

Sediment budget for a small catchment in mountainous terrain

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with 6 figures and 3 tables

Zusammenfassung. Geomorphologische Prozesse sind oft sehr veränderlich in Zeit und Raum. Angaben von kurzfristigen Messungen an wenigen Orten können nicht einfach extrapoliert werden, um die Sedimentbilanz für ein Abflußgebiet aufzustellen. Bevor das Netzwerk der Meßpunkte eingerichtet wird, um die Komponenten der Sedimentbilanz zu messen, kann eine annähernde Bilanz durch Kartierung, Experimente im Labor, Gesteinsanalyse und Prozeßdaten von ähnlichen Abflußgebieten aufgestellt werden. Besonders sollten die Geländebeobachtungen der Beziehungen zwischen dem Transport und der Ablagerung der Sedimente, sowie die Veränderung der Partikel betont werden. Wir haben eine annähernde Sedimentbilanz für ein kleines Abflußgebiet berechnet. Lösung verursacht 60% der Denudation; die durchschnittliche Verweildauer des Bodens an den Hängen ist 20000 Jahre, und die Bodeneigenschaften ändern sich systematisch während der Bewegung hangabwärts. Die geschätzten Geschwindigkeiten des Bodenkriechens und die Flußfracht sind ausbalanciert. Die Hälfte der Bodenmasse, die die Abflußrinnen erreicht, wird als Schwebfracht weggetragen, während der Rest in den Äuflüssen, in Schwemmfächern und in der Talsohle vorübergehend abgelagert wird. Die Ruhezeit der Durchgangsaufschüttungen nimmt flußabwärts von Jahrzehnten auf etwa 10000 Jahre zu. Während des Transports verkleinern Verwitterung und Abschleifung 80% der Bodenfracht auf Schwebfracht.

Summary. Geomorphic processes are often slow and highly variable in time and space. Data from short-term monitoring at a few localities are not easily extrapolated to compute the sediment budget for a catchment. Before designing a monitoring network to measure components of the sediment balance, the investigator can construct an approximate budget from data that can be obtained quickly through field mapping, laboratory experiment, petrographic studies, and measurements in similar catchments. Emphasis should be placed on field cognition of the linkages between processes of sediment transfer and storage, as well as the changes which particles undergo. We have computed an approximate sediment budget for a small catchment. Dissolution is responsible for 60% of the denudation; the average residence time of soil on hillsides is 20,000 yrs; and the soil characteristics change systematically as the soil moves downhill. The estimated rates of soil creep and sediment discharge from the basin are in balance. One half of the soil discharged to channels is carried away as suspended load while the remainder is stored temporarily in tributaries, debris fans, and the floodplain. The residence times of sediment in these storage elements increase downvalley from decades to about 10,000 yrs. During the migration, weathering and attrition alter 80% of the bedload to suspended load sizes.

Résumé. Certains processus géomorphologiques sont souvent lents et ils varient beaucoup dans le temps et dans l'espace. Donc, il est difficile d'extrapoler les résultats des mesurages courts de quelques endroits pour calculer le bilan sédimentologique d'un bassin versant. Avant qu'il fasse le plan d'un réseau pour mesurer les composantes du bilan, le géomorphologue peut établir un bilan approximatif à partir de données qu'il peut vite obtenir en faisant une carte géomorphologique, par des expériences en laboratoire, par des études pétrographiques et par l'utilisation de mesures dans des bassins analogues. On devrait insister sur l'identification faites dans le terrain concernant les liens entre les processus de transfert et de conservation de sédiment, ainsi que les changements subis par les parcelles pendant le transfert. Nous avons établi un bilan approximatif de sédiments dans un petit bassin. La dissolution cause 60% de la dénudation; en moyenne, une parcelle de sol reste sur une pente pendant 20,000 ans; et les caractéristiques pédologiques se modifient alors que le sol glisse. Les taux estimés du glissement du sol et de l'apport sédimentologique du bassin sont égaux. Une moitié du sol déversée dans le lit fluvial est entraînée en suspension et le reste est mis temporairement dans les tributaires, les cônes d'alluvions et les plaines alluvionnaires. La durée de séjour du sédiment dans ces accumulations augmente graduellement le long de la vallée et passe de quelques décennies à 10,000 ans. Pendant le transfert, la décomposition et l'usure réduisent 80% de la charge de fond en dimension de charge de suspension.

Statement of the problem

To make a sediment budget for a drainage basin, one must quantify and relate the major processes responsible for the generation and transport of sediment. This task is difficult because the processes are often slow and highly variable in space and time. Data from short-term monitoring at a few localities are not easily extrapolated to compute average values for large areas. We have compiled a sediment budget for the various denudation processes in a small catchment on the basis of field mapping of sediment accumulations and from data available in the geomorphic literature. Our estimates can be checked against measured sediment yields from nearby basins. The computed sediment budget suggests the design for a set of measurements to refine our knowledge of the rates at which sediment is generated, moved and stored intermittently through the catchment, and is modified as it travels. We present a qualitative description of the sediment production and transport processes and a quantitative sediment budget. The results are then used to make predictions about sediment transport and soil formation processes.

The catchment

Rock Creek basin in the Coast Range of Oregon is a deeply dissected, 16.2 km² catchment developed in porphyritic basaltic lavas that are intruded by dikes of aphanitic basalt. Under the wet (3400 mm of precipitation per year), temperate climate of the region, this rock weathers to a 0.8 m thick reddish-brown gravelly sandy loam with poorly developed structure and horizons. The hillslopes of the basin are convex with narrow ridgetops, long 20–30 ° slopes and no concave footslopes. The drainage density is 4.0 km km⁻² and the average main channel gradient is 0.067. The drainage basin is mantled by a uniform cover of old-growth Douglas fir (*Pseudotsuga menziesii*) and hemlock (*Tsuga heterophylla*).

During storms, rainwater infiltrates and flows through the soil, returning

to the ground surface in the lower parts of hollows or on the valley floor as saturation overland flow (DUNNE, in press). Erosion processes, therefore, are dominated by mass wasting.

Weathering and hillslope processes

Two sets of processes cause weathering in the Coast Range of Oregon. Chemical weathering by the voluminous throughflow of dilute groundwater leaches the rocks and leaves a residue which is densely fractured and laced with alteration zones. The weathered residue is then disrupted by roots, burrowing animals and expansion during hydration to form a soil.

Water penetrates fractures, vesicles, cleavage planes and microfractures and reacts with the iron-rich glassy matrix of the basalts to form kaolinite, goethite and amorphous oxides along a network of closely spaced alteration zones. The greyish brown rock residue can be easily crumbled by hand. It contains clasts that vary in size from clay to gravel, but the relict textural and structural features of the rock such as bedding and phenocryst lineations are still discernible. Bedrock weathered to this condition is called saprolite and typically has a low bulk density (averaging 1.0 g/cc in this case) due to extensive leaching and expansion of minerals. Aphanitic dike rocks which lack significant pathways for water penetration tend to resist weathering and survive as coarse gravel in the soils.

Weathered rock and saprolite are mixed to produce a soil by a variety of physical and biological processes. There is overturning and mixing during wetting and drying and during slow plastic deformation of the base of the soil under shear. Root-growth and animal boring are also important, but the most important process of soil formation from the weathered substrate seems to be tree fall. On steep hillsides with thin soils, tree roots penetrate to the weathered bedrock or saprolite and when the tree falls substantial amounts of these materials are torn loose and mixed into the soil column. Measurements of the depths of roots on trees in the basin indicate that the primary structural support roots of even young trees grow on or into the underlying weathered rock.

The soils are mineralogically similar to saprolite but disruption lowers the bulk density and destroys any residual cohesion from the bedrock (as shown by shear strength measurements that we have made on similar material from western Washington). These last mentioned changes are important in allowing soil creep and landsliding within the soil, while the saprolite remains immobile.

The intensity of mixing increases downslope with the gradient as shear deformation becomes more intense and tree falls more frequent. Saprolite survives near the ridgetops and is the dominant source material for soil formation there. The saprolite becomes thin or absent, however, on the steeper parts of hillsides, where the mechanical disturbance is great enough to incorporate less weathered bedrock into the soil. The bulk density of weathered bedrock currently being incorporated into the base of the soil profile increases continuously with gradient and distance downslope from 0.8 to 1.2 g cm⁻³ in porphyritic vesicular basalt, from 1.0 to 2.3 g cm⁻³ in breccia, and from 1.1 to 2.1 g cm⁻³ in aphanitic

basalt. This variation is reflected in the soil properties. Ridgetop soils contain about 35% by weight of gravel and have a mean bulk density of 0.8 g cm^{-3} . At the base of hillslopes, gravel comprises 50% of the soil which has a mean bulk density of 0.95 g cm^{-3} .

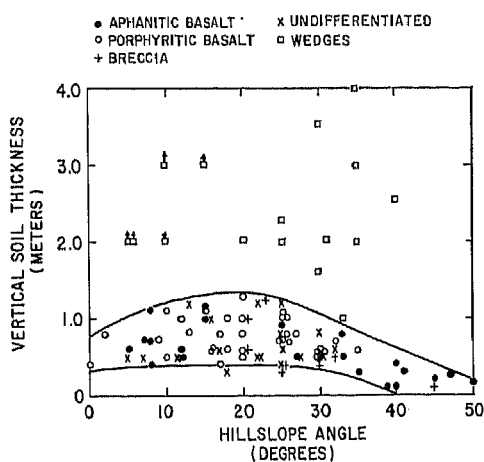


Fig. 1. Relationship between local hillslope gradient and depth for soils in the Rock Creek area. Measurements were made along road cuts. The upper curve separates shallow soils formed by mixing processes from soils referred to here as wedges which are formed by deposition in old landslide scars. If wedge depth extended below the road cut, an arrow on its symbol indicates that it is a minimum thickness.

The depth of soil also varies systematically as hillslope gradient increases downslope (fig. 1). The depth of soil is controlled by the effective depth and intensity of mixing, and the balance between mixing and downslope transport of soil. Somewhat deeper soils could theoretically remain stable with respect to landsliding because of the high permeability of the soil and the cohesion supplied by tree roots. Slides on the hillslopes are rare except for shallow footslope failures and debris flows from hollows (discussed later). PIERSON'S (1977) measurements of pore water pressures in a nearby area support our claim; he showed that the factor of safety is unlikely to approach unity on the main hillsides even in large storms.

Hillslope transport and discharge of soil to channels

In thin soils the mixing processes of wetting and drying, tree throw and animal burrowing which produce the soil also move it downslope. Because these processes are intense near the surface, the rates of soil creep are typically greatest near the top of the soil column and diminish rapidly with depth (KIRKBY 1967). In thick soils, however, shear deformation predominates (KOJAN 1967) and extends to depths of at least 10 m in parts of the Oregon Coast Range (SWANSTON & SWANSON 1976).

In order to compute the discharge of soil into stream channels by creep it is necessary to know the distribution of soil thickness and the average velocity of soil movement. The former, though highly variable, can be obtained quickly by direct observation; field measurements of soil creep require many years of observation. To construct an approximate sediment budget before the measure-

ments become available, we have used data from the geomorphic literature to estimate creep rates.

In Rock Creek basin, soils could be separated into two categories: thin soils, the depth of which varied with hillslope gradient; and thick soils, referred to here as wedges, which are generally thicker and coarser than the other soils (fig. 1). The wedges are filled landslide scars whose origin we will discuss in the next section. Transects along road cuts and stream channels revealed that the wedges intersected approximately 10% of the length of first-order channels, 30% of second-order channels and 50% of third and fourth-order channels.

Few measurements of soil creep rates have been made under conditions similar to those of the Oregon Coast Range; i.e. on steep, forested hillslopes with an annual rainfall of 3000 mm. Fortunately, the few measurements are consistent (table 1). We therefore accepted averages of the two sets of data shown therein as being representative of our shallow and deep soli profiles. These estimates can be refined as our own field measurements become available.

We can now compute separately the sediment discharge from thin soils and deep wedges, using data from table 1. The average soil depth in the Rock Creek

Table 1. Summary of creep rates and sediment yields from steep, wet catchments with an undisturbed forest cover.

Source	Total depth of movement (m)	Average soil creep (mm/yr)	Gradient	Annual precipitation (mm/yr)
Deep soils*				
SWANSTON & SWANSON (1976)	7.3	10.9	0.15	1500–2500
	7.3	10.7	0.20	
	5.6	7.1	0.15	
	5.2	11.7	0.30	
	2.6	10.4	0.30	
Shallow soils				
LEWIS (1976)	0.38	2.3	0.35	4500
	0.45	1.8	0.32	3050
EYLES & Ho (1970)	0.40	3.2	0.18	2540
Source	Suspended sediment yield (t/km ² /yr)	Drainage area (km ²)	Runoff (mm/yr)	Location
DUNNE (unpublished data)	18.3	108	150	Kenya
	20.0	157	182	
	22.2	98	278	
	26.1	62	352	
DOUGLAS (1967)	1–80	"small"	90–3000	Eastern Australia
LEAF (1966)	4.4**	2.7	300	Colorado
	2.6**	1.2	300	
FREDERIKSEN (1970)	28.6	0.6	1500–2000	Cascade Mts., Oregon

* "Old landslide deposits", gradients unpublished (SWANSTON & SWANSON, personal communication).

** Total sediment yield.

basin is 0.8 m, the bulk density of soil discharging into the channel is 0.95 t m^{-3} and the average annual sediment discharge from shallow soils is

$$(0.8 \text{ m}) \times (0.95 \text{ t m}^{-3}) \times (0.0025 \text{ m y}^{-1}) \times (119,215 \text{ m of hillside } 16.2 \text{ km}^{-2}) \\ = 14.0 \text{ t km}^{-2} \text{ y}^{-1}$$

Shallow soils discharge into channels primarily by small failures caused by undercutting of the hillside. Occasional large footslope failures occur. Typically these slides leave scars which are elongate parallel to the channel and rarely affect more than 30 m of hillside adjacent to the channel.

Wedge soils are on average about 2 m thick and have a bulk density of 1.1 t m^{-3} . They probably creep at about 0.010 m y^{-1} (table 1), supplying sediment to channels at the rate of

$$(2.0 \text{ m}) \times (1.1 \text{ t m}^{-3}) \times (0.010 \text{ m y}^{-1}) \times (12,640 \text{ m } 16.2 \text{ km}^{-2}) \\ = 17.2 \text{ t km}^{-2} \text{ y}^{-1}$$

Wedge soils discharge into the channel by frequent small failures along the base of the slope and by infrequent but spectacular debris flows which may extend more than 70 m up the hillslope. The local acceleration of soil discharge by debris flow from a wedge is counterbalanced by a long period of diminished contribution from the slide scarp as the wedge is reformed by creep discharge from adjacent hillsides. During the early stages of filling, some soil is washed into the channel (see later), but with this relatively small exception, the rare debris flows simply transport material which would otherwise move by deep soil creep and which is included in the last calculation.

The total soil discharge into the channel by creep is $31.2 \text{ t km}^{-2} \text{ y}^{-1}$. The steepness of the basin and the widespread occurrence of bedrock in the valley floors indicate that there is no long-term aggradation of sediment within the catchment. Therefore, the creep rate should equal the total sediment yield at the mouth of the basin. The few measurements we could find from undisturbed forested watersheds, listed in table 1, show that our estimate of sediment yield is reasonable. In particular, our results compare well with the $35 \text{ t km}^{-2} \text{ y}^{-1}$ from ANDERSON's (1954) map of suspended load discharge in this part of western Oregon and with $28.6 \text{ t km}^{-2} \text{ y}^{-1}$ reported by FREDRIKSEN (1970) from a small basin in the Cascade Mts. of Oregon.

Role of debris flows

In spite of heavy rains and steep slopes, landslides high on valley sides are rare. This is due in part to the cohesion provided by the numerous tree roots which penetrate the thin soil column and anchor the soil to the underlying bedrock (SWANSTON 1970). Also, the coarse textured soils are extremely permeable (YEE & HARR 1977) and readily transmit water, preventing the buildup of high pore water pressures on the valley sides. In contrast, small slides are common along the base of slopes where soils are undermined by streams.

Hillslope failures are usually initiated in partially filled hollows bordering the channel. Hollows occur in the headwaters of all first-order channels and on hillsides along channels of any order. These unchanneled swales typically are filled with a regolith 3 m thick. As a result of their subsurface topography, hollows

concentrate water, developing high pore pressures (PIERSON 1977) which can eventually lead to failure of the soil mass. Because of high water content the failures become flows and travel rapidly downslope, scouring the hollows and often the stream channel for hundreds of meters before coming to rest in a fan. For example, during a storm in 1974, two small slides which were initiated in a hollow at the head of a first-order stream 8 km north of Rock Creek rushed down the channel at speeds of $2\text{--}4\text{ m sec}^{-1}$ (from equations of JOHNSON 1970), scouring the 1000 m tributary channel and causing 6 small hillslope failures along the channel. About 4000 m^3 of poorly sorted gravel, sand silt and clay were deposited as a debris fan at the mouth of the tributary. About 1700 m^3 of detritus were deposited along the main channel, and only 130 m^3 of material were deposited in the tributary channel upstream of a road crossing. Of the 5800 m^3 of sediment transported by this flow, approximately 35% was scoured from the tributary channel bed, 3% was discharged into the tributary channel by slides and the rest was scoured from the soils and terrace deposits along the valley walls.

Contribution of debris flows to the sediment budget

A continual problem in attempts to reconstruct sediment budgets for mountainous catchments in the northwest U.S. has been the difficulty of determining the frequency of debris failures. We determined from sequential air photographs and from field observations that 12 debris failures occurred from 1961 to 1973 in the Rock Creek basin, discharging $15,000\text{ t}$ of soil into channels. However, in the winter of 1964–1965, a major flood event with a recurrence interval between 6 and 86 years (depending upon the catchment) occurred along the Oregon Coast Range. During this flood, 9 of the 12 slope failures were initiated. Because the relation between the frequency of floods and of landslides is not well known, it is not possible to assign a frequency to these landslides. More studies of the frequency of hillslope failures as related to hydrologic conditions are needed to resolve this problem.

The $10,000\text{ t}$ of soil discharged during the flood event of 1964–65 were equivalent to 36 years of creep discharge from wedges and 20 years of creep discharge from the entire watershed. The flood discharge did not, however, produce the equivalent of 20 years of sediment yield from the basin because much of the soil released was deposited in tributary channels and fans, which are still being eroded and transported down channel.

Significance of wedges

The areal density of topography hollows from which debris flows originate is $2/\text{km}^2$, half of which lie at the heads of first-order streams. Less obvious are a large number of hollows in bedrock, which are filled to different degrees with deep wedge-shaped pockets of soil (fig. 2). Many of these bedrock hollows are filled completely and have no topographic expression; they are recognizable only on road cuts and stream banks. We interpret the bedrock hollows as products of infrequent landslides. We have previously calculated the yield of soil creep from

these wedges, and now will consider their origin and role in the sediment budget of the catchment.

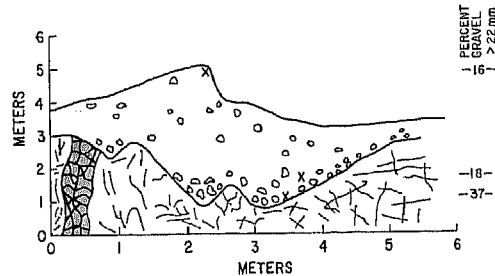


Fig. 2. Sketch of a wedge exposed in a road cut north of Rock Creek. The crosses indicate locations of samples collected for the textural analysis, the data from which are shown on the right. The top line represents the soil surface.

Deep road cuts are the best places to observe wedge soils. Their most striking feature is the 'u' or 'v' shaped depression in bedrock on which they rest and their thick depth of soil. Typically at the base of the wedge the soil material is gravelly and individual planar clasts tend to lie subparallel to the bedrock boundary. The saprolite at the base of the wedge is often thicker than under nearby areas with a thin soil mantle. Gravels at the base of the wedge consists almost entirely of porphyritic basalt similar in appearance to the underlying bedrock. A short distance up from the base of the wedge the proportion of aphanitic basalt increases dramatically. Soil is coarsest at the base of the wedge. Towards the surface, its texture gradually becomes identical to that of adjacent shallow soils.

These filled depressions in the bedrock can remain stable for long periods of time. We found a 357 year old hemlock in one filled hollow, and on another wedge, the deep burnt roots of a hundred-year old tree that was probably burnt in a fire 140 years ago.

Occasionally, these wedge soils fail as debris flows, stripping the soil and scouring the weathered bedrock. As water flows across the freshly exposed bedrock surface, more material may be removed. Small failures, replenished by creep from adjacent soils, begin the process of refilling the depression. Initially the finer material is washed away and a gravel lag is aligned subparallel to the weathered bedrock surface. In a hollow scoured about 10 to 20 years ago, the bed is now composed of patches of coarse gravel and bedrock (fig. 3). If the coarse material is not washed out by large flood events or another debris flow, further creep from the surrounding hillsides will mantle the coarse gravel with a finer deposit. This material diminishes the ability of water, which must now flow partly below the surface, to transport debris. As more soil creeps into the hollow, more of the water flows below the surface. Thus, less fine sediment is washed out and the rate of accumulation increases with time. Fig. 3 shows the textural characteristics of wedge deposits of different ages, which illustrate changes in particle sorting as a hollow fills. In fig. 4 we have plotted the amount of fill in a hollow for a range of times since scour as estimated from the ages of trees. Note that the slope of the line is 1.3, indicating that the rate of accumulation increases with time.

Eventually, water flows entirely as subsurface runoff in filled hollows, and they become concentrated zones of subsurface storm flow. Because surface washing has discontinued, the hollow is finally filled by material that is texturally

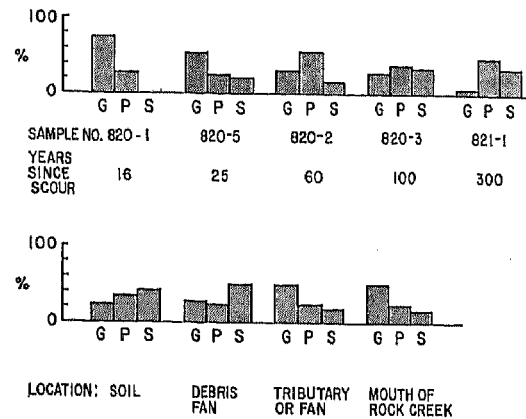


Fig. 3. Textural variations of sediment in various stages of transport. The fraction greater than 22 mm was measured by the pebble-count method of WOLMAN (1954) and the data were combined with sieve analyses of the finer particles. The letters indicate: Gravel (greater than 22 mm); Pebbles (from 2 to 22 mm); and Sand (less than 2 mm). The upper diagram shows the successive finding of material during filling of hollows. The lower diagram illustrates that soil can be transported to the debris fan essentially unsorted, or it can be deposited in the tributary where it experiences breakdown and sorting during transport to the debris fan. Further transport to the mouth of Rock Creek occurs without further significant textural change. The soil texture shown is an average of texture resulting from the combined discharge of shallow and thick soils.

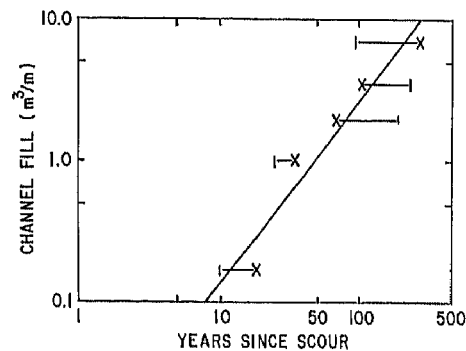


Fig. 4. Rate of filling of scoured hollows. Time since scour was determined from ages of vegetation on fills. (X) is our best estimate, (—) represents the range of uncertainty in our estimate. Vegetation consisted of salmonberry and alder on more recently scoured hollows, and large alder, hemlock and Douglas fir on older, partially filled hollows.

similar to adjacent soil from which it is derived. The dominant form of downslope transport becomes deep creep under shear.

As the hollow fills and a wedge is formed, the probability of failure increases with increasing soil depth for two reasons. First, the factor of safety for the soil decreases. Second, the probability of root penetration to bedrock declines, and the cohesion provided by tree roots becomes less effective. Because this component accounts for approximately 40% of the shear strength of shallow (1 m) soils close to failure on forested the mountains of the west coast, its decrease greatly increases the probability of failure.

Whether in the long term a depression grows and becomes a hollow or a channel depends on the erodibility of the underlying bedrock and the amount of subsurface flow that it receives. In the catchment a full spectrum of bedrock depressions exists from channels to hollows to wedges. It is not clear what eventually determines the number of 'permanent' unfilled hollows versus hollows which occasionally fail and then fill.

Fluvial transport

Soil that is discharged into channels may experience immediate sorting when fine material is washed away as suspended or wash load, or it may be carried to a debris fan by a debris torrent and be deposited essentially unsorted. In fig. 3, the average textures of material discharged by soil creep tributary channel sediment, debris fan sediment and floodplain sediment near the mouth of Rock Creek are compared. Comparison of the texture of the sediment in the tributary with that of the soil discharged by creep reveals that if few of the particles greater than 22 mm are washed away, about 80% of the original less-than-2-mm size fraction must be carried away as suspended load, and about 50% of the 2-to-22 mm size fraction is either carried away as suspended load or quickly breaks down to suspended load size and washes from the channel. The proportions of these two size classes which are thus carried away constitute 50% of the soil that enters the channel.

Despite continued transport to the mouth of Rock Creek, channel sediment shows no further textural change because attrition during weathering and transport are approximately balanced by the influx of coarse particles along the stream. There is abundant evidence that attrition is intense. In soils, gravel-size rock fragments are angular, measuring 1.0–1.5 on the POWERS (1953) scale. In tributaries, rounding increases from angular to subangular to subrounded (1.5–2.0). In the main channel rounding increases only slightly from subrounded to rounded (2.5–3.0). The spatial distribution of gravels of varying degrees of hardness also suggests breakdown of intensely weathered particles. At 68 sites along the channels of the basin, 15 particles ranging in size from 4 to 22 mm were tested for hardness by attempting to break them by hand. A map of the results (fig. 5) shows that the proportion of hand-breakable gravel generally decreases as drainage increases downstream. Pebble counts of rock types support this observation by showing that although the basin is approximately 80% underlain by porphyritic basalts, 50% of the sediment in the fourth order channel consists of hard aphanitic basalt.

Sand-size sediment in the Rock Creek channel is 86–97% weathered rock fragments and readily breaks down to silt and clay during tumbling experiments, as MOBERLY (1968) found for Hawaiian sands. The rate of breakdown of Rock Creek sand in a tumbling mill was 10% reduction of weight in a particular size class in 100 km of 'transport' (DIETRICH, in preparation). This rate is substantially higher than the value of 0.1% per 100 km reported by KUENEN (1960) for quartz and feldspar.

Values typically reported for the ratio of bedload to total solid load discharge are from 5–10% (ROSGEN, U.S. Forest Service, personal communication on the basis of extensive bedload measurements in the Rocky Mts.) and 10–19%

Because breakdown of sediment is significant during transport, it was necessary to estimate the downstream change in bedload discharge. As stated earlier, about 50% of the detritus which is discharged into the stream from soils is quickly removed as suspended load. Thus, on a first or second order tributary, 50% of the sediment is carried as fluvial bedload or debris flows, and we have already tentatively accepted a bedload component of 10% for the mouth of Rock Creek. Until better data become available, we also assume that this proportion is related to a power function of drainage area, and we can therefore estimate the approximate relation between bedload discharge rate (Q_b) and drainage area (A) (table 2).

Table 2. Coefficients and exponents for power functions.

System	Relation	a	b	c	m	n	p
All systems	$Q_b \propto A$		10.0			0.57	
Tributaries	$A \propto X$			$1.67(10^{-6})$			2.08
	$V \propto A$	7.25			0.83		
Main channel	$A \propto X$			$1.1(10^{-4})$			1.32
valley floor	$V \propto A$	1.53			1.35		
active sediment	$V \propto A$	0.391			1.07		
gravel bars	$V \propto A$	0.00871			1.37		
Debris fans	$V = 970 A^{0.93}$						

For all regression equations $p < 0.02$, $10 < n < 28$.

Residence times in channels and flood plains were computed as follows: Residence time per meter equals the mapped volume of sediment per meter of reservoir (V) divided by the bedload discharge rate. If the volume and discharge are written as power functions of drainage area:

$$(1) \quad V = a(A^m) \text{ and } Q_b = b(A^n)$$

and drainage area is a function of reservoir length (X):

$$A = c(X^p)$$

then residence time per meter is

$$\frac{dt}{dX} = \frac{a}{b} \frac{c^{(m-n)}}{X^{(m-n)p}}$$

Integrating this function between two points in the reservoir gives the residence time through that section of reservoir:

$$(2) \quad \int_{t_1}^{t_2} dt = \int_{X_1}^{X_2} \frac{a}{b} \frac{c^{(m-n)}}{X^{(m-n)p}} dX$$

$$t_2 - t_1 = \frac{\frac{a}{b} c^{(m-n)}}{1 + (m-n)p} X^{(1 + (m-n)p)} \bigg|_{X_1}^{X_2}$$

Velocity of sediment through a reservoir (U) is simply the inverse of the residence time per meter of reservoir length.

Residence time in debris fans was computed by dividing the volume of sedi-

ment in a fan by the estimated bedload discharge rate of the tributary (V and Q_B from equation 1). Results of the above procedure are given in Tables 2 and 3. As expected, this residence time increases with tributary order for both tributaries and debris fans.

Residence time is much longer once particles enter the valley floor of the main channel. A particle spends about 10^4 years in the valley floor, most of this time removed from the channel. It is leached by water rich in carbon dioxide and organic acids. The length of the residence time further supports our contention that thorough attrition occurs as the bedload material travels to the sea.

Perhaps the most interesting predictions shown in table 3 are the velocities of particles within the gravel bars and the bankful channel down to a scouring depth of 0.3 m. These results suggest that if the velocity of all particles moving in a mobile bed is considered, rather than just in the gravel bars and surface of the bed, the average travel distance per year of all channel sediment is much lower than if the particles were to move just in the gravel bars. Painted rock studies of average transport rate typically measure just the transport rate of surface particles rather than of particles which are under the armor layer but still move from time to time. Our data suggest that these studies overestimate the average transport rate of particles in the bed.

Table 3. Velocity and residence time of sediment in storage systems.

System	Order	Velocity, U (m/y) ($\frac{Q_B}{V}$ from eq. 1)	Residence time, (y) (t_1-t_2) from eq. 2)	T_2^* (y)	T^{**} (y)
Tributaries		$1.38 A^{-0.26}$	$0.015 [X_2^{1.54} - X_1^{1.54}]$	—	—
	1			19	
	2			114	
	3			231	
Debris fans					$97 A^{0.36}$
	1				33
	2				67
	3				87
Main channel					
valley					
floor	4	$6.5 A^{-0.78}$	$6.2(10^{-5}) [X_2^{2.03} - X_1^{2.03}]$	4933	
active					
sediment	4	$25.6 A^{-0.60}$	$2.5(10^{-4}) [X_2^{1.60} - X_1^{1.60}]$	619	
gravel					
bars	4	$1150 A^{-0.80}$	$2.9(10^{-7}) [X_2^{2.00} - X_1^{2.00}]$	31	

* T_2 is the residence time for a particle which enters a channel of a given order at one-half of the way along its segment.

** T is the total residence time of a particle in a fan of a given order.

Because the gravel bars in Rock Creek are only 0.3 m thick, our calculations of velocity through these features should give values similar to the surface velocities measured with painted rocks. For the data of MILHOUS (1972, 1973) on a similar creek we have calculated a velocity through gravel bars of 236 m y^{-1} , whereas his field measurements yield a value of 234 m y^{-1} . Although the similarity of our numbers is fortuitous, it suggests that our equation has the correct general form.

Denudation and soil formation

The rate of lowering in the Rock Creek basin is

$$L = \frac{Q_s + Q_D}{d_b}$$

where L is the lowering rate in m y^{-1} , Q_s is the solid load discharge in t m^{-2} and Q_D is the dissolved load discharge in $\text{t m}^{-2} \text{ y}^{-1}$. The average bedrock density (d_b) is approximately 2.5 t m^{-3} . The solid load discharge equals the creep discharge of $31.2 \text{ t km}^{-2} \text{ y}^{-1}$ and the measured dissolved load is $52.5 \text{ t km}^{-2} \text{ y}^{-1}$. The lowering rate of the bedrock is $33.5 \text{ mm } 10^{-3} \text{ y}^{-1}$.

We have assumed that the processes and products in this basin area are in approximate equilibrium. Therefore, for a point on the hillside, the soil thickness is constant through time. The total volume of soil in the basin, divided by the discharge rate should indicate the average residence time of soil on the hillsides. For the Rock Creek basin, the erosion rate by creep for an 0.8 m thick soil with a bulk density of 0.8 t m^{-3} (the average value for the whole soil mantle) is $20,500$, equivalent to a soil lowering rate of $39 \text{ mm } 10^{-3} \text{ y}^{-1}$.

The lowering rate for the soil is 16% greater than that calculated for the bedrock. This discrepancy may be within the errors in our procedures, but the rate of soil lowering should exceed that of the bedrock because of the expansion that takes place when saprolite and weathered bedrock are torn up by biogenic mixing processes. In Rock Creek basin the average bulk density of saprolite is 1.0 t m^{-3} and that of overlying soil is 0.8 t m^{-3} . Some of this density change is due to dissolution during the $20,500$ years the soil is on a hillside, and the rest results from mechanical disruption by tree roots.

Summary

In fig. 6 we summarize the significant processes controlling the sediment budget in Rock Creek basin. The diagram emphasizes the importance of the linkages that must be identified before one can use information from measurements of individual processes. The definition of the linkages requires intensive field investigation to estimate the frequency of catastrophic processes such as debris flows, to define the magnitude of sediment stores and to recognize subtle features such as hollows filled with wedge soils.

We have developed an approximate sediment budget based upon field mapping and process rates estimated from other studies. This approximation indicates where our emphasis should be placed when installing instruments to monitor various processes.

A good estimate of rates of soil creep is vital in a basin in this region, but transport in the wedge soils is so important that a special effort must be made to monitor deep creep as well as the more widespread and easily measured shallow movement.

Slides and slumps are not widespread, except for the small failures in the stream banks, and warrant less attention. Debris flows, however, are extremely

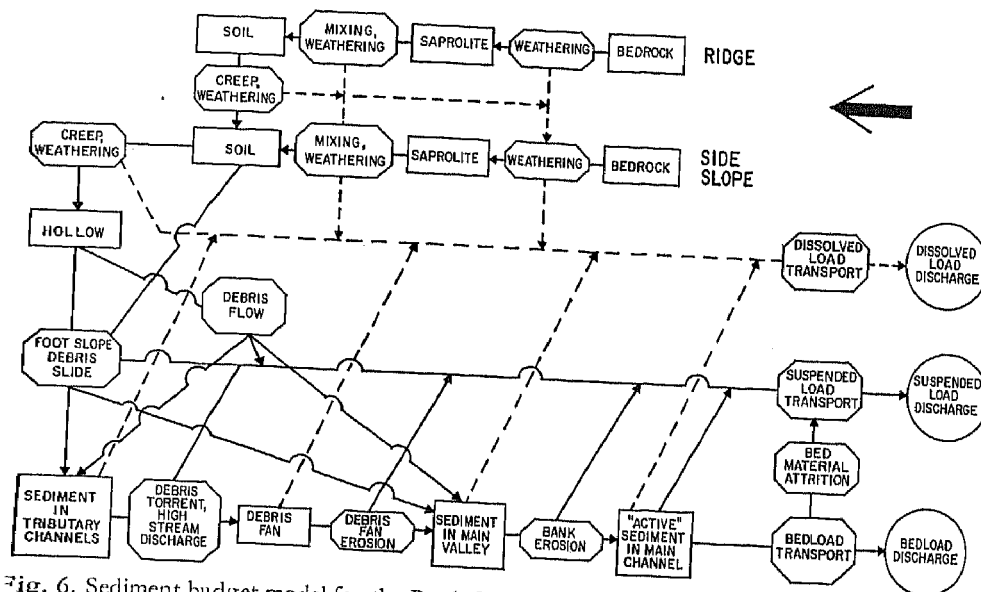


Fig. 6. Sediment budget model for the Rock Creek basin. Rectangles represent storage systems. Octagons indicate transfer processes. Circles represent outputs. Solid lines represent the transfer of sediment and dotted lines represent the migration of solutes.

important agents of landform development in the region for a number of reasons. They carry into the stream a portion of the soil transported down hillsides by creep. They carve out hollows in the bedrock, some of which are completely filled and others which survive as topographic features because of their sizes and frequency of failure. By occasionally scouring the hollows and the tributary channels to bedrock, debris flows assist in (and perhaps control) incision of the channel system and the fixing of the drainage density.

Our calculations indicate that more work needs to be done on the frequency of debris flows by techniques such as tree-ring dating. The 16-year photographic record is much too short for defining average frequencies of these catastrophic events.

The approximate sediment budget also emphasizes the storage role of the valley floor sediments. The debris fans at tributary mouths, for example, store the equivalent of 800 years of bedload discharge and 87 years of suspended load discharge from the drainage basin. Relatively small interannual changes in these accumulations would have large effects upon the sediment discharge of the basin. The changes in yield would not be related to changes in hillside conditions, and could not be predicted by sediment transport formulae. The alluvial features, therefore, warrant careful monitoring.