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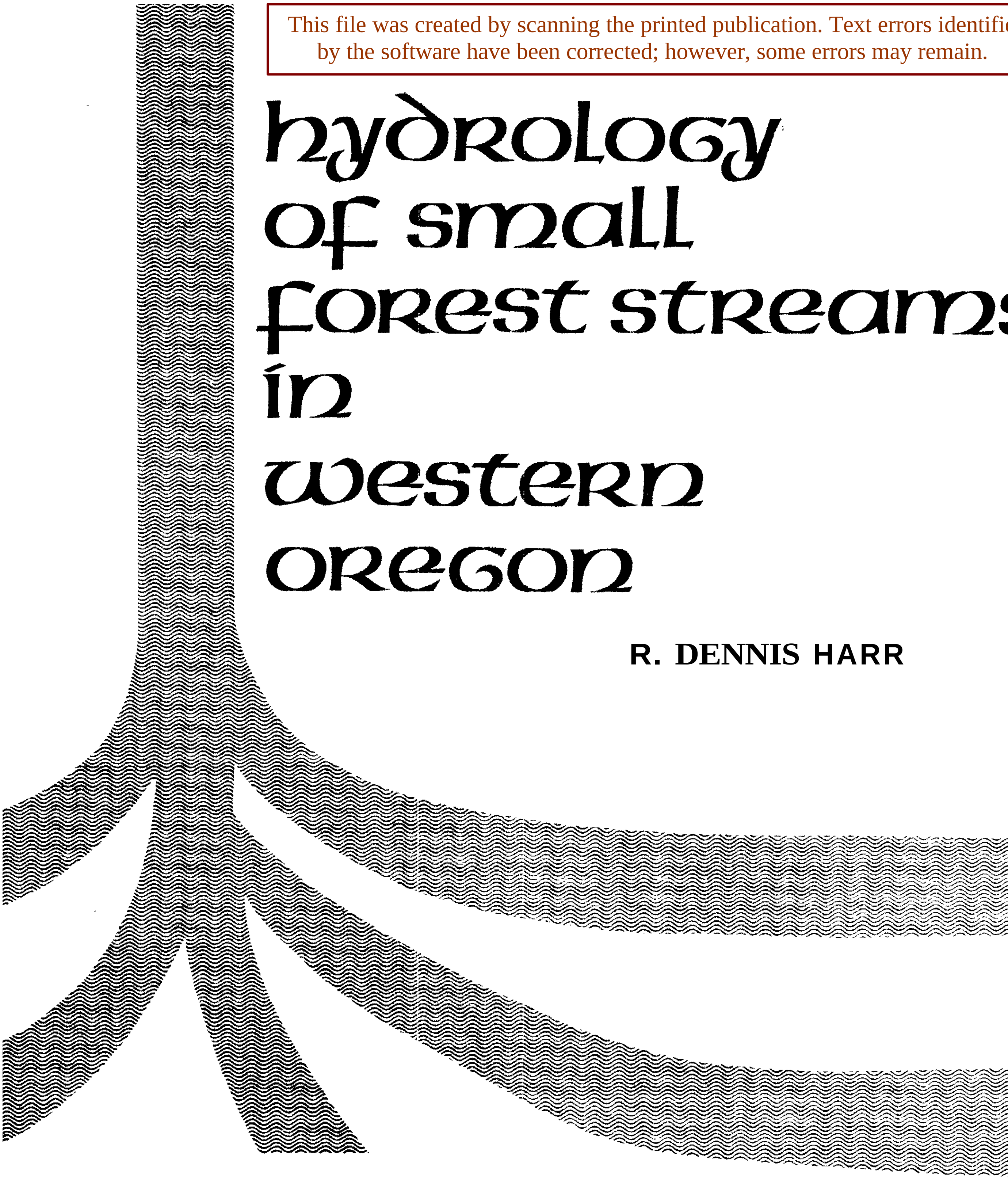

hydrology of small forest streams in western oregon

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Abstract

The hydrology of small forest streams in western Oregon varies by time and space in terms of both streamflow and channel hydraulics. Overland flow rarely occurs on undisturbed soils. Instead, water is transmitted rapidly through soils to stream channels by displacement of stored soil water. Drainage networks expand and contract according to the interaction between precipitation characteristics and soil's capability to store and transmit water. Drainage networks are more extensive in winter than in summer. Streamflow may he 1,000 to 5,000 times greater during winter storms than during summer low flow. A stream's kinetic energy varies along with streamflow. Channel width and depth, heterogeneity of bed materials, and the accumulation of large, organic debris affects the dissipation of kinetic energy. Clearcutting can increase relatively small peak flows, but forest roads and extensive areas of soil compacted by other means may increase larger peak flows. Both roadbuilding and clearcutting can cause soil mass movements which can drastically alter a stream's channel hydraulics by adding debris or scouring the channel to bedrock. Removal of naturally occurring organic debris that has become part of a stable channel can accelerate bed and bank erosion.

KEYWORDS: Streamflow, runoff cycle, clearcutting (-streamflow.

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Introduction

Small forest streams are a valuable resource of the Pacific Northwest. High quality stream water is used by municipalities and agriculture. Small streams also serve as spawning and rearing areas for anadromous fish and habitat for resident trout. These streams originate in and flow through highly productive forests whose wood products are vital to the economy of the Pacific Northwest. In many cases, the value of these small streams lies primarily in their effect on downstream waters, a fact recognized by Oregon's Forest Practices Rules (State of Oregon 1974). Moreover, the quality and quantity of water in these small streams may be drastically altered by forest land management practices (Brown 1973).

DEFINITION OF SMALL STREAM

Small streams are those in headwater areas where most streamflow originates. Small streams in western Oregon refers to streams of orders 1, 2, and 3. The order number reflects the stream's position in a drainage network. The smallest streams are designated order 1 (fig. 1). They have no surface tributaries and are either intermittent or permanent. Where two first-order streams join, a channel segment of order 2 begins, and so forth until the highest order channel segment is formed. Some of the small streams in the H. J. Andrews Experimental Forest are shown in figure 1. In this drainage network the highest stream order is 5. This stream (Lookout Creek) flows into Blue River (order 6), which in turn flows into the McKenzie River (order 7), a major tributary of the Willamette River. There are conservatively 250,000 small stream channel segments in western Oregon totaling several hundred thousand miles and collectively draining more than 80 percent of our forest land.

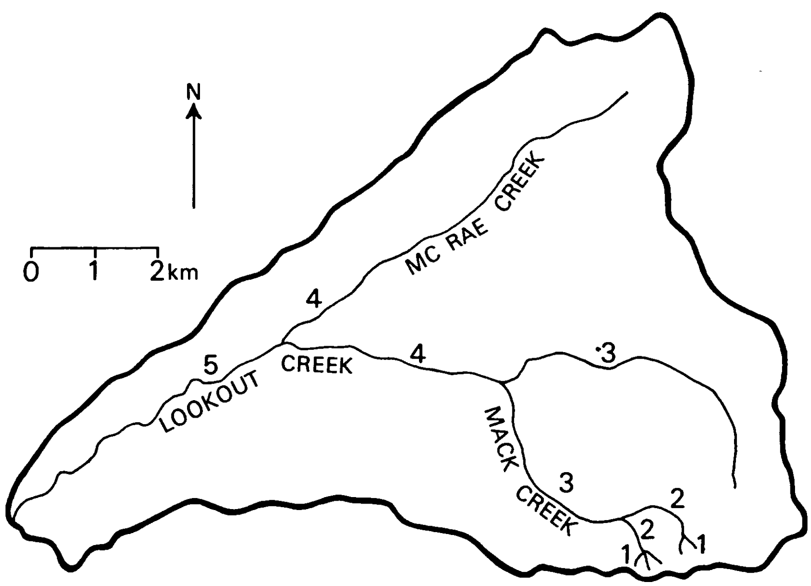

Figure 1.--Stream order numbers of some small streams in the H. J. Andrews Experimental Forest near Blue River, Oregon. Not all small streams are shown.

THE REGION

The climate of western Oregon is influenced by the proximity of the Pacific Ocean. Winters are cool and wet, and summers are warm and dry. Over most of the region, precipitation occurs mainly as rain in frequent, long-duration, low- to moderate-intensity frontal storms. About 75 percent of annual precipitation occurs between October 1 and April 1. Elevations above 915 m (3,000 ft) receive much of their annual precipitation in the form of snow. Frontal storms interact with topography to provide annual precipitation in excess of 300 cm (120 in) on the west side of the Coast Ranges and the Cascade Range. The lower limit of annual precipitation for most forest watersheds in this region is about 50-75 cm (20-30 in).

The topography of most forest land in western Oregon is steep. This topography reflects the nature of the climate, primarily the large amounts of precipitation and its influence on runoff and erosion. Much of the land is highly dissected, and many slopes are extremely steep. These steep slopes exert a major influence on rapid movement of water to streams and through a stream system.

If there is a term that can most appropriately be applied to the hydrology of small, forest streams, it is "variability," not only in streamflow and hydraulic characteristics but also over time and space. This report should make it clear that variability is the rule rather than the exception., It is essential to keep variability in mind while attempting to develop and apply guidelines for managing forest lands without damaging water resources.

Subsequent sections of this report deal with the processes involved in streamflow production, channel flow and associated hydraulic characteristics, and the effects of timber harvesting activities on streamflow. The information presented applies mainly to rain-dominated hydrologic systems and is based largely on results of hydrology studies conducted in western Oregon. Similarities in climate, vegetation, and physiography indicate that, with few exceptions, these hydrologic processes also apply to watersheds in western Washington and the northern coast of California where rain is the dominant form of precipitation.

Hydrologic Processes

Streamflow is generated mainly by processes operating outside the stream channel. Thus, when discussing the hydrology of a small stream, it is imperative to include the hydrology of that stream's watershed. Although the latter is far from being completely understood, bits of information can be assembled to describe what most likely occurs during runoff production--the multitude of processes which collectively transform precipitation to streamflow.

FOREST HYDROLOGIC CYCLE

It may be helpful to briefly review the various processes influencing runoff production for a rain-dominated system (fig. 2). Rain falling on a forest is

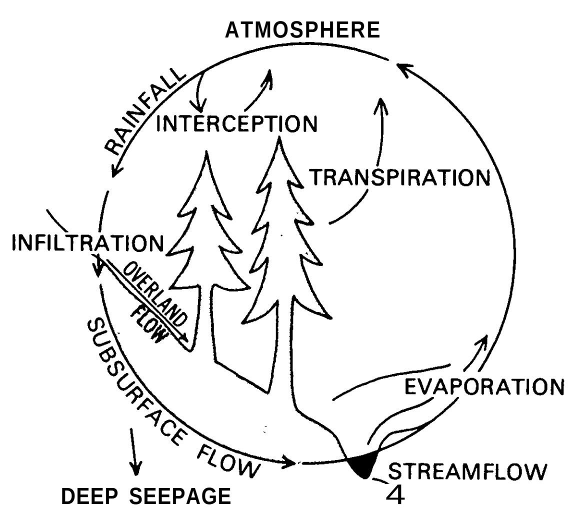

Figure 2.--Rain-dominated hydrologic cycle.

subject to interception by vegetation surfaces. Some of this intercepted water evaporates, while the remainder falls to the forest floor. The relative amount of rain evaporated depends on the size and duration of the storm producing the rain. For example, in an old-growth Douglas-fir forest, 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, totaling more than 20 cm (8 in) was intercepted and evaporated. In other words, the larger the storm and the longer it lasts, the smaller the proportion of rain lost to evaporation.

Interception not evaporated falls to the forest floor along with rain not initially intercepted. On the forest floor water is subject to infiltration. 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 soil and some may become overland flow. In undisturbed forest soils in western Oregon, infiltration capacities far exceed the maximum rates of rainfall so that all water enters the soil. Overland flow rarely occurs where forest soils are undisturbed. Lack of overland flow is one of the most important characteristics of undisturbed forest land in the Pacific Northwest as well as in most forested regions with a humid climate.

Once in 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 forest land 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 water moves depends 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 are low and frequently about equal to the average rate of rainfall during a storm.

Relatively slow-moving soil water is subject to evaporation and to removal by plants through the process of transpiration. The rate at which plants withdraw water from the soil is largely a function of energy available for water vaporization in leaves and the ease with which water may be withdrawn from the soil. As soil moisture decreases during the growing season, remaining water becomes more tightly held by the soil. As the growing season progresses, many plants exhibit reduced rates of transpiration and in some cases transpiration may cease altogether.

Streamflow, on an annual or longer basis, is the difference between precipitation and the evaporation losses described above. There are changes in soil moisture storage from time to time. Some water may also seep deep into the subsoil and bedrock and not appear as streamflow in a small headwater basin. But generally, water not removed by plants moves downslope to supply streams. It is this movement to stream channels that we want to examine in more detail.

COMPONENTS OF STREAMFLOW

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. Although streamflow cannot be separated into distinct quantities from these different sources, 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. Classical hydrology has been characterized by its emphasis on overland flow and the importance of the difference between rates of precipitation and infiltration on the generation of streamflow. In well-vegetated areas with humid climates, however, overland flow is the exception rather than the rule (Hewlett and Hibbert 1967). In western Oregon, overland flow in undisturbed forests rarely occurs. If flowing water is observed in this region, it almost always will be in relatively well-defined channels or over seriously disturbed soils such as those in roads, skid trails, or severely burned areas. In undisturbed areas, infiltration capacities of soils are extremely high and, in many cases, may be several hundred times greater than maximum sustained rates of rainfall (Dyrness 1969, Ranken 1974, Yee 1975, Harr 1977). With the low- to moderate-intensity storms common to western Oregon, it is highly unlikely that an undisturbed soil would not accept all water delivered to it as rainfall. Moreover, soils are able to transmit water rapidly, such that overland flow over saturated soils is also highly unlikely (Harr 1977, Yee 1975). The lack of overland flow leaves channel interception and subsurface flow as the only sources of water flowing in stream channels.

TRANSLATORY FLOW

Small streams in western Oregon respond quickly to rainfall, even in the absence of overland flow, which is by far the most rapid process of water movement. For example, in the H. J. Andrews Experimental Forest, 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). A relatively small runoff event in a 10-hectare (25-acre) watershed is shown in figure 3. The initiation of the storm hydrographs and the occurrence of peak flows closely follow the beginning and the end of rainfall.

How small streams respond quickly to rainfall without overland flow is not completely understood. Perhaps the best explanation available today is what Hewlett and Hibbert (1967) have termed "translatory flow" or flow by displacement. In this process, two types of flow are distinguished although they are not entirely separable. If a line is extended from the point of initiation of a storm hydrograph until it intersects the falling limb of the storm hydrograph (fig. 4), the flow above the line becomes, by definition, quick flow and consists of channel interception and subsurface flow. The flow below the separation line is denoted delayed flow. The relative amount of quick flow is a measure of a watershed's efficiency in producing streamflow.

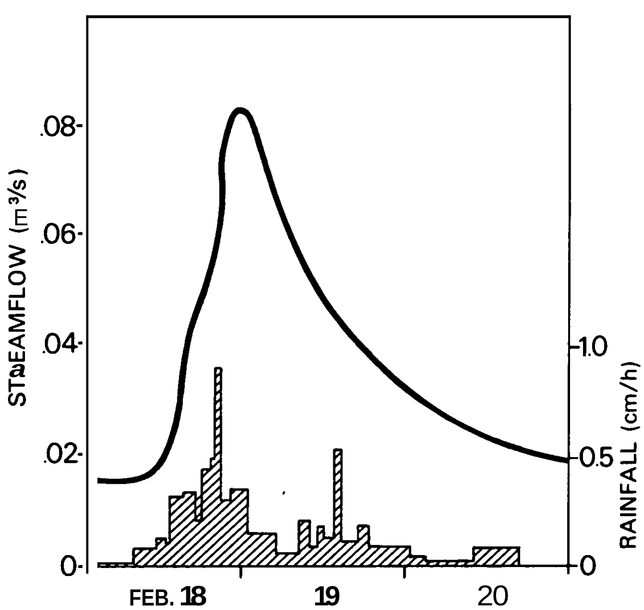

Figure 3.--Precipitation and streamflow at Watershed 10, H. J. Andrews Experimental Forest, February 18-20, 1974.

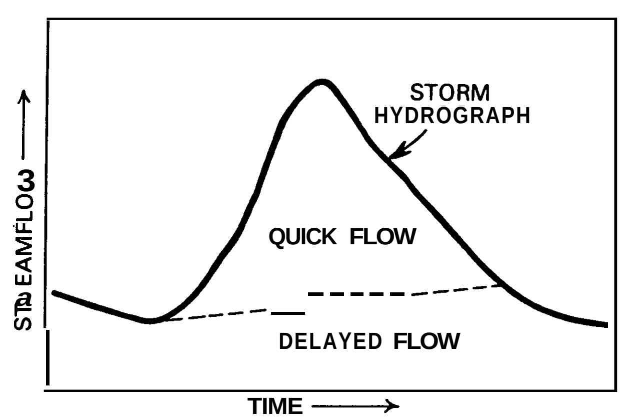

Figure 4.--Method of separating quick flow and delayed flow.

Rainfall on a slope will contribute to quick flow and delayed flow. Quick flow is composed of new rain and old rain from water stored in the soil and displaced by new rain. The wetter the soil, the greater the contribution of stored water to streamflow. Rainfall on that portion of a slope closest to a stream produces more quick flow than rainfall on a higher slope position because soil moisture content usually is greater in soil adjacent to streams due to continual drainage of soil water from upslope. Thus, these lower slopes are more conducive to translatory flow.

Understanding this theoretical process of flow by displacement may be aided by examining the analogy of water movement through a garden hose which contains some water. The three faucets in figure 5 represent different slope positions, and the length of hose attached to each represents soil of uniform texture. Water in the hoses is analogous to soil water storage with different levels of water representing different soil moisture contents among slope positions, The dashed line represents the level of soil moisture necessary for substantial translatory flow. Once the water level in a hose reaches this level, the contribution from a slope position to outflow from the sloping hose will be greatly increased.

Rainfall on the slope would be simulated by simultaneously turning on all faucets. It should be evident that water will flow almost immediately from the sloping hose (start of storm runoff), but that this flowing water will not be coming from any of the faucets (new rain). Rather it will be water that had been stored in the lower portion of the sloping hose (soil moisture storage) and was displaced by water from the lowermost faucet (new rain). It should also be evident that water from the lowermost faucet (lower slope) will reach the host outlet (stream) first, followed by that from the middle faucet (middle slope) and finally by that from the upslope faucet. The more water in

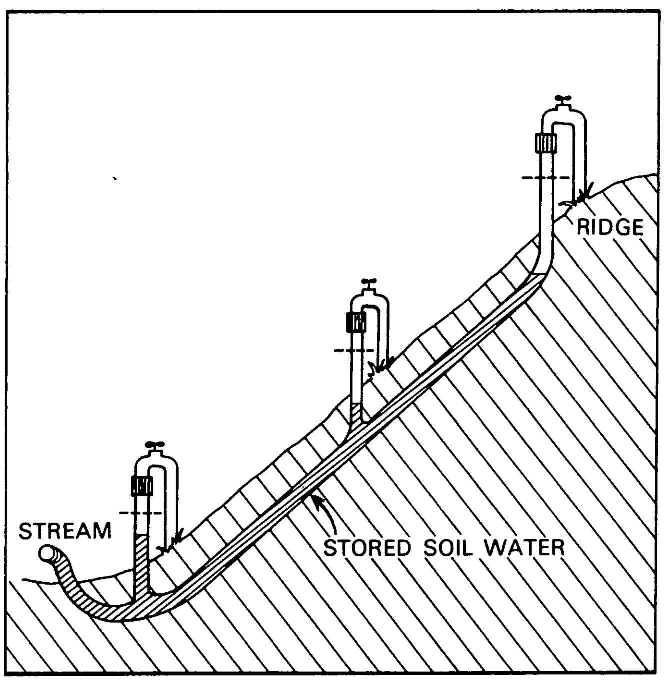

Figure 5.--Stored soil water on a slope represented by water and garden hoses. Turning on faucets simultaneously represents rain falling on the slope.

the hose (the wetter the soil at a certain slope position), the more likely it will be for a certain faucet to contribute to hose outflow (streamflow). For the upslope faucet to contribute to outflow, it must be left on (rainfall must continue) for a period of time sufficient for water from it to fill its hose to the dashed line (increased soil moisture content) so that this water can also contribute to the displacement process.

Two studies in western Oregon tend to to support this displacement theory of rapid response of small, upland watersheds to precipitation (Yee 1975, Harr 1977). In both studies, soil moisture contents remained quite high between storms. (In terms of the garden hose analogy, hoses remained nearly full.) These high moisture contents allowed the soils to respond quickly to subsequent storms; and even 1 meter below the soil surface, soil water movement increased greatly during rainfall.

VARIABLE SOURCE AREA

Perhaps the most important concept in the hydrology of small streams is the variable source area of storm runoff. This concept relates storm runoff to a dynamic source area that expands and contracts according to rainfall characteristics and the capability of the soil mantle to store and transmit water (Hewlett and Nutter 1970). In this way, a channel network grows to many times its perennial dimensions (fig. 6) and streams become both longer and wider. A smaller degree of growth is also exhibited over the course of individual storm events. This degree of change during winter storms is best illustrated by the bottom two watershed conditions in figure 6.

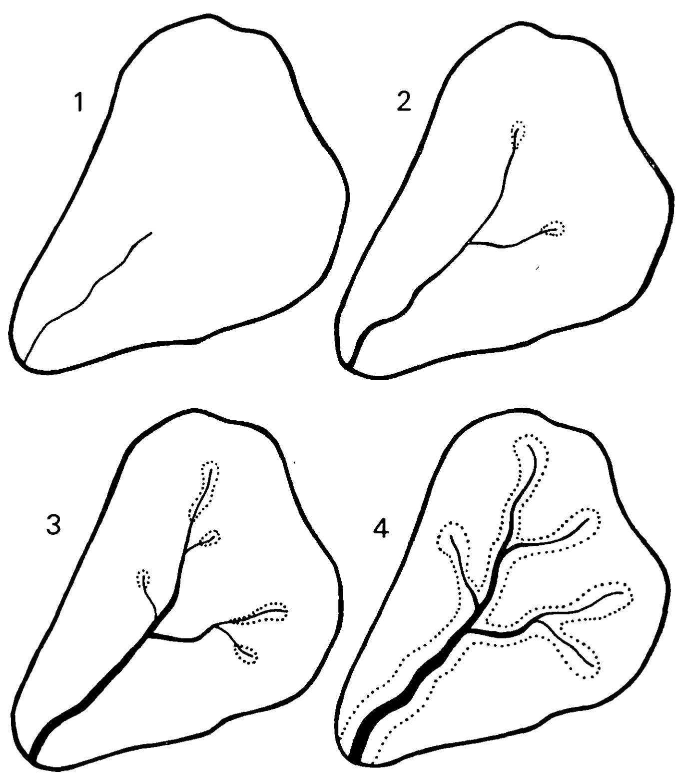

Figure 6.--Time-lapse view of a small watershed showing the expansion of the channel network and source area of storm runoff (dotted lines).

The variable source area concept has been supported by numerous hydrologic studies primarily in the eastern United States (Hewlett 1974). This concept is also consistent with field observations made in the Pacific Northwest during storm runoff periods.

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. In terms of organic debris, the stream represents a depository of variable surface area as well as a mode of transportation. During extreme runoff events, debris not falling directly into the channel of the permanent stream may be subject to transport as the channel system expands. Consequently, channel expansion has important implications for proper disposal of logging debris and for application of fertilizer and pesticides. Unless one is aware of the variable source area of streamflow and its associated channel expansion, it is difficult to determine what the length and width of stream channels will be during winter high flows by simply looking at them during summer low flows. It is even more hazardous to rely solely on maps to determine stream length, because most maps are very inaccurate in their delineation of the channels of small streams. Thus, on-the-ground experience during storm runoff periods is essential to appreciate the dynamic nature of drainage networks.

PEAK FLOW

As the watershed responds to storm rainfall, streamflow increases to a maximum level known as "peak flow." Each storm hydrograph has its own peak flow reflecting the interaction of rainfall with the physical characteristics of the watershed. Higher 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 both the antecedent moisture content of the watershed's soils (how much water is in the garden hose) 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. Many such peak flows occur each year with substantial variation among size of peaks in 1 year as well as among years. Maximum peak flows have resulted from rain-on-snow events during which a substantial portion of streamflow comes from rapid snowmelt concurrent with the downslope flow of rainwater we have been discussing.

Some understanding of the frequency of occurrence of various sized peak flows is helpful to those dealing with small streams. Frequency information is obtained through analysis of peak flow data. The results of one such analysis is shown in figure 7 where magnitude of peak flow is plotted over return period. Return period (T_r) is given by

$$T_r = \frac{n + 1}{m},$$

where, n = years of record and m = ranking number of a peak flow among other peak flows (Linsley et al. 1958). The resulting frequency curve may be used to estimate either the magnitude of a peak flow with a certain return period or the return period of a peak flow with a certain magnitude. For example, the magnitude of a peak flow expected to occur <u>on the average</u> once in 10 years would be about 1.37 m^3/sec-km^2 (125 ft^3/sec-mi^2). Larger peak flows occur less frequently.

The magnitude of peak flow is inversely related to frequency of occurrence (figure 8). As magnitude of peak flow increases, the number of peaks exceeding the magnitude decreases. About 95 percent of peak flows larger than the base value of 0.11m^3/sec-km^2 (10 ft^3/sec-mi^2) are

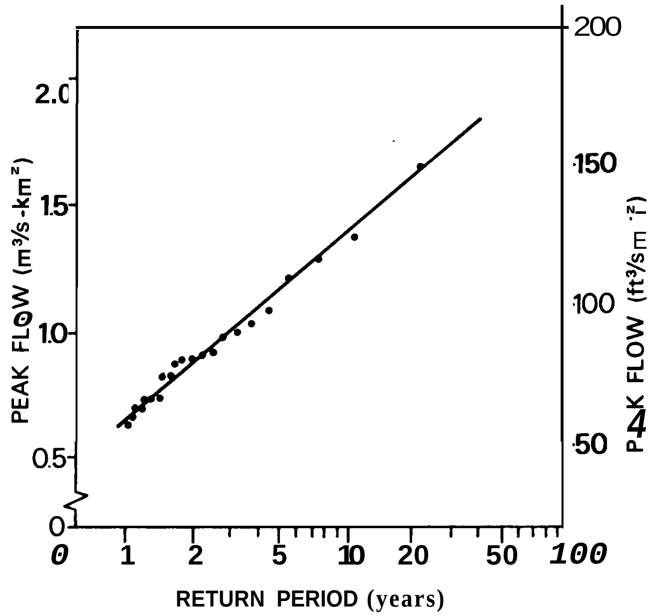

Figure 7.--Frequency curve for storm peak flows at Watershed 2, H. J. Andrews Experimental Forest, 1953-73.

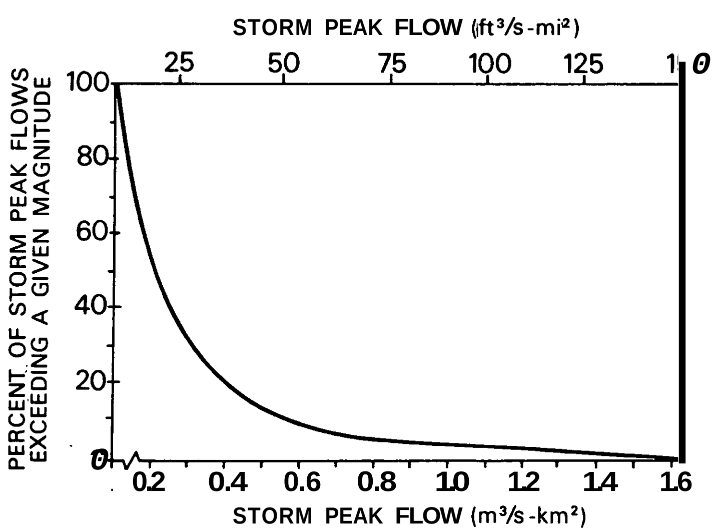

Figure 8.--Size distribution of storm peak flows greater than 0.11 $m^3/s\text{-}km^2$ at Watershed 2, H. J. Andrews Experimental Forest, 1953-73.

smaller than the average annual peak flow as estimated above. Thus, the vast majority of peak flows occurring in a small stream are relatively small. The largest peak flow shown in figure 7 resulted from rapid snowmelt during prolonged rainfall and has a return period of about 25 years. This event, which occurred in December, 1964, is estimated to have had a return period of 50 to 100 years on larger streams having a longer period of record (Rothacher and Glazebrook 1968). Although a minor rainfall-snowmelt runoff event may occur once in 2 years, the heavy rain on a deep snowpack occurs only rarely.

MINIMUM FLOWS

The distribution of annual streamflow for most small watersheds in western Oregon closely follows the distribution of annual precipitation (fig. 9). Maximum monthly flows occur

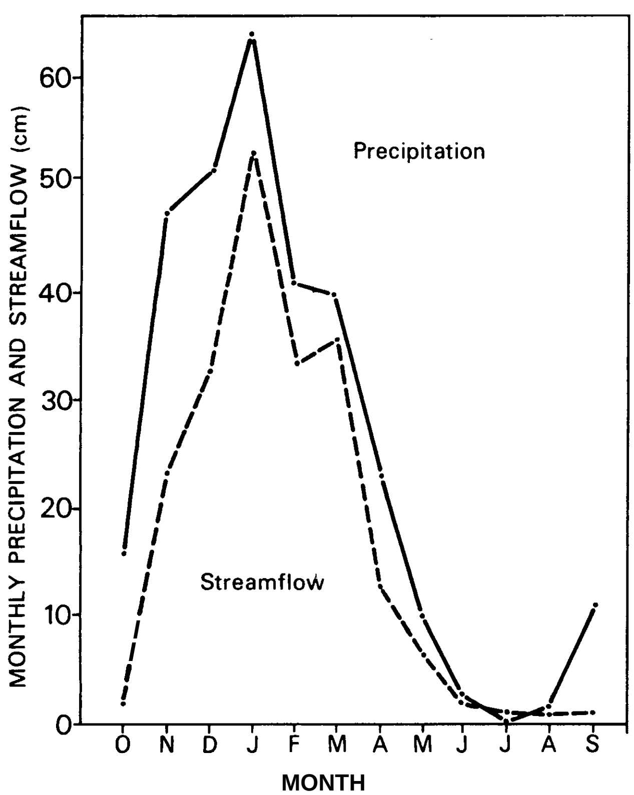

Figure 9.--Monthly precipitation and streamflow at Watershed 2, H. J. Andrews Experimental Forest for 1972 water year.

during the winter months when storms are frequent and relatively large. Streamflow gradually decreases throughout the growing season as sloping soil masses slowly drain and soil water is withdrawn and transpired by plants. Minimum streamflow in perennial streams usually occurs in late summer to early fall. Minimum flow rates may be 1,000 to 5,000 times smaller than maximum peak flows that occur during winter months. Flow may cease entirely in many first-order streams. Extreme variation in flow rates has important implications for channel hydraulics which in turn influence sediment transport and the disposal of organic debris.

Channel Hydraulics

So far this discussion has been restricted to routing water from the watershed slopes to the stream. We have seen that both channel length and width increase markedly over time according to the nature of the interaction among certain hydrologic processes. This section examines the water in a stream channel and shows how its movement through the channel is related to certain characteristics of the channel, that is its gradient, width, depth, and velocity of flow. (The width discussed in terms of expansion of channel networks referred to distance across a stream at a fixed point. This width generally increases and decreases over a water year or during storm runoff. The width to be described in this section refers to the variation in distance across a stream from one channel segment to the next.)

Two basic relationships are involved in water movement through a channel system. First, gravitational and frictional forces affect streamflow. Gravitational forces cause water movement and frictional forces resist flow. The gravitational component depends on the inclination of the channel bed; the steeper the channel gradient, the greater the gravitational component and the greater the tendency for water to flow.

The frictional forces depend on the nature of the bed and banks, i.e., size and shape of bed materials, presence of brush along banks, etc.

The second relationship is between the rate of streamflow or discharge and the hydraulic characteristics, namely, channel depth and width and streamflow velocity. Discharge Q at a fixed point may be expressed as

$$Q = WDV$$

where W = width, D = depth, and V = average velocity of water past a point. From this equation, you can see that for any constant discharge Q, a change in one characteristic on the right-hand side of the equation must be accompanied by a change in at least one other characteristic. For example, if width W doubles from one channel segment to another immediately downstream, depth D, velocity V or both must be decreased such that Q remains constant.

Within a particular drainage network, average channel gradient is inversely related to stream order number. Thus, the small streams we are dealing with are characterized by relatively steep gradients. These gradients are, however, averages; and there is usually considerable deviation from them over the length of a channel. Many streams exhibit a pool-riffle type of appearance in which relatively deep, low-gradient, slow-moving pool segments are separated by relatively shallow, high-gradient, high velocity riffle segments. Such pools and riffles appear to depend on some degree of heterogeneity of bed material size (Leopold et al. 1964) or heterogeneity in resistance of channel bedrock to abrasion. Pools tend to form behind coarser bed materials or behind ridges of rock more resistant to abrasion. In some instances, pools may also form behind logs or accumulations of other organic debris (Heede 1972).

Flowing water possesses considerable kinetic energy directly proportional to the square of flow velocity. As velocity increases with an increase in streamflow or channel gradient, energy of flowing water in a uniform channel, that is, one with no change relative to distance in gradient, depth, or width, will also increase. Energy is dissipated by the resistance between the stream and its bed and banks; by internal resistances caused by boundary features such as bends, bed topography, junctures, etc. that set up eddies and secondary circulations; and by spill resistance caused by water flowing rapidly over an obstruction into flow of much lower velocity. In many small streams, logs and other large organic debris serve as energy dissipators by acting as flow deflectors and as dams over which water must flow. The kinetic energy of flowing water increases and decreases with velocity in a downstream direction as the gradient constantly changes and as water flows through a pool-riffle series of which large organic debris may be an integral part.

Seasonality of Runoff Processes

The seasonality of runoff production may be summarized by describing the sequence of streamflow conditions for a small watershed. This watershed is drained by a permanent, second order stream.

In late summer prior to the fall rains, soil moisture levels are at their lowest because water has been removed by evapotranspiration and slow drainage to streams. Storms are infrequent and small and a large proportion of rainfall, when it does occur, is intercepted by forest vegetation and evaporated with little water reaching the soil. Any storm runoff during this period results almost entirely from channel interception, and peak flows are extremely small, Consequently, the stream has little capacity to move sediment or transport more than small, floatable debris. Streamflow is maintained by slow drainage from soil into isolated saturated zones from which water seeps slowly into the adjacent stream. Streamflow is at a very low rate, and stream network dimensions are at annual minimums, Most first order channels have no flow, and those containing water are short with their flow occupying only a small part of their widths. Flow may consist of a slow trickle between relatively isolated pools.

With the advent of fall rains, the forest hydrologic system becomes more active. Interception losses become proportionally smaller as storms become larger and more frequent. Gradually, rain falling on the moist lower slopes is translated to streamflow to supplement channel interception. As rains continue, soil moisture levels increase upslope, and the stream channel network begins to expand. Because soil becomes wetter and more conducive to translatory flow and because the stream channel network has tapped more of the watershed's slopes, an increasingly greater proportion of each succeeding 'storm is translated into stormflow. Peak flows have also increased.

By late fall or early winter, the runoff production system is near maximum efficiency. Soils have been recharged, and they remain wet between storms. A large proportion of rain falling on the entire watershed is translated into storm runoff. During the larger storms, the extent of the drainage network is at a maximum. Main channels have extended into the more remote parts of the watershed, and even intermittent channels now carry water. Soil water storage in the slopes feeding these intermittent streams, however, is insufficient to maintain streamflow between storms, and the channel network shrinks back to the channel segments which flow continually in the winter. During most winter storms, interception by vegetation is of little consequence. Transpiration rates during the winter are also quite low. Generally, decreases in soil moisture result from drainage of soil between storms. Peak flows are relatively large, and the stream is capable of moving large amounts of debris.

By early spring, evapotranspiration losses become more important and, in terms of runoff production, the watershed becomes less efficient. Storms gradually become smaller and less frequent. Forest vegetation begins to withdraw soil water such that increasing amounts of water are required to recharge the soil water reservoir. Consequently, the portion of the watershed contributing to storm flow contracts toward stream channels. In cases where soil water storage is insufficient to maintain streamflow, flow ceases altogether, and the watershed's drainage network contracts.

Finally, as summer comes, the efficiency of the watershed's runoff production decreases still further. Storms are small, relatively infrequent and interception losses are high. Translatory flow becomes less prevalent, and storm runoff results largely from channel interception. Drainage of soil and subsoil to stream channels continues but at a steadily decreasing rate. Transpiration by forest vegetation reaches a maximum, also decreasing soil water storage, Stream channels shorten until the drainage network is at an annual minumum size.

Effects of Management Activities

Watershed studies in western Oregon have evaluated the effects of timber harvesting activities on streamflow. A brief review of these effects on various aspects of streamflow follows.

ANNUAL STREAMFLOW

Increases in the amount of water flowing from a watershed after cutting forest vegetation has been shown many times in many places (Hibbert 1967). In western Oregon, such increases are near the maximum observed. For example, after extensively clearcutting a small watershed in the Oregon Coast Ranges, annual streamflow increased 60 cm (24 in) (Harris 1973). Similarly, clearcutting an entire watershed in the western Cascade Range increased annual streamflow 46 cm (18 in) (Rothacher 1970). Watersheds less extensively cut showed smaller increases. In terms of the hydrologic system described in this report, removal of forest vegetation drastically reduces interception and transpiration losses such that more soil water drains into streams. As vegetation returns, however, such increases in annual streamflow decrease as both interception and transpiration increase.

MINIMUM FLOWS

Removal of forest vegetation considerably increases streamflow during the summer low flow period. Although absolute increases are small, relative increases are large. For example, minimum flows were tripled after a small watershed in the western Cascade Range was clearcut and burned (Rothacher 1971). Similar increases in low flow were observed in the Coast Ranges (Harr and Krygier 1972). Again, such increases result from drastically reduced transpiration making more soil water available for streamflow. Such increases are relatively short-lived as revegetation occurs.

PEAK FLOWS

Removal of forest vegetation increases average peak flows. Greatest increases have been noted for fall peaks (Rothacher 1973, Harr et al. 1975). Such increases have resulted primarily from changes in soil moisture contents on areas where trees are cut. When forest vegetation is removed, soil moisture contents remain higher than on a forested area so that less fall rain is required to recharge soil moisture, and more rainfall can be translated into streamflow. By winter, however, when soil moisture in a cut area has been recharged, the cut area responds nearly the same hydrologically as it did before its trees were removed. Interception is less after trees are removed; but as we saw previously, interception during larger winter storms is minimal. Thus, timber removal has little effect on the size of large peak flows (Rothacher 1971 and 1973, Harris 1973, Harr et al. 1975).

Soil disturbance, on the other hand, may have a pronounced effect on the size of even large peak flows. For example, road surfaces have virtually no infiltration capacity, and road cuts may intercept slow subsurface flow and rapidly transport water through a ditch-culvert system. As a result, peak flows for small, headwater streams may be increased when roads occupy at least 12 percent of a watershed (Harr et al. 1975). There is some indication that severe soil disturbance resulting from tractor skidding and windrowing slash may also increase larger peak flows (Harr 1976).

CHANNEL HYDRAULICS

Certain forest activities may indirectly affect channel hydraulics of small streams by accelerating the occurrence of soil mass movements, a major form of natural erosion in western Oregon as well as in other areas of extremely steep topography. For example, most soil mass movements in the H. J. Andrews Experimental Forest have been associated with roads (Dyrness 1967). Roads were also the major source of sediment in study in the Coast Ranges (Brown and Krygier 1971). Clearcutting on steep slopes with shallow, residual soils can also increase the frequency of occurrence of soil mass movements (Swanston 1974). Large amounts of soil, rock, and organic debris frequently are deposited in small streams. Channel hydraulics are seriously altered not only by this deposition, but also by the sudden release of water and debris when a dam formed by such deposition fails and a stream channel is scoured to bedrock (Fredriksen 1970).

Channel hydraulics may be adversely affected by removal of naturally occurring large organic debris. In many instances, such debris may have become part of a stable stream environment and, as stated previously, serves as dissipator of a stream's kinetic energy. After such naturally occurring debris is removed in stream cleanup, during or after timber harvest, severe erosion of stream banks and beds may occur. Where debris is keyed into banks, removal may reduce lateral support of soil masses and expose soil to flowing water. Kinetic energy, formerly dissipated by water flowing over debris dams, may be used to erode stream banks and move bed material.

Summary

1. Subsurface flow of water through the soil mantle is the primary mechanism by which water is transmitted from the soil surface to streams. Overland flow rarely occurs because infiltration capacity of undisturbed forest soils is rarely limiting.
2. A watershed's channel system expands and shrinks according to the interaction between rainfall characteristics and the ability of the soil mantle to transmit water. Consequently, a drainage network is much larger during winter storms than during late summer.
3. Flow rates vary considerably. Peak flows in winter may be 1,000 to 5,000 times greater than summer low flows. Peak flow during winter storms may be 100 times greater than flow before storms. The larger a peak flow, the less its frequency of occurrence.
4. Channel hydraulic characteristics vary greatly over time and space due to heterogeneity of bed materials and to the accumulations of large organic debris.
5. Annual streamflow, summer flow, and small peak flows (notably those in the fall) may be increased after clearcut logging. Large peaks are not affected by forest cutting but may be increased when seriously compacted soil occupies at least 12 percent of the watershed.
6. Both clearcutting and roadbuilding may cause soil mass movements in some areas. Such mass movements, as well as those occurring naturally, may drastically change the channel hydraulics of small streams by damming or scouring them.
7. Large organic debris may be part of a stable channel. Removal of such naturally occurring debris during stream cleanup after timber harvest may accelerate erosion of stream banks and channels.

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