 2000, May 2002, Richardson et al. 2005) and reducing the amount of storage by reducing wood recruitment (Richardson et al. 2005).

Nutrient production and delivery are particularly complex because of the high potential for uptake and transformation (Table 1). Stream water chemistry and nutrient availability are strongly influenced by geologic weathering, particularly in areas with younger soils derived from noncrystalline rocks (Feller 2005). Geologic weathering is an important source of potassium, magnesium, and calcium. Atmospheric deposition is an important source of compounds such as sulfate, nitrogen (N), and mercury (Schuster et al. 2002, Feller 2005). Nutrients can be leached from organic material such as litterfall, but the relative importance of nutrient inputs from hillslope leaching rapidly diminishes in the downstream direction as streamflow increases (Feller 2005). Nitrogen, and especially phosphorus (P), are of primary concern because they typically limit aquatic productivity (Goldman et al. 1990).

Nitrogen and phosphorus fluxes into and from headwater streams are controlled by runoff processes, chemical reactions, and biological uptake and transformations in the soils, groundwater, and channels. Most of the nitrogen transported from forests to streams is in the form of nitrate (Vitousek et al. 1979), but the amounts are relatively small in most undisturbed forest ecosystems (Feller 2005). An increase in nitrogen inputs to headwater streams will not necessarily increase primary productivity because light is often the limiting factor (Bisson 1982).

Phosphorus concentrations in streams are typically very low due to the rapid biological uptake and ease of chemical bonding (Hem 1970, Feller 2005). Most of the phosphorus entering into aquatic ecosystems in forested areas will be in the form of orhophosphates and either sorbed onto soil particles or incorporated into organic compounds (MacDonald et al. 1991). Any processes or management activities that increase the delivery of organic matter and sediment—particularly fine sediment—to the stream channel will increase the input of phosphorus.

Once in the stream, nutrients can be adsorbed by mineral and organic surfaces, oxidized by organisms or light, undergo chemical transformation, used by primary producers such as algae, and microbially transformed by processes such as nitrification (Table 1) (Feller 2005). The concept of nutrient spiraling refers to the uptake, transformation, and transport of nutrients in the downstream direction. In headwater channels the majority of inorganic N is typically removed or transformed in a few minutes or hours, or within tens or hundreds of meters (Peterson et al. 2001). The time and distance necessary for nitrogen uptake or transformation increases in the downstream direction because of the increase in water depth and discharge (Alexander et al. 2000). The rapid uptake of P within streams also results in little downstream transport of dissolved P (Feller 2005). Most of the downstream delivery of P will occur in particulate forms during high flows (Newbold et al. 1983).

The delivery of N and P is of greatest concern when there are downstream oligotrophic lakes and reservoirs, as the accumulation of these nutrients can lead to water quality degradation and changes to aquatic ecosystems (Dunne and Leopold 1978). The complexities of channel processes, when combined with the diversity of nutrient sources and inputs, makes it very difficult to link specific management activities in forested areas to downstream nutrient fluxes or concentrations (Feller 2005). In most cases the cumulative delivery of nutrients from forested areas is very small relative to the inputs from agricultural and urban areas (Dunne and Leopold 1978, EPA 1997, Carpenter et al. 1998).

Downstream Trends in Stream Temperature

Stream temperature typically increases in the downstream direction (Moore et al. 2005). The cooler temperatures in headwater reaches are due to the high proportion of canopy cover, the preponderance of cooler subsurface stormflow, and cooler air temperatures at higher elevations (Moore et al. 2005). Localized cooling can occur in downstream reaches due to an increase in shading, groundwater inputs, hyporheic exchange, and thermal stratification in pools (Nielsen et al. 1994, Moore et al. 2005).

Timber harvest, wildfire, grazing, windthrow, tree mortality, agriculture, urbanization, and debris flows can decrease the amount of canopy cover and increase summer water temperatures (Johnson and Jones 2000, Poole and Berman 2004, Moore et al. 2005). The magnitude of the temperature increase is generally proportional to the decrease in riparian shade. Riparian management zones often are designed to limit the decrease in stream shading and resulting temperature increases (Brown and Krygier 1970), but narrow riparian zones may allow the stream to be exposed (Jackson et al. 2001). In forested headwater streams, complete clearing can increase the maximum annual stream temperature by up to 16°C (Brown and Krygier 1970), but most studies in the Pacific Northwest have found that forest harvest increases the maximum stream temperature by no more than 5°C (Moore et al. 2005). In some cases clearcutting to the edge of a stream had no significant effect on water temperatures, and the absence of a significant increase was attributed to the shading provided by large accumulations of logging slash plus cooler post-harvest air temperatures (Jackson et al. 2001). If stream temperatures do increase due to forest harvest, studies show that from 5 to more than 20 years are required for stream temperatures to recover to preharvest levels. Larger streams usually require a longer time to recover (Moore et al. 2005).

An increase in stream temperature induced by forest harvest can be transmitted downstream, or it can quickly diminish once the stream enters a forested reach with extensive shading (Table 1) (Moore et al. 2005). The rate of downstream cooling in a well-shaded reach can be quite rapid due to the combination of groundwater inputs, hyporheic exchange, and heat conduction to the substrate (Story et al. 2003). One study in the interior of British Columbia found more than 4°C of cooling in the first 150 m below a clearcut (Story et al. 2003).

The potential for upstream temperature increases to affect downstream water temperatures is limited by the cooling effects associated with groundwater inflows, inflows from tributaries that have not been subjected to management-induced increases in temperature, hyporheic exchange, heat conduction to the streambed, and evaporative cooling (Table 1) (Moore et al. 2005). One study showed that stream temperature increases at the mouth of a 325 km^2 catchment were correlated with a cumulative harvest index (Beschta and Taylor 1988), but this was before the requirements for riparian buffers (Moore et al. 2005). The effects of forest management on water temperatures continues to be an active topic for research, but the increasing requirements for buffer strips and retaining riparian cover mean that increases in maximum water temperatures due to forest harvest generally are becoming less frequent, smaller in magnitude, and more localized.

Downstream water temperatures also can increase due to the indirect effects of timber harvest and other land use activities. For example, aggradation from management-induced mass wasting can lead to a wider, shallower channel that is more susceptible to solar radiation warming (Moore et al. 2005). An increase in fine sediment may clog channel substrate, thereby reducing hyporheic exchange and its associated cooling. Alternatively, extreme aggradation may result in subsurface flow and cooler water temperatures when the flow reemerges further downstream (McSwain 1987).

Detecting Headwater Effects on Downstream Reaches and Implications for Adaptive Management

The reviews in the previous sections highlight the diversity and complexity of headwater-downstream interactions for water, sediment, large wood, particulate organic matter, nutrients, and water temperature. These sections also summarized how management activities in headwater areas might affect each of these constituents, and the extent to which a change in the headwaters might be transmitted downstream and affect some downstream resource of concern. A strong understanding of headwater-downstream interactions is necessary under the National Environmental Protection Act, as this explicitly requires federal agencies to evaluate and consider the environmental impacts of their actions in decisionmaking. More specifically, federal agencies are required to consider the cumulative effect of their actions when added to other past, present, and reasonably foreseeable future actions (CEQ 1997). Many states have laws that require similar environmental assessments for actions taken by state agencies. Similarly, the Clean Water Act regulates both point and nonpoint sources of pollution, and the associated regulations provide an explicit process for controlling pollution on a watershed scale when water quality standards are not being met (MacDonald 2000). A watershed-scale approach also may be necessary when developing habitat conservation plans for aquatic species under the Endangered Species Act.

The difficulty is that in each case there can be a tremendous amount of uncertainty in how a given policy or management activity in an upslope or upstream area will affect aquatic resources. This uncertainty stems from the wide variability of site conditions, the variability in how a given activity is carried out, the uncertainty with respect to future storm events, and the inability to adequately characterize all of the controlling processes and site factors. The uncertainty increases as one attempts to predict the effects of multiple activities over space and time, and how the effects of these activities are accumulated and transmitted downstream (MacDonald 2000). The implication is that it can be very difficult to predict accurately the effects of policies and management activities at the watershed scale.

Given this uncertainty, regulators and land managers are increasingly turning to an adaptive management process (Gray 2000). Adaptive management refers to the iterative process of initiating one or more sets of activities, monitoring the effect of those activities on the resource(s) of concern, and then adjusting management actions in response to any observed change (Walters 1986) (Figure 9).

Adaptive management implicitly assumes that (1) there is a direct linkage between the management actions being implemented and the resource of concern; (2) any adverse change in the resource of concern can be detected within a reasonable time frame; (3) if an adverse change is observed, a change in management can be rapidly implemented; and (4) the management change will rapidly lead to the desired change in the resource of concern. Adaptive management is a very effective process for situations that meet these criteria, and examples can range from managing one's bank balance to weed control in a newly established forest plantation. However, the use of adaptive management at the watershed scale can be much more problematic because these four criteria may not be easily satisfied (Figure 9).

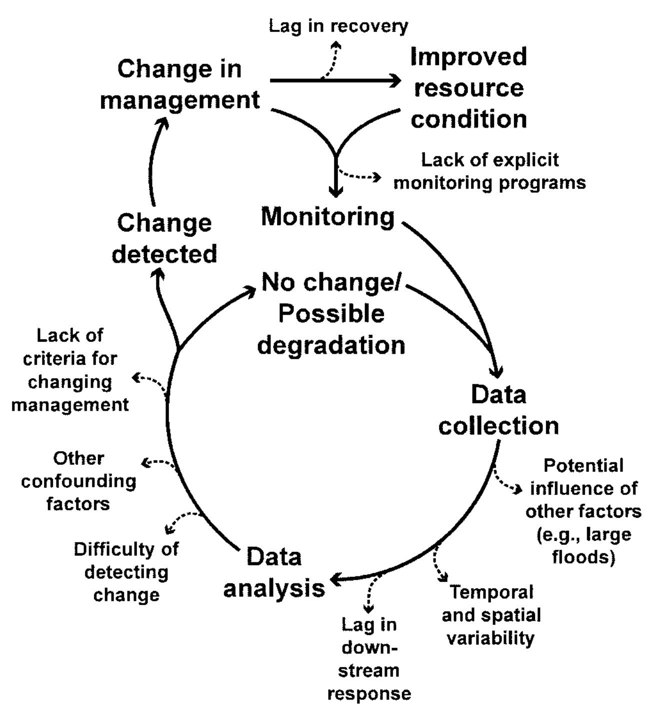

Figure 9. Schematic of the adaptive management cycle (solid arrows). Dashed arrows indicate potential breaks or failures.

The first assumption presumes that there is a direct linkage between the management actions being implemented and the resource of concern. This linkage is often relatively clear at the headwater scale, but as watershed scale increases there is an increase in the number of management activities as well as the number and complexity of the controlling processes. It therefore becomes increasingly difficult to link a given management activity to the condition of a downstream resource (Figure 9). For example, a decline in anadromous fish populations could be due to a degradation of spawning habitat and lower fry emergence rates, a degradation of rearing habitat and a reduction in escapement, competition with introduced species, changes in streamflow, the presence of dams or other migration barriers, an increase in mortality while the fish are in the ocean, overfishing, climate change, disease, or a combination of these and other factors. Even if one can document that an increase in fine sediment is causing a decline in spawning habitat, one must then determine the relative role of each possible sediment source, and identify which management actions need to be altered and where. If the increase in fine sediment can be attributed to an increase in bank erosion, for example, one must then determine whether the increase in bank erosion is due to an increase in the size of peak flows, an exceptionally large storm event, a decrease in bank vegetation, or some combination of factors. In most cases there will not be one obvious cause, and this makes it very difficult to definitively link particular management actions to the condition of a downstream resource. The difficulty of making such linkages increases almost exponentially with increasing spatial scale due to both the increase in the number of processes that must be considered and the increase in the number and type of management activities.

The second implicit assumption is that change can be detected within a reasonable time frame. The first problem with this assumption is that monitoring data usually are very limited because there are so few resources devoted to monitoring. Yet the simple truism is that, "If you aren't monitoring, you aren't managing." (S. Swanson, University Nevada at Reno, personal communication, 1992). In other words, if one isn't qualitatively or quantitatively evaluating the effects of one's actions, it's difficult to claim that one is actually managing a particular resource or activity. It clearly is not possible to monitor everything everywhere all of the time, so monitoring programs have to be carefully designed and focused (MacDonald et al. 1991). The development and implementation of a monitoring program also can be hindered by the lack of rigorous methods and criteria for evaluating key resources, such as the quality of spawning habitat, the quality of rearing habitat, or the amount, location, and size of large woody debris in different environments and channel types. The different stakeholders and resource managers must also agree a priori on what type and magnitude of change is needed to trigger a specified change in management. In the absence of an explicit monitoring program and explicit criteria for decisionmaking, adaptive management is not a viable option (Figure 9).

The second problem is that significant change may be very difficult to detect. In most cases significant change is defined by a P value of 0.05, which means that there is less than a 5% chance that a difference is due to chance. For adaptive management to be an effective strategy, one must be able to detect rapidly an adverse change so that management can be adjusted (Figure 9). One also wants high power, which is the likelihood of detecting change when there really is a change. Conceptually, the minimum detectable change depends on the magnitude of the change due to management activities relative to the temporal variability, measurement uncertainty, specified level of significance, and desired power (Figure 10). It is beyond the scope of this article to discuss quantitatively each of these components, but managers typically underestimate the magnitude of natural variability and overestimate their ability to detect a statistically significant change.

The difficulty of detecting change can be illustrated by calculating how many years of annual sediment data are needed pre and post-treatment to detect a specified percentage change. In undisturbed basins the typical coefficient of variation (CV, where the CV is equal to the standard deviation divided by the mean) for annual sediment yields is 70 to 100% (Bunte and MacDonald 1999). If we ignore the relatively large measurement uncertainty and assume the CV to be 100%, a 0.05 level of significance, and a power of

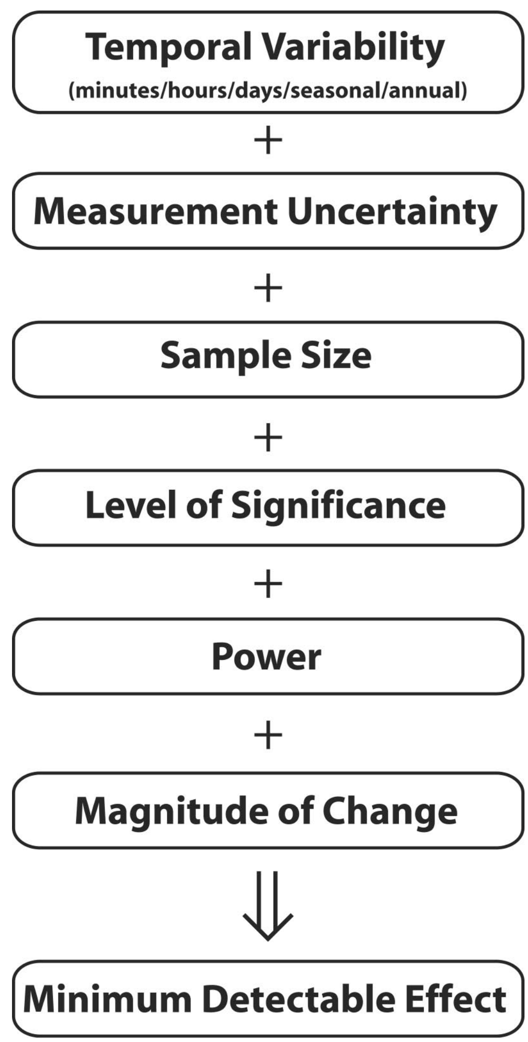

Figure 10. Schematic of the various factors that control the minimum detectable effect.

80%, one needs 17 years of pretreatment data and 17 years of posttreatment data to detect a 100% increase in annual sediment yields (Loftis et al. 2001). With only 5 years of pretreatment and 5 years of posttreatment data, the minimum change that can be detected under these same conditions is a 250% increase. Change can be detected more rapidly if some of the variability in the parameter of interest can be accounted for by another variable, such as annual precipitation or the sediment yields from a comparable control watershed (Loftis et al. 2001).

Because most resources are highly variable over time and space, resource managers can expect to detect only relatively large changes within any reasonable length of time (Loftis et al. 2001). Current environmental regulations and best management practices (BMP) were developed in response to the more obvious declines in resource condition, and as regulations become more stringent we are trying to detect progressively smaller changes. The implication is that it will become increasingly difficult to detect significant changes at the watershed scale, and to link these changes to specific management actions. An important caveat on the adaptive management process is that the inability to detect a significant change does not necessarily mean that no change has occurred, as many monitoring programs have relatively low statistical power.

The third assumption underlying adaptive management is that the resource of concern will respond rapidly to a change in management. The validity of this assumption will vary widely with stream type, the type of change imposed by management actions, and the resource of concern (Montgomery and MacDonald 2002). Some water quality parameters, such as turbidity, change rapidly in response to a storm event, and these changes are rapidly transmitted downstream. However, many decades may be required before a change in riparian management affects the amount of LWD in a stream channel. A long lag in resource response will preclude the use of adaptive management, or result in an extended period of adverse effects before there is a change in management and a positive effect on resource condition (Figure 9).

Watershed scale also can affect the magnitude and timing of a resource response. The overall tendency is for downstream resources to become less responsive to management changes as watershed size increases because of the greater potential for time lags (storage), dilution, and uptake or transformations (Table 1). As one example, the deposition of sediment in a first-order channel due to a road crossing may exhibit little or no time lag between a change in management (e.g., armoring a fillslope or installing waterbars just before the crossing) and a decrease in the amount of sediment being deposited. The initial deposit also may be rapidly removed by high flows. In larger watersheds there is more likely to be a substantial time lag between efforts to decrease sediment inputs in headwater streams and the amount of sediment in a downstream location.

These examples suggest that resource response times depend on at least three potential time lags. The first time lag is the rate at which the causal process recovers. If the causal process is surface erosion from unpaved forest roads, erosion can be rapidly reduced by building new waterbars or paving. However, if the problem is an increase in stream temperature due to a loss of riparian canopy or a change in the size of peak flows due to forest harvest, the time lag for recovery will probably be quite slow because it depends on the rate of forest regrowth. Bank erosion may recover relatively quickly if it is due to overgrazing, but not if it is caused by an increase in the size of peak flows.

The second time lag is the time needed to transport the constituent of concern to the location of interest. This time lag can be negligible in the case of peak flows or certain parameters such as turbidity, but much longer in the case of coarse-textured sediment. The third lag is the amount of time needed for the resource to recover. Macroinvertebrates can quickly recover from a toxic chemical spill if there is a healthy community upstream, but the recovery of a resident fish population may take several generations. If a possible consequence of management is the extinction of a desired species or population, adaptive management is not a viable option.

Each of these time lags and potential consequences must be evaluated to determine whether adaptive management is a viable approach (Figure 9). In general, all three of these time lags must be relatively short for adaptive management to be effective in minimizing resource damage. However, if society is willing to accept some resource degradation, adaptive management may be a viable option across larger time and space scales. For example, the progressive development and application of forestry-related BMP in the Pacific Northwest has been an adaptive management process. The problem is that each iteration of this adaptive management cycle required a certain amount of resource degradation to trigger another round in the development and imposition of BMPs. The acceptability of this degradation is a political and social issue, but one also has to consider the extent to which this degradation is reversible. Certain management activities are largely irreversible, such as urbanization or the construction of high-value infrastructure projects (e.g., highways, dams, or ski resorts). The listing of numerous salmonid populations in the Pacific Northwest under the Endangered Species Act indicates a failure of at least some of the regulatory systems designed to protect these populations. Both failures and successes must be critically examined to determine the conditions under which adaptive management can be an effective strategy for protecting aquatic resources.

Conclusions

The hillslopes draining to headwater streams account for most of the catchment area and generate the majority of the streamflow. In most cases the streamflow from headwater channels is efficiently routed to downstream areas. This means that management-induced changes in streamflow will accumulate downstream, but the changes in peak flows at the headwater scale generally will be diminished in the downstream direction because of dispersion, dilution, storage, and desynchronization. Similarly, turbidity, fine sediment, dissolved organic carbon, and particulate organic matter can be readily transported downstream, and for these constituents it may be possible to directly link a change in headwater inputs to downstream conditions.

There is a much weaker link between upstream inputs and downstream fluxes for coarse sediment, large woody debris, nutrients, and stream temperatures. The poorer connectivity is due to a combination of in-channel storage, biological uptake, chemical and physical transformations, dilution, and physical, chemical, and biological breakdown. The relative importance of these in-channel processes increases with increasing watershed scale, so it becomes increasingly difficult to link a specific change in headwater streams to downstream resource conditions as watershed size increases. The linkage between headwater streams and downstream conditions also is complicated by the high temporal and spatial variability in the delivery of materials from hillslopes into headwater streams, and from headwater streams to downstream reaches.

The connectivity between headwater and downstream reaches directly affects the extent to which adaptive management can be applied at the watershed scale. The viability of adaptive management at the watershed scale depends on the ability to rapidly detect change, a strong linkage between management actions and instream conditions, the time lags between management actions and the condition of a given resource, and the reversibility of adverse change. The complex and variable linkages between headwater streams and downstream areas suggest that these conditions will be rarely satisfied in larger watersheds, and this will limit the usefulness of adaptive management at the watershed scale.

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