

7 Hydrology of Cold-region Montane Forested Watersheds

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Abstract

This chapter explores the snow and hydrological processes characteristic of steep, cold-region montane forested watersheds, particularly in western North America, which rely heavily on snowmelt as a critical water source for ecosystems and human use. While many hydrological principles discussed in other chapters apply to these high-elevation regions, the unique interactions among surface runoff, shallow lateral flow and baseflow merit special attention, particularly in the context of snow processes and steep terrain. The principles outlined here are broadly relevant to forested ecosystems that transition from chaparral or agricultural landscapes at lower elevations to subalpine forested ecosystems at higher elevations.

7.1 Introduction

High-elevation forests are often considered the ‘water towers’ for much of the world. From the Americas to Europe, Africa, Asia and Australia, large urban areas and irrigation projects look to runoff from higher-elevation forests to meet their water supply needs. These areas not only provide surface water to keep streams flowing and lakes and reservoirs filled but also are often important areas for recharging groundwater reservoirs that are later tapped by wells to supply critical water needs. Worldwide, it is estimated that more than a billion people rely on snow-covered areas and glaciers for their water supply (Immerzeel *et al.*, 2020).

The accumulation of snow in high-elevation forests is especially important in arid, semi-arid and Mediterranean climates, as melting snow recharges groundwater and provides low flows throughout the dry season. In continental climates, convective storms can also contribute to snow accumulation in the winter; however, high elevations typically receive more summer precipitation due to orographic effects. This chapter describes how precipitation is routed through forested watersheds as lateral flow and baseflow, rather than surface runoff, which is more common in non-forested or disturbed watersheds.

In most high-elevation forests, the key hydrological processes include winter snow accumulation and melt, higher summer

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precipitation compared with lower elevations, shallow subsurface flow from gradually melting snow in the spring, and continuous baseflow from deep seepage that persists through the dry season. Some of these processes were introduced in Chapters 1 and 2 (this volume). This chapter focuses mainly on hydrological processes in the mountainous and subalpine temperate coniferous forests of western North America, including the states of Idaho, Oregon, Washington, northern California and Montana (west of the Continental Divide) in the USA, and southern British Columbia in Canada. This region is characterized mainly by a maritime climate on the west coast and a continental climate inland. The transition zone between these climates can experience either maritime or continental conditions, depending on prevailing air masses (Hubbart *et al.*, 2007). The main difference between watersheds in these climates is the amount and form of winter precipitation, which not only influences hydrological processes but also has cascading effects on the development of soil and vegetation. For example, higher winter rainfall and peak snow accumulation in maritime climates contribute to deeper soil moisture retention and support more extensive root systems, while drier conditions in continental climates can lead to shallower soils and drought-tolerant vegetation (Zhang *et al.*, 2024).

The maritime climate, found along the Pacific coast, is dominated by warm, moist air masses from the Pacific Ocean. Similar climates are present in countries bordering the Mediterranean Sea and other regions influenced by maritime air masses in winter. Further east, continental convective storms dominate the hydrology, characterized by significant temperature variations, with colder winters and hotter summers. While this chapter primarily addresses hydrological processes in western North America, the principles apply to other steep, snow-affected forested areas worldwide that experience similar climates.

7.2 Snow Processes

Snow processes in forested watersheds occur primarily in the atmosphere and canopy and on the ground. In the atmosphere, snow crystals form

in the clouds and, as they fall, they may either melt into rain, remain as snow or become a mix of both, depending on upper and lower atmospheric temperatures and vapor pressures. Upon reaching the forest canopy, the snow can be intercepted, where it may eventually be released to the ground, melt or sublimate. On the ground, snow can accumulate or melt, with melt rates influenced by energy inputs such as shortwave radiation from the sun, longwave radiation from the atmosphere and/or the canopy, convective heat from warm air or soil, and latent heat from humid air. Sublimation is influenced by wind speed and vapor pressure deficit, and can also occur from the snow under certain conditions, especially when energy inputs are high, the air is dry and/or blowing snow occurs.

Snow accumulation is influenced by the amount of precipitation falling as snow, wind speed and direction, and interception by the forest canopy, which can reduce snowpack by up to 40% due to the combined effects of interception and subsequent sublimation (Varhola *et al.*, 2010). Snowmelt is driven primarily by shortwave radiation; however, forest canopy structure, including tree height, canopy density and tree spacing, affects the degree of shading and thus the amount of incoming shortwave radiation that reaches the forest floor. In dense forests, reduced shortwave radiation may be offset by increases in the longwave radiation from the canopy (Pomeroy *et al.*, 2009). In less dense forests, gaps in the canopy result in highly variable snowmelt rates, creating ‘hotspots’ where total incoming radiation exceeds open areas, or ‘cold holes’ where radiation is less than under dense canopies (Lawler and Link, 2011). Additionally, shading reduces shortwave radiation, and lower albedo from leaves and dust increases the absorption of solar energy by the snowpack. This dual effect, combined with enhanced longwave radiation from trees, contributes to the ‘radiative paradox,’ where net incoming radiation increases with forest canopy cover (Seyednasrollah *et al.*, 2013). Sensible and latent energy fluxes are also affected, with forests reducing wind speed and thereby diminishing the convective transfer of heat and moisture.

Altering the forest canopy can increase streamflow discharge, but the response of snow accumulation and melt to management activities varies among high-elevation forests. These

responses depend on interactions between the canopy, snow interception, snowmelt rates and runoff, which are influenced by elevation, slope, aspect and the dominant air mass (maritime or continental), as discussed by Dickerson-Lange *et al.* (2021). This categorization is further complicated by year-to-year variability, as a location may exhibit characteristics of different regimes or a combination of responses within a single year. Reducing the canopy can also decrease evapotranspiration, potentially increasing water yield under certain climatic conditions. Additionally, snow redistribution by wind plays a critical role in forested landscapes, shaping snowpack distribution and water availability. Disturbances such as wildfire or tree die-offs can amplify wind-driven snow redistribution, further altering snow retention and potentially reducing water yield from affected watersheds.

Building on the understanding of how forest canopy structure influences snow accumulation and melt, recent modeling work by Sun *et al.* (2022) further explored the effects of canopy density on snowpack duration across different climate types, offering insights for forest-management strategies aimed at enhancing snow retention. The authors found that in wet/warm (winter precipitation >300mm and winter temperatures >−1°C) and dry/cold (winter precipitation <300mm and winter temperatures <−1°C) climates, reducing canopy density generally increased snowpack duration, with low-density forests maintaining snow longer than open areas. From a forest-management perspective, thinning the canopy in these climates is more effective than clearing for enhancing snowpack duration and storage. In wet/cold climates (winter precipitation >300mm and winter temperatures <−1°C), snowpack duration is less sensitive to changes in canopy density.

7.2.1 Maritime climates

Maritime climates, with their proximity to large bodies of water, experience milder winters characterized by frequent fluctuations above and below freezing. Abundant precipitation, driven by atmospheric rivers carrying moisture from nearby large bodies of water, often falls as rain or

a mix of rain and snow. These atmospheric rivers deliver substantial snowfall over short durations, especially as moisture-laden air encounters mountain ranges and undergoes orographic lifting, resulting in rapid snowpack buildup. The warm, humid air masses also contain sensible and latent heat, which can quickly melt the snowpack through convective heat transfer and condensation (Marks *et al.*, 1998).

The snowpack in these regions is dynamic, with frequent accumulation and melting throughout winter. The relatively warm temperatures and high ablation rates lead to an early onset of snowmelt, typically between March and April, as higher atmospheric temperatures and increased solar irradiance as the year progresses accelerate the melting process (Trujillo and Molotch, 2014). Additionally, frequent temperature swings near the freezing point make maritime climates particularly susceptible to rain-on-snow (ROS) events. Many ROS events are driven by atmospheric rivers, where warm rains fall on shallow snowpacks, triggering sudden and intense snowmelt. This can lead to significant runoff from saturated soils and increase the risk of flooding, which requires careful monitoring and management.

At the smaller scale of forest canopies, these warmer, moisture-rich conditions further affect how snow is intercepted, retained and released from the canopy. Warmer temperatures during snowfall events can lead to the formation of cohesive snow crystals, potentially enhancing interception efficiency (Lundquist *et al.*, 2021). These conditions, however, also increase the likelihood of snow unloading from the canopy through melting or ROS events. While initially promoting snow interception, warmer temperatures accelerate unloading through melt and drip, reducing the total snow retained in the canopy. Additionally, the removal of canopy cover, such as through forest clearing, increases wind speed within forest clearings, enhancing turbulent energy exchanges at the snow surface and further accelerating snowmelt (Marks *et al.*, 1998). This can result in either shorter or longer durations of snow cover in forested areas compared with open areas. Finally, high humidity, mild temperatures and frequent cloud cover contribute to lower sublimation rates, as the moist air reduces the vapor gradient and cloud cover limits shortwave radiation.

7.2.2 Continental climates

Continental climates, typically located farther inland and away from large bodies of water, experience more extreme temperature variations between winter and summer. Winters in these regions are colder, with extended periods of subfreezing temperatures, leading to prolonged snow accumulation seasons. Snowfall is often steady but less frequent compared with maritime climates, with less influence from atmospheric rivers. Snow processes in continental climates are driven primarily by cold, dry air masses, resulting in lighter, less dense snowpacks. Snow accumulation can persist for months due to the consistently low winter temperatures and reduced ablation rates. In these regions, snowmelt typically occurs later in the season, often starting in late spring or early summer, as temperatures rise gradually. Snowmelt is influenced primarily by shortwave radiation, which increases as the days lengthen, and warmer daytime temperatures lead to a slower, more controlled melt compared with maritime regions. ROS events are less common in continental climates due to the colder temperatures, but when they do occur, they are often associated with unusual warm spells or transitional periods in late winter or early spring. The slower, steady snowmelt and reduced frequency of ROS events generally lower the risk of rapid runoff and flooding, although snowmelt floods can still pose challenges in low-lying areas.

Snow–vegetation interactions in continental forests differ significantly from maritime forests. For example, during snowfall events, lower temperatures produce less cohesive snow crystals, such as needles and columns, which reduce initial interception efficiency. However, colder temperatures also increase branch stiffness, enabling canopies to support larger snow loads (Klamerus-Iwan *et al.*, 2020). This structural support may offset the initial lower interception efficiency, allowing significant snow accumulation in the canopy over time. The dry conditions typical of continental climates promote snow sublimation, reducing the snowpack in both the canopy and on the ground. Wind further enhances sublimation losses, especially in open areas, by increasing

snow unloading and redistributing intercepted snow, often leading to uneven snow distribution.

While topography and the size and amount of forest canopy gaps are important for snow accumulation, the snowmelt is driven primarily by the net available radiant energy, which accounts for a large proportion of the total snowmelt energy. The radiative fluxes for snowmelt are incoming shortwave radiation (direct and diffuse) and longwave radiation (diffuse). In addition, longwave radiation from the forest canopy and trunks is also significant (Lundquist *et al.*, 2013). The proportion of each of these components that reaches the snow surface is dependent on many factors such as elevation, aspect, latitude, day length, cloudiness and solar angle.

7.3 Forests and Avalanches

Landscapes in snowy regions share complex relationships with avalanches that are driven by interconnections of terrain, weather, climate, ecology and land use. Forests may ameliorate or exacerbate avalanche issues, and avalanches may influence forest dynamics.

7.3.1 Snow avalanche anatomy and characteristics

Avalanche paths have a starting zone, track and runout zone (Fig. 7.1). The starting zone is the area of snow accumulation and is where failure usually occurs. In this zone, an internal failure plane with less stable snow on the top and frozen snow, rock or soil beneath the failure plane forms. An increase in downhill forces exceeding resistive forces on this plane and/or lubrication from meltwater at the failure plane will release the top layer, creating a slab of snow cascading downhill (Fig. 7.1, zone 1). The track is a portion of the path where snow is transported from the starting zone, often gaining additional mass (Fig. 7.1, zone 2). The runout zone is where the avalanche decelerates, stops and deposits debris (Fig. 7.1, zone 3). In large, well-defined paths, especially in forested areas, these zones are obvious (Fig. 7.1, path B). In smaller paths or those above the treeline, the delineation of different zones may be subtle (Fig. 7.1, path A).



Fig. 7.1. Typical avalanche paths near Loveland Pass, Colorado. Path A is located entirely above the treeline. Path B intersects the treeline and deposits debris in the valley bottom. Both paths have starting zones (1), tracks (2) and runout zones (3), delineated approximately by black lines in this figure. The transition between the zones and the paths' margins and extents vary with the magnitude and season of individual events. In extreme events, all of path A may become part of the starting zone for path B. In minor events, the starting zone alone of path B may contain a starting zone, track and runout zone, like path A.

Starting zones are often located above the treeline in alpine areas where wind loading on lee slopes enhances snow deposition. These starting zones feed tracks that intersect the treeline and flow through a well-defined path and runout zone without significant tree cover. Paths entirely below the treeline often have clear starting zones, tracks and runout zones defined by forest margins or avalanche gullies developed over time. Snowpack character, terrain and vegetation can affect path morphology and extent, with variations in avalanche paths occurring across different snow climates from maritime to continental and high arctic.

In maritime climates, avalanches tend to be a two-step process. The first step is the accumulation of a snowpack of sufficient depth to

initiate an avalanche. This is generally followed by a rainfall event on top of melting snow, until the snowpack is sufficiently saturated such that, as with the slab avalanche, the internal stresses from the heavy wet snow exceed the resistance to stress from the rainfall-weakened snowpack (Conway and Raymond, 1989). Such avalanches more typically are confined to historical chutes free of large trees, maintained by frequent landslides.

7.3.2 Forest effects on avalanches

Forests affect avalanches through both passive and active processes. Passive processes include

forest influences on snowpack accumulation, distribution and snowpack energy balance. Trees may act as anchors for snow on slopes but may also create weak points where avalanches can start. Forests also significantly affect the local energy balance of the snowpack by diminishing incoming shortwave radiation, increasing longwave radiation and reducing turbulent energy exchanges due to suppressed wind. This energy balance drives snow metamorphism, ultimately controlling snowpack structure and stability, which varies greatly between forested and unforested slopes.

Active processes affected by trees include avalanche track delineation, roughness or friction on moving snow, entrainment of woody material, mass in the avalanche flow, and runout zone dynamics that resist or arrest flow. Trees provide significant friction or flow impediment for avalanches, dissipating kinetic energy and altering avalanche velocity. Small trees may bend or break with little impact, while large trees can completely stop an avalanche flow. Large trees on the margins of runout zones may experience frequent events that are incapable of causing significant damage, but longer-return events may periodically devastate the established boundaries (Schläpky *et al.*, 2014).

7.3.3 Avalanches and hydrology

Snow avalanches effectively move large volumes of snow from one location to another, causing two notable hydrological changes. First, snow is moved from one energy balance regime to another, such as from zone B1 to B3 in Fig. 7.1. Even small elevation changes can result in late-lying snow deposits if the accumulation is significant, as in zone A3 in Fig. 7.1. Snow in high-alpine zones experiences high incoming shortwave radiation and large outgoing longwave radiation losses, with cold temperatures and high winds. In valley bottoms, reduced shortwave radiation and longwave losses occur due to atmospheric transmittance and shading from valley walls, with warmer temperatures and lower winds, causing faster snowmelt in these areas. Second, the structure and distribution of the snowpack are altered. Snowpacks in the starting zone are typically shallow,

low-density and spread over a large area, except in wind deposits. Avalanching transforms this into a high-density, deep snowpack with a smaller surface area.

The hydrological effect of avalanching depends on changes in snowpack properties and the energy balance between the starting zone and deposition area. Snow typically melts faster and earlier at lower elevations, but deep avalanche deposits in mountain valley bottoms can last well into the melt season, often outlasting the surrounding snowpack by weeks or months. Martinec and de Quervain (1971) observed that an avalanche in the Swiss Alps increased early-season streamflow and decreased late-season streamflow compared with modeled non-avalanche results. However, other researchers, such as Sosedov and Seversky (1965), found different outcomes. The effects of avalanching on runoff regimes vary and require local path- or event-specific investigations.

Avalanches change the hydrological pathways of snow that would otherwise have melted in the starting zone. Snowmelt in upper basin areas undergoes hillslope processes like overland flow, infiltration, subsurface flow and possible evapotranspiration (see Section 7.4). In contrast, snowmelt from avalanche deposits in valley bottoms has a short, direct path to the stream channel. The snowmelt from these locations may have different biogeochemical signatures due to variations in pathways, residence time, and exposure to soils, vegetation and geology (Sánchez-Murillo *et al.*, 2015).

Avalanches can also alter runoff by damming streams, leading to impounded flow that is later released catastrophically, potentially causing downstream flooding. This can change downstream vegetation and channel morphology, and impact life and infrastructure. Additionally, avalanches may affect ice-covered lake surfaces, causing a plunger effect that rapidly expels stored water, leading to downstream flooding (Williams *et al.*, 1992).

7.4 Hydrology in Steep Watersheds

High-elevation steep watersheds, including montane and subalpine regions with or without forests, have similar soil water and baseflow

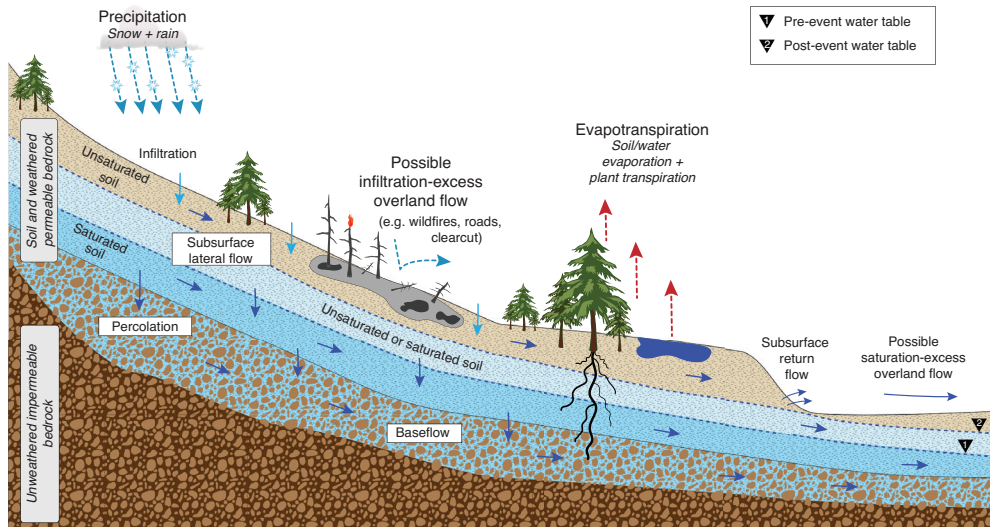


Fig. 7.2. Schematic of hydrological pathways on a steep, concave, forested hillslope.

processes but differ in vegetation impacts on hydrology. Chapter 2 provided a good introduction to forest hydrological processes, and this chapter expands on those principles, focusing on watersheds where snow accumulation and melt dominate the hydrological response. Fig. 7.2 illustrates the dominant flow processes described in this section.

Upland forested watersheds are characterized by steep slopes, shallow soils, the absence of floodplains, high precipitation and low evapotranspiration, unlike lower-elevation watersheds. Consequently, streams respond quickly to storms. Water from rainfall or snow-melt moves through hillslopes into streams via overland flow, subsurface lateral flow and baseflow (Fig. 7.2), although in real-world applications or hydrological modeling, it is difficult to accurately quantify the contributions of each of these components to the streamflow hydrograph (Beven, 2021). As slope steepness increases, so does flow velocity. These pathways respond differently to snowmelt versus rainfall in terms of runoff amount, peak flows and timing of contributions to streamflow. Each pathway is influenced by the interactions among climate, vegetation, topography, soil characteristics and geology. In forest soils, macropores, soil pipes from decayed tree roots, lateral and vertical tree root systems, animal burrows and large cracks

provide permeable conduits for rapid water movement, enhancing subsurface lateral flow processes (Bronstert *et al.*, 2023).

7.4.1 Rain- versus snow-dominated hydrological systems

Both rain and snow are forms of precipitation; however, they affect the hydrological cycle in different ways. Rain increases soil moisture, promotes evaporation and can lead to runoff, impacting river flows and water availability. Snow, in contrast, accumulates on the ground, increasing surface albedo, which reflects more sunlight and can delay melting until warmer seasons. This delayed melting affects streamflow timing and water supply, particularly in regions dependent on snowmelt. The temperature threshold at which precipitation is partitioned between rain and snow is influenced by atmospheric conditions, although it generally varies between -0.4°C and $+2.4^{\circ}\text{C}$. Continental climates tend to have warmer thresholds, while maritime climates have cooler ones (Jennings *et al.*, 2018).

In steep mountainous regions, precipitation increases and temperatures decrease with elevation, affecting the phases of precipitation (rain,

sleet or snow). Higher-elevation watershed processes are driven by snowfall and snowmelt, while moderate elevations are dominated by rain or ROS events. Lower-elevation runoff is mostly constrained by low precipitation. Both snow- and rain-dominated processes show large seasonal variability and interact with vegetation. In snowmelt-dominated regions, runoff usually occurs in late spring or early summer, with significant water percolating into the bedrock (Wilson and Guan, 2004). In rainfall-dominated watersheds, peak flows are linked to seasonal weather patterns and can occur at any time, including summer. In watersheds dominated by more maritime climates, peak flows are typically associated with ROS events driven by atmospheric rivers or seasonal increases in shortwave radiation.

7.4.2 Infiltration and runoff processes

The two primary mechanisms that generate surface runoff are infiltration excess and saturation excess (Fig. 7.2). In steep, undisturbed forested watersheds, infiltration-excess or Hortonian overland flow is rare due to the high infiltration capacity of organic litter and permeable soils, which typically exceed rainfall intensity. Water preferentially moves vertically into subsurface layers instead of over the surface. However, infiltration-excess overland flow can occur when rainfall or snowmelt rates exceed infiltration capacity, particularly on forest roads, logging trails (due to soil compaction), after wildfires (due to reduced infiltration from litter loss and water repellency) and on frozen soils. Infiltration-excess runoff is also common in areas with continental climates experiencing high-intensity thunderstorms or monsoonal rainfall events.

Saturation-excess runoff occurs whenever the soil water-storage capacity is exceeded. This occurs most often in topographically convergent landscape positions in steep, humid ecosystems. The saturation-excess source areas vary in size depending on overall wetness of the watershed, which has led some to refer to these landscapes as dominated by ‘variable source area’ hydrology. The soil becomes saturated due to prolonged wet conditions, snowmelt, high infiltration rates in

shallow soils or its location in convergent areas of the landscape, such as at the base of a hill. At the base of steep hills, the slope often levels out, and the convergence of subsurface flow leads to increased soil moisture that results in saturated conditions and exfiltration of water (Fig. 7.2). Under these saturated conditions, soil cannot absorb additional rainfall or snowmelt, resulting in surface runoff, even if the melt rate or rainfall intensity is below the soil’s saturated hydraulic conductivity. In some cases, the subsurface lateral flow can clearly be seen seeping out of road cuts or at steep breaks in that landscape, a phenomenon known as ‘subsurface return flow’ (Fig. 7.2). These seepage locations are often referred to as spring lines.

Baseflow is generally the return of water that has percolated below the root zone into fissured bedrock beneath forest soils and/or water accumulated on top of the soil–bedrock interface forming a perched water table that drains slowly to the nearest channel or seeps into the groundwater. This water can take from days to years to find its way to forest streams. The rate of return depends on the amount of water that has percolated and the geology (Sánchez-Murillo *et al.*, 2014; Brooks *et al.*, 2016).

In steep forested areas, infiltration-excess overland flow is rare due to the high infiltration capacity of forest soils. Instead, saturation-excess runoff and subsurface return flow are the primary mechanisms of runoff generation, while baseflow contributes to sustained streamflow during non-storm periods. (Brooks *et al.*, 2004; Srivastava *et al.*, 2013). The contribution of each runoff pathway depends on site-specific conditions, such as local climate, soil properties, bedrock permeability, topography and soil water content.

7.4.3 Soil water

In steep forest hydrology, important soil properties include thickness, bedrock permeability, drainable porosity (the difference between soil water content at saturation and field capacity), topography, antecedent soil water conditions and plant rooting depth. The spatial variability of these properties affects soil storage capacity, water availability for evapotranspiration, and

vertical and lateral water movement, thereby defining runoff generation processes. Soil texture interacts with soil thickness to influence hydrological responses on steep forested hillslopes. Coarse-textured soils have higher infiltration rates and hydraulic conductivities but retain less water than finer-textured soils. In steep watersheds, rocky soils retain less water than those with fewer rocks (van Wesemael *et al.*, 2000).

Soil thickness and properties affecting soil water availability are important influences on the relative proportion of subsurface lateral flow, bedrock percolation and vegetation water uptake. Topography influences the variability of soil thickness and texture, with shallower soils common on upland slopes and deeper soils in lowlands (van Wesemael *et al.*, 2000).

Shallower soils, having less water-holding capacity, saturate faster, leading to more subsurface lateral flow, bedrock infiltration and lower evapotranspiration due to limited water for shallow-rooted vegetation. A shallow soil is more likely to become saturated during a wet spell, leading to saturation-excess runoff. In contrast, deeper soils, with higher water-storage capacity, saturate less often and retain more water, resulting in less subsurface lateral flow and greater evapotranspiration for deep-rooted vegetation.

Figure 7.3 shows the comparison of observed and simulated hillslope hydrological responses from two watersheds with different soil thickness located in the Lake Tahoe Basin, Nevada (Brooks *et al.*, 2016). The Blackwood Creek watershed, with shallow soils, showed

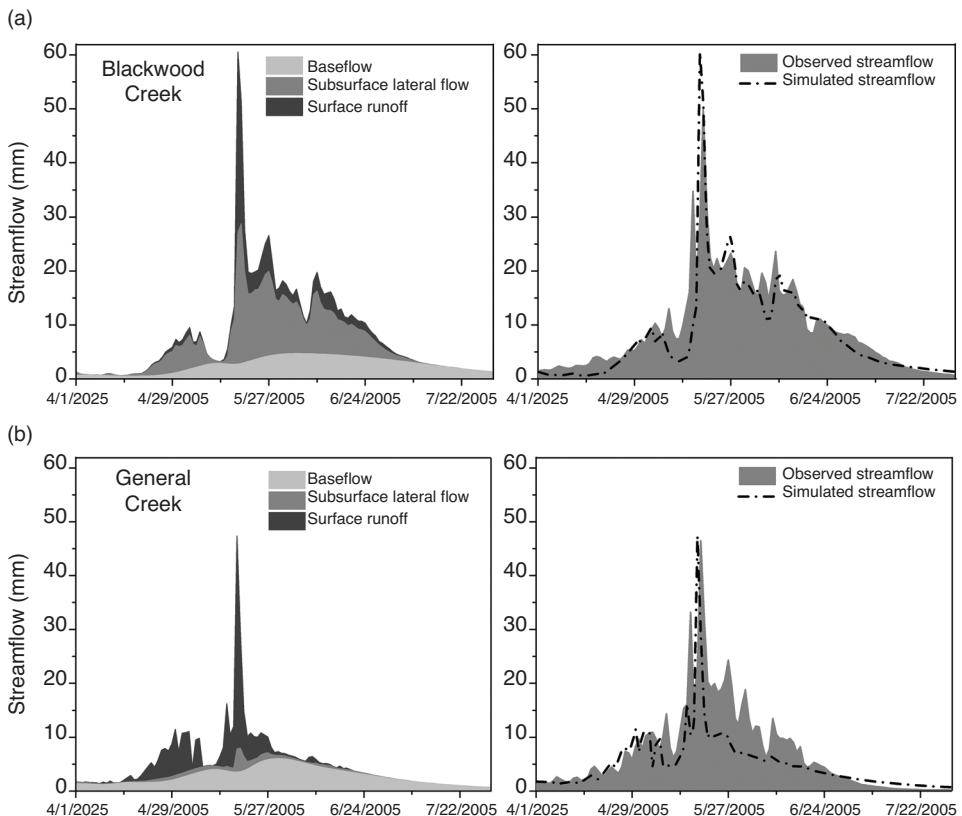


Fig. 7.3. Simulated and observed hydrological responses from two watersheds located in the Lake Tahoe Basin: (a) Blackwood Creek watershed (shallow soils); (b) General Creek watershed (deep soils). Reprinted from Brooks *et al.* (2016), with permission from Elsevier.

major contributions to streamflow from subsurface lateral flow in May and June (Fig. 7.3a), while the General Creek watershed, with deep soils, showed minimal subsurface lateral flow and primarily baseflow contributions (Fig. 7.3b).

Runoff generation mechanisms depend on soil water content at the time of rainfall or snowmelt, known as ‘antecedent soil water.’ In general, higher soil water content leads to a faster and more intense runoff response. Continuous processes such as soil evaporation, plant transpiration, subsurface lateral flow and bedrock percolation gradually reduce soil water content between events. Thus, runoff mechanisms depend on current soil water conditions and the site’s hydrological history. Soil water conditions are influenced by climatic factors (e.g. precipitation, temperature, shortwave radiation) and seasonal weather variability.

Soil water content during winter is often low because the previous growing season’s evapotranspiration depleted it, and temperatures dropped below freezing before the onset of winter precipitation. In spring, rising temperatures cause snowmelt, increasing soil water content and activating runoff pathways (Fig. 7.2), often aided by early-spring ROS or later-spring rain on wet soil. In the summer or dry season, evapotranspiration reduces soil water content, especially in regions with dry summers like the western USA and southern Europe. Soil water content is dynamic, influenced by climatic conditions, soil properties (drainable porosity, thickness, texture), vegetation (species, age, rooting depth), underlying geology (bedrock permeability, soil depth) and topography (van Wesemael *et al.*, 2000).

The above-mentioned complex interactions make it challenging to fully understand high-elevation steep hydrological processes and explain different hydrological responses in similar watersheds. In recent decades, researchers have focused on studying small watersheds in detail, leading to new theories on variability in hydrological responses. These theories emphasize the importance of spatial variability in soil and vegetative properties, microclimates, snowmelt dynamics, and surface and groundwater interactions. Current research in snow-dominated high-elevation mountainous relies heavily on complex computer models

(see Chapter 16, this volume), stable isotopes or machine learning techniques to understand these interactions.

7.4.4 Bedrock permeability

The role of bedrock permeability in controlling water movement can be understood, in a simplistic view, by comparing watersheds with weathered bedrock with those with unweathered bedrock. In areas with weathered bedrock, which is more permeable due to fractures and porosity, water is more likely to move from the soil into the bedrock. As a result, less water flows laterally through the soil because gravity quickly pulls the water down into the permeable weathered bedrock. In this condition, seasonal evapotranspiration may be lower because the soil retains less water for plants to access later in the growing season. In contrast, areas with unweathered bedrock, which is less permeable and more solid, will have less water percolating into the bedrock. This leads to more water remaining in the soil layer, resulting in greater lateral flow through the soil and more water available for plants during the growing season.

However, realistically, as shown in Fig. 7.2, a systematically varying weathering zone that develops within bedrock beneath soils and above unweathered bedrock transforms the poorly conductive bedrock into a dynamic water reservoir. During the wet season, infiltrating precipitation accumulates in the weathered bedrock, where excess water either moves laterally through fractures or percolates deeper to recharge groundwater. The weathered bedrock, often referred to as a source of ‘rock moisture,’ stores water in the unsaturated zone, distinct from soil moisture (Dralle *et al.*, 2018; Rempe and Dietrich, 2018). In the dry season, vegetation gradually depletes rock moisture, sustaining transpiration and evapotranspiration later into the season than would be possible with soil moisture alone.

7.4.5 Topography

The hillslope topography plays a major role in mountainous hydrological processes. Based on

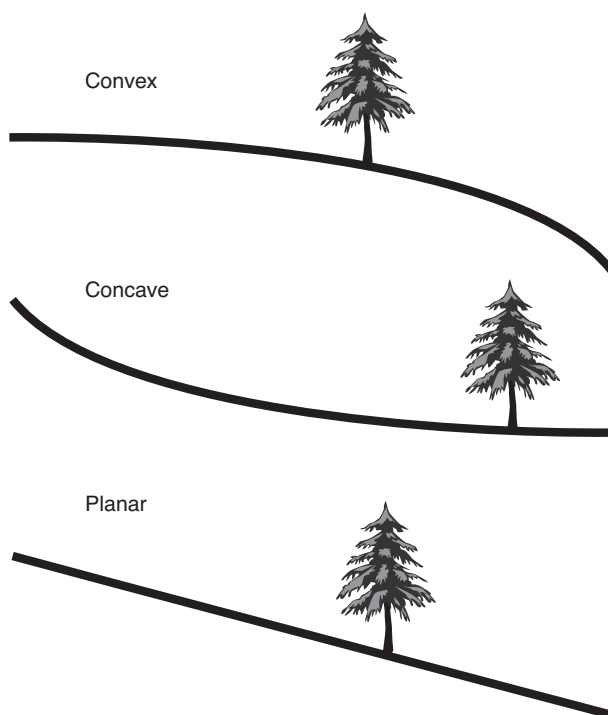


Fig. 7.4. Convex, concave and planar slope shapes.

slope length and the variability in steepness, most of the hillslopes can be characterized as convex, concave or planar (Fig. 7.4). Different sections within a hillslope profile significantly affect soil saturation and flow behavior. Fig. 7.2 depicts an idealized concave hillslope with well-drained soils of uniform thickness and low-permeability bedrock. The slope is steep at the top and gentler at the bottom, illustrating subsurface lateral flow and saturation-excess overland flow or subsurface return flow to the stream. Before a storm, baseflow defines a water table near the stream. During or after a storm, direct precipitation on the valley bottom and subsurface lateral flow elevate the water table and soil water content. For small to moderate events, soil is unlikely to become saturated, and therefore saturation-excess runoff will be rare and subsurface lateral flow will dominate the volume of runoff to streamflow. However, during large events, more of the soil profile saturates, leading to saturation-excess runoff and subsurface return flow as dominant pathways. Soil thickness influences soil saturation near

the valley bottom and runoff pathways. Convex and planar slopes are less likely to experience saturation-excess or subsurface return flows unless geology diverts subsurface flow to the surface.

At a landscape scale, soil depth is not uniform but rather tends to be deeper in upland swales and less deep on upland ridges. Soils tend to be deeper on the lower parts of concave slopes and less deep on shoulder areas of convex slopes. Underlying geology can further affect soil depth, leading to a diversity of potential seepage points as lateral flows are forced to the surface or as surface runoff seeps back into the soil profile (McDonnell *et al.*, 2007).

7.4.6 Impacts of forest access on hydrology

Forest access is associated with nearly all forests to support fuel management, timber harvest, fire suppression and for recreation (Gucinski *et al.*,

2001; Elliot, 2013). Access includes ‘temporary’ skid trails, low-volume forest roads, paved roads, off-road vehicle trails and other trails. All access corridors reduce vegetative cover, increase soil compaction and decrease infiltration rates. The increased runoff may simply infiltrate into nearby forested areas, or may increase watershed runoff, particularly from 2 to 5 year return period storms (Gucinski *et al.*, 2001; DeArmond *et al.*, 2021). The impact of roads on runoff is less noticeable during larger or smaller storms.

Access corridors may become vegetated, but compaction and low hydraulic conductivity can persist for years on lightly compacted skid trails and for decades on more compacted trails and roads (DeArmond *et al.*, 2021). Off-road vehicle trails are particularly problematic due to soil pulverization from traffic followed by high erosion rates, which can expose bedrock and significantly increase runoff potential (Meadows *et al.*, 2008).

In addition to road compaction generating runoff, roads excavated into steep hillsides can intercept subsurface lateral flow, increasing surface runoff and peak flows. This interception may extend stream networks further into steep forested hillsides, further increasing peak flows.

Mitigation practices such as revegetation and mulching may decrease surface erosion but are unlikely to reduce other hydrological impacts. Mechanical ripping of closed roads can increase infiltration from lower-intensity storms. Cross-drains can reduce runoff concentration, downslope channel erosion and sediment delivery, especially if they direct runoff into wetlands or forests instead of streams (Copstead *et al.*, 1998). Paving roads decreases sediment delivery from the road surface, but the increased runoff can cause downslope or downstream erosion. Removing unneeded roads is a common watershed improvement practice.

7.5 Projected Impacts of Climate Change

The climates that are the basis of the knowledge presented in this chapter are not constant. Predictions of future climates around the world suggest a continued increase in temperatures and an increase or decrease in annual

precipitation (Stocker *et al.*, 2013). During the 20th century, air temperatures have increased, decreasing the mountain snowpack, especially at elevations below 1800 m, and threatening water resources (Milly *et al.*, 2008). Higher temperatures are expected to decrease snow accumulation, affect the timing and rate of spring snowmelt (Musselman *et al.*, 2017), and may reduce annual runoff (Berghuijs *et al.*, 2014).

Snowmelt during the snow accumulation season is increasing and is projected to continue rising under warmer conditions, leading to a loss of seasonal snow water storage in North America’s mountainous regions and elsewhere (Musselman *et al.*, 2021; Scaff *et al.*, 2024). However, greater snowfall amounts can mediate these effects. Higher snowfall is associated with reduced melt during the accumulation season, particularly in the continental interior of the western USA, which may mitigate some impacts of warming on snowmelt. Conversely, decreased snowfall in maritime regions will likely amplify the effects of warming on winter melt (Marshall *et al.*, 2020). These findings underscore the importance of accurately capturing changes in precipitation amounts and distributions in climate change projections.

Major changes are expected in the extent and distribution of snow-covered areas and the timing of peak snow water equivalent (Klos *et al.*, 2014; Marshall *et al.*, 2019). While high-elevation mountains will experience temperature-related changes, mid- and low-elevation mountains in the rain-snow transition zone are most sensitive to climate change. In the western USA, a shift from snow to rain by the mid-21st century is estimated to decrease winter snow-dominated areas by 30% and shorten snow cover duration from 5 months to 3 months (Klos *et al.*, 2014). This will likely increase early-spring flows and decrease summer flows, impacting overall hydrology in maritime climates. Watershed managers may need to increase storage within watersheds using reservoirs to mitigate the effects of a shorter snowfall season (Marshall *et al.*, 2019). Additionally, more streams may dry up before the end of the dry season (Rauscher *et al.*, 2008). A decrease in snowpack will also reduce albedo, complicating energy balance analyses in climate change models and potentially leading

to greater temperature increases (Manninen and Stenberg, 2009).

Increasing temperatures and more variable snowmelt will cause vegetation to transpire earlier, leading to earlier soil water depletion (Wieder *et al.*, 2022). In drier forests, this could shift plant communities toward drought-tolerant species (Rehfeldt *et al.*, 2006). Earlier soil water loss will also increase the risk of insect damage, wildfire and changes in plant communities, impacting hydrology (Westerling *et al.*, 2006). This trend is already evident with more frequent and intense fires in both hemispheres. Wildfire and tree mortality often increase peak flows and sediment delivery from degraded and burned watersheds (see Chapter 10, this volume; Elliot, 2013).

7.6 Conclusions

Mountainous and snow-dominated forests are major sources of water on every continent. Snow accumulation and melt rates are big

drivers of hydrological processes. Temperature, radiation, and the interaction between rain and snow are major factors in generating runoff. Surface and subsurface flow processes route water from saturated hillsides to deeper ground-water storage or directly to streams, resulting in hydrographs with long durations and low peak flows. Avalanches can redistribute large amounts of snow on steep landscapes, but the effects on watershed hydrology are site specific and small on a large or regional watershed scale. Flooding is associated with high rainfall and snowmelt rates, and on ROS events occurring when soils are saturated. Future climates will likely result in more flooding from rainfall events and less snow accumulation, especially at lower elevations. The impact that future climate scenarios have on streamflows cannot be generalized but will depend on elevation, source of air masses and complex interactions among the climate, topography, soil and vegetation. Hydrological models will continue to play an important role in better understanding these complex interactions.

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