ABSTRACT

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Sediment budgets quantify the transport and storage of soil and sediment in drainage basins or smaller landscape units. Studies of sediment routing deal with the overall movement of soil and sediment through a series of landscape units. The 14 papers and 5 summaries from discussion groups in this volume report results of sediment budget and routing studies conducted principally in forested drainage basins. Papers also deal with sediment routing studies using computer models, physical models, and field observations in nonforest environments.

This work emphasizes methods for judging the relative importance of sediment sources within a basin, the many roles of biological factors in sediment transport and storage, and the importance of recognizing changes of sediment storage within basins when interpreting sediment yield. Sediment budget and routing studies are important tools for both research scientists and land managers.

Keywords: Sedimentation, watershed management, sediment budget, drainage basins, geomorphology.

CONTENTS

Introduction: Workshop on Sediment Budgets and Routing in Forested Drainage Basins

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Geomorphologists have traditionally studied erosion of landscapes by analysis of individual erosion processes, measurement of sediment yield at one or more points along a river system, and stratigraphic analysis of deposits. These approaches, however, have rarely been used collectively to provide a framework for under standing soil and sediment movement through drainage basins. Many current problems in basic and applied geomorphology and ecology can be better addressed when placed in the broad, conceptual framework provided by analysis of sediment budgets and routing. Such analysis provides an understanding of geomorphic systems, just as similar studies have been fruitful in analysis of nutrient cycling, hydrology, and other complex systems.

Movement of soil and sediment through drainage basins commences with sediment production from bedrock weathering augmented by dust input from the atmosphere and biological production of organic matter. The organic and inorganic components of soil and sediment move down hillslopes by a great variety of processes, some operating more or less continuously but others only seasonally or episodically. Movement may occur by a sequence of linked processes or by several independent processes acting simultaneously. Soil eventually enters stream channels where it is subject to transport by fluvial processes. Between periods of movement, soil and sediment are temporarily stored at a variety of sites. They may undergo chemical and physical change while in storage and during transport. Both movement and storage are strongly influenced by live and dead vegetation.

A sediment budget is a quantitative description of this movement of sediment through a single landscape unit. Here, sediment refers to erodible materials at or near the earth's surface, including both organic and inorganic matter on hillslopes and in channels. A complete budget considers input rate, storage volume, type of modification, and discharge rate of sediment. The landscape unit may be an entire drainage basin, a hillslope segment, a first-order stream channel, or other feature. For example, the sediment budget of a first-order stream channel may include measurement of material transport across the channel margins by hillslope processes, changes in storage, and total sediment discharge from the channel.

The term "sediment routing" is used both generally and specifically. Many investigators, including most of those in this volume, use the term in a nonquantitative sense to convey the concept of sediment movement through a drainage basin. Parallel to use of the term "flood routing," sediment routing, in a stricter sense, is the computation of the movement of sediment through a series of landscape units and of the modification of the sediment during its passage. The computation is based on the continuity equation for sediment transport through successive landscape units along the flow path. Output from one landscape unit becomes input to the next unit downslope or downstream. Thus, if the conveyance of sediment through a first-order channel is of interest, the channel might be divided into increments, and input, output, and change in storage of each size fraction in each increment would be observed. Input would include transport from upstream as well as transport across the channel margins. How well a budget and routing procedure characterizes and quantifies a geomorphic system depends on how well transfers, storages, and linkages in the system are understood; the precision and duration of field measurements; and the scale of spatial and temporal segmentation of the drainage basin under analysis.

Sediment transport and storage constrain development, structure, and composition of terrestrial and aquatic ecosystems. Conversely, vegetation and, to some extent, fauna influence sediment routing in most landscapes by affecting sediment transport and storage and by directly contributing an important organic component to the total sediment load.

Several features distinguish sediment routing in forest environments from that in other vegetation types. Forest vegetation plays a prominent role in many phases of sediment transport and storage. Transport of soil by windthrow of trees, the binding of soil by tree roots, and other vegetative factors affect soil movement in forests. Standing trees and fallen logs on hillslopes and in channels trap and store sediment temporarily. Sediment-routing systems are influenced by forest-managememt activities, including road construction and removal of vegetation. Short-term studies of sediment routing in forested drainage basins should be put in the context of vegetative succession on this time scale; changes in forest plant communities that directly control sediment routing typically occur over a period of several decades to a century or more after severe disturbances. On a longer time scale, evaluation of management impacts on sedimentation should consider the frequency and erosional consequences of disturbances of soil and vegetation by human activities, as well as natural forest disturbances such as wildfire and windthrow.

Sediment-routing studies and budgets have been used for a variety of purposes. Early studies centered on the basic problem of how sediment moves through landscapes. Even within these basic studies, different areas have been emphasized. Dietrich and Dunne (1978), for example, were particularly interested in changes in properties of soil and sediment moving through drainage basins, and they analyzed sediment transport from the long-term perspective of sedimentary petrology and steady-state geomorphology. Rapp (1960) and Caine (1976), on the other hand, were more concerned with the relative importance of high magnitude, episodic events and low magnitude, more frequent or continuous processes. Leopold et al. (1966) measured rates of slope and channel processes in a semiarid area to aid analysis of arroyo filling and cutting. Swanson et al. (1982) developed a sediment budget to analyze the relative importance of physical processes in the transport of organic and inorganic matter through a steep, forested drainage basin and compared this budget with one for a less steep drainage basin forested with a younger stand.

Sediment-budget studies have also been useful in analyzing effects of management practices on soil erosion, sediment yield, and water quality. Kelsey (1977) and Lehre (1980) observed that land use altered the relative importance of some storage sites and transfer processes in sediment-routing systems in northern California. Fredriksen (1970), Janda (1978), and others have analyzed partial sediment budgets with the specific purpose of interpreting sediment contributions from human-influenced sources in forest environments. Consideration of sediment-routing concepts can also help managers design effective networks for monitoring management impacts on soil, stream channels, and water quality.

The diversity of objectives, approaches, and professional disciplines of those who conduct sediment-routing and budget studies in forest environments has resulted in poor communication among scientists and land managers with this common interest. To encourage communication and a useful commonality of approaches, we convened a workshop on Sediment Budgets and Routing in Forest Drainage Basins on May 30 through June 2, 1979, at the Forestry Sciences Laboratory, Corvallis, Oregon. The workshop was cosponsored by the National Science Foundation and the Pacific Northwest Forest and Range Experiment Station, USDA Forest Service. The workshop format provided for roughly equal division between formal talks and small group discussions. Participants were encouraged to emphasize theory and to speculate on new approaches in study design and problem analysis.

This volume contains papers presented at the workshop and summaries of some discussion groups. Some papers do not deal explicitly with forest environments but are included because they offer principles or techniques useful to sediment-routing and budget studies. The papers are grouped into those dealing with general issues in sediment-routing and budgets and those dealing with case studies of sediment movement and storage on hillslopes and in channels. First, Dietrich, Dunn, Humphrey, and Reid suggest basic rules for developing a sediment budget with examples from forested areas of the Pacific Northwest. Simons, Li, Ward, and Shiao follow with a paper discussing computer-simulation modeling, as both a tool for understanding general characteristics of sediment-routing systems and for predicting impacts of some management activities. Using examples from steep, chaparral-covered basins, Rice examines difficulties in modeling sediment routing in landscapes that are greatly modified by randomly occurring, extreme, and infrequent events. Vegetation-free sediment-routing systems of badland environments and laboratory models provide opportunity to examine some general, systematic patterns of sediment routing as described by Harvey and Bergstrom.

Two case studies deal primarily with slope processes. Lehre presents a sediment budget that emphasizes slope processes in a small drainage basin in the central California coast. Hupp and Sigafoos describe use of dendrochronologic methods to assess slope processes on block fields in the Appalachian Mountains.

Papers by Kelsey, Madej, Taylor, and Smith and Hicks address sediment transport along stream channels and interpret transport characteristics in terms of sediment supply by hillslope processes.

Two papers analyze the characteristics of large woody debris in streams as they affect sediment storage. Megahan's studies of channel storage in small streams in the Idaho Batholith concern the effects of forest-management practices on channel stability and sediment yield. Likens and Bilby use examples from the White Mountains in central New Hampshire to describe the evolution of organic-debris dams in relation to forest stand development.

In the final paper, Swanson and Fredriksen discuss application of sediment-routing and budget concepts to analysis of erosional consequences of forest management.

Five brief reports of discussion sections follow the formal papers. Groups of 7 to 16 participants met in afternoon sessions for informal discussions of identifying and mapping sources and storage areas of sediment; tracing and dating the movement and storage of sediment; the role of weathering and soil formation in sediment routing; the influence of magnitude, frequency, and persistence of various types of disturbance on geomorphic form and process; and the use of flowcharts in sediment-routing studies. These discussions provided scientists of diverse backgrounds opportunity to exchange ideas and concerns, to identify unresolved problems, and to focus on future research needs. These discussions did not result in consensus or state-of-the-art statements for each topic because of time limitations, the broad scope of each topic, and our limited knowledge and techniques for addressing the problems. Lack of common definitions of processes and storages and common measurement techniques frustrated some attempts to compare geomorphic systems in different physiographic and ecologic settings.

Readers of this volume may become aware of several recurring themes: interactions between vegetation and sedimentation systems are numerous and complex; forest vegetation makes these environments highly retentive of sediment; and sediment-routing studies provide a useful approach to holistic analysis of drainage-basin function. These issues, along with identified research needs, are discussed more fully in a concluding statement. Analysis of sediment budgets and routing offers great promise in study of drainage-basin evolution and impacts of management practices on sedimentation and on forest and stream ecosystems.

LITERATURE CITED

Caine, N.
1976. A uniform measure of subaerial erosion. Geol. Soc. Am. Bull. 87:137-140.

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

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, Oreg.

Janda, R. J.
1978. Summary of wastershed conditions in the vicinity of Redwood National Park, Calif. U.S. Geol. Surv. Open-File Rep. 78-25, 82 p.

Kelsey, Harvey M.
1977. Landsliding, channel changes, sediment yield and land use in the Van Duzen River basin, north coastal California, 1941-1975. Ph. D. thesis. Univ. Calif., Santa Cruz. 370 p.

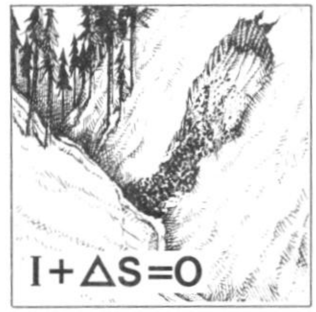
Lehre, Andre.
1981. Sediment mobilization and production from a small Coast Range catchment: Lone Tree Creek, Marin Co., California. Ph. D. thesis. Univ. Calif., Berkeley.

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

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

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.

Papers



Construction of Sediment Budgets for Drainage Basins

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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.

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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 deep-seated 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 and 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 (Yair 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

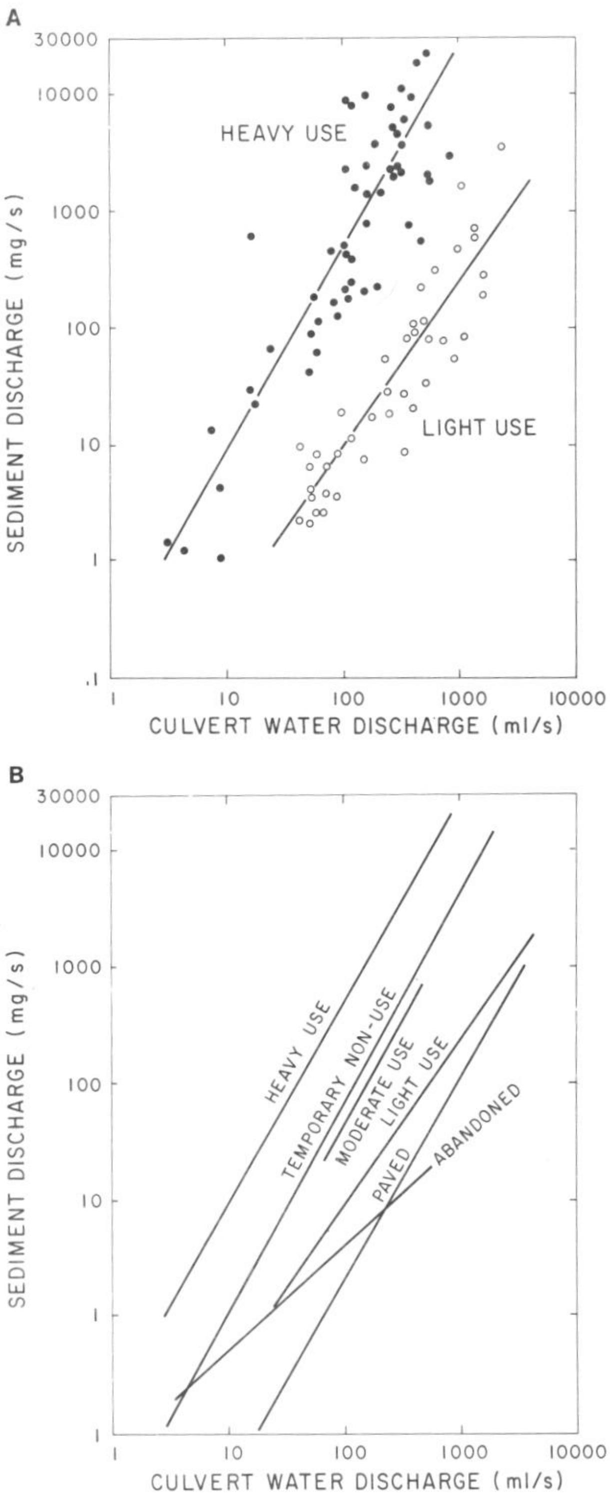


Figure 1.--Effects of road-surface conditions on sediment-discharge rating curves of drainage culverts for gravel-surfaced logging roads in the Clearwater drainage basin of coastal Washington. A. Comparison between "heavy use" (16 to 32 logging trucks/day) and "light use" (no logging trucks in 3 to 60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

western Washington, Reid (1981) developed a technique 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 culvert during 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.

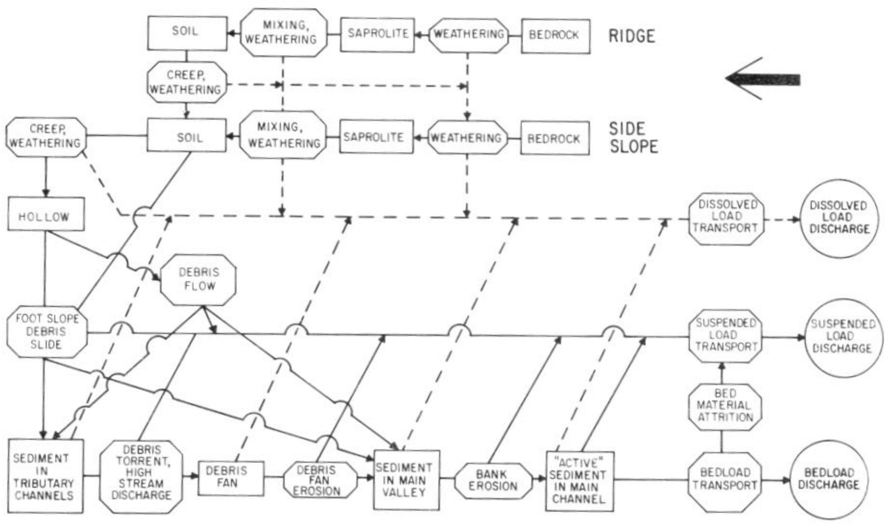


Figure 2.--Sediment-budget model for the Rock Creek basin, central coastal Oregon: rectangles, storage sites; octagonals, transfer processes; circles, outputs; solid lines, transfer of sediment; and dotted lines, migration of solutes (from Dietrich and Dunne 1978).

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

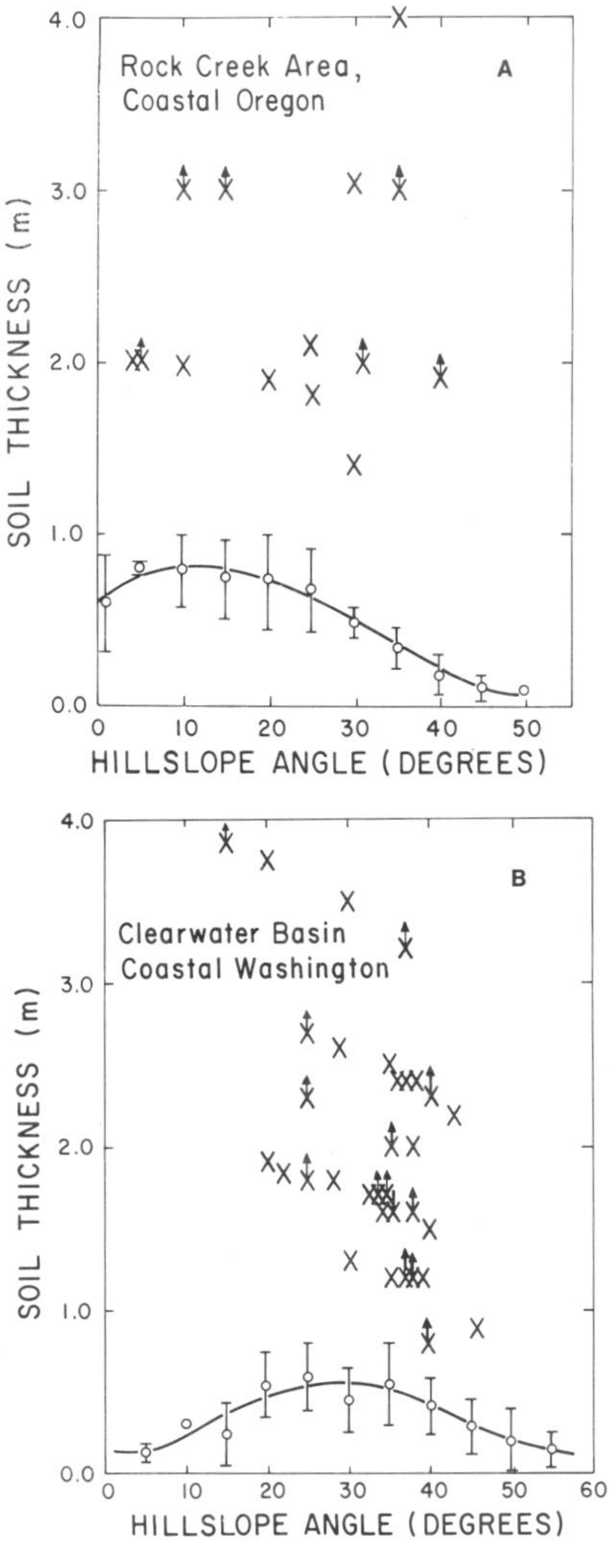


Figure 3.--A. Relationship of hillslope gradient to soil depth for soils in the Rock Creek area of coastal Oregon. B. Clearwater basin of the Olympic Peninsula of Washington. Measurements were made along road cuts. Circles represent the average of depths for 5-degree classes, and the bars are the standard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

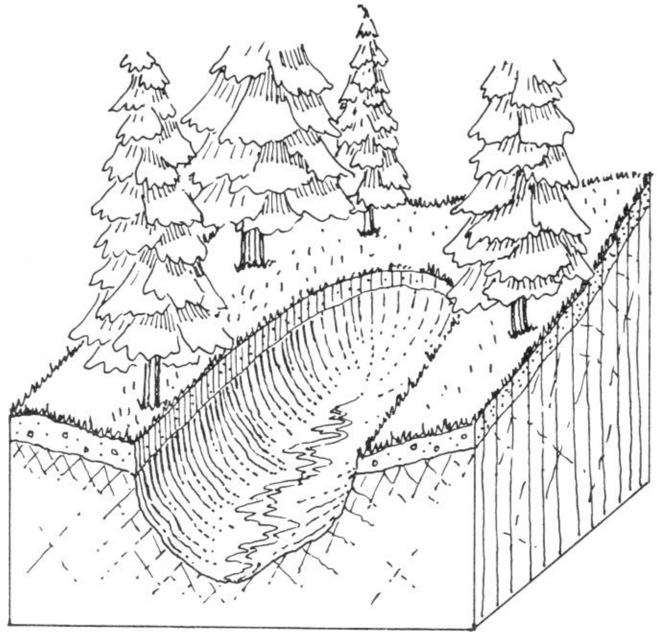
Figure 4.--Soil wedge partially filling depression in sedimentary bedrock, western Olympic Peninsula. Note the characteristic "U"-shaped depression and the great soil depth compared to adjacent areas.

average soil-creep rate. This relationship 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. 6a), we estimate that this period



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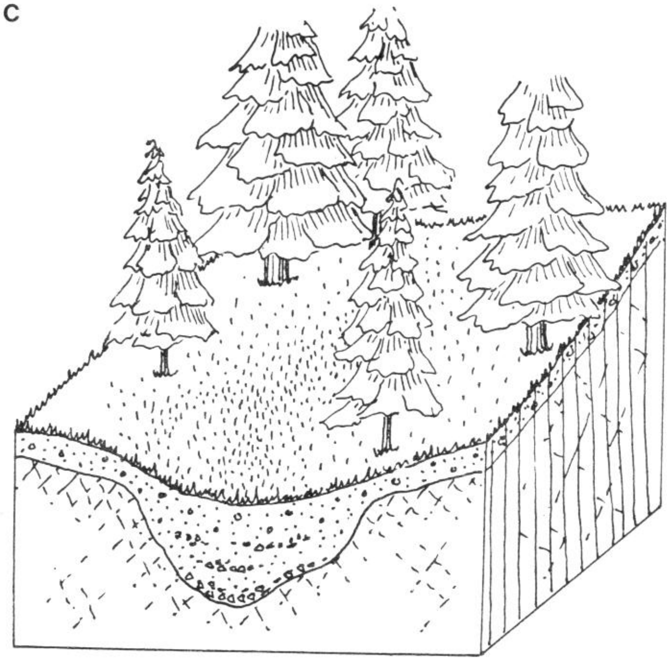

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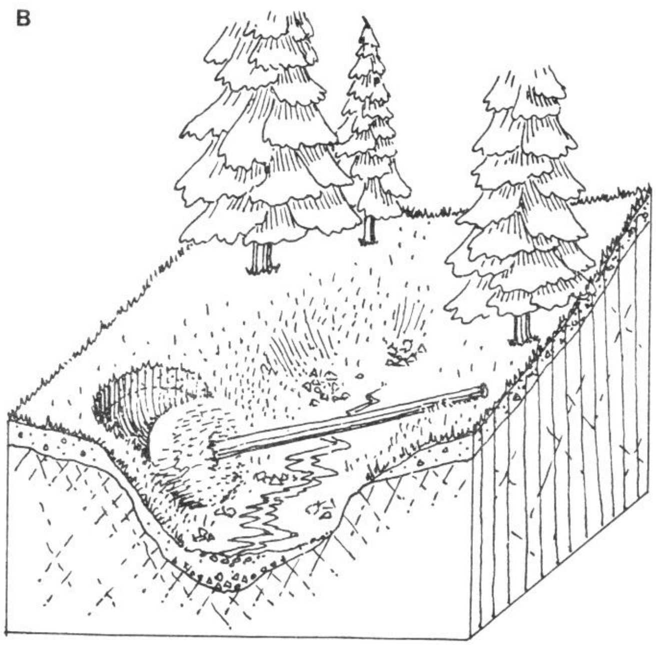


Figure 5.--Evolution from a landslide scar in a bedrock depression to a soil wedge. A. After the landslide, the exposed bedrock surface forms an impermeable horizon shedding rainwater and subsurface discharge into the depression as overland flow. B. Sediment eroded from the over-steepened soil perimeter 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 the surrounding soil from which it is derived.

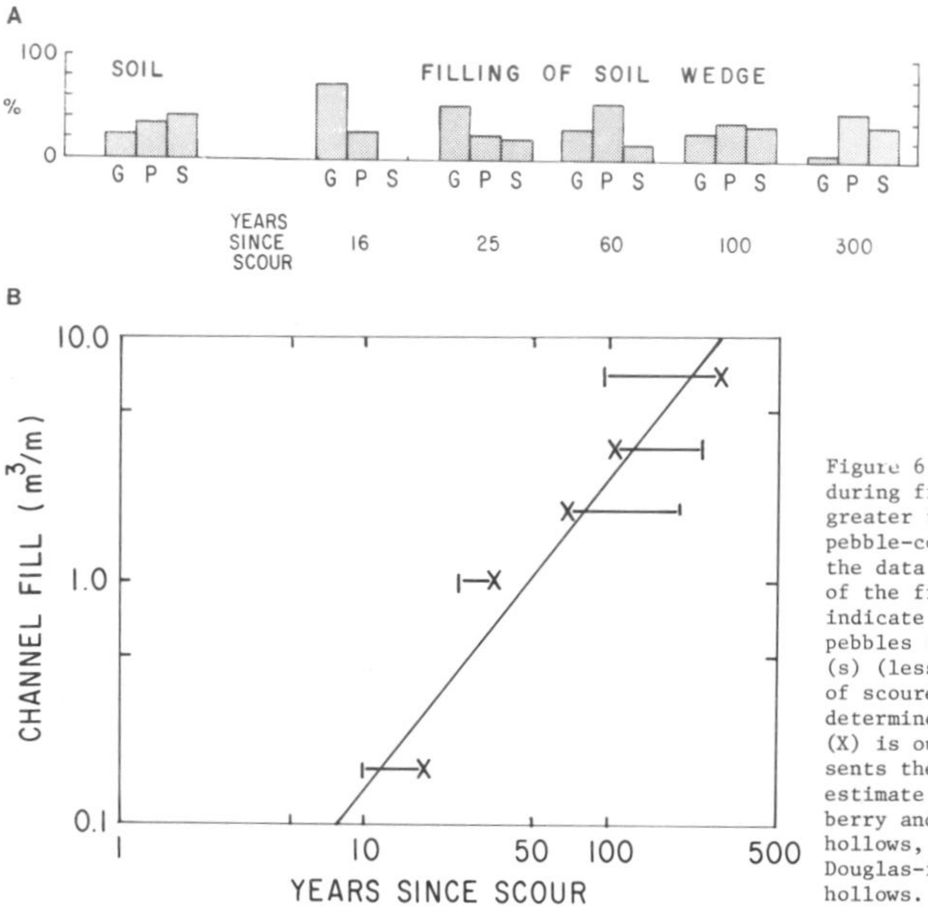


Figure 6.--A. Successive fining of material during filling of hollows. The fraction greater than 22 mm was measured by the pebble-count method of Wolman (1954), and the data were combined with sieve analyses of the finer particles. The letters indicate gravel (g) (greater than 22 mm), pebbles (p) (from 2 to 22 mm), and sand (s) (less than 2 mm). B. Rate of filling of scoured hollows. Time of scour was determined from ages of vegetation 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 scoured hollows, and large alder, hemlock, and Douglas-fir on older, partially filled hollows.

will last about 100 years. By the end of this period, 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.

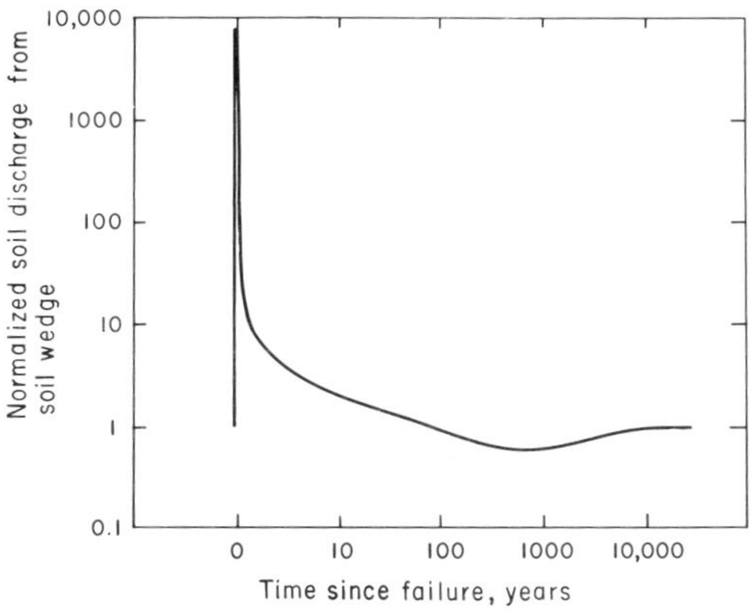


Figure 7.--Chronologic variation of sediment discharge into a stream channel as a result of failure of a soil wedge and eventual refilling of the bedrock depression. The annual sediment discharge for each year was divided by the average annual sediment discharge to normalize the curve.

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 will 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 (τ) 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 (τ) such that

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

where M_o 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

$$\psi(\tau) = \frac{1}{M_o} \frac{dM(\tau)}{d\tau} \quad . \tag{2}$$

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):

$$T_a = \frac{1}{M_o} \int_o^\infty \tau dM(\tau) \tag{3}$$

$$= \int_0^\infty \tau\psi(\tau)d\tau \quad . \tag{4}$$

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 τ becomes very large, $F(\tau)$ approaches the total flux per unit time out of the reservoir:

$$\lim_{\tau \to \infty} F(\tau) = F_o \quad . \tag{5}$$

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

$$\Phi(\tau) = \frac{1}{F_o} \frac{dF(\tau)}{d\tau} \quad . \tag{6}$$

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:

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

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

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

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

The left side of equation 8a gives the amount of sediment greater than age τ that leaves the resèrvoir per unit time. To maintain a steady-state age distribution, the amount of sediment with an age greater than τ 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 τ of each side of the equation yields the transit-time density function,

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

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.

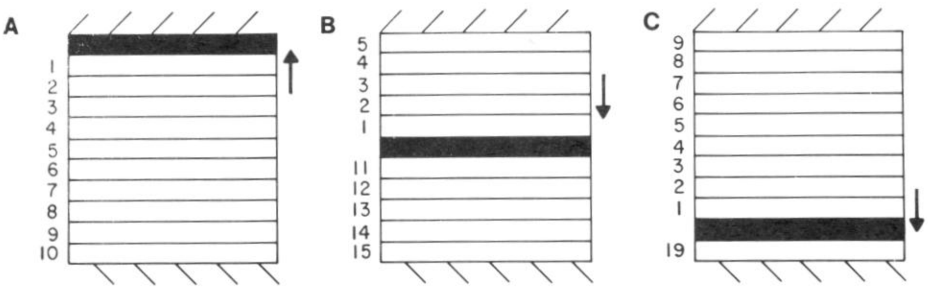


Figure 8.--A-C. Change in age distribution of flood-plain sediment reservoir resulting from uniform lateral migration of the active channel (darkened area) at one channel-width per year. Numbers in columns give age in years for each increment of flood-plain deposit.

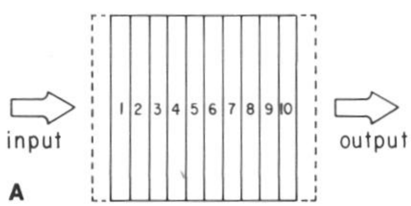
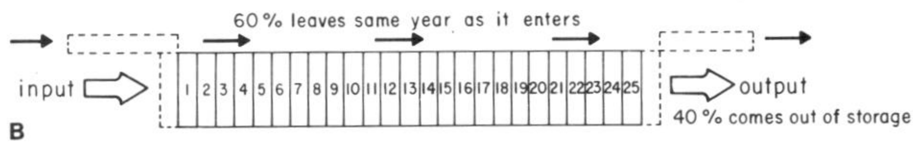


Figure 9.--A-B. Two models of transfer of sediment through a gravel bar. Numbers are the age of the sediment in particular parts of each reservoir. Total sediment flux rate is the same for both models but, in the second, 60 percent sediment entering the reservoir leaves immediately, and the remaining 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_o) divided by the total flux rate (F_o) equals the residence time of sediment in a reservoir in the steady-state condition:

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

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, $M(\tau)$, 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., $F(\tau) = 0$ when $\tau = 0$), equation 8a can be used to obtain F_o, 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, $M(\tau)$, which was integrated to obtain a total volume of sediment, M_o, of 1.08×10^7 m^3. Differentiating $M(\tau)$ with respect to τ yields the right side of equation 8a and, therefore, at $\tau = 0$, a flux rate, F_o, of 1.08×10^5 m^3/year. Substituting this value and M_o into equation 9 leads to a residence time of 100 years.

The transit-time function for Everitt's data (obtained by twice differentiating $M(\tau)$, 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

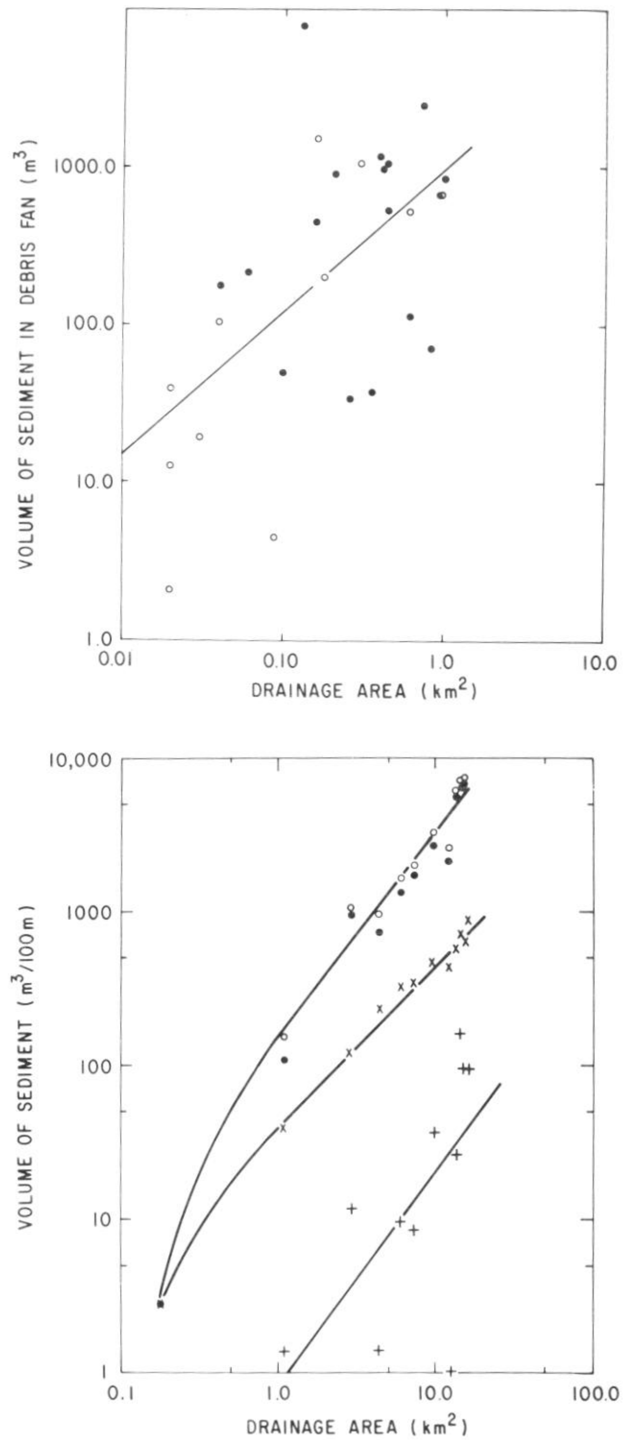


Figure 10.--A. Volume of debris fans as a function of tributary drainage area along the main channel of the 16.2-km^2 Rock Creek drainage basin, Oregon. Open circles are debris fans covered by salmonberry and young alder trees. Dots are fans covered by more mature trees. B. Volume of sediment in gravel 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 10A and B are given by Dietrich and Dunne (1978).

amount of sediment in storage can be related to distance along the 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_1 and Q_o, 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

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

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

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

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_o 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 $Q_o = Q_i$, such that

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

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

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

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 steady-state 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

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

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

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 $j \cong 0$. 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. A new 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. Res. Inst., Univ. Wash., Seattle. Final Rep. FRl-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.

Modeling of Water and Sediment Yields From Forested Drainage Basins

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ABSTRACT

Wise management of land and water resources in forested drainage basins requires knowledge of the interrelations within the forest environment and the numerous controlling physical processes. One aspect of drainage-basin response to management activities is changes in water and sediment yields. An important approach for predetermining these yields is through use of water and sediment routing and yield models. Models based on the physical processes controlling erosion provide a realistic simulation of natural and management influences on drainage-basin response. We discuss general concepts of mathematical modeling with particular emphasis (including examples from forested drainage basins) on surface erosion physical-process simulation models developed at Colorado State University.

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INTRODUCTION

Need for Modeling

Increases in development of land and water resources have promoted beneficial and adverse activities in forest drainage basins. One aspect of drainage-basin response is changes in water and sediment yields brought about by different management activities. Drainage-basin sediment yield reflects the change in land-surface erosion, which contributes much of the fine sediments carried in larger rivers. Recent studies have shown that, in addition to other impacts, fine sediment is extremely important for transporting adsorbed substances such as pesticides and other organic compounds.

Accurate methods for estimating changes in water and sediment yields from forested lands are needed to help evaluate related adverse impacts on aquatic and terrestrial ecosystems. One approach is through use of mathematical models that simulate the movement of water and sediment from forested drainage basins. With such models, a forest planner can estimate changes in water and sediment yields brought about by different management activities. Insights gained from model simulations can help the forest planner in formulating development strategies that meet short- and long-term goals.

Mathematical models also aid the conceptualization and organization of drainage-basin system processes. Developing a model emphasizes the strong and weak links in our knowledge of forested drainage basins. Recognition of the weak links, such as subsurface flow, helps direct future research.

Although many computer models are available for predicting water yield, few models have been developed for sediment routing and yield. Water routing is essential for determining sediment movement and resultant aggradation, degradation, and water quality; therefore, water routing and yield are necessary for determining sediment routing and yield.

Scope of Paper

This paper pesents general concepts of mathematical modeling of water and sediment yields from drainage basins with emphasis on sediment yield from surface erosion of the sheet-rill type. The three other major types, however--landslide, gully, and channel erosion--are discussed in their context with the overall sediment yield of drainage basins. Physical-process models have been singled out and developed in preference to other modeling approaches because they are more representative of the system being simulated, they require fewer sets of data to develop, they are not specific to a particular site or period, and they more readily reflect land-use changes or other activities in the drainage basin. Selected physical-process models are presented with applications to forested drainage basins.

CONCEPTS OF MATHEMATICAL MODELS

Model Types

Several types of mathematical models may be used by forest hydrologists. These include "black box" or lumped parameter, regression, stochastic, and physical-process types. A physical-process approach is developed in this paper; however, certain criteria are required for all types of mathematical models.

Criteria of Useful Mathematical Models

Mathematical models useful for predicting response of drainage basins to natural or management changes should meet the following criteria: (1) The temporal (time) resolution should be adjustable to both short- and long-term response of the drainage basin; (2) the spatial (space) resolution should be flexible from small to large drainage basins; (3) the modeling approach should be transferrable between drainage basins; (4) the model should correctly represent the effects of management activities with selected input variables; (5) uncertainties caused by variability of climate and space should be considered in the model; (6) the model should be usable within the constraints of reasonably available data; (7) the model should be usable by management personnel and scientists; (8) the model should accommodate different levels of accuracy and resolution in the simulations, based on the accuracy and resolution of the input variables; (9) model development should adopt a modular approach that builds a coordinated nucleus of standardized system components for use in a wide spectrum of drainage-basin and river systems; and (10) the model should be well documented so that it can be easily understood and correctly applied. In addition to these attributes, model complexity and resultant data needs should be considered.

Model Complexity and Data Needs

Potential users of a model are concerned that it may be too complex to use, may not correctly reflect the controlling processes in a drainage basin, or may require data that are extremely difficult to obtain. The flexibility of physical-process models provides a means for resolving these common problems.

To gain precision, a model must simulate the physical processes in the finest possible scale. This often leads to a complex division of the drainage basin and rigorous mathematical treatment of the physical processes. Complex mathematical treatments are not easily understood by many users, who may reject it for a more comfortable but less sophisticated approach. This need not happen if the model is clearly stated and formulated so that the user can examine each physical process. In many applications, a complex and extremely detailed model is not necessary and simpler versions can be used. Model simplification, if the basic physical processes are left intact, need not reduce the applicability of the model.

If a model is organized as a group of modules or subroutines, it can easily be modified to account for variations in the controlling physical processes in a study area. For example, in the Beaver Creek Watershed near Flagstaff, Arizona, runoff processes are dominated by overland flow, and subsurface flow is negligible. Thus, only overland flow and channel flow modules are necessary for adequate modeling of runoff. In contrast, runoff from the H. J. Andrews Experimental Watershed is almost entirely a result of subsurface flow. Surface runoff modules are minor and may be removed, with subsurface-flow routines being substituted. An alternative to a modular-type approach would be an all-purpose, generalized model, but this type of model often requires inputs for routines that do not reflect the controlling processes. Such "overkill" in modeling can lead to user dissatisfaction, increased data preparation costs, and erroneous results. The modular approach also lends itself to efficient updating of a program. As applied research in drainage-basin hydrology and hydraulics advances, new or improved mathematical descriptions of the controlling processes appear. If, for example, a better subsurface flow routine is developed, then it can be substituted as needed into the appropriate models. A modular-type model allows the user to select the physical-process components that best describe the controlling phenomena or update those processes as needed.

Another criticism of many models concerns data requirements. Regression-type models, because of their simplicity in form, are often composed of few variables. This makes them attractive to potential users. The physical significance of each variable, however, or--more important--its relation to land-management activities is often questioned. Although the input to physical-process models may be more extensive than for other model types, each variable can be related to measurable characteristics of the drainage basin, thus avoiding many problems in interpreting the meaning of the different model variables. Recently, Simons et al. (1979a) have reported useful techniques for estimating input parameters from commonly gathered data. With such manuals, the field user can relate drainage-basin information and model parameters to provide the inputs needed to run models that would otherwise appear complex. Although inputs may need to be adjusted to reflect the judgment of the field user, model results using this approach will usually be adequate for planning. More detailed investigations will require a better defined and more accurate data set.

Physical-process models can and do avoid many of the concerns often expressed by potential users. If the model is linked in a modular format, it can be upgraded or adjusted to reflect the complexity or simplicity of the controlling physical processes. In addition, input variables to physical-process models can be more readily related to drainage-basin characteristics, and therefore management activities, than can variables for other types of models. All of these attributes make physical-process models more appealing for general use in drainage-basin studies.

SEDIMENT SOURCES IN DRAINAGE BASINS

Selection of an appropriate sediment-yield model is dependent upon the sources of sediment in the drainage basin. Components of the mathematical model must match the physical processes controlling the movement of water and sediment from the drainage basin. Four major sources of sediment in drainage basins are sheet and rill erosion, gullies, landslides, and channels (Simons et al. 1979b). Each of these natural sources can be modified by human activities. In addition, sediment sources are created by agricultural tillage, grazing of domestic animals, roadway construction and maintenance, timbering, mining, urbanization, and development of recreational land. Sheet-rill erosion and channel sources are probably best understood in terms of modeling. Gullying has been described by regression and physical-process models. Regression models usually relate gully growth to drainage area, soils, and precipitation (such as Thompson 1964). Physical-process approaches examine gully growth using processes of mass wasting and hydraulic transport (Bradford et al. 1973, Piest et al. 1975). Similarly, landslide sources have been described throughout the literature, but accurate, quantitative models are still lacking. Mathematical descriptions of landslide sediment sources have been made by Ward (1976) and Simons et al. (1978a). More study is needed to estimate delivery of landslide-produced sediment to stream channels. Sediment in a drainage basin may be from a single type of source or a combination of types. The contribution from each source is often hard to identify without detailed field surveys. Sheet-rill and channel-erosion models will be the primary physical-process models until better mathematical descriptions of gullying and landsliding can be developed.

FORMULATION OF PHYSICAL-PROCESS MATHEMATICAL MODELS

Segmenting the Drainage Basin

Drainage-basin characteristics must be subdivided and digitized both spatially and temporally before they can be mathematically analyzed. Drainage-basin segmentation is spatial digitization (Simons et al. 1978b). Segmentation allows a drainage basin to be represented by single or multiple planar surfaces for the overland part and straight-line segments for the channel reaches (fig. 1). Segmenting the drainage basin allows computation of slope and azimuth of each planar surface and channel segment. The azimuth is then used to define flow direction and sequence to be used in flow computations. If data on vegetation type, soil type, canopy cover, and ground cover are available, they can be encoded at grid nodes to depict the spatial variation of these important factors. Such information is necessary in computing variations in water and sediment yields. Segmentation of the drainage basin can be as detailed as deemed necessary by the user. If information is required every 100 m, then the grid should be of a corresponding size. Some small drainage basins, however, may be represented by two planes and one channel.

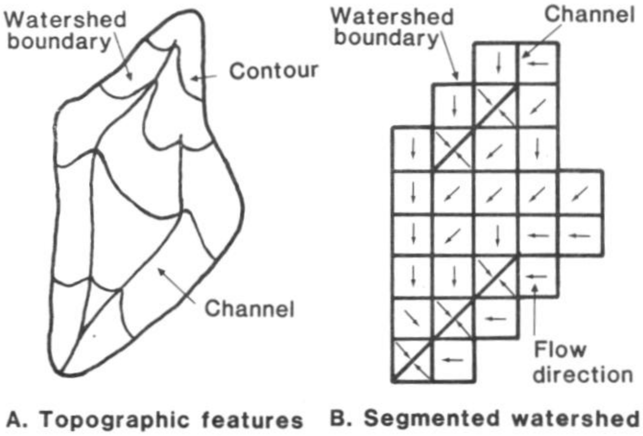


Figure 1.--Example of drainage-basin segmentation.

In addition to the above drainage-basin descriptors, channel cross-sectional geometry is usually represented by power relationships between parameters of wetted perimeter, flow area, top width, and depth.

Temporal subdivision takes place when precipitation and runoff are considered. Although rainfall and runoff are continuous processes, they must be divided into workable periods for analyses. This often depends on the size of drainage basin and the purpose of the model. If a long-term water balance is needed, periods of a day to a week may be adequate. If storm runoff from a small drainage basin is desired, rainfall and runoff periods from one to a few minutes may be required. Each use of the model and each drainage basin necessitate reevaluation of an appropriate time frame. Along these same lines, physical-process models can be used to estimate the response of drainage basins to rainfall and help determine a realistic sampling period for gaging installations and further model simulations.

Model Structure

Once drainage basin and other factors have been numerically defined by a segmentation procedure, the drainage basin can be modeled. Simons et al. (1975) developed a drainage-basin sediment model that is primarily applicable for surface-erosion simulation (fig. 2). It simulates surface flow and sediment production in small drainage basins. The basic framework of this model has been subsequently used to develop simplified and other models (e.g., Simons et al. 1977b, Simons et al. 1977d). The drainage basin is conceptually divided into the land-surface or overland-flow loop and a channel-system loop, each with different physical processes. The overland-flow loop simulates processes of interception, evaporation, infiltration, detachment of soil by impact of raindrops, erosion by overland flow, and overland water flow and sediment routing to the nearest channel. A channel-system loop determines water and sediment contributed by overland flow, which are routed in the channel, and the amount of channel erosion or sediment deposition throughout the channel system.

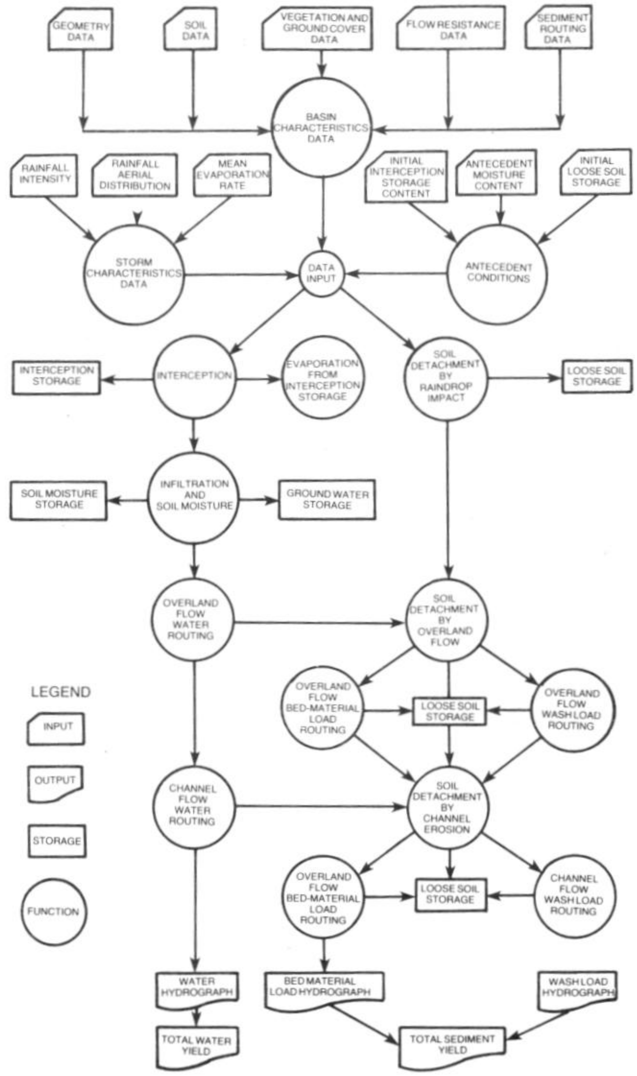


Figure 2.--Flow chart for the drainage-basin sediment and routing model.

Model Components

Rainfall excess is that amount of rainfall available for overland flow after all losses have been subtracted. Commonly considered losses are interception, depression storage, infiltration, and evapotranspiration. In storm-runoff simulation, losses from evapotranspiration are considered small and assumed negligible during the storm. Losses from depression storage can be combined into ground-cover interception. The two remaining losses are interception and infiltration.

The amount of interception loss depends on the percentage of the drainage-basin surface covered by canopy and ground cover, their respective water-holding capacities, and their initial moisture levels. The total intercepted volume can be written as

$$V_i = C_c V_c + C_g V_g \tag{1}$$

in which V_i is the total intercepted volume in depth, C_c is the canopy-cover density ($0 \leq C_c \leq 1$), V_c is the potential volume of canopy-cover interception in depth, C_g is the ground-cover density ($0 \leq C_g \leq 1$), and V_g is the potential volume of ground-cover interception in depth. The value of V_i is dependent on the type of vegetation, height of vegetation (related to leaf areas), and the initial moisture level. The value of V_g is a function of forest litter, grass, and the initial moisture level. Methods for estimating interception depths, as well as other variables, have been presented by Simons et al. (1979a).

As the model computations proceed through time, the rainfall is all intercepted until the potential volumes for ground and canopy cover are satisfied. This would constitute a series of net rainfall rates. Although interception losses are continuous over the storm period, the losses are assumed to occur during the beginning of the storm, and effective rainfall does not begin until the interception volume is satisfied.

The infiltration equation of Green and Ampt (1911) used in the drainage-basin model is a simple, physically based, two-parameter equation. It may be derived by direct application of Darcy's Law under the following assumptions: (1) a distinct piston wetting-front exists; (2) the hysteresis effects in the soil properties are negligible; and (3) the pressure effect of ponded water is negligible.

A Green-Ampt type equation may be written as

$$\frac{F}{\delta} - \ell n \left(1 + \frac{F}{\delta}\right) = \frac{Kt}{\delta} \tag{2}$$

in which F is the infiltrated volume, K is the hydraulic conductivity of the soil in the wetted zone, t is the time, and δ is the potential head parameter and defined as

$$\delta = (\theta_w - \theta_i)\,\psi_{ave} \tag{3}$$

in which θ_w is the moisture content of the soil after wetting, θ_i is the antecedent moisture content, and ψ_{ave} is the average suction head across the wetting front.

If at any time, t, the infiltrated volume is F(t), then at some later time $t + \Delta t$

$$F(t + \Delta t) = F(t) + \Delta F \tag{4}$$

in which ΔF is the change in infiltrated volume that occurred during the time increment Δt. An expression for ΔF is obtained from equation 4 as

$$\Delta F = F(t + \Delta t) - F(t) \quad . \tag{5}$$

Li et al. (1976a) developed the following method of solving for the infiltration rate. Their derivation yields

$$\frac{\Delta F}{\delta} - \ell n \left[\frac{\delta + F(t) + \Delta F}{\delta + F(t)}\right] = \frac{K}{\delta}\Delta t \quad . \tag{6}$$

Equation 6 is implicit with respect to ΔF. The equation is simplified, however, by expanding the logarithmic term in a power series (Li et al. 1976c)

$$\ell n\left(1 + \frac{\Delta F}{\delta + F}\right) = \ell n\ 1 + \frac{\dfrac{2\Delta F}{\delta + F}}{2 + \dfrac{\Delta F}{\delta + F}} + \ldots \tag{7}$$

Truncating equation 7 after the second term and substituting into equation 6 gives

$$\frac{\Delta F}{\delta} - 2\left[\frac{\dfrac{\Delta F}{\delta + F}}{2 + \dfrac{\Delta F}{\delta + F}}\right] = \frac{K}{\delta}\Delta t \quad . \tag{8}$$

Equation 8 is simplified into a quadratic with a solution of (Li et al. 1976a)

$$\Delta F = -\frac{(2F - \Delta K\ t)}{2} + \frac{[(2F - \Delta K\ t)^2 + 8K\Delta t(\delta + F)]^{1/2}}{2} \quad . \tag{9}$$

Only the positive root of equation 9 has any physical significance. The average infiltration rate, $\bar{f}$, is obtained by dividing ΔF by Δt, or

$$\bar{f} = \frac{\Delta F}{\Delta t} \quad . \tag{10}$$

After subtracting the interception and infiltration losses, the rainfall excess, i_e, can be determined. A rainfall event is commonly reported as an hyetograph, that is, a series of net rainfall intensities, I, each lasting for a time increment, Δt. Thus, if the net rainfall intensity is greater than the infiltration rate, the infiltration rate is subtracted from the net rainfall intensity to give the excess rainfall. If the net rainfall intensity is less than the corresponding infiltration rate, the infiltration rate equals the net rainfall intensity, and there is no excess.

Water Routing

Water runoff can be described by the equation of continuity, the equation of motion, and equations describing resistance to flow.

Continuity Equation.--The equation of continuity for water is

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} = q_\ell \tag{11}$$

in which Q is the discharge, x is the downslope distance, A is the cross-sectional area of flow, t is time, and q_ℓ is lateral inflow or outflow rate per unit length of channel. For overland flow, q_ℓ is rainfall excess, A is depth of flow, and Q is the discharge per unit width of channel.

Momentum Equation.--The momentum equation for a prismatic channel can be expressed as (Henderson 1966)

$$S_f = S_o - \frac{\partial y}{\partial x} - \frac{1}{gA}\frac{\partial Q}{\partial t} - \frac{1}{gA}\frac{\partial}{\partial x}\left(\frac{Q^2}{A}\right) \qquad (12)$$

in which S_f is the friction slope, S_o is the channel bed slope, y is the depth of flow, and g is the gravitational acceleration.

The assumption of the kinematic wave approximation is that the friction slope is equal to the channel bed slope; that is, the gradients resulting from local and convective accelerations are assumed to be negligible, and the water-surface slope is assumed to be equal to the bed slope. Then the simplified momentum equation can be expressed as

$$S_o \approx S_f = f\frac{Q^2}{8gRA^2} \qquad (13)$$

in which f is the Darcy-Weisbach friction factor, and R is the hydraulic radius. By definition

$$R = \frac{A}{P} \qquad (14)$$

in which P is the wetted perimeter. Usually the wetted perimeter can be expressed as a power function of flow area

$$P = a_1 A^{b_1} \qquad (15)$$

where a_1 and b_1 are constants.

If Manning's equation is used, the simplified momentum equation is

$$S_o \approx S_f = \frac{n^2Q^2}{2.21\ R^{4/3}A^2} \qquad (16)$$

in which n is Manning's roughness coefficient.

Resistance Equations.--In a natural drainage basin, form resistance from ground cover is a very important component of the resistance to flow. The dependence of flow resistance on ground cover becomes further complicated depending on whether or not the ground cover is submerged. Rarely is the ground cover submerged in overland-flow units, and only resistance caused by flow through ground cover is considered. In channel-flow units, the probability of submerging the ground cover is larger. The resistance is then considered as the resistance caused simultaneously by flow through and flow over ground cover.

Resistance to flow for the overland-flow response units is generally expressed as a function of surface-material type, vegetation type, and density of vegetation. Palmer (1946), Ree (1949), and Ree and Palmer (1949) conducted a series of experiments in channels with various types of grasses. More recently, Kouwen and Unny (1969), Phelps (1970), Wenzel (1970), Li and Shen (1973), and Chen (1976) carried the experimental studies further. Results indicate a functional relationship between the overal Darcy-Weisbach friction factor and flow characteristics as

$$f = \frac{K_t}{N_r} \qquad (17)$$

in which f is the overall Darcy-Weisbach friction coefficient, N_r is the flow Reynolds number, and K_t is a constant describing the overall resistance. According to Chen (1976), this type of relationship can be used for a Reynolds number up to 100,000. This would cover practically all of the possible overland-flow conditions on natural surfaces.

The flow Reynolds number is defined as

$$N_r = \frac{QR}{\nu A} \qquad (18)$$

in which ν is the kinematic viscosity of water.

Assuming that the factors describing resistance to flow are independent, the overall resistance can be expressed as

$$f = f_g + f_d \qquad (19)$$

in which f_g is the Darcy-Weisbach friction factor for grain resistance only, which is a function of grain size, flow Reynolds number, and rainfall intensity, and f_d is the Darcy-Weisbach friction factor resulting from form drag resistance, which is a function of ground-cover density, size of ground cover, drag coefficient, depth of flow, and flow Reynolds number. In overland-flow cases, the value of f_g can be further assumed as

$$f_g = \frac{K_g}{N_r} \qquad (20)$$

in which K_g is the parameter describing Darcy-Weisbach friction factor for grain resistance only. Laboratory experiments show that K_g is between 30 and 60 (Chow 1959, Woolhiser 1975). K_g is assumed to be about 45 to 50.

The overall overland-flow resistance can be assumed as a function of ground-cover density as

$$K_t = K_\ell + (K_h - K_\ell)\ C_g^2 \qquad (21)$$

in which K_ℓ is the parameter describing the minimum resistance (C_g = 0.0), and K_h is the parameter describing the maximum resistance (C_g = 1.0).

In channel flow, the Darcy-Weisbach friction factor is often considered as a constant. Sometimes, the Chezy or Manning's equations are used in evaluating the overall resistance for overland flow and flow in rivers.

Relation of Discharge and Flow Area.--In general, the cross-sectional area of the flow can be expressed as a power function of discharge

$$A = \alpha Q^{\beta} \tag{22}$$

in which α and β are coefficients whose values depend on the shape of the channel, the friction slope, and the roughness of the wetted perimeter.

If the Darcy-Weisbach friction factor is used, the values of α and β can be determined by substituting equations 13, 14, 15, 17, and 18 into equation 22. The solutions are

$$\alpha = \left(\frac{K_t \nu a_1^2}{8gS_f}\right)^{1/(3-2b_1)} \tag{23}$$

and

$$\beta = \frac{1}{3-2b_1} \quad . \tag{24}$$

For overland flows or flows in very wide channels, the wetted perimeter is constant, so that $b_1 = 0$ and $\beta = 1/3$. If Manning's and Chezy's equations are used, the corresponding α and β values can similarly be determined.

The problem of water routing requires solving equations 11 and 22. These two equations can be solved either by an analytical method or a numerical method. The analytical solutions of equations 11 and 22 are available for some special cases (Eagleson 1970, Harley et al. 1970, Kibler and Woolhiser 1970, Li et al. 1975). They are not repeated here, but more detail can be found in Li (1979). Analytical approaches use the method of characteristics in x (distance) and t (time) space to describe water flow with no upslope inflows, only lateral inputs. Lateral inputs could be rainfall on planes or planes discharging into a channel. An analytical solution is desirable because no problem is encountered in computational stability and convergence. Unfortunately, the analytical solution is often restricted for practical applications because of the formulation of kinematic "shock." The shock is represented by the intersections of characteristics in the x-t plane. This produces an abrupt computational increase in flow depth (Kibler and Woolhiser 1970). Li et al. (1976b) give an indepth discussion of this limitation. For a small drainage basin with simple geometry having two planes and one channel, however, the above analytical procedure can be used. Simons et al. (1977a) report a simple drainage-basin model using small programmable calculations based on the two-planes and one-channel representation of drainage basins.

For the numerical solution, a nonlinear scheme was developed using an iterative procedure to solve a four-point implicit formulation of equations 11 and 22.[1] A linear scheme is used to obtain the initial estimate of the unknown discharge for the nonlinear scheme. The linear scheme may be used with no iterations to solve for the unknown discharge, providing the accuracy of the answer is satisfactory.

Sediment Routing

The sediment-routing portion of the model is similar to the formulation used by Simons et al. (1975). The model is further developed to consider the routing of sediment by sizes, however. Increasing attention is being paid to routing sediment by sizes because different sizes of sediment have different uptake rates of other pollutants.

Movement of sediment from drainage basins is governed by the equation of continuity for sediment and sediment-transport equations. The amount of soil that could be transported is described by equations of sediment supply, which are the amount of soil detachment from raindrop impact and surface runoff. The equations used in the model are described below.

Continuity for Sediment.--The equation of continuity for sediment can be expressed as

$$\frac{\partial G_s}{\partial x} + \frac{\partial CA}{\partial t} + (1-\lambda)\frac{\partial PZ}{\partial t} = g_s \tag{25}$$

in which

$$C = \frac{G_s}{Q} \tag{26}$$

and G_s is the total sediment-transport rate by volume per unit time, C is the sediment concentration by volume, Z is the depth of loose soil, λ is the soil porosity, P is the wetted perimeter, and g_s is the lateral sediment inflow.

Sediment load can be subdivided into different sizes, and the continuity equation can be solved for N parts as

$$\frac{\partial G_{si}}{\partial x} + \frac{\partial C_i A}{\partial t} + (1-\lambda)\frac{\partial PZ_i}{\partial t} = g_{si} \tag{27}$$

$$i = 1, \ldots, N$$

where i is the size-fraction index and N is the number of size fractions. The other terms are related to the individual size fractions as

$$C_i = \frac{G_{si}}{Q} \tag{28}$$

$$C = \sum_{i=1}^{N} C_i \tag{29}$$

and

$$Z = \sum_{i=1}^{N} Z_i \quad . \tag{30}$$

[1]Li, R. M., R. K. Simons, and D. B. Simons. A generalized kinematic wave approximation for floodrouting. Submitted to Am. Soc. Civ. Eng. Hydraul. Div. for possible publication.

Sediment-Transport Equations.--Sediment-transport equations are used to determine transport capacity of a specific flow condition. Different transport capacities can be expected for different sediment sizes. For each sediment size, the transport rate includes the bedload-transport rate and the suspended-load transport rate.

The Meyer-Peter, Müller equation is a simple and commonly used bedload-transport equation (U.S. Bureau of Reclamation 1960). It is

$$q_b = \frac{12.85}{\sqrt{\rho}\ \gamma s} (\tau_o - \tau_c)^{1.5} \tag{31}$$

in which

$$\tau_c = \delta_s (\gamma_s - \gamma) d_s \quad . \tag{32}$$

Here, q_b is the bedload transport rate in volume per unit width for a specific size of sediment, τ_o is the boundary shear stress acting on the grain, τ_c is the critical tractive force, ρ is the density of water, γ_s is the specific weight of sediment, γ is the specific weight of water, d_s is the size of sediment, and δ_s is usually 0.047 for most flow conditions (Gessler 1965). If rilling develops on the overland-flow surface, the value of δ_s should be lower.

The Meyer-Peter, Müller equation is not necessary, but an equation should be chosen that is applicable to field conditions. Other bedload-transport equations follow a form similar to equation 31.

The flow discharge, Q, and flow area, A, are determined in time and space by the water-routing procedure previously described. The corresponding value of τ_o is computed as follows. If the mean flow velocity is

$$V = Q/A \quad , \tag{33}$$

then the boundary shear stress acting on the grain (using equation 20) is

$$\tau_o = \frac{1}{8} \rho f_g V^2 = \frac{1}{8} \rho \frac{K_g}{N_r} V^2 \quad . \tag{34}$$

A commonly used and theoretically sound method for estimating suspended load was presented by Einstein (1950). Other methods are available, but the Einstein method has been found acceptable in many modeling applications.

The suspended-sediment concentration profile relating sediment concentration with depth above the bed (Einstein 1950) can be written as

$$\frac{C_\xi}{C_a} = \left(\frac{R-\xi}{\xi} \frac{a'}{R-a'}\right)^w \tag{35}$$

in which C_ξ is the sediment concentration at the distance ξ from the bed, C_a is the concentration at a distance a' above the bed, and w is a parameter defined as

$$w = \frac{V_s}{\kappa U_*} \tag{36}$$

where V_s is the settling velocity of the sediment particles, κ is the von Karman constant (assumed 0.4), and U_* is the shear velocity of the flow defined as

$$U_* = \left(\frac{\tau_*}{\rho}\right)^{1/2} \quad . \tag{37}$$

Note that

$$\tau_* = \frac{1}{8} f \rho V^2 \tag{38}$$

where f is the overall Darcy-Weisbach resistance factor previously described.

A logarithmic-velocity profile is commonly adopted to describe the vertical velocity distribution in turbulent flows such as

$$\frac{u_\xi}{U_*} = B + 2.5 \ln \left(\frac{\xi}{\eta_s}\right) \tag{39}$$

in which u_ξ is the point mean velocity at the distance ξ from the bed, B is a constant dependent on roughness, and η_s is the roughness height.

The integral of suspended-sediment load above the a' level in the flow is obtained by combining equations 35 and 39 or

$$q_s = \int_{a'}^{R} u_\xi C_\xi d\xi = C_a U_* \int_{a'}^{R} \left[B + 2.5 \ln\left(\frac{\xi}{\eta_s}\right)\right] \left(\frac{R-\xi}{\xi} \frac{a'}{R-a'}\right)^w d\xi \tag{40}$$

substituting

$$\sigma = \frac{\xi}{R} \tag{41}$$

and

$$G = \frac{a'}{R} \tag{42}$$

results in

$$q_s = C_a U_* a' \frac{G^{w-1}}{(1-G)^w} \left\{ \left[B + 2.5 \ln\left(\frac{R}{\eta_s}\right)\right] \int_G^1 \left(\frac{1-\sigma}{\sigma}\right)^w d\sigma + 2.5 \int_G^1 \ln\sigma \left(\frac{1-\sigma}{\sigma}\right)^w d\sigma \right\} \quad . \tag{43}$$

According to Einstein (1950), the concentration near the "bed layer," C_a, is related to the bedload transport rate, q_b, by the expression

$$q_b = 11.6\ C_a U_* a' \tag{44}$$

in which a' is defined as the thickness of the bed layer or twice the size of a representative sediment particle.

The average flow velocity, V, is defined by the equation

$$V = \frac{\int_0^R u_\xi \, d\xi}{\int_0^R d\xi} \qquad (45)$$

Using equation 39

$$\frac{V}{U_*} = B + 2.5 \ln(\frac{R}{\eta_s}) - 2.5 \qquad (46)$$

Einstein (1950) defined the two integrals in equation 43 as

$$I_1 = \int_G^1 (\frac{1-\sigma}{\sigma})^w \, d\sigma \qquad (47)$$

and

$$I_2 = \int_G^1 (\frac{1-\sigma}{\sigma})^w \ln\sigma \, d\sigma \qquad (48)$$

The integrals, I_1 and I_2, cannot be integrated in closed form for most values of w, so a numerical integration is necessary. An efficient numerical method of determining I_1 and I_2 in the model was developed by Li (1974).

The substitution of equations 44, 46, 47, and 48 into equation 43 yields

$$q_s = \frac{q_b}{11.6} \frac{G^{w-1}}{(1-G)^w} [(\frac{V}{U_*} + 2.5)I_1 + 2.5 \, I_2] \qquad (49)$$

The total bed-material transport per unit width of channel is

$$q_t = q_b + q_s \qquad (50)$$

Equation 50 gives the total bed-material transport per unit width of channel for a uniform size of sediment in the bed. When transport by different sizes and the entire width of channel are considered, the sediment-transport capacity of the ith size fraction of a sediment mixture would be

$$G_{ci} = P \, F_{ai} \, q_t \qquad (51)$$

in which G_{ci} is the sediment-transport capacity for the ith size, and F_{ai} is the adjusted fraction of the ith size sediment, which is determined in the next section.

Equations for Sediment Supply.--The sediment-supply rate is another determining factor in the actual sediment-transport rate. Simons et al. (1975) stated that the sediment supply depends on the initial depth of loose soil left from previous storms, the amount of soil detachment by raindrop impact, and the amount of soil detachment by flow.

Soil Detachment by Raindrop Impact.--Raindrop impact is a primary source of kinetic energy for detaching soil from any unprotected land surface. Ellison (1944) made a comprehensive study of raindrop splash. Laws and Parson (1943), Mutchler (1967), and Young and Wiersma (1973) contributed to the study of detachment by raindrop impact. The conclusion drawn from these studies is that soil detachment is a function of erosivity of rainfall and the erodibility of the soil particles. The erosivity is directly related to the energy produced by raindrop impact and is generally formulated as a power function of rainfall intensity, size of droplet, cover condition, and terminal velocity of the drop. Carter et al. (1974) reported that energy produced by raindrop impact is also a function of air temperature, season of the year, and storm duration. In the model, the potential rate of soil detachment by raindrop impact in inches per hour is assumed as a power function of rainfall intensity as given by Meyer (1971)

$$D_r = a_5 I^{b_5} (1 - \frac{Z_w}{Z_m})(1-C_g)(1-C_c) \qquad (52)$$

in which I is the rainfall intensity, a_5 is a parameter depending on soil characteristics, and b_5 is a constant (b_5 = 2.0, Meyer (1971)). The term Z_w is the depth of water plus the loose soil. The term Z_m is the maximum penetration depth of raindrop splash. According to Mutchler and Young (1975), Z_m can be equal to three times the median raindrop size. The median raindrop size is often expressed as a power function of rainfall intensity, I. Therefore, Z_m can be written as

$$Z_m = 3(2.23 \, I^{0.182}) \qquad (53)$$

in which Z_m is in millimeters.

Equation 52 is valid when Z_w is less than Z_m. When the depth of loose soils plus the water depth is greater than Z_m, D_r is zero; that is,

$$D_r = 0, \text{ if } Z_w > Z_m \qquad (54)$$

If trees are very tall, the drops from leaves can regain terminal velocity and have the same erosive potential. Ground-cover density under trees is usually very high, however, thus protecting the surface. The canopy-cover density, C_g, in equation 52 should be the portion that can protect the surface effectively from raindrop impact.

The potential rate of loose-soil detachment, D_r, is expressed in units of depth per unit time. Thus, the new amount of loose soil available for transport is

$$Z_{i_{New}} = Z_{i_{Old}} + F_i \, D_r \, \Delta t \qquad (55)$$

where Z_i is the amount of loose soil available and F_i is the original percentage of sediment in a given size fraction designated by i.

The percentage in each size fraction on the surface changes over time because of armoring. Water transports the smaller sizes of sediment more easily and leaves larger sized fractions behind. Thus, the percentages of surface material need adjustment at each time step. If the total depth of loose soil is greater than D_a (the thickness of the armor layer), the adjusted percentages, F_{ai}, can be written as

$$F_{ai} = \frac{Z_i}{Z} \quad . \tag{56}$$

If the total loose-soil depth, Z, is less than D_a, the adjusted percentages must account for the layer of undisturbed soil that is distributed according to the original percentages plus the loose soil that covers it,

$$F_{ai} = \frac{1}{D_a} [Z_i + F_i (D_a - Z)] \quad . \tag{57}$$

The thickness of the armor layer can be determined as the maximum size of particles in motion. The size of sediment for which 84 percent of the sample is finer by weight, D_{84}, is usually assumed to be representative of the armor layer.

Soil Detachment by Surface Runoff.--The amount of soil detachment by surface runoff is determined by comparing the total sediment-transport capacity to the total available amount of loose soil. The total sediment-transport capacity is

$$G_c = \sum_{i=1}^{N} G_{ci} \quad . \tag{58}$$

By substituting the total sediment-transport capacity, G_c, into the transport rate in equation 25, the total potential change in loose soil, Z^P, can be determined as

$$\Delta Z^P = \frac{\partial Z}{\partial t} \Delta t \quad . \tag{59}$$

If $\Delta Z^P > -Z$ (that is, positive aggradation or negative degradation is less than Z), the loose-soil storage is enough for transport and no detachment of soil by surface runoff is expected. Soil is detached if $\Delta Z^P < -Z$, and the amount of detachment is

$$D = -D_f [\Delta Z^P + Z] \tag{60}$$

in which D is the total amount of detached soil and D_f is defined as a detachment coefficient with values ranging from 0.0 to 1.0 depending on soil erodibility. As an example, if the flow were over a nonerodible surface such as gravel, the value for D_f would be zero. If the flow were in a river where the riverbed is always loose and available for transport, the value for D_f would be unity.

The new amount of loose soil should be further modified as

$$Z_{i_{Total}} = Z_{i_{New}} + D\, F_i \tag{61}$$

in which Z_i is calculated for each size fraction of sediment.

Numerical Procedure for Sediment Routing.--Similar to the water-routing scheme in its finite difference structure, the numerical routing of sediment by sizes allows a constant check on continuity at each time and space point. This constant mass balance permits tracking of aggradation and degradation on the land surface and in stream channels. Such information defines lateral loadings to stream channels from overland flow. Channel segments where storage or scour is occurring can be delineated to help in defining the movement of sediment. Segments with excessive aggradation or degradation indicate areas where fisheries may be harmed. The movement and storage of sediment is particularly useful in relating land-use activities to sediment yields. Channel storage will sometimes mask actual increases in sediment production until a runoff event of significant magnitude "flushes" the stored material. By numerical routing, storage areas and the size of the "flushing" event can be determined.

APPLICATIONS OF METHODOLOGY

The methodology presented in the previous sections has been used to develop numerous physical-process models that estimate water and sediment yields from land surfaces, roadways, and drainage basins. The abundance of models precludes inclusion of all of them in this paper.

Instead, a single drainage-basin application is presented with a description of the approach used in parameter estimation and model adjustment. Two levels of models are used. One is a complex geometry, finite-difference routing scheme (Simons et al. 1975), and the other is a simplified geometry, analytical routing scheme (Simons et al. 1977c). Watershed 17 in the Beaver Creek Experimental Forest near Flagstaff, Arizona, was chosen for application because of its relatively good data base and because it is characterized by surface runoff and erosion. The drainage basin is fairly flat, with most slopes less than about 10 percent. Total area is 116.3 ha. Soils are stony clays, silt loams, and stony silt loams. Saturated hydraulic conductivities are about 1.3 mm/hour. Ground cover is about 34 percent, and canopy cover is about 10 percent. Interception values as derived from Zinke (1965) are estimated at 2.5 mm for ground cover and about 7.5 mm for the canopy. The minimum-resistance parameter estimated for a gravel surface is 500 and the maximum estimated is about 42,000. These values provide a reasonable calibrated flow resistance for the ground-cover conditions in the area. Manning's n for the channel is estimated at about 0.05.

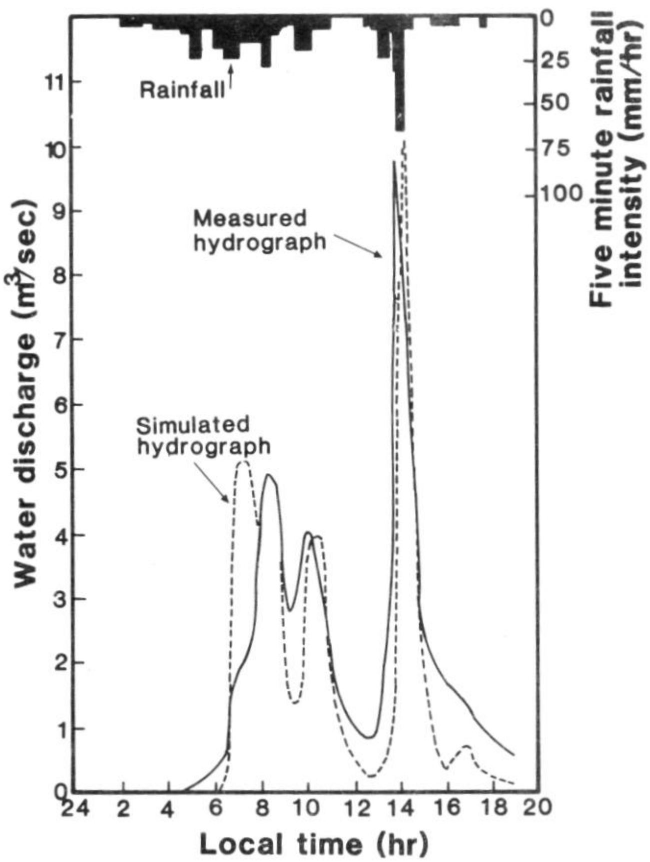


Figure 3.--Water hydrograph from Watershed 17 for the September 5, 1970 storm (after Simons et al. 1975).

The model was calibrated with four parameters and one storm on Watershed 17. The storm was on Labor Day, September 5, 1970. Initial moisture content was varied within reasonable limits for field conditions to match runoff volume, and overland-flow resistance was varied to match timing of the peaks; the rainfall-detachment coefficient, a_5, and overland-flow detachment coefficients, D_f, were adjusted to account for the measured sediment yields. The a_5 coefficient is closely related to and adjusted by considering the fine fraction of the yield, and D_f can be adjusted with respect to the total yield. The results obtained by the water calibration, using the initial moisture conditions and flow resistance, are shown in figure 3. Simulated and measured hydrographs match well. Simulated sediment yield for the storm was within 15 percent of measured values.

The parameters from Watershed 17 were then used in predicting the response of a nearby 127-ha drainage basin, Watershed 1, to five rainfall storms. Watershed 1 is a little steeper than Watershed 17 but had been clearcut so that canopy cover was zero. Other parameters related to soils and flow resistance are similar for both drainage basins. The model was used to simulate outflow hydrographs and sediment yields for five storms from Watershed 1. The only variable that was adjusted between the storms was the antecedent moisture. All other variables remained the same. Table 1 lists the results of the simulations as compared to the measured values for the water hydrographs. Only two sediment-yield values were available, and both of those were predicted within 25 percent.

Table 1--Errors in simulation

Storm	Measured surface runoff	Percentage difference between simulated and measured values: Water yield	Peak flow	Time to peak
	Millimeters			
WATERSHED 17				
Sept. 5, 1970 (calibration storm)	101	-14.6	+11.4	+2.1
WATERSHED 1				
Sept. 5, 1970	41.7	+12.9	-0.2	-3.1
Sept. 6, 1967	0.76	+7.2	-7.7	+6.7
Nov. 22, 1965	2.79	+1.3	-0.4	+2.1
Nov. 24, 1965	2.03	-23.7	+31.0	-7.7
Nov. 25, 1965	6.10	-23.2	+23.6	-6.3

The good agreement between the predicted and measured values obtained by only adjusting a single model parameter indicates the transferability of this model from one drainage basin to another with a minimum of recalibration. This occurs because the model adequately simulated the controlling physical processes of overland-flow and surface-erosion characteristics of the drainage basin. If the physical processes had been different, the model would not be applicable.

Application of Simplified Water- and Sediment-Yield Model to Watershed 17

In this example, the water- and sediment-yield model (SEDWAT) developed by Simons et al. (1977c) is used to assess changes in streamflow and sediment yield resulting from different drainage-basin practices. In SEDWAT, drainage-basin geometry is represented by two planes and one channel, and water is analytically routed. Simulated practices are mechanical site preparation, timber harvesting, and grazing. Watershed 17 is chosen for use in these hypothetical examples. The base event was the storm of September 5, 1970. Peak recorded discharge was about 9.77 cm, and runoff volume was about 101 mm. Sediment production was 0.67 t/ha. Using the same data base and similar calibation, model simulations produced a peak of 10.27 cm, a water yield of 5.25 mm, and sediment yield was again 0.67 t/ha. All three of these simulated values are quite good because the controlling processes are correctly incorporated in the simplified model. Because the simulated and measured responses are comparable, the simulated results will be used as initial or baseline conditions.

To demonstrate the use of physical-process models for analyzing future management activities, four planning alternatives were simulated.

Alternative 1. Mechanical Site Preparation

The major modification to the drainage basin is compaction of the soil. The percent of area with ground and canopy remains the same, but the hydraulic conductivity is reduced to 0.25 mm/hour over the entire drainage basin (table 2). The simulated sediment yield is what would be expected after the loose material has been flushed out of the system during an earlier storm (table 3); that is, the sediment-detachment coefficients have not been changed.

Alternative 2. Removal of Canopy Cover

Although economical, clearcutting may be accompanied by high rates of erosion and subsequently higher yields of sediment, depending on the field conditions. The increase in sediment yield in this case is simulated by combining the effects of mechanical site preparation from alternative 1 with removal of all canopy cover. The hydraulic conductivity is again 0.25 mm/hour, and the percent of area covered by ground cover and canopy cover is set to zero (table 2). A ground cover of zero is extreme and is used here for comparative purposes.

Table 2--Parameters used in SEDWAT simulation of management alternatives

	Baseline and alternative parameters		
Alternative	Hydraulic conductivity	Canopy cover	Ground cover
	mm/hour	-------Percent--------	
Baseline condition	1.27	10	34
1	.25	10	34
2	.25	0	0
3	1.02	0	17
4	.76	10	0

Table 3--Simulated drainage-basin responses using SEDWAT for four land-use alternatives. Baseline conditions: peak discharge = 10.27 m^3/second, water yield = 95.25 mm, sediment production = 0.67 t/ha.

	Ratio of simulated result to baseline condition		
Alternative	Peak discharge	Water yield	Sediment production
1	1.17	1.21	1.06
2	1.17	1.21	3.25
3	1.13	1.04	1.62
4	1.15	1.09	2.79

Alternative 3. Removal of Trees with Less Ground Disturbance

If care is taken to minimize the disturbance of ground cover, the canopy can be removed with a considerably lower sediment yield (table 3). This may be seen by reducing the ground-cover percent to half the original and the hydraulic conductivity to 1.02 mm/hour (table 2). The canopy is completely removed.

Alternative 4. Overgrazing

Livestock or large populations of native animals can often cause severe damage to the ground cover by overgrazing. The effects of overgrazing can be included in the model by simulated removal of the ground cover and reduction of the hydraulic conductivity through compaction caused by trampling. In this example, the conductivity is reduced to 0.76 mm/hour and the ground cover is zero (table 2).

These alternatives are summarized in table 2. The results of model simulation of the baseline condition and all four alternatives are listed in table 3. Obviously, land-use alternative 2 creates conditions allowing the highest production of sediment. In every alternative, the peak discharge was increased, primarily because of a drop in hydraulic conductivity and partially from a decrease in flow resistance. Water yield also increased in every alternative because of decreases in interception and infiltration.

Table 4--Simulated sediment-water concentration for four land-use alternatives

Alternative	Concentration of sediment in mixture	
	Parts per million	Ratio to baseline
Baseline conditions	706	1.00
1	619	.88
2	1897	2.69
3	1100	1.56
4	1808	2.56

The relative effects of each alternative can also be compared in terms of average concentration of the sediment-water mixture (table 4). These comparisons represent a key point for assessing management alternatives. If increases in water and sediment yields are primary criteria, alternative 2 would be the worst; however, if sediment concentration is used, alternative 4 is just as undesirable (table 4).

Note that the parameters chosen in all of these examples reflect particular responses to land use. These numbers were arbitrarily chosen and could reflect, in some instances, the most severe impact. One of the strengths of using a physical-process simulation model is that effects of different levels of land use may be examined.

Numbers generated from the simulations can be compared as relative ratios or marginal values. Effects of land-use alternatives can then be discussed in terms of percentage increase or decrease. Simultaneously, the absolute simulated values can be checked to see if they are physically realistic and, if so, useful as water-quality indicators.

Planning decisions based on state-of-the-science techniques such as mathematical models have two other attributes. First, they quantify drainage-basin response in a manner that can be readily related to measurable drainage-basin characteristics. Second, they are becoming accepted as the best method for determining system response to drainage-basin change. These two reasons alone indicate that mathematical modeling will become, by necessity, a widely used technique in estimating water and sediment yields from forested drainage basins.

SUMMARY AND CONCLUSIONS

Mathematical models based on physical processes governing runoff and sediment yield from drainage basins can be developed and used for a variety of purposes. The key is correct simulation of the controlling processes in the chosen drainage basin. If properly formulated and documented, models can be a valuable aid to forest managers. When correctly applied, physical-process models can provide insights into the movement of water and sediment within and from drainage basins for initial and disturbed conditions. This paper has attempted to bridge gaps between model builders and field users to show the applicability and utility of physical-process models.

LITERATURE CITED

Bradford, R. G., D. A. Farrell, and W. E. Larson.
1973. Mathematical evaluation of factors affecting gully stability. Soil Sci. Soc. Am. Proc. 37:103-107.

Carter, C. E., J. D. Greer, H. J. Brand, and J. M. Floyd.
1974. Raindrop characteristics in south-central United States. Trans. Am. Soc. Agric. Eng., p. 1033-1037.

Chen, C. L.
1976. Flow resistance in broad shallow grassed channels. J. Hydraul. Div., Am. Soc. Civ. Eng. 102, HY3, p. 307-322.

Chow, V. T.
1959. Open channel hydraulics. McGraw-Hill, New York. 680 p.

Eagleson, P. S.
1970. Dynamic hydrology. McGraw-Hill, New York. 462 p.

Einstein, H. A.
1950. The bed load function for sediment transportation in open channel flows. USDA Tech. Bull. 1026, 77 p. Washington, D.C.

Ellison, W. D.
1944. Studies of raindrop erosion. Agric. Eng., April, p. 131-136, May, p. 181-182.

Gessler, J.
1965. The beginning of bedload movement of mixtures investigated as natural armoring in channels. Translated by E. A. Prych, W. M. Keck Lab. Hydraul. and Water Resour., Calif. Inst. Technol., Pasadena.

Green, W. H., and G. A. Ampt.
1911. Studies on soil physics, part 1: The flow of air and water through soils. J. Agric. Sci. 4:1-24.

Harley, B. M., F. E. Perkins, and P. S. Eagleson.
1970. A modular distributed model of catchment dynamics. Mass. Inst. Tech., Dep. Civ. Eng., Hydrodyn. Lab. Rep. 133. Boston, Mass.

Henderson, F. M.
1966. Open channel flow. Macmillan, New York. 522 p.

Kibler, D. F., and D. A. Woolhiser.
1970. The kinematic cascade as a hydrologic model. Hydrol. Pap. 39, Colo. State Univ., Fort Collins.

Kouwen, N., and T. E. Unny.
1969. Flow retardance in vegetated channels. J. Irrig. Drain. Div., Am. Soc. Civ. Eng. 95(IR2):329-342.

Laws, O. J., and D. A. Parson.
1943. The relation of raindrop size to intensity. Trans. Am. Geophys. Union 24:452-459.

Li, R. M.
1974. Mathematical modeling of response from small watershed. Ph. D. thesis. Dep. Civ. Eng., Colo. State Univ., Fort Collins. 212 p.

Li, R. M.
1979. Water and sediment routing from watersheds. In Modeling of rivers, p. 9-1-9-88. H. W. Shen, ed. John Wiley and Sons, New York.

Li, R. M., and H. W. Shen.
1973. Effect of tall vegetation on flow and sediment. J. Hydraul. Div., Am. Soc. Civ. Eng. 99(HY5):793-814.

Li, R. M., D. B. Simons, and M. A. Stevens.
1975. On overland flow water routing. In Proceedings, National Symposium on Urban Hydrology and Sediment Control, Lexington, Ky., p. 237-244.

Li, R. M., D. B. Simons, and K. G. Eggert.
1976a. Process model of infiltration and subsurface flow routing. Colo. State Univ. Rep. CER76-77RML-DBS-KGE20. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Li, R. M., D. B. Simons, L. S. Shiao, and Y. H. Chen.
1976b. On kinematic wave approximation for flood routing. Rivers 76, vol. 1, p. 377-398. In 3rd Annual Symposium on Inland Waterways for Navigation Flood Control and Water Diversion, Colo. State Univ., Fort Collins.

Li, R. M., M. A. Stevens, and D. B. Simons.
1976c. Solution to Green-Ampt infiltration equation. J. Irrig. Drain. Div., Am. Soc. Civ. Eng. 102(IR2):239-248.

Meyer, L. D.
1971. Soil erosion by water on upland areas. In River mechanics, vol. II., p. 27-1-27-25. Hsieh Wen Shen, ed. Fort Collins, Colo.

Mutchler, C. K.
1967. Parameters for describing raindrop splash. J. Soil Conserv., May-June, p. 91-94.

Mutchler, C. K., and R. A. Young.
1975. Soil detachment by raindrops. In Proceedings of the Sediment-Yields Workshops, Oxford, Miss., p. 113-117. USDA Agric. Res. Serv., ARS-S-40. New Orleans, La.

Palmer, V. J.
1946. Retardance coefficient for low flow in channels lined with vegetation. Trans. Am. Geophys. Union 27:187-197.

Phelps, H. O.
1970. The friction coefficient for shallow flows over a simulated turf surface. Water Resour. Res. 6(4)1220-1226.

Piest, R. F., J. M. Bradford, and R. G. Spomer.
1975. Mechanisms of erosion and sediment movement from gullies. In Present and prospective technology for predicting sediment yield and sources, p. 162-176. USDA Agric. Res. Serv., ARS-S-40. New Orleans, La.

Ree, W. O.
1949. Hydraulic characteristics of vegetation for vegetated waterways. Agric. Eng. 30(4):184-189.

Ree, W. O., and V. J. Palmer.
1949. Flow of water in channels protected by vegetative linings. U.S. Soil Conserv. Tech. Bull. 967, 115 p. Washington, D.C.

Simons, D. B., R. M. Li, and M. A. Stevens.
1975. Development of models for predicting water and sediment routing and yield from storms on small watershed. Colo. State Univ. Rep. CER74-75DBS-RML-MAS24. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, and L. Y. Shiao.
1977a. Simple water routing and yield model using a small programmable calculator. Colo. State Univ. Rep. CER76-77DBS-RML-KGE52. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, and L. Y. Shiao.
1977b. Formulation road sediment model. Colo. State Univ. Rep. CER76-77DBS-RML-LYS50. Prepared for USDA For. Serv., Rocky Mt. For. and Range Ext. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, L. Y. Shiao, and K. G. Eggert.
1977c. Storm water and sediment runoff simulation for upland watersheds using analytical routing technique. Volume 1--Water routing and yield, Colo. State Univ. Rep. CER77-78DBS-RML-LYS-KGE16. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, and T. J. Ward.
1977d. A simple procedure for estimating on-site erosion. In Proceedings, International Symposium on Urban Hydrology, Hydraulics, and Sediment Control, Lexington, Ky., p. 95-102.

Simons, D. B., R. M. Li, and T. J. Ward.
1978a. Mapping of potential landslide areas in terms of slope stability. Colo. State Univ. Rep. CER78-79DBS-RML-TJW19. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, and T. J. Ward.
1978b. Watershed segmentation by a digital computer for mathematical modeling of watershed response. Colo. State Univ. Rep. CER78-79DBS-RML-TJW24. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., R. M. Li, T. J. Ward, and K. G. Eggert.
1979a. Estimation of input parameters for modeling of water and sediment yields. Colo. State Univ. Rep. CER78-79DBS-RML-TJW-KEG50. Prepared for USDA For. Serv., Rocky Mt. For. and Range Exp. Stn., Flagstaff, Ariz.

Simons, D. B., T. J. Ward, and R. M. Li.
1979b. Sediment sources and impacts in the fluvial system. In Modeling of rivers, p. 7-1-7-27. H. W. Shen, ed. John Wiley and Sons, New York.

Thompson, J. R.
1964. Quantitative effect of watershed variables on rate of gully head advancement. Trans. Am. Soc. Agric. Eng. 7:54-55.

U.S. Bureau of Reclamation.
1960. Investigation of Meyer-Peter, Müller bedload formulas. Sedimentation Section, Hydrology Branch, Division of Project Investigations, U.S. Dep. Inter., Bur. Reclam., Denver, Colorado. 22 p.

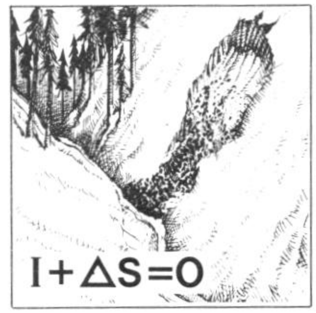
Ward, T. J.
1976. Factor of safety approach to landslide potential delineation. Ph. D. thesis. Colo. State Univ., Fort Collins. 119 p.

Wenzel, H. G.
1970. The effects of raindrops impact and surface roughness on sheet flows. WRC Res. Rep. 3. Water Resour. Center, Univ. Ill., Urbana.

Woolhiser, D. A.
1975. Simulation of unsteady overland flow. In Unsteady flow in open channels, p. 485-508. K. Mahmood and V. Vevjevich, eds. Water Resour. Publ., Fort Collins, Colo.

Young, R. A., and J. L. Wiersma.
1973. The role of rainfall impact on soil detachment and transport. Water Resour. Res. 9:1629-1936.

Zinke, P. J.
1965. Forest interception studies in the United States. In International Symposium on Forest Hydrology, p. 137-161. W. E. Sopper and H. W. Lull, eds. Pergamon Press, Oxford.



Sedimentation in the Chaparral: How Do You Handle Unusual Events?

Raymond M. Rice

ABSTRACT

Processes of erosion and sedimentation in steep chaparral drainage basins of southern California are described. The word "hyperschedastic" is coined to describe the sedimentation regime which is highly variable because of the interaction of marginally stable drainage basins, great variability in storm inputs, and the random occurrence of brush fires.

The difficulties and advantages of describing chaparral sedimentation using either empiric or process models are dicussed. Modeling based on Monte Carlo simulation is suggested as a way of capturing many of the benefits of empiric and process models.

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INTRODUCTION

Processes of erosion and sedimentation in steep chaparral drainage basins in southern California are described, as the basis of a wider discussion of sediment budgets, sediment routing, and drainage-basin modeling in general. A simulation approach to sediment budgets and sediment routing is proposed which incorporates some desirable features of process models and empiric models.

THE CHAPARRAL ENVIRONMENT

Chaparral occurs in a wide variety of environments. This discussion treats a narrower range of conditions occurring in the steep "front country" drainage basins of the San Gabriel and San Bernardino Mountains in the Transverse Range in southern California. These drainage basins present the greatest flood and management problems, and this limitation in scope permits a sharper definition of relationships than would be possible in a more general discussion.

Climate

Chaparral grows in a Mediterranean climate. In southern California, most chaparral stands receive between 400 and 700 mm of rain annually with about three-quarters of the precipitation falling during December through March. Rainfall varies greatly from year to year. In Glendora, rain in the wettest year of a 97-year record was over seven times that in the driest year (fig. 1).

Winter temperatures average 9°C. The summer months average about 20°C and are normally without rain. Fire is favored by severe heat waves, accompanied by humidities of 15 to 20 percent, which occur several times each season. Maximum temperatures during these heat waves are typically around 40°C. Another fire-promoting element of the climate is the strong, desiccating, foehn winds that occur most frequently from mid-September through December. Chaparral vegetation has evolved in response to repeated fire and great annual and seasonal variations in moisture availability and evaporative stress.

Geology and Topography

The eastern portion of the Transverse Range is mainly underlain by Precambrian to Cretaceous igneous and metamorphic rocks; to the west, Cenozoic sedimentary formations predominate (California Division of Mines and Geology 1969). In both areas, poorly developed, coarse-textured, erodible soils are the rule. Soil depths average about 60 cm, but depth is of little hydrologic importance because the parent material is typically so weathered or fractured that for a depth of 1 to 3 m its ability to store and transmit water is little different from the overlying soil (Krammes 1967). As a consequence, chaparral drainage basins normally have a thick, hydrologically active mantle.

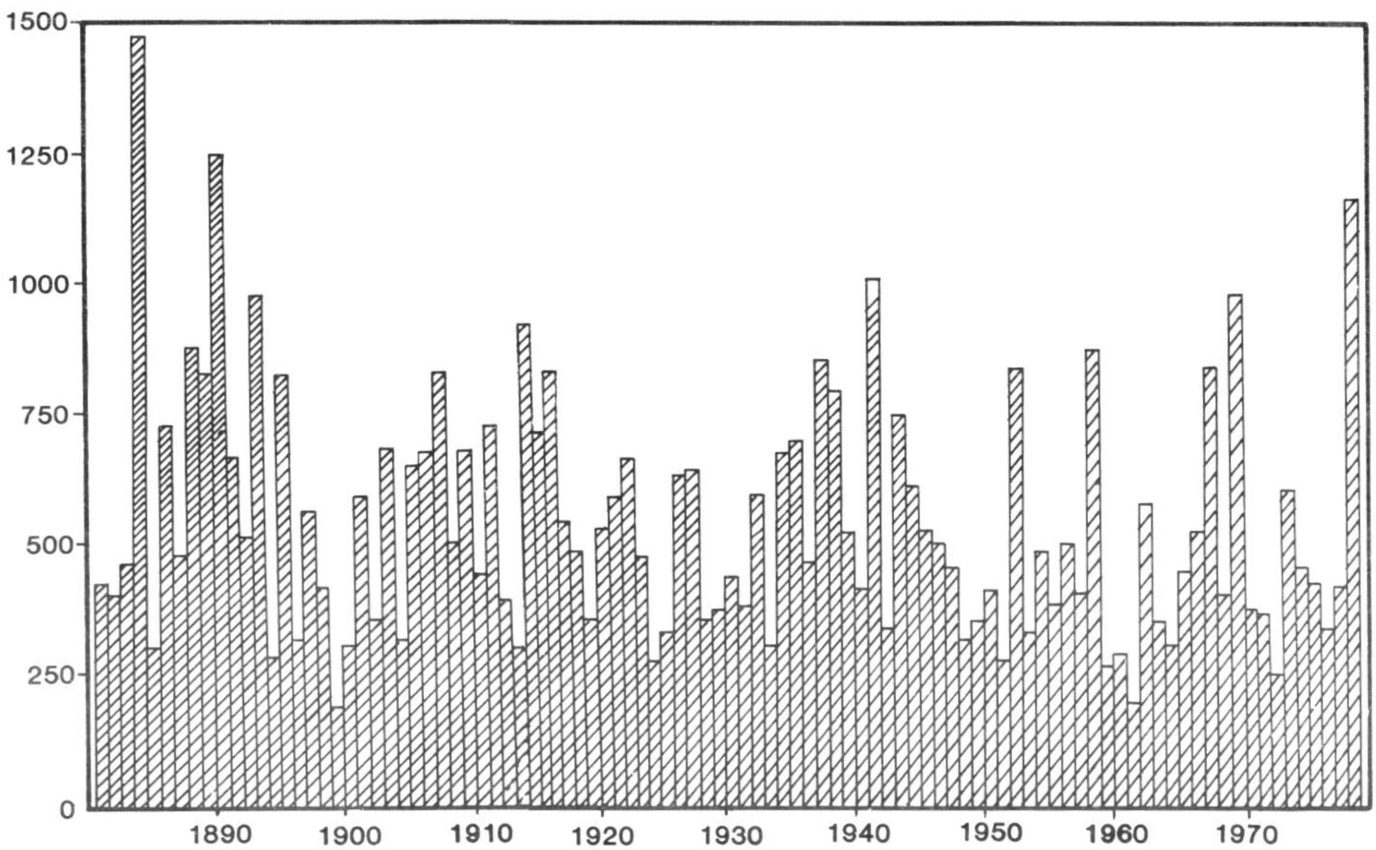

Figure 1.--Annual rainfall in the West rain gage. Glendora, California.

The mountains of the Transverse and Coast Ranges are currently being uplifted at a rate of about 60 cm a century (Scott and Williams 1974). This rapid uplift has rejuvenated the drainage system, producing average channel gradients of about 15 percent and average hillslope gradients of about 60 percent. The steep gradients cause the hillslopes to be very efficient at delivering soil to the stream channels, and the channels can also move large amounts of sediment if flows are high enough.

Vegetation

Chaparral plants are adapted to the Mediterranean climate and to the associated fires. Adaptations to drought, such as leaf morphology and the accumulation of extractive compounds in chaparral, contribute to high flammability. Because the plant community is specialized for postfire regeneration, the vegetative and hydrologic recovery of chaparral drainage basins is rapid after fire. As a result of rapid vegetative regrowth, Kittredge (1939) found that, 17 years after a fire, litter accumulation was 13 t/ha and, in an area not burned for over 50 years, the accumulation was 37 t/ha. This litter accumulation protects the soil surface, increases its roughness, and temporarily stores rain falling at rates greater than the infiltration rate. These actions insure that surface erosion and overland flow rarely occur.

Fire

Ecological evidence suggests that fire has played a prominent role in chaparral communities for millions of years. Keeley (1977) proposes that the existence of the genera Arctostaphylos and Ceanothus, the majority of whose species do not sprout, indicates that the interfire interval formerly may have been longer than it is now. He calls this the "Stochastic-Fire Hypothesis." He argues that because lightning fires in southern California occur mainly in summer and mainly at high elevations in the interior portions of the mountain ranges, for most of the chaparral to be ignited by lightning would be difficult. He suggests that relatively few lightning fires would continue to burn long enough to be swept downslope and coastward by the autumn foehn winds. Aschmann (1977) believes that increased fire frequency resulted both from accidental and purposeful burning on the part of aborigines. Preliminary analyses of varved cores taken from the Santa Barbara basin support this idea (Byrne et al. 1977). That record indicated that very large fires may have occured every 20 to 40 years. A 150-year core taken from the 16th and 17th centuries showed considerably more variability in charcoal concentrations than one spanning the 40 years between 1931 and 1970. This suggests the hypothesis that the greater temporal and areal homogeneity of modern fire occurrence is a result of the population increase.

Figure 2.--Dry-ravel deposits which had been blocking channel of Monroe Canyon are eroded by storm of October 9-10, 1960.

EROSIONAL PROCESSES

Dry Ravel

Dry ravel is the most ubiquitous erosional process occurring in the chaparral. This process, sometimes termed "dry creep" (Krammes 1965), is the downslope movement by gravity of individual grains or aggregates of soil. The slope threshold for dry ravel is about 30 degrees. Kittredge (1939) estimated that dry ravel occurred on about 17 percent of his study area. My own inventory of the chaparral zone of southern California indicates that about 25 percent of the area is subject to dry ravel. The proportion of the front-country drainage basins that are steep enough for dry ravel is higher. In Englewild Canyon above Glendora, 40 percent of the slopes are steeper than 30 degrees. Englewild has gentler topography than the majority of the front-country drainage basins. Therefore, about two-thirds of the slopes of front-country drainage basins are probably steep enough for dry ravel.

Dry ravel is initiated by many types of small disturbances. Small sand runs can start when deer, small rodents, and even birds walk on a slope. One of the principal triggers, however, seems to be the movement of vegetation during periods of strong foehn winds. Five years of monitoring the Arroyo Seco drainage showed an average dry-season erosion rate of 0.96 m^3/ha, which was 55 percent of the total surface erosion measured (Anderson et al. 1959). Later studies of postfire erosion revealed that about half of the wet-season erosion was actually dry ravel occurring between winter rainstorms (Krammes and Osborne 1969). Consequently, I believe that, with unburned chaparral, the dry-ravel rate is about 1.4 m^3/ha per year. In the autumn of 1959, a wild fire swept over the Arroyo Seco study area resulting in about 39 m^3/ha of dry-ravel erosion during the next 3 months. The immediate effect of fire on dry ravel is to consume the forest litter that has been serving as temporary barriers to the downslope movement. Accelerated ravel occurs within minutes of the passage of the fire and can produce debris cones blocking stream channels within a few hours (fig. 2). Fire also accelerates dry ravel by the creation of a water-repellent layer (DeBano 1969). The waterrepellent layer traps precipitation in the surface soil where it is more easily evaporated. In addition to reducing percolation, it also prevents the capillary rise of water to the surface from deeper soil strata. By inhibiting the vertical movement of water, the water-repellent layer causes the soil surface to dry more quickly between storms and, therefore, to be subject to dry ravel for a greater portion of the wet season. Although fire-related water repellency is detectable for a long time, it ceases to affect runoff or erosion significantly within 2 or 3 years after a fire.

Surface Erosion by Flowing Water

Surface erosion is a rarity in unburned chaparral drainage basins because litter usually protects the soil surface. Even without that protection, the thick, hydrologically active mantle can store and transmit large volumes of water. Furthermore, rainfall intensities rarely exceed infiltration rates. In none of the 300 largest storms during 24 years on the San Dimas Experimental Forest did the rainfall rate exceed the average hydraulic conductivity of the soil for 1 hour; in only 1 percent of the storms was it exceeded for 30 minutes (Reimann and Hamilton 1959). In all, only about 2.5 percent of the precipitation fell at a rate higher than the infiltration rate of the soil, and some of this rain was dispersed in time so that it had little effect on runoff. Both Colman (1953) and Troxell (1953) reported that almost no overland flow occurred during the flood of 1938. During that storm, less than 1 percent of the precipitation was measured as surface runoff from research plots, even though the streamflow from various drainage basins amounted to 16 to 38 percent of storm precipitation. Surface erosion remains almost nil without the transporting medium of overland flow, even though some minor erosion may result from drop impact of rainfall or throughfall.

The passage of a fire through chaparral usually greatly increases surface erosion by altering hillslope hydrology. A water-repellent layer is often created. This layer presents a relatively impervious barrier to deeper percolation. As a result, a drainage basin that might have a hydrologically active mantle 2 m thick when unburned may have its effective thickness reduced by fire to a few centimeters (Krammes and Osborne 1969). A recently burned slope can become "saturated" by just a few centimeters of precipitation.

Characteristically, dry ravel after fire accumulates during summer and fall on the flatter portions of burned slopes and in ephemeral channels. Overland flow the following winter scours most of its sediment load from these erodible deposits. The scenario may be repeated several winters after the fire as the rills formed during each runoff period are refilled by dry ravel between storms and during summers.

Landslides

Landslides were not considered as an important erosional process in the chaparral until the past decade when they were recognized as both an important erosional mechanism (Bailey and Rice 1969) and as a serious threat to life and property (Campbell 1975). They were probably ignored in earlier years because of their infrequency. Campbell (1975) placed the return period of the average landslide-producing storm in the Santa Monica Mountains at between 10 and 25 years. The return period seems to be less than 10 years for the front-country portions of the San Gabriel and San Bernardino Mountains. After being alerted to the significance of landslides by the storms of 1966 and 1969, I found records of soil slips (as they were called) on the San Dimas Experimental Forest in 1933, 1938, 1943, and 1965. These landslides occurred during storms producing about 500 mm of precipitation in a 5-day period, with 150 to 200 mm falling in 24 hours late in the storm. Campbell (1975) suggests that about 250 mm of antecedent precipitation followed by a period of rainfall in excess of 6 mm/hour constitute the conditions necessary for soil slips and debris flows in the Santa Monica Mountains. Bailey (1967) belives that the 12-day (before the end of the landslide-producing storm) precipitation amount is a good descriptor of the landslide-producing capabilities of a storm.

Data from 1966 and 1969 storms on the San Dimas Experimental Forest (Rice et al. 1969, Rice and Foggin 1971) give some appreciation for the effect of storm size on landslide erosion in the chaparral. These data may be affected, however, by the fact that the area burned in 1960. By Bailey's (1967) criterion, the 1966 and 1969 storms have return periods of about 11 and 43 years, respectively. Based on the 24-hour amount, the return periods would be 7 and 22 years. In the 1966 storm, 1.7 percent of the area produced landslides; in 1969, landslides were found on 5.5 percent of the area. In 1966, the erosion rate was 21.1 m^3/ha; in 1969, it was 298 m^3/ha. In 1966, the gentlest slope upon which a landslide was found was 40 degrees. In 1969, the threshold had lowered to 31 degrees. Campbell (1975) identified a range of 26 degrees to 45 degrees as the most common slope for the occurrence of soil slips. Slips on the San Dimas Experimental Forest were measured on slopes ranging from 31 to 48 degrees.

Channel Processes

Stream channels constituted a buffer in the routing of sediment from chaparral drainage basins. Virtually all dry-ravel deposits stop there (fig. 2). They serve as magazines of sediment which insure that most high flows occurring will transport sediment from the drainage basin at near their capacity. A good proportion of the landslide deposits will also come to rest in the stream channel because much of the material they transport is not readily carried by the stream. Consequently, sediment tends to accumulate in the stream channel and is flushed out infrequently by high flows, often in the first

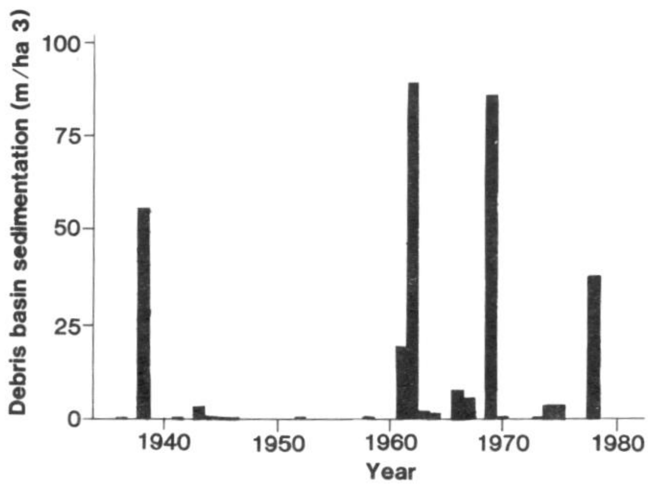

Figure 3.--Annual accumulate in Bell IV debris basin. (Hydrologic Year 1961 and 1962 record combined, sediment distributed between them based on records from nearby basins. Hydrologic Year 1978 is occular estimate, survey data not yet available.)

postfire years. Observed postfire erosion rates depend on the particular sequence of fire and winter rains. For example, in the Bell IV drainage basin on the San Dimas Experimental Forest, about 108 m^3/ha of sediment were discharged in the first 2 postfire years (fig. 3). The first year was very dry and the second had normal rainfall. In contrast, nearby Harrow Canyon--which burned in 1968--produced 438 m^3/ha of sediment as the result of a storm having a return period of about 35 years occurring the following winter. These interactions between fires and storms tend to cause erratic sediment delivery. Ninety-two percent of the sediment yielded by Bell IV during its 41-year record occurred in 5 years (fig. 3). The dependence on high flows for sediment transport makes fire and soil-water repellency important in chaparral drainage basins.

Sources of Sediment

Identifying the sources of sediment in the chaparral drainage basins is often difficult. During large storms, when most of the sediment is produced, events that take place late in the storm tend to obscure evidences of earlier erosion. The sources of sediment from Harrow Canyon durng the 1969 flood are an exception. As part of a study after the 1968 burn in Harrow Canyon and Englewild Canyon (Rice and Osborne 1970), I had extensively reconnoitered all of the canyons in the vicinity a few weeks after they had burned. Most of the channels were deeply filled with old alluvial and colluvial sediments. Channel-stabilizing check dams had been constructed on these deposits in some of the canyons. In Englewild Canyon, the pipeline of a private water development ran 600 m up one of the main forks lying on the bank of the creek. After the flood, practically all the channels were scoured to bedrock, the check dams had vanished, and remnants of the pipeline were found, in places, nearly 5 m above the new stream channel. Identifying sediment sources after the flood would have been difficult had it not been for the early reconnaissance and measurements of other erosional mechanisms (column 1, table 1).

Table 1--Sources of postfire sediment

Erosional processes	Harrow Canyon 1969 1/ Observed erosion	Harrow Canyon 1969 1/ Adjusted erosion	Estimated typical pattern
	Percent		
Wind	2/1	1	1
Dry ravel	3/1	20	40
Landslides	4/2	2	1
Surface erosion	3/22	27	18
Channel scour	5/74	50	40

1/Measurements and estimates cover the period from the fire (8/23/68) until the end of the rainy season (5/9/69).

2/Estimate .3 mm (based on deposits collected in Glendora).

3/Measured by 10 erosion plots (each .004 ha, Rice and Osborne 1970).

4/Mapped from a helicopter and converted to volume using equation in Rice et al. (1969).

5/Residual obtained by subtracting above estimates from the 438 m^3/ha measured in debris basin.

The Harrow Canyon measurements, however, probably underestimated dry ravel because most of it would have occurred before the installation of the plots. Surface erosion is also probably underestimated because of the length of the plots (14.4 m). Allowing for these assumed errors, half of the sediment produced in 1969 came from channel deposits (column 2, table 1). Prefire conditions in Harrow Canyon were unusual. Climatic and vegetative evidence indicated that the winters after the previous fire (more than 70 years earlier) were mild, permitting vegetation to become established on postfire alluvial and colluvial deposits. This vegetation was robust enough to stabilize the channel during the large floods of 1938 and 1943 when many southern-California channels were cleaned out. Apparently, Harrow Canyon channels were scoured in 1969 only because an exceptionally large flood occurred on a freshly burned drainage basin. More typically, erosion after a fire would come about 40 percent from the channel and an equal amount from dry ravel (column 3, table 1).

The previous discussion related to immediate origin of sediment in postfire runoff, not the original source of the sediment. Data on the relative magnitude of various sources of erosion on hillslopes is virtually nonexistent. I believe that about 80 percent of the sediment leaving chaparral drainage basins is initially eroded by gravitational processes of dry ravel and landslides. Channel scour of bedrock (as contrasted to remobilization of deposits) is a minor source of new sediment.

THE STOCHASTIC PROPERTIES OF SEDIMENT TRANSPORT IN CHAPARRAL DRAINAGE BASINS

Fire Occurrence

Wind and humidity are importance determinants of fire occurrence. Humidity determines whether ignition takes place, and wind is the primary meteorological control of the rate of spread and ultimate size of the fire.

Topography plays a role in determining the ultimate consequences of a fire. Fires starting low on the slopes usually have a better chance of spreading and inflicting drainage basin damage than those starting on ridges.

Another determinant of effects of fire is the condition of the vegetation (Rothermal and Philpot 1973). The age of vegetation affects both the fuel loading and the proportion of the stand that is dead. Age also affects seasonal changes in green-fuel moisture. These factors, in turn, interact with wind velocities to change the rate of spread. As the result of the interaction of these functions in a manner that is partially random, partially deterministic, a source of ignition may result in nothing--or in a conflagration. The chance of a conflagration increases markedly about 20 years after a fire because of accumulation of enough fuel to carry fire readily (Rothermal and Philpot 1973).

The Impact of Fire

What a fire does to the soil and vegetation depends both on the characteristics of the fire and the conditions of the vegetation and soil. Impact on soil centers mainly on whether or not a water-repellent layer is created. The degree of water repellency is determined by the temperature and duration of the fire and by the texture and water content of the surface soil (DeBano et al. 1970). A coarse-textured soil with a low moisture content which is subject to a hot (700°C) surface temperature for a relativey short period (10 to 15 minutes) is most likely to develop severe water repellency. The effect of a fire on vegetation depends on its age and vigor (older or weaker plants are less able to regenerate after fire). Less seed survives fires in older stands which tend to burn hotter than younger stands. Plant vigor depends both on the general health and the physiologial stress to which the plant is subject before the fire. Because of opposing seasonal trends of burning condition and plant vigor, the likelihood of a fire is positively correlated with the likelihood of damaging the chaparral and creating soil-water repellency.

The responses of several front-country drainage basins to the storms of 1969 indicate the magnitude of the effects of fire. Average peak discharge from 11 burned drainage basins was seven times greater than the average peak discharge from four drainage basins which had not burned for 9 years or more. Sediment yield from five unburned drainage basins increased by a factor of 14 and for two burned drainage basins by a factor of 120.

Storms

A storm, whatever its intensity, interacts with drainage basin vulnerability. Vulnerability of a drainage basin to excessive erosion and sedimentation is determined by its soil-water status, density of its vegetation and litter, the degree of soil-water repellancy, and recent storm history. The same storm can have quite a different effect on different drainage basins. For example, during the 1969 storm, a drainage basin that had not burned for 50 years produced landslide erosion at a rate of 16 m^3/ha. A similar drainage basin burned the previous summer produced 10 m^3/ha, and one burned 9 years before produced 298 m^3/ha. In the first drainage basin, good infiltration, and perhaps some decadence in the vegetation, may have led to reduced slope strength. The root system of the former vegetation in the freshly burned drainage basin had not decayed significantly, and the water-repellent layer created by the fire minimized percolation of water to potential failure surfaces. Most of the slides were associated with stream channels and probably triggered by undercutting. Although storm variability might account for the differences in landsliding in these two drainage basins, it is unlikely to be the explanation for the high erosion in the 9-year-old drainage basin. More likely, enough time had elapsed since its fire that infiltration had been restored sufficiently to allow saturation above potential failure surfaces. The regrowing vegetation, however, was not large enough for its root system to restore slope strength. It also could not transpire at a rate that depleted soil water to the same extent as mature chaparral.

Hyperscedasticity

I would like to propose the term "hyperscedasticity" to describe the sedimentation regime of chaparral drainage basins. It is a regime in which a sensitive terrane reacts to highly variable disturbances producing erratic outputs of sediment. The climate is highly variable, with extremes of rainfall and drought. Many slopes are steep and composed of deeply weathered parent material. Consequently, they are very sensitive to disturbances. Fire is a largely random component playing a prominent role in the environment. So prominent, in fact, that definition of the hydrologic and sedimentologic processes of chaparral drainage basins hinges on an understanding of the role of fire. In addition to the major random components, many minor ones are related to fire behavior, fire effects, vegetative recovery, seismic activity, and channel stability. The sediment regime is further complicated by the existence of thresholds in many important processes, most notably fire and mass wasting. The conditional and independent probabilities associated with these processes and their linkages have resulted in a sediment regime dominated by a few very large events, many insignificant ones, and low predictability.

MODELING, SEDIMENT BUDGETS, AND SEDIMENT ROUTING

Process Models

Process models, to even a greater extent than statistical models, are creatures of the computer age. As computers became larger and more economical to use, approximating erosional processes in the memory of a machine became possible. A three-dimensional array could describe the physiography of the drainage basin to whatever detail was deemed necessary. Based on physical laws or empiric data, rules could be incorporated governing how water or sediment moved from node to node within the array. Such models have proved useful, especially in well-behaved cases. They have the intuitive appeal of being similar in form to the prototype system.

Just how strong is the similarity between actual conditions and process modeling of erosion and sedimentation? From a practical point of view, processes must be approximated by lumping various components in time, space, and function. After these comparisons are made, the model is then usually "tuned" until it will reproduce a set of empiric data. Such a procedure invalidates the claim of generality. Proponents argue that process models are extrapolatable to different areas because they are based on physical relationships, and physics does not change with time and distance. This claim of robustness has been compromised by the approximations and by incorporating calibration procedures into process-model building.

The calibration of process models carries another risk. Because their data requirements are exacting, the available suitable data sets tend to be small. When a small sample is drawn from a time series that produces significant effects only infrequently, the sample is likely not to include such an event. This attribute of the calibration of process models tends to cause underestimation of the variance of the dependent variable, which is particularly undesirable because accurate reproduction of large events is critical to model utility. If, on the other hand, the sample happens to include a large event, the large variance computed may be misinterpreted as "typical."

Empiric Models

Multiple regression is the principal empiric model used. Certainly, it has the longest history as a predictive tool in hydrology and sedimentation. This discussion will, in the main, be relevant to other statistical procedures as well. When using such models, the hydrologist breathes a prayer: "Dear Lord, let the world be linear and Gaussian." He desires linearity so that the structural form of the analysis is correct and normality so that he can estimate the precision of coefficients. The prayer is rarely answered. Nonetheless, empiricists are forced to jam the world into the regression straightjacket and hope that--even if their prayer is not answered--the central-limit theorem will come to their rescue. To make that rescue more likely, they are tempted to include more and more observations of poorer and poorer quality in their analyses so that $\underline{n}$ may more closely approach infinity. Most modern computer utilities offer programs that assist the analyst in selecting an optimum set of independent variables. But, no resort to "heroic statistics ... offers a real substitute for solid data" (Philip 1975).

Assuming that the previously mentioned problems have been more or less satisfactorily resolved, the empiricist still has only what could be charitably termed a pseudophysical model. His partial-regression coefficients express the relation between each independent and the dependent variable. The "independent" variables, however, are rarely independent. Some are correlated (perhaps an artifact of the data base), and others have functional relationships (such as soil with slope, aspect, and elevation). This means that the regression coefficients are only hinting at a functional relationship. To their credit, regression analyses facilitate the consideration of a much larger set of casual variables than would be practical for most process models. They also may have a slight advantage over process models in that if extreme events are included in the data set, the least-squares analysis will tend to cause them to dominate the form of the resulting regression equation. They share the same sampling problems that cause process models to be unlikely to include important events, however.

Monte Carlo Models

Models based on Monte Carlo simulation may incorporate desirable features of both process models and empiric models. Similar to process models, a Monte Carlo model would form a chain of conditions or processes that feed one another. The chain would begin with meteorological inputs and terminate with the delivery of sediment. Unlike in a process model, the nodes in the model would not have a spatial relation to one another and would not form a grid in two or three dimensions. Rather, the nodes would define a series of relations linking inputs to the output of sediment. Both the mean and variance of a function would be defined at most nodes. In this fashion, both the estimate of the presumed relation and the uncertainty associated with that estimate enter into the simulation. Although conceiving these nodes as bivariate relations is easiest, no inherent reason prevents nodes representing multivariate relations.

Each time the flow of the program reaches one of the nodes, it branches to a subroutine that generates random numbers. The number so acquired is then applied to the probability distribution of that function to determine the value of the function to be used in that simulation. The uncertainty related to various functional relationships is thus retained explicitly in a model, which is otherwise deterministic. The model could be cycled many times in the analysis of any particular set of circumstances so as to yield a spectrum of possible outputs. The analyst would have an estimate both of the expected response to a particular situation and of the probability of very beneficial or very disastrous consequences. Management decisions might hinge more on the extreme outcomes than upon the expected outcomes.

A much-simplified case will be used to give a little further appreciation for the functioning of a Monte Carlo model. It treats mass wasting as a function of storm size and fire history. Sedimentation is handled by a delivery ratio, and none of the interactions among sediment processes is considered. The greatest simplification, however, is in treating the process as an annual series rather than considering within-year temporal variation in many of the functions.

In the example, the objective is to estimate mean long-term production of sediment from mass movements. A further assumption is that the functions shown in figure 4 have a sound basis in theory and also in empiric data. The first step in the simulation is to estimate the size of the largest storm of the season by sampling from the probability distribution of maximum annual storm precipitation (fig. 4a). If the storm size obtained from the samping is less than some critical value (x), the remainder of the simulation is irrelevant and can be bypassed because the storm is so small that the probability of landslides is essentially zero regardless of vegetative condition. In this case, the occurrence of the trial would be recorded and a new drawing made from the storm size distribution.

If the drawing from the storm size distribution results in the storm larger than the threshold size, the remainder of the simulation comes into play. The next random drawing is from the population distribution of possible ages of chaparral (fig. 4b). Vegetation age--more correctly, its size and vigor--largely determines drainage-basin response to the storm.

The age is related to resistance to landslides in figure 4c. Resistance is highest immediately after a fire because the root system is still intact and reduced infiltration is likely because of water repellency. Resistance declines in the next few years as infiltration is restored and root decay sets in. Then, after about 8 years, the slope becomes more resistant as roots develop on sprouting plants and new seedlings. Between 25 and 30 years, a slight decline takes place as the pioneer species such as Ceanothus are crowded out by more long-lived species. The decline soon levels out as the stand enters senescence, with a gradual turnover of plants as individuals die and are replaced. The curve in figure 4c is patterned after root biomass of redwood (Sequoia sempervirens (D. Don) Endl.) in northwestern California (Ziemer, R. R., 1700 Bayview Drive, Arcata, California, personal communication). It is also similar to curves developed for the uprooting resistance of trees in Japan (Kitamura and Namba 1966). I have assumed that the variance of this trend in landslide resistance would increase markedly after a fire and then decline. The basis of that assumption is that, in the early years, recovery would be dependent on postfire weather, the age of vegetation at the time of the previous burn, and the severity of the previous burn. As the stand gets older, it comes to express the ecological potential of the site, regardless of the point at which it began its recovery. Consequently, the variance in very old stands becomes rather small.

From the resistance obtained from the relationship of figure 4c, the expected volume of erosion from an average landslide-producing storm is computed using a function relating landslide volume to resistance (fig. 4d). This mean response is then increased or decreased according to storm size, using the relationships in figure 4e. In this case, the variance results from the fact that storms that are nominally identical, in terms of figure 4a, actually have different capacity to produce landslides because of the distribution of rainfall intensity within the storm. Between-storm variability in erosion is assumed to decline for larger storms.

The last step in this model is to estimate sedimentation using a delivery ratio (fig. 4f). The mean response assumes debris resulting from the landslides during small storms will have low fluidity and, consequently, low delivery ratios. As the debris increases, the delivery ratio increases to a maximum and then declines as landslide debris becomes so prevalent that it greatly exceeds the ability of the stream to carry it from the drainage basin. The variance responds to those same factors. In small storms, the delivery ratio depends on the location of slides throughout the drainage basin and is, therefore, highly variable. As storm size increases, delivery to the stream increases and the variance declines. With large amounts of debris, the system becomes stream-power dependent and the variance is correspondingly reduced.

The model could be run for a specified number of cycles or halted when estimates of the mean and variance of sediment production had stabilized. In addition to generating information about the distribution of possible sediment outputs, a Monte Carlo model could yield data concerning the stochastic properties of the time series of sediment outputs. These possibilities make the application of Monte Carlo simulations to sediment outputs worthy of investigation. Bear in mind, however, that although sampling from probabililty distributions is frequent in the use of Monte Carlo models, the accuracy and precision of the estimates of sediment production are unknown. Even if all the variates used in their development were normally distributed, no general or exact way exists for computing the parameters of the output distribution based on the estimates of the various inputs. Usually a Monte Carlo model will rest on a more uncertain foundation. Some functions will be well defined by empiric data (figs. 4a, 4b), others will be built of sketchy information and extrapolation (fig. 4c), and still others will reflect only theory or hypothesis (figs. 4d, 4e, 4f). Therefore, in spite of the intuitive appeal of Monte Carlo models, their results have to be accepted on faith or calibrated with empiric data.

Some statistical problems are relevant to all types of models: process, empiric, or Monte Carlo. Paramount among them is the hyperscedastic

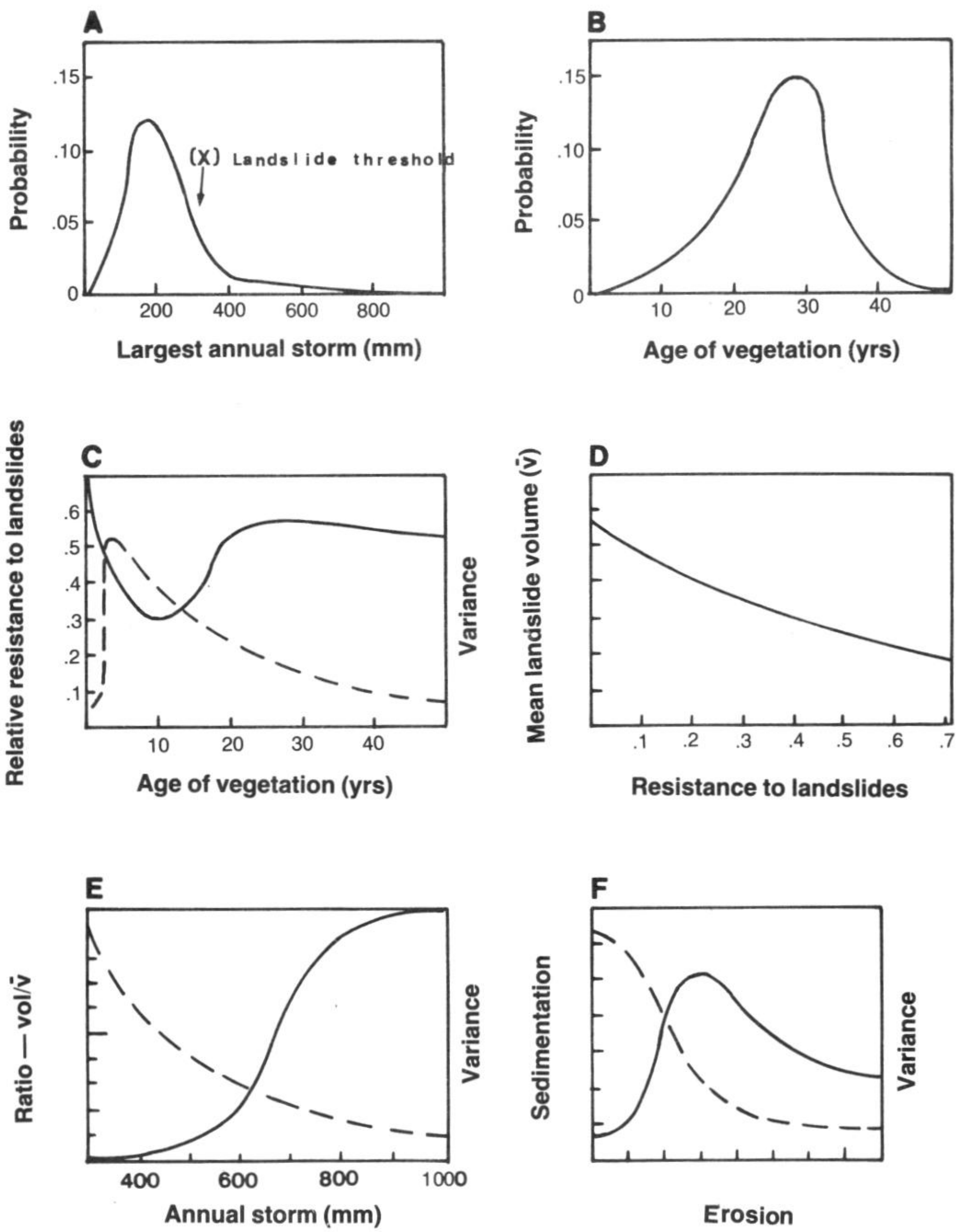

Figure 4.--Functions (solid lines) and their variances (broken lines) used to sketch the structure of a simulation of landslide erosion in Chamise chaparral.

nature of most of the phenomena, especially the outputs. This has the effect of making the relevant data base extremely small. The information from the Bell IV drainage basin probably represents something like 4.5 "years" of record for use in building a sedimentation model--even though it took 41 years to collect. This small, relevant data base places a heavy burden on modelers; they have been able to observe very few of the possible interactions of a complex process.

Because of the hyperscedasticity of the phenomena and the way data are collected, the robustness of the coefficients should be doubted: To what extent do they estimate their parametric values, and to what extent are they the result of spurious correlations in the developmental data? And what about the developmental data; are they from a stationary time series? Are current estimates of parameters (even if accurate) valid estimates of their future values? Models are often constructed because changes are planned for a drainage basin. Normally, we had no a priori way of knowing how well models will hold up under changed conditions. Are the processes defined in sufficient detail? These questions, too, will have to be resolved on faith. Each model builder must wield Occam's razor, attempting to include all the complexity necessary for accurate prediction, but nothing more.

The small, relevant data base for modeling hyperscedastic processes presents a paradox: Good models require a large amount of good data for their development, but the lack of a long and abundant data stream is just what makes necessary the development of models of erosion, sediment transport, and other hydrologic phenomena. Clearly we are in a "trans-scientific" environment. Weinburg (1972) has introduced this adjective "trans-scientific" to describe "questions which can be asked of science and yet cannot be answered by science." Although posing such questions as rigorously testable hypotheses is possible, conducting the research necessary to answer them is impractical (if not impossible). In response to this dilemma, "It remains our obligation to insure that our methods are as scientific and objective as possible. Let us at least work towards a situation where the trans-scientific judgements which practical hydrologists are forced to make are informed and sustained by a truly scientific hydrology: a skeptical science with a coherent intellectual content firmly based on real phenomena (Philip 1975)."

Process models, because of their data requirements, will ususally have applicability only to the management of drainage basins where high values are at stake--values which can justify the expenditure of the resources required for their use. In most circumstances, to quote from Philip (1975), "The task of performing an accurate detailed physical characterization and using this in a reliable predictive calculation is scientifically feasible; but would almost always demand for its proper performance, the expenditure of resources out of all proportion to the benefits." Empiric models are at the other extreme. Although requiring only a modest expenditure of resources, their rigid structure dictates that they lack a firm basis in real phenomena.

Monte Carlo models may offer a feasible solution to some trans-scientific hydrologic problems. Their resource requirements are intermediate between process models and empiric models. Like empiric models, they consider random variation explicitly. In common with processed models, they permit an amalgam of empiric information, theory, and (if necessary) conjecture. Hopefully, the result can satisfy Philip's (1975) desire for "a coherent intellectual content firmly based on real phenomenon."

LITERATURE CITED

Anderson, H. W., G. B. Coleman, and P. J. Zinke. 1959. Summer slides and winter scour--wet-dry erosion in southern California mountains. USDA For. Serv. Tech. Pap. 36, 12 p. Pac. Southwest For. and Range Exp. Stn., Berkeley, Calif.

Aschmann, H. 1977. Aboriginal use of fire. In Proceedings of the symposium on environmental consequences of fire and fuel management in mediterranean ecosystems. USDA For. Serv. Gen. Tech. Rep. WO-3, p. 132-141. Washington, D.C.

Bailey, R. G. 1967. Soil slips on the San Dimas Experimental Forest. M. A. thesis. San Fernando Val. State Coll., Northridge, Calif. 83 p.

Bailey, R. G., and R. M. Rice. 1969. Soil slippage: an indicator of slope instability on chaparral watersheds of southern California. Prof. Geogr. 21:172-177.

Byrne, R., J. Michaelson, and A. Soutar. 1977. Fossil charcoal as a measure of wildfire frequency in southern California: a preliminary analysis. In Proceedings of the symposium on environmental consequences of fire and fuel management in mediterranean ecosystems. USDA For. Serv. Gen. Tech. Rep. WO-3, p. 361-367. Washington, D.C.

California Division of Mines and Geology. 1969. Geologic map of California: Los Angeles sheet.

Campbell, R. H. 1975. Soils slips, debris flows, and rainstorms in the Santa Monica mountains and vicinity, southern California. U.S. Geol. Surv. Prof. Pap. 851, 51 p. Washington, D.C.

Colman, E. A. 1953. Fire and water in southern California's mountains. USDA For. Serv. Pap. 3, 7 p. Pac. Southwest For. and Range Exp. Stn., Berkeley, Calif.

DeBano, L. F.
1969. The relationship between heat treatment and water repellency in soils. In Proceedings of the symposium on water-repellent soils, p. 265-279. Riverside, Calif.

DeBano, L. F., L. D. Mann, and D. A. Hamilton.
1970. Translocation of hydrophobic substances into soil by burning organic litter. Soil Sci. Soc. Am. Proc. 34(1):130-133.

Keeley, J. E.
1977. Fire-dependent reproductive strategies in Arctostophylos and Ceanothus. In Proceedings of the symposium on environmental consequences of fire and fuel management in mediterranean ecosystems. USDA For. Serv. Gen. Tech. Rep. WO-3, p. 391-396. Washington, D.C.

Kitamura, Y., and S. Namba.
1966. A field experiment on the uprooting resistance of tree roots. Proc. 77th Meeting Jap. For. Soc., p. 568-570. Tokyo, Megro.

Kittredge, J., Jr.
1939. The annual accumulation and creep of litter and other surface materials in the chaparral of the San Gabriel Mountains, Calif. J. Agric. Res. 58(7):537-541.

Krammes, J. S.
1965. Seasonal debris movement from steep mountainside slopes in southern California. Proc. of the Fed. Inter-Agency Sediment. Conf., Jackson, Miss. USDA Misc. Publ. 970, p. 85-88. Washington, D.C.

Krammes, J. S.
1967. Exploratory investigation of the permeability of granitic material in the San Dimas Experimental Forest. Ph. D. thesis. Oreg. State Univ., Corvallis.

Krammes, J. S., and J. F. Osborne.
1969. Water repellent soils and wetting agents as factors influencing erosion. In Proceedings of the symposium on water-repellent soils, p. 177-187. Riverside, Calif.

Philip, J. R.
1975. Some remarks on science and catchment prediction. In Prediction in catchment hydrology. Austral. Acad. Sci., Canberra.

Reimann, L. F., and E. L. Hamilton.
1959. Four hundred and sixty storms. USDA For. Serv. Misc. Pap. 37, 101 p. Pac. Southwest For. and Range Exp. Stn., Berkeley, Calif.

Rice, R. M., E. S. Corbett, and R. G. Bailey.
1969. Soil slips related to vegetation, topograpy, and soil in southern California. Water Resour. Res. 5(3):647-659.

Rice, R. M., and G. T. Foggin III.
1971. Effect of high intensity storms on soil slippage on mountainous watersheds in southern California. Water Resour. Res. 7(6):1485-1496.

Rice, R. M., and J. F. Osborne.
1970. Wetting agent fails to curb erosion from burned watersheds. USDA For. Serv. Res. Note PSW-219. Pac. Southwest For. and Range Exp. Stn., Berkeley, Calif.

Rothermal, R. C., and C. W. Philpot.
1973. Predicting changes in chaparral flammability. J. For. 71(10):640-643.

Scott, K. M., and R. P. Williams.
1974. Erosion and sediment yields in mountain watersheds of the Transverse Ranges, Ventura and Los Angeles Counties, California-- analysis of rates and processes. U.S. Geol. Surv., Water Resour. Invest. 47-73, 128 p. Washington, D.C.

Troxell, H. C.
1953. The influence of ground-water storage on the runoff in the San Bernardino and eastern San Gabriel Mountains of southern California. Trans. Am. Geophys. Union 34(4):552-562.

Weinburg, A. M.
1972. Science and trans-science. In Civilization and science, p. 105-122. Elsevier, Amsterdam.

Use of a Physical Model to Determine Effects of Periodic Erosion in Steep Terrain on Sediment Characteristics and Loads

Michael D. Harvey

ABSTRACT

A "similarity-of-process" (Hooke 1968) model was constructed in the Rainfall Erosion Facility (R.E.F.) at Colorado State University, Ft. Collins, Colorado. Mass slope failure and the character of resulting sediments were investigated. Total sediment yields immediately after slope failure were dominated by the suspended load. Through time, sediment yields diminished but then increased again and were dominated by bed-material load. Total sediment loads were subdivided into the components of traction, saltation, and suspension load using the methods proposed by Visher (1969) and Middleton (1976). During a single storm event in which water discharge varied from 0.3 to 2.8 liter/second, sediment discharge varied by an order of magnitude and exhibited hysteresis. Suspended loads peaked on the rising limb, and traction loads peaked on the falling limb of the hydrograph. Sediment discharge behavior observed in the R.E.F. is compared to behavior demonstrated by data collected in the field in both New Zealand and Colorado.

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INTRODUCTION

Hardware modeling in geomorphology may be defined as the study, under closely monitored or controlled experimental conditions, of a physical representation of a selected geomorphic feature (Mosley and Zimpfer 1978). Hooke (1968) considered that a model could be treated as a small prototype in its own right--not representing a specific location, but rather the general features of many locations. He referred to this approach as "similarity of process." He considered that gross scaling relations must be met, the model must reproduce some morphologic characteristic of the prototype, and the processes producing this characteristic in the model must have the same effect in the prototype. Using this approach, the generally recognized disadvantages of physical models--scaling ratios, initial conditions, and boundary conditions--can be countered. Physical modeling, although used predictively and descriptively, is inherently incapable of providing the last link in the chain of scientific investigation--the theory or law--because the results can only be qualitatively and cautiously applied to the real world (Mosley and Zimpfer 1978). The advantages and disadvantages of hardware modeling in geomorphology have been summarized by Mosley and Zimpfer (1978). The use of models can be justified in the generation and testing of hypotheses that can then be investigated in the field.

The interactions between slopes and channels as a result of low-frequency, high-magnitude catastrophic events are difficult to measure in the field. An experiment was therefore conducted in the R.E.F. to test the following hypotheses: After mass slope failure, sediment yield from a drainage basin is dominated initially by suspended load; through time, sediment yield is dominated by bed-material load; the coarser fractions move downstream in waves or pulses; and the changing character of the sediments results in changes in downstream channel morphology. This paper presents the results of sedimentologic investigations of these hypotheses and reviews some of the pertinent literature.

Adams (1978) reported on the effects of the Inangahua earthquake of 1967, which delivered $52 \times 10^6\ m^3$ of material to the Buller River on the west coast of the South Island of New Zealand; he considered that 95 percent of the debris introduced to the channel was transformed rapidly into suspended load, and the remaining 5 percent remained in the channel and was transported as bedload. Pain and Bowler (1973) concluded that about half of the $27 \times 10^6\ m^3$ of sediment delivered to the channel as a result of the Madang earthquake in New Guinea reached the sea in 6 months. Both authors consider that suspended loads were double those of the preslope failure period. On the other hand, Hayward (1978), reporting on the Torlesse Stream in the South Island of New Zealand, concluded that, during 1974-77, suspended sediments contributed less than 10 percent of the total annual yield of sediment and that the bulk of the sediment transported is moved as bedload derived from in-channel storage and bank failures. The stored sediments represent the coarser fractions of the debris introduced to the channel system by catastrophic slope failures as the result of a severe thunderstorm in 1957. The work cited probably encompasses the variations in proportion of suspended load to bedload through time after catastrophic slope failures.

EXPERIMENTAL METHODS

Using Hooke's (1968) similarity-of-process approach, a drainage basin of 5 x 10 m dimensions was constructed in the R.E.F. The basin was connected by a hydraulically efficient conduit (Z) to a 12 x 2.2 m straight channel (main), which was intended to simulate a downstream, nonheadwaters channel reach (fig. 1). The morphometric characteristics of the drainage basin and main channel are summarized in table 1. During the construction of the divides, a subsurface layer was compacted to provide a failure surface to facilitate mass movements of the slope materials. The material used in slope construction was moderately cohesive and ranged in size from -3 to 5 (Folk 1965); the material was considered by Mosley (1975) to allow accurate reproduction of most aspects of natural river morphology. The grain-size distribution curve for "parent material" is shown in figure 2. Using Shields' (1936) approach, the channels were determined to be able to transport all sediment size classes in the basin. Precipitation was only applied to the drainage basin from a set of overhead sprinklers. Rates could be varied as required, and distribution over the basin was reasonably uniform.

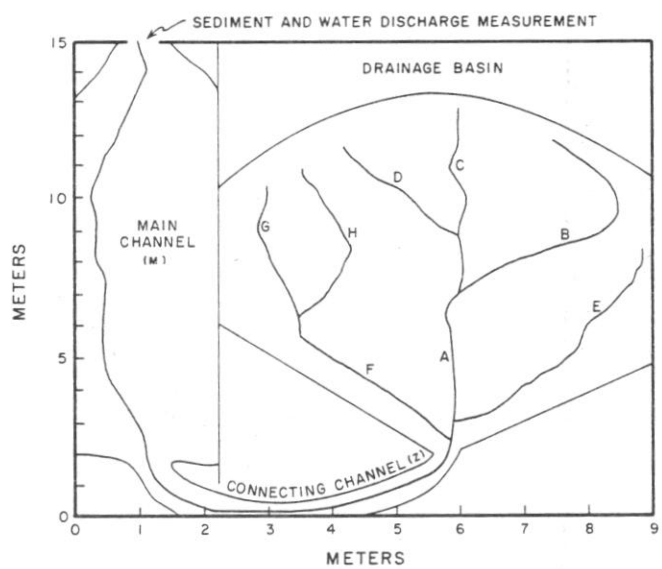


Figure 1.--Plan view of R.E.F. showing drainage basin, connecting channel (Z), main channel (M), and measuring point.

At the end of the main channel, total discharge of water and sediment was measured at closely spaced intervals during the course of each run, which was intended to simulate a storm event. Initially, low-intensity precipitation was applied to the drainage basin to simulate low-flow conditions. A 5-minute burst of high-intensity precipitation was then applied to the drainage basin to simulate a

Table 1--Morphometry of R.E.F. model

Drainage basin or channel	Initial channel slope	Valley slope ranges	Area	Basin shape	Relief ratio	Drainage density	Ruggedness
		Degrees	Square meters				
Entire drainage basin	--	22-44	51.97	0.499	0.066	0.699	0.461
A	0.0858	21-43	27.63	0.257	0.064	0.664	0.441
F	0.0771	25-44	14.63	0.266	0.093	0.844	0.581
E	0.0769	20-44	9.84	0.271	0.084	0.574	0.290
Main (M)	0.0045	--	--	--	--	--	--
Connecting (Z)	0.0038	--	--	--	--	--	--

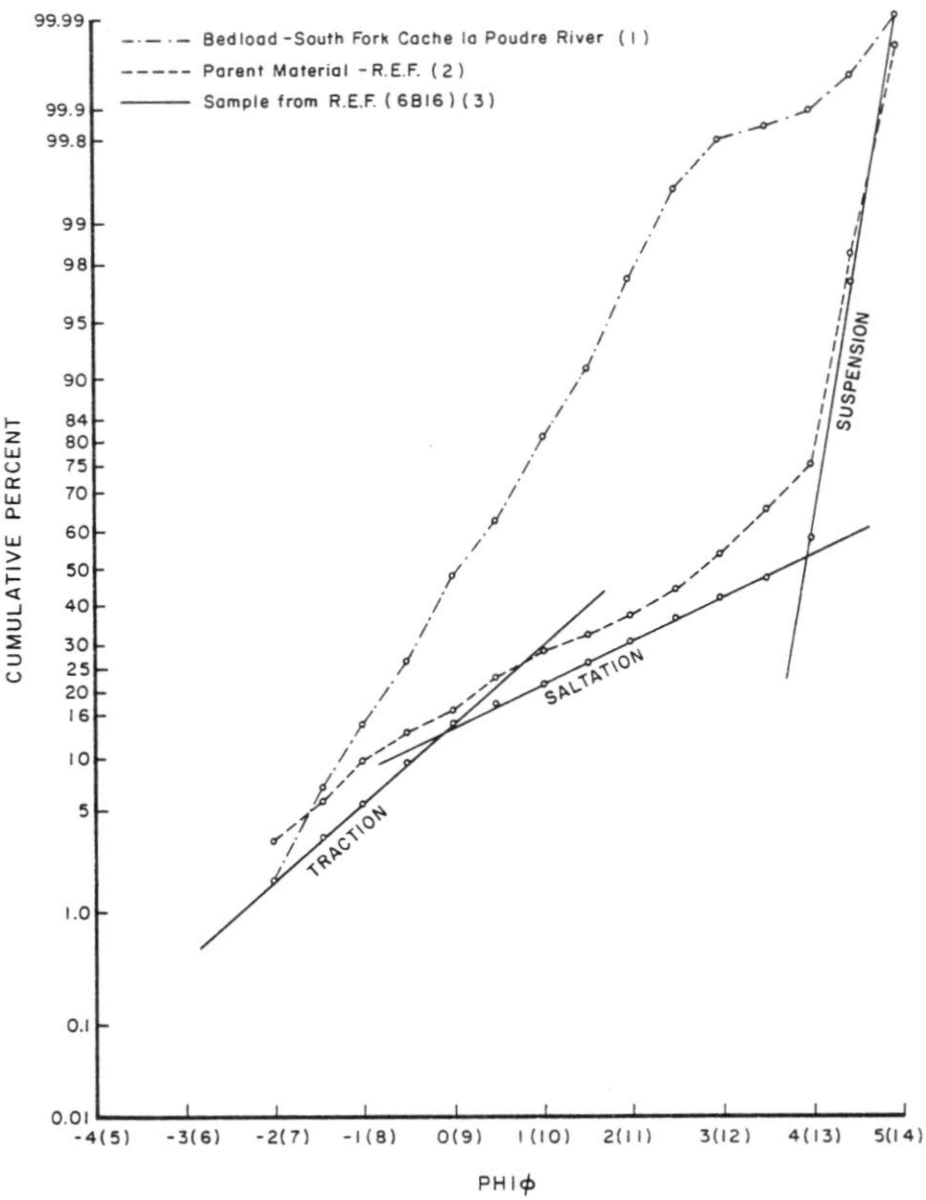
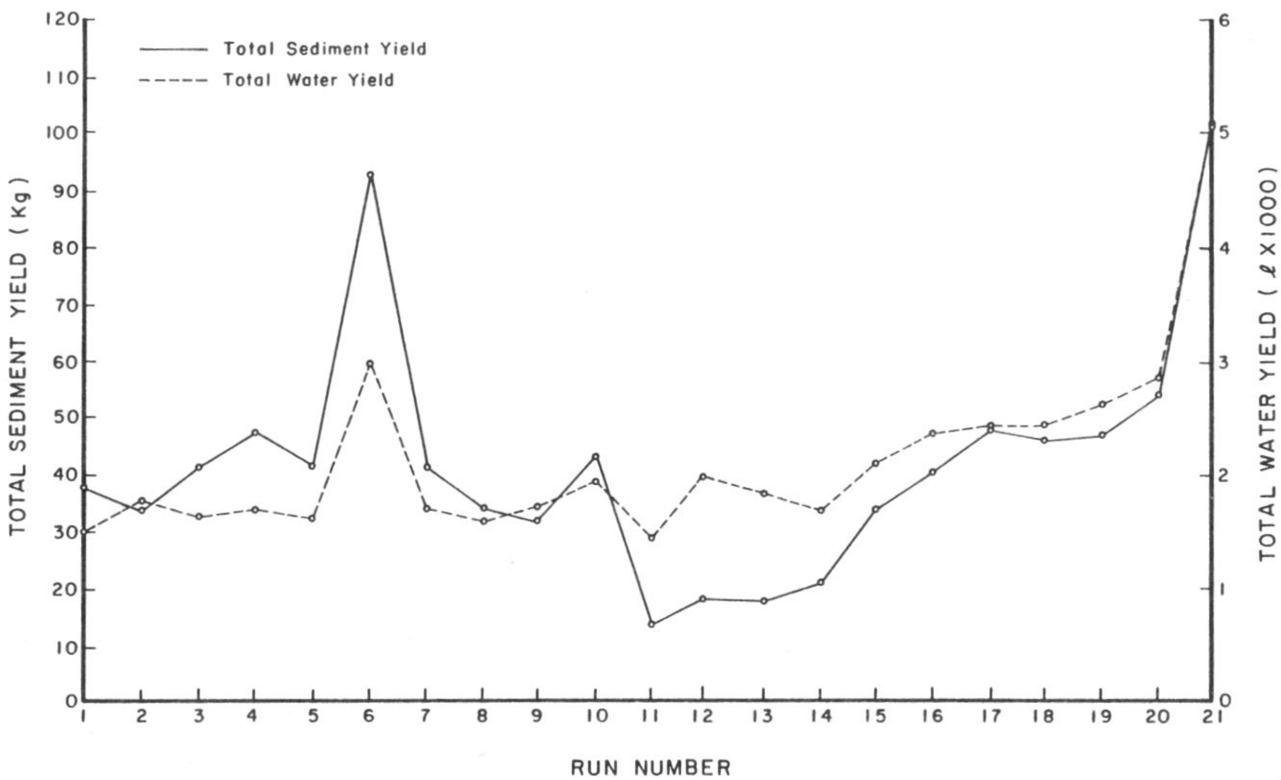


Figure 2.--Log probability plots of: (1) bedload sample from the South Fork of the Cache la Poudre River, Colorado; (2) "parent material" used in R.E.F.; (3) sample 6B16 from the R.E.F.



major storm. After the 5-minute burst, precipitation intensity was again reduced until steady discharge conditions were attained, and the system was then shut down. Sediment samples were ovendried and then sieved according to standard techniques. Grain-size statistics were obtained using the methods determined by Folk (1965).

Sediment was delivered from slopes to channels by three principle means: slope wash, debris slides, and raindrop splash. The debris-slide sites generally developed into gullies with time. Lateral migration of channels caused the undercutting of slopes, which also resulted in slope failures.

To understand the sedimentology of the system, we need to know what fraction of the total sediment yield in each run was transported by the three transport mechanisms of traction, saltation, and suspension. The scale of the operation prevented direct measurement of modes of transport, and therefore the method proposed by Visher (1969) and Middleton (1976) for subdividing the total load into the transport components was investigated. The method was tested using grain-size analyses of bedload collected with a Helley-Smith sampler in the South Fork of the Cache la Poudre River, Colorado. If the method is valid, bedload data should plot as a single straight line on log probability paper. An example of the results is shown in figure 2. Statistical analysis of the bedload data indicated that a single population was being investigated and provided further support for the method. On the basis of the results from the bedload data, I decided to use the method to subdivide the total sediment samples obtained in the R.E.F. Grain-size data from 100 R.E.F. samples were plotted on log probability paper. An example is provided in figure 2.

Figure 3.--Total sediment and water yields for 21 runs in the R.E.F.

RESULTS

Sediment yields (fig. 3) and transport rates for 21 runs in the R.E.F. were calculated. Runs 1 to 3 probably represent the response of the system to the effects of model construction. During Run 6, mass movement failures occurred in the drainage basin. Total sediment yield for the run was 93.5 kg, and suspended sediment (size range of 4 and less) constituted 62.6 percent of the yield. Runs 7 through 11 showed a general decrease in total sediment yield per run (41.7 kg to 13.4 kg), but the yields were still dominated by suspended sediments (58.6 percent to 59.7 percent). The decrease in total sediment yields per run can be attributed to a reduction in supply of sediment from the mass-movement detritus. The suspended sediments were rapidly transported, but the coarser fractions were stored in channels as evidenced by an overall increase in mean bed-elevations of all the headwater channels. Total sediment yields per run increased from Run 12 through Run 19 (18.2 kg to 46.9 kg). The contribution of suspended sediment to total yields fell from 53.5 percent to 43.9 percent. The increase in total sediment yields and the reduction in the proportion of suspended sediment can be attributed to the mobilization of coarser fractions stored in the channel, as demonstrated by the overall reduction in mean bed-elevation of headwater channels and the construction of fans in the main channel. Small mass-movement failures occurred in the drainage basin during Run 20, and this is expressed by an increase in total sediment yield (54.0 kg) but, more importantly, in the proportion of suspended sediment (50.6 percent).

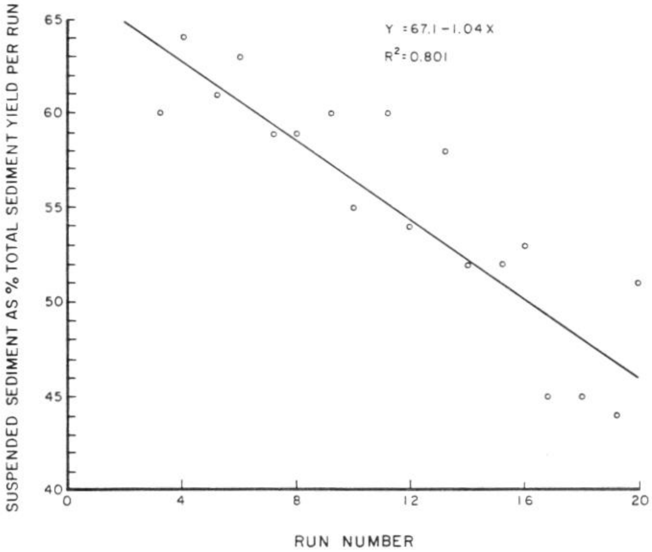
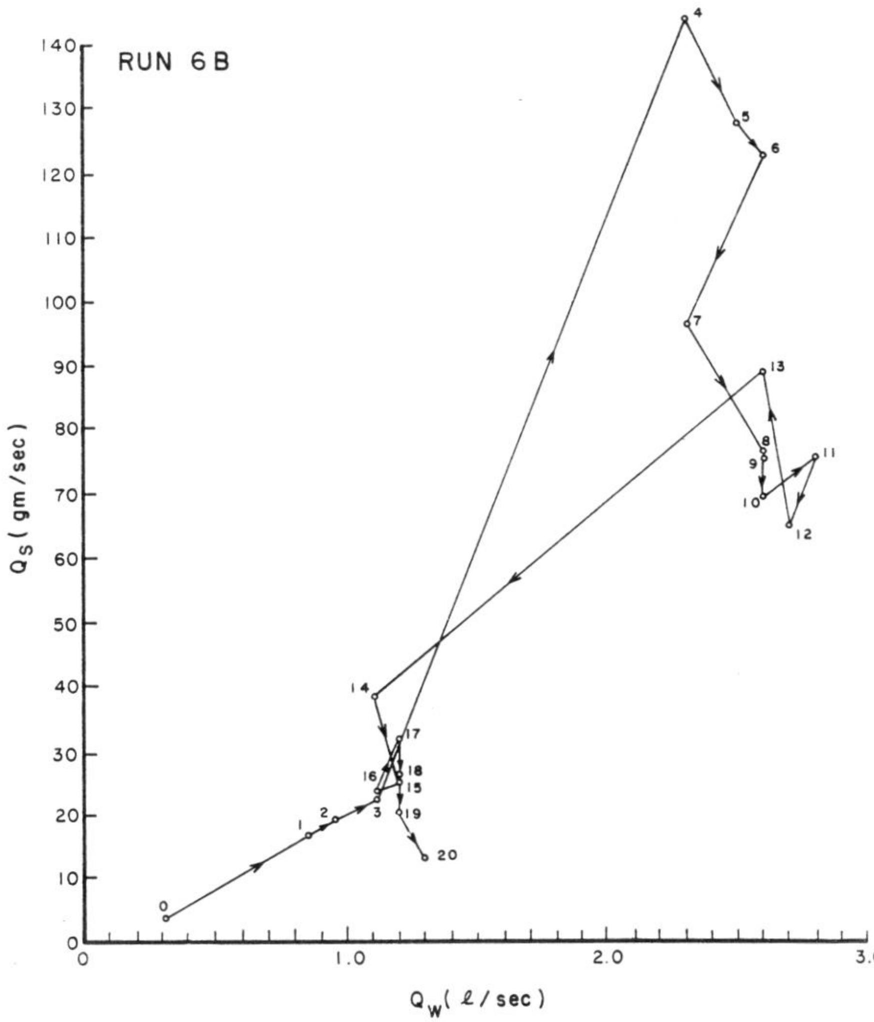


Figure 4.--Suspended sediment as percent of total sediment yield per run versus run number.

Figure 4 shows the relation of the proportion of total sediment yield delivered as suspended sediment (Y) to run number (X), described by the equation: $Y = 67.1 - 1.04X$ ($R^2 = 0.8$)(= 0.001).

Sediment-transport rates (Q_s) during the course of individual runs did not increase linearly with water-discharge rates (Q_w). Figure 5 summarizes the discharge data for Run 6, which is presented as a general example. The discharge rates are characterized by a series of hysteresis loops. Gregory and Walling (1973) report similar behavior for suspended sediments. Transport rates were from 5 to 20 g/second (points 0-3) at Q_w values up to 1.3 liter/second. As Q_w increased to 2.3 liter/second, Q_s increased to 140 g/second (point 4), and sediment type was dominated by suspended sediment. Transport rates fell to 68 g/second (point 10) while Q_w remained essentially constant. As Q_w increased to a maximum of 2.8 liter/second, Q_s increased to 76 g/second (point 11) and was the result of an increase in the amount of bed-material load being transported. On the waning stage, Q_s increased to 88 g/second (point 13) before falling substantially to 37 g/second (point 14). The increase was from increased transport of the coarsest fractions. Figure 6 presents the changes in grain-size distribution during the course of Run 6.

Figure 5.--Hysteresis loops of total sediment and water discharge for Run 6.



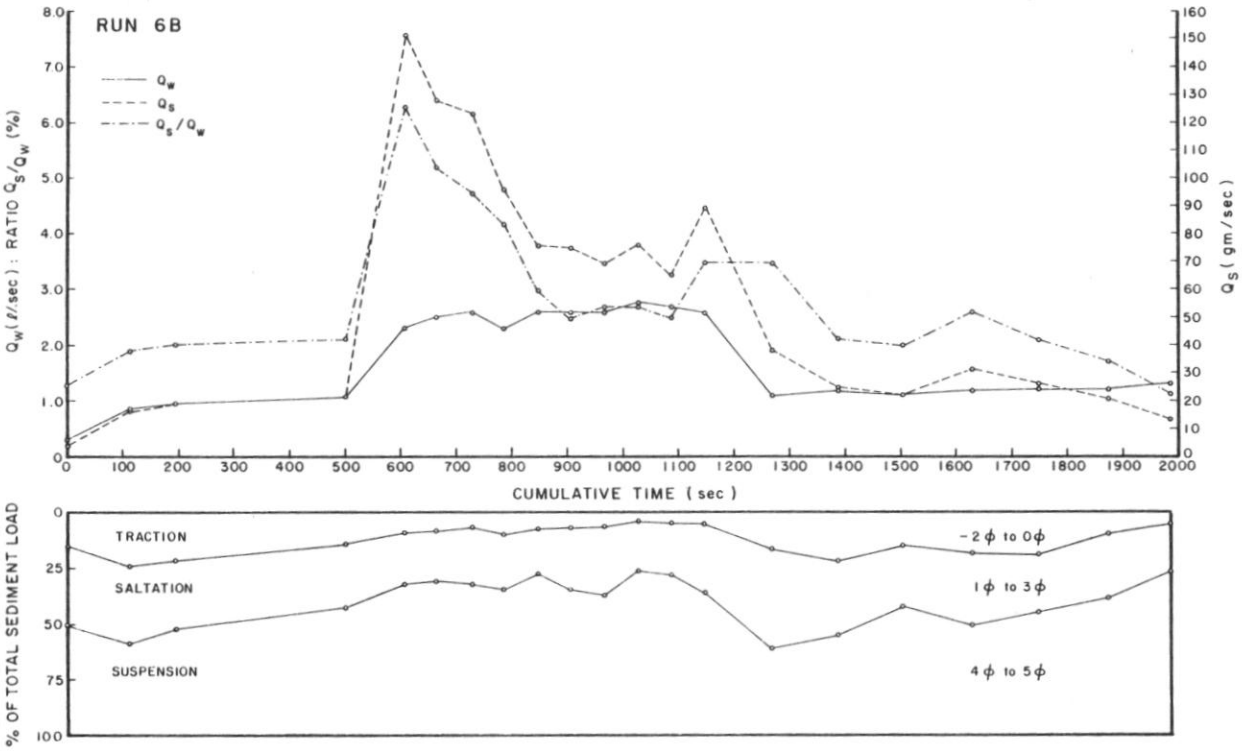


Figure 6.--Summary of discharge data and sedimentology, Run 6.

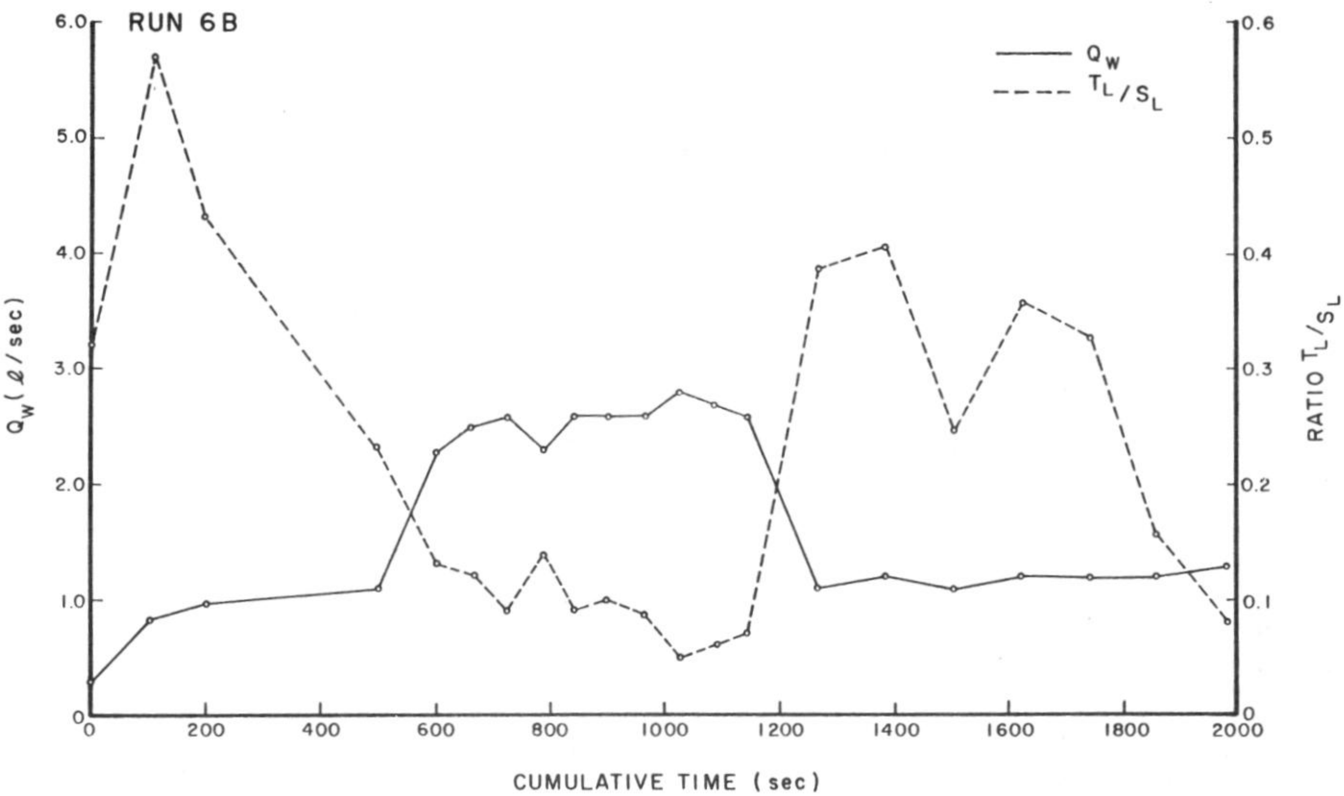


Figure 7.--Water discharge and ratio of traction load to suspended load (T_L/S_L) for Run 6.

In small drainage basins sculptured by fluvial processes, suspended sediment loads tend to peak before the peak of the hydrograph (Porterfield 1972, O'Loughlin et al. 1978). This phenomenon was observed in the R.E.F. during many runs, but the peaks did coincide in some runs--especially the earlier ones where large volumes of material in the suspended-load size range (4 and less) were available for transport. In contrast to the suspended load, the coarser fractions (-2 to 3) peaked after the peak of the hydrograph. W. W. Emmett (personal communication, U.S. Geological Survey, Denver, Colorado) has noted the same phenomenon in the East Fork River in Wyoming.

These phenomena may be attributed to three possible causes. Heidel (1956) established that the peak of sediment concentration progressively lagged behind the flood peak, and this may be what occurred in the R.E.F. Second--at the peak of the discharge--large volumes of sediment are being transported, which can lead to localized aggradation of some channel reaches. Reduction in sediment delivery from the drainage basin on the waning stage of the hydrograph leads to scouring of the locally aggraded reaches, thereby increasing the amount of material available for transport. Third, sediment supply falls off more rapidly than the water discharge. One feature in all runs was that with the low water discharges on the rising stage, a relatively high proportion of coarser material occurred in the sediment samples. This is attributed to continuation of scour of locally aggraded reaches before sufficient sediment is moved out of the drainage basin to prevent it. The relative contributions of traction load and suspended load over the course of Run 6 are presented in figure 7, in which the ratio of traction load to suspended load is plotted with the hydrograph.

Hayward (1978) showed similar trends of early and late coarse material discharge during the course of a storm in the Torlesse Stream, South Island, New Zealand (fig. 8). Bedload data collected in the South Fork of the Cache la Poudre River, Colorado (Doehring and Ethridge 1979) showed similar trends (fig. 9).

SUMMARY AND CONCLUSIONS

Sediment yields and transport rates resulting from periodic mass movements were measured for 21 runs in the R.E.F. Slope failures first occurred in Run 6 and were reflected in the total sediment yield (93.5 kg), but more importantly in the composition of the total yield (62.2 percent suspended sediment). Total sediment yield per run fell through Run 11 (13.4 kg), but suspended load was still the dominant component of total load. Reduction in total sediment yields was accompanied by increases in mean bed-elevation of all headwater channels where mass movements had occurred. In Runs 12-19, total sediment yields per run again increased (46.9 kg), but bed-material load became the dominant component of total load (56.1 percent). The change in composition of the total yield was accompanied by reduction in mean bed-elevation of headwater channels and the formation of fans in the main channel. During Run 20, small mass movements occurred in the drainage basin and were expressed as an increase in total sediment yield (54.0 kg), but more importantly by an increase in the proportion of suspended sediment (50.6 percent).

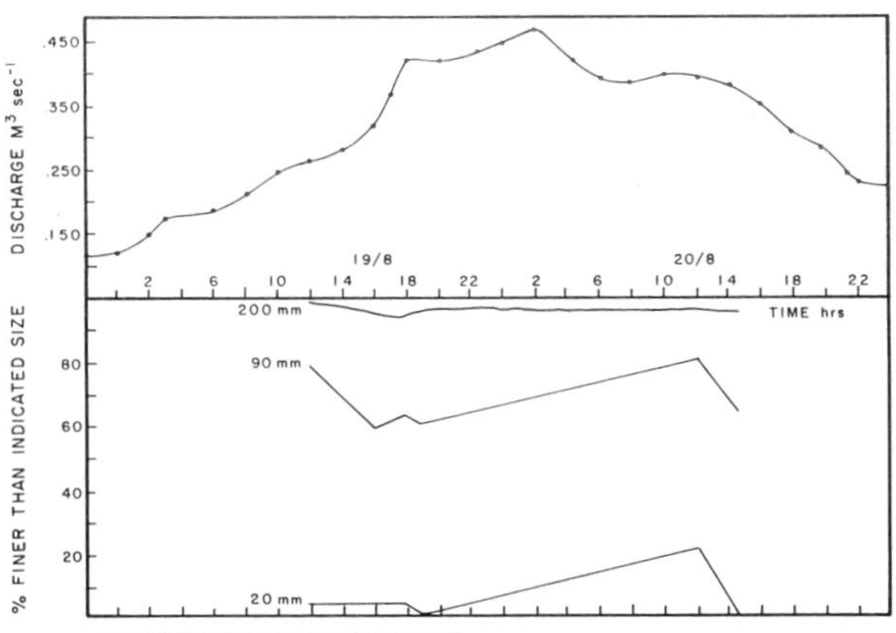


Figure 8.--Water discharge and sedimentology for storm event of 19/8 to 20/8 in the Torlesse Stream, New Zealand (after Hayward 1978).

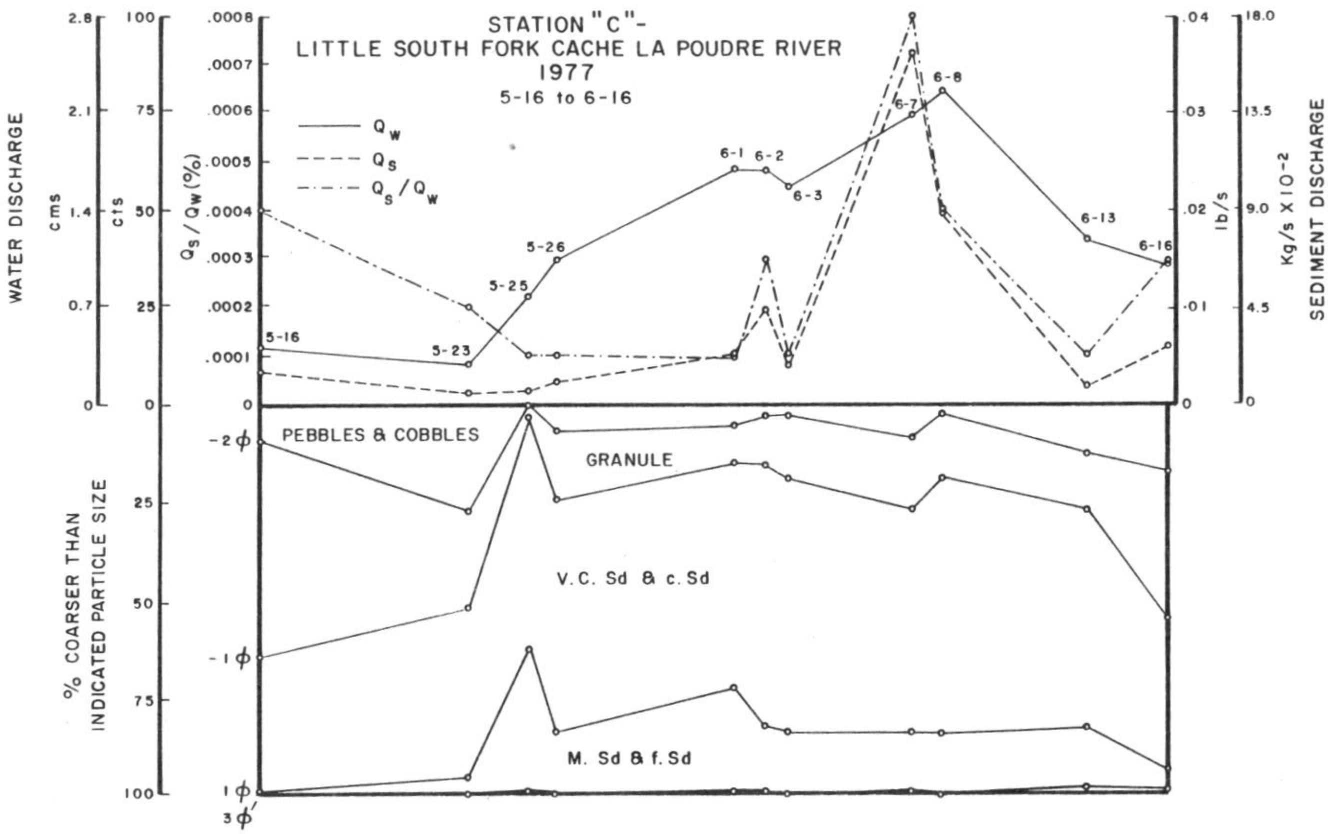


Figure 9.--Water and sediment discharge and sedimentology for 1977 runoff season, South Fork, Cache la Poudre River, Colorado.

The data generated in the R.E.F. agree in general with the results reported by Pain and Bowler (1973), Adams (1978), and Hayward (1978). The data also confirm that the quantities and composition of sediments transported after major slope failures change with time. The decrease in the proportion of suspended load is not as dramatic as would be expected in the field because the unvegetated slopes in the R.E.F. produce high background levels from the impact of raindrops.

Sediment-transport rates during the course of individual runs exhibited hysteresis. The first peak of sediment was dominated by suspended sediments and occurred before the peak of the hydrograph as a result of maximum sediment production or availability in the early stages of the storm event. As the hydrograph peaked, a second sediment peak occurred and was related to an increase in the transport of bed-material load. The third peak of sediment occurred on the waning stage and resulted from increased transport of the coarsest fractions, possibly because of the scour of locally aggraded and oversteepened reaches in the channel. An early increase in the percentage of coarse material at low discharges on the rising limb was observed and is attributed to continuing scour of locally aggraded reaches. Early and late peaks of coarse material were also observed in rivers in New Zealand and Colorado.

Hooke's (1968) similarity-of-process approach to physical modeling in geomorphology has been used in this study. Although model results must be viewed with caution when applied to the field situation, some observations indicate that the R.E.F. is behaving as a natural system, and may be used to indicate areas where field research could be directed. The similarities in the sedimentology of the R.E.F., field data collected in New Zealand and Colorado, and that reported in the literature suggest that sediment yields and composition change through time after major slope failures in headwater zones. Changes in sediment-rating curves with time may reflect this and could possibly be used to predict the downstream movement of waves or slugs of coarse material. S. Schumm (personal communication, Colorado State University, Ft. Collins, Colorado) considers that the destabilization of previously stable river reaches may be the result of this phenomenon. A study of the Niobrara River is currently underway to test this hypothesis.

ACKNOWLEDGMENTS

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LITERATURE CITED

Adams, J. E.
1978. Late Cenozoic erosion in New Zealand. Ph. D. thesis. Victoria Univ. Wellington, N.Z. 109 p.

Doehring, D. O., and F. G. Ethridge.
1979. Estimating coarse sediment transport in mountain watersheds: final report for Eisenhower Consortium for Western Environmental Forestry Research, Contract No. 16-692-9.R. Colo. State Univ., Fort Collins. 34 p.

Folk, R. L.
1965. Petrology of sedimentary rocks. 159 p. Univ. Texas, Hemphills Publ., Austin.

Gregory, K. J., and D. R. Walling.
1973. Drainage basin form and process. 449 p. Halstead Press, New York.

Hayward, J. A.
1978. Hydrology and stream sediments in a mountain catchment. Ph. D. thesis, Vol. I. Univ. Canterbury, N.Z. 63 p.

Heidel, S. G.
1956. The progressive lag of sediment concentration with flood waves. Trans. Am. Geophys. Union 37:56-66.

Hooke, R. L.
1968. Model geology: prototype and laboratory streams: discussion. Geol. Soc. Am. Bull. 79:391-394.

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

Mosley, M. P.
1975. An experimental study of channel confluences. Unpubl. Ph. D. thesis. Colo. State Univ., Fort Collins.

Mosley, M. P., and G. L. Zimpfer.
1978. Hardware models in geomorphology. Prog. Phys. Geogr. 2(3):438-461.

O'Loughlin, C. L., L. K. Rowe, and A. J. Pearce.
1978. Sediment yields from small forested catchments, North Westland-Nelson, New Zealand. J. Hydrol. (N.Z.) 17(1):1-15.

Pain, C. F., and J. M. Bowler.
1973. Denudation following the November 1970 earthquake at Madang, Papua New Guinea. Z. Geomorphol. Suppl. 18:92-104.

Porterfield, G.
1972. Computation of fluvial sediment discharge. In Techniques of water resource investigation. Book 3, chapter C3, 66 p. U.S. Geol. Surv.

Shields, A.
1936. Arwendung der Anhlickkeits - mechanik und der Turbalenzforschung auf die Geschiebebewegung. Mitteilungender Preuss. Versuchganst Fur Wasser, Erd, und Schiffshau: Berlin, Heft 26, 26 p.

Visher, G. S.
1969. Grain size distributions and depositional processes. J. Sediment. Petrol. 39(3):1074-1106.

Episodic Behavior in Badlands: Its Effects on Channel Morphology and Sediment Yields

Frank W. Bergstrom

ABSTRACT

Variable sediment yield and valley and channel morphology in rapidly eroding steepland drainage basins is referred to as episodic behavior. Three zones along channels can be distinguished on the basis of channel and valley morphology. Sediment production is discontinuous as headwater storage elements are periodically depleted and refilled. During a sediment production event, stored bed sediment is eroded from the headwaters and transported downstream, overloading the channel system. The headwater zone (zone 1) is a zone of material flushing. The transition zone (zone 2) braids during flushing of material from zone 1 and trenches during nonflushing periods. the lower, braided zone (zone 3) is a site of deposition as in-channel alluvial fans are locally built and reworked. Morphologic changes among the three zones are out of phase. Maximum sediment storage at a station progresses downstream. The types of sediment-storage elements vary from zone to zone. Drainage-basin sediment yield varies in response to these morphologic changes. Sediment delivery ratios range from less than 1 to 3 during a major sequence of sediment production, deposition, and erosion.

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INTRODUCTION

Erosion and sedimentation in mountain environments have characteristics distinctive from areas of lower relief. Helley and LaMarche (1973) showed that exceptionally large storm events are predominantly responsible for stream valley development in mountain basins. Their investigations in north coastal California indicate that only several of these valley-modifying events have occurred in the last 400 years. These episodic storms produce and transport large quantities of sediment. Such behavior can be expected to produce highly variable sediment yield at the basin outlet.

A major sediment production event integrates widespread mass wasting of slope-stored debris and sediment stored as headwater valley fill. This can increase sediment load within the channel system beyond stream carrying capacity, and deposition results. These deposits will subsequently be eroded and moved downstream.

In such a sedimentation sequence of production, deposition, and remobilization, sediment is temporarily retained in storage, but it eventually makes its way downstream, causing alternating episodes of deposition and erosion in stream channels.

INVESTIGATIONS

This study was undertaken to investigate the effects of episodic sediment production on stream morphology in steepland drainage basins. Once a qualitative concept of sediment transport processes and a quantitative assessment of sediment storage changes had been made during a sedimentation sequence, a model of steepland sediment production and channel change in space and time can be designed.

Several researchers have noted a distinct zonation of bed sediment transport processes and channel and valley morphology along the longitudinal profile of a steepland drainage basin. O'Loughlin (1969), Kelsey (1977), and Hayward (1978) all described a three-zone system. In this system, the upper basin acts as a source of debris for downstream reaches and consistently tends toward a negative sediment mass budget. The middle of the basin acts as a conduit or transmitter of debris and maintains a relatively balanced sediment mass budget. The upper and middle zones feed sediment into the lower reaches where decreased channel gradient and broad valley floors commonly result in net sediment accumulation. Harvey (this volume) describes these same zones in an experimental drainage basin in the Rainfall Erosion Facility at Colorado State University.

Previous conclusions regarding changes in channel morphology and sediment transport in steeplands have been derived from short-term studies. To substantiate these hypotheses, long-term studies are required, unless a time-space substitution can be achieved. Rapidly eroding badlands provide an opportunity to do this. Some prerequisites for such an area are high relief, abundant bedload production, and sufficient runoff to transport bedload.

The Kraft Badlands in east-central Wyoming is an appropriate area in which to perform such a study. These badlands lie just north of Pine Ridge, the southern boundary of the Cheyenne River Basin, at 42°58' north latitude and 104°9' west longitude. The Kraft Badlands are typical of the area below Pine Ridge Escarpment where the Chadron and Brule Formations of the White River Group crop out. The rolling grassland of this region supports about 10 percent groundcover of mixed grass and sagebrush with grass predominating. Soils are thin and residual with no profile development. When the vegetation and soils are breached, badlands form in the underlying bedrock. The study area is underlain by a sandy-clayey siltstone of the Brule Formation. Sediment is produced from bedrock by shrink-swell weathering. This process fractures the siltstone into coarse sand and granule-size particles, which are then transported to headwater channels and lower-slope storage locations by winter winds and dry raveling. Sediment is stored at these locations until summer thunderstorms deliver it to the main channel network. This constitutes an annual sediment-production event. The badlands display all the characteristics of steepland basins: wide, braided valley fills; long straight slopes; high relief; and elongate drainages.

A detailed investigation was conducted in the Kraft Badlands in the summer of 1978.

MORPHOLOGY AND PROCESSES OF STUDY BASINS

Preliminary studies revealed that three zones could be distinguished along the longitudinal profile of the study basins, as described in steeplands elsewhere (O'Loughlin 1969, Hayward 1978). Distinct breaks in valley morphology generally mark boundaries between the headwater, transitional, and lower braided zones--or, for simplicity, zones 1, 2, and 3, respectively. Zone 1 is separated from zone 2 by an abrupt increase in valley width and decrease in channel slope. Zone 2 terminates at zone 3, where valley width dramatically increases, usually where third-order tributaries join. Slope slightly increases at this point because of localized fan-building at the tributary junction.

Debris flows and debris torrents move sediment downstream through zone 1. Debris flows occur in steep first- and second-order streams, and scour bedrock. Mass failure is initiated when pore water pressures and rainfall intensities reach critical levels. Debris flows in the Kraft Badlands generally deposit steep fans at the first stream junction encountered; however, the flows may stop at any location within a channel. Debris torrents bring sediment down more gently inclined second- and third-order channels and deposit lobes of sediment in lower zone 1 and at the boundary between zones 1 and 2. Torrents are very wet slurries, similar to large flows described by Klock and Helvey (1976).

Below second- and third-order junctions, in zone 2, sediment moves as bedload in a braided channel confined between valley walls. At peak flows, the bedrock walls act as rigid boundaries. After an event with high bed-sediment flux, alternate reaches exhibit evidence of a raised channel bed in a regular fashion downstream. The spacing of these aggraded reaches corresponds with riffle spacings in pool-riffle streams (Keller and Melhorn 1978).

Valleys widen downstream and channels become permanently braided in zone 3. Energy is generally insufficient to transport continuously the quantity of material delivered during a storm event without temporary storage within the zone.

STORM PERIODS

Precipitation is used as a surrogate for runoff in this study. Storm characteristics during the study period are described in table 1. All storms were short-duration convective thunderstorms. Several consisted of a series of thunderstorms of various magnitudes which were associated with passing frontal systems. Four storm periods are distinguished beginning with May 27, 1978 (day 58) designated as the beginning of storm period 1 (table 1). The final column in this table signifies whether mass wasting of headwater channel stored sediment occurred only during storm periods 2 and 3 (table 1).

SEDIMENT BUDGETS FOR FIVE BASINS

Sediment production and yield were estimated or measured for basins 1100, 1200, 1300, 0200-0250, and 0300 (table 2; fig. 1). From these data, sediment-delivery ratios (SDR) were calculated.

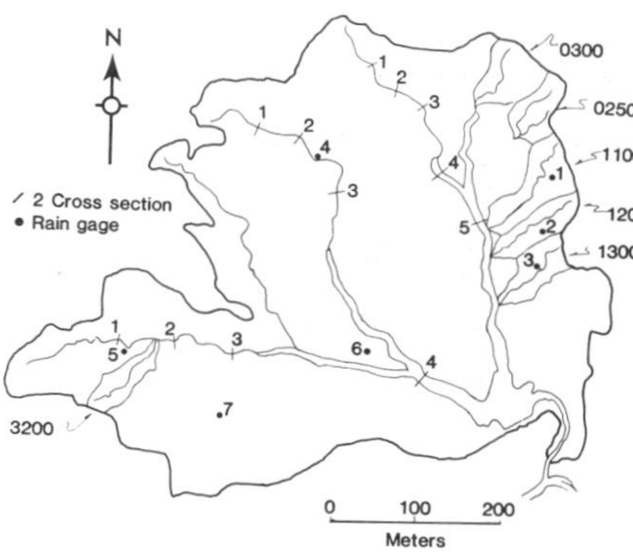


Figure 1.--Location of storage rain gages, tributary basins, and main drainages with cross sections in the Kraft Badlands, Wyoming.

The sediment-delivery ratio (SDR), the ratio of sediment yield to sediment production from hillslopes for the total basin, is usually less than 1.0. A value of unity occurs when all the material produced in the basin moves directly to the outlet or when erosion of downstream channel-stored sediment compensates for that sediment deposited upstream. Unity for the SDR then becomes the reference point from which net gain or loss from storage within the channel network can be quantified relative to sediment production.

A mass-balance equation was used to calculate sediment yield (Y) for the three storm periods. Assuming continuity of mass, yield can be calculated from the equation, $Y = P - \Delta S$, where Y = yield, P = production, and ΔS = change in storage along the main channel.

Sediment Delivery Ratios with Time

The sediment-delivery ratio (SDR) increased through time and with larger storms during the study (fig. 2). For storm period 1 (day 64, fig. 2), the basin 1100 SDR was greater than 1.0. Gross sediment volume transported was small relative to events in storm periods 2 and 3. Zone 3 alluvium was gullied and transported out of the drainage basin unaccompanied by headwater bed material production. After storm period 2, the SDR in basins 1200 and 1300 was less than in basins 0300 and 1100, all being less than 1.0 (fig. 2).

For period 3 storms, the most erosive storms of the study period (fig. 2), the SDR decreases for the low-relief basins (1100 and 0300), but increases for the high-relief basins (1200 and 1300). In all cases, the SDR is less than 1.0. Therefore, sediment delivery increased more rapidly--relative to main stream storage--in steep, deeply incised basins (high relief) than in less steep, more shallowly incised, low relief basins.

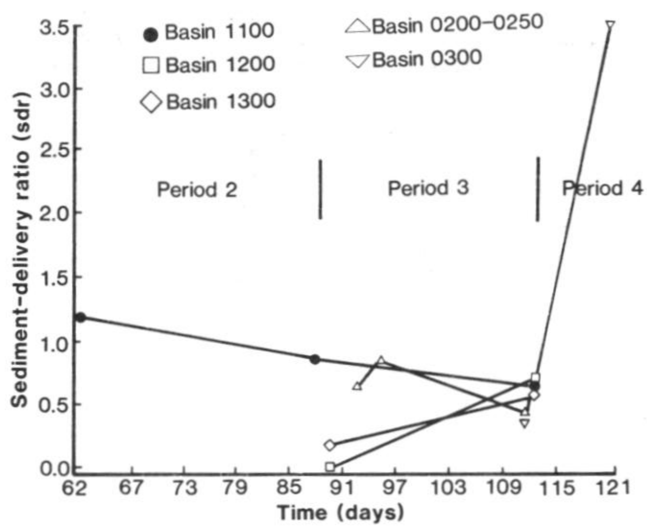


Figure 2.--Change of sediment-delivery ratios with time for five tributary basins.

Table 1--Precipitation data for period May 27, 1978 (day 58) to August 12, 1978 (day 134)

Storm period	Date	Storm number	Peak rainfall intensity	Mass wasting of headwater channel sediment?
			cm/h	
	May 27 (day 58)			
1		1	0.2	NO
1		2	0.5	NO
	June 4 (day 64)			
2		3	0.5	YES
2		4	0.5	YES
2		5	1.0	YES
2		6	1.0	YES
2		7	0.2	YES
2		8	2.3	YES
2		9	1.5	YES
	June 3 (day 94)			
3		10	2.0	YES
3		11	0.5	YES
3		12	0.2	YES
3		13	0.5	YES
3		14	20.6	YES
	July 23 (day 114)			
4		15	1.5	*
4		16	2.3	*
4		17	0.5	*
4		18	0.2	*
	Aug. 12 (day 134)			

*Headwater channel sediment storage completely depleted.

Table 2--Geomorphic characteristics of the six tributary basins

Drainage basin	Relief, H	Basin length parallel to main stream	Length of main stream	Area, A	Total stream length, L	Drainage density, Dd=L/A
	Meters	Meters	Meters	$Meters^2$	Meters	$Meters/meters^2$
0300	3.7	64.7	77.4	2512	489	0.19
0200	4.8	73.1	89.0	2097	492	0.24
1100	8.7	112.3	143.0	4574	867	0.19
1200	7.2	100.6	107.4	2662	492	0.18
1300	9.1	99.9	102.1	2661	658	0.25
3200	9.4	90.6	99.5	2031	444	0.22

During storm period 4, the SDR increased dramatically and again exceeded 1.0 in basin 0200-0250. The main channel in zone 2 was the source of transported sediment. This is similar to period 1 in basin 1100. The difference in the magnitude of the SDR between periods 4 and 1 results from the greater intensity and duration of storm 14 relative to period-1 storms.

The variability of main-channel storage and sediment-delivery ratio over the study period within an individual basin illustrates that drainage-basin morphology is not the only factor controlling these variables. For better understanding of the great variability of steepland erosion and sedimentation implied by figure 2, the linkages between processes of production and transport themselves must be taken into consideration.

MORPHOLOGY CHANGES AND SEDIMENT ROUTING

General Description

Basin 1200 is very similar to basins 1300 and 3200 (fig. 3). Figure 4 shows the longitudinal profile for basin 1200. Headwater channels are steep, deep, narrow, and long, whereas channels in zones 2 and 3 are wider, flatter, and maintain permanent alluvial fills. Snowmelt runoff incised an initial, sinuous channel in the alluvium of zone 2. This channel shallowed and became discontinuous in zone 3, and zone 1 was unaffected.

Behavior During Storm Period 2

No storm runoff occurred during period 1, but period-2 storms were of sufficient magnitude and intensity to initiate debris flows in headwater channels. Debris torrents eroded the upper zone-1 main channel to bedrock, and a large mass of sediment was delivered to the boundary between zones 1 and 2 (fig. 5). Sections of fill in the upper 20 m of channel in zone 1 were scoured during period 2 (the July 4 survey).

At the boundary between zones 1 and 2, thalweg elevation rose abruptly (fig. 5). A secondary peak at cross section 6 corresponded with an increase in valley width. Downstream from these peaks, the thalweg bed elevation rose above the original snowmelt channel elevation (fig. 6), but it did not reach the reference survey elevation, so a loss of elevation is recorded on figure 5.

Cross section 7, 27 m downstream from the head of the channel, shows the deposition at the boundary between zones 1 and 2 (fig. 7). The initial, U-shaped fill inherited from winter accumulation still existed in the June 3 survey (fig. 7). During period 2, debris-torrent deposition increased mean bed elevation 9.7 cm, the maximum recorded increase in mean bed elevation throughout the study basin.

As sediment storage increased at cross section 7, the valley floor of zone 2 was reworked, and the snowmelt channel filled--with little change in mean bed elevation. Channel morphology changed to braided, with the channel covering about half the valley width.

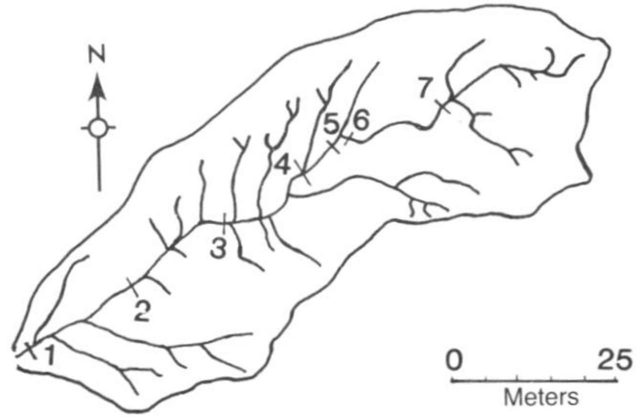


Figure 3.--Basin 1200 showing drainage divide, drainage pattern, and cross-section locations.

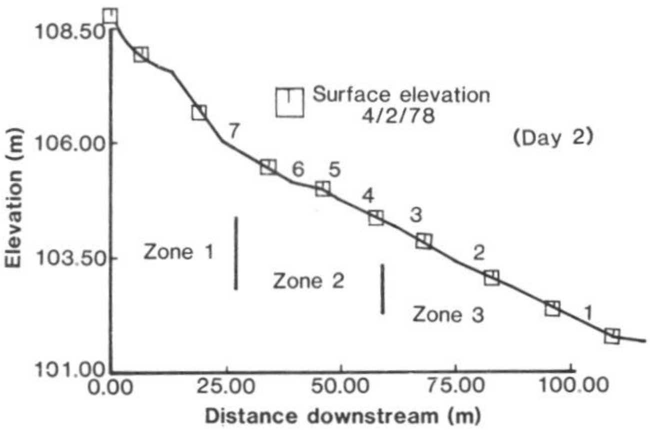


Figure 4.--Longitudinal profile of channel in basin 1200. Numbers refer to cross-section locations.

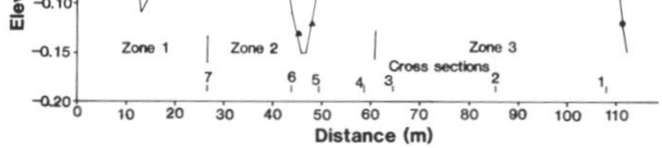


Figure 5.--Thalweg bed-elevation change relative to the April 2, 1978 valley floor (terrace) elevation, basin 1200.

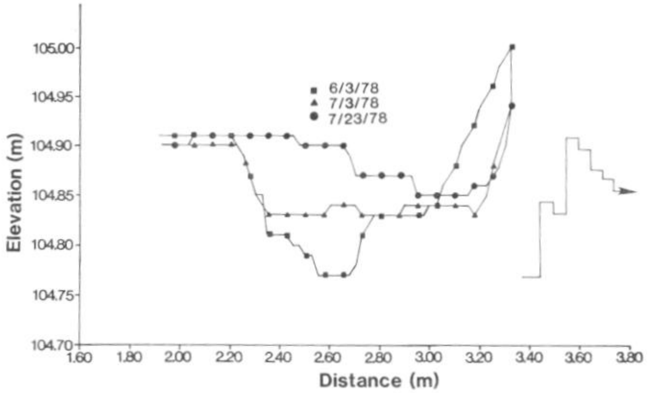


Figure 6.--Cross section 5, zone 2, basin 1200. Schematic diagram at the right illustrates channel-bed behavior during the study period.

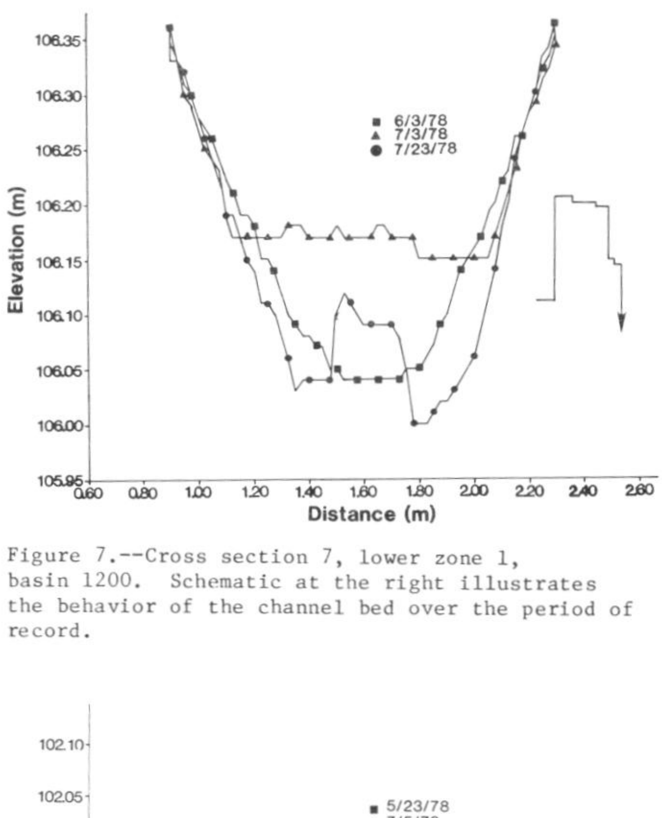


Figure 7.--Cross section 7, lower zone 1, basin 1200. Schematic at the right illustrates the behavior of the channel bed over the period of record.

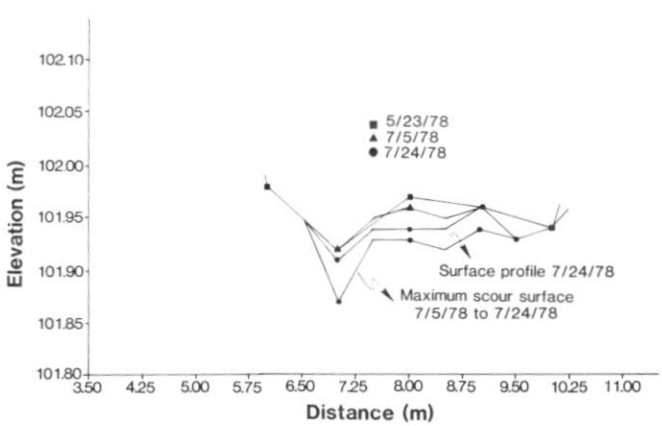


Figure 8.--Cross section 1, lower zone 3, drainage basin 1200. Lower profile for 7.24.78 is the surface of greatest scour during the survey interval. Upper profile for 7.24.78 is existing surface at time of survey.

During period 2, local filling of reaches of zone 3 (fig. 5) reflects the formation of fans in the channel. The profiles of cross section 1 (fig. 8) illustrate the trenching and subsequent reconstruction of an in-channel fan. Multiple channels incised the fill during storm period 2. This was followed by minor refilling, as documented by the surface profile on July 5 (fig. 8). During storm period 3 (July 24 survey), the greatest recorded incision was followed by channel filling and general valley aggradation as the fan began to rebuild.

Behavior During Storm Period 3

Period-3 storms moved the sediment stored at cross section 7 downstream (fig. 5), but the accumulation 40 m downstream from the channel head was not removed. This accumulation shifted downstream, filling and widening the channel below the point of valley widening at cross section 6.

During period 3, the channel at cross section 7 degraded to the level of the central terrace of the July 23 profile in figure 7. The channel then continued to incise, leaving this terrace and finally contacting bedrock at both valley sides. The overall changes in periods 2 and 3 were characterized by abrupt filling followed by incision (fig. 7). In contrast, the zone-2 channel filled during storm period 3 and subsequently retrenched a channel in a series of steps, forming unpaired terraces with an overall increase in sediment storage relative to the early June condition (fig. 6). Therefore, the two sections were out of phase in fluctuations of bed-sediment storage. In-channel fan building continued in zone 3.

MODEL OF CHANNEL CHANGES IN BASIN 1200

Observations in basin 1200 provide a basis for describing the general behavior of the streambed during a sedimentation sequence.

Zone 1 (Headwater Zone)

Figure 9 shows an idealized sequence of channel cross section development in lower zone 1. Winter sediment production produces the initial spring conditions of a U-shaped valley fill (line 1). Debris-torrent deposition abruptly aggrades the cross section (line 2). Precipitation on hillslopes of exposed bedrock yields flashy, sediment-free runoff, which erodes the deposit to the terrace level of line 3. Subsequent late-storm streamflow further incises the remaining deposit, forming a valley-side terrace. This illustrates an excess of transport energy relative to sediment availability in zone 1.

Zone 2 (Transition Zone)

Over the same sedimentation sequence, the zone-2 channel fluctuates by eroding initially and then shifting between aggrading and retrenching. Discontinuous terraces and bars characterize the initial fill (line 1, fig. 9). Storms that precede sediment production from zone 1 discontinuously incise a narrow channel and

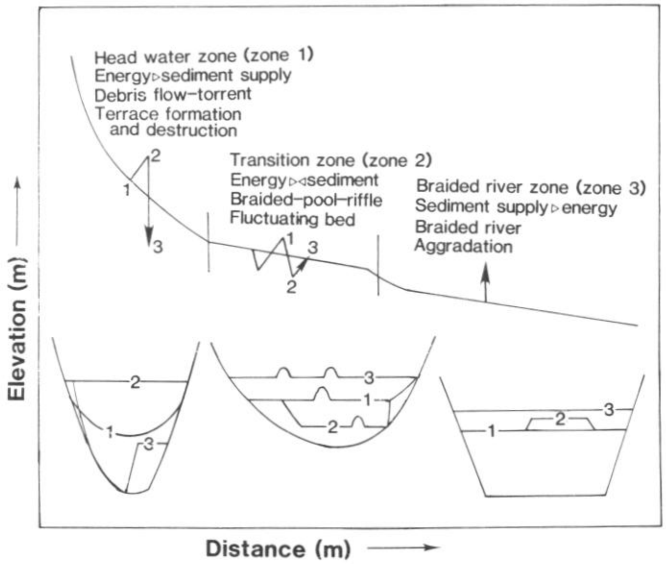


Figure 9.--Idealized longitudinal profile divided into headwater, transitional, and braided zones. Arrows denote mean behavior of bed through a sedimentation cycle. Cross sections correspond to each zone over time, progressing from 1 to 3.

produce multiple, paired or unpaired terraces. As bed sediment produced in zone 1 is transported through zone 2, the bed elevation rises and flow covers the entire valley floor. That is, during a series of major runoff events that follow a major upstream sediment production event, valley bed sediment storage temporarily increases. Scour of recently deposited sediment occurs after sediment delivery from zone 1 ceases, yet runoff events continue.

Zone 3 (Lower Braided Zone)

In zone 3, mean bed elevation tends to increase with time during a sedimentation sequence. The general trend is for material to shift from the headwaters to zone 3 and to be stored there for periods exceeding a year under moderate hydrologic conditions. Only during the largest hydrologic events did basins produce runoff and transport sediment throughout the three zones. During smaller storms (period 1), when zone 1 and zone 2 or both were actively producing and transporting material, zone 3 received little flow capable of modifying the valley fill. The time available to produce bed changes in zone 3 was limited during the summer storm season. Duration of runoff was very important. Zone 3 acts as a sink for sediment until a sufficiently large hydrologic event causes transport in zone 3.

Bar formation and general aggradation in discrete reaches form fans. Beginning at line 1 (fig. 9), channels incise and possibly widen to encompass most of the valley width when water and bed-sediment discharge are low. These channels quickly fill when bed-sediment discharge increases and in-channel fan building begins. Bars build and coalesce (lines 2 and 3, fig. 9) until the entire valley floor is raised. Trenching through the fan by headward extension of a trench should eventually remove most of this deposit.

SUMMARY

In steep terrain, as exemplified by the Kraft Badlands, sediment production is not uniform in time or space. A sediment production, deposition, erosion sequence occurs when large quantities of coarse sediment are episodically produced and delivered to the channel network in steepland drainage basins. Widespread mass wasting of slope and channel-stored material produces large quantities of sediment that are progressively delivered to the main stream. Transport capacity of the channel network is overloaded by a production event and a sequence of channel filling and erosion downstream ensues.

Three zones along the longitudinal profile were recognized: (1) headwater, (2) transition, and (3) lower braided zones. Different bed-sediment transport processes and resulting channel and valley forms characterize each zone.

The zones within a basin are usually out of phase with one another. Zone 2 initially acts as a source of sediment while zone 3 acts as a sink. As zone 1 begins to produce sediment, zones 2 and 3 fill. Finally, zone 2 acts as a conveyor of material temporarily stored at the boundary between zones 1 and 2, and zone 3 acts as both a sink and source of sediment as in-channel fans build and erode. Bed-sediment availability at the basin outlet increases late in this sequence. Therefore, sediment yield measured at the basin outlet from a storm of given magnitude varies greatly, depending on when the storm occurs relative to channel-bed changes upstream.

Of great significance is the large magnitude of the changes, and their nonuniformity in time. This suggests that engineering structure designs based on short-term records of sediment yield and existing channel morphology in steepland basins may prove to be inadequate.

LITERATURE CITED

Hayward, J. A.
1978. Hydrology and stream sediments in a mountain catchment. Ph. D. thesis. Univ. Canterbury, Christchurch, N.Z. 3 vol., 147 p.

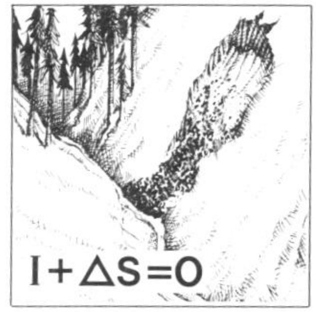
Helley, E. J., and V. C. LaMarche.
1973. Historic flood information for northern California streams from geological and botanical evidence. U.S. Geol. Surv. Prof. Pap. 485-E, 16 p.

Keller, E. A., and W. M. Melhorn.
1978. Rhythmic spacing and origin of pools and riffles. Geol. Soc. Am. Bull. 89(5):723-730.

Kelsey, H.
1977. Landsliding, channel changes, sediment yield and land use in the Van Duzen River basin, north coastal California 1941-1975. Ph. D. thesis. Univ. Calif., Santa Cruz. 370 p.

Klock, G. O., and J. D. Helvey.
1976. Debris flow following wild fire in north central Washington. Proc. Third Fed. Interagency Sediment. Conf., Denver, Colo. p. 1.91-1.97.

O'Loughlin, C. L.
1969. Streambed investigations in a small mountain catchment. N.Z. J. Geol. Geophys. 12:684-706.



Sediment Budget of a Small Coast Range Drainage Basin in North-Central California

Andre K. Lehre

ABSTRACT

Detailed measurements of erosion and sediment discharge in the drainage basin of Lone Tree Creek, a 1.74-km^2 drainage 14 km northwest of San Francisco, showed that landslides in colluvium-filled swales are the most important erosional agent and account for most of the sediment yield from the drainage basin. The bare scars act as loci for sheetwash and gully development; they refill by backwearing of slide scarps and by soil creep. During the study period, only 53 percent (2068 t/km^2) of all sediment mobilized was discharged from the basin; the remainder was stored in slide scars, on foot slopes, and in gully and channel banks and beds. Sediment is removed from storage by storms with recurrence intervals greater than 10 years. Comparison of current rates of landsliding and scar filling suggests that colluvium is currently being stripped and that the rate of sliding has increased approximately tenfold in the last 100 years.

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INTRODUCTION

The sediment budget of a drainage basin is a quantitative statement of relations between sediment mobilization and discharge, and of associated changes in storage. Such budgets are useful for land management and are a first step in understanding landscape evolution. Proper formulation of a sediment budget requires: identification of erosional processes and understanding of their controls and interrelations; measurement of magnitude and frequency of sediment mobilization by each process, and of sediment discharge from the drainage basin; and identification of sediment-storage elements and quantification of the volume, residence time, and changes in volume of stored sediment. Of particular interest are the circumstances under which material is removed from or added to storage.

Erosional processes in steepland drainage basins of the California Coast Range are complex, involving sliding, soil creep, gullying, sheetwash, and channel-bed and bank erosion. Detailed studies are few. This paper reports results of 3 years of measurements in a small, mountain drainage basin north of San Francisco and the model of process linkages they suggest.

THE DRAINAGE BASIN

Location and Land Use

Lone Tree Creek is a small, steep, relatively undisturbed drainage basin in the California Coast Range about 14 km northwest of San Francisco. It lies on the southwest slope of Mt. Tamalpais and drains directly to the Pacific Ocean. Study was restricted to the 1.74 km^2 upstream of Highway 1, an area including the drainages of Main Lone Tree Creek (MLTC) and South Branch (SOBR) (fig. 1). The drainage basin was grazed by cattle from the mid-1800's until 1967, when it was incorporated into Mt. Tamalpais State Park. Today it is used only by occasional hikers.

Geology

The drainage basin is underlain by greywacke melange of the Franciscan Assemblage (Blake et al. 1974), consisting predominantly of a matrix of sheared and locally comminuted greywacke and shale embedded with isolated blocks of hard coherent greywacke and "exotic" blocks of greenstone, chert, altered rhyolite, and metamorphic rocks. These blocks or "knockers" range from 0.3 to 150 m across. The broken and weathered greywacke matrix is exposed only in stream channels, some landslide scars, and as barren, steep-fronted steps on hillslopes and ridgecrests.

Topography

Lone Tree Creek drains narrow, steep-walled canyons separated by gently rounded upland ridgecrests (fig. 1). Average channel slopes range from 15 percent on MLTC to 36 percent on some tributaries. Upland slopes range from 5 to 30 percent. Valley side-slopes are 35 to 85 percent and tend to be convex or straight in profile and strongly convex in plan. Most hillslopes are indented by shallow swales spaced 30 to 150 m apart across the slope. These topographic hollows are typically 60 to 90 m long and 30 to 60 m wide. Upslope and at their heads, swales are U-shaped and lack a defined channel; downslope, some remain U-shaped, without a channel, but many become narrowly V-shaped and merge into a gully at their lower end. Soil augering and exposures in gullies and landslide scars reveal that the U-shaped portions contain from 1 to 5 m of colluvium filling a U- or V-shaped hollow scoured in bedrock. Bedrock is often exposed at the downslope end of the swale, particularly if it narrows and becomes a gully. Similar features have been observed elsewhere in California (Wahrhaftig 1974, p. 87) and Oregon (Dietrich and Dunne 1978). Swales carry runoff during heavy winter rains and are the first-order drainages of the basin. They are also the main loci of landsliding and hillslope gullying.

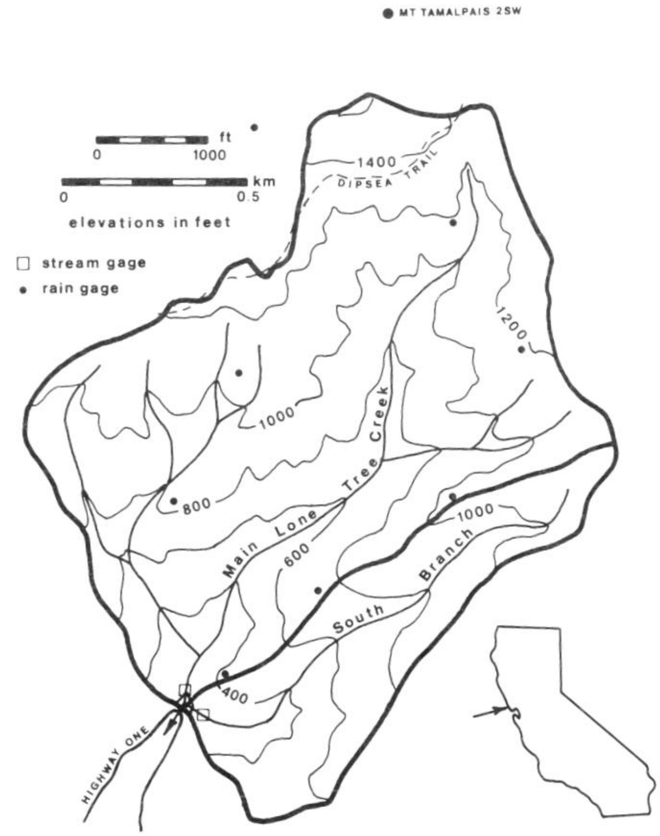

Figure 1.--Lone Tree Drainage Basin.

Including first-order tributaries mapped on air photos and in the field, the drainage basin contains 10.8 km of channels. Drainage density is 6.21 km/km^2. Relief ratio is 190 m/km.

Vegetation

Grassland covers about 50 percent of the drainage basin. It occupies chiefly ridgecrests and exposed upper and west-facing slopes. Brushland, composed primarily of rabbitbrush (Baccharis spp.), sage (Artemisia spp.), and poison oak (Rhus diversiloba T. & G.), covers about 30 percent of the basin. It occupies the steeper, rockier, drier slopes and lies most typically between grassland and the forest of the canyon bottoms. Local concentrations of brush also occur in and along the lower portions of many gullies draining landslides. Forest occupies most of the valley bottom and north-facing slope of MLTC and the bottoms and lower slopes of its tributaries in the upper half of the basin. In the lower half of the drainage basin, forest consists mainly of California laurel (Umbellularia californica (H. & A.) Nutt.), some oak (Quercus spp.) and tanoak (Lithocarpus densiflorus (H. & A.) Rehd.), and infrequent Douglas-fir (Pseudotsuga menziesii (Mirb.) Franco); in the upper half, forest is mostly 30- to 50-year-old Douglas-fir and Laurel. Understory is usually either dense bracken fern or nearly nonexistent.

Climate and Hydrology

Lone Tree Creek has a Mediterranean climate moderated by sea-fog which blankets much of the basin in summer. Mean annual rainfall is 865 mm (Rantz 1971), about 90 percent of which falls between November 1 and April 30. The abrupt rise of the basin from 61 m at the outlet to 427 m at its head 1.9 km away produces strong orographic effects. Data from eight gages in the basin show that rainfall at the outlet is only half that received at Mt. Tamalpais 2SW, a U.S. Weather Service gage 0.4 km north of the basin head.

Rainfall intensities measured at Mt. Tamalpais 2SW from October 1971 to September 1974 exceeded 5 mm hour less than 5 percent of the time. Regional depth-duration-frequency relations (Rantz 1971) indicate that intensities of 25 mm/hour have a 15- to 20-year recurrence interval. Ninety-five percent of 24-hour rainfalls are less than 65 mm. The maximum 1-hour rainfall during the study was 20 mm on January 12, 1973, which also produced the maximum 24-hour rainfall of 165 mm. These represent the wettest day of a very wet year and were associated with a 15- to 20-year flood (Lehre 1981).

Total annual precipitation may differ by a factor of 2 in successive years. Based on the 34-year record at the nearby Muir Woods gage, the 3 study years 1971-72, 1972-73, and 1973-74 rank 22, 3, and 4, respectively, in terms of total annual rainfall. Although the rainfall totals for 1973 and 1974 are similar (table 1), the 1974 rainfall was spread more evenly over the rainy season.

Mean annual runoff of Lone Tree Creek, estimated through several methods (Rantz 1967, 1974; Crippen and Beall 1971), is about 330 mm or about 38 percent of estimated mean annual precipitation. For the study period, runoff ranged from 34 to 74 percent of annual rainfall (table 1). Runoff produced by individual storms of similar size ranged from 2 to 94 percent of rainfall, depending on rainfall intensity and antecedent precipitation.

During and immediately after large storms, overland flow occurs down the center and lower slopes of most grassland swales; these areas may remain saturated or nearly so for several days after a storm. Overland flow also occurs on bare soil in landslide scars and in saturated zones immediately adjacent to stream channels. The openness and unrilled appearance of forest litter suggest that overland flow is relatively unimportant there. The proportion of drainage basin contributing direct runoff in storms, estimated by the technique of Dunne et al. (1975), increases from 1 to 3 percent in light storms to 25 to 33 percent in heavy ones and on SOBR approached 70 percent in the storm of January 12, 1973.

HYDROLOGIC INSTRUMENTATION AND DATA COLLECTION

Long-term precipitation data are available from two nearby U.S. Weather Service gages, Mt. Tamalpais 2SW (recording) and Muir Woods (nonrecording). Eight cans installed as gages at several elevations within the basin were read after major storms.

Streamflow was measured at staff gages on MLTC and SOBR near Highway 1 (fig. 1). A crest-stage gage at the MLTC site and an auxiliary staff gage at the inlet of the highway culvert permitted indirect measurement of flood flows. From October 1971 to October 1974, gages were read at 5-minute to 6-hour intervals during nearly all major storms and about half of the minor ones.

Table 1--Annual rainfall, runoff, and water loss for water years 1972-74 on Lone Tree Creek

Water year	Rainfall	Runoff	Water loss	Runoff, as % of rainfall
	Millimeters			
1972	602	206	396	34
1973	1184	874	310	74
1974	1046	770	277	74
Average for period	945	617	328	65

Suspended-sediment and dissolved-solids concentrations were determined from depth-integrated samples taken at the gaging sites using a DH-48 sampler (Guy and Norman 1970). Time between successive samples in a storm ranged from 5 minutes to 6 hours. A total of 184 samples were taken on MLTC, 172 on SOBR, and 79 on their combined flow. Of these, 30 on MLTC and 29 on SOBR were analyzed for total concentration of dissolved solids. Sediment concentrations were determined by procedures outlined by Guy (1969). A very small number of bedload samples (six on MLTC, one on SOBR) were taken with a hand-held Helley-Smith bedload sampler. High velocities, high turbidity, large bed-particle sizes, and time constraints made bedload sampling generally impractical. Bedload estimates are thus uncertain.

WEATHERING, SOIL, AND COLLUVIUM

Weathering in the drainage basin is dominantly chemical. Rainwater penetrating joints and fractures in the greywacke bedrock reacts with feldspars and unstable dark minerals to form illite, vermiculite, and random mixed-layer clays (Schlocker 1974). The end product is a rock residue of silty sand or sandy silt containing from 10 to 85 percent gravel-size (2- to 150-mm) clasts of less-weathered bedrock. Its bulk density is 1.45 to 1.9 g/cm^3 compared to 2.4 to 2.65 g/cm^3 for unweathered rock.

Weathered rock becomes colluvium through mixing and downslope transport by processes both biological (burrowing, root growth, tree fall) and physical (creep, sliding, sheet erosion, gullying). Colluvium forms the parent material of many drainage-basin soils.

Drainage-basin soils are generally thin (except in swales), stony, and poorly developed. Texturally, both soil and colluvium are stony or very stony silt loams (20 to 80 percent fragments larger than 2 mm). Soil clay content in the less than 2-mm fraction is 15 to 25 percent. Soil bulk density under grass cover ranges from 1.15 to 1.5 g/cm^3.

Soil and colluvium thicken dramatically from less than 0.6 m on divides between swales to 1.3 to 5 m in swale centers. This colluvium commonly contains local concentrations of coarse gravel aligned subparallel to either the bedrock or ground surface; surface wash and rilling produce nearly identical concentrations of gravel on the surface of present-day scars. Downslope elongation of weathered and fractured bedrock blocks contained in the colluvium, together with short, crude stone lines extending downslope from bedrock irregularities, suggest differential shear or creep. Surface creep is suggested by the dark, overthickened (0.6 to 1.6 m thick), apparently cumulative A-horizon of soils in grassland swale centers. In recent slide scars, dark topsoil accumulates below eroding scarps and is spread downslope by wash and creep.

Landslides and gullies on hillslopes are restricted almost entirely to swales. These contain all ages of landslide scars, from fresh, raw scars to old, nearly filled and totally revegetated ones visible only at low sun angles. From these observations, I interpret the swales as the product of infrequent landslides and gullying, and their colluvial fills as the result of subsequent creep, scarp-backwearing, surface wash, and (sometimes) slides upslope. Slide debris either enters channels directly or is stored on hillslopes below the scar.

A MODEL OF PROCESS RELATIONSHIPS

The first step in making a sediment budget for a drainage basin is to identify the processes mobilizing sediment, their relations, and their controls. Dietrich and Dunne (1978) have done this for a forested western Oregon drainage basin. Like theirs, the model proposed below for Lone Tree Creek involves filling and episodic evacuation of swales; it differs, however, by the inclusion of gullying and in the importance assigned to sheet and scarp erosion. Evidence for the model is discussed in detail in Lehre (1981).

General Erosion Model

Erodible material in the drainage basin--weathered rock, soil, colluvium, and alluvium--resides in four storage elements: swales, hillslope mantle outside of swales, channel and gully banks, and channel bed. In a gross sense, erosional processes can thus be divided into those acting primarily to move sediment from point to point on hillslopes, and those that mainly transfer sediment from the hillslopes to the channel system or from one part of the channel system to another. Linkages between storage sites and transfer processes are diagramed in figure 2.

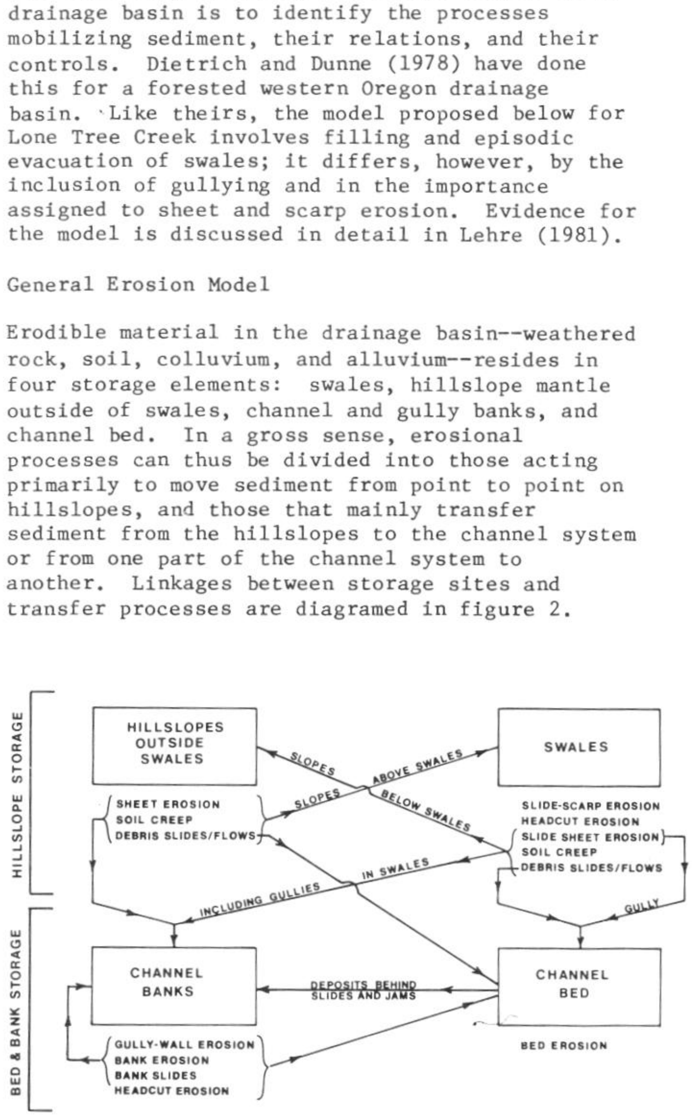

Figure 2.--General linkages between sediment-storage sites and erosional processes. Boxes indicate storage elements; listed below each box are erosional processes mainly responsible for mobilizing sediment in that element. Arrows show transfers between elements. Labels on arrows qualify or restrict location of transfers.

Swales are initiated when weathered rock fails as a debris slide, leaving a long, scoop-shaped depression. Sheet erosion and rilling begin operating on the floor of the scar, winnowing out fines and leaving a gravel lag. Water concentrated by the scar may initiate a gully downslope, which works headward and joins the scar to the channel system. The gully commonly widens and deepens significantly after connection to the scar. Sediment is contributed to scar and gully by spalling, sliding, and sheet erosion of their scarps; the gully also receives sediment eroded from the scar upslope. Soil creep on adjacent slopes helps replenish the eroding scarps. Within the scar, sediment is redistributed by surface wash, rilling, dry sliding, and creep. Much of the debris entering the gully is deposited either in a fan downslope (if the gully is discontinuous) or in the bed of the channel system.

Filling of the swale begins as coarse lag accumulates in rills and gully, forcing more water to infiltrate and move as subsurface flow. At the same time, shrubs and grass toppled in from scarps become established and stabilize the accumulating debris. Filling increases as erosion in scar and gully-bottom are reduced. With sufficient filling and revegetation, scarp erosion slows and rate of filling declines. Ultimately, scar and gully are healed, producing a smooth, U-shaped, colluvium-filled swale. Soil-water seepage and subsurface stormflow drain toward the swale axis, increasing moisture content and favoring production of saturation overland flow and high pore pressures in the swale center. The colluviul fill is gradually destabilized as it is thickened by continued creep and weakened by weathering (abetted by high moisture content). Finally, the colluvium fails in a heavy rainstorm, scouring the weathered bedrock, and the whole process begins over.

Sediment transported downslope out of swales by sliding either enters a channel directly, where it is usually entrained by stormflow, or is deposited on the slope below and returns to hillslope storage.

Sediment in storage on hillslopes outside of swales is transported downslope by creep (including dry sliding and burrowing), surface wash, and, less commonly, by landslides. It ultimately enters the channel system through gully, bank, or headcut erosion. Most sediment in this element represents material weathered in place or involved in slow downslope transport. Locally, usually below swales, slide and gully deposits may cause large accretions to hillslope storage.

All sediment not transferred directly from hillslopes to the channel system by debris flow or gully transport must enter channels by erosion of channel banks, gully walls, and headcuts. Although this erosion is actually accomplished by sheetwash, scour, spalling, and sliding, it is more useful to consider only (with the exception of large bank slides) their combined effect. Thus figure 2 lists as the transfer processes gully-wall erosion, headcut retreat, bank erosion, and bank slides. Sediment mobilized by these processes is transferred to the channel bed or redeposited lower on channel banks.

Material reaching the channel bed is either removed immediately from the drainage basin as bedload and suspended load, stored temporarily as bars and channel deposits, or converted to bank deposits through accumulation behind a slide deposit or debris jam.

PROCESSES AND PROCESS MEASUREMENT

The second step in creating a sediment budget is quantifying the contributions from each erosional process. In this section, each process is briefly discussed, together with the means used to measure its rate. Here I use "sediment mobilization" to mean total amount of sediment moved any distance by a process, "sediment production" to mean the amount of mobilized sediment reaching or given access to a channel, and "sediment yield" to designate the amount of sediment actually discharged from the drainage basin. Note that sediment production by a process can never exceed mobilization by that process. Yearly mobilization and production for each process are listed in table 2.

Hillslope Failures

Landslides occupy 2.7 percent of basin area and are the most conspicuous erosional features in the drainage basin. I mapped scars into three categories: raw (more than 80 percent unvegetated), recovering (10 to 80 percent unvegetated), and healed (less than 10 percent unvegetated). I identified a total of 404 individual scars, of which 320 (68 raw, 158 recovering, 94 healed) are on hillslopes, chiefly in swales; the remaining 84 (28 raw, 35 recovering, 21 healed) occur on channel banks. Most hillslope slides occur in grass and brush; only five were identified in forest.

Hillslope failures are almost exclusively debris slides and flows. Failure occurs during heavy rains, either on the bedrock-colluvium contact or above a denser or more clay-rich layer in the colluvium. During the study period, 29 slides were triggered by the storm of January 11-12, 1973, when 165 mm of rain fell in 24 hours, producing a 15- to 20-year flood. In the storm of March 31-April 1, 1974 (115 mm/24 hours, producing a 2- to 3-year flood), only two landslides occurred, both in swales undercut in 1973 by small, discontinuous gully heads. The high percentage of debris reaching channels in 1973 (53 percent) and 1974 (68 percent) (table 2) reflects the great fluidity of the debris flows and, to a much lesser extent, proximity to channels. Unlike the debris torrents reported in Oregon and northern California (e.g., Pierson 1977), these debris flows are not highly erosive. They commonly pass over grass and through brush with little damage; I attribute this to their high clay and water content.

Table 2--Sediment mobilization and production from individual processes in Lone Tree drainage basin

Source	1971 - 1972 Mobilization (on slopes)	1971 - 1972 Production (reaching channels)	1972 - 1973 Mobilization (on slopes)	1972 - 1973 Production (reaching channels)
	Metric tons/km^2			
Landslides	0	0	1049	559
Slide scarp erosion	34	21	59	37
Slide scar sheet erosion	20	15	76	59
Headcut erosion	22	20	25	24
Soil creep	6*	Included in gully-	6*	Included in gully-
Hillslope sheet erosion	4*	scarp erosion	4*	scarp erosion
Gully-scarp erosion		82		194
MLTC bank erosion		8		20
Bed erosion		2		92
Total	86	148	1219	985
	1973 - 1974		1971 - 1974 average	
Landslides	1795	1223	948	594
Slide scarp erosion	41	26	45	28
Slide scar sheet erosion	107	84	68	53
Headcut erosion	7	6	18	17
Soil creep	6*	Included in gully-	6*	Included in gully-
Hillslope sheet erosion	4*	scarp erosion	4*	scarp erosion
Gully-scarp erosion		201		159
MLTC bank erosion		20		16
Bed erosion		15		37
Total	1960	1575	1089	904

*Indicates effective mobilization. Effective mobilization, computed as (total length of channel banks in drainage basin) X (mobilization rate per unit width of slope), gives rate at which sediment can be supplied to channel banks by processes acting continuously over drainage-basin hillslopes. In contrast, total mobilization (sediment moved any distance) by such processes is given by (thickness of moving layer) X (total area affected by process) and is independent of downslope transport velocity.

Sequential comparison of aerial photos taken in 1952, 1965, and 1973, together with a volume-area relation for the scars, yields an average 22-year rate of sediment mobilization by sliding of 525 t/km^2 per year. Most of this sliding occurred in the storms of December 1955-January 1956, which produced 100-year flooding in most of northern California.

In an attempt to estimate long-term slide erosion rates, I have assumed arbitrarily that all raw and recovering landslides are under 100 years old. Their aggregate volume, divided by 100, yields a long-term slide mobilization rate of 187 t/km^2 per year. This rate is 7 to 10 times larger than rates reported from the Oregon Coast Range (Dietrich 1975, Swanson et al. 1976, Pierson 1977), 2.5 to 3.5 times rates in the western Oregon Cascades (Morrison 1975, Swanson and Dyrness 1975), and about one-sixth of the minimum 80-year rate in the San Gabriel Mountains of southern California (data of Rice et al. 1969 and Rice and Foggin 1971).

Slide-Scarp Erosion

Slide scarps erode through spalling, sliding, and sheetwash. I obtained measurements of scarp backwearing from 149 sets of stakes installed around 14 raw or recovering landslides. Each set consists of two survey stakes driven to define a line approximately normal to the scarp. A steel tape was used to measure the distance along this line from the outer stake to the scarp; measurements are accurate to 5 to 10 mm. Stake-scar distances were remeasured after the rainy season. Retreat was converted to volume by multiplying by scarp height.

Retreat rates at individual stakes ranged from 0 to 240 mm in a single year. The 3-year mean retreat rate (all stakes) was 20 mm/year, equivalent to an average volumetric rate of 0.014 m^3/m per year. Standard error (s_e) of the retreat rate is 2 mm/year. Mobilization was computed by multiplying the volumetric rate by total length of slide scarps in the drainage basin; this yields a 3-year total of 233 t.

Sediment production from landslide scarp-retreat is substantially less than mobilization for two reasons: about 40 percent of total scarp length surrounds slides not connected to the drainage net; and some unknown, but probably large, proportion of the mobilized sediment does not find its way directly to a channel but rather remains temporarily or permanently in the scar. An upper bound on sediment production can be calculated by subtracting from total mobilization the mobilization occurring in scars unconnected to the drainage net (table 2).

Slide-Scar Sheet Erosion

Sheet erosion in slide scars is caused by sheetwash, rainsplash, micromudflows, and rilling. I measured this process with 173 nail-and-washer erosion pins (Emmett 1965) installed in six landslide scars. Average erosion in rilled scars was 5 mm/year (s_e = 1 mm/year); in unrilled slides it was 2 mm/year (s_e = 0.7 mm/year). Mobilization is calculated by multiplying bare slide area by the appropriate pin-erosion rate. In making this calculation, I assumed that "raw" slides contributed from 100 percent of their area, but "recovering" slides contributed from only 50 percent. "Healed" slides were ignored. Total sediment mobilized from 1971-74 was 353 t. An upper bound on sediment production is given by subtracting from this total the mobilization occurring in scars unconnected to the channel system (table 2).

Headcut Retreat

Headcut retreat was measured with 58 stake sets installed around 19 headcuts. Average retreat rate for the 3-year period was 140 mm/year (s_e = 26 mm/year), corresponding to an average volumetric rate of 0.041 m^3/m per year. Applying this to the total length of headcuts in the drainage basin yields 3-year sediment mobilization of 18 t/km^2 per year. Adjusting this for headcuts in scars not connected to channels yields net production (table 2).

Soil Creep

Soil creep on drainage-basin hillslopes is being measured by 102 creep test-pillars consisting of columns of 25-mm-long, 10-mm-diameter wooden dowel segments injected vertically into the soil through a 13-mm-diameter metal tube. Depending on soil depth and stoniness, pillar lengths range from 0.6 to 1.1 m. Creep is measured by excavating a pit next to the pillar and measuring offsets from the original line of insertion, defined by the lowermost pillar segments. Uncertainty in original insertion angle is $\pm$ 0.5°. Note that this technique measures only shallow soil creep and can yield only a minimum estimate of creep rate.

One pillar in forest and one in grass were excavated in 1974, 2 years after insertion. Rates at both sites are similar, but results are ambiguous because of the half-degree insertion uncertainty. They averaged a minimum of 0.5 and a maximum of 1.1 mm/year over a depth of 0.4 m, corresponding to volumetric rates of 0.0002 and 0.0004 m^3/m per year, respectively. Applying these maximum rates to all slopes bordering channels in the drainage basin yields a yearly effective mobilization of 5.3 to 12.2 t/km^2 (see note, table 2). Because creep rates are certainly not uniform over the drainage basin, the error in this estimate is large.

In grassland, creep caused by vole burrowing is locally important, especially in swales. Although this creep affects only the upper 0.15 m of soil, measurements of tunnels and debris piles yield volumetric rates of 0.0001 to 0.006 m^3/m per year (Lehre 1981). Assuming 10 percent of the grassland is burrowed (visual estimate), and using an average rate of 0.002 m^3/m per year, about 3 t/km^2 per year is mobilized by voles.

Hillslope Sheet Erosion

I estimated sheet erosion on hillslopes by means of 53 erosion pins installed on both well-vegetated and poorly vegetated grassland hillslopes. No pins were installed in brush or forest. Over 3 years, most pins showed either no erosion or slight net deposition; the pattern of erosion and deposition suggests hillslope sheet erosion takes place as rainsplash-driven creep of a layer about 1.5 mm thick (Lehre 1981). To calculate the corresponding sediment mobilization, I have assumed the sheet moves downslope 0.1 m/year--a volume flux of 0.000 15 m^3/m per year. Applying this rate to all slopes bordering channels yields an effective mobilization estimate of 4 t/km^2 per year (see note, table 2). Although this rate is certainly incorrect in detail, I believe it reflects the proper order of magnitude.

Gully Scarp Erosion

Erosion of gully side-scarps and walls, excluding headcuts, was measured with 133 stake sets arrayed about 12 gullies. Average retreat of raw gully walls for the period 1971-73 was 101 mm/year (s_e = 12 mm/year); that of vegetated walls was 37 mm/year (s_e = 6 mm/year). Corresponding volumetric rates are 0.033 m^3/m per year and 0.0096 m^3/m per year. These rates are significantly different at the 5-percent level. Assuming that the ratio of raw to vegetated walls in this sample is representative for the drainage basin, the combined weighted average rate of 0.013 m^3/m per year multiplied by the total length of gully walls in the drainage basin yields a total mobilization and production of 833 t for 1971-74, or 160 t/km^2 per year.

MLTC Bank Erosion

Bank erosion on MLTC could not be measured directly. The height (average 2.5 m) and indefinite edge of most of its banks prevented use of scarp-retreat stakes, and monumented cross sections were prohibited by limitations of time, staff, and equipment. Instead, bank erosion was estimated by comparison of two traverses of the channel, one at the beginning and one near the end of the study. During the first traverse, all bank slides were mapped. In the second traverse, all slides were again mapped, and volumes of new sliding determined. I also examined channel banks for evidence of fresh spalling, sliding, and undercutting (e.g., vegetation and moss trim lines; exposed roots; overhanging root mats, raw grooves) and estimated the volumes removed. On this basis, the average volumetric erosion rate of MLTC banks is 0.011 m^3/m per year, and the total 1971-74 erosion is 82 t. I believe this estimate is within 50 percent of the true amount.

Table 3--Sediment and dissolved-solids yield of Lone Tree Creek. See text for discussion.

	1971-72	1972-73	1973-74	Average 1971-74	Estimated long-term average
	Tons/km^2			Tons/km^2 per year	
Suspended load	17	1227	576	607	180
Bedload	7.1	193	54	85	34
Total particulate	24	1420	630	692	214
Dissolved load	16	50	48	38	23
Total denudation	40	1470	678	730	237

Bed Erosion

The largest uncertainty in this study is the amount of sediment that either entered or was removed from bed storage by aggradation and degradation of channels within the drainage basin. Ideally, these amounts could be determined by surveying a large number of monumented cross sections spaced at regular intervals along MLTC, SOBR, and major tributaries. This was impossible in the study. As a poor expedient, I have assumed that erosion and deposition measured at the MLTC and SOBR gaging sites are applicable to all alluvial reaches of those channels. Although this is certainly in error (see Lehre 1981 for discussion), the resulting estimates may suggest the proper order of magnitude. With this proviso, total 1971-74 bed erosion on MLTC and SOBR amounts to 192 t.

SEDIMENT YIELD AND DISSOLVED LOAD

Suspended Load

Suspended-sediment concentrations measured in Lone Tree Creek range from 1 to 16 000 mg/liter. Concentrations during storms are typically 200 to 1000 mg/liter; after storms they drop rapidly to 10 to 30 mg/liter or less. Silt and clay make up over 50 percent of the sample, even at the highest sediment concentrations. Sediment concentration is controlled chiefly by rate of supply from extrachannel erosional processes, particularly sheet erosion in slide scars and on gully walls, bank spalling and caving, and slides and debris flows (Lehre 1981). Total suspended load discharge from the drainage basin during 1971-74 was 3157 t, or 607 t/km^2 per year (table 3).

Bedload

Where sediment is present, the bed of Lone Tree Creek is gravel or boulders; mean bed-particle diameter (d_{65}) ranges from 28 to 142 mm in reaches near the gaging sites. By contrast, mean diameter (d_{65}) of material in my bedload-transport samples ranged from 3.5 to 18 mm, and 25 to 70 percent by weight consisted of particles less than 4 mm. This fine material has two sources: part of it resides in the bed beneath a surface armor layer; the remainder is directly contributed to the flow by slides, gullies, and bank erosion. Measured unit-transport rates are about 0.01 to 0.10 kg/second per meter of channel width.

Bedload discharge was computed from a rating curve which, though defined by only five points, yields estimates of Q_b at high discharge (more than 4 m^3/second) that are only 15 percent lower than those predicted by the Meyer-Peter and Müller (MPM) equation (Vanoni 1975). At lower flows, the equation overestimated bedload transport by 3.5 to 15 times.

Total bedload discharge for the period 1971-74 was 422 t (85 t/km^2 per year), or about 14 percent of the suspended-load discharge (table 3). I believe the bedload estimates are within $\pm$ 50 percent of the true bedload discharge.

Dissolved Load

The most distant part of the Lone Tree drainage basin is less than 2.4 km from the ocean, suggesting that "cyclic salts" (Janda 1971) carried inland by wind and rain may be an important constituent of the dissolved load. Janda (1971, table 2) found that nondenudation components accounted for 55 percent of total dissolved load in two coastal streams in San Mateo County. I also obtained a value of 55 percent for nondenudation components in a water sample from Pike County Gulch, a small coastal drainage basin similar to Lone Tree Creek, but 8 km north. I assume this percentage applies to the Lone Tree drainage basin as well. For the period 1971-74, I computed a total denudation dissolved load of 196 t (37.5 t/km^2 per year), or about 6 percent of total denudation in the drainage basin (table 3).

Table 4--Sediment mobilization on slopes, production to channels, and yield (discharge) from drainage basin for years 1971-74 on Lone Tree Creek. See text for discussion.

Year	Rainfall	Recurrence interval of peak flow	Mobilization on slopes (1)	Production to channels (2)	Redistribution on slopes (3)	Yield: bed + susp. load (4)	bed + bank storage (5 = 2 - 4)
	Millimeters	Years	Metric tons km^2				
1971-72	602	1.5	86	148	-71	24	+124
1972-73	1184	15-20	1219	985	+317	1420	-435
1973-74	1046	3-5	1960	1575	+389	630	+945
1971-74	2832	--	3265	2708	+635	2074	+634

Long-Term Estimates

"Long-term" estimates of suspended-sediment, bedload, and dissolved-solids yields (table 3) were computed by using their average 1971-74 rating curves in conjunction with an "average" duration curve synthesized from a dimensionless regional flow-duration curve for Marin County streams (Lehre 1974, fig. 12; Lehre 1981). These estimates assume that the present period is representative of past conditions. If landslide frequency has increased in the last 100 to 150 years (see discussion), this assumption is invalid.

SEDIMENT BUDGET

Reconciliation

The third and final step in construction of a sediment budget is setting up a balance sheet showing mobilization, production, sediment yield, and storage changes. I have done this in table 4, using the data in tables 2 and 3.

Table 4 represents a simplification of the relations shown in figure 2. Storage elements are lumped simply into hillslope and bed-and-bank sites; available data are insufficient for finer discrimination, especially in the channel component. Processes included in mobilization on slopes are those listed in the upper (hillslope) half of figure 2. Production to channels includes all material mobilized by the processes listed in the lower (bed and bank) half of figure 2, together with all material mobilized by hillslope processes that reached or had access to a channel. This estimate of production assumes that all material mobilized by scarp and sheet erosion in slide scars connected to the channel system is delivered to the channels. In fact, some unknown proportion of this sediment remains behind in the scar.

In calculating production, creep and hillslope sheet erosion are not explicitly included. Material moving downslope by these processes enters the channel system through bank and gully-wall erosion and is accounted for there (table 2).

Redistribution on slopes is the difference between the total amount of sediment mobilized by slope processes and that produced. This represents the amount of material moved from one position of hillslope storage to another farther downslope. Negative values indicate transfer of sediment to bed and bank storage exceeds new mobilization on hillslopes. This implies growth of gullies and scars at the expense of the slope.

The difference between production to channels and sediment yield (transport out of drainage basin) gives the change in bed and bank storage. Negative values of this quantity indicate that removal of material from bed and bank storage exceeds resupply from hillslope processes. Positive values represent new accretions to storage (from slopes above) and possibly redistribution of material already in bed and bank storage.

Discussion

The sediment budget of table 4 demonstrates convincingly that in dry years and in wet years without extreme flow events most of the sediment mobilized goes into storage, chiefly on the lower parts of slopes and in channel and gully beds and banks. Large net removal of sediment from storage occurs in flow events with recurrence intervals greater than 10 to 15 years. Process measurements suggest a reason for this. In an "average" or a dry year, gully-wall erosion and slide-scarp retreat are the chief mobilizers of sediment (table 2). Most of this material is moved by spalling, local sliding, or rainbeat, and is not transported far before being deposited and returned to storage. Only sheet erosion in slide scars and, to a lesser extent, on gully walls and gully headcut retreat can effectively entrain and remove sediment; these are, in fact, the main sources of suspended load in such years. By contrast, in years with extreme rainfall and flow events (such as 1973), debris slides and flows are responsible for over 80 percent of all sediment mobilized and between 55 and 65 percent of all sediment produced. In these years, gully-wall, bank, and bed erosion are also much increased. Because nearly all of this material is contributed directly to the channel during very high flow, when the stream has the capacity to transport it, most of it is carried out of the drainage basin.

The importance of very large flows in preventing sediment from re-entering storage can be seen by comparing 1973 and 1974. In 1973, a year with a 15- to 20-year flow event, about 1550 t were contributed to channels by slope processes (55 percent from debris flows) and measured bank erosion, yet 2465 t were discharged from the drainage basin. The difference must have come from unmeasured bed and bank erosion. In 1974, a year with only a 3- to 5-year flow event, about 2720 t were contributed to channels, 78 percent came from two large landslides prepared for failure by undercutting the previous year. Only 1090 t were discharged from the drainage basin, however, primarily because the peak flows were insufficient to transport the amounts of sediment supplied (Lehre 1981).

Ideally, a sediment budget should specify the magnitude and frequency (recurrence interval) of sediment mobilization by each process and the average residence time of sediment in each storage element. These are difficult questions; their general answer requires both longer records of process activity, and better quantitative models of relations between process activity and driving forces and of movement of sediment through storage elements, than are currently available. My data do, however, provide insights into some of the problems of process relations and residence time in swales.

Debris slides and flows are the single most important erosional agent in the drainage basin, accounting for at least 53 percent (112 t/km^2 per year) of its "long-term" particulate yield (214 t/km^2 per year). The magnitude and frequency of sliding are thus clearly important to understanding sediment routing in the drainage basin, but their determination is a vexing and yet unsolved problem.

Recurrence intervals of rainfall events (or, as a surrogate, flow events) and sliding are not simply related for four reasons. First, extreme events in 1 year (e.g., 1973) may undermine or weaken slopes so that they fail in subsequent smaller events (e.g., 1974). This destroys the independence of events commonly assumed in recurrence interval calculations. Second, a particularly large rainfall event may evacuate a large proportion of the sediment in storage in swales, thus reducing the amount of sediment available for mobilization in subsequent large storms. Third, as the colluvial fill in a swale accumulates, thickens, and weathers, its probability of failure in a storm of a given size increases. The volume of slides produced by a storm of specified frequency will thus depend not only on storm magnitude, but also on the filling history of the swales. Finally, changes in drainage-basin vegetation from fire, storm, land use, or natural succession may markedly alter the susceptibility of slopes to failure, and thus change magnitude-frequency relations. Until these problems can be adequately dealt with, landslide "recurrence intervals" will not be particularly meaningful.

Slide scars fill through creep and scarp erosion. Creep and hillslope sheet erosion supply sediment to slide scarps at a maximum rate of 0.0009 m^3/m per year; scarps, however, are eroding at 0.0054 m^3/m per year and thus must be wearing back at the expense of the slope. Assuming then that scars fill primarily by scarp erosion, that all sediment so mobilized remains within the scar, and that the linear rate of scarp retreat is constant (20 mm/year), simple calculations (Lehre 1981) suggest a minimum of 540 years are required to fill a 2.5 m deep scar to half its depth. Reasonable changes in assumptions increase healing time by a factor of 5 to 10.

Continued maintenance of colluvial fills in swales demands that long-term rates of filling by creep and emptying by slides be in equilibrium. Density of swales is 129/km^2; assuming a "typical" swale length of 60 m and a 0.6-m-thick moving mantle, a creep rate of 13 mm/year is required to maintain equilibrium with the "long-term" slide rate of 187 t/km^2 per year (Lehre 1981). Observed creep rates are 0.75 to 1.5 mm/year over a depth of 0.4 m, suggesting colluvium is currently being stripped. For the colluvial fills to have accumulated, then, past creep rates were either 10 times higher than at present, or rates of landslide erosion were 10 times lower. This hypothesis will be more rigorously tested as additional slide and creep data are collected.

Under the assumption, used to calculate the long-term slide rate, that all raw and recovering slides are less than 100 years old, the frequency of slide occurrence in the past century is 226/100 years. Assuming that all 94 healed slides are between 100 and 540 years old, their frequency of occurrence is about 21/100 years, suggesting a tenfold increase in slide frequency in the past century. (Note that increasing the age of the oldest recognizable healed slide, or decreasing the age of the oldest raw/recovering slide, only increases the disparity in rates. To have about equal frequencies, the raw and recovering slides would have to have ages of 0 to 250 years, and healed scars would be 250 to 350 years old. These ranges do not seem realistic, particularly for healed slides, which may actually range in age from less than 100 to more than 550 years.)

Taken together, creep rates and slide numbers strongly suggest that landslide frequency has increased tenfold in the past 50 to 150 years, and that sliding is currently mining relict colluvium formed in equilibrium with creep rates about equal to current creep rates. The timing of this increase corresponds closely with the introduction of cattle to the drainage basin, and the resultant conversion of the grasslands from the longer rooted, perennial, native bunch grasses to shallower rooted, annual, introduced grasses. This, combined with trampling and heavy grazing, may have been sufficient to weaken slopes already near the limits of stability.

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LITERATURE CITED

Blake, M. C., J. A. Bartow, V. A. Frizzell, J. Schlocker, D. Sorg, C. M. Wentworth, and R. H. Wright.
1974. Preliminary geologic map of Marin and San Francisco Counties, and parts of Alameda, Contra Costa, and Sonoma Counties, California. U.S. Geol. Surv. Misc. Field Stud. Map MF-574.

Crippen, J. R., and R. M. Beall.
1971. A proposed streamflow data program for California. U.S. Geol. Surv. Open-File Rep. 46 p. Menlo Park, Calif.

Dietrich, W. E.
1975. Sediment production in a mountainous basaltic terrain in central coastal Oregon. M.S. thesis. Univ. Wash., Seattle. 81 p.

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

Dunne, T., T. R. Moore, and C. H. Taylor.
1975. Recognition and prediction of runoff-producing zones in humid regions. Hydrol. Sci. Bull. 20(3):305-327.

Emmett, W. W.
1965. The Vigil Network: methods of measurement and a sampling of data collected. Int. Assoc. Sci. Hydrol. Publ. 66:89-106.

Guy, H. P.
1969. Laboratory theory and methods for sediment analysis. U.S. Geol. Surv. Tech. Water Resour. Invest. Book 5, Ch. C1, 58 p.

Guy, H. P., and V. W. Norman.
1970. Field methods for the measurement of fluvial sediment. U.S. Geol. Surv. Tech. Water Resour. Invest. Book 3, Ch. C2, 59 p.

Janda, R. J.
1971. An evaluation of procedures used in computing chemical denudation rates. Geol. Soc. Am. Bull. 82(1):67-80.

Lehre, A. K.
1974. The climate and hydrology of the Golden Gate National Recreation Area. Report to U.S. Natl. Park Serv., Golden Gate Natl. Rec. Area. 61 p. San Francisco, Calif.

Lehre, A. K.
1981. Sediment mobilization and production from a small Coast Range catchment: Lone Tree Creek, Marin Co., California. Ph.D. thesis. Univ. Calif., Berkeley. 370 p.

Morrison, P. H.
1975. Ecological and geomorphological consequences of mass movements in the Alder Creek watershed and implications for forest land management. B.S. honors thesis. Univ. Oreg., Eugene. 102 p.

Pierson, T. C.
1977. Factors controlling debris-flow initiation on forested hillslopes in the Oregon Coast Range. Ph.D. thesis. Univ. Wash., Seattle. 166 p.

Rantz, S. E.
1967. Mean annual precipitation-runoff relations in north-coastal California. U.S. Geol. Surv. Prof. Pap. 575-D:D281-D283.

Rantz, S. E.
1971. Mean annual precipitation and precipitation depth-duration-frequency data for the San Francisco Bay Region, California. U.S. Geol. Surv.-HUD San Francisco Bay Region Environ. and Resour. Plann. Study Basic Data Contrib. 32, 23 p. Menlo Park, Calif.

Rantz, S. E.
1974. Mean annual runoff in the San Francisco Bay Region, California, 1931-1970. U.S. Geol. Surv. Misc. Field Stud. Map MF-613.

Rice, R. M., E. S. Corbett, and R. G. Bailey.
1969. Soil slips related to vegetation, topography, and soil in southern California. Water Resour. Res. 5(3):647-659.

Rice, R. M., and G. T. Foggin III.
1971. Effect of high intensity storms on soil slippage on mountainous watersheds in southern California. Water Resour. Res. 7(6):1485-1496.

Schlocker, J.
1974. Geology of the San Francisco North Quadrangle, California. U.S. Geol. Surv. Prof. Pap. 782. 109 p.

Swanson, F. J., and C. T. Dyrness.
1975. Impact of clear-cutting and road construction on soil erosion by landslides in the western Cascade Range, Oregon. Geology 3(7):393-396.

Swanson, M., F. J. Swanson, and C. Woods.
1976. Preliminary report of a cooperative study of landslide erosion in the Mapleton Ranger District. 10 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Vanoni, V. A., ed.
1975. Sedimentation engineering. 745 p. Am. Soc. Civ. Eng., New York.

Wahrhaftig, C.
1974. Geology of the Golden Gate National Recreation Area. Report to U.S. Natl. Park Serv., Golden Gate Natl. Rec. Area. 122 p. San Francisco, Calif.

Plant Growth and Block-Field Movement in Virginia

Cliff R. Hupp and Robert S. Sigafoos

ABSTRACT

Tree-ring dating provides information on temporal and spatial aspects of related geomorphic processes. Studies in Massanutten Mountains in Virginia involve effects of mass movement on trees and how these effects can be used to date movement of block fields. Block fields occur below resistant Massanutten sandstone in Passage Creek basin. Slopes are boulder-strewn, and locally vegetation is absent or sparse. Trunk ages yield a minimum age for a geomorphic/hydrologic event. Corrasion scars yield exact year of occurrence. Dates correlate with extreme flows on Passage Creek. Pleistocene frost action may have facilitated original formation of block fields, but some now spread during intense rainfall, though climatic conditions are different from those at formation.

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INTRODUCTION

Dendrochronologic methods have had wide use in dating occurrence and determining rates of both hillslope and fluvial geomorphic events and processes. These methods are a valuable tool in quantifying some erosion processes and sites of sediment storage in forested areas. Most work of this type has occurred where geomorphic activity is conspicuous. Here, we examine the character of more subtle geomorphic processes, which can locally provide a source of coarse particles for streambed armouring and influence bedload characteristics downstream of block-field areas.

The block fields were identified in the field and on aerial photographs. The primary site is in the northern part of the Massanutten Mountain Range near Passage Creek in northwestern Virginia (fig. 1). The range is a syncline underlain by resistant Massanutten sandstone consisting of white quartz sandstone and quartzite, locally conglomeratic. The Martinsburg shale formation occurs immediately below, and many block fields have formed at this contact (Rader and Biggs 1976). They have slopes between 32 and 42 degrees, with boulders ranging from several centimeters to a few meters in diameter (fig. 2). The slopes are a mosaic of block fields and forested patches; the predominant forest species are chestnut oak (Quercus prinus L.), cherry (Prunus serotina Ehrhart), black gum (Nyssa sylvatica Marshall), birch (Betula lenta L.), American ash (Fraxinus americana L.), eastern hemlock (Tsuga canadensis (L.) Carr.), American basswood (Tilia americana L.), and black locust (Robinia pseudoacacia L.). Common subcanopy species are dogwood (Cornus florida L.), shadbush (Amelanchier arborea Fernald), ironwood (Carpinus caroliniana Walter), witch hazel (Hamamelis virginiana L.), and pawpaw (Asimina triloba (L.) Dunal). Taxonomy follows Radford et al. (1968).

Trees growing on, near, or below the block field provide evidence of recent movement. Movement can be dated through tree-ring analysis of leaning trees with adventitious sprouts, stem deformations, and corrasion scars. Tree ages on new deposits provide a minimum time since deposition, and tree ages above the block field but below the outcrop indicate the minimum time since relative stability was reached.

Some geologists believe that block fields or scree in the central Appalachians are Pleistocene relicts and not now subject to frequent movement (Hack, J. T., U.S. Geological Survey, Reston, Virginia, personal communication). Block fields occur downslope from resistant sandstone outcrops from which individual blocks were probably first formed by frost action during the Pleistocene. The block field studied is currently spreading downslope, and the time of movement correlates with recorded floods.

Block fields are a common feature of the Massanuttens. Although some appear to be stable, most have vegetative indications of movement. Analyzing the process by which the block fields move poses many problems; however, our study suggests moving water is a prime factor.

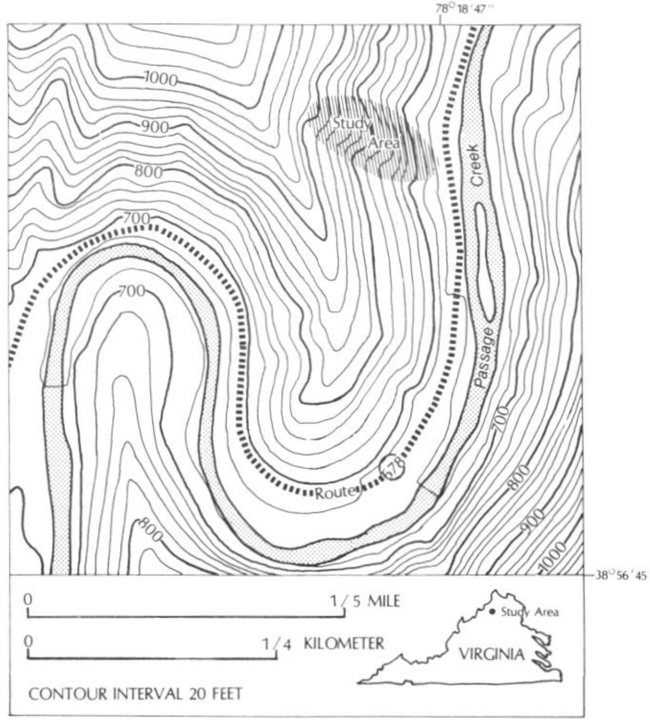


Figure 1.--Part of the Massanutten Mountains, Virginia, showing study area.

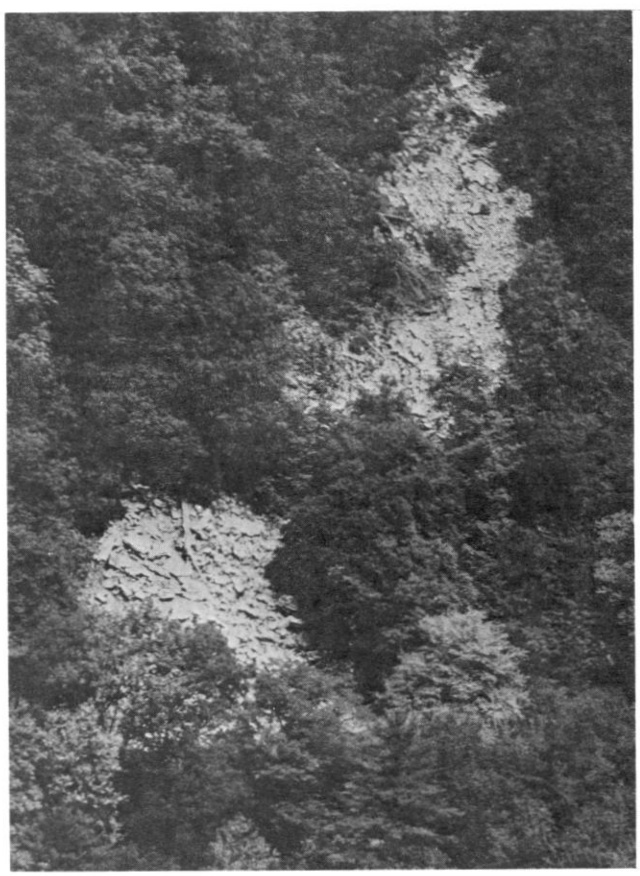


Figure 2.--The block field studied. Visible part of fallen log is 6 m long.

METHODS

General Principles

The basic principles of tree-ring dating have been outlined extensively (Sigafoos 1964, LaMarche 1968, Alestalo 1971, Sigafoos and Hendricks 1972, Shroder 1978). Slope movement and rockfall can have many effects on woody vegetation, including partly felling trees, scarring stems, and creating bare areas where plants can become established. Tree-ring analysis allows for fairly rapid and reliable dating of recent geomorphic activity (Alestalo 1971). The ability of individual trees to recover from damage and still record the event in their annual-ring chronology provides a valuable research tool for geomorphic studies. Use of dendrochronology in dating geomorphic phenomena should shed light on some incompletely understood processes.

Tree trunks grow radially during spring and early summer. Wood formed in spring is different from wood formed the previous summer. This difference produces an annual ring, so each year's growth can be identified and dated. Some species have ring morphologies that allow for ease of measurement. These trees are "ring porous"; that is, the wood that forms in the spring has large pores that form a conspicuous ring when viewed in cross section. Others are "diffuse porous," with uniform pores throughout the annual ring. In dating geomorphic events, one does not have the luxury of picking select species and individuals. Land-use history, age of stand, life history of the species, and diffuse-porous species all pose problems. On the slopes we studied, two common species (birch and black gum) exhibit the effects of mass movement in their deformations, but because both are diffuse porous, the annual-growth increments are difficult to see. Some species also have extra or missing rings. Considerably more care has to be exercised in ring-boundary determinations and ring counts than would be necessary if we could use only selected species (Phipps, R. L., U.S. Geological Survey, Reston, Virginia, personal communication).

Eccentric annual growth can be seen in stem wood where one side of the trunk produces more wood than another. Should this occur for several years, the center of the section or core appears to be skewed to one side of the trunk, or in other words, radii in opposite directions will be unequal in length. Eccentric growth can result from an unequal direction of incident light (a phototropic response) or tilting (a geotropic response). Determination of onset, magnitude, and duration of eccentric growth can indicate the date, direction, and duration of a geomorphic event. The cause of the eccentric growth must be ascertained, however, before this technique is reliable (Phipps 1974). Angiosperms--deciduous, broad-leaved flowering plants--produce thicker annual rings on the uphill side; gymnosperms--coniferous plants--produce thicker rings on the downhill side. Reaction-wood formation producing eccentric rings is often difficult to analyze in angiosperms.

Severe tilting can cause adventitious sprouting along the parent trunk, and such sprouting is common in angiosperms. Sprouts from trunks of tilted trees are found often around block fields and along stream courses. Because sprouting occurs rapidly after a geomorphic or flood event, the age of the sprout will date that event to the year (Sigafoos 1964).

The corrasion of tree stems by rockfall damages the cambium (wood-producing region). This can result in termination of radial growth where the tree was struck. In subsequent years, the scar will be increasingly covered by callus growth until the damaged area is completely covered and the cambium is once again continuous around the trunk (fig. 3). Cross sections through the scarred portion of a trunk can yield the exact year of damage and often the season. LaMarche (1968) and Shroder (1978) have used this technique to date active slope movement. Increment cores taken at different angles on either side of a scar make dating possible without destroying the tree (fig. 4).

Increment cores taken at the base of a tree growing on a landform can yield the minimum time lapsed since the landform assumed some measure of stability. Sigafoos and Hendricks (1961, 1972) used this technique extensively in dating glacial deposits. Serial age determination of trees downslope on a block field from head to leading edge may yield an approximate rate of downslope movement. We are currently testing this procedure.

Methods Used

All of the techniques outlined in the preceding paragraphs were employed on the studied block-field slope. Figure 5 illustrates samples of three major tree-ring techniques used in this study.

Increment cores for age determination were taken from 39 sprouts growing from inclined trunks (fig. 6). Cross sections or cores from 55 tree stems were taken to date scars, and 48 core samples were taken to date the age of the landform surface. These data were then compared to existing streamflow records obtained from the Virginia Water Control Board for the period 1933-78. Dates of high flow were determined and used to ascertain possible correlations of block-field movement.

Field observations of lichen coverage and degree of weathering of the rock were made. Although variations in these characteristics of block fields can support tree-ring dating techniques, we believe they cannot date movement as accurately. In this study, they were used as supportive evidence of movement.

A map of the block field was drawn from aerial and oblique ground photographs and from field measurements with a compass and range finder. Areas of movement on the block field were identified by absence of lichen cover and unweathered surfaces of rock, and movements were dated using the trees below or near the movement.

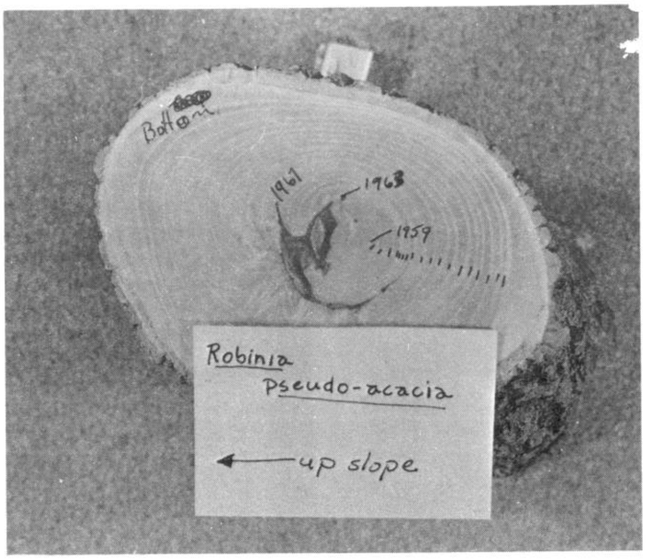


Figure 3.--Multiple scarring and annual callus growth in black locust.

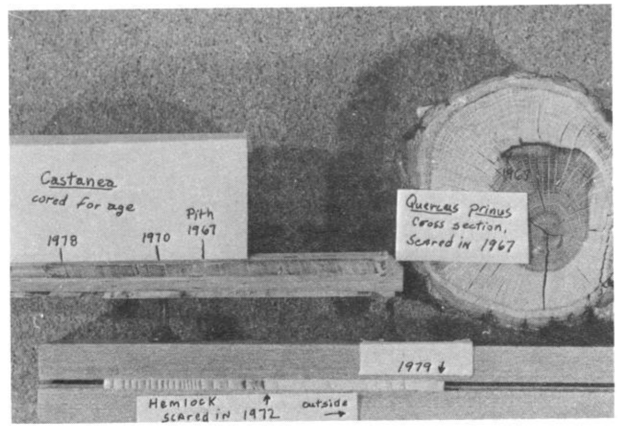


Figure 5.--Examples of three types of information obtained by tree-ring analysis. Tree ages may be determined from increment cores (upper left: chestnut sprout from a parent damaged by a rock fall). Damage date may be determined from increment cores (bottom: hemlock, cored to date scar formation, 1972) and from cross sections (upper right: chestnut oak, sectioned to date scar, 1967).

Figure 4.--Growth of callus wood indicates older scar on chestnut oak. Core samples from just outside calus can date the scar.

Figure 6.--Adventitious sprouts from severely tilted parent. Age of the sprouts can indicate time since tilting. Note core hole in center sprout.

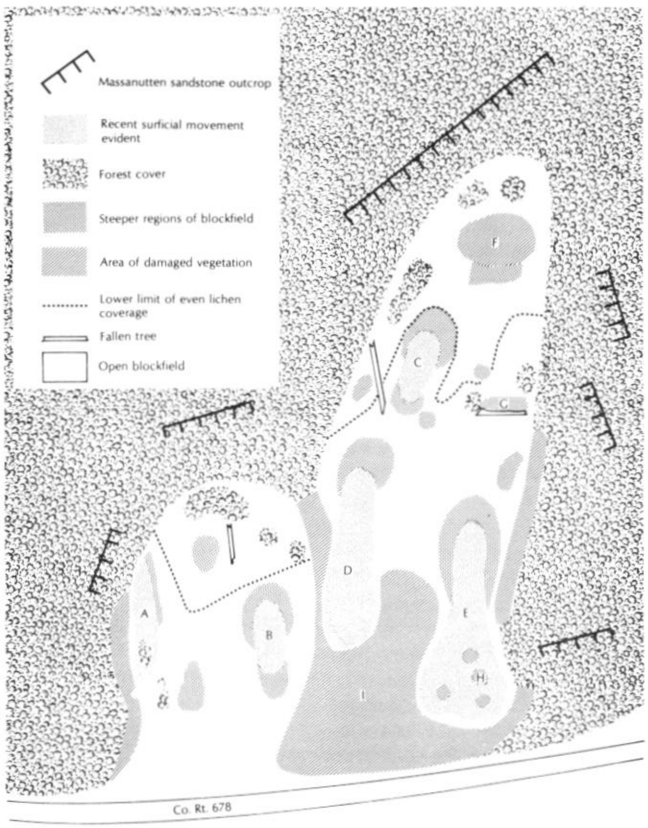


Figure 7.--Map of block field:

A. and B. Depression in block field with unweathered smaller boulders. A. has tree-ring evidence of movement in 1979 and 1972.
C. Recent slope failure. Tree-rings date movement in 1967.
D. Recent slope failure. Tree-rings date movement in 1967, 1972, and 1979.
E. Recent slope failure. Tree-rings date movement in 1942, 1954, 1967, 1972, and 1979. Smaller boulders at lowest portion may move with every substantial rain event.
F. Pawpaws growing in block field. Scars on stem below surface do not correspond to flood events.
G. Fallen tree damming debris upslope, with depression downslope.
H. Severely damaged vegetation. Boulders subject to frequent movement during most rain events.
I. Scattered boulders among vegetation with extensive tree damage.

RESULTS

The map of the block field (fig. 7) shows where recent movement was indicated by either vegetation or fresh rock exposure. Boulders in the most active lower portion are considerably smaller than those upslope, ranging from 10 to 25 cm in diameter, with a few larger. In depressions of recent slope failures (C and D, fig. 7), boulders ranged from 20 to 35 cm in diameter. In the relatively stable upper portions of the block field (area of even lichen coverage, fig. 7), the boulders are generally greater than 35 cm. Throughout the block field, isolated very large (up to 3-m) boulders are found. In the depressions and the lowest part of the block field, boulders are found with lichens on the underside and recently exposed upper surfaces.

A fallen log (G, fig. 7) created a debris dam. Downslope from the log, material has been removed, producing a depression in the surface material, while debris was piled against the upslope side of the log. LaMarch (1968, p. 372-373) has used this type of evidence to document movement. An attempt was made to cross-date the log with living vegetation so that time in that position could be ascertained. Fallen timber in the East is not preserved well; a minimum time since death was determined by ring counting to be 25 years, but whether the tree fell the same year it died is not known. The methods of cross-dating are described by LaMarche (1968, p. 351).

One of the 18 pawpaws (F, fig. 7) was excavated to 1.5 m and was visible for another half meter into the block field. All of the excavated portion of the trunk was stemwood, indicating that the point of germination was deeper in the block field. A corrasion scar at 30 cm below the block-field surface indicates internal movement. The surface boulders near the pawpaws were loose but had even lichen coverage, suggesting no recent, rapid, surficial downslope movement.

The trees between the block field and the outcrop ranged from 176 years old to seedlings. Trees are oldest near the outcrop and are progressively younger toward the block field. The age distribution from outcrop to block field indicates that the forest is becoming established in the stable upper part of the block field. The vegetated area below the block field has young trees (fig. 8) of ages that correlate with periods of little tree deformation and no major floods. Many trees are dead or in poor condition, presumably as a result of corrasion from numerous rock falls. Boulders are scattered on the forest floor, commonly lodged on the upslope side of a tree trunk (fig. 9).

Of 152 trees analyzed on this block field, growth of 94 correlated with major floods on Passage Creek (fig. 10, table 1). Not all tree-ring evidence correlated with floods on Passage Creek. Many trees predated the existing flood record or did not indicate a particular year with enough trees damaged to be considered significant (greater than three).

Figure 8.--General view of lower lobe of block field.

Figure 9.--Active lower part of block field. Small trees here exhibit many stem deformations and eccentric growth.

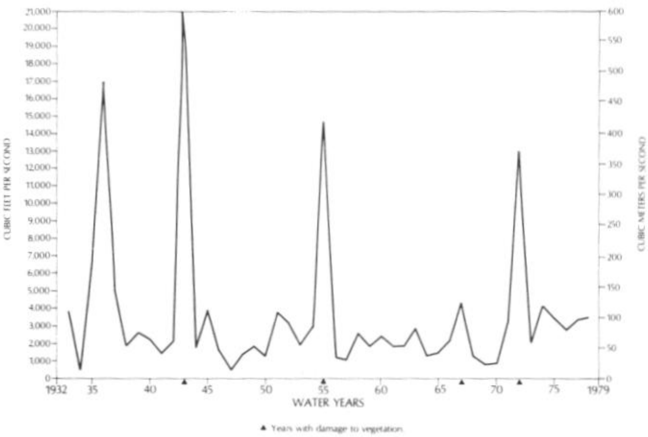


Figure 10.--Annual peak discharges. Passage Creek near Buckton, Virginia (1933-78). A water year is from October 1 through September 30. A flood in October of 1942 would be in water year 1943.

Table 1--Numbers and type of tree-ring evidences corresponding to floods on Passage Creek; of the 58 in nonflood years, 39 predated existing flood record

	Floods on Passage Creek					
Evidence	1979	1972	1967	1954	1942	Nonflood years
Corrasion scars	20	9	19	4	1	12
Sprouts from tilting	0	5	10	5	2	17
Age of main stems	0	4	4	7	4	29
Total	20	18	33	16	7	58

CONCLUSIONS AND DISCUSSION

Our research has demonstrated that some block fields are currently spreading downslope. This movement can be recorded on the trees growing near the block fields, and dates of recent activity can be determined by tree-ring analysis. The correlation of dates of movement with high streamflow suggests that high precipitation is a prime factor in current block-field activity.

Botanical Evidence

Analysis of corrasion scars is the most accurate method of dating block movement. The trees, however, in a relatively short time can obscure outward evidence of corrasion. The method is thus limited to dating more recent events (note the general decline in importance of corrasion scars with age in table 1). Undoubtedly, if older trees were cross sectioned, a longer scar chronology could be developed, but this would kill the trees. Adventitious sprouting from tilted trees (fig. 6) occurs within 1 year after tilting; this sprouting is therefore accurate to within 1 year and is generally easy to identify. Assurance that the tilting is a result of slope failure cannot be made as easily as corrasion is analyzed. Age of trees growing on degraded or aggraded areas can be determined as long as coring is done at the base of the tree. Dating the geomorphic event in this manner yields an estimate rather than an exact date because time to seedling establishment after an event is variable (Sigafoos and Hendricks 1961). Dating older events by tree ages is limited by the life span of the tree species (note the increase in importance of tree age as a dating technique with time in table 1).

Absence of lichens on a block's surface was used in this study to pinpoint areas of recent slope failure, and regions of the block field with even lichen coverage were interpreted as stable or moving slowly as a unit. Lichenometric techniques have not been used in the Massanuttens to date block-field surfaces. Until lichen growth rates are established in this region, reliable estimates cannot be made. The climate in Virginia, compared to colder regions, may increase the rate and variability of lichen growth, and the action of herbivores on the lichen thallus may also be increased. These factos may reduce the reliability of lichenometric dating at this site (Lawry, J. D., Biology Department, George Mason University, Fairfax, Virginia, personal communication).

Movement Mechanism

Block fields in the Massanuttens may have formed as talus below sandstone outcrops or have accumulated above the rock from which they were derived (Hack 1965, p. 32). Traditionally, the block fields have been considered stable relicts of Pleistocene climate, where frost action facilitated their formation.

Our research and that of Hack (1965, p. 32-44) indicate that some block fields are still actively moving downslope. Block fields below outcrops are locally many meters downslope from the outcrop. A reasonable speculation is that separation of the block field from the outcrop indicates a cessation or slowing of frost action on the outcrop, while the increased rainfall of Holocene climate has moved the bulk of the block field downslope from the outcrop. Without the continued supply of material from the outcrop, active block-field slopes may be considered out of equilibrium.

A possible mechanism for current movement would be the buildup of a hydrostatic head within the block field during intense rainfall. Given sufficient slope angle and hydrostatic pressure, parts of the lower end may blow out, producing small debris avalanches or lobes. The more stable upper reaches then shift downward, much less dramatically, adjusting to the removal of material downslope (fig. 11). Eisenlohr (1952) has described "blowouts" of this nature in Pennsylvania. The block field rests on relatively impermeable shale, which would concentrate runoff deeper and lower in the block fields, causing water to accumulate at the base of the block field. The presence of pawpaws, which grow in wet areas, on a south-facing (usually dry) slope such as the studied block field, indicates wetter conditions that would be expected.

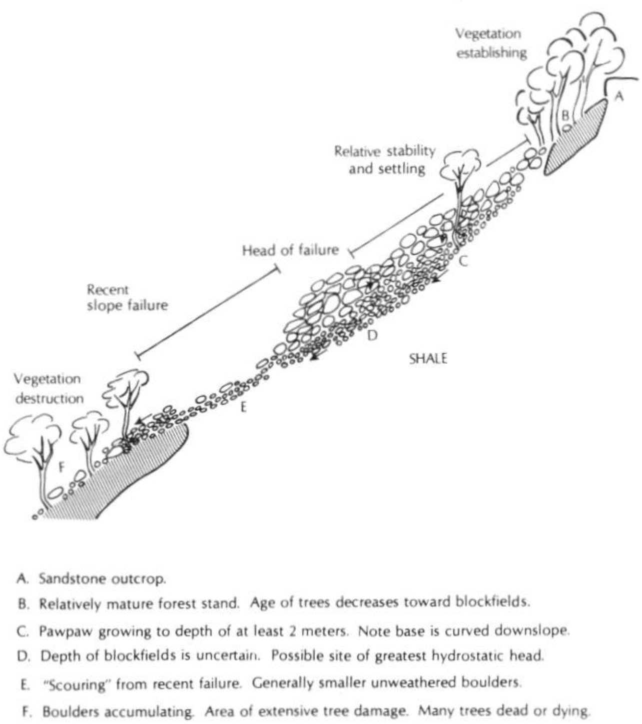


Figure 11.--Cross section of block field:
A. Sandstone outcrop.
B. Relatively mature forest stand. Age of trees decreases toward block fields.
C. Pawpaw growing to depth of at least 2 m. Note base is curved downslope.
D. Depth of block fields is uncertain. Possible site of greatest hydrostatic head.
E. "Scouring" from recent failure. Generally smaller unweathered boulders.
F. Boulders accumulating. Area of extensive tree damage. Many trees dead or dying. Arrows indicate probable direction of boulder movement.

The lack of fine material in the block fields has caused some problems with the idea that a hydrostatic head is created during intense rainfall. J. T. Hack (U.S. Geological Survey, Reston, Virginia, personal communication) believes, as we do, that hydrostatic pressure during intense rainfall is enough to create a blowout, considering the slopes have an angle of 32 to 42 degrees. Smaller blocks, pebbles, and sand are commonly found in or below the most recent blowouts.

Ice lenses in the pore spaces may block runoff, thus increasing the hydrostatic pressure. This process, however, would be most effective during colder seasons. More research is needed before the season of greatest block-field movement can be determined. We have dated movement during October 1942, probably caused by a severe flood. Movement during 1967 correlates with a flood in March, and fresh corrasion scars on trees were found in March of 1979. Corrasion scars older than the flood record indicate damage just prior to early wood formation which suggests winter or early spring activity.

Slope failure from intense storms is common in the Appalachians, often with devastating effects. We believe that the correlation between tree-ring dates of block-field activity and flood records is more than coincidence. Some lack of correlation could result from intense local convectional storms over a block-field area which may not produce general flooding on Passage Creek.

The periodic mass movement of block fields in the Massanuttens may represent a major means of downwasting in these highly resistant areas. The block fields investigated are responding to current climatic conditions different from those at the time of formation (Hack and Goodlett 1960, p. 62-63). Block fields, at least those studied, are an active part of today's landscape, rather than a relict of the Pleistocene.

LITERATURE CITED

Alestalo, J.
1971. Dendrochronological interpretation of geomorphic processes. Fennia 105:1-140.

Eisenlohr, W. S., Jr.
1952. Floods of July 18, 1942 in north central Pennsylvania. U.S. Geol. Surv. Water-Supply Pap. 1134-B, 99 p.

Hack, J. T.
1965. Geomorphology of the Shenandoah Valley, Virginia and West Virginia, and origin of the residual ore deposits. U.S. Geol. Surv. Prof. Pap. 484, 84 p.

Hack, J. T., and J. C. Goodlett.
1960. Geomorphology and forest ecology of a mountain region in the central Appalachians. U.S. Geol. Surv. Prof. Pap. 347, 66 p.

LaMarche, V. C., Jr.
1968. Rates of slope degradation as determined from botanical evidence, White Mountains, California. U.S. Geol. Surv. Prof. Pap. 352-I, p. 341-377.

Phipps, R. L.
1974. The soil creep-curved tree fallacy. J. Res., U.S. Geol. Surv. 2(3):371-377.

Rader, E. K., and T. H. Biggs.
1976. Geology of the Strasburg and Tom's Brook Quadrangles, Virginia. Va. Div. Miner. Resour. Rep. Invest. 45, 104 p.

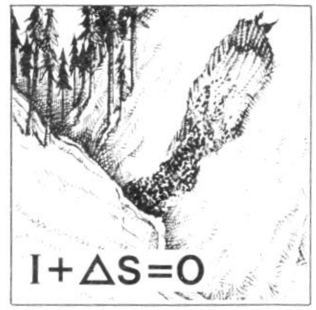
Radford, A. E., H. E. Ahles, and C. R. Bell.
1968. Manual of the vascular flora of the Carolinas. 1,183 p. Univ. N.C. Press, Chapel Hill.

Shroder, J. F., Jr.
1978. Dendrogeomorphological analysis of mass movement of Table Cliffs Plateau, Utah. Quat. Res. 9:168-185.

Sigafoos, R. S.
1964. Botanical evidence of floods and flood-plain deposition. U.S. Geol. Surv. Prof. Pap. 485-A, 35 p.

Sigafoos, R. S., and E. L. Hendricks.
1961. Botanical evidence of the modern history of Nisqually Glacier, Washington. U.S. Geol. Surv. Prof. Pap. 387-A, 20 p.

Sigafoos, R. S., and E. L. Hendricks.
1972. Recent activity of glaciers of Mount Ranier, Washington. U.S. Geol. Surv. Prof. Pap. 387-B, 24 p.



Hillslope Evolution and Sediment Movement in a Forested Headwater Basin, Van Duzen River, North Coastal California

Harvey M. Kelsey

ABSTRACT

The study was made on the 160-km^2 headwater basin of the Van Duzen River. Erosion in steep, forested first-order drainage basins is mainly by debris avalanches, any one of which can encompass a large proportion of the drainage-basin area. After initial deposition of debris in the headwater channel, the rate of subsequent downstream sediment movement depends on the sites of deposition, the occurrence of major climatic events that generate more debris avalanching, and the recovery times of hillslopes affected by avalanching. Residence time of debris on headwater slopes ranges from 15,000 to 50,000 years. Once debris enters perennial stream courses (first-order through third-order streams), sediment transit time out of the study area ranges from less than 10 years to about 5,000 years, depending on whether sediment is temporarily stored in the active channel, in thick channel fills, on strath surfaces isolated by downcutting, or on wide alluviated valley bottoms.

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INTRODUCTION

The purpose of this paper is to discuss sediment movement in the uppermost portion of a steep, forested drainage basin in the Coast Ranges of northern California. The study area is the 160-km^2 upper drainage basin of the 1111 km^2 Van Duzen River basin (fig. 1). The Van Duzen is the northernmost tributary of the Eel River; it flows into the Eel 22 km upstream of its mouth 480 km north of San Francisco.

The Van Duzen and other California north coast rivers have the highest sediment yield of basins of comparable size in the United States (Judson and Ritter 1964, Brown and Ritter 1971). In other papers (Kelsey 1977, 1978, 1980), I have discussed in detail the major geomorphic processes that operate in the Van Duzen River and I have presented a quantified sediment budget for the Van Duzen above the U.S. Geological Survey gaging station that covers the upper 575 km^2 of the drainage basin. Neither the measured sediment yield at the gaging station nor the sediment budget for the basin, however, adequately describes in detail the processes of erosion and deposition in any one part of the drainage basin. This paper describes hillslope erosion, sediment movement in channels, and landscape evolution in the uppermost part of the Van Duzen. The small drainage area and fairly homogeneous bedrock create a nearly ideal study site that is not complicated by many geomorphic processes and numerous landform types.

GEOLOGIC AND CLIMATIC SETTING

The Van Duzen drainage basin drains rocks of the Franciscan assemblage (Bailey et al. 1964), which is a Mesozoic to early Cenozoic accumulation of folded, sheared, and faulted continental margin deposits. Geologic mapping (Kelsey and Allwardt 1975) shows both melange and sandstone units occur in the lower two-thirds of the drainage basin, but in the upper basin study area the principal rock types are slightly metamorphosed, competent graywacke sandstones or interbedded sandstones and siltstones. Although I will not discuss the lower drainage basin in this paper, the presence there of Franciscan melange as well as Franciscan sandstone is a notable distinction from the study area. Franciscan melange hillslopes range in morphology from smooth, undulating grassland or grass-oak woodland slopes to hummocky, boulder-strewn, poorly drained grasslands sculpted by creep or earthflow landslides. Forested sandstone and siltstone slopes of the study area, by contrast, are smooth and straight (fig. 2). The highly sheared and faulted character of these Franciscan rocks, in combination with their present tectonic setting and regional climate, are the cause of the high rates of erosion.

Distinctive landforms testify to recent tectonic uplift of the area. These landforms include narrow, deeply cut river canyons incised below more moderately dipping upper slopes; abundant young strath terrace surfaces elevated above the major rivers; and actively downward and headward cutting steep, headwater channels.

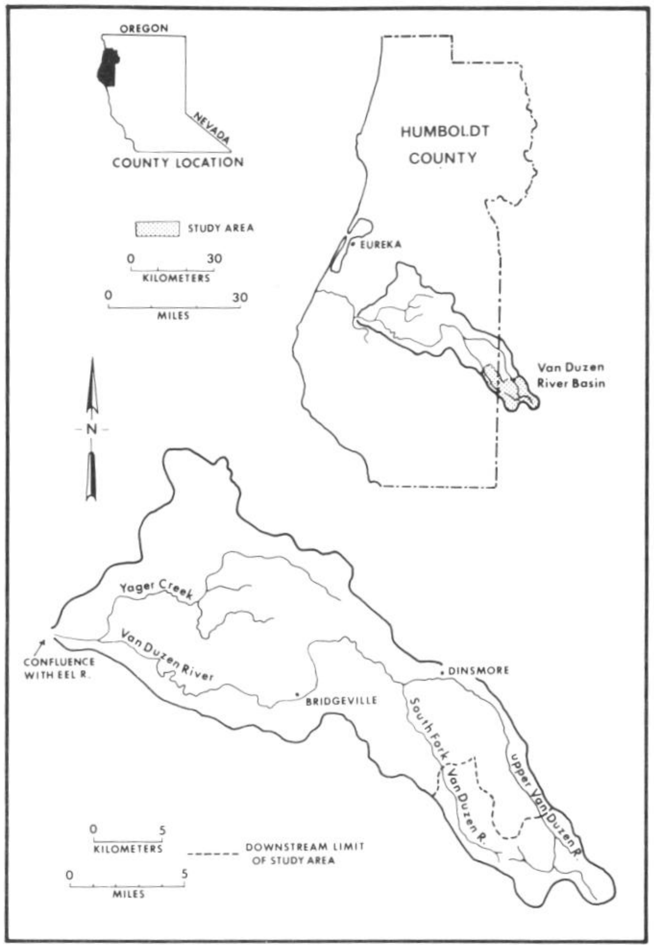

Figure 1.--The 1111-km^2 Van Duzen basin, showing the 160-km^2 study area in the upper portion of the basin.

Figure 2.--The steep, forested study area in the headwaters of the South Fork Van Duzen drainage basin (right). Typical melange terrain, vegetated with grassland and grass-oak woodland (left). Slope morphology on the two terrain-types is controlled by notably different geomorphic processes.

The climate is characterized by high annual rainfall (125-250 cm) that occurs mostly from October through April; the majority of the sediment transport occurs each winter during the two to six most intense storms. Infrequent high-intensity storms of long duration, which are major sediment-transporting events, recur every 100-500 years. One such storm, which occurred in December 1964 (Waananen et al. 1971), caused widespread landsliding and channel aggradation in numerous drainage basins in the California northern Coast Ranges, including the Van Duzen; the 1964 storm serves as a major focus of this study.

The two drainages in the study area are the upper Van Duzen River and the upper portion of the South Fork Van Duzen River (fig. 1). These rivers have classical concave profiles with steep, narrow headwater drainages. At their downstream ends, both rivers have wide alluviated valleys. A distinguishing geomorphic feature is that both rivers are perched above the lower drainage basin of the Van Duzen by a prominent knickpoint that forms a baselevel midway down the drainage basin. An erosionally resistant channel of Franciscan melange boulders from lower drainage-basin hillslopes causes this knickpoint. The knickpoint impedes downcutting of the upper drainage-basin rivers, which effectively maintains the wide alluviated valley at a midbasin location (Kelsey 1980).

LANDSCAPE FORMATION BY DEBRIS AVALANCHING

Rapid, episodic, and shallow debris avalanches are the main type of slope failure in the steep headwater basins of the study area. The avalanches scour down into fractured and usually partially weathered bedrock; avalanche thickness ranges from 1.0 to 4.5 m, and averages about 2.5 m. The failures leave a raw, unvegetated slide scar that may take as much as a century to revegetate because all soil is removed (fig. 3). Parent material is predominantly bedded metasiltstones which become fluid upon failure. The avalanches occur on mid or upper slopes and coalesce into channel-confined debris torrents as the material moves downslope (Swanston 1970). The avalanche scar generally has a straight profile below the crown scarp. Little colluvium remains perched on the scar, and most of the coarse debris initially deposited in the headwater channel moves further downstream in a more gradual manner by fluvial transport.

The straight profile of forested slopes in the study area, the lack of extensive fluvial dissection of these slopes, and the substantial amount of landsliding that occurred as a result of the 1964 storm (Kelsey 1980), all strongly suggest that debris avalanching is the mode of landscape formation on forested sandstone slopes of the study area. During the intense storm of December 1964, many of the slopes showing the most obvious evidence of geologically young, prehistoric sliding did not fail; most of the 1964 failures were on older slopes. The one large (0.34 km^2), raw, prehistoric landslide scar (fig. 3) and the

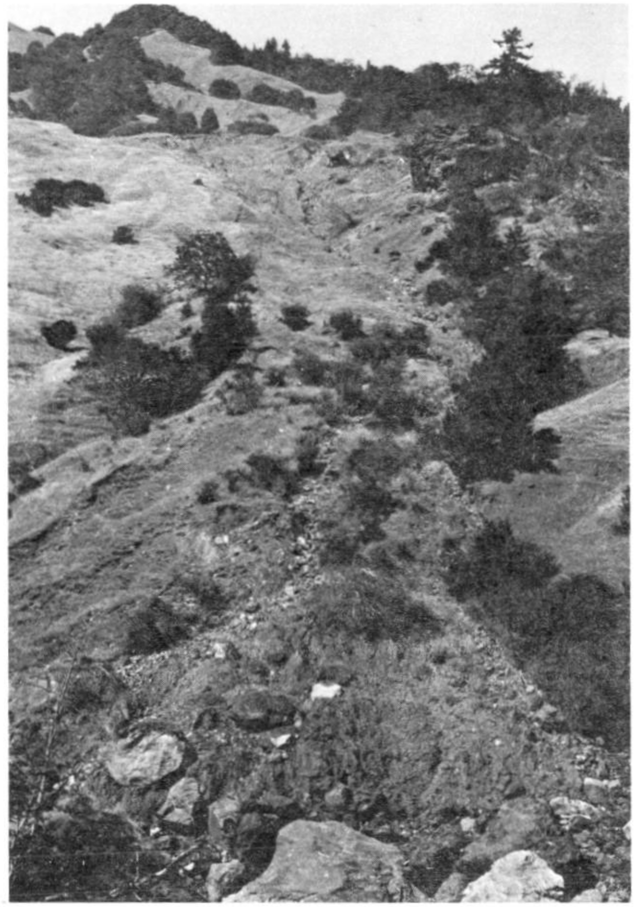

Figure 3.--Mule Slide (0.34 km^2) in the headwaters of the Van Duzen River. This debris avalanche failed in prehistoric time--possibly around 1790, based on an aggradation event that occurred 3-4 km downstream at about then. Note the bare patches along the tributary immediately downstream of the avalanche scar, which attest to the debris torrent that was generated by the avalanche. This tributary has a flat valley floor, 45-90 m wide, that is filled with alluvium from the debris torrent. Despite the fact this avalanche is at least 150 years old, it shows little evidence of revegetation, suggesting long recovery times for such slope failures.

basins with the morphologically youngest crown scarps and the most recent debris torrent deposits all remained stable during the storm, whereas other slopes not showing prominent features of recent prehistoric sliding did fail in 1964. The fact that the oldest slopes appear most susceptible to failure suggests that major climatic events are more likely to affect slopes that are revegetated and becoming progressively more weathered, rather than recently failed slopes that are in the process of recovery.

MOVEMENT OF DEBRIS AVALANCHE-DERIVED SEDIMENT DOWN STREAM CHANNELS

The upper drainage basin of the Van Duzen basin offers an excellent opportunity to study some aspects of the movement of sediment from slopes to headwater channels where complicating factors of large drainage area and heterogeneous rock units are eliminated. In the headwater area, sediment moves rapidly off the hillslopes, then much more slowly in the channels. The rate and path of sediment transport from slope through upper drainage-basin channels to the lower drainage basin are dependent on the frequency of mass slope failures and on the sites of initial and subsequent deposition of landslide debris in the valley bottom.

Debris avalanches greater than 30 000 m^3 occurred at the heads of nine headwater tributary basins during the December 1964 storm (fig. 4). Avalanche sites were all in virgin forest unaffected by timber harvesting or road building. The avalanches ranged in area from 0.5 to 5.5 ha, scoured into weathered bedrock, and were between 1.0 and 4.5 m thick. Total amount of avalanche debris discharged into the headwater channels of the two upper drainage basins was 4 041 000 metric tons. In addition to sediment delivered from the slopes by avalanching, severe bank erosion of older 18- to 30-m-thick debris-torrent fill-terraces on the channel margins of the upper South Fork occurred, which further contributed to channel aggradation. I studied the movement of sediment derived from these avalanches and channel fills for the 10 years after the flood by a combination of field mapping and interpretation of time-sequential sets of aerial photos. I analyzed individually the first-order and second-order headwater channels of the Van Duzen and the South Fork Van Duzen, and next highest order channel reaches, to determine the amount of debris contributed by avalanching and by erosion of previously stored channel fill to downstream aggradation, the depth of aggradation in headwater channels immediately after the flood, and the rate at which alluvium moved farther downstream once initially deposited (table 1). Methods of mapping along the channels are described in Kelsey (1980).

Sediment delivery from steep headwater slopes to sites of initial deposition occurs rapidly, but subsequent downstream movement of sediment is much more gradual. In the South Fork and Upper Van Duzen headwater basins respectively, 67 and 81 percent of the sediment mobilized from slopes and stored channel fill in December 1964 was initially deposited in the headwater channels (table 1), and the remainder was transported downstream by the flood. Aggradation in headwater channel reaches immediately after the flood ranged between 1.0 and 4.5 m. In both headwater basins, 75 percent of the aggraded sediment moved farther downstream in the next decade (fig. 5), and its migration could be clearly traced by field mapping and by inspection of sequential aerial photos. The longitudinal profiles of the South Fork Van Duzen and upper Van Duzen drainage basins (figs. 6 and 7) show the sites of initial debris torrent deposition and the sites of maximum channel aggradation at different times after the

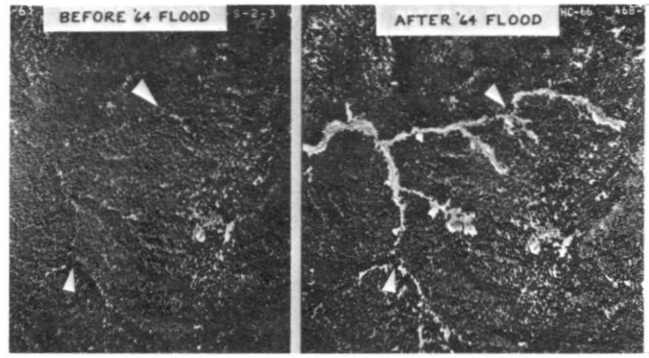

Figure 4.--Comparative aerial photographs of the headwaters of the South Fork Van Duzen River showing slope conditions before and after the 1964 storm. The photos were taken in the summers of 1963 and 1966; the white arrows show identical locations on the two photos. The large, elongate debris-avalanche scars grade downslope into channel-confined debris torrents; these torrents scour and widen channels and deposit thick alluvial fills at their downstream depositional ends (see fig. 5). Bank erosion on channel margins downstream from initial debris-torrent deposition occurred as the coarse sediment was subsequently mobilized downstream; this erosion has caused smaller debris slides along the channel that are visible in the postflood photo.

Figure 5.--Upstream view (August 1975) at the confluence of the South Fork Van Duzen River with Red Lassic Creek. Note sheared-off stumps of trees originally buried with alluvium after the December 1964 storm (3.6-m stadia rod leaning against middle stump for scale). The source of the alluvium was debris avalanches immediately upstream. Behind the stumps is an alluvial fill deposit from a former debris torrent; fills such as this one are now preserved along the valley margins at localities protected from frequent stream corrasion. Alluvial fill from the 1964 debris-avalanche events was in turn nested alongside the older fill shown here.

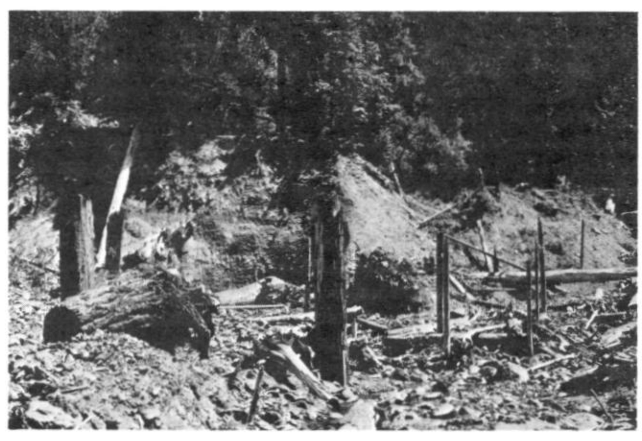

Table 1--Mobilization of alluvium in the headwater reaches of the Van Duzen River, 1964-75, as a consequence of the December 1964 storm

Channel reach, going in downstream direction	Amount of landslide debris discharged directly into reach during the December 1964 flood ρ = 1.92 g/cm^3	Estimated amount of alluvial channel fill, stored in floodplain and fill terraces, remobilized by the December 1964 flood ρ = 1.92 g/cm^3	Initial aggradation as a direct consequence of the December 1964 flood: Average depth of fill	Initial aggradation: Amount of fill ρ = 1.92 g/cm^3	Amount of degradation as of 1975: Average depth of scour	Amount of degradation: Amount of fill removed ρ = 1.92 g/cm^3
	Metric tons	Metric tons	Meters	Metric tons	Meters	Metric tons
South Fork Van Duzen River:						
Headwater Reach A 2d order	218 000	--	1.8	34 000	1.7	31 000
Headwater Reach B 1st order	490 000	--	2.1	19 000	1.7	15 000
Headwater Reach C 2d order	783 000	1/210 000	2.7	132 000	2.1	100 000
Headwater Reach D 2d order	36 000	--	3.4	27 000	2.9	24 000
Channel Reach E 3d order	75 000	1/249 000	4.3	1 059 000	3.2	794 000
Channel Reach F 3d order	49 000	36 000	1.4	264 000	1.1	198 000
Tributary Reach G 2d order	56 000	--	0.5	9 000	0.5	9 000
Channel Reach H 3d order	22 000	--	1.1	127 000	0.8	89 000
Tributary Reach I 2d order	490 000	32 000	1.7	209 000	1.2	157 000
Channel Reach J 3d order	8 000	9 000	0.8	80 000	0.6	56 000
Tributary Reach K	163 000	--	2.1	2 000	1.5	1 000
Total	2 390 000	536 000		1 962 000		1 474 000
Upper Van Duzen River:						
Headwater Reach A 3d order	479 000	--	2.1	228 000	1.5	184 000
Headwater Reach B 1st order	197 000	--	2.9	87 000	2.0	68 000
Headwater Reach C 1st order	90 000	--	3.1	47 000	2.1	37 000
Headwater Reach D 3d order	767 000	--	1.8	176 000	1.5	157 000
Channel Reach E 3d and 4th order	91 000	--	1.8	451 000	1.4	350 000
Channel Reach F 4th order	27 000	--	1.3	110 000	0.9	92 000
Channel Reach G	0	--	0.5	70 000	2/0.7	2/101 000
Total	1 651 000			1 169 000		888 000

1/Values for alluvium stored in fill terraces derived from deposition of the Red Lassic debris avalanche.

2/Contrary to the rest of the column, these values indicate amount and depth of AGGRADATION in the downstream-most reach, which aggraded as alluvium moved downstream.

1964 storm. Migration rates for the aggradation maxima in the upper channel reach of each drainage basin were 8.9 km/10 years in the upper Van Duzen (gradient = 0.015) and 13.7 km/10 years in the steeper South Fork (gradient = 0.027). As the pulse of sediment moved downstream, bank cutting occurred, and this bank erosion was especially extensive along the wide floodplains of the South Fork at the lower end of the study area (fig. 6).

In the wide, 6.5-km long, alluviated valley of the South Fork, which starts 8 km below the site of the 1964 avalanching, aggradation-induced lateral corrasion eroded 35 ha of floodplain between December 1964 and 1974. Floodplain banks retreated as much as 110 m as the active channel widened. Seventy-nine percent of this erosion occurred by June 1966, indicating most of the bank retreat was a direct consequence of the flood and the initial pulse of aggradation. After 1966, reaches of maximum bank retreat moved downvalley with the locus of maximum aggradation (Kelsey 1977).

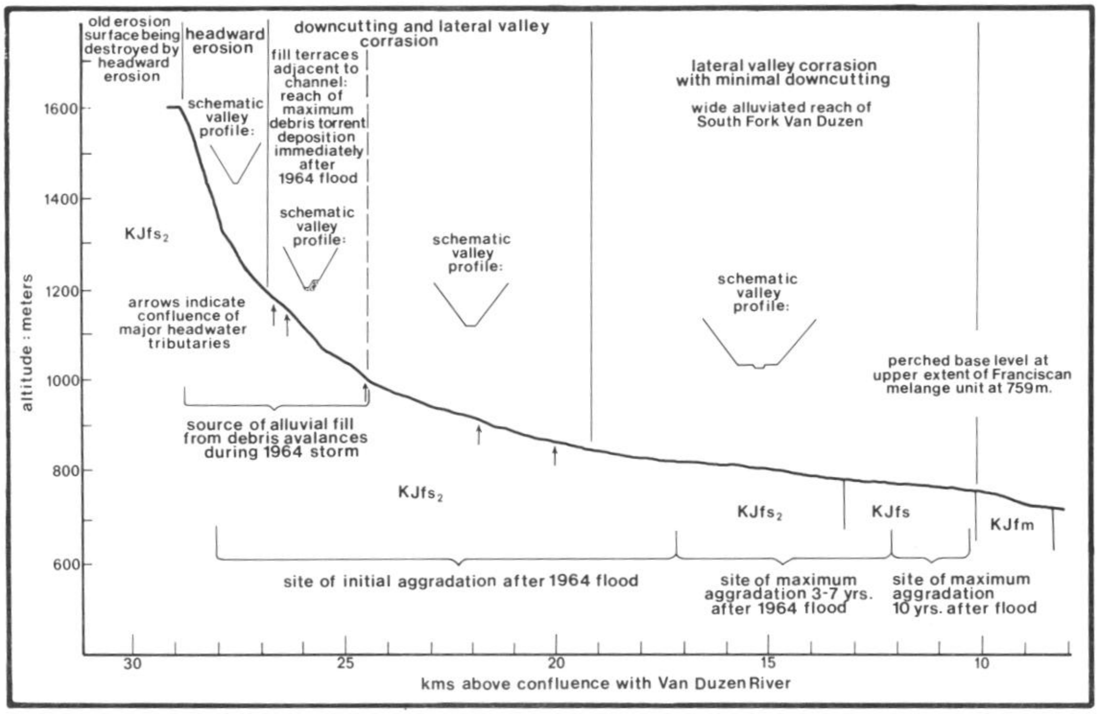

Figure 6.--Longitudinal stream profile of the South Fork Van Duzen River showing the tributary sources of debris avalanching, sites of initial debris-torrent deposition after the December 1964 flood, and migration of maximum channel aggradation after the 1964 storm. The channel profile is segmented into reaches of similar channel form and geomorphic process. ($KJfs_2$ is metasandstone; KJfs is sandstone; KJfm is melange.)

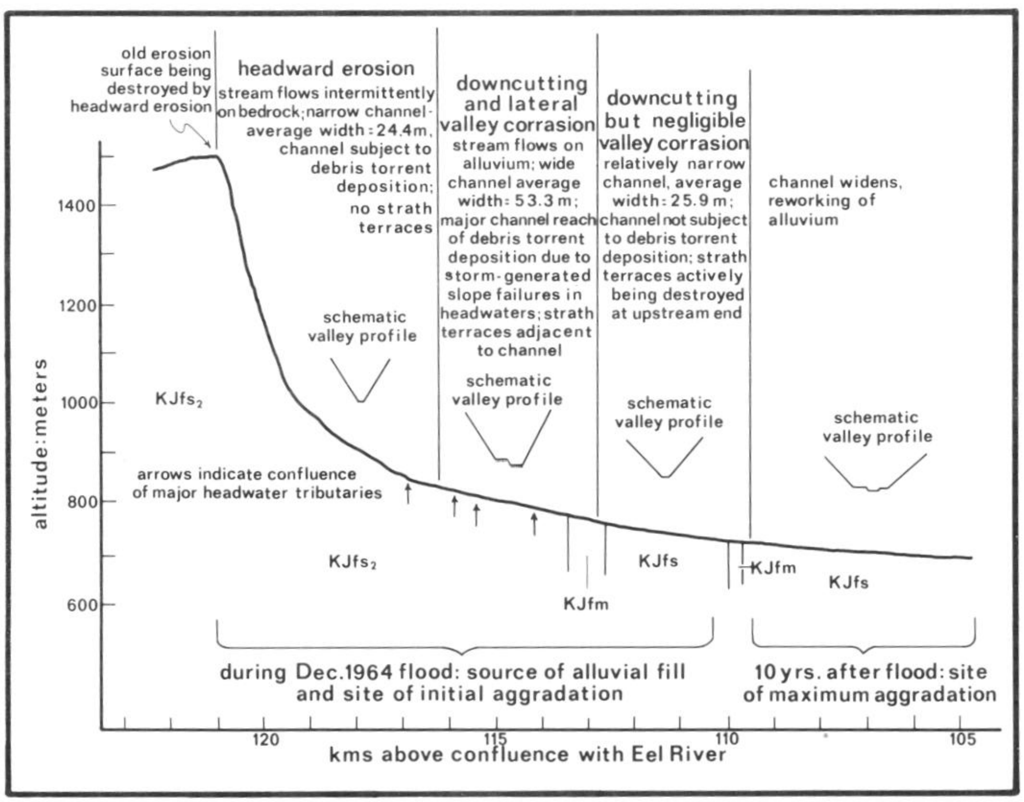

Figure 7.--Longitudinal stream profile of the upper Van Duzen River showing the dominant geomorphic processes operating along the channel, and the tributary sources of avalanching and sites of deposition during and after the December 1964 storm and flood. ($KJfs_2$ is metasandstone; KJfs is sandstone; KJfm is thin shear zones of melange.)

Aggradation and bank erosion in the South Fork occurred concurrently with changes in channel pattern from a single meandering stream to a braided system of channels. A comparison of channel widths measured by the U.S. Land Survey in 1872 with 1963 and 1974 channel widths along the wide valley of the South Fork indicates that in the period 1872-1974, channels increased in width anywhere from 200 to 2,000 percent and practically all this erosion occurred between 1963 and 1974.

Channel recovery from such increases in width (i.e., channel downcutting, decreased channel width and revegetation of floodplains) is a slow process that along the South Fork Van Duzen probably will take longer than 25 years and may take more than a century. Along channels elsewhere in the California Coast Ranges, I have observed that after 15 years, many 1964 flood bars are becoming revegetated and are being isolated by downcutting. On other channel reaches where landslide contributions were greater or landslide activity continued after 1964, downcutting is a slower process and flood bars are still being overtopped by discharges that recur every year or two. Such long recovery times for the South Fork Van Duzen or other channels that carry landslide-delivered pulses of sediment stand in marked conflict with Wolman and Gerson's (1978) contention that stream channels in mountainous, temperate climates recover relatively quickly (10 years) after a major climatic event. Infrequent events of large magnitude, such as the 1964 storm and flood, can leave a long-lasting imprint on channel morphology when such a climatic event triggers slope failures that deliver massive amounts of alluvium to stream channels.

Movement of coarse sediment through upper drainage-basin channels results in a systematic downstream change in channel morphology and valley profile (figs. 6 and 7). Because the headwater basin area is underlain by a fairly homogeneous Franciscan sandstone unit, geomorphic processes operate in a reasonably simple and predictable manner, given stable climatic conditions.

The headwater tributaries of both drainage basins are actively eroding into an old upland surface, and the dominant geomorphic process is active headward erosion along first-order and second-order tributaries where stream gradients are steepest, stream channels are narrowest, streams flow on bedrock, and little alluvium is stored in the channels. The tributary slopes periodically undergo debris-avalanche failures, which generate debris torrents. Where the first- and second-order tributary streams enter the third- and fourth-order South Fork and upper Van Duzen Rivers, channel width increases markedly. Voluminous amounts of coarse alluvium enter these river reaches from tributary debris torrents. These pulses of alluvium promote lateral corrasion of the valley walls, which leads to the formation of strath surfaces (fig. 7). At this point of channel widening, the dominant geomorphic process in both rivers changes from headward cutting to alternating downcutting and lateral valley corrasion. Stream gradient decreases noticeably where the channel widens (figs. 6 and 7).

Alluvial terraces border upper reaches of the South Fork and upper Van Duzen only in their wide channel reaches where lateral corrasion occurs. In the upper Van Duzen, the terraces are strath terraces preserved by uplift of the planated bedrock channel floor. The straths have an alluvial veneer that averages 7.5 m thick, which is about the depth of channel scour and fill during large floods; the surfaces range from 15 to 45 m above the present channel elevation (Kelsey 1977). Alluvial terraces in the South Fork channel are all fill terraces primarily associated with one exceptionally large prehistoric debris avalanche.

Downstream from the reach of alternative downcutting and lateral corrasion, the South Fork and upper Van Duzen channels have differing geomorphic characters because of geologic contrasts in the lower portion of the drainage basins. The South Fork continues to corrade laterally, and downcutting is minimal because of the proximity of the upper drainage basin baselevel on the South Fork (fig. 6).

The upper Van Duzen channel, downstream of the reach of alternating downcutting and lateral valley corrasion, narrows to about 26 m in response to the decreased sediment contribution to the channel (fig. 7). Coarse alluvium supplied by upstream debris torrents appears to have been sorted and winnowed out, and mapping of alluvial deposits suggests as much as 45 percent of the original channel alluvium has been transported downstream. The remaining bed material is no longer sufficient to cause substantial corrasion of the bedrock valley walls. Potential transport capability exceeds sediment supply, so downcutting is the dominant process along this channel reach. At the upstream end of this reach, strath terraces are being destroyed by downcutting as erosion continues headward (fig. 7). Further downstream on the upper Van Duzen, the gradient moderates, the valley floor widens, and the river flows in a broad alluviated valley similar to the South Fork.

This downstream progression of changes in both valley profile and geomorphic process in a headwater basin that is periodically supplied with slugs of coarse alluvial fill from avalanching is precisely the progression described by Bull (1979) in his discussion of the interplay of stream power (power available to transport the sediment load) and critical power (power needed to transport the sediment load) in determining the efficiency of sediment transport and the shape of the valley profile. In steep, v-shaped headwater tributaries where stream power exceeds critical power, sediment movement downstream is rapid. Where gradient begins to moderate and low-order streams converge and supply large quantities of sediment to higher order channels, critical power equals stream power in the long-term average. Downcutting and lateral valley corrasion at these sites alternate as the dominant geomorphic processes and both strath and fill terraces can form and be preserved. Some coarse sediment is moved rapidly through such channel reaches, but other coarse alluvium, which I estimate to be between 5 and 20 percent, becomes entrapped in erosionally isolated fills, and residence times range from hundreds to many thousands of years.

In the lowest gradient, highest order stream reaches of both upper drainage basins, wide alluviated valleys occur where stream power is substantially less than critical power most of the time. As a result of infrequent, high discharges, the wide alluviated valleys aggrade, but in the long term, the stream gradually scours as it reworks and transports coarse alluvium downstream. Residence times of different pockets of alluvium on wide alluviated valley floors are difficult to determine and vary greatly depending on the position of the alluvial deposits.

RECURRENCE INTERVALS OF DEBRIS AVALANCHING AND RESIDENCE TIME OF ALLUVIUM IN CHANNELS

Large-scale mobilization of channel fills, such as that of the December 1964 storm, only happen when storm runoff is charged with slugs of sediment generated by hillslope avalanching. A large flood-producing storm alone is not sufficient to initiate substantial alluvial reworking in higher order channels. The storm must trigger hillslope failures that contribute coarse debris to channel. A 1955 storm in the Van Duzen was of comparable magnitude to the 1964 storm (Harden et al. 1978) but the absence of significant slope failure in 1955 resulted in negligible channel change (Kelsey 1980).

Recurrence-interval estimates of debris-avalanche events in headwater channels are valuable in determining the average interval of channel stability between major sediment-transporting events. These recurrence intervals can best be approximated by studying the alluvial stratigraphy of debris-torrent fills at the mouth of headwater tributaries, where series of truncated alluvial fans from prehistoric torrents are preserved as forested fill-terraces. Detailed studies using dendrochronology and C_{14} dating at two such sites where a series of three fills are nested along valley wall margins suggest episodes of major avalanching occurred in these headwater drainages at intervals greater than 300 years but probably less than 2,000 years, though minor avalanches and smaller debris torrents occurred more frequently (Kelsey 1977, 1980). The dating of channel fills in the upper Van Duzen, combined with data collected by Helley and LaMarche (1973) from a similar study in nearby drainages, shows that storms capable of triggering large landslides and mobilizing substantial amounts of alluvium occur, on the average, once every 100 years in at least one large drainage basin in the northern Coast Ranges. Especially severe storms that cause widespread avalanching and aggradation throughout the northern Coast Ranges are more infrequent, and the data suggest these storm events do not occur as frequently as 100 years, and probably occur only once every 500 or more years. Convincing evidence on hillslopes and in channels of such large magnitude events is not preserved, however. Note that the rate of recurrent slope failure is greater adjacent to third- and fourth-order streams in the lowest reaches of the headwater area where streambank corrasion accompanies each episode of channel aggradation, causing frequent but much smaller debris sliding on the footslopes.

These recurrence-interval estimates are a first step in estimating the residence time of alluvium stored in Van Duzen headwater channels. The following rough estimates are for second- and third-order channels of the study area. The estimates are based on the areal distribution of fills in the channels, as well as on the estimated debris-avalanche recurrence intervals. If sediment remains in the active channel, defined by the mean annual flood, residence time is short, probably less than ten to tens of years. Sediment may also be deposited in thick debris torrent fills. Fills adjacent to the active channel will probably be removed by bank erosion from large storms that recur, on the average, once every century. Fills that are further removed from the active channel and are nested against the valley margin may become isolated by incision and stabilized by forest cover. These fills may remain intact for many hundreds or a few thousand years until avalanches from subsequent large storms send a sediment-charged flood surge through the valley, removing portions of the fills and replacing them with new material. Sediment deposited on strath surfaces that are subsequently isolated from the stream by downcutting into the bedrock have much longer residence times. Alluvium stored on these strath surfaces can survive from a few thousand to as much as ten thousand years until destroyed by debris sliding on the valley sideslopes. Further work is needed to determine what percentage of the coarse alluvium is subject to the different residence times.

HILLSLOPE EVOLUTION AND THE RESIDENCE TIME OF DEBRIS ON HEADWATER SLOPES

My hillslope model for the Van Duzen headwater area, based on observed slope morphologies, sizes of recent avalanches, and alluvial stratigraphy, is that headwater slopes are sculpted by relatively large debris avalanches, and these debris avalanches over time affect all the headwater drainage areas (Kelsey 1980). Both the extent of avalanching during the 1964 storm and the presence of relict avalanche scars that encompass entire first-order basins argue strongly for the model. This model stands in contrast to that of Dietrich and Dunne (1978) and Lehre (this volume), in which debris avalanches and debris flows are initiated from bedrock hollows or swales which occur in only a small part of the landscape, and that other processes, such as creep, surface erosion, and tree throw, feed debris to the swale areas. Both models probably do represent actual conditions for forested drainage basins in different environments. The duration of time between major failures at a specific site is dependent in one case on the rates of hollow filling by creep, surface erosion, and tree throw, and in the other case, by the rate of soil formation and weathering of the regolith at the site of avalanching. Frequency of failure at a site is probably greater in the case of hollows because hollow-filling processes are faster than *in situ* weathering of newly exposed material. I believe that differences in uplift rate are important in determining which mode of landscape evolution is in operation, and relatively high uplift rates in the Van Duzen favor debris-avalanche sculpted topography. Important, but poorly understood variables, such as the rate of creep and the frequency of debris avalanche or debris flow failures, make further refinement of either model difficult at this time.

The residence time for slope debris is the approximate interval for the entire surficial slope mantle to be mobilized by catastrophic landsliding at least once, assuming the persistence of present climatic and tectonic conditions. The estimate of residence time therefore applies only to forested headwater slopes from ridge tops to a boundary about 50 m above perennial streams, which is that boundary above the immediate influence of channel corrasion and downcutting. I have estimated a residence time for the surficial mantle by knowing the area unvegetated from recent avalanching, and applying an estimate of recurrence interval for this amount of avalanching. In 1975, between 2 and 3 percent of the headwater slope area was exposed from avalanching, and--based on alluvial stratigraphic studies cited above--I assume this extent of avalanching occurs every 500 years or more. Table 2 shows alternative estimates of residence time assuming that either 2 or 3 percent of the headwater basin area is denuded by avalanching every 500, 750, or 1,000 years, and that a different 2 or 3 percent is affected each time until the entire weathered slope mantle is removed once. The last assumption is of course not entirely accurate, because one slope may fail twice while an adjacent one does not fail at all, but indications are that the slopes that have been stable the longest are most prime for failure. Estimates of residence times range from 17,000 to 50,000 years, which is one to two orders of magnitude greater than the residence time of debris once it reaches second- and third-order perennial stream channels.

Using the estimates of residence time for the surficial slope mantle, an erosion rate for the headwater basin area can be calculated if we know the average thickness of the surficial layer lost by periodic avalanching. Avalanche failures in the headwater basins ranged between 1.0 and 4.5 m thick, and 2.5 m is a representative thickness. Assuming the average density of the rock and soil to be 1.92 g/cm^3, the above five different residence times (table 2) give the following range of erosion rates:

Residence time	Erosion rate
Years	Tons/km^2 per year
16,700	280
25,000	190
33,300	140
37,500	125
50,000	95

Table 2--Determinations of residence time of the surficial slope mantle for six cases in the steep headwater region of the Van Duzen drainage basin.

Case	Recurrence interval of major episode of avalanching	Portion of basin affected	Residence time of surficial mantle
	Years	Percent	Years
I	500	2	25,000
II	750	2	37,500
III	1,000	2	50,000
IV	500	3	16,700
V	750	3	25,000
VI	1,000	3	33,300

Sediment-discharge measurements in a 1.17-km^2, second-order forested drainage basin in the lower Van Duzen drainage basin gave a sediment yield of 45-180 t/km^2 per year (Kelsey 1977). This basin does not experience debris avalanching as frequently as the upper drainage-basin headwater region. Erosion rates in the headwater region could be 1.5 to 3 times as great, and probably range between 80 and 540 t/km^2 per year. Estimates of residence time of slope debris under current climatic and tectonic conditions therefore appear reasonable in light of the limited measured erosion rates.

THE ROLES OF PERSISTENT VERSUS EPISODIC EVENTS IN MOBILIZING SEDIMENT

Persistent processes, such as fluvial sediment transport in gullies or streams by storm runoff, are low-magnitude events; episodic processes occur infrequently but generally mobilize relatively large amounts of sediment in short intervals. Episodic processes play a highly significant role in the Van Duzen headwater study area. Fluvial hillslope-erosion rates are quite low (Kelsey 1980). Periodic debris sliding or avalanching is the main process of sediment delivery to major channels and is the primary mode of landscape sculpture. In contrast, fluvial sediment transport, creep, and earthflow landslides are persistent processes that are particularly important in melange terrain in the lower Van Duzen basin. In this terrain, persistent processes are not only the most important sediment transporters but are also largely responsible for shaping the landscape. Only near large channels do episodic processes, such as debris sliding, slumping, or bank erosion, become significant in transferring debris from slopes to channels.

One of the long-standing questions in geomorphology is the importance of high-frequency/low-magnitude events versus low-frequency/high-magnitude events on transporting sediment and sculpting landforms. Geomorphic processes in the headwater study area point to the important role of infrequent, large-magnitude climate events, as compared to their role in other stream systems that flow through more stable geologic settings. In contrast to conclusions of Wolman and Miller (1960) that the low-magnitude geomorphic events appear to be by far the most important for sediment transport and landform evolution, large events in the Van Duzen headwaters have substantially modified landforms, both on the slope and in the channel, and have contributed large quantities of sediment to channels. Wolman and Miller's analysis was largely based on sediment yields from gaging stations, whereas this study addresses hillslope erosion rates caused by avalanching and investigates channel changes from the interplay of avalanching and the movement of sediment stored in channels. The latter approach documents the importance of infrequent episodic events in erosion of the Van Duzen headwater basins. In this study, sediment delivered to channels by avalanching apparently can account for almost the entire estimated erosion rate of headwater basins. In addition, the infrequent and high-magnitude runoff events are also responsible for the major changes in sediment storage, which account for the significant long-term changes in channel morphology.

DRAINAGE-BASIN RECOVERY AND HIGH-MAGNITUDE, EPISODIC EVENTS

Where episodic, high-magnitude events are important, the recovery times on hillslopes and in channels after large landslides or aggradation events become significant. Recovery means revegetation of hillslopes and restitution of channel width and depth to that configuration prior to rapid, storm-induced change.

The recurrence interval of large-scale geomorphic events is influenced by whether or not major storms occur, on the average, while hillslopes and channels are still recovering. If a large part of the basin is in early vegetative recovery, high-magnitude events may have greater effect. In the case of the 1964 storm in the Van Duzen headwaters, there were no areas just beginning to revegetate. Therefore, failure of recently exposed slopes was negligible. The disturbance frequency in a basin probably increases as the proportion of a basin in early vegetative recovery increases. However, once a failed slope has revegetated but prior to significant weathering of the newly exposed surface material, the slope is not highly prone to failure. For this reason, young, revegetated landslides in the Van Duzen headwaters were not affected by the 1964 storm. The effect of recovery time on disturbance frequency is not easily determined because of the short historic record, but if we substitute space for time, that is, look at a large area, the 1964 storm apparently did not hit an area in the process of recovery from a previous widespread, high-magnitude event. This pattern is true both for the 160-km^2 study area and for a much larger area, about 1500 km^2, that encompasses the upper drainage basins of many northern California Coast Range basins. Had the same storm hit an area in widespread recovery from a previous storm, disturbance probably would have been significantly greater.

The frequency of disturbance is probably increased by landuse. Intensive landuse in the Coast Ranges since 1860--especially road building, timber harvesting, and grazing--has exposed bare soil or weathered bedrock and ruptured the vegetation mat. In addition to the vegetative and channel recovery from climatically induced slope failures on undisturbed slopes, the disturbed areas may also be prone to further failure, especially along major channels and on headwater slopes. The amount of area in some stage of vegetative recovery is therefore much greater than a century ago. Major climatic events are now likely to trigger more frequently both small landslides directly attributable to landuse as well as large landslides only indirectly related to landuse.

SUMMARY

Two third-order, forested basins that comprise the headwaters of the Van Duzen River provide an opportunity to study the movement of sediment on hillslopes and in channels. Sediment is mobilized on hillslopes primarily by debris avalanches that deliver material to first- and second-order channels by debris torrents. After a debris avalanche occurs, residence time of the newly exposed surficial slope material ranges from 15,000 to 50,000 years. Sufficient soil weathering occurs during this residence time so that slope failure is likely to recur during a large storm event.

After sediment reaches a stream channel by avalanching, its residence time is dependent on the site of sediment deposition and the recurrence interval of upslope debris avalanching that serves to remobilize channel-stored fill and deposit new fill from the slopes. Debris avalanches in a first- or second-order basin recur about every 300-500 years. Therefore, flushing of the more accessible fills adjacent to the active channel occurs with the same frequency. Channel fills nested along valley walls further removed from the active channel move less frequently and may remain in place for hundreds or a few thousand years. Active channel deposits move on the scale of every year to at least once a decade. Alluvium perched on strath terraces that have been isolated from the channel by uplift only becomes reincorporated in the active channel by sideslope debris sliding. Such sliding probably recurs at any one site every 1,000-5,000 years. The proportion of channel alluvium that is active compared to that alluvium stored in fills or on bedrock straths is hard to estimate, and more work is clearly needed. Residence time of debris on hillslopes, however, is usually two orders of magnitude longer than residence time of alluvium stored as fill in channels.

Infrequent, high-intensity storms trigger avalanches that sculpt the landscape and set in motion the redistribution of sediment stored in channels. More persistent and smaller storm events probably are responsible for transporting the greater volume of sediment in channels. Infrequent, episodic events, however, are responsible for major shifts in sediment storage, and therefore these events are the catalysts for the significant changes in channel morphology that occur on a time scale of thousands of years.

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LITERATURE CITED

Bailey, E. H., W. P. Irwin, and D. L. Jones.
1964. Franciscan and related rocks and their significance in the geology of western California. Calif. Div. Mines and Geol. Bull. 183, 177 p.

Brown, W. M., III, and J. R. Ritter.
1971. Sediment transport and turbidity in the Eel River basin. U.S. Geol. Surv. Water Supply Pap. 1986, 70 p.

Bull, W. B.
1979. Threshold of critical power in streams. Geol. Soc. Am. Bull., Part 1, 90:453-464.

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

Harden, D. R., R. J. Janda, and K. M. Nolan.
1978. Mass movement and storms in the drainage basin of Redwood Creek, Humboldt County, California--a progress report. U.S. Geol. Surv. Open-File Rep. 78-486, 161 p.

Helley, E. J., and V. C. LaMarche.
1973. Historic flood information for northern California streams from geological and botanical evidence. U.S. Geol. Surv. Prof. Pap. 485-E, 16 p.

Judson, S., and D. F. Ritter.
1964. Rates of regional denudation in the United States. J. Geophys. Res. 69:3395-3401.

Kelsey, H. M.
1977. Landsliding, channel changes, sediment yield and land use 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.

Kelsey, H. M.
1980. A sediment budget and an analysis of geomorphic process in the Van Duzen River basin, north coastal California, 1941-1975. Geol. Soc. Am. Bull., Part II, 91:1119-1216.

Kelsey, H. M., and A. O. Allwardt.
1975. Geologic map of the Van Duzen River basin. In Van Duzen River basin environmental atlas. Calif. Dep. Water Resour., North. Dist., Plate 8, Sheets A-1.

Swanston, D. N.
1970. Principle mass movement processes influenced by logging, road building and fire. Proc., Symposium on forest land uses and stream environment. School For., Oreg. State Univ., Corvallis.

Waananen, A. O., D. D. Harris, and R. C. Williams.
1971. Floods of December 1964 and January 1965 in the far western states - part 1, description. U.S. Geol. Surv. Water Supply Pap. 1866-A, 265 p.

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

Wolman, M. G., and J. P. Miller.
1960. Magnitude and frequency of forces in geomorphic processes. J. Geol. 68(1):54-74.

Sediment Transport and Channel Changes in an Aggrading Stream in the Puget Lowland, Washington

Mary Ann Madej

ABSTRACT

Recent land-use changes increased sediment yield in a fourth-order stream in western Washington from 22 t/km^2 per year to 185 t/km^2 per year. In response, channel width increased and depth decreased, but channel gradient and mean flow velocity remained about constant. As a result of channel changes, sediment transport rates increased from 500 t/year to 4200 t/year. Active sediment constitutes 5 percent of stored sediment, and much sediment will be flushed from the system in about 20 to 40 years.

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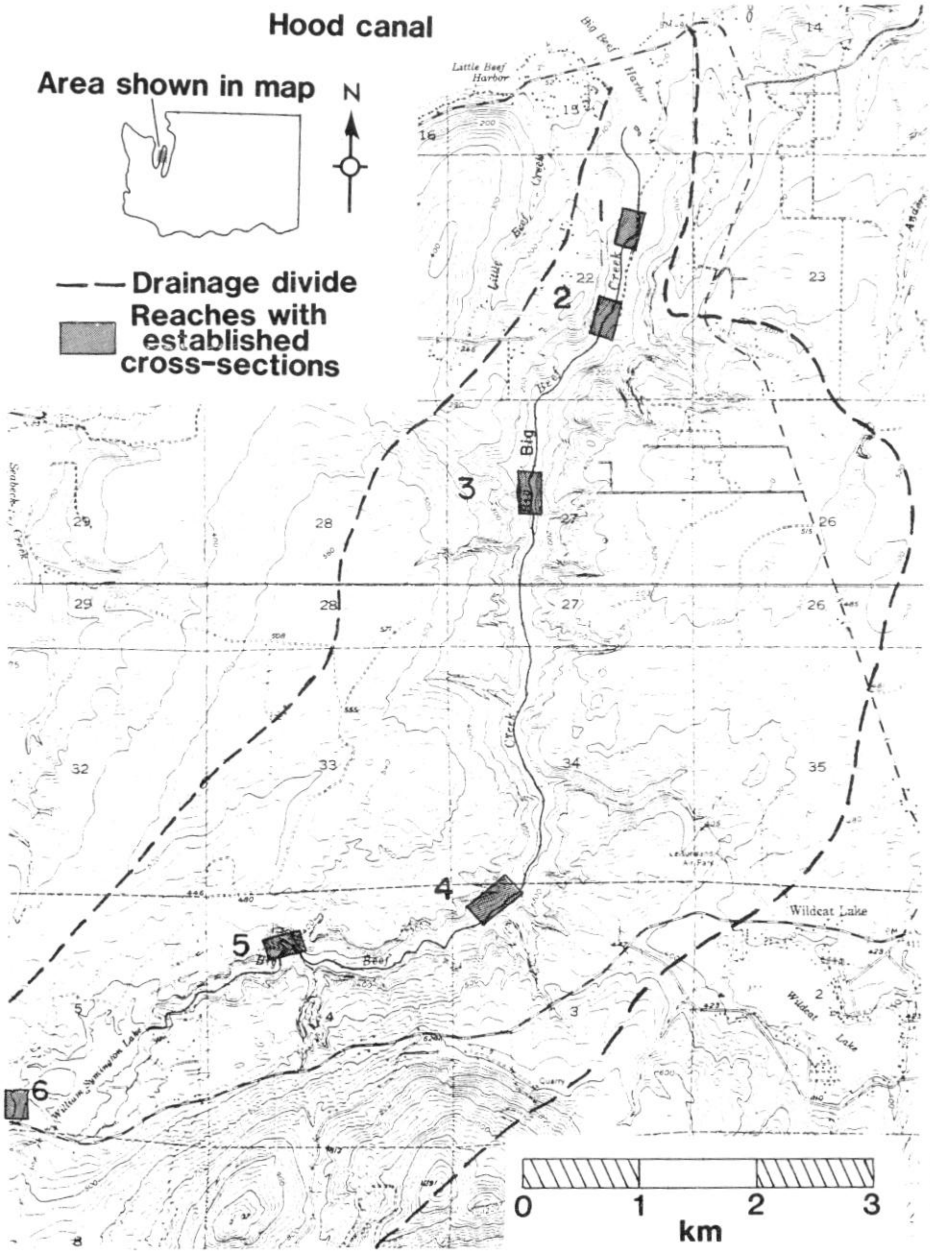

Figure 1.--Map of lower Big Beef Creek drainage basin showing reaches with established cross sections.

INTRODUCTION

The sediment load of a stream is determined by such characteristics of a drainage basin as lithology, vegetation, precipitation, and land use, and it enters the stream system by a variety of erosional processes. If stability is to be maintained, an equilibrium must exist between sediment entering the stream and sediment transported through the channel. Changes in sediment load can upset this balance and result in physical and biological changes in the stream system.

The purpose of this study is to determine the source and volume of sediment introduced into a stream, and its subsequent distribution, rate of movement, and influence on channel geometry. Three approaches were used. First, a sediment budget quantifies the amount of sediment introduced into the stream channel. Second, sediment-transport equations predict how much sediment moves under conditions of various channel morphologies. Third, survey measurements indicate the volume and distribution of the deposited sediment.

FIELD AREA

Big Beef Creek, a gravel-bedded stream, was selected for study because high rates of sedimentation and channel instability during the last 10 years were evident and baseline data were available from records kept at the University of Washington Fisheries Research Station at the mouth of the creek. This fourth-order stream drains an area of 38 km^2 on the Kitsap Peninsula in Puget Sound, western Washington (fig. 1). More than 85 percent of the drainage basin is forested with second-growth western red cedar (Thuja plicata Donn ex D. Don), Douglas-fir (Pseudotsuga menziesii (Mirb.) Franco), and associated species. The climate is characterized by cool, dry summers and mild, wet winters. Annual precipitation (predominantly rain) ranges from 1300 to 1700 mm, and most major storms occur between December and March.

The dominant surficial geologic unit in the drainage basin dates from the Vashon (late Wisconsin) glaciation. A mantle of compact till, composed of a bluish-gray mixture of cobbles, pebbles, and silt and clay, covers much of the area. In addition, poorly sorted recessional outwash sands and gravels form many of the surface deposits in the Big Beef Creek drainage basin.

The upper 15 km^2 of the Big Beef Creek basin has a low channel gradient (0.2 percent), an average hillslope gradient of 5 percent, and drains into an artificial lake. Because the lake acts as a sediment trap for suspended and bedload particles from the upper basin and most runoff and sediment discharge are generated below the lake, this study only considers a sediment budget for the lower drainage basin, where average channel gradient is 1 percent and hillslope gradients range from 10 to 70 percent. Drainage density below the lake is 2.5 km/km^2, as measured on 1:12,000 aerial photographs.

In 1969, the U.S. Geological Survey established a gaging station 1.6 km above the mouth of Big Beef Creek. The frequency of annual peak discharges was compiled from daily discharge records. The 10- and 50-year floods are about 30 and 56 m^3/second, respectively. No flood has exceeded 30 m^3/second in the past 10 years, and regional records indicate that no 50-year floods have occurred in the last 25 years. Thus recent changes in the Big Beef Creek basin occurred under fairly normal climatic patterns.

The principal activities in the drainage basin since the 1850's have been logging and some cattle grazing. Most virgin timber was logged by 1930 and a resurgence of logging activity occurred in the mid-1950's, when harvest of second-growth timber was initiated. Urban development is becoming widespread and this decreases the amount of forested land as well. In 1942, 3 percent of the land was deforested, by 1965 the figure was 12 percent, and today it is almost 20 percent. Likewise, road density tripled, from 1.2 km/km^2 in 1942 to 3.5 km/km^2 at present.

Personnel at the University of Washington Fishery Research Station noted heavy sediment deposition at the mouth of Big Beef Creek during the winters of 1966-67 and 1967-68; in the summer of 1969, the lower 500 m of the creek were channelized artificially. In 1970, a weir 2 m high was built across the mouth of the stream, which has acted as a small dam and has caused deposition of bed material for several hundred meters upstream of the weir.

METHODS

Cross-Section Surveys

Fifty-five cross sections along the main channel of Big Beef Creek were surveyed periodically from 1969-78 by Cederholm (1972) and me (fig. 1). Gradients of channel thalweg and water surface also were determined in several locations.

Detailed records of channel changes at one cross section were obtained from U.S. Geological Survey gaging station records for 1969-79. These records (Forms 9-207 and 9-275) consist of bimonthly measurements of width, cross-sectional area, mean velocity, discharge, and stage relative to gage datum.

Regional Channel Geometry Survey

Because the channel geometry of Big Beef Creek was not determined until after the drainage basin had been disturbed by road construction and timber harvest, predisturbance geometry was predicted from a regional channel-geometry relation developed from relatively undisturbed drainage basins in similar geologic and geographic settings. Estimates of regional channel geometry were obtained by measuring bankfull width and depth, channel gradient, and sediment-size distribution and correlating them with drainage area at 20 sites on nearby streams (Madej 1978). Comparison of dimensions of the present Big Beef Creek channel with those from undisturbed channels draining the same area indicates the extent of changes in Big Beef Creek.

Sediment Studies

Size distribution of sediment in the main channel and tributaries was characterized with the pebble-count method described by Wolman (1954). Samples of channel gravel, and hillslope and bank material were also collected, sieved, and weighed.

Sediment distribution within the main valley of the drainage basin was determined by mapping the location and volume of gravel bars, overbank sand deposits, terraces, tributary deltas, and debris fans. Aerial photographs at a 1:12,000 scale from seven flights between 1942 and 1976 allowed me to compare the locations and sizes of large gravel bars, log jams, and braided reaches through time. Area and location of mass movements were also mapped from photographs. Field checking provided more accurate estimates of volumes of material, as well as ages of mass-movement features dated by vegetation.

The size of the largest particle of bed material moved from a site by high flows during 1977 was obtained by observing marked rocks. Over 2,000 painted rocks were scattered on the upstream end of six gravel bars in the main channel and tributaries. The painted rocks ranged from 16 to 180 mm (D_{50} = 64 mm). After high flows, the size of each rock found and the distance it had moved were recorded.

Bedload-transport rates were estimated by the methods proposed by Einstein (1950), Meyer-Peter and Müller (1948), and Emmett (1976). Dr. H. W. Shen of Colorado State University supplied a computer program of the Einstein Bed Load Function. This program was used to compute bed-material load under various conditions representative of the changing channel geometry from 1969 to 1978.

RESULTS

Measurements

Table 1 lists the results of cross-sectional surveys of six reaches during three periods. Generally, aggradation occurred in lower reaches, degradation in middle reaches, and little change occurred in upper reaches, as illustrated in selected cross sections (fig. 2).

Net aggradation occurred in Reach 1 above the weir. Between 1969 and 1976, the bed aggraded 0.5-1 m in the lower 300 m. Overbank flow deposited material over an area 70 m wide and built small levees to form new banks about 0.5 m above old banks. Alders rapidly revegetated overbank deposits and helped stabilize new banks. Overbank deposition is active, as indicated by several centimeters of sand burying young alders.

Net aggradation occurred in Reach 2 (fig. 2). The channel responded by assuming a more braided pattern in this reach and forming midchannel gravel bars. Bankfull width increased about 25 percent from 1973 to 1978, and bankfull depth decreased by a similar amount. Bankfull velocity (from U.S. Geological Survey records) remained about constant throughout this period.

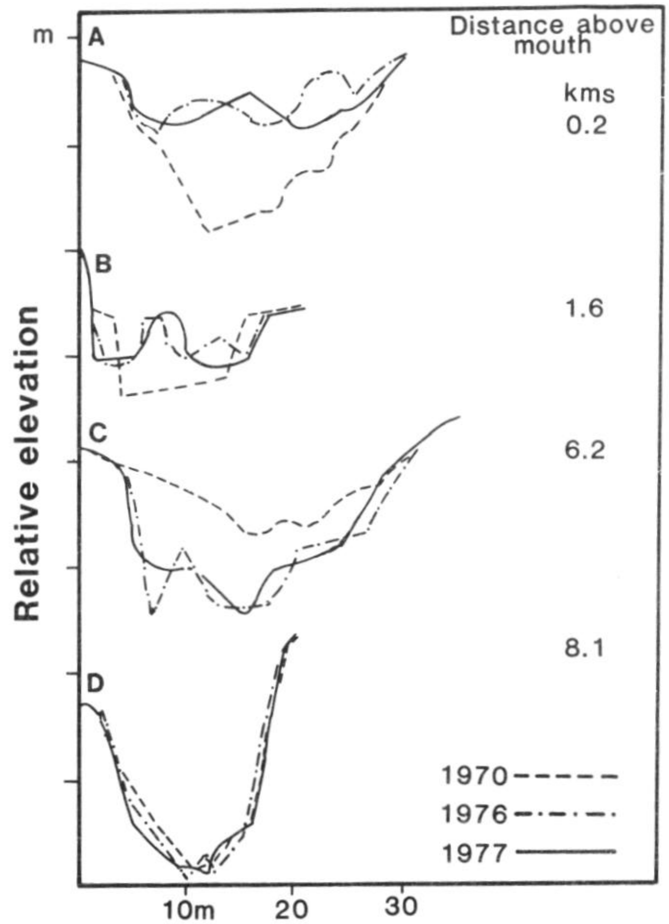

Figure 2.--Selected vertically exaggerated cross-sectional surveys in the main channel of Big Beef Creek.

Table 1--Results of channel cross-sectional surveys, Big Beef Creek, 1969-77

Reach	Approximate distance above mouth	Change[1] in volume of sediment per unit length of channel (m^3/m of main channel per year) 1969-71 [2]	1971-76	1976-77
	Kilometers			
1	0-0.8	+5.05	+0.88	+1.26
2	1.1-1.6	+1.05	+0.34	-0.86
		1969-76		
3	2.9-3.2	-0.91		-0.05
4	6.2-6.6	-1.33		+0.86
5	7.7-8.0	-0.17		-0.01
6	9.6	-0.16		+0.50

[1]Net deposition is indicated by a plus, net erosion by a minus.

[2]Surveys conducted by Cederholm (1972).

Bank erosion and scour took place at several locations in Reaches 3 and 4, which may be a response to previous aggradation. Where roads are still contributing large volumes of sediment, aggradation and growth of gravel bars is apparent. In contrast, the channel just below the lake (Reach 5), which is located in a relatively undisturbed area, showed little change.

Longitudinal surveys indicate that, in the lower 2 km, channel gradient decreased from 0.090 to 0.085 between 1970 and 1977. Total channel length increased from 9785 to 10 010 m. Sinuosity increased from 1.30 to 1.32, and overall channel gradient decreased slightly.

Patterns of channel length changes correlate approximately with trends of cross-sectional changes. Where little cross-sectional change occurred, channel length remained constant. In degrading reaches, channel length increased; in the aggrading lower reaches of Big Beef Creek, channel length decreased.

The predicted and actual values of width for lower Big Beef Creek differ considerably (fig. 3), although the difference is negligible for the upper stream below the lake. Values for bankfull depth, channel gradient, and mean particle size are within limits of scatter in the regional pattern.

After one winter, 53 percent of the painted rocks were found. Large particles ($\geq$128 mm) did not move as far as intermediate sizes (45-90 mm). Surprisingly, small particles (45 mm) did not move as far as those in the intermediate size class. Many of these small particles were found lodged behind large ones and were protected from further movement. Mean distance moved by intermediate-sized particles (45-90 mm) was 80 m.

Distribution of sediment stored in the channel is listed in table 2. In side channels, the volume of sediment above bedrock was measured. In the main channel, sediment was only measured to the depth of scour which was taken to be 0.4 m, the maximum depth of degradation in cross-sectional surveys.

Log jams are temporary sediment traps for bed material moving downstream. Logging activity may increase the number and magnitude of log jams in a stream channel, which in turn may increase the deposition of gravel and sand. In a relatively undisturbed reach, four log jams were noted, or 2.5 jams/km of channel. In reaches with recent logging, the concentration was 12 jams/km and these jams were larger than those in undisturbed reaches. Log jams are associated locally with braided channels as the stream bends around the jams.

Suspended sediment was sampled periodically at 0, 0.8, and 1.5 km above the mouth. Using U.S. Geological Survey daily discharge data, I constructed three suspended-sediment rating curves (Madej 1978). Average suspended-sediment load near the mouth of Big Beef Creek is 2325 t/year (110 t/km^2 per year), assuming the basin above the lake does not contribute any suspended sediment to downstream reaches.

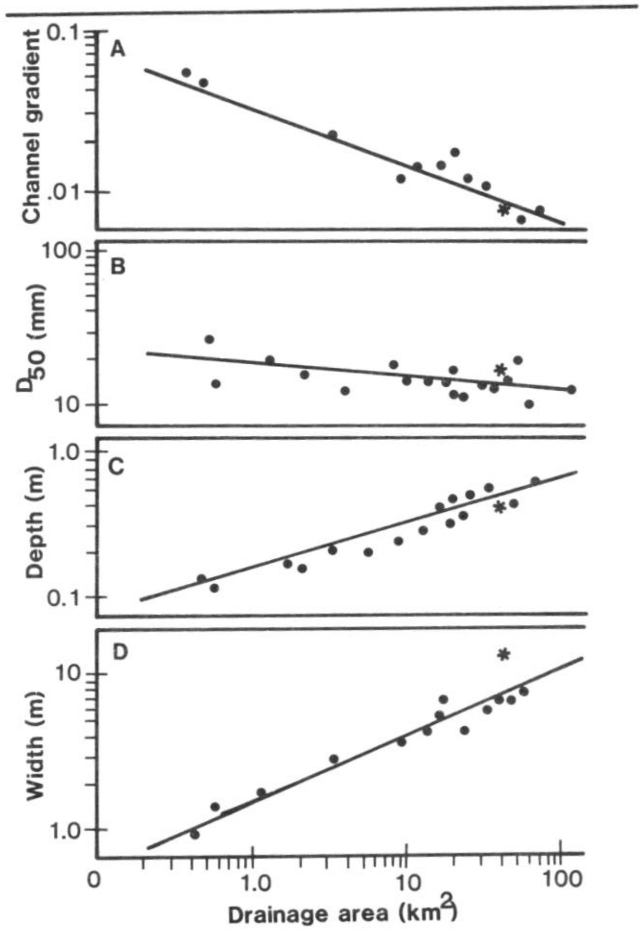

Figure 3.--Results of regional channel-geometry survey for Puget Lowland Streams (present dimensions for Big Beef Creek at 1.5 km are indicated by an asterisk).

Table 2--Distribution of alluvium in Big Beef Creek Basin (lower 23 km^2)

Site of storage	Length of channels	Volume of sediment
	Kilometers	Cubic meters
3d-order tributaries	23	2.2×10^4
4th-order channel (to depth of scour)	8	4.5×10^4
(Active sediment in main channel[1/])	--	(2.3×10^4)
Floodplain (to depth of scour)	--	4.5×10^5
Total		5.2×10^5

[1/]Material that has been transported recently and is temporarily stored in unvegetated gravel bars and behind log jams.

CALCULATIONS

Sediment Budget

The amount of sediment entering Big Beef Creek can be quantified by examining the contribution of processes supplying hillslope material to channels. Sediment input in an undisturbed area was considered to enter channels by soil creep and hillslope failures. Rate of soil creep under edaphic and climatic conditions in the Big Beef Creek drainage basin was estimated from published creep rates for other areas (table 3). Three studies have been conducted in temperate areas receiving about the same annual precipitation as Big Beef Creek (1300-1500 mm), and the coarse soil textures in these study areas are similar to the Big Beef Creek basin. Average rate of creep for the top 15-25 cm of soil for all the studies lies within a narrow range, and a value of 1.5 mm/year was chosen as a reasonable estimate of creep rate in the Big Beef Creek drainage. In addition, tree throw (Dietrich et al., this volume) contributes the creep equivalent of 1.5 mm/year on steeper, wetter hillslopes. The extent of tree throw in the Big Beef Creek basin is not as great, and I estimate 1.0 mm/year of soil movement by tree throw, or a total rate of 2.5 mm/year.

Table 3--Rates of soil creep in temperate, humid climates

Investigator	Precipitation	Gradient	Average soil creep (upper 15-25 cm)
	Millimeters/ year	Percent	Millimeters/ year
Everett (1963)	990	25-65	0-2.2
Young (1960)	1400	35-55	1-2
Kirkby (1967)	1710	15-35	1-3

The most active layer of soil in terms of creep is the upper zone, which is influenced by such processes as root growth and decay, tree fall, and burrowing of animals. In the Big Beef Creek basin, the observed thickness of this active layer is about 0.5 m on hillslopes of 25-45 percent, and it has a density of about 1.7 t/m^3.

Total length of first- to fourth-order channels in the lower drainage basin is about 57 500 m. Because soil transported downhill by creep enters channels through shallow bank failures on both sides of a channel, the total amount of material introduced by creep is: 5.75×10^4 m x 2 x 0.5 m x 0.0025 m/year x 1.7 t/m^3 = 245 t/year = 10.6 t/km^2/year. This is a minimum estimate, assuming soil deeper than 0.5 m contributes little material in comparison to the more active surface layer.

Material discharged through soil creep is transported either as bedload or suspended load, depending on the size of material. Suspended-sediment samples collected in Big Beef Creek have a maximum particle size of 1 mm. Textural analyses of soil in the Big Beef drainage basin indicate that 40 percent of the active layer is greater than 1 mm. Thus, 147 t of material discharged into the channel through soil creep become suspended load, and 98 t is transported as bedload.

Some bedload may break down into particles small enough to be carried as suspended load. Tumbling experiments with rocks of similar lithology and size as material in Big Beef Creek (Kuenen 1956) show that 10 km of transport is insufficient to break down much material. Average rate of attrition is 0.1 percent/km of transport.

Large discrete hillslope failures are the second major contributor to a stream's sediment load. Field measurements of depth of slides averaged 0.6 m. During the last 10 years, 880 m^2 of land was affected by landslides that were not associated with roads, logging, or residential lots. So, the rate of mass wasting for undisturbed parts of the basin is: 880 m^2 x 0.6 m x 1.7 t/m^3 : 10 years = 90 t/year. These slides, which occurred on slopes >15 percent or at the headcuts of first-order channels, were representative of an area of 6 km^2 in the lower drainage basin, so the rate can be expressed as (90 t/year)/(6 km^2) or 15 t/km^2 per year. This rate is lower than estimates from other drainage basins in wetter, steeper environments (Swanston 1969, Dietrich 1975, Pierson 1977).

Because human activity significantly affected the remainder of the lower drainage basin, the rate of naturally occurring slides for the entire lower basin must be extrapolated from the rate of landsliding in the undisturbed sections. An area of 6 km^2 has hillslopes <15 percent, and probably does not contribute sediment through landsliding. The remaining 17 km^2 contributes 15 t/km^2 per year, or 255 t of sediment per year. This sediment is characterized by the size distribution of the average soil column. Thus, 153 t/year is suspended load, and the remainder becomes bedload.

Consequently, in an undisturbed state, sediment input into Big Beef Creek was:

Soil creep - 245 t/year	147 t/year of suspended load
	98 t/year of bedload
Mass movements - 255 t/year	153 t/year of suspended load
	102 t/year of bedload
Total load - 500 t/year (22 t/km^2 per year)	300 t/year of suspended load
	200 t/year of bedload

This value for total sediment yield lies at the lower end of the range of sediment yields reported from drainages wetter, steeper, or both--such as 28 t/km^2 per year in western Oregon (U.S. Environmental Protection Agency 1971), 73 t/km^2 per year in southwest Alberta (Hollingshead 1971), and 105 t/km^2 per year in British Columbia (Slaymaker and McPherson 1977).

The Big Beef Creek drainage basin is no longer an undisturbed system because of changes in land use. In areas undergoing logging and road construction, rates of occurrence of mass movement may increase (Anderson 1954, Fredriksen 1970). In addition, processes such as sheetwash and rainsplash erosion--normally insignificant in humid, well-vegetated areas--may transport material from roads to channels.

Annual sediment contribution by disturbance-related landslides was quantified by mapping volumes of slides originating on roads and cut areas, and dividing volumes by the age of slides. These slides were smaller but more numerous than naturally occurring slides on valley sides. Between 1968 and 1977, an average of 845 t/year entered the stream channel through debris slides induced by land-use changes.

Roads are an additional source of sediment. Sheetwash on roadbanks and roadbeds, erosion of roadside ditches and sidecast material, roadbank gullying, and dry ravel from roadbanks are all active on roads in the Big Beef Creek basin. The volume of material eroded from roadside ditches was estimated in the field. Alluvial fans are not found at outlets of culverts draining ditches, and the largest painted rock moved in roadside ditches had a diameter of 64 mm. Thus, most eroded material probably reaches the stream channel. Roadside ditches contribute 440 t/year to the stream system.

Sheetwash carries sediment from roadbanks and roadbeds to stream channels. Few studies have directly measured amounts of sediment originating from roads, but these studies indicate the order of magnitude of sediment eroded from unpaved roads: Hornbeck and Reinhart (1964) - 9900 t/km^2 per year for fresh roads and 125 t/km^2 for abandoned roads; Diseker and Richardson (1961) - 20 000 to 60 000 t/km^2 per year; and Wolman and Schick (1967) - 20 000 to 50 000 t/km^2 per year. L. Reid and T. Dunne (University of Washington) are currently studying erosion on logging roads, and their results for moderately used and abandoned roads are of the same magnitude as those of Hornbeck and Reinhart. Of these estimates, the 9900 t/km^2 of Hornbeck and Reinhart seems the most reasonable to apply to the Big Beef Creek drainage basin, because these authors studied tractor-yarding skid roads under slope and precipitation conditions similar to those in the Big Beef drainage basin.

Although soil loss in the above studies was measured under a variety of conditions, sediment yields obtained were always far above the sediment yield in undisturbed areas. To apply these values directly to the Big Beef Creek drainage basin is risky because of variation in local conditions. Nevertheless, these values are useful estimates of soil loss from roads in the Big Beef Creek drainage.

Road prisms in the Big Beef Creek basin average 8 m in width. In 1977, 30 of the 65 km of unpaved roads were actively used. Using Hornbeck and Reinhart's values of sediment yield, I computed soil loss on roads to be: .008 km [(30 km x 9900 t/km^2 per year) + (35 km x 125 t/km^2 per year)] or 2410 t/km$_2$ per year. By examining the size distribution of lag deposits of the road surface, I estimated that 80 percent of the eroded material (1930 t) was transported as suspended sediment. Scour of roadside ditches contributes an additional 440 t/year. Thus a total of 2850 t/year is eroded from roads, of which 685 t is bedload.

The total sediment load of the stream is the sum of the contributions of natural hillslope processes and erosion processes from disturbed areas (table 4). Although these estimates in the sediment budget are approximate, their concordance with other measurements indicates they are of the correct magnitude. The estimate of 3000 t of suspended sediment/year compares favorably with the 2325 t/year obtained from suspended sediment rating curves. Estimated bedload input is 1200 t/year; the volume of sediment deposited in the lowest kilometer of Big Beef Creek between 1970 and 1977 was 4700 m^3, or 9000 t, an average load of 1400 t/year. Some of the 1400 t/year may represent sediment moving out of storage from upstream reaches, rather than bedload input from hillslope processes.

Table 4--Annual sediment input to Big Beef Creek

	Suspended load	Bedload
	-Metric tons/year-	
Soil creep	147	98
Naturally occurring landslides	153	102
Disturbance-related landslides	507	338
Sheetwash on road surfaces	1928	482
Scour of inboard ditches	264	176
Total	3000	1200

The computed sediment budgets suggest that total load of the stream increased from 500 t/year under undisturbed conditions to over 4200 t/year at present. Even if absolute values of sediment input are contested, the apparent contrast of sediment loads under undisturbed and disturbed conditions are large and significant.

Sediment Transport

Once sediment enters a channel, it is transported downstream during high flows. Bedload transport represents work done by a stream and is a function of the boundary shear stress generated at any given moment and place. Boundary shear stress, τ_b, produced by a fluid is calculated as:

$$\tau_b = u_* \rho \qquad (1)$$

where u_* is related to velocity u at distance z from the channel boundary by

$$u = (u_*/0.4)\ (\ell n\ z/z_o) \qquad (2)$$

where ρ = density of the fluid, u_* = shear velocity, and 0.4 = von Karman's constant. The depth at which velocity apparently goes to zero as the channel bed is approached is the roughness length of the boundary, z_o, and is dependent upon particle size. It can be estimated by $D_{65}/30$, where D_{65} is the grain diameter that is greater than 65 percent of the bed material (Graf 1971).

Bedload transport rates for the gaging site at Big Beef Creek (fig. 1) before and after disturbance were calculated from Einstein (1950), Meyer-Peter and Müller (1948), and Emmett (1976) (table 5). I have discussed their equations and the applicability to Big Beef Creek (Madej 1978). Dimensions of the undisturbed channel were estimated from regional channel-geometry results; those of the disturbed channel were obtained from survey measurements. Measured frequencies of discharge for each year were used. The three equations predict that an average of between 600 and 1400 t of bedload/year would be transported past the gaging site.

INTERPRETATION

Present channel morphology allows greater bedload transport than the undisturbed channel (table 5). Analysis of boundary shear stress in the channel leads to the same conclusion. An increase in stream-channel capacity must be caused by an increase in τ_b. Using U.S. Geological Survey velocity and discharge measurements, I compared magnitude and distribution of shear stresses at four sites near the gaging station at equivalent discharges over a 4-year period (Madej 1978). Figure 4 shows the results of this analysis at one site, where erosion occurred at the right bank and the channel was 2 m wider in 1975 than in 1971.

Several common characteristics were found in the four analyses. The channel became wider and shallower through time (also shown by periodic cross-sectional surveys). Large gravel bars have low shear stresses above them and high shear stresses around them, perhaps caused by flow accelerating around the barrier. Most important, high shear stresses are found in areas of bank erosion, and the high stresses can be maintained over a shallower cross section. Thus, the ability of the channel to transport sediment is increased under the new channel configuration.

Channel Changes: Possible Modifications

The geometry of alluvial rivers can adjust to changing conditions imposed upon them (Mackin 1948, Leopold and Maddock 1953). In Big Beef Creek, an increase in sediment load was accommodated by changes in several interrelated hydraulic variables. Grain size and lithology of sediment supplied to a stream are variables independent of the stream system, as is discharge entering a stream. Other factors are semidependent and can undergo some modification by a stream in response to changes, i.e., flow resistance, mode of sediment transport, sediment-particle size, and channel pattern. Some parameters are almost totally functions of processes of the stream itself and can be considered to be dependent: velocity, width, depth, and channel gradient.

Table 5--Prediction of annual bedload transport in metric tons/year

	Investigator		
	Einstein (1950)	Meyer-Peter and Müller (1948)	Emmett (1976)
Undisturbed channel	645	580	340
1971	1510	1430	1030
1972	660	1270	630
1973	500	1350	470
1974	1940	3255	995
1975	70	55	30
1976	715	1090	395
Mean annual transport after disturbance	900	1410	590

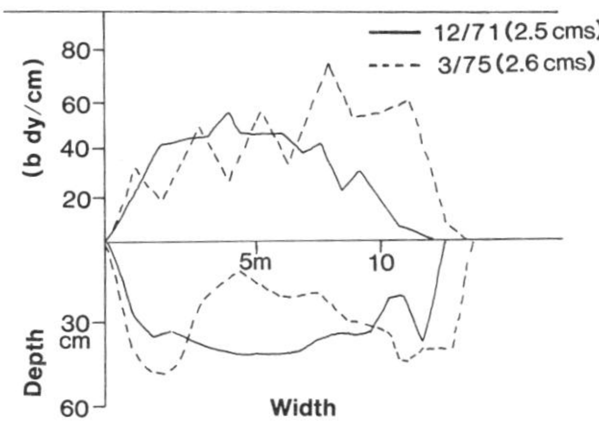

Figure 4.--Distribution of boundary shear stress in main channel, 1.6 km above the mouth of Big Beef Creek.

In response to an increase in sediment load, the more dynamic factors in a stream change to accommodate the increase. Adjustments are limited by certain physical relationships, however. The first of these is the continuity equation:

$$Q = w \bar{d} \bar{u} \tag{3}$$

where Q = discharge, w = width, $\bar{d}$ = mean depth, and $\bar{u}$ = mean velocity.

The Manning equation (4) defines the relationship between $\bar{u}$ and $\bar{d}$:

$$\bar{u} = \frac{d^{2/3} s^{1/2}}{n} \tag{4}$$

where s = slope of the stream-energy line and n = turbulent-friction coefficient (roughness). In addition, sediment transport, Q_s, is here assumed to be dependent upon stream power:

$$Q_s \propto w\bar{d}\bar{u}s \quad . \tag{5}$$

The manner in which the Big Beef Creek channel changed is summarized in a previous section. Theoretically, however, these changes do not represent the only possible changes, and it is important to explain why the channel has been modified in this manner. Because transport of sediment represents work accomplished, the amount of sediment transport was used as the criterion for capacity for work among various channel configurations. The Einstein method was used to compare the transport capability of several channel configurations.

Several combinations of width, depth, and velocity (restricted by continuity and roughness relationships) were tested with the Einstein computer program (fig. 5). An increase in width/depth ratio indicates increased bedload sediment transport (Schumm 1968, Pickup 1976). This holds true if velocity is held constant (line A in fig. 5), implying a decreasing roughness coefficient.

In a second set of calculations, velocity changed. If slope and roughness are considered to be a constant, k, the value of velocity from the Manning equation is $kd^{2/3}$, and discharge is defined as: $Q = kwd^{5/3}$. In this case (line B in fig. 5), rates of sediment transport decrease with increasing width/depth ratio. Therefore a change in the roughness 'n' apparently is needed to increase transport rates.

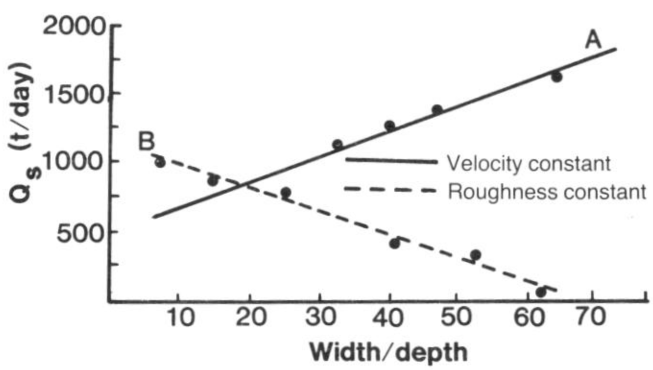

Figure 5.--Rates of sediment transport at various width-to-depth ratios.

Discharge measurements indicate that bankfull velocity remained about constant over the last 8 years, although the width-to-depth ratio, w/d, increased from 19 to 27. Line A in figure 5 suggests that w/d can increase indefinitely to transport increasingly larger sediment loads; however, in reality, stream width does not increase indefinitely. To understand what mechanism limits the increase and where these limits may be expected to occur, the effect of bed-particle roughness on flow must be considered.

Coarse surficial particles on a river-channel bed cause turbulence and energy loss in water flowing over them. The effect of these particles is most pronounced at low flows. Leopold et al. (1964, p. 160) described this phenomenon in gravel-bedded streams using a ratio of depth of water (d) to particle size (D_{84}) of the armor layer. The relationship is described as:

$$n_p = d^{1/6}\,[8g\,(2\log d/D_{84} + 1.0)^2]^{-1/2} \qquad (6)$$

where n_p is roughness from bed particles and g is the acceleration from gravity, 980 cm/second^2.

Hence roughness from bed particles is dependent only on depth of water and not on width or velocity. Values of n predicted by this method are minimum values, because other forms of roughness such as gravel bars, log jams, bank roughness, and meanders increase n. But water cannot flow over a rough bed with values of n below that predicted by the d/D_{84} relationship (equation 6).

At any given discharge, if width and depth are defined, velocity is known through the continuity equation. If velocity and slope remain constant as depth changes, roughness as defined by the Manning equation must change. Figure 6 shows the relationships of width, depth, velocity, and roughness at bankfull discharge (10 m^3/second) in Big Beef Creek. Because no streams in the regional survey show a w/d less than 18, only those greater than 18 are considered. Actual ratios are below 27, but hypothetical cases with higher ratios are considered as well. Because survey measurements show little change in channel gradient, gradient is assumed constant in this analysis. Manning's definition represents total roughness, n_t, contributed by several factors; the value of n_t decreases with increasing width and is given for each location considered.

With decreasing depth, roughness from bed particles, n_p, increases (from equation 6), and horizontal contours of figure 6 indicate the value of n_p at different depths of flow.

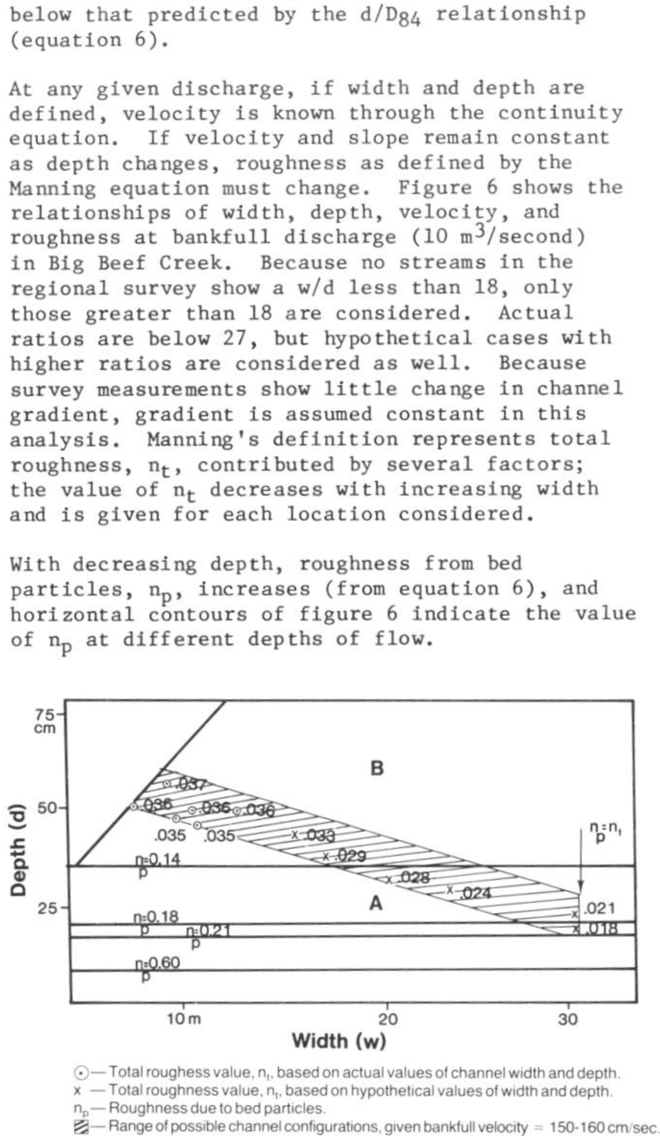

Figure 6.--Velocity and roughness constraints on the Big Beef Creek channel configuration at bankfull discharge (relationships determined from equations 3, 4, and 6).

Records from the U.S. Geological Survey gaging station show that mean velocity at bankfull stage remained between 150 and 160 cm/second over an 8-year period. Only a small portion of the diagram represents channel configurations where bankfull discharge is accommodated within this velocity range (the stippled area of fig. 6). Because velocity remained constant despite changes in channel form, I assumed velocity would remain constant if the width/depth ratio increased, and so the stippled area also includes hypothetical configurations.

As long as discharge can flow with a roughness term greater than that predicted by the d/D_{84} relationship, the channel configuration is possible. Conversely, once bed-particle roughness is greater than the roughness needed to keep velocity constant, satisfying conditions of continuity, $Q = wdu$, becomes impossible. Thus, a hydraulic constraint prevents unlimited increases in channel width.

This relationship implies that the amount of sediment that can be transported as bedload is limited. Sediment transport can increase with width-to-depth ratio only as long as roughness-velocity constraints are met. If all sediment supplied to the channel cannot be transported at the maximum possible width/depth configuration, aggradation occurs.

Configurations that are not within the constant velocity envelope probably do not exist, because the amount of work necessary to create those situations is large. For example, point A in figure 4 is associated with a velocity of 213 cm/second, and $n_t = 0.0195$. Such a low value for total channel roughness is rarely, if ever, found in gravel-bedded streams (Barnes 1967). Alternately, point B in figure 4 represents a situation where velocity is 76 cm/second and $n_t = 0.087$. Because n_p is only 0.008 at this depth, the n_t value of 0.087 would be difficult to reach even with the contribution of other causes of roughness, and this configuration is improbable.

Big Beef Creek does not display the maximum theoretical width of 30 m. Several reasons may account for this. A stable channel must have boundaries that can withstand shear stresses imposed by the flow. Using Graf's method (1971), critical shear stress on channel walls in Big Beef Creek is between 55 and 150 dy/cm^2, above which erosion occurs. These stresses are approached and probably exceeded at high flows (fig. 4); however, the resistance of banks caused by vegetation must be considered. A cover of Bermuda grass can resist stresses up to 480 dy/cm^2 (Leopold et al. 1964). The thick mat of tree and grass roots on the banks of Big Beef Creek probably inhibits bank erosion.

Undercutting of banks below the root zone does occur, but not rapidly. The channel at the U.S. Geological Survey gaging station underwent the largest changes in 1973-74, when discharges were in excess of 20 m^3/second. Maximum discharge since that time has been 13 m^3 second. More widening may occur when another high discharge is reached, but it seems unlikely under present vegetation and discharge conditions that the channel will erode 15 m to reach 30 m.

The valley widens near the mouth of Big Beef Creek, and channel width does reach 30 m, but this partly reflects weir and tidal action effects on channel geometry.

Storage of Sediment in the Channel

In several surveyed as well as unsurveyed reaches, much sediment eroded from roads and recently harvested hillslopes is obviously stored in the channel in the form of gravel bars. Field mapping indicates that the total volume of stored channel material is about 520 000 m^3 (1 092 000 t), of which 23 400 m^3 (49 200 t) is active material (i.e., in unvegetated bars down to a scour depth of 0.4 m) (table 2). Most active material lies in the lower 6.5 km, where channel braiding and log jams are prevalent.

An estimate of potential residence time for active material under present conditions is obtained by dividing the amount by the bedload-transport rate after disturbance, 1000 t/year, or 50 years. An average rate of movement can be estimated by dividing channel length, 10.3 km, by 50 years, or 206 m/year. Milhous (1975) obtained a particle-transport rate of 234 m/year through bedload measurements in an undisturbed, steeper channel with coarser gravel.

Relation of Channel Changes to Land Use

Changes in the Big Beef Creek channel can be related to land use in the drainage basin. Intense logging in the mid-1950's 2000 m above the mouth dramatically increased sediment load in that reach, and aggradation occurred. Assuming bed material moves 200 m/year, effects of excess sediment would be felt near the mouth in (2000 m)/(200 m per year), or 10 years, about 1965. In fact, local residents noted excessive sedimentation in the winters of 1966-68.

Present cross-sectional surveys show degradation in the logged reach and aggradation near the mouth, suggesting the slug of sediment is being transported downstream. A high bedload rate in this reach is also indicated by field observations. Channel length decreased, and the number and size of midchannel bars increased.

A similar trend is seen 6.2 km upstream (fig. 2C). Road construction and logging in the late 1960's caused aggradation of the adjacent stream channel. Presently, net degradation is occurring there, and channel widening and braiding occur downstream for 0.8 km. This accumulation constitutes a second distinct wave of sediment in Big Beef Creek, and it will affect the stream for several years to come.

CONCLUSION

In forested drainage basins, sediment is supplied to stream channels by soil creep and mass movements. Channel form is adjusted to the amount of sediment supplied. Logging, road construction, and urban development remove vegetation, disrupt the ground surface, and cause accelerated erosion from sheetwash, mass movements, and rainsplash erosion, and hence increase the sediment load. Under undisturbed conditions, Big Beef Creek received 500 t of sediment per year. Land-use changes increased the load to 4200 t/year. Construction of a weir at the mouth of the stream caused the bedload fraction to be caught above the weir, where an average of 1400 t of sand and gravel are deposited annually.

Stream-channel dimensions changed in response to increased sediment load. The channel in 1978 was wider and shallower than the 1970 channel, more gravel bars were present, and sinuosity decreased slightly in areas of high sediment transport. Mean flow velocity remained about constant. The changes result in an increase in shear stress along bed and banks, which in turn results in a higher rate of sediment transport. Before disturbance, sediment-transport rate of bedload was probably 300-600 t/year; at present, it is 600-1400 t/year.

Channel adaptations are restricted by hydraulic constraints described in the continuity and Manning equations. Velocity and slope are relatively conservative parameters, and most change is accommodated by width and depth.

The stream channel presently has about 1 000 000 t of sediment in storage. Active sediment, 49 000 t, moves an average of 200 m/year in the main channel. Thus sediment placed in the channel by present disturbances will take on the order of 20 to 40 years to be removed.

LITERATURE CITED

Anderson, H. W.
1954. Suspended sediment discharge as related to streamflow, topography, soil, and land use. Trans. Am. Geophys. Union 35(2):268-281.

Barnes, H. H., Jr.
1967. Roughness characteristics of natural channels. U.S. Geol. Surv. Water Supply Pap. 1849, 213 p.

Cederholm, C. J.
1972. The short-term physical and biological effects of stream channelization at Big Leaf Creek, Kitsap County, Washington. M. S. thesis. Univ. Wash., Seattle. 80 p.

Dietrich, W. E.
1975. Sediment production in a mountainous basaltic terrain in central coastal Oregon. M. S. thesis. Univ. Wash., Seattle. 81 p.

Diseker, E. G., and E. C. Richardson.
1961. Roadside sediment production and control. Am. Soc. Agric. Eng. 4:62-68.

Einstein, H. A.
1950. The bedload function for sediment transportation in open channel flows. U.S. Dep. Agric. Tech. Bull. 1026, 71 p.

Emmett, W. W.
1976. Bedload transport in 2 large gravel-bed rivers, Idaho and Washington. In Proc. 3d Fed. Interagency Sediment. Conf., Water Resourc. Counc., p. 4-101-4-114. Denver, Colo.

Everett, V. R.
1963. Slope movement, Neotoma Valley, southern Ohio. Rep. Inst. Polar Stud. No. 6, 16 p. Ohio State Univ., Columbus.

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, Oreg.

Graf, W. H.
1971. Hydraulics of sediment transport. 513 p. McGraw Hill, New York.

Hollingshead, H.
1971. Sediment transport measurements in gravel rivers. J. Hydraul. Div., Am. Soc. Civ. Eng. 97(HY11):1817-1834.

Hornbeck, J. W., and K. G. Reinhart.
1964. Water quality and soil erosion as affected by logging in steep terrain. J. Soil Water Conserv. 19(1):23-27.

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

Kuenen, P. H.
1956. Experimental abrasion of pebbles. II. Rolling by current. J. Geol. 64:336-368.

Leopold, L. B., and T. Maddock.
1953. The hydraulic geometry of stream channels and some physiographic implications. U.S. Geol. Surv. Prof. Pap. 252, 57 p. Washington, D.C.

Leopold, L. B., M. G. Wolman, and J. P. Miller.
1964. Fluvial processes in geomorphology. 522 p. W. H. Freeman, San Francisco.

Mackin, J. H.
1948. Concept of the graded river. Geol. Soc. Am. Bull. 59:463-512.

Madej, M. A.
1978. Response of a stream channel to an increase in sediment load. M. S. thesis. Univ. Wash., Seattle. 111 p.

Meyer-Peter, E., and R. Müller.
1948. Formulas for bedload transport. Proc. 2d Int. Assoc. Hydr. Struct. Res., p. 39-64. Stockholm.

Milhous, R. T.
1975. Sediment transport in a gravel-bottomed stream. Ph. D. thesis. Oreg. State Univ., Corvallis. 232 p.

Pickup, G.
1976. Adjustment of stream channel shape to hydrologic regime. J. Hydrol. 30:365-373.

Pierson, T.
1977. Controlling factors and geomorphic significance of debris flows on forested hillslopes in the central Oregon Coast Range. Ph. D. thesis. Univ. Wash., Seattle. 166 p.

Schumm, S. A.
1968. River adjustment to altered hydrologic regimen-Murrumbidgee River and paleochannels, Australia. U.S. Geol. Surv. Prof. Pap. 598, 65 p. Washington, D.C.

Slaymaker, O., and H. J. McPherson.
1977. An overview of geomorphic processes in the Canadian Cordillera. Z. Geomorphol. 21(2):169-186.

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

U.S. Environmental Protection Agency.
1971. Studies on effects of watershed practices on streams. Water Pollut. Control Ser., 56 p. Oreg. State Univ. Press, Corvallis.

Wolman, M. G.
1954. A method of sampling coarse river bed material. Trans. Am. Geophys. Union 35(6):951-956.

Wolman, M. G., and A. P. Schick.
1967. Effects of construction on fluvial sediment, urban and suburban areas of Maryland. Water Resour. Res. 3:451-464.

Young, A.
1960. Soil movement by denudation processes on slopes. Nature 188:120-122.

Upland Sediment Yield and Shoreline Sediment Delivery in Coastal Southern California

Brent D. Taylor

EXTENDED ABSTRACT

In coastal southern California (fig. 1), the natural sediment system involves the continual relocation of surface geologic materials. Surface materials are eroded hydraulically from inland areas with sufficient relief and precipitation and then deposited on low-gradient coastal plains and at the shoreline where the velocity of surface runoff decreases. Generally, coarse sand, gravel, and larger particles are deposited near the base of the mountains and hills on alluvial fans, and finer sediments are deposited further downstream on floodplains, in bays or lagoons, and along the shoreline. Very fine silt and clay particles, which make up a significant part of the eroded material, are carried offshore where they eventually accumulate in deeper areas. Finally, sand delivered to the shoreline is moved along the coast by waves and currents until it is lost to offshore areas through submarine canyons and other paths.

Human developments in this region have substantially altered the natural sediment-transport system--through timber harvest, grazing, and sand and gravel mining; the construction and operation of water conservation facilities and flood-control structures; and shoreline developments.

Almost always, these developments have grown out of recognized needs and have served their primary purposes well. Possible deleterious effects on the local or regional sediment balance were initially unforeseen or felt to be of secondary importance.

Since 1975, the Environmental Quality Laboratory (EQL) at the California Institute of Technology and the Shore Processes Laboratory at Scripps Institute of Oceanography have conducted a joint study to define quantitatively:

- The natural system of inland and coastal transport of sediment in southern California.

- The extent to which human activities have altered this natural system.

A comprehensive EQL report, currently being prepared, is scheduled for publication during 1981 and will be available for general distribution.[1]

Inland areas can be identified as either erosional or depositional according to recent geological activity. Erosional areas have been subdivided into drainage basins, and estimates of average annual sediment yield have been computed based on

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[1] Sediment Management for Southern California Mountains, Coastal Plains and Shoreline, Environ. Qual. Lab. Rep. No. 17, Calif. Inst. Technol. 1981, 100 p. 17-A: Regional geological history, Edward W. Fall, 1981, 30 p., 4 maps; 17-B: Inland sediment movements by natural processes, Brent D. Taylor, 1981, 100 p., 2 maps; 17-C: Coastal sediment delivery by major rivers in southern California, William R. Brownlie and Brent D. Taylor, 1981, 292 p.; 17-D: Special inland studies, William W. Brown III, Oded C. Kolker, Nancy Palmer, Wade Wells, 1981, 150 p., 4 maps; 17-E: Shoreline sedimentation processes, Douglas L. Inman, Abraham Golick, and Martha Shaw, Scripps Inst. Ocenogr., 1981, 250 p.; 17-F: Special coastal studies, Martha Shaw, Scripps Inst. Oceanogr., 1981, 150 p., 2 maps.

debris-accumulation data from 36 entrapment structures located in the study area and a simple regression model relating sediment yeild to dominant land type and drainage-basin areas.

Measured denudation rates vary from about 0.01 mm/year on plains areas to 3 to 5 mm/year in small, mountainout drainage basins. Denudation rates for each land type are remarkably uniform throughout the study area despite significant variations in rock types, geological history, and the size distribution of eroded materials.

Study results indicate that in these coastal drainage basins, an average of more than 12 million m^3 of sediment leave upland drainage basins each year. This material (6 million m^3 fines, 5 million m^3 sand, and 1 million m^3 of coarser material) is first delivered to inland valleys, coastal plains, or directly to the shoreline--depending on drainage-basin location.

Estimates have also been obtained for average annual sediment delivery to coastal areas from each of the larger hydrographic drainage units (fig. 1, table 1). Specific effects on upland erosion and shoreline sediment delivery of human developments and artificial control structures are also determined. Available data indicate that while upland erosion on 10 major rivers that drain 64 percent of the 31 000-km^2 coastal area has been altered very little, shoreline sediment delivery of beach-sized sand from these rivers has been reduced some 50 percent over the past 50 years.

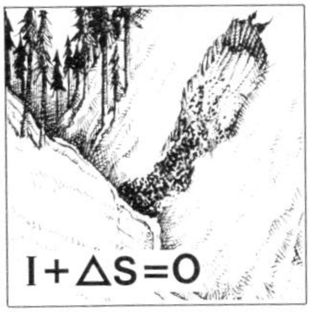
Comparisons of sediment yield from upland drainage basins with coastal delivery by major rivers suggest that, under natural conditions, alluvial rivers in the southern part of the study area are depositional along their floodplains, with only a fraction of upland sediment production being delivered to the shoreline. The three northern rivers, on the other hand, are erosional on their floodplains and deliver more sediment to the shoreline than is yielded from their respective upland basins.

Figure 1.--Coastal southern California hydrographic drainage units identified in table 1.

Table 1--Major drainage units in the sediment-management study area

Map symbol	Principal basin or group of small basins	Basin type	Controlled drainage area	Total drainage area	Percent of drainage area controlled
			- - - Square kilometers - - -		
A	Santa Ynez Mountains Group	G	--	901	--
B	Ventura River Basin	SMD	243	585	42
C	Ventura Group	G	--	52	--
D	Santa Clara River Basin	SMD	1 527	4 219	37
E	Oxnard Group	G	--	159	--
F	Calleguas Creek Basin	SMD	--	837	--
G	Santa Monica Mountains Group	G	166	1 493	11
H	Los Angeles River Basin	SED	3/866	2 155	40
I	Long Beach Group	G	--	120	--
J	San Gabriel River Basin	SED	1 400	1 663	84
K	Huntington Beach Group	G	--	234	--
L	Santa Ana River Basin	SED	3 950	4 406	90
M	Lake Elsinore Basin	SC	1 989	1 989	100
N	Laguna Hills Group	G	--	1 737	--
O	Santa Margarita River Basin	SMD	958	1 927	50
P	San Luis Rey River Basin	SMD	531	1 450	37
Q	Escondido Creek Group	G	--	568	--
R	San Dieguito River Basin	SMD	785	896	88
S	San Clemente Canyon Group	G	--	437	--
T	San Diego River Basin	SMD	686	1 119	61
U	San Diego Group	G	--	157	--
V	Sweetwater River Basin	SC	471	567	83
W	Otay River Basin	SC	255	370	69
X	Tijuana River Basin	SMD	3 175	4 483	72
Total			17 002	32 524	53

1/G, drainage basin group; SED, single, extensively developed basin; SC, single confined basin; SMD, single, moderately developed basin.
2/Calculated by adding the drainage areas controlled by the major water-retention structures that are farthest downstream in each basin.
3/Whittier Narrows Flood Control Basin controls both Los Angeles and San Gabriel Rivers. This estimate assumes that 35 km^2 of the drainage area controlled by the Whittier Narrows structure lies within the Los Angeles River drainage basin.
4/Excludes Lake Elsinor Basin (M).
5/Closed interior basin. Overflow into Santa Ana River basin has not occurred since 1916.



Ashland Creek Drainage Basin Sediment Budgets and Routing Studies

Richard D. Smith and Bill G. Hicks

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EXTENDED ABSTRACT

The Ashland Creek drainage basin which supplies water to the city of Ashland, Oregon, has experienced major problems with erosion and landslides. Debris avalanches, originating in weathered granitic rock, have moved down the major drainages, damaging the water-supply system and partially filling the water-supply reservoir.

Two studies to develop partial sediment budgets have been conducted to assess the relative effects of natural and management-related sediment production. The first was completed by the Rogue River National Forest[1] in 1975 and specifically dealt with the material deposited in Reeder Reservoir at the base of the 5020-ha drainage basin by a major storm in January 1974. The second study was completed in 1977 by J. M. Montgomery Consulting Engineers,[2] Boise, Idaho. This study, prepared for the City of Ashland, was initiated because of the 1974 storm, but included an analysis of sediment production in the Ashland Creek drainage for the period 1955-76.

At the time of the January 1974 storm, 85 km of road had been constructed in the basin, covering about 90 ha or 2 percent of the area including cut and fill slopes. Clearcutting and partial cutting had occurred in 7 percent of the basin and another 1 percent had been thinned.

The USDA Forest Service study included mapping of all failures in the basin, subdividing them into management-related and natural events. Rough estimates were made of failure volumes released into the system. These volumes were then compared with the volume deposited in the reservoir during the storm event. The difference between the landslide failure volume and volume collected in the reservoir was then subdivided into stream-channel erosion of natural and management-related, pre-1974 material in channels. Estimates of surface erosion from roads were about equal to the estimated natural surface erosion for the 1974 storm. Additional assumptions and interpretations were made to arrive at comparative estimates of management-related and natural sources of material deposited in the reservoir (table 1). About 60 percent of the material in the reservoir was attributed to management-related activities and 40 percent to natural events.

The Montgomery study covered the full range of erosional factors in the basin. It developed a simulated history of sediment production for the entire Ashland Creek drainage. To approximate erosion rates related to roads and timber harvest, data from granitic terrane of the Idaho batholith were used. Significant volumes were attributed to road-related erosion, including effects of annual road maintenance. The study stressed that the technique used analyzes only surface and mass erosion for the entire drainage basin. The data and conclusions do not deal with routing through the drainage system to the reservoir. The approach was designed to approximate sediment supply to channels in response to various management activities within the drainage basin. This portion of the Montgomery study highlights the various factors affecting the total erosion rate and thus forms a basis for making decisions on where to intensify field investigations. The study identifies the sources of an estimated 270 000 m^3 of soil erosion from 1955 to 1977: natural (3 percent), surface erosion from 1- to 2-year-old road surfaces (6 percent), surface erosion from older road surfaces (58 percent), surface erosion from cut and fill slopes of roads (1 percent), mass movement from roads and timber harvest sites (20 percent), surface erosion from timber-harvest units (3 percent), and surface erosion from ski areas (9 percent).

Contrasts between the two studies result from the use of different data bases, recognition of channel-sediment sources in the Forest Service study and not in the Montgomery study, and differences in the time period analyzed.

Table 1--Estimates of sediment derived from various sources in Ashland Creek drainage basin[1/]

Source	Volume
	Cubic meters
Management-related:	
Mass failure of road fills	21 700
Other mass failures from roads	6 200
Mass failures from clearcuttings	150
Stream-channel erosion (debris-slide effect)	15 300
Management-related, pre-1974 material in channels	15 300
Management-related subtotal	58 650
Natural:	
Stream-channel erosion (major storm-flow effect)	35 600
Stream-channel erosion (debris-slide effect)	1 700
Debris slides	3 600
Natural subtotal	40 900
Total sediment delivery to reservoir	99 550

1/See text footnote 1.

[1]Wilson, S., and B. G. Hicks. 1975. Slope stability and mass wasting in the Ashland Creek watershed. Unpubl. rep., on file Rogue River Natl. For., Medford, Oreg.

[2]J. M. Montgomery Consulting Engineers. 1977. Water resources management plan and facility study. Unpubl. rep., on file Rogue River Natl. For., Medford, Oreg.

Channel Sediment Storage Behind Obstructions in Forested Drainage Basins Draining the Granitic Bedrock of the Idaho Batholith

Walter F. Megahan

ABSTRACT

Data on sediment storage behind obstructions were collected on seven forested, mountain drainage basins in the Idaho Batholith for a 6-year period from 1973-78. Four of the drainage basins were undisturbed throughout the study period, one contained an old road, and two were logged during the course of the study. The total volume of sediment stored behind obstructions varied between drainage basins and years in response to changes in bankfull channel width and annual peak-flow rates, respectively. Logs were the most important type of obstruction because they had the greatest longevity and stored the greatest amount of sediment. An average of 15 times more sediment was stored behind obstructions than was delivered to the mouths of the drainages as annual average sediment yield. Logging reduced total channel-sediment storage behind obstructions because many natural obstructions were destroyed by felling and subsequent clearing operations to remove logging debris from channels. Storage behind obstructions is an important component of the overall sediment routing through forested drainage basins. Accordingly, erosion and sedimentation monitoring must be carefully designed to avoid misinterpretation. Also, some guidelines are presented to help minimize the change in channel-sediment storage caused by timber harvest.

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INTRODUCTION

Development of a sediment budget for a drainage basin requires an accounting of: (1) onsite erosion; (2) the amount and rate of transport of eroded material between source and stream channels; (3) channel-sediment storage; and (4) sediment outflow from the basin. Channel-sediment storage is especially important on forested drainage basins because of additional storage potential provided by obstructions. Obstructions from logs and other debris occur naturally on forested areas, but their size and abundance may be strongly influenced by stand conditions and by disturbances such as logging (Froehlich 1973, Swanson et al. 1976, Swanson and Lienkaemper 1978).

Information on the volume of sediment storage and the type, size, number, and longevity of the obstructions causing the storage is needed to provide better understanding of channel-storage processes in forested drainage basins. Megahan and Nowlin (1976) reported on a study of sediment storage in channels draining seven small forested drainage basins in the mountains of central Idaho. Part of that study included an inventory of sediment trapped behind obstructions in 1973-74. This paper summarizes the progress made through 1978 in the channel obstruction portions of the study.

DESCRIPTION OF THE STUDY AREA

The seven drainage basins are in the Silver Creek study area in the Middle Fork of the Payette River drainage near Crouch, Idaho.

Descriptive data for the study drainage basins are presented in table 1. Four of the seven study drainage basins are undisturbed; the Ditch Creek drainage contains a low-standard road constructed during the 1930's and Control and K-1 Creeks were logged by helicopter in November 1976. The forest cover is dominated by Douglas-fir (*Pseudotsuga menziesii* [Mirb.] Franco) and ponderosa pine (*Pinus ponderosa* Laws.) on the slopes and grades into grand fir (*Abies grandis* [Dougl.] Forbes) and subalpine fir (*Abies lasiocarpa* [Hook.] Nutt.) near drainage bottoms. Timber volumes average about 117 m^3/ha in the vicinity of drainage bottoms.

The area is representative of conditions in the Idaho Batholith, a 41 400-km^2 expanse of granitic bedrock in central Idaho (fig. 1). This mountainous area is characterized by steep slopes; shallow, extremely erodible, coarse-textured soils; and large climatic events resulting from rainfall, snowmelt, or both. Erosion hazards are high and soil disturbances, both natural and caused by human activities, can greatly accelerate erosion and consequent sedimentation (Megahan 1975). Much of the total sediment discharge occurs as bedload, because of the coarse texture of the granitic parent materials. Streamflow and sediment yield exhibit marked seasonal variation in response to large winter storms, spring snowmelt, and long, dry periods in summer.

METHODS

Data on sediment storage behind obstructions were collected at sample reaches in each study drainage basin. Each sample reach is 43 m long. The first sample reach is located 30 m above a sediment basin at the mouth of each drainage. Additional sample reaches are located at 152-m intervals along the dominant channel until no indicators of perennial streamflow are obvious (such as a well-defined channel cross section or aquatic vegetation). The dominant channel was defined by following the mainstream up from the drainage-basin outlet during the period of baseflow in August and selecting the tributary with the greatest flow at each confluence.

Table 1--Descriptive data for seven drainage basins on the Silver Creek study area

Drainage	Area	Mid-elevation1/	Dominant aspect	Mean channel gradient2/	Total channel length3/	Stream order3/	Average bankfull width
	Sq. kilometers	Meters		Percent	Kilometers		Meters
C	1.94	1779	SE	21.5	9.5	3	1.5
D	1.22	1765	SE	24.2	5.5	2	1.1
Eggers	1.29	1733	SE	22.0	5.8	3	1.1
Ditch	1.06	1631	SE	20.8	2.9	2	0.8
Cabin	1.04	1533	SE	14.9	5.1	3	1.0
Control	2.02	1597	SE	15.2	6.8	3	1.1
K-1	0.26	1623	NW	31.5	1.0	1	1.0
Average	1.27	1666	--	21.4	5.1	--	1.1

1/(Maximum elevation + minimum elevation)/2.

2/(Total relief/length main channel to the upper ridge) x 100.

3/Taken from a 1:31,680 planimetric map.

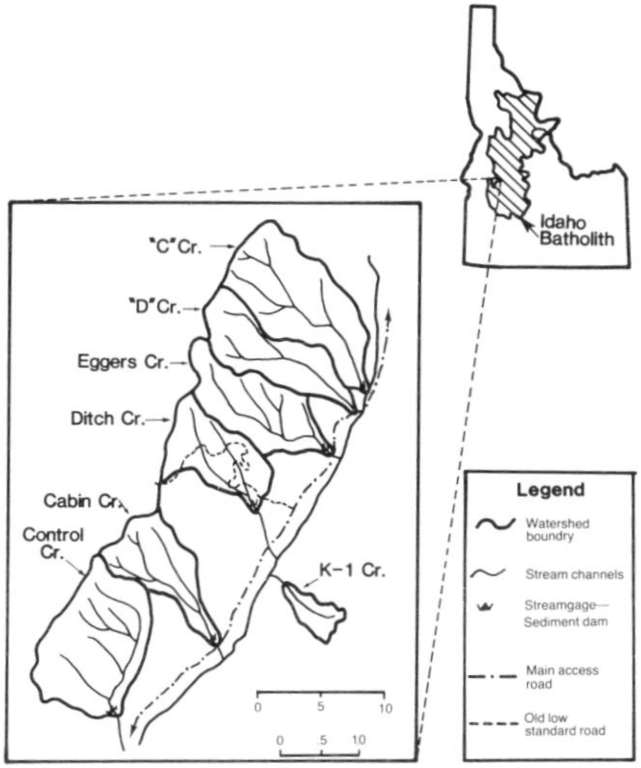

Figure 1.--Study area in the Idaho Batholith.

An average of 10 reaches was sampled on each study channel for an average of 27 percent of the study channel sampled. Because sample reaches are located only on the dominant channel in each drainage basin, only about 8.7 percent of the total channel length in each drainage basin is sampled (table 2). Data are collected annually during low flow in late July and August. All reaches on all streams were sampled only during 1973, 1974, and 1978. In 1975, all samples were taken except the upper four reaches on C Creek. C, D, Eggers, Ditch, and Cabin Creeks were not sampled in 1976; C and D were omitted again in 1977.

Data for sediment deposited behind obstructions are available for 1973-78. Obstructions are defined as any material in the channel causing sediment accumulations because of discontinuities in channel gradient. In 1973, all discernible obstructions and associated sediment accumulations were measured, a procedure that proved to be too time consuming. Sampling during the following years was restricted to obstructions with the following minimum dimensions: height, 0.2 m; average width, 0.3 m; and length, 0.6 m. The effect of the more restrictive sampling is minimal; 31 percent of the total number of obstructions fell below the allowable limit in 1973 but accounted for only 11 percent of the total volume of stored sediment. Height is defined as the difference between a level rod reading taken on the bed at the downstream side of the obstruction (the rod is raised if necessary to correct for any scouring at this point) and a rod reading taken on the sediment deposit immediately upstream from the obstruction. Rod readings are taken to the nearest 0.4 cm using an abney level. Length is the distance from the upstream end of the obstruction to the upstream end of the accumulated sediment. Width of the sediment accumulation is the average of 3 widths taken normal to the length at distances of 0.17, 0.5, and 0.83 of the length from the obstruction. The upstream end and edges of sediment accumulations are defined by breaks in channel gradient, differences in the particle-size distribution of bottom sediments, and differences in composition of bottom materials. A third rod reading is taken at the upstream end of the accumulated sediment to allow calculation of the slope of the deposited sediments. The most apparent cause of the obstruction is defined by type, as log (woody material over 10 cm in diameter), rock, root, stump, or debris, which includes branches, twigs, and leaves. Finally, the location of the obstruction is mapped.

A continuously recording streamgage and a sediment basin for determining annual sediment yields are operated at the mouth of each drainage basin. Sediment basins are surveyed twice a year in June and October using a network of closely spaced

Table 2--Sampling intensity on study drainage basins

Drainage	Number of sample reaches	Length of study channel	Percent of study channel sampled for obstructions	Percent of total channel length sampled for obstructions
		Meters		
C	14	2198	27.2	6.3
D	8	1210	28.2	6.2
Eggers	12	1868	27.4	8.8
Ditch	11	1704	27.5	16.1
Cabin	9	1539	25.0	7.4
Control	11	1704	27.5	6.9
K-1	4	552	30.9	17.7
Average	9.9	1539	27.3	8.7

Table 3--Average number of obstructions and volume of sediment stored behind obstructions in study streams per 30 m of channel sampled

Year	C	D	Eggers	Ditch	Cabin	Control[1/]	K-1[1/]	Mean
	Number/volume cubic meters							
1973	4.5/2.3	6.2/2.3	3.2/0.6	3.9/0.7	3.9/0.7	2.8/0.4	3.6/0.7	4.0/1.1
1974	3.1/1.9	3.6/1.6	2.4/0.5	3.1/0.8	2.8/0.7	2.8/0.6	4.3/1.2	3.0/1.0
1975	2.2/1.4	3.4/1.1	2.6/0.4	3.7/0.4	3.4/0.5	2.7/0.6	4.5/0.5	3.1/0.7
1976	--/--	--/--	--/--	--/--	--/--	4.2/1.0	7.5/0.9	4.9/0.9
1977	--/--	--/--	3.9/0.4	4.4/0.5	4.9/0.7	2.4/0.3	4.6/0.6	3.9/0.5
1978	3.6/1.4	4.4/0.6	3.5/0.2	3.4/0.4	5.8/0.4	3.8/0.6	5.9/0.7	4.1/0.6
Mean	3.1/1.8	4.4/1.4	2.9/0.4	3.5/0.6	4.0/0.6	3.0/0.5	5.0/0.8	3.6/0.8

[1/]Drainage basins logged in November 1976.

cross sections. Generally, more than 95 percent of the total sediment yield for the year is measured during the June survey. Trap efficiencies of the basins are estimated to average more than 75 percent because of the coarse texture of the soils on the study drainage basins (Megahan 1975). Two weather stations and a raingage network are also operated on the study area.

RESULTS AND DISCUSSION

A total of 1,715 obstructions were sampled during the 6 years of data collection. Averaging all years and streams, 3.6 obstructions were found per 30 m of channel with 0.8 m^3 of sediment storage per obstruction. This amounts to 2.9 m^3 of sediment storage per 30 m of channel or 493 m^3 for the average total channel length on the study drainage basins.

Number of Obstructions and Sediment Storage

As might be expected, the number of obstructions and stored sediment volumes vary between streams and between years (table 3). Analysis of variance tests show significant (α = 0.01) differences between streams for both number of obstructions and volume of stored sediment.

Heede (1972) reported an increase in number of obstructions with increasing channel gradient on two study streams in Colorado. A similar analysis proved unsuccessful for the streams in this study. Reasons for the lack of agreement are not clear, although the fact that the Idaho streams are smaller and steeper is probably a contributing factor. As with number of obstructions, volume of sediment stored behind obstructions was not related to channel gradient. Average volume of sediment (m^3) behind obstructions showed a weak, positive relationship to average bankfull channel width (m), however. A linear regression analysis had a coefficient of determination of 0.62 and a standard error of 0.3 m^3/30 m. The regression coefficient was 0.526 (significant at the 0.05 level) and the y intercept was -0.54 m. Other, presently unexplored, factors--such as quality, type, and condition of streamside vegetation--would likely provide better predictors of number of obstructions and sediment storage behind obstructions.

Assuming no major changes in the factors introducing obstructions to the channel (such as windstorms), variations in streamflow probably account for some of the annual changes in number of obstructions and sediment storage behind obstructions. This is illustrated by comparing the frequency distribution of sediment stored behind obstructions for years of high and low flow (fig. 2). During the high-flow year of 1974, when peak flows averaged 0.20 m^3/sec per ha, only the large, stable obstructions remained in the channel. In contrast, more obstructions with a smaller volume were found when flow energies were low, as in 1978 when average peak flows were about half those recorded in 1974.

Type of Obstruction

Debris was the most common type of obstruction, averaging about 42 percent of the total. Logs formed 34 percent of all obstructions; rocks, roots, and stumps made up an additional 13, 10, and 1 percent, respectively.

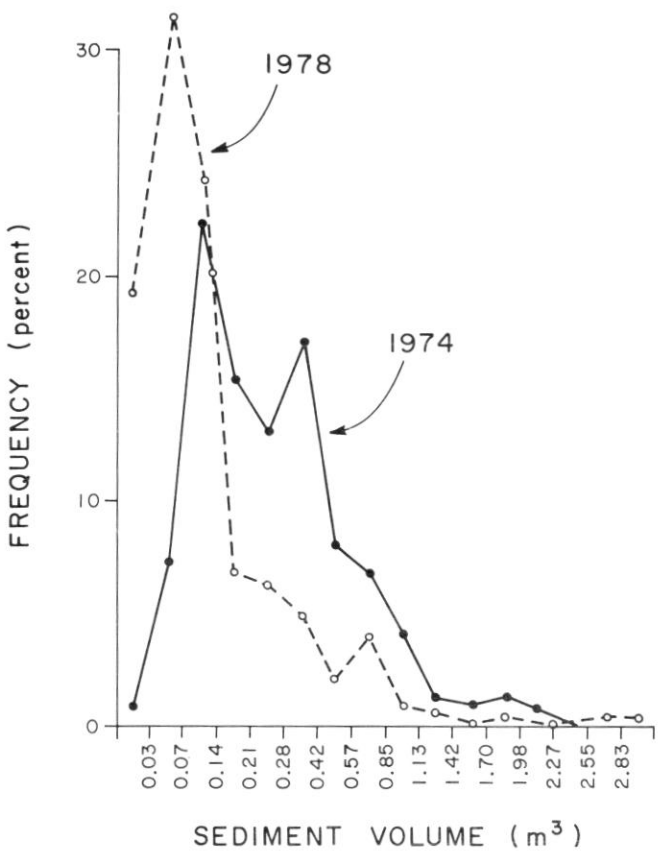

Figure 2.--Frequency of occurrence of sediment volume behind obstructions by size classes.

Because of differences in amount of stored sediment, frequency of occurrence alone does not reflect the importance of obstruction type. A comparison of the frequency of occurrence of volume of sediment storage behind the two most common types of obstructions illustrates this (fig. 3). Although fewer obstructions are logs than debris, the logs store greater amounts of sediment, which greatly magnifies the importance of logs. Logs account for 49 percent of the total sediment stored, but organic debris accounts for only 29 percent. An additional 16, 6, and 0.2 percent of the sediment storage is caused by rock, roots, and stumps.

Longevity of Obstructions

The number of obstructions in a channel at any time is a function of the rate of introduction of new obstructions to the channel from adjacent slopes, the longevity of obstructions once in place, and the rate of supply of obstruction material from upstream. Introduction of new obstructions occurs as a long-term, relatively constant supply of material from normal ecological processes in the forest--such as litterfall and random mortality--plus a stochastic component caused by natural catastrophic disturbances--such as forest fire, windstorm, and logging. Once introduced, longevity of obstructions (defined as the length of time an obstruction was found at a given location) is a function of the rate of decay of organic materials and the tendency for movement

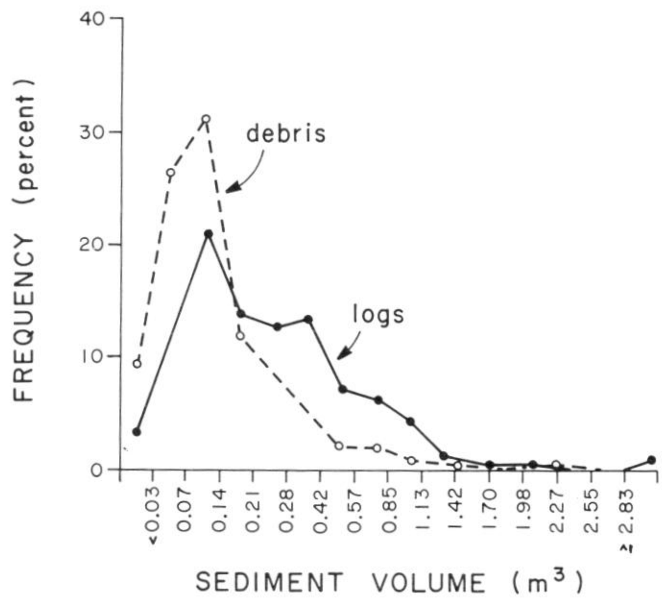

Figure 3.--Frequency of occurrence of sediment volume by size classes stored behind debris and log obstructions.

of the obstruction, erosion around it, or both. If the obstruction material moves, it then has the potential for forming a new obstruction further downstream.

Development of a model for occurrence of obstructions is beyond the scope of the present data. Information is available to illustrate the longevity of obstructions once they have formed, however. During the field survey, the distance along the channel is measured from a permanent reference point. This makes it possible to define the longevity of obstructions by comparing data for successive years. Data are available for 6 years for Control and K-1 Creeks. The longevity of all types of obstructions tends to decrease rapidly with age, but the rate of decrease for logs is lower than for the other types (fig. 4). This further emphasizes the importance of logs as channel obstructions; they not only retain the most sediment but also last longest.

Decay is probably an important process influencing the rate of decline of debris obstructions. Decay rates are slow for logs in a wet environment, however, so erosion under and around the log (or both) is probably a more important cause for failure. Logs often serve as indirect causes of obstructions by acting as channel obstacles that do not in themselves restrict flow but rather form a base for the collection of debris that does. Then, the longevity of the obstruction depends on the longevity of the secondary material rather than the log itself, even though the obstruction is classified as a log.

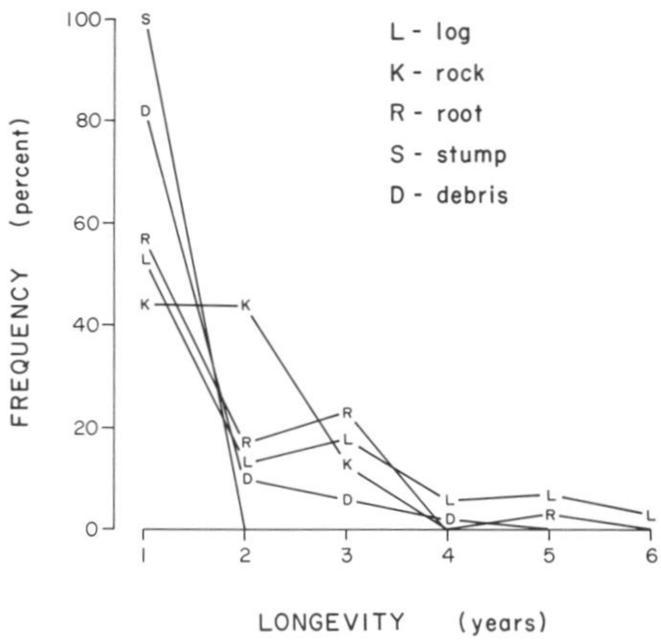

Figure 4.--Frequency of occurrence of the number of years an obstruction remains in place by obstruction type.

Volume of Sediment Storage Behind Obstructions

Channel-sediment storage is an important component of the overall sediment budget for a forested drainage basin. The three types of channel storage are: (1) temporary storage in channel bedforms as a function of flow conditions and sediment-particle size; (2) longer term storage caused by obstructions; and (3) very long-term valley storage in flood-plain deposits. The importance of storage behind obstructions is illustrated by comparing sediment storage with annual sediment yield. Annual sediment yield measured in debris basins at the mouth of each study drainage basin is much less than the average annual sediment storage for study reaches extrapolated to the total channel length (table 4). The ratio of storage to yield ranges from a high of 33 for the largest drainage (C Creek) to a low of 3.0 for Ditch Creek, and it averages 15. The ratio for Ditch Creek is much lower than that for any other drainage, because Ditch Creek contains an old, low-standard road that doubles sediment yield from the drainage basin (Megahan 1975).

Table 4--Average annual volume of sediment stored behind obstructions compared to average annual sediment yield in debris basins

Stream	Sediment storage behind obstructions	Sediment measured in debris basin	Obstruction storage/ debris basin
	Cubic meters		
C	534	16	33
D	282	11	26
Eggers	75	8	9
Ditch	52	17	3
Cabin	101	11	9
Control	127	11	12
K-1	24	3	8
Average	171	11	15

These observations emphasize the need to account for channel storage when working with sediment yields from small drainage basins in forested areas. On the average, about 15 years of annual sediment yields are stored behind obstructions. Given a large enough hydrologic event, much of this material could be flushed out of the system. Unless the previous channel storage were accounted for, this sudden increase in sediment yield might be attributed to recent onsite hillslope erosion when in fact it originated from channel-erosion processes.

Effects of Logging in the Study Drainage Basins

Control and K-1 Creeks were logged during this study. All trees down to 25 cm diameter breast high were clearcut in Control Creek on three major cutting units totaling 35 ha. The entire area of K-1 Creek was selection logged. Logs were removed from both drainage basins by helicopter. Only trees that were likely to die before the next timber harvest were marked for cutting within 15 m of active stream channels on both drainage basins. The logging contract stipulated that logging debris entering stream channels was to be removed.

The logging was done in November 1976. Channel surveys the following summer showed that logging debris had encroached on stream channels during timber harvest; in extreme cases, entire trees had rolled lengthwise into the channel. All logging slash was removed from the channel in accordance with the timber sale contract, however. Contrary to results at other locations, this logging temporarily decreased sediment storage behind obstructions, chiefly because some of the natural obstructions were destroyed by the tree felling and channel cleaning operations. This is illustrated by comparing double mass curves for the logged drainage basins to adjacent unlogged drainages (figs. 5 and 6). Although the trends shown are not statistically significant, numbers of obstructions and stored volumes tended to decrease in 1977 after logging. These results are reinforced by comparing the change in volume of sediment stored behind obstructions in a stream from one year to the next (table 5). Sediment storage decreased from 1976 to 1977 for both Control and K-1 Creeks, followed by a net increase from 1977 to 1978.

Table 5 also illustrates the need to consider all components of the sediment budget when developing drainage-basin sediment budgets. Positive and negative values in the table indicate annual increases or decreases in the total volume of sediment storage behind obstructions. A constant outflow of sediment occurs from the drainage basins each year into the sediment basins. Assuming that the figures in the table are correct, other forms of channel sediment storage and erosion processes both in the channel and on the slopes obviously must be evaluated to balance the annual sediment budget.

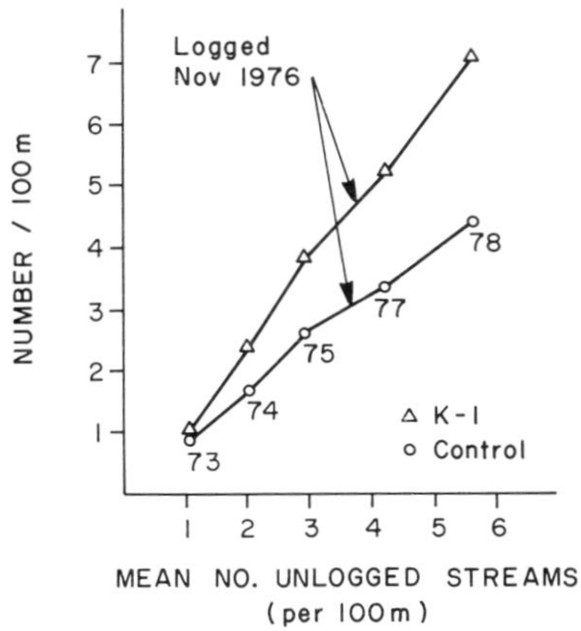

Figure 5.--Double mass analysis showing number of obstructions before and after logging.

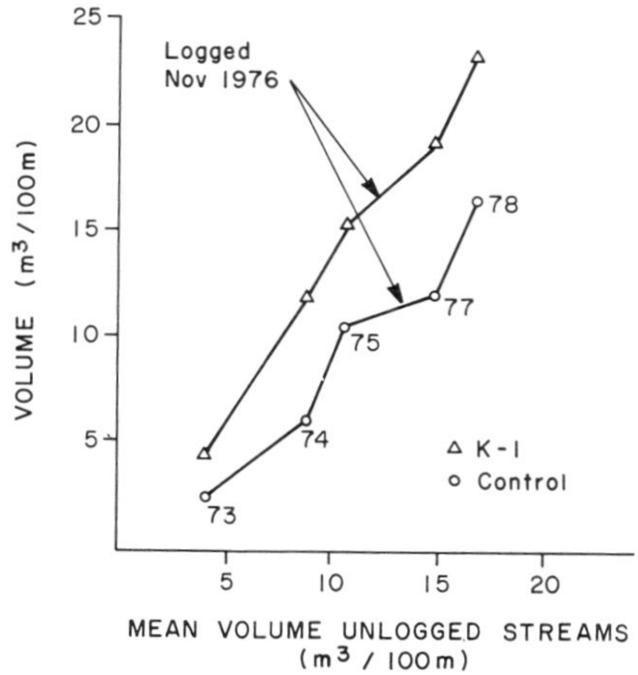

Figure 6.--Double mass analysis showing volume of sediment storage behind obstructions before and after logging.

CONCLUSIONS AND MANAGEMENT IMPLICATIONS

Both numbers of obstructions and volume of sediment stored behind obstructions varied by drainage basins and by years. Differences between drainage basins were partly accounted for by a direct relationship to bankfull channel width, and differences between years were inversely related to annual instantaneous peak flow rates. Obstructions caused by organic debris were most numerous. Log obstructions were by far the most important, however, because logs store more sediment per obstruction and last longer. Longevity of obstructions is influenced by the decay rate of the material forming the obstruction and by movement of and erosion around obstructions. Most obstructions in the study area lasted less than 2 years. Even the most permanent type of obstruction, logs, lost 97 percent of their effectiveness for storing sediment within 6 years after emplacement.

Extrapolation of storage data to entire channels shows that, on the average, about 15 times more sediment is stored behind obstructions than is yielded annually at the drainage-basin mouth. This illustrates the importance of sediment storage to overall erosion-sediment budgeting for forested drainage basins.

Clearcutting and selection logging by helicopter had little effect on channel obstructions in the study area because of minimal streamside cutting and postlogging channel clearing. The net effect of logging was to cause a small decrease in the number of obstructions and in sediment stored behind obstructions for 1 year afterward.

The total amount of sediment stored behind obstructions fluctuated between years; however, sediment outflow was measured from all basins for all years. These data further emphasize the need to consider other types of channel storage, including bedforms and flood-plain storage, in addition to streambank erosion and erosion on slopes for a complete understanding of sediment budgets for forested drainage basins.

Some important land-management implications can be derived from this study for timber harvesting and methods to monitor erosion and sedimentation responses to land-management activities.

Table 5--Change in sediment stored behind obstructions in study streams

Year	C	D	Eggers	Ditch	Cabin	Control	K-1
	Cubic meters						
1973-74	-129.2	-123.9	-14.2	+ 7.4	+ 2.4	+ 46.5	+17.8
1974-75	-350.9	- 88.7	-34.2	-38.5	-25.5	+ 13.1	-21.9
1975-76	--	--	--	--	--	+ 74.0	+12.1
1976-77	--	--	--	--	--	-150.0	-10.9
1977-78	--	--	-41.2	-13.0	-49.5	+ 76.7	+ 1.9

Timber Harvest

Timber-harvest activities should be designed to minimize changes in channel-sediment storage during and after disturbance by:

- Wherever possible, logging debris should be kept out of stream channels.

- If logging debris does get into stream channels during logging, it should be carefully removed to avoid disturbance of natural obstructions.

- An additional point to consider when evaluating the desirability of buffer strips is that they provide a source of natural channel debris that helps to stabilize channel-sediment storage over time after timber harvest.

Monitoring Erosion and Sedimentation Effects

Monitoring erosion and sedimentation from land uses is often necessary to help minimize impacts on fish, assure compliance with water-quality standards, or because of legislation calling for monitoring of the environmental effects of land-management activities. The objectives of the monitoring effort must be carefully defined and the monitoring designed accordingly. If onslope erosion is the concern on a particular area, then erosion should be measured at that point. Similarly, downstream sediment yield should be measured at a relevant downstream point if water-quality standards or fishery impacts are important. Inferences about erosion from sediment data or about sediment from data on erosion should be made with extreme caution, unless the data are collected to evaluate the third component of the erosion-sedimentation continuity equation--sediment storage. This is particularly important in areas where a large proportion of the sediment moves as bedload.

LITERATURE CITED

Froehlich, H. A.
1973. Natural and man-caused slash in headwater streams. Loggers Handb., vol. 33, 8 p. Pac. Logging Congr., Portland, Oreg.

Heede, B. H.
1972. Flow and channel characteristics of two high mountain streams. USDA For. Serv. Res. Pap. RM-96, 12 p. Rocky Mt. For. and Range Exp. Stn., Fort Collins, Colo.

Megahan, W. F.
1975. Sedimentation in relation to logging activities in the mountains of central Idaho. Proc. Sediment-Yield Workshop, USDA Sediment Lab., Oxford, Miss., Nov. 28-30, 1972. U.S. Agric. Res. Serv. Rep. ARS-S-40, 285 p.

Megahan, W. F., and R. A. Nowlin.
1976. Sediment storage in channels draining small forested watersheds in the mountains of central Idaho. In Proc. 3d Fed. Interagency Sediment. Conf., p. 4-115-4-126. Water Resour. Counc. Denver, Colo.

Swanson, F. J., and G. W. Lienkaemper.
1978. Physical consequences of large organic debris in Pacific Northwest streams. USDA For. Serv. Gen. Tech. Rep. PNW-69, 12 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Swanson, F. J., G. W. Lienkaemper, and J. R. Sedell.
1976. History, physical effects, and management implications of large organic debris in western Oregon streams. USDA For. Serv. Gen. Tech. Rep. PNW-56, 15 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Development, Maintenance, and Role of Organic-Debris Dams in New England Streams

Gene E. Likens and Robert E. Bilby

ABSTRACT

We propose that the formation of organic-debris dams on streams depends primarily on the size of tree (log) available. After disturbance, organic-debris dams are at first diminished and then form on larger and larger stream channels as the terrestrial ecosystem develops, and as a result, the regulation of erosion and transport of dissolved and particulate material from the landscape is enhanced. The species composition and phase of development of hardwood forests also may affect the occurrence and longevity of organic-debris dams. Steady-state amounts of organic matter in stream channels may reflect the stream order, as well as the developmental phase of the terrestrial ecosystem.

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INTRODUCTION

An organic-debris dam is an accumulation of organic matter that obstructs water flow in a stream. In general, these structures form when a piece of large woody debris falls into the stream. Unless the woody piece is extremely large, such as a tree bole, the wood will be carried downstream until it reaches a point in the channel where rocks and boulders, protruding from the streambed, catch and hold it against the current. Gradually, smaller sticks begin to collect against the larger piece, providing a framework on which leaves can accumulate. Ultimately, the structure becomes almost watertight, creating an upstream pool.

Only recently has the role of organic-debris dams in stream ecosystems been recognized (Zimmerman et al. 1967, Heede 1972, Swanson et al. 1976, Swanson and Lienkaemper 1978, Bilby 1979, Keller and Swanson 1979, Keller and Tally 1979). Debris dams are important in the development and maintenance of stream-channel morphology, and they also help to regulate the transport of particulate--and to a lesser extent, dissolved--materials through a stream system.

In the White Mountain region of New England, first-order (Strahler 1957) streams generally are 1.6 to 2.4 m wide and contain from 20 to 40 organic-debris dams per 100 m of stream channel. The frequency decreases to between 10 and 15 dams in second-order (2.5- to 2.8-m-wide) streams and from 1 to 6 in third-order (3.7- to 6.6-m-wide) streams (Bilby 1979). Organic debris dams are rare in streams larger than third-order. Debris dams contain 75, 58, and 20 percent of the total standing stock of organic matter in first-, second-, and third-order streams, respectively. Therefore, these structures are currently important in regulating sediment routing only in headwater streams. The questions we wish to address in this paper are: What factors are responsible for creating the relationship between stream size and frequency of organic-debris dams in streams in mountainous areas of New England? How variable is this pattern over time?

DEVELOPMENT OF ORGANIC-DEBRIS DAMS

The effect of precipitation and vegetative cover on erosion are well known (e.g., Ursic and Dendy 1965; Ralston and Hatchell 1971; Bormann et al. 1974; Patric 1976, 1977), but the role of organic-debris dams in regulating erosion and transport of particulate material over a long period (for example, as the terrestrial ecosystem recovers after major disturbance) is not known. In the absence of many data, we suggest that organic-debris dams may be just as important, if not more so at certain developmental periods, in controlling sediment transport from a mountainous landscape than are other environmental factors such as amount of precipitation and vegetative cover. We believe this because, in the absence of appreciable amounts of overland runoff, most of the eroded material originates from the stream channel. The roles and relative dominance of amount of precipitation, vegetative cover, and organic debris dams may be separated in time, however. For example, after some major disturbance which results in the removal of most organic matter (as would occur during catastrophic events such as glaciation or a very hot fire), amount and intensity of precipitation would be the most important factors until vegetation becomes well established. With the development of vegetation and the formation of root systems, a canopy cover, and a litter layer over the surface of the ground, erosion and transport of particulate matter would be reduced. The amount of liquid water available for runoff obviously is a critical factor affecting erosion and transport of particulate matter. The amount of water is reduced by evapotranspiration (mostly transpiration in northern hardwood forests), and the amount available for overland runoff is regulated by infiltration rate and storage capacity of the soil. Studies at the Hubbard Brook Experimental Forest in the White Mountains of New Hampshire (e.g., Bormann et al. 1974) and elsewhere, however, have shown that even in forested areas where biotic factors may function at a maximum, transport of particulate matter from drainage basins is exponentially and directly related to streamflow. Initially, then, after a major disturbance, biotic factors would tend to diminish erosion by absorbing the erosive power of raindrops and by reducing the amount of liquid water available for runoff, and this regulation could occur BEFORE trees became very large. For example, Marks (1974) has shown that the leaf area of early successional species in the White Mountains may be nearly equivalent to that of later, shade-tolerant dominants in only 4-6 years after clearcutting.

The size of woody debris produced by these early successional forests, however, is quite small. The average diameter of the major log found in organic debris dams in the White Mountains increases as stream width increases (fig. 1). As a result, debris dams would form in only the smallest channels during early developmental phases of the terrestrial ecosystems, because the trees would be too small to block larger channels. Although a young tree may be tall (long) enough to span a stream channel, its diameter (i.e., mass) appears to be the critical factor in determining whether it can maintain stability of the debris dam during extreme discharge events. As the trees become larger, dams would form in larger channels farther downstream and further reduce the transport of particulate matter (fig. 2).

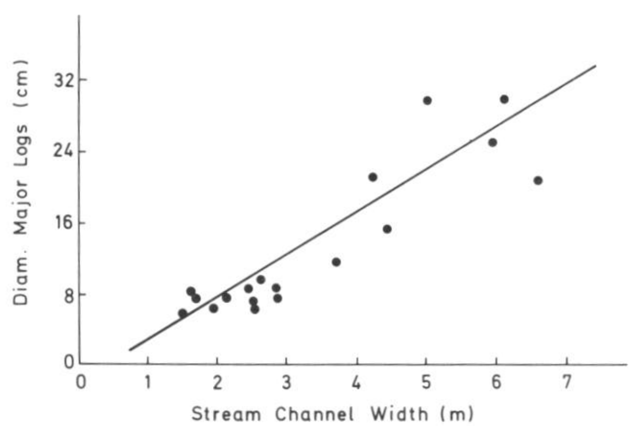

Figure 1.--Relation of stream-channel width to the average diameter of the largest member of an organic debris dam sampled in the White Mountains of New Hampshire ($Y = 4.50X - 1.73$; $r^2 = 0.79$).

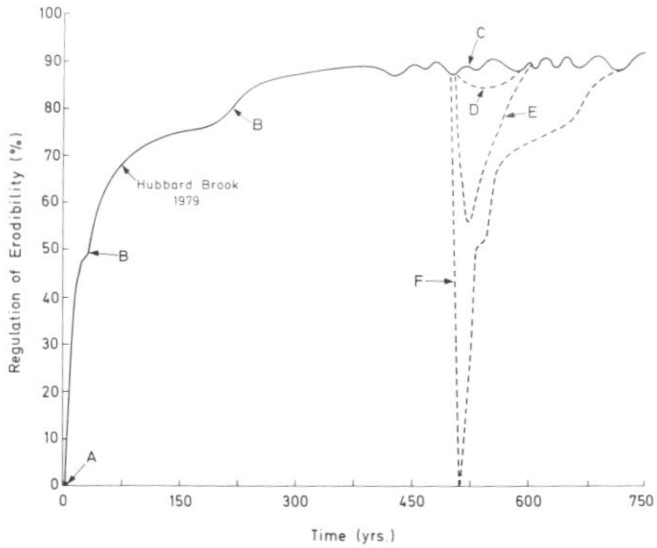

Figure 2.--Change in the regulation of erodibility with time:

A. Development of a young forest (small trees, low density).
B. Successional events when organic-debris dam formation is common.
C. Vegetative cover and organic-debris dam development complete, regulation of erodibility in steady state (characteristic of climate, vegetation type, and topography).
D. Short-term disturbance to vegetative cover only.
E. Sustained disturbance to vegetative cover and loss of organic-debris dams.
F. Sustained disturbance to vegetative cover with loss of surface humic layers (litter) and organic-debris dams.

At Hubbard Brook, a 15.6-ha drainage basin (Watershed 2 (W2)) was experimentally deforested in the autumn of 1965 and treated with herbicide for 3 years afterward to prevent any vegetative regrowth (Likens et al. 1970). Vegetation was allowed to regrow after 1969. Although no physical damage was done to the streambed during this operation, the greatly diminished input of terrestrial organic matter to the stream and increased discharge of water resulting from greatly reduced transpirational losses (cf. Hornbeck and Gee 1974) must have led to the destruction of many of the organic-debris dams on this stream. Increases in sediment yields from this drainage basin were relatively small during 1965-66 and 1966-67, but in 1967-68, they were more than 6-fold greater than on nearby uncut drainage basins and rose to 11-fold greater in 1968-69 (Bormann et al. 1974). Because essentially no overland flow occurred in the deforested drainage basin, much of this increased sediment yield must have originated from erosion of the stream channel and adjacent forest floor.

In the summer of 1976, the frequency of debris dams on W2 was still about 50 percent less than would be expected for a stream of this size in the White Mountains. The larger pieces of woody debris forming the framework of the dams, which were present in 1976, appeared old and weathered. These logs also were commonly larger than any of the trees now growing on the drainage basin. These dams thus originated before the clearcut. New dams now have formed on the first-order tributaries, but not in the second-order stream, because the woody debris produced by the forest is not yet of sufficient size to block the second-order streams. Erosion and transport of particulate matter, however, have now been reduced almost to uncut levels, primarily because of reestablishment of biotic regulation in the terrestrial ecosystem by the regrowing vegetation. In contrast, the experimental removal of all organic-debris dams from a 175-m stretch of a second-order stream in forested Watershed 5 (W5) of the Hubbard Brook Experimental Forest led to a 6-fold increase in the export of organic and inorganic particulate matter and a 6-percent increase in the export of total dissolved substances during a 12-month period (Bilby 1979). Thus, the increase in sediment yield observed from W5 after removal of dams was about the same as that observed on W2 during 1967-68. These results support our hypothesis that organic-debris dams are a major factor in regulating the loss of particulate matter from forested ecosystems.

These patterns suggest that the amount of organic matter stored in the streambed may change appreciably after disturbance and during the subsequent developmental sequence of the terrestrial ecosystem and that this organic matter in the stream channel plays a critical role in regulating erosion and transport of materials from the landscape.

A STEADY-STATE CONDITION?

We currently believe that, within the Hubbard Brook Experimental Forest, the amount of organic debris in the first- and second-order stream channels is near steady state on an annual basis (Fisher and Likens 1973, Bormann et al. 1974), but how did this come about? According to our hypothesis, streams have not always been in steady state relative to the accumulation of organic matter. That is, after glaciation, organic matter accumulated in stream channels as debris dams formed or possibly developed downstream when larger and larger structural components became available; after major disturbance (e.g., deforestation), initial losses of organic matter and debris dams are followed by accumulation as the terrestrial ecosystem recovers. Currently, no trees are large enough to span and maintain position during flood conditions in the main Hubbard Brook channel (a fifth-order stream).

Bormann and Likens (1979) have proposed a model for biomass accumulation in northern hardwood forested ecosystems after exogenous disturbance such as clearcutting (fig. 3). According to this model, living biomass accumulates for about 170 years, during which time the trees increase greatly in size. After about 170 years or so, these trees begin to die and fall over in increasing numbers. Eventually a condition called the shifting-mosaic steady state is reached, where patches of young, intermediate, and some old trees are interspersed on the landscape. These small, even-aged patches are caused by the endogenous disturbance as trees fall over.

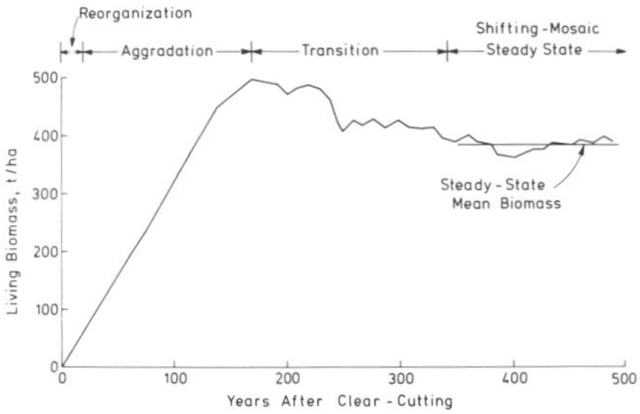

Figure 3.--Changes in living biomass during ecosystem development after clearcutting of northern hardwood forests. Phases of ecosystem development are: Reorganization (I), Aggradation (II), Transition (III), and Steady State (IV) (modified from Bormann and Likens 1979).

An important question relative to the role of organic-debris dams in streams is whether they respond to these changes by "moving" up and down the stream depending upon the size, shape, and species of tree available to produce the principal member in the organic-debris dams. Presently at Hubbard Brook, organic-debris dams are rare in streams with a mean channel width greater than 7 m. The last major cutting of the Hubbard Brook forest was in 1900-17 (Bormann et al. 1970), however, and most of the trees have not yet reached full size. The diameter and height expected of mature individuals of the dominant tree species at Hubbard Brook are listed in table 1. Assuming the relationship between diameter of the major member of an organic-debris dam and stream-channel width (fig. 1) holds true for larger logs, and that when one of these boles falls into a stream the diameter of the section crossing the channel is about equal to the diameter of the bole at half the height of the tree, we can estimate the width of channel that could be blocked by mature trees at Hubbard Brook (table 1).

Thus, before extensive logging took place in the Hubbard Brook Valley, a significant number of organic-debris dams may have occurred in stream channels up to almost 10 m in width (fifth-order). At the present time, stream channels of this size are completely devoid of debris dams. Therefore, even 60 years after a logging operation, which presumably resulted in the loss of many of the organic-debris dams from the stream systems, the drainage basin-ecosystem still has not regained its full potential to regulate sediment routing.

Obviously critical to our hypothesis are conditions that result in the falling over or blowing down of large trees. Bormann and Likens (1979) suggested that this is an important type of endogenous disturbance in the natural development of northern hardwood ecosystems. Is the falldown or blowdown of trees a random event on an areal basis? Obviously, if a large tree falls across the stream channel, the potential to form an organic-debris dam is much greater than if the tree falls in some other direction. We do not yet know whether more trees fall along a stream channel and toward a stream channel than elsewhere in a forest. If the stream channel is more than a few meters wide, this treeless zone could act somewhat like the edge of a forest where trees are more vulnerable to blowdown. Increased treefall along a stream channel might also result from increased erosion of the banks. If so, then this would be a type of positive feedback with the terrestrial system in the development and maintenance of organic-debris dams. In fact, the very presence of organic-debris dams may put more erosive pressure on the streambank (Zimmerman et al. 1967) and, thereby, enhance treefall at that spot.

Table 1--Diameter at breast height (d.b.h.) and maximum height commonly reached by the most important tree species at Hubbard Brook, and the steam channel width that could be blocked by a piece of debris equal to the diameter at half the tree height

Species	D.b.h. commonly[1/] reached by mature trees	Height commonly[1/] reached by mature trees	Diameter at half[2/] the height	Stream channel width that can be blocked
	Centimeters	Meters	Centimeters	Meters
Fagus grandifolia Ehrh.	90	24	48	9.5
Fraxinus americana L.	90	24	48	9.5
Tsuga canadensis (L.) Carr.	90	21	48	9.5
Pinus strobus L.	90	27	47	9.3
Betula alleghenienisis Britton	75	21	40	8.1
Acer saccharum Marsh.	60	24	32	6.7
Picea rubens Sarg.	60	21	32	6.7
P. glauca (Moench) Voss	50	20	27	5.8
Abies balsamea (L.) Mill.	45	18	24	5.2

[1/]Data from Harlow and Harrar (1950).

[2/]Calculated from d.b.h. and height, assuming a conical bole.

ROLE OF BOULDERS

Another factor rarely considered is the role of boulders in the formation of organic-debris dams. We have found that boulders are critical to the formation of organic-debris dams in stream channels in the Hubbard Brook Valley. We measured the volume of rocks protruding 10 cm or more above the streambed at randomly chosen cross sections of the stream channel and at cross sections on the downstream side of organic-debris dams. The average volume of rocks was significantly greater (students t-test at 0.05 level) behind dams than at randomly located cross sections. In streams containing large boulders, smaller logs may be able to form organic-debris dams than in streams without large boulders, everything else being equal. Very large boulders essentially divide a stream into smaller channels. We suggest that the role of boulders, so common in New England streams, should be examined before the development and role of organic-debris dams in streams can be fully evaluated.

HISTORICAL EVIDENCE

The sediment profile in Mirror Lake provides a long-term record of erosion and transport of particulate matter in the Hubbard Brook Valley (Likens and Davis 1975). This record suggests that about 4,800 years ago major mortality occurred in the extant hemlock (Tsuga canadensis (L.) Carr.) forests surrounding Mirror Lake. Normally, after such a major disturbance to the vegetation in a drainage area, erosion and transport of particulate matter from the landscape would be expected to increase. We suggest, however, that the hemlock decline may have been a factor in ultimately reducing transport of particulate matter by providing a source of large, dead tree trunks to form the structure for organic-debris dams. As large trees, they could have blocked much larger streams in the Hubbard Brook Valley than are presently dammed (table 1), and at the same time successional vegetation could have provided appreciable biotic regulation of erosion on the landscape. In fact, Likens and Davis (1975) found that the input of material to Mirror Lake did decrease after the hemlock decline, but they attributed the decrease to focusing of the sediments on the lake bottom. At present, we cannot resolve this problem. The blight of American chestnut (Castanea dentata (Marsh.) Borkh.) in New England and the current widespread death of American beech (Fagus grandifolia Ehrh.) also may represent periods when formation of organic-debris dams are enhanced.

The sediment record in more recent times (last 200 years) suggests a more uniform or increased transport of particulate matter to Mirror Lake (Likens and Davis 1975, Moeller and Likens 1978, Von Damm et al. 1979). This might be the result of the ultimate decay of old logs that were formed in the Hubbard Brook area before extensive cutting in 1900-17 (Bormann et al. 1970). Only now are the living trees becoming large enough to again form "stable" organic-debris dams on the larger stream channels in the Hubbard Brook Valley.

How long do organic-debris dams last in New England streams once formed? A cesium-137 profile to a depth of 55 cm in an accumulation of sediment behind a large organic-debris dam on Bear Brook in the Hubbard Brook Experimental Forest indicated no difference in age from top to bottom, implying that the sediment has been either homogenized frequently by biological or physical activity, or that it was formed within the last 25 years and all at once. How long might a log be expected to last under such conditions in an organic-debris dam? In streams in Oregon with coniferous vegetation, organic-debris dams may last for a hundred years or so (Swanson and Lienkaemper 1978).

TEMPORAL PATTERNS OF ORGANIC-DEBRIS DAMS

What role does the type and form of vegetation play in forming organic-debris dams? How does the density of wood affect the formation and maintenance of organic-debris dams as terrestrial ecosystems develop with time? How does the growth form (height and diameter of bole) of trees affect the development and maintenance of organic-debris dams? How does susceptibility to blowdown (e.g., life history and type of root system) of trees change with ecosystem development and affect the formation of organic-debris dams? How does variable resistance to decay affect organic-debris dams? Normally, such factors are not considered when questions about the history of erosion or land use are raised, but we suggest that these factors are of utmost importance in evaluating long-term landscape interactions. Indeed, the vegetation of the White Mountains has changed rather dramatically since glacial retreat some 10,000 to 15,000 years ago (Likens and Davis 1975). Is there a characteristic steady-state condition for amount of organic debris stored in the stream channels of the ecosystem for each dominant type of vegetation?

We suggest that the type of tree can be very important in the formation of organic-debris dams. In some western U.S. forests dominated by large, Douglas-fir (Pseudotsuga menziesii (Mirb.) Franco) trees, organic-debris dams are found on larger streams than in hardwood forests on the east coast merely because the Douglas-firs are larger in diameter and taller (Swanson et al. 1976, Bilby 1979). Thus, with ecosystem development whereby trees increase in size and species composition changes, we would expect organic-debris dam formation and longevity to change accordingly.

Hypothetical patterns for the formation of debris dams over time in three different sizes of stream channels after logging of a previously undisturbed forested ecosystem in the White Mountains of New Hampshire are shown in figure 4. In small streams (fig. 4A), the slash produced during the cutting is large enough to block the channel; hence, the frequency of relatively small dams would increase immediately after logging. The slash decomposes and dam frequency drops until pin cherry (Prunus pensylvanica L.f.) trees, which are an extremely common and short-lived (~30 years), early successional species in the White Mountains (fig. 4D)

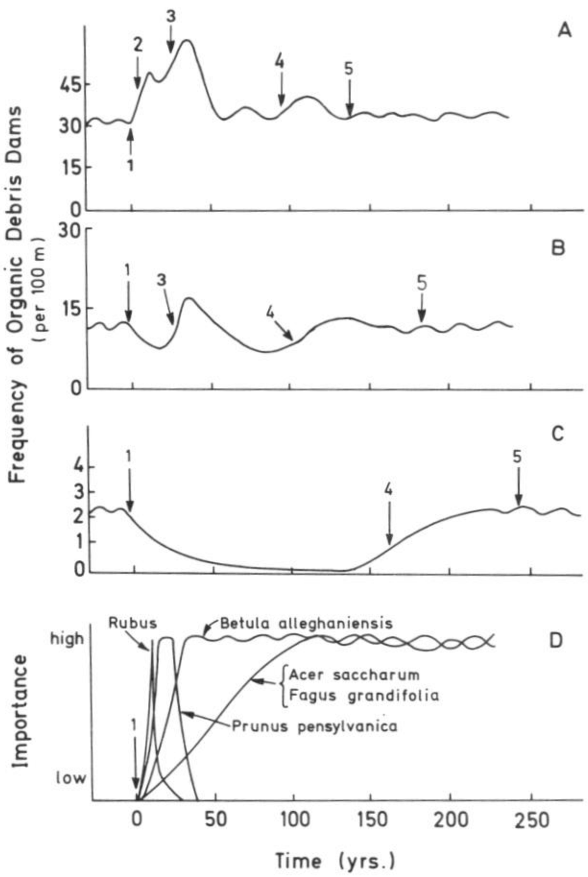

Figure 4.--Hypothetical changes in the frequency of organic-debris dams over time in first-order (A), second-order (B), and third-order (C) streams in the White Mountains of New Hampshire and in the relative importance of major plant species over a successional sequence (D) (from Marks 1974):

1. Clearcut logging of the drainage basin-ecosystem.
2. Slash enters channel and causes increase in organic-debris dam formation.
3. Falling pin cherry trees enter stream channel and form debris dams.
4. Mature, late-successional tree species begin to fall and enter stream channel, forming debris dams.
5. Return to precutting, steady-state condition.

begin to die and fall over. This input of organic matter produces a second peak in dam frequency. The dams formed from pin cherry and remaining slash then decompose as the forest grows. Eventually most of the early successional dams disappear and conditions previous to cutting are reestablished.

In slightly larger streams (fig. 4B), slash is not of sufficient size to block the channel. As a result, frequency of organic-debris dams decreases after clearcutting because of the reduction of organic matter. When pin cherry trees are dying, an increase in dam frequency occurs. The pin cherry dams decompose and dam frequency drops until the forest around the stream begins to produce woody debris large enough to block the stream.

In still larger channels (fig. 4C), neither slash nor pin cherry trees are large enough to span the stream channel. As a result, frequency of dams decreases for a long period after logging. Trees must reach large size before dam formation can be initiated in these larger streams. Over 100 years may be required to return to precutting conditions in these systems. Thus, even though the upper tributaries of Hubbard Brook are now in steady state relative to organic matter, the larger tributaries are likely to be accumulating (or will be in the next few decades if the forest is not cut again) organic matter in debris dams.

We have raised more questions than answers in this short discussion, but we think these questions are provocative and important in terms of evaluating the role of organic-debris dams as regulators of erosion and transport of materials from mountainous landscapes. Clearly organic-debris dams are a common and important feature of headwater streams in forested areas. The occurrence and changing role with time of organic-debris dams in larger tributaries are less well known, but would appear to be of potentially great importance in regulating erosion and sediment transport.

ACKNOWLEDGMENTS

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LITERATURE CITED

Bilby, R. E.
1979. The function and distribution of organic debris dams in forest stream ecosystems. Ph. D. thesis. Cornell University, Ithaca. 143 p.

Bormann, F. H., and G. E. Likens.
1979. Pattern and process in a forested ecosystem. Springer-Verlag, New York. 253 p.

Bormann, F. H., G. E. Likens, T. G. Siccama, R. S. Pierce, and J. S. Eaton.
1974. The export of nutrients and recovery of stable conditions following deforestation at Hubbard Brook. Ecol. Monogr. 44(3):255-277.

Bormann, F. H., T. G. Siccama, G. E. Likens, and R. H. Whittaker.
1970. The Hubbard Brook ecosystem study: composition and dynamics of the tree stratum. Ecol. Monogr. 40(4):373-388.

Fisher, S. G., and G. E. Likens.
1973. Energy flow in Bear Brook, New Hampshire: an integrative approach to stream ecosystem metabolism. Ecol. Monogr. 43(4):421-439.

Harlow, W. M., and E. S. Harrar.
1950. American forestry series: textbook on dendrology. McGraw-Hill Inc., New York. 555 p.

Heede, B.
1972. Flow and channel characteristics of two high mountain streams. USDA For. Serv. Res. Pap. RM-96, p. 12. Rocky Mt. For. and Range Exp. Stn., Fort Collins, Colo.

Hornbeck, J. W., and G. W. Gee.
1974. Forests and floods. N.H. For. Notes (117):18-21.

Keller, E. A., and F. J. Swanson.
1979. Effects of large organic material on channel form and fluvial processes. Earth Surf. Processes 4(4):361-380.

Keller, E. A., and T. Tally.
1979. Effects of large organic debris on channel form and fluvial processes in the coastal redwood environment. In Adjustments of the fluvial system. p. 169-197. D. D. Rhodes and G. P. Williams, eds. Kendall/Hunt Publ. Co., Dubuque, Iowa.

Likens, G. E., F. H. Bormann, N. M. Johnson, D. W. Fisher, and R. S. Pierce.
1970. Effects of forest cutting and herbicide treatment on nutrient budgets in the Hubbard Brook watershed-ecosystem. Ecol. Monogr. 40(1):23-47.

Likens, G. E., and M. B. Davis.
1975. Post-glacial history of Mirror Lake and its watershed in New Hampshire, U.S.A.: an initial report. Verh. Int. Ver. Limnol. 19(2):982-993.

Marks, P. L.
1974. The role of pin cherry (Prunus pensylvanica L.) in the maintenance of stability in northern hardwood ecosystems. Ecol. Monogr. 44:73-88.

Moeller, R. E., and G. E. Likens.
1978. Seston sedimentation in Mirror Lake, New Hampshire, and its relationship to long-term sediment accumulation. Verh. Int. Ver. Limnol. 20:525-530.

Patric, J. H.
1976. Soil erosion in eastern forests. J. For. 78:671-677.

Patric, J. H.
1977. Soil erosion and its control in eastern woodlands. North. Logger and Timber Processes 25(11), 6 p.

Ralston, C. W., and G. E. Hatchell.
1971. Effects of prescribed burning on physical properties of soil. In Prescribed burning symposium proceedings. p. 68-84. USDA For. Serv., Southeast. For. Exp. Stn., Asheville, N.C.

Strahler, A. N.
1957. Quantitative analysis of watershed geomorphology. Trans. Am. Geophys. Union 38:913-920.

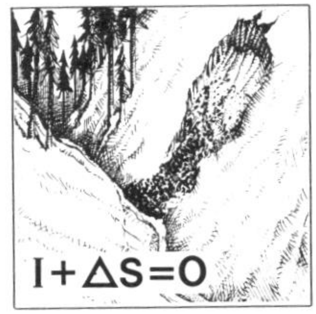
Swanson, F. J., and G. W. Lienkaemper.
1978. Physical consequences of large organic debris in Pacific Northwest streams. USDA For. Serv. Gen. Tech. Rep. PNW-69, 12 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Swanson, F. J., G. W. Lienkaemper, and J. R. Sedell.
1976. History, physical effects and management implications of large organic debris in western Oregon streams. USDA For. Serv. Gen. Tech. Rep. PNW-56, 15 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Ursic, S. J., and F. E. Dendy.
1965. Sediment yields from small watersheds under various land uses and forest covers. In Federal interagency sediment conferences proceedings. U.S. Dep. Agric. Misc. Publ. 970:47-52.

Von Damm, K. L., L. K. Benninger, and K. K. Turekian.
1979. The ^{210}Pb chronology of a core from Mirror Lake, New Hampshire. Limnol. Oceanogr. 24(3):434-439.

Zimmerman, R. C., J. C. Goodlett, and G. H. Comer.
1967. The influence of vegetation on channel form of small streams. Symposium on river morphology. Int. Assoc. Sci. Hydrol., Publ. 75:255-275.



Sediment Routing and Budgets: Implications for Judging Impacts of Forestry Practices

Frederick J. Swanson and Richard L. Fredriksen

ABSTRACT

Sediment budget and routing studies offer some improvements over traditional studies of small drainage-basin manipulations and individual erosion processes for analysis of impacts of forestry practices on soil erosion from hillslopes and sedimentation in streams. Quantification of long-term (century) and short-term (decadal) impacts awaits more detailed analysis of the dynamics of sediment storage in stream channels and at hillslope sites prone to failure by debris avalanches.

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INTRODUCTION

Sediment routing can be considered the conceptual or quantitative description of the movement of soil and sediment down hillslopes and through the fluvial system from one temporary storage site to another. A sediment budget quantifies the input, change in storage, modification, and output of sediment for a landscape unit. Analysis of sediment routing and budgets has been used in a variety of ways ranging from basic geomorphology research (Rapp 1960, Leopold et al. 1966, Dietrich and Dunne 1978) to analysis of land-management impacts on sedimentation (Janda 1978, Pearce and O'Loughlin 1978). Application of sediment routing and budget studies in basic research has been rare, and their use in applied geomorphology has been even more limited.

With further development, these approaches to understanding geomorphic systems will greatly aid in analyzing and mitigating effects of forest practices on soil erosion and sedimentation in streams. A sediment budget provides measures of the relative importance of both natural sediment sources and sources induced by human activities. The persistence of sediment sources is dependent on the volume of sediment stored at a site and the rate of sediment resupply, which can be described by sediment budgets. Efficient, economic solution of erosion problems begins with identifying the major sediment sources so corrective actions can be applied at the most beneficial points in the system.

Current land-management issues on a broad scale concern identification of cumulative sedimentation impacts of progressive development of forest drainage basins and use of timber-harvest scheduling to minimize these impacts. Some understanding of sediment movement through a whole drainage basin is an essential starting point in evaluating cumulative, long-term impacts of forest practices. This whole-basin perspective should also be an important part of planning future research on effects of forest management on sedimentation.

Traditional assessments of erosional impacts of forest practices have taken a more narrow approach, emphasizing studies of individual erosion processes and small drainage basins. A process, such as surface erosion or shallow, rapid, soil mass movement,[1] may be considered in isolation. The rate of a particular process may be measured in forested and disturbed areas and compared. Small drainage basins are treated as "black boxes" and their water and sediment yields are compared before and after treatment and with a control basin. Linking studies of processes and small drainage basins for better interpretation of sediment sources is a first step toward understanding sediment routing in a landscape.

In this paper, we discuss examples of results and limitations of studies of certain individual processes and of small drainage basins for quantifying impacts of forest practices on sediment routing. Reexamination of these studies leads to suggestions for improved design of future investigations of management effects on sediment routing. These suggestions are generally summarized in the basic rules for developing a sediment budget.

Dietrich et al. (this volume) outline requirements for quantifying sediment routing: identify and quantify storage sites in the landscape; identify and quantify processes that transport material between storage sites; and determine linkages among transfer processes and storage sites. These are the necessary and sufficient steps for quantifying sediment budget, assuming the system is in steady state. In studies of long-term sediment budgets for natural forest and landscape conditions, this assumption may be reasonable. In assessing effects of management activities on sediment routing, however, it is commonly necessary to account for large, relatively short-term changes in sediment storage, which preclude the steady-state assumption (Pearce and O'Loughlin 1978). Management-induced changes in sediment storage may occur in more than one type of storage area, and the changes may not all have the same sign.

Here we argue that analysis of changes in sediment storage provides useful understanding of some short-term and many long-term impacts of management practices on sediment routing. To make this argument, we first offer an overview of the sediment-routing system for small, steep, western Cascade drainage basins and then discuss analysis of management impacts on crucial, but poorly understood, parts of this system.

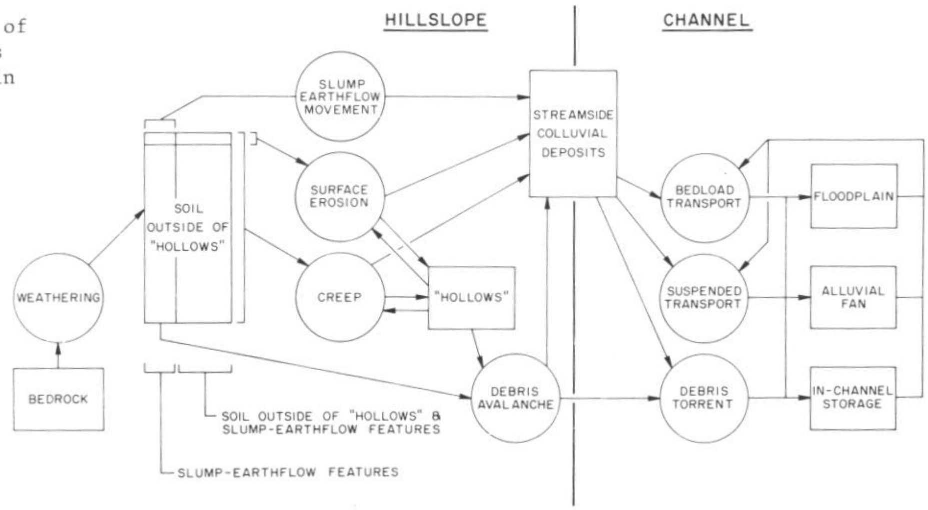
SEDIMENT ROUTING REGIME IN STEEP, WESTERN OREGON FOREST LAND

The sediment-routing system of a drainage basin may be viewed as a variety of transport processes moving soil and sediment through a series of temporary storage sites. An example of linkages among storage sites by transport processes are shown in simplified conceptual form in figure 1 for steep, forested landscapes in western Oregon. This routing scheme is based on work at intensive study sites in the western Cascades of Oregon. The area is underlain by lava flow and clastic volcanic bedrock and forested with Douglas-fir (*Pseudotsuga menziesii* (Mirb.) Franco), western hemlock (*Tsuga heterophylla* (Raf.) Sarg.), and other coniferous and a few deciduous species. Most of the more than 230 cm of average annual precipitation falls as rain during long, low-intensity frontal storms between November and April.

In this area, creep, surface erosion, root throw, debris avalanches, slump, and earthflow are all potentially significant processes of particulate matter transport down slopes and into channels. Once in the channel, this material either enters temporary storage sites or moves as suspended sediment and bedload and in debris torrents.

[1]Here we use the term "debris avalanche" to refer to all such mass movements, recognizing that *sensu strictu* debris flows, avalanches, and slides (Varnes 1978) are involved.

Figure 1.--Simplified flow chart of relationships among storage sites (boxes) and transport processes in steep, volcanic terrane in the western Cascades, Oregon.



Hillslope and channel processes have a variety of serial interactions in which one process may (1) directly trigger another, (2) supply sediment for transfer by another process, and (3) increase the potential for occurrence of another process. These interactions complicate sediment budgets by making it difficult to attribute sediment delivery to a point in a drainage basin to one transport process. Creep, for example, carries soil to locations adjacent to channels, but delivery to the channel occurs by surface erosion, bank erosion by debris torrents, or small mass failures of streambanks. Debris avalanches deliver sediment to channels from steep microdrainages or "hollows" (Dietrich and Dunn 1978). Debris avalanches also initiate at the oversteepened headwall and toe areas of recently active slumps and earthflows and on some planar slopes, particularly where root throw triggers events. The hollows are slowly refilled by surface erosion, root throw, and creep before being catastrophically evacuated again by debris avalanching. Sometimes, streambank cutting contributes to stream side failures, especially from toes of earthflows. Other interactions among transport processes in this landscape are discussed in Swanson et al. (1982).

Temporary storage of material occurs in a great variety of sites in drainage basins (fig. 1). The soil mantle can be considered an area of storage and divided into subunits on the basis of types of transport processes involved. Surface erosion by dry ravel, rain splash, and freeze-thaw processes, for example, affect the upper centimeter or so of the soil surface. Surface movement is faster than soil creep, which affects the entire soil column. Creep, surface processes, and rotational translational failure are superimposed in slump-earthflow terrain (fig. 1).

Storage sites for alluvial material vary in relative importance along a river system. Large organic debris commonly forms dominant storage sites in first-, second-, and third-order channels in old-growth forests. Deposits in channels not related to organic debris and flood plain deposits are the principal storage sites for alluvium in larger streams. Alluvial fans are potentially important long-term storage sites located at junctions of low-order (generally first- or second-order) channels and higher order rivers. Fans accumulate where flood plains are broad enough to provide sites for storage (Swanson and James 1975).

The sediment-routing system described above and in figure 1 is simplified and ignores important aspects of system behavior. Much of the soil movement by hillslope processes, for example, involves redistribution on slopes rather than delivery to a channel. Transfer of sediment between slope and channel areas is also far more complex than described here. Furthermore, important feedback mechanisms, such as acceleration of slope-transport processes by bank cutting and streamside mass failures, are not treated explicitly.

DIFFICULTIES IN INTERPRETING MANAGEMENT IMPACTS ON SEDIMENT ROUTING

Studies of individual erosion processes and manipulations of small drainage basins in areas with this general type of sediment-routing system have revealed many-fold increases in soil and sediment movement after logging and road construction (Fredriksen 1970, Fredriksen and Harr 1979). Several problems arise in isolating effects of different management practices and distinguishing between short-term (decadal) and possible long-term (several cutting rotations) management effects on erosion. Crucial problems are understanding and quantifying the dynamics of two important storage sites in the system: (1) sites on hillslopes from which debris avalanches orginate and (2) channel storage sites, particularly those related to large organic debris.

Debris-Avalanche Sites

Impacts of forest practices on soil erosion by debris avalanches are commonly measured with inventories of soil movement by debris avalanches in forest, clearcut, and road right-of-way areas (Dyrness 1967, Swanson and Dyrness 1975, and others). Dietrich and Dunne (1978) and Dietrich et al. (this volume) have critically reviewed some aspects of this procedure. Analyses of debris-avalanche inventories in steep, unstable land generally have documented increased soil erosion by debris avalanches in the first few decades after clearcutting and road construction (Swanston and Swanson 1976). The increase in failure frequency in clearcut areas has been attributed mainly to reduced root strength when root systems of killed vegetation have decayed significantly, but before roots of incoming vegetation are well established (Swanston 1970, O'Loughlin 1974, and others). Road failures generally result from altered distribution of soil, rock, and water on a slope.

The effects of cutting on debris-avalanche erosion over an entire rotation (80 to 100 years in much Federal land in the Pacific Northwest) and over several rotations are unknown. H. A. Froehlich (School of Forestry, Oregon State University, Corvallis, personal communication) and others have argued informally that the 10- to 15-year period of increased debris-avalanche erosion is followed by an extended period of debris-avalanche occurrence significantly below the rate observed in the areas of older, established vegetation usually sampled to determine a reference "natural" rate. If this is true, clearcutting may alter the timing of debris-avalanche erosion, but may not necessarily increase the overall rate on the time scale of one or more timber rotations. This hypothesis cannot be tested with existing inventories of debris-avalanche occurrence because of complexities of land use and storm histories and shortness of record.

Interpreting the effects of management on debris-avalanche erosion on the time scale of a century or more depends on understanding the recharge and storage dynamics of sites that fail by debris avalanching. Disregarding roads, debris avalanches in many areas of western Oregon originate predominantly from (1) hollow sites defined and described by Dietrich and Dunne (1978) and Dietrich et al. (this volume), and (2) sites locally oversteepened by slump-earthflow movement. Hollows are recharged by surface erosion, root throw, creep, and weathering of bedrock. Debris avalanches associated with slump-earthflow features occur on headwall scarps, at breaks in slope in midslope positions, and at toes of earthflows. Continued slump-earthflow movement creates opportunities for repeated failure at these sites.

The relative importance of debris-avalanche initiation at hollow and slump-earthflow sites varies greatly from one landscape to another. Debris avalanches from hollows predominate in many steep, highly dissected areas, but earthflow activity determines the incidence of debris avalanches in terrain of lower relief sculpted by slow, deep-seated, mass movements. Both types of sites are important in the volcanic terrane of the western Cascade Range. About 30 percent of soil moved by debris avalanches in the 62 km^2 of forested and clearcut areas in the H. J. Andrews Experimental Forest (1950-1979) originated from slump-earthflow features.

Effects of clearcutting on the rate of debris-avalanche erosion in a landscape containing numerous sites that repeatedly fail by debris avalanching is related to the rate of recharge of those sites and effects of management practices on

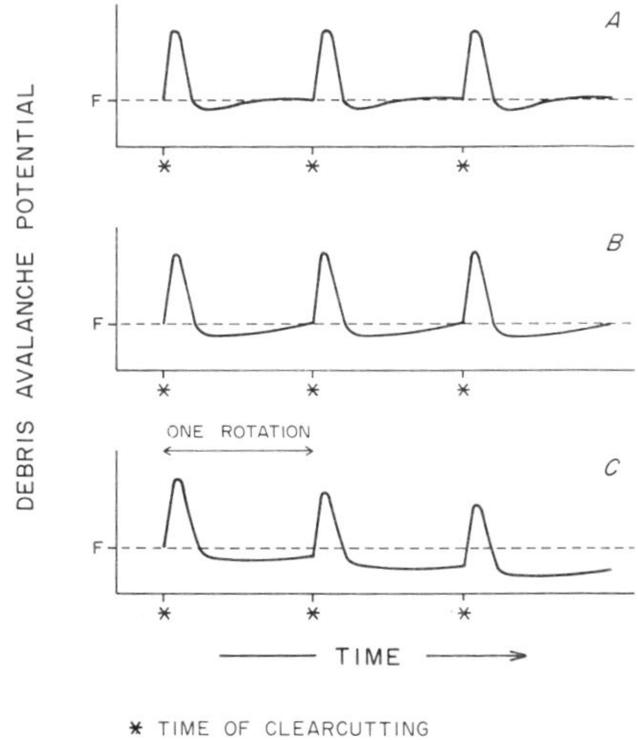


Figure 2.--Hypothetical variation in debris-avalanche potential in a landscape with many sites for debris-avalanche failure, only a few of which fail in the first 10 to 15 years after clearcutting.

A. Refilling of failed sites is fast relative to cutting frequency.
B. Refilling occurs in about one rotation.
C. Refilling takes much longer than cutting frequency.

Debris-avalanche potential rather than erosion rate is shown because actual erosion occurs in brief, infrequent periods.

processes that recharge the sites. If recharge time is much shorter than the period between cuttings, the rate of debris-avalanche erosion between the period of accelerated erosion and the next clearcutting is similar to the background forest rate (fig. 2A). Under these conditions, successive cuts will have an impact on debris-avalanche erosion similar to the first cut because sites of recent failures will be recharged at the time of subsequent cuts. Where recharge typically occurs in the period of one rotation, subsequent cuts may have the same impact as earlier cuts, but the rate of debris-avalanche erosion after the period of accelerated erosion may drop significantly below the background forest rate during each rotation (fig. 2B). If recharge occurs over periods much longer than the cutting rotation, several successive cuts may progressively have reduced impact on debris-avalanche erosion because some sites that failed after earlier cuts are not ready to fail again when subsequent cuts occur (fig. 2C). This effect may also result in debris-avalanche erosion below the background forest rate between the period of accelerated erosion and the next cut.

Filling rates of debris-avalanche scars are poorly known. Dietrich et al. (this volume) estimate that refilling of hollows occurs on the time scale of 1,000 years, based on estimates of creep rate for forested areas. The rate may be appreciably faster if root throw, animal activity, and various surface-erosion processes are also taken into account. Furthermore, the rate of each of these processes--except root throw--may be accelerated by removal of vegetation. Wildfire, logging, and slash burning trigger periods of accelerated soil movement (summarized in Swanson 1981) which presumably causes accelerated hollow filling. How important are such periods of accelerated erosion in filling a hollow? Swanson (1981) attempted such an analysis for sediment yield in the central western Cascades of Oregon and estimated that about 25 percent of long-term sediment yield occurred in periods of accelerated erosion after wildfire. Although this estimate contains great uncertainties, it suggests that hollow filling during periods that include severe disturbances of vegetation may be significantly faster than the rate estimated for forested conditions only.

Current knowledge of the recurrence of debris avalanches from sites related to slump-earthflow features is beset with similar uncertainties. Many slump-earthflow features in this area move at rates of centimeters to meters per year (Swanston and Swanson 1976), so sites of debris-avalanche failures in slump-earthflow deposits may be recharged as quickly as in a few years to decades. Other slump-earthflows move more slowly or infrequently, so recharge of associated debris-avalanche sites is slower. Effects of clearcutting on slow, deep-seated, mass movement features have not been documented quantitatively. Gray (1970) and others hypothesize that the major effect is that reduced evapotranspiration results in increased availability of soil moisture, which may prolong seasonal periods of movement.

In summary, short-term (decadal) increases in debris-avalanche erosion after clearcutting have been documented, but effects over a whole rotation or multiple rotations (centuries) are unknown. These longer term effects are determined by rates of processes that prepare sites to fail again. All of these processes are ultimately limited by the rate of rock weathering and soil formation. Before we can assess long-term management impacts on debris-avalanche erosion, we need more information on (1) rates and mechanisms of refilling of hollows, (2) rates and mechanisms by which slump-earthflows prepare associated debris-avalanche sites for repeated failure, and (3) effects of management practices on these mechanisms and rates. Field measurements of recharge processes should be made in appropriate geomorphic contexts. It is essential to analyze debris avalanches in their overall sediment-routing context, including the storage dynamics of sites of debris-avalanche initiation.

Small Drainage Basin Studies--Channel Storage

Manipulation of small drainage basins has been used to measure erosional consequences of forest practices. USDA Forest Service researchers have conducted this type of research on 10- to 100-ha drainages in the H. J. Andrews Experimental Forest in the western Cascade Range, Oregon. In a series of paired-basin experiments, sediment yields from control and manipulated basins are monitored and compared for periods before and after logging and road construction. Originally these studies were designed to measure impacts of forest practices on sediment yield and nutrient loss in different terrains (Fredriksen 1970, 1972; Swanson et al., 1982). As these studies have progressed, we increasingly recognized the need to understand sediment routing through each basin.

Channel-storage dynamics are a particularly important, but commonly neglected, element in the response of basins to forest cutting (Pearce and O'Loughlin 1978). The potential significance of sediment stored in channels is revealed by estimates that average annual export of coarse particulate material from small basins is less than 5 or 10 percent of sediment stored in the few channel systems analyzed (Megahan and Nowlin 1976; Megahan, this volume; Swanson and Lienkaemper 1978). Consequently, moderate changes in volume of stored sediment can account for large year-to-year changes in sediment yield, even if sediment supply from hillslopes is constant. Accelerated erosion from hillslopes may not show up as increased sediment yield if sediment is stored in channels (Pearce and O'Loughlin 1978).

Management practices can alter channel storage by (1) altering rates of sediment input and output by changing peak flows, availability of erodible sediment, and rates of hillslope erosion, (2) altering storage capacity by changing quantity and distribution of large organic debris, and (3) increasing potential for debris torrents, which can flush stored sediment and large organic debris from steep channels. Studies of experimental basins in the Andrews Forest provide examples of a broad range of changes in channel storage in response to management activities.

Unfortunately, we have insufficient data at this time to compute complete sediment budgets. Only fragmentary data exist for change in channel storage and sediment input to channels by processes other than debris avalanches. The variety of channel-storage changes, however, emphasizes the importance of quantifying channel storage in future studies of management impacts. The history of debris torrents over the past 40 years has strongly influenced patterns of sediment yield from Watersheds 1, 2, and 3, while analysis of Watershed 10 over a shorter period when torrents occurred reveals other effects of channel storage.

Studies on Watersheds 1, 2, and 3

Measurement of suspended sediment and sediment trapped in ponding basins (here termed bedload[2]) began in 1957 at the 96-ha Watershed 1 (WS1), 60-ha WS2, and 101-ha WS3 (Fredriksen 1970). WS1

[2]Some of the material caught in sediment basins includes suspended sediment. About 25 percent of material collected in the sediment basin at WS10 after logging has been less than 2 mm in diameter.

was completely clearcut without roads between 1962 and 1966, and the slash was burned in a hot fire in the fall of 1966. WS2 has been maintained as a control; it is forested with 400- to 500-year-old Douglas-fir and western hemlock, and a mix of younger trees established after a light wildfire about 135 years ago. Roads covering 6 percent of WS3 were constructed in 1959, and three areas totalling 25 percent of the drainage basin were clearcut and broadcast-burned in 1963.

WS1 and WS3 have responded very differently to their respective treatments (fig. 3) because of contrasts in types of treatments, timing of treatments with respect to major storms, and roles of the two channel systems as sediment sources and sinks. WS3 was freshly logged and burned when the two extreme storms of Water Year 1965 (WY1965) occurred. The storm of late December 1964 triggered a series of debris torrents, most of them initiated from roadfill failures, that sluiced out much of the drainage network of WS3 (Fredriksen 1970). These torrents carried about 20 000 t of organic and inorganic material out of the drainage basin. Over 90 percent originated from the roadfills, and most of the remainder was material stored in the channel before logging. During the following 13 years (WY1966 through WY1978), about 900 t of bedload material were exported from WS3 (annual bedload yield from WS2 has been 9.9 t/km^2 for WY1957-WY1976). Thus a few momentary events of WY1965 transported more than 22 times as much sediment as the entire next 13 years. The torrents greatly reduced both the volume of material in storage and the storage capacity of WS3 channel system by removing large organic debris.

The history of WS1 has been very different. The basin was only partially cut--and burning had not yet occurred--when the WY1965 storms struck, so the absence of roads and earlier stage of cutting made WS1 less sensitive to these storms than WS3. No debris torrents occurred in WS1, and most of the 800 m^3 of soil moved to channels by debris avalanches in WY1965 collected temporarily behind the abundant, large organic debris in the channel. Broadcast burning and some clearing of debris from the channel in 1966 initiated a period of accelerated export of bedload that totalled about 2900 t in WY1966 through WY1978. Thus bedload yield for this period from WS1 is over 3 times the yield from WS3. From measurements of channel cross sections, we estimate that about 4300 t of the material that entered the WS1 channel after logging remains in temporary storage in the channel system. The channel is now undergoing net decrease in storage. The large volume of sediment stored in the WS1 channel and unstable channel conditions suggest that bedload yield derived from these readily available sources can remain high for another decade or so.

Presence or absence of debris torrents has been an important factor in the contrasting sediment export between WS1 and WS3. Roadfills that were poorly constructed and poorly located by today's standards failed in the heads of long, straight, steep channels of WS3. These are ideal conditions for initiating debris torrents that move long distances down channels (Swanson and Lienkaemper 1978).

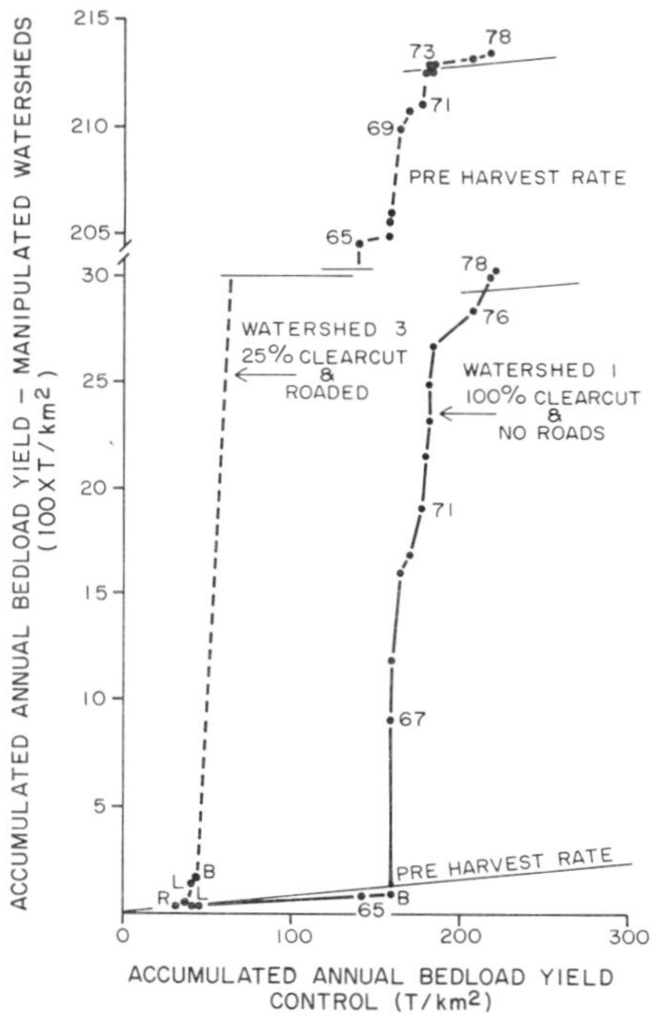


Figure 3.--Double mass plot of sediment collected in ponding basins at Watersheds 1 and 3, H. J. Andrews Experimental Forest. Preharvest rates are based on relationships between manipulated drainage basins and the control established in the predisturbance period. L = year of logging, R = road construction, B = broadcast burning.

Eight debris avalanches, each of which transported more than 75 m^3 of soil, have occurred in WS1 since clearcutting, but none triggered a debris torrent because they did not enter the main channel with sufficient mass and velocity and sufficiently straight trajectory to maintain momentum down the main channel.

Much of the contrast in sediment yield between WS1 and WS3 over the period of several decades after logging and road construction results from differences of channel-storage factors. WS3 was flushed and now has relatively low volume of stored material and low capacity for additional storage because of low quantities of large organic debris. Bedload export from WS3 is now limited by sediment supply from hillslopes rather than from release from channel storage. On the other hand, the timing of sediment release from channel storage is a dominant factor controlling persistent, high bedload yield from WS1, although continued sediment supply from hillslope sources is also important.

These observations point up the need in future drainage-basin studies to quantify changes in channel storage and, if possible, to distinguish material that entered the channel before and after disturbance. The mass budget equation for the channel should be: output = input + change in volume of material that entered the channel before disturbance + change in volume of material entering the channel after disturbance. Surveyed and monumented cross sections combined with stratigraphic analysis of deposits encountered on cross-section lines can be used to measure these aspects of channel-storage dynamics.

Studies on Watershed 10

Studies at WS10 in the H. J. Andrews Experimental Forest reveal the need to account for changes in channel storage when evaluating management impacts on sediment yield in basins where torrents have not dominated the recent history of sediment export. This steep, 10-ha drainage basin was studied intensively under forested conditions from 1970 to 1975 and since clearcutting and skyline yarding in summer 1975 (Fredriksen 1972; Swanson et al., 1982). Large slash was yarded to the ridge-top landing; the basin was not broadcast burned. About half of the 50 logs that had been in the main channel of WS10 before logging were removed, and slash larger than about 5 cm diameter and 50 cm length was hand-cleaned from the channel.

Measurement of effects of logging on sediment yield is based on samples of successive storms at manipulated WS10 and 9-ha control WS9. Unfortunately, sediment-basin collections before logging were of short duration and marginal quality because of intense research activity in lower WS10, so bedload yields are compared for the postcutting period only.

Four storms during WY1976 transported 18.9 t of particulate material into the sediment pond (here termed bedload) at the outlet of WS10.[2] The first two storms produced peak flows that typically occur several times a year, yet combined they exported about 6.8 t of bedload--about 7 times the average annual bedload yield for small, old-growth forest basins (Swanson et al., 1982). The third and fourth storm events produced successively higher peak flows and exported 8.4 and 3.7 t of bedload, respectively.

WY1977 was the driest in the 86-year history of precipitation records in central western Oregon; no significant bedload transport occurred in WS10. Several major events during WY1978 exported a total of 8.8 t of bedload, although this period included two peak flows that exceeded those of WY1976. Over this 3-year period after cutting, WS10 exported 27.6 t of bedload, while WS9 yielded only 0.8 t.

These results follow two general patterns: an increase in total yield after clearcutting and a decline in total yield for a given peak-flow magnitude through the sequence of storms. Greater total export after disturbance could be attributed to increased transport capability of the system (such as increased peak flow), to increased availability of material to be transported, or both. After clearcutting of WS10, the magnitude of peak flows from snowmelt actually decreased relative to control WS9, and no detectable change occurred in peak flows for events with rainfall only (Harr and McCorison 1979). Therefore, changes in sediment export from WS10 primarily reflect changes in sediment availability and storage rather than altered basin hydrology.

Based on measurements of hillslope erosion and qualitative observations of the amount and type of material stored in the channel, export from WS10 appears to come from three sources: (1) soil and organic matter--mainly green twigs and needles--moved into the channel during felling and yarding operations, but not removed during channel cleaning, (2) material that entered the channel by natural processes before logging and had been in temporary storage behind logs in the channel, but was released from storage when logs were removed, and (3) material transported to the channel by hillslope erosion after logging. Each of these sources makes sediment available at different times. Source 1 was most significant in the first few major storms after cutting. By the fourth storm of WY1976 much of this readily transported material rich in organic matter had been flushed downstream to the basin or deposited in more stable debris accumulations within the channel. Source 2 gained importance in the first few years after logging and after material in Source 1 had been moved. Postlogging hillslope erosion (Source 3) will probably not become dominant in WS10 until several years after cutting. The timing of sediment availability from Source 3 in WS10 is a result of (1) absence of roads feeding sediment-laden water directly into the drainage system, which could supply sediment even before cutting occurs, and (2) the effect of hand-piled slash along the stream channel in retarding movement of soil to the channel. These sediment traps become less effective as they collapse from decay and snow loading. Sediment supply by debris avalanches and possibly creep is believed to increase several years after cutting in response to decay and loss of strength of roots (Swanston 1970).

This scenario could, of course, be altered in other drainage basins if, for example, accelerated surface erosion from broadcast burning or occurrence of debris avalanches soon after cutting quickly flood the channel system with material from Source 3. In WS10, though, we have measured only 1.2 t of material transported into the channel system between October 1975 and February 1976, although 19.8 t were exported. The inputs resulted from surface erosion by dry ravel, rain splash, and needle ice. Transport rates to the channel were sampled in 34 0.5-m-wide boxes located along the stream perimeter. No debris avalanches have transported soil to the channel since cutting.

The results from WS10 indicate that an understanding of channel-storage dynamics is essential to interpreting short-term (few years) data on sediment yield from disturbed drainage basins. Furthermore, changes in storage of material that entered the channel before and after logging must be distinguished. This distinction would provide better resolution of the quantity and fate of soil eroded after logging. Too often, changes in sediment yield are interpreted only in terms of altered hillslope erosion.

Channel Storage--Long-Term Considerations

Forest-management practices can have long-term effects on quantities of large organic debris in channels and associated channel-storage capacity and aquatic habitat. Although poorly quantified, the strong positive correlation between amounts of large organic debris and stored sediment in small, steep, V-notch channels is obvious in field reconnaissance. Presence of large debris in steep channels also benefits aquatic ecosystems by providing cover, a source of nutrients, diversity of aquatic habitats, and depositional sites where organic matter can accumulate and be available for consumption by aquatic organisms. The sediment storage of large debris may also benefit aquatic organisms by buffering areas downstream from sites of pulses of sediment by processes such as debris avalanches. Downstream movement and subsequent accumulation in higher order channels may cause damage to structures, blocks to fish passage, and other problems.

When a channel such as in WS3 is flushed by a debris torrent that removes large debris, the period of recovery of debris loading and associated capacity for sediment storage may span several decades to a century or more if a source of large woody debris is available. Clearcutting without leaving trees along the channel removes the future source of large debris. Unless we specifically manage streamside stands to produce large debris for streams, little significant woody material will enter streams in managed stands. Intensive silviculture and harvesting practices produce no large woody residues.

Concentrations of large debris have persisted in streams affected by natural wildfire disturbances in western Oregon forests (Swanson and Lienkaemper 1978). Large pieces carried over from the previous stand had residence time greater than the time it took the postfire stand to grow trees large enough to produce large debris. Consequently, debris loading and associated sediment storage was likely to be maintained through the period of recovery after natural forest disturbances.

Unless the ecosystem is consciously managed otherwise, the net effect of intensive forest management is likely to be a gradual, widespread decrease in large organic debris in streams. The sediment-storage capacity of high-gradient, low-order portions of channel systems would decline greatly, and travel time of coarse particulate matter through such stream reaches presumably would be reduced. Reduced diversity and area of prime aquatic habitat is also a likely result.

Further quantification of the role of large organic debris in sediment storage throughout a river network would help strengthen arguments for or against this hypothesis. Analysis of rates of input of large debris to channel sections with different histories of flushing and disturbance of adjacent stands are also essential to predicting long-term impacts of management activities on roles of channel storage in sediment-routing systems.

CONCLUSIONS

We have traditionally measured effects of forest practices on soil erosion and sedimentation with studies of individual processes and small drainage basins. Viewing the problem of impact assessment from the perspective of overall sediment routing suggests specific ways to strengthen our understanding of impacts on soil and aquatic resources. Sediment-routing concepts encourage analysis of storage sites as well as transfer processes and analysis of each in the context of the whole system. Use of this approach to reexamine studies of management effects on debris-avalanche erosion and sediment yield from small basins reveals numerous unanswered questions, particularly in terms of impacts over periods greater than a few decades.

Debris-avalanche inventories document short-term (decadal-scale) increases in debris-avalanche erosion after clearcutting. Determining longer term (century-scale) impacts is contingent on understanding the types and rates of processes that refill storage sites subject to failure by debris avalanche. Field installations to measure these recharge processes should be placed in an appropriate geomorphic setting. For example, the role of creep and other processes in filling hollows should be based on measurements in microdrainages contributing soil to hollows, as well as on smooth or hummocky slopes where convergent soil and subsurface water movement does not occur or is more unpredictable than in hollows. Measuring effects of management practices on recharge processes is essential because rate of recharge is a long-term control on both the kind and degree of management impact on debris-avalanche erosion. Soil formation is the ultimate controlling factor.

Changes in channel storage regulate sediment yield from small drainage basins affected by management practices. Altering the location, size, or replenishment of large organic debris alters the sediment storage and yield characteristics of channel systems. Where basins have been treated as "black boxes," changes in sediment yield have been mainly attributed to variation in hillslope-erosion processes. Future studies should assess dynamics of channel-storage systems by using repeated surveys of monumented cross sections and, where possible, distinguishing between stored sediment that entered the channel before and after disturbance of the adjacent stand. Assessment of long-term effects of management practices on sediment storage in low-order forested channels is keyed to understanding (1) relations between large organic debris and sediment storage and (2) management influences on the role of adjacent stands as sources of pieces of large wood in streams.

These examples of research needs emerge from using sediment routing and budgeting concepts to analyze shortcomings of earlier approaches. Ultimately, management impacts will be quantified with detailed sediment budgets. For the present, however, routing and budgeting concepts provide new perspectives for analyzing management effects on geomorphic systems and help to identify important problems for future research.

LITERATURE CITED

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

Dyrness, C. T.
1967. Mass soil movements in the H. J. Andrews Experimental Forest. USDA For. Serv. Res. Pap. PNW-42, 12 p. Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

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, Oreg.

Fredriksen, R. L.
1972. Nutrient budget of a Douglas-fir forest on an experimental watershed in western Oregon. In Proc. Symp. Research on Coniferous Forest Ecosystems. p. 115-131. J. F. Franklin, L. J. Dempster, and R. H. Waring, eds. USDA For. Serv., Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

Fredriksen, R. L., and R. D. Harr.
1979. Soil, vegetation, and watershed management. In Forest soils of the Douglas-fir region. p. 231-260. P. E. Heilman, H. W. Anderson, and D. M. Baumgartner, eds. Coop. Ext. Serv., Wash. State Univ., Pullman.

Gray, D. H.
1970. Effects of forest clearcutting on the stability of natural slopes. Bull. Assoc. Eng. Geol. 7:45-66.

Harr, R. D., and F. M. McCorison.
1979. Initial effects of clearcut logging on size and timing of peak flows in a small watershed in western Oregon. Water Resour. Res. 15:90-94.

Janda, R. J.
1978. Summary of watershed conditions in the vicinity of Redwood National Park, Calif. U.S. Geol. Surv., Open-File Rep. 78-25, 82 p.

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

Megahan, W. F., and R. A. Nowlin.
1976. Sediment storage in channels draining small forested watersheds in the mountains of central Idaho. In Proc. Third Fed. Interagency Sediment. Conf. p. 4-115--4-126. Denver, Colo.

O'Loughlin, C. L.
1974. A study of tree root strength deterioration following clearfelling. Can. J. For. Res. 4:107-113.

Pearce, A. J., and C. L. O'Loughlin.
1978. The forest scene: Hydrologic and geomorphic processes on forested lands - A selective overview. In Proc. Conf. on Erosion Assessment and Control in New Zealand. p. 62-76. N.Z. Assoc. Soil Conservators. Christchurch, N.Z.

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

Swanson, F. J.
1981. Fire and geomorphic processes. In Fire regime and ecosystem properties. H. A. Mooney, T. M. Bonnicksen, N. L. Christensen, J. E. Lotan, and W. A. Reiners, eds. USDA For. Serv. Gen. Tech. Rep. WO-26. p. 401-420.

Swanson, F. J., and C. T. Dyrness.
1975. Impact of clearcutting and road construction on soil erosion by landslides in the western Cascade Range, Oregon. Geology 3:393-396.

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 M. E. James.
1975. Geomorphic history of the lower Blue River - Lookout Creek area, western Cascades, Oregon. Northwest Sci. 41:1-11.

Swanson, F. J., and G. W. Lienkaemper.
1978. Physical consequences of large organic debris in Pacific Northwest streams. USDA For. Serv. Gen. Tech. Rep. PNW-69. 12 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. USDA 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. D. R. Coates, ed. Dowden, Hutchinson, and Ross, Inc., Stroudsburg, Pa.

Varnes, D. J.
1978. Slope movement types and processes. In Landslides - analysis and control. R. L. Schuster and R. J. Krizek, eds. Spec. Rep. 176, p. 11-33. Natl. Acad. Sci., Washington, D.C.

Discussion Group Summaries

Evaluating and Mapping Sources and Temporary Storage Areas of Sediment

Leslie M. Reid

INTRODUCTION

Legislation to regulate forest practices, water quality, and management of federal lands has increased the land managers' need for efficient methods of identifying and mapping sources of sediment in forested basins. At the same time, theoretical analysis of landscape evolution has led research geomorphologists to the consideration of many of the same sedimentation problems as those confronting land managers. This discussion session demonstrated that, although many of the fundamental questions addressed by these groups are identical, the approaches taken in answering those questions differ widely.

APPROACHES AND GOALS

Land managers are interested in the effects of varying land use on sediment production. Often they are asked to map the distribution of and predict sediment production from natural as well as management-induced and management-enhanced sources. They may also wish to predict distribution of sediment sources and rates of sediment production for different management alternatives in specific areas. They are interested in such questions as how timber harvesting or roadbuilding affects slope stability in an area, how the amount of sediment currently in storage in flood plains will be affected by a change in land use, and what happens to sediment introduced into a stream system by a change in land use. Usually answers are desired quickly.

Research geomorphologists may try to answer these same questions from a theoretical standpoint. They are interested in the distribution of erosional processes, for example, in order both to understand the importance of various external controls on those processes and to infer the importance and specific influence of the processes on landscape form and development. The goal of a land manager's investigation is a prediction specific to a locale, but the research geomorphologist's goal is usually an understanding of a general principle. Such general principles may be useful to a land manager as tools for making predictions in a specific area, but if other methods give adequate results more quickly or more economically, there is no compelling reason to use a theoretical approach.

The difference in approach stems from a basic difference in intent. The land manager must answer site- or area-specific questions quickly, but the researcher is interested in elucidating general principles and has the luxury of working on a more relaxed time scale.

Land managers have used the landscape stratification approach--also referred to as terrain mapping or morphogenetic classification--to characterize erosion processes over wide areas quickly. Underlying this approach is the assumption that areas of similar climate, topography, bedrock, vegetation, and land use experience the same kinds of erosion processes, produce sediment at similar rates, and respond in a similar manner to a given land use. The criteria used to define stratification units are chosen not only because of their genetic importance in controlling distribution and rate of erosion processes, but also because they are visible on aerial photographs

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or, as is often the case with bedrock geology, have already been mapped. A representative area in each defined unit can then be examined in detail by using field mapping, transects, monitoring programs, or more detailed aerial-photograph interpretation, and the results of the studies may be used to determine a characteristic sediment yield, process distribution or rate, or, if a disturbed area is selected, response to land use. The values for a representative area are then used to typify that unit or to provide the basis for revised stratification criteria that are more appropriate for the area of interest. This method has been widely used to evaluate slope stability (examples from T. Chamberlin, B. Hicks, and W. Megahan) and has also proved useful in assessing and predicting effects of logging on channel stability (example from O. Williams). In this last example, topographic maps and aerial photographs were used to classify the channels into one of seven stability classes. Observations of the response of representatives of each stability class to a given logging practice were then used to predict the response of a specific channel to that practice.

The stratification approach cannot be used when site-specific answers are needed. This approach can predict the kind of landslide likely to occur and the average failure frequency in an area, but it cannot be used to determine the locations of individual failures or the conditions under which a given site will fail. Nor can the stratification approach be used reliably to predict process distribution or response beyond the range of the impacts already existing on each stratification unit. For example, a stratification-based approach may use data from a 10-year-old clearcutting to predict that clearcutting of a certain basin will result in a threefold increase in yearly sediment production from landslides. The approach can give no indication, however, of how long the effect will last unless a series of progressively older clearcuts exists in the same stratification unit. Similarly, the effect of a management plan for which no prototype exists on that stratification unit cannot be evaluated. Finally, the method is restricted to areas and processes for which controlling variables are easily mapped. In areas where debris-avalanche occurrence is controlled by the distribution of deep soil wedges or of localized bedrock joints, stratification by hillslope gradient, vegetation, and bedrock lithology, each of which may have little relation to the primary controls, would be of little value.

For such areas, an approach frequently adopted by researchers may be useful. Individual erosion processes are conceptually isolated and the controls on the rate and distribution of each process are quantified (Dietrich et al., this volume). Long-term impacts of current management activities or the effects of a proposed management activity may then be predicted by analyzing the response of the processes to the changing primary controls. Sediment production from gravel road surfaces, for example, is largely controlled by factors such as overland flow length, road gradient, and road use. These relationships have been defined in the Clearwater basin, Washington, and can be used to estimate sediment yields from new road designs and use patterns by determining the effects of the changes on the factors controlling sediment production (L. Reid). If process controls are adequately understood, site-specific process rates may be predicted using a combination of this technique and a probabilistic treatment of the driving variables.

The process approach thus can be used not only to solve the same kinds of problems that the stratification approach deals with, but also to make predictions for specific locations and for conditions without prototypes. The process approach requires an initial investment of effort that may make it impractical for short-term reconnaissance work, but if more specific questions are anticipated or if the necessary scale of stratification is expected to become more detailed, the stratification criteria will need to be more process-oriented. In such a case, an initial commitment of time toward the development of a quantitative understanding of primary controls of process distribution and rate may be most economical in the long run.

METHODS: DISCRETE SOURCES

The methods used to evaluate sediment sources can be applied to both the stratification and process approaches. Methods such as field mapping and aerial photograph interpretation are standard techniques for evaluating discrete sediment sources such as landslides. If these techniques are to be used quantitatively, however, care must be taken to avoid comparing areas of different resolution without making allowances for that difference. For example, debris-avalanche frequencies cannot be compared by counting landslide scars if the scars heal more quickly in one area than in another, and a count of scars in a logged area compared to one in a forested area has little meaning unless the size of the smallest scar recognizable beneath the tree canopy is known.

Analysis of sequential aerial photographs is of particular value in determining frequencies and recovery times for different processes, but in many areas, few photograph sets are available. A source of aerial photographs frequently overlooked is county tax records in courthouse files: since the early 1940's, governments of many timber-producing western counties have been taking aerial surveys at 4- to 10-year intervals to assess timber holdings (R. Janda, D. Hardin, M. Nolan).

Ground photographs, too, may facilitate the evaluation of past geomorphic and vegetative change or even allow the reconstruction of process distribution and rate in areas before disturbance. In addition, both sequences of aerial and ground photos may be used in conjunction with climatic records to assess the impact of specific storms (S. Duncan, D. Hardin). Historical photographs of the area of interest may exist locally, and during the 1930's the Soil Conservation Service compiled a series of high-quality photographs documenting erosion problems in each State. Though the emphasis of the series was on agricultural and range lands, other land use categories were included. The photograph collection is preserved in the National Archives (S. Trimble). The U.S. Geological Survey also maintains a collection of historical photographs (R. Janda).

Other remote-sensing techniques have recently been applied to geomorphic problems. Both aerial infrared and radar imagery have been used successfully to map areas of landslide hazard. Lineations observed on infrared photographs in New Zealand were found to correspond to moisture differences between coherent bedrock and more slide-prone, crushed zones, and thus correlated well with landslide distribution (M. Harvey). In Idaho, lineations made visible by radar imagery also correlated very well with landslides (W. Megahan). Computer-rectified side-scanning radar images, because of their fine resolution and low-angle illumination, are useful in discerning subtle differences in relief such as those produced by landslides, but like visible-light photography, radar imagery cannot penetrate forest cover. These techniques can be used not only to map sediment-source distribution, but also to help map the distribution of rate-controlling variables such as vegetation, hillslope gradient, and bedrock geology. Other controls, such as soil depth and ground-water level, are less amenable to analysis by remote-sensing techniques.

Wedges--areas of deep soil on hillslopes with otherwise uniform soil depth--are very important in determining the location of debris avalanches in some areas, yet because wedges may have little or no surface expression, they are very difficult to recognize in the absence of road cuts. Portable seismic units have shown promise in measuring soil depth if the soil-bedrock interface is distinct (S. Duncan), and a portable conductivity meter is being developed for this purpose by the Bureau of Land Management. A prototype already on the market has been used successfully in southern Oregon. The conductivity meter is used to measure the depth to a subsurface contrast in conductivity, and because such a contrast may be the result of any of several factors including material, density, and moisture differences, readings must be calibrated for each area by digging test pits (B. Hicks). Eventually an understanding of how wedges form may make possible the prediction of their frequency and distribution in specific areas.

Mapping of peak or time-averaged piezometric levels is fraught with similar difficulties. Because the temporal occurrence of landslides is so closely tied to the occurrence of major storms and thus to high piezometric levels, maps showing the probable duration of a given piezometric level may prove useful in determining landslide potential in an area (T. Chamberlin). But as yet, the use of piezometers is the only widely available method for measuring ground-water pore pressures, and the time and effort necessary to install piezometers has restricted their use to networks covering only small areas. In this case, a more process-oriented approach, where the controls on piezometric levels are determined in detail in a small area and then projected to the larger area of interest, may be most useful (S. Duncan). The technology necessary for remote sensing of soil thickness, ground-water depth, and other process controls may already exist; it may require only a closer working relationship with specialists in remote sensing to link the geomorphologist's needs with the capabilities of remote sensing.

METHODS: DISPERSED SOURCES

In many areas, dispersed sources such as surface wash, rill erosion, and small-scale bank erosion may contribute as much or even more sediment to stream channels than large, discrete sources. Even in areas where the net contribution from dispersed sources is small, they may be important as persistent sources of fine material. In addition, dispersed sources may prolong the impact of a landslide by preventing its revegetation, and in the case of bank erosion, may locally create the potential for discrete failures.

Because of the more continuous nature of their sediment production and because they are not discernible by remote sensing, dispersed sources require a more process-oriented approach to analysis even for purposes of stratification. Empirical correlations of sediment yield with different types and intensities of land use have in the past dealt implicitly with such sources, but because these studies require a large data base and do not distinguish specific source types, they are generally useful only in broad stratification schemes.

Direct monitoring of individual sources provides information that can be projected to stratification units of any size based on the distribution of rate-controlling variables. Monitoring methods depend, of course, on the types of processes in question. Surface erosion by such processes as rainsplash and rilling has been monitored using erosion pins on slide scars and gully walls (Lehre, this volume); erosion pins have also proved useful for measuring back-cut retreat on roads. With frequent observations it is possible to isolate the effects of specific storms and freeze-thaw, and careful selection of study plots allows the isolation of other variables such as aspect, lithology, slope angle, and slope length. Gerlach troughs are effective in monitoring sediment loss from the same sources (R. Janda), as are repeated measurements of surface elevation and sediment character on erosion plots (R. Rice). Erosion-plot studies using measurement of erosional landforms such as rainsplash pedestals and gullies have also been used to estimate effects of timber harvest and road construction on soil movement (Hauge 1977). Surface erosion by overland flow on gravel roads is also a significant source of fine sediment and can be monitored by measuring sediment concentrations in surface or culvert drainage. Sites must be selected carefully to isolate rate-controlling variables such as road gradient, drainage area, back-cut contribution, road use, and storm intensity. Use of a portable rainfall simulator may add flexibility in site selection (L. Reid).

In some areas, surface erosion can be measured by observing the depth of soil profile truncation. Because profile truncation is a criterion used in describing soil series, the extent of surface erosion may on occasion be mapped directly from existing soils maps (S. Trimble). Similarly, buried profiles indicate net accumulation or storage.

SEDIMENT STORAGE

Once sediment has entered the transport system, it may be stored for periods ranging from minutes to millions of years in landscape elements such as hollows, streambeds, gravel bars, and terraces. Though measurements of sediment production rates are important in determining the amount of sediment entering stream systems, an evaluation of sediment storage is necessary if the location, magnitude, and duration of the long-term impacts resulting from sediment production are to be determined. In addition, many sediment-related impacts of major concern involve a change in the location of sediment storage or in the character and amount of the material stored; such impacts include stream siltation, infiltration of fine sediment into spawning gravels, filling of pools, and channel destabilization. Finally, any accounting of sediment production that attempts to relate production rates from individual sources to a basin sediment yield must also take account of changes in volume of material in storage, because sediment may be either gained from or lost to storage during transport.

Trimble (1981) demonstrated the importance of this effect for the Coon Creek basin in Wisconsin, and further suggested that the present sediment load of many streams was determined by basin conditions over past decades or even hundreds of years. At Coon Creek, sediment originally produced by poor agricultural practices in the late 1800's was stored in the uplands for several decades. By the 1930's, it had been transported downstream as far as the main valley, resulting in massive aggradation of flood plains, even though agricultural practices had improved markedly by that time.

Similar effects are seen in the movement of sediment waves through stream systems. Coarse sediment introduced into a river at a point (as from a landslide) may travel as a slow-moving wave that may be recognizable only as an otherwise anomalous local change in channel morphology. This sediment is effectively in storage and is moving independently of any processes occurring at its source; its major impact at a downstream point will not be felt until the wave of sediment reaches that point. Such effects have been described in western Washington (Madej, this volume), northern California (Kelsey 1980), New Zealand (M. Harvey), and elsewhere. Thus, an analysis of sediment storage is necessary to answer such questions as what happens to sediment once it enters the transport system, what areas will be affected by an increase in sediment production, how long will the impact persist, and how will material already in storage respond to a change in land use.

A major problem in the evaluation of sediment storage is that the distinction between a source and a sink for sediment is not always clear. Because of the continuum between transport and storage, a meaningful assessment of either must consider residence time of material (D. Harden). Most landscape elements can act in either capacity, depending on recent climatic conditions, location in a basin, magnitude of a specific climate event, or local environmental conditions. This last point is of special interest, because it implies that a change in land use, and thus in factors such as local water balance or apparent soil cohesion from root strength, may be large enough to convert storage elements into active sources.

Location within the drainage network is also important in determining the response of a storage element to changing conditions. In the northern California Coast Range, for example, small headwater streams tend to aggrade their beds during small storms and degrade during large, peak-flow events, but sediment aggrades in larger streams during large events and is gradually eroded during the smaller ones (Janda 1978).

A further complication to the analysis of sediment storage is the discontinuous nature of sediment transport. Transport of gravel is accomplished by the alternation of long periods of stationary storage on the streambed with brief episodes of rapid transport during storms. Similarly, flume studies in Colorado suggest that coarse material introduced into a stream is transported as a series of prograding fans in channels (Harvey, this volume). A clast is stored in any given fan for a relatively long period before it is once again eroded out and transported to the toe of the fan. In each case, the sediment in "active" transport spends a large proportion of its transport time in storage.

On a larger scale, entire landscape elements may be undergoing similar cycles. Hollows, for example, are thought to gradually fill with colluvium until the fill--or wedge--reaches a critical depth. At this point, high pore-water pressure during a large storm can cause it to fail. The colluvial fill is transported downslope, and the slightly enlarged bedrock depression is left to refill (Dietrich et al., this volume).

The importance of time scale is evident in each of these examples. Wedges and low-order stream channels, which appear to be stable sediment traps during most of their cycles, are seen to be net sources of sediment when observed over several cycles. In addition, the dependence of geomorphic response on storm magnitude adds a probabilistic variable to any long-term evaluation of a landscape element.

SUMMARY

Maps based upon quantitative, process-based studies and those based upon empirical correlations of source distribution with landform parameters both have been used to evaluate sediment production. Which of these approaches is most useful depends on the purpose of the specific study, but methods that have been developed for measuring production rates can be used in conjunction with either approach. These methods include various mapping techniques, monitoring programs, and experiments. Analysis of sediment production in a drainage basin must take into account changes in sediment storage, but this is made difficult by the sensitivity of storage elements to factors such as climatic history, magnitudes of individual storms, and local physical and biological conditions. In addition, because the distinction between storage and transport is to some extent arbitrary, and because landscape elements may undergo natural cycles alternating between net sediment input and net output, the definition of sediment-storage elements is partially dependent on the time scale of interest.

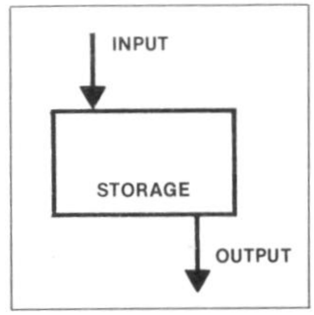
LITERATURE CITED

Hauge, C. J.
1977. Timber harvesting and soil erosion. Interim Report, Dep. For., State of Calif., Sacramento. 55 p.

Janda, R. J.
1978. Summary of watershed conditions in the vicinity of Redwood National Park. U.S. Geol. Surv. Open-File Rep. 78-25, 82 p. Menlo Park, Calif.

Kelsey, H. M.
1980. A sediment budget and an analysis of geomorphic process in the Van Duzen River basin, north coastal California, 1941-1975. Geol. Soc. Am. Bull., part II, April 1980, 98 p.

Trimble, S. W.
1981. Changes in sediment storage in the Coon Creek Basin, Driftless Area, Wisconsin, 1853 to 1975. Science 214(4517):181-183.



Tracing and Dating the Movement and Storage of Sediment

Randall B. Brown and Linda B. Brubaker

STATEMENT OF THE PROBLEM

Participants agreed that the most recent sediments, including those that are currently active, are of primary concern in tracing and dating the movements of sediment in forest drainage basins. Thus, the discussion revolved around techniques that might be applicable in dating and tracing sediments up to tens of thousands of years in age. Of interest are the rates of movement and the magnitude and frequency of sediment-transport processes. A particularly acute problem is the dating of recent alluvium in channels and terraces, and of colluvium on hillslopes and fans.

Participants further agreed that the technology is more advanced for DATING sediments than for TRACING the movement of sediment. Thus, for discussion, the two problems were separated.

Dating of Sediment

Dendrochronology holds great promise for application in sediment budgets and routing. The central technique of dendrochronology is counting annual growth-rings in wood to date the tree; the age of the tree may then place limits on the onset or duration of some geomorphic process--such as sheet or gully erosion, deposition, or flooding at the site. Other information extracted from patterns of tree-ring width or density provide data on environmental conditions, such as drought or cold weather, which are germane to the interpretation of sedimentary records.

The range of time over which these methods are useful extends from the present as far back as several thousand years. Living or dead trees can be used and compared with accurately dated, tree-ring chronologies on file at research centers, to date sequences of ring widths. Recent developments in the measurement of density by X-rays aid in finding late-wood density patterns that are otherwise unrecognizable.

Sigafoos (1964), for example, has shown how the counting of tree rings outside of healed scars--and the bowing of trees and consequent adjustments in wood anatomy and sprouting--can indicate flood damage to trees at various elevations on a valley floor. Careful documentation and dating of these indicators allow the construction of a flood-frequency curve. Sigafoos also dated changes in rooting patterns after partial burial of tree trunks to document sedimentation on a floodplain. LaMarche (1968) measured the height of exposed roots on bristlecone pine (Pinus aristata Engelm.) trees in the White Mountains of California. The trees, which are several thousand years old, were dated by counting growth rings. Dunne et al. (1978, 1979) have used the method on much younger trees and bushes to map profiles of erosion along hillslopes in Kenya, and to indicate recent acceleration of erosion. Dietrich and Dunne (1978) used dendrochronology to define the rate of infilling of hillside hollows that had been excavated by landslides. Other studies have considered damage to live trees by fire (Heinselman 1973) and

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splitting by differential ground movement (Schroder 1978), which both produce rings that are not continuous around the trunk circumference. Alestalo (1971) used dendrochronology to study a variety of geomorphic processes.

One intriguing approach is suggested by the tendency for exposed tree roots to take on the character of tree stems for the duration of the exposure. Dating the sequence of change from root to stem character on cut faces can aid in the reconstruction of cut and fill sequences. Relative dating techniques might then be used to relate such cut faces and escarpments with associated geomorphic surfaces.

The width of lichen colonies is an index of the age of exposed raw materials under a constant climate (Benedict 1967). Before this approach can be applied in a given area, however, age-size relationships must be defined by the measurement of lichens on surfaces of known age. Because of the possibility of lichens invading a fresh deposit or exposure before trees, lichenometry can be a valuable tool in dating, especially in the range of tens to hundreds of years before the present.

Several isotopes can be used as markers/tracers in sediment movement studies. Lead-210, for example, is a natural isotope that falls from the atmosphere at a more or less constant rate at a given latitude, and is deposited in reservoirs and other depositional sites. Changes in ^{210}Pb levels in depth profiles reveal periods of accelerated sedimentation rates that, in at least one instance, have been shown to correspond with historical records of human disturbance (Nevissi and Schell 1977). The technique is useful over the range of about 10 to 200 years before the present.

Cesium-137 is a nuclear-fission product with fallout levels that peaked in the early 1960's as a consequence of atmospheric testing of nuclear weapons (Ritchie et al. 1975). This isotope is attached readily to soil particles and serves as a tag on fine sediments. Its best use seems to be as a datable stratigraphic marker. Overthickening or thinning of ^{137}Cs-rich surface layers likewise has utility in estimation of sedimentation/erosion rates.

Applications also may exist for some other geochemical techniques, such as the use of $^{12}C/^{13}C$ ratio changes in organic materials as a function of increased burning of fossil fuels in recent times, the worldwide increases in Pb content of organisms from accelerated gasoline use over recent decades, or changes in deuterium ratios in wood.

Classical ^{14}C dating has severe limitations over the range of tens to hundreds of years. Imprecision over this range, fluctuations in atmospheric ^{14}C levels, contamination problems, and the complexity of interpreting ^{14}C dates for decomposed organic materials combine to make ^{14}C dating on very recent sediment a risky undertaking (Campbell et al. 1967, Ruhe 1975).

Historical records are a powerful tool in assessing recent geomorphic events. Stream cross sections taken in bridge design and construction work, topographic information collected in railroad construction, aerial photographs taken over the last 50 years, and other photographic records have been used with great success in assessment of modern erosion and sedimentation rates, stream-channel changes, and gullying. Historical and archeological artifacts are also of use (Leopold and Snyder 1951; Trimble 1970, 1981). Knowledge of fire and earthquake activity and major climatic events can give important clues to periods of high landscape instability in some areas.

Several other dating techniques were discussed in this session. Changes in pollen percentages in a sediment core sample, the arrival of exotic species, and other uses of time boundaries in pollen profiles can be helpful in sedimentation studies. Pedogenic processes and the degree of development of soil features can serve as indicators of age on recent geomorphic surfaces or buried surfaces. Particle-size distribution and the depth variation of grain size can help in assessing erosional/depositional history. Natural stratigraphic markers--such as volcanic ash deposits and paleosols--are useful in identifying buried, exposed, or relict geomorphic surfaces.

Tracing of Sediment

The task of tracing sediment movement is more difficult than dating, because the conceptual methods and technology for tracing sediments are not as well developed. Nevertheless, some tracing techniques do exist and several were discussed.

Rock and mineral provenance studies, including study by scanning electron microscope of quartz-grain morphology (Glasmann and Kling 1980), are useful techniques in identifying source areas of sediment. The influence of deep-seated failures or deep gullying may be detectable with provenance studies in those instances where source landscapes have depth variations in mineralogy or degree of weathering (Youngberg et al. 1971). Thompson et al. (1980) have reviewed how the analysis of various magnetic properties of sediment can be used to identify sediment sources both after deposition of the sediment and during its transport.

Application of other sedimentological techniques developed by petrologists and engineers also have use in sediment routing. Areal/depth variation in particle-size distribution, and analysis of sedimentary structures, can be useful in assessing rates and types of transport processes.

Tracing experiments that use painted, fluorescent, or exotic rocks have limitations in natural systems. Because only a small percentage of the tagged objects are usually recovered, the fate of the majority of tagged objects is uncertain and so, thus, are the major transport pathways of the system. Also, the dynamics of the mass or pulse of sediment as a whole are not well elucidated by this technique. Participants suggested that simplified versions of such experiments, as in a rock-floored channel or a controlled, areally limited environment would be easier to interpret at the current stage of knowledge. More complex experiments could be designed later.

Use of radioactive tracers presents other problems. The controversial nature of such materials and their dilution in transport and consequent low detectability make them almost prohibitive in field studies. Tracers are more useful in controlled settings such as small flumes or small drainage-basin models. Radioactive minerals such as uraninite are traceable in small-scale bedload-transport studies.

We now have a dearth of conceptual models to describe the fate of a marked particle in a sedimentary system. Until such models are developed, designing useful tracing experiments and interpreting the results of experiments already done will be difficult (e.g., Laronne and Carson 1976, p. 82; Mosley 1978; Dietrich et al., this volume).

LITERATURE CITED

Alestalo, J.
1971. Dendrochronological interpretation of geomorphic processes. Fenn., Soc. Geogr. Fennaie 105:1-140.

Benedict, J. B.
1967. Recent glacial history of an alpine area in the Colorado Front Range, USA. I. Establishing a lichen-growth curve. J. Glaciol. 6:817-832.

Campbell, C. A., E. A. Paul, D. A. Rennie, and K. J. McCallum.
1967. Applicability of the carbon-dating method of analysis to soil humus studies. Soil Sci. 104:217-224.

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

Dunne, T., W. E. Dietrich, and M. J. Brunengo.
1978. Recent and past erosion rates in semi-arid Kenya. Z. Geomorphol. Suppl. 29:130-140.

Dunne, T., W. E. Dietrich, and M. J. Brunengo.
1979. Variation and controls of soil erosion in the grazing lands of Kenya. Int. Assoc. Hydrol. Sci., Proc. Canberra Symp. on the Hydrology of Areas of Low Precip., p. 421-428.

Glasmann, J. R., and G. F. Kling.
1980. Origin of soil materials in foothill soils of Willamette Valley, Oregon. Soil Sci. Soc. Am. J. 44:123-130.

Heinselman, M. L.
1973. Fire in the Boundary Waters Canoe Area, Minnesota. Quat. Res. 3:329-382.

LaMarche, V. C.
1968. Rates of slope degradation as determined from botanical evidence, White Mountains, California. U.S. Geol. Surv. Prof. Pap. 352-I, p. 341-377.

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.

Leopold, L. B., and C. T. Snyder.
1951. Alluvial fills near Gallup, New Mexico. U.S. Geol. Surv. Water-Supply Pap. 1110-A, 17 p.

Mosley, P. M.
1978. Bed material transport in the Tamoki River near Dannevirke, North Island, New Zealand. N. Z. J. Sci. 21:614-626.

Nevissi, A., and W. R. Schell.
1977. Sedimentation rates in Puget Sound and the application to heavy metals pollution. In The use, study, and management of Puget Sound, p. 81-87. A. Duxbury, ed. Univ. Wash., Seattle.

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

Ruhe, R. V.
1975. Geomorphology, geomorphic processes and surficial geology. Houghton Mifflin Company, Boston. 246 p.

Schroder, J. F.
1978. Dendrogeomorphological analysis of mass movement on Table Cliffs, Utah. Quat. Res. 9:168-185.

Sigafoos, R. S.
1964. Botanical evidence of floods and floodplain deposition. U.S. Geol. Surv. Prof. Pap. 485-A, 35 p.

Thompson, R., J. Bloemendal, J. A. Dearing, F. Oldfield, T. A. Rummery, J. C. Stober, and G. M. Turner.
1980. Environmental applications of magnetic measurements. Science 207:481-486.

Trimble, S. W.
1970. The Alcovy River Swamps: the result of culturally accelerated sedimentation. Bull. Ga. Acad. Sci. 28:131-141.

Trimble, S. W.
1981. Changes in sediment storage in the Coon Creek Basin, Driftless Area, Wisconsin, 1853 to 1975. Science 214(4517):181-183.

Youngberg, C. T., M. E. Harward, G. H. Simonson, D. Rai, P. C. Klingeman, D. W. Larson, H. K. Phinney, and J. R. Bell.
1971. Hills Creek Reservoir turbudity study. Water Resour. Res. Inst. and Dep. Soil Sci., Oreg. State Univ., Corvallis. WRRI-14, 252 p. plus append.

Weathering and Soil Profile Development as Tools in Sediment Routing and Budget Studies

Deborah R. Harden, Fiorenzo Ugolini, and Richard J. Janda

INTRODUCTION

Studies of soil morphology and genesis have, for the most part, focused on sites where profile characteristics result primarily from vertical pedogenic processes and are little influenced by repeated erosion or deposition. Soil investigations at such sites may be used to place limits on the age of some relatively stable parts of the landscape (Birkeland 1974), but they are of little value in studying processes and rates of sediment movement through drainage basins. On most sloping landscapes, periodic erosion and deposition are integral factors in soil formation. Additionally, weathering and soil formation strongly influence the hydrologic and erosional characteristics of the regolith. The interactions between geomorphic and pedogenic processes have been discussed in a variety of intriguing papers (e.g., Zinke and Colwell 1965, Conacher and Dalrymple 1977, Parsons 1978); however, workshop participants believe that these concepts will have to be developed more fully before they can be used routinely in developing sediment budgets for forested drainage basins. Nonetheless, participants did cite several ways in which soils and weathering characteristics can increase understanding of some aspects of sediment routing.

Pedological and geochemical processes concurrently operating near the earth's surface contribute to the production of material that is potentially erodible. In some instances, as in the case of areas affected by podzolization, the distinction between pedological and geochemical processes is strikingly distinct (Ugolini et al. 1977). Common to these processes is weathering which changes the strength, grain size, and mineralogical composition of geological materials.

The discussion revealed that recent attempts by soil scientists to achieve professional consensus on terminology used to classify and to describe soil profiles (U.S. Department of Agriculture, Soil Survey Staff 1960, Soil Conservation Service 1975) have resulted in a complex vocabulary that is not yet widely accepted by geologists and hydrologists. Most workers discuss concepts of soil morphology and genesis in terms that strongly reflect their professional specialization and field experience in specific types of terrain. Thus, communication between researchers with different specializations or experience in contrasting terrains is sometimes difficult. The greatest difficulty is in describing profiles that reflect episodic deposition or erosion as well as weathering; such conditions are widespread on forested hillslopes. Additional complications in terminology result when stratigraphic studies indicate that deep soil horizons reflect past climatic and vegetation conditions different from existing conditions (Janda 1979). Soil horizons, e.g., B3 horizons in ultisols, considered crucial in some soil stratigraphic studies are regarded to be below the solum and in some studies of nutrient cycling and erosion processes, considered saprolite or parent material.

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Table 1--Soil properties and diagnostic horizons useful as age indicators (adopted from Yaalon 1971)

Easily reversible, rapidly developed (10^1-10^3 years)	Less easily reversible (10^2-10^4 years)	Irreversible, slowly developed (10^3-10^5 years)
pH	Clay illuviation	Oxic horizon
Organic matter	Iron oxide accumulation	Plinthite
Base saturation	Physical properties	Fragipan
Folic horizon	(bulk density, engineering properties)	Placic horizon
C:N ratios	Argillic horizon	Duripan
Cambic horizon---------	Cambic horizon	Calcrete
Gleying	Histic horizon (fire area)	Histic horizon
Spodic horizon	Clay mineralogy	Clay mineralogy
Mollic horizon	Calcic horizon	Weathering rinds
	Umbric horizon	Mineral etching

NOTE: Many of the properties may be more or less reversible in different environments.
Absolute age determination methods: ^{14}C, fission-track, thermoluminescence, ^{137}Cs, U-series, amino acid racemization.

STRATIGRAPHIC STUDIES

Soil stratigraphy (the use of surface and buried soils to subdivide and to correlate sediments primarily of Quaternary age (Birkeland 1974)) may be useful in constructing sediment budgets by identifying the relative stability of different landscape elements and by placing limits on frequency and rate of erosion. Participants in the discussion group compiled a table of soil properties that can serve as possible age indicators and their response times to environmental change (table 1).

Easily reversible properties, such as pH, organic material, carbonates, and others, are probably in equilibrium with existing climatic and vegetation conditions at a site, but irreversible properties, such as plinthite, oxic horizons, duripan, and others, may persist under conditions that are drastically different from those under which they developed (Yaalon 1971).

Soil stratigraphy may be used to identify relative contributions of different landscape units to the overall sediment yield from a drainage basin. Low erosion rates on hillslopes and ridges are indicated by soils displaying strongly developed profiles and by clay mineralogy and pollen reflecting climatic and vegetation conditions that have not existed for tens or even hundreds of thousands of years. In contrast, hillslopes bearing weakly developed soil profiles or active colluvium indicate more rapid erosion rates. Downslope movement of weathered colluvium or intact blocks of strongly developed soils complicate the use of soil-profile characteristics for estimating rates of erosion.

Application of erosion rates computed from measured stream-sediment discharges to only those parts of drainage basins where soils suggest active erosion provides more realistic estimates of rates of landscape development than traditional computations of drainage basin-wide denudation. In addition, areas of temporary sediment storage may be identified in the landscape, and rates of accumulation or residence time may be estimated by soil-profile characteristics.

Another potential use of soil stratigraphy in studying erosion processes is as an indicator of recurrence intervals of episodic erosion events. Different ages of landslide episodes can often be recognized by differences in soil depth and weathering intensity, or by buried soil horizons. Workshop participants believe that this type of investigation in terrain sculpted by frequent (that is, where frequency of failure at a site is on the order of 10^1-10^3 years), shallow debris slides would require a refinement of existing soil stratigraphic techniques and probably the use of rapidly developing, easily reversible soil properties (table 1). In constrast, age difference between larger, more infrequent slumps and translational slides can probably be established through use of existing soil stratigraphy techniques based on slowly developing, irreversible soil properties (table 1).

Several participants pointed out that such time-dependent weathering characteristics as thickness of weathering rinds of clasts, ratios of fresh to weathered clasts, and etching of heavy minerals are diagnostic and often easier to use than soil-profile development. In some settings, differences in specific surface and subsurface rock-weathering characteristics actually give a clearer separation of Quaternary sediments than differences in soil-profile development (Burke and Birkeland 1979).

PROCESS STUDIES

Map distributions and profile characteristics of soils in forested areas can help distinguish erosion and deposition processes operating on hillslopes. For example, some hillslopes in northwestern California that appear sculpted predominantly by creep and streamside translational slides show progressively less weathered colluvium and younger, shallower soils in a downslope direction. In contrast, some slopes that experience episodic debris avalanches show imbricate or interfingering arrangements of fresh colluvium over saprolite, or of strongly developed soils over weakly developed ones. Some hillslopes that appear sculpted by large-scale rotational slumping show intact blocks of deep, mature soils that moved downslope from stable interfluve positions. Detailed soil maps may help identify the prevalent erosional processes operating on individual hillslopes and the frequency with which those processes occur.

One model of the linkage between soil formation and hillslope erosion in regions sculpted primarily by shallow, episodic mass movement and creep assumes that rates of weathering determine availability of sediment (Dietrich and Dunne 1978; Dietrich et al., this volume). As the regolith becomes deeper and increasingly weathered, it becomes increasingly susceptible to catastrophic removal by debris slides. In terranes where the rate of production of saprolite is exceeded by its rate of removal, the rate of soil formation determines sediment production. However, in landscapes where rocks do not have to be weathered to be eroded or where mass-movement processes extend deeper than the regolith, erosion rates may be independent of rates of weathering. Workshop members also pointed out that the type and rate of revegetation may be a controlling factor in reconstituting eroded soils.

Workshop participants agreed that better understanding of weathering and soil-forming processes is necessary for developing a model of hillslope erosion. Much recent work has been done on chemical weathering in soil formation, but physical weathering processes in forested environments have been studied much less intensively. Production of fine-grained particles from exposed and buried rock surfaces disrupts primary rock structure (stratification, foliation, etc.). Some members believe this disruption is caused solely by physical disaggregation, but others think that volume changes associated with chemical weathering may aid in disrupting primary rock and saprolite structure. Some scientists use disruption of rock structure to define the boundary between the solum and the undisturbed saprolite or bedrock. In a landslide scar, accretion by sloughing into the bare surface followed by revegetation may be necessary before rock disaggregation and soil formation can proceed. The model of physical disruption of saprolite to produce the solum may be valid if shallow erosion processes are prevalent. Other landscapes may have different regimes of soil erosion and formation and of regolith formation, however. Workshop members agreed that a better understanding of these regimes would aid in applying soil studies to studies of sediment production and budgets.

Several participants pointed out that residence times in some temporary sediment-storage compartments in forested drainage basins are long enough to allow weathering to modify grain-size distributions and other physical properties of sediment and thereby influence subsequent erosion. In areas of relatively slow erosion and rapid weathering, weathering may also significantly reduce the total mass of sediment moving through the drainage basin. In these circumstances, landscape evolution can probably be understood best by working with mass budgets that include transport of dissolved solids as well as sediment (Cleaves et al. 1970; Bormann et al. 1974; Swanson et al., 1982).

RESEARCH NEEDS

1. Refine techniques of using soil-profile development and rock-weathering characteristics to assign ages to deposits. Emphasis should be placed on the use of rapidly developing soil properties to provide finer resolution and greater applicability to younger deposits.

2. Expand criteria for distinguishing those soil properties that are in equilibrium with existing climatic and vegetative conditions from those properties that are a relic of former conditions.

3. Develop criteria for distinguishing between sediment weathered in place after transport and sediment that was weathered primarily before its most recent movement.

4. Integrate more observations of water chemistry and soil-profile development into sediment- (or mass-) routing studies in forested drainage basins.

LITERATURE CITED

Birkeland, P. W.
1974. Pedology, weathering and geomorphological research. 285 p. Oxford Univ. Press, London.

Bormann, F. H., G. E. Likens, T. G. Siccama, R. S. Pierce, and J. S. Eaton.
1974. The export of nutrients and recovery of stable conditions following deforestation at Hubbard Brook. Ecol. Monogr. 44(1):255-277.

Burke, R. M., and P. W. Birkeland.
1979. Reevaluation of multiparameter relative dating techniques and their application to the glacial sequence along the eastern escarpment of the Sierra Nevada, California. Quat. Res. 11(1):21-51.

Cleaves, E. T., A. E. Godfrey, and O. P. Bricker.
1970. Geochemical balance of a small watershed and its geochemical implications. Geol. Soc. Am. Bull. 81(10):3015-3032.

Conacher, A. J., and J. B. Dalrymple.
1977. The nine unit landsurface model: an approval to pedogeomorphic research. Geoderma 18(1):1-154.

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

Janda, R. J.
1979. Geologic look at soils of the Douglas-fir region. In Forest soils of the Douglas-fir region, p. 75-77. P. E. Heilman, H. W. Anderson, and D. M. Baumgartner, eds. Wash. State Univ. Coop. Ext. Serv., Pullman.

Parsons, R. B.
1978. Soil-geomorphology relations in mountains of Oregon. USA. Geoderma 21(1):25-40.

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.

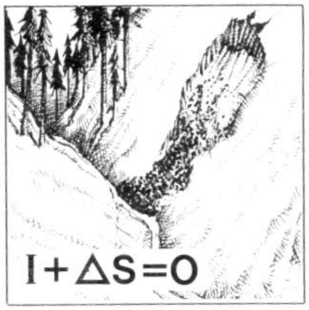
Ugolini, F., R. Minden, H. Dawson, and J. Zachara.
1977. An example of soil processes in the *Abies amabilis* zone of the central Cascades, Washington. Soil Sci. 124(5):291-302.

U.S. Department of Agriculture, Soil Survey Staff.
1960. Soil classification, a comprehensive system (7th approximation). 265 p. U.S. Gov. Print. Off., Washington, D.C.

U.S. Department of Agriculture, Soil Conservation Service.
1975. Soil taxonomy, a basic system of soil classification for making and interpreting soil surveys: Agriculture handbook 436. 754 p. U.S. Gov. Print. Off., Washington, D.C.

Yaalon, D. H.
1971. Soil forming processes in time and space. In Paleopedology, p. 29-39. D. H. Yaalon, ed. Internatl. Soc. Soil Sci. and Israel Univ. Press, Jerusalem.

Zinke, P. J., and W. L. Colwell.
1965. Some general relationships among California forest soils. In Forest-soils relationships in North America, p. 353-365. Second North Am. Soils Conf., Oreg. State Univ., 1963. Oreg. State Univ. Press, Corvallis.



Influence of Magnitude, Frequency, and Persistence of Various Types of Disturbance on Geomorphic Form and Process

Harvey M. Kelsey

The relative importance of events of various magnitude and frequency on geomorphic form and process is one of the earliest recognized basic questions of geomorphic theory, and this workshop discussion seemed to point out that the subject is not much better developed than it was when Wolman and Miller (1960) published "Magnitude and Frequency of Forces in Geomorphic Processes." We tried to focus the discussion on defining data limitations, asking the proper questions, narrowing the scope of the problem(s) to workable limits, identifying statistical sampling problems, and suggesting investigative techniques. As one participant pointed out, we are all victims of our own experience, and though our observations and backgrounds are the foundations for moving forward, they also can cause tunnel vision. We hope the range of participants to some extent removed these limitations.

The best way to summarize the discussion is to pose a set of questions and tentative statements that reflect the major concerns of the participants. Obviously, we reached no conclusions, but we did cover a number of important questions that should be confronted. We began by asking if we are interested in the effects of magnitude and frequency of events on (1) processes that sculpt landforms (shape the landscape) or (2) processes that mobilize the greatest amount of sediment out of a basin. In other words, are we to address the landform-evolution problem or the sediment-routing problem. To this end, the group posed the question, "Should we apply the problem of the effects of magnitude and frequency of events to different drainage-basin segments or sites?" For example, the following basin segments or sites can be studied:

1. Hillslope processes (e.g., debris-slide events, surface erosion),
2. "Fingertip" tributaries,
3. Main channels (higher order channels in a basin),
4. Points of output of a basin (gaging station at end of basin--the Wolman and Miller (1960) approach).

When dealing with the problem of the effects of magnitude and frequency on geomorphic process and form, must we define first whether we assume a landscape is in dynamic equilibrium (Hack 1960), whether a landscape evolves by exceeding critical thresholds (Schumm 1973, Bull 1979), or whether another model is more appropriate? Are different landscape models mutually exclusive, or are they pertinent in different time frames?

Thinking of the geomorphic event/climatic event relationship in terms of force and response is useful: climatic event = driving force; geomorphic event = response. Perhaps we need to apply magnitude-frequency analysis separately in different climatic regimes because the relationship between driving force and response varies for different climatic areas. Therefore, climatic regime as well as basin segment must be defined. Complicating the picture, significant climatic changes on the scale of 10^2-10^4 years will alter the magnitude-frequency character of geomorphic response. Perhaps a cutoff point exists in terms of time period beyond which pursuing magnitude-frequency analysis is not useful because of climatic changes.

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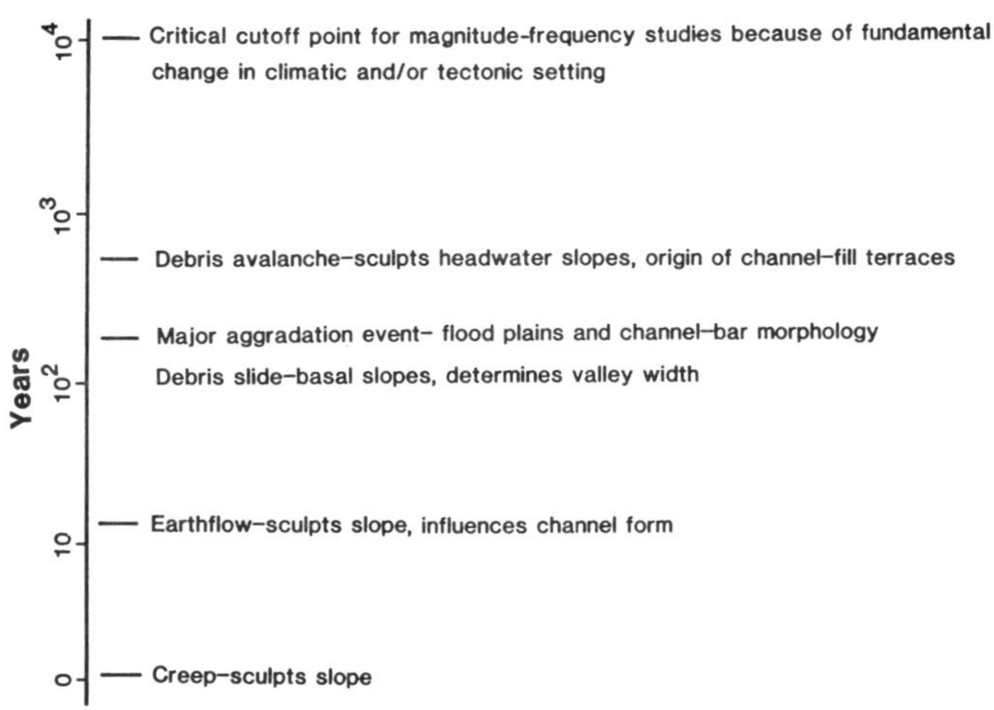
A basic problem of magnitude-frequency studies is the need to extrapolate for long time spans from a limited period of scientific observation. This problem raises the question, can space be substituted for time? Using the example of soil hollows and related debris avalanches (Dietrich and Dunne 1978; Lehre, this volume), frequency of failure and the rate of soil infilling of the hollows can be estimated by examining the depth of soil in hollows and the occurrence of debris avalanches over a broad area for a relatively short period of time. Alternatively, the estimate can be made by analyzing the filling rate and failure probability of one hollow over a long period. If space is substituted for time in studying magnitude-frequency relations of geomorphic processes and/or form, two criteria, in most cases mutually exclusive, should be met: (1) a homogeneous sample area (similar geology, tectonic setting, and climate) and (2) a large sample area. If we cannot take a look at a long record, an alternative is the deterministic approach of applying existing theory in an attempt to extrapolate from a limited sample population.

How should we deal with channel-storage compartments in space and time as compared to slope processes in space and time, and how can changes in these compartments best be monitored? Maybe we should first ask how related geomorphic features (headwater slopes, flood plains, footslopes, interfluves) form. Then the problem might be divided among different drainage-basin segments. For example, sediment is stored in major channels at sites with different residence times: active channel, channel down to depth of scour for major events, and the part of the channel that is storing alluvium for longer periods. Each channel-storage compartment is most influenced by a discharge of a different magnitude and frequency.

When dealing with the geomorphic importance of events of various magnitudes and frequencies, distinguishing whether the system (drainage basin, hillslope, higher order channel, or other geomorphic unit) is material-limited or energy-limited or transitional between the two is important. For example, debris avalanching on steep headwater slopes in resistant, competent rock types may be limited by availability of material if the rate of weathering is slow relative to the frequency of large storms that trigger debris avalanches. A contrasting example is bedload transport in a channel reach that has recently aggraded. Here, the system is limited only by transport capability because abundant sediment is available.

Is there a characteristic frequency for each process and is there a most important frequency for each type of landscape change? Perhaps attempting to characterize the most important frequency of activity for various geomorphic processes and events in a given area over a log time scale (fig. 1) would be instructive. The recovery of various components of geomorphic systems and ecosystems to severe disturbance, such as some of those depicted in figure 1, may occur on an exponential basis, so characteristic time scales might best be described in terms of half-life. Refilling of hollows after a debris slide, for example, may occur first rapidly and then progressively slower.

Figure 1.--Log time scale showing characteristic frequencies for different geomorphic processes.



When investigating effects of magnitude and frequency of events on process and form, we must ask whether the events are independent of each other. Magnitude-frequency analysis is most useful when independence of successive events can be assumed. Recurrence intervals cannot be used for prediction if the recurrence of one event is dependent on the last. Where records of a sufficient number of events are available, statistical techniques can aid analysis of relationships among successive geomorphic events. As Rice (this volume) and others point out, however, geomorphic change in steep lands is typically dominated by infrequent events, so a record of even 40 to 50 years is too brief for this type of analysis.

In magnitude-frequency analysis, we must distinguish between geomorphic events and climatic events. Frequency of geomorphic events and frequency of climatic events are often confused. One problem in magnitude-frequency analysis is that successive climatic events may be considered independent of one another, but successive geomorphic events may be highly related. The second of a pair of back-to-back major floods, for example, will operate on a different set of sediment-storage and streamside-vegetation conditions and therefore have different geomorphic consequences than the first flood. Independence of successive events is a basic assumption in the magnitude-frequency analysis of Wolman and Miller (1960).

Another way of addressing the same problem is to ask what is the length of memory or recovery rate of a system relative to the frequency of climatic events that trigger the geomorphic change. Expressed graphically, changes in a system may all originate from the same base level (case I, fig. 2) when recovery is fast relative to event frequency. Where events may occur so frequently that recovery from previous events is not complete (case II), the effect of any one event is determined in part by the magnitude and timing of past events. A third alternative (case III) is a combination of independent events and events that are influenced by previous disturbance.

What is human influence on magnitude and frequency of events? Land use may well be fundamentally changing the resilience and shape of the earth's surface. As a consequence, a driving force of the same magnitude may now trigger events of greater magnitude more frequently than before widespread ground disturbance. Is land use changing the character of geomorphic response to climatic events? If so, is land use likely to trigger geomorphic change in a nonlinear fashion? What is the relative importance of land use in influencing the frequency of high-magnitude as compared to low-magnitude events?

The discussion group sketched a summary figure showing some possible relationships of the magnitude and frequency of driving force and geomorphic response at different locations within a fluvial system (fig. 3). The different locations are expressed by position on a simplified longitudinal stream profile. Positions on the profile are delineated as being material-limited, energy-limited, or transitional, in terms of geomorphic process.

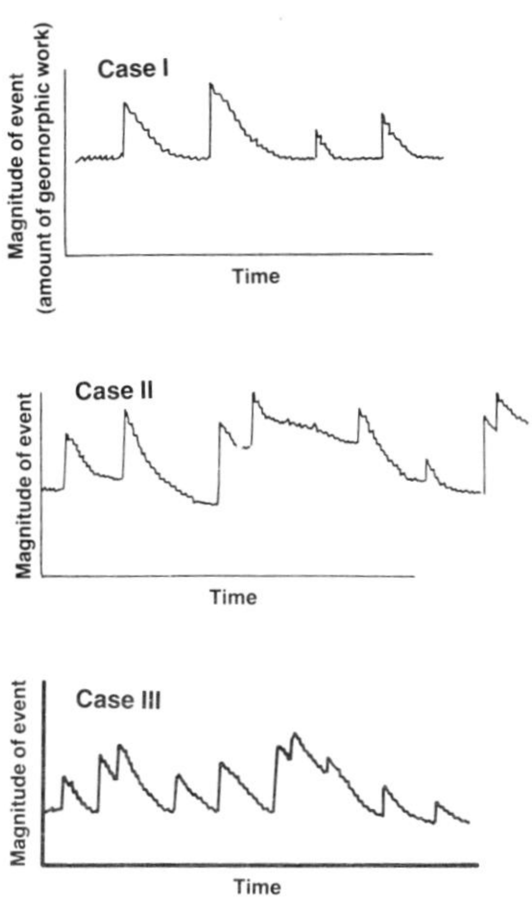


Figure 2.--Different case example of the effect of recovery rate in a geomorphic system on the magnitude of individual events in a series of geomorphic events. See text for discussion.

Participants generally concluded that magnitude-frequency analyses hold much promise for analyzing and contrasting geomorphic systems and determining minimal sampling periods for episodic events. Magnitude-frequency analysis has not been rigorously applied and examined since Wolman and Miller's original paper. Problems may arise in using this approach in systems that have strong memory (interdependence of successive events). This may be particularly true in steep, geomorphically active, forested terrain.

To better understand the magnitude-frequency characteristics of geomorphic systems, long-term data collection systems such as the Vigil Network and water resources monitoring program of the U.S. Geological Survey should be maintained. Such efforts provide both essential records of low-magnitude, frequent events, as well as some opportunity to document major, infrequent events. Further work using stratigraphic, dendrochronologic, and other methods should be employed to supplement efforts to monitor infrequent events directly, because records based on direct observation are far from adequate to answer important questions concerning magnitude-frequency characteristics of geomorphic change in geomorphically active areas.

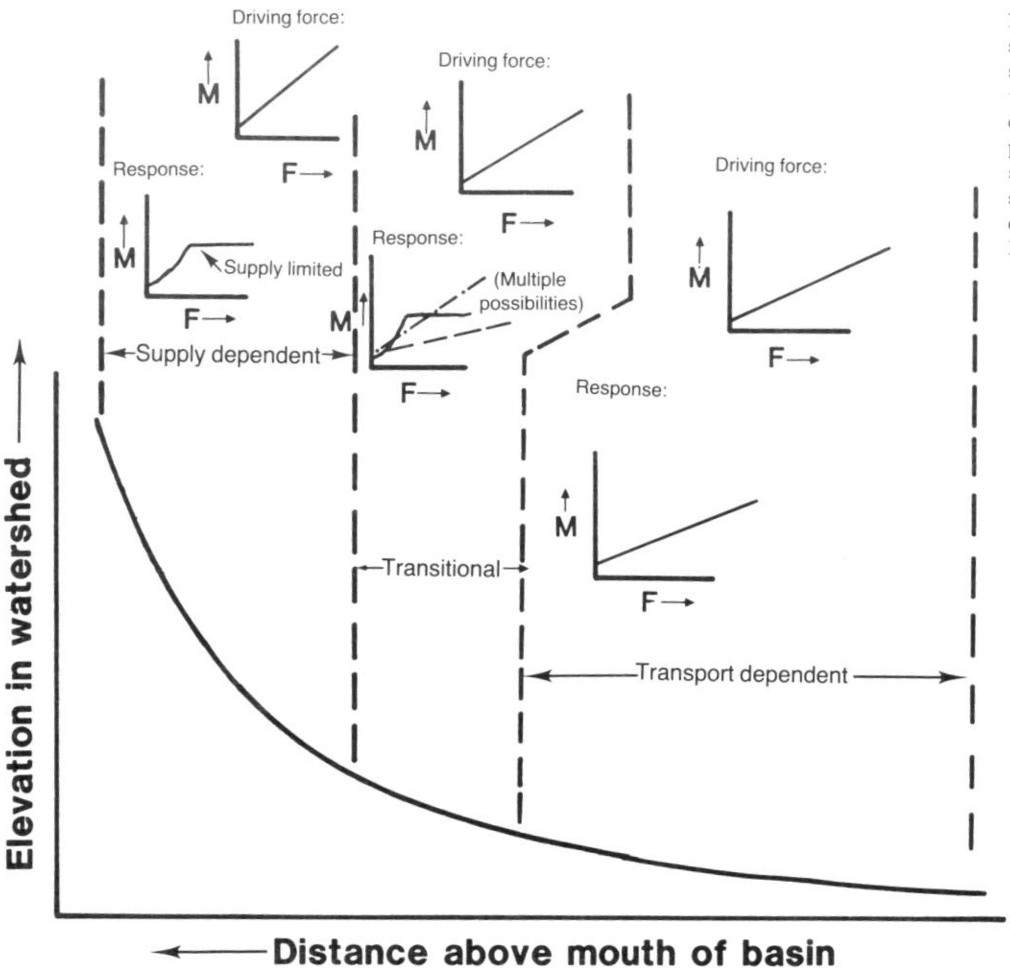


Figure 3.--Summary diagram showing possible relationships of magnitude of magnitude (m) and frequency (f) of driving force and geomorphic response at different sites within a fluvial system. Sites are depicted on a schematic version of a longitudinal stream profile.

LITERATURE CITED

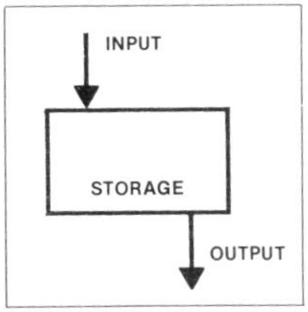
Bull, W. B.
1979. Threshold of critical power in streams. Geol. Soc. Am. Bull. Part I, 90:453-464.

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

Hack, J. T.
1960. Interpretation of erosional topography in humid temperate regions. Am. J. Sci. 258-A:80-97.

Schumm, S. A.
1973. Geomorphic thresholds and the complex response of drainage systems. In Fluvial geomorphology, p. 299-310. M. Morisawa, ed. Publications in Geomorphology, State Univ. N.Y., Binghamton.

Wolman, M. G., and J. P. Miller.
1960. Magnitude and frequency of forces in geomorphic processes. J. Geol. 68:54-74.



The Use of Flow Charts in Sediment Routing Analysis

Leslie M. Reid

THE USE OF FLOW CHARTS IN SEDIMENT ROUTING

Flow charts are a widely used means of diagramming relationships among transport processes and storage sites during analyses of sediment routing. Because they have taken so many different forms, however, it is very difficult to use published flow charts to compare geomorphic systems. Though they generally are constructed to achieve the same purpose, flow charts differ in the definition of the material being transported, type of transport processes involved, and the kinds of landscape elements considered to be sediment storage sites. In addition, the structure of flow charts may vary greatly. During this discussion, we attempted to probe the advantages and disadvantages of the different types of charts in hopes of arriving at a general form that would facilitate direct comparison of sediment budgets from different environments. Though a general form was not attained, many criteria necessary for a general flow chart, and for flow charts in general, were recognized.

Formally, a flow chart is a schematic representation of a series of operations. In the context of geomorphology, these operations are generally processes such as landslides, sheetwash, and solution transfer, which transport material from one storage site to another. Operations might also include processes such as chemical weathering and abrasion which change the character of the material, or information flow, which controls interactions between processes. Which operations are selected depends ultimately on the understanding of how the system operates, the amount of information available about the system being diagrammed, and the use to which the flow chart will be put.

Flow charts can be used to organize information, to describe a system, or to provide a framework for predictive calculations. At the least sophisticated level, a flow chart can be used to expose conceptual holes in our understanding of the relationships in a system. Such a chart need only be qualitative, though as the level of understanding increases and the recognized conceptual gaps occur at finer scales of resolution, the complexity of the flow chart and the need for quantification also increase. The same kinds of charts may be used to pinpoint critical, rate-controlling steps deserving more careful evaluation.

Flow charts are also useful for describing a specific system once data have been collected. Here, too, the level of sophistication may vary from a qualitative diagram of process interactions to a quantitative sediment budget. A corollary of this function is the use of flow charts to compare systems in different environments, and it is for this use that a need for some uniformity in structure and definitions becomes apparent. The most common means of comparison has been the single number index of sediment yield per unit area, both because it is the only parameter that has been widely reported and because it is relatively easy to measure. But important differences in processes and rates in different basins cannot be described by a single number. More useful would be a comparison based directly on sediment production and transport processes. A standardized flow-chart form would provide a means of organizing data for such a comparison.

At the most sophisticated level, a flow chart may allow prediction of the response of a basin to

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changes in environmental variables. The construction of such a model requires a quantitative understanding of all process rates and feedback controls operating in the system, as well as of the dependence of those rates on the driving variables.

The need for three types of flow charts can thus be recognized. Descriptive-qualitative charts organize information about relationships, and descriptive-quantitative charts attach average or instantaneous transfer rates to the qualitative framework. The third type of chart, a predictive-quantitative flow chart, includes measurements of feedback mechanisms and dependencies upon driving forces and thus allows calculation of the effects of changing conditions on the parameters of interest. Construction of such a chart, however, requires data not currently available for any natural geomorphic system.

The descriptive-qualitative flow chart has been the most widely used in geomorphology, largely because difficulties in measuring the rates of gain and loss of material to storage have prevented quantification of the charts. The traditional form of a descriptive flow chart is a series of boxes joined by arrows showing a temporal or spatial sequence. Boxes, however, can stand for many different items or concepts. Recent geomorphic flow charts have used boxes to represent (1) idealized locations as geometrically defined landscape units (e.g., units of a segmented landscape discussed by Simons et al., this volume), (2) geomorphically defined locations (swale--channel banks--channel bed of Lehre, this volume), (3) sediment deposits (soil--wedge--fan--channel bed, as described by Dietrich and Dunne 1978), (4) transport processes (creep--slump--debris slide--debris torrent shown by Swanson et al., 1982), and (5) material character (bedrock--saprolite--soil). The arrows then correspond to transport processes, weathering processes, or merely indicate sequential development.

Appropriate definitions of elements of a flow chart depend on the purpose of the flow chart. The limitations of each approach to flow-chart development must be recognized. Over long periods, for example, landforms and idealized locations may evolve from one kind to another (swales may fail, form channels, and gradually refill) and deposits may erode away or new kinds form. Storage elements, too, are difficult to evaluate, because storage and transport form a continuum that may be distinguished only by an arbitrary selection of a minimum residence time to define storage. Charts based on processes tend to be complex because most processes interact with others. Complexity is also introduced because a change in material character may be interpreted as an indirect expression of mass transfer; weathering may change bedrock to saprolite in situ, for example. In any case, care must be taken to avoid mixing different types of elements in one diagram, unless the elements are distinguished symbolically.

The more complete the flow chart, the more difficult it is to interpret, and the less useful it becomes as a graphical description of a system. Increasing complexity may be avoided by breaking the flow chart into several charts, each dealing with a separate subsystem. The subsystems may then be recombined by way of a master chart that treats each subsystem as a separate compartment. Each subsystem may represent an individual process, landform, or time segment.

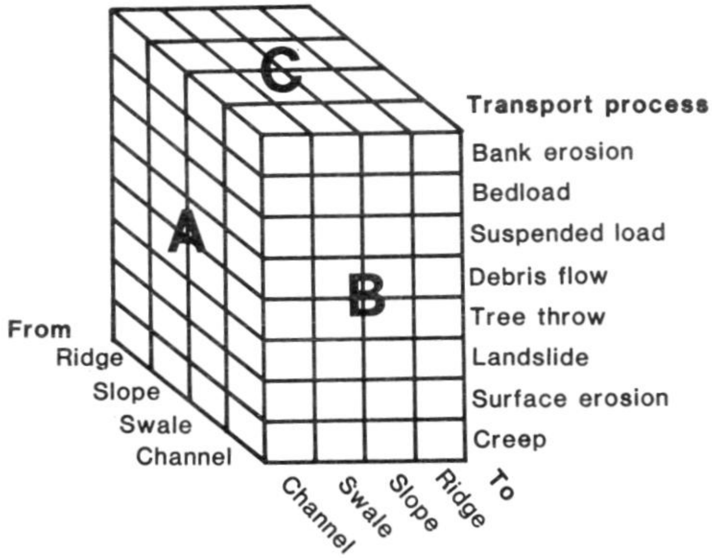
A similar approach may be used to handle the change of process type and rate as channel order increases. Debris torrents, for example, are an important transport process in steep, low-order basins. But as the channel gradient decreases with increasing basin order, torrents become less important and eventually disappear altogether. Such an effect cannot be shown on a simple chart, but if a separate flow chart is constructed for each order of channel, the effect becomes evident. These separate charts are then combined by way of a master chart that specifies the transfers between channels of different orders.

Several approaches have been used to show quantitative sediment budget data on flow charts. The relative importance of different elements or transport routes may be shown by the relative sizes of the element boxes or the thickness of interconnecting arrows, even if the actual values are not known.

If process rates and storage volumes have been measured, a quantitative, descriptive flow chart may be constructed. Rates of material transport can be shown by appending the actual transport rates to the flow chart. This method is useful in that it preserves the visual impact of the generalized chart and facilitates direct comparison of flow charts, but it becomes unwieldy as the complexity of the chart increases.

Highly complex systems can be quantitatively described using transitional matrices. With this method, locations (absolute or idealized), landforms, or storage elements are used as labels for columns and rows of a matrix (fig. 1). Sediment is considered to be transported from the location noted in a row to that noted in a column, and transport processes are differentiated by planes in the third dimension; mass transfers are filled in according to source, destination, and transport process. Sequence--an important aspect of flow charts and a critical factor in understanding a geomorphic system--is difficult to discern, but may be partially reconstructed by comparing inputs and outputs from different locations. Though visually obscure, the result makes the organization of information relatively easy and is useful in accounting; change in storage volume may be easily determined by comparing the totals in the rows and columns for each location. This operation also makes apparent the restrictions of this budget and any other descriptive-qualitative budget that is based on short-term measurements and does not deal explicitly with probabilistic variations in driving variables; the budget can reflect only what is happening at a specific time or the average result over a longer period. Otherwise, loss of material in transport to storage would imply infinite aggradation.

The third type of flow chart--the predictive-quantitative model--requires a quantitative understanding of process rates, driving variables,



A Destination (From swale)

	Channel	Swale	Slope	Ridge
Bank erosion	3	1	–	–
Bedload	17	1	–	–
Suspended load	30	.5	–	–
Debris flow	40	3	–	–
Tree throw	5	12	–	–
Landslide	12	4	–	–
Surface erosion	1	1	–	–
Creep	–	38	–	–

B Source (To swale)

	Ridge	Slope	Swale	Channel
Bank erosion	–	–	1	–
Bedload	–	–	1	–
Suspended load	–	–	.5	–
Debris flow	–	2	3	–
Tree throw	–	15	12	–
Landslide	3	19	4	–
Surface erosion	–	1	1	–
Creep	–	63	38	–

C Landslides

From \ To	Channel	Swale	Slope	Ridge
Ridge	–	3	1	–
Slope	28	19	23	–
Swale	12	4	–	–
Channel	–	–	–	–

Figure 1.--Hypothetical three-dimensional matrix for tabulation of material transfer rates between storage locations in a forested basin. Material is moved from source A to location B by transport process C. Hypothetical tabulations, in t/km^2 per year, are shown in 1A, 1B, and 1C. Change in storage at a location can be calculated by subtracting total output from total input at that location. In the example given, the rate of change of the volume of material stored in swales would be (total (1B) - total (1A)) = $-5\ t/km^2$ per year.

and information flow; each is incorporated into the chart to form a model of the real system. When values for the driving variables are specified, the chart can be used to predict the response of any part of the system or of the system as a whole. The complexity of a realistic model, and the necessity to handle driving variables probabilistically, usually requires the model to be programmed onto a computer, leaving only the framework to be represented as a chart.

After discussing flow charts in general, we attempted to develop a general qualitative chart to describe sediment movement through steep, forested landscapes. We failed. After several hours, we had just begun to resolve the difficulties and confusion that arose in settling on a particular type of flow chart, definitions of storage sites and transport processes, and rules for showing linkages among them. The makeup of the flow chart was found to depend heavily on the specific objectives for its use, even though a general form was desired.

The consensus among discussion participants was that flow charts are extremely useful in organizing thoughts about a sediment-routing system and in comparing different landscape units. Such comparisons, however, require that flow charts be developed using a common set of rules and definitions, so that differences in flow charts for different geomorphic systems reflect real geomorphic differences. Guidance for developing a useful flow-chart form could come from other fields, such as hydrology, nutrient and energy cycle modelling for ecosystems, and network theory. Developing common definitions of landscape units and processes requires extensive discussion and a knowledge of what is geomorphically important in a wide variety of environments.

LITERATURE CITED

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

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.

Summary: Sediment Budget and Routing Studies

Frederick J. Swanson, Richard J. Janda, and Thomas Dunne

ABSTRACT

Sediment budget and routing studies have been useful in dealing with a variety of basic and applied problems over a wide range of scales in time and space. Further research and application is needed to: define and quantify sediment storage; improve knowledge of mechanisms of sediment-transport processes; quantify frequency and magnitude of episodic processes; integrate biological factors into quantitative analysis of sediment budgets and routing; improve knowledge of effects of weathering on sediment routing; and mesh better the computer simulation of sediment routing with field studies of conditions in forested mountain land.

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INTRODUCTION

Studies of sediment budgets and routing have increased understanding of a broad spectrum of geomorphic and ecological problems (Jäckli 1957; Rapp 1960; Leopold et al. 1966; Caine 1976; Dietrich and Dunne 1978; Kelsey 1980; Swanson et al., 1982). By sediment budget, we mean the quantitative description of sediment movement through a single landscape unit; sediment routing is either the computation of sediment movement through a series of units or the more qualitative concepts of sediment movement through a drainage basin. Much early work centered on computing total denudation and assessing the relative importance of individual erosion processes. Other workers have applied sediment budget studies to practical problems associated with the effects of human activities on geomorphic processes and associated landform changes (Kelsey 1980, Reid 1981). Applications of sediment budget and routing analysis in Redwood Creek basin (California) (Kelsey et al. 1981), areas near Mount St. Helens (Washington), and elsewhere deal with a variety of management-related issues, including the persistence of high rates of sediment transport.

The basic ingredients of complete budgets are: identification of storage sites, transport processes, and linkages among them, and the quantification of storage volumes and rates of transport processes. Despite these common elements, papers on sediment budgets published here and elsewhere display marked differences in the time scales considered and the relative emphasis placed on storage and transfer processes. These differences reflect contrasts in objectives and site-specific conditions such as vegetation, land-use history, and dominant erosion process. Of particular concern at this workshop were the distinctive effects of forest vegetation on sediment storage and transport.

Although the utility of sediment budgets has been amply demonstrated, they continue to be little used. Increased application of the sediment-budget studies, even where data are seriously limited, would increase understanding of geomorphic and ecological systems--this volume was compiled to encourage such increased use. Here, we summarize some advantages of the sediment-budget approach over studies focused on individual processes or on sediment-yield data alone, six major themes that recur throughout the papers and workshop discussions, and new and continuing research needs.

RATIONALE FOR MAKING SEDIMENT BUDGET AND ROUTING STUDIES

Enthusiasm and need for quantification of erosion and sediment transport have led to many field measurements and computer simulation models of sediment production on hillslopes and along streams. Monitoring of drainage-basin sediment yield provides an integrated, "black box" view of how sediment output from basins responds to average conditions, fluctuations in weather, and management and other disturbances. Monitoring of hillslope erosion can provide more detailed information on processes than can measuring basin sediment yield alone. This approach, however, has often been weakened by lack of attention to sampling problems and by uncertainties of how particular mechanisms fit into the sequence of processes that transport sediment out of a drainage basin. Measurement of both sediment yield and rates of processes offers little information on the role of temporary storage or the linkage between transport processes.

Some of these problems are reduced by drawing up a conceptual sediment budget or sediment-routing scheme early in the investigation of sediment movement. This requires explicit recognition of how sediment is generated, transferred, and modified during its passage through drainage basins. The initial conceptual model may be only qualitative or approximately quantitative, but it must be based on field work. Dietrich and Dunne (1978), Dietrich et al. (this volume), Lehre (this volume), and Kelsey (this volume) have all stressed the need for careful classification of transport processes and storage sites in the landscape under investigation. This preliminary analysis draws attention to the most important processes and aids the design of appropriate measurement strategies. No single approach to the definition of a sediment budget or routing scheme exists because of the variety of processes, materials, and disturbing factors even in the restricted setting of drainage basins covered by temperate forests.

Field observations of sediment transfer and storage also suggest how the accounting of sediment can be carried out. The investigator is forced to consider how sediment moves between sites and whether errors can result from adding together contributions of sediment to a channel by processes that act in series. An example occurs where soil creep merely supplies sediment to sites of landsliding which then conveys it to a channel. Dietrich and Dunne (1978) have pointed out that adding these two inputs would be double accounting and would overestimate the sediment flux to the channel.

Processes should be classified to help define the temporal and spatial requirements of a sampling scheme. Do the processes operate over extensive areas or in a limited number of restricted sites? Which processes operate persistently and which are episodic? Whether a process is viewed as persistent or episodic in part reflects the time reference of the budget. Most geomorphic measurements have been carried out over too short a time and at too few sites for adequate definition of long-term average sediment yields, their sources, controls, and response to disturbance. Field work necessary for construction of a sediment budget or routing scheme focuses attention on the need for spatial stratification of measurement sites and for lengthening the period of record by means of dendrochronological and other techniques (Brown and Brubaker, this volume).

A sediment budget or routing procedure and attendant assumptions may apply to a particular time scale only. The procedure of Dietrich and Dunne (1978), for example, is founded in the long-term view of sedimentary petrology and steady-state geomorphology. The steady-state

Table 1--Channel storage in relation to mean annual sediment discharge. Definition of channel storage varies from study to study, but generally includes readily mobilizable material. Note that (storage/annual discharge) is a measure of residence time of sediment in storage in years

Stream (author)	Drainage area	Storage	Annual sediment discharge	Storage - annual discharge
	Square kilometers			
Watershed 10, OR (Swanson et al., 1982)	0.10	510 t (subject to debris torrent)	6.6 t Total particulate transport including episodic debris torrents	77
Average of seven streams in Idaho batholith (Megahan, this volume)	1.27	171 m^3 "behind obstructions"	11 m^3 debris basin accumulation	15
Rock Creek, OR (Dietrich and Dunne 1978)	16.2			
First-order tributaries				1/19 yr
Third-order tributaries				1/231 yr
Main channel, active sediment				1/619 yr
Big Beef Creek, WA (Madej, this volume)	38	49,200 t "active, to depth of scour"	1,000 t bedload	50
Redwood Creek, CA (M. A. Madej, Redwood National Park, personal communication)	720	7,200,000 main channel only	1,860,000 total sediment load	3.9
Otamarahu River, New Zealand (Mosley 1977, cited in Pearce and O'Loughlin 1978)				35-40

1/Residence time for particle entering channel of this order at midpoint of channel segment.

assumption is invalid for other time scales, however. Nonsteady-state behavior occurs where significant change in storage takes place, whether as a result of human activities or major storms (which affect geomorphic systems on the scale of years and decades), climatic change (on the 10^3 year scale), or drainage development (on the scale of 10^3 years and longer). Swanson and Fredriksen (this volume), for example, stress the value of not assuming steady-state conditions when examining sediment routing on the time scale of vegetation disturbance and recovery.

The sediment budget and routing approach led Dietrich and Dunne (1978) to consider the relation of weathering to transport of soil entering channels and to grain-size distribution of sediments transported down stream channels. The grain-size distribution of stream sediments is important because it influences channel form, fish-spawning habitat, sorption of nutrients and other chemicals, and rates at which large, short-lived influxes of sediment can be flushed from a system.

MAJOR CONCLUSIONS AND RESEARCH NEEDS

Sediment Storage

Sediment storage is an essential but poorly understood part of the geomorphic system. Geomorphologists have traditionally emphasized transport processes. Storage of sediment, however, is an equally important aspect of long-term movement of material through drainage basins. Total stored sediment, duration of storage at various sites, changes in volume of material stored, and changes in the physical properties of materials while in storage have important implications in analysis of sediment routing.

Quantitative data on sediment storage in channels is meager, but more abundant than information on storage of material on hillslopes. Available quantitative studies of sediment storage in channels in forested areas (table 1) define and measure storage in different ways. Nonetheless, these data indicate that the volume of temporarily stored alluvium is commonly more than 10 times larger than the average annual export of total particulate sediment. Mean residence times are on the order of decades and centuries. Thus, moderate changes in storage can cause major changes in sediment yield even if sediment supply from hillslopes remains constant.

Rough estimates of mean residence time of soil on hillslopes are several orders of magnitude greater. For a soil loss rate of 100 t/km^2 per year (on the low side of typical rates for mountain land), the mean residence time of a 1-m soil profile is 10,000 years. This is somewhat less than the residence time of soil on hillslopes computed by Dietrich and Dunne (1978) and Kelsey (this volume) for very different terrains.

Changes in sediment storage can drastically affect interpretations of erosional conditions within a drainage basin based on sediment-yield data alone (Janda 1978). The timing and magnitude of erosion of soil and its ultimate delivery as sediment to a downstream point may be very different. Kelsey (1980), Trimble (1981), and others have demonstrated that storage in channels and flood plains may delay and subdue the peak of downstream delivery of sediment introduced from hillslope sources. Conversely, increased peak flows because of altered hillslope hydrology can result in increased erosion and downstream transport of stored sediment without increased hillslope erosion (Park 1977). Management impacts on hydrology or sediment availability can therefore have cumulative effects on sediment routing downstream, an issue of growing concern in many areas of forest-land management.

Residence time of sediment in storage determines the opportunity for stabilization by rooting of vegetation on the scale of years and decades and for changes in size distribution of material by weathering over centuries and millenia. Weathering of deposits in gravel bars, flood plains, and other storage sites facilitates breakdown of sediment, thus changing the relative importance of transport as dissolved, suspended, and bedload (Bradley 1970, Dietrich and Dunne 1978). Effects of weathering are particularly important in tectonically active areas, such as the Pacific Rim--much of which is underlain by mechanically weak rocks. On the time scale of significant weathering, geomorphologists can learn much through application of the techniques developed by Quaternary stratigraphers and sedimentologists (Birkeland 1974, Tonkin et al. 1981).

Erosion Processes--Mechanisms and Linkages

The mechanics of erosion processes and their controls are not well understood. These problems limit efforts to model movement of sediment through drainage basins. Several examples of these limitations sparked vigorous discussion at the workshop. One example is the widespread application of infinite slope assumptions to stability analysis of shallow debris slides, despite knowledge that many sites prone to debris sliding may not be well represented in hydrologic and other respects by infinite slope assumptions (Pierson 1980). This is true for "hollow" (Dietrich and Dunne 1978) and "swale" (Lehre, this volume) types of failure sites, which appear to be essentially the same and are hereafter referred to as hollow/swale sites.

Mechanics of surface-erosion processes in steep forest land in areas with low to moderate intensities of rainfall are also poorly understood, as evidenced by continued interest in applying to steeplands the Universal Soil Loss Equation (USLE) developed empirically by Wischmeier and Smith (1955) in lowland environments. The USLE was developed to estimate surface erosion by rainsplash and sheetwash on gradients of less than 20 percent. Overland flow is rare in forested landscapes. The surface-erosion processes that do operate, such as dry ravel and splash, may have very different relationships between transfer rate and slope length, rainfall characteristics, soil characteristics, and gradient than those described by the USLE.

Analysis of individual erosion processes in their overall geomorphic context is also critical. The rate or frequency of one process may be closely linked to the rates of other processes. Long-term erosion by debris slides, for example, may be limited by recharge of slide-prone hollow/swale sites by soil creep, root throw, and other processes. Thus measuring soil creep into hollow/swale sites would help in estimating rates of filling of slide-prone portions of the landscape. These data could then be used to judge effects of management practices on refilling rates as well as on initiation of debris slides. Similarly, progressive downslope movement of streamside earthflows may temporarily buttress the toe of the slope, impeding further movement. Subsequent stream erosion of the earthflow toe can remove support and accelerate movement, hence interpretations of earthflow movement rates should consider recent stream history.

Sediment transported as suspended load or bedload has been the subject of extensive, sophisticated analyses by hydraulic and civil engineers. Application of their equations in sediment-routing studies in steep forest lands is difficult, however, because sediment transport there is commonly limited by the rate of sediment supply from hillslopes, fans, and the streambed below an armor layer. These equations are not well suited for dealing with the coarse, poorly sorted sediment and large woody debris that forms the complex roughness elements typical of forested mountain streams. Channel form and pattern in these environments may be controlled by vegetation, bedrock, and hillslope mass movements rather than channel hydraulics and sediment properties that predominate in lowland streams.

Problems Posed by Episodic Processes

Many sediment-transport processes, such as creep and bedload transport in sand channels, are persistently active, although at widely varying rates. Debris slides and other processes are episodically active for only short periods. The distinction between persistent and episodic processes are muted in dealing with periods much longer than the time between episodes of activity. Geomorphologists have had a long-standing interest in the importance of episodic processes in long-term transfer of material and landscape sculpture (e.g., Wolman and Miller 1960). Episodic processes are generally considered to be the dominant mode of sediment transport in steep forest land, but no concensus exists on how to deal with them quantitatively in studies of sediment budgets and routing. Estimating sediment transport by episodic processes is an essential but difficult part of computing a sediment budget. For example, alternative approaches to estimating debris-slide erosion have been proposed. Dietrich and Dunne (1978) and Dietrich et al. (this volume) attempt to quantify frequency of failure at a particular site; others (Swanson et al., in press; Lehre, this volume) apply a geographically broader inventory method of computing erosion per unit area and time.

An important difficulty in dealing with episodic events is evaluating the interaction of successive major disturbances within a drainage basin (Bevan 1981; Kelsey, discussion group report, this volume). The approach to magnitude-frequency analysis proposed by Wolman and Miller (1960) assumes independence of successive events, but studies in a variety of areas indicate that major floods may change the quantity of sediment available for transport (Brown and Ritter 1971) and channel conditions (Ritter 1974, Baker 1977) encountered by subsequent events.

"Geomorphic recovery" after major disturbances is an essential part of judging effects of episodic events on sediment yield, but it is a concept interpreted in many ways. Wolman and Gerson (1980) discuss channel recovery in terms of return to predisturbance geometry, but judge recovery of landslide scars on the basis of degree of revegetation. Geomorphic recovery from a sediment-routing standpoint could be viewed as the refilling of storage sites and their readiness to fail again. The rate of such recovery for landslide scars varies greatly depending on the scale of the feature. Massive slope failures of essentially entire, first-order drainage basins (Kelsey, this volume) recover by rock weathering and soil formation. A much smaller proportion of a drainage basin fails in hollow/swale sites. Hollow/swale sites are recharged by transport of colluvium from adjacent areas, so their recovery is likely to be more rapid than that of the more massive failures where recharge is limited by weathering rate. Over several episodes of sliding, however, weathering must be the rate-limiting process in both systems.

Clearly the long-term significance of episodic events in sediment budgets and routing systems is difficult to quantify because the meager record of past events is dominated by the most recent one or two and the longer term sequence and timing of past events is important. Here again, the geomorphologist may have to rely on the tools of the dendrochronologist and Quaternary stratigrapher to place limits on the timing of past events.

Biotic Factors in Sediment Routing

Biological parts of landscape systems contribute important components of sediment, act as agents of sediment transfer, form sediment-storage structures, and record forest and geomorphic history. Biological influences on geomorphology are particularly well developed in forest vegetation--reflecting, in part, the massive size and relatively slow decomposition rate of woody material in many forest environments.

The role of organic matter as a soil component is better understood than its role as sediment. Organic matter in soil is an essential part of both nutrient cycling and mineral weathering, which strongly influences soil stability. Sediment-transport studies by hydrologists and geomorphologists typically disregard the importance of organic matter in sediment both in deposits and in transit, but ecological research on drainage basins has emphasized the importance of streams in exporting organic matter from ecosystems (Arnett 1978). Organic matter may comprise a large proportion of sediment in transport, thus potentially complicating sediment sampling and confounding interpretation of the data (Arnett 1978). Sedimentologists have long been interested in the alteration of sediment characteristics during transport through a drainage basin, but interest among geomorphologists in extending these concepts to soil-sediment relations is rather new (Dietrich and Dunne 1978). Aquatic ecologists are becoming increasingly interested in the parallel issue of variation in the quantity and type of organic matter transported or temporarily stored throughout a drainage network (Naiman and Sedell 1979, 1980). Interactions between dissolved and fine particulate organic and inorganic matter (Jackson et al. 1978) present problems for sampling, distinguishing, and interpreting dissolved and suspended sediment yield from a sediment-routing standpoint. These interactions affect the fate and persistence of pollutants in ecosystems.

Plants and animals also affect soil and sediment movement and temporary storage in a variety of ways, many of which are described in papers in this volume. Effects of fauna and flora on individual erosion processes have been quantified in some detail, ranging from Darwin's (1881) work on erosion by earthworms to recent studies of tree-root effects on the potential for shallow mass movements (O'Loughlin 1974, Ziemer 1981). Where vegetation decreases the effectiveness of sediment-transport processes, it enhances sediment storage and increases the residence time of sediment by dissipating the energy of sediment-transporting media and by holding sediment in place. Large woody debris in streams and on hillslopes and tree roots are examples of biological materials that retain sediment at temporary storage sites.

The multiple, cumulative effects of vegetation on sediment routing through small (less than 100 ha) drainage basins have been demonstrated by studies in both forested and disturbed conditions (Bormann et al. 1969, 1974; Fredriksen 1970; Swanson et al., 1982; and others).

Dendrochronology may also be an integral part of sediment-routing studies by placing limits on the date of an event, rate of a process, and residence time of material in storage (Alestalo 1971; Schroder 1978; Hupp and Sigafoos, this volume; Brown, this volume; and others). Furthermore, general aspects of vegetation history can be interpreted by dendrochronologic analysis of events, such as wind storms and wildfire, and by palynological analysis of vegetation response to change in climate. This knowledge affects extrapolation of sediment-budget information to periods longer than those covered by direct observational data.

Interest in these long-term interactions between biotic and geomorphic systems goes beyond basic, academic concerns, and forms a foundation for interpreting and predicting impacts of management activities on forest ecosystems and landscapes. The successional development of ecosystems after disturbance determines the pace of recovery of vegetative control of sediment movement and storage. Likens and Bilby (this volume), Swanson et al. (1982), and others have argued that analysis of geomorphic systems should be placed in the context of vegetative succession and disturbance history.

Weathering

Study of weathering and its effect on availability of plant nutrients, soil development, and soil stratigraphy is advanced compared with knowledge of weathering as a regulator of sediment routing. Weathering affects the availability of material for transport and the types and rates of transport processes operating in an area (Dietrich and Dunne 1978). Geological material enters the sediment-routing system by weathering of bedrock, which makes it available for transport.

Dietrich and Dunne (1978) suggest that weathering is a critical rate-limiting factor in the long-term movement of sediment in many mountain environments. This may be particularly true in steeplands with shallow soils over competent bedrock; examples are the Oregon Coast Ranges (Dietrich and Dunne 1978) and the mountains of Hawaii (Scott and Street 1976). Weathering is less crucial in determining the availability of erodible material where primary sediment sources are deep soils, unconsolidated sediments, or tectonically shattered rock, such as recently glaciated terrain (Madej, this volume) and the tectonically active California Coast Ranges (Kelsey, this volume).

Once material is available for transport, the types and rates of movement depend strongly on soil depth and on physical properties of the material determined by parameters such as grain-size distribution and clay mineralogy (Dietrich and Dunne 1978). Dietrich et al. (this volume) argue that change in soil depth with refilling of "hollows" results in increasing susceptibility of the site to failure by additional debris sliding. Weathering processes and their interaction with biota alter soil cohesion, bulk density, and mechanical properties, consequently controlling the rates of virtually all hillslope transport processes.

Weathering changes material while it is in temporary storage at various sites within a drainage basin. The rates and types of weathering reactions may vary from one storage site to another, depending on characteristics of the local weathering environment, such as hydrology, temperature fluctuations, pH, and oxidation-reduction conditions. Glancy (1971) and others, for example, have noted the break up of pebbles of sedimentary rock on gravel-bar surfaces over a period of months. They suggest that this is a bar-surface phenomenon only, so stones buried at shallow depths within the bar would not undergo this partial conversion from bedrock to suspended-load particle sizes.

The residence time of material in some storage sites stretches to the time scale of significant weathering and soil-profile development. Workshop participants (Harden et al., this volume) argued that soil stratigraphic techniques could profitably be used to determine residence time of storage sites and sometimes to set limits on the time since the last mass movement at a site.

These few examples indicate that weathering studies have an important, but little used, place in sediment-routing research.

Modeling of Sediment Budgets and Routing

Computer simulation of sediment routing holds great promise for aiding compilation of sediment budgets and for simulating sediment-routing systems in ways useful for predicting system change in response to disturbances. A simulation model provides a rigorous statement of a sediment-routing system and highlights the kind and quality of field data needed for prediction. Existing simulation models require much more development before they meet this promise, however.

Two types of models were discussed at the workshop: a model by Simons et al. based on physical processes and Rice's Monte Carlo simulation of sediment production in response to a long sequence of fires and rainstorms. Each approach has its benefits and limitations.

Limitations of the model described by Simons et al. and of similar models developed for agricultural lands include: lack of treatment of mass wasting processes that can dominate sediment transport in steep land; uncertainty about the accuracy of some components of the model, such as the use of equations developed for sediment transport in deep stream channels to estimate transport by sheetflow; and the need for calibration of the model against a set of field data to obtain several parameters of the equations. Calibrations include such physically ill-defined concepts as soil "detachment coefficients" for rainfall and overland flow. Finding a set of coefficients that produce a good fit between predicted and observed water and sediment discharge does not necessarily lead to understanding of what is actually happening in the landscape, or even of where most of the sediment originates in a heterogenous landscape. Nor do such fits promote confidence in predictions of the consequences of some disturbance by climate or land use. More field experiments need to be conducted and generalized so that model parameters can be estimated a priori and tested against a few measured outputs. Nevertheless, information organized in such models is useful for developing other models and conducting field experiments to refine them.

Rice proposes the application of Monte Carlo simulation to describe the response of an erosion-sedimentation system to random meteorological events that drive the sequence of fires and rainstorms and interactions among them. He questions the utility of process-based mathematical models because of the need for calibration. Particularly in a region where the sediment budget is strongly affected by random phenomena that vary greatly from year to year, the model should be calibrated against a large number of events covering the most important combinations of parameter values. Rice also points out, however, that in addition to frequently discussed difficulties with the structure of geophysical data, empirical-statistical models suffer the same drawback as process models; the instrumental record is unlikely to contain sufficient important events to include adequate combinations of the most effective factors.

The refinement of studies of sediment budgets requires a combination of: field monitoring of processes coupled with measurements of the controlling variables, so that physically based process models can be developed; field experiments under controlled conditions to extend the range of observations on which the process models are founded; and the development of deterministic models of processes and their linkages. These models can then be used in Monte Carlo simulations, as suggested by Rice. A precedent in hydrology is the recent work of Freeze (1980) on runoff processes.

CONCLUSIONS

In this workshop, we took an interdisciplinary look at the state of knowledge on development and use of sediment budgets and routing studies for forest drainage basins and identified important directions for future research. Most analyses have considered channels as the major storage sites and budgets have included hillslope processes, changes in channel storage, and outflow by fluvial processes. Temporal scales range from one year to millenia, spatial scales from less than a hectare to tens of thousands of square kilometers. Objectives of current studies range from purely basic questions of how geologic materials move through drainage basins to analyses of impact of management practices on sedimentation and a variety of resources.

Central points identified and further research needs are:

- Sediment storage is an essential but poorly understood and poorly quantified component of sediment budgets or routing analyses.

- Our ability to conduct field studies and to develop computer simulation models of sediment transport processes is limited by our knowledge of mechanisms of transport and the geomorphic context in which they operate.

- Episodic processes dominate sediment transport in many steep terrains, but theory and quantification of these processes are not well developed, particularly the interactions between successive events.

- Biota play a variety of essential roles in the production, transport, and storage of sediment, but knowledge of biological functions is poorly integrated into quantitative analysis of sediment budgets and routing.

- Weathering affects the availability and properties of sediment, but because significant weathering commonly occurs over long periods relative to traditional studies of processes, weathering has been little studied or used in sediment budget and routing studies.

- Computer simulation modeling is useful for predicting system behavior and for integrating concepts, process mechanisms, and field data. Modeling efforts, however, have not yet dealt with the types of storage sites and erosion processes that dominate in many forested mountain lands.

Advance of knowledge in each of these areas would be facilitated by improving theoretical analyses of processes, accumulation of long-term data sets, and more standardization of procedures and terminology. Economic considerations, in part, are leading to declining support by science managers for collecting long-term data sets, although scientists at the workshop were unanimous in their support of the need for such records. Computer simulation modeling, although a useful way of maximizing the value of field data, does not by itself provide useful surrogate records in geomorphic systems where infrequent events dominate sediment transport and where knowledge of interactions between successive events is weak.

Standardization of procedures and terminology in field and modeling efforts would facilitate future efforts to compare and contrast budgets and routing in diverse geomorphic systems. Efforts to standardize, however, must be tempered by the need to express adequately the sediment routing characteristics of particular terrain, climate, and vegetation types. The inability of the discussion group on use of flow charts in sediment routing studies to develop a single flow chart common to many landscapes reflects the difficulty of balancing details of local knowledge with the general need of achieving a basis for comparing diverse systems.

Use of sediment budget and routing analysis of drainage basins is in its infancy. Continued application of these methods in a variety of environments in studies with diverse objectives attests to increasing recognition of the value of sediment budgets and routing studies.

LITERATURE CITED

Alestalo, J.
1971. Dendrochronological interpretation of geomorphic processes. Fennia 105:1-140.

Arnett, R. R.
1978. Regional disparities in the denudation rate of organic sediments. Z. Geomorphol. Suppl. 29:169-179.

Baker, V. R.
1977. Stream-channel response to floods, with examples from central Texas. Geol. Soc. Am. Bull. 88:1057-1071.

Bevan, K.
1981. The effect of ordering on the geomorphic effectiveness of hydrologic events. In Erosion and sediment transport in Pacific Rim steeplands. Christchurch, New Zealand, January 1981. Int. Assoc. Hydrol. Sci. Publ. 132, p. 510-526.

Birkeland, P. W.
1974. Pedology, weathering, and geomorphological research. Oxford Univ. Press, New York. 285 p.

Bormann, F. H., G. E. Likens, and J. S. Eaton.
1969. Biotic regulation of particulate and solution losses from a forest ecosystem. Bioscience 19:600-610.

Bormann, F. H., G. E. Likens, T. G. Siccama, R. S. Pierce, and J. S. Eaton.
1974. The export of nutrients and recovery of stable conditions following deforestation at Hubbard Brook. Ecol. Monogr. 44:255-277.

Bradley, W. C.
1970. Effect of weathering on abrasion of granitic gravel, Colorado River (Texas). Geol. Soc. Am. Bull. 81:61-80.

Brown, W. M., and J. R. Ritter.
1971. Sediment transport and turbidity in the Eel River Basin, California. U.S. Geol. Surv. Water-Supply Pap. 1986. 71 p.

Caine, N.
1976. A uniform measure of subaerial erosion. Geol. Soc. Am. Bull. 87:137-140.

Darwin, C.
1881. The formation of vegetable mould through the action of worms with observations on their habits. John Murray, Edinburgh. 362 p.

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

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, Oreg.

Freeze, R. A.
1980. A stochastic-conceptual analysis of rainfall-runoff processes on a hillslope. Water Resour. Res. 16(2):391-408.

Glancy, P. A.
1971. Sediment transport by streams in the Chehalis River basin, Washington, Oct. 1961 to Sept. 1965. U.S. Geol. Surv. Water-Supply Pap. 1798-H. 53 p.

Jäckli, H.
1957. Gegenwartsgeologie des bünderischen Rheingebietes: Ein Beitrag zur exogenen Dynamik Alpiner Gebirgs-landschaften; Beiträge zur Geologie der Schweiz, Geotechnische Ser., Leiferung 36, Bern. 126 p.

Jackson, K. S., I. R. Jonasson, and G. B. Skippen.
1978. The nature of metals-sediment-water interactions in freshwater bodies, with emphasis on the role of organic matter. Earth Sci. Rev. 14:97-146.

Janda, R. J.
1978. Summary of watershed conditions in the vicinity of Redwood National Park, Calif. U.S. Geol. Surv. Open-File Rep. 78-25. 82 p.

Kelsey, H. M.
1980. A sediment budget and an analysis of geomorphic process in the Van Duzen River basin, north coastal California, 1941-1975. Geol. Soc. Am. Bull., Pt. 2, April, 1980. 98 p.

Kelsey, H., M. A. Madej, J. Pitlick, M. Coghlan, D. Best, R. Belding, and P. Stroud.
1981. Sediment sources and sediment transport in the Redwood Creek Basin: A progress report. Redwood Nat. Park, Arcata, Calif., Tech. Rep. 3. 114 p.

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

Naiman, R. J., and J. R. Sedell.
1979. Characterization of particulate organic matter transported by some Cascade Mountain streams. J. Fish. Res. Board Can. 36:17-31.

Naiman, R. J., and J. R. Sedell.
1980. Benthic organic matter as a function of stream order in Oregon. Arch. Hydrobiol. 87(4):404-422.

O'Loughlin, C. L.
1974. A study of tree root strength deterioration following clearfelling. Can. J. For. Res. 4:107-113.

Park, C. C.
1977. Man-induced changes in stream channel capacity. In River channel changes. K. J. Gregory, ed. John Wiley and Sons, New York. p. 121-144.

Pearce, A. J., and C. L. O'Loughlin.
1978. Hydrologic and geomorphic processes on forested lands - a selective overview. In Proceedings, conference on erosion assessment and control in New Zealand, August 1978, p. 62-76. Christchurch, N.Z.

Pierson, T. C.
1980. Piezometric response to rainstorms in forested hillslope drainage basins. J. Hydrol. (N.Z.) 19(1):1-10.

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

Reid, L. M.
1981. Sediment production from gravel-surfaced forest roads, Clearwater Basin, Washington. Final Rep. FRI-UW-8108, 247 p. Fish. Res. Inst., Univ. Wash., Seattle.

Ritter, J. R.
1974. The effects of the hurricane Agnes flood on channel geometry and sediment discharge of selected streams in the Sesquehanna River basin, Pennsylvania. J. Res. U.S. Geol. Surv. 2(6):753-761.

Schroder, J. F.
1978. Dendrogeomorphological analysis of mass movement on Table Cliffs Plateau, Utah. Quat. Res. 9:168-185.

Scott, G. A. J., and J. M. Street.
1976. The role of chemical weathering in the formation of Hawaiian amphitheatre-headed valleys. Z. Geomorphol. Suppl. 20:171-189.

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.

Tonkin, P. J., J. B. J. Harrison, I. E. Whitehouse, and A. S. Campbell.
1981. Methods for assessing late Pleistocene and Holocene erosion history in glaciated mountain drainage basins. In Erosion and sediment transport in Pacific Rim steeplands. Christchurch, New Zealand. Int. Assoc. Hydrol. Sci. Publ. 132, p. 527-543.

Trimble, S. W.
1981. Changes in sediment storage in the Coon Creek Basin, Driftless Area, Wisconsin, 1853 to 1975. Science 214(4517):181-183.

Wischmeier, W. H., and D. D. Smith.
1955. Predicting rainfall-erosion losses from crop east of the Rocky Mountains. USDA Agric. Res. Serv. Agric. Handb. 282.

Wolman, M. G., and R. Gerson.
1980. Relative scales of time and effectiveness of climate in watershed geomorphology. Earth Surf. Proc. 3:189-208.

Wolman, M. G., and J. P. Miller.
1960. Magnitude and frequency of forces in geomorphic processes. J. Geol. 68(1):54-74.

Ziemer, R. R.
1981. Roots and the stability of forested slopes. In Erosion and sediment transport in Pacific Rim Steeplands. Christchurch, New Zealand. Int. Assoc. Hydrol. Sci. Publ. 132, p. 343-361.