Soil, Vegetation and Watershed Management of the Douglas-Fir Region

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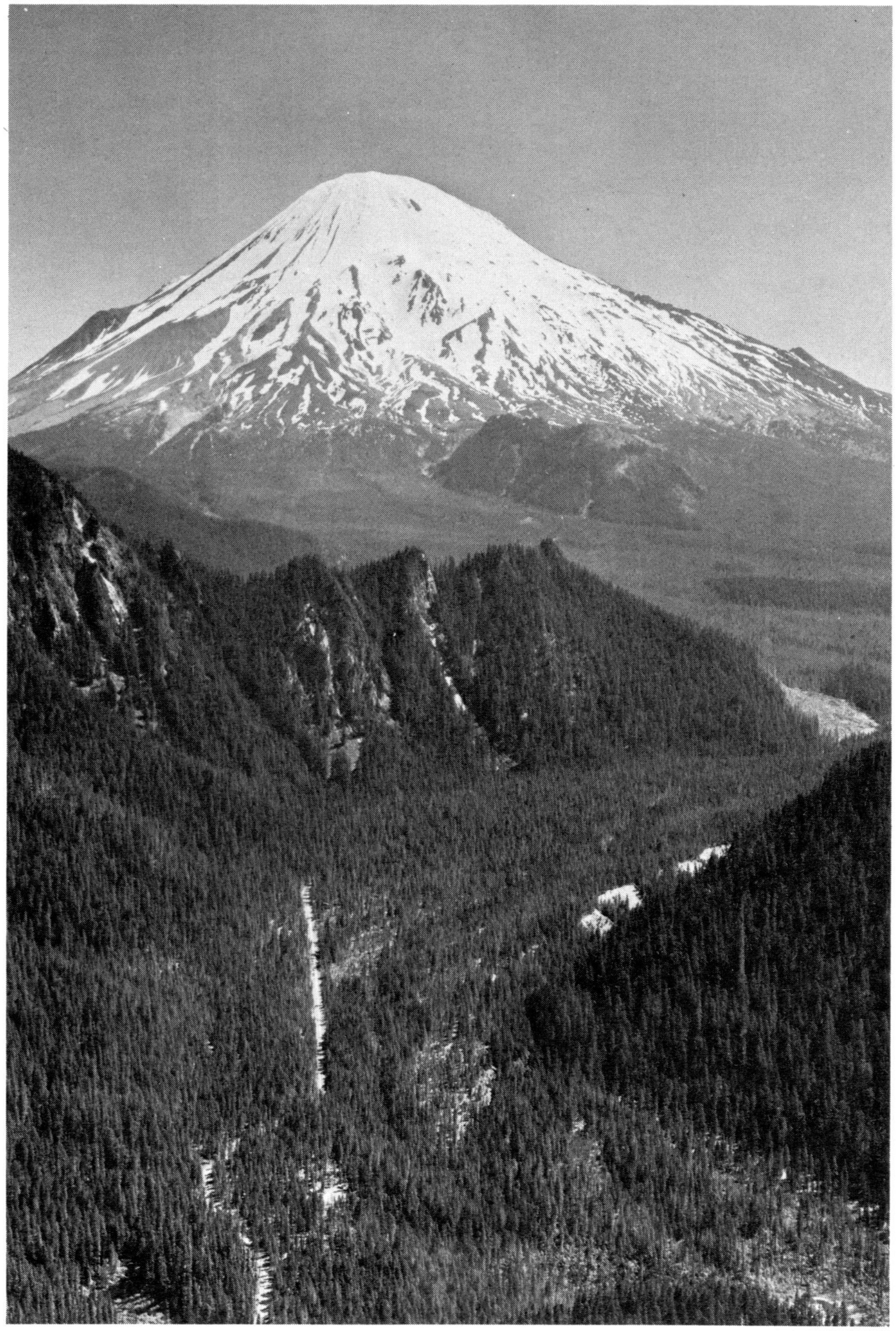

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CHAPTER XIII

SOIL, VEGETATION, AND WATERSHED MANAGEMENT

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INTRODUCTION

Water, soil, and nutrients are valuable resources of forest land. In the Douglas-fir region, large amounts of high quality water are consumed by cities, industry, and agriculture, and streams are used for spawning and rearing of native and anadromous fish species and for recreation. Soil is a medium for growth of forest vegetation, but becomes a water pollutant when put in streams by erosion. Similarly, nutrients, which are so vital for forest productivity, may degrade streams if present in water in high concentrations. The temperature of stream water determines the quality of habitat for valuable salmon, steelhead, and trout.

Timber harvesting alters the quality of water and the supply of soil and nutrients by changing the processes that control their movement from the land into streams. The loss of soil and nutrients from forests may be as detrimental to future productivity of forests as to present quality of stream water. The purpose of this chapter is to describe current knowledge about the impacts of forest management on streams and forests. The subjects covered here are hydrology, soil erosion, nutrient balance and stream temperature.

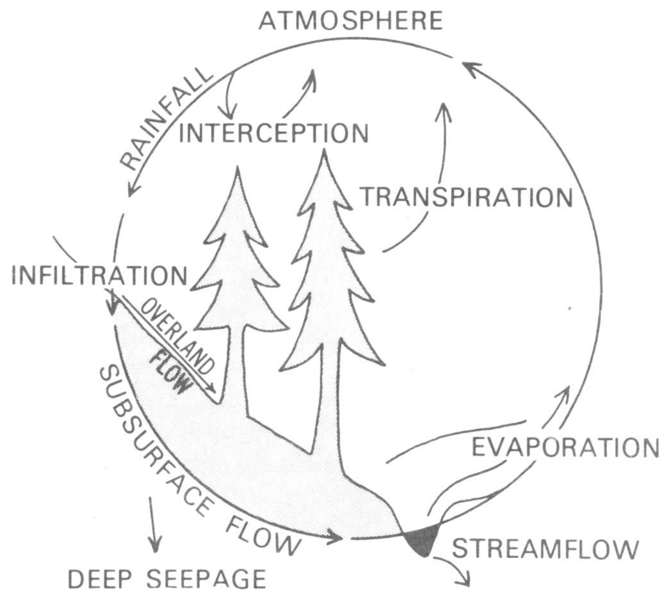

Figure 1. Forest hydrologic cycle.

HYDROLOGY

Hydrologic Processes

First of all, let's examine the hydrology of a forested watershed to see how man's forest activities might alter the hydrologic system. Water moves from the atmosphere through the forest system and back to the atmosphere by processes collectively known as the forest hydrologic cycle (Fig. 1). Some precipitation falling on a forest is intercepted by vegetation surfaces and forest litter. Some of this intercepted water evaporates, and the remainder reaches the mineral soil. The relative amount of intercepted rain depends on the size and duration of the storm producing the rain. During large winter storms, a small proportion of total precipitation is intercepted by forest vegetation. Conversely, during small storms, a large proportion of total precipitation is intercepted by forest vegetation and evaporated. Over the course of a year in the Douglas-fir region, 10-15 percent of annual precipitation in forested areas may be intercepted by plant surfaces and returned to the atmosphere by evaporation.

Water reaching mineral soil undergoes infiltration or entry into the soil. If water arrives at the soil surface faster than the soil can accept it, some water becomes overland flow, that is, water that flows over the soil surface. Whether or not overland flow will occur depends on the nature of the soil, primarily the macroporosity (amount of larger pores) of not only the surface soil, but also the lower depths of soil. The greater the amount of macroporosity, the more easily water can enter the soil and percolate to lower depths.

Water entering the soil is subject to *detention* storage and *retention* storage. These storages depend on the relative amounts and sizes of pores in the soil. Detention storage refers to water detained or slowed down as it moves through the soil, whereas retention storage refers to water retained in small pores for a relatively long period. Large pores aid infiltration and percolation but are incapable of longer term retention storage. Conversely, small pores favor retention storage at the expense of infiltration and percolation. Thus, a uniform sand of single grain structure has high infiltration and percolation capacities but low retention storage. Sandy soils tend to be droughty. Streams draining sandy soils may be intermittent, and plants frequently lack water during a substantial part of their growing season. On the other hand, unaggregated clay soils have very low infiltration and percolation capacities but will retain moderate amounts of water. Loam soils of crumb structure have a wide range of pore sizes that allow high infiltration and percolation capacities and retention of moderate amounts of water. Most forest soils in the Douglas-fir region are loams.

Another important physical characteristic of soil that affects streamflow and tree growth is soil depth. Soil depth controls in large measure the amount of water that can be stored in the soil or on a watershed and slowly released to streams or used by forest vegetation.

In undisturbed forest land in the Douglas-fir region, virtually all water arriving at the soil surface enters the soil. This is particularly true in forest land west of the Cascades. Overland flow occurs rarely and generally only under special circumstances, such as where soil is extremely shallow and easily saturated or where soil is frozen. This lack of overland flow is perhaps the most important hydrologic characteristic of undisturbed forest land in the Douglas-fir region (Fig. 2). This characteristic may also be seriously altered by certain of man's forest activities or by wildfire.

Water in retention storage may be evaporated from the soil or removed by plants through the process of transpiration. These processes are dependent on the energy available for water vaporization. Evaporation from soil under a forest is minimal because of the small amount of available energy. The rate of transpiration depends on the amount of energy available for water vaporization in leaves and the ease with which water may be withdrawn from the soil. As soil moisture contents decrease during the growing season, remaining water becomes more tightly held by the soil. Thus, as the growing season progresses, rates of transpiration may be reduced and, in some cases, transpiration may cease altogether.

Water not evaporated or transpired (evapotranspiration) by forest vegetation may eventually become streamflow. Thus, streamflow is, at least on an annual basis, largely the difference between precipitation and the evaporation losses described above. There are changes in soil moisture storage from year to year, and some water also seeps deep into the subsoil and into bedrock to become ground water.

At any given time and place, streamflow is comprised of water from channel interception (i.e., rain falling directly on the water surface or streambed within a channel), overland flow, and subsurface flow. Streamflow, of course, cannot be separated into distinct quantities from these different sources; but relative amounts of water from these sources can have a great influence on the time distribution of water arriving at the stream channel and the watershed outlet. Generally, the greater the proportion of overland flow, the more rapid the rate of increase in streamflow.

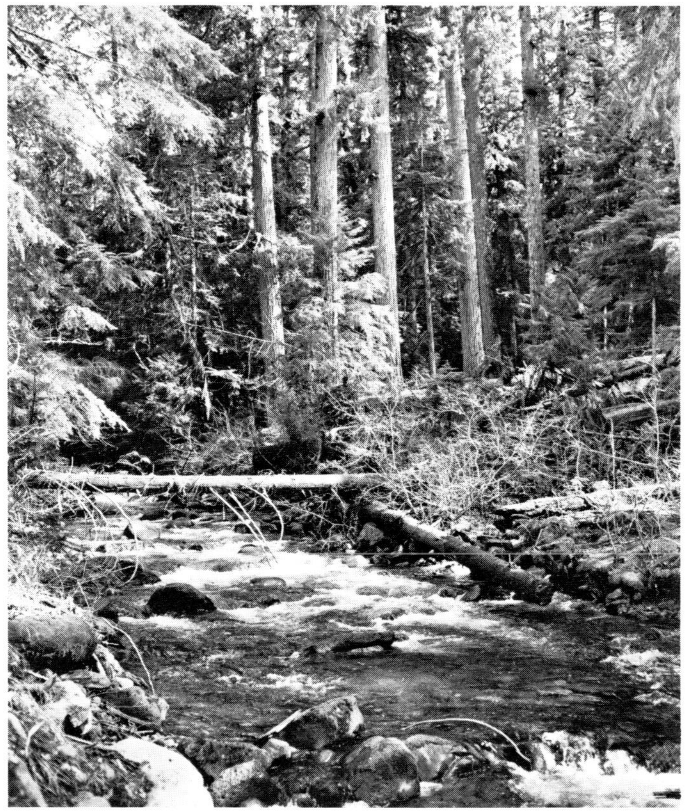

Figure 2. Small streams in the Douglas-fir region respond quickly to rainfall. Virtually all water passes through the soil mantle on its way to streams instead of flowing over the soil surface. This lack of overland flow is perhaps the most important hydrologic characteristic of undisturbed forest land and one of the characteristics most easily altered by timber harvest activities.

The source area of runoff in a small watershed expands and contracts according to rainfall characteristics and the capability of the soil mantle to store and transmit water. By means of this "variable source area," a channel network grows to many times its perennial dimensions and streams become both longer and wider (Fig. 3). A smaller degree of growth also occurs over the course of individual storms. This latter degree of change is illustrated by the bottom two watershed conditions shown in Figure 3. As a result of the variable source area of streamflow, both quantity and quality of streamflow can change drastically over a given period of time because the proportion of a watershed actively involved in streamflow production changes. Consequently, channel expansion has important implications for application of fertilizer and pesticides and for proper disposal of logging debris.

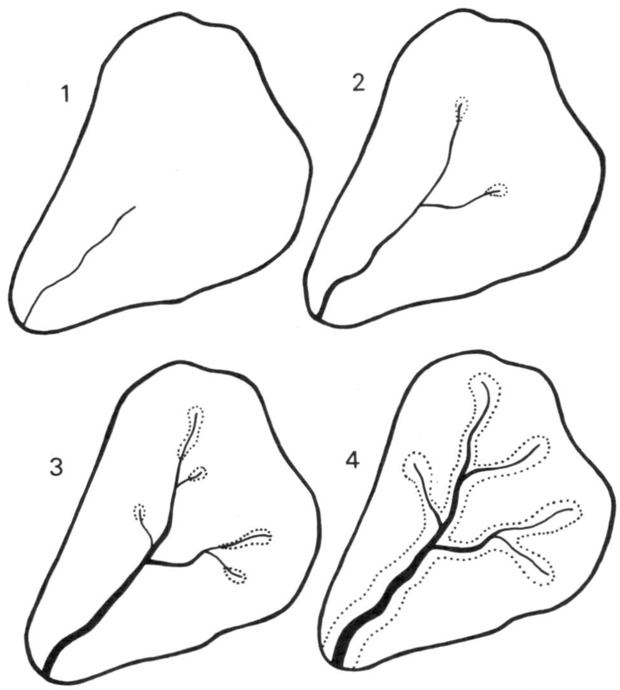

Figure 3. Time-lapse view of a small watershed showing the expansion of the channel network and source area of storm runoff (dotted lines).

Streamflow Characteristics

Characteristics of streamflow from forest land are of major importance in watershed management. The amount and timing of streamflow have important implications for erosion in headwater areas, downstream flooding, summer irrigation, as well as certain nonconsumptive uses of water. Moreover, these implications may be susceptible to change by some of man's activities in forests of the region.

Annual streamflow or water yield is of considerable importance in watershed management. Water yield is the total amount of water which flows from a watershed in a year. As stated above, it is largely the difference between precipitation and evapotranspiration. Examples of annual water yield are given in Table 1 for selected watersheds of the Douglas-fir region. Streams deriving significant portions of flow from glaciers, from areas above the Douglas-fir zone, or from low-land agricultural areas have been omitted. Annual yields are expressed as a uniform depth of water over entire watershed areas.

As can be seen in Table 1, there is considerable variation in annual water yield among watersheds of the region, primarily because of differences in climate. The range in yield is reflected in the Wynoochee River in Washington and the South Umpqua River in southwestern Oregon. The Wynoochee River drains the southern Olympic Mountains, one of the wettest areas of the Douglas-fir region, whereas the South Umpqua watershed in southern Oregon receives only about one third as much precipitation as the Wynoochee watershed. Furthermore, the weather during the growing season in the South Umpqua watershed is much more conducive to water vaporization so that evapotranspiration losses are greater. Thus, mean annual yield for the South Umpqua is only about 20 percent of the yield for the Wynoochee watershed.

Another important characteristic of streamflow is the timing of flow. In most of the Douglas-fir region, the annual distribution of streamflow is well correlated with annual distribution of precipitation. The vast majority of precipitation and streamflow occurs during the October through March period as the result of frequent, long-duration, low intensity frontal storms. A major deviation from this pattern occurs in watersheds where snowpacks develop. In these basins, precipitation falling as snow is temporarily stored; and annual streamflow may show a primary peak during spring months as this snow melts and another peak coinciding with rain in fall or early winter.

A third set of important streamflow characteristics is magnitude and timing of maximum flows. In the Douglas-fir region, maximum annual flows generally occur in December or January (Table 1), but may occur as early as November or as late as April where snowmelt causes maximum flows. During long-duration storms common to these months, soils are extremely wet and are able to transmit large quantities of water during and shortly after rainfall. Sizes of source areas of runoff are at annual maximums, and water is efficiently transported out of watersheds. The maximum streamflows of record shown in Table 1, however, have nearly all resulted from heavy rainfall with concurrent rapid snowmelt. This rapid snowmelt releases large amounts of water which, when added to heavy rainfall, provides water input to soils at a rate much higher than that provided by rainfall alone. As detention storages are exceeded, the result can be devastating floods in the lowlands and landslides and extensive damage to stream channels, roads, bridges, and culverts in upland areas.

The maximum streamflows given in Table 1 are not directly comparable among watersheds because of differences in watershed size. For example, the maximum streamflow for the South Santiam River in Oregon is over four times as large as that for Lookout Creek in Oregon (782 m^3/sec versus 189 m^3/sec) because the South Santiam watershed is much larger. If we divide these by watershed area, we see that each square kilometer of area in the Lookout Creek

Table 1. Mean annual, maximum, and minimum streamflows for selected watersheds of the Douglas-fir region through the 1974 water year.

Location and stream	Area (km^2)	Length of record (yr)	Mean of annual yields[1] (mm)	Mean of maximum streamflows (m^3/sec)	(month)	Mean of minimum streamflows (m^3/sec)	(month)
WASHINGTON							
Olympic basins							
Duckabush River	172	36	2189	254	Nov.	1.27	Sept.
S. Fork Skokomish River	198	43	3345	612	Jan.	1.76	Sept.
Wynoochee River	192	49	3845	668	Nov.	1.61	Sept.
Soleduck River	217	40	2589	665	Nov.	1.44	Sept.
Humptulips River	337	33	3595	935	Dec.	2.32	Sept.
West Cascade basins							
S. Fork Nooksack River	267	41	2491	547	Nov.[2]	0.55	Sept.
Alder Creek	28	31	1128	20	Dec.[2]	0.12	Sept.
Pilchuck Creek	135	25	1871	195	Jan.[2]	0.014	Aug.
S. Fork Stillaguamish River	308	46	3133	918	Feb.[2]	1.56	Sept.
Snoqualmie River	971	44	2393	1730	Nov.[2]	0.27	Aug.
Cedar River	105	29	2339	269	Nov.[2]	0.57	Nov.
Southwest Washington basins							
E. Fork Lewis River	324	45	2111	544	Jan.	0.82	Nov.
Salmon Creek	47	31	1187	42	Jan.	0.034	Aug.
Grays River	103	19	3008	263	Jan.	0.48	Sept.
OREGON							
Coastal basins							
Nehalem River	1728	35	1441	1328	Jan.	0.96	Aug.
Wilson River	417	44	2629	1020	Jan.	0.91	Sept.
Siletz River	523	55	2719	1160	Nov.	1.36	Sept.
Alsea River	865	35	1622	1180	Dec.	1.27	Sept.
S. Fork Coquille River	438	55	1633	1380	Dec.	0.34	Sept.
East Coast Range basins							
S. Yamhill River	344	40	1649	374	Dec.	0.07	Oct.
Luckiamutte River	89	40	2122	157	Dec.	0.11	Sept.
Marys River	412	34	1035	385	Dec.	0.02	Aug.
West Cascade basins							
Molalla River	251	39	1974	688	Dec.	0.51	Oct.
S. Santiam River	451	39	1653	782	Dec.	0.65	Dec.
Lookout Creek	62	17	1903	189	Dec.	0.18	Nov.
Southwest Oregon basins							
S. Umpqua River	1163	36	809	1700	Dec.	0.57	Sept.
Grave Creek	57	29	963	177	Dec.	0.003	July

[1]Mean annual yield expressed as a uniform depth over the entire area of a watershed.
[2]Streamflow frequently has a secondary annual peak during spring snowmelt.

watershed produced about 3.0 m^3/sec while each square kilometer of area in the South Santiam River watershed produced only about 1.7 m^3/sec. Larger watersheds nearly always exhibit lower unit area peak flows than small watersheds because (1) a larger basin generally experiences a lower average rainfall intensity; (2) peaks from small subwatersheds are seldom synchronized and so are not additive at the outlet of the larger watershed; and (3) a larger basin has greater channel storage that attenuates storm runoff.

Another major characteristic of streamflow is minimum annual flow. These minimum flows depend on the distribution of annual precipitation, the relative water use by forest vegetation, and the relative contribution of groundwater to streamflow. Of considerable importance here are soil properties such as depth, porosity, and pore size distribution. Over the region, the time of minimum flows varies from middle summer to late fall. Magnitude and month of minimum streamflows are shown in table 1. Again, for comparison among watersheds, minimum flows must be divided by watershed area.

The maximum and minimum recorded streamflows shown in table 1 indicate the wide variation in streamflow that may occur on any one watershed. If soils in a given watershed have low detention storage, then maximum flows tend to be high. If retention storage is low, then minimum flows also tend to be low. Also, if a large portion of a watershed's annual precipitation occurs as snow, water will be released during spring and summer seasons and minimum flows will be relatively high.

Changes in Streamflow Characteristics

Several studies have been conducted to determine the effects of timber harvest activities on streamflow characteristics in small headwater basins. Such activities can drastically alter the hydrologic cycle by reducing interception and transpiration losses (Fig. 4). On an annual basis, more water reaches the soil, transpirational demands are reduced, and an increased amount of soil water is available for streamflow. Severe soil disturbance, such as soil compaction by tractors, road construction, and high intensity slash burning, may drastically reduce infiltration capacities of forest soils and cause overland flow. But, as we shall see later, changes in streamflow in these (less than 3 km^2) headwater basins have little effect on streamflow characteristics in much larger watersheds such as those listed in table 1. This is because only small parts of these larger watersheds are in a drastically altered hydrologic condition at any one time.

Timber cutting can increase water yields from small watersheds. Completely clearcutting a small watershed can increase annual yield by more than 50 cm (Fig. 5); clearcutting 25-30% of a watershed in patches after road construction can increase annual yield up to about half that of complete clearcutting (Fig. 6). Greatest seasonal increases occur in fall and spring when timber removal causes soil water contents to be greater and less precipitation is required to satisfy soil moisture storage. Minimum flow may more than double in late summer or fall, but the increases are small in absolute terms.

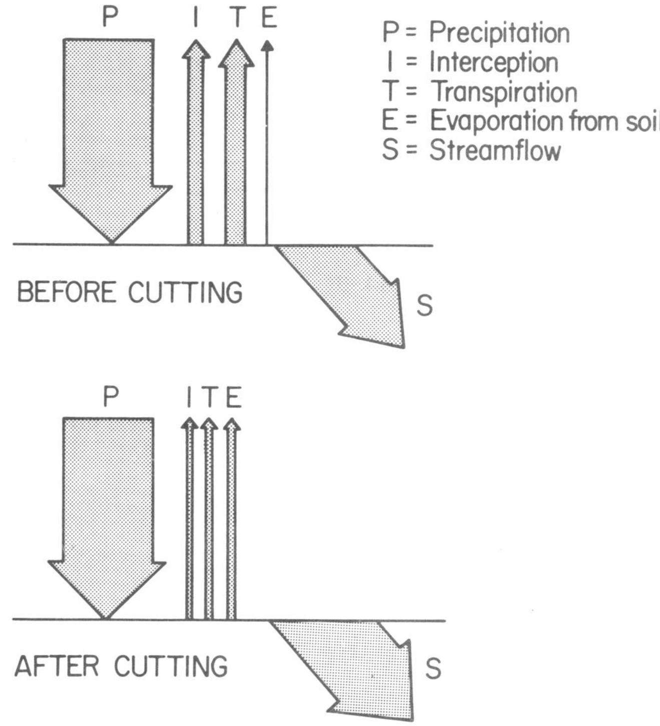

Figure 4. Relative sizes of components of the forest hydrologic cycle before and after timber cutting.

An increase in streamflow is only as permanent as the change that produced the increase. As cutover areas revegetate, more water is returned to the atmosphere by interception and transpiration and less is available for streamflow. Thus, water yields begin to diminish. Summer increases are particularly short-lived and may largely disappear after 2 to 3 years as streamside vegetation returns (Fig. 7).

Under sustained yield forest management, only a small percentage of a large watershed is logged each year. Revegetation begins almost immediately in most areas and whatever initial changes in streamflow that have occurred may begin to diminish. Many years are required for altered streamflow characteristics to return to their prelogging levels, perhaps more than 20 years in the case of annual yield. Under probable harvesting schedules for the future, increases in streamflow from logged areas in these large watersheds will be diluted by unaltered streamflow from unlogged areas and smaller increases from revegetating areas. The net effect is that at the outlet of large parent watersheds like those in table 1, the effects of timber harvesting in headwater areas

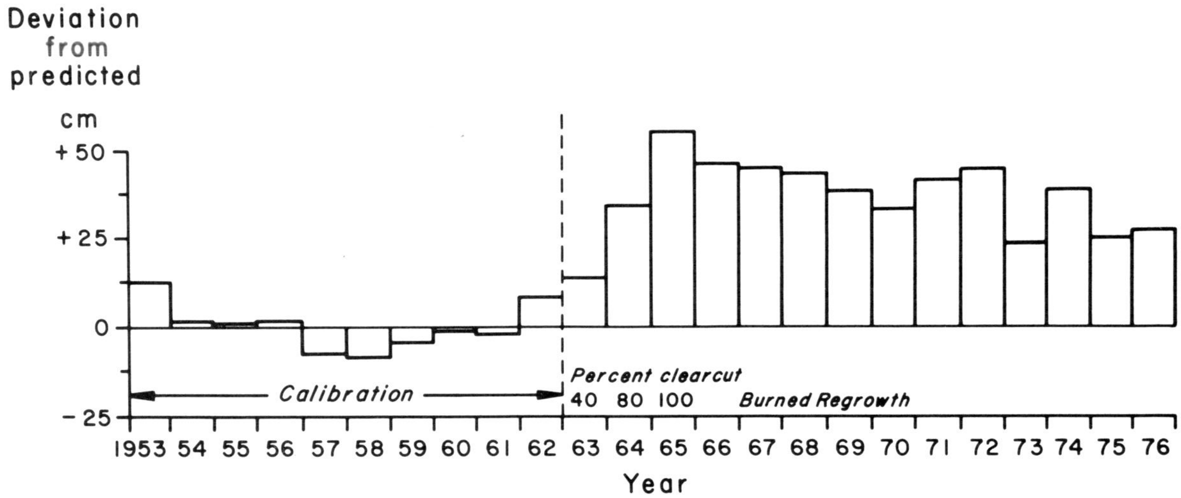

Figure 5. Increases in annual water yield after clearcut logging in Watershed 1, H. J. Andrews Experimental Forest, Oregon.

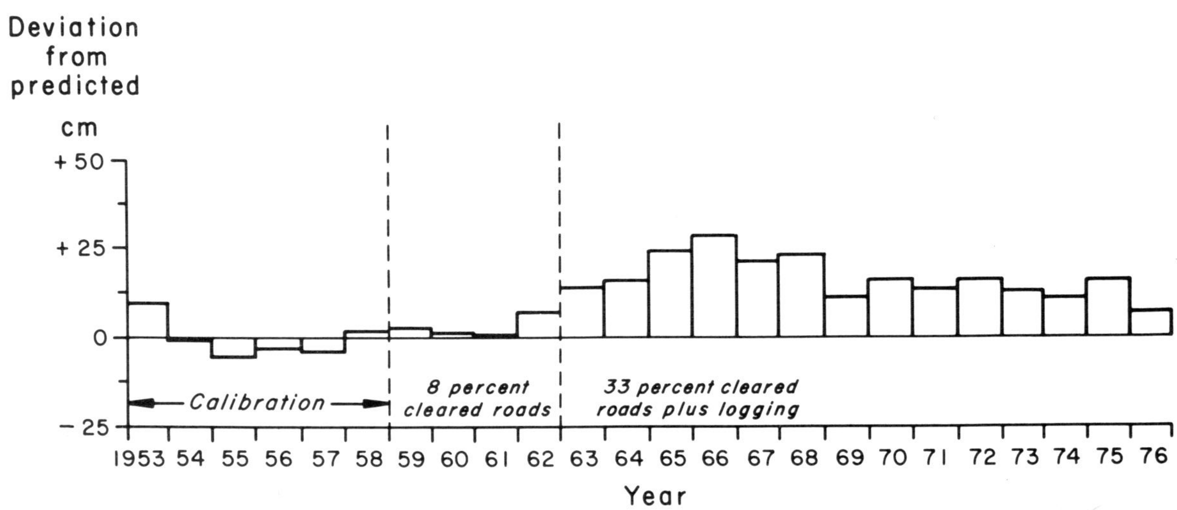

Figure 6. Increases in annual water yield after patch-cut logging in Watershed 3, H. J. Andrews Experimental Forest, Oregon.

will probably continue to be undetectable, particularly in view of the normal accuracy of streamflow measurements.

There is continuing speculation about the influence of timber harvesting on the magnitude and frequency of floods in the Douglas-fir region. A frequent argument is that harvesting activities which so drastically alter the forest hydrologic cycle must increase floods. The effects on floods, however, cannot be explained this simply. When one considers possible effects, he must also consider headwater or upland areas separately from downstream areas which include floodplains used for agriculture and municipal and industrial development. Whereas timber harvesting occurs in the headwater areas, damage from maximum streamflows occurs in either or both regions.

Peak flow is one characteristic of a runoff event in small headwater basins which may cause damage in that basin. Peak flow—the highest instantaneous rate of streamflow attributable to a particular rainstorm or snowmelt period—is largely responsible for damage to stream channels, bridges, and culverts in headwater areas. Thus, any increase in the magnitude of the large, infrequent peak flows could cause damage in these headwater areas. Increasing the magnitude of smaller, much more frequent peak flows should not cause appreciable damage in headwater areas.

The second damage-causing characteristic of a runoff event in small headwater basins is stormflow volume. It is this volume from a small subwatershed which, when synchronized with volume from similar watersheds at a point downstream, results in the peak flow in the parent watershed at that downstream point. Clearly, for timber harvesting to increase magnitude of the downstream peak flow requires that a major portion of the parent watershed be sufficiently altered hydrologically and that the time of arrival of increased storm runoff from all the altered upland subwatersheds be at least partially synchronized at the downstream point. This is very unlikely to occur.

Research has shown that removal of forest vegetation increases the magnitude of smaller peak flows. Greatest increases result primarily from changes in soil water contents on areas where trees are cut. In winter when the soil mantle has been recharged in the forested and the cut areas, both areas will respond nearly the same hydrologically. Thus timber removal alone has little effect on the magnitude of peak flows or of stormflow volume during large runoff events—the events most likely to cause damage in upland or downstream areas.

Soil disturbance may have a pronounced effect on the size of large peak flows. For example, road surfaces have virtually no infiltration capacity, may intersect slow subsurface flow, and rapidly transport water in a ditch-culvert system to streams. As a result, peak flows for small headwater streams may be increased when roads occupy at least 12 percent of a watershed area (Harr 1976a). There is also some indication that severe soil disturbance from tractor skidding and windthrowing of slash may increase the magnitude of larger peak flows by compacting soil to the point where infiltration is virtually eliminated and overland flow occurs (Fig. 8).

We have discussed certain characteristics of streamflow in small headwater basins that may be altered by timber harvesting activities. Although timber harvesting operations on upland watersheds may not appreciably alter the quantity or timing of water downstream, these harvesting operations may seriously impact not only water quality in these upland basins, but also certain aspects of water quality far downstream. This water quality and how it is influenced by man is the subject of the next part of this chapter.

SOIL EROSION

The soil mantle is the basic forest resource we are managing. Future forest productivity will depend on the depth of soil available for rooting and the ability of the soil to store and supply water to provide nutrients to roots and to exchange gases between the soil atmosphere and the above-ground atmosphere.

The Pacific Northwest has high natural rates of soil erosion because it is geologically young and characterized by steep mountainous topography. Soil erosion processes are divided into two general types. Soils move as individual particles or *aggregates* from the soil surface (surface erosion) or as entire sections of the soil mantle—often including underlying weathered bedrock (mass erosion). In both types, the effect of gravity, the force moving the soil downslope, increases with the slope of the land. Forest cover on mountain slopes is an important natural control of soil erosion. The canopy of this vegetation and the organic layer covering the soil dissipates the energy of falling raindrops, and the roots strengthen the soil mantle.

Before the advent of European man to the region, the soil erosion cycle was controlled by natural events such as wildfire that destroyed forests over large areas and catastrophic flood storms that caused extensive soil erosion. Natural wildfires have been largely controlled and timber harvesting activities have replaced them as the dominant agent of soil erosion. Most of the present soil losses result from catastrophic storms with roads and recent clearcuts often the sites of erosion.

Soil erosion processes differ between west and east sides of the Cascade Range because of differences in climate, vegetation, soil properties, and land use. On the west side of the Cascade Range, the climate is maritime, while on the

Figure 7. Removing forest vegetation from entire small watersheds can increase the amount of water that flows from the watersheds. Such increases diminish over time as reestablished vegetation uses greater amounts of water.

Figure 8. Severe soil compaction on forest roads, landings, skid trails, and other areas disturbed during logging can cause overland flow. As a result, size of peak flows of streams draining logged watersheds may be increased.

east side the climate is more continental. Most of the precipitation in the region comes in winter months from frontal storms moving landward from the Pacific Ocean.

On the warm-moist west side, the soils are porous and most of the precipitation passes through the soil mantle on its way to streams. Because rapid vegetation growth covers bare soil with leaf cover and litter within a few years of disturbance and overland flow seldom occurs, surface erosion is of minor importance in forested areas. Therefore, dominant erosion processes on the west side are landslides—a process which moves masses of the soil mantle to streams.

The east side is much drier because of the rain shadow effect of the Cascade Range. Vegetation is a mosaic of grassland and forest with conifer stands occurring at higher elevations. Because soils are less porous than those on the west side, overland flow is more common with surface erosion being particularly important on grazing lands. The effects of catastrophic runoff events are most severe when the surface soil is sealed by freezing.

SURFACE EROSION

Two types of surface erosion are recognized in the Pacific Northwest. One is a *wet process* associated with surface runoff. The second type is a *dry process* or ravel occurring in the summer months. Although both processes operate throughout the region, the wet process probably accounts for most of the soil erosion east of the Cascade crest, and the dry process is probably the principal surface erosion process west of the Cascades. Both processes operate when the soil is essentially bare of organic cover.

Wet Type

Surface erosion is a two-step process involving detachment of soil particles and their subsequent transport downslope. The size and density of soil particles and the degree of soil protection afforded by plant and litter cover control the detachment process.

Two types of wet surface erosion are recognized, depending on the mode of transport—flowing water or raindrop splash. Water flowing over the soil surface causes rilling and gullying which are easily recognized from the scars left on the ground. Where soils are porous and water can infiltrate the surface, particles can also be detached and moved downslope by raindrop splash. This is called sheet erosion—a form of erosion that may go virtually unnoticed. Indicators of sheet erosion are formation of pedestals of soil under impervious materials such as flat stones, wood chips, or beneath exposed roots.

Soil structural and hydrologic properties are indicators of *erosion potential*—the probability that erosion will occur. Surface erosion does not occur on most forest soils because individual particles are cemented together by clay and organic matter forming larger structural units called *aggregates* and more energy is required to move these larger objects. The openings between aggregates conduct water into soil pores. But surface runoff often occurs on poorly aggregated soils when the organic layer is removed. Raindrops striking weakly aggregated soils break down the aggregates, and the fine particles of soil thrown into suspension seal the soil surface and reduce infiltration.

Freezing and *non-wettability* also decreases infiltration in well-structured soils. Ice in surface pores can seal the soil surface and can cause high rates of surface erosion when warm rains break the cold spell. High soil erosion rates from such events are observed east of the Cascades roughly every 10 years. At times, certain soils will not accept water because organic substances in them repel water when the soil is dry. Water is forced over the surface causing surface runoff. Heating of soils by fire intensifies the non-wettable condition.

The strength of aggregation is often indicated by the type of bedrock from which soils are formed. Studies in several western states have shown that soils derived from granite, quartz diorite, granodiorite, and certain high quartz sandstones are most erodible. Relatively non-erodible soils are formed from basalt, andesite, and gabbro. The higher the quartz content of the parent material, the greater the potential erosion hazards of the resultant soil.

The greater erosion potential of soils high in quartz is largely the result of poor aggregation. Such soils tend to be coarse textured in the surface layers (loamy sands or sandy loams) and deficient in silts and clays. Because coarse textured soils have little clay, they also have few stable aggregates. On the other hand, basalts, andesites, and gabbro contain minerals which decompose to form clays. Greater organic matter contents, that have resulted from the higher productivity of these soils, also contribute to stability of their aggregates.

The rate that water can flow through soil can also determine erodibility. Soils with a high infiltration rate can produce surface runoff if the percolation capacity is much less than the maximum sustained rate or precipitation received by the soil. (See page 4 of this chapter for a discussion of soil porosity.)

Disturbance by harvest activities can alter infiltration capacities of forest soils. Shallow disturbance may only mix the aggregated surface soil and litter, but deep disturbance caused by dragging large logs typical of the region commonly exposes poorly aggregated subsoils. Sealing of subsoils in skidroads which run up and down the slope often causes rill or gully erosion.

Similar effects on infiltration capacities are often caused by logging road construction and by tractors used to skid logs. Compaction destroys most of the non-capillary pore space necessary for conducting storm runoff. Tractor roads on sloping land often create ephemeral streams where drainage was entirely subsurface before harvesting.

Soil water contents which maximize compaction are typically moderate; soils are neither dry nor wet. The greatest degree of compaction occurs midway between field capacity and the permanent wilting point. Soils are moderately moist in late spring and early summer and again in fall as precipitation recharges the soil mantle. Consequently, soils may be dry enough to support tractors without serious and nearly irreversible compaction only in late July, August, and September.

Compaction causes long term reductions in soil porosity and infiltration capacity. One study found that 5 years after logging, infiltration capacities on skidroads were only 4 percent of those for undisturbed soils (Tackle 1962). Another study in western Oregon examined depth of compaction and root penetration into compacted soils on four sites harvested within the past 55 years (Power 1974). Roots penetrated to depths of only 10 cm on compacted roads and landings. Roots from trees in uncompacted areas grew up to margins of compacted areas where they either turned and paralleled the compacted areas or occupied the loose soil over compacted areas. Except for the surface 10 cm, soil porosity was apparently improved little during the 55-year period. Thus compaction may last for more years than are required to produce a new forest.

Dry Type

Another form of surface erosion is dry ravel—the downslope movement of single particles or aggregates of soil on steep, sparsely vegetated slopes. Dry ravel is common on dry streambanks, roadbanks, and fresh landslide scars during the dry summer period in the Pacific Northwest. Detachment is caused by the lack of *cohesion* between structural parts as the soil dries. Because transport is by gravity, dry ravel is often classified as a form of mass movement.

Minimizing Surface Erosion

In forest settings, surface erosion is virtually nonexistent where vegetation and the forest litter cover the soil surface. Maintenance of organic and vegetative cover is the ultimate strategy of erosion control. As a practical consequence, control involves minimizing the area of bare soil created during harvest operations and rapidly establishing vegetation cover on bare soil surfaces. Erosion control measures are most effective when they are made part of project plans for harvest operations.

Surface soil erosion potential should be considered for each site when silvicultural methods are selected for regeneration and methods are chosen for transporting logs to landings. Specifics of the project plan for controlling soil erosion are determined by requirements for protecting streams from siltation and maintaining soil productivity. If erosion hazard is unusually high, the optimum decision might be to forego timber harvesting. Such decisions are often justified to maintain high levels of water quality in municipal watersheds or in important salmon or trout streams. On the other hand, if the site is designated for timber harvest, the manager may control soil disturbance by selecting among several logging systems including tractor, high lead, skyline, helicopter, and balloon. Shelterwood cutting is preferred if additional site protection is required. Where soils are stable and the erosion risk is low, clearcutting with replanting is an acceptable practice (Fig. 9).

Figure 9. Soils, deeply disturbed along high lead skid roads, remain bare and susceptible to surface erosion for many years (A, B, C, and D above). Vegetation finally covered more than half of the plot by the 14th year (E).

Figure 9A. Bare soil 2 years after yarding.

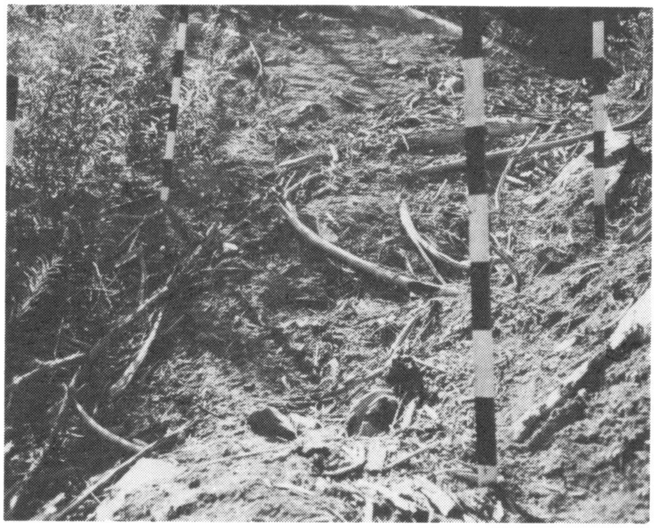

Figure 9B. Bare soil 5 years after yarding.

Figure 9D. Bare soil with fireweed and thimbleberry (*Rubus parviflorus*) 10 years after yarding.

Figure 9C. Bare soil with a trace of fireweed (*Epilobium angustifolium*) 8 years after yarding.

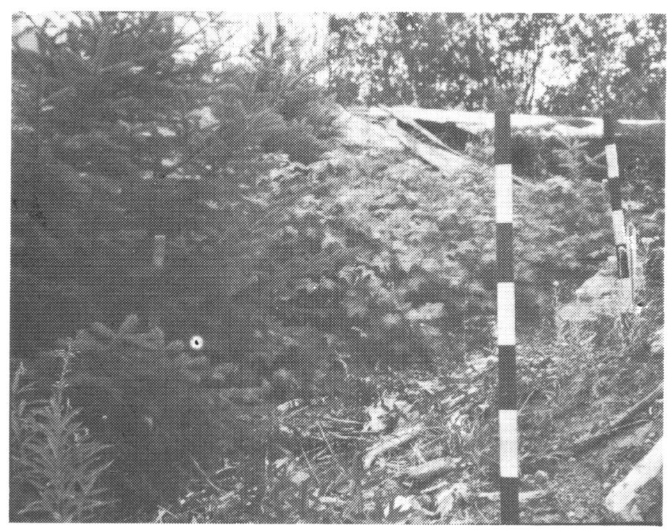

Figure 9E. Thimbleberry and Douglas-fir (*Pseudotsuga menziesii*) trees begin to cover the soil 14 years after logging.

Surface erosion can be minimized in steep terrain by limiting surface disturbance and compaction during harvest operations. As shown in table 2, soil disturbance and compaction vary among methods of clearcutting in western Oregon. All cable operations caused much less soil damage compared to the tractor operation. Damage was reduced by cable systems in proportion to the lift given the logs while being moved to landings. The area devoted to high-lead skid roads is much less than for tractors (Fig. 8 and 9). Helicopter logging should give results similar to those for balloon. In terms of lasting soil damage, reductions in soil compaction are more important than reductions in disturbance since bare soil is often necessary for establishing conifers but soil porosity is essential for full root development.

Table 2. Disturbance and compaction of soil caused by four harvesting systems in the H. J. Andrews Experimental Forest.

Harvest Methods	Bare soil (%)	Compacted (%)
Tractor	35.1	26.4
High lead	14.8	9.1
Skyline	12.1	3.4
Balloon	6.0	1.7

Surface erosion can also be caused indirectly by disposal of logging residue to reduce the fire hazard or by site preparation for planting of trees. Fire hazard is particularly high after clearcutting old-growth forests because tremendous volumes of residue remain on the ground. Methods of hazard reduction are burning or removal of residue from the site. Site preparation for planting includes application of herbicides to kill competing vegetation followed by broadcast burning after a suitable drying period.

Broadcast burning of residue removes the protective cover of the remaining vegetation and organic debris. But the effect of burning on soils may be somewhat less important than a burned area's appearance would suggest because the fire usually burns only 40-60 percent of the area. Detrimental effects of burning on soil aggregates are confined mainly to 3-8 percent of the area where extremely hot fires beneath large concentrations of fuel consume most of the organic matter in the surface soil. Soil structure is not seriously affected by light burning (Fig. 10).

Hand or tractor piling for burning reduces the area affected by fire. But hand piling is costly and practical only for small sizes of residue and therefore is not a common practice in the Pacific Northwest. Tractor piling increases soil disturbance and compaction often causing long term erosional impacts on the site and loss of soil productivity.

Removal of residue from the site is typically done with the same machinery used to remove merchantable logs. Cull logs, tops, and large branches are piled on landings and burned.

Figure 10. Lightly burned areas (A) are soon covered by herbs beginning with woodland groundsel (*Senecio sylvaticus* L.) (B). Groundsel is replaced the next year (C) by fireweed (*Epilobium angustifolium* L.) and other species. Snowbrush (*Ceanothus velutinus* var. *velutinus* Dougl. ex Hook.) seedlings 3 years after yarding in (B) are competitors of fireweed by the 6th year (D). Snowbrush is a nitrogen fixer and dominates the site by the 9th year (E). Nutrient cycling is promptly re-established by the vegetation and surface soils are stabilized by year 3 (B).

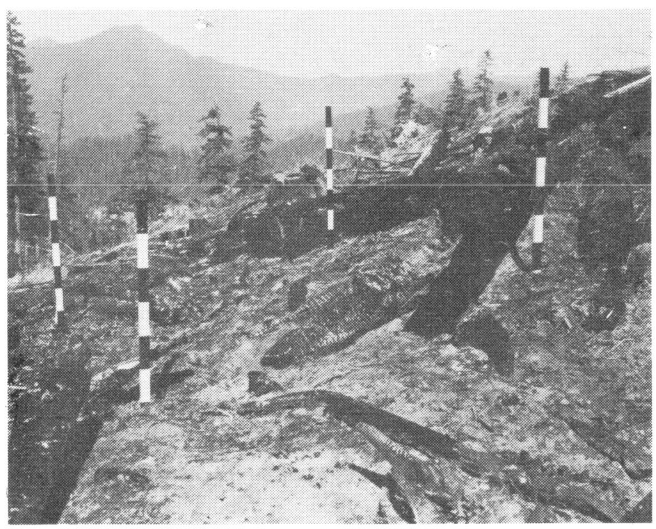

Figure 10A. Year 2, after broadcast burning.

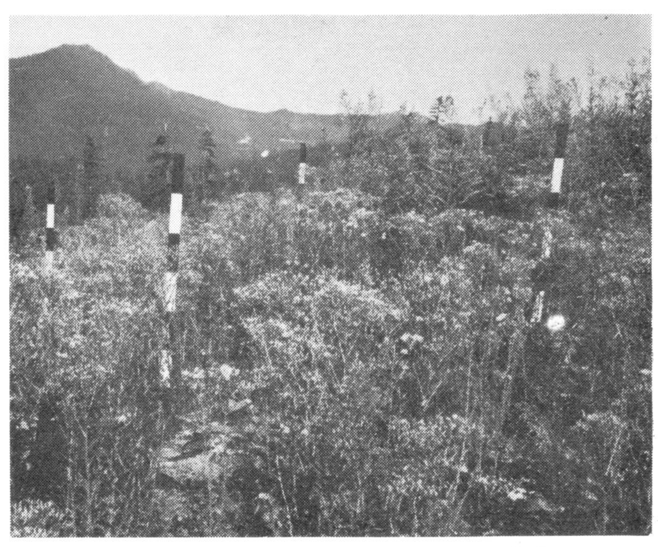

Figure 10B. Year 3, the woodland groundsel stage.

Figure 10C. Year 4, the fireweed stage.

Figure 10D. Year 6, shrubs (snowbrush ceanothus) overtops the fireweed.

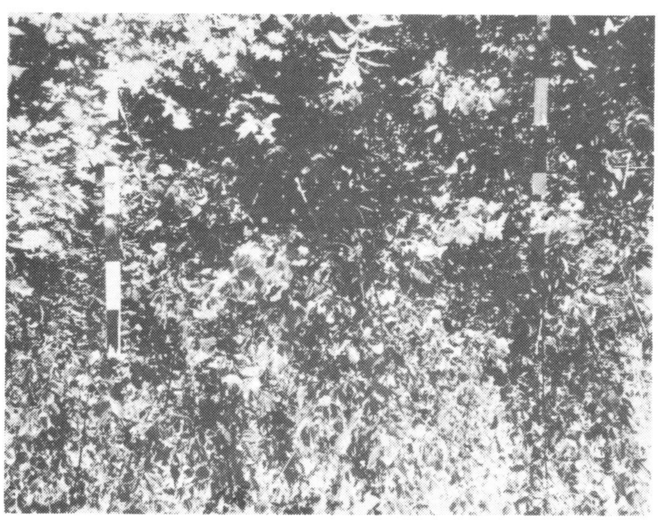

Figure 10E. Year 9, shrubs, predominantly snowbrush ceanothus, dominate the site.

Information on the impact of broadcast burning on soils derived from basic igneous parent materials was gathered from two sites in western Oregon. Broadcast burning after skyline logging increased the amount of bare soil from 12 percent before burning (table 2) to 55 percent after burning (Mersereau and Dyrness 1972). Rapid invasion of vegetation, however, essentially halted surface soil erosion by the second season after burning. At the second site, tractor piling and burning of logging residue or removal of cull logs by high lead cable reduced fire hazard. Table 3 shows the effects of skidding both merchantable and cull logs. Tractor piling of residue caused 89.2 to 94.6 percent disturbance of the surface soil.

The area compacted (26.4–27.3 percent) is similar to that for tractor harvesting. The high lead operation disturbed the soil surface on more than half of the area, and nearly as much as the disturbance caused by high lead and broadcast burning (55 percent) at the first site.

Except for the loss of nitrogen by volatilization, broadcast burning appears to produce fewer and shorter term impacts on soil compared to the more energy-intensive tractor piling of residue for burning. Soils bared by skidding logs with overhead cable systems or by broadcast burning are rapidly covered by vegetation on moist and mesic sites. This rapid revegetation is undoubtedly the most effective control of surface soil erosion. Broadcast burning is not advisable where soils are shallow and poorly aggregated on steep slopes. The erosion problem may be compounded by nonwettability on such steep sites.

Road slopes erode rapidly by both dry and wet surface processes and can be ready sources of sediment for streams. Surface erosion on these slopes can be quickly controlled by the application of grass seed, mulch, and nitrogen fertilizer. Timing of application is important; temperature and moisture conditions are optimum in late summer or early fall. The mulch is often needed to protect the soil and help establish vegetative growth; this is especially true at higher elevations where low temperatures limit rates of growth during the fall and early winter.

MASS EROSION

Mass erosion processes, often termed landslides, can be separated into four distinct but related types, creep, slump-earthflow, debris avalanche-debris flows, and debris torrent. The internal frictional resistance of the soil and underlying porous material retards downslope movement. Increasing amounts of water can decrease the frictional resistance of the material, often triggering landslides. Forest vegetation can increase frictional resistance by removing water from the failure zone by evapotranspiration or by reinforcing and anchoring the porous layers with a web of roots.

Table 3. Disturbance and compaction of soil caused by skidding merchantable and cull logs from large and small clearcuts at the South Umpqua Experimental Forest.

	Size of clearcut	Percent bare soil	Percent compacted soil
Tractor piled	large	89.2	27.3
and burned	small	94.6	26.4
Culls removed	large	55.9	6.8
by high lead	small	57.7	5.6

Characteristics of soils and bedrock will determine which mass erosion process will occur on a given site. Soils high in clay have plastic properties, that is, the soil mass tends to stick together. Such materials are termed cohesive. Cohesive soils, because of strength and plasticity, tend to creep or flow and often deform without forming cracks or zones of failure (Fig. 11). When shear failure does occur,

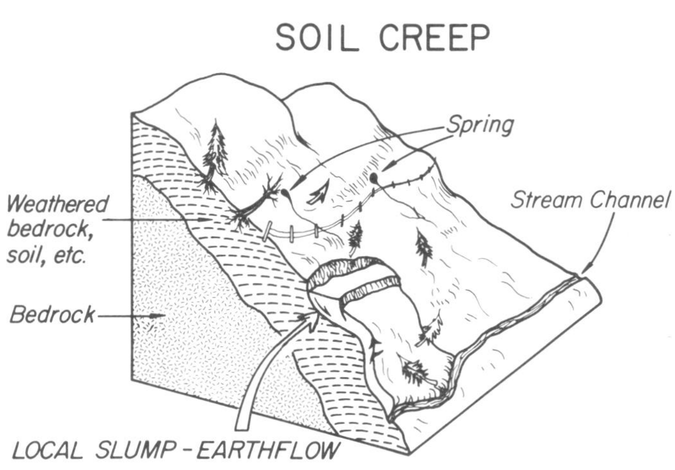

Figure 11. Creep is the slow downslope movement of clayey material.

movement usually involves backward rotation of a single block or several blocks in a lobate form called a slump-earthflow (Fig. 12).

Soils that do not readily stick together are termed noncohesive. The texture of noncohesive soils varies from silt loams to sands. These materials do not exhibit the deformational movements common to cohesive materials and tend to develop well defined, almost instantaneous failures of the debris avalanche type which move rapidly downslope (Fig. 13).

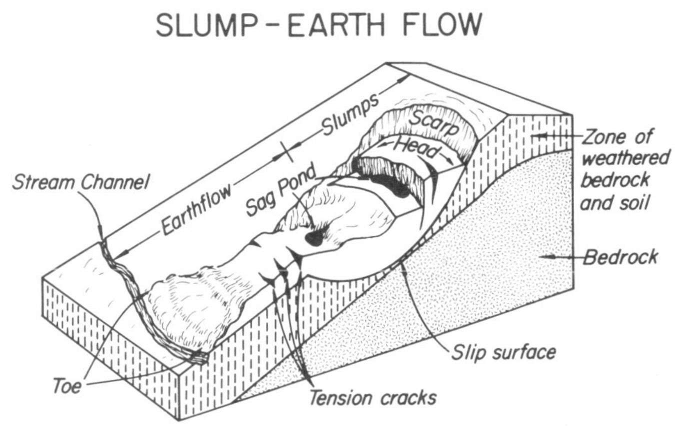

Figure 12. Slump-earthflow involves two distinct but related features; 1) shear failure of a slump block which deposits earth on the slope below, and 2) flowage of the earth downslope to stream channels.

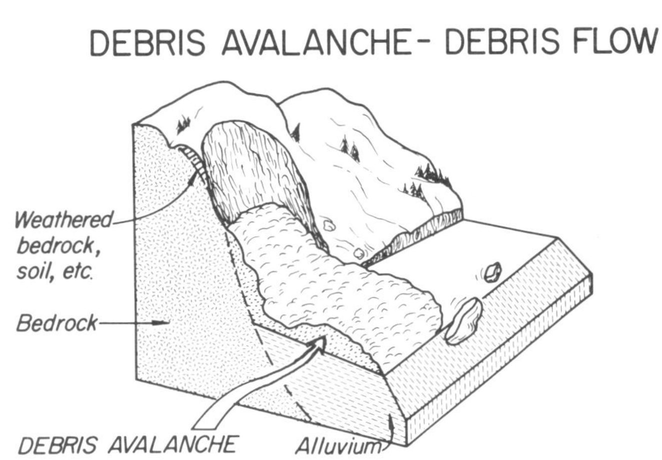

Figure 13. Debris avalanches are the rapid and shallow mass movements of coarse textured material downslope.

Mass erosion also varies depending upon the strength of bedrock underlying soils. Weak bedrock, such as weathered sandstone or deeply weathered igneous rock such as tuff and breccia, does not possess the strength to support its own mass; creep and slump-earthflows are the most common result. Where landslides often include the bedrock layers, the bedrock is termed incompetent. Competent bedrocks (basalt, andesite, granite) have the strength to support their own masses. Landslides over competent bedrock include only the overlying soil and are of the debris avalanche type.

Creep

Creep is the slow downslope movement of clayey cohesive materials (Fig. 11). Movement is much like a thick viscous fluid with stresses insufficient to produce distinct failures. Water contributes to creep movement by reducing the internal frictional resistance of the soil mass. The depth of creep movement varies with the depth of weathered material and movement of water through the mass. Most of the movement occurs during the rainy season except in isolated instances where the mass is subirrigated by an external source such as springs.

Creep movement occurs extensively in clay-rich soils and weathered bedrock of Washington, Oregon, and northern California. Measured rates of movement are variable but as a general rule range up to 15 mm per year.

Creep is the most persistent of all mass erosion processes. Movement may occur over nearly the entire area of small drainage basins in deep cohesive materials and on slopes inclined at only a few degrees. The result is a continuing supply of soil to streams by encroaching streambanks. The quantity of soil delivered to streams is large. For example, a creep rate of 10 mm per year with a bulk density of 1.6 would deliver 64 metric tons per kilometer of stream channel per year. During high flow events, this material is carried into channels by undercutting and slumping of streambanks. Creep is an important contributor of sediment to the Eel and Mad Rivers in northern California which have the largest sediment discharges for basins of similar size in North America.

Minimizing Creep

Forests probably have the least control over creep movement compared to other erosion processes except where the depth of movement is shallow (less than 5 m). Since creep movement on areas where vegetation has been removed is driven mainly by increased amounts of soil water caused by decreased evapotranspiration, increased movement will probably continue for the 20 to 30 years required to establish another forest. Removal of water by vegetation or rerouting water away from creep areas by surface streams is the most effective control of mantle creep.

Slump-earthflow

Slump-earthflows occur in areas where stresses are great enough to cause discrete failures. Slumps occur as a rotational movement of a block of earth over a broadly concave slip surface. Slumps are often called rotational failures (Fig. 12). Simple slumps often cause little break-up of the moving material, but where materials break-up and flow downslope or form a series of gliding blocks, the movement is called earthflow. The term, slump-earthflow, is used in the Pacific Northwest because both features often occur together. Lower ends of earthflows typically terminate at stream channels. Transfer of earthflow material to stream channels takes place by shallow landslides (debris avalanches) or gullying, depending upon soil and vegetation conditions. Therefore, an active slump can cause general instability such that material can be delivered to streams by several other processes.

In the Pacific Northwest, slump-earthflows may range in area from 1 ha to 2-3 km^2. The zone of failure may occur at depths of several meters to several tens of meters below the surface. Scarps—upslope margins of the failure (Fig. 12)—occur where the rotating block of earth slips away from the original land surface. A series of scarps may be formed where there are multiple slip surfaces.

Slump-earthflows are poorly drained features. Ponded water (sag ponds) is commonly observed on the surface of rotating blocks and within earthflow masses (Fig. 12).

Geologic factors control the occurrence of slump-earthflows in the humid and wet portions of the Pacific Northwest. Deep cohesive soils and clay-rich bedrock are especially prone to slump-earthflow failure. Such failures are frequently noted where these materials are located on the slope beneath hard, competent bedrock units or caprock. Runoff from the caprock and flows within the bedrock units are often deposited directly onto the slump-earthflow. Vegetation has little influence where failures are more than several meters deep and beyond the reach of roots. Vegetation, however, may assist in stabilizing movement by dewatering the soil through evapotranspiration.

Many slumps in the Pacific Northwest appear to be inactive at the present time. But areas of active earthflow movement are common and can be recognized by fresh ground breaks caused by tension cracks and shear zones or by tipped and bowed trees. Movement rates vary from imperceptibly slow to more than 1 meter per day in extreme cases. Movement rate, however, may be extremely variable over an earthflow area with one part of the area moving rapidly while another part may exhibit no movement.

The occurrence of slump earthflows can be illustrated by surveys at sites in western Oregon and northern California. At the H. J. Andrews Experimental Forest in Oregon, 25.6 percent of the area underlain by soft incompetent volcanic tuff-breccia bedrock exhibits slump-earthflow landforms. Less than 1 percent of areas of the competent basalt and andesite flow rock shows evidence of slump-earthflows. At the Redwood Creek basin in California, 27.4 percent of the area shows slump-earthflow features, many of which are associated with clay-rich bedrock. Areas underlain by more competent rock ,are much less prone to slump-earthflow. Erosion by slump-earthflow can be expected in other areas of the Pacific Northwest underlain by incompetent clay-rich bedrock.

Minimizing Slump-earthflow Erosion

Construction activities involving excavation, filling, or rerouting of streams including the surface drainage from forest roads may have dramatic impacts on slump-earthflow erosion. Since a slump is a huge block of earth which, like a teeter-totter, rotates to assume a more stable configuration, a change in the balance of the block may induce further rotation. There are numerous examples of accelerated or reactivated slump-earthflow movement after construction of forest roads. Placement of large road fills on the head of a slump block or excavation of earth from the toe are common practices that have caused renewed movement of slump-earthflows (Fig. 12).

Stability [of slump blocks] is also affected by routing surface drainage onto the slump block or onto the earthflow below the slump block. Slumps are poorly drained, and sag ponds are evidence of poor drainage. Routing of additional water onto the block increases the mass of the block and also decreases the shearing resistance of material at the failure plane. Singly or in combination, the above errors of road location or design often trigger slump-earthflows that may have remained stable for many centuries. Rejuvenation of movement of small slumps can often be prevented by loading the toe with a road fill or rip-rap or by routing surface water away from the slump-earthflow and draining sag ponds by shallow ditches.

Although the effects of clearcutting on slump-earthflow erosion have not been studied, indirect evidence suggests that clearcutting may influence the stability of slumpy terrain. Because of the great depth of the failure zone, the strength of roots are thought to be ineffective. Hydrologic impacts, however, may be important. Increased availability of water owing to reduced evapotranspiration will cause increased water content of soils on a slump-earthflow. Results from clearcutting a small watershed at the H. J. Andrews Experimental Forest indicate that increased drainage from deeply weathered and clay-rich terrain including areas of slump-earthflow movement may be large and prolonged. Over an 11-year period since timber cutting was completed, increased annual drainage has averaged 40 cm and exceeded 50 cm on three occasions (Harr 1976b). On slump-earthflows, the increased available water is likely to be stored for longer periods and may contribute to increased rate and/or duration of earthflow movement during the wet season.

Debris Avalanches

Debris avalanches are rapid, shallow mass movement (Fig. 13). These planar failures commonly occur in areas of competent bedrock overlain by non-cohesive soils. Debris avalanches leave elongated scars in shallow drainage depressions, or they may leave spoon-shaped failure surfaces. Volumes of earth oved by this process can range from 1 to more than 10,000 cubic meters.

Debris avalanches are usually triggered by infrequent and intense storms which, on the average, occur at 5- to 7-year intervals along the north Pacific Coast. Mobilization of the soil is thought to be the result of the concentration of drainage water in shallow depressions in the bedrock or on surfaces of compacted glacial till. Because debris avalanches are shallow, factors such as strength given the soil by the root web, anchoring effects of roots in bedrock, and transfer of wind stress to the soil mantle by stems of trees are all potentially important influences. Factors that determine the water supply to the soil, such as snow distribution and the rate of rainfall and snowmelting, all exert control over when and where debris avalanches occur.

Although the movement rate of debris avalanches has never been measured, evidence from the few eye witness accounts and from the steepness of slopes where debris avalanches occur would indicate that velocities of up to 20 m/sec are not unreasonable. The rates of occurrence of debris avalanches vary depending upon the stability of the landscape and the frequency of storm events which trigger them. The annual rates of debris avalanche erosion in forested areas among four study sites in Oregon, Washington, and British Columbia (Swanston and Swanson 1976) have ranged from 11 to 72 m^3/km^2/year.

Debris avalanches are the most commonly occurring form of mass erosion on steep slopes where timber has been harvested. Impacts of harvesting activities can be estimated by comparing the rates of debris avalanching in clearcuts and road rights-of-way with the rate from forested areas during the same time period. These analyses indicate that clearcutting increases the natural rate by 2 to 4 times. Roads have produced even more profound increases in debris avalanche erosion. Among the four sites, erosion activity after road building was 25-340 times greater than the rate in forested areas. These increases in volumetric terms ranged from 282-15, 565 m^3/km^2/year (Swanston and Swanson 1976).

Although studies in the Pacific Northwest are too brief to assess the general level and duration of erosional activity from disturbed sites by debris avalanches, historical records from the H. J. Andrews Experimental Forest suggest that erosion may continue for at least 12-16 years. In addition, the evidence indicates that duration of debris avalanche erosion from road rights-of-way may persist for more than twice as long as that from clearcut areas, but the annual rates per unit area of land are more than 13 times greater from roads than clearcuts. Because clearcuts in the H. J. Andrews Experimental Forest cover about six times the area of those from road rights-of-way, the total soil loss by debris avalanching from roads has been about double that from clearcuts.

Minimizing Debris Avalanche Erosion

Measures for minimizing debris avalanching from roads include location, design, construction, and maintenance. On harvested areas, maintaining the strength given soils by roots is the only feasible means of control. Some degree of root strength in soils can be maintained by harvesting methods such as shelterwood, thinning, or narrow strip cuts.

High potentials for debris avalanching occur where excavated material from road construction is sidecast onto slopes steeper than 65 percent. This material is in tenuous balance–held by the frictional resistance with the slope, stumps, and brush beneath it. These overloaded slopes frequently fail when winter rains saturate the deposited material. If such steep terrain must be crossed by roads, the road right-of-way must be narrowed, and excavated material must be hauled to disposal sites on stable terrain.

Steep headwall basins are particularly susceptible to debris avalanching. Shallow soils and converging subsurface drainage make these landforms very unstable as evidenced by tension cracks, debris avalanche tracks, and exposed bedrock. Roads should not be located in these areas. Forests on headwalls are often unproductive and, therefore, better left for erosion control.

Surface erosion on debris avalanche scars often remains active for several decades. Continuing sheet and dry ravel erosion often prevents reestablishment of vegetation and natural control of soil movement on such slopes (Fig. 14).

Debris Torrents

Debris torrents are the rapid movement of water-charged soil, rock, and organic material down steep stream channels. Debris torrents typically originate in steep headwater channels. They are triggered during extreme runoff events either by debris avalanches originating on the slopes above the stream or by the breakup and mobilization of material deposited in the channel. As the mass moves down the stream much like a flood wave, the volume of the torrent is augmented by capture of organic and mineral matter stored in the channel. The moving mass may scour the entire stream channel to bedrock and the banks to a height of 6 m or more depending on the volume of the mass (Fig. 15A). In some cases, debris torrents triggered by less than 10 m^3 of material may ultimately move volumes in excess of 10 000 m^3. Torrents lose momentum as channel gradients lessen, and a tangled mass of unsorted sediment and large organic debris may be deposited over several hectares (Fig. 15B) or may disappear into the next larger stream.

The main factors controlling the occurrence of debris torrents are the quantity and stability of debris in channels, steepness of the channels, stability of the adjacent hillslopes, and the size of peak flows in the stream. Unfortunately, at this time we have only limited capability of predicting the risk of debris torrents in channels.

Figure 14. Revegetation of bare soil on landslide scars is often slower in clearcuts (A) compared to sites within the forest (B). Scars from debris avalanches may remain bare and susceptible to surface erosion for several decades.

Figure 14A-1. First season after a debris avalanche in a clearcut.

Figure 14A-2. Year 4 after a debris avalanche in a clearcut.

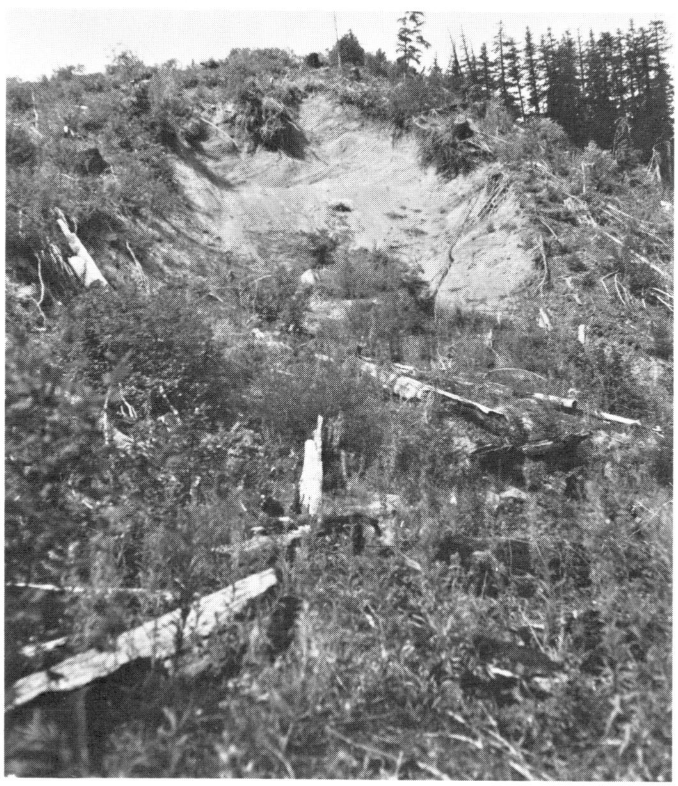

Figure 14A-3. Year 8 after a debris avalanche in a clearcut.

Figure 14B-1. First season after a debris avalanche in the forest.

Figure 14B-2. Year 4 after a debris avalanche in the forest.

Figure 14B-3. Year 8 after a debris avalanche in the forest.

Figure 15. Debris torrents (A) are often triggered by debris avalanches near the source of small streams. Although debris torrents are a natural process, frequency of their occurrence is increased by timber harvesting. Torrents scour all of the mineral and organic material deposited in channels to a height of several meters above the channel. Debris torrent deposits (B) are often destructive to capital investments such as roads and bridges. The value of streams for fish habitat is degraded when torrents are deposited directly into stream channels.

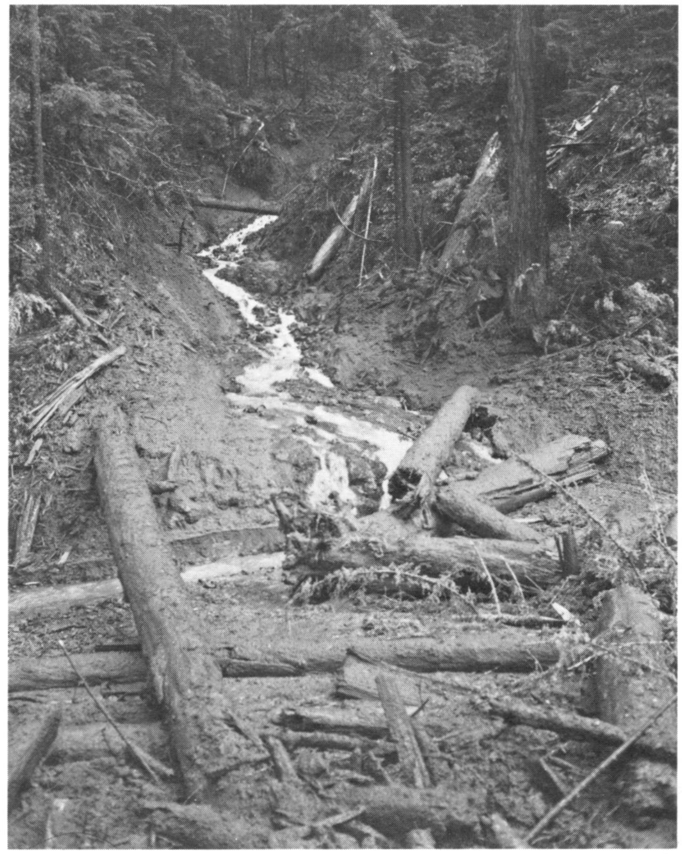

Figure 15A. Debris torrent channel with deposited material in the foreground.

Figure 15B. Debris torrent deposit.

Debris torrents have received little study in the mountains of the Pacific Northwest. Velocities of debris torrents are estimated at 20-30 m/sec. In the two small areas in western Oregon where debris torrent have been documented, annual rates of occurrence were 5-8 events per 1 000 km^2 from forested areas (Swanston and Swanson 1976). Because debris avalanches have played an important role in triggering debris torrents (83 percent of all inventoried debris torrents were triggered by debris avalanches), susceptibility of an area to debris avalanching is one indicator of potential for debris torrents in steep stream channels.

Timber harvesting activities dramatically accelerate the occurrence of debris torrents primarily by increasing the frequency of debris avalanches. Two other factors, increased debris accumulations in channels and increased peak streamflow, may also increase the occurrence of torrents.

The relative impacts of clearcutting and road building on debris torrents may be assessed by comparing the frequency of occurrence of debris torrents in natural forest settings with the rates of occurrence observed in clearcut areas and areas with roads. From a record of 15 and 25 years at the two areas, clearcutting was found to increase the occurrence of torrents by 4 and 9 times and roads were responsible for increases of 42 and 133 times.

Minimizing Debris Torrents

The history of debris avalanches at the two study sites clearly indicates that debris avalanches trigger most debris torrents and the occurrence of debris avalanches is greatly increased by road building and clearcutting. Therefore, reducing the number of debris avalanches will also reduce the number of debris torrents.

SOLUTION EROSION

Solution erosion is the process whereby bedrock is dissolved, and soluble materials are removed in drainage water. Rain and snowmelt water passing through the soil carries weakly dissociated organic acids. The acids arise from the breakdown of organic materials or are metabolic byproducts of organisms living in the forest or the soil. These weak acids decompose minerals by a process called *weathering*, and soluble chemical elements are transported from the site in drainage water. The insoluble materials left behind, called weathering products, are the parent materials from which soils are formed.

The rate of solution erosion is controlled by characteristics of climate and bedrock. Weathering increases with increase in temperature and volume of water flowing through subsoil and bedrock layers. Mineral composition of bedrock and accessibility to percolating water determines the potential for solution erosion. Of minerals common to the earth's crust, quartz is most stable, olivine is least stable, and the micas and sodium and potassium feldspars are intermediate. Accessibility of minerals to flowing ground water is determined by the porosity of the bedrock which in turn depends on the packing of elements of the rock during formation or subsequent fracturing of the rock.

Some examples may serve to illustrate the principles given above. A quartz-rich bedrock such as granite will weather very slowly even in the presence of an intense weathering environment because of quartz, a highly stable mineral. Basalt will weather at several times the rate of granite in the same environment because of a mineral complement that is less stable. But an andesitic tuff composed of loosely packed fragments of volcanic ash will weather several times faster than massive andesite flow rock of the same mineral composition because minerals in the tuff are more accessible to weathering agents.

SEDIMENT TRANSPORT IN STREAMS

Soil materials deposited in channels are transported downstream by flowing water. The capacity of a stream to move these materials is a function of the volume, velocity, and turbulence of the stream. The energy required to move deposited materials increases exponentially with increasing diameter of the material. Therefore, very small soil particles such as fine clays, silts, and fine sands remain suspended in high-gradient streams typical of mountainous regions of the Pacific Northwest. Coarse sands and small gravels bounce along the stream bottom, and large gravels and cobbles roll down the bed in contact with the bottom. Still larger angular objects may slide down the bed without rolling. Fine particulate fractions carried in suspension are termed suspended load and coarser materials on the bottom are called bedload.

The proportion of the bed materials and size categories in motion depends upon the energy of the stream. At low flow in winter or summer, streams carry only solution load. As streamflow increases, suspended load increases; and finally bedload movement occurs as streamflow rises to peaks which typically occur a few days every year. But the amounts of suspended load and bedload carried by streams also vary depending on the rate that eroded materials enter streams and their textural composition. Thus fine textured soils delivered by creep erosion will be transported as suspended load, while debris avalanche erosion of coarse materials such as glacial colluvium will be transported as bedload.

Sediment transport is modified by woody debris deposited in channels. Trees falling across streams collect floating organic debris, reduce the velocity of the stream, and trap sediment. As bedload accumulates behind debris jams, the level of the streambed rises and water falling over the downstream end of the jam dissipates energy and reduces the velocity of streamflow. The profile of first, second, and third order channels in western Oregon and Washington is often a stairstep of debris jams and plunge pools. (A first order stream has no tributaries, a second order stream has at least one first order tributary, a third order stream has at least one second order tributary, etc.) The riffles and pools behind debris jams make ideal spawning and rearing habitat for native and anadromous fish when the jams do not restrict fish passage.

Management of debris in streams presents a mix of benefits and risks. Improved diversity of habitat for biota associated with debris dams is clearly a benefit, but the volume of organic and bed material in channels heightens destructive effects downstream when channel material is incorporated in debris torrents. Destruction of natural channel banks, roads, and bridges is often the result. Because removal of large natural debris during clearcutting may deprive the stream of check dams and the stream habitat may be degraded for a century or more, stream debris management during logging should include leaving the natural debris in the channel undisturbed. Roads and bridges can sometimes be protected by allowing sufficient clearance in channels to allow debris torrents to pass through.

MEASUREMENTS OF TOTAL EROSION RATE

Studies of land-stream systems are conventionally made on small watersheds. Water leaving the basin is measured continuously, and solution and suspended loads are determined by sampling stream water. Bedload is collected and measured in special catch basins.

Study Areas

The rate that the sediment load moves through channels and the composition of the load varies with topography, composition of soils and bedrock, and climate. Erosion rates from three forested sites in the western Cascade Range of Oregon illustrate the variation in rate and form of transport (Fig. 16). The characteristics of the sites are given on Table 4.

The Fox Creek site (FC-2) is gently sloping, stable terrain underlain by a massive, competent and unfractured andesite. As a result, weathering agents have low accessibility to bedrock and coarse glacial colluvium in the soil mantle. Sediment supply to the channel is predominantly from streambank cutting and tree uprooting along streambanks. Annual erosion at this site is roughly 140 kg/ha, the lowest among the three sites. Transport is principally by solution and bedload, although the bedload value is approximate.

The H. J. Andrews site (HJA-2) is a steeply sloping watershed (average slope of 61 percent) with the stream channel deeply cut into incompetent welded tuff bedrock.

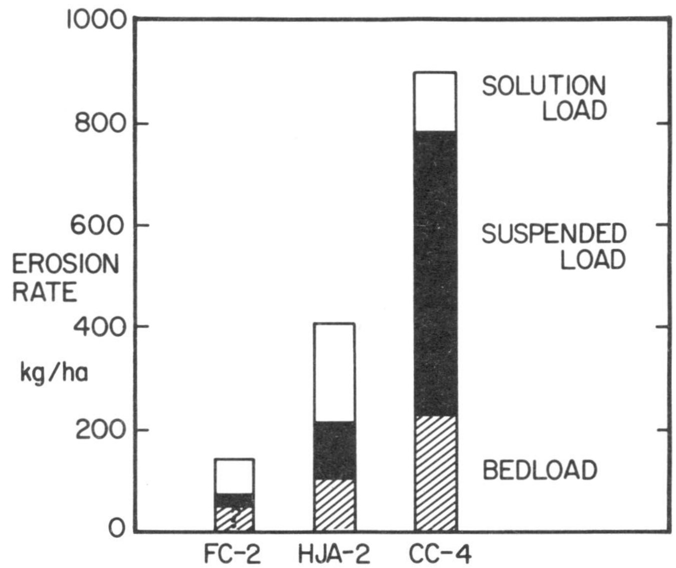

Figure 16. Mean annual erosion by solution, suspended sediment, and bedload from three small watersheds with undisturbed Douglas-fir forest cover. See Table 4 for description of the sites.

Debris avalanching is the dominant soil erosion process. The total annual rate of 410 kg/ha is nearly three times greater than FC-2. Half of the material is transported by solution erosion, and the remainder is nearly equally divided between suspended sediment and bedload.

The Coyote Creek site (CC-4) is less steeply sloping (average slope of 38 percent) compared to HJA-2, but has an annual rate of erosion (900 kg/ha) about 6.5 times that at FC-2. An indication of the size fractions of material supplied to the stream by mantle creep is the transport of 61-percent suspended load and 25-percent bedload. Accessibility of the bedrock to solution erosion is nearly equal to that of HJA-2, but the solution loss is less at CC-4 because there is less water to remove soluble materials from this drier site.

The erosion rate varies among the three sites because of different dominant erosion processes, the stability of the soil and parent materials, and the effectiveness of forest vegetation to control the rate of these processes.

Increased Suspended Sediment Load Due to Timber Harvesting

The increase in soil erosion among sites currently under study in western Oregon has been variable, and increases are due to a variety of causes. Soil erosion often increases as a result of timber harvesting and construction of logging roads.

The increase in suspended sediment yield (Fig. 17) has ranged from barely perceptible at Fox Creek (FC-1 and 3) to

Table 4. Characteristics of experimental watershed study sites in western Oregon.

Watershed Study	Code	Harvest Treatment	Roads	Average Slope	Parent Material	Mean Annual Streamflow	Dominant Erosion Process	Forest Type
		---------- Percent	----------	----------		cm		
Fox Creek	FC-1	25 PC, B	2[2]	7-12[3]	andesite (competent)	265	streambank	over-mature Douglas-fir western hemlock
(55 km E Portland, Oregon)	FC-2	0 undisturbed						
	FC-3	25 PC, NB	1					
Alsea (Needle Branch)	AL-1	82 CC, B	5	35-40	marine sand & mudstone (incompetent)	190	debris avalanching	mature Douglas-fir Red alder
(12 km S Toledo, Oregon)	AL-2	25 CC, B	4					
Deer Creek								
Coyote Creek	CC-1	50 SH, NB	5	23-38	non-welded tuff (incompetent)	70	creep and debris avalanching	mature sugar pine, Douglas-fir
(50 km ESE Roseburg, Oregon)	CC-2	30 PC, YUM	5					
	CC-3	100 CC, YUM	1					
	CC-4	0 undisturbed						
H. J. Andrews Experimental Forest	HJA-1	100 CC, B	0		welded tuff & breccia (incompetent)	105	debris avalanching	mature to over-mature Douglas-fir western-hemlock
(67 km ENE Eugene, Oregon)	HJA-2	0 undisturbed		53-63				
	HJA-3	25 PC, B	6					

[1]Percent harvested: harvest by CC-clearcut, PC-clearcut in patches, SH-shelterwood; residue removal by B-broadcast burning, NB-no burning, and YUM-yarding unmerchantable material.
[2]Percentage of the area in roads.
[3]Percent slope of the area.

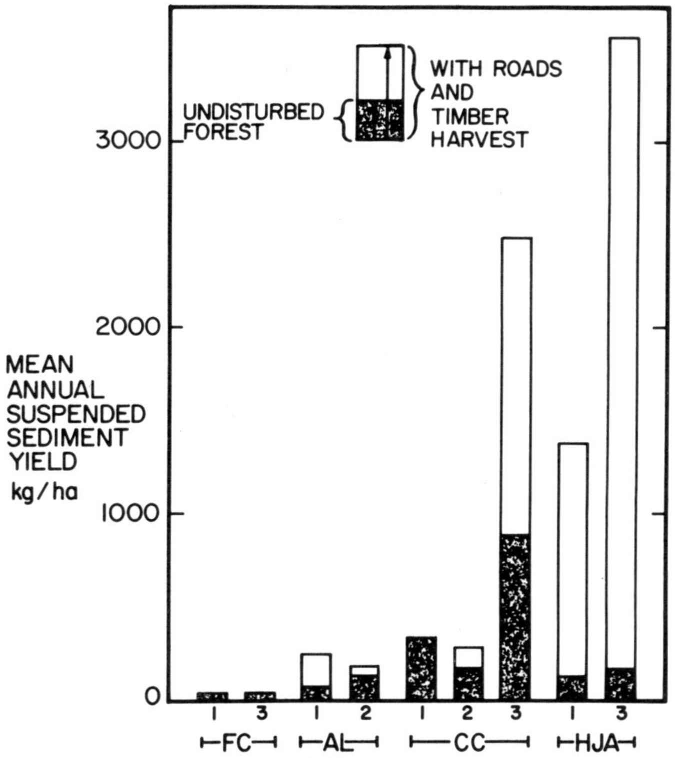

Figure 17. Mean annual suspended sediment yield from four sites in western Oregon. See Table 4 for description of sites and harvest activities.

nearly 23 times the natural rate in the H. J. Andrews Experimental Forest (HJA-3). In Figure 17, the general level of soil erosion increases with increasing average slope of the contributing basins from left to right. Also, the general level of erosion varies according to the erosion processes that deliver soil to streams and the occurrence of large runoff events from heavy precipitation and snowmelt. The greatest natural rates of erosion occur at Coyote Creek (CC) where mantle creep is the dominant process. Natural rates are notably lower where sources of erosion materials are debris avalanches such as at the H. J. Andrews Experimental Forest and the Alsea Study and least where erosion was from stable streambanks at Fox Creek.

Standards of road construction largely determine the rates of soil erosion from sites with high erosion potentials. At HJA-3, construction in 1959 was typical of the 1950's and 1960's; road fills were not compacted and soil was sidecast downslope—some of it reaching channels at the time of construction on slopes that frequently exceeded 60 percent. Also, in December 1964 soon after timber was harvested, these experimental watersheds experienced one of the most severe runoff events in their recorded history. About 81 percent of the material lost from HJA-3 over a 16-year period occurred in a 2-day period during this runoff. Roads were the source of most of this material. In CC-1 and CC-2, logging roads caused little erosion because of careful location, design, and construction even though the bedrock is deeply weathered and incompetent. The design called for compaction of subgrades and road fills and waste soil was transported to disposal sites. Taken together, these improved standards have undoubtedly contributed to the lack of erosion from these roads in CC-1 and CC-2. To date, however, these roads have not been impacted by a storm of the magnitude experienced earlier at the H. J. Andrews. Roads have not been a source of sediment to streams at the Fox Creek site.

Suspended Sediment Concentration and Water Quality

The suspended sediment concentration of streams closely follows streamflow as shown in Figure 18. Maximum concentrations often occur just before peak flows and minimum concentrations occur at low to moderate flows. The close correspondence of the hydrograph and the suspended sediment trace indicates that in many instances sediment transport is correlated with streamflow. By contrast, suspended sediment traces that do not follow storm hydrographs are often the result of soil mass movements deposited directly into the channel. Streams reach maximum concentration of 10 to 50-percent soil by volume when debris torrents pass down channels.

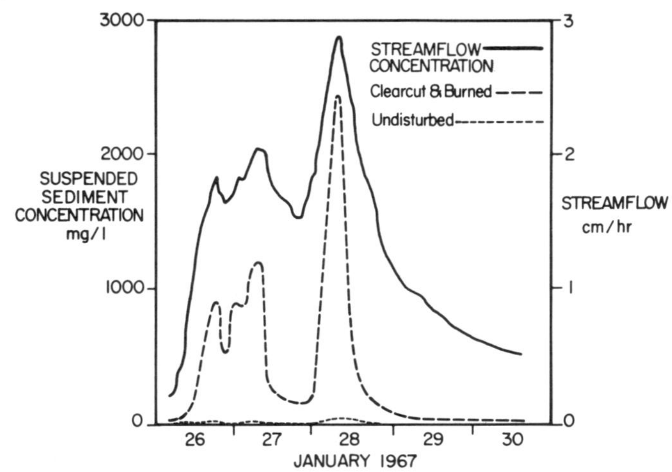

Figure 18. Storm hydrograph and suspended sediment trace for a clearcut watershed (HJA-1) where logging residue was broadcast burned and for the same watershed in an undisturbed condition, H. J. Andrews Experimental Forest. See Table 2 for a description of the site.

Clearcutting causes little soil erosion on gently to moderately sloping land surfaces with competent parent materials; but on steeper sites with higher erosion potentials with incompetent parent materials, much greater soil losses can often result. At Fox Creek, increased soil erosion was easily prevented because of stable terrain and competent bedrock. Soil losses from the Alsea watersheds (incompetent) were relatively small although slopes are relatively steep. But complete clearcutting, such as was done at HJA-1 and CC-3 (incompetent) (table 4), often cause much greater soil losses. Combined losses from roads and clearcutting on steep land not only may degrade stream habitat by deposition of sediment, but also such losses may eventually reduce productivity of forests on these landscapes. Harvest by shelterwood or very small clearcut (CC-1 and CC-2) can greatly reduce soil losses. Depletion of soil also may be reduced by growing forests on longer rotations on steep land such as that of HJA-1, 2, and 3.

There have been variable erosional responses to disposal of logging residue by broadcast burning. Where surface soils are well structured such as at the Fox Creek and H. J. Andrews sites, burning has had no discernible effect on soil loss (Fig. 17). But the principal source of soil lost from AL-1 was apparently surface erosion after an unusually hot slash fire. Such hot slash fires have been shown to break down soil aggregates (Dyrness and Youngberg 1957), and soil losses from surface runoff erosion are possible.

Increased suspended sediment concentrations are the result of increased sediment supply to the channel caused by timber harvesting (Table 5). As a point of reference in evaluating table 5, 10 mg/l of mineral sediment will not be noticed in a glass of drinking water. Increasing concentrations, however, soon make the water undesirable for such uses as drinking, fish rearing and spawning, and recreation. Where timber harvesting has increased soil erosion, increased sediment concentrations in the stream are also evident. The streams with the highest concentrations are also those that continue to carry the highest concentrations for the longest time among the sites (Table 5). Fox Creek water meets the highest criteria for water quality at all times, whereas sediment concentrations in the other study watersheds frequently exceed the 10 mg/l limit for drinking water. While undisturbed, Coyote Creek watersheds produced water of the poorest quality, but important increases in sediment concentration occurred only at CC-3. The greatest increases in concentration have occurred at HJA-1 and HJA-3.

Several general conclusions can be drawn from these studies. Examination of Figure 17 shows that forest vegetation effectively prevents soil erosion where the dominant process is debris avalanching. Examples are the Alsea and the H. J. Andrews Experimental Forest sites. Forest cover was much less effective in reducing erosion at Coyote Creek where erosion processes are earthflows and creep.

The time required for vegetation to stabilize soils appears to be related to the erosion potentials at the sites. Erosion potential generally increases with angle of slope but is also controlled by erosion processes. On gently sloping topography at Fox Creek with competent bedrock, there was essentially no increase in erosion after timber harvest. Sites with incompetent bedrock are often dependent on forest vegetation for this stability, particularly where debris avalanches predominate. At AL-1, a vegetation stand composed of shrubs and herbs effectively controlled soil erosion

Table 5. Mean annual and mean maximum concentration of suspended sediment in the streamflow from undisturbed forested watersheds and watersheds where timber was harvested. See Table 4 for description and location of sites.

Watershed sites	Site code	Duration of Increase	Mean: Undisturbed forest	Mean: After harvesting	Approximate Mean Maxima: Undisturbed forest	Approximate Mean Maxima: After harvesting
		yrs	mg/l			
Fox Creek	FC-1	0	0.9	1.1	4.9	4.3
	FC-3	0	1.0	1.2	4.5	5.8
Alsea Study	AL-1	5	4	10	300	1260
	AL-2	0	6	9	1000	1110
Coyote Creek	CC-1	0	44	41	150	155
	CC-2	1	25	35	100	200
	CC-3	6+[1]	137	261	400	1400
H. J. Andrews	HJA-1	13+	11	103	1000	4400
	HJA-3	18+	13	308	2100	>15000

[1]+ indicates that increased concentrations are continuing.

5 years after harvest. Soil erosion has continued at a rapid rate on the much steeper terrain at HJA-1 and HJA-3 and remains unaffected by a dense stand of shrubs and Douglas-fir regeneration 15 years after commencement of clearcutting in HJA-1 and HJA-3 and 18 years after road construction in HJA-3. The soil erosion by slump, earthflow, and creep has continued unabated on CC-3 for 7 years since clearcutting and stabilization of soils may be as long term as that observed at the H. J. Andrews site.

NUTRIENT BALANCE

Conifer forests of the Pacific Northwest are well organized to conserve essential nutrient capital received from atmosphere and bedrock. The nutrient capital is recycled by the forest to meet the forests' nutrient requirements. The quantity of essential nutrients lost from the cycle to streams is very small and much less than the forest receives each year in precipitation. But destruction of the forest vegetation interrupts the nutrient cycle, and nutrient loss to streams often remains increased until the forest is reestablished. There is much interest in this nutrient loss. The relevant questions being asked are: (1) do concentrations of nutrients increase in streams after timber harvesting, (2) are these concentrations toxic to living organisms, (3) are nutrient losses in streamflow important to productivity of streams or forest sites?

The Nutrient Cycle in Forests

The nutrient cycle is basic to all soil-plant systems. Nutrients are taken up from the soil by roots, returned to the soil in canopy drip, litterfall, and root sloughing–made soluble by decomposition, and taken up again. Early in the development of the stand, a large part of the nutrient capital in vegetation is retained in needle and limb biomass which in turn enters the nutrient cycle after senescence and litterfall. But as annual growth rings accumulate, more and more of the nutrient capital is retained in non-recyclable wood and bark. Without nutrient inputs, the cycle would run down because of the incorporation of nutrient capital in biomass (See also chapter XI).

The relative importance of the atmosphere and bedrock as sources of nutrients depends on whether the nutrient has a volatile phase. Nutrients with a volatile phase (C, N, S) come principally from the atmosphere whereas bedrock is the major source of those without a volatile phase (Na, K, Ca, Mg, P, Mn, Fe, Mo, B, Cu).

Input from the atmosphere is by two means, (1) in precipitation (including dry fallout) and (2) fixation directly from atmospheric gases. C is fixed by photosynthesis, N is fixed by micro-organisms (bacteria or algae) either free living in soil or living in symbiotic association with a host plant. Red alder (*Alnus rubra* Bong.) and snowbrush ceanothus (*Ceanothus velutinus* Doug. ex Hook.) are important symbiotic N fixers, and lichens in the forest also fix N. N is also present in precipitation, and precipitation is thought to be the principal source of S, although there are also several minerals in bedrock that contain S.

Weathering is the process that releases nutrients from the minerals in bedrock and converts them into soluble forms for entry into the nutrient cycle. The rate that earth-derived nutrients are released depends upon the weathering rate of bedrock which was discussed earlier in the section on solution erosion. Nutrients released by weathering are dissolved in soil water and brought into the nutrient cycle by root uptake.

Nutrients are stored in soil as components of organic matter, in soil organisms (fungi, bacteria, soil animals), and as ions on the surfaces of colloidal organic matter and clay minerals. Cations, Na, K, Ca, Mg, and trace metals, are absorbed to the surfaces of these organic and mineral colloids. Uptake by roots is accomplished by displacement from the exchange site by another cation released by roots or micro-organisms (see chapter VI). Nutrients stored in organic matter are released by decomposition.

Nutrients are removed from the nutrient cycle (outflow) by (1) leaching of soluble ions and organic matter in drainage water, (2) soil moved to streams by erosion, (3) volatilization caused by fire or micro-organisms, and (4) by timber harvest.

Leaching losses are most important for those nutrients which form mobile anions (nitrate, phosphate, sulfate) in soil solution. The same three nutrients are also transported in drainage water as components of dissolved organic matter. Nitrate (NO_3^-) is readily leached from the soil, but leaching of sulfate and phosphate ions is controlled by anion exchange or physical adsorption. See chapter XI for additional discussion of leaching of nutrients.

The source of NO_3^- is the soil ammonium ion (NH_4^+) and plant proteins. Decomposition of proteins liberates NH_4^+ which is oxidized to NO_3^- by nitrification, a process mediated by specialized soil micro-organisms. (See chapter VIII, a discussion of nitrification.) Leaching losses from Douglas-fir forests are reduced by the amount of precipitation lost to the atmosphere by evapotranspiration because this water is not available to leach nutrients from the soil mantle.

Nutrient loss by soil erosion is determined by the processes supplying soil to streams and the rates of those processes. Since most of the nutrient capital is stored in the surface horizons of soils, more nutrient capital will be removed there than an equal rate of mass erosion which commonly removes entire segments of the soil mantle instead of just the surface. Nutrient outflow by soil erosion may be more important on the east side of the Cascade Range where surface erosion is more common, than on the west side where mass erosion processes predominate.

Volatilization of C, N, and S is the result of fire or the slower action of micro-organisms. C and N are volatilized when organic matter is burned. Nitrogen and S are volatilized by microbial action under reducing conditions associated with poor drainage; but since most forest soils are relatively well drained in the Douglas-fir region, these losses of N and S are probably minor. Fire losses of N are very important in a burned forest.

Clearcutting and the Nutrient Balance

Timber harvesting affects the conservation of nutrients in four ways: (1) uptake by vegetation is virtually stopped; (2) living non-merchantable residue is converted to decomposable detritus; (3) decomposition of residue, dead roots, and forest floor is accelerated by a warmed and moistened environment; and (4) soil erosion is often increased. Removal of vegetation causes a concomitant decrease in evapotranspiration. Therefore, there is more water to leach soluble nutrients from decomposing organic matter and from soil.

The species composition and rate of growth of vegetation on clearcuts may determine the proportion of the nutrient capital lost after logging. Species composition is influenced by species composition prior to logging, and whether the residue is burned. Herb species such as fireweed (*Epilobium angustifolium* L.) are early occupants of burned areas together with sprouting shrubs (Fig. 9C). Nitrogen fixers that often follow burning on such sites are snowbush ceanothus and red alder. Surviving shrubs and herbs often make up the stand on unburned and lightly disturbed clearcuts, particularly on mesic and cool moist sites.

Nutrient outflows after clearcutting are influenced by soil drainage which is highest in winter when most precipitation occurs. Although cation (Ca, Mg, K, Na) outflows may increase up to 2-3 times after burning, such losses are of little consequence because cations are not known to limit forest productivity. N and P are of most concern because of their limited supply in soils and stream water and because of the potential for increased growth of algae and other aquatic plants in streams, lakes, and reservoirs after timber harvest. N and P are carried as dissolved ions, components of dissolved organic matter, and suspended sediment in streamflow.

Nitrate concentration in streamflow is closely related to soil hydrology. Maximum NO_3^- concentration is often measured early in winter after the soil water storage capacity has been exceeded by fall rains (Fig. 19). Minimum concentrations occur in the dry summer months when soil drainage is at a minimum and active biological communities in streams utilize the small amounts of NO_3^- which do enter streams.

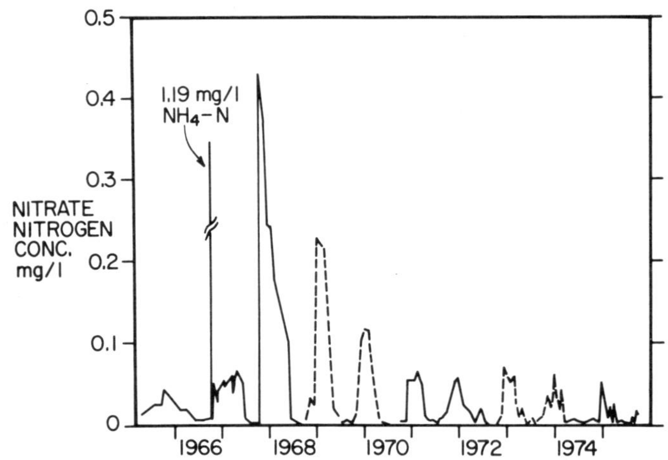

Figure 19. Nitrate and ammonia concentration in streamflow following clearcutting and broadcast burning of logging residue at the H. J. Andrews Experimental Forest (HJA-1). Dashed lines are estimated data.

Decline of NO_3^- in stream water as clearcuts revegetate results from dwindling supplies of readily decomposable logging residues and litter and from increasing uptake of nutrients by vegetation. Where revegetation is most rapid, as on the Oregon and Washington coasts, increased NO_3^- outflow from clearcutting may not persist for more than 5 years. On drier sites in the western Cascade Range, increased NO_3^- outflow may persist more than 15 years after clearcutting (Figure 19). No information is available for high elevation forests in the Pacific Northwest.

The concentration of dissolved organic N increases after clearcutting, but the increases in concentration are much less than those observed for NO_3^-. Dissolved organic N concentrations generally follow the pattern of NO_3^- outflow (Figure 19) and are undoubtedly also controlled by the supply of readily decomposable organic matter.

The greatest N outflows are caused by soil erosion from clearcut areas with high soil erosion potentials. Such large N outflows have been observed from two sites in western Oregon, the HJA-1 and CC-3 watersheds described in table 4.

The concentration of P in streams often increases after clearcutting, but the outflows are small compared to the available supply of P in soils.

The Nitrogen Balance and Clearcutting

The N balance is most important in the Pacific Northwest because N most often limits the productivity of Douglas-fir and other conifer forests. The N supply for the next generation of Douglas-fir forests depends in part on the management of the logging residue. Although the soil usually contains 80 to 90 percent of the total N capital in Douglas-fir forests, only a small part of the soil N is actively cycled. Most of the N in soil is held in stable forms which are resistant to decomposition. Nitrogen cycled from the residue should be more than ample to satisfy the N requirements of vegetation in the new stand as it did for the previous forest. Between 300 and 700 kg N/ha of logged forest should be made available by decomposition. If the residue is burned to reduce the fire hazard, a large part of the N capital will be lost by volatilization (Fig. 20).

Over the eons, Douglas-fir forests have been destroyed periodically by wildfires and symbiotic fixation of N has been a necessary adaptation to replace the N volatilized by fire and removed in streamflow. Measured annual fixation rates for red alder have varied from 4 to 325 kg/ha (Newton *et al.* 1968) and rates for snowbrush ceanothus from 70 to 110 kg/ha (Youngberg and Wollum 1976). Both species have natural life spans of more than 25 years; and more N can probably be fixed in this time than was lost by volatilization, soil erosion, and leaching (Fig. 20).

Chemical herbicides, commonly used to reduce the competition of shrubs and herbs with conifer seedlings, are equally effective on N-fixing trees and shrubs. Thus, effective N_2 fixation may not occur on stands treated with herbicide (Fig. 20). Increased growth of tree seedlings from release of competition may be followed by eventual decline in growth

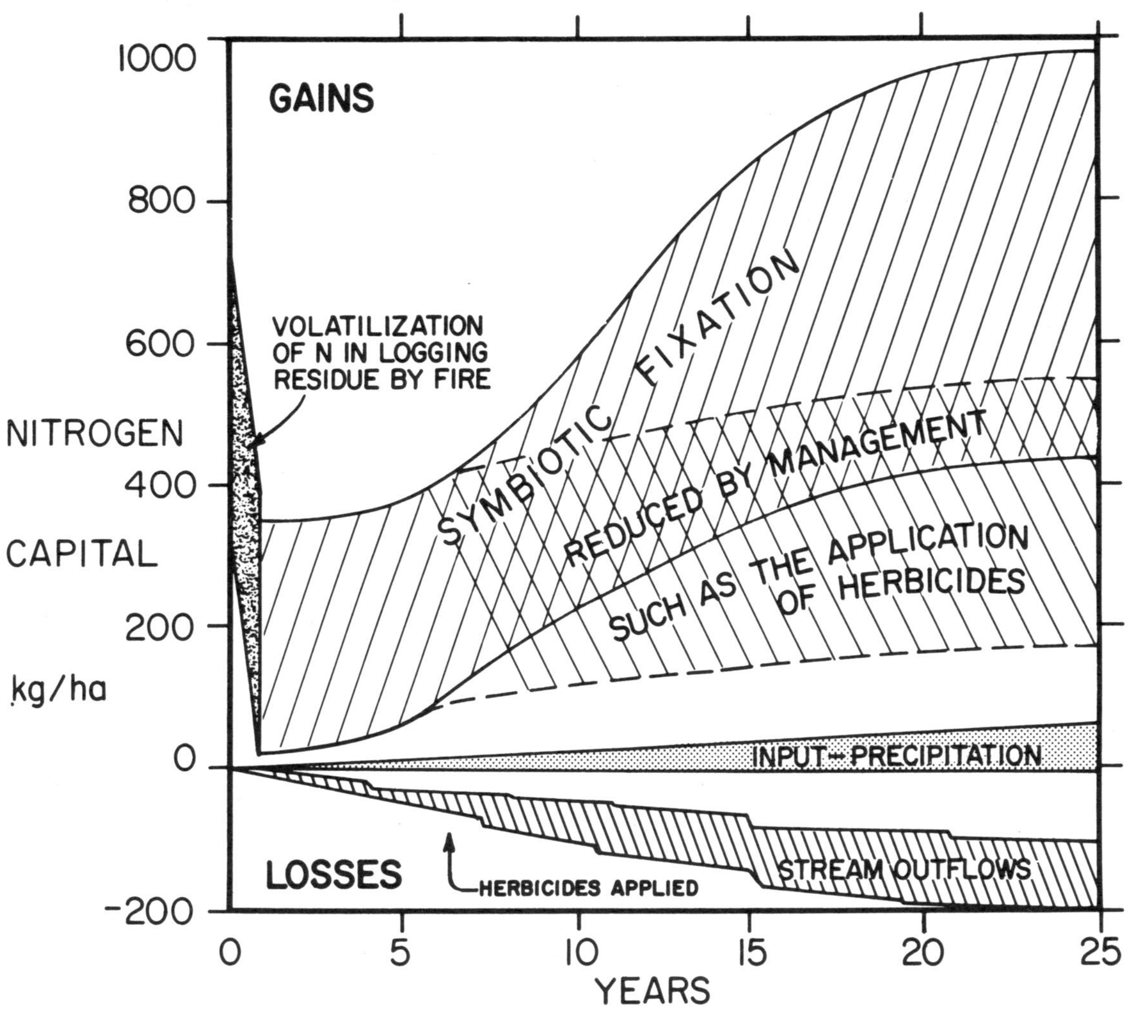

Figure 20. Idealized nitrogen balance for Douglas-fir clearcuts showing N volatilization losses caused by burning of logging residue and replacement by symbiotic fixation of volatilized N by snowbrush ceanothus or red alder. Fixation is capable of replacing both volatilization and stream outflows caused by leaching and soil erosion. N fixation will be reduced by practices such as application of herbicides which reduce the biomass of N fixers.

caused by N deficiency later in the life of the stand. Application of chemical fertilizer may be necessary to sustain growth. Much more research is needed on the management of N capital in forests of the Douglas-fir region.

Nitrogen outflows probably rarely exceed 200 kg/ha in 25 years, even in steep mountainous topography where soil erosion rates are high (Fig. 20). These outflows (reduced by the input from precipitation) probably range from 50 to 140 kg/ha in 25 years. Such rates of N outflow will not seriously reduce the N capital (Fig. 20).

Nutrient outflows are much lower with timber harvesting methods other than clearcutting. Nutrient outflows were increased only slightly or not at all by shelterwood cutting on one stand (CC-1) and removal of one-third of the volume of another stand in small (0.5-to 14.-ha) clearcuts (CC-2). (See Table 4 for a description of the sites.) The overall effect on nutrient outflows must await completion of the cutting cycle on these experimental stands.

Nutrient Enrichment and Water Quality

Water quality can be impaired if the concentration of chemicals or nutrients is raised to toxic levels, or if increased concentrations raise the productivity of aquatic algae and plants to noxious levels. Toxic concentrations of nutrients and heavy metals have been found only where logging residue was burned in stream channels (Fredriksen, 1971). Toxic concentrations of dissolved NH_3 and Mn probably do not persist for more than a few weeks after burning. These toxic concentrations can be prevented by not burning logging residue in live stream channels. Increased growth of aquatic plants, not a common occurrence in western Oregon and Washington, may be of concern in the eastern portions of the two states.

Nitrogen and P often limit the growth of aquatic algae and plants although trace metals may also limit growth. Leaching of forest or agricultural soils often increases supply of these nutrients. The impact of forestry practices will be of concern mainly in drainage basins which also serve as sources for municipal water supplies. Aquatic algae may affect the taste and clarity of the water for human use.

The effect on aquatic plants of increased N and P loading of streams after clearcutting of forests should be minimal because the loading occurs in the cool seasons of the year when the energy for photosynthesis is minimal (Fig. 19). In the warm seasons, the relatively low levels of N and P released from the soil, are utilized by vegetation and algae living in the bed or banks of streams in clearcuts. Because lakes in forest settings are often deficient in N and P, relatively small increases in P and N from clearcutting may be beneficial by raising the level of algae production. The production of higher forms of life, including fish, may be raised as a consequence.

STREAM WATER TEMPERATURE

Clearcutting frequently increases stream water temperature. The source of heat is the increase in direct solar radiation reaching the stream after the removal of shade provided by the forest.

Elevated stream temperatures can be detrimental to populations of resident trout and anadromous fish. Although temperatures above 25°C may cause mortality, particularly for fish in juvenile and embryonic stages, reduced growth, vigor, and resistance to disease are probably the main effects of high temperature. Elevated water temperature can stress salmon and trout since, as the temperature rises, the amount of oxygen that the water can hold declines. At the same time, the oxygen requirement for the respiration of the fish increases. Because fish must migrate to cooler water to survive, the result of severe increases in temperature of streamwater is loss of habitat for juvenile fish.

Water-temperature increases have ranged from 16°C following slash burning and stream clearance in a slow-moving portion of a coastal stream in Oregon to no detectable increase. Logging debris and understory vegetation left after logging can provide sufficient shade to prevent appreciable increase in temperature. In one case, water temperature of a stream increased 7°C after logging residue and peripheral shade were removed by a debris torrent caused by a debris avalanche. In a nearby stream flowing through an unburned clearcut, residual vegetation and logging residue over the stream kept the water temperature increase to 2°C. After slash burning, water temperature increased 8°C in this stream. In the Oregon Coast Ranges, the maximum temperature measured after logging and slash burning was 29°C, 15°C above the predicted maximum for this stream with forest cover.

Maximum water temperatures decline as clearcut areas revegetate. Environmental conditions are conducive to rapid revegetation in the humid, temperate climate west of the Cascade Range in Washington and Oregon. Near the Oregon coast, a mean monthly maximum temperature increase of 8°C decreased to 3°C 2 years later because of the regrowth of vegetation over the stream.

Controlling water temperature through forest management requires the ability to predict increases in water temperature that would result from a range of management practices. The temperature rise ($\triangle T$) expected from harvest operations can be predicted from the stream surface area (A) receiving solar heat (H), and the stream discharge (D). The equation is $\triangle T = CAH/D$ where C is a constant. (For details concerning the use of the equation see Brown 1971.) The general relationship among the variables indicates that water temperature will increase in proportion to the area of the stream surface exposed, and that the temperature increase will be inversely proportional to stream discharge. When streamside vegetation is managed to provide optimum habitat for fish, the maximum temperature attained may be more important than the temperature rise. Where ground water temperatures are near the temperature of melting snow, a temperature rise of $10^{o}C$ would be less important than in a rain-fed stream at lower elevation where normal water temperatures are higher.

Salmon is an important resource in the Pacific Northwest, where an acre of stream suitable for spawning and rearing salmon has been estimated to equal or exceed the value of an acre of timber. Unfortunately, the land base for spawning and rearing fish has shrunk because of the construction of high power dams. Water temperature can be controlled to improve the habitat for these anadromous fish by managing streamside vegetation to provide shade for streams. A brush cover of vine maple, salmonberry, or elderberry may be sufficient cover on narrow streams; but on wider streams, a buffer strip of trees is often required. Buffer strips must be windfirm, for windthrow of trees along channel banks may cause a secondary impact on fish habitat from excessive sediment loading. Where no shade remains after clearcutting, solar heat may be reduced by planting rapidly growing vegetation such as alder, cottonwood, or various species of willow along the stream margins.

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