3
Hydrology

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Overview

- Streamflow is highly variable in mountainous areas of the Pacific coastal ecoregion. The timing and variability of streamflow is strongly influenced by form of precipitation (e.g., rainfall, snowmelt, or rain on snow).
- High variability in runoff processes limits the ability to detect and predict human-caused changes in streamflow. Changes in flow are usually associated with changes in other watershed processes that may be of equal concern. Studies of how land use affects watershed responses are thus likely to be most useful if they focus on how runoff processes are affected at the site of disturbance and how these effects, hydrologic or otherwise, are propagated downstream.
- Land use and other site factors affecting flows have less effect on major floods and in large basins than on smaller peak flows and in small basins. Land use is more likely to affect streamflow during rain on snow events, which usuallv produce larger floods in much of the Pacific coastal ecoregion than purely rainfall events.
- Long-term watershed experiments indicate that clear-cutting and road building influence rates and modes of runoff, but these influences are stronger for some areas, events, and seasons than for others. Logging and road building can increase areas that generate overland Row and convert subsurface flow to overland fow, thereby increasing rates and volumes of stormflow. Logging and road building can also increase runoff rates and volumes from transient snow packs during rain on snow events.
- Removal of trees, which consume water, tends to increase soil moisture and base streamflow in summer when rates of evapotranspiration are high. These summertime effects tend to disappear within several years. Effects of tree removal on soil moisture in winter are minimal because of high seasonal rainfall and reduced rates of evapotranspiration.
- The rate of recovery from land use depends on the type of land use and on the hydrologic processes that are affected.

Introduction

Streamflow is an essential variable in understanding the functioning of watersheds and associated ecosystems because it supplies the primary medium and source of energy for the movement of water, sediment, organic material, nutrients, and thermal energy. Changes in streamflow are almost invariably linked to changes in other watershed processes such as erosion, sedimentation, woody debris dynamics, and heat transfer-processes that are also important to aquatic communities and discussed in other chapters.

How forest management practices and other land uses affect hillslope runoff and streamflow has long been debated and remains controversial. The controversy is intensified by the difficulty of extrapolating results of watershed studies from one basin to the next because of the variability of hydrologic response with basin size, flow magnitude, season, climate, geology, and type and intensity of land use, Furthermore, given a certain amount of timber to harvest and the associated road systems, it is more difficult to prevent or mitigate potential hydrologic impacts than sediment impacts. The location and manner in which an area is logged or roads are built can effectively control erosion from the small areas that potentially produce disproportionately large volumes of sediment in a basin, but changes in vegetation and soil compaction may affect hydrologic response pervasively. Often such hydrologic responses are more a function of the extent of the area of logging or length of road than the methods of harvesting forests or building roads. Thus, hydrologic response and associated impacts focus the debate on ***how much*** timber to harvest or road to build.

This chapter provides some general information on hydrologic processes and discusses the influences of land use, particularly forestry, on runoff and streamflow in the Pacific coastal ecoregion. A more comprehensive general hydrologic background can be found in texts by Dunne and Leopold (1978) Gordon et al. (1992), Mount (1995), or Black (1996).

Hydrologyof the Pacific Coastal Ecoregion

The Pacific coastal ecoregion ranges from a cool maritime climate with a rather equal seasonal distribution of precipitation in the north to a warm Mediterranean climate with dry summers and wet winters in the south. Annual precipitation and runoff generally increase from south to north (Figure 3.la) (Naiman and Anderson 1996). The amount of precipitation is also strongly influenced by mountain ranges, which have a general north-south orientation. On the windward (western) side of the mountains, precipitation increases with elevation; whereas, on the lee (eastern) side, precipitation drops abruptly because of a pronounced rain shadow which, in extreme cases, produces desert conditions.

Seasonality of runoff is influenced by temperature as well as precipitation. At high elevations and latitudes north of about 48°N, much of the winter precipitation is stored in snowpacks. In the Olympic Range, the North Cascades of Washington, and the coastal ranges of British Columbia and southeast Alaska, melting of winter snowpacks and glaciers produces peak streamflows in spring and early summer and maintains moderate flows throughout the summer (Figure 3.1c). Glaciers are present in the Olympics and North Cascades of Washington, the high coastal ranges of British Columbia and Alaska, and on the highest volcanoes in the Cascades of Oregon and California. Meltwater runoff from glaciers commonly peaks during the warmest periods of the summer in Julv and August.

Runoff from snowmelt is limited by how rapidly thermal energy in the air can supply enough heat to melt snow. It takes only 1 cal to warm 1 g of water 1 °C, but it takes 80cal to melt 1 g of ice at 0°C. In addition, the low density and specific heat of air compared to that of ice (approximately 0.1% and 20%, respectively) severely limit the direct transfer of heat from air to snow. However, the transfer of latent heat from air to snow (i.e., condensation) can produce rapid snow melt. Nightly cooling often limits snowmelt runoff rates that are generated over several days. Maximum rates of snowmelt approach approximately 4cm/day (Dunne and Leopold 1975). and resulting runoff rates tend to be less, because of mixing of runoff from areas of varying rates of snowmelt. These rates compare to runoff rates of approximately 9cm/day during very large floods generated at least partly by rainfall in the Pacific coastal ecoregion.

The highest rates of runoff in the world, other than from dam-break floods, are generated by rainfall. However, in the Pacific coastal ecoregion, rapid snowmelt commonly accompanies the influx of warm subtropical air masses that also produce some of the highest sustained rates of rainfall, and the combination rain on snow events produce most of the largest floods. Some snowmelt during rainfall occurs every year usually without serious consequence. However, rapid snowmelt during

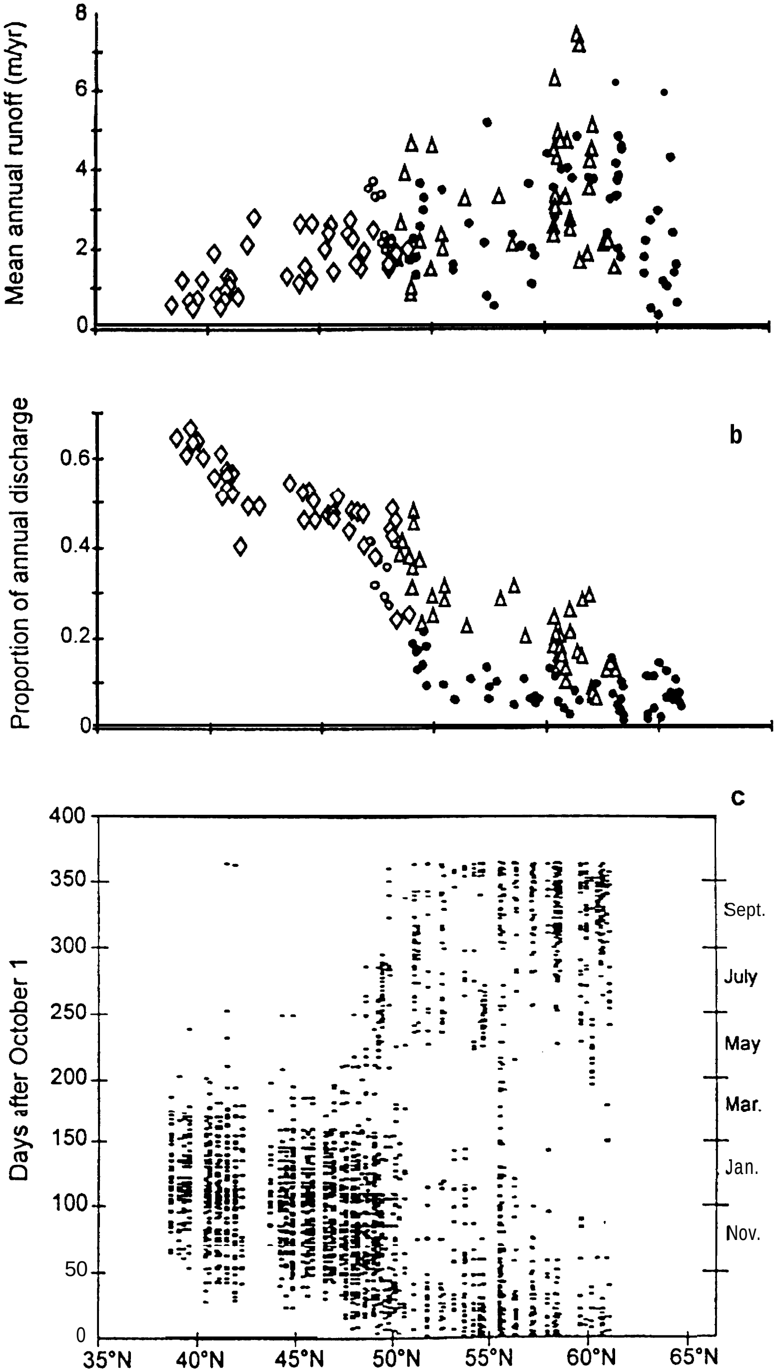

FIGURE 3.1. Variation with latitude of hydrologic characteristics for 151 rivers draining four subregions of the Pacific coastal ecoregion: (◇) southern coastal mountains, (o) Olympic Mountains, (∧) northern lowlands and islands, (•) northern mainland mountains. (a) Mean annual runoff as a function of latitudc. (b) Proportion of annual discharge carried during the three months of winter. (c) Temporal distribution of peak annual discharge. Each point represents the date and latitude of a maximum daily discharge for a given river in a single water year (93 rivers total having at least 10 years of record) (Naiman and Anderson 1996 with permission).

rainfall contributed to all but two of the 23 largest annual peak flows of the Willamette River at Salem, Oregon, between 1814 and 1977 (Harr 1981). In the 60-ha Watershed 2 of the H.J. Andrews Experimental Forest, major peak flows of 10L/s/ha are five times more likely to result from rain on snow than from rain atone (Harr 1981). Although the phrase "rain on snow" implies the snow is melted directly by warm rain, the snow is primarily melted by heat transferred to the snowpack by convection and condensation of water vapor on the snowpack surface. While this distinction may seem trivial, it is important to identify the appropriate process before questions concerning the influence of land use on the magnitude of rain on snow floods can be answered. Detailed analyses of such runoff events are rare because information is almost atwavs tacking about snow depth and density, air temperature, and form of precipitation during any given storm.

South of latitude 48°N (California, Oregon, and much of Washington), about 80% of the total annual precipitation, which ranges from 750 to 3,500mm, falls during the six-month period between the beginning of October to the end of March. During winter, frequent frontal storms move eastward from the Pacific, typically producing relatively low-intensity precipitation (e.g., less than about l0mm/hr for periods of 18 to 72 hours). Temperatures, regulated by latitude and elevation cause precipitation to fall as rain, snow, or a combination of the two and govern the magnitude and timing of associated peak streamflows. Patterns of streamflow discharge reflect the strong contrast in precipitation between summer and winter (Figure 3.lb). Except at high elevations, up to 70% of the annual streamflow occurs during the three months of winter (Naiman and Anderson 1996). Large streamflow peaks are generated when a winter storm is particularly strong or when several storms follow in rapid succession and produce moderately intense rainfall over a period of several days. Lack of rainfall during the summer results in low summer streamflows. Small headwatcr streams commonly become dry before the onset of the fall rains, except at the highest elevations where streams arc fed by late-melting snowpacks or glacicrs.

The proportion of annual precipitation falling as snow varies greatly with elevation and latitude. For example, in western Oregon, snow is uncommon below about 350m. At intermediate elevations from 350 to 1,100m snow is intermittent, tasting only a week or two between warm periods. Above 1,000m, one-third to three-fourths of the annual precipitation may fall as snow, which begins to accumulate in November and usually lasts until late May. Further south, in northern California, these elevational zones are about 1,000m higher. In California, high-elevation lands occupy about 3% of the state but produce about 13% of the annual streamflow (Colman 1955). The high-elevation lands are of even greater importance in Utah, where 60% of the state's streamflow comes from the upper Uinta and Wasatch mountains which occupy nly 10% of the land. In the Rockv Mountains, 85% of the annual streamflow occurs from May through July, with less than 5% occurring during the winter months (Leaf 1975).

In the southern Pacific coastal ecoregion, the general transition from rain to snow with higher elevation strongly affects the magnitude and timing of runoff events. Figure 3.2 depicts typical seasonal variations in streamflow between three streams in Washington that drain basins having different elevations. In the Wynoochee River (gaging station elevation of 12m), streamflow reflects rainfall, which is concentrated in winter. In the Middle Fork Snoqualmie River (elevation 240 m), winter precipitation falls as either rain or snow. The largest floods occur when heavy rainfall is produced by a warm subtropical air mass that also raises the freezing level and results in a rapid melt of an existing snowpack. Smaller peak flows occur as the residual snowpack melts in late spring. In the Twisp River (elevation 850m), winter precipitation falls as snow and produces peak flows when the snowpack melts in late spring. Consequently, the hydrology is dominated by three precipitation types: rain, rain on snow, and snow. Lower and more southern parts of the Pacific coastal ecoregion arc exclusively within the rain zone, and

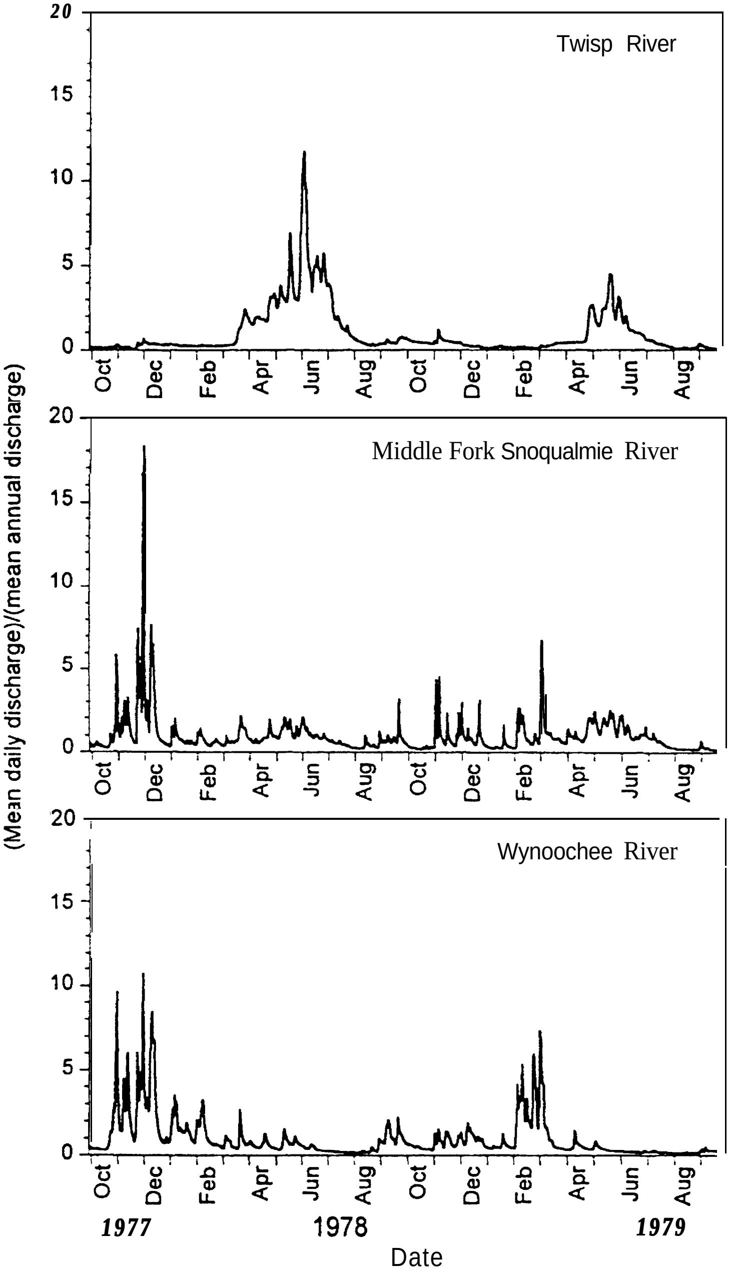

FIGURE 3.2. Hydrographs for three Washington rivers (Twisp River, drainage area = 680km^2; Middle Fork Snoqualmie River, 430 km^2; Wynoochee River, 430km^2) for water-years 1978 and 1979. Mean daily discharge is divided by mean annual discharge to normalize magnitudes (discharge data from USGS 1997).

northern, high-elevation areas are always within the snow zone. However, high-flow events are the result of varying combinations of rain and snow. Thus, the relationship between climate, runoff, and land use is highly complex

Runoff Processes

Runoff processes are linked throughout a watershed from hillslopes to the mouth of the main channel. Rates of moisture movement along these pathways depend on the volume of water introduced to the system and the efficiency with which it is transported through the system. A simple mass balance equation is useful for understanding runoff processes throughout a drainage:

Outflow = Inflow - Losses - Change in storage

where outflow is runoff from a hillslope or reach of channel; inflow is rainfall or streamflow into the channel reach; losses are processes such as evapotranspiration and deep seepage; and change in storage includes, for example, pool volume and soil moisture. Therefore, the rate and capacity for filling and depleting stored water is a key element in the timing, magnitude, and duration of runoff rates in all parts of a watershed. Reduced water storage in one part of a basin leads to increased release of water downstream. For example, there is much less "storage" available for water moving over the surface of hillslopes than there is below the surface. This is in part why water flows down hillslope surfaces more than 10 times faster than it does through the soil mantle. Thus, increasing surface runoff at the expense of subsurface runoff can increase peak flows in channels downstream. Reviews of the effects of land use on the hydrology of hillslopes and channels are provided by Kirkby (1978) and Reid (1993).

Hillslope Runoff

It is useful to focus on the effects of land-use practices on hillslope runoff processes for two reasons. First, detecting and then evaluating the causes of changes in streamflow is more difficult further downstream. Second, information gained by evaluating how and where land use affects runoff processes can help determine how other processes, such as erosion and sediment delivery, are affected. Once these processes are understood, ways to prevent altered runoff and erosion may be identified. For example, whether rain or meltwater runs over the soil surface or through the soil mantle strongly influences how quickly it arrives at a stream channel and how much sediment and solutes arc produced. Figure 3.3 depicts some important hydrologic pathways in the Pacific coastal ecoregion that are described below, A more comprehensive review of runoff processes is provided by Dunne and Leopold (1978) and Kirkby (1978).

Subsurface Flow

Subsurface flow accounts for nearly all of the water that is delivered to stream channels from undisturbed forested hillslopes (Harr 1977). Precipitation infiltrating the soil surface travels through the soil as either shallow subsurface flow or deep seepage that replenishes groundwater storage. Groundwater storage maintains base flows during dry periods. Water is transmitted within the soil along two different flow paths: through the soil matrix (micropores) and through macropores including root holes, soil cracks, animal burrows, and soil pipes. Subsurface flow velocities vary widely (Table 3.1). Flow in macropores is very slow (10^{-7}-10^{-6}m/hr), while flow in soil pipes (10^{-1}-10^{2} m/hr can be as rapid as unchannelized overland flow. Normally, subsurface flow contributes little to erosion. However, if subsurface flow paths become obstructed, water can build up in the soil mantle and cause slope failure.

In many locations, large macropores or structural voids can occupy as much as 35% of the total soil volume in a forest soil (Aubertin 1971). In the coastal mountains of British Columbia, Chamberlin (1972) observed that roots made up about 50% of the upper 0.5m

Table 3.1. Order of magnitude of runoff velocities by different processes.

Runoff process	Characteristic velocity (m/hr)
Subsurface	
undifferentiated	10^{-7}-10^{-4}
micropore	10^{-7}-10^{-6}
macropore	10^{-5}-10^{-4}
soil pipes	10^{-1}-10^{2}
Surfacc	
overland flow (unchannelizcd)	10^{1}-10^{2}
channel flow (gullies and stream channels)	10^{2}-10^{3}

Modified from Dunne and Leopold 1978, Kirkby 1978.

of the forest soil. When voids resulting from root decay, animal activity, and subsurface erosion by chemical (solution) and physical processes become interconnected, they form soil pipes capable of transporting subsurface water rapidly (Table 3.1) (Kirkby 1978). Where piping networks are extensive, the hydraulic conductivity of the soil matrix is of secondary importance in generating flow during storms (stormflow) (Whipkey 1965, Mosley 1979). For example, pipeflow accounted for nearly all of the stormflow from l-ha headwater swales in northern California (Ziemer 1992) and Japan (Tsukamoto et al. 1982). In a 50-ha coastal drainage in central California, Swanson et al. (1989) reported that during a 25-year recurrence-interval storm, nearly 70% of the water was discharged through the subsurface piping network. Jones (1987) found that pipeflow was responsible for 49% of the stormflow from the Maesnant catchment in Wales.

Saturated Overland Flow

Overland flow can occur where the soil becomes fully saturated and subsurface flow emerges as *return flow* (or ***exfiltration);*** additional rainfall or meltwater flows over the surface as *saturated overland flow.* Soils commonly become saturated where shallow subsurface flow converges in topographic depressions or accumulates in areas of decreasing hillslope gradient. Zones of saturated overland flow are most common in valleys and swales. Zones of saturated overland flow commonly occupy small but expandable areas of drainage basins and contribute disproportionately to flows during storms; they expand during wet periods and contract during dry periods. This phenomenon is known as the *partial-area concept* of storm runoff (Betson 1964, Dunne and Leopold 1978).

In undisturbed Pacific coastal ecoregion forests, areas that generate saturated overland flow are usuallv confined to the base of hillslopes, near stream channels, and in swales (Figure 3.3). They can also occur where soils thin downslope over impermeable bedrock. Where hillslopes are straight, steep. and highly permeable, there is little tendency for return flow to occur. However, mechanical disturbance of areas that generate overland flow can

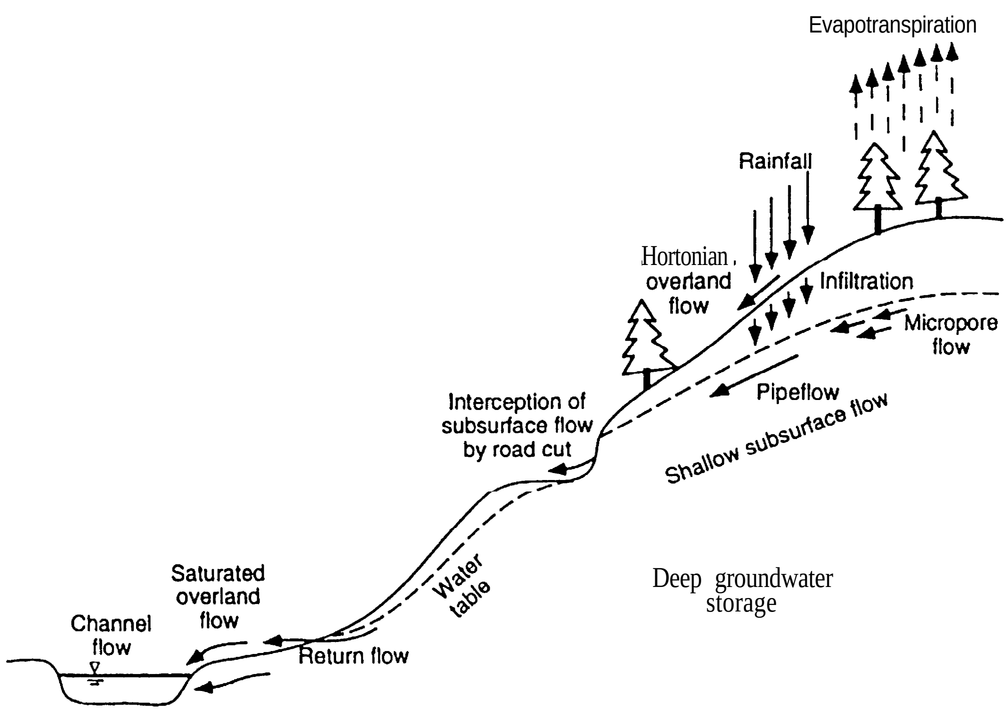

FIGURE 3.3. Distribution of hydrologic processes on an idealized hillslope in the Pacific coastal ecoregion.

create large sediment sources by decreasing soil stability where surface flow energies are high.

Hortonian Overland Flow

When water encounters the ground surface more rapidly than it can infiltrate into the soil, the excess water runs over the surface as *Hortonian overland flow* (Figure 3.3) (after Robert E. Horton, who proposed principles of surface runoff). Because infiltration rates of wetted forest soils typically exceed rainfall rates in the Pacific coastal ecoregion, Hortonian overland flow is unusual in undisturbed forests in the region. It is more likely to occur in more arid climates and agricultural lands where infiltration is less and maximum rainfall intensities are greater. Nonetheless, Hortonian overland flow can occur locally following fires in which volatilized organic molecules coat soil particles producing a water-repellent layer that prevents water from infiltrating into coarse textured soils (DeBano 1969, Beschta 1990, McNabb and Swanson 1990). These hydrophobic conditions can extend to a depth of 15cm and persist for six or more years after the fire (Dyrness 1976, DeBano 1981).

Where and when overland flow occurs governs many of the impacts of watershed disturbances. Firstly, overland flow can travel at much greater velocities (10^1-10^2m/hr) than undifferentiated subsurface flow (10^{-7}-10^{-4} m/hr) (Table 3.1). Therefore, increased areas of Hortonian overland flow directly contribute to streamflow peaks during storms in headwater channels, which respond to the most rapid components of runoff. Secondly, overland flow has a much greater capacity to erode and transport sediment. An increase in Hortonian overland flow in mountainous terrain is likely to be accompanied by soil loss and an increase in sediment load to streams.

Land use can increase areas of Hortonian overland flow and saturated overland flow and therebv increase hillslope erosion and stormflow magnitude in headwater channels. In rangeland. soil compaction and loss of ground cover from heavy grazing can decrease soil permeability (the soil's capacity to transmit water), and thereby increase the area and frequency of Hortonian overland flow (Horton 1933, Dunne and Leopold 1978). Hortonian overland flow is increased in urbanized areas by the expansion of impervious surfaces (e.g., roofs, streets, parking lots) (Leopold 1968, Dunne and Leopold 1978, Sauer et al. 1983). In forested lands of the Pacific coastal ecoregion, Hortonian overland flow is most commonly restricted to areas of compacted soils, such as roads, skid trails, and landings. Subsurface flow emerging in road cuts directly augments surface flow to streams through road ditches (Figure 3.3), and can contribute to erosion of road cuts and surfaces as well as road ditches (Megahan 1972). Therefore, by intercepting subsurface flow and causing Hortonian overland flow, roads can expand the channel network in a basin and thereby increase the rate of stormflow runoff (Wemple 1994). Converting subsurface flow to overland flow is especially effective in increasing stormflow volumes and rates where subsurface flow is dominated by flow through micropores, but less so where pipeflow is intercepted. When subsurface flow through the soil matrix is converted to surface flow. runoff velocity is increased by as much as five orders of magnitude (Table 3.1). But when pipes or large macropores are present, differences in runoff velocity between subsurface pipeflow and surface runoff indicate that a shift to surface runoff mav result in an increase in runoff velocity of onk order of magnitude or less. Mechanisms for changes in runoff processes resulting from logging and road building are conceptualized in Figure 3.4.

Although it is relatively easy to understand and detect human disturbance of runoff processes on hillslopes. evaluating downstream effects of altered runoff processes becomes increasingly difficult as the size of the basin increases. Increased surface runoff is usually easiest to detect nearest the disturbed area. One reason is that an enhanced spike in stormflow runoff attenuates downstream as the flood wave spreads out and mixes with unaffected or less-affected runoff from other parts of the basin. Also, the arrival of runoff peaks from sources upstream may or may not coincide at some point in the trunk stream. Altered

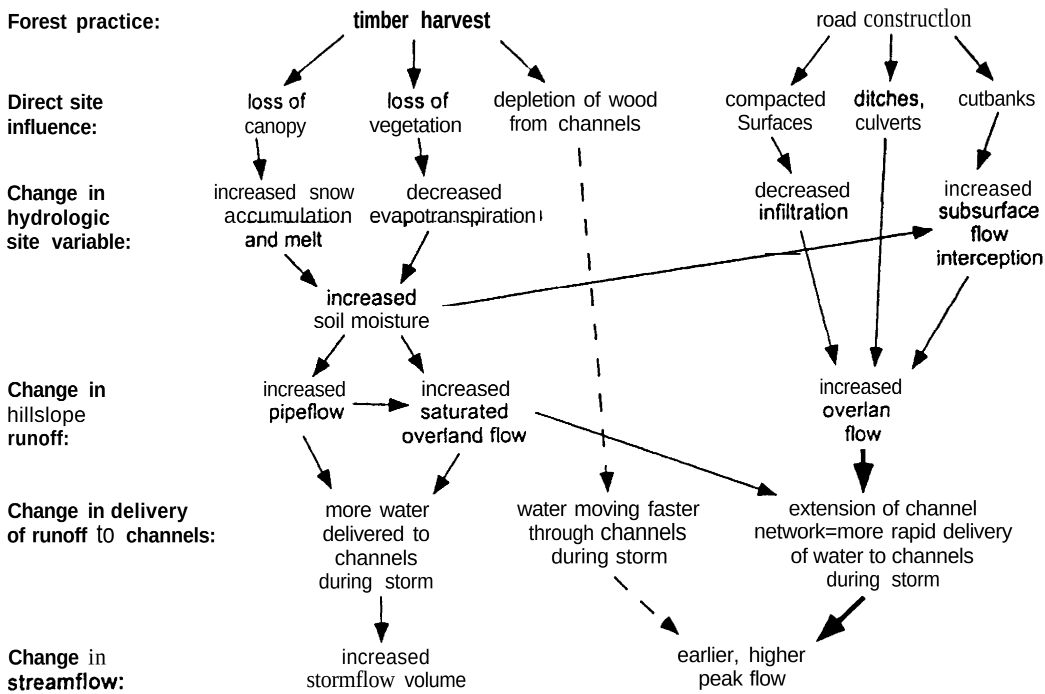

FIGURE 3.4. Conceptual hierarchical mechanisms for alteration of hillslope and channel runoff by timber harvesting and road building. The strength of linkages to higher runoff rates increases from dashed to solid to heavy lines (modified from Jones and Grant 1996).

runoff may enhance or decrease these coincidences and thus either increase or decrease downstream peak flows.

Increased erosion and sediment transport usually accompany increased runoff. For example, in the 1960s and 1970s diverted road drainage was one of the primary sources of sediment in areas of the Redwood Creek basin. California, that were logged and had roads (Figure 3.5), (Weaver et al. 1981). Plugged culverts and other failures of road drainage diverted runoff onto hillslopes where high rates of surface flow never had occurred before. The result was deep gullying-an outcome of extreme local increases in surface runoff. Eroded material was added to the already high sediment loads of the affected tributaries. but whether changes in streamflow could have been detected is debatable.

Runoff in Channels

Streamflow, in contrast to hillslope runoff, pertains only to surface flow in the channel- although subsurface flow below channels and floodplains is verv important to benthic and hyporheic organisms Differences in runoff processes between basins or regions can strongly affect the variability of peak flow in channels. For example, Pitlick (1994) compared the ratio (Q_{100}/Q_m of the magnitude of the 100-yr flood (Q_{100} discharge with a recurrence interval of 100 years) to the magnitude of the mean annual flood (Q_m) in five climatic zones in the western United States. In alpine areas in Colorado where snowmelt dominates flood runoff, Q_{100}/Q_m is approximately 2. In the California Coast Ranges. where large frontal storms dominate flood runoff, Q_{100}/Q_m is 3 or more, and in the Klamath Mountains, where rain on snow is more common, Q_{100}/Q_m is approximately 5.

The variation of flow during and between seasons is a key selective pressure on aquatic and riparian organisms and a primary control on channel form and process. Each season has a characteristic flow frequency which is vital to ecosystem function. Some of the seasonal

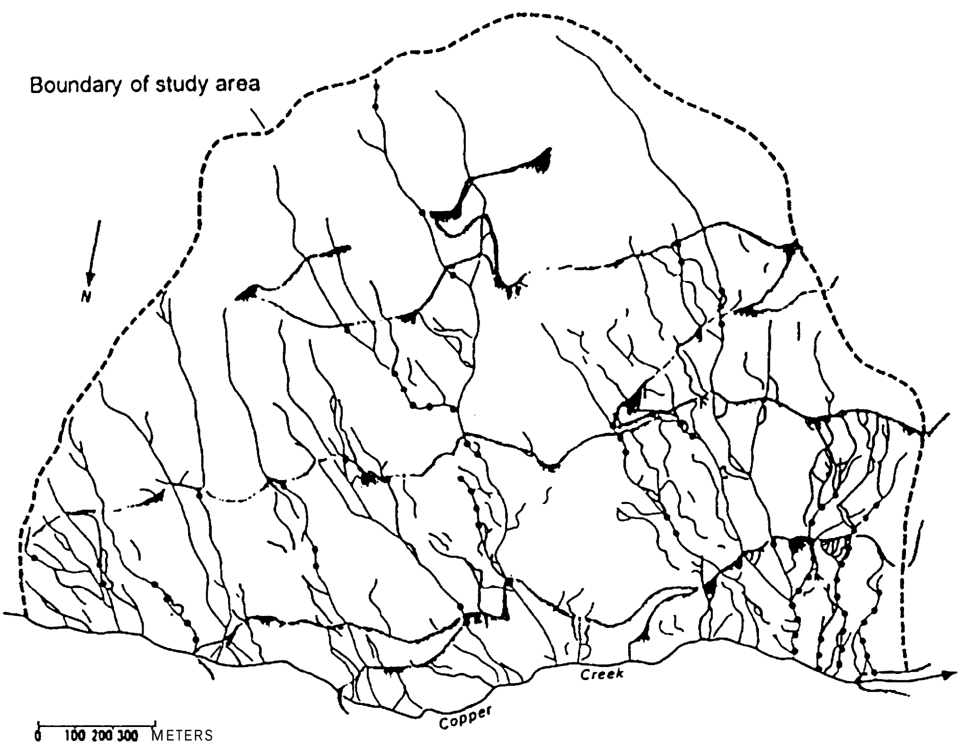

FIGURE 3.5. Gully systems that resulted from road drainage failures on a 246-ha study site in South Copper Creek sub-basin, Redwood Creek basin, California (mapped in 1979). Rills less than 0.1 m^2 in cross-sectional area are not shown. (Weaver et al. 1995).

effects of flow variability in the Pacific coastal ecoregion are summarized below:

- The timing and extent of fish spawning runs, which commonly begin in the fall, depend on flows high enough to allow fish to enter and penetrate the channel network.
- Annual floods, which commonly occur in winter and spring, distribute sediment and organic debris through the system, scour the bed, and remove newly established vegetation in the active channel. Floods can increase mortalitv of incubating fish embryos, and depending on incubation periods of different species and the timing of floods, they may strongly affect relative cohort populations of different species. Floods cause mortality in certain benthic invertebrates, and alter food webs, and therebv affect the entire trophic structure of aquatic communities (Wootton et al. 1996).
- Extreme flood events create new surfaces by erosion and deposition. Aquatic and riparian ecosystems in mountainous areas depend on extreme events (e.g., floods, landslides. windstorms, fire) to renew dynamic processes and maintain a mosaic of surfaces that are at various stages of evolution since the previous disturbance (Grant and Swanson 1995).
- Recessional flows in spring and early summer are occasionally punctuated bv peak flows. Streamflow during this period controls the success of riparian plant seeds to germinate in channels and on streambanks and floodplains. Seeds of riparian trees commonly disperse over a time frame of a few we&s. In order to successfully germinate, seeds must be deposited high enough to avoid drowning or scour, and low enough to avoid desiccation as water tables drop with recessional stages (McBride and Strahan 1985, Lisle 1989 Segelquist et al. 1993). Seeds may be swept away or germinate, and seedlings may be drowned, desiccated, or survive depending on water stages in spring and early summer.

- Summer low flows allow sediment to settle, water to clear, and low-energy habitats to expand. Low flows also limit total aqueous living space.

Thus, the entire annual sequence of flows governs the trajectory of aquatic and riparian ecosystems. As a result, stream channels and ecosystems are in a constant state of flux within a wide range of variability, and the duration and frequency of occurrence for any given state span decades or centuries (Chapter 11). The following discussion of the relative influence of flows of different magnitude on channel form assumes relatively constant supplies of woody debris and sediment. However in nature, woody debris and sediment are supplied at widely varying rates and greatly influence the immediate and enduring effects of high-runoff events on channels and floodplains (Chapter 11).

A wide range of flows that entrain bed material and erode banks is responsible for forming stream channels. Low sediment-transporting flows occur frequently, but are too weak to move much material over a short time span. Extreme floods cause large volumes of erosion and deposition, but occur infrequently (e.g., once every several decades). A channel is a legacy of the history of flows that it has carried. If an extreme event has occurred recently. the channel retains most of the forms and dimensions created by that flood; the cumulative effects of the succeeding smaller floods are insufficient to significantly alter the channel. However. if the last extreme event occurred a decade or more ago, channel characteristics are likely to be adjusted to moderate floods that are both large enough and frequent enough to move significant volumes of material and reshape the channel. Thus, over the long term. the magnitude and frequency of a given discharge determine its effectiveness in altering the channel, and in the short term, the occurrence of a given discharge in the sequence of preceding flow events determines its role in shaping the channel's current form (Wolman and Miller 1960).

The *effective* discharge is the flow magnitude that, over a period of years, transports the most scdimcnt or does the most work in forming the channel (Wolman and Miller 1960, Benson and Thomas 1966, Andrews 1980). Effective discharge is measured by finding the maximum product of flow frequency and sediment transport rate among equal ranges of discharge. The range of flow that transports the most sediment is commonly the one that fills the channel to the top of its banks. However, equating sediment transport to channel formation is complicated by the usual dominance of suspended sediment in the sediment load, which, because it comprises a minor fraction of bed material, may play a minor role in channel-forming processes (Wolman and Miller 1960, Wolman and Gerson 1978, Ritter 1988). The frequency of effective discharge varies widely (Nash 1994), but it commonly corresponds to bankfull discharge (the discharge that fills the channel to the top of its banks and just begins to overflow onto the floodplain) in magnitude (Andrews 1980). However, in mountainous areas of the Pacific coastal ecoregion. effective discharge can be difficult to define statistically and often exceeds bankfull discharge (Nolan et al. 1987. Grant and Wolff 1991). Moreover, many of the large bed particles that for-m the structure of mountain channels are moved only during rare floods (Grant 1987).

Bankfull discharge is a useful reference because it can be measured in the field, it is theoretically related to channel-forming processes, and it has a characteristic frequency (Wolman and Leopold 1956). For example. bankfull discharge is a key component of strategies to maintain channel forming discharges in water-rights decisions. Bankfull discharge is usually equaled or exceeded, on average, every one to five years in channels that are neither aggrading nor degrading (Williams 1978). However, a consistent bankfull stage may be difficult to recognize in channels in mountainous areas of the Pacific coastal ecoregion.

Flood Routing

How do changes in hillslope runoff processes translate to changes in streamflow, and how do these propagate downstream? Downstream changes in streamflow are best understood and most easily quantified by the conservation principle introduced earlier:

Outflow = Inflow - Losses - Change in storage

To illustrate this further, imagine a reservoir (Figure 3.6). At some initial time (t = 0), the equation is balanced: there are no changes in any of the components, and the lake level is in equilibrium. At some later time (t = 1), an increase in flow at the inlet is not transmitted immediately to produce an equal increase at the outlet. Instead, the lake level rises gradually as it fills, and this rise causes a slow increase in the outflow. Later (t = 3), as the inflow decreases, the outflow exceeds inflow and decreases slowly as the lake level gradually falls. During this period of transition (t = 2), equilibrium between inflow and outflow may be achieved for a short time. The effect of storage is that the inflow hvdrograph has a higher peak flow of shorter duration than the outflow hvdrograph.

Although the principles outlined above are still valid, a channel differs from a reservoir in that flow is retarded, and water is thereby stored by frictional resistance along the entire channel rather than by a dam. For a section of channel, the inflow includes flow from the upstream channel plus runoff from hillslopes and tributaries entering the channel along the reach. “Reservoir” storage is provided by the channel itself along with its floodplain. Storage within the channel primariIy depends on channel size and roughness: increased channel roughness causes the flow to slow and deepen: decreased roughness causes water to evacuate the channel more quickly. Simplifying channels and removing roughness elements such as riparian vegetation and large woody debris reduce channel storage of runoff and contribute to higher peak flows downstream.

Flood storage outside of the channel can vary from near zero in channels tightly constricted by vallev walls to huge volumes in reaches bordered by extensive floodplains and wetlands. In mountainous areas of the Pacific coastal ecoregion. floodplains are commonly small and infrequent, and valley flats are dominated bv rarely tloodcd terraces. Wide floodplains, on the other hand, provide a buffer to flows greater than bankfull for downstream channels. An increase in flow greatly increases the storage which dampens the increase in flow downstream. Channelization exacerbates downstream flooding by removing roughness elements and isolating a channel from its floodplain where floodwaters can be stored. One of the greatest threats to flood control, paradoxically, is confining flood waters to channels because the reduced upstream storage increases the potential for more serious flooding downstream. Increased rates of hillslope runoff or channel straightening can cause streambed erosion, and the resulting increase in channel depth can confine high flows to channels.

The exchange of flood water and sediment between channels and floodplains provides a vital link between aquatic and riparian ecosystems (Gregory et al. 1991). Flood water from the channel carries suspended sediment that settles out on floodplains and in backwaters, thus the return flow to the channel can be partially cleansed of sediment. The floor of floodplains is usually rich in organic matter that can be transported to the channel by the return flow and enrich the channel ecosystem.

Subsurface Flow in Channels and Riparian Zones

Subsurface flow in channels and floodplains performs vital ecological functions (Chapter 16). Flow in the hyporheic zone-defined by Edwards (Chapter 16) as the saturated sediments beneath and beside a river channel that contain both surface and ground water-rarely attains the discharge or velocity of channel flow, but it can become a large component of total water discharge when surface summer flows become extremely low, even in the largest rivers. Partial filling of channels with coarse sediment can cause a greater proportion of channel runoff to become subsurface flow. Subsurface discharge in Little Lost Man Creek, a small, pristine cobble-armored channel in northern coastal California, was approximately one-quarter of surface flow during the summer (Zellweger et al. 1989). Nominal subsurface flow velocities under the channel and

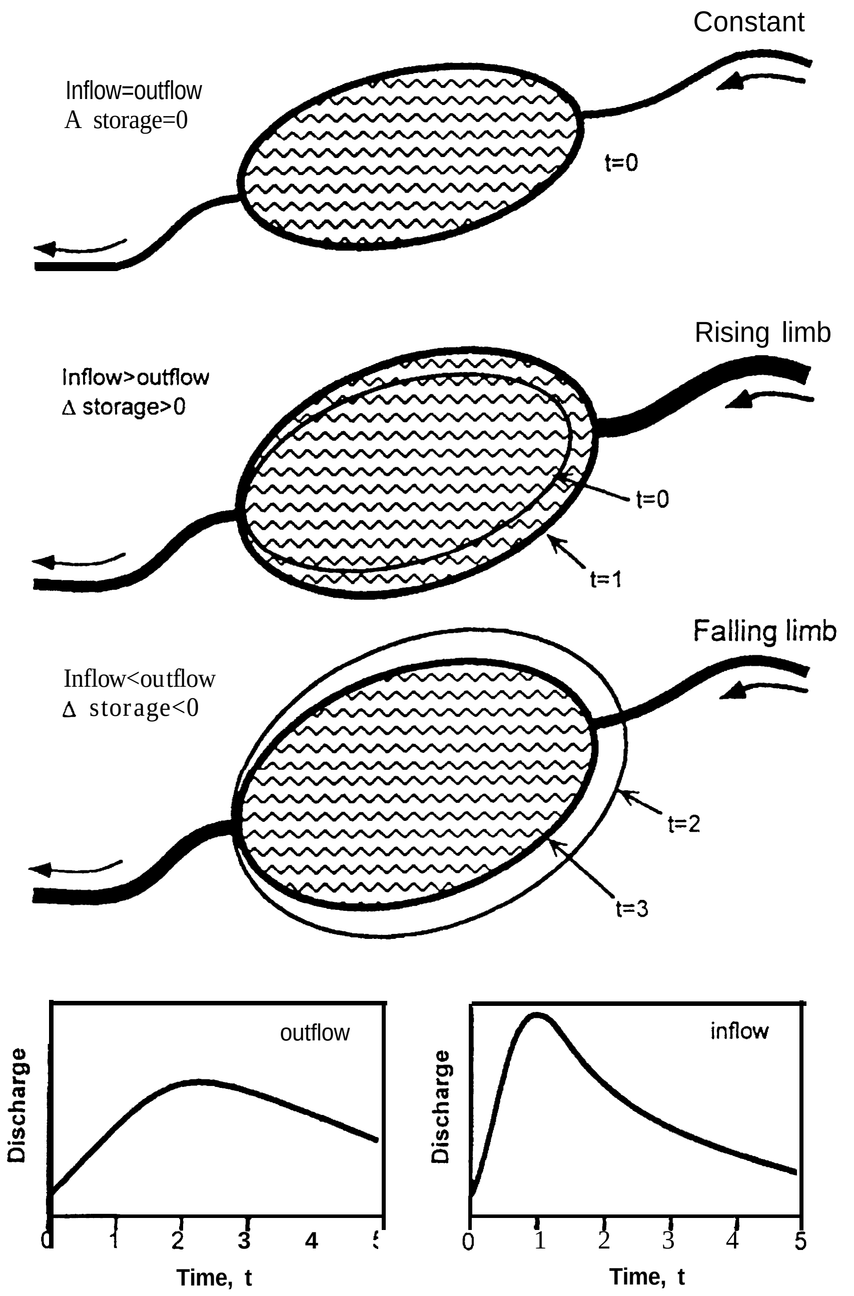

FIGURE 3.6. Flood routing in a channel with a reservoir during constant flow (t = 0) and the rising (t = 1) and falling (t = 3) limbs of a peak flow event. Volumes of storage in the reservoir during four time periods (t = 0 to 3) are represented by oval size: finer lines outline the water surface area in the reservoir during the preceding time interval. The patterned oval depicts reservoir level at corresponding inflow and outflow rates. At t = 2 (not shown). outflow equal inflow.

riparian corridor ranged from 0.4 to 13m/hr (Triska et al. 1993).

Emerging subsurface flow in channels moderates more extreme seasonal water temperatures (Bilby 1984, Nielsen et al. 1994, Keller et al. 1995). For example, intergravel flow through gravel bars can emerge in still pools during low summer flow and provide cold-water refuges for salmonids when ambient stream temperatures exceed lethal limits (Nielsen et al. 1994). Other sources of cool water include surface and subsurface flow from tributaries and seeps from streambanks. Water in stratified pools in summer is commonly 3 to 9°C cooler than surface water. Similarly, seeps in off-channel habitats can provide warmer water in midwinter (Hetherington 1988).

Effects of Land-Use Practices on Runoff

Land uses such as forest practices. road building, and livestock grazing have important implications for peak flows and floods, water yield, and hvdrologic recovery. The effects of land uses varies with basin size and the magnitude of flows, and recovery processes, which vary in space and time, depend on the type of disturbance and the hydrologic processes affected.

Peak Flows and Floods

Anthropogenic influences on the magnitude of Roods is a recurring issue largely because of the high natural variability of flows, especially flood flows. The difficulty of detecting changes in flood size caused by land use is illustrated by Hirsch et al. (1990):

> It is not uncommon for annual floods to have a coefficient of variation (ratio of standard deviation to mean) of one or even more. Suppose the coefficient of variation of annual floods was one and the frequency distribution changed abruptly halfway through a 40-year annual flood record; in order for the change to be discernible in a statistical test with 95% power, the change in the mean would have to be at least 45%. Discriminating such a modification is further complicated by the fact that watershcd change, which may modify flow, maybe gradual rather than abrupt (p. 329).

Although a 45% change in flood magnitude would be undetectable in this example, it does not mean that such a change would be ecologically and culturally benign.

To address the question of whether humans influence high flows in certain basins often involves a conundrum. If an impact exists, it is costly; if an impact is assumed to exist, but really does not, it can be costly to take unnecessary preventative measures. However, determing the presence of impacts is difficult because variability in runoff makes many hydrologic impacts unpredictable and even undetectable from a statistical standpoint. How society responds to human-caused changes in floods depends on how risks that are difficult to evaluate are perceived and weighed (Chapter 26).

A debate over the beneficial influence of forests in flood protection has continued for at least a centurv in the United States. The arguments being made today are more moderate, but not unlike those made in the early part of the twentieth centurv. Concern about overexploitation of forests and the idea that forest conservation could reduce floods resulted in passage of the Weeks' Law in 1911. The Weeks' Law authorized the federal government to purchase private land to establish National Forests in the eastern United States for the protection of the watersheds of navigable streams. In the early 1900s. Chittenden (1909) stated that forest cutting alone does not result in increased runoff. During the early part of the twentieth century there were many opinions but little data to test the relationship between forests and floods. To address the varied opinions, watershed research was initiated in the 1930s in southern California (San Dimas), Arizona (Sierra Ancha), and south Carolina (Coweeta). The studies at Coweeta resulted in the first scientific evidence that conversion from a forest to a mountain farm greatly increased peak flows, but clear-cutting the forest without disturbing the forest floor did not have a major effect on peak flows (Hoover 1945). By the 1960s, there were 150 forested experimental watersheds throughout the United States. When Lull and Reinhart (1972) released their definitive paper summarizing what was known about the influence of forcsts and floods, about 2,000 papers had been publishcd reporting research results about the hydrology of forested watcrshcds. Lull and Reinhart (1972) focused on the eastern United States. A decade later Hewlett (1982) extended the evaluation to the major forest regions of the world. Hewlett concluded, as did Chittenden (1909) and Lull and Reinhart (1972), that the effect of forest operations on the magnitude of major floods is minor in comparison with the influences of rainfall and basin storage.

Results from the Pacific coastal ecoregion are variable. Rothacher (1971, 1973) found no appreciable increase in peak flows for the largest floods as a result of clear-cutting. Paired watershed studies in the Cascades (Harr et al. 1979), Oregon Coast Range (Harr et a1. 1975), and coastal northwestern California (Ziemer 1981, Wright et al. 1990) similarly suggest that the magnitude of large floods that occurred when the ground was saturated were not increased significantly by logging.

Using longer streamflow records of 34 to 55 years. Jones and Grant (1996) evaluated changes in peak flow from timber harvest and road building from a set of three small basins (0.6-1 km^2) and three pairs of large basins (60-600km^2) in the Oregon Cascades. In the small basins. they reported that changes in small peak flows were greater than changes in large flows. In their category of “large” peaks (recurrence interval greater than 0.4yr), flows were significantly increased in one of the two treated small basins. but the ten largest flows were apparently unaffected by treatment. They also reported that forest harvesting increased peak discharges by as much as 100% in the large basins over the past 50 years (Jones and Grant 1996). However, independent analysis of the same data set used by Jones and Grant indicated that a relationship could not be found between forest harvesting and peak discharge in the large basins (Beschta et al. 1997).

Variation with Basin Size

Effects of forest practices on storm runoff are generally more pronounced and easier to detect in small basins than in large basins. The reasons for this are both statistical and physical. The ability to detect changes in large basins is limited not only by the quality of data available, but also by the sample size of appropriate basins to study in a given hydrologic province. The scale of management units is commensurate with the area of low-order basins of 0.1 to 1 km^2, so at any time there are numerous small watersheds with very high or very low percentages of affected area that can be tested for effects on runoff. In contrast, only a small percent of the area of a large basin (drainage area >$10^2 km^2$) is usually affected at any one time, while the rest of the basin is either pristine or recovering from past effects. Consequently, effective comparisons between treated and nontreated large basins for the same event are unlikely. The best available method to address this problem is to analyze records from small watersheds where the type and timing of land use activities can be controlled and flows can be measured accurately.

Observed effects in small basins cannot be accurately extrapolated to large basins. because processes of flood generation and routing are not represented in the same proportions. Storm peaks originating from small tributaries are lagged. damped, and desynchronized as they move downstream to contribute to flood stages in larger basins (Hewlett 1952). Stormflow response of small basins is governed primarily bv hillslope processes, which are sensitive to forest practices. In contrast, stormflow response of large basins is governed primarily by the geomorphology of the channel network (Robinson et al. 1995), which is less likely to be affected by forest practices. Logging and road building commonly affect stormflow by causing the network of open-channel flow to extend upslope. This extension of the channel network is proportionately small in large basins (Beven and Wood 1993). Increases in peak runoff from hillslopes and headwaters tend to be attenuated by storage in downstream channels and floodplains. Progressing downstream, changes in channel storage of runoff (e.g., from impoundments, channel incision, widespread removal of woody debris, channelization) influence peak runoff in channels more thnn changes in runoff from hillslopes and headwaters.

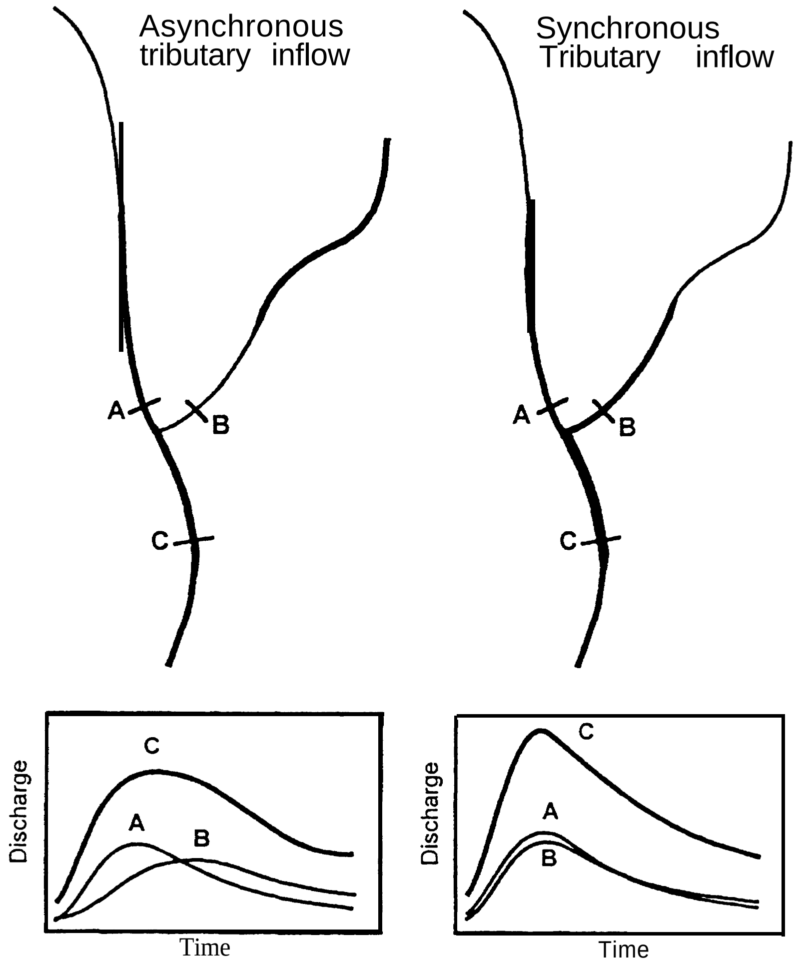

FIGURE 3.7. Effect of svnchronicity of tributary hydrographs on peak discharge in the main stem. With the same volume of total flood runoff, synchronous tributary inputs create greater peak discharge in the mainstem than asynchronous inputs.

Peak discharges in mainstem channels depend in part on the synchronicity of peak discharges from tributaries. Changing the lag time between rainfall and peak runoff from tributaries as a result of hillslope disturbances without influencing flow routing in channels may or may not enhance synchronicity of the tributary inputs (Figure 3.7). Thus, no general relationship between changes in peak flows of tributaries and mainstem channels is expected. Instead, the peak flow response depends on the channel-network hydrology of each basin. However, a consistent shortening of tributary lag times tends to conpress peak-flow arrival times in the mainstem and increase the probability of synchronous inputs.

Variation with Type of Precipitation

Rain on snow. There is little evidence that forest practices significantly affect large floods produced by rain. However, it is possible that clear cutting exacerbates some rain on snow Roods, although the magnitude of such an effect is highly variable and difficult to measure or detect. Snow that is intercepted by a forest canopy in the Pacific coastal ecoregion is apt to melt in the canopy and reach the forest floor as melt water or as wet snow. In areas where trees have been removed, more snow accumulates on the ground. Berris (1984) reported that, in the Oregon Cascades, snow water in a clear-cut was twice that in the forest. In contrast, in colder regions where the snow is drier, wind often sweeps snowfall from large exposed openings resulting in deeper accumulations of snow in protected forests. The convective transfer of latent and sensible heat is often greater in clear cuts because wind speed at the ground is greater. Berris reported that the energy available to melt snow during a rainstorm was about 40% greater in clear-cut areas than in unlogged forests (1984).

Snow accumulation and melt. The demand for water in the west has resulted in assessments of the potential for increasing water yields by delaying deliverv of melt water from the snow zone (Anderson 1963, Kattelmann et al. 1983, Ponce 1983). Management proposals have included various patterns of cutting forests (Anderson 1956), snow fences (Martinelli 1975, Tabler and Sturges 1986), intentional avalanching (Martinelli 1975), application of chemical evaporation suppressants (Slaughter 1970), and weather modification (Kattelmann and Berg 1987). Anderson (1956), for example, designed a “step and wall” forest cutting pattern of alternating cut strips and residual forest in the higher elevation snow zone to maximize snow accumulation in the openings and shade at the forest margin in order to minimize melt rate and thereby provide more water later in the summer season. Although some of these measures have technical merit, serious constraints prevent implementation. Not only do the costs of such measures generally outweigh the value of increased runoff, but also much of the high elevation land is located within National Parks, designated wilderness, or areas administratively reserved from active management. Within those few areas where manipulative land management is possible, concerns about water quality, visual impacts, wildlife, and other resource values preclude serious consideration of water yield improvement projects in the snowpack zone. In the 1950s, grand plans were developed to increase water yields from the mountains in the west (Anderson 1960, 1963) but by the 1980s none of these programs had been implemented.

Variation with Season

Effects of forest practices on streamflow vary strongly with season because of wide variations in hydrologic conditions. In much of the Pacific coastal ecoregion, summers are characterized by long, rainless periods. Since little soil moisture recharge occurs during the growing season in the west, large differences in soil moisture can develop during the summer because of differences in evapotranspiration rates between logged and unlogged watersheds. For example, a single mature pine tree in the northern Sierra Nevada depleted soil moisture to a depth of about 6m and a distance of 12m from the trunk (Ziemer 1968). This one tree transpired about $88 m^3$ more water than a surrounding logged area. This summer transpiration loss is equivalent to about 180mm of rain over the affected area.

With the onset of the rainy season in the fall, the soil profile begins to be recharged with moisture. In Oregon, Rothacher (1971, 1973) reported that the first storms of the fall produced streamflow peaks from a 96-ha clear-cut watershed in the H.J. Andrews Experimental Forest that ranged from 40 to 200% larger than those predicted from the prelogging relationship. In the Alsea watershed near the Oregon coast, Harris (1977) found no significant change in the mean peak flow after clear cutting a 71-ha watershed or patch cutting 25% of an adjacent 303-ha watershed. However, when Harr (1976) added an additional 30 smaller earlv winter runoff events to the data, average fail peak flow increased 122%. In northwestern California. Ziemer (1981) reported that selection cutting and tractor yarding of an 85-year-old second-growth redwood (***Sequoia sempervirens***) and Douglas-fir ***(Pseudotsuga menziesii)*** forest in the 424-ha Caspar Creek watershed increased the first streamflow peaks in the fall about 300% after logging. The effect of logging on peak flow was best explained by a variable representing the percentage of the area logged divided by the sequential storm number that began each fall. These first rains and consequent streamflow in the fall are usually small and geomorphically inconsequential in the Pacific coastal ecoregion. The large peak flows, which tend to modify stream channels and transport most of the sediment, usually occur during mid-winter after the soil moisture deficits have been satisfied in both the logged and unlogged watersheds. These large events were not significantly affected by logging in the H.J. Andrews (Rothacher 1973), Alsea (Harr 1976, Harris 1977), or Caspar Creek (Ziemer 1981) studies.

Variation with Flood Magnitude

On a regional basis, flood magnitude is usually significantly correlated with only a few variables, such as basin size and the amount and intensity of precipitation. For example, the magnitude of the largest rainfall-runoff floods from diverse areas worldwide correlates very well with basin area alone (Costa 1987). In each of five mountainous areas of the United States (Colorado alpine. Colorado foothills, Sierra Nevada, Klamath Mountains, and California Coast Ranges), the magnitude of the mean annual flood in basins with drainage areas greater than $10 km^2$ can be predicted well using only drainage area and mean annual precipitation; variables describing slope, drainage density. and percent forest cover do not significantly improve the relationship (Pitlick 1994).

Although the largest floods are most important from a flood hazard standpoint, the influence of smaller more frequent floods cannot be discounted from a channel condition or ecological standpoint. High flows occurring on average every one to five years are most important for transporting sediment and forming channels in many regions (Wolman and Miller 1960, Andrews 1980). although the less frequent large Hoods can have greater geomorphic effect in the Pacific coastal ecoregion, particularly in mountain channels (Nolan et al. 1987, Grant et al. 1990, Grant and Swanson 1995). Increases in the magnitude of moderate floods tend to increase sediment transport and enlarge channels either by eroding them or building higher banks. However, the response is complex and difficult to detect, because watershed delivery of sediment to channels also occurs during periods of high runoff (Chapters 2 and 11).

Different usage of the term "flood" between the general public and hydrologists can confuse public debate about effects of land use on peak streamflow. To the public, use of the term flood usually evokes the idea of a rare major discrete event that inundates and causes damage to roads, homes, businesses, or agriculture. A "normal" high streamflow event that is expected to occur each year or once every couple of years is usually not considered by laypeople to be a "flood." Human infrastructure is usually constructed to cope with such "normal" events, so property damage from these events seldom occurs. To an hydrologist, the term "flood" loosely refers to a wide range of magnitude of hydrograph peaks, including those that are contained within streambanks as well as extreme events. To avoid confusion, hvdrologists should take care to state the size and frequency of the streamflow event being discussed and should exercise caution in using terminology that can be misinterpreted bv the public.

There is a fundamental problem in determining whether forest practices increase size (i.e., magnitude and extent) of large floods. The problem is greater when attempting to determine whether forest practices increase the size of large floods in large river basins. First. the greater the size of the flood (or basin) being investigated. the less likely that there will be any changes caused by forest practices. Second, any such changes become harder to detect because the available sample size decreases as the size of the flood and the size of the basin increases. To evaluate changes in hydrologic response associated with land use, enough streamflow events must be observed to obtain sufficient statistical power for determining significance. This usually forces the inclusion of smaller events as "floods" to increase the number of observations. Within a 50-year. record, it would be extremely fortunate to measure a 25-year streamflow event before land treatment to compare with a 25-year event after treatment. Even so, there would be little to say statistically about the events because of the small sample size. Only about five 10-year events would be expected during that 50-year record, amd those events probably would be scattered throughout the record, before, during, and after treatment.

There arc physical reasons why forest practices arc less likely to influence large floods than small floods. While logging and road building may affect flow magnitudes by increasing the extent of more rapid surface runoff at the expense of slower subsurface runoff (discussed below), effects on runoff processes vary less with storm size as land becomes saturated (Dunne 1983) and surface runoff caused by human activity (e.g., from roads) becomes a smaller proportion of total stormflow. Moreover, as the duration of a rainfall event increases, any change in the delivery rate of runoff from hillslopes to channels resulting from forest practices becomes less important in flood magnitude.

Water Yield

Throughout much of the arid west, the lack of water during the summer growing season has been a severe constraint on agriculture, power production, urbanization, and virtually all forms of human enterprise. With the establishment of the Wagon Wheel Gap studies in Colorado in the early 1900s. serious scientific thought began to be directed toward evaluating the effect of forest manipulation on water yield. Bosch and Hewlett (1982), summarizing the results of 94 catchment experiments world wide, found extreme variation between areas. but in no case did clearing vegetation reduce water yield. In each case, clearing vegetation resulted in water yields that either remained constant or increased. In cases where water yields increased, the regrowth of vegetation following clearing returned water yields to those observed before clearing. Bosch and Hewlett concluded that the potential for increasing water yield by removing vegetation was greatest in areas having coniferous forests, less in deciduous hardwoods, and least in brush and grasslands. In addition, water yield increases following vegetation removal were greatest in high rainfall areas, and within a given area, tended to be greater in wet years than in dry years (Ponce and Meiman 1983). These small watershcd studies indicate that there is no potential for increasing water yield by manipulating vegetation in areas when precipitation is less than about 40cm, and marginal potential when precipitation is between 40 and 50cm (Clary 1975, Hibbert 1983).

Much of the Pacific coastal ecoregion is within a climate zone where logging might be expected to result in an increase in streamflow during the summer. For example, in a paired watershed study at Caspar Creek in northern coastal California (Figure 3.8) selective logging of 67% of the stand volume in the 484-ha South Fork watershed in the early 1970s increased the summer lowflow about 120% or about 0.3L/s/km^2 (170m^3 of water per dav). This increase in summer flow declined with regrowth of the vegetation and returned to prelogging levels within about eight years (Keppeler and Ziemer 1990). When 12% of the 473-ha North Fork watershed was clear-cut in 1985, summer streamflow increased about 150% for one year before returning to near prelogging levels (Ziemer et al. 1996). Three to five years later, an additional 42% 0f the stand volume was clear cut. Summer lowflow again increased about 200% (0.4 L/s/km’). This increased summer lowflow is anticipated to return to prelogging levels within 8 to 10 years. Similar patterns have been reported elsewhere in which water yield is observed to increase immediately after forest cutting and then return to precutting levels within a few years (Hewlett and Helvey 1970. Ursic 1986, Stednick 1996).

In the 1950s, regional proposals promised to deliver more water in water-deficient regions by clearing vegetation over large areas. Before these proposals could be implemented. not only did more detailed studies show that manv of the earlier assumptions were not generally applicable, but social and environmental concerns about increased erosion, degraded aesthetic values, and habitat destruction associated with vegetation conversion began to emerge. By the mid-1980s, it had become clear that the options for increasing water yield by manipulating vegetation over large areas were quite limited (Ziemer 1987) For western Washington and Oregon, estimated sustained increases in water

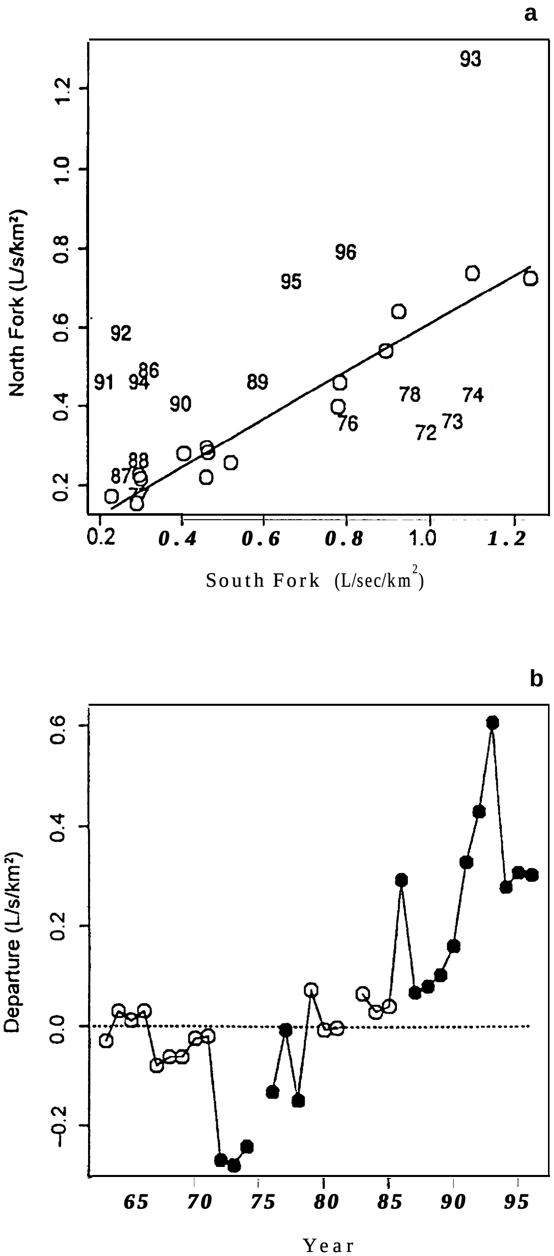

FIGURE 3.5. Relationship between minimum summer streamflow in the North and South Forks of Caspar Creek, 1963-1996. (a) The regression line is based on nondisturbance years (O); the numbers are the disturbance years. (b) Departure is the difference between observed streamflow in nondisturbance (O) and disturbance years (•) and that predicted by the nondisturbance regression (modified from Ziemer et al. 1996).

yield from most large watersheds subjected to sustained-yield forest management are at best only 3 to 6% of unchanged flows (Harr 1983).

Hydrologic Recovery from Land-Use Impacts

The period of time needed for recovery from impacts of land use is an important consideration for resource management (Figure 3.9). Recovery processes vary widely in space and time. In many cases, a return to predisturbance conditions is difficult to define or evaluate because of a lack of data, inherent variability, and an asymptotic decline in effects after a period of rapid recovery.

Recovery of hydrologic conditions following severe land use impacts depends on rates of establishment and growth of vegetation. For example, recovery following grazing is relatively rapid because grassland vegetation is small and grows and spreads rapidly enough to quickly affect runoff processes. Regrowth of vegetation and subsequent loosening of compacted soils in previously grazed areas of the Pacific Northwest can result in increased capacity for infiltration and substantially decreased overland flow velocities within a few years after removal of stock. Further, much wildland grazing occurs at high elevations where frequent freeze-thaw cycles loosen compacted soils. In the Pacific coastal ecoregion regrowth of shrubs and small trees commonly returns rates of evapotranspiration to prelogging levels within about five years (Harr 1979). However, recovery of the tree canopy to levels that restore natural rates of forest snow retention and melt rate takes several decades.

Roads are nearly permanent features on the landscape. Runoff diversion caused by roads cannot be restored by revegetation, but requires erosion or human intervention to reestablish natural drainage patterns. Abandoned roads, skid trails, and landings remain impervious to water infiltration for decades until vegetation slowly becomes established, roots begin to penetrate and break up the compacted soil, and a litter layer and soil profile develops. Road cuts continue to intercept and reroute shallow subsurface flow. Without judicious maintenance, culverts eventually plug, ditches fill, waterbars fail, and the probability of the occurrence of an hydrologic event that exceeds the road's design flows increases with time. Because road maintenance has become a low priority due to reduced funding for many agencies and landowners, risks of hydrologic or erosional impacts from roads will likely increase over time.

Removal of large trees along streams has important consequences for runoff routing in channels. Without replenishment, woody debris disappears from channels over periods of decades or longer, and the loss of channel roughness can increase channel runoff velocities and peak discharges downstream. Furthermore, loss of root strength following the removal of riparian trees may initiate accelerated channel erosion (Chapter 2).

Although water yield changes return to pretreatment levels relatively quickly, changes in the phvsical condition of the watershed that affect streamflow generation and routing may remain for decades. For example, following timber harvesting and site preparation in the Alsea watersheds in coastal Oregon, annual water yield increases returned to pretreatment levels within a few years, but the physical condition of the watersheds is still significantly changed 28 years after treatment (Stednick 1996).

Because land management benefits from a better understanding of the benefits and risks of watershed practices, it remains worthwhile to attempt to measure the magnitude of changes in runoff from land use practices and other influences at a variety of spatial and temporal scales and conditions, even though the outlook to accomplish this with statistical confidence is bleak. Attention to relationships between site conditions and runoff processes can provide answers where changes in river flow are elusive. The motive for investigating changes in runoff is to predict downstream impacts, not necessarily as changes in streamflow alone, but as

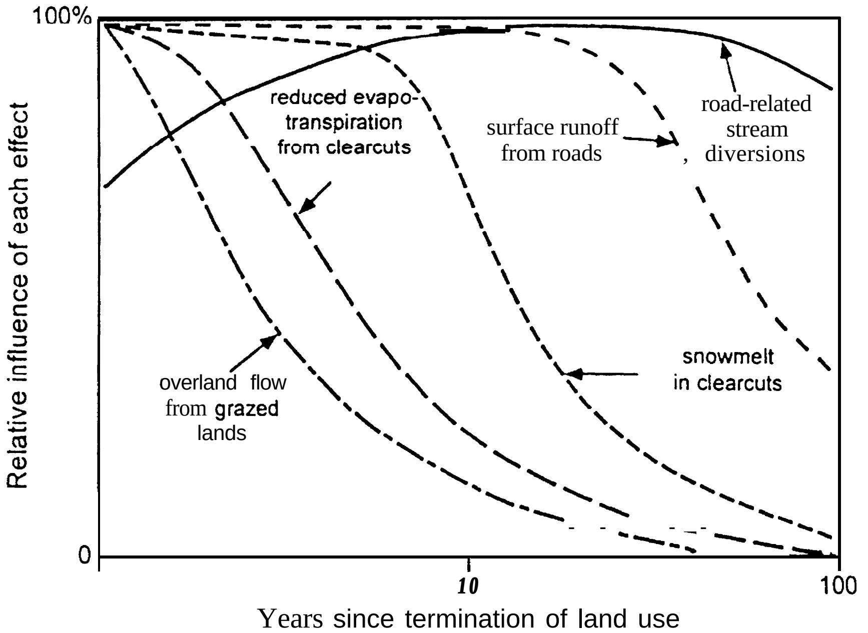

FIGURE 3.9. Characteristic temporal variation of the relative influence of various land uses upon on-site conditions aftcr the land-use activity has terminated. Recovery of channels affected by downstream propagation of these cffects is not reprcsented. The state of recovery at any time is represented as the magnitude of the site effect relative to the maximum, which most cases occurs just after the use is completed or terminated.

changes in the flow of all watershcd products and their subsequent effects on aquatic and riparian ecosystems that often begin as a local disturbance in runoff.

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