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Hydrogeomorphic linkages of sediment transport in headwater streams, Maybeso Experimental Forest, southeast Alaska

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Abstract:

Hydrogemorphic linkages related to sediment transport in headwater streams following basin wide clear-cut logging on Prince of Wales Island, southeast Alaska, were investigated. Landslides and debris flows transported sediment and woody debris in headwater tributaries in 1961, 1979, and 1993. Widespread landsliding in 1961 and 1993 was triggered by rainstorms with recurrence intervals (24 h precipitation) of 7·0 years and 4·2 years respectively. Occurrence, distribution, and downstream effects of these mass movements were controlled by landform characteristics such as channel gradient and valley configuration. Landslides and channelized debris flows created exposed bedrock reaches, log jams, fans, and abandoned channels. The terminus of the deposits did not enter main channels because debris flows spread and thinned on the unconfined bottom of the U-shaped glaciated valley. Chronic sediment input to channels included surface erosion of exposed till (rain splash, sheet erosion, and freeze–thaw action) and bank failures. Bedload sediment transport in a channel impacted by 1993 landslides and debris flows was two to ten times greater and relatively finer compared with bedload transport in a young alder riparian channel that had last experienced a landslide and debris flow in 1961. Sediment transport and storage were influenced by regeneration of riparian vegetation, storage behind recruited woody debris, development of a streambed armour layer, and the decoupling of hillslopes and channels. Both spatial and temporal variations of sediment movement and riparian condition are important factors in understanding material transport within headwaters and through channel networks. Copyright © 2004 John Wiley & Sons, Ltd.

KEY WORDS headwater streams; landslides; debris flows; debris fans; bedload transport and storage; riparian vegetation; southeast Alaska

INTRODUCTION

Various hydrological processes and topographic attributes govern hydrogeomorphic linkages in headwater systems. The frequency of storm events influences the timing, magnitude, and spatial distribution of runoff processes, and resulting sediment transport and deposition from hillslopes to channels and from headwaters to downstream reaches (Dunne, 1991; Dhakal and Sidle, 2004). Subsurface and groundwater hydrology alter the stability of hillslopes and zero-order basins (Tsukamoto *et al*., 1982) and thus relate to the initiation of landslides, which are dominant geomorphic processes in the Pacific Northwest (Sidle *et al*., 1985). Sediment produced by landslides is transported to downstream reaches as channelized debris flows and bedload. Transport distances of sediment depend on tributary junctions, channel gradient, valley configuration, and formation of log jams (Nakamura, 1986; Benda and Cundy, 1990, Hogan *et al*., 1998); these, in turn, affect

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sediment transport and storage throughout channel networks (Wolman, 1977; Benda and Dunne, 1997a, b; May and Gresswell 2003).

Bedload and suspended sediment transport in steep-gradient headwater streams depends significantly on the occurrence of mass movement and their recovery processes (Grant and Wolff, 1991). Sources of bedload and suspended sediment are small bank failures, bank erosion, root throw, dry ravel, surface erosion due to rain splash, sheet erosion, and soil creep (Dietrich and Dunne, 1978). Availability of sediment is modified by changes in vegetation coverage and riparian stand regeneration after mass movement disturbance (Gomi *et al*., 2001). Vegetation cover in scour zones of landslide and debris flow tracks reduces the sediment delivery from hillslopes to streams (Shimokawa, 1984; Nakamura *et al*., 2000). Regenerated riparian stands contribute to the recruitment of woody debris; such woody debris creates sites for sediment storage (Gomi *et al*., 2001).

Timber harvesting and logging roads affect the hydrogeomorphic linkages in headwater streams. The probability of mass movement on marginally stable sites increases because root strength decreases 3 to 15 years after timber harvesting (Sidle *et al*., 1985). Roads intercept precipitation and subsurface flow from hillslopes, and thus modify water flow pathways (Sidle *et al*., 2004). Because of topographic changes, debris flows are both intercepted and initiated at logging roads (Wemple *et al*., 2001). Logging slash provides temporary sites for storage of sediment in small headwater streams after clear-cutting (Gomi *et al*., 2001; Sidle *et al*., 2004). In contrast, decreases in the number of woody debris dams due to riparian vegetation removal increase sediment movement (Likens and Bilby, 1982).

Spatial and temporal variation in sediment transport occurs due to changes in the amount of sediment and woody debris related to timber harvesting, the occurrence of mass movement, and the vegetation recovery processes (e.g. Shimokawa, 1984; Grant and Wolff, 1991; Benda and Dunne, 1997a). Wolman (1977) pointed out the importance of linkages among source areas, storage, and transport for understanding sediment routing processes in channel networks. Sediment transport from hillslopes to channels and in-channel storage have been extensively studied for estimating dynamics of materials and impacts of timber harvesting on stream ecosystems in British Columbia, in Oregon, in Washington, and in Japan (e.g. Dietrich and Dunne, 1978; Nakamura *et al*., 1995; Benda and Dunne, 1997b; Slaymaker, 2000). Although sediment transport due to landslides and debris flows has been investigated in southeast Alaska (e.g. Swanston, 1970; Johnson *et al*., 2000), few studies have documented sediment transport and storage within headwater streams and from headwaters to downstream reaches of U-shaped glaciated valleys. In particular, sediment movement in such landforms may be distinctive in terms of linkages from hillslopes to streams and from headwaters to downstream reaches compared with unglaciated terrain. Indeed, because hillslopes, zero-order basins, and channels are tightly coupled in headwater streams (Sidle *et al*., 2000; Gomi *et al*., 2002), timber harvesting and mass movement strongly affect sediment supply, transport, and storage.

Thus, this study focuses on documenting sediment routing processes, such as transport and storage in headwater systems, particularly related to the occurrence of mass movement and vegetation recovery after timber harvesting in a southeast Alaskan watershed. The objectives of this study are: (1) to describe sediment transport and storage due to episodic mass movement and more regular bedload and suspended sediment movement; and (2) to demonstrate hydrogeomorphic linkages between headwater streams and downstream systems for various sediment movement conditions.

STUDY SITES

This study was conducted in the Maybeso Experimental Forest within the Tongass National Forest, on Prince of Wales Island, southeast Alaska (Figure 1). The drainage area of the Maybeso watershed is 39 km^2. About 25% of the watershed (10 km^2) was clear-cut logged from 1953 to 1957 using cable yarding methods. Riparian corridors along the main stem of Maybeso Creek and its tributaries were also logged during that period (Bryant, 1980). Logging residue was not removed from streams and slash burning was not conducted after harvesting.

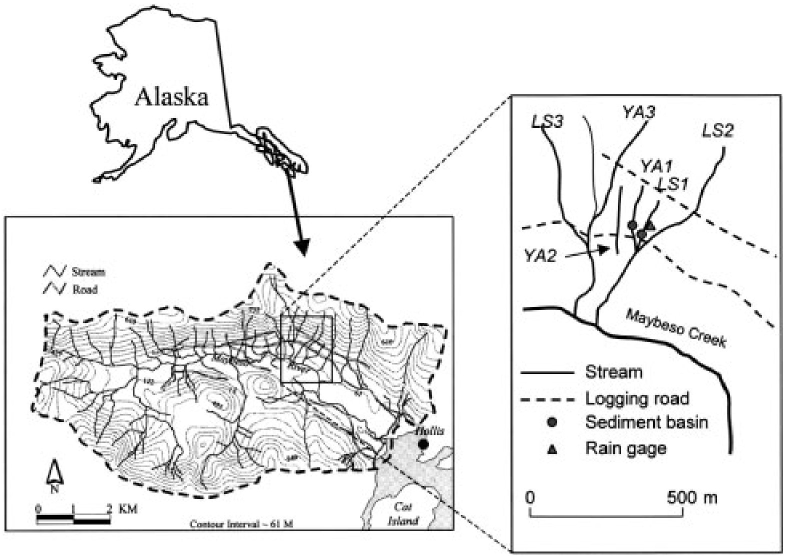


Figure 1. Study sites in Maybeso valley, southeast Alaska: weirs were located in LS1 and YA1 channels to capture bedload sediment. A tipping-bucket rain gauge was located near LS1 and the long-term weather station is located in Hollis. Broken line shows watershed boundary. See Table I for definition of the stream types

The Maybeso Experimental Forest was established in 1982 to monitor the effects of watershed-wide logging on soil erosion, fish and wildlife habitat, and forest stand succession.

The climate in this area is cool and temperate. The mean annual temperature is 10 °C and the mean annual precipitation is 2800 mm. The basin is characterized by a U-shaped glaciated valley covered by a veneer of clay-rich till of varying thickness below 500 m elevation (Swanston, 1970). Major geologic units are metasedimentary mudstone, greywacke, shale, slate, diorite, and granodiorite (Johnson *et al*., 2000). Tolstoi soils dominate 90% of the Maybeso watershed (above 500 m in elevation); at lower elevations, below 500 m, Karta soils are found. The Tolstoi mapping unit is comprised of shallow, well-drained soils developed on steep slopes, covered with colluvium and underlain by fractured bedrock. Karta soils are podzols developed on impermeable, unweathered glacial tills (Swanston, 1967). Dominant riparian forest vegetation includes western hemlock (*Tsuga heterophylla*), Sitka spruce (*Picea sitchensis*), western red cedar (*Thuja plicata*) and red alder (*Alnus rubra*). Occurrence of alder in riparian zones is highly influenced by past disturbances, such as landslides and debris flows. Resident fish in the lower reaches of headwater streams include cutthroat trout (*Oncorhynchus clarki*) and Dolly Varden (*Salvelinus malma*).

Streams were selected based on disturbance classes related to the history of mass movement and riparian stand conditions. Young-growth alder [mean diameter at breast height (DBH): 0·2 m] dominated riparian zones in streams affected by mass movement in 1962 (one case in 1979); these streams were identified as YA. Small alder stands (mean DBH: 0·03 m) or partially bare slopes were observed in riparian corridors affected by 1993 landslides; these streams were identified as LS (Table I; Figure 1). Three headwater streams were investigated in both LS and YA channels in 1998 and 1999. All streams in LS and YA were affected by timber harvesting, but they had different histories of mass movement (Table I). All streams were incised <5 m into the sidewalls of the U-shaped glaciated valleys. Material along stream sides and channel heads immediately mobilized due to landslides. Subsequently, channelized debris flows transported and deposited in the lower gradient reaches of both the LS and YA channels. Based on field observations of scour and deposition, both LS and YA channels were divided into upper (scour) and lower (deposition) sections (see figure 2 in Gomi *et al*. (2000)). Scour zones constituted both sources and run-out zones of sediment because of the difficulty of distinguishing these features. The YA1 and LS2 channels joined at the lower reaches of LS1, and the

YA3 channel joined to the lower reach of LS3 (Figure 1). YA2 did not join any other channels because it disappeared in the debris flow deposits downstream. Mean gradients of the headwater channels ranged from 9 to 45%, and mean bankfull widths ranged from 0·6 to 3·7 m. The drainage area at the downstream end of the study reaches ranged from approximately 0·14 to 0·35 km^2 (Table I).

METHODOLOGY

The occurrence and distribution of mass movement, such as landslides and debris flows, after timber harvesting was documented through field surveys and by examining aerial photographs taken in 1959 (1 : 15 840), 1962 (1 : 15 840), 1980 (1 : 12 000), and 1998 (1 : 63 360). Unvegetated slope areas that were formed by mass movement were mapped and counted. Scour and deposition zones of six streams in the LS and YA channels were intensively investigated in the field. Stream channels, abandoned channels, and sediment deposition along channels, such as debris fans and terraces, were mapped using a tape and compass. The length of the headwater reaches surveyed varied from 100 to 400 m, depending on the length of scour and deposition zones (Table I). The reference point (0 m point) of the stream survey was the boundary between the scour or run-out and deposition zones of mass movement. Stream gradient and cross-sectional profiles were surveyed using a level, stadia rod, and tapes. Exposed bedrock length and its position in the channel were also measured. The wetted and bankfull widths of streams were estimated every 5 m. Bankfull width was defined by the presence of moss and rooted vegetation along the channel margins and the top of banks.

Locations, number, and volumes of large woody debris (LWD) pieces (diameter $\geq$0·1 m, length $\geq$0·5 m) and numbers of fine woody debris (FWD) pieces (diameter 0·03–0·1 m, length $\geq$0·5 m) were measured and calculated using methods described by Gomi *et al.* (2001). Fine organic debris (FOD: accumulations of leaves and branches) was also measured if sediment was stored behind these features. Sediment storage behind woody debris and other obstructions (e.g. boulders and bedrock) was measured based on the geometry of sediment wedges (width, length of the wedge, and average depth at the front of the wedge). Average depth of the sediment wedge was measured using a sediment probe. The cause of sediment deposition was categorized according to the formation elements of the sediment wedge: LWD, FWD, FOD, boulders, and bedrock. The

Table I. Characteristics of the study sites[a]

	Landslides and debris flows	Drainage area (km^2)	Length of sampled reach (m)	Mean channel gradient[b] (%)	Mean bankfull width[b] (m)	Exposed bedrock (% length)	D_{50} (mm)
LS1 upper	1993/1979	0·21	340	40·3 (11·0)	1·2 (0·6)	26·7	30
LS1 lower			400	9·9 (5·8)	2·6 (1·5)	0·0	25
LS2 upper	1993	0·27	150	31·0 (12·3)	1·6 (0·4)	40·7	31
LS2 lower			250	13·7 (6·0)	1·8 (0·6)	0·0	30
LS3 upper	1993	0·35	300	31·7 (11·5)	2·8 (1·3)	59·8	43
LS3 lower			350	9·8 (8·1)	3·7 (1·3)	0·0	29
YA1 upper	1961	0·22	125	36·6 (9·3)	1·1 (0·5)	26·0	36
YA1 lower			100	17·5 (6·8)	1·4 (0·5)	2·0	35
YA2 upper	1961	0·14	225	42·7 (9·1)	0·9 (0·3)	38·0	40
YA2 lower			125	18·6 (7·1)	0·9 (0·3)	0·0	27
YA3 upper	1979/1961	0·21	150	28·9 (11·2)	2·0 (0·6)	47·1	38
YA3 lower			300	17·4 (7·3)	1·9 (0·6)	0·0	28

[a] LS, recent landslide (logged from 1953 to 1957 and landslide and debris flows in 1979 and (or) 1993); YA, young alder riparian forest (logged from 1953 to 1957 and landslides and debris flows in 1961 and (or) 1979). Sites are divided into scour/run-out (upper section) and deposition (lower section) zones of landslides/debris flows.
[b] Standard deviations are given in parentheses.

volume of sediment stored behind woody debris and other obstructions was computed based on a rectilinear pyramid (Gomi *et al.*, 2001).

The age of sediment deposits was estimated from even-age stands of alder growing in these scour and deposition zones (e.g. Nakamura, 1986). Three cross-sections in each stream were surveyed to describe valley and channel profiles. The size of mobile streambed material was measured based on the median diameter of 100 pebbles in a $0{\cdot}2 \times 0{\cdot}2$ m^2 grid at the three cross-sections (Wolman, 1954). Large cobble and boulder components ($>0{\cdot}2$ m) were excluded because of their relative immobility, except during landslides and debris flows. Drainage areas of study streams were measured from a topographic map (USGS Crag C-3) using a digital planimeter.

Bedload and suspended sediment movement was monitored in the LS1 and YA1 channels from September to November 1999. Most of the sediment transport in southeast Alaska occurs in the fall storm season from September to December (Sidle, 1988; Sidle and Campbell, 1985). Therefore, characteristics of sediment movement and relative differences in sediment transport between LS1 and YA1 channels can be evaluated. V-notch weirs (120°) were installed in LS1 and YA1 in the summer of 1999 to measure discharge and sediment yield. Sediment deposited behind the weirs was sampled when significant volumes accumulated after storm events. Sampled sediment was sieved into size classes of 11·2, 16·0, 25·0, 32·0 45·0, and 115·0 mm, and weighed in the field. Sediment smaller than 11·2 mm was taken back to the laboratory, dried, and then weighed for smaller sieve classes (4·75, 2·00, 1·00, and 0·425 mm). Sub-samples of finer sediment ($<11{\cdot}2$ mm) were burned for 2 h at 500 °C to remove organic matter in each sieve class. The bulk density of the sampled sediment was 1·2 g cm^{-3}.

Suspended sediment samples were collected during selected storm events in the LS1 and YA1 channels with an automated pumping sampler. The intake of the pumping sampler was attached to a metal rod and suspended in mid-stream. This minimized the effects of flow separation and clogging by organic matter for sampling efficiency (Sidle and Campbell, 1985). Sampling was automatically activated at a predetermined storm stage. A 500 ml sample was collected every 10 to 15 min. Because only 24 sampling bottles were contained in the sampler, suspended sediment samples were typically only collected during the rising limbs of storms. Samples were transported to the Hollis field station (Figure 1) where they were analysed. Total suspended solids (TSS) were determined by passing a known volume of stream water through a glass-fibre filter with an effective retention of 1·2 μm; filters were oven dried at 100 °C and residual sediment was weighed. Suspended sediment, the mineral portion of the sample, was weighed after the TSS sample was burned at 550 °C for approximately 2 h (Sidle and Campbell, 1985). Precipitation was also monitored every 10 min using a tipping-bucket rain gauge at an open-canopy area near the LS1 channel (Figure 1). Daily precipitation was collected at Hollis from 1947 to 1994, although data are missing from 1951 to 1952 and from 1963 to 1987.

RESULTS AND DISCUSSION

Occurrence and distribution of mass movement

Both the occurrence and spatial distribution of mass movement processes are important for sediment transport in headwater streams and the downstream linkages of sediment transport. Hillslope gradient, distribution of glacial till, bedrock structure, and faulting are important factors affecting mass movement initiation in Maybeso Experimental Watershed (Swanston, 1967). Based on topographic maps, most of the landslides initiated on slopes $>30°$. Slides mainly occurred in the elevation range from 450 to 550 m, the upper boundary for the distribution of glacial till. Most landslides occurred along the north slope of Maybeso Experimental Watershed; a few landslides were found on the south slope (Figure 2). Swanston (1967) noted that the north slope of Maybeso was more extensively faulted than the south slope; this affects differences in occurrence of mass movement between slope aspects. In addition, hillslopes were relatively steeper on the north compared with the south side of the watershed. More landslide scour occurred on the east side than on the west side of the logged portion of the north slope (Figure 2). Since bedrock and slope gradient were similar on both sides of the north slope, differences in the distribution and compaction of glacial till may alter the occurrence of landslides.

More of the scour and deposition related to mass movement after timber harvesting occurred in the logged area than in the unlogged area of the Maybeso watershed (Figure 2). Despite the absence of canopy cover in the logged area, there was little evidence of mass movement in most of the headwater channels in 1959; only 14 events were detected on the unvegetated slope using aerial photographs (Figure 2). Following the storm events in 1961, scour and deposition from landslides (number of events on unvegetated slope areas: 30) and subsequent debris flows were clearly visible in the 1962 photographs. Multiple landslides and large depositional areas along channels were observed in deeply incised (>15 m) landforms (Figure 2). Single landslides and subsequent debris flows were found in headwater systems with shallowly incised (<15 m) landforms.

Scour and deposition from landslides (number of events in unvegetated slope areas: 38) observed in the 1980 photographs were associated with a 1979 storm event (Figure 2). Because the riparian vegetation imparted an element of roughness along the lower reaches of headwater streams, the travel distances of debris flows in 1979 may be shorter than for the events in 1961 (Johnson *et al.*, 2000). Of the three mass erosion episodes covered in this study, the 1993 event produced the smallest number of landslides (number of events in unvegetated slope areas: 18) and debris flows, as observed in the 1998 small-scale photographs. Despite the frequent mass movements after timber harvesting, no landslides and debris flows directly reached the main stem of Maybeso Creek (Figure 2). Because resolution, scales, and vegetation coverage differed among the photographs, the minimum detectable size of landslides on photographs may differ. Therefore, the distribution and number of landslides based on photographic interpretation may be conservative.

The sizes of storms are significantly related to the occurrence of mass movement (Dhakal and Sidle, 2004). Landslides in southeast Alaska typically occur during the autumn storm season from September to December (Swanston, 1967; Sidle, 1984; Johnson *et al.*, 2000). The amounts of 24 h precipitation on 14 October 1961 and on 27 October 1993 were 128·4 mm and 116·7 mm respectively; the third and fifth highest values of an incomplete 20 year record. Because average soil depth was 0·96 m in hillslopes and zero-order basins,

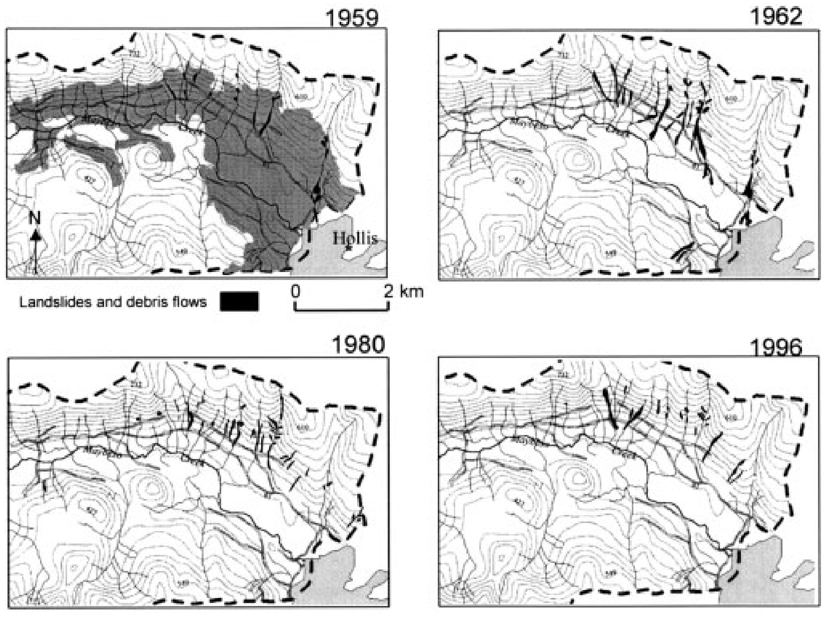


Figure 2. Distribution of landslides and debris flows after timber harvesting in the Maybeso watershed. Spatial distribution of mass movement was estimated from aerial photographs. Timber harvesting was conducted in 1953 and 1957. Most of the scour and deposition due to mass movement was observed within the logged area

soil within drainage depressions (i.e. zero-order basins) was believed to be fully or nearly saturated during the 1961 and 1993 storms, based on the relationship between piezometric head and 24 h precipitation in Maybeso Experimental Forest (Swanston, 1967). The relationship between rainfall return period T_r and 24 h precipitation P was estimated from precipitation data at Hollis using the Weibull plotting position formula:

$$P = 108{\cdot}7 \log(T_r) + 65{\cdot}6 \qquad (R^2 = 0{\cdot}87;\ p < 0{\cdot}001) \tag{1}$$

The return intervals for the 1961 and 1993 storm events were calculated as 7·00 years and 4·20 years respectively. This finding agrees with the landslide-triggering storm estimates (5 to 8 year recurrence interval based on 24 h precipitation) noted by Swanston (1970) for this area. Montgomery *et al.* (2000) also found that a 24 h rainfall event with a recurrence interval >4 years might initiate shallow landslides in the decade after logging in Oregon and Washington. Storms with recurrence intervals >2 years were sufficient to initiate landslides in Queen Charlotte Islands, British Columbia (Schwab, 1998). More importantly, however, the combinations of slope gradient and precipitation intensity, as well as antecedent moisture (Sidle, 1984; Dhakal and Sidle, 2004), govern the initiation of landslides.

Changes in vegetation and in geomorphic and hydrologic conditions due to timber harvesting and logging roads can modify slope stability. Because the soils in the Maybeso watershed are essentially cohesionless, roots anchor the soil to bedrock and provide a long-filament binder to soil substrate (Swanston, 1970). Therefore, reduced root strength in the period of 3 to 15 years after logging (Sidle *et al.*, 1985) significantly affected the stability of slopes at the time of the 1961 storm. The two highest 24 h precipitation amounts in Maybeso were recorded during and immediately after logging [31 December 1953 (264 mm) and 7 December, 1959 (178·2 mm)]. Although soil moisture in December is typically greater than in other seasons, these two extreme precipitation events did not trigger mass movements. This is probably because anchoring and reinforcing effects of roots were still contributing to stability of hillslopes. Because most of the landslides occurred just below mid-slope logging roads in 1961, unstable road-fill material and/or channelling of water from road surfaces, cut slopes, and ditches may have also contributed to landslide initiation (Schwab, 1998; Wemple *et al.*, 2001).

Scour, run-out, and deposition of mass movement

Channel gradient and valley configuration are important determinants of the location of scour, run-out, and deposition of mass movement. Following initiation by landslides, sediment and woody debris moved as channelized debris flows toward the valley bottom of Maybeso Creek. Distances of scour and run-out zones in the LS and YA channels ranged from 275 to 700 m (Table II). Within the scour and run-out zones, bedrock was exposed in 27 to 60% of the channel length (Table I). Landslides initiated on hillslopes with gradients ranging from 36° to 58°. Swanston (1970) noted that landslides commonly occurred on hillslopes at or near 37° in the Maybeso watershed. The sediment deposition zone occurred near the foot of the slope down to the main stem, where channel gradient was in the range of 4·9 to 7·5° (Table II). Depositional volume (sediment and woody debris transported by mass movement) was greater in LS sites (750 to 1871 m^3) than in YA channels (309 to 869 m^3; Table II). Because landslides and debris flows in LS2 and LS3 had longer scour and run-out distances and initiated in second-growth conifer forests, deposited materials contained greater amounts of sediment and woody debris.

Valley configuration sets the template for sediment transport and deposition. Debris flow materials were typically deposited at the base of the U-shaped glacial valley walls, because channel configuration of tributaries changes from one of narrow and steep to one of wide and gentle channels (Figure 3a). The decreased gradient reduced the momentum of debris flows. Wider valleys result in spreading and thinning of debris flows and increasing resistance to flow (Nakamura, 1986; Benda and Cundy, 1990). Separation of water and sediment due to percolation into alluvium may also reduce the momentum of debris flows. Debris flows encountered an old logging road at the foot of the hillslopes (Figures 3b and 4), thus the road played a trivial role in stopping debris flows. Such changes caused the formation of debris fans and log jams in the bottom of the

Table II. Sediment transport during mass movement and in-channel storages of sediment[a]

Site	Source and run-out zones			Deposition zones		
	Initiation area	Length (m)	Channel gradient[b] (degrees)	Total volume of sediment (m^3)	Channel gradient[c] (degrees)	Length (m)
LS1 upper	Road fill	350	47			
LS1 lower				750	5	100
LS2 upper	Second-growth forest	600	43			
LS2 lower				1045	5	85
LS3 upper	Second-growthforest	700	55			
LS3 lower				1871	5	300
YA1 upper	Road fill	330	52			
YA1 lower				309	8	150
YA2 upper	Below road	275	36			
YA2 lower				440	9	85
YA3 upper	Road fill	350	50			
YA3 lower				869	6	230

[a] see Table I for stream type codes.
[b] Hillslope gradient was measured near the initiation of landslide.
[c] Channel gradient measured at the beginning of deposition zones.

valley. Because the landslide in LS3 was initiated at a higher altitude, the resultant debris flow there may be transported farther (Table II). In contrast, debris flow material may deposit in steeper reaches of YA channels than in the LS channels because the landslides initiated at lower elevations in YA channels. No debris flows directly entered the main channel of Maybeso Creek due to a wide and flat valley floor, despite the frequent mass movements in headwaters (Figure 2). Roberts and Church (1986) noted that the amount of sediment transported from the hillslope to main channels due to debris flows depends upon the presence or absence of a valley flat in Queen Charlotte Islands.

Because streams were not entrenched into fans, several poorly sorted sediment deposits were observed at the foot of hillslopes (Figures 4 and 5). Debris fans formed prior to 1961 may also affect sediment deposition on the foot of hillslope. Owing to sediment and woody debris deposition in 1961, 1979, and 1993, debris fans formed in the transition zone from hillslopes to floodplains (on the logging road in Figures 4 and 5). Channels shifted dramatically due to the formation of debris fans and log jams. Small fans and bars were also found along channels and on terraces about 200 m downstream of the logging road in LS1 (Figure 4) and YA3 channels (Figure 5).

Sediment deposits behind in-channel obstructions (woody debris and boulders) 200 m downstream of the reference point (logging road) were greater than at the foot of hillslopes (near the reference point) in LS1 and YA3. Because channels in LS2 and YA1 eventually merged into the LS1 channel, sediment from LS2 and YA1 was transported to the downstream reaches of LS1. Such small deposits along channels, as well as behind obstructions in downstream reaches, may have been formed by sediment transported as bedload movement during the 1993 event or subsequent storms (Figures 4 and 5). The volume of each deposit along and within channels ranged from 1 to 10 m^3. Owing to such sediment movement, fan-shaped deposits formed along lower reaches of the headwaters. Sequential sediment movement also partially damaged riparian stands and caused local aggradation and side-channel formation in downstream reaches.

In-channel storage of sediment

Sediment produced during and after mass movement was stored in channels and fans. Total sediment stored behind LWD, FWD and other obstructions (boulders and bedrock) in stream channels ranged from 0·2

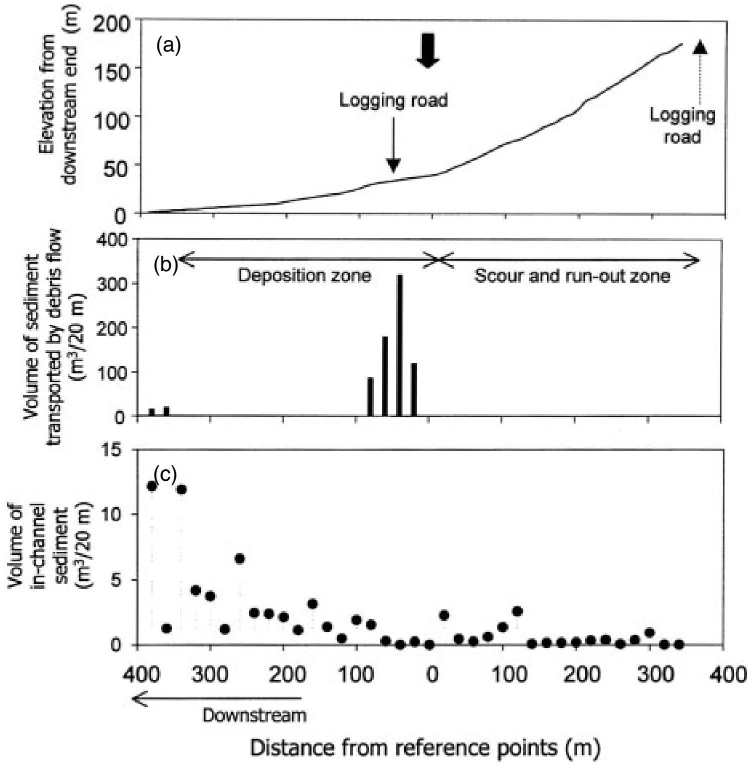
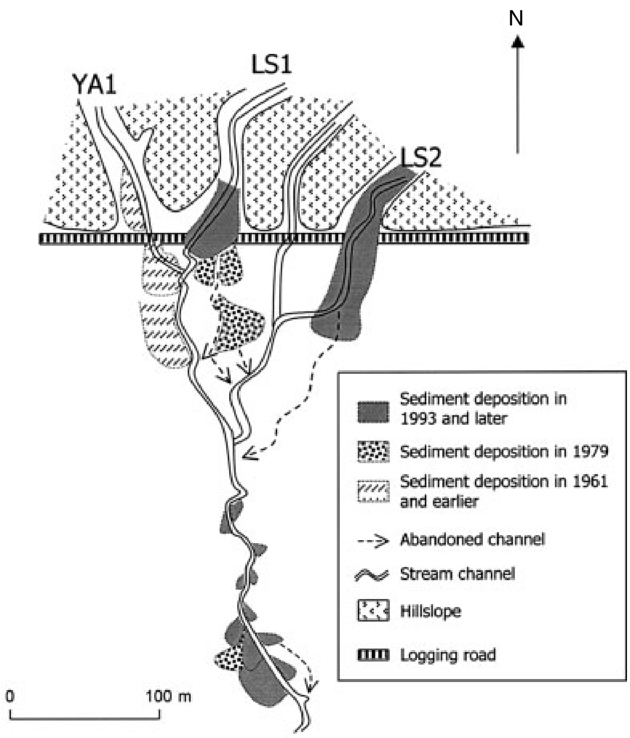


Figure 3. (a) Longitudinal profile along LS1. Small arrows show old logging roads located at the foot (solid arrow) and middle (dotted arrow) of hillslopes; large arrow shows the reference point (0 m) for the stream survey. (b) Estimated volume of sediment transported and deposited due to mass movement. (c) Volume of sediment stored behind woody debris and the other channel obstructions (e.g. boulders) due to debris flows

to 10·1 m^3 per 100 m (Table III; Figure 3c). Sediment stored in upper sections of the LS and YA channels was either remnants of mass movement deposits or was transported from hillslopes after recent mass movement. Sediment stored in the lower sections of LS and YA channels (Table II) was transported during the mass movements and subsequent events. Larger volumes of sediment were found in the lower sections of the LS channels than in other sections. Sheet erosion, rain splash, and creep within landslide scour zones of weathered till govern sediment supply from hillslopes to LS channels. Soil erosion pedestals developed on bank slopes. Bare soil and till were affected by frost-heaving during winter and, thus, produced much coarser sediment that was temporarily stored in channels. Eroded material on hillslopes was transported directly to streams. In contrast, evidence of surface erosion and direct connections of sediment transport from hillslopes to streams in the YA channels was lacking due to vegetation and litter cover. Alder roots penetrated into bedrock and till and expanded on exposed substrate, thus binding soil and organic matter.

Different histories of mass movement relate to different vegetation recovery processes and in-stream sediment storage in LS and YA channels. The volume of woody debris and in-stream function of woody debris characterized the differences between the LS and YA channels (Table III and Figure 6). Because both branches and leaf litter were recruited from regenerating riparian alder stands, the amount of LWD and FWD increased in YA channels 40 years after mass movement. The volumes of in-channel woody debris in YA channels (0·7 to 1·3 m^3 per 100 m) were greater than in LS channels (0·2 to 0·5 m^3 per 100 m; Figure 6). Sediment storage ratio, which was calculated as the total volume of sediment divided by sediment stored behind woody debris, was significantly higher in YA (0·55) channels than in LS channels (0·22) (Gomi *et al.*, 2001). Therefore, increases in woody debris pieces recruited from alder stands created channel roughness and



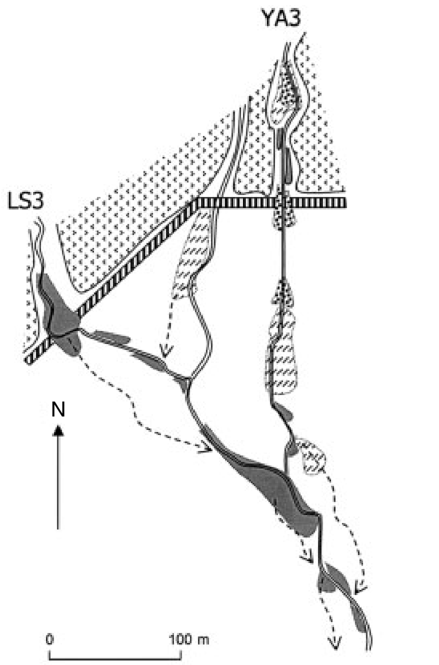
Volumes of sediment (m^3)	1961	1979	1993	downstream
LS1	-	548	160	
LS2	-	-	1045	> 42
YA1	309	-	-	

Figure 4. Sediment deposition due to debris flows and subsequent sediment movement in the LS1, LS2, and YA1 channels. Debris fans formed at the bottom of the glaciated valley near the logging road. The table shows the volume of mobile sediment in different years estimated in the field

turbulence and thus contribute to sediment storage (Gomi *et al*., 2001). Such sediment storage due to channel obstructions may be an important source for more regular sediment transport processes, such as bedload and suspended sediment (Roberts and Church, 1986). Because of woody debris recruitment and in-channel storage of sediment, the percentage of exposed bedrock was slightly lower in the YA channels (Table I). For much longer periods (>50 years), woody debris inputs and colluvial material are stored in hollows and channels until landslides and debris flows remobilize the material. Larger sizes of materials are likely stored in the channels because of the limited transport capacity in small streams.

Transport of bedload and suspended sediment

Bedload and suspended sediment transport in streams is linked to sediment supply from hillslopes and in-stream storage. Rain splash and sheet erosion on glacier till and forest soil produce mostly finer sediment, such as clay, silt, and fine sand, most of which is transported as suspended sediment. Small bank and slope failures typically produce coarser sediment and alter channel morphology. During the 21 October 1999 storm, a small bank failure occurred in LS1, 30 m upstream from the weir. According to Equation (1), the return interval of the 24 h precipitation (71 mm) was 1·5 years. Based on Swanston's (1967) equation, 71 mm



Volumes of sediment (m^3)	1961	1979	1993	downstream
LS3	-	-	1871	> 20
YA1	501	249	119	

Figure 5. Sediment deposition due to debris flows and subsequent sediment movement in the LS3 and YA3 channels. More sediment deposits due to debris flows in LS3 were found approximately 200 m downstream of logging roads. Sediment deposits of varying ages were found along YA3. The table shows the volume of mobile sediment in different years estimated in the field. See Figure 4 for the code on the map

of rainfall in 24 h precipitation would cause a 0·65 m rise in piezometric head. Thus, sections of shallow soil near the channel bank may have been nearly saturated during this storm. Nevertheless, bank and slope undercutting due to higher discharge may have played an important role in creating unstable conditions in the side-slopes. The small bank failure produced 1 to 1·5 m^3 of sediment, including soil and weathered till, based on dimensional estimates. Failed material consisted of weathered till (clay and silt) and coarse fragments (gravels and cobbles). Since the capacity of the weir pond in LS1 was less than 1 m^3, a maximum of 85 to 90% of the sediment was captured behind the weir. Because of the sudden supply of saturated sediment from the bank failure during the high discharge, the sediment was probably mobilized and transported downstream during the storm. Such a channelized mass wasting process with a large sediment flux and higher water content is described as a hyperconcentrated flow.

Sediment produced from hillslopes and stored in channels is subsequently transported downstream as bedload and suspended sediment, depending on the size of storm events, sediment supply, and channel roughness. The relationships between sediment deposition at weir ponds in LS1 and YA1 and maximum 24 h rainfall during the various sampling intervals from 1 September to 15 November 1999 are shown in Figure 7. Bedload sediment transport in LS1 was two- to ten-fold greater than in YA1. Significant exponential relationships were derived between the maximum 24 h precipitation P (mm) and the total volume of sediment V (m^3) in LS1 and in YA1 (Figure 7).

$$\text{LS1:} \quad V = 7{\cdot}8 \times 10^{-4}\ e^{0{\cdot}104P} \qquad (R^2 = 0{\cdot}90,\ p < 0{\cdot}001) \tag{2}$$

$$\text{YA1:} \quad V = 7{\cdot}4 \times 10^{-4}\ e^{0{\cdot}056P} \qquad (R^2 = 0{\cdot}66,\ p = 0{\cdot}008) \tag{3}$$

Table III. Characteristics of woody debris and sediment storage[a]

	Number of LWD per 100 m		Number of FWD per 100 m	In-channel storage of sediment (m^3 100 m)		
	Total	In-channel		Behind LWD	Behind FWD/FOD	Behind boulder and bedrock
LS1 upper	15	4	15	0·4	0·2	2·5
LS1 lower	40	29	106	2·8	1·6	10·1
LS2 upper	37	27	67	1·1	0·9	2·6
LS2 lower	54	45	71	3·3	0·9	3·8
LS3 upper	57	27	53	2·3	0·5	4·7
LS3 lower	80	46	116	4·1	0·2	9·6
YA1 upper	24	21	41	1·9	1·3	3·7
YA1 lower	52	37	81	7·9	0·8	0·6
YA2 upper	20	16	55	0·8	1·2	2·0
YA2 lower	14	14	48	4·5	0·2	7·8
YA3 upper	29	23	27	1·7	0·7	1·9
YA3 lower	66	59	105	4·1	0·4	3·4

[a] See Table I for stream type codes.

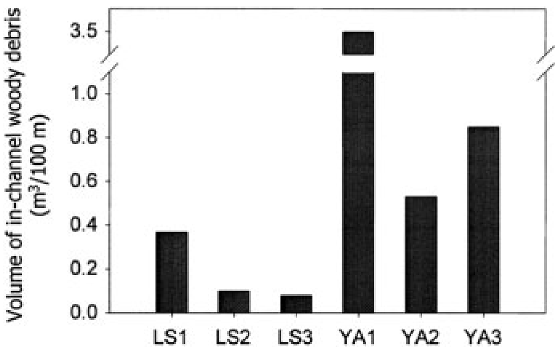


Figure 6. Volume of in-channel large woody debris in the upper reaches of LS and YA channels

The exponent in Equation (2) (LS1) is significantly different than in Equation (3) (YA1; $F = 5{\cdot}71$, $p = 0{\cdot}03$) at the $\alpha = 0{\cdot}05$ confidence level, whereas the constants are similar. For small rainfall events (e.g. 10 mm in 24 h), bedload movement was similar in both channels. During larger events, total bedload movement in LS1 was significantly greater than in YA1. In addition, the relationship between maximum 24 h precipitation and volume of sediment in LS1 was stronger ($R^2 = 0{\cdot}90$) than in YA1 ($R^2 = 0{\cdot}66$; Figure 7). These findings imply that bedload sediment response to rainfall input in LS1 was less complex than in YA1. This result may relate to the higher availability of sediment in the LS1 channel due to its constant supply from unvegetated bank slopes. Development of an armoured layer after disturbances in the YA channel also reduced sediment transport. Variations in bedload transport in YA1 during storms may be caused by sediment storage behind woody debris and boulders or the sudden release of sediment from such storage sites (Adenlof and Wohl, 1994). The median diameter D_{50} of the bedload was approximately 10 mm and was smaller than channel bed substrate (Table I). Although there was no replication of LS1 and YA1, we assumed that similar relations between bedload transport and rainfall might be observed in the other LS and YA channels.

Total sediment yield during the selected storm events was estimated from bedload and suspended sediment yields, although we could not collect suspended sediment samples during all phases of storm events. Total

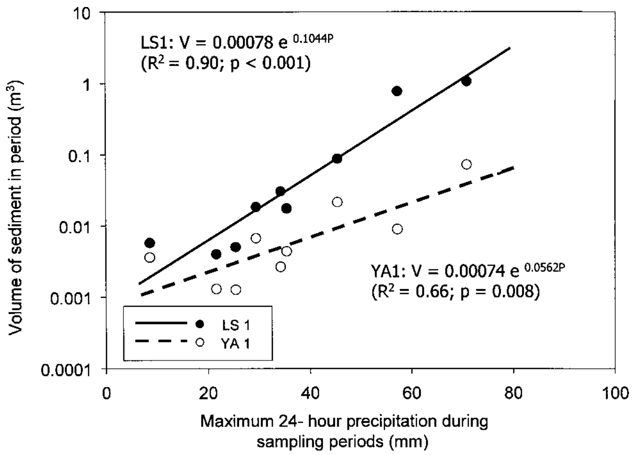


Figure 7. Relationships between the volumes V (m^3) of sediment deposited in weir ponds and maximum (event) 24 h precipitation P (mm) for sediment sampling periods in LS1 and YA1 from September to November 1999

suspended sediment yield from LS1 (19 to 207 kg per event) was 10- to 20-fold greater than from the YA1 (2 to 37 kg per event). Suspended sediment accounted for a smaller percentage to the total sediment yield in YA1 (26%, 32%, and 68%) compared with LS1 (49%, 67%, and 70%) during the storm events on 8 and 22 September amd 1 November respectively. Because less bedload was transported late in the storm season, suspended sediment comprised 68% of the total sediment yield in YA1 during the 1 November storm. Lenzi and Marchi (2000) estimated that suspended sediment contributed from 16 to 100% of the total sediment yield in a small stream in Italy. These large variations in the percentage of suspended fractions of total sediment load are affected by sediment sources, seasonal changes in availability of sediment, and the instantaneous supply from the slope and the channel bed (Sidle and Campbell, 1985; Lenzi and Marchi, 2000). Cumulative bedload yield during the 1999 monitoring period was 2·0 m^3 and 0·1 m^3 in the LS1 and YA1 channels respectively. Therefore, assuming an average proportion of suspended sediment to total sediment of 62% in LS1 and 42% in YA1 throughout the storm season, total sediment yields in LS1 and YA1 channels were approximately 5·3 m^3 and 0·2 m^3 respectively.

Sediment transport and hydrogeomorphic linkages of headwater streams

Temporal variation. Based on results of sediment transport and storage in the LS and YA channels of Maybeso watershed, temporal variation in sediment transport occurs along with changes in linkages from hillslopes to streams. These changes are related to mass movement and vegetation recovery processes. Ground vegetation cover on previously exposed soil and till, and the regeneration of alder stands in the riparian areas after mass movements modify soil erosion processes and linkages from hillslopes to streams (Shimokawa, 1984). After the regeneration of alder riparian stands, the nitrogen-supplemented soils may facilitate regeneration of other successional plants (Bormann and Sidle, 1990). Soil under alder riparian stands may be improved with the accumulation of organic matter, increases in flora and fauna (possibly contributing to the formation of macropores), development of soil structure, and increases in infiltration capacity and hydraulic conductivity (Bormann and Sidle, 1990). The partial decoupling of side-slopes and streams gradually occurs as slopes revegetate and consequently decrease the supply of sediment from hillslopes to channels in YA.

The role of woody debris related to the transport and storage of sediment may change with the succession of riparian stands after mass movement. Woody debris recruitment may change from early establishment of alder to the replacement of alder with conifers in later stages of succession. Succession patterns of riparian forests after timber harvesting differ with and without mass movement in headwater streams (Gomi *et al.*, 2001). Likens and Bilby (1982) noted that the succession of riparian stands after logging modified the amount of woody debris dams for a periods of 50 to 100 years after disturbances. Although woody debris recruited from alder riparian stands contributed to sediment storage 40 years after mass movement, alder typically decomposes and fragments more rapidly than conifers (Harmon *et al*., 1986).

Dominant sediment transport modes after timber harvesting in headwater streams of the Maybeso watershed can be grouped in relation to triggering storm events (Table IV). Landslides and debris flows after timber harvesting were triggered by storm events with recurrence internals $\geq$5 years. Relatively unsorted materials in large quantity ($\geq$100 m^3), including cobbles and boulders ($D_{50} \geq 200$ mm) and LWD pieces, were transported during these events. Hyperconcentrated flows may occur when large amounts of sediment are transported from the hillslope and accumulate in channels (Table IV). Although only one case was observed, approximately 1 m^3 of sediment from a slope failure was transported as a hyperconcentrated flow during a storm with a recurrence interval of 1·5 to 5 years. A flood surge may also occur if a log jam breaks (Nakamura *et al*., 2000). Gravel, cobbles, and small boulders ($10 < D_{50} < 200$ mm) and FWD pieces were transported during such intermediate events (Table IV). Smaller quantities of sediment ($\leq$1 m^3) were transported more regularly (daily rainfall return period $\leq$1·5 years) as bedload and suspended sediment (Table IV). Materials smaller than gravel, as well as organic matter (e.g. leaves and branches), were typically transported during these smaller events.

Spatial variation. A shift in the dominant geomorphic processes from up- to down-stream reaches within headwater streams causes various types of deposition and erosion features along channels. Sequential changes from colluvial- (mass movement) to alluvial-dominated (bedload and suspended sediment) processes occurred within 1 km in the Maybeso watershed due to changes in channel gradient, material size, and valley configuration. Such changes also modified riparian structure and channel form (Nakamura *et al*., 2000). Most of the sediment produced by the mass movement formed debris fans at the bottom of the U-shaped glacial valley. A portion of this deposited material was subsequently transported as flood surges, bedload, and suspended sediment due to break up of log jams, bank erosion, and channel avulsions on the debris fans, but most of deposited sediment remained in the same location for longer periods. For example, materials that entered from hillslopes to the bottom of valleys in Rock Creek, coastal Oregon, were stored on the valley floor for a long period (Dietrich and Dunne, 1978). Sediment stored in channels mobilizes more frequently depending on the stability of obstructions. Bedload sediment may gradually be transported from downstream reaches of debris fans to the main stem of Maybeso Creek. Suspended sediment is transported directly to the main stream of Maybeso Creek during storm events. The dominant channel-reach morphology changes from bedrock and cascade to step–pool and pool–riffle in headwater streams (Montgomery and Buffington, 1997).

Table IV. Mode of sediment movements in headwater streams in the Maybeso watershed

	Return periods in 24 h precipitation (years)	Total sediment volume (m^3)	Mobile material	Geomorphic changes
Landslides and debris flows	$\geq$5	$\geq$100	$D_{50} > 200$ mm, LWD	Fans and new channels form Channel aggradation Channel reach types rearranged
Hyperconcentrated flow flood surge	1.5–5	1–100	$10 < D_{50} < 200$ mm, FWD	Small fans, lobes, side channels form Minor steps collapse Local aggradation
Bedload and suspended movement	$\leq$1·5	$\leq$1	$D_{50} = 10$ mm, leaves and branches	Moderate channel erosion and deposition Local changes in pool depth

Sediment movement related to mass wasting and the recovery process alters the distribution of such channel reach types.

Although tight linkages from hillslopes to streams within headwater systems were observed, routing processes from the downstream portion of headwaters to the main channel of Maybeso Creek were not strongly coupled (Figure 8). Because of decreases in the channel gradient and increased valley width, sediment movement is more dispersed in downstream reaches. Unit stream power is lower in wider and shallower channels, and this induces sediment deposition. Stream water percolates into substrate in some reaches and, conversely, groundwater recharges through the streambed. Wide floodplains, beaver ponds, and bogs intercept and dissipate material transport from headwaters to downstream reaches (Figure 8). Bryant (1980) documented channel changes in Maybeso Creek before and after timber harvesting and found that log jam formation, bank instability, and sediment accumulation occurred after logging. Because mass movements did not directly enter the main channel (Figure 8), most of sediment in Maybeso Creek was likely the result of bedload transported from tributaries, bank erosion, and side-channel formation in floodplain materials. Schwab (1998) estimated that a greater amount of sediment was transported by mass movement in the early 1900s than after timber harvesting in the 1950s in Queen Charlotte Islands. Hogan *et al*. (1998) estimated that the most log jams were formed during and after mass movement. Such results may be related to linkages of sediment and woody debris movement between headwater streams and downstream systems at various spatial and temporal scales (Hogan *et al*., 1998; Slaymaker, 2000) (Figure 8).

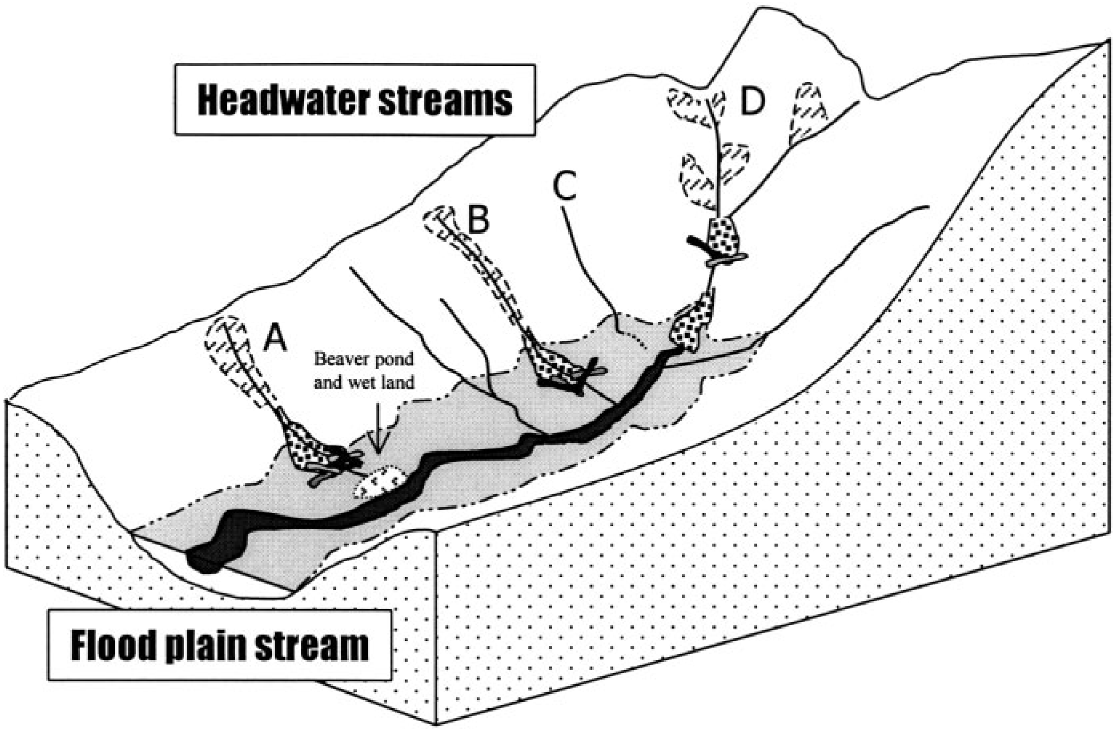


Figure 8. Schematic view of linkages from hillslope to channels, and from headwater streams to main channels. Multiple landslides are often found in headwater systems with ravine landforms, whereas single landslides affect channels with shallowly incised landforms. Headwater A: headwater channel eventually merges into wetlands and beaver ponds in both shallowly incised and ravine landforms. Headwater B: single landslides and channelized debris flows in headwater systems with shallowly incised landforms; sediment and water diffuse near the foot of hillslopes and do not directly enter the main channel. Headwater C: because of alluvial material in the floodplain, the stream becomes influent during the dry season. Headwater D: multiple landslides are found in headwater systems. Because of the larger amount of sediment, sediment is transported near or in the main channel

SUMMARY AND CONCLUSIONS

Hydrological processes with varying magnitudes and frequencies alter sediment movement after timber harvesting in headwater systems of the Maybeso watershed. The history of mass movement controls the availability of sediment in channels and the structure of riparian vegetation. Recovery of vegetation after episodic mass movement modifies bedload and suspended sediment transport and in-channel storage of sediment in headwater streams. The dynamics of sediment movement in managed forest headwaters can be summarized according to the following sequence: (1) mass movement after timber harvesting is triggered by storm events with recurrence intervals >5 years in logged areas; (2) sediment and woody debris redistributes from upper to lower reaches of channels; however, debris flows do not enter main channels; (3) subsequent sediment movement is transported from debris fans formed at the bottom of the U-shaped glacial valley; (4) greater amounts of bedload and suspended sediment are transported immediately after mass movement; (5) bedload and suspended sediment transport largely depends on vegetation recovery processes and woody debris recruitment after mass movements; and, finally, (6) the sequence of sediment transport and transformation of sediment movement modes from headwaters to downstream reaches alter riparian stands and channel morphology; this, in turn, creates heterogeneous riparian and in-stream landscapes in the Maybeso watershed.

Landforms strongly influence sediment transport with respect to hydrologic and geomorphic linkages from hillslopes to streams and from headwaters to downstream reaches in the Maybeso watershed. Landform characteristics, such as valley incision, valley floor topography, and vegetation cover, control the spatial and temporal distribution of sediment transport and storage (Nakamura *et al*., 1995; Benda and Dunne, 1997b). Such landform characteristics are also important for understanding the patterns of disturbances in riparian zones and channel morphology throughout the channel network (Nakamura *et al*., 2000). Therefore, hydrogeomorphic linkages in different landforms must be evaluated to understand the dynamics of water, sediment, nutrients, and organic matter in channel networks and riparian zones.

Knowledge of sediment transport and storage and the interaction of sediment movement with riparian vegetation and woody debris may aid in the understanding of dynamic material fluxes and processing in headwater streams. Since the total area of headwater systems comprises a major portion of the channel network, material transport in headwater systems in critical to understanding the dynamics of channel networks (Benda and Dunne, 1997a; Sidle, 2000; Gomi *et al*., 2002). In particular, the importance of headwater streams for the habitat and food supply of stream biota are a major concern in forest management.

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REFERENCES

Adenlof KA, Wohl EE. 1994. Controls on bedload movement in a subalpine stream of the Colorado Rocky Mountains, USA. *Arctic and Alpine Research* **26**: 77–85.

Benda LE, Cundy TW. 1990. Predicting deposition of debris flows in mountain channels. *Canadian Geotechnical Journal* **27**: 409–417.

Benda L, Dunne T. 1997a. Stochastic forcing of sediment supply to channel networks from landsliding and debris flows. *Water Resources Research* **33**: 2849–2863.

Benda L, Dunne T. 1997b. Stochastic forcing of sediment routing and storage in channel networks. *Water Resources Research* **33**: 2865–2880.

Bormann BT, Sidle RC. 1990. Changes in productivities and distribution of nutrients in a chronosequence at Glaicer Bay National Park, Alaska. *Journal of Ecology* **78**: 561–568.

Bryant MD. 1980. Evolution of large organic debris after timber harvesting: Maybeso Creek, 1949–1978. USDA Forest Service, General Technical Report PNW-101.

Dietrich WE, Dunne T. 1978. Sediment budget for small catchment in mountainous terrain. *Zeitschrift für Geomorphologie, NF* **29**: 191–206.
Dhakal AS, Sidle RC. 2004. Distributed simulation of landslides for different rainfall conditions. *Hydrological Processes* **18**: this issue.
Dunne T. 1991. Stochastic aspect of the relations between climate, hydrology and landform evolution. *Transactions, Japanese Geomorphological Union* **12**: 1–24.
Gomi T, Sidle RC, Bryant MD, Woodsmith RD. 2001. Characteristics of woody debris and sediment in headwater streams, southeast Alaska. *Canadian Journal of Forest Research* **31**: 1386–1399.
Gomi T, Sidle RC, Richardson JS. 2002. Understanding processes and downstream linkages of headwater systems. *BioScience* **52**: 905–916.
Grant GE, Wolff AL. 1991. Long-term patterns of sediment transport after timber harvesting, western Cascade mountains, Oregon. In *Sediment and Stream Water Quality in a Changing Environment: Trends and Explanation*, Peters NE, Walling DE (eds). International Association of Hydrological Science Publication 203. IAHS Press: Wallingford; 31–40.
Harmon ME, Franklin JF, Swanson FJ, Sollins P, Gregory SV, Lattin JD, Anderson NH, Cline SP, Aumen NG, Sedell JR, Lienkaemper GW, Cromack K, Cummins KW. 1986. Ecology of coarse woody debris in temperate ecosystems. *Advances in Ecological Research* **15**: 133–302.
Hogan DL, Bird SA, Rice S. 1998. Stream channel morphology and recovery processes. In *Carnation Creek and Queen Charlotte Islands Fish/Forest Workshop: Applying 20 Years of Coast Research to Management Solutions*, Hogan DL, Tschaplinski PJ, Chatwin S (eds). British Columbia, Ministry of Forestry, Research Branch, Victoria, B.C., Land Management Handbook No. 41; 77–96.
Johnson AC, Swanston DN, McGee KE. 2000. Landslide initiation, runout, and deposition within clear-cuts and old-growth forests of Alaska. *Journal of the American Water Resources Association* **36**: 1–14.
Lenzi MA, Marchi L. 2000. Suspended sediment load during floods in a small stream of the Dolomites (northern Italy). *Catena* **39**: 267–282.
Likens GE, Bilby RE. 1982. Development, maintenance, and role of organic debris dams in New England streams. In *Sediment budgets and routing in forested drainage basins*, Swanson FJ, Janda RJ, Dunne T, Swanston DN (eds). USDA Forest Service, General Technical Report PNW-141; 122–128.
May CL, Gresswell RE. 2003. Processes and rates of sediment and wood accumulation in headwater streams of the Oregon coast range, USA. *Earth Surface Processes and Landforms* **28**: 409–424.
Montgomery DR, Buffington JM. 1997. Channel-reach morphology in mountain drainage basins. *Geological Society of America Bulletin* **109**: 596–611.
Montgomery DR, Schmit KM, Greenberg HM, Dietrich WE. 2000. Forest clearing and regional landsliding. *Geology* **28**: 311–314.
Nakamura F. 1986. Analysis of storage and transport processes based on age distribution of sediment. *Transactions, Japanese Geomorphological Union* **7**: 165–184.
Nakamura F, Maita H, Araya T. 1995. Sediment routing analyses based on chronological changes in hillslope and riverbed morphology. *Earth Surface Processes and Landforms* **20**: 333–346.
Nakamura F, Swanson FJ, Wondzell SM. 2000. Disturbance regimes of stream and riparian systems—a disturbance–cascade perspective. *Hydrological Processes* **14**: 2847–2860.
Roberts RG, Church M. 1986. The sediment budget in severely disturbed watershed, Queen Charlotte Ranges, British Columbia. *Canadian Journal of Forest Research* **16**: 1092–1106.
Schwab JW. 1998. Landslides on the Queen Charlotte Islands: processes, rates and climatic events. In *Carnation Creek and Queen Charlotte Islands Fish/Forest Workshop: Applying 20 Years of Coast Research to Management Solutions*, Hogan DL, Tschaplinski PJ, Chatwin S (eds). British Columbia, Ministry of Forestry, Research Branch, Victoria, B.C., Land Management Handbook No. 41; 41–47.
Shimokawa E. 1984. A natural recovery process of vegetation on landslide scars and landslide periodicity in forested drainage basin. In *Proceeding of Symposium on Effects of Forest Land Use on Erosion and Slope Stability*, Hawaii; 99–107.
Sidle RC. 1984. Shallow groundwater fluctuations in unstable hillslopes in coastal Alaska. *Zeitschrift für Gletscherkunde und Glazialgeoleologie* **20**: 79–95.
Sidle RC. 1988. Bedload transport regime of a small forest stream. *Water Resources Research* **24**: 207–218.
Sidle RC. 2000. Watershed challenges for the 21st century: a global perspective for mountain terrain. In *Land Stewardship in the 21st Century: The Contributions of Watershed Management*, Ffolliot PF, Baker MB, Edminster CB, Dillon MC, Mora KL (Technical coordinators), 13–16 March, Tucson, AZ. Proceedings Rocky Mountain Research Station RMRS-P-13, USDA Forest Service, Fort Collins, CO; 45–56.
Sidle RC, Campbell AJ. 1985. Patterns of suspended sediment transport in a coastal Alaska stream. *Water Resources Bulletin* **21**: 909–917.
Sidle RC, Pearce AJ, O'Loughlin CL. 1985. *Hillslope Stability and Landuse*. American Geophysical Union, Water Resources Monograph 11. AGU: Washington, DC.
Sidle RC, Tsuboyama Y, Noguchi S, Hosoda I, Fujieda M, Shimizu T. 2000. Streamflow generation in steep headwaters: a linked hydrogeomorphic paradigm. *Hydrological Processes* **14**: 369–385.
Sidle RC, Sasaki S, Ostuki M, Noguchi S, Baharuddin K. 2000. Sediment pathways in a tropical forest: effects of logging road and skid trails. *Hydrological Processes* **18**: this issue.
Slaymaker O. 2000. Assessment of the geomorphic impact of forestry in British Columbia. *AMBIO* **29**: 381–387.
Swanston DN. 1967. *Geology and slope failure in the Maybeso valley, Prince of Wales Island, Alaska*. Unpublished PhD thesis, Michigan State University.
Swanston DN. 1970. *Mechanics of debris avalanching in shallow till soils of southeast Alaska*. USDA Forest Service, Research Paper PNW-103.
Tsukamoto Y, Ohta T, Noguchi H. 1982. Hydrological and geomorphological study of debris slides on forested hillslope in Japan. In *Recent Developments in the Explanation and Prediction of Erosion and Sediment Yield*, Walling DE (ed.). International Association of Hydrological Science Publication 137. IAHS Press: Wallingford; 89–98.
Wemple BC, Swanson FJ, Jones JA. 2001. Forest roads and geomorphic process interaction, Cascade Range, Oregon. *Earth Surface Processes and Landforms* **21**: 191–204.
Wolman MG. 1954. A method of sampling coarse riverbed material. *Transactions, American Geophysical Union* **35**: 951–956.
Wolman MG. 1977. Changing needs and opportunities in the sediment field. *Water Resources Research* **13**: 50–54.

