
## Papers

# Construction of Sediment Budgets for Drainage Basins

William E. Dietrich, Thomas Dunne, Neil F. Humphrey, and
Leslie M. Reid

## ABSTRACT

A sediment budget for a drainage basin is a quantitative statement of the rates of production, transport, and discharge of detritus. To construct a sediment budget for a drainage basin, one must integrate the temporal and spatial variations of transport and storage processes. This requires: recognition and quantification of transport processes, recognition and quantification of storage elements, and identification of linkages among transport processes and storage elements. To accomplish this task, it is necessary to know the detailed dynamics of transport processes and storage sites, including such problems as defining the recurrence interval of each transport process at a place.

William E. Dietrich, Assistant Professor, Department of Geology and Geophysics, University of California, Berkeley, California 94720; Thomas Dunne, Professor; Neil F. Humphrey, Graduate Student, Leslie M. Reid, Graduate Student, Department of Geological Sciences and Quaternary Research Center, University of Washington, Seattle, Washington 98195.

## INTRODUCTION

A sediment budget for a drainage basin is a conceptual simplification of the interaction of processes that convey debris from bedrock down hillslopes and channels and out of the basin. Monitoring the creation and movement of all particles in a drainage basin is impossible, and the recurrence and progress of many processes exceed the lifespan of the researcher. These obstacles force us to develop conceptual models of sediment generation and transfer that can guide field measurements. Such a framework is valuable in decisions about where to take representative measurements, about how to extrapolate data to unmonitored parts of a basin, and about how to measure debris transport.

Definition of sediment budgets has been inhibited by a lack of models for the mechanics of processes, by failure to define processes clearly, and by the great difficulty in measuring rates of sediment transfer. We cannot propose a single general method for construction of a budget. Instead we suggest an essential list of elements that all sediment budgets must have to be meaningful. We begin by proposing a definition of a sediment budget that leads to the requirements in construction of a sediment budget. The rest of the paper is devoted to an examination of these requirements, using as illustrations problems we have encountered in our fieldwork in the coastal mountains of Oregon and Washington. A central theme of this paper is that, with a small, but essential amount of fieldwork, an approximate sediment budget can be constructed that satisfies our suggested requirements and can be used as a guide for a monitoring program.

## DEFINITION OF A SEDIMENT BUDGET

A sediment budget for a drainage basin is a quantitative statement of the rates of production, transport, and discharge of detritus. In most studies, measurement of sediment production is assumed equivalent to quantification of sources of sediment discharged into channels. Sediment is also produced by the chemical degradation and physical mixing of weathered bedrock to form a soil, however, and material is transferred between size fractions as a result of attrition during transport. Also, the rate and nature of sediment production from the bedrock by soil-forming processes may affect significantly the rate of hillslope erosion and mode of fluvial debris transport. In a sediment budget, the soil mantle should be treated as a separate element of sediment storage having a definable rate of inflow and outflow of detritus. Quantification of debris transport requires not only defining rates of sediment transfer between major storage elements-such as between soil and stream channels--but also computing rates of movement through these sediment reservoirs. The latter is particularly difficult to do with present understanding of processes, but it is from such a quantification that predictions can be made about the change in size of storage elements and the rapidity and magnitude of the response of sediment discharge from a drainage basin after some disturbance. Discharge of detritus is simply the rate of transport of sediment past a monitoring station.

To construct a sediment budget for a drainage basin, the temporal and spatial variations of transport and storage processes must be integrated, and to do so, these requirements should be fulfilled: (1) recognition and quantification of transport processes, (2) recognition and quantification of storage elements, and (3) identification of linkages among transport processes and storage elements. To accomplish this task, it is necessary to know the detailed dynamics of transport processes and storage sites, including such-problems as defining the recurrence interval of each transport process at a place.

## RECOGNITION AND QUANTIFICATION OF TRANSPORT PROCESSES

A careful, long-term monitoring program designed to sample the spatial and temporal variations in elements of a sediment budget is an expensive venture. Such programs have usually been undertaken to answer questions about rates of landform evolution (Rapp 1960), effects of climatic change (Leopold et al. 1966), and effects of land use (Fredriksen 1970, Janda et al. 1975, Swanson, et al. (1982). Yet, as illustrated by Leopold and his associates, a careful examination of the processes is needed both to interpret the results of measurements and to evaluate the effects of changes in the major controls of processes. Construction of predictive sediment budgets is hindered by the lack of understanding of the mechanics of processes. At present, no transport process is sufficiently well understood that field data on major controls can be used to predict transport rates. Until studies of such processes as soil creep, landslides, sheetwash, and sediment transport in river channels produce realistic mathematical models, we must rely on extensive monitoring to provide data on transport rates.

Before a monitoring program is begun, it is essential to recognize and quantify approximately the major processes that generate and transport sediment. An initial field study should be made to identify the distribution of dominant transport processes and interrelationships among them. From such a survey and from currently available theory for the physics of transport processes, a program of monitoring can be established to sample the range in values of major controls and the spatial variation of processes as a function of these controls. The only way to generalize from a few measurement sites to a landscape is to develop predictive models of the relation of each transport process to its controls.

An important step in the simplication of a sediment budget to a form that can be examined in the field is the recognition of the dominant hillslope erosion processes. Striking differences in dominant processes occur in forested areas of different climate, geology, and land use. The dominance of a process can also depend on stand history. A dramatic example is the period of intense rain splash, sheetwash, and dry ravel that occurs after a major burn in the chaparral of Southern California (Rice, this volume). A central issue in this case becomes the assessment of the transport of sediment (on hillslopes, into and along channels) after specific burns, relative to the total transport evaluated over a period sufficiently long to accommodate the spatial and temporal variation in magnitude and intensity of burns. In this paper, we refer to the recognition and measurement of only a few processes that have been shown to be important in wet, forested environments.

## Slumps and Earthflows

In some landscapes underlain by mechanically weak rocks, massive earthflows and slumps move large quantities of debris to channel perimeters where shallow debris slides discharge sediment into the channel (as described by Swanson and Swanston 1977). These deep-seated movements involve soil, saprolite, weathered bedrock, or fresh bedrock and can extend to tens of meters below the soil surface. Engineering geologists who have measured relatively slow rates of movement in these deepseated features have referred to the movement as creep (e.g., Wilson 1970, Swanston and Swanson 1976). This has clouded the distinction between earthflow and creep. For clarity, here we will refer to slow mass movement of just the soil mantle as soil creep. This distinction matters: measured engineering "creep" tends to be much faster than soil creep and to occur by a different process.

Slumps and earthflows can persist for thousands of years as topographic features and occasional sources of sediment, according to Swanston and Swanson (1976) and Gil and Kotarba (1977), who have begun to investigate the conditions under which this form of hillslope transport occurs. Kelsey (1978) has emphasized that earthflows covering only a small proportion of a basin may provide a major input of sediment to the channel. Therefore, careful mapping of these features is necessary at an early stage of budget studies. This is not an easy task in densely forested regions where the surface expression of the slump-earthflow may be subdued after periods of relatively slow movement. Measurement of the discharge of sediment from earthflows requires drilling and the installation of flexible tubing anchored in a stable substrate. Deformation of the tube is measured with an electrical inclinometer (Kojan 1968, Swanston and Swanson 1976) to provide a vertical profile of velocity. The difficulty and cost of installation limits the number of measurement sites, so that the velocity field is poorly defined. Inclinometer measurements can be augmented by repeated survey of stakes on the surface of the earthflow, however. Even without inclinometer measurements, the discharge can sometimes be adequately defined from data on depths of movement and surface velocities (Gil and Kotarba 1977, Kelsey 1977).

## Debris Slides

The importance of debris slides in forested mountains has been emphasized by Swanston (1969, 1970), O'Loughlin and Pearce (1976), and many others. Such failures typically occur in at least three physiographic locations. On the footslopes of hills, relatively small debris slides convey soil and weathered bedrock into adjacent channels.

Much of the soil transported by these debris slides may have been moved to the footslopes from further upslope by soil creep and earthflows. The volume pnd frequency of this transfer across the hillslope-channel interface must be measured if this component of the sediment budget must be isolated. In some budgets where only the total sediment flux is required, however, it may suffice to assess the sediment transport by other processes on the hillside and to ignore this last step in the transfer to channels (Dietrich and Dunne 1978). Debris slides and avalanches also occur along hillslope profiles and convey all or a portion of displaced sediment to the channel. Again, it is necessary to obtain volume, proportionn reaching the stream, the centroid displacement of the remainder, and the frequency of transport. This may be done through a combination of fieldwork and aerial photographic interpretation (e.g., Simonette 1967). The relatively short aerial photograph record can sometimes be extended by using dendrochronology to date older landslides. Finally, debris slides and debris flows often originate in topographic hollows at stream heads or in thick soils in bedrock hollows, which Dietrich and Dunne(1978) have called "wedges." Their role in the sediment budget must be defined in the same way as slides on planar hillsides.

Increased debris sliding caused by forest clearing and road construction has been examined in many different forests (see Gresswell et al. 1979, for a recent example and for references). Typically, the frequency of debris-sliding increases in the cut areas, and a fourth location for debris slides--the road-affected region--is added to the landscape. Here, distinguishing between debris slides from road fill, cut bank, and channel crossing may be necessary because each may have different volume, distance of travel, frequency of occurrence, and sensitivity to predisturbance conditions and management practices.

## Soil Creep and Biogenic Transport

The term "soil creep" is generally used to refer to a group of processes that result in relatively slow downhill movement of the soil mantle (Young 1971, Carson and Kirkby 1972). These processes include the mass flow of debris under continuous shear stress imposed by the downslope component of the soil weight; displacement of soil by expansion and contraction from wetting and drying or freezing and thawing; expansion, collapse, and settling of debris caused by weathering (Young 1978); and downslope movement by raveling and spalling of surface debris. The relative importance of these processes in the sediment budgets of forested drainage basins is generally unknown because few measurements have been attempted, and no theory exists to predict movement based on soil properties. In environments in which slumping, earthflow, and midslope debris slides are rare or nonexistent and rain splash and overland flow are unimportant, however, the rate of supply of sediment to channels must be controlled by soil creep and the equally subtle processes associated with biogenic transport.

The difficulty of making field measurements of soil creep has prevented quantification of its role in the sediment budget and has impeded the development of predictive models of the processes involved. A simple, accurate method for monitoring soil creep and its controls, which allows rapid installation of a large number of monitoring sites, has not been developed. Further, different processes require different measurement techniques. A review of the various methods that have been used can be found in Young (1971), Carson and Kirkby (1972), and Anderson and Cox (1978). Continued work like the study by Fleming and Johnson (1975) on soil creep in northern California would be valuable in the context of a sediment-budget study. They used flexible tubing to monitor the seasonal rate of creep as controlled by moisture content of the soil. They successfully demonstrated the importance of rate and process theory (Mitchell 1976) to the understanding of the mechanics of soil experiencing mass flow, although they were unable to use the theory quantitatively in its present form. From field measurements and mapping of soils and bedrock, they also examined the geologic and topographic controls of soil movement.

Biogenic transport is the movement of debris as a result of biological activity. Biological activity is the cause of the transport rather than simply a control or influence and can be divided into two categories according to rate. Such processes as tree throw and animal burrowing cause local catastrophic transport, but pervasive processes such as root growth and wind stress on trees result in the slow transport of debris. Biogenic transport has traditionally been subsumed under the term soil creep; it is caused, however, by a set of processes unlike those generating soil creep. Although often mentioned as causes of downslope movement of debris, biogenic transport processes have been the object of only a few careful studies. These include studies of animal burrowing by gophers (Thorn 1978, and articles referred to therein), termites (Williams 1973), voles and moles (Imeson 1976), and by beetles (Pair 1976), as well as studies of tree throw (Denny and Goodlett 1956, Kotarba 1970, and Swanson et al. (1982).

These studies have largely focused on the direct movement of soil by the organism; the annual transport rate is calculated as the product of the volume of soil displaced, distance it has moved, and annual frequency of disturbance on a hillside divided by the area of the hillslope. The distance moved is simply the downslope distance between the centroids of the soil mass in its original and displaced position. This transport rate can only be computed from direct observation of the volume, centroid displacement, and frequency. Traditional creep-monitoring techniques will not detect it.

Displacement of soil by animals is followed by filling of the burrow or hole and the decay of the exposed mound, which may both be caused by abiotic processes. Complete evaluation of biogenic transport should include quantification of these secondary processes. One such study by Imeson (1977) suggests that rain splash of exposed soil mounds formed by voles is an important contributor of sediment to channels in the deciduous forests of Luxembourg.

Although Kirkby (1967) and others have generally considered biogenic transport to be unimportant, recent studies cited above suggest that in some regions it may dominate transport over much of the hillslope. In particular, in some forested landscapes the importance of tree throw is clearly expressed by the microtopography created by decaying tree-throw mounds.

Estimates of tree-throw transport rate in terms of equivalent rate of soil creep also suggest its importance. Denny and Goodlett (1956) computed a transport rate equivalent to 1.5 mm/year in the central Appalachians. Kotarba (1970) found that a single wind storm in the Tatra Mountains of Poland caused transport approximately equivalent to 20 cm of soil creep on some hillslopes (assuming average soil depth to be 50 cm). The trees that fell were on the average about 135 years old, which would give an average soil creep rate of roughly 2 mm/year over the lifespan of the tree as a result of this single storm. In the Olympic Mountains of coastal Washington, we have computed a tree-throw rate of 1.8 mm/year for a roughly 80-year period (Reid 1981). These transport rates are comparable in magnitude to soil-creep measurements in forested hillsides (Barr and Swanston 1970, Eyles and Ho 1970, Lewis 1976); if representative, they strongly suggest the need to incorporate tree-throw transport into hillslope sediment budgets.

## Sheetwash and Rain Splash

Sheetwash and rain splash are relatively ineffective in transporting sediment in undisturbed forested basins in the Pacific Northwest. The high soil permeability and thick humus layer confine such activity to areas of recent disturbance--such as landslide scars, rootwads, tree-throw scars, and areas of bank erosion--all of which are soon revegetated. In logged basins, however, these processes take on a new significance. Not only is the area laid bare by landslides and bank erosion increased, but new forms of disturbance such as road building, log yarding, and slash-burning expose even larger areas of mineral soil. On skid trails and slash-burn surfaces, infiltration capacities often remain high enough that overland flow does not occur; but gravel road surfaces are well compacted, their permeabilities are lowered, and sheetwash is frequency generated.

The contribution of sediment to streams from surface erosion from roads can be added as a separate element to-the sediment budget. For a sediment budget to be of value for management decisions, it should include an evaluation of the role of road design and usage on sediment yield. Unlike for many of the processes described above, quantitative predictions can now be made of rates of sediment transport from roads based on simple field measurement of major controls. For example, in a study of the effects of roads on the sediment discharge into streams of the Clearwater basins in western Washington, Reid (1981) developed a technique

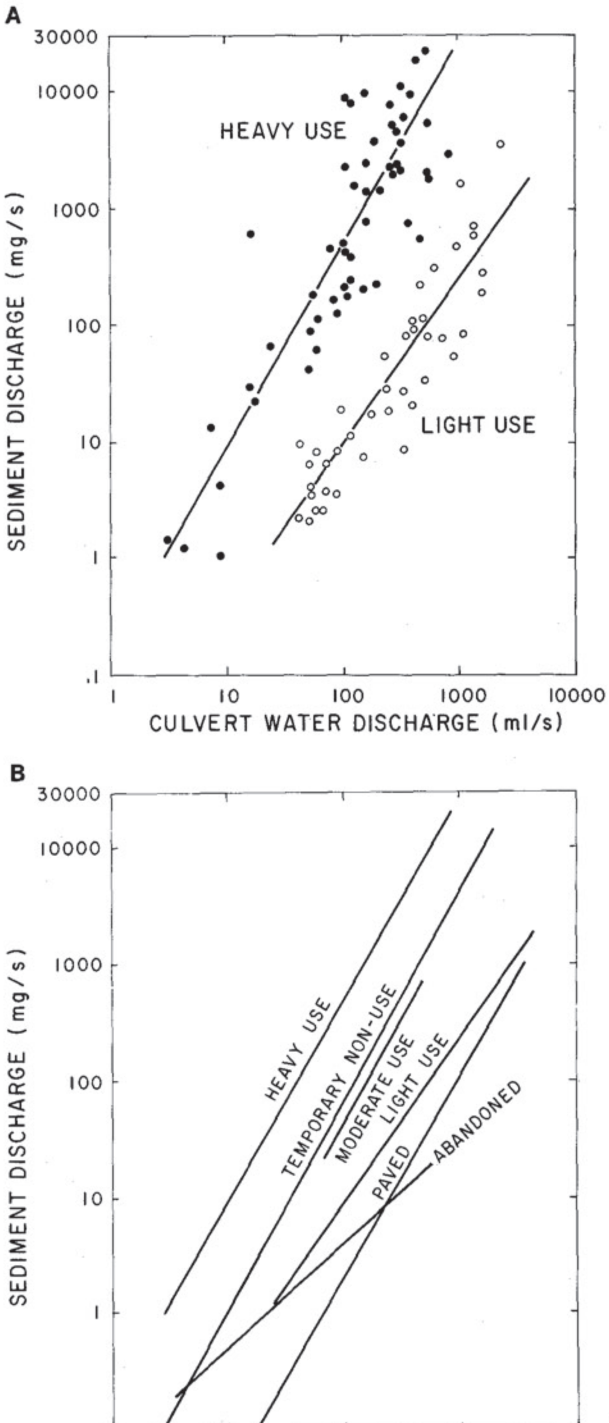



Figure 1—Effects of road-surface conditions on sediment-discharge rating curves of drainage culverts for gravel-surfaced logging roads in the Clearwater drainage basin of coastal Washington. A. Comparison between “heavy use” (16 to 32 logging trucks/day) and “light use” (no loggin trucks in 3 to 60 days). B. Summary diagram comparing different use conditions, including “moderate” (1 to 2 trucks per day) and “temporary nonuse” (logging trucks within past two days).

for computing sediment influx to streams in a drainage basin that contains a variety of roads experiencing a range of usage. She monitored culvert discharge from 10 road segments of average length and gradient, taking care to avoid segments with significant drainage from a back-cut or hillslope. The roads represented six surface conditions: two segments were paved, two were gravel-surfaced but abandoned, and the remaining six gravel-surfaced segments were used with varying intensity, from heavy (more than eight logging trucks per day) to light (without truck traffic for 3 to 60 days before measurement). Culvert discharge, sediment concentration, and rainfall intensity were measured at each culvertduring a series of storms and used to construct unit hydrographs and sediment rating curves for each use-level on each road (fig. 1). The unit hydrographs were then used to generate a continuous record of runoff from a year’s continuous rainfall record measured in the basin. Applying the sediment-rating curve to the generated runoff record, Reid computed the annual sediment yield from the road segments under different surface and road-use conditions.

## IDENTIFICATION OF LINKAGES AMONG PROCESSES AND STORAGE ELEMENTS

Linkages among processes and storage elements establish the general form of a sediment budget, which can be expressed in a flow diagram--such as figure 2. Identification of causal linkages in the budget highlights the effects of successive transfers on the characteristics and quantity of sediment moved. For example, in drainage basins free of extensive valley-floor deposits, the particle-size distribution of alluvium is controlled by soil-formation processes and by attrition and sorting during fluvial transport. In low-order channels, stream transport has little opportunity to sort or comminute sediment supplied by slope processes. Residence time of sediment in small, steep tributaries in many areas is probably 100 years or less (Dietrich and Dunne 1978), so chemical weathering probably has little effect on particle hardness. In the long term, the relative proportions of bedload-size and suspended-size particles discharged from these channels must largely reflect the texture of the soil, with the coarsest fraction being transported by debris flows in some regions. Along the floors of higher order valleys, the residence time of sediment in storage is much longer, so that chemical weathering and fluvial transport can lead to dramatic shifts in the particle-size distribution of the load, as coarse particles break into finer sizes. The rate of breakdown is greatly influenced by the degree of weathering of debris in the parent soil. Dietrich and Dunne (1978, p. 200) have summarized field and laboratory evidence for this breakdown, but quantitative applications of laboratory studies to the field setting are still lacking.

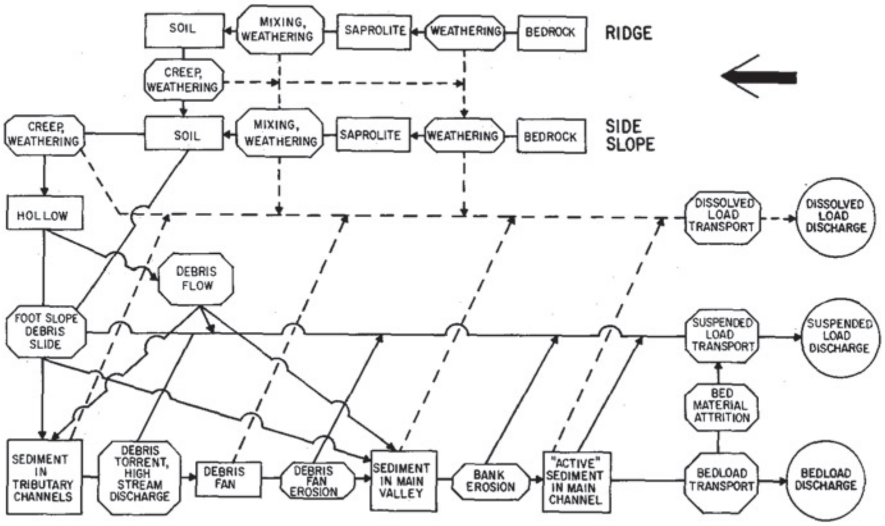



Figure 2—Sediment-budget model for the Rock Creek basin, central coastal Oregon: rectangles, storage sites; octagonals, transfer processes; circles, outputs; solid lines, transfer of sediment; and dotted lines, migration of solutes (from Dietrich and Dunne 1978).

In a previous paper (Dietrich and Dunne 1978), we attempted to show the importance of breakdown of weathered sediment in the 16-km$^2$ Rock Creek basin of Oregon. Sediment with settling velocities less than the shear velocity tends to stay in suspension (e.g., Middleton 1976); and in steep channels in our study area, the settling velocity of 2-mm-diameter sediment (about 21 cm/second) is generally exceeded at high flows by the local shear velocities estimated from the downstream component of the weight of the fluid. About 43 percent of the soil mass discharged into stream channels is less than 2 mm in diameter and would, therefore, travel primarily as suspended load. In gravel-bedded streams of the western United States, measured bedload transport ranges from 4 to 10 percent of the total solid load (Rosgen, D., USDA Forest Service, personal communication, based on many measurements in the Rocky Mountains; Emmett and Seitz 1973, 1974). Higher ratios have been estimated in some small basins (Larson and Sidle 1980). If the ratio of bedload discharge to total solid-load discharge is 10 percent at the mouth of Rock Creek, about 82 percent of the bedload-size material discharged from the soils must break down to suspended load during transport to the basin outlet. This calculation assumes that, over the long time scale, no change occurs in the volume of coarse sediment in storage in the basin. The presence of bedrock at or near the surface throughout most of the Rock Creek drainage network points to the lack of important storage.

The identification of linkages is critically important in distinguishing those erosion processes that are separate additive components of sediment discharge from the processes that are carrying the same material along sequential portions of its path through the basin. In forested mountainous landscapes, a fundamental and yet unresolved problem is the definition of the relation of the ubiquitous—but slow—processes of soil creep and biogenic transport to the rapid transport of sediment by landslides, so that the contribution of each process to the sediment budget can be separated. To investigate this problem for a drainage basin, it is necessary to estimate the relative importance of types of hillslope-transport processes and their distribution within the basin. We have examined this problem in two regions of the low coastal mountains of Oregon and Washington underlain by volcanic and sedimentary rocks.

We first recognized two distinctive soil types. Thin, gravelly soils cover most of the landscape and vary systematically with hillslope angle (fig. 3). Thick, sandy and gravelly soils in U-shaped bedrock depressions, which we have called soil "wedges" (Dietrich and Dunne 1978), show no clear variation with hillslope angle. The wedges (fig. 4), which appear to be filled landslide scars, occur at the headwaters of.most first-order tributaries and intersect 10 percent of the perimeter of the first-order channels and up to 50 percent of third- and fourth-order channels. We see no evidence of water erosion in the basins represented in figure 3. In both drainage basins, deep-seated landslides incorporating significant amounts of saprolite or bedrock are rare. Instead, most landslides involve just the soil mantle and commonly emanate from wedges. Because the form of the bedrock depression concentrates subsurface water into wedges, landsliding occurs more frequently there than in surrounding thin soils. The shallow, coarse-textured soils overlie competent weathered bedrock, and we have seen no evidence for active earthflows. In a previous paper (Dietrich and Dunne 1978), we suggested that the creep rate in soil wedges is four times faster than in the surrounding thin soils. This proposal was based on our misinterpretation of field measurements by Swanston and Swanson (1976). Their data pertain only to earthflows associated with large landslides rather than to soil creep. At present, we know of no program of measurements on the discharge from wedges by soil creep.

Soil creep, biogenic transport, and debris sliding are not separate additive components to total discharge of soil into channels. Soil that is finally discharged to the channel by one process may have been carried along most of the slope by other processes. Therefore, the total flux of sediment to channels should not be computed by adding the discharge from debris slides to an average soil-creep rate. This rela-

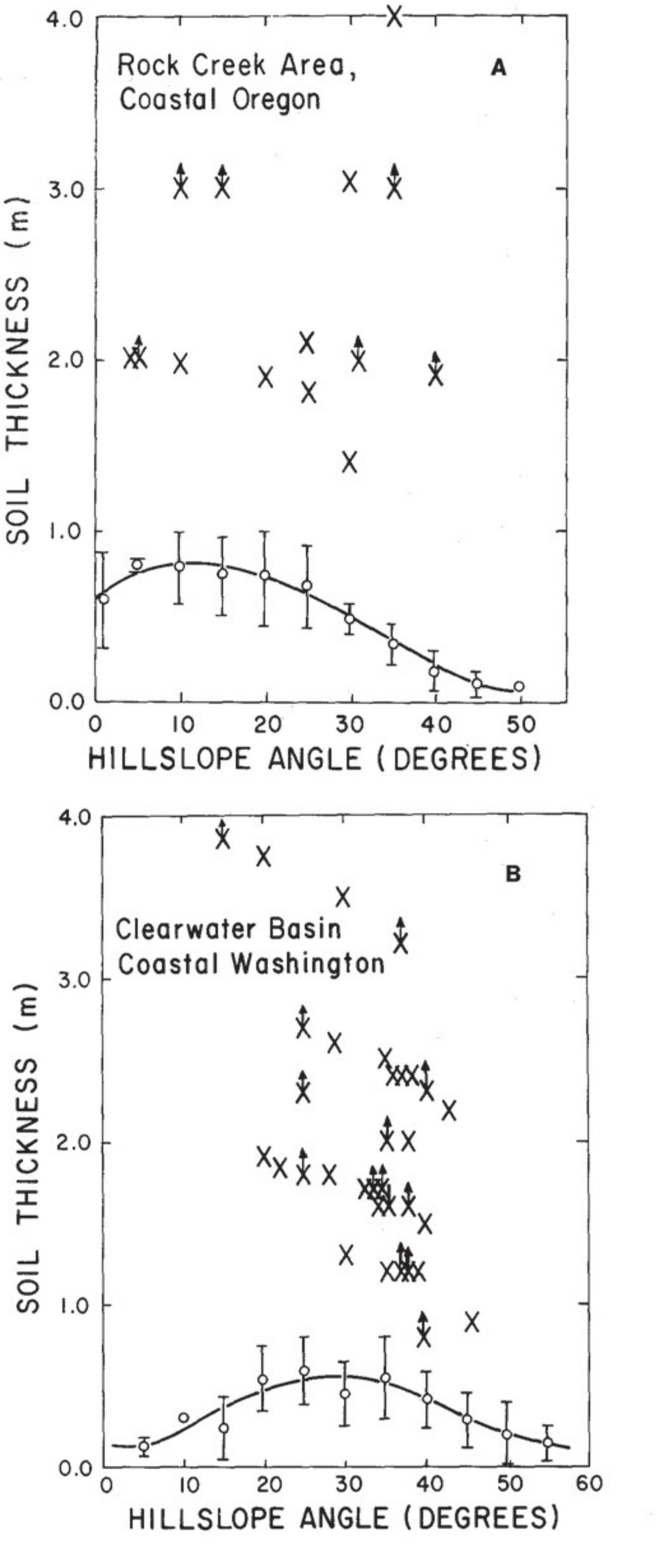



Figure 3—A. Relationship of hillslope gradient to soil depth for soils in the Rock Creek area of coastal Oregon. B. Clearwater basin of the Olympic Peninsula of Washington. Measurements were made along road cuts. Circles represent the average of depths for 5-degree classes, and the bars are the standard deviations. The crosses represent the depth for wedges and, there the total depth was not exposed, an arrow was attached to the symbol.

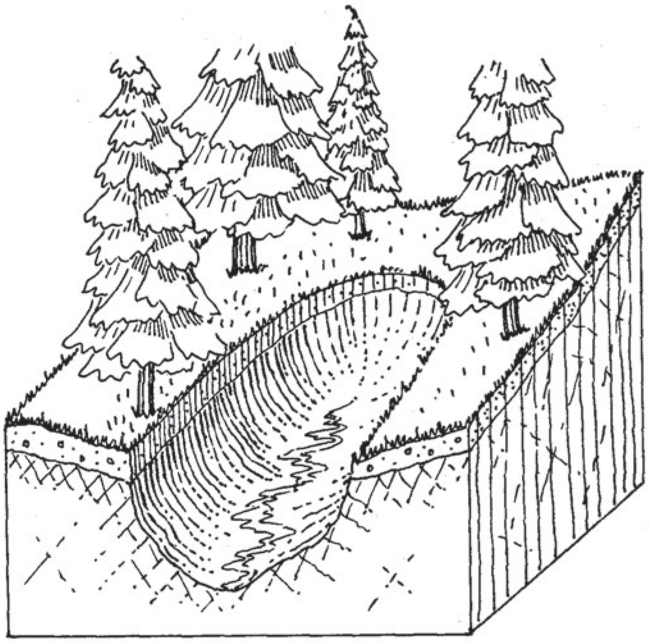
Figure 4—Soil wedge partially filling depression in sedimentary bedrock, western Olympic Peninsula. Note the characteristic “U” -shaped depression and the great soil depth compared to adjacent areas.

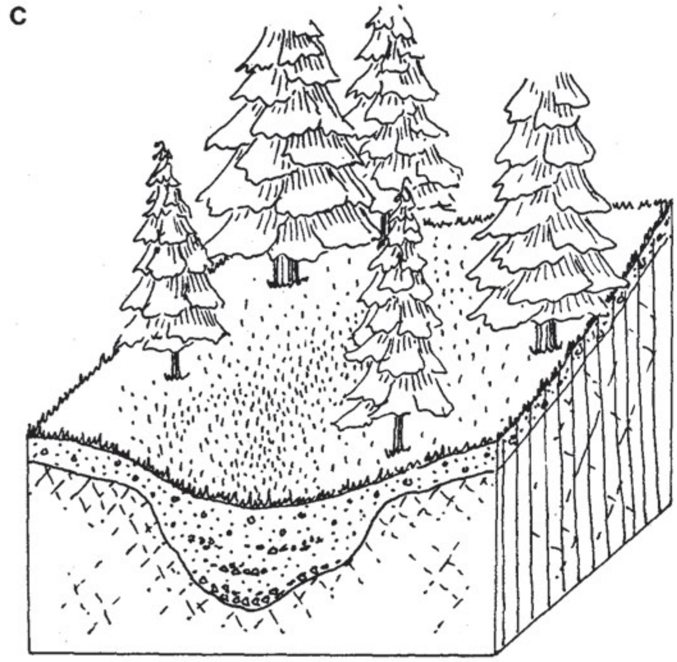
tionship can be visualized, for example, by examining what happens when a debris slide erupts from a soil wedge.

A debris slide may develop into a debris torrent and scour the channel and footslopes of first- and second-order basins. On the bed and walls of the slide scar, fresh and weathered bedrock are exposed and near-vertical walls are left in the surrounding soils (fig. 5a). After the slide, the bedrock weathers rapidly along fractures and joints; and the vertical soil face degrades through rain splash, raveling, sloughing, and small-scale slumping of soil into the scar. Removal of soil from the face and the steepening and destabilizing of soil upslope also lead to accelerated tree-throw around the margins of the scar and accelerated soil creep. The distance over which accelerated soil movement extends upslope depends on the time and amount of soil required to refill the scar.

Soil initially discharged into the bare scar will be subject to concentrated water flow, and only the coarsest particles will remain in the depression as a gravel layer (fig. 5b). Continued discharge of poorly sorted soil into the scar forms a thickening wedge of partially reworked sediment from which saturation overland flow (Dunne 1978) is generated progressively less frequently. When saturation overland flow can no longer be produced in the wedge, soil discharged into the scar accumulates without reworking by water and thus retains the textural characteristics of the surrounding profile from which it is derived (fig. 5c). Figure 6a illustrates the changes in texture of deposits with time, and figure 6b portrays the rate of filling in hollows as estimated from dendrochronology.

During the early phases of refilling, the scar will be a source of high sediment discharge to the stream. Based on the dendrochronology shown in figure 6b and the corresponding change in sediment texture (fig. ba), we estimate that this period will last about 100 years. By the end of this period,





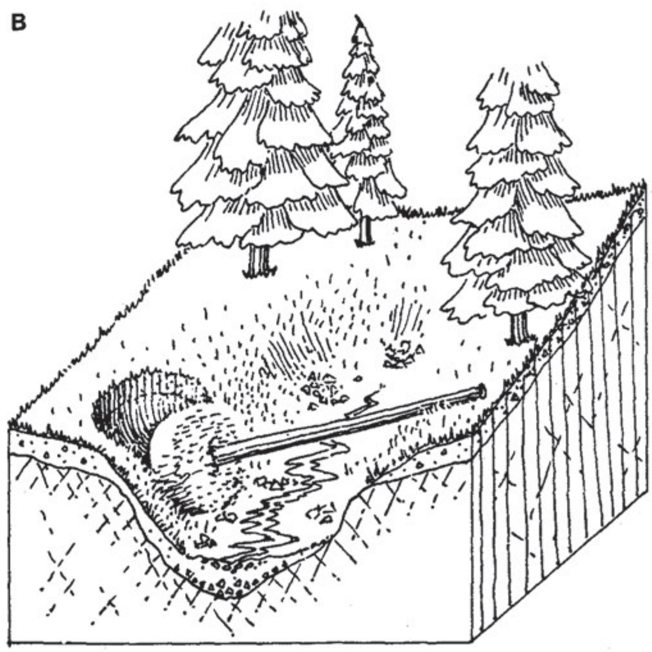




Figure 5—Evolution from a landslide scar in a bedrock depression to a soil wedge. After the landslide, the exposed bedorck surface forms an impermeable horizon shedding rainwater and subsurface discharge into the depression as overland flow. B. Sediment eroded from the over-steepened soil perimeter into the depression is washed of its fine component, leaving a gravel-lag deposit covering the rock surface. C. Continued deposition leads to less frequent saturation overland flow and less surface transport. Eventually, the lack of surface wash causes the soil near the surface of the soil wedge to become similar in texture to surrounding soil from which it is derived.

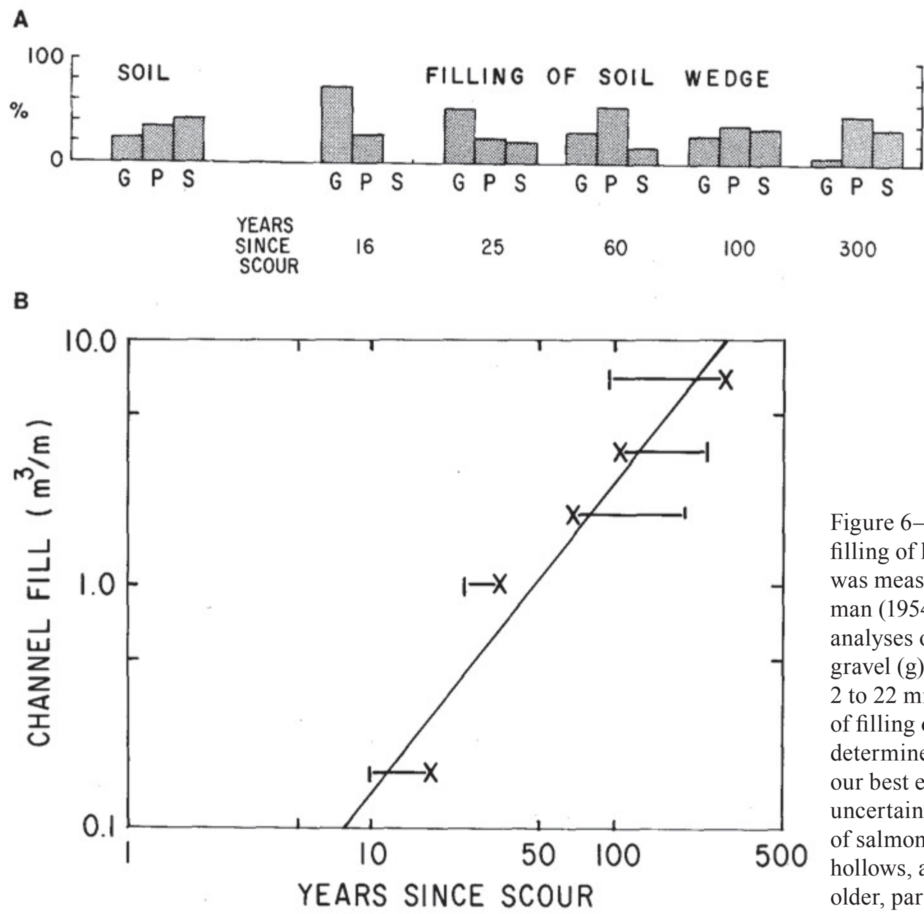



Figure 6—A. Successive fining of material during filling of hollows. The fraction greater than 22 mm was measured by the pebble-count method of Wolman (1954), and the data were combined with sieve analyses of the finer particles. The letters indicate gravel (g) (greater than 22 mm), pebbles (p) (from 2 to 22 mm) and sand (s) (less than 2 mm). B. Rate of filling of scoured hollows. Time of scour was determine by ages of vegetationm on fills. (X) is our best estimate, error bar represents the range of uncertainty in our estimate. Vegetation consisted of salmonberry and alder on more recently sourced hollows, andlarge alder, hemlock, and Douglas-fir on older, partially filled hollows.

the texture of the upper layer of the wedge is similar to that of the surrounding soil. As the scar fills, sediment discharge from the wedge decreases. The scar may fill completely so that no topographic expression remains, or it may fail during a large storm that occurs as it is filling. During accumulation, sediment discharge from the wedge must be less than prefailure levels. When the filling is complete or when an equilibrium depth is reached that balances the influx and discharge of soil, the latter attains its prefailure rate (fig. 7).

The time required to refill a depression can be estimated by computing a creep discharge rate into the slide scar across its exposed perimeter. For example, a creep rate of 3 mm/year in a 50-cm-thick soil will refill a bedrock depression 5 m wide, 20 m long, and 2 m deep in 3,000 years. This estimate represents a minimum because much of the initial soil discharge into the exposed hollow will be washed out. An increase in the frequency of failure because of a climatic change or management activity would accelerate soil movement towards the scar and thin the surrounding soil over periods of hundreds to thousands of years. Debris-slide scars then probably fill between 1,000 and 10,000 years after initial failure. Although transport into the scar will be accelerated by the sloughing of soil and gullying of exposed soil, the ultimate transport rate into the scar will be limited by the supply of soil and the rate of soil transport toward the location of failure.

On the short time scale for which a quantitative sediment budget might be developed from a monitoring program, debris-slide scars will generally be sources of high sediment discharge. That the period of increased erosion extends for a long time after the slide occurs is well illustrated by Tanaka's (1976) measurements in the Tanzawa Mountains of Japan. His repeated topographic surveys showed rates of sediment discharge from 50-year-old debris scars to be about 100 times greater than the estimated undisturbed rate of sediment discharge from hillslopes. Lundgren (1978) has also reported that in the subhumid mountains of Tanzania, erosion during a 7-year period after formation of landslide scars was as great as the initial loss from the landslides.

If debris slides lead to accelerated weathering of the underlying bedrock during exposure and burial, or if they cause loss of debris from the weathered bedrock in the scar or its surroundings, then a component of their discharge can be defined as a contribution to the sediment budget separate from creep and biogenic transport. Otherwise, debris slides emanating from the soil mantle act more as periodic fluctuations in the rate of discharge of sediment from the hillslope by soil creep and biogenic transport.

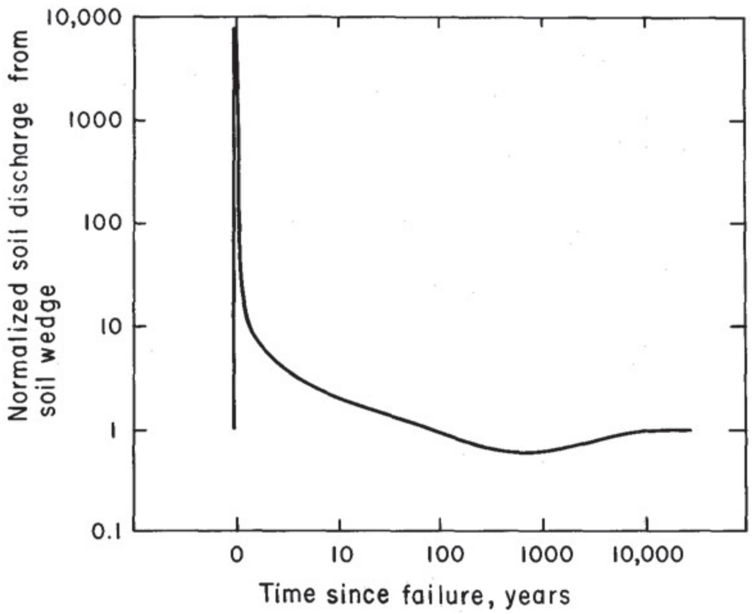



Figure 7—Chronologic variation of sediment discharge into a stream channel as a result of failure of a soil wedge and the eventual refilling of the bedrock depression. The annual sediment discharge for each year was divided by the average annual sediment discharge to normalize the curve.

## DEFINITION OF RECURRENCE INTERVALS

The degree to which one process affects another and the importance of a transport process to the sediment budget are both dependent on the magnitude and frequency of recurrence of the process at a place. Definition of the probability distribution for some transport processes (e.g., sheetwash from roads) is relatively easy. Definition of a recurrence interval of landslides is a difficult problem in steep, forested drainage basins, however. Frequency of landsliding depends on the stochastic properties of external controls—such as storm intensity, antecedent weather, and earthquakes. The importance of these controls at a site will depend on specific site conditions such as soil permeability, vegetation cover, groundwater flow path, and soil strength. The frequency of landsliding, then, needs to be specified at a site. In the Coast Ranges of Oregon and Washington, landsliding occurs with varying frequency from topographic hollows, filled bedrock hollows, and planar slopes. A regional landslide frequency cannot be applied to a specific site, unless all the failures used to define the regional frequency originate from the same type of site (e.g., filled bedrock depressions at the heads of valleys). Then space can be substituted for time in the calculation of recurrence interval.

Estimates of regional rates of landsliding can be obtained from sequential aerial photographs and from dendrochronology, but the occurrence of major storms confounds a simple computation of long-term frequency of these events. No simple correlation can be made between recurrence interval of the hydrologic event and the recurrence of landsliding because of site-specific controls on the hydrologic events and because of lag effects resulting from long recovery times from previous landsliding. Similarly, landsliding caused by earthquakes cannot be simply related to the frequency of occurrence of the earthquakes, as done by Garwood et al. (1979). Long periods of record and large samples wall eventually allow the definition of useful mean recurrence intervals from each type of site, but—at present—the aerial photographic record is too short for accurate determination of recurrence intervals.

Definition of recurrence interval becomes particularly difficult when the landslide is not an instantaneous discharge of soil into a channel but is a deep-seated failure that moves episodically. Not only is relating the movement to the probability of occurrence of the specific site conditions and hydrologic events difficult, but to generalize the probability of failure to similar sites is extremely difficult. This problem is further complicated by the tendency for slumping to generate debris avalanches at the channel perimeter (Swanson and Swanston 1977), creating an interdependency of frequency and magnitude between different processes.

Landslide frequency is traditionally defined for a large area from sequential aerial photographs without specification of landslide site conditions. This frequency is assumed to apply to basins of any size in the area. Not only does this approach neglect effects of major storms, but it also ignores the effect of basin size, which clearly influences the range and frequency of site conditions. The effect of basin scale is an important one and has been discussed by Wolman and Gerson (1978) as it applies to stream morphology.

Quantitative definition of the role of landsliding in a sediment budget can be approached in several ways. For the simplest case in which landsliding involves just the soil mantle and essentially represents the last step in the transport of soil to the stream channel, the problem of recurrence interval can be ignored as long as some estimate can be made of the rates of hillslope transport by other processes. This is only appropriate, however, in a sediment budget for a long period during which it can be assumed that the transport processes are in equilibrium and major landforms are not changing substantially.

Another approach would be to construct a sediment budget for a specific interval only and not to develop a predictive model of the transport processes. Only landslides generated within the study basin would be considered. Similarly, only measurements of other processes within the basin would be used. The value of this procedure depends greatly on the length of time over which measurements can be taken. This method, however, allows semiquantitative examination of the effects on landsliding of major differences in basic controls such as geology, climate, or land use.

We have suggested above that a regional landslide frequency analysis could be made for specific sites, particularly if the time base can be expanded using dendrochronologic or other dating techniques. The frequency of site types within a basin would then be used to construct a sediment budget. This problem is also discussed by Kelsey in his Discussion Group Summary for this workshop. Although confusing effects of unequally distributed meteorologic events and stand history would degrade the quantitative results, this procedure would yield a meaningful, predictive sediment budget.

The most general approach in the long term would be to develop models for hillslope instability. In the simple case in which debris slides are caused primarily by adverse combinations of soil strength and high pore-water pressure, recurrence interval of landsliding at a site may eventually be defined through a combination of a deterministic model of hillslope stability, a deterministic model of pore-water pressure generated by precipitation, and a stochastic expression for precipitation occurrence. Included in this model would be the change of soil depth with time as the site refilled from the previous landslide. Regrettably, definition of recurrence of landsliding is not an easy problem.

## QUANTIFICATION OF STORAGE ELEMENTS

Sediment storage elements in the landscape are the medium through which transport processes act, and therefore their quantification is of great importance in the construction of a sediment budget. To discuss the characteristics of different types of storage elements (or sediment reservoirs) in a drainage basin and to examine their interaction, we briefly define a few essential terms used in reservoir theory as presented most simply in Eriksson (1971) and Bolin and Rodhe (1973). A variety of sediment reservoirs exist in a drainage basin, ranging from vast quantities of debris stored in the soil mantle on a hillslope to small accumulations in gravel bars. Each accumulation can be characterized by the age distribution of sediment in the reservoir and the age distribution of sediment leaving the reservoir. Age is used here to mean the time ($\tau$) since the sediment entered the-reservoir. The first distribution could be defined in the field by dating the deposits in a reservoir and by developing the cumulative curve of mass $M(\tau)$ in the reservoir, less than or equal to a certain age ($\tau$) such that

$$\lim_{\tau \to \infty} M(\tau) = M_0 \tag{1}$$

where $M_0$ is the total mass of the reservoir. Everitt (1968), for example, aged flood-plain deposits on the Little Missouri River using dendrochronology. Taking the derivative of this cumulative curve, with respect to the age of sediment, creates an age-distribution function $\Psi(\tau)$, where

$$\Psi(\tau) = \frac{1}{M_0} \frac{dM(\tau)}{d\tau} \tag{2}$$

The average age of sediment in a reservoir, $T_a$, can be computed by integrating the cumulative curve with respect to mass and dividing by the total mass, which is equivalent to computing a weighted mean age (Eriksson 1971):

$$T_a = \frac{1}{M_0} \int_0^\infty \tau \, dM(\tau) \tag{3}$$

$$= \int_0^\infty \tau \Psi(\tau) \, d\tau \tag{4}$$

For many sediment reservoirs, the age-distribution function will be impossible to define unless the reservoir contains natural tracers, such as vegetation and tephra, or artificial tracers, such as isotopic lead (Barnes et al. 1979), radioactive fallout (Ritchie et al. 1975), painted rocks (e.g., Laronne and Carson 1976), and anthropogenic debris (Costa 1975). These tracers need to be distributed in time for periods of centuries and in space for sites representing the range in age in a reservoir to yield meaningful results about most sediment reservoirs. If sediments of different ages become well mixed, however, quantitative evaluation of the age-distribution function will be difficult.

Reservoirs can also be characterized by the age distribution of sediment leaving the reservoir. If it were possible to evaluate the time spent in storage for each particle that leaves the reservoir, then for a particular time increment, a cumulative curve of discharged mass of sediment leaving the reservoir less than or equal to a certain age, $F(\tau)$, could be constructed. This curve is called a "transit time" function because it gives the distribution of transit times through the reservoir for the discharged sediment. As $\tau$ becomes very large, $F(\tau)$ approaches the total flux per unit time out of the reservoir:

$$\lim_{\tau \to \infty} F(\tau) = F_0 \tag{5}$$

As in equation 2, the derivative of the cumulative curve with respect to age yields the distribution function of transit time:

$$\Phi(\tau) = \frac{1}{F_0} \frac{dF(\tau)}{d\tau} \tag{6}$$

The average transit time of sediment discharged from the reservoir, which is defined to be the residence time of the sediment in the reservoir (Bolin and Rodhe 1973), is the integral of the cumulative curve of transit times divided by the total flux per unit time:

$$T_t = \frac{1}{F_o} \int_o^{\infty} \tau \, dF(\tau) \qquad (7a)$$

$$= \int_o^{\infty} \tau \Phi(\tau) d\tau \qquad (7b)$$

Characterization of transit times from field measurements of sediment discharged from a reservoir is extremely difficult. Most studies that attempted to define transit times have been for fluvial transport processes and used painted rocks. These studies have failed to recognize three important aspects of sediment reservoirs, however, and as a result they have not yielded useful information.

First, it is necessary to recognize that measurements of the movement of marked particles is a transit-time study of a particular reservoir or set of interacting reservoirs, and, therefore, defining the various types of reservoirs in an area and their boundaries is also necessary. In fluvial studies, at least two types of reservoirs may be recognized: flood plains, bars, and active streambed. The computed residence time of particles in these reservoirs will depend partly on the exchange rate between reservoirs and on the definition of reservoir boundaries. Second, transit time of sediment depends on where in a reservoir the marked particles are placed. In stream channels, the particle-size distribution of the bed material across the channel is a function of the boundary shear-stress distribution. If the goal is to typify the transit-time characteristics of the reservoir, introduced marked particles must duplicate the particle-size distribution across the channel. Further, the particles must be placed throughout the active depth of the channel sediment, not just at the surface as is normally done. This consideration may make a difference of greater than 20 times in the measured transit times of particles (based on computations by Dietrich and Dunne 1978). Third, to compute transit times, all particles must be recovered. Unfortunately, recovery rate of marked particles is generally very low; following all particles without seriously disturbing the reservoir would be nearly impossible. Regrettably, this last problem is a major constraint. We will address the problem again later. Further discussion of the theory of tracers can be found in Eriksson (1971) and Nir and Lewis (1975).

Definition of the residence time of sediment in a reservoir from measurement of transit times is difficult. As an alternative, residence time can sometimes be computed.from the age distribution or estimated average age of sediment in the reservoir. Eriksson (1971) has shown that under steady-state conditions, as sediment of a particular age leaves the reservoir, a corresponding aging of sediment remaining in the reservoir must occur, such that

$$F_o - F(\tau) = \frac{dM(\tau)}{d\tau} \qquad (8a)$$

The left side of equation 8a gives the amount of sediment greater than age $\tau$ that leaves the reservoir per unit time. To maintain a steadystate age distribution, the amount of sediment with an age greater than $\tau$ that remains in the reservoir must increase at a rate equal to this rate of efflux, as given by the right side of the equation. The derivative with respect to $\tau$ of each side of the equation yields the transit-time density function,

$$\frac{dF(\tau)}{d\tau} = -\frac{d^2M(\tau)}{(d\tau)^2} \qquad (8b)$$

Although the transit-time function can be computed from the age distribution of sediment in the reservoir from equation 8b, the function is very sensitive to small errors in definition of the age distribution and can only be done practically for poorly mixed or unmixed reservoirs. Such a reservoir might be a flood plain constructed predominantly by lateral accretion of bedload on point bars. A simple example, however, illustrates that even in a poorly mixed reservoir such as a flood plain, the age distribution may be time- and space-dependent.

Consider the simple case, illustrated in figure 8, of a stream channel migrating uniformly across its flood plain at the rate of one channel width per year. If the deposits were sampled when the channel was in the three positions in figure 8a, b, and c, the average age according to equation 3 would be 5.5, 8.0, and 6.4 years, respectively. This time dependency may be overcome in the field by expanding the upstream and downstream boundaries of the reservoir so that the channel position ranges back and forth across the valley floor at any time.

To examine the relation between residence time and average age, consider two steady-state models of a gravel bar illustrated in figures 9a and b. This first is a piston model in which gravel enters at one end of the reservoir and leaves exactly 10 years later. Clearly, the residence time of the gravel in the bar would be 10 years, and the average age of the sediment would be 5 years. By contrast, in the second model, although the annual discharge rate is the same, 60 percent of the gravel in any year enters and leaves the reservoir, while the other 40 percent becomes deeply incorporated in the bar and takes 25 years to leave the reservoir. The average transit time of sediment leaving the bar is (0 x .6) + (25 x .4) = 10 years, the same as in the first model. But the average age of sediment in the reservoir will be 13 years.

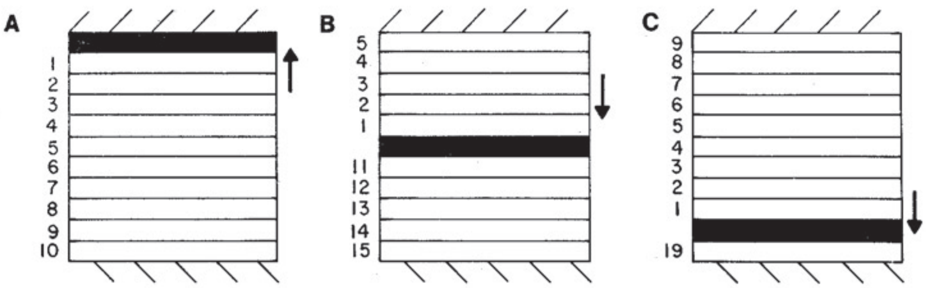



Figure 8—A-C. Change in age distribution of flood-plain sediment reservoir resulting from uniform lateral migration of the active channel (darkened area) at one channel-width per year. Numbers in columns give age in years for each increment of flood-plain deposit.

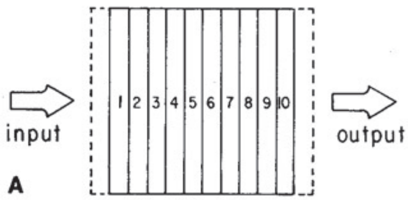
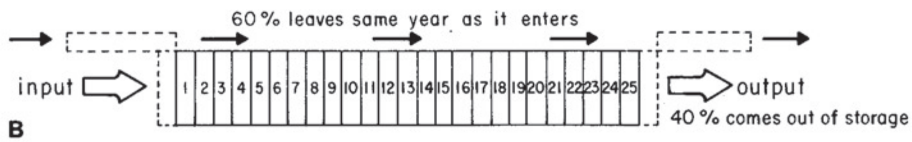



Figure 9—A-B. Two models of transfer of sediment through a gravel bar. Numbers are the age of the sediment in particular parts of each reservoir. Total sediment flux rate is the same for both models but, in the second, 60 percent sediment entering the reservoir leaves immediately, and the remaining 40 percent travels slowly through the entire reservoir.




Bolin and Rodhe (1973) have demonstrated that even in steady-state reservoirs, average age of sediment may be less than, equal to, or greater than the residence time. They also showed mathematically that the turnover time of a reservoir defined as the total mass in the reservoir ($M_0$) divided by the total flux rate ($F_0$) equals the residence time of sediment in a reservoir in the steady-state condition:

$$T_r = \frac{M_o}{F_o} \qquad (9)$$

Where the reservoir can be treated as if in steady state, the estimated residence time becomes a much more tractable problem. In the short period of a few years, most storage elements in a landscape are probably very poorly represented by the steady-state assumption. Over the long period, however, the average quantity of sediment in storage in some reservoirs may be considered relatively constant. For example, in small alluvial valleys with steep hillslope boundaries in which the channel is close to bedrock, the assumption of steady-state storage of sediment in the flood plain is probably reasonable over a period of a few decades, particularly if the sediment flux into and out of the foodplain occurs primarily by bank erosion and channel migration rather than by episodic deposition or scour by debris torrents.

A river-valley segment of the Little Missouri River of western North Dakota which approximately meets these requirements was intensively surveyed by Everitt (1968). He mapped the age distribution of cottonwood trees on a 34,000-m-long length of flood plain formed by a freely meandering, 91-m-wide, sand-bedded river in a 900-m-wide valley. At the time of the study, the channel had recently cut off two large bends and was subject to ice jams that caused it to shift rapidly across its flood plain. As a result, the density function of flood-plain area occupied by sequential 25-year age classes of sediment was negatively exponential. From Everitt's data on the area and elevation of sediment above low water in various age classes, we have computed an approximate age-distribution function, M(τ), for the volume of sediment in the flood plain. The derivative of this function is the right side of equation 8a; if one can assume that no flood-plain sediment enters and leaves the reach in the same year (i.e., F(τ) = 0 when τ = 0), equation 8a can be used to obtain $F_0$, the flux rate of sediment through the reservoir. Everitt fitted a power function to the height above low water of flood-plain sediment of various ages and obtained results similar to the following by a different technique. We used the power function of height and the exponential function of area to compute the age-distribution function, M(τ), which was integrated to obtain a total volume of sediment, $M_0$, of $1.08 \times 10^7$ m$^3$. Differentiating M(τ) with respect to τ yields the right side of equation 8a and, therefore, at τ = 0, a flux rate, $F_0$, of $1.08 \times 10^5$ m$^3$/year. Substituting this value and $M_0$ into equation 9 leads to a residence time of 100 years.

The transit-time function for Everitt's data (obtained by twice differentiating M(τ), substituting into equation 8b, and integrating the left side) suggests that most of the sediment spends less than 1 year in the flood plain. This counterintuitive result may be an artifact of using a power function to express the height of the flood plain above low water for a given age because this combines with the exponential curve for the area-distribution function to place the bulk of the flood-plain sediment in a very young age-class.

Another reservoir that might be treated as having a constant volume in storage over long periods is the debris fan. The volume of a fan has generally been found to be a power function of the drainage area of the basin contributing to it (Bull 1964). We have also found this relationship for a small drainage basin in coastal Oregon (fig. 10a). If the long-term average annual discharge into or out of the fan can be estimated from short-term measurements, then the average residence time in a debris fan can be computed approximately from equation 9.

A third reservoir that might more commonly be in steady state is the "active" channel sediment that comprises all of the sediment in the channel bed, including gravel bars, down to a rarely attained maximum scour depth and across to channel boundaries marked either by a distinct bank or by vegetation changes. "Active" sediment is not a meaningful term in some low-order tributaries where sediment is moved primarily by episodic debris torrents. Unless channel storage is changing dramatically from modifications in channel geometry in response to a significant change in sediment load or flow characteristics, the "active" stored sediment in, say, a length of channel of about 10 channel widths is roughly constant, even in aggrading and degrading streams. If the flux rate along the channel and into or out of the flood plain can be estimated, then residence time can be computed.

The time scale that defines the period over which steady state is to be assumed is proportional to the residence time of sediment in the reservoir. If disturbances in flux rate occur over the same time scale, then steady state is a poor approximation. Also, many disturbances from forest management occur on a time scale shorter than the period over which steady state might be assumed. Nonsteady state implies the sediment reservoir is not only characterized by age distributions but also by absolute time. We know of no simple general theory for dealing with the nonsteady state although work has been done toward that end (Nir and Lewis 1975, Lewis and Nir 1978).

In both the steady- and nonsteady-state cases, another approach to computing residence time and developing a sediment budget is to construct a transition probability matrix to define the flux into and out of one reservoir and into another. For example, consider the transport of sediment along an alluvial valley floor. Two general sediment-storage elements, the active channel (defined above) and all the other flood-plain deposits, can be quantified by surveying; the amount of sediment in storage can be related to distance along the

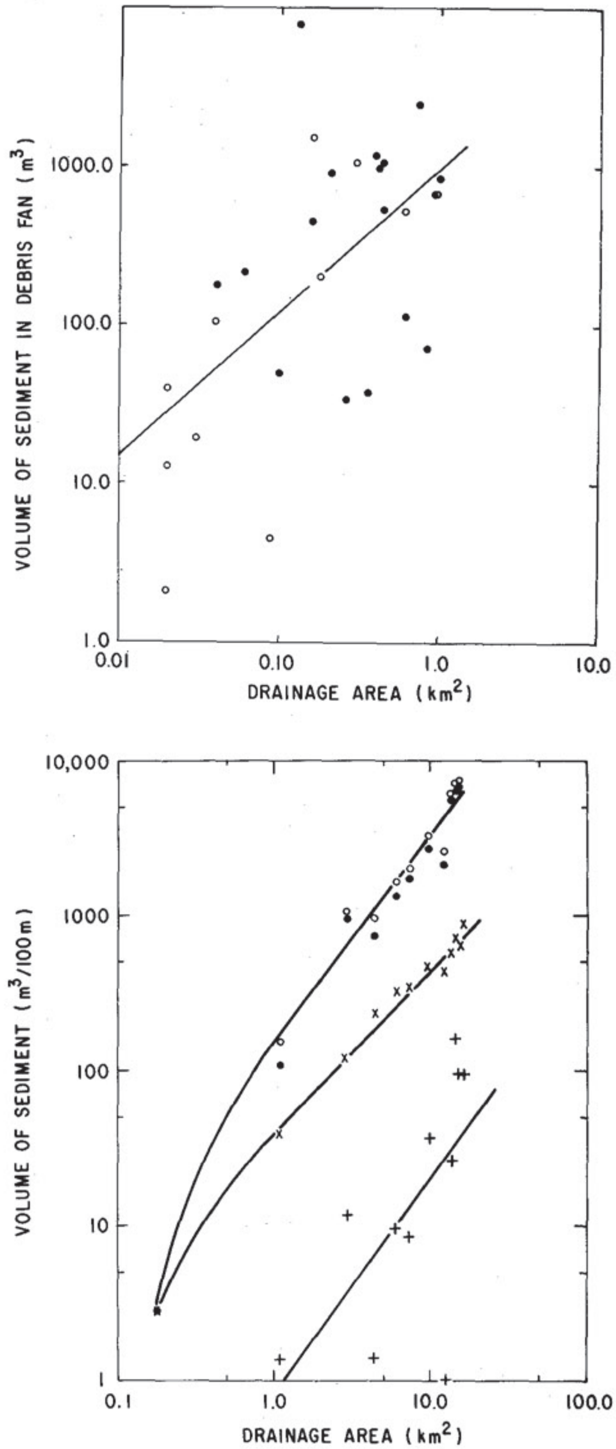



Figure 10—A. Volume of debris fans as a function of tributary drainage area along the main channel of the 16.2-km² Rock Creek drainage basin, Oregon. Open circles are debris fans covered by salmonberry and young alder trees. Dots are fans covered by more mature trees. B. Volume of sediment in gravel bars (pluses), "active" channel (crosses)., flood plain (dots), and total valley sediment (open circles) along the main channel of Rock Creek. Equations for least squares fit to the data in l0A and B are given by Dietrich and Dunne (1978).

channel from headwaters (fig. 10b). The sediment in storage per unit length of channel can be depicted by the column matrix

$$V = \begin{vmatrix} x \\ y \end{vmatrix}$$

where x and y are the volume of sediment in storage per unit length of valley in the active channel and the flood-plain deposits, respectively.

The exchange between the two reservoirs caused by such channel changes as migration, widening or narrowing, or downcutting can be represented by the transition probability matrix

$$S = \begin{vmatrix} a & b \\ c & d \end{vmatrix}$$

where the first row represents the proportion per unit time of the total sediment in the active channel that either remains in the active channel (a) or is transferred to the flood plain (b). Similarly, (c) is the proportion of flood-plain sediment eroded into the active channel, and (d) is the proportion that remains. If the transpose of S is taken and this new matrix $S^T$ is multiplied times V, a column matrix is formed by the product expressing the amount of sediment in the two reservoirs as a result of the exchanges specified by S.

The input and output of sediment can be represented by column matrices $Q_i$ and $Q_0$, respectively,

$$Q_i = \begin{vmatrix} i \\ j \end{vmatrix} \quad \text{and} \quad Q_o = \begin{vmatrix} k \\ m \end{vmatrix}$$

where i and j are the volumes per unit time added to the active channel and flood plain; k and m are the volumes discharged from each, per unit time. Values other than zero for j and m occur as a result of overbank flow. The net increase or decrease of sediment stored in each reservoir after a unit time is given by the column matrix

$$\Delta V = \begin{vmatrix} r \\ s \end{vmatrix}$$

The continuity of mass per unit length of channel can now be written as

$$V - S^T V + Q_i = Q_o + \Delta V \qquad (10)$$

This is a general equation for any system of reservoirs in steady or nonsteady state. In many sediment routing problems, each term can be quantified approximately from fairly simple field measurements. A valuable way to use equation 10 would be to establish a series of surveyed cross sections across the valley floor. Repeated survey at these sections, recording of channel scour depths with scour chains, and measuring of sediment transport both in the channel and on the flood plain at the upstream and downstream end of the channel length should allow computation of a sediment budget using equation 10. In this form, equation 10 has very little predictive value. In the nonsteady-state problem particularly, it is best thought of as the basic mass balance into which field measurements are placed to define the routing of sediment.

At steady state, however, no change takes place in storage ($\Delta V = 0$); although (i + j) must equal (k + m), $Q_0$ need not equal $Q_i$. This inequality could occur by the deposition of flood-plain sediment during overbank flow being compensated for by bank erosion of flood-plain sediment into the active channel. In other cases, equation 10 can be greatly simplified if a steady state exists and $Q_0 = Q_i$, such that

$$V - S^T V = 0 \qquad (11)$$

Given that V is known, an infinite number of $S^T$ matrices can solve equation 11, but the actual values of a, b, c, and d are related by the simplification of the two linear equations represented by the equation

$$xb = yc \qquad (12)$$

This simply states that volumes exchanged between the flood plain and the active channel must be equal. Thus, in the steady state described above, if the volume of sediment stored in each reservoir is known and if a single term can be defined in S, such as the annual rate of bank erosion, then the other three terms can be computed. In the steadystate case, the sediment budget can be defined from simple field measurements and, according to equation 9, the residence time for sediment in the active channel ($T_{ra}$) and in the flood plain ($T_{rf}$) is

$$T_{ra} = \frac{x}{xb + i} \qquad (13a)$$

$$T_{rf} = \frac{y}{yc + j} \qquad (13b)$$

These residence times are the average time a particle spends in the reservoirs. If the question is how long on the average sediment will take to move some distance along the valley, however, flux rate must be defined in terms of transport across the upstream and downstream ends of the reach, including transport along the flood plain. Exchange between the flood plain and the active channel has no effect on the residence time in the valley floor. Channel migration influences residence time of sediment in the separate reservoirs, but it does not affect the downstream transit time unless downstream flux rates are dependent on rates of bank erosion.

Consider three possible transport relationships for the two major reservoirs defined above. In the first case, no exchange occurs between reservoirs. $S^T$ becomes the identity matrix, and the residence time, which is equal to the average travel time per unit length of the valley, is x/i and y/j for the transport through the active channel and the flood plain, respectively. In a more realistic case, some exchange rate occurs between the two reservoirs, typically by channel migration, but the transport of sediment along the flood plain occurs at an insignificant rate. Thus the residence time of sediment in the valley floor ($T_{rv}$) is

$$T_{rv} = \frac{x + y}{i} \qquad (14)$$

because j ≈ 0. The residence time for sediment in the flood plain alone is given by equation 13b which simplifies to 1/c, the inverse of the probability of transition from flood plain to channel in S. The residence time of a particle in the active channel alone is given by equation 13a.

Thus, the residence time in the active channel is always much less than the residence time of sediment in the valley floor. The residence time for sediment in the flood plain can be greater than or less than residence time for the valley floor because, in the present analysis, residence time in the flood plain refers to the time since the previous transfer of a particle into the flood plain, either from the channel or from upstream. In many rivers, flood-plain sediment transport and reservoir exchange are common. In this case, the average transit time becomes

$$T_{rv} = \frac{x + y}{i + j} \qquad (15)$$

and the residence time in the flood plain decreases to that given by equation 13b.

Sediment storage and sediment transport rate are functions of distance downstream (fig. 10), and equation 14 can be integrated to compute the average travel time between two positions along a channel. The integrated equation is given in Dietrich and Dunne (1978). Integration of equation 15 will yield a slightly more complicated form of equation 2 in Dietrich and Dunne (1978).

These results also suggest a way of overcoming some of the problems associated with marked particle studies in stream channels. If transit time between two positions along the channel is computed for several years for a group of marked particles, part of the transit-time function can be defined. Using calculated average transit time from the integral forms of equations 14 and 15, a function might be fitted to the data to extend it to the longer period needed for complete determination. The transit-time function given by equation 6 probably has an exponential form for most rivers and, therefore,

$$\Phi(\tau) = \frac{dF(\tau)}{F_o d\tau} = me^{-n\tau} \qquad (16)$$

where m and n are constants that can be determined experimentally from successive years of measurement of the proportion of total marked particles passing the downstream position. Substitution into equation 7b and integration yields:

$$T_r = \frac{m}{n^2} \qquad (17)$$

Equation 17 can then be used to improve estimates made from short-term, marked-particle studies.

The transit time between two positions along a small channel in a narrow valley can be defined using marked particles in the following manner. At two-or more sections along the channel, oblique troughs across the bed of the stream could be installed that trap all bedload and generate a vortex current that transports the sediment across the channel to a pit (vortex bed sampler, Milhous 1973). All particles in the upstream pit would be painted, or otherwise marked, and replaced in the channel downstream. Anew color code could be used for each flood or season. Sediment returned to the channel would subsequently be captured in the next downstream vortex sampler, yielding the transit time between sections. All marked particles need not be followed because only the marked particles that pass into the lower trough must be counted. The form of the transit-time function would depend on distance between sampling troughs, and establishing more than two sections would permit examination of this dependency. Although such a monitoring scheme would probably only work on small streams, the general form of the transit-time function might be generalized to larger streams.

Because transit-time functions for sediment transport in river valleys are probably not normally distributed, the average transit time (residence time) may be a very poor indication of time spent in a reservoir for the bulk of sediment that moves through a reservoir. For example, sediment passing downstream along a valley floor may have a long residence time because of a small exchange rate with a large flood plain deposit, although most of the sediment leaving the valley-floor section may have traveled quickly through the reservoir along the surface of the channel bed. To quantify the lag times between input of sediment and discharge from a reservoir, and to examine such problems as quantifying the period over which sediment experiences different chemical weathering environments, one must attempt to define the transit-time distribution.

## CONCLUSION

A sediment budget for a drainage basin provides a quantitative accounting of the rates of production, transport, storage, and discharge of detritus. Its construction requires: recognition and quantification of transport processes, recognition and quantification of storage elements, and identification of linkages. among transport processes and storage sites. To accomplish this task, it is necessary to know the detailed dynamics of transport processes and storage sites, including such problems as defining the recurrence interval of each transport process at a place.

Qualitative and semiquantitative fulfillment of these requirements help in designing preliminary field studies and determining the general form of a sediment budget for a particular basin. This approximate budget can then be used to design long-term studies. Much progress is still needed in making useful field measurements and developing physically based models before complete, quantitative sediment budgets can be constructed.

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## LITERATURE CITED

**Anderson, E. W., and N. H. Cox. 1978.** A comparison of different instruments for measuring soil creep. Catena 5:81-93.

**Barnes, R. S., P. B. Birch, D. E. Spyridakis, and W. E. Schell. 1979.** Changes in the sedimentation histories of lakes using Lead-210 as a tracer of sinking particulate matter. In Isotope hydrology, vol. 2, p. 875-898. Int. Atomic Energy Agency, Washington, D.C.

**Barr, D. J., and D. N. Swanston. 1970.** Measurement of soil creep in a shallow, slide-prone till soil. Am. J. Sci. 269:467-480.

**Bolin, B., and H. Rodhe. 1973.** A note on the concepts of age distribution and transit time in natural reservoirs. Tellus 25:58-62.

**Bull, W. B. 1964.** Geomorphology of segmented alluvial fans in western Fresno County, California. U.S. Geol. Surv. Prof. Pap. 352-E, p. 79-129.

**Carson, M. A., and M. J. Kirkby. 1972.** Hillslope form and process. 474 p. Cambridge Univ. Press, London.

**Costa, J. E. 1975.** Effects of agriculture on erosion and sedimentation in the Piedmont Province, Maryland. Geol. Soc. Am. Bull. 86:1281-1286.

**Denny, C. S., and J. C. Goodlett. 1956.** Microrelief resulting from fallen trees. _In Surficial geology and geomorphology of Potter County, Pennsylvania. U.S. Geol. Surv. Prof. Pap. 288, p. 59-68.

**Dietrich, W. E., and T. Dunne. 1978.** Sediment budget for a small catchment in mountainous terrain. Z.. Geomorphol. Suppl. 29:191-206.

**Dunne, T. 1978.** Field studies of hillslope flow processes. In Hillslope hydrology, p. 227-293. M. J. Kirkby, ed. Wiley, London.

**Emmett, W. W., and H. R. Seitz. 1973.** Suspended- and bedload-sediment transport in the Snake and Clearwater Rivers in the vicinity of Lewiston, Idaho, March 1972 through June 1973. U.S. Geol. Surv. Basic-Data Rep., 78 p.

**Emmett, W. W., and H. R. Seitz. 1974.** Suspended- and bedload-sediment transport in the Snake and Clearwater Rivers in the vicinity of Lewiston, Idaho, July 1973 through July 1974. U.S. Geol. Surv. Basic-Data Rep., 76 p.

**Eriksson, E. 1971.** Compartment models and reservoir theory. Annu. Rev. Ecol. and System. 2:67-84.

**Everitt, B. L. 1968.** Use of the cottonwood in an investigation of recent history of a flood plain. Am. J. Sci. 266:417-439.

**Eyles, R. J., and R. Ho. 1970.** Rates of seasonal creep of silty clay soil. Q. J. Eng. Geol. 8:1-29.

**Fleming, R. W., and A. M. Johnson. 1975.** Rates of seasonal creep of silty clay soil. Q. J. Eng. Geol. 8(1):1-29.

**Fredriksen, R. L. 1970.** Erosion and sedimentation following road construction and timber harvest on unstable soils in three small western Oregon watersheds. USDA For. Serv. Res. Pap. PNW-104, 15 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oregon.

**Garwood, N. 'C., D. P. Janos, and N. Brokaw. 1979.** Earthquake-caused landslides: a major disturbance to tropical forests. Science 205:997-999.

**Gil, E., and A. Kotarba. 1977.** Model of slide slope evolution in flysch mountains (an example drawn from the Polish Carpathians). Catena 4:233-248.

**Gresswell, S., D. Heller, and D. N. Swanston. 1979.** Mass movement response to forest management in the central Oregon Coast Ranges. USDA For. Serv. Resour. Bull. PNW-84, 26 p., illus. Pac. Northwest For. and Range Exp: Stn., Portland, Oreg.

**Imeson, A. C. 1976.** Some effects of burrowing animals on slope processes in the Luxembourg Ardennes. Geogr. Ann. 58A:115-125.

**Imeson, A. C. 1977.** Splash erosion, animal activity and sediment supply in a small forested Luxembourg catchment. Earth Surf. Processes 2:153-160.

**Janda, R. J., K. M. Nolan, D. R. Harden, and S. M. Colman. 1975.** Watershed conditions in the drainage basin of Redwood Creek, Humboldt County, California, as of 1973. U.S. Geol. Surv. Open-File Rep. 75-568, 266 p.

**Kelsey, H. M. 1977.** Landsliding, channel changes, sediment yield, and landuse in the Van Duzen River Basin, North Coastal California, 1941-1975. Ph. D. thesis. Univ. Calif., Santa Cruz. 370 p.

**Kelsey, H. M. 1978.** Earthflows in Franciscan melange, Van Duzen River Basin, California. Geology 6:361-364.

**Kirkby, M. J. 1967.** Measurement and theory.of soil creep. J. Geol. 75:359-378.

**Kojan, E. 1968.** Mechanics and rates of natural soil creep. Proc. 1st Sess. Int. Assoc. Eng. Geol., Prague. p. 122-154.

**Kotarba, A. 1970.** The morphogenic role of foehn wind in the Mts. Stud. Geomorphol. Carpatho-Balancia 4:171-186

**Laronne, J. B., and M. A. Carson. 1976** Interrelationships between bed morphology and bed-material transport for a small, gravel-bed channel. Sedimentology 23:67-85.

**Larson, K. R., and R. C. Sidle. 1980.** Erosion and sedimentation data catalog of the Pacific Northwest. USDA For. Serv., Pac. Northwest Reg. R6-WM-050-1981.

**Leopold, L. B., W. W. Emmett, and R. M. Myrick. 1966.** Channel hillslope processes in a semi-arid area, New Mexico. U.S. Geol. Surv. Prof. Pap. 352-G, p. 193-253.

**Lewis, L. A. 1976.** Soil movement in the tropics--a general model. Z. Geomorphol. Suppl. 25:132-144.

**Lewis, S., and A. Nir. 1978.** On tracer theory in geophysical,systems in the steady and non-steady state. Part II. Non-steady state--theoretical introduction. Tellus 30:260-271.

**Lundgren, L. 1978.** Studies of soil and vegetation development on fresh landslide scars in the Mgeta Valley, Western Uluguru Mts., Tanzania. Geogr. Ann. 60A:91-127.

**Middleton, G. V, ‘ 1976.** Hydraulic interpretation of sand size distributions. J. Geol. 84:405-426.

**Milhous, R. T. 1973.** Sediment transport in a gravel-bottomed stream. Ph. D. thesis. Oreg. State Univ., Corvallis. 232 p.

**Mitchell, J. K. 1976.** Fundamentals of soil behavior. 422 p. Wiley and Sons, New York.

**Nir, A., and S. Lewis. 1975.** On tracer theory in geophysical systems in the•steady and non-steady state. Part I. Tellus 27:372-382.

**O’Loughlin, C. L., and A. J. Pearce. 1976.** Influence of Cenozoic geology on mass movements and sediment yield response to forest removal, North Westland New Zealand. Bull. Int. Assoc. Eng. Geol. 14:41-46.

**Rapp, A. 1960.** Recent developments of mountain slopes in Karkevagge and surroundings, northern Scandinavia. Geogr. Ann. 42:71-200.

**Reid, L. R. 1981.** Sediment production from gravel-surfaced roads, Clearwater basin. Wash. Fish. Re s. Inst., Univ. Wash., Seattle. Final Rep. FR1-UW-8108, 247 p.

**Ritchie, J. C., Hawks, P., and R. J. McHenry. 1975.** Deposition rates in valleys determined using fallout Cesium-137. Geol. Soc. Am. Bull. 86:1128-1130.

**Simonett, D. S. 1967.** Landslide distribution and earthquakes in the Bewani and Torricelli Mountains, New Guinea. _In Landform studies from Australia and New Guinea, p. 64-84. J. N. Jennings and J. A. Mabbutt, eds. Aust. Nat. Univ. Press, Canberra.

**Swanson, F. J., R. L. Fredriksen, and F. M. McCorison. 1982.** Material transfer in a western Oregon forested watershed. In Analysis of coniferous forest ecosystems in the western United States. R. L. Edmonds, ed. Hutchinson Ross Pub. Co., Stroudsburg, Pa. p. 233-266.

**Swanson, F. J., and D. N. Swanston. 1977.** Complex mass movement terrain in the western Cascade Range, Oregon. Geol. Soc. Am. Rev. Eng. Geol. 3:113-124.

**Swanston, D. N. 1969.** Mass wasting in coastal Alaska. USDA For. Serv. Res. Pap. PNW-83, 15 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

**Swanston, D. N. 1970.** Mechanics of debris avalanching in shallow till soils of southeast Alaska. U.S. For. Serv. Res. Pap. PNW-103, 17 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

**Swanston, D. N., and F. J. Swanson. 1976.** Timber harvesting, mass erosion, and steepland forest geomorphology in the Pacific Northwest. In Geomorphology and engineering, p. 199-221. Coates, ed. Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pa.

**Tanaka, M. 1976.** Rate of erosion in the Tanzawa Mountains, central Japan. Geogr. Ann. 58A:155-163.

**Thorn, C. E. 1978.** A preliminary assessment of the geomorphic role of pocket gophers in the Alpine Zone of the Colorado Front Range. Geogr. Ann. 60A:181-187.

**Williams, M. A. J. 1973.** The efficacy of creep and slopewash in tropical and temperate Australia. Aust. Geogr. Stud. 11:62-78.

**Wilson, S. D. 1970.** Observational data on ground movements related to slope instability. J. Soil Mech. and Found. Eng., Am. Soc. Civ. Eng. 96:1521-1544.

**Wolman, M. G., and R. Gerson. 1978.** Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surf. Processes 3:189-208.

**Yair, A. 1976.** Sources of runoff and sediment supplied by the slopes of a first order drainage basin in an arid environment (Northern Negev). _In Jerusalem Elat Symposium 1974, reviews and summaries of Israeli research projects. R. Gerson and M. Inbar, compilers. Z. Geomorphol. Suppl. 20:5-6.

**Young, A. 1971.** Slopes. 288 p. Longman Group Ltd., New York.

**Young, A. 1978.** A twelve-year record of soil movement on a slope. Z. Geomorphol. Suppl. 29:104-110.