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Influence of Forest and Rangeland Management on Anadromous Fish Habitat in Western North America

IMPACTS OF NATURAL EVENTS

DOUGLAS N. SWANSTON



ABSTRACT

Natural events affecting vegetative cover and the hydrology and stability of a stream and its parent watershed are key factors influencing the quality of anadromous fish habitat.

High intensity storms, drought, soil mass movement, and fire have the greatest impacts. Wind, stream icing, and the influence of insects and disease are important locally.

Keywords: Anadromous fish habitat, hydrology, stability.

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PREFACE

This is the second in a series of publications summarizing knowledge about the influences of forest and rangeland management on anadromous fish habitat in western North America. This paper addresses the effects on fish habitat of naturally occurring watershed disturbances and sets the scene for future discussions of the influences of human activities.

Our intent in presenting the information in these publications is to provide managers and users of the forests and rangelands of western North America with the most complete information available for estimating the consequences of various management alternatives.

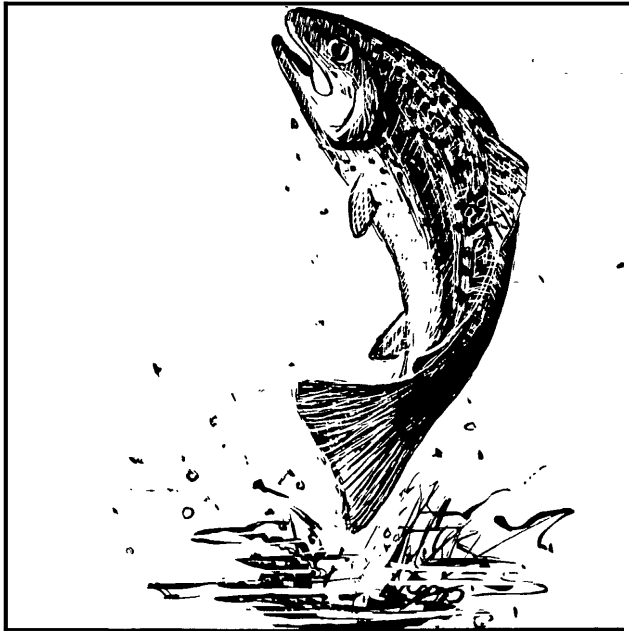
In this series of papers, we will summarize published and unpublished reports and data as well as the observations of resource scientists and managers developed over years of experience in the West. These compilations will be valuable to resource managers in planning uses of forest and rangeland resources, and to scientists in planning future research. The extensive lists of references will serve as a bibliography on forest and rangeland resources and their uses for western North America.

Previous publications in this series include:

1. "Habitat requirements of anadromous salmonids,"
by D. W. Reiser and T. C. Bjornn.

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Not all of these impacts are necessarily damaging to the habitat. For example, movement and redistribution of gravel by scour and streambed overturn may flush finer sediment from the gravels, improve access to gravel and intragravel waterflow, and create larger areas of usable spawning gravel. Stream-channel scour and excess streamflow may also remove blocking organic debris and improve access. The actual effect on habitat acceptability is largely dependent on peripheral damage to other habitat variables and on timing. Scour and streambed overturn after spawning can effectively destroy an entire generation by burial and mechanical grinding of eggs and alevins.

INTRODUCTION

Acceptable habitat for anadromous salmonids is the product of interaction among the geologic, climatic, and vegetative factors that control channel gradient and configuration, bottom composition, base flow, quantity and timing of maximum discharge, sediment and organic loading, and ground and overstory vegetative cover.

Under "normal" or "average" climatic conditions, those factors are in virtual equilibrium, and acceptable habitat is maintained with few or no short-term changes. Any alteration in this normal condition, however, has the potential for initiating events that may drastically alter the state of equilibrium. The result may be removal or redistribution of spawning gravels, addition of substantial quantities of sediment and organic debris, alteration of access, destruction of viable eggs and alevins, removal or redistribution of food organisms, and increased temperatures from removal of streamside vegetation.



HYDROLOGIC IMPACTS

Seasonal and short-term changes in weather and hydrologic conditions and the amount and distribution of vegetative cover are major factors controlling streamflow and occurrence of natural events that may be damaging to anadromous fish habitat.

BASIC HYDROLOGIC PROCESSES

The quantity and timing of streamflow in anadromous fish spawning channels is largely determined by water input and the hydrologic processes operating within the contributing watershed. Harr (1976) clearly describes these hydrologic processes for small forested watersheds.

Water is introduced into the hydrologic cycle as rain, atmospheric moisture (fog), or snow. Rain falling on a watershed may reach the ground and stream surfaces directly or be intercepted by vegetation (fig.1). Some of

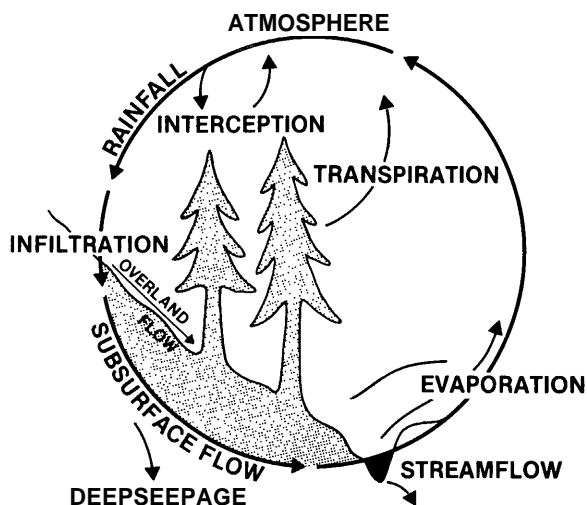


Figure 1.--Diagram of the rain-dominated hydrologic cycle (from Harr 1976).

this intercepted water evaporates; the remainder reaches the ground as stem-flow and drips through the canopy and leaf cover. The relative amount of rain "lost" to evaporation or delayed in its passage to the watershed by this interception process depends on the amount and extent of vegetative cover and the intensity and duration of the storm producing the rain. In the humid, heavily vegetated, old-growth forest areas within the influence of the Pacific storm systems, interception losses from rainfall are significant in the spring, summer, and fall when major storms are infrequent. Interception losses are substantially reduced during the winter when evapotranspiration is low

and frequent high-intensity, long-duration storms occur. For example, in old-growth Douglas-fir forests in the western Cascades of Oregon, Rothacher (1963) found that nearly 100 percent of storms less than 0.13-cm (0.05-in) was intercepted and evaporated, but less than about 5 percent of storms of more than 20 cm (8 in) was intercepted and evaporated. In the drier interior areas of the West, annual rainfall is low, and high-intensity, long-duration storms are infrequent, trees are less closely spaced, and total vegetative cover may be sparse. Under these conditions, interception losses are substantially less, ranging from 5 to 14 percent of total annual rainfall (Anderson et al. 1976), but are significant in controlling the total amount of water entering a watershed. When major storms do occur, interception losses become negligible.

Snow falling on the watershed is subject to the same interception as rain and in about the same proportions (Dunford and Niederhof 1944, Rowe and Hendrix 1951, Sartz and Trimble 1956, Hart 1963). In addition, where substantial snow packs develop, water may be detained for considerable periods in its passage through the watershed and into the stream system. Snow packs contribute to such "surface storage" both in the frozen phase and as free water held in the pore spaces (Anderson et al. 1976). The volume and duration of detention depend on such conditions as snow depth, air temperature, pore size and initial free water content (Corps of Engineers 1956, Smith and Halverson 1969, Smith 1974). Greatest detention is in the warm snow packs of the Sierra Nevada and the Pacific slopes of the northern Rocky Mountains where most of the water is detained over the winter and released when the snow melts. In the Cascade and Coast Ranges of

northern California, Oregon, Washington, British Columbia, and Alaska, although most precipitation occurs as rain, snow is common--particularly at higher elevations. Occasionally, a snow pack may remain for 1 to 3 months, but in most years it usually melts within 1 to 3 weeks. The highest runoff in this region has resulted from rapid snowmelt during warming trends coupled with prolonged heavy rainfall (Waananen et al. 1971, Harr et al. 1979). Watershed vegetation delays snowmelt by shading and thus extends the time that snow is held in surface storage. It also controls, to a certain extent, the rate and timing of snowmelt. Although snow under forest cover melts later and persists longer than snow in the open, it may melt more rapidly once melting begins, because melting begins later in the season when temperatures may be much higher (Anderson et al. 1976). A mixture of vegetated and open areas on a watershed may promote snowmelt at different times and reduce the total quantity released at any one time. Topography also desynchronizes melting; snow on southerly aspects may disappear before much of the snow melts on northerly aspects.

Once the water reaches the watershed floor, it infiltrates into the soil. Depending on the difference between the rate of water arrival at the soil surface and the soil's infiltration capacity or ability to allow water to enter, some water enters the subsurface and some may become overland flow. In undisturbed forests, infiltration rates of soils generally far exceed the maximum rates of rainfall so that all water enters the soil. Overland flow occurs under such conditions of vegetation, primarily as the result of limited capacity for storage, percolation, and channel expansion. Overland flow may also result from rapid addition

of stored water (melting snow pack) coupled with rainfall, which together may exceed the infiltration capacity of the soil.

Upon entering the soil, water is subject to gravitational and capillary forces that cause it to move and frictional forces that tend to restrict movement. Because of the slope of most watersheds and because soil conductivity generally decreases with depth, water entering the soil begins to move downslope as it moves deeper into the soil. The direction and rate at which the water moves depend on rainfall rates and soil properties. Both rate and direction vary considerably over the course of a storm (Harr 1977). Maximum velocities of soil water flow are low and frequently about equal to the average rate of rainfall during a storm. This slow-moving soil water is subject to evaporation and depletion by plants through the process of transpiration. The rate at which plants withdraw water is largely a function of energy available for water vaporization in leaves and the availability and ease with which water may be withdrawn from the soil. Thus, during the spring and early summer when soil water content is high and the growing season is at its peak, evapotranspiration withdrawals are high. If no additional water is added to the soil as the growing season progresses, soil moisture decreases and the remaining water becomes more tightly held by the soil. The result is a decrease in subsurface flow to the channel and a net reduction in streamflow. This condition is alleviated after the first storms in the fall recharge the soil water deficit.

In simplified terms, streamflow--on an annual or longer basis--is the difference between precipitation and evapotranspiration losses (Harr 1976). Soil moisture storage changes from time to time. Some water may

also seep deep into the subsoil and bedrock and not appear as streamflow in a small headwall basin. Generally, however, water not removed by plants ultimately moves downslope as saturated or unsaturated flow to supply streams.

As the watershed responds to rainfall, streamflow increases to a maximum known as "peak flows." Each storm hydrograph has its own peak flow, reflecting the interaction of rainfall with the physical characteristics of the watershed. High, sustained rates of rainfall contribute to greater storm runoff and higher peak flow. The magnitude of the increase in streamflow between the start of storm runoff and the peak is highly variable. It depends on the antecedent moisture content of the watershed's soils and the characteristics of the storm. Increases in streamflow of at least two orders of magnitude are not infrequent between the start and the peak of storm runoff. Maximum peak flows have resulted from rain on snow, during which a substantial portion of streamflow comes from rapid snowmelt concurrent with the downslope movement of water.

The quantity and rate at which water reaches the channel and passes through a watershed 'system during a particular hydrologic event is influenced by storm and watershed size and certain topographic considerations. Obviously, the larger a storm, the greater is the amount of water going into the system and the larger the potential streamflow. The influence of vegetation on streamflow resulting from small storms is greater than that from large storms because interception and evapotranspiration account for a larger proportion of small rainfalls, and soil water retained against gravity by plant roots accounts for a greater proportion of water entering the soil. As rainfall increases,

these withdrawals become less important. Stormflow resulting from extreme events is minimally influenced by these withdrawal processes, although a forest cover does detain some portion of any rainstorm and thus somewhat reduces flood discharge and peak flows.

Watershed size influences the quantity of streamflow and the size and timing of peak flows during any particular storm. Generally, the smaller the watershed, the more rapid are the streamflow increases in response to rainfall. For example, in small watersheds on the H. J. Andrews Experimental Forest in western Oregon, maximum rates of runoff have approached 80 percent of the average rate of rainfall for the previous 12 to 24 hours and 75 percent of the maximum 6-hour rainfall (Rothacher et al. 1967). As watersheds increase in size, total water yields increase, but peak flows and response time to storm events become somewhat reduced.

Quantity and timing of local stormflow are also related to elevation. Higher elevation watersheds generally receive a larger quantity of water per storm. Because these watersheds are small, soil water retention and evapotranspirational losses are less influential and runoff and peak flows tend to be higher than in the larger, lower watersheds. Studies in North Carolina (Hewlett 1967) showed that forested primary ridges at 1 524 m delivered almost 457.2 mm of direct runoff per year, but forest land at lower elevations delivered only 63.5 mm. Rainfalls of 213.4 to 274.3 mm in a December storm on three watersheds above 914.4 m produced 127.0 to 223.5 mm of direct runoff and maximum peaks of 68-167 m³/s. The rainfall on three watersheds below 91.4 m was 172.7 to 177.8 mm; direct runoff was 40.6 mm to 58.4 mm and peaks were 22 to 32 m³/s (Hoover and Hursh 1944).

Perhaps the most important concept in understanding the

hydrology of watersheds is the variable source area of storm runoff described by Harr (1976). This concept relates storm runoff to a dynamic source area that expands and contracts according to rainfall characteristics and the capacity of the soil mantle to store and transmit water (Hewlett and Nutter 1970). Thus, as a storm progresses, the channel network expands to many times its perennial dimensions (fig. 2),

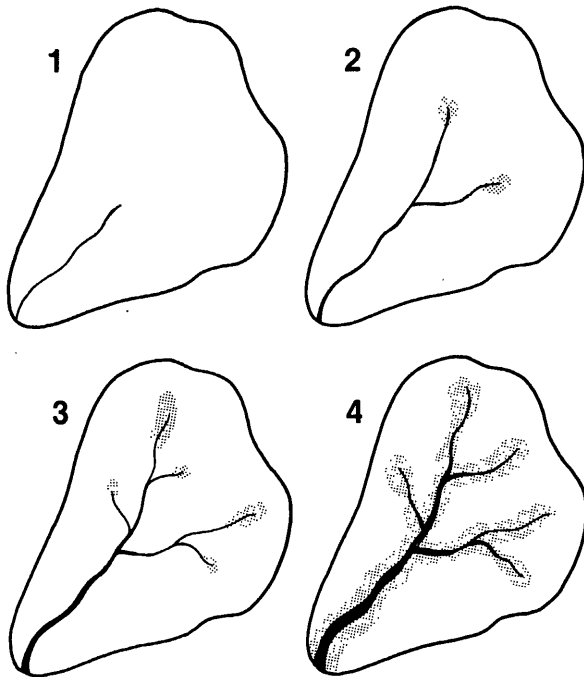


Figure 2.--Time lapse view of a small watershed showing the expansion of the channel network and source area of storm runoff (shaded area) (from Harr 1976).

and streams become both longer and wider. As a result of the variable source area of streamflow, both quantity and quality of streamflow can change drastically over a given period because the proportion of a watershed active in streamflow production changes. For sediment and organic debris, the stream represents a depository of variable surface area as well as a mode of transportation. During extreme runoff, sediment stored along the channel and debris not falling directly into the channel of the permanent stream may be subject to transport as the channel system expands.

MINIMUM FLOWS

Monthly precipitation and corresponding monthly streamflow vary greatly by season in the mountainous regions of the Northwest. The heaviest rainfall usually occurs from high-intensity storms during the late fall and winter in the Pacific coastal mountains and during the late spring along the west slope of the Rocky Mountains.

Extended dry weather is infrequent in the northern portions of the Pacific coastal mountain system (Washington, British Columbia, and Alaska). In these areas, rainfall is sufficiently frequent throughout the year to maintain adequate streamflow even during the summer. Elsewhere in the forested regions of the Northwest, extended summer drought is common; most of the annual precipitation occurs during the late fall and early spring storms. In such areas, minimum streamflows occur during the late summer and early fall and may be 1,000 to 5,000 times smaller than maximum peak flows in winter. Flow may cease entirely in many first-order streams (Harr 1976).

During low-flow periods, soil moisture is at its lowest because water has been removed by evapotranspiration and slow subsurface drainage to streams. Storms are infrequent and small; a large proportion of rainfall, when it occurs, is intercepted by forest vegetation and evaporates, with little water reaching the soil. Any storm runoff results almost entirely from channel interception, and peak flows are extremely small. The net result is that streams have little capacity to move sediment or transport debris. Streamflow is maintained by slow drainage from local ground-water sources or isolated saturated zones. Streamflow is at a very low rate, and stream network divisions are at annual minimums. First-order channels have no flow, and higher

order channels are short, with flow occupying only a small part of their widths. Flow may consist of a slow trickle between relatively isolated pools. Under these conditions, large areas of stream gravel are exposed; intra-gravel water movement is reduced and limited to considerable depths below the surface. During droughty periods, the potential for fire in the surrounding vegetation is also greatly increased,

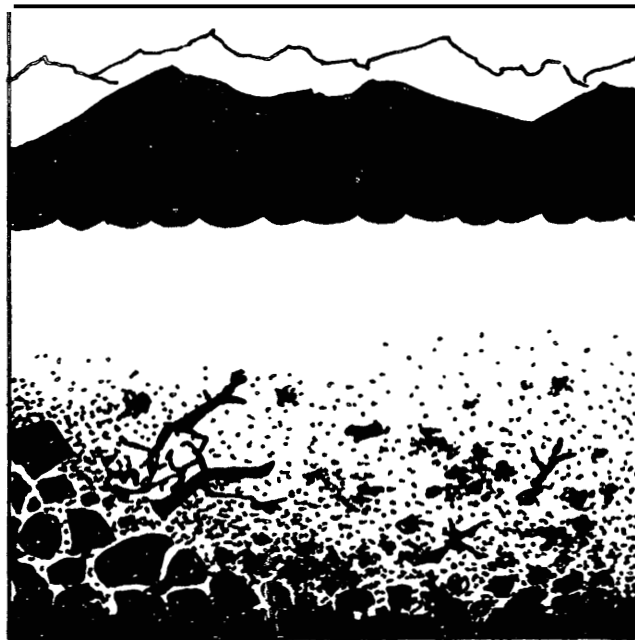
MAXIMUM FLOWS

Storms of high intensity or long duration in undisturbed watersheds cause increased peak flows and local flooding,, frequently resulting in accelerated mass soil movement, streambank erosion, and surface erosion of nonvegetated areas with attendant increases in sediment discharge to the channel 1/ (Rothacher 1959, 1973; Anderson 1975; Fredriksen et al. 1975). Patton and Baker (1976) have defined flash flood or high peak-flow potential in drainage basins by a regional indexing technique computed as the standard deviations of the annual maximum streamflows, Basins with high flash-flood potential tend toward greater relief and greater drainage densities combined with steep hillslopes and stream-channel gradients, These are characteristics of the majority of the Pacific Coast anadromous fish stream and river systems,

In the heavily forested Pacific coastal mountain systems, highest peak flows come during the late fall and winter as a result of heavy rainfall on wet soil mantles, At lower elevations, these peak flows are

1-/Unpublished report, "Forest practices and streamflow in western Oregon," by R. D. Harr. Paper presented at Symposium on Watershed Management, Am. SOC. Civ. Eng. , Logan, Utah, 1975.

produced almost entirely by high-intensity rainstorms, Within the snowpack zone, both rain alone and warm rain on snow associated with warming temperatures are important generators of peak flow. In western Oregon, peak flows during major winter storms may be as much as 100 times greater than flows immediately before such storms.1/ Comparable increases can be postulated for the rest of the Pacific coastal mountain belt based on the similarity in intensity and frequency of storms throughout the region. Along the west slopes of the Rocky Mountains, peak flows develop during the spring as a result of melting snow and occasionally as the result of upslope or convective storms, which dump large quantities of rain in the foothills during May and June (Anderson 1975). If snowmelt and storm rainfall are synchronized, substantial increases in peaks and accompanying flooding are produced (Goodell 1959).



SEDIMENT AND ORGANIC DEBRIS LOADING

Heavy sedimentation of streams and channel damage from flood flow and soil mass movements are closely linked to the pulsing action of streamflow caused by fall and winter storm fronts from the

Pacific Ocean and by the rapid snowmelt and upslope convective storms generated in the interior mountain areas. Streams rise and fall rapidly in relation to precipitation, storm duration, snowmelt, and the amount of water already in the soil.

The suspended sediment content of stormflow generally lags behind the storm hydrograph, with the highest sediment concentrations in transport occurring as streamflow increases to its peak. Sediment transported during these stormflow periods can be substantial. Mean annual concentrations of sediment from undisturbed watersheds are relatively low. For example, Fredriksen et al. (1975) reported mean annual concentrations of sediment for undisturbed watersheds in the western Cascades of Oregon and in the Oregon Coast Ranges from 1.3 to 44.6 mg/l and from 1.4 to 21.4 mg/l, respectively. This contrasts sharply with mean maximum increases in concentration occurring primarily during stormflow, which ranged from 11 to 52 times the mean annual rate for the western Cascades watershed and from 39 to 83 times the mean annual rate for the Coast Ranges watershed. Much of this increase is the direct result of soil mass movements into the stream channels and mobilization of sediment temporarily stored within or along the channel margins.

Large organic debris deposited and redistributed during stormflow and soil mass movements is a common and important channel feature, with both physical and biological consequences to anadromous fish habitat.

Most often debris--including material resulting from fire, disease, and decomposition--is delivered to the stream by a combination of processes including windthrow, streambank undercutting, and soil mass movements. Once in the stream, it may be deposited almost immediately in or along the channel margins or

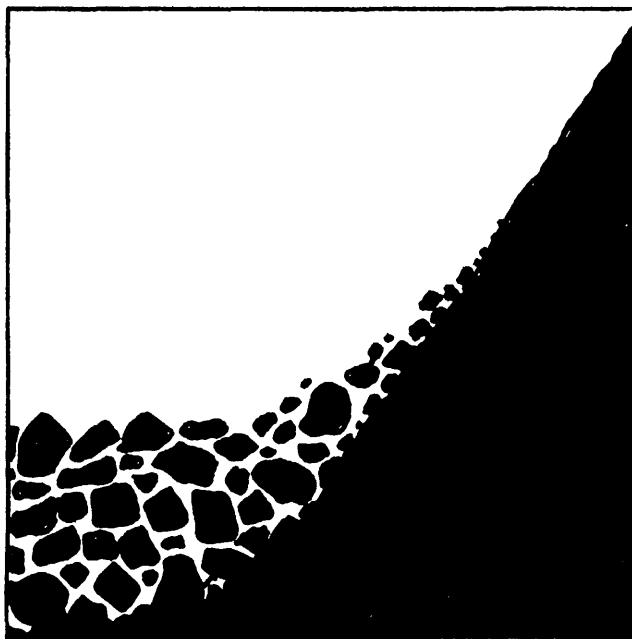
transported for considerable distances down the channel to be deposited as windrows of small organic detritus and piles or jams of mixed logs and smaller organic debris. The point of deposition and the distance of travel depend largely on the originating-process and the volume of streamflow in the channel at the time of deposition. Streams of all sizes are affected by debris, but loading tends to decrease as stream width increases. In small headwater channels, trees generally lie where they fall and are only moved as a result of decomposition or by catastrophic events, such as extreme stormflow or debris torrents.^{2/} As streams within a watershed expand in response to storm runoff, they become wider and deeper. Large organic material may float and accumulate at channel obstructions, forming debris accumulations or debris dams. In rivers, where a tree bole of any size can be floated, large organic debris usually accumulates at bends or along the channel margin as a result of high flow (Swanson et al. 1976).

Large jams can block the passage of migrating fish and effectively close areas to spawning (Meehan 1974). In addition, such jams may form a temporary base level for the affected channel, resulting in deposition on the upstream side and extensive scour downstream. If the jam is large enough, the stream may form a new channel, causing extensive bank erosion and effectively bypassing the jam and sections of spawning gravels downstream. In headwater streams, debris torrents and extreme stormflows carry large amounts of wood through the system causing extensive scour, redistribution of stream gravels, and damage to culverts, roads, and bridges.^{2/}

^{2/}Manuscript in preparation, "some management impacts of organic debris in Pacific Northwest streams," by F. J. Swanson and G. W. Lienkaemper. USDA For. Serv., Pac. Northwest For. and Range Exp. Stn., Portland, Ore.

Debris accumulations in small- to medium-sized streams (third- and fourth-order) often trap gravel that may provide excellent spawning habitat for anadromous fish. Also, fine sediment trapped by debris in headwater streams is routed through the system slowly as the wood accumulations decompose. Wood in small streams frequently provides a "steepened channel profile," in which much of the drop in the stream is over wood-created falls alternating with longer, low-gradient pools. This gives a large portion of the stream a lower gradient than the overall stream channel.

The biological community can benefit from such stream debris. Debris accumulations provide cover for resident and anadromous fish (Narver 1971, Hall and Baker 1975), while serving to retain detritus entering the stream system. Before detritus feeders find detritus of terrestrial origin palatable, it must be conditioned by the microbes in the stream (Triska and Sedell 1975). Debris accumulations keep the conditioned detritus, the insects that eat it, and the fish that eat the insects all in the same area.



MASS SOIL MOVEMENT

Increased water in the soil-- both seasonal and during storm

periods--produces saturated soil and a rising water table on steep slopes, frequently initiating soil mass movements that transport large volumes of sediment and organic debris into and through the channel systems. Such processes add substantial quantities of sediment and organic debris to the stream channels over short periods (minutes, hours, days), causing channel alteration, rapid increases in bed and suspended loads, siltation of gravels, and partial or complete blocking of the stream channel through debris-dam formation. If entry velocities and channel gradient are high enough, the tremendous bulking effects of contained soil, rock, and organic debris commonly scour the channel below the point of entry, removing and redistributing bottom gravels and destroying substantial portions of the streamside vegetation.

The mechanics of failure, types of soil mass movement, and the factors controlling and contributing to development of soil mass movements on forested terrain are well described in the literature (Bishop and Stevens 1964; Swanston 1967, 1969, 1971, 1974, Swanston and Swanson 1976). Three groups of processes have a major impact on habitat.

SLUMP-EARTHFLOWS

Earthflow processes (fig. 3) are for the most part slow moving; where they intersect a stream, they provide continuous long-term sources of sediment to the channel. Measured rates of movement in Oregon and northern California, where these processes are most common, range from 2.5 cm/year to as high as 2 720 cm/year with the higher rates predominantly occurring along the major anadromous fish rivers draining the northern California Coast Ranges (table 1). Based on studies of 19 earthflows entering the Van Duzen River basin in northern California, Kelsey (1978) estimated that the total

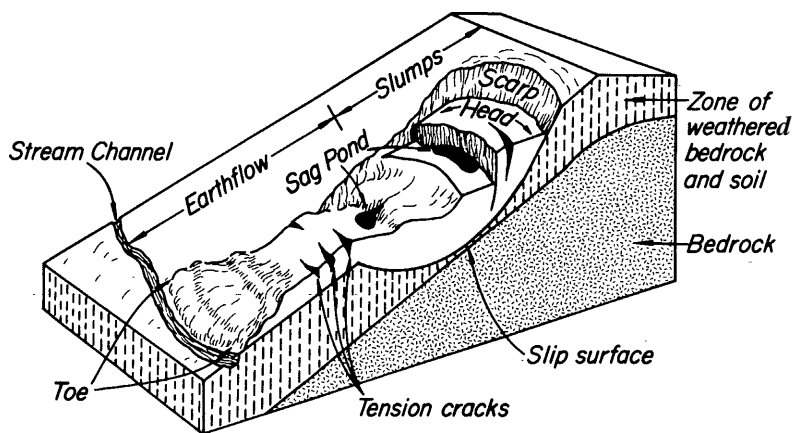


Figure 3.--Diagram of a typical slump-earthflow developed in deeply weathered bedrock and surficial materials. Slumping is the backward rotation of a block of soil along a curved failure surface with little downward displacement. Earthflows usually begin with a slump or series of slumps and, through a combination of true rheological flow of the clay fraction and slumping and sliding of individual blocks, move downslope with a continuity of motion resembling the flow of a viscous fluid.

Table 1--Rates of movement of active earthflows in the western Cascade Range, Oregon (Swanston and Swanson 1976), and Van Duzen River basin, northern California (Kelsey 1978)

Location	Period of record	Movement rate	Method of observation
	<u>Years</u>	<u>Cm/yr</u>	
Landes Creek ¹ (Sec. 21, T.22 S, R.4 E)	15	12	Deflection of road
Boone Creek ¹ (Sec. 17, T.17 S, R.5 E)	2	25	Deflection of road
Cougar Reservoir ¹ (Sec. 29, T.17 S, R.5 E)	2	2.5	Deflection of road
Lookout Creek ¹ (Sec. 30, T.15 S, R.6 E)	1	7	Strain rhombus measurements across active ground breaks
Donaker Earthflow ² (Sec. 10, T.1 N, R.3 E)	1	60	Resurvey of stake line
Chimney Rock Earthflow ² (Sec. 30, T.2 N, R.4 E).	1	530	Resurvey of stake line
Halloween Earthflow ² (Sec. 6, T.1 N, R.5 E)	3	2 720	Resurvey of stake line

¹Swanston and Swanson 1976.

²Kelsey 1978.

sediment discharge to the river by earthflow processes between 1941 and 1975 was 1 409 500 m³. This is equivalent to an annual yield of 41 455 m³ or about 2 182 m³ per failure. In contrast, studies at Lookout Creek in the western Cascades of Oregon showed annual movement rates of only 10 cm/year (Swanston and Swanson 1976), with estimated annual yields from a single earthflow of only 340 m³.

Earthflow movement is predominantly seasonal--with most movement occurring after fall and winter rains have thoroughly wetted the slopes--although movement may be continuous in areas where groundwater maintains the water content of the moving mass.

During periods of movement, the individual earthflow toe

protrudes into the channel and is gradually eroded away by high winter flows, Slumping (fig. 4), from the undercutting of the protruded material by stormflow, may abruptly add large quantities of soil, rock, and organic debris to the channel. If flows are high enough, this soil and some of the finer organic debris may be transported out of the system almost immediately to be distributed downstream as blankets and windrows of sediment and organic material. The larger rock and organic debris may remain behind as a residual "lag," forming a tangled mass of rocks and logs behind which gravels and sediment from up-channel may accumulate. This lag frequently creates a sharp increase in channel gradient through the accumulation zone.

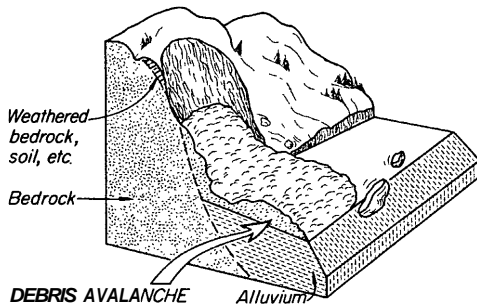


Figure 4.--The "Drift Creek Slide" in the west-central Coast Ranges of Oregon, a slump in nearly horizontal sandstone and siltstone. The lower end has been transformed into an earthflow and has dammed Drift Creek to form a lake more than 12 meters deep.

DEBRIS AVALANCHES AND DEBRIS FLOWS

Debris avalanches, flows, and torrents constitute the most damaging of the soil mass-movement processes to anadromous fish habitat. Debris avalanches and debris flows (fig. 5A and B) are rapid, shallow mass movements that develop on steep hillsides and generally contribute 60 percent or more of their initial failure volume almost immediately to the channel. Debris avalanches generally grade into debris flows as water content of the sliding mass increases.

A DEBRIS AVALANCHE- DEBRIS FLOW



B



Figure 5.--Debris-avalanche debris-flow failures. Debris avalanches are rapid downslope movements of soil, rock, and organic debris, which generally grade into debris flows as water content of the sliding mass increases: A, Diagram of typical debris avalanche or debris flow in shallow, semi-cohesionless soils; B, massive debris avalanche developed in granite-derived soils near Wrangell, Alaska.

Debris avalanches have rather consistent characteristics in the various geologic and geomorphic settings from northern California to southeast Alaska and eastward into the Idaho batholith country (Bishop and Stevens 1964; Dyrness 1967; Swanston 1970, 1974; Gonsior and Gardiner 1971; O'Loughlin 1972; Colman 1973; Fiksdal 1974). In all these areas, debris avalanches are usually triggered during high-intensity storms. For example, in the H. J. Andrews Experimental Forest, Oregon, storms of a 7-year or less return period initiate debris avalanching in forest areas. Swanston (1969) has correlated storms with a 5-year return interval with accelerated debris avalanching in coastal Alaska.

Debris avalanches leave scars as spoon-shaped depressions with long tails extending downslope toward the channel, from which less than 10 to as much as 10 000 m³ of soil and organic debris have been transported. Average volumes of individual debris avalanches in forested areas in the Pacific Northwest range from about 1 540 to 4 600 m³. Areas prone to debris avalanches are typified by shallow, low-cohesion soils on steep slopes where subsurface water may be concentrated by subtle topography on impermeable bedrock or glacial till surfaces. Because debris avalanches are shallow failures, factors such as timber and other vegetation (which control root-anchoring effects and transfer of wind stress to the soil mantle, as well as rate of water supply to the soil during rainfall and snowmelt) have significant effect on where and when debris avalanches occur. An added factor in areas of even occasional seismic activity is the lateral stress applied to the soil mantle by ground shaking during earthquakes. Seismic activity has been noted as a factor contributing to soil mass-movement initiation in Alaska

(Bishop and Stevens 1964) and in the central Rocky Mountains (Bailey 1971). In the Queen Charlotte Islands, British Columbia, which is noted as the most seismically active area in Canada (Sutherland-Brown 1968), a direct correlation with high soil mass-movement activity has been postulated by Alley and Tomson (1978).

The rate of occurrence is controlled by stability of the landscape and the frequency of storms severe enough to trigger them. Swanston and Swanson (1976 and table 2) report annual rates of debris-avalanche erosion from six forested study sites in Oregon, Washington, and British Columbia, ranging from 11 to 72 m³/km² per year.

Table 2--Debris-avalanche erosion in forest, clearcut, and roaded areas (Swanston and Swanson 1976)

Site	Period of record	-----Area-----		Slides	Debris-avalanche erosion	Rate of debris-avalanche erosion relative to forested areas
		Years	Percent			
Stegualeho Creek, Olympic Peninsula, Washington (Fiksdal 1974)						
Forest	84	79.0	19.3	25	71.8	1.0
Clearcut	6	18.0	4.4	--	--	--
Road	6	3.0	.7	83	11 825.0	165.0
			<u>24.4</u>	<u>108</u>		
Alder Creek, western Cascade Range, Oregon (Morrison 1975)						
Forest	25	70.5	12.3	7	45.3	1.0
Clearcut	15	26.0	4.5	18	117.1	2.6
Road	15	3.5	-6	75	15 565.0	344.0
			<u>17.4</u>	<u>100</u>		
Selected drainages, Coast Mountains, S.W. British Columbia						
Forest	32	83.9	246.1	29	11.2	1.0
Clearcut	32	9.5	26.4	18	24.5	2.2
Road	32	1.5	4.2	11	1282.5	25.2
			<u>276.7</u>	<u>58</u>		
H. J. Andrews Experimental Forest, western Cascade Range, Oregon (Swanson and Dyrness 1975)						
Forest	25	77.5	49.8	31	35.9	1.0
Clearcut	25	19.3	12.4	30	132.2	3.7
Road	25	3.2	2.0	69	1 772.0	49.0
			<u>64.2</u>	<u>130</u>		

1Calculated from O'Loughlin (1972, and personal communication), assuming that area of road construction in and outside clearcuttings is 16 percent of the area clearcut. Colin L. O'Loughlin is now at Forest Research Institute, New Zealand Forest Service, Rangiora, New Zealand.

When a debris avalanche occurs, large quantities of soil, rocks, and organic debris are dumped directly into the channel. Because debris avalanches occur primarily during high stormflow, a large part of the soil and organic debris is transported almost immediately away from the entry point to be deposited downstream as blankets and windrows. Boulders and large organic debris remaining at the entry point may temporarily dam the channel and cause an increase in channel gradient for a variable distance downstream.

DEBRIS TORRENTS

Debris torrents are the rapid movement of water-charged soil, rock, and organic debris down steep stream channels (fig. 6). Debris torrents typically occur in steep, intermittent, first- and second-order channels. These events are triggered during extreme stormflow by debris avalanches from adjacent hillslopes, which enter a channel and move directly downstream, or by the breakup and mobilization of debris accumulations in the channel. The initial slurry of water and associated debris commonly entrains large quantities of additional inorganic and living and dead organic material from the streambed and banks. Some torrents are triggered by debris avalanches of less than 100 m³, but may ultimately include 10 000 m³ of debris entrained along the track of the torrent. As the torrent moves downstream, hundreds of meters of channel may be scoured to bedrock (fig. 7). When the torrent loses momentum, a tangled mass of large organic debris is deposited in a matrix of sediment and fine organic material covering areas up to several hectares (Swanston and



Figure 6.--Debris torrent developed in pumice on the north slope of Entiat Valley, east-central Cascades, Washington. Torrent developed as the result of temporary damming of the stream at near peak flow by debris avalanches along the channel slopes.



Figure 7.--Channel scoured to bedrock by debris torrent passage in the Oregon Coast Ranges.

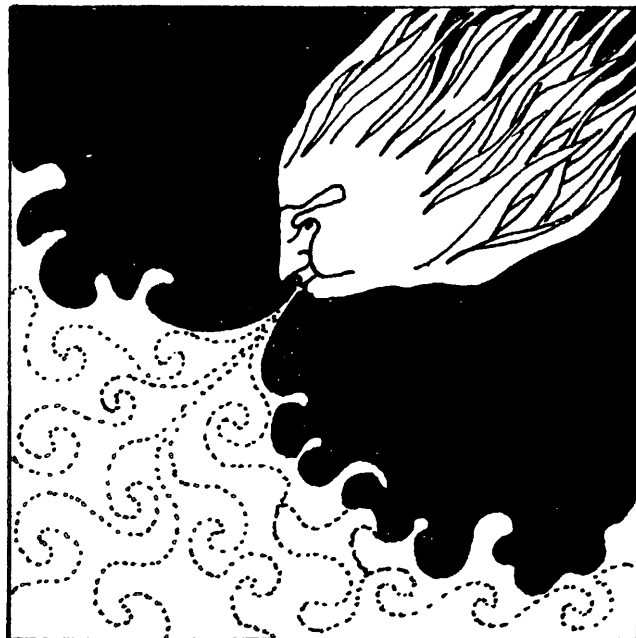
Swanson 1976, and fig. 8). The main factors controlling the occurrence of debris torrents are the quantity and stability of debris in channels (supplied by earlier debris avalanches and debris torrents), steepness of channel, stability of adjacent hillslopes, and peak-discharge characteristics of the channel. The concentration and stability of debris in channels reflect the history of stream flushing and the health and stage of development of the surrounding timber stand (Froelich 1973),



Figure 8.--Debris jam in an anadromous fish stream in the Oregon Coast Ranges. The jam resulted from a debris torrent entering the stream from an adjacent slope.

Although debris torrents pose significant hazards to anadromous fish habitat, they have received little study (Fredriksen 1963, 1965; Morrison 1975; Swanson et al. 1976). Velocities of debris torrents, estimated to be up to several tens of meters per second, are known from only a few spoken and written accounts. The rates of occurrence of torrents have been systematically documented in only two small areas of the Pacific Northwest, both in the western Cascade Range

of Oregon (Morrison 1975, Swanson and Swanson 1976). In these studies, rates of debris-torrent occurrences were observed to be 0.005 and 0.008 event per square kilometer per year for forested areas (table 3). Torrent tracts initiated in forested areas ranged from 100 to 2 280 m and averaged 610 m of channel length. Debris avalanches played a dominant role in triggering 83 percent of inventoried torrents (Greswell et al, 1979). Mobilization of stream debris not immediately related to debris avalanches (debris-dam failure within the channel) has been an important local factor in initiating debris torrents in headwater streams (Swanson 1969).



WIND

Strong winds, primarily during storms, frequently uproot trees, which drastically disturbs forest soils and may effectively alter the immediate channel environment.

Windthrow has been recognized for many years as a widespread natural phenomenon in forested regions throughout the United States (Shaler 1891, Holmes 1893, Van Hise 1904, Lutz and Griswell 1939, Stephens 1956). Storms with winds of hurricane force may cause an entire stand to be uprooted; more commonly, however, scattered

Table 3--Debris-torrent occurrence for selected areas in western Oregon
(Swanston and Swanson 1976)

Site	Area of watershed	Period of record	Debris torrents triggered by debris avalanches	Debris torrents with no associated debris avalanche	Total		Rate of debris-torrent occurrence relative to forested areas
					No.	No./km ² /yr	
H. J. Andrews Experimental Forest, western Cascade Range, Oregon ¹ /							
Forest	49.8	25	9	1	10	0.008	1.0
Clearcut	12.4	25	5	6	11	.036	4.5
Road	2.0	25	17	--	17	.340	42.0
	64.2		31	7	38		
Alder Creek drainage, western Cascade Range, Oregon							
Forest	12.3	90	5	1	6	.005	1.0
Clearcut	4.5	15	2	1	3	.044	8.8
Road	.6	15.	6	--	6	.667	133.4
	17.4		13	2	15		

¹/Frederick J. Swanson, unpublished data, on file at Forestry Sciences Laboratory, USDA Forest Service Pacific Northwest Forest and Range Experiment Station, Corvallis, Oregon.

individuals or groups of trees are knocked down (fig. 9). In the Pacific Northwest and along the North Pacific coast, windthrow is an important initiator of soil mass movement and, excluding soil mass-movement activity, is probably the most significant natural phenomenon providing organic materials to the stream system.

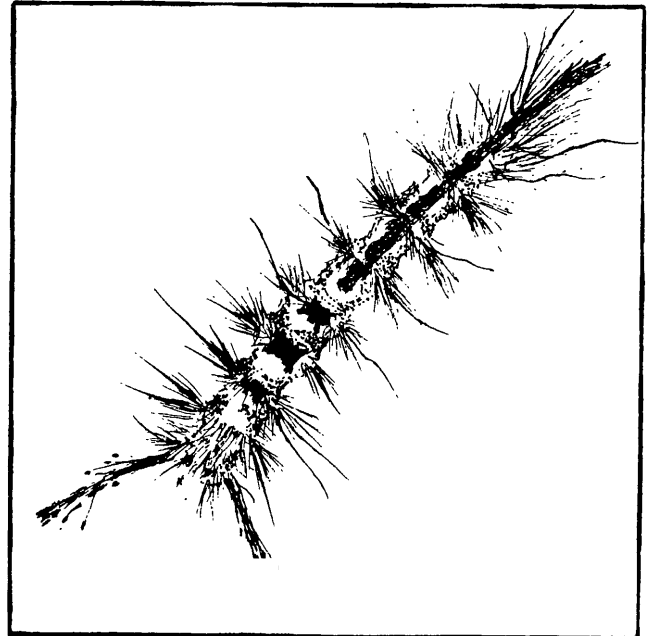


Figure 9.--Natural windthrow along Kook Creek, Chichagof Island, south-east Alaska.

Studies of windthrow in virgin forests of the Oregon Coast Ranges (Ruth and Yoder 1953) indicate it is most severe on areas with a high water table or very shallow soils. Windthrow on such forested slopes tends to open the mineral soil to direct ingress of water, destroys the anchoring and reinforcing effect of tree roots on unstable sites, and--if windthrow occurs on a section of slope at or near saturation--may initiate soil mass movement directly through impact and instantaneous pore-water pressure development (Swanston 1967). Windthrow has been identified directly as the

probable initiating cause of debris avalanches in coastal Alaska (Swanston 1967) and the central Oregon Coast Ranges, and it is believed to be a contributor to debris-avalanche development in the western Cascades (Swanson and Dyrness 1975).

Windthrow adjacent to channels adds large organic debris directly to them and may open large sections of the stream to direct sunlight. Large organic debris in streams (logs and branches) may create debris dams, allowing gravel and fine sediment to accumulate from upstream, at least as long as the dam remains in place. Such dams may be removed during stormflows or, if the dam remains in place, may cause local flooding or alteration of the channel to bypass the obstruction. The exposure of the channel to direct sunlight may increase water temperatures in that section during the summer, and the lack of insulating canopy may decrease winter temperatures.



INSECTS AND DISEASE

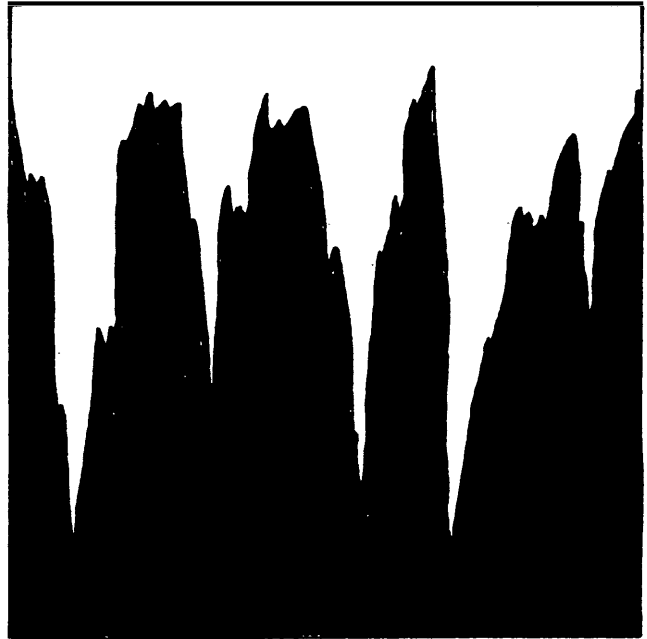
The killing of trees by insects and disease affects the forest and ultimately the stream environment much the same as windthrow. Interception and evapotranspiration are

reduced or stopped, but infiltration is usually not affected (Anderson et al. 1976). Because of the loss of overhead cover near stream channels, water temperature may be locally increased and water chemistry altered temporarily by leaf fall and accumulation of organic vegetation in channels. The weakened root systems in dead trees also make the affected timber much more susceptible to windthrow (with its accompanying soil disturbance and accumulation of large organic debris in channels). Although no direct data are available, the destruction of viable root systems on unstable slopes by these epidemics would greatly reduce the slopes' ability to resist failure when soil water content is high and in periods of stormflow.

Forest Service data (USDA Forest Service 1958, 1975) indicate that insects kill far more trees than disease does. Damage from insect and disease infestation is scattered. Mortality in any given area is usually confined to a single tree species and not all the individuals of that species are infested. In mixed forest types, this has practically no impact on erosion and streamflow because living forest cover and its attendant root systems are maintained. Under those conditions, the only damage to the stream environment might be from death of streamside trees, which would open the channel to direct sunlight and increase organic debris loading.

Occasionally, large areas do become infested, and substantial portions of the timber cover are destroyed. On the White River drainage in Colorado (Love 1955, Bethlahmy 1974), bark beetles killed most of the trees in a 1 974-km² area with a resultant increase in total streamflow of about 22 percent over the next

25 years. Peak flows were 27 percent higher. The actual impacts on the channel system are unknown.



FREEZING AND ICE FORMATION

In northern latitudes, freezing temperatures and development of ice on hillslopes and within stream channels may substantially reduce rate of streamflow and increase sediment contributions from bare hillslope areas under certain conditions. If the channel freezes over, ice formation and subsequent melting and breakup may result in flooding and extensive bank and channel erosion by mechanical plowing and formation of "anchor ice."

In areas where extensive channel-ice formation is rare, freezing temperatures tend to have their greatest impact on accelerated discharge of surface sediment to the channel.

"Concrete frost"--wet soil solidly frozen--probably occurs only sporadically on forested slopes in the interior areas of the Northwest (Anderson et al. 1976). In the Pacific coastal mountains, it is generally absent. Where present, it may prevent infiltration and cause local overland flow; its

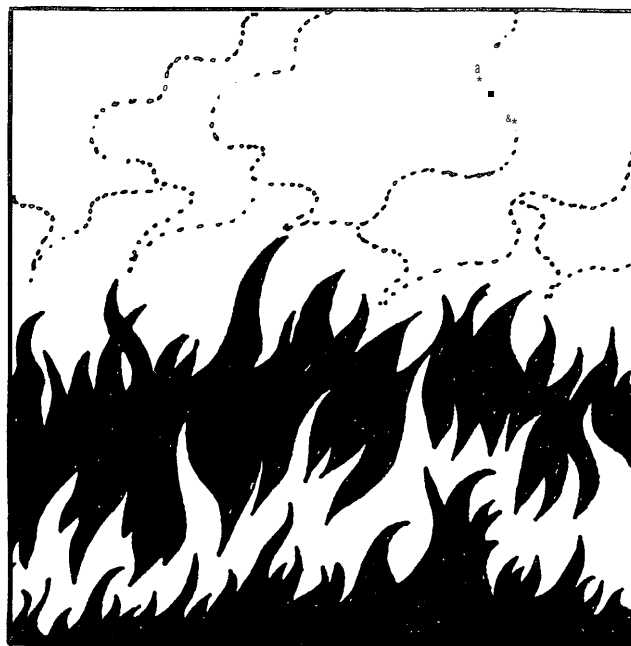
frequency is so low, however, that it probably has very little effect on water in the channel. Much more important in terms of soil freezing effects is the development of "needle ice." Needle ice is produced by the growth of frost crystals beneath pebbles and soil particles on unvegetated slopes during diurnal cycles of freezing and thawing (Sharpe 1960). The particles are lifted perpendicular to the slope surface: when the needle ice begins to melt, the ice crystals and their load of earth, pebbles, and organic debris fall downslope and may continue to slide and roll for some distance. Such "surface creep" is an important local contributor to sediment transport from bare mineral soil areas to stream channels throughout the mountainous areas of Western North America,

In areas where extensive channel ice is formed, freezing results in supercooling of the water, nucleation of "frazil ice" particles (spicules and thin plates of ice formed in supercooled, turbulent waters) and formation of anchor ice around stones and gravel particles along the channel bottom (Michel 1973, Gillfilain et al. 1973). Static ice begins to form along the stream banks in areas of non-turbulent flow. Through accumulation of frazil ice along the rough streamward edges of the static ice, slush and iceflows eventually form a continuous ice cover. Anchor ice forms along the channel bottom from the accumulation of frazil ice particles on the rough surfaces of coarse bottom sediments and on the lee sides of pebbles, rocks, and boulders. During ice formation, anchor ice frequently breaks loose from the bottom and is carried to the surface or downstream, with gravel and coarse bottom sediments still adhering. The result may be an extensive redistribution and

downstream transport of bottom materials.

In small streams used by anadromous fish in the arctic and subarctic, such ice formation diverts a substantial volume of water from winter streamflow and results in rapid transport of channel sediment downstream, Kane and Slaughter (1973) have estimated that winter icing of Gold Stream-- a stream near Fairbanks, Alaska, that is used by anadromous fish-- accounts for nearly 40 percent of the winter streamflow,

The breakup of ice cover in the spring generally follows ablation of the seasonal snowpack when rising water in the channel causes cracking of the ice from vertical hydrostatic pressures, and the resultant blocks and plates of ice are carried downstream as ice floes. The movement of these floes is intermittent and jerky, resulting in periodic damming and flooding of low areas near the channel, extensive gouging and mechanical erosion of the channel banks, and sedimentation and redistribution of bottom gravels,



FIRE

Burning of forests destroys the covering vegetation on slopes

and along stream channels and may locally alter the physical properties of the surface layers of soil. The immediate impact is to increase both water yield and stormflow discharge from the watershed. Fire also exposes the bare mineral soil to increased surface runoff and erosion. Surface runoff from burned areas generally increases dissolved nutrient transport and loading of the channel systems. Surface erosion processes--including raindrop and rill erosion and dry ravel during wetting/drying and freezing/thawing cycles--transport large volumes of sediment and debris from the watershed slopes to the channel (fig. 10). From 1 to 5 years afterward, fire may increase the potential for accelerated landslide activity through decay of anchoring and reinforcing root systems (Bishop and Stevens 1964, Swanston 1974, Ziemer and Swanston 1977).



Figure 10.--Dry ravel, or dry creep and sliding, of surface materials downslope after a fire along the Klamath River in northern California.

Intensive drying of soil, combustion of organic matter that binds soil aggregates, loss of litter cover, and strong convective winds produced by the fire's heat all contribute to debris movement down steep slopes during hot fires. In steep terrain, rolling rocks and logs released by burning of roots and other supportive organic matter trigger downslope movement of additional material, greatly increasing concentration of large woody debris on slopes and within stream channels.^{3/} Hydrophobic (water repelling) soils have been reported in a great variety of ecosystems after wildfire and slash fires (DeBano 1969, DeByle 1973, Megahan and Molitor 1975, Dyrness 1976, Campbell et al. 1977). The relative importance of this phenomenon

on increased rill, sheet, and soil mass-movement erosion commonly observed after intense fire has not been well documented. Hot ground fire can reduce water storage capacity of surface organic matter (Dyrness et al. 1957). Reduced interception and evapotranspiration may result in decreased summer drawdown of soil water by vegetation (Klock and Helvey 1976a), although sometimes, as in heath vegetation, water loss can increase when the soil surface is exposed (C. H. Grimingham, Univ. of Aberdeen, personal communication). Effects of reduced interception and evapotranspiration may be offset in part by increased overland flow in response to reduced infiltration from loss of litter layer, development of hydrophobic soil, compaction by raindrops, plugging of pores by fine soil

^{3/}Manuscript in preparation, "Fire and geomorphological processes," by F. J. Swanson. USDA For. Serv., Pac. Northwest For. and Range Exp. Stn., Portland, Oreg.

material, and sometimes actual fusing of the soil surface (Dyrness et al. 1957; Ahlgren and Ahlgren 1960; Brown 1972; Helvey 1972, 1973; Rice 1973; Anderson et al. 1976; Campbell et al. 1977). In general, these factors lead to increases in both soil water-storage and runoff from burned sites.

Contrasts in snow hydrology of burned and unburned ecosystems have received little study, particularly in terms of fire-induced changes in groundwater regime (see footnote 3). Snow accumulation and melt in open (clearcut or naturally treeless) and forested areas have been the subject of extensive research; a stand of blackened snags presents a vastly different environment from either forested or treeless areas. Speculation on snow hydrology of burned areas is complicated by the great contrasts between cold/dry and warm/wet snow types and between snowpack and multiple accumulation/melt seasonal regimes. Work on warm snowpacks by Smith (1974) and others does suggest, however, that formation of melt zones around blackened snags and rapid-condensation melting may add more meltwater to the soil in burned areas than in forests and snag-free open areas. Forests may have greater loss by evaporation; snowpacks in open areas may contain continuous, relatively impermeable horizons that carry meltwater directly to streams.

Destruction of covering vegetation by fire opens the channel to direct sunlight, increasing water temperatures and perhaps producing substantial increases in organic debris loading from falling limbs, litter, and trees. The increase in stormflow and acceleration of

erosion depends on the intensity, severity, and frequency of burning and how much of a particular watershed burned (Anderson et al. 1976). If much foliage is destroyed, interception and evapotranspiration are reduced, and the potential for accelerated landslide activity from root deterioration is increased. Where the organic layers of the forest floor are also consumed and mineral soil is exposed, infiltration and soil water-storage capacities are reduced, greatly increasing surface erosion and runoff potential.

Massive wildfires in western forests have accelerated both erosion and sedimentation as a result of these processes. Studies of fire-denuded watersheds in west-central Washington (Klock and Helvey 1976a and b) showed that maximum streamflows were double the rate of flows before the fire. In addition, as the result of combined rapid snowmelt, high intensity rainstorms, and the destruction of covering and anchoring vegetation, massive debris torrents occurred 2 years after the burn with a frequency 10 to 28 times greater than before the fire. Noble and Lundeen (1971) report a postfire erosion rate of 413.3 m³/km² per year for a portion of the South Fork Salmon River in the Idaho Batholith. This amounts to an acceleration seven times greater than sediment yield for similar, but unburned, lands in the vicinity (Megahan and Molitor 1975).

In northern California, Wallis and Anderson (1965) reported sediment discharge 2.3 times greater from burned than from unburned areas. Seventeen years after the Tillamook Burn in the Wilson River watershed of Oregon, the annual rate of sediment discharge was 175.3 t/km², 5-8 times that of nearby unburned forested watersheds with similar geology (Anderson 1954).

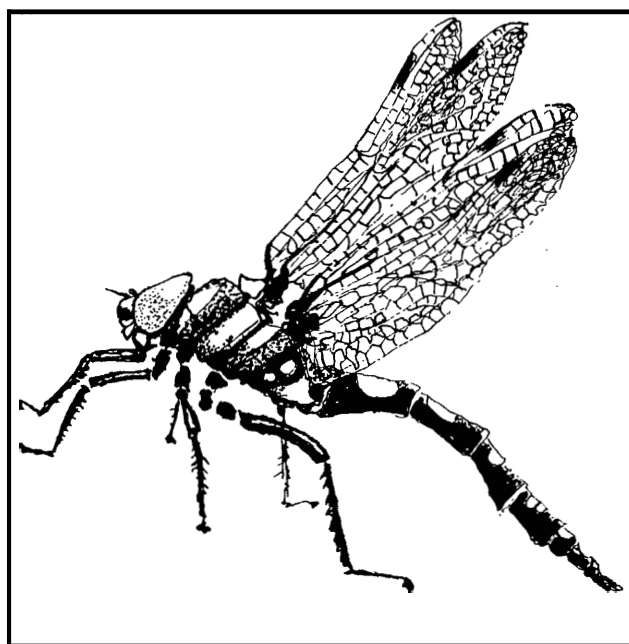
Fire affects nutrient availability and subsequent nutrient loading of streams in several ways. Nutrients incorporated in vegetation, litter, and soil can be volatilized during pyrolysis and combustion, mineralized during oxidation, or lost by ash convection (Grier 1975). After the fire is out, nutrients can then be redistributed by leaching of the ash layer and soil and transported to the stream by surface erosion, soil mass movement, or solution transport. Studies on nutrient transfer to streams after wildfire and controlled fire in the western Cascades of Oregon (R. L. Fredriksen, Forestry Sciences Laboratory, Corvallis, personal communication) and in north-central Washington (Tiedemann 1973, Grier 1975), indicate that after a fire, free nitrogen concentrations were lower than in nonburned areas, nitrate increased, and phosphorus may have increased slightly. In western Oregon, Fredriksen (personal communication) reported that pH changes generally last less than 1 month, increased phosphorus concentrations no more than 2 years, and increased nitrate concentrations from 1 to 10 years.

The impact of these changes in nutrient concentration on salmonid productivity is not well known. The levels of increased nutrients reported in streams after fire appear to be below toxic thresholds for aquatic organisms and dissipate rapidly with stream dilution and flushing.

Gibbons and Salo (1973) pointed out that the addition of nutrients to a stream may be beneficial, especially to relatively sterile streams, by supporting additional plant and animal life; such results remain difficult to predict, however, and excessive nutrient loading may result in eutrophication.

CONCLUSIONS

The natural events described may operate separately or in combination to create limiting habitat characteristics in a particular stream section or system. Human activities in the stream and its parent watershed may profoundly affect these events, their frequency, and magnitude. A firm understanding of the natural processes is thus essential for a clear understanding of the effects of forest and rangeland management on the habitat of anadromous salmonids.



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