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Frederick B. Pierson and C. Jason Williams

Abstract

Fire can dramatically influence rangeland hydrology and erosion by altering ecohydrologic relationships. This synthesis presents an ecohydrologic perspective on the effects of fire on rangeland runoff and erosion through a review of scientific literature spanning many decades. The objectives are: (1) to introduce rangeland hydrology and erosion concepts necessary for understanding hydrologic impacts of fire; (2) to describe how climate, vegetation, and soils affect rangeland hydrology and erosion; and (3) to use examples from literature to illustrate how fire interacts with key ecohydrologic relationships. The synthesis is intended to provide a useful reference and conceptual framework for understanding and evaluating impacts of fire on rangeland runoff and erosion.

Keywords: ecohydrology, fire effects, infiltration, rangeland hydrology, runoff, soil erosion, soil water repellency, ecohydrology resilience, hillslope hydrology, hydrologic response, postfire response, soil response

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Acknowledgments

Cover Photo. Prescribed fire at the Reynolds Creek Experimental Watershed, Idaho (photo: USDA Agricultural Research Service, Northwest Watershed Research Center, Boise, Idaho).

CONVERSION FACTORS

Multiply	By	To obtain
degree Celsius (°C)	1.8, and then add 32	degree Fahrenheit (°F)
millimeter (mm)	3.94×10^{-2}	inch (in)
centimeter (cm)	3.94×10^{-1}	inch (in)
meter (m)	3.28	feet (ft)
kilometer (km)	6.21×10^{-1}	mile (mi)
gram (g)	3.53×10^{-2}	ounce (oz)
kilogram (kg)	2.205	pound (lb)
square meter (m^2)	10.76	square foot (ft^2)
square meter (m^2)	2.47×10^{-4}	acre (ac)
hectare (ha)	2.47	acre (ac)
square kilometer (km^2)	3.86×10^{-1}	square mile (mi^2)
square kilometer (km^2)	247	acre (ac)
cubic meter (m^3)	35.3	cubic foot (ft^3)
liter (L)	3.53×10^{-2}	cubic foot (ft^3)
millimeter per hour (mm h^{-1})	3.94×10^{-2}	inch per hour (in h^{-1})
liter per second (L s^{-1})	3.53×10^{-2}	cubic foot per second (ft^3 s^{-1})
cubic meter per second (m^3 s^{-1})	35.3	cubic foot per second (ft^3 s^{-1})
gram per square meter (g m^{-2})	4.46×10^{-3}	ton per acre (t ac^{-1})
kilogram per hectare (kg ha^{-1})	4.46×10^{-4}	ton per acre (t ac^{-1})
megagram (Mg)	1.102	ton (t)

Contents

Introduction

Fire initiates disturbances that alter soil properties and the condition and structure of vegetation and ground cover (Miller et al. 2013), potentially resulting in amplified runoff and soil loss (Shakesby and Doerr 2006; Robichaud et al. 2010; Pierson et al. 2011; Williams et al. 2014b). (Note that in this synthesis, citations are listed in chronological order to reflect the progression of knowledge on a given topic.) The hydrologic and erosional responses to fire vary with the intensity, severity, and spatial scale of the disturbance, the stability and resilience of the affected ecosystem, and the prevailing precipitation regime. Our historical understanding of these responses for rangeland ecosystems comes mostly from anecdotal reports and from short-term, small (0.25 to 1 m^2) field plot studies of hydrologic behavior on gently sloping, semiarid shrub and grassland sites (Pierson et al. 2002a).

In recent years, researchers have expanded the inference space by studying wildfire and prescribed burning effects on runoff and erosion rates from steeply sloped shrublands (Pierson et al. 2001a, 2002b, 2008a,b, 2009), woodlands (Pierson et al. 2013, 2014; Williams et al. 2014a), and shrub-forest interface sites (Johansen et al. 2001). These studies, however, represent a minor portion of the diverse plant communities and soil types that occur on sloping rangelands. Despite these limitations, this review presents fire impacts on rangeland hydrology by (1) explaining fundamental hydrologic processes necessary for understanding hillslope and watershed runoff and erosion responses; (2) describing how climate, vegetation, and landscape properties interact with these principal processes to influence hydrologic and erosional behavior; and (3) providing examples from literature to illustrate these relationships with respect to fire and postfire hydrologic recovery. The hillslope scale is emphasized, given the lack of literature on watershed-scale fire effects for rangelands. Our geographic focus is the western United States, although we include some literature from other semiarid regions around the globe. We include literature from dry forests because of similarities in the postfire runoff and erosion processes across the rangeland-dry forest continuum in the Intermountain West (Williams et al. 2014b). The goal of this review is to provide the reader a background for understanding rangeland hydrology processes and a conceptual framework from which to understand how fire affects rangeland hydrology and erosion across a range of site and vegetative conditions and across precipitation regimes.

Section 1: Hillslope and Watershed Hydrology

The Hydrologic Cycle

The hydrologic cycle (fig. 1) refers to the continuous pathways in which water moves in different phases through the atmosphere; on, into, through, and across the land surface; to oceans and storage reservoirs; and upwards back into the atmosphere (Dunne and Leopold 1978; Branson et al. 1981; Maidment 1993; Hornberger et al. 1998; Dingman 2002; Brutsaert 2005). Knowledge of the hydrologic cycle provides a useful framework from which to conceptualize and understand vegetation-soil-climate-hydrology interactions and to understand how fire and other disturbances influence runoff and erosion behavior (Dunne and Leopold 1978). For a particular hillslope or watershed, the cycle consists of water inflow, transit and storage, and outflow. Inflow primarily occurs as precipitation, overland flow and streamflow from upslope areas, and ground water returns from springs and into streambeds and lakebeds. Transit and storage components of water arriving at the land-atmosphere interface include precipitation interception, infiltration, and water storage on and underneath the land surface. Outflows include gaseous phase losses to the atmosphere through evaporation and transpiration, and liquid water losses to plant use, watershed runoff, and deep drainage to aquifers. The remainder of this section explains the fundamental components of the hydrologic cycle.
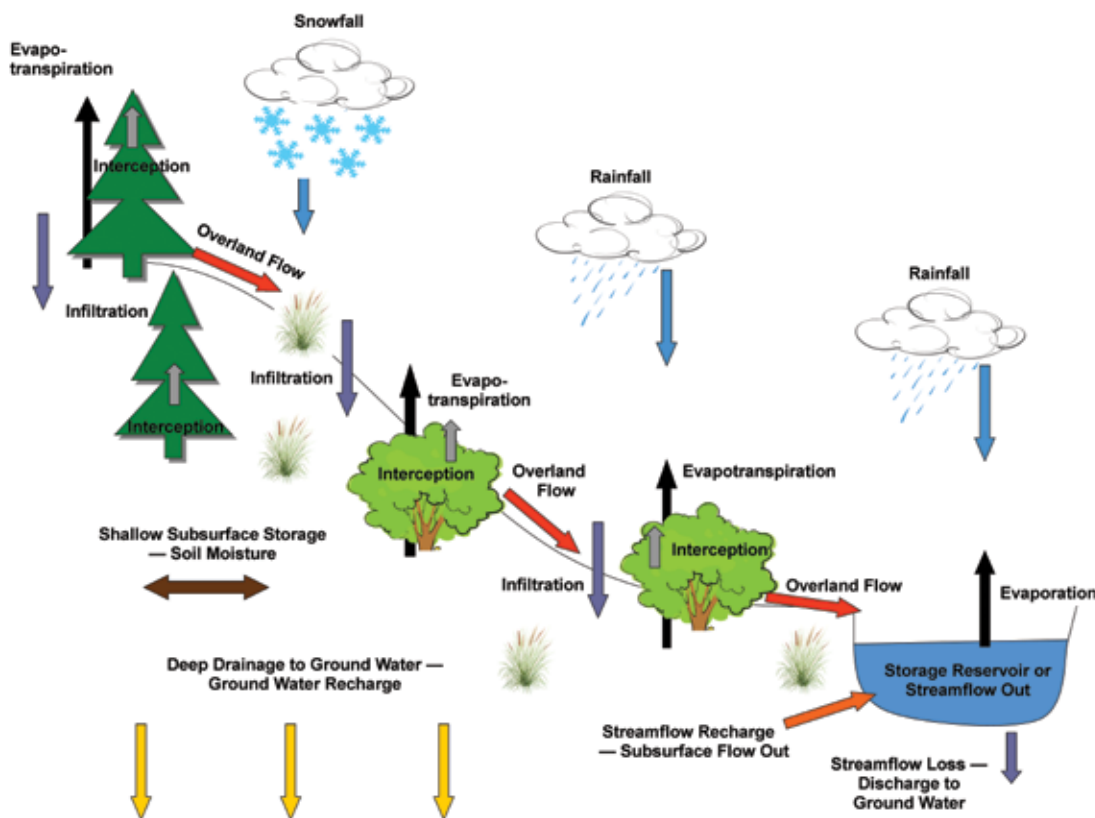
Figure 1—Illustration of generalized hydrologic cycle for a hillslope showing directional inflows (rainfall, snowfall, infiltration) and outflows (evapotranspiration, losses to interception, overland flow, deep drainage, streamflow, and subsurface flow out).

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Precipitation

The primary water input to the land-atmosphere interface is precipitation. Precipitation forms when warm, moist air at the Earth's surface rises into the atmosphere and cools by adiabatic expansion (Branson et al. 1981; Dingman 2002). During cooling, water vapor condenses on small particles of matter, forming water droplets. The droplets remain suspended until gravity overcomes the upward force of the rising air mass, resulting in precipitation. Precipitation falls as snow or ice where the air temperature above the ground surface is 0 °C; otherwise precipitation falls as rain. Uplifting of warm air occurs through either frontal, convective, or orographic lifting, and the lifting process dictates the resulting storm type.

Frontal lifting occurs when warm air and cool air masses collide, forcing the warm air mass upwards and over the cool air mass. Warm fronts advancing toward cool air masses generate prolonged low-intensity (quantity per unit of time), gentle rainfall over large land areas, whereas large cold fronts advancing toward large moist warm air masses facilitate high-intensity rainfall of shorter duration in a narrow advancing band. Occluded fronts occur when a cold air mass overtakes a warm air mass (and collides with another cold air mass), resulting in cold air everywhere at the surface and warm air above. The rising warm air over the cool air at the surface generates precipitation, commonly at an extremely high intensity (Dingman 2002).

Convective lifting is associated with localized heating of surface air. Warm surface air becomes buoyant, rises, and expands due to lower atmospheric pressure. The air mass cools as it expands and convective clouds form. As the air mass cools, moisture particles begin to coalesce therein. Convective lifting is most common during warm moist periods, and generates intense rainfall and hailstorms over small areas scattered across the landscape. However, large-scale convective events may occur and, when over a large enough area, facilitate flash flood events.

Orographic lifting occurs when a warm air mass is forced upward along the windward side of a topographic barrier (such as a mountain range). Precipitation generated from orographic lifting occurs on the windward side of the barrier and usually increases with elevation. Rain shadows form on the leeward side of the barrier as the air parcel crests and warms, and clouds dissipate. Orographic events are often associated with frontal or convective events that encounter a topographic barrier.

Precipitation is measured as a depth over some duration or period of time (daily, monthly, seasonally, or annually) and is reported for individual storms as an intensity. Precipitation data for towns, watersheds, and other locales are available from area-specific climate stations, precipitation gauges, and snow surveys with varying periods of record (years of data). The various types of precipitation gauges, surveys (fig. 2), and methods for extrapolating and applying precipitation data from available sources can be found in most hydrology textbooks (see Dunne and Leopold 1978; Branson et al. 1981; Dingman 2002; McCuen 2004; Brutsaert 2005). Here we focus our discussion on the types of precipitation measurements commonly reported.

Rainfall depth is the quantity of accumulated rainfall (expressed as a length measurement such as mm or cm) at a point on the landscape (such as at a rain gauge). Rainfall duration is the time period over which a specified event occurred. Rainfall or storm intensity is the rainfall rate expressed as depth of accumulation over a specified

Figure 2—(A) Examples of shielded (right) and unshielded (left) dual precipitation gauge system used by the USDA Agricultural Research Service, Northwest Watershed Research Center, at the Reynolds Creek Experimental Watershed, Idaho, and (B) snow water equivalent measurement along a snow course transect of the USDA Natural Resources Conservation Service (photo A: Agricultural Research Service; photo B: Ron Nichols, Natural Resources Conservation Service).

interval of time (for example, mm h^{-1} or cm h^{-1}). Depth and duration variables are often expressed together to define a storm event in terms of a depth-duration relationship (48 mm during a 45-min interval) or an intensity-duration relationship (64 mm h^{-1} for 45 min) and are related to specified recurrence intervals (frequency) or return periods (for example, a 100-year event).

A recurrence interval is an estimate of the interval of time between events of a certain intensity or size. A recurrence interval is not the actual time between events of the specified intensity or size, but rather represents the probability of that event occurring. For example, the 100-year precipitation event has a 1 in 100 chance, or 1 percent probability, of occurring each year. Most climate stations report depth-duration and intensity-duration relationships in graphical or tabulated form for a range of return interval storms (depth-duration-frequency or intensity-duration-frequency; fig. 3). Average annual rainfall is calculated from gauges with long-term records as the total rainfall catch

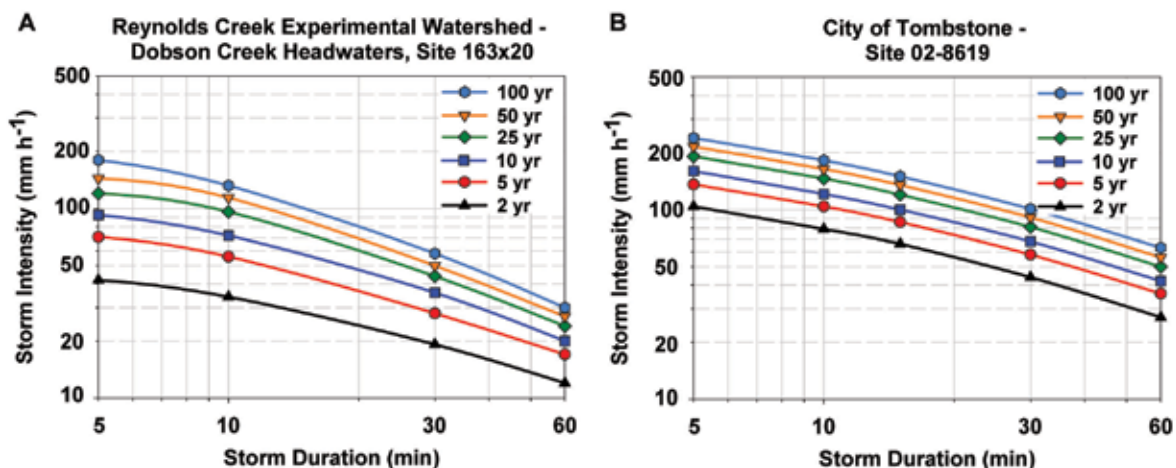**Figure 3**—Intensity-duration-frequency graphs for (A) climate station 163x20 at 2,170 m elevation in the snowfall-dominated Reynolds Creek Experimental Watershed, Idaho, 1963 to 1998 (Hanson and Pierson 2001) and (B) climate station 02-8619 at 1,400 m elevation, Tombstone, Arizona, near the summer monsoon rainfall-dominated Walnut Gulch Experimental Watershed, 1893 to 2000 (Bonnin et al. 2006).

for a number of complete years of record divided by the number of years used. Rainfall for a given year may be reported as a percentage of the average annual rainfall (for example, 120 percent).

Snowfall is most often measured as a storm-specific depth (accumulation) of newly fallen snow at a point, snowpack accumulation at the land surface over some duration (daily, monthly, seasonally, or annually), or the snow water equivalent (depth) of newly fallen snow or the snowpack. In contrast to rainfall, snowfall accumulation stores water at the land surface and releases it for other hydrologic processes more gradually than rainfall. Hydrologists are most interested in the snow water equivalent (SWE) within accumulated snow, as it represents water availability for the hydrologic cycle (fig. 1) (see Dingman 2002).

Snow water equivalent is the depth of water that would result from the complete melting of the snowpack of a specified area and is a function of the snowpack density and depth over the defined area of interest. The quantity and timing of water delivery released from the snowpack depend on SWE and the net energy input into the snowpack (Dingman 2002). Snowpack melting begins after the snowpack temperature is isothermal at 0 °C. Melt water is retained within the snowpack until the water holding capacity is exceeded, initiating delivery of snowmelt. Snowmelt (reported as depth of water) refers to the amount of liquid water leaving the snowpack during a given time period. Other commonly used terms for snow processes include snowpack ablation and water input. Ablation (measured as a depth of water) is the total loss of water substance (snowmelt and evaporation) from the snowpack in a given time period. Water input is the total liquid water (measured as depth of rain and snowmelt) leaving the snowpack during a given time period. As with rainfall, SWE and other snow measurements are available from numerous regional and local climate and precipitation stations. Snowstorms are reported in terms of recurrence intervals, and snow water equivalent or snow depth may be reported as a depth or percentage of average accumulation for various time steps (such as daily, monthly, or annually).

The timing, type, and quantity of precipitation falling at the land-atmosphere interface are driven mostly by elevation and geography. Rangelands in the northern and central United States receive substantially more annual precipitation than rangelands in the desert Southwest (figs. 4 and 5) (see also Branson et al. 1981). In addition to annual quantity differences, the timing or seasonality of precipitation inputs varies significantly across U.S. rangelands (fig. 5). Precipitation inputs on southwestern U.S. rangelands (fig. 5E and 5F) occur mostly during the summer monsoonal season (up to 60 to 70 percent of annual precipitation can occur during July and August) as intense convective thunderstorms (Branson et al. 1981; Osborn 1983a,b; Mendez et al. 2003; Wainwright 2006; Goodrich et al. 2008). Most of the annual precipitation in the northwest and north-central United States falls November through May and is snowfall-dominated in mountain locations and rainfall-dominated in valley locations (figs. 6 and 7) (Branson et al. 1981). Rainfall patterns at these locations usually occur as low-intensity, long-duration events, in late autumn or winter, and during a 4- to 8-week spring rainy season. Rain-on-snow or rain-on-frozen-soil events are common at mountainous locations in early winter and during transitional snow cover periods in spring (Wilcox et al. 1989; Pierson et al. 2001b). Mountain and valley locations experience high-intensity convective storms during the dry summer months. Annual precipitation is usually greater at higher elevations than on
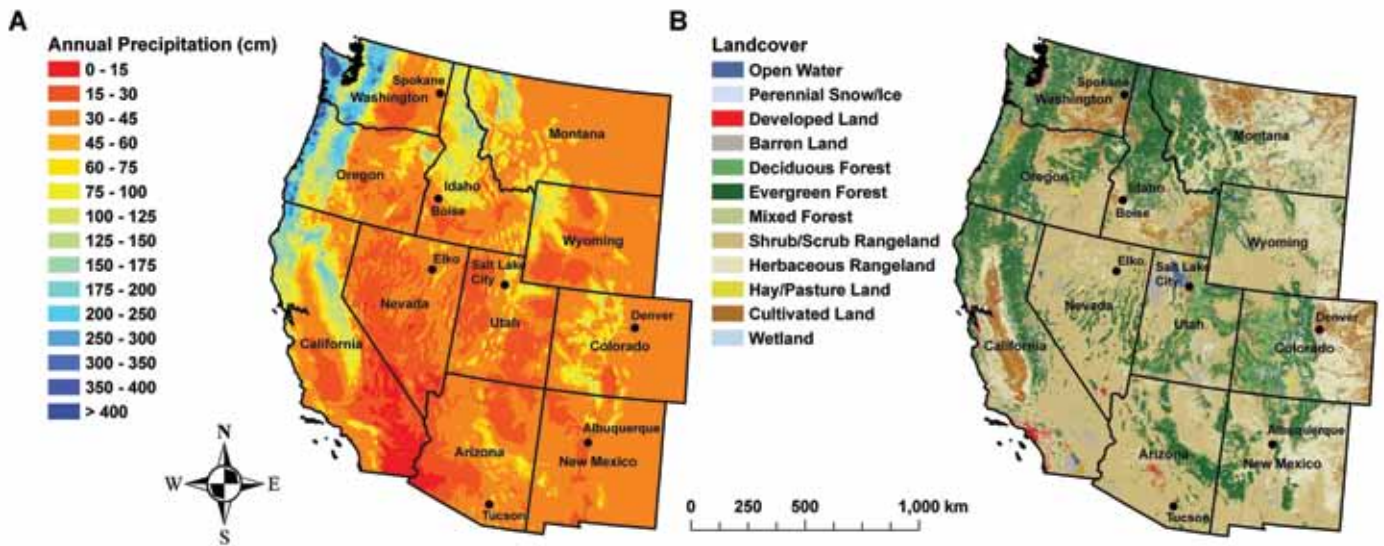
Figure 4—(A) Annual precipitation (PRISM Climate Group 2011) and (B) landcover (U.S. Geological Survey 2011) for the western United States.

valley floors due to adiabatic processes (fig. 6) (see Branson et al. 1981; Dingman 2002). Precipitation trends are bimodal for some south-central U.S. rangelands (see Romme et al. 2009), with peaks occurring in mid-to-late winter and during summer monsoon months (during which 30 to 40 percent of annual precipitation occurs; see Bowen 1996). Elevational precipitation trends for the central United States are similar to those occurring on northwest and north-central U.S. rangelands.
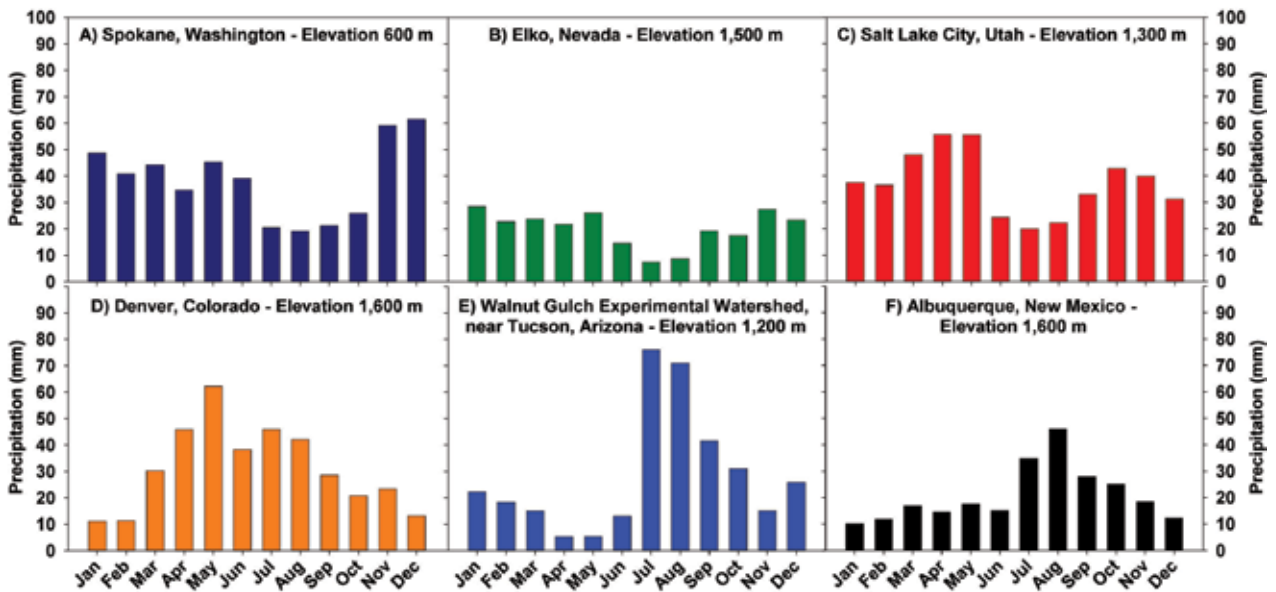
Figure 5—Average monthly precipitation for western U.S. urban centers immediately adjacent to rangelands (PRISM Climate Group 2011). Panels A–C are indicative of annual precipitation in the Inland Northwest. Panel D is indicative of annual precipitation trends on central U.S. rangelands. Panels E-F demonstrate the influence of the monsoonal season (July–August) on rainfall in the desert Southwest.
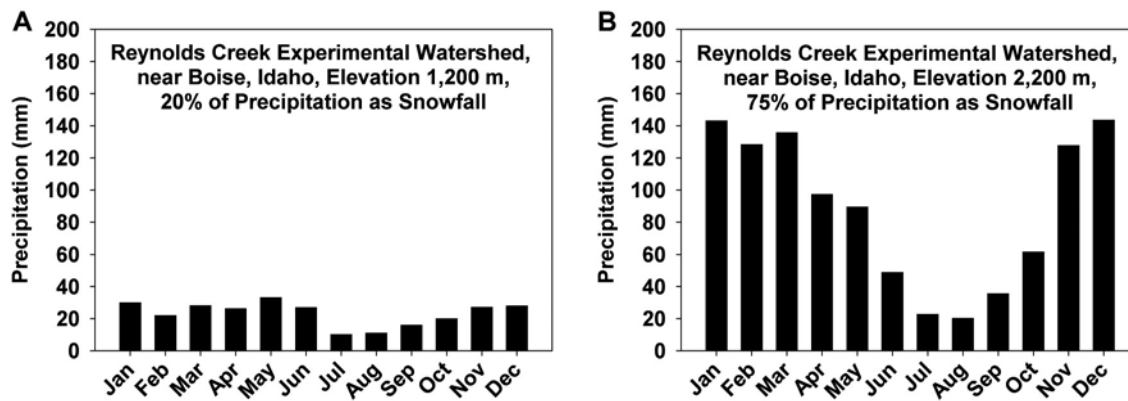
Figure 6—Average monthly precipitation for two climate stations (A – Site 076x59, low elevation; B – 163x20, high elevation) located in the Reynolds Creek Experimental Watershed, Idaho (Hanson and Pierson 2001). Precipitation differences between the two sites are primarily driven by the differences in elevation.

Figure 7—Reynolds Creek Experimental Watershed near Boise, Idaho, where precipitation is snow dominated in the uplands and rainfall dominated in the lowlands (photo: USDA Agricultural Research Service).

Interception

Precipitation arriving at the land-atmosphere interface either is intercepted by vegetation, rocks, or litter debris or falls unimpeded to the soil surface. Intercepted precipitation evaporates into the atmosphere (interception loss) or is transferred as liquid water to the soil surface as throughfall or stemflow (Thurow et al. 1987; Návar and Bryan 1990; Martínez-Meza and Whitford 1996; Whitford et al. 1997; Wainwright et al. 1999; Abrahams et al. 2003; Bhark and Small 2003; Carlyle-Moses 2004; Owens et al. 2006). Throughfall reaches the ground surface by passing directly through spaces

within and between canopies, and includes canopy drip. Canopy drip begins once canopy liquid water interception and surface evaporation capacities are exceeded. Stemflow is precipitation input that reaches the ground surface by running down stems and trunks of vegetation. Gross or bulk precipitation is the precipitation measured in the open, or above the canopy. The net precipitation is the gross precipitation arriving at the land-atmosphere interface minus total interception loss. Intercepted precipitation retained from throughfall and stemflow is termed "static canopy storage" (Dunkerley 2000).

Interception is strongly influenced by the precipitation frequency and intensity and by the type and structure of the vegetative community (Branson et al. 1981; Owens et al. 2006; Dunkerley 2008). Interception losses from a series of small storm and clearing events are proportionately greater than those from large, prolonged rainfall events (Rowe 1948; Hamilton and Rowe 1949; Owens et al. 2006). Prolonged events saturate plant surfaces and the resulting interception rate becomes equal to the evaporation rate. Intercepted precipitation is evaporated during clear periods between short events, reestablishing a portion of the interception capacity. This results in a large proportion of each small storm's gross precipitation being applied to interception as compared to a prolonged or large event. Interception is usually greater from conifer than broad-leaved tree species and is greater from trees species than from shrubs and grasses.

Variations in the canopy density of individual plants and the vegetative community complicate quantification of interception losses over large scales. Therefore interception terms are commonly reported as a depth or volume of water or as the percentage of precipitation falling on an individual plant or at a point during the period of interest. Interception losses over large areas are determined by spatially aggregating interception losses by plant species or area representations (West and Gifford 1976). Dunkerley (2000) provides a brief overview of interception terminology, measurement methods, and estimation approaches.

Infiltration

The rate at which water infiltrates into the soil profile is influenced by the amount and arrival rate of water at the ground surface, the ability of the soil to conduct water into and through the soil profile, and the slope, roughness, and chemical characteristics of the soil surface (Dunne and Leopold 1978; Knapp 1978; Branson et al. 1981; Selby 1993; Hillel 1998; Dingman 2002; Brutsaert 2005). Infiltration is reported on a point scale as a rate (depth per unit of time) or the cumulative depth of water (for example mm or cm) that infiltrates into the soil profile over some period of time (such as a storm event). For rainfall events, water availability is a function of the intensity and duration of the storm (water input rate), interception losses, and the ability of the surface to detain or pond water (surface detention).

Surface detention is a function of the land surface slope, the roughness or microtopography of the soil surface, and the quantity and structure of litter and woody debris present. Gentle slopes with rough surfaces and substantial amounts of litter and debris generally detain more water on the ground surface than steep slopes that are bare or devoid of ground cover. Snowmelt contributions to infiltration are also influenced by the water input rate, interception losses, and surface detention. The snowpack generally

provides a more gradual release of water to the ground surface than does rainfall. However, rapid releases of water from a snowpack may occur during peak snowmelt or rain-on-snow events (Dingman 2002).

Water infiltrates soil mainly due to a negative pressure gradient or suction (matrix suction) into the soil matrix and secondarily due to gravity (Knapp 1978; Branson et al. 1981; Selby 1993; Hillel 1998; Hornberger et al. 1998; Dingman 2002; Brutsaert 2005). Matrix suction results from the physical affinity of water to soil-particle surfaces and pores, and decreases with increasing soil wetness (Hillel 1998). In general, infiltration is high in the early stages of water input into dry soil, then decreases as the surface soil becomes increasingly wet, and approaches a relatively steady state (steady state infiltration rate) as soil becomes saturated. Decreased infiltration over time following rainfall results mainly from decreased matrix suction with wetting, but also may occur due to surface sealing (Assouline 2004) and compaction from raindrop impact or shrink/swell soil properties.

Porous or rough surfaces are usually more conductive than uniform surfaces. However, high-intensity rainfall can break apart highly conductive porous structure or aggregates of surface soils, facilitating infilling of soil pores with fine soil particles (Thornes 1980; Selby 1993). Infilling of pores creates a hydrologic barrier. Compaction reduces pore size and can also create a hydrologic barrier and reduce the surface soil infiltration capacity. Infiltration capacity refers to the maximum rate that water can enter soil in a given condition. Hydrologic barriers may also form from freezing of surface soils or swelling of clay soils. Frozen soils near saturation can form "concrete frost" layers with very low conductivity (Blackburn and Wood 1990; Blackburn et al. 1990).

Swelling properties of clay soils may decrease the conductivity of pore spaces upon wetting, reducing infiltration with increased soil wetness. The presence of organic matter can increase or decrease the infiltration capacity. Organic matter is associated with greater aggregate stability (Cerdà 1998b), low bulk density, and formation of large pores (macropores) or cracks in the soil surface. Aggregate stability and low bulk density values facilitate maintenance of large pore voids and macropores that transfer water rapidly downward into or laterally through the soil profile. In contrast, organic matter may contribute to water-repellent (hydrophobic) conditions, which impede infiltration (Meeuwig 1971; DeBano and Rice 1973; Doerr et al. 2000, 2009). The formation of water-repellent soils and the effects on infiltration are discussed in more detail in Section 3, *Effects of Soil Water Repellency on Runoff Generation*.

Infiltration into wet soils is significantly influenced by the saturated hydraulic conductivity of the soil profile (Knapp 1978; Branson et al. 1981; Selby 1993; Hillel 1998; Dingman 2002). Once the soil profile becomes wet, any further water input is partially dependent on the redistribution or downward transmittance (percolation) of existing soil water. Hydraulic conductivity (measured as length per unit time) refers to the rate at which water is redistributed through the soil profile and is a function of pore space connectivity (soil porosity and soil structure) and soil wetness. Hydraulic conductivity increases with increasing soil wetness due to greater connectivity of wet pores (wet flowpaths). Under saturated conditions, all pore spaces are filled, and higher conductivity occurs where large and continuous pores represent most of the pore volume. For example, sandy, coarse-grained soils with extensive large pore connectivity will transmit water downward more rapidly under saturated conditions than fine-grained clayey soils with

numerous micropores. Therefore, infiltration rates for wet soil conditions approximate saturated hydraulic conductivity and are generally greater for coarse-grained or well-aggregated soils. Saturated hydraulic conductivity commonly decreases with soil depth due to decreases in porosity in deeper portions of soil profiles.

Finally, redistribution and, hence, infiltration are also influenced by the structure of the soil profile. The presence of an impeding/restrictive layer (layer of low hydraulic conductivity) may retard water movement during infiltration. The rate of water movement through restrictive layers is reduced relative to the remainder of the soil profile. Impeded flow is often overcome through bypass (macropore or preferential) flow. The processes and measurement methods for water infiltration and redistribution in unsaturated and saturated soils are described in detail by Hillel (1998) and Dingman (2002).

Soil Water Storage and Ground Water Recharge

Some of the water infiltrating the ground surface is retained in the unsaturated zone as soil water storage, and some passes through the profile into the saturated zone as deep drainage or ground water recharge. The unsaturated zone encompasses the soil water or rooting zone, an intermediate zone, and the capillary fringe immediately overlying the saturated zone (Hornberger et al. 1998; Dingman 2002). The soil water zone is the uppermost portion of the soil profile where soil water is extracted and used by plants or evaporated into the atmosphere. The intermediate zone is often referred to as the zone of aeration and is the area between the soil water zone and the capillary fringe. Water enters the intermediate zone by percolation from above and exits by gravity drainage. Pore spaces in the capillary fringe are saturated or are near saturation and pore water is held there by capillary forces (Hornberger et al. 1998; Dingman 2002). Generally, less than 1 percent of water entering the soil profile passes through the soil water and intermediate zones into deep storage below the capillary fringe. In this review we restrict our discussion on soil water movement and storage to the unsaturated zone because most surface responses are associated with soil water content in the rooting zone of the soil profile. Explanations of the soil physics that dictate soil water movement and retention in the unsaturated and saturated zones are provided in Kramer and Boyer (1995), Hillel (1998), and Dingman (2002).

The water or moisture content of the soil profile dictates water availability for plants and biological processes, and is a function of the soil texture and structure (Kramer and Boyer 1995). Soil water content at any point in time is measured as volumetric or gravimetric water content (see Dingman 2002). Volumetric soil water content is the ratio of water volume to soil volume, and gravimetric soil water content is the mass of water per unit mass of dry soil. Soil water content between precipitation events or periods is referred to as the "antecedent water/moisture content." The distribution, connectivity, and size of the soil voids greatly influence the water content and degree of wetness (the ratio of water content to porosity). Water in capillaries or micropores is tightly held to soil particles by matrix potential, whereas large and connected pores drain rapidly due to gravity (Kramer and Boyer 1995). Excluding evaporative demands, total soil water storage (product of volumetric water content and thickness of the layer) results from the resolution of the matrix and gravitation forces, and plant water use.

Water content available for plants is the difference between the field capacity content and permanent wilting point water content (Kramer and Boyer 1995; Dingman 2002). Field capacity is the water content held against gravity, and refers to the relatively stable soil water content at which continued downward drainage is negligible. The permanent wilting point (Kramer and Boyer 1995; Dingman 2002) is the soil water content at which water availability is too low to support plant transpiration demands and the point at which plants wilt (see Section 1, *Evapotranspiration*). Field capacity and water retention are generally greater for clay soils than for sandy soils and are intermediate for loams. Clay soils are compact and cohesive (with numerous micropores) and drain slowly. In contrast, sandy soils are noncohesive, have large and connective voids, drain rapidly, and possess limited water storage capacity. The presence of organic matter in soils usually increases the water storage capacity.

Evapotranspiration

Evapotranspiration is the primary loss mechanism for precipitation inputs in most geographic areas of the United States and may constitute 50 to almost 100 percent of incoming annual precipitation (see Branson et al. 1981; Kramer and Boyer 1995; Hornberger et al. 1998; Dingman 2002; Brutsaert 2005). Evapotranspiration losses arise from two primary processes of liquid water conversion to water vapor: evaporation and transpiration. Evaporation is the process by which liquid or solid water from ground and vegetative surfaces, from rivers and lakes, and from ice and snow is converted to water vapor and transferred back into the atmosphere. Evaporation occurs when atmospheric vapor pressure is lower than vapor pressure of the evaporative surface. In the most basic sense, evaporation is equal to a coefficient for barometric pressure and wind velocity multiplied by the difference in maximum vapor pressure at the surface and the vapor pressure in the air above the evaporative surface (Dalton's Law) (see Branson et al. 1981 and Dingman 2002).

Transpiration is the direct evaporation of liquid water from within the leaves of plants. During transpiration, water vapor diffuses to the atmosphere from leaf surfaces, forming a water deficit within foliage cells. This deficit is transmitted from foliage via water columns within plant xylem tissue through branches, stems, and large roots to fine roots, where soil water uptake occurs. Transpiration is a biological evaporative process that is influenced by plant leaf, stem, and root structures, water use strategies, soil microclimate, water availability, and plant influences on wind and the aboveground microclimate (Branson et al. 1981; Kramer and Boyer 1995; Hornberger et al. 1998; Dingman 2002).

In order for evapotranspiration to occur there must be (1) a positive net flow of energy to the evaporative or transpiring surface, (2) water available for conversion to water vapor, and (3) a flow of vapor away from the surface (see Hornberger et al. 1998 and Dingman 2002). The conditions that control the net flow of energy determine the energy available for vaporization of liquid or solid water. Energy arrives at a respective surface as incoming solar energy and is either reflected or absorbed, depending on the surface reflectivity (albedo). Light-colored surfaces like freshly fallen snow have much higher albedo than do dark-colored surfaces such as black soil, and reflect much of the arriving

solar energy back into the atmosphere. Some of the arriving energy is consumed to heat (sensible heat) the surface and air around it. The evaporation process consumes additional energy (latent heat) in the conversion of liquid or solid water to water vapor. Therefore the net flow of energy required for evapotranspiration must satisfy energy reflected and energy transferred as sensible and latent heat (Kramer and Boyer 1995; Hornberger et al. 1998; Dingman 2002).

During water-limited conditions, incoming energy is consumed more in the heating of the surface and the air than as latent heat for water vaporization. In contrast, in wet conditions, evapotranspiration is dictated by the quantity of incoming radiant energy, the dryness of the air, and the efficiency of wind to transport water vapor away from the surface. Evapotranspiration may be reported as potential or actual evapotranspiration and is usually provided as a depth of water at a point per period of time (for example, annual evapotranspiration). Potential evapotranspiration is water loss that occurs when a surface is fully wet and no soil water deficiencies exist relative to plant water use. Actual evapotranspiration refers to water loss that occurs when water availability is reduced below that found for a wet surface. Methods to measure and calculate evapotranspiration, along with more detailed explanations of evapotranspiration processes, are found in Kramer and Boyer (1995), Hornberger et al. (1998), and Dingman (2002).

Surface Runoff and Streamflow

The mechanisms for surface runoff generation include Hortonian and saturated overland flow generation, direct precipitation into stream channels, and ground water returns to the land surface or stream channels (Horton 1933; Dunne 1978; Selby 1993; Hornberger et al. 1998; Dingman 2002). Hortonian overland flow (Horton 1933) is generated when water input on land exceeds the rate at which water can infiltrate the soil. Hortonian flow (also called infiltration-excess flow) is most common under intense rainfall in sloping semiarid to arid regions (water-limited) or where surface conductivities are low. Saturation overland flow (saturation-excess flow) results from continued water input at the surface of a saturated soil profile. Ponded and saturated soil surfaces shed any additional water inputs from the atmosphere. The hillslope or watershed area (source area) contributing to saturated overland flow varies seasonally or during precipitation events. Hydrologists commonly refer to this variable zone of saturation overland flow as the "variable source area" (see Selby 1993; Hornberger et al. 1998; Dingman 2002).

Infiltration- and saturation-excess overland flows may create sheetflow (interrill) or concentrated flow (rills), or a combination thereof, on sloping terrain. Sheetflow refers to overland flow as a thin, relatively spatially connected film or sheet on the land surface. Concentrated flow is runoff that accumulates or converges into well-defined microchannels or rills. Direct precipitation into stream channels occurs during all precipitation events and can contribute significantly to peak and total event streamflows. Subsurface return or event flow constitutes only a small portion of event streamflow. However, areas of ground water mounding or ridging and hillslopes with extensive macropore networks may contribute substantially to runoff generation and streamflow response (Wilcox et al. 1997; see Dingman 2002).
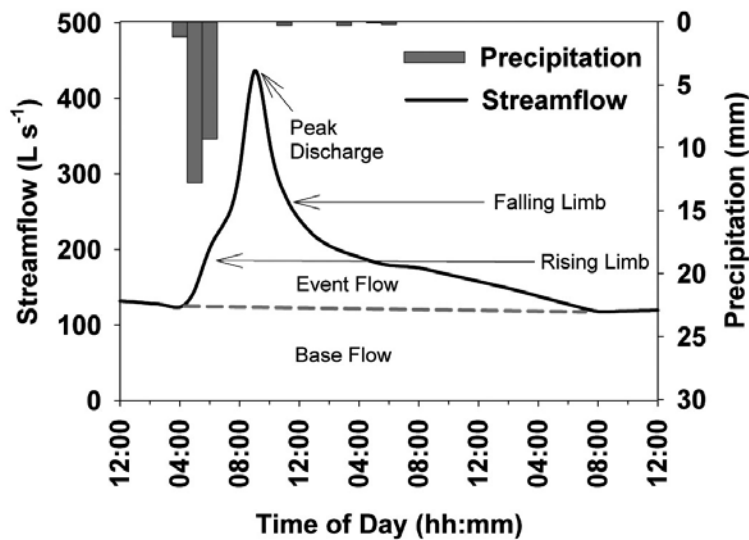
Figure 8—Example streamflow hydrograph and precipitation hyetograph. Precipitation and streamflow shown are from a convective thunderstorm event (~22 mm rainfall during 2 hours) occurring July 21, 1971, in the Reynolds Creek Experimental Watershed as recorded at the Tollgate Weir Site 116x83 (USDA Agricultural Research Service 2011a). Event and base flow contributions to streamflow are separated by the dashed gray line. Arrows and callouts indicate rising limb, peak discharge, and falling limb of the hydrograph as responses to the storm event.

Streamflow is the flow rate or discharge of water (measured as volume per unit of time) along a defined natural channel and is partitioned as either base or event flow (Dunne and Leopold 1978; Dingman 2002; McCuen 2004). Base flow refers to the portion of the streamflow that cannot be attributed to a particular precipitation event and is generally assumed to be ground water return flow into stream channels. Base flow of a particular stream or drainage network (pattern of streams within a watershed) is relatively consistent from year to year and depends mostly on the availability of ground water returns. Stream baseflow may be spatially variable where exchanges of surface water and ground water facilitate streamflow gains or losses. Event flow is the portion of streamflow directly resulting from event effective water input, and may also be referred to as storm runoff or storm flow. Effective water input is the water input from a particular precipitation event, usually in the form of direct precipitation into streams and Hortonian or saturated overland flow. Time variability and space variability in event flow generally increase with increasing watershed size due to temporal and spatial variations in precipitation water input, overland flow generation, and streamflow routing.

Watershed or hillslope response to precipitation events is commonly quantified graphically and analyzed by using streamflow hydrographs and precipitation hyetographs (fig. 8) (Hornberger et al. 1998; Dingman 2002; McCuen 2004). Hydrographs depict stream discharge versus time for a point along a stream channel. Hyetographs quantify water input (precipitation or snowmelt) at a point versus time and may be shown on a secondary x- and y-axis of a streamflow hydrograph to view respective stream responses to water-input events (fig. 8). The shape of the streamflow hydrograph provides qualitative and quantitative interpretation of watershed response to varying water input. The event response is evident on the hydrograph by an increase in discharge (rising limb) greater than base flow, to a peak (peak discharge), followed by a decrease to baseflow (falling or recession limb) (fig. 8). Hydrologists are particularly interested in watershed response time, time to peak runoff, peak discharge, and cumulative runoff. The response time and

time to peak refer to the time it takes for runoff to occur and then to peak following water input initiation. Peak discharge is the maximum discharge that occurs during a particular event. Cumulative runoff is the integration of runoff rates with respect to runoff duration. A short response time, a short time to peak runoff, and a steep rising limb indicate rapid or flashy watershed response to a water input event.

The overall response to a particular event, and the resultant hydrograph shape, depends on the size of the drainage area, soils and geology, slope, and land use/vegetation patterns (Dingman 2002; McCuen 2004). Large watersheds commonly exhibit delayed responses to precipitation events unless the event is near the stream outlet or occurs over a large, contiguous area. Watersheds or hillslopes with low infiltration rates or steep slopes, or both, typically have short response times, high peak discharge, and steep rising limbs. Runoff response is delayed and discharge is usually low where extensive vegetation and ground cover exist and land use favors water retention. Shorter event and peak response times, higher peak discharges, and steepened rising limbs with respect to similar rainfall events postdisturbance indicate degraded surface conditions.

An individual hydrograph does not specifically identify the condition eliciting the response. However, hydrographs can be compared for similar precipitation events over a range of watershed conditions to infer cause-and-effect relationships where supportive watershed/hillslope data are available. Infiltration responses to water-input events may also be quantified in hydrograph form, as the inverse of the runoff relationships (see Meeuwig 1971; Dunne 1978; Selby 1993; Hillel 1998; Hornberger et al. 1998; Dingman 2002).

Water Balance

The components of the water cycle with respect to a specified watershed and time period represent the area water balance (see Branson et al. 1981; Dingman 2002; Wilcox et al. 2003b). The generalized water balance is expressed as:

$$P + G_{in} - Q - ET - G_{out} = \Delta S \qquad \text{(Equation 1)}$$

where P is precipitation, G_{in} is incoming ground water, Q is streamflow or surface runoff, ET is evapotranspiration, G_{out} is outgoing ground water, and ΔS is the change in water storage over the period of interest. Over annual scales, the water balance provides an accounting of the annual water budget. For rangeland ecosystems, runoff usually amounts to less than 10 percent of the annual water budget (Wilcox et al. 2003b). Nearly all of the remainder of precipitation falling on rangelands is lost to evapotranspiration. Evapotranspiration may exceed precipitation during dry years. Deep drainage of soil water beyond the rooting zone as ground water recharge, G_{out}, is usually less than a few millimeters (Wilcox et al. 2003b). The net change in water storage is generally considered to be zero over long time periods (such as several years). Section 3 provides more detail on the water budget components with respect to rangeland ecosystems.

Section 2: Sediment Detachment, Transport, and Mass Wasting

Soil erosion refers to the detachment and transport of soil by raindrop impact, running water, wind, ice flows, and other mass-movement processes (Selby 1993). Soil detachment is a function of the erosive energy acting on the soil surface and the resistance of the surface to erosion (Thornes 1980; Selby 1993; Toy et al. 2002). In hydrologic terms, erosive energy acting on the soil surface (shear stress) results from raindrop impact or the flow of water, or a combination of both. The resistance of the surface (critical shear stress) is a function of the soil properties, soil cohesiveness, and other surface characteristics that define the soil erodibility. Erodibility is defined as the vulnerability of a soil in its current condition to erosion by rainfall, runoff, and wind. Detachment occurs when the shear stress acting on a soil exceeds the critical shear stress or resistance of the soil (Foster and Meyer 1972; Nearing et al. 1999; Kinnell 2005).

As defined, the removal of detached sediment requires entrainment by a transport mechanism. Transport may occur by displacement from raindrop impact or entrainment into flowing water, or both (Kinnell 1988, 1990, 2005). Transport capacity of flowing water is dependent on the volume of water, mass of solids versus mass of water, energy loss as the flow moves downslope, and efficiency of transport (Thornes 1980; Toy et al. 2002; Kinnell 2005). In this section we summarize these fundamental sediment detachment and transport mechanisms. Our emphasis is on hillslope processes given the context of this review. Julien (1998), Knighton (1998), and McCuen (2004) provide detailed explanations of channel or fluvial sediment entrainment, transport, and routing processes beyond the scope of this review. Soil erosion through wind, creep, weathering, and other mechanisms not directly involving rainfall and flowing water are discussed in Selby (1993), Toy et al. (2002), Ravi et al. (2007), Sankey et al. (2009), and Field et al. (2012).

Rainsplash and Sheetflow Processes

The combined effects of rainsplash and sheetflow/sheetwash processes are termed "rainwash" or "interrill" processes (Selby 1993; Toy et al. 2002). Rainwash processes often represent the dominant erosion processes where microchannel formation is substantially limited by soil aggregation, soil cohesion, and surface protection. Here, we describe the two components of rainwash—rainsplash and sheetflow—separately because their co-occurrence is dependent on overland flow generation. Rainsplash erosion is the transfer of sediment resulting from raindrop impact. The effects of raindrop impact include detachment and displacement of soil (see Kinnell 2005), disaggregation of soil aggregates, and reduced infiltration due to surface sealing (Moss 1991; Assouline 2004; Kinnell 2005). Sediment detachment and transport by rainsplash occurs in a "splash-crown" (fig. 9A) with sediment mass declining exponentially outward from the point of impact on flat surfaces (Thornes 1980; Toy et al. 2002; Kinnell 2005). On sloping terrain, downslope transport may exceed three times the mass of upslope transport where slopes are greater than 10 percent (Thornes 1980).

Raindrop detachment rates are commonly highest within several minutes after the onset of rainfall, followed by an exponential decrease to a steady rate; however, rates
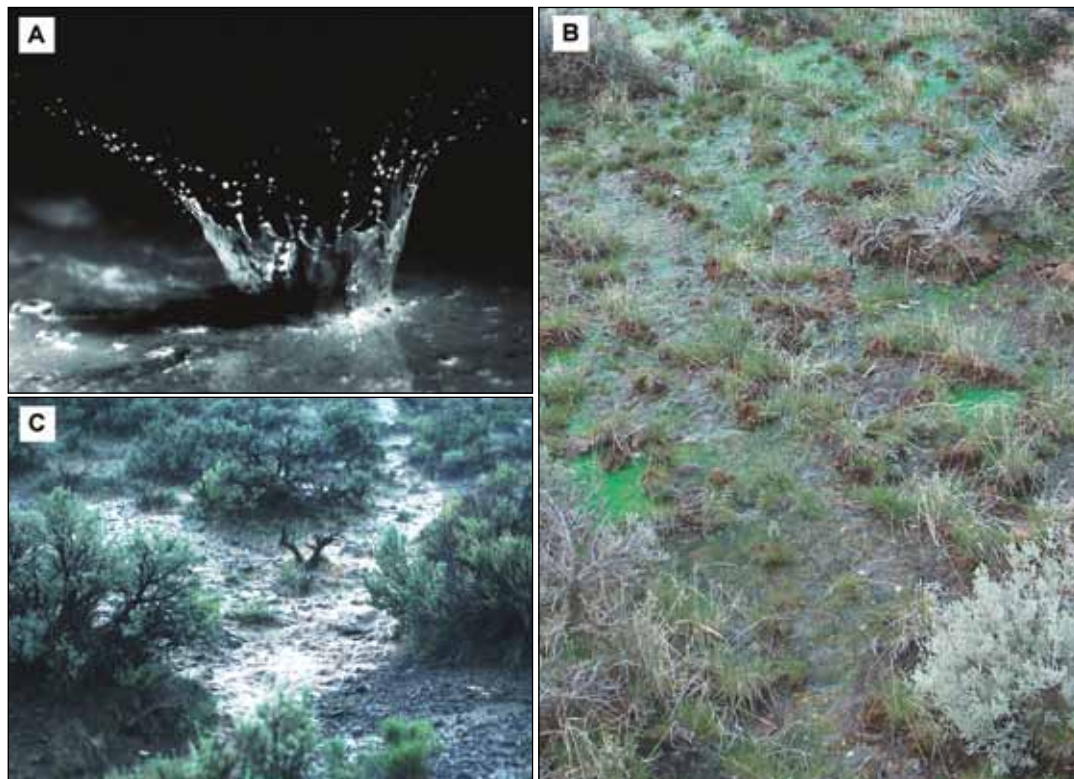
Figure 9—Soil erosion by (A) rainsplash, (B) sheetflow, and (C) concentrated flow processes (photo A: USDA Natural Resources Conservation Service; photos B and C: USDA Agricultural Research Service, Northwest Watershed Research Center).

may be highest at rainfall onset if the supply of detachable sediment is low (see Parsons et al. 1994). The main source of energy in this process is the kinetic energy of rainfall as dictated by the raindrop mass and terminal velocity (Gilley and Finkner 1985; Sharma et al. 1991; Bryan 2000; Salles and Poesen 2000; Salles et al. 2002). Terminal velocity increases with increasing drop size, and drop size generally increases with increasing rainfall intensity (Gunn and Kinzer 1949; Salles and Poesen 2000; Salles et al. 2002; Van Dijk et al. 2002). Thus, the impact energy of rainfall and sediment detachment generally increases with increasing rainfall intensity. However, cumulative sediment detachment may be significant from low-intensity storms over long durations.

The decrease in detachment by raindrops during rain events is often associated with an increase in the depth of sheetflow (Thornes 1980; Moss and Green 1983; Ferrera and Singer 1985; Gilley et al. 1985; Toy et al. 2002). Flow depths equal to the diameter of approximately three drops greatly reduce the impact of rainfall (Moss and Green 1983; Kinnell 1990, 1991, 1993). The shear stress and additional transport capacity of sheetflow may, however, offset the decrease in raindrop impact, resulting in a net increase in total sediment production, or yield (Kinnell 2005). In addition to increasing sediment yield, drop impact from high-intensity rainfall may break apart soil aggregates and facilitate particle sorting by size and infilling of surface pores with fine material. This process, referred to as surface sealing (Bradford et al. 1987; Assouline 2004), may create a surface crust or compact the surface soil, resulting in reduced infiltration (by a factor of 1 to 10) and increased runoff and sediment transport by sheetflow. Soils with high clay and organic matter content are generally less influenced by raindrop effects than are sandy or sandy loam soils, and the net effect may be related to soil bulk density.

Sheetflow generates shear stress that is minor (about 100-fold less) compared to raindrop impact, but it serves as an additive detachment and transport mechanism for interrill sediment yield (Thornes 1980; Selby 1993; Toy et al. 2002; Kinnell 2005). Sheetflow does not usually occur as a broad flow across a hillslope; rather, it occurs in isolated irregular- flow patches of several millimeters depth separated by flow obstacles (fig. 9B) (Emmett 1970, 1978). Raindrops falling into shallow flow depths create turbulence and detachment, facilitating sediment entrainment. Soil detachment from sheetflow results from drag created by differential shear stress on the upslope and downslope faces of the particle, Bernoulli lift in the horizontal direction, and vertical turbulence (see Thornes 1980; Kinnell 2005). The amount of sediment and the size of particles entrained depend on the flow velocity and turbulence, both of which generally increase with increasing slope and flow depth. The rate of transport generally increases with flow depth to a maximum of about one to three raindrop diameters (Selby 1993). Entrained particles remain in suspension until a deposition velocity occurs (less than 0.015 m s^{-1}). The presence of ice in surface soils may amplify interrill erosion by slightly raising portions of the surface soils, making them more susceptible to detachment and entrainment by rainsplash and sheetflow processes (Blackburn and Wood 1990; Blackburn et al. 1990).

Concentrated Flow Processes

Sediment yield from concentrated flow (rill) processes is several orders of magnitude greater than that of sheetflow and rainsplash (Thornes 1980; Wainwright et al. 2000; Pierson et al. 2008a). Concentrated flow processes may account for 50 to 90 percent of total sediment yield on slopes with very little vegetation. Concentrated flowpaths or rills are microchannels of several to tens of centimeters in width and several to 300 millimeters in depth that are easily obliterated between storm events. These microchannels form when surface roughness elements (microtopography) concentrate sheetflow into narrow, deeper flowpaths, increasing the velocity and erosive energy of runoff (fig. 9C) (Emmett 1970, 1978). Concentrated flow detachment and incision occur when the incoming interrill sediment load is less than the concentrated flow's transport capacity, and the shear stress applied to the soil is greater than soil surface critical shear stress or erodibility (Nearing et al. 1989, 1999; Al-Hamdan et al. 2012b). The shear stress applied is a function of the density, depth, and velocity of the flowing water, the friction imposed by the soil and cover, land surface slope, and acceleration due to gravity (see Toy et al. 2002; Al-Hamdan et al. 2012a,b, 2013).

Deposition occurs when the flow transport capacity is exceeded. Sediment detachment and transport are commonly highest at the initiation of concentrated flow and decrease gradually as more resistive materials are exposed with flowpath incision or where sediment supply is limited (Nearing et al. 1997; Wainwright et al. 2000; Pierson et al. 2008a; Al-Hamdan et al. 2012b). On freshly exposed surfaces, parallel concentrated flowpaths may cross grade or merge by breaking down the divides between microchannels (micropiracy). This process diverts the flow into the deeper, more dominant flowpaths and generally increases the spacing of concentrated flowpaths in the downslope direction.

Gully Erosion Processes

Gullies are recently channelized drainage features that transmit ephemeral flow, usually have steep sides and a head scarp (leading upslope area of exposed soil and rock), and are more than 30 cm wide and 60 cm deep (see Selby 1993; Toy et al. 2002). These erosional features commonly form when a master rill deepens and widens its channel, especially where changes in slope or vegetation patterns occur on unconsolidated materials (fig. 10) (Neary et al. 2012). Gullies may also form where debris and mud flows exit unstable drainage basins or where large subsurface drainage features collapse. The most common cause of gully formation is a loss in surface protection associated with a change in the overlying vegetation or soil disturbance.

Gullies forming from rills often have no head scarp, increase in width and depth downslope toward a master gully, and end in a deposition zone of coalescing fans at the base of toe slopes. Gullies with head scarps maintain the scarp where soils are cohesive, but upslope or upstream headcutting occurs if soils are weak (unconsolidated). Peak discharges from gullies typically far exceed the peak discharge of the previously unchanneled valleys in which they occur. Erosion from gully processes may be severe where high-intensity rainfall events occur over poorly vegetated surfaces with weakly consolidated or unconsolidated sediments. Gully erosion most commonly occurs in pulses, and sediment supply comes mostly from head scarp erosion and bank failures or sidewall sloughing (Selby 1993).

Figure 10—Gully erosion on an unprotected soil (photo: Lynn Betts, USDA Natural Resources Conservation Service).

Mass Movement Processes

Mass movement erosion occurs when the shear stress applied to a body of soil material on a slope exceeds the resistance or critical shear stress (shear strength) of the material to downslope movement and may result directly or indirectly from a particular water input event (see Sidle et al. 1985; Selby 1993). Shear stress is increased by removal of lateral soil support, soil profile shifting, overburdening of soils with rain or snow, ground vibrations, undercutting of banks or ridges, and increased slope steepness. Shear strength is reduced by loosening of soils with soil shifting, increased buoyancy and capillary tension associated with pore water changes, alteration of soil structure, decreased root anchoring and elevated water tables due to vegetation alteration, and the presence of relict weakness planes (such as faults and joints). Mass soil movement processes include creep, falls/topples, slides, and flows. Here, we briefly summarize the common types and causes of soil mass movement. Sidle et al. (1985) and Selby (1993) provide extensive description and explanation of the types, causes, and occurrences of soil mass movements and present approaches to slope stability assessment and analysis.

Soil creep and fall processes generally constitute minor soil loss relative to other erosion processes. Creep of soil downslope may occur as individual soil particles (particle creep) or en masse (slope creep). Particle creep occurs due to gravity, particle expansion and contraction with heating and cooling, and wetting-drying and freeze-thaw processes. Slope creep refers to the slow downslope creep of large soil masses and is a function of the creep rate and the depth of material in movement. Creep rates usually range between 0.1 and 15 mm y^{-1} on well-vegetated slopes, but may be as high as 500 mm y^{-1} on exposed slopes and areas with frequent freeze-thaw cycles (Selby 1993). Falls result from the undercutting of slope faces or toe slopes by flowing water or from cliff-top sloughing after freeze-thaw or wetting-drying periods. Falls may contribute significantly to the downslope transport of rocks, but sediment contributions from this process are generally minimal on annual time cycles.

Slides occur on failure planes that are either straight (translational) or curved (slumps), and are the most common form of landslide (Sidle et al. 1985; Selby 1993). Translational slides are more common than slumps. Translational slides usually occur due to reduced soil strength with saturation and they form in long, shallow (1 to 4 m) linear features. High-intensity rainfall saturates the soil profile, reducing the soil strength along soil material boundaries of different permeability or density. The soil-bedrock interface is a common translational failure plane where saturated soils are underlain by shallow bedrock. Overburdening, slope steepening, and ground vibrations are also causes of translational slides. The rate of movement for translational slides is commonly several meters per day. Similar to translational slides, slumps form when overburdening under wet conditions weakens the shear strength of the soil. Slumps, however, are rotational or curved failure planes, and may initiate long after water input has ceased. They usually occur in cohesive soils derived from soft rocks like shales, mudstones, and overconsolidated clays. Downslope progression usually occurs at a rate of a few millimeters per year to several meters per day. However, slumps that occur in soils with high water content may generate more substantial downslope transfer of sediment and often result in an earthflow event at the toe of the failure.

Flows are gravity-induced mass movements that are intermediate between sliding and water flows. They occur as debris, earth, or mud flow resulting from wet or dry liquefaction of coarse debris, fine-grained soil, or clay soil, respectively (Selby 1993). These fast-moving events (a few meters per day to tens of meters per second) are promoted by steep slopes, high soil water content, remolding of soil material following other mass movement events, presence of soil with low liquid limits (easily liquefied soil), ground vibrations, and the occurrence of soils with open fabrics that facilitate soil movement. Flows typically occur with abundant wetness, but may also occur as dry rock avalanches or rock fragment flows. They often occur subsequent to an upslope slide that contributes substantial debris and sediment downslope at a high velocity.

Debris flows (fig. 11) are flow events consisting of large quantities of debris and runoff. Debris flows that contain organic matter in large forms, such as trees and logs, are referred to as "debris torrents." Here, we consider the discussion that follows to be similar for debris torrents and flows, and thus refer to both simply as "debris flows." The high bulk density and viscosity of debris flows facilitate flow shear strength substantial enough to transport large boulders and debris. Debris flows progress downslope with a boulder- and debris-laden front followed by slurry and hyperconcentrated flow of coarse- and fine-soil materials. Flowpaths can extend for many kilometers and commonly cease in low-gradient alluvial fans with boulder levees (see Selby 1993).

Debris flows may occur in 20 to 100 waves during a single event, with thinner fluid pulses occurring between waves. These events can occur suddenly and pose significant risk to life, property, and resources due to the high-impact force (5 to 100 times that of floods) and velocity and the sediment/debris loading possible. For example, an extreme rainstorm event in central California in 1982 generated more than 200 mm of rainfall over 32 hours, resulting in more than 18,000 slides (see Ellen and Wiezorek 1988). Debris flows from the slides damaged at least 100 homes and killed 14 people, of whom 10 were buried in their homes. The total cost of the damage was estimated at more than $280 million. Selby (1993) provides a brief review of this and other catastrophic debris

Figure 11—Debris flow following a thunderstorm event occurring on burned forest land (photo: USDA Forest Service).

USDA Forest Service RMRS-GTR-351. 2016.

flow events of similar magnitude. Cannon et al. (1998, 2001a,b, 2008) explain runoff- and infiltration-driven triggers for debris flow initiation on burned landscapes and provide additional examples of the potential impact of debris flows on values-at-risk.

Earthflows and mudflows are slow- to rapid-moving viscous flows of fine sand, clay, and silt particles mixed with water. As with debris flows, they often result from upslope slides, particularly slumps. Slumps of wet soil bulging forward often take the form of bulbous toes or tongue-like rolls of earth and mud. The downslope movement is dependent on the weight of the material, slope steepness, shear strength of the material, and pore water pressures. The rate of movement for earthflows usually ranges from less than several meters a day to hundreds of meters per hour. Earthflows may affect areas from several square meters to hectares, but these impacts may require several years and are commonly of minor degree. Mudflows are highly mobile (with velocities of several meters per second) and pose greater threat than earthflows to life, property, and resources. For example, the volcanic eruption of Mt. Saint Helens in Washington State in 1980 generated lahar flows (volcanic mudflows) 120 km down the Toutle River and contributed more than 50 million m^3 of sediment into the Lower Columbia River (Pierson 1986). The Mt. Saint Helens event illustrates changes in flow behavior and deposition from mass movement initiation to streamflow delivery (see Scott 1988).

Spatial and Temporal Variations of Processes

Sediment detachment and transport may vary dramatically in space and time (Thornes 1980; Toy et al. 2002). The spatial scaling of sediment yield is a function of the arrangement and connectivity of surface susceptibility, driving forces (such as rainfall distribution), and erosion processes occurring within the area of interest (Pierson et al. 2011; Williams et al. 2016a). Rainsplash and sheetflow processes (fig. 9A and 9B) dominate at the small-plot scale (1 to 2 m^2) and erosion highly depends on the susceptibility of the soil surface to raindrop impact. Over large-plot scales (tens of square meters), sediment yield is more influenced by the fluid-flow entrainment of raindrop- and flow-detached sediment in sheetflow and concentrated flow (fig. 9C) and the connectivity of these processes (Williams et al. 2016a).

At the hillslope scale, the landscape often has a heterogeneous arrangement of susceptible conditions and driving forces, resulting in a poorly connected spatial organization of processes and erosion (Pierson et al. 1994a,b; Abrahams et al. 1995; Bergkamp 1998; Puigdefábregas et al. 1998; Reid et al. 1999; Wilcox et al. 2003a; Puigdefábregas 2005; Williams et al. 2014b). For example, small perturbations on a hillslope may create small patches of exposed bare soil highly susceptible to rainsplash erosion. High-intensity rainfall on these patches may generate substantial erosion from raindrop impact, but the protected surfaces between the perturbations create a disconnect at the larger hillslope scale, resulting in minor sediment yield.

The same landscape with uniform disturbance may undergo substantially more soil erosion from a similar storm due to an increase in the spatial connectivity of surface susceptibility, rainsplash detachment, and formation of well-organized sheetflow or concentrated flow (Williams et al. 2016a). At watershed scales, the distribution of rainfall or other driving forces is often highly variable, as is erodibility, facilitating even greater

disconnect than observed at hillslope scales. In-channel processes also play a role in sediment delivery over landscape scales. In general, sediment yield per unit area decreases with increased spatial area due to the inherent loss in connectivity of processes, susceptibility, and driving forces (Pierson et al. 1994a,b; Wilcox et al. 2003a). The collective arrangement creates a spatially dynamic environment of sediment detachment, transport, and deposition that is dependent on the respective magnitude and extent of each of these components' influence.

Temporal variability comes from event oscillations, climate variations, and changes in land use or disturbance regimes (see Thornes 1980). Short-term variations in erosion usually refer to changes occurring during a single storm event or over seasonal to annual time scales. During a storm, the most readily available sediment is eroded first, usually resulting in an initial pulse of sediment much greater than sediment delivery near the conclusion of the storm (Pierson et al. 2008a; Al-Hamdan et al. 2012b). The availability of sediment is a complex dynamic of sediment detachment and transport that, as previously mentioned, is highly variable in space. Oscillations in the magnitude and spatial arrangement of driving forces during individual events may create spatially and temporally variable sediment sources that further influence the temporal response.

Over seasonal and annual timescales, erosion may be influenced by changes in vegetation, soil conditions, animal activity, or climatic factors. Semiarid areas, for example, commonly have low-intensity, long-duration rainfall in winter and high-intensity, short-duration rainfall in summer (see Branson et al. 1981). These different climate regimes elicit different erosional responses. The seasonal responses are further influenced by co-occurring changes in the vegetation cover and soil wetness that may increase or decrease site erodibility. For example, denser canopy cover on semiarid rangelands during summer months, as compared to winter months, offers greater surface protection from high-intensity rainfall through increased interception. During winter months, low ground cover and freeze-thaw soil processes may facilitate reduced infiltration and high erosion rates during low-intensity rainfall (Blackburn and Wood 1990; Blackburn et al. 1990; Seyfried and Flerchinger 1994; Wilcox 1994).

Over timescales that cover multiple years or longer, erosion is more influenced by climatic changes and land-use disturbances. Prolonged drought conditions may limit plant growth and recruitment of surface-protecting litter. A series of wet years may stimulate canopy and ground cover recruitment. Likewise, effects of land use, disturbance, or postdisturbance rehabilitation may take many years or even decades to influence sediment yield and may be linked with climatic influences (Allen and Breshears 1998). Finally, all temporal responses are strongly linked to the time dependence of spatial links in erosion processes or cumulative effects.

Section 3: Climate-Vegetation-Soil-Hydrology Interactions

Precipitation processing and the resultant runoff and erosion processes at point to landscape scales are a function of climate, vegetation, and soil interactions (Seyfried and Wilcox 1995; Puigedfàbregas et al. 1999; Pierson et al. 2002a; Robichaud et al. 2010; Pierson et al. 2011; Williams et al. 2014b). Climate acts as a driving force for runoff generation and erosion, whereas vegetation and soil properties act as resistive forces. Climate influences the timing, quantity, type, and intensity (erosive energy) of precipitation falling at the land-atmosphere interface. Air temperatures and evaporation rates affect available water and plant establishment and productivity. Canopy interception controls the amount and erosive energy of precipitation passing through to the ground surface, and influences the immediate canopy area, climate, and water use. Ground cover, organic matter, and soil fauna recruitment are a function of plant productivity and climate, and their interactions regulate soil stability and retention. Ground cover reduces the erosive energy of raindrops and surface flow, increases aggregate stability, traps and stores sediment, mediates infiltration rates, and modifies the soil climate and soil fauna activity. Plant root tissues and soil fauna activity influence the soil bulk density and infiltration capacity. Organic matter and byproducts of fauna activity may inhibit or facilitate soil wettability. Soil development is strongly related to parent materials, erosion processes, climate, and the vegetative community. Soil porosity and structure affect infiltration, percolation, and throughflow, all of which influence soil water storage and antecedent moisture conditions. Soil water storage and climate regulate plant productivity and vegetation recruitment. Collectively, these relationships control the spatial and temporal arrangement of runoff and erosion processes. In this section, we discuss these relationships in detail with respect to runoff and erosion in preparation for assessing hydrologic impacts of fire.

Climate as a Driver of Hydrologic Response

The type, intensity, duration, and timing of a precipitation event greatly influence rangeland hydrologic response. Rainfall events in excess of soil infiltration capacity may produce substantial runoff, whereas the same quantity of precipitation falling as snow may generate little or no event runoff. Snow accumulation on the land surface stores precipitation and delays runoff, allowing more time for evaporative losses and infiltration (Dingman 2002; McNamara et al. 2005). For rainfall events, the intensity is an important measure for predicting rangeland response (Branson et al. 1981). High-intensity storms often greatly exceed the infiltration and storage capacity of the land surface and facilitate rapid runoff generation and large quantities of streamflow. Runoff from low-intensity storms is often minimal as long as the rainfall rate does not significantly exceed the infiltration capacity of the soil (Wilcox et al. 2003a). The duration of a storm event also influences runoff generation. High-intensity storms over long durations present the greatest risk for elevated runoff and erosion, whereas low-intensity events over short durations pose lower risks for runoff and soil loss. However, long-duration low-intensity events may also generate substantial runoff, especially under saturated soil conditions (Castillo et al. 2003; Wilcox et al. 2003a). In general, the risk for high runoff and erosion

is greater the longer the infiltration and storage capacity of the land surface is exceeded. The response is amplified with increasing rainfall intensity.

Seasonal variation in runoff behavior from rangelands is related to the prevailing precipitation regime. In the mountainous north and central United States, high-elevation range sites are mostly snow dominated while low- to mid-elevation sites are rainfall dominated with transient winter snowpacks (Branson et al. 1981; Seyfried and Wilcox 1995). The greatest runoff rates from snow-dominated uplands occur during the spring snowmelt/runoff period or during winter rain-on-snow events (fig. 12A, table 1) (Branson et al. 1981; Wilcox et al. 1991; Seyfried and Flerchinger 1994; Marks et al. 2001; Pierson et al. 2001b). The spring runoff season at these sites begins when air temperatures warm above freezing and the precipitation trend shifts from snowfall to rainfall. Large quantities of available water and saturated soil conditions amplify runoff (McNamara et al.
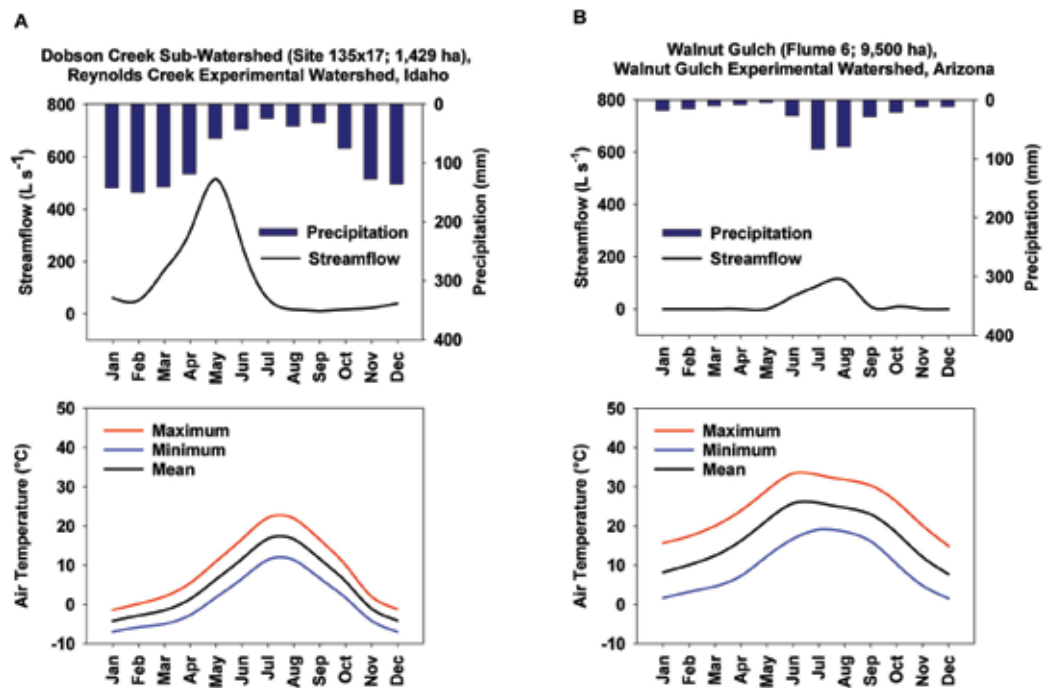

Figure 12—Mean monthly streamflow, precipitation, and air temperatures for (A) a semiarid, snow-dominated drainage in the Reynolds Creek Experimental Watershed, Idaho (USDA Agricultural Research Service 2011a) and (B) an arid, rainfall-dominated drainage in the Walnut Gulch Experimental Watershed, Arizona (USDA Agricultural Research Service 2011b) (photos: Agricultural Research Service).

Table 1—Ten largest peak flows of record measured on the snowfall-dominated Reynolds Creek (Outlet Weir, 1963 to 1996), Reynolds Creek Experimental Watershed , Idaho, and rainfall-dominated Walnut Gulch (Flume 1, 1953 to 2010), Walnut Gulch Experimental Watershed, Arizona.

Reynolds Creek, Idaho (Snowfall dominated, 23,372 ha)		Walnut Gulch, Arizona (Rainfall dominated, 14,932 ha)	
Date	**Peak flow[a] ($m^3 s^{-1}$)**	**Date**	**Peak flow[b] ($m^3 s^{-1}$)**
23 Dec 1964	109.03	17 Aug 1957	318.61
31 Jan 1963	66.02	02 Aug 1957	183.54
15 Feb 1982	58.97	12 Aug 1972	171.52
11 Jan 1979	47.09	11 Aug 2000	147.50
11 Jun 1977	31.70	10 Sep 1967	132.54
28 Jan 1965	31.53	22 Jul 1964	121.48
21 Jan 1969	25.48	22 Aug 1961	111.26
11 Apr 1982	24.40	10 Aug 1971	102.38
27 Jan 1970	20.64	16 Aug 1958	95.93
02 Mar 1972	19.19	09 Sep 1964	89.58

[a] Streamflow summaries provided by the USDA Agricultural Research Service, Northwest Watershed Research Center (USDA Agricultural Research Service 2011a).

[b] Streamflow summaries provided by the USDA Agricultural Research Service, Southwest Watershed Research Center (USDA Agricultural Research Service 2011b).

2005; Seyfried et al. 2009). Snowmelt runoff is generated mainly from subsurface return flow in or near stream channels rather than sheetflow or concentrated flow (Flerchinger et al. 1992; McNamara et al. 2005; Seyfried et al. 2009; Williams et al. 2009).

Runoff at low-to-mid elevations occurs primarily as overland flow due to rainfall on shallow snowpacks (less than 30 cm snow) and frozen soils (Johnson and Smith 1978; Wilcox et al. 1989; Blackburn and Wood 1990; Blackburn et al. 1990; Wilcox et al. 1991; Seyfried and Flerchinger 1994; Seyfried and Wilcox 1995; Marks et al. 2001; Pierson et al. 2001b). Runoff during summer months usually results from short-duration, high-intensity rainfall in excess of infiltration (Branson et al. 1981; Wilcox et al. 1991; Pierson et al. 2001b). For snow-dominated sites and valley locations, infiltration-excess runoff from high-intensity storms is often limited to small areas resulting from isolated rainfall patterns and heterogeneous soils and canopy/ground cover (Blackburn 1975; Blackburn et al. 1992; Pierson et al. 1994a,b).

In the desert southwestern United States, some mountainous locations present intermediate precipitation-runoff patterns, in contrast with the purely snow-dominated or rainfall-dominated northwest and central U.S. uplands (Wilcox et al. 2003a). Such sites exhibit both winter snow-dominated and summer rainfall-dominated precipitation-runoff regimes, but the largest runoff events are usually related to intense monsoon summer thunderstorms (Wilcox 1994; Breshears et al. 1998; Wilcox et al. 2003a). Valley range-lands in the desert Southwest also experience the greatest catchment runoff (90 percent of annual) during the summer monsoon season (fig. 12B, table 1; Osborn and Lane 1969; Branson et al. 1981; Osborn and Renard 1988; Wilcox et al. 2003a; Stone et al. 2008). Monsoonal storms occur as variable high-intensity (up to 250 mm h^{-1}), short-duration (5 to 30 min), and spatially limited (10 to 100 km^2) events (Osborn 1964; Branson et al. 1981; Osborn 1983a,b; Osborn and Renard 1988; Renard 1988; Renard et al. 1993; Goodrich et al. 2008). The more intense (more than 100 mm h^{-1}) storms produce

significant point and hillslope-scale runoff, yielding watershed-scale runoff equivalent to several years of normal runoff (Branson et al. 1981; Renard et al. 1993).

Sediment generation from rangelands is often related to the amount of runoff. For snow-dominated uplands, peak erosion occurs during snowmelt runoff (Branson et al. 1981; Pierson et al. 2001b). Streamflow sediment concentrations during snowmelt runoff are most influenced by streambank sloughing and streambed-sediment entrainment processes, rather than rainsplash and sheetflow. Low- to mid-elevation mountainous sites may experience the greatest erosion rates from low-intensity, long-duration rain events during freeze-thaw periods (Seyfried and Flerchinger 1994). Freeze-thaw processes increase surface erodibility, and, combined with lower vegetation and ground cover in winter months, facilitate increased surface erosion from rainsplash and sheetflow (Blackburn and Wood 1990; Blackburn et al. 1990; Seyfried and Flerchinger 1994; Wilcox 1994).

Summer convective, high-intensity events at snow-dominated sites may generate substantial point and hillslope erosion from rainsplash, sheetflow, and concentrated flow. Sediment eroded at the point and hillslope scales during these events usually remains onsite, stored locally or in ephemeral stream channels, as runoff is commonly too low for offsite transport (Slaughter and Pierson 2000). Where summer runoff occurs, sediment concentration per unit of runoff is usually several magnitudes greater than during the spring runoff period (Johnson and Smith 1978). As with runoff, sediment yield in the desert Southwest peaks during the summer monsoon season (Branson et al. 1981; Wilcox et al. 1996; Reid et al. 1999; Wilcox et al. 2003a; Nearing et al. 2007). High-intensity rainfall during this period magnifies rainsplash, sheetflow, concentrated flow, and flash flood erosion, generating most of the annual surface erosion from southwestern U.S. rangelands (Branson et al. 1981; Renard et al. 1993).

Vegetation Influences on Water Availability at the Surface and Near-Surface

Interception Effects on Water Availability

Canopy and ground cover interception at the land-atmosphere interface primarily influence rangeland hydrology by limiting the amount of water available for infiltration and runoff (fig. 13). For low-intensity, short-duration rainfall events, most of the precipitation is captured by plant canopies, litter, and other ground cover, and results in evaporative losses (Branson et al. 1981; Owens et al. 2006; Dunkerley 2008). High-intensity rainfall events usually exceed cover storage capacities, resulting in some precipitation routing by throughflow and stemflow processes in addition to the evaporative losses (Thurow et al. 1987; Návar and Bryan 1990; Martinez-Meza and Whitford 1996; Whitford et al. 1997; Wainwright et al. 1999; Dunkerley 2000; Abrahams et al. 2003; Bhark and Small 2003; Carlyle-Moses 2004; Owens et al. 2006; Dunkerley 2008). The percentage of event gross rainfall captured by cover elements generally decreases as rainfall intensity increases (Branson et al. 1981; Carlyle-Moses 2004; Owens et al. 2006). Cumulative interception losses over multistorm and annual time periods vary with the frequency and magnitude of precipitation events and meteorological conditions (Dunkerley 2008).
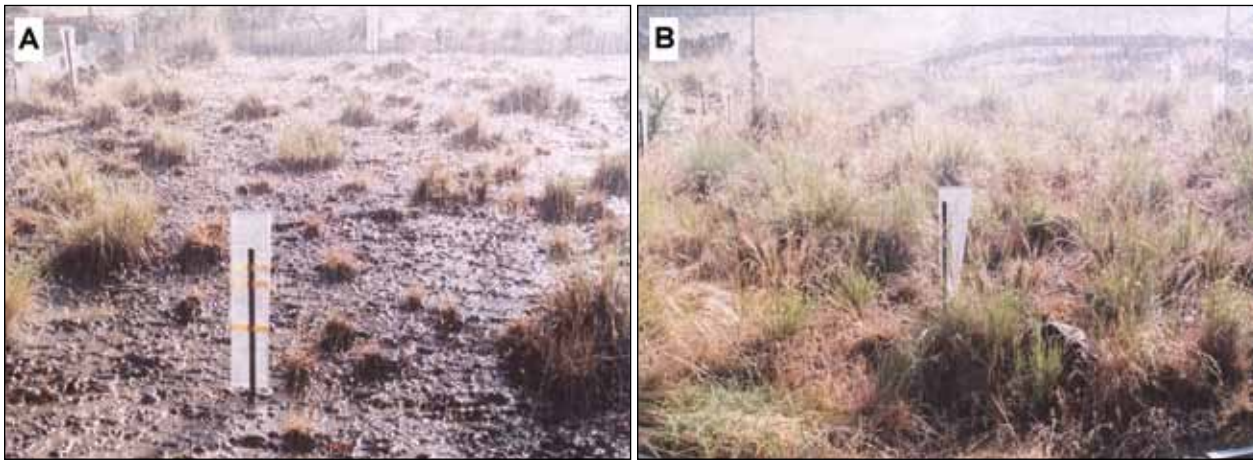
Figure 13—Ponding and interception of artificial rainfall applied on (A) a sagebrush site invaded by western juniper and (B) the same site 10 years following juniper removal (photos: USDA Agricultural Research Service, Northwest Watershed Research Center, adapted from Pierson et al. 2007a).

Interpreting estimates of rainfall interception loss, throughfall, and stemflow from literature is confounded by the variability in measurement approaches, reported units, and spatial and temporal experimental scales (Dunkerley and Booth 1999; Dunkerley 2000; Abrahams et al. 2003; Llorens and Domingo 2007; Dunkerley 2008). Estimates have been reported from event, seasonal, and annual timescales over varying rainfall intensities, storm patterns, and cumulative precipitation (Branson et al. 1981; Dunkerley and Booth 1999; Dunkerley 2000; Carlyle-Moses 2004; Llorens and Domingo 2007; Dunkerley 2008). Results are commonly expressed as a depth of water or percentage of gross precipitation falling on individual plants or over an entire plant community. Extrapolation of these results to other like plants and communities is tenuous given variability in vegetation and climate characteristics from one rangeland site to another (Dunkerley and Booth 1999; Dunkerley 2000, 2008). General estimates suggest tree, shrub, and grass foliage can store about 1.3 mm of water (Bonan 2002).

Branson et al. (1981) and others (Rowe 1948; Hamilton and Rowe 1949; Hull and Klomp 1974; West and Gifford 1976; Tromble 1983; Thurow et al. 1987; Tromble 1988; Návar and Bryan 1990; Martinez-Meza and Whitford 1996; Wood et al. 1998; Dunkerley and Booth 1999; Dunkerley 2000; Abrahams et al. 2003; Wilcox et al. 2003b; Carlyle-Moses 2004; Owens et al. 2006; Llorens and Domingo 2007; Dunkerley 2008; Taucer et al. 2008) provide estimates of event and annual rainfall interception and stemflow for a variety of rangeland plants and communities. Here, we summarize these data to provide the reader an idea of the general magnitude of precipitation that returns to the atmosphere via interception loss. We summarize ranges in the percentage of gross rainfall intercepted at the event and longer term (seasonal and annual) timescales by individual cover type and by plant community (see table 2). Gross rainfall interception by rangeland individual shrubs (Rowe 1948; Hamilton and Rowe 1949; West and Gifford 1976; Branson et al. 1981; Tromble 1983) and conifer trees (Skau 1964; Slaughter 1997; Owens et al. 2006; Taucer et al. 2008) averages from 50 to 60 percent for low-intensity storms to 5 to 35 percent for high-intensity or large events. Gross rainfall interception by individual shrubs over multistorm to annual timescales ranges from 5 to 46 percent (Hull 1972; Hull and Klomp 1974; Thurow et al. 1987; Tromble 1988; Návar and Bryan 1990; Martinez-Meza

Table 2—Event and annual interception rates reported in literature for various individual rangeland plant and community types (see Branson et al. 1981).

Cover type	Event interception as % of gross rainfall	Annual interception as % of gross rainfall
Individual conifer or shrub	50–60 for Low-Intensity 5–35 for High Intensity	5–50, 5–15 more common
Litter	2–20	2–20
Shrub or woodland community	5–50	5–25
Herbaceous community	15–80	10–55

and Whitford 1996; Domingo et al. 1998; Serrato and Diaz 1998), with most reported values around 5 to 15 percent.

Shrub- and woodland-community rainfall interception on the annual scale ranges between 5 and 25 percent (Rowe 1948; Hamilton and Rowe 1949; Pressland 1973; West and Gifford 1976; Tromble 1983; Thurow et al. 1987; Tromble 1988; Dunkerley and Booth 1999; Carlyle-Moses 2004). Fewer data are available for herbaceous vegetation and litter interception. Clark (1940) measured rainfall interception by native prairie grasses in Nebraska, USA, at levels of 29 to more than 80 percent under low-intensity artificial rainfall. Thurow et al. (1987) summarized several studies that found grassland interception of gross annual rainfall ranges from 13 to 56 percent. Thurow et al. (1987) estimated that interception of gross annual rainfall at two Texas grassland sites with 56 percent (shortgrass) and 62 percent (midgrass) cover was 11 percent and 18 percent, respectively. Dunkerley and Booth (1999) reported a 32 percent interception of gross annual rainfall by grass in Australia.

Branson et al. (1981) provides a summary of literature on litter interception with estimates of 2 to 17 percent of gross annual rainfall. These estimates are quite variable in part because of the methods used to identify when the litter layer stops and the soil layer begins. For example, Owens et al. (2006) estimated that 5 percent of gross annual rainfall on Ashe juniper (*Juniperus ashei* Buchholz) trees at a Texas woodland site was intercepted by the coarse litter beneath trees, whereas Thurow et al. (1987) determined that litter (all dead plant material above the mineral soil) intercepted 20 percent of annual rainfall in a live oak (*Quercus virginiana* Mill.) motte of the Edwards Plateau, Texas. Routing of intercepted water as stemflow has been estimated at 5 to 17 percent of gross rainfall for individual rangeland conifers (Thurow and Hester 1997; Owens et al. 2006; Taucer et al. 2008) and at 3 to 10 percent of gross rainfall for individual shrubs and shrub communities (Thurow et al. 1987; Návar and Bryan 1990; Martínez-Meza and Whitford 1996; Wainwright et al. 1999; Abrahams et al. 2003; Bhark and Small 2003; Carlyle-Moses 2004).

Canopy interception of snowfall is generally considered of minor hydrologic importance because most intercepted snow reaches the ground as meltwater or is shed as large snow masses (Dingman 2002; Storck et al. 2002). General interception estimates suggest trees can store about 3.8 mm of water as snow (Bonan 2002). Literature on snowfall canopy interception specific to rangeland plants is extremely limited. The most commonly cited references are Hull (1972) and Hull and Klomp (1974). They found dense shrub cover (2.2 plants per m^2) intercepted 37 percent of snowfall at an Idaho rangeland site.

Reports of snow interception measurements of conifer species come from woodland- and forest-dominated sites. Breshears et al. (1997b) reported that snow accumulation during each of three winter seasons was much greater in areas between tree canopies than underneath canopies at a twoneedle pinyon/oneseed juniper (*Pinus edulis* Englem./*J. monosperma* (Engelm.) Sarg.) woodland site in New Mexico. Snow water equivalent in the first year of the 3-year study was about 80 percent greater in openings between trees than in areas underneath canopies. Schmidt and Gluns (1991) reported 45 to 50 percent canopy interception of snowfall for three conifer species in a forested setting when snow water equivalent was 10 mm and snow specific gravity was 0.06. Interception decreased to 10 percent for an equivalent storm with snow specific gravity of 0.13 (Schmidt and Gluns 1991). In another forested study, Storck et al. (2002) measured approximately 60 percent canopy interception of snowfall by four different conifer species, with minimal differences between species.

Whole plant interception of windblown snowfall by rangeland vegetation is paramount in retaining snow against wind scour (Seyfried and Flerchinger 1994; Pomeroy and Gray 1995; Seyfried and Wilcox 1995; Flerchinger et al. 1998; Liston and Sturm 1998; Flerchinger and Cooley 2000; Marks and Winstral 2001; Marks et al. 2001; Sturm et al. 2001; Liston et al. 2002; Marks et al. 2002; Winstral and Marks 2002). Wind and topography interact to redistribute fallen snow on undulating terrain, while vegetation reduces wind velocities and facilitates deposition (Marks et al. 2001; Marks and Winstral 2001; Marks et al. 2002; Winstral and Marks 2002). Deeper snow accumulations provide greater insulation for surface soils and plant productivity and prolong snow-covered periods (Sturm et al. 2001; Liston et al. 2002).

The vegetation snow-holding capacity is a function of the vegetation height, density of plants, and snowpack conditions (Pomeroy and Gray 1995; Liston and Sturm 1998; Sturm et al. 2001). Overall, accumulation of windblown snow is maximized at the height of the canopy. Any deposition in excess of canopy height is readily windblown between events, potentially transported offsite, or lost to wind-driven sublimation (Sturm et al. 2001; Liston et al. 2002). Hutchinson (1965) found that a shrub stand 50 cm in height stored 25 mm more water than an adjacent area void of shrubs. Flerchinger et al. (1998) reported that snow depth at a wind-driven rangeland site in Reynolds Creek Experimental Watershed, Idaho, typically varied by plant community from less than 60 cm in low sagebrush/grass (*Artemisia arbuscula/Poa secunda* J. Presl) to 100 cm in mountain big sagebrush/snowberry (*A. tridentata* Nutt. ssp. *vaseyana* (Rydb.) Beetle/*Symphoricarpos* spp.) and 100 to 800 cm in an aspen/willow (*Populus tremuloides*/*Salix* spp.) stand. Streamflow over a 10-year period at the site highly depended on vegetation retention of windblown snow (Flerchinger and Cooley 2000). Numerous other studies have reported similar results from water-limited, snow-dominated rangelands, indicating the retention of windblown snow is a dominant influence on the timing and quantity of water available for infiltration, soil storage, plant use, and streamflow generation (Seyfried and Wilcox 1995; Flerchinger et al. 1998; Luce et al. 1998; Flerchinger and Cooley 2000; Marks et al. 2001; Winstral and Marks 2002; Seyfried et al. 2009; Williams et al. 2009).

Canopy Influence on Soil Microclimate, Evapotranspiration, and Antecedent Soil Water

In addition to intercepting water, vegetation and litter may further moderate the soil microclimate and water availability by shading solar radiation and insulating surface soils (Belsky et al. 1989; Pierson and Wight 1991; Joffre and Rambal 1993; Breshears et al. 1997b, 1998; Domingo et al. 2000; Lebron et al. 2007). The near-surface microclimate influences soil evaporation and affects soil moisture regimes (Branson et al. 1981; Breshears et al. 1998; Hillel 1998). Pierson and Wight (1991) reported interspace locations (areas between shrub canopies, also called intercanopy) in a sagebrush community (*A. tridentata* Nutt.) had higher (by 5.2 °C) maximum and lower (by 1.5 °C) minimum near-surface (0 to 10 cm depth) soil temperatures than coppice locations (areas underneath and immediately adjacent to canopies, also called subcanopy) during the spring season. Small grass clumps and moss clumps within interspaces had little influence on near-surface soil temperatures. Pierson and Wight (1991) inferred that shrub cover and the associated litter mounds (coppices) insulated the soil surface from incoming solar radiation during daylight hours and from sensible heat loss at night. Breshears et al. (1997b) found that interspaces between tree canopies of a twoneedle pinyon/oneseed juniper woodland exhibited greater (40 to 50 percent more) near-surface solar radiation than tree coppices, and that preferential shading on the northern side of tree coppices significantly reduced near-surface solar radiation. Solar radiation differences between coppice and interspaces were much greater during the summer solstice.

Breshears et al. (1997b) also determined that snow water equivalent was greater in interspace locations than under tree canopies and that the differential accumulation resulted in temporal variability in the spatial arrangement of soil water. Soils underneath tree canopies were wetter than interspace soils in early winter following complete melt of coppice snowpacks and during the monsoon season immediately after intense runoff-generating rainfall events. Wetter soil conditions on the edges of coppices compared to interspaces following intense rainfall were assumed partially related to lateral redistribution of surface runoff from interspace locations to coppices as runon. Interspace soils were wetter, by 3 percent volumetric moisture content, than coppice soils later in the winter and in early spring during the interspace snowmelt period. The differential snow accumulation and melt patterns, largely related to canopy snow interception, exerted a greater influence on the spatial distribution of soil water (canopy versus interspace locations) than did the effects of preferential shading (Breshears et al. 1997b). The primary effect of solar radiation on soil moisture patterns was observed within interspace patches; north edges with more solar radiation were wetter than south edges during winter and spring. Breshears et al. (1997b) attributed within-interspace differences to the canopy drip effects (melting snow) on the warmer south side of trees (north edge of interspaces).

Breshears et al. (1998), working at the same site as Breshears et al. (1997b), found that maximum air temperature was as much as 10 °C greater on interspaces than tree coppices during late spring through summer and that the associated differences in spatial temperature produced differences in soil evaporation. Breshears et al. (1998) determined that spatial differences in soil temperature influenced soil evaporation only when soils were thawed and were amplified at lower soil water contents (as expressed by soil water potential). Joffre and Rambal (1993) reported greater water storage under tree canopies

(with grass) than in unshaded interspaces on three Mediterranean rangeland sites when precipitation exceeded mean annual levels. Soil water storage was low in interspace and coppice locations when precipitation was limited, resulting in the loss of about 60 percent and more than 95 percent of precipitation to evapotranspiration in interspaces and coppices, respectively. Domingo et al. (2000) found that canopy shading in a semiarid Mediterranean shrubland created a milder microclimate in shrub areas and that diurnal temperature fluctuations were greater in interspaces.

The above noted studies indicate that canopy effects on the near-surface microclimate may influence soil water availability, but that the overall impact is highly dependent on the quantity of precipitation and other spatial effects (interception, lateral redistribution). Furthermore, microclimate effects may have greater implications for biological processes (seed germination and emergence, nutrient and microbial processes) and spatial vegetation structure than for direct runoff generation (Pierson and Wight 1991; Ludwig and Tongway 1995; Scholes and Archer 1997; Breshears et al. 1998; Breshears and Barnes 1999; Reynolds et al. 1999; Belnap et al. 2005; Huxman et al. 2005; Ludwig et al. 2005; D'Odorico et al. 2007).

The direct effects of evapotranspiration on rangeland runoff generation vary by precipitation regime and are usually minor relative to their influence on annual and seasonal water balances (Branson et al. 1981). Annual runoff from rangeland sites usually represents 0 to 10 percent (but can be as high as 50 percent) of annual precipitation depending on the type and structure of the plant community, meteorological patterns, and soils/geology (Carlson et al. 1990; Wilcox et al. 1991; Joffre and Rambal 1993; Wilcox 1994; Weltz and Blackburn 1995; Carlson and Thurow 1996; Wilcox et al. 1996, 1997; Flerchinger et al. 1998; Flerchinger and Cooley 2000; Wilcox et al. 2003b, 2006; Nearing et al. 2007; Wilcox et al. 2008). Runoff generation from rainfall-dominated rangelands primarily occurs as infiltration-excess overland flow and is minimally influenced by evapotranspiration demands (Branson et al. 1981; Pierson et al. 2001b; Wilcox et al. 2003a; Stone et al. 2008). Exceptions occur following multistorm events or prolonged low-intensity storms that wet up the near-surface environment, shifting the runoff process to saturation excess (Wilcox 1994; Castillo et al. 2003).

Saturation excess runoff is a function of available soil water storage and rainfall intensity. Storage capacity is related to antecedent moisture conditions and soil structure and depth. Under these conditions, evapotranspiration dictates the storage capacity or degree of saturation in the near-surface environment, strongly influencing, along with rainfall intensity, the timing and quantity of runoff. Saturation excess overland flow is more common at snow-dominated sites, during and immediately after peak snowmelt, and therefore, evaporation of water from the snowpack or saturated soils in these settings plays an important role in reducing water availability for surface and subsurface flow (Seyfried and Flerchinger 1994; Marks et al. 2001; Pierson et al. 2001b; Wilcox et al. 2003a; McNamara et al. 2005; Seyfried et al. 2009; Williams et al. 2009).

Evapotranspiration demands at snow-dominated sites strongly influence, along with seasonal water input, the seasonal duration of ephemeral streamflow (McNamara et al. 2005; Williams et al. 2009). Annual actual and potential evapotranspiration, inclusive of interception losses, make up more than 90 percent of annual precipitation from rangeland sites, and are limited primarily by the amount of precipitation and available soil water (Campbell and Harris 1977; Branson et al. 1981; Carlson et al. 1990;
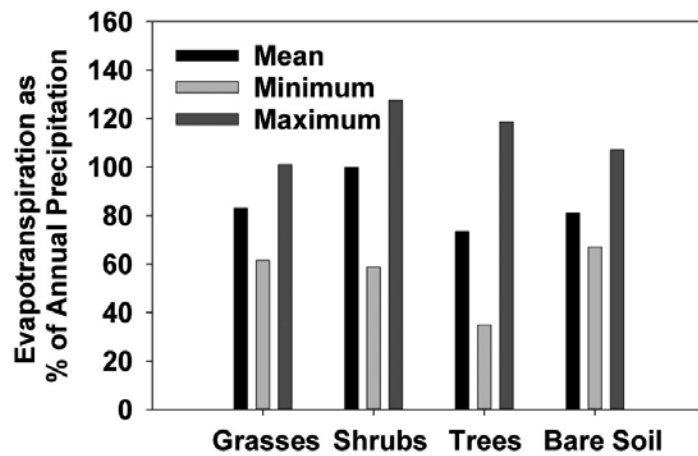
Figure 14—Estimated averages and ranges (as maximum and minimum) of annual evapotranspiration as percentage of annual precipitation for various rangeland plant types as reported in literature (Carlson et al. 1990; Joffre and Rambal 1993; Weltz and Blackburn 1995; Carlson and Thurow 1996; Flerchinger et al. 1998; Yoder and Nowak 1999b; Brandes and Wilcox 2000; Flerchinger and Cooley 2000).

Weltz and Blackburn 1995; Flerchinger et al. 1996, 1998; Yoder and Nowak 1999b; Flerchinger and Cooley 2000; Zhang et al. 2001; Huxman et al. 2005; Wilcox et al. 2006; Wilcox and Thurow 2006; Wilcox et al. 2008). Estimates from literature indicate actual evapotranspiration from rangeland herbaceous plants, shrubs, trees, and bare soil ranges from 60 to 100 percent, 60 to 130 percent, 35 to 120 percent, and 70 to 110 percent of annual precipitation, respectively (fig. 14). A review of literature on evapotranspiration rates by various rangeland plant communities is provided by Branson et al. (1981). The percentage of evapotranspiration occurring as transpiration varies considerably (7 to 80 percent) between plant communities and depends on the amount and timing of precipitation, available energy, plant growth form, and water availability throughout the rooting depth of the soil profile (Reynolds et al. 2000; Wilcox et al. 2003b; Huxman et al. 2005; Scott et al. 2006; Stannard and Weltz 2006; Moran et al. 2009). Evaporation from the soil surface is dependent on surface soil moisture conditions and available energy, and generally increases with increasing exposure of bare ground (Breshears et al. 1998; Scott et al. 2006; Moran et al. 2009).

Cover Influences on Infiltration, Runoff, and Water Transfer and Storage

The heterogeneous vegetative and ground cover structure and soil characteristics across rangeland communities exhibit a high degree of spatial organization and integration relative to water and soil resource recruitment (Tongway et al. 1989; Tongway and Ludwig 1990; Pierson et al. 1994b; Dunkerley and Brown 1995; Ludwig et al. 1997; Scholes and Archer 1997; Breshears and Barnes 1999; Puigdefábregas et al. 1999; Reynolds et al. 1999; Belnap et al. 2005; Ludwig et al. 2005; Puigdefábregas 2005; Rango et al. 2006; D'Odorico et al. 2007; Turnbull et al. 2012). The interaction of vegetation, ground cover, soil properties, climate, and resultant hydrologic processes on water-limited sites creates stable patches of water, nutrient, and soil accumulation and retention. Shrub/tree coppices and herbaceous or litter-covered areas create surface and subsurface conditions that favor infiltration and soil and nutrient retention, whereas bare areas exhibit higher rates of runoff and soil loss (figs. 15 and 16; Blackburn 1975; Abrahams et al. 1988; Schlesinger et al. 1990; Seyfried 1991; Blackburn et al. 1992;

Pierson et al. 1994a,b; Wilcox 1994; Abrahams et al. 1995; Ludwig and Tongway 1995; Wilcox and Breshears 1995; Parsons et al. 1996; Schlesinger et al. 1996; Wilcox et al. 1996; Whitford et al. 1997; Bergkamp 1998; Davenport et al. 1998; Reid et al. 1999; Schlesinger et al. 1999, 2000; Dunkerley 2002; Wainwright et al. 2002; Bhark and Small 2003; Wilcox et al. 2003a,c; Pierson et al. 2010; Turnbull et al. 2010a,b; Pierson et al. 2011, 2013, 2014; Williams et al. 2014a).

Canopy and ground cover influence the soil microclimate and the recruitment of soil microbes and microfauna that aid nutrient recycling and further improve infiltration and soil water storage (Pierson and Wight 1991; Blackburn et al. 1992; Breshears et al. 1997b, 1998; Imeson et al. 1998; Reynolds et al. 1999; Belnap et al. 2005; Ludwig et al. 2005). Heterogeneous vegetation patterns yield horizontally and vertically differential water use and soil water storage (Walter 1971; Belsky et al. 1989; Belsky et al. 1993; Joffre and Rambal 1993; Ryel et al. 1996; Breshears et al. 1997a,b, 1998; Breshears and Barnes 1999; Reynolds et al. 1999; Ludwig et al. 2005). These surface and subsurface interactions, in an undisturbed condition, result in an organized plant community that facilitates a positive feedback of biological productivity and hydrologic processes to conserve water, soil, and nutrient resources (Ludwig and Tongway 1995; Ludwig et al. 1997; Davenport et al. 1998; Cammeraat and Imeson 1999; Puigdefábregas et al. 1999; Reid et al. 1999; Reynolds et al. 1999; Ludwig and Tongway 2000; Ludwig et al. 2000; Pyke et al. 2002; Wilcox et al. 2003a; Ludwig et al. 2005; Puigdefábregas 2005; Pierson et al. 2010). The stability of the system is defined by its resistance to reduction of these capacities and by its resiliency to perturbations (Schlesinger et al. 1990; Ludwig and Tongway 1995; Ludwig et al. 1997; Ludwig and Tongway 2000; Pyke et al. 2002; Ludwig et al. 2005; Williams et al. 2014a).

Vegetation and Cover Influences on Infiltration and Runoff Generation

Higher infiltration rates on coppice mounds versus interspaces are attributed to deeper surface soil horizons, greater organic matter accumulation and aggregate stability, lower bulk density, macropores, canopy interception and stemflow, and surface retention of throughflow and runon underneath and immediately adjacent to the canopy area. Litter amassment and decomposition underneath shrub and tree canopies (fig. 15) and differential rainsplash contribute to soil, organic matter, and nutrient accumulation (Blackburn 1975; Blackburn et al. 1992; Parsons et al. 1992; Ludwig and Tongway 1995; Schlesinger et al. 1996; Puigdefábregas et al. 1999; Reynolds et al. 1999; Schlesinger et al. 1999; Belnap et al. 2005; Ludwig et al. 2005). Litter and organic matter promote aggregate stability, macropore formation, and low bulk densities associated with higher infiltration rates and retain surface water, prolonging time for infiltration (Meeuwig 1970; Blackburn and Skau 1974; Tromble et al. 1974; Roundy et al. 1978; Wood et al. 1978; Wood and Blackburn 1981; Beven and Germann 1982; Devaurs and Gifford 1984; Thurow et al. 1986; Johnson and Gordon 1988; Wilcox et al. 1988; Blackburn et al. 1990; Dunne et al. 1991; Seyfried 1991; Pierson et al. 1994a,b; Abrahams et al. 1995; Parsons et al. 1996; Seyfried and Wilcox 1995; Cerdà 1998b; Wilcox et al. 2003a; Puigdefábregas 2005; Pierson et al. 2010, 2013, 2014; Williams et al. 2014a). Soil fauna activity is enhanced by the microclimate, moisture regimes, and nutrient availability underneath canopies. The associated biological activity further improves soil aggregation, macroporosity, and
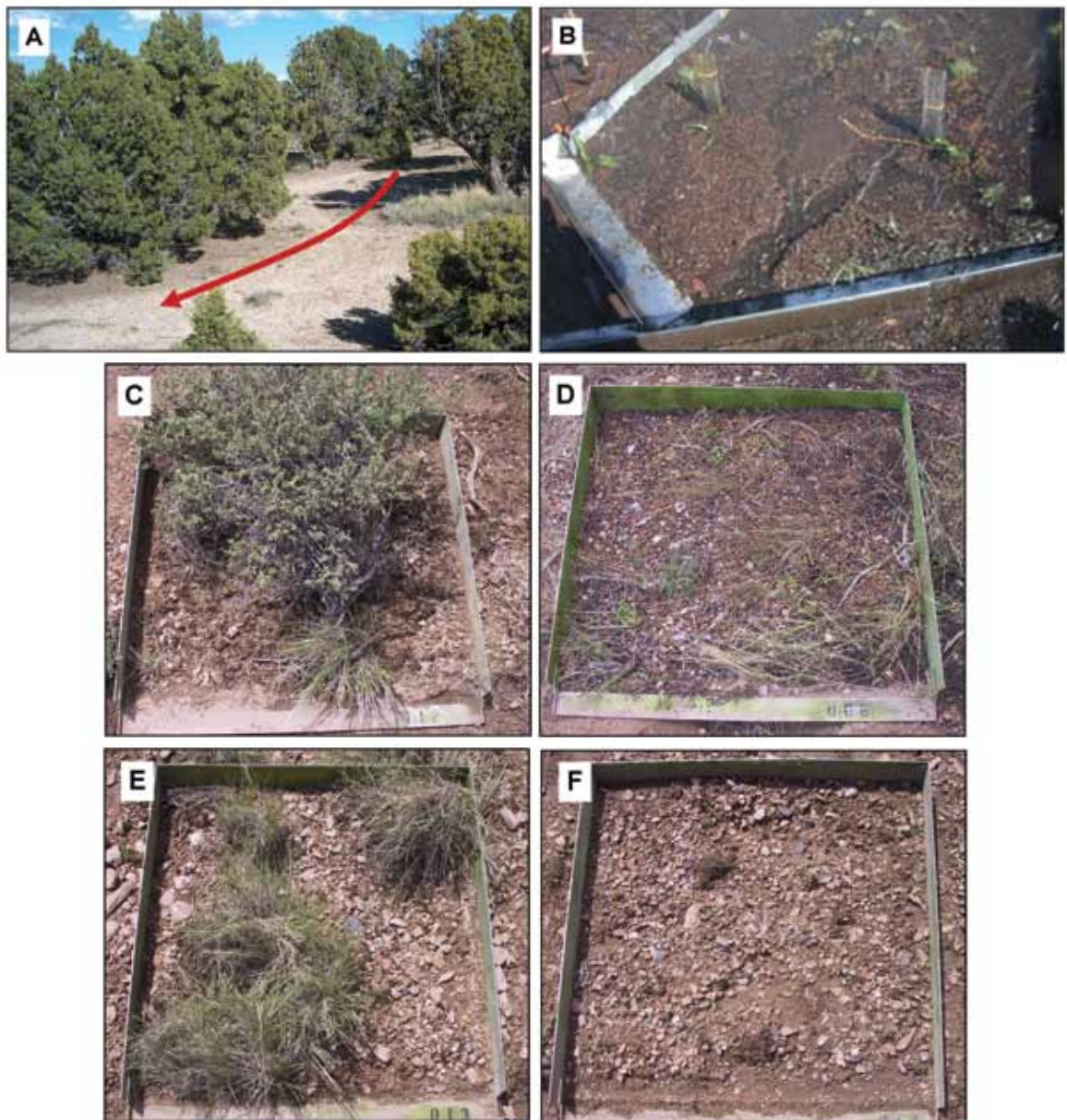
Figure 15—Photographs from a Utah juniper woodland site in Utah, showing (A) general direction for overland flow over contiguous bare interspace, (B) interspace concentrated flowpath during artificial rainfall simulation, (C) hydrologically stable shrub and (D) tree coppice microsites, (E) moderately stable vegetated interspace microsite; and (F) hydrologically unstable bare interspace microsite. High-intensity artificial rainfall simulations performed at the site by Pierson et al. (2010) produced 38, 46, 133, and 313 cumulative grams of sediment per m^2 area from 0.5 m^2 plots on the shrub coppice (C), tree coppice (D), and vegetated (E) and bare (F) interspaces, respectively (photos: USDA Agricultural Research Service, Northwest Watershed Research Center).

infiltration (Cammeraat and Imeson 1998; Imeson et al. 1998; Puigdefábregas et al. 1999; Dunkerley 2002; Belnap et al. 2005; Ludwig et al. 2005). Stemflow concentrates water input at plant bases, allowing rapid vertical recharge of the soil profile via preferential flow along root channels (Thurow et al. 1987; Návar and Bryan 1990; Martinez-Meza and Whitford 1996; Newman et al. 1997; Whitford et al. 1997; Devitt and Smith 2002; Abrahams et al. 2003; Bhark and Small 2003; Carlyle-Moses 2004; Owens et al. 2006; Lebron et al. 2007).

Plant growth form also influences infiltration processes. Infiltration rates are generally higher for bunchgrasses than sod-forming grasses (Wood and Blackburn 1981; Knight et al. 1984; Thurow et al. 1986, 1988; Blackburn et al. 1992; Pierson et al. 2002a). Greater vegetative biomass and organic matter accumulation on bunchgrasses than sodgrasses result in greater rainfall and runoff interception (Knight et al. 1984; Thurow et al. 1986, 1988). Additionally, biomass and organic matter accumulations under bunch-grasses most likely favor infiltration-increasing microbial activity (Blackburn et al. 1992). Infiltration under shrub canopies is usually greater than under grass canopies (Wood and Blackburn 1981, Schlesinger et al. 1999), but the relationship may be reversed depending on grass biomass (Wilcox et al. 1988). The overall greater infiltration in canopy patches on shrublands and grasslands increases water availability beneath canopies, which in turn stimulates biological activity, plant growth, and organic matter and nutrient recruitment. This creates a continuous positive feedback (Schlesinger et al. 1990; Belnap et al. 2005; Puigdefábregas 2005; D'Odorico et al. 2007).

Interspace areas on rangelands, particularly shrublands, are often associated with surface and subsurface characteristics that inhibit infiltration and soil water storage, and promote rapid ponding (fig. 13) and runoff initiation. Interspaces occur with various amounts of herbaceous cover, or exist as contiguous bare patches (fig. 15; Blackburn et al. 1992; Pierson et al. 1994a,b; Abrahams et al. 1995; Seyfried and Wilcox 1995; Wilcox and Breshears 1995; Parsons et al. 1996; Reid et al. 1999; Wilcox et al. 2003a). Well-vegetated interspaces may exhibit similar surface characteristics as canopy areas to some degree, but usually generate more surface runoff (table 3; Reid et al. 1999; Bhark and Small 2003; Wilcox et al. 2003a). On more water-limited or degraded sites, interspaces have low plant biomass and organic matter (fig. 15F) and thin surface soil accumulations (Blackburn 1975; Abrahams and Parsons 1991a; Abrahams et al. 1995; Parsons et al. 1996; Wilcox et al. 1996; Pierson et al. 2010). These characteristics result in poor ag-gregate stability and soil structure, and high bulk densities relative to coppices. They also facilitate low infiltration rates (Blackburn and Skau 1974; Blackburn 1975; Roundy et al. 1978; Wood et al. 1978; Thurow et al. 1986; Johnson and Gordon 1988; Wilcox et al. 1988; Blackburn and Wood 1990; Blackburn et al. 1990; Abrahams and Parsons 1991a; Seyfried 1991; Blackburn et al. 1992; Pierson et al. 1994a,b; Abrahams et al. 1995; Seyfried and Wilcox 1995; Wilcox and Breshears 1995; Abrahams et al. 1996; Parsons et al. 1996; Wilcox et al. 1996; Reid et al. 1999; Wilcox et al. 2003a; Pierson et al. 2010, 2011, 2013, 2014; Williams et al. 2014a). In general, surface characteristics of interspace areas are consistently different from coppices throughout the year, but the magnitude of the differences and respective influences on infiltration exhibit some seasonality. The spatial differences in vegetation cover and surface characteristics exert a greater influence than do seasonal differences on infiltration and runoff generation from sparsely covered shrublands, whereas seasonal differences in spatially arranged plant biomass might be of greater influence on infiltration patterns on well-vegetated grass-dominated sites (Blackburn et al. 1992).

Infiltration in interspace locations is strongly influenced by the expanse of bare ground, rock cover, or vesicular crusts (Blackburn 1975; Wood et al. 1978; Johnson and Gordon 1988; Abrahams and Parsons 1991a; Parsons et al. 1992; Abrahams and Parsons 1994; Pierson et al. 1994a; Parsons et al. 1996; Reid et al. 1999; Pierson et al. 2010). Exposure of bare ground to raindrop impact increases potential for surface sealing or

Table 3—Site characteristics, runoff coefficients, and sediment yield reported for rainfall simulation and natural rainfall experiments in the western United States.

Study	Climatic region (location)	Microsite/ plant community	Treatment/ burn severity	Plot size (m²)	Slope (%)	Time postfire (mth)	Rain rate (mm h⁻¹) /duration (min)	Rain type	WDPT (s)[a]	Soil water (%)[b] conditions	Bare soil (%)	Canopy cover (%)	Ground cover (%)	Surface roughness (mm)	Runoff coef. (%)[c]	Sed. yield (g m⁻²)
Craddock and Pearse (1938)	Cold desert (Great Basin, Idaho)	Bunchgrasses	Unb	20.2	30	-	46/60	Artificial	-	Dry	-	-	~35	-	1	1
		Bunchgrasses	Unb	20.2	30	-	92/30	Artificial	-	Dry	-	-	~35	-	1	1
		Bunchgrasses	Unb	20.2	40	-	46/60	Artificial	-	Dry	-	-	~35	-	<1	1
		Bunchgrasses	Unb	20.2	40	-	92/30	Artificial	-	Dry	-	-	~35	-	<1	1
Craddock and Pearse (1938)	Cold desert (Great Basin, Idaho)	Grassland (*Bromus tectorum*)	Unb	20.2	30	-	46/60	Artificial	-	Dry	-	-	~20	-	10	31
		Grassland (*Bromus tectorum*)	Unb	20.2	30	-	92/30	Artificial	-	Dry	-	-	~20	-	5	11
		Grassland (*Bromus tectorum*)	Unb	20.2	40	-	46/60	Artificial	-	Dry	-	-	~20	-	26	49
		Grassland (*Bromus tectorum*)	Unb	20.2	40	-	92/30	Artificial	-	Dry	-	-	~20	-	53	331
Craddock and Pearse (1938)	Cold desert (Great Basin, Idaho)	Grassland (*Lupinus* spp./*Stipa lettermanii*)	Unb	20.2	30	-	46/60	Artificial	-	Dry	-	-	~25	-	44	274
		Grassland (*Lupinus* spp./*Stipa lettermanii*)	Unb	20.2	30	-	92/30	Artificial	-	Dry	-	-	~25	-	57	511
		Grassland (*Lupinus* spp./*Stipa lettermanii*)	Unb	20.2	40	-	46/60	Artificial	-	Dry	-	-	~25	-	44	253
		Grassland (*Lupinus* spp./*Stipa lettermanii*)	Unb	20.2	40	-	92/30	Artificial	-	Dry	-	-	~25	-	55	468
Johnson and Gordon (1988)	Cold desert (Great Basin, Idaho)	Shrub coppice (*Artemisia tridentata*)	Unb	~0.9	6	-	127/30	Artificial	-	Dry	15	41	85	25	15	5
		Interspace	Unb	~0.9	6	-	127/30	Artificial	-	Dry	29	4	71	13	48	39
Johnson and Blackburn (1989)[d]	Cold desert (Great Basin, Idaho)	Shrub steppe (*A. tridentata*)	Unb	32.6	10	-	64/60	Artificial	-	Dry	16	39	84	-	2	5
		Shrub steppe (*A. tridentata*)	Unb	32.6	6	-	64/60	Artificial	-	Dry	19	12	81	-	11	11
		Shrub steppe (*A. tridentata*)	Unb	32.6	9	-	64/60	Artificial	-	Dry	45	16	55	-	17	15
Pierson et al. (1994b)	Cold desert (Great Basin, Idaho)	Shrub coppice (*A. tridentata*)	Unb	~0.1	7	-	67/60	Artificial	-	4	0	100	100	-	3	1
		Vegetated interspace	Unb	~0.1	7	-	67/60	Artificial	-	4	4	-	96	-	23	23
		Bare interspace	Unb	~0.1	7	-	67/60	Artificial	-	2	94	-	6	-	65	381
Reid et al. (1999)	Temperate steppe mountains (Pajarito Plateau, New Mexico)	Tree coppice (*Pinus edulis* or *Juniperus monosperma*)	Unb	1.0	6-12	-	Variable[e]	Natural	-	-	0	-	100	-	8[e]	39[e]
		Vegetated interspace	Unb	~2	6-12	-	Variable[e]	Natural	-	-	34	-	66	-	25[e]	298[e]
		Bare interspace	Unb	~2	6-12	-	Variable[e]	Natural	-	-	84	-	16	-	37[e]	1,007[e]

Table 3. Continued

Study	Climatic region (location)	Microsite/plant community	Treatment/burn severity	Plot size (m²)	Slope (%)	Time postfire (mth)	Rain rate (mm h⁻¹)/duration (min)	Rain type	WDPT (s)[a]	Soil water (%)[b]/conditions	Bare soil (%)	Canopy cover (%)	Ground cover (%)	Surface roughness (mm)	Runoff coef. (%)[c]	Sed. yield (g m⁻²)
Benavides-Solorio and MacDonald (2001, 2002)[f]	Temperate steppe mountains (Colorado Front Range, Colorado)	Dry forest (*Pinus ponderosa*)	Low-Unb	1.0	20-25	1-3	79/60	Artificial	~65	2	1	-	99	-	55	80
			Mod	1.0	20-35	1-3	79/60	Artificial	~50	2	12	-	88	-	58	179
			High	1.0	20-45	1-3	79/60	Artificial	~60	2	77	-	23	-	66	1,280
Johansen et al. (2001)	Temperate steppe mountains (Pajarito Plateau, New Mexico)	Dry forest (*P. ponderosa*)	Unb	32.5	5	3	60/60[g]	Artificial	-	~5	48		52	-	23	36
			High	32.5	7	3	60/60[g]	Artificial	-	~5	74		26	-	45	912
Pierson et al. (2001a, 2008a)	Cold desert (Great Basin, Nevada)	Shrub coppice (*A. tridentata*)	Unb	0.5	30-40	1	85/60	Artificial	200	7	1	100	99	-	30	12
			High	0.5	30-40	1	85/60	Artificial	102	1	99	1	1	-	37	41
		Interspace	Unb	0.5	30-40	1	85/60	Artificial	220	5	6	74	94	-	49	24
			High	0.5	30-40	1	85/60	Artificial	97	1	99	4	1	-	30	21
Pierson et al. (2002b)[h]	Cold desert (Great Basin, Idaho)	Shrub coppice (*A. tridentata*)	Unb	0.5	35-60	12	67/60	Artificial	-	~14	7	88	93	18	11	2
			Mod	0.5	35-60	12	67/60	Artificial	-	~5	97	11	3	12	34	30
			High	0.5	35-60	12	67/60	Artificial	-	~5	98	13	2	12	37	22
		Interspace	Unb	0.5	35-60	12	67/60	Artificial	-	~14	89	18	12	18	24	4
			Mod	0.5	35-60	12	67/60	Artificial	-	~5	95	16	5	12	26	12
			High	0.5	35-60	12	67/60	Artificial	-	~5	99	5	1	10	49	148
O'Dea and Guertin, (2003)	Hot desert (Sonoran Desert, Arizona)	Grassland	Unb	30	1-3	-[j]	Variable	Natural	-	Dry	-	21	-	-	6	1,700[i]
			Mod	30	1-3	-[j]	Variable	Natural	-[j]	Dry	-	11	-	-	6	2,800[l]
		Grassland	Unb	30	1-3	-[k]	Variable	Natural	-	Dry	-	37	-	-	14	1,800[k]
			Mod	30	1-3	-[k]	Variable	Natural	-[j]	Dry	-	29	-	-	14	2,200[k]
Pierson et al. (2007a)	Cold desert (Great Basin, Oregon)	Shrub-interspace zone	Unb	32.5	19	-	55/60	Artificial	-	-	84	6	16	24	25	118
		Shrub-interspace zone – trees removed[l]	Unb	32.5	19	-	55/60	Artificial	-	-	64	23	36	36	2	1
Pierson et al. (2007b)[m]	Cold desert (Great Basin, Idaho)	Grassland (*B. tectorum*)	Unb	32.5	2	-	64/60	Artificial	-	4	34	54	66	1	2	1

Table 3. Continued

Study	Climatic region (location)	Microsite/plant community	Treatment/ burn severity	Plot size (m²)	Slope (%)	Time postfire (mth)	Rain rate (mm h⁻¹)/ duration (min)	Rain type	WDPT (s)[a]	Soil water (%)[b] conditions	Bare soil (%)	Canopy cover (%)	Ground cover (%)	Surface roughness (mm)	Runoff coef. (%)[c]	Sed. yield (g m⁻²)
Michaelides et al. (2009)	Hot desert (Chihuahuan Desert, New Mexico)	Shrub coppice (*Larrea tridentata*)	Unb	1.5	7	-	125/16-30	Artificial	-	7	79	65	21	-	55	97
		Interspace	Unb	1.5	3	-	125/16-30	Artificial	-	8	89	1	11	-	49	122
		Grassland (*Bouteloua eriopoda*)	Unb	1.5	2	-	125/16-30	Artificial	-	12	96	58	4	-	49	89
Pierson et al. (2009)	Cold desert (Great Basin, Idaho)	Shrub coppice (*A. tridentata*)	Unb	0.5	35-50	1	85/60	Artificial	286	7	2	84	98	34	39	17
			Mod-High	0.5	35-50	1	85/60	Artificial	261	3	42	10	58	11	76	183
		Interspace	Unb	0.5	35-50	1	85/60	Artificial	110	3	25	31	75	18	63	195
			Mod-High	0.5	35-50	1	85/60	Artificial	117	4	84	0	16	11	55	705
		Shrub-interspace zone	Unb	32.5	35-50	1	85/60	Artificial	-	2	24	57	76	21	4	8
			Mod-High	32.5	35-50	1	85/60	Artificial	208	4	76	0	24	11	27	988
Pierson et al. (2010)	Cold desert (Great Basin, Nevada)	Shrub-interspace zone (intercanopy)	Unb	13	9	-	64/45	Artificial	-	6	26[n]	39	74	17	24	45
		Tree coppice (*P. monophylla* or *J. osteosperma*)	Unb	13	9	-	64/45	Artificial	-	11	3	27[o]	97	20	3	18
		Shrub-interspace zone (intercanopy)	Unb	13	9	-	102/45	Artificial	-	wet	26[n]	39	74	17	47	222
		Tree coppice (*P. monophylla* or *J. osteosperma*)	Unb	13	9	-	102/45	Artificial	-	wet	3	27[o]	97	20	5	36
Pierson et al. (2010)	Cold desert (Great Basin, Utah)	Shrub-interspace zone (intercanopy)	Unb	13	14	-	64/45	Artificial	-	7	29[p]	19	71	26	10	37
		Tree coppice (*J. osteosperma*)	Unb	13	15	-	64/45	Artificial	-	9	8	21[o]	92	31	2	13
		Shrub-interspace zone (intercanopy)	Unb	13	14	-	102/45	Artificial	-	wet	29[p]	19	71	26	44	296
		Tree coppice (*J. osteosperma*)	Unb	13	15	-	102/45	Artificial	-	wet	8	21[o]	92	31	10	66

Table 3. Continued

Study	Climatic region (location)	Microsite/plant community	Treatment/ burn severity	Plot size (m²)	Slope (%)	Time postfire (mth)	Rain rate (mm h⁻¹) /duration (min)	Rain type	WDPT (s)[a]	Soil water (%)[b]/ conditions	Bare soil (%)	Canopy cover (%)	Ground cover (%)	Surface roughness (mm)	Runoff coef. (%)[c]	Sed. yield (g m⁻²)
Pierson et al. (2013), Williams et al. (2014a)	Cold desert (Great Basin, Idaho)	Shrub coppice (*Artemisia* spp.)	Unb	0.5	20	-	102/45	Artificial	<5	wet	25	117	75	13	20	6
			High	0.5	18	12	102/45	Artificial	11	wet	57	21	43	9	23	143
		Interspace	Unb	0.5	14	-	102/45	Artificial	<5	wet	46^q	20	54	9	63	36
			High	0.5	16	12	102/45	Artificial	<5	wet	49^q	21	51	8	51	135
		Tree coppice (*J. occidentalis*)	Unb	0.5	21	-	102/45	Artificial	42	wet	0	17^0	100	12	23	6
			High	0.5	17	12	102/45	Artificial	54	wet	50	5^0	50	8	58	206
		Shrub-interspace zone	Unb	13	19	-	102/45	Artificial	-	wet	28^r	18	72	16	50	272
			High	13	16	12	102/45	Artificial	-	wet	39^r	32	61	17	50	572
		Tree coppice (*J. occidentalis*)	Unb	13	16	-	102/45	Artificial	-	wet	7	26^0	93	23	13	48
			High	13	18	12	102/45	Artificial	-	wet	25	15^0	75	21	58	1,083
Pierson et al. (2014)	Cold desert (Great Basin, Nevada)	Shrub coppice (*Artemisia* spp.)	Unb	0.5	11	-	102/45	Artificial	<5	wet	12	93	88	14	4	6
			High	0.5	13	10	102/45	Artificial	<5	wet	49	53	51	8	10	48
		Interspace	Unb	0.5	9	-	102/45	Artificial	<5	wet	44^s	33	56	9	41	23
			High	0.5	10	10	102/45	Artificial	<5	wet	49^s	30	51	8	46	41
		Tree coppice (*P. monophylla* or *J. osteosperma*)	Unb	0.5	12	-	102/45	Artificial	48	wet	0	7^t	100	12	0	0
			High	0.5	15	10	102/45	Artificial	65	wet	10	4^t	91	12	28	46
Pierson et al. (2014)	Cold desert (Great Basin, Utah)	Shrub coppice (*Artemisia* spp.)	Unb	0.5	17	-	102/45	Artificial	<5	wet	18	69	82	13	8	33
			High	0.5	16	10	102/45	Artificial	<5	wet	44	28	56	11	29	220
		Interspace	Unb	0.5	19	-	102/45	Artificial	<5	wet	52^u	19	48	11	56	233
			High	0.5	16	10	102/45	Artificial	<5	wet	38^u	7	62	9	64	351
		Tree coppice (*J. osteosperma*)	Unb	0.5	20	-	102/45	Artificial	88	wet	7	22^t	93	12	22	98
			High	0.5	21	10	102/45	Artificial	125	wet	9	2^t	91	12	18	294

[a] Water drop penetration time (WDPT) is an indicator of strength of soil water repellency as follows: <5 s wettable, 5 to 60 s slightly repellent, 60 to 600 s strongly repellent (Bisdom et al. 1993).

[b] Measured near the soil surface (<5 cm depth).

[c] Runoff coefficient is equal to cumulative runoff divided by cumulative rainfall applied. Value is multiplied by 100 to obtain percentage.

[d] Data shown are for dry run experiments at Coyote Butte, Nancy, and Summit sites, respectively by row.

[e] Runoff and erosion following natural rainfall events were monitored over a 14-month period.

[f] Data presented for Bobcat Fire only.

[g] Rainfall applied for 60 min under dry conditions, followed by 24-h hiatus, 30 min of rainfall, 30-min hiatus, and 30 min rainfall. Total rain applied was 120 mm.

(footnotes continued on next page)

h Data presented from south-facing slopes solely.

i Cumulative runoff and sediment yield for period of 1 July 1998 to 1 October 1998 resulting from natural rainfall events (100 mm). Fire was in May 1998.

j Informal water drop tests showed no postfire soil water repellency at soil surface (O'Dea and Guertin 2003).

k Cumulative runoff and sediment yield for period of 1 July 1999 to 1 October 1999 resulting from natural rainfall events (106 mm). Fire was in May 1998.

l *J. occidentalis* trees removed from site by chainsaw cutting 10 years before rainfall simulation experiments.

m This is a multiple year study; means shown for soil, cover, runoff, and sediment yield are average of 3 simulation years.

n Bare ground was about 64 percent (38 percent rock cover and 26 percent bare soil).

o Trees removed from plot by chainsaw immediately before simulations.

p Bare ground was about 79 percent (50 percent rock cover and 29 percent bare soil).

q Bare ground was about 90 percent across unburned (42 percent rock cover and 46 percent bare soil) and burned (43 percent rock cover and 49 percent bare soil) plots.

r Bare ground was about 88 percent (60 percent rock cover and 28 percent bare soil) for unburned plots and 84 percent (45 percent rock cover and 39 percent bare soil) for burned plots.

s Bare ground was about 73 percent (29 percent rock cover and 44 percent bare soil) for unburned and 87 percent for burned (38 percent rock cover and 49 percent bare soil) plots.

t Trees removed from plot about 12 months before simulation as part of earlier study (Pierson et al. 2010).

u Bare ground ≥90 percent across unburned (38 percent rock cover and 52 percent bare soil) and burned (56 percent rock cover and 38 percent bare soil) plots.

development of infiltration-inhibiting surface crusts (Branson et al. 1981; Puigdefábregas et al. 1999). Decreasing infiltration and increasing runoff with increasing expanse of bare or vesicular surfaces are well documented in literature (Branson and Owen 1970; Blackburn et al. 1992; Pierson et al. 1994a; Abrahams et al. 1995; Parsons et al. 1996; Wilcox et al. 1996; Schlesinger et al. 1999, 2000; Pierson et al. 2002a, 2007a, 2010, 2014; Williams et al. 2014a,b). The effects of rock cover (more than 2 mm) depend on the size, amount, and embeddedness of the rocks (Wilcox et al. 1988; Poesen et al. 1990; Poesen and Ingelmo-Sanchez 1992). Infiltration is generally positively correlated with rocks lying on top of the soil matrix due to increased surface roughness and greater porosity and aggregation around rocks; surface rock extends time to ponding and runoff, increasing time for infiltration (Poesen et al. 1990; Abrahams and Parsons 1994; Poesen et al. 1994; Valentin 1994; Cerdà 2001; Martínez-Zavala and Jordán 2008). Infiltration is negatively correlated with embedded rock cover due to a decrease in nonabsorbing area. Wilcox et al. (1988), Abrahams and Parsons (1991a), and Pierson et al. (2010, 2013) reported negative correlations between rock cover and infiltration in interspace areas, but did not explicitly evaluate embeddedness. The studies by Wilcox et al. (1988) and Abrahams and Parsons (1991a) indicate interspace areas occurred in swales and were more compacted and crusted than coppice areas. Wilcox et al. (1988), Abrahams and Parsons (1991a), and Pierson et al. (2010) suggested that the negative correlations were not exclusively associated with rock cover; instead, the relationship was due to co-occurring low infiltration rates of the bare interspace areas and extensive rock cover. Wilcox et al. (1988) further indicated infiltration was negatively correlated with smaller size rock cover (2 to 12 mm) and positively correlated with rock cover of intermediate sizes (26 to 150 mm). Tromble et al. (1974) also reported a negative relationship in infiltration and small-size rock cover (less than 10 mm). These studies suggest rock cover can facilitate infiltration and that negative effects of rock cover on infiltration most likely occur when smaller rocks dominate and the rock cover is embedded rather than freely lying atop the soil surface (Brakensiek and Rawls 1994).

The vegetation- and soils-driven hydrologic heterogeneity (Puigdefábregas 2005) of rangeland ecosystems creates a mosaic of runoff source and sink areas at the hillslope scale (figs. 15 and 16) (Schlesinger et al. 1990; Wilcox and Breshears 1995; Schlesinger et al. 1996; Bergkamp 1998; Puigdefábregas et al. 1999; Reid et al. 1999; Wilcox et al. 2003a; Ludwig et al. 2005; Turnbull et al. 2012; Williams et al. 2014a). The timing and quantity of overland flow generation on rangeland sites is strongly correlated to the quantity and arrangement of canopy and ground cover and bare interspace (Branson et al. 1981). Runoff occurs more rapidly after the onset of rainfall in sparsely vegetated interspaces than in vegetated interspace and coppice locations. These relationships, along with rainfall distribution, are responsible for spatial variability in runoff generation during intermediate storm events, but may be dampened by high-intensity or long-duration rainfall, creating more uniform runoff (Reid et al. 1999; Puigdefábregas 2005).

The hydrologic connectivity and downslope surface hydraulic conductivity dictate the progression or decay of surface runoff with increased slope length (Abrahams et al. 1991; Dunne et al. 1991; Pierson et al. 1994b; Wilcox 1994; Abrahams et al. 1995; Cerdà 1997; Bergkamp 1998; Davenport et al. 1998; Puigdefábregas et al. 1999; Reid et al. 1999; Wainwright et al. 2000; Wilcox et al. 2003a; Puigdefábregas 2005). Well-connected flowpaths develop in consecutive source areas where overland flow is routed

Figure 16—Concentrated flow formed in interspace areas during a high-intensity rainfall on a shrub steppe site in the Reynolds Creek Experimental Watershed, Idaho. The lack of runoff from shrub microsites clearly demonstrates the commonly observed hydrologic stability observed for areas underneath shrub or tree canopies on rangeland sites (photo: USDA Agricultural Research Service, Northwest Watershed Research Center).

around topographically elevated coppice mounds, grass clumps, or roughness elements (fig. 16) (Emmett 1970, 1978; Dunne et al. 1991; Seyfried 1991; Parsons et al. 1992; Thornes 1994; Abrahams et al. 1995; Wilcox and Breshears 1995; Parsons et al. 1996; Schlesinger et al. 1996, 1999; Wilcox et al. 2003a). Where concentrated, these flowpaths transfer large volumes of water laterally at greater overland flow depths and velocities than occur in sheetflow processes (Abrahams et al. 1995; Parsons et al. 1996; Pierson et al. 2007a). These effects are amplified on steep slopes (Al-Hamdan et al. 2013), although infiltration and slope steepness have been shown to have a positive correlation on some rangelands (Wilcox et al. 1988). The interception of flowpaths (mostly due to ponding behind coppices or topographic features) and subsequent re-infiltration (runon) in coppice or vegetated hydrologic sinks are thought to stimulate biological productivity and further facilitate coppice-interspace structure (Schlesinger et al. 1990; Joffre and Rambal 1993; Dunkerley and Brown 1995; Ludwig and Tongway 1995; Wilcox and Breshears 1995; Breshears et al. 1997b; Tongway and Ludwig 1997; Bergkamp 1998; Puigdefábregas et al. 1999; Reid et al. 1999; Bhark and Small 2003; Wilcox et al. 2003a; Ludwig et al. 2005). Abrahams et al. (1995) and Parsons et al. (1996) observed that fine-scale vegetative heterogeneity of a hydrologically stable grassland facilitated runon processes and that the coarseness of vegetative structure in a degraded shrubland community amplified runoff with increasing slope length (up to 35 m). Reid et al. (1999) estimated that runon from bare interspaces (sources) to vegetated interspace areas (sinks) in a hydrologically stable twoneedle pinyon/oneseed juniper woodland accounted for 12 percent of precipitation over the course of 77 rainfall events.

These studies illustrate that a coarsely arranged source-sink structure, as observed on degraded sites, potentially generates and releases more surface runoff than a finely structured source-sink community (Schlesinger et al. 1990; Abrahams et al. 1995; Parsons

et al. 1996; Wilcox et al. 1996; Davenport et al. 1998; Bhark and Small 2003). Studies by Abrahams et al. (1995), Parsons et al. (1996), Wainwright et al. (2000), Michaelides et al. (2009), and Turnbull et al. (2010b, 2012) provide comparative examples of these relationships for fine (grassland) versus coarsely arranged (shrubland) rangeland communities in southern Arizona. Pierson et al. (2010, 2013) and Williams et al. (2014a) present examples of similar relationships following conifer encroachment into Great Basin shrub steppe.

Plant Use of Subsurface Water and Its Influence on Runoff

A positive feedback exists between subsurface water acquisition, plant community structure, and infiltration, and the maintenance of this arrangement (Richards and Caldwell 1987; Dawson 1993; Burgess et al. 2001; Ludwig et al. 2003; Ryel et al. 2003, 2004; Muñoz et al. 2008; Scott et al. 2008). Herbaceous and woody plants differ in their ability to obtain the limited soil water on most rangeland sites (Walter 1971; Breshears and Barnes 1999; Schenk and Jackson 2002). Rangeland plant community density and structure therefore reflect the ability of the plants to obtain and efficiently use available soil water (Caldwell 1985; Dawson 1993; Schenk and Jackson 2002). Walter (1971) proposed a two-layered model of soil water use in water-limited ecosystems based on rooting depth partitioning among herbaceous and woody species. The model suggests that herbaceous plants primarily extract soil water from upper soil layers and that woody plants have the sole access to deeper soil water.

Subsequent research has shown woody plants vary in the depth at which they extract soil water (Peláez et al. 1994; Montaña et al. 1995; Breshears et al. 1997a; Schenk and Jackson 2002) and that they are capable of competing laterally with herbaceous species for shallow soil water in interspace areas (Caldwell et al. 1985; Ansley et al. 1991; Peláez et al. 1994; Montaña et al. 1995; Breshears et al. 1997a; Breshears and Barnes 1999). Schenk and Jackson (2002) summarized rooting depths for herbaceous and woody plants over a wide range of precipitation regimes. They observed that differences in rooting depths between herbaceous and woody species tend to decrease with increasing precipitation. This suggests that as water becomes limited, a more distinct vertical separation in water use by woody and herbaceous plants emerges (Schenk and Jackson 2002). Breshears and Barnes (1999) pointed out that horizontal as well as vertical gradients exist due to woody plant lateral acquisition of soil water from interspace areas and that decreased near-surface soil water favors woody plant recruitment.

Similar processes for horizontal acquisitions of surface water via preferred infiltration or runon have been proposed to explain maintenance of vegetated islands (Joffre and Rambal 1993; Dunkerley and Brown 1995; Wilcox and Breshears 1995; Ludwig et al. 1997; Bergkamp 1998; Puigdefábregas et al. 1999; Bhark and Small 2003; Wilcox et al. 2003a; Ludwig et al. 2005) and desertification of water-limited landscapes (Schlesinger et al. 1990). Prolonged periods of dry soil conditions coarsen the vegetative structure in favor of woody plants and shrub/tree islands whereas wet periods facilitate a more vertically and horizontally heterogeneous community of herbaceous and woody plants (Schlesinger et al. 1990). Coarsening of the plant community (such as transitions from grassland to shrubland and shrubland to woodland) through drought or disturbance commonly increases bare ground area, hydrologic connectivity of runoff source areas, and

surface runoff from point to hillslope scales (Abrahams et al. 1995; Parsons et al. 1996; Wilcox et al. 1996; Davenport et al. 1998; Pierson et al. 2010; Turnbull et al. 2010a,b; Pierson et al. 2011; Turnbull et al. 2012; Williams et al. 2014a, 2016a).

Plant community structure may also be influenced by plant-specific belowground water conservation strategies that vertically redistribute (via hydraulic redistribution) soil water (Caldwell et al. 1998; Horton and Hart 1998; Jackson et al. 2000; Meinzer et al. 2001). Hydraulic redistribution is the passive transfer of soil water through roots upward as hydraulic lift (Richards and Caldwell 1987; Dawson 1993; Wan et al. 1993; Emerman and Dawson 1996; Caldwell et al. 1998; Horton and Hart 1998; Yoder and Nowak 1999a; Mendel et al. 2002; Ludwig et al. 2003; Zou et al. 2005; Muñoz et al. 2008) or downward as hydraulic descent (Burgess et al. 1998; Schulze et al. 1998; Smith et al. 1999; Jackson et al. 2000; Burgess et al. 2001; Leffler et al. 2002; Ryel et al. 2002; Hultine et al. 2003a,b; Ryel et al. 2003; Hultine et al. 2004; Ryel et al. 2004; Leffler et al. 2005; Scott et al. 2008) from wetter to drier soil layers along a gradient in water potential. Lateral redistribution has also been observed (Brooks et al. 2002; Burgess and Bleby 2006), but its occurrence is less documented than vertical transfers. Hydraulic lift has been documented in more than 50 woody taxa and herbaceous species (Jackson et al. 2000), but is most common in deeper-rooted shrubs and trees. During dry periods, plants undergoing hydraulic lift absorb soil water from moist deep soil layers during evening hours when transpiration demands are low. The absorbed water is then transferred upwards and released via roots to the drier near-surface soil during the night. The released water is subsequently reabsorbed during the next day to meet daily transpiration demands.

Burgess et al. (1998) demonstrated that the reverse of hydraulic lift also occurs. They found that as soils wet up following the dry season, the roots of silkoak (*Grevillea robusta* A. Cunn. ex R. Br.) and river redgum (*Eucalyptus camaldulensis* Dehnh.) redistributed soil water from the wetter near-surface to drier soil pockets at depth. They termed the process hydraulic redistribution. In other studies, Ryel et al. (2003, 2004) discovered that pulse rain events in a stand of big sagebrush (*A. tridentata* Nutt.) delivered rain at different depths (downward distribution) simultaneously rather than sequentially from upper to lower soil layers. This process might be expected with macropore or preferential flow; however, the arrival times at the different depths were simultaneous over a few days rather than hours as commonly reported for macropore and other preferred flowpaths. Hydraulic descent has also been reported for other water-limited plants including a cheatgrass (*Bromus tectorum* L.) monoculture in Utah, (Leffler et al. 2005), velvet mesquite (*Prosopis velutina* Woot.) in southern Arizona (Hultine et al. 2004, Scott et al. 2008), Utah juniper (*J. osteosperma* (Torr.) Little) in northern Utah (Leffler et al. 2002), and Arizona walnut (*Juglans major* Torr.) in southeastern Arizona (Hultine et al. 2003a).

The decreased water stress associated with hydraulic redistribution provides dryland vegetation numerous ecological benefits thought to increase ecosystem primary productivity. These include enhanced water-use efficiency and transpiration (Richards and Caldwell 1987; Caldwell and Richards 1989; Dawson 1993, 1996; Emerman and Dawson 1996; Caldwell et al. 1998; Brooks et al. 2002; Ryel et al. 2002; Hultine et al. 2004; Ryel et al. 2004; Muñoz et al. 2008; Scott et al. 2008), greater fine-root longevity (Richards and Caldwell 1987; Dawson 1993; Meinzer et al. 2004), increased microbial activity and nutrient acquisition (Caldwell et al. 1998; Dawson 1993, 1996; McCulley et al. 2004), prolonged symbiotic mychorrizal associations during drought (Richards and Caldwell

1987; Caldwell et al. 1998; Horton and Hart 1998; Querejeta et al. 2003, 2007), reduced carbon consumption (Dawson 1993; Caldwell et al. 1998), extension of the growing season (Ryel et al. 2004; Muñoz et al. 2008; Scott et al. 2008), and decreased competition for limited water resources (Smith et al. 1999). The primary benefit is, of course, water conservation.

Few long-term assessments of the ecosystem benefits associated with hydraulic redistribution exist (Scott et al. 2008). In a 2-year study, Scott et al. (2008) found that velvet mesquite on a southern Arizona rangeland redistributed soil water from the near surface to deep soil locations throughout the year, including the dormant season. Downward redistributed winter-season precipitation allowed trees to transpire more during the dry pre-monsoon period. Hydraulic descent of monsoonal summer rainfall extended the growing season and allowed for greater photosynthesis and plant productivity during the seasonal drought. Similar results were reported in another study of velvet mesquite in southeastern Arizona (Hultine et al. 2004).

Plants neighboring hydraulic-lifting species may benefit from soil water redistribution as well (Caldwell and Richards 1989; Dawson 1993; Yoder and Nowak 1999a; Smith et al. 1999, Brooks et al. 2002; Leffler et al. 2005; Zou et al. 2005), but these benefits are probably limited by plant competition and the dry surface soil moisture conditions in which hydraulic lifting occurs (Caldwell et al. 1998; Ludwig et al. 2004; Muñoz et al. 2008). Dawson (1993) reported that hydraulic lift by sugar maple (*Acer saccharum*) in a mesic forest influenced soil moisture conditions up to 5 m from the tree base. Within this distance, plants neighboring sugar maple used 3 to 60 percent of the lifted water. Neighboring plants that used a high percentage of lifted water exhibited increased water-use efficiency and greater aboveground growth (Dawson 1993). Yoder and Nowak (1999a) suggested that hydraulic lift by Mojave yucca (*Yucca schidigera*) plants in the water-limited Mojave Desert (located in parts of California, Nevada, Utah, and Arizona), provided daytime near-surface soil water for neighboring creosote (*Larrea tridentata*), rough joinfir (*Ephedra nevadensis*), burrobush (*Ambrosia dumosa*), pale desert-thorn (*Lycium pallidum*), and Indian ricegrass (*Achnatherum hymenoides*) plants. All six species exhibited hydraulic lift, but only Mojave yucca (a crassulacean acid metabolism, or CAM, species) lifted water during daytime hours and transpired at night.

Such benefits are not always observed for neighboring plants (Caldwell et al. 1998; Brooks et al. 2002; Ludwig et al. 2004; Muñoz et al. 2008). In an east African savanna, Ludwig et al. (2004) determined that any benefit of hydraulic lift by umbrella thorn (*Acacia tortilis*) observed for neighboring plants was overwhelmed by near-surface competition with the water-lifting trees. Ludwig et al. (2004) suggested that the trees outcompeted the grasses for lifted water and that grasses were then limited by the lack of additional water availability under the more xeric conditions relative to those in the Dawson (1993) study. Muñoz et al. (2008) also reported that near-surface water use by xeric community shrubs hydraulically lifting soil water mitigated potential losses to neighboring plants. Ishikawa and Bledsoe (2000) reported that hydraulic lifting by blue oak (*Q. douglasii*) occurred too late in the growing season for neighboring grasses to benefit.

Plants in dry climates with shallow and fibrous roots generally complete seasonal physiological processes during periods of high near-surface soil water content. As the near-surface environment dries out, shallow-rooted species die off and competition for

near-surface resources is reduced. Deep-rooted species delaying hydraulic lift to the near-surface can then extend the growing season and, by wetting the near-surface, improve near-surface nutrient acquisition during periods of the year when competition is low (Caldwell et al. 1998). Benefits to neighboring plants then appear to vary with precipitation regime, type of plant community, and the timing of hydraulic redistribution.

The volume of water redistributed is largely a function of available soil water and the daily evapotranspiration requirements. In a semiarid climate, Ryel et al. (2003) estimated that 74 percent of precipitation from a 36 mm event and 100 percent of precipitation from small (less than 8 mm) rainfall events that infiltrated 30 to 150 cm into the soil profile resulted from hydraulic redistribution by big sagebrush roots. Richards and Caldwell (1987) reported mountain big sagebrush (*A. tridentata* Nutt. ssp. *vaseyana* (Rydb.) Beetle) hydraulically lifted one-third of its daily evapotranspiration demand. Caldwell and Richards (1989) found that artificial suppression of hydraulic lift reduced mountain big sagebrush daily transpiration by 25 to 50 percent. Hultine et al. (2003a) reported daily hydraulic descent amounted to 10 to 60 percent of daily transpiration for Arizona walnut. Hultine et al. (2004) estimated that diurnal hydraulic descent rates of velvet mesquite during the winter dormant season were 70 percent of that during the growing season following monsoon rainfall.

Leffler et al. (2005) found that about 6 percent of soil water at 10 to 20 cm depth underneath a cheatgrass monoculture was hydraulically lifted by cheatgrass during flowering and seed set. In the same study, senesced cheatgrass in a greenhouse-stored pot lifted 17 percent of soil water measured in an upper soil layer. Brooks et al. (2002) determined 28 to 35 percent of water removed daily from the upper 2 m of the soil profile in coniferous forests of moist Douglas-fir (*Pseudotsuga menziesii* [Mirb.] Franco) in Washington and dry ponderosa pine (*Pinus ponderosa* Dougl. ex Laws) in Oregon, was replaced by nocturnal hydraulic redistribution. In a mesic climate, Emerman and Dawson (1996) found that an individual sugar maple tree was capable of lifting 100 L day^{-1} and that hydraulic lift provided 25 percent of tree daily water use. Dawson (1996) reported sugar maple was capable of lifting 25 percent of daily transpiration demand. The studies cited above are only a sample from literature to demonstrate that hydraulic redistribution occurs across a range of plant communities (herbaceous and woody) and climate regimes and may have a substantial impact on ecosystem water balances (Meinzer et al. 2001). Hydraulic redistribution may account for as much as 70 percent of daily transpiration; however, most values are in the 20 to 35 percent range depending on plant type, antecedent moisture conditions, time of year, and evapotranspiration demand.

Effects of Soil Water Repellency on Runoff Generation

Soil water repellency is a naturally occurring soil condition that inhibits infiltration. Its occurrence has been well documented on shrubland, chaparral, woodland, and semi-arid forest ecosystems (Meeuwig 1971; Scholl 1971, 1975; DeBano 1981, 1991; Doerr et al. 1996, 2000; Jaramillo et al. 2000; DeBano 2001; Huffman et al. 2001; Hubbert et al. 2006; Lebron et al. 2007; Verheijen and Cammeraat 2007; Woods et al. 2007; Madsen et al. 2008; Pierson et al. 2008b; Doerr et al. 2009; Pierson et al. 2009, 2010; Robinson et al. 2010; Bodí et al. 2013; Pierson et al. 2013, 2014; Williams et al. 2014a). Water-repellent soils form by the coating of particles with hydrophobic compounds leached

from organic matter accumulations, microbial by-products, or fungal growth under litter and duff (Savage et al. 1972; Imeson et al. 1992; Bisdom et al. 1993; Doerr et al. 2000). The strength of soil water repellency and its influence on infiltration are a function of the quantity and type of overlying vegetation, soil texture, and soil water content (Burcar et al. 1994; Dekker and Ritsema 1994; Bauters et al. 2000; Doerr and Thomas 2000; Shakesby et al. 2000; Dekker et al. 2001; Huffman et al. 2001; MacDonald and Huffman 2004; Verheijen and Cammeraat 2007; Madsen et al. 2008; Pierson et al. 2008b, 2009, 2010). The type and quantity of vegetation dictate the amount and type of hydrophobic compounds potentially available. Coarse-textured soils generally are more susceptible to soil water repellency than fine-textured soils due to their greater particle surface area (DeBano 1991, Bisdom et al. 1993, Huffman et al. 2001); however, recent research has demonstrated that strong soil water repellency can occur in fine-textured soils (Doerr et al. 2000, 2006, 2009). Doerr et al. (2000, 2009) provide a review of occurrence, causes, hydrologic and erosional effects, and measurement methods of soil water repellency.

The strength and persistence of soil water repellency is highly variable in time and space (DeBano 1971; Witter et al. 1991; Shakesby et al. 1993; Dekker and Ritsema 1994; Doerr and Thomas 2000; Dekker et al. 2001; Huffman et al. 2001; MacDonald and Huffman 2004; Leighton-Boyce et al. 2005; Verheijen and Cammeraat 2007; Woods et al. 2007; Madsen et al. 2008; Pierson et al. 2008b, 2009, 2010). Soil water repellency for a particular soil may be present under dry conditions, decrease with soil wetting, and reappear with soil drying (Shakesby et al. 1993; Doerr et al. 2000). Dekker et al. (2001) demonstrated that critical soil-water thresholds demarcate wettable and water-repellent soil conditions. Doerr et al. (2009) suggest from literature that the critical threshold ranges from 5 percent for organic dune sands to more than 30 percent for fine-textured soils. Huffman et al. (2001) reported that water repellency in sandy loam soils at semiarid-forested sites in Colorado became wettable at soil water contents of 12 to 25 percent. Doerr and Thomas (2000) reported that temporal variability in soil water repellency was associated with seasonal rainfall patterns, biological productivity, and wetting and drying regimes. Pierson et al. (2008b, 2009) found that soil water repellency and the magnitude of its influence on infiltration and runoff exhibited significant annual variability at multiple steeply sloped mountain big sagebrush sites in the Inland Northwest, but the study did not explicitly track soil moisture patterns (fig. 17).

In addition to temporal variance, the strength of soil water repellency may be spatially variable (horizontally and vertically), owing to its presence mostly under or immediately adjacent to canopy- and litter-covered areas and spatial soil-moisture gradients (Imeson et al. 1992; Ritsema and Dekker 1994; Dekker et al. 2001; Verheijen and Cammeraat 2007; Woods et al. 2007; Madsen et al. 2008; Pierson et al. 2008b, 2009, 2010, 2013, 2014; Williams et al. 2014a). On unburned sites, soil water repellency is commonly stronger at the soil surface and degrades with depth below the mineral surface (Huffman et al. 2001; Leighton-Boyce et al. 2007; Pierson et al. 2008b, 2009, 2010). The effects of fire on the occurrence and hydrologic impacts of soil water repellency are discussed in Section 4, *Exacerbation, Alteration, and Formation of Soil Water Repellency*.

Soil water repellency facilitates runoff initiation either by inhibiting infiltration at the surface (infiltration-excess runoff) or causing saturation of a shallow soil layer (saturation-excess runoff) immediately overlying a water-repellent zone (Doerr et al. 2000). In either case, runoff initiation may occur rapidly, but infiltration generally increases as
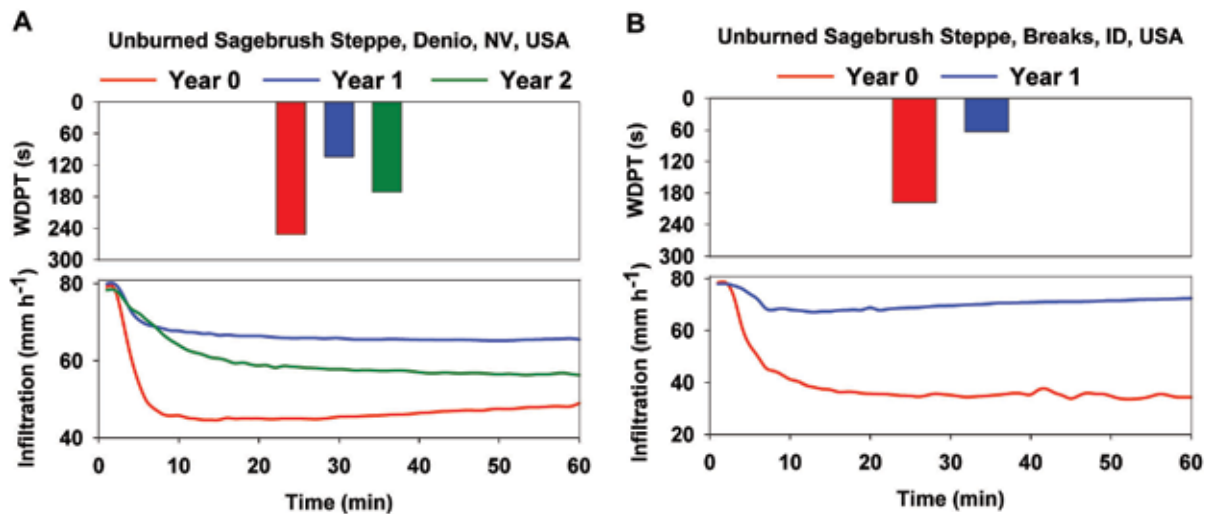
Figure 17—Temporal variability of soil water repellency (measured by using water drop penetration time, WDPT) effects on infiltration of artificial rainfall into unburned, coarse-textured soils at two sagebrush sites: (A) Denio Fire (wildfire), Pine Mountain Range, Nevada, (Pierson et al. 2008a) and (B) Breaks Prescribed Fire, Reynolds Creek Experimental Watershed, Idaho (Pierson et al. 2009). WDPT is an indicator of strength of soil water repellency as follows: <5 s wettable, 5 to 60 s slightly repellent, 60 to 600 s strongly repellent (Bisdom et al. 1993).

soils become wet (Letey et al. 1962; Meeuwig 1971; Burch et al. 1989; DeBano 2000; Robichaud 2000; Pierson et al. 2008b, 2009). Increasing infiltration rates over time (minutes to hours) occurs due to (1) a gradual decrease in repellency with wetting, or (2) lateral and vertical water transfer through preferential infiltration and flow via macropores or breaks in the repellent layer (Meeuwig 1971; DeBano 1981; Burch et al. 1989; Imeson et al. 1992; Hendrickx et al. 1993; Ritsema et al. 1993; Dekker and Ritsema 1994; Ritsema and Dekker 1994, 1995; Dekker and Ritsema 1995, 1996; Ritsema and Dekker 1996; Ritsema et al. 1997, 1998a, 1998b; Doerr et al. 2000; Jaramillo et al. 2000; Leighton-Boyce et al. 2007; Pierson et al. 2008b). Vegetation and ground cover store intercepted rainfall, as well as slow and retain overland flow, allowing more time for infiltration and soil wetting (Meeuwig 1971; Leighton-Boyce et al. 2007; Pierson et al. 2008a, 2009, 2010, 2013; Williams et al. 2014a). Vegetation and ground cover also promote macropore development along root channels, animal burrows, and other soil voids (Belnap et al. 2005; Ludwig et al. 2005). Preferential flow into macropores bypasses water-repellent layers to wettable soil within the root zone (Meeuwig 1971; DeBano 1981; Burch et al. 1989; Imeson et al. 1992; Hendrickx et al. 1993; Ritsema et al. 1993; Dekker and Ritsema 1994; Ritsema and Dekker 1994, 1995; Dekker and Ritsema 1996; Ritsema et al. 1998a, 1998b; Doerr et al. 2000; Jaramillo et al. 2000; Leighton-Boyce et al. 2007; Pierson et al. 2008b; Robinson et al. 2010). In some cases, the vertical bypass of water through repellent zones via macropores leaves a dry layer at the soil surface and wet conditions in the root zone (Meeuwig 1971; Burch et al. 1989; Imeson et al. 1992). The overall effect of preferential flow depends on the extensiveness of the macropore network and the strength of soil water repellency.

The efficacy of naturally occurring soil water repellency on infiltration can be significant at point scales, but quantification of the impacts over larger scales is confounded by spatial variability in hydrophobicity and cover as well as soil properties (Imeson et al. 1992; Doerr et al. 2000; Shakesby et al. 2000; Doerr and Moody 2004; Leighton-Boyce

et al. 2007; Pierson et al. 2009). DeBano (1971) found that horizontal infiltration was 25 times faster in a soil under wettable conditions as compared to a similar soil under hydrophobic conditions. Leighton-Boyce et al. (2007) determined that runoff from small-plot (0.36 m^2) rainfall simulations was 16 times higher under water-repellent conditions than when the same soils were wettable. Madsen et al. (2008) found that pre-wetting water-repellent surface soils underneath Utah juniper and twoneedle pinyon litter yielded hydraulic conductivities (as measured by an infiltrometer) 6 to more than 30 times greater than under water-repellent conditions. Madsen et al. (2008) observed (without taking specific measurements) that tree coppices retained surface water and routed it laterally toward preferential wet spots under the tree canopy.

Vertical preferential flow along wet spots has been referred to as fingered flow (Ritsema and Dekker 1994; Dekker and Ritsema 1995; Ritsema et al. 1997). Dekker and Ritsema (1996) reported that fingered flow into dry, strongly water-repellent conditions generated significant differences up to nearly 30 percent volumetric moisture content between closely spaced samples of fine-textured soils. In multiyear rainfall simulation studies of two steeply sloping mountain big sagebrush sites in Nevada and Idaho, Pierson et al. (2001a, 2008a,b, 2009) found that minimum- and steady-state infiltration rates (0.5 m^2 rainfall simulation plots) on unburned shrub coppices increased 25 to 65 percent after a between-years decrease in soil water repellency strength by 55 to 75 percent. Minimum- and steady-state infiltration rates on unburned interspaces in the studies by Pierson et al. (2001a, 2008a) increased by 65 and 55 percent, respectively, after a 55 percent between-years decrease in soil water repellency strength. Pierson et al. (2009) reported that threefold stronger soil water repellency on shrub coppice than interspace plots resulted in 31 mm and 49 mm of runoff from shrub coppices and interspaces, respectively. The contradiction in runoff rates with strength was attributed to interception, surface retention, and preferential flow (inferred) associated with greater canopy and ground cover on coppices. Soil moisture and cover conditions for respective coppice and interspace areas were similar for the unburned condition throughout the Pierson et al. (2001a, 2008a,b, 2009) studies. Clearly, soil water repellency can significantly reduce infiltration rates over small scales, but the heterogeneity of soil and cover conditions on undisturbed sites and preferential flowpaths most likely subdue the effects at hillslope and catchment scales (Meeuwig 1971; Burch et al. 1989; Imeson et al. 1992; Doerr et al. 2000; Shakesby et al. 2000; Doerr and Moody 2004; Verheijen and Cammeraat 2007; Pierson et al. 2009).

Soil water repellency may provide water conservation and increased plant productivity for some woody species and may indirectly mitigate runoff generation (Doerr et al. 2000; Jaramillo et al. 2000; Lebron et al. 2007; Madsen et al. 2008; Robinson et al. 2010). Imeson et al. (1992) suggested that preferential flow to deep storage beneath the surface water-repellent layer trapped soil water and prevented it from evaporation and upward capillary transfer. Lebron et al. (2007) and Madsen et al. (2008) observed (in field observations) that surface water on water-repellent soils under Utah juniper and twoneedle pinyon was routed to preferential wet spots. They postulated that these locations provide fingered flow through the water-repellent layer to deep soil storage. Roundy et al. (1978) hypothesized similar behavior to explain rapid infiltration of simulated rainfall into water-repellent soils of Utah juniper. Other researchers have proposed soil water repellency as a routing mechanism to preferential flowpaths and deep soil

recharge (see Doerr et al. 2000; Lebron et al. 2007). The recharge of deeper soil layers through preferential flow indirectly influences runoff behavior through increased plant productivity (Ryel et al. 2003). Water availability deep in the soil profile favors woody plant recruitment and facilitates a coppice/interspace structure (Breshears and Barnes 1999). Increased plant productivity through greater water availability and transpiration rates (Ryel et al. 2003) recruits surface plant and litter biomass associated with higher infiltration rates (Ludwig et al. 1997; Wilcox et al. 2003a, Huxman et al. 2005; Ludwig et al. 2005). Therefore, surface flow routing by soil water repellency may function similar to the lateral surface transfers of overland flow (runon) in maintaining shrub, grass, and tree islands of higher biological activity and water retention (Schlesinger et al. 1990; Joffre and Rambal 1993; Pierson et al. 1994a,b; Dunkerley and Brown 1995; Ludwig and Tongway 1995; Seyfried and Wilcox 1995; Wilcox and Breshears 1995; Breshears et al. 1997b; Tongway and Ludwig 1997; Bergkamp 1998; Puigdefábregas et al. 1999; Reid et al. 1999; Bhark and Small 2003; Wilcox et al. 2003a; Huxman et al. 2005; Ludwig et al. 2005; Robinson et al. 2010).

Cover Influences on Sediment Detachment and Transport

Surface Protection from Raindrop Detachment

The primary effects of cover on rainsplash erosion are dissipation of rainfall energy, direct prevention of rainfall contact with the soil surface, and soil stabilization. Recall that for sediment detachment to occur the erosive energy (shear stress) applied to the soil surface must exceed the detachment resistance of soil (critical shear stress or shear strength) (Foster and Meyer 1972; Sharma et al. 1991; Nearing et al. 1999; Kinnell 2005). Canopy and ground cover dissipate the erosive energy of rainfall via interception, thereby reducing the shear stress applied to the soil surface (Al-Hamdan et al. 2013). Recent studies have estimated that rangeland canopy and ground cover can reduce rainfall erosivity approximately 50 percent (Martinez-Mena et al. 1999; Wainwright et al. 1999). Plants and organic material also contribute to the soil shear strength by anchoring soils and promoting aggregate stability (Blackburn 1975; Pierson et al. 1994a,b; Cammeraat and Imeson 1998; Cerdà 1998b; Puigdefábregas et al. 1999; Pierson et al. 2010, 2013, 2014; Williams et al. 2014a). The surface protection and soil stabilization by cover elements (figs. 15C and 15D) are paramount in minimizing erosion given that raindrop impact is the primary sediment contributor to shallow overland flow (Young and Wiersma 1973; Wainwright et al. 2000; Kinnell 2005). Cover may significantly reduce soil loss even where surface runoff is substantial (for example during intense rainfall or water-repellent soil conditions) (Pierson et al. 2009, 2010).

Rainsplash erosion rates are seldom quantified separately from overall interrill erosion rates in rangeland field studies. Parsons et al. (1992, 1994) found the rainsplash erosion rate on Arizona rangelands was 0.01 to 0.04 g m^{-2} min^{-1} on grassland (73 to 86 mm h^{-1} intensity) and 0.34 g m^{-2} min^{-1} on shrubland (145 mm h^{-1} intensity) during artificial rainfall experiments (see Wainwright et al. 2000). Rainsplash during the shrubland experiments eroded about 1.6 times more sediment from areas between plant canopies than from areas underneath plant canopies (Parsons et al. 1992). The Parsons et al. (1992, 1994) and Wainwright et al. (1999) studies demonstrate the potential influence of cover

on rainsplash erosion; however, the net effects vary with soil properties (see Bryan 2000; Kinnell 2005), cover amount/type (Gabet and Dunne 2003), and rainfall characteristics (Bryan 2000; Salles and Poesen 2000; Salles et al. 2000).

Cover Effects on Sheetflow Erosion

Canopy and ground cover reduce sheetflow erosion by controlling the water available for sediment transport and by recruiting surface roughness elements that disperse overland flow (Emmett 1970, 1978; Branson et al. 1981; Thurow et al. 1986; Seyfried 1991; Abrahams et al. 1995; Wainwright et al. 2000; Pierson et al. 2002b, 2010, 2013; Williams et al. 2014a, 2015). Shallow sheetflow (fig. 9B) has little erosive energy, but is the primary transport mechanism for soil detached by raindrops or by other pre-event processes (such as freeze-thaw or weather) (Kinnell 1990, 2005). Progressively deeper flowpaths dampen the erosive energy of raindrops (Moss and Green 1983; Kinnell 1990, 1991, 1993), but may exert enough shear stress on the soil surface to detach and entrain soil (Foster and Meyer 1972; Kinnell 2005). Collectively, rainsplash and sheetflow are primary conduits for hillslope sediment delivery except where concentrated flow (fig. 9C) or rills occur (Kinnell 2005). Higher infiltration rates and rainfall interception associated with cover elements reduce water availability for transport of eroded material (Blackburn 1975; Branson et al. 1981; Abrahams et al. 1991; Blackburn et al. 1992; Pierson et al. 1994a,b; Abrahams et al. 1995; Reid et al. 1999; Wainwright et al. 2000; Pierson et al. 2008a, 2009, 2010; Williams et al. 2014a). Vegetation, litter, and rocks promote surface roughness, which dissipates the velocity and energy of runoff where it does occur (Emmett 1970, 1978; Seyfried 1991; Abrahams and Parsons 1994; Abrahams et al. 1995; Parsons et al. 1996; Wainwright et al. 2000; Pierson et al. 2002b, 2007a; Al-Hamdan et al. 2012a, 2013). Reduced flow velocities have lower detachment and allow surface runoff to disperse and sediment to fall out of suspension. Ponding behind shrub mounds, grass clumps, and litter dams further dissipates rainsplash and facilitates deposition of sediment delivered from upslope runoff (Emmett 1970, 1978; Seyfried 1991). Within-event soil loss from well-vegetated areas is generally 2- to 10-fold less than that from sparsely covered or bare interspaces (table 3), but differences can exceed three orders of magnitude (Pierson et al. 1994b). Actual differences vary with cover, soil, rainfall, and topography characteristics. The net effect of cover on interrill processes can reduce rangeland within-storm soil loss 8- to 10-fold across the plant (less than 1 m^2) to patch (tens of square meters) scales (Pierson et al. 1994b, 2009) and, where cover exceeds 50 to 60 percent, results in minor hillslope soil loss (Gifford 1985; Pierson et al. 1994b, 2008a, 2009, 2010, 2013; Williams et al. 2014a).

Cover Effects on Concentrated Flow Erosion

The effects of cover elements on concentrated flow erosion (fig. 9C) are similar to those in sheetflow erosion. The main effects are reduced runoff discharge, flow velocity, and sediment detachment (Emmett 1970, 1978; Branson et al. 1981; Thurow et al. 1986; Seyfried 1991; Abrahams et al. 1995; Wainwright et al. 2000; Pierson et al. 2007a, 2008a, 2009; Al-Hamdan et al. 2012a,b, 2013; Williams et al. 2014a). Vegetation and ground cover reduce water available for concentrated flow formation and thereby decrease

concentrated flow discharge (Pierson et al. 2007a, 2008a, 2009). Soil detachment by concentrated flow is well correlated with flow velocity (Pierson et al. 2008a, 2009) and discharge (Nearing et al. 1997, 1999; Govers et al. 2007; Al-Hamdan et al. 2012b), and flow velocity is strongly related to discharge (Govers 1992; Nearing et al. 1997, 1999; Giménez and Govers 2001; Govers et al. 2007; Al-Hamdan et al. 2012a). Grass clumps, plant bases, coppice mounds (fig. 16), and litter dams create topographic highs that may concentrate flow where runoff occurs, but concentrated flowpaths on well vegetated/covered sites generally flow a short distance and disperse (Emmett 1970, 1978; Seyfried 1991; Parsons et al. 1996; Bryan 2000; Wainwright et al. 2000).

The erosive energy and transport capacity of concentrated flow are greatly reduced when flow intersects ground cover elements (Al-Hamdan et al. 2012a,b, 2013). Roughness created by ground cover counteracts flow energy by amplifying hydraulic friction until the flow submerges the ground cover (Emmett 1970, 1978; Abrahams and Parsons 1991b; Abrahams and Parsons 1994; Abrahams et al. 1994, 1995; Parsons et al. 1996; Nearing et al. 1997; Wainwright et al. 2000). Studies from a western juniper (*J. occidentalis* Hook.) woodland (Pierson et al. 2007a) and a mountain big sagebrush rangeland (Pierson et al. 2009) reported concentrated flow velocities that were 1.5- (woodland interspaces) to more than 2-fold (recently burned mountain big sagebrush) greater on degraded hillslopes with 80 percent bare ground than on adjacent hillslopes with 60 percent and 20 percent bare ground, respectively. Sediment yield from concentrated flow processes was fourfold (Pierson et al. 2009) to eightfold (Pierson et al. 2007a) greater from the degraded sagebrush and woodland slopes. Sediment transported by concentrated flow where it does occur on well-vegetated sites often forms miniature alluvial fans adjacent to vegetative clumps (Emmett 1970, 1978; Seyfried 1991). These features indicate that concentrated flow does redistribute surface soil from bare areas to vegetated zones on healthy rangelands, but hillslope soil loss from this process is minor under such conditions (Pierson et al. 2007a, 2009). In contrast, concentrated flow becomes the dominant erosion mechanism on degraded rangelands where ground cover is sparse (Moffet et al. 2007; Pierson et al. 2008a, 2009, 2011, 2013; Williams et al. 2014a,b).

Section 4: Impacts of Fire on Rangeland Runoff and Erosion

Fire Behavior and Regimes on Rangelands

The environmental effects of a particular fire or series of fires are often placed in the context of fire behavior, intensity, severity, and regime (DeBano et al. 1998; Shakesby and Doerr 2006; Brooks 2008; Keeley 2009). Terms explaining these relationships are frequently used inconsistently (Lentile et al. 2006; Keeley 2009), especially among nonfire-specific disciplines like hydrology. Therefore, we present a brief summary of the fire terms used here to discuss fire effects on rangeland hydrology. Fires are categorized as ground, surface, or crown type, and each is associated with a particular behavior (DeBano et al. 1998; Shakesby and Doerr 2006; Brooks 2008). Fire behavior refers to the rate of spread, residence time, and flame dimensions, and is related to fuel, weather, and topographic conditions at the time of burning (DeBano et al. 1998, Brooks 2008). Ground fires are mostly flame-free and burn slowly through duff or decayed organic matter on the soil surface. Surface fires burn rapidly and consume litter, woody dead material at or near the surface, herbaceous fuels, shrubs, and small trees. Crown fires burn rapidly through canopies of trees and tall shrubs, leaving most of the stem and land surface fuels unburned. An individual fire may comprise one or more of these three primary fire types (see DeBano et al. 1998).

The term "fire regime" refers to the pattern of repeated burning within a large spatial expanse (landscape scale) over long time periods and is defined by a characteristic combination of fire type, frequency, intensity, severity, size, and seasonality (Brooks 2008; Baker 2009). Fire frequency is the number of fires that occur over a specified period of time for a particular area, and may be expressed as a return interval/cycle or the length of time necessary for the area of interest to burn. Seasonality is the period of the year when fires are most likely to occur. Fire intensity refers to the amount of heat released. Severity is the degree of impact to soils or vegetation, or to both (see Keeley 2009).

Fire behavior, intensity, and severity are affected by the vertical and horizontal continuity and density of fuels. Fuels, as with fire types, are commonly divided vertically into layers termed "ground" (duff, roots, and buried or partially buried woody dead materials), "surface" (litter, herbaceous plants, and low shrubs), and "canopy" (tall shrubs and trees) fuels (DeBano et al. 1998). Extensive horizontal continuity of surface fuels, canopy fuels, or both, facilitates large fires, especially under dry and windy conditions. Breaks in fuel continuity retard fire progression. Rangeland fuels usually have low horizontal continuity of surface and canopy fuels—except on more productive sites or following long fire-free periods (Brooks 2008; Keane et al. 2008). The spread of fire from surface fuels to the canopy often requires ladder fuels (horizontally connected heterogeneous vertical continuity). Fire intensity is largely dictated by the amount or density of fuel, or both. Fuel density (amount of fuel per unit volume of space) influences combustion and the duration of burning. Loosely packed fuels provide sufficient air supply for higher combustion and rapid rates of spread relative to densely packed fuels (DeBano et al. 1998; Baker 2009). Fuel moisture conditions and temperature also influence fire intensity by regulating

flammability and combustion. Dry, hot fuels exhibit higher flammability, combustion, and rates of spread than moist, cool fuels (DeBano et al. 1998; Baker 2009).

The variable most commonly referenced in assessing fire impacts on hydrology is severity (Pierson et al. 2001a, 2002b; Lewis et al. 2006; Shakesby and Doerr 2006; Pierson et al. 2008a, 2009; Parsons et al. 2010; Pierson et al. 2014; Williams et al. 2014a,b), although it is not quantified consistently (Keeley 2009). In general, fire severity is a function of fire intensity and residence time (DeBano et al. 1998). Numerous severity classification systems exist, but three classes are common: low, moderate, and high severity (DeBano et al. 1998; Baker 2009; Parsons et al. 2010). Low-severity fires cause low soil heating (100 to 250 °C), light charring or minor reduction of litter, and virtually no consumption (but some charring) of duff. Moderate-severity fires char the ground surface without visible alteration of the mineral soil surface, leave behind some gray- to black-colored ash, consume litter and most woody debris (except logs), and have surface temperatures (up to 1 cm depth) of 300 to 400 °C. High-severity fires produce surface temperatures in excess of 500 °C, create deep ground charring where duff is completely consumed, visibly affect the upper mineral soil layer (leaving it reddish in color), and result in white ash from consumption of grasses, litter, and most shrub stems.

Wells et al. (1979) extended severity classification over large expanses based on total area burned at low, moderate, or high intensities. They suggested the following framework: (1) low-severity burn (less than 2 percent severely burned, less than 15 percent moderately burned, remainder is unburned or burned at low severity); (2) moderate-severity burn (less than 10 percent severely burned, more than 15 percent burned moderately, remainder unburned or burned at low severity); and (3) high-severity burn (more than 10 percent with patches of high burn severity, more than 80 percent moderate to severely burned, remainder burned at low severity). Brown (2000) categorized fire regimes loosely related to recurring uniformity of fire severity over a specified area of interest. The categories are: (1) stand-replacement regime (lethal burning of 80 percent or more of dominant vegetation), (2) understory fire regime (generally nonlethal to dominant vegetation), and (3) mixed-severity fire regime (recurring fire produces variable response in time and space, from nonlethal understory to stand-replacement fire). The uniform fine fuels on most grassland communities produce a stand-replacement fire regime (DeBano et al. 1998; Rice et al. 2008). These fires may be of high intensity, but generally cause only minor surface soil heating (100 to 400 °C) (Wright and Bailey 1982) due to the lack of woody fuels and low residence times (DeBano et al. 1998). In contrast, shrubland ecosystems may exhibit a mixed-severity or stand-replacement fire regime (Rice et al. 2008) and yield soil surface temperatures of 260 to 700 °C (Wright and Bailey 1982). Parsons et al. (2010) provide guidance on field mapping of burn severity.

Fire Effects that Dictate Hydrologic Response

Reduced Interception and Surface Protection

Consumption of canopy and ground cover by fire reduces interception capacity and surface water retention and thus increases the quantity and intensity of water input at the soil surface and the flow volume and velocity across it (DeBano et al. 1998; Shakesby and Doerr 2006). The amount of additional water input made available by burning
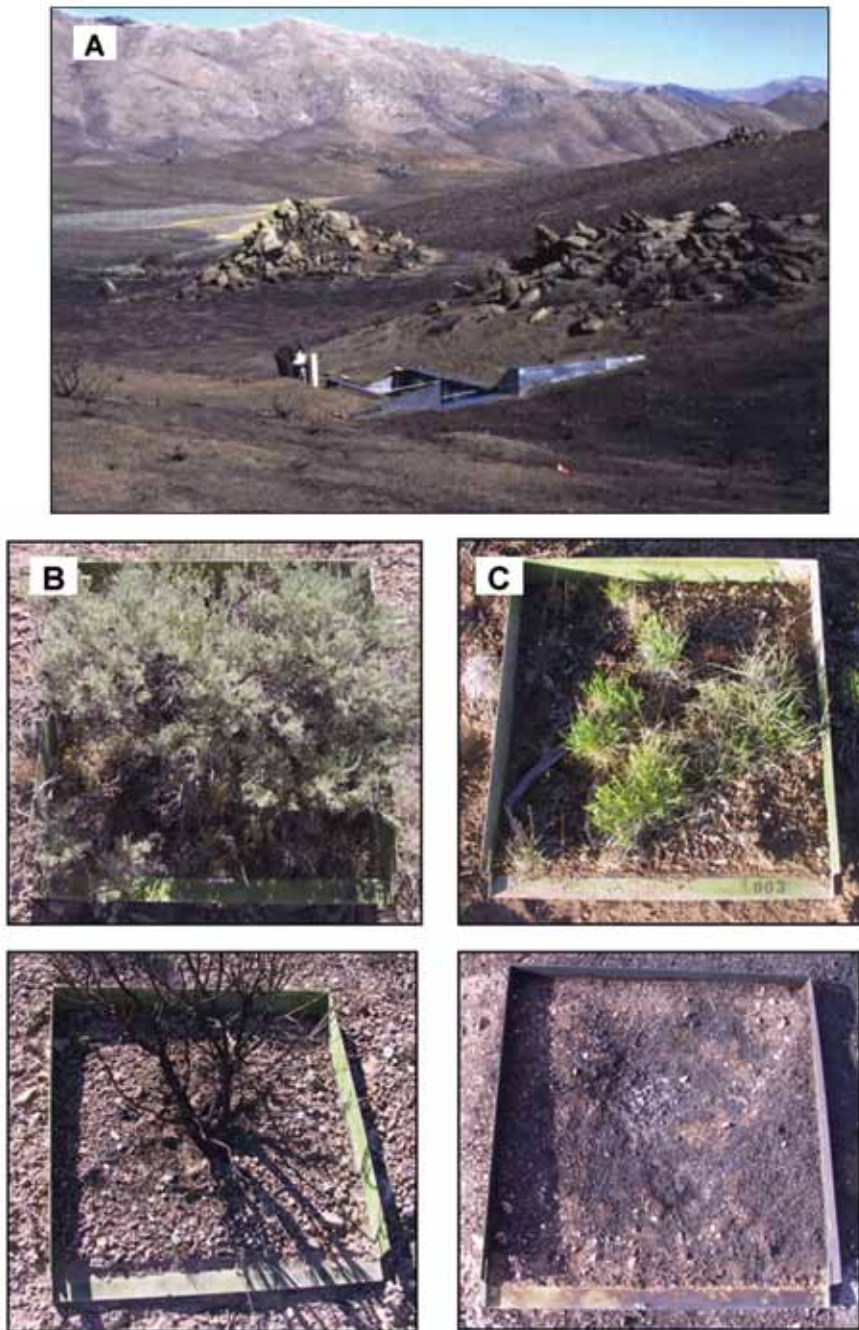
Figure 18—(A) Postfire landscape (3 months after wildfire) at the Denio Fire, Pine Mountain Range, Nevada, and cover removal by fire for (B) shrub coppice and (C) vegetated interspace experimental plots (0.5 m²) (photos: USDA Agricultural Research Service, Northwest Watershed Research Center).

is dependent on the interception and storage capacity of the postburn cover. General estimates suggest that the quantity of interception by unburned rangeland trees, shrubs, and grasses approximates 1 to 2 mm of rainfall per storm (Bonan 2002) depending on the cover biomass, rainfall intensity and duration, cover moisture content, and the vertical and horizontal arrangements of cover elements (see Section 3, *Vegetation and Cover Influences on Infiltration and Runoff Generation*).

The conversion of interception loss (table 2) and stemflow rates to rainfall arrival at the soil surface is nearly 100 percent where severe burning uniformly removes all canopy and ground cover elements (fig. 18). Postfire reductions in raindrop dissipation (increases

in drop intensity) are as important as the increases in the quantity of water (Shakesby and Doerr 2006). Raindrop energy dictates splash sediment detachment under disturbed conditions whereas the quantity of overland flow and the downslope velocity govern flow detachment and transport capacity (Pierson et al. 2008a, 2009; Al-Hamdan et al. 2012b). Greater raindrop impact after canopy and ground cover removal results in increased soil detachment from rainsplash processes (Shakesby and Doerr 2006; Pierson et al. 2001a, 2002b, 2008a, 2009, 2013, 2014; Williams et al. 2014a, 2016a). Reductions in ground cover (decreased surface roughness) abate surface retention of overland flow, allowing flow to concentrate and move downslope with greater velocity, erosive energy, and transport capacity (Shakesby and Doerr 2006; Pierson et al. 2008a, 2009; Al-Hamdan et al. 2012a,b, 2013; Pierson et al. 2013; Williams et al. 2014a, 2016a). The potential overall effect is a decrease in the time to runoff generation and an increase in cumulative runoff and sediment yield over the duration of a storm event.

In snow-dominated environments, removal of vegetation by fire may alter snow accumulation, the timing of runoff initiation, cessation, and peak flow within the year, and the amount of snowmelt runoff. Burning may also result in increased surface temperatures and snowmelt rates due to greater incoming solar radiation postburn (Tiedemann et al. 1979). Any reduction in vegetation, therefore, reduces snow accumulation and water availability for biological processes and streamflow generation. Reduced snow retention also potentially alters runoff characteristics from summer thunderstorms on water-limited sites by inhibiting vegetation production and ground cover recruitment. Where snow does accumulate, runoff responses to mid-winter rain-on-snow events may be substantial after burning (see Marks et al. 2001; Pierson et al. 2001b).

Physical, Chemical, and Biological Alteration of Soils

Hydrologically important soil properties (porosity, soil moisture, stability, structure, and water repellency) are strongly influenced by vegetation, organic debris, and micro-organisms, which can be removed at varying degrees by burning (DeBano et al. 1998; Shakesby and Doerr 2006). The magnitude at which these properties are influenced by burning depends on the degree of soil heating and the amount of organic matter removed. The impact of burning on soils is at a maximum when the entire canopy and all surface organics are consumed and the mineral surface is exposed. The primary effect of fire relative to organic matter is to expedite the mineralization process. Postfire recovery of consumed organic matter may take as long as 5 years on some rangelands (Wright and Bailey 1982). Soil organic matter is combusted at temperatures above 200 ˚C and is completely consumed at 500 ˚C (DeBano et al. 1998). These temperatures are well within the range of surface temperatures commonly reported for grassland and shrubland fires (Wright and Bailey 1982; Miller et al. 2013). Subsurface soil temperatures on grasslands usually are unaffected by burning, but can range from approximately 100 to 250 °C to depths of 5 cm on shrublands (Wright and Bailey 1982; DeBano et al. 1998). The combustion of organic matter from surface and subsurface soils can alter soil structure, increase bulk density, and decrease porosity (Giovannini and Lucchesi 1997; Giovannini et al. 1988; Pierson et al. 2001a,b; Hubbert et al. 2006; Shakesby and Doerr 2006).

Soil stability and aggradation can also be reduced through alteration of soil particles during burning (DeBano et al. 1998; Neary et al. 1999; Andreu et al. 2001; Giovannini et

al. 2001; Hubbert et al. 2006; Shakesby and Doerr 2006; Mataix-Solera et al. 2011). Soil temperatures between 150 and 400 °C can drive off hydroxyl groups in clays and increase erodibility (Giovannini and Lucchesi 1997; DeBano et al. 1998). Collectively, these soil alterations inhibit infiltration and promote runoff generation and erosion (Hester et al. 1997; DeBano et al. 1998; Shakesby and Doerr 2006). Ash at the soil surface may clog surface soil pores and further accentuate the runoff response (Campbell et al. 1977; Wells et al. 1979; Lavee et al. 1995; Neary et al. 1999), although contrary results have been reported recently for the immediate postfire period (Cerdà and Doerr 2008; Woods and Balfour 2008; Larsen et al. 2009; Bodí et al. 2012; Ebel et al. 2012).

Burning may reduce the role of invertebrates, microorganisms, and mycorrhizae fungi in facilitating infiltration (Ahlgren and Ahlgren 1960; Wright and Bailey 1982; DeBano et al. 1998; Shakesby and Doerr 2006). Mycorrhizae and the by-products of soil fauna promote soil aggregation. Voids created by fauna movement within the soil profile increase porosity. Soil temperatures above 40 to 210 °C are fatal for most fungi and soil organisms; organic matter consumption by fire reduces the primary food source for soil fauna (Wright and Bailey 1982; DeBano et al. 1998; Mataix-Solera et al. 2009). Recovery of soil fauna and fungi may be rapid (Neary et al. 1999) depending on postfire soil microclimate and available nutrients, but can take as much as 3 to 5 years where resources are limited (Wright and Bailey 1982). Finally, soil moisture retention, a key component of plant and soil fauna productivity in water-limited ecosystems, is also adversely affected by the loss of pore structure and surface insulation (against evaporation) by litter (Wright and Bailey 1982; DeBano et al. 1998).

Exacerbation, Alteration, and Formation of Soil Water Repellency

Soil heating during burning may enhance, reduce, or create hydrophobic soil conditions (DeBano and Krammes 1966; DeBano et al. 1970; Savage 1974; DeBano et al. 1976; DeBano 1981, 1991; Shakesby et al. 1993; Doerr et al. 1996; DeBano et al. 1998; DeBano 2000; Doerr et al. 2000; Robichaud 2000; Shakesby et al. 2000; Benavides-Solorio and MacDonald 2001; Huffman et al. 2001; Doerr et al. 2004; MacDonald and Huffman 2004; Doerr et al. 2006; Hubbert et al. 2006; Fox et al. 2007; Pierson et al. 2008b; Arcenegui et al. 2008; Doerr et al. 2009; Shakesby 2011). The key determinants of whether soil water repellency is enhanced, reduced, or created during burning are the presence of organic matter and the soil temperature reached during burning (DeBano and Krammes 1966; DeBano et al. 1970; Doerr et al. 1996; DeBano et al. 1998; Doerr et al. 2004, 2009). Naturally occurring soil water repellency (fig. 19) is typically unaltered by soil temperatures less than 175 °C (DeBano 1981). Soil temperatures between 175 and 270 °C enhance "background" water repellency or may form hydrophobic soil conditions (DeBano 1981; Doerr et al. 2000, 2009). Water repellency breaks down or is destroyed at soil temperatures above 270 to 400 °C (Savage 1974; DeBano et al. 1976; Giovannini and Lucchesi 1997; Doerr et al. 2004).

During fires, combustion of organic matter at the soil surface radiates heat downward into the soil profile and vaporizes organic substances. Some of these substances are translocated downward along temperature gradients until they cool and condense, forming a variable-thickness hydrophobic layer parallel to the soil surface (DeBano 1981; Doerr et al. 1996; DeBano et al. 1998; DeBano 2000; Doerr et al. 2000, 2004, 2009). The

Figure 19—Dry soil underneath a 2- to 3-cm thick layer of water repellent soil observed immediately after application of 85 mm of rainfall (see Pierson et al. 2008b) over a 1-h period following the Breaks Prescribed Fire, Reynolds Creek Experimental Watershed, Idaho. Soils were strongly water repellent at the mineral soil surface and to a depth of 2 to 3 cm immediately before and after the prescribed fire (photo: USDA Agricultural Research Service, Northwest Watershed Research Center).

depth at which the water-repellent layer occurs is related in part to the degree of heating or fire severity (Huffman et al. 2001). Generally, higher surface temperatures, up to that at which repellency is destroyed, increase the depth at which soil water repellency is found (DeBano et al. 1998). Fire-enhanced or fire-induced soil water repellency is commonly found at or within the first 5 cm of the soil surface, and rapidly decreases in strength with increasing soil depth (Benavides-Solorio and MacDonald 2001; Huffman et al. 2001; MacDonald and Huffman 2004; Benavides-Solorio and MacDonald 2005; Pierson et al. 2008b, 2009).

The spatial and temporal persistence (figs. 19 and 20) of postburn soil water repellency are highly variable (Imeson et al. 1992; Doerr et al. 2000; Huffman et al. 2001; MacDonald and Huffman 2004; Hubbert et al. 2006; Woods et al. 2007; Pierson et al. 2008b, 2009, 2013; Bodí et al. 2013; Williams et al. 2014b). The strength of postburn soil water repellency varies with the spatial continuity of the respective burn. Repellency strength is usually positively correlated with burn severity, the amount of organic material present during combustion, and the steepness of the downward soil temperature gradient during soil heating (DeBano et al. 1998). Steep temperature gradients enhance downward translocation (fig. 21). The steepness of the gradient is regulated in part by soil water content and its influence on heat transfer. The persistence of fire-induced soil water repellency commonly ranges from less than 1 to 6 years (DeBano et al. 1976; Huffman et al. 2001; MacDonald and Huffman 2004).

The influence of repellency on infiltration is highly variable over seasonal and annual time scales (fig. 22) and is related to variations in soil moisture content (Shakesby et al. 1993; Burcar et al. 1994; Dekker and Ritsema 1994; Ritsema and Dekker 1994; Dekker and Ritsema 1996; Doerr et al. 2000; Shakesby et al. 2000; Benavides-Solorio and MacDonald 2001; Dekker et al. 2001; MacDonald and Huffman 2004; Hubbert et
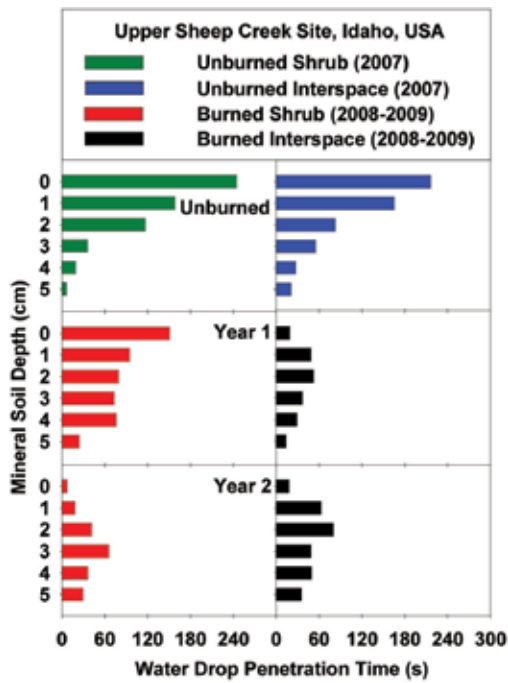
Figure 20—Spatial (by microsite) and annual variation in soil water repellency (measured by water drop penetration time, WDPT) measured on unburned and burned soils underneath shrubs and in interspaces areas at the Upper Sheep Creek Prescribed Burn site in the Reynolds Creek Experimental Watershed. Year 1 and year 2 measurements are 1 and 2 years postfire, respectively. WDPT is an indicator of strength of soil water repellency as follows: <5 s wettable, 5 to 60 s slightly repellent, 60 to 600 s strongly repellent (Bisdom et al. 1993).
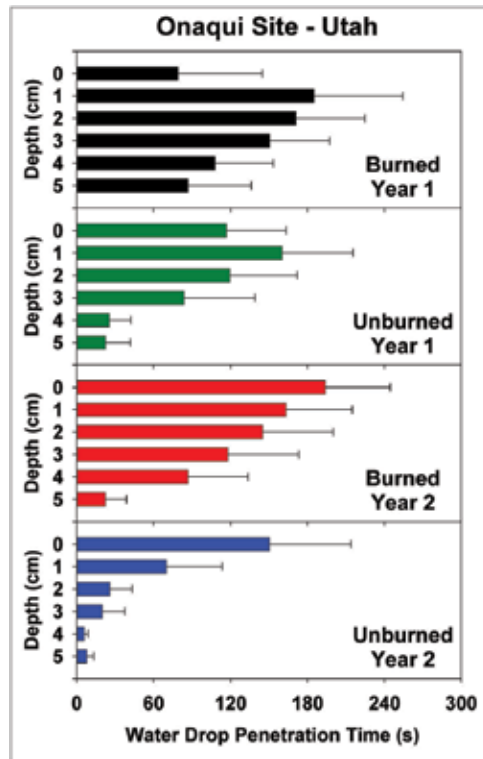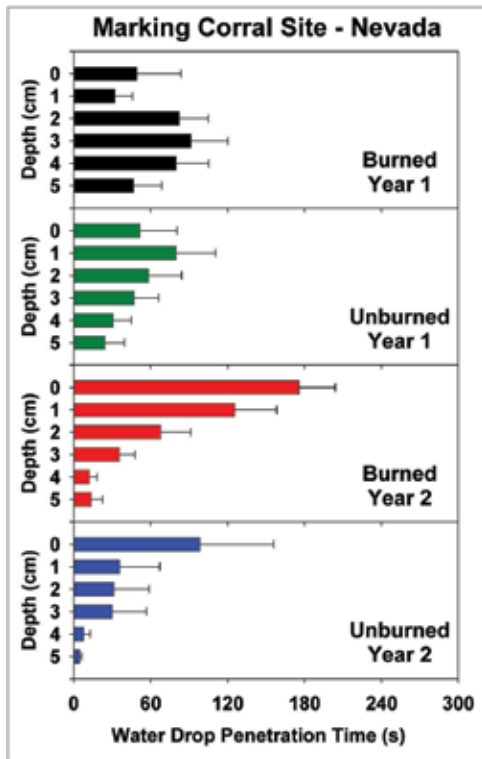
Figure 21—Soil water repellency measured by using the water drop penetration time (WDPT) method on unburned and burned areas underneath Utah juniper and pinyon (Marking Corral site) and Utah juniper (Onaqui site) (see Pierson et al. 2010, 2014). WDPT is an indicator of strength of soil water repellency as follows: <5 s wettable, 5 to 60 s slightly repellent, 60 to 600 s strongly repellent (Bisdom et al. 1993). WDPTs at both sites indicate translocation of prefire repellency from the surface to deeper soil layers (~ 2 to 4 cm) through the first year postfire. The soil water repellency profile 2 years postfire was similar to that of prefire conditions with stronger repellency at the mineral soil surface.

al. 2006; Pierson et al. 2008b; Bodí et al. 2013) and seasonal wetting and drying trends (Doerr and Thomas 2000; Dekker et al. 2001). The overall effect of preexisting or fire-induced soil water repellency is exacerbated by the removal of vegetation and ground cover (Benavides-Solorio and MacDonald 2001, 2002; Pierson et al. 2008b, 2009, 2013, 2014; Williams et al. 2014a). The increased availability of water arriving at the soil surface, lack of surface protection, and presence of hydrophobic soil conditions promote runoff generation and surface soil erosion by raindrop-induced flow, sheetflow, and concentrated flow processes (DeBano and Krammes 1966; DeBano et al. 1979; Morris and Moses 1987; Scott and van Wyk 1990; Terry and Shakesby 1993; Inbar et al. 1997; DeBano 2000; Shakesby et al. 2000; Benavides-Solorio and MacDonald 2001, 2002, 2005; Shakesby and Doerr 2006; Pierson et al. 2008b; Larsen et al. 2009; Pierson et al. 2009, 2014; Williams et al. 2014a,b).
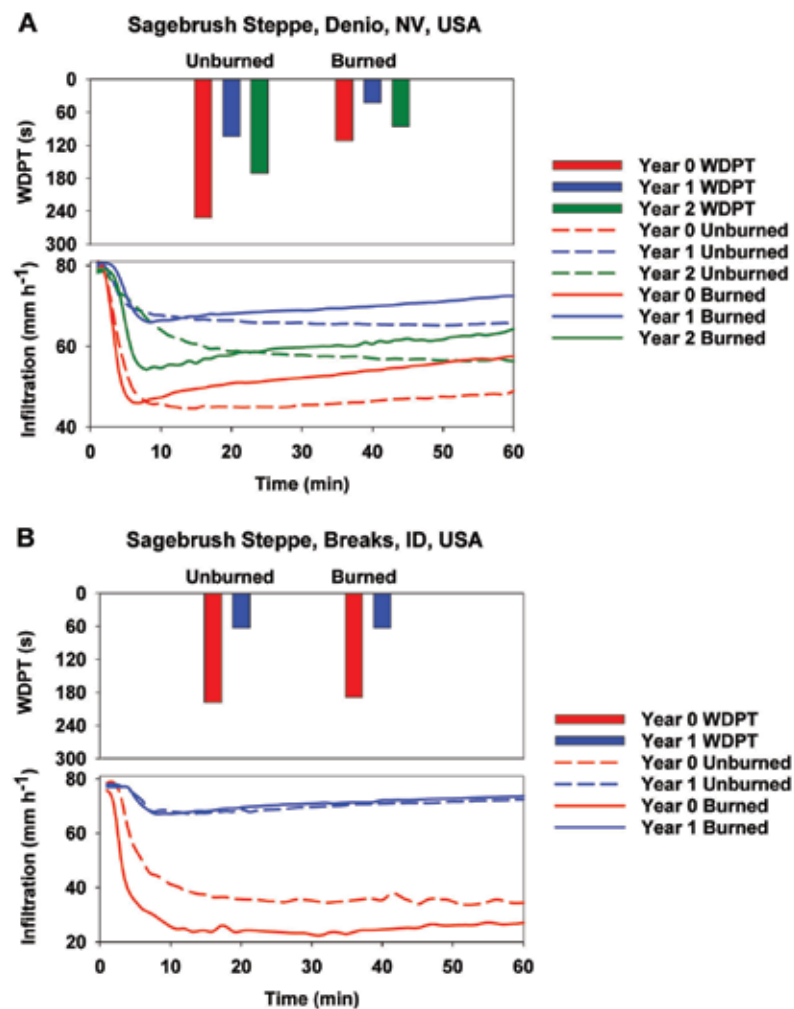
Figure 22—Temporal variability of soil water repellency (measured by using water drop penetration time, WDPT) effects on infiltration of artificial rainfall into unburned and burned, coarse-textured soils at two sagebrush sites: (A) Denio Wildfire, Pine Mountain Range, Nevada (Pierson et al. 2008a) and (B) Breaks Prescribed Fire, Reynolds Creek Experimental Watershed, Idaho (Pierson et al. 2009). WDPT is an indicator of strength of soil water repellency as follows: <5 s wettable, 5 to 60 s slightly repellent, 60 to 600 s strongly repellent (Bisdom et al. 1993).

Fire Effects on Infiltration and Runoff Generation

The degree to which fire affects infiltration and runoff processes depends on the magnitude of alterations to soil properties, vegetation, and ground cover (DeBano et al. 1998; Shakesby and Doerr 2006; Pierson et al. 2011; Williams et al. 2014b, 2016a). Runoff generation further depends on water arrival in excess of aboveground storage, surface storage, and infiltration. Thus, the rainfall intensity and duration as well as the conditions of the vegetative community and ground surface are key determinants of the runoff response (Benavides-Solorio and MacDonald 2005; Spigel and Robichaud 2007). Hydrologic response is further influenced by topographic attributes such as hillslope steepness (Al-Hamdan et al. 2013). The occurrence and hydrologic value of soil fauna, organic matter, macropores, and soil structure are closely related to the presence of vegetation and ground cover (Belnap et al. 2005), particularly in water-limited environments like rangelands (Ludwig et al. 1997, 2005). Therefore, many assessments of fire effects on infiltration and runoff consider alteration of vegetation, ground cover, and surface properties solely and exclude investigation of fire effects on soil organisms, macropores, and soil organic matter.

Literature clearly indicates that fire influences soil fauna, macropores, and soil organic matter (Wright and Bailey 1982; DeBano et al. 1998; Neary et al. 1999; Mataix-Solera et al. 2009; Certini et al. 2011; Mataix-Solera et al. 2011), but their respective alteration and direct relation to postfire hydrologic responses are rarely specifically measured or cited. Alteration of canopy, ground cover, and surface soil properties commonly serve as a surrogate for the collective fire impact and exhibit strong correlations with postfire hydrologic response (Benavides-Solorio and MacDonald 2001; Johansen et al. 2001; Benavides-Solorio and MacDonald 2005; Shakesby and Doerr 2006; Cerdà and Doerr 2008; Pierson et al. 2008a, 2009, 2013, 2014; Williams et al. 2014a,b). Vegetation and ground cover intercept rainfall and overland flow, surface roughness promotes ponding, and surface soil characteristics influence infiltration rates. Ponding delays runoff generation and allows water to infiltrate through breaks in hydrophobic soils or through macropores created by root channels, organic matter, or soil fauna. Organic matter input through plants and soil fauna promotes aggregate stability and infiltration by enhancing soil structure. From literature, three primary points emerge relative to these relationships: (1) canopy cover and surface protection are paramount in reducing water availability for runoff (Cerdà 1998a; Johansen et al. 2001; Pierson et al. 2001a; Benavides-Solorio and MacDonald 2005; Pierson et al. 2008a, 2009, 2013; Williams et al. 2014b); (2) the effects of decreased surface protection on water availability are markedly influenced by soil conditions (such as soil water repellency) that inhibit or promote infiltration and storage (Pierson et al. 2001a,b; Shakesby and Doerr 2006; Pierson et al. 2008a,b, 2009, 2013, 2014; Williams et al. 2014a); and (3) the effects of decreased canopy cover, surface protection, or the co-occurrence of these effects, are intensified with increasing rainfall intensity (Inbar et al. 1998; Benavides-Solorio and MacDonald 2005; Spigel and Robichaud 2007; Larsen et al. 2009; Williams et al. 2014b) and hillslope angle (Al-Hamdan et al. 2013).

Rainsplash and Sheetflow Processes

Over the last decade, a number of researchers have measured fire effects on infiltration and runoff over small-plot scales (0.25 to 1 m^2) (table 3) (Cerdà 1998a; Robichaud 2000; Benavides-Solorio and MacDonald 2001; Pierson et al. 2001a; Benavidez-Solorio and MacDonald 2002; Pierson et al. 2002b; Cerdà and Doerr 2005; Groen and Woods 2008; Pierson et al. 2008a,b; Woods and Balfour 2008; Pierson et al. 2009, 2013, 2014; Williams et al. 2014a,b, 2016a). Small-plot sizes limit inferences to rainsplash and sheetflow processes solely (Mutchler et al. 1988). These studies have, however, provided valuable insight into infiltration and runoff generation for both burned and unburned conditions, and elaborate on the magnitude of runoff response to burning. A series of rainfall simulation studies on small (0.5 m^2) plots (Pierson et al. 2001a, 2002b, 2008a, 2008b, 2009) (see table 3) on steeply sloped (35 to 60 percent) sagebrush rangelands demonstrate the effects of cover removal, surface alteration, and soil water repellency on postfire rangeland infiltration and runoff.

Pierson et al. (2002b) (see table 3) investigated the hydrologic effects of moderate- and high-severity burning of sagebrush rangelands on north- and south-facing hillslopes 1 year after the Eighth Street Fire in 1996 near Boise, Idaho. We restrict our discussion of the study to the south-facing hillslopes for brevity. Prefire total live and litter masses were 32,519 kg ha^{-1} and 14,372 kg ha^{-1}, respectively, on shrub coppices and 519 kg ha^{-1} and 1,721 kg ha^{-1} in interspaces. Moderate- and high-severity burning reduced shrub coppice total live and litter mass by nearly 100 percent. Moderate burning of interspaces had no effect on total live mass, but reduced litter mass by 90 percent. High-severity burning of interspaces reduced total live and litter masses by 74 percent and 96 percent, respectively. Percent bare ground (see table 3) increased on shrub coppices from 7 percent prefire to nearly 100 percent for the moderate- and high-severity conditions. Interspace bare ground was high prefire (89 percent) and increased to nearly 100 percent following the burn. Soil organic matter (2 to 6 percent) was nearly equal across all plots and was not affected by burning. Near-surface bulk density (0 to 2 cm) was slightly higher on interspace plots (1.35 g cm^{-3}) than on coppices (1.21 g cm^{-3}) and was not significantly altered by burning. Gravimetric soil-water content ranged from 5 to 14 percent. Surface roughness decreased by 30 percent after moderate- and high-severity burning of coppice microsites and by 30 and 45 percent after moderate- and high-severity burning, respectively, on interspaces.

Runoff in the Pierson et al. (2002b) study was measured during simulation of a summer season convective storm with an intensity of 67 mm h^{-1} and duration of 60 min. Runoff coefficients (runoff per unit of applied rainfall) from unburned hillslopes were 11 percent for coppice areas and 24 percent for interspace areas. Runoff coefficients for coppices increased threefold following moderate- and high-severity burning. Runoff coefficients for interspaces were nearly equal (24 to 26 percent) for unburned and moderate-severity plots, but increased twofold to nearly 50 percent for high-severity burn plots. Greater infiltration and lower runoff on coppices prefire were attributed to interception and surface water retention associated with high live canopy and litter biomass. Runoff generally increased with the burn severity, potentially owing to enhanced background soil water repellency. Soil water repellency was not directly assessed, but the shape of the infiltration curves clearly indicates the presence of water repellency (steeply decreasing followed by gradual increase throughout simulation) (fig. 23) (see Meeuwig 1971;
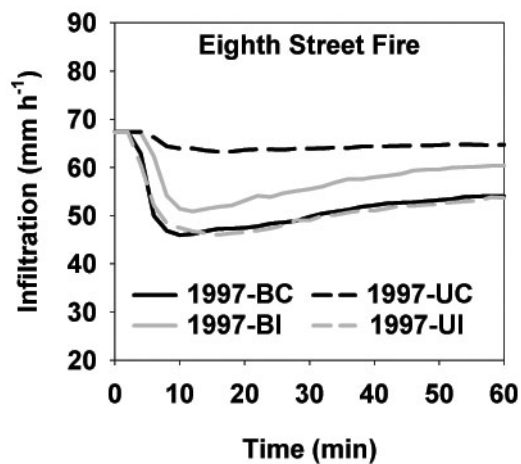
Figure 23—Infiltration of artificial rainfall (67 mm h⁻¹, 60 min) applied on burned (coppice – BC, interspace – BI) and unburned (coppice – UC, interspace – UI) sagebrush hillslopes following the Eighth Street Wildfire in 1996 near Boise, Idaho. Data from Pierson et al. (2002b).

Imeson et al. 1992; Robichaud 2000; Pierson et al. 2001a, 2008b). Postfire increases in runoff were also associated with decreasing surface roughness.

A 3-year investigation by Pierson et al. (2001a, 2008a,b) (see table 3, figs. 22A and 24) measured infiltration and runoff of rainsplash and sheetflow following the Denio Fire in 1999 in Nevada. The fire burned steeply sloping (30 to 40 percent) mountain big sagebrush at high severity. A convective-type storm was simulated by applying 85 mm h⁻¹ rainfall intensity for 60 min to plots 0.5 m² in size. Before the fire, total ground cover was nearly 100 percent in shrub coppice and interspace areas. Shrub canopy cover dominated in coppice areas whereas grasses dominated interspaces. Canopy cover was consumed entirely by the wildfire, and bare ground increased from less than 10 percent to more than 95 percent. Ground cover recovered to about 60 percent (40 percent bare ground) following the second growing season, but remained significantly different from the unburned condition (90 percent). Shrub cover was slow to recover (5 percent after two growing seasons) and litter after the second growing season amassed 30 percent coverage.

Surface bulk densities (0 to 4 cm) were not significantly different between prefire coppice (0.93 g cm⁻³) and interspace (0.94 g cm⁻³) areas in the Pierson et al. (2001, 2008a,b) studies. Burning increased bulk density (to 1.21 g cm⁻³) on both microsites, presumably by decreasing organic matter content (Pierson et al. 2001a), although organic matter was not specifically measured. Soil water repellency was strong on all unburned plots the year of the fire and was reduced more than 50 percent by high-severity burning on shrub coppice and interspace microsites (Pierson et al. 2008b). Soil water repellency across all plots (fig. 22A) decreased by 55 percent from the year of the fire (year 0) to 1 year postfire (year 1) and then increased across all plots approximately 40 to 50 percent from year 1 to year 2 (2 years postfire). Runoff coefficients (table 3) were slightly greater, but not significantly different, for burned (37 percent) than unburned (30 percent) shrub coppices immediately postfire, and were significantly higher for the unburned (49 percent) than burned (30 percent) interspaces.

Pierson et al. (2001, 2008a,b) explained that slightly greater runoff and significantly lower minimum infiltration on burned coppices resulted from the removal by fire of canopy and ground cover over strongly water-repellent soils. Canopy and litter cover on unburned coppices mitigated the effects of strong soil water repellency whereas the removal of cover accentuated the effects of the persistent, but reduced, soil water repellency on burned coppices (Pierson et al. 2008b). The decrease in runoff from
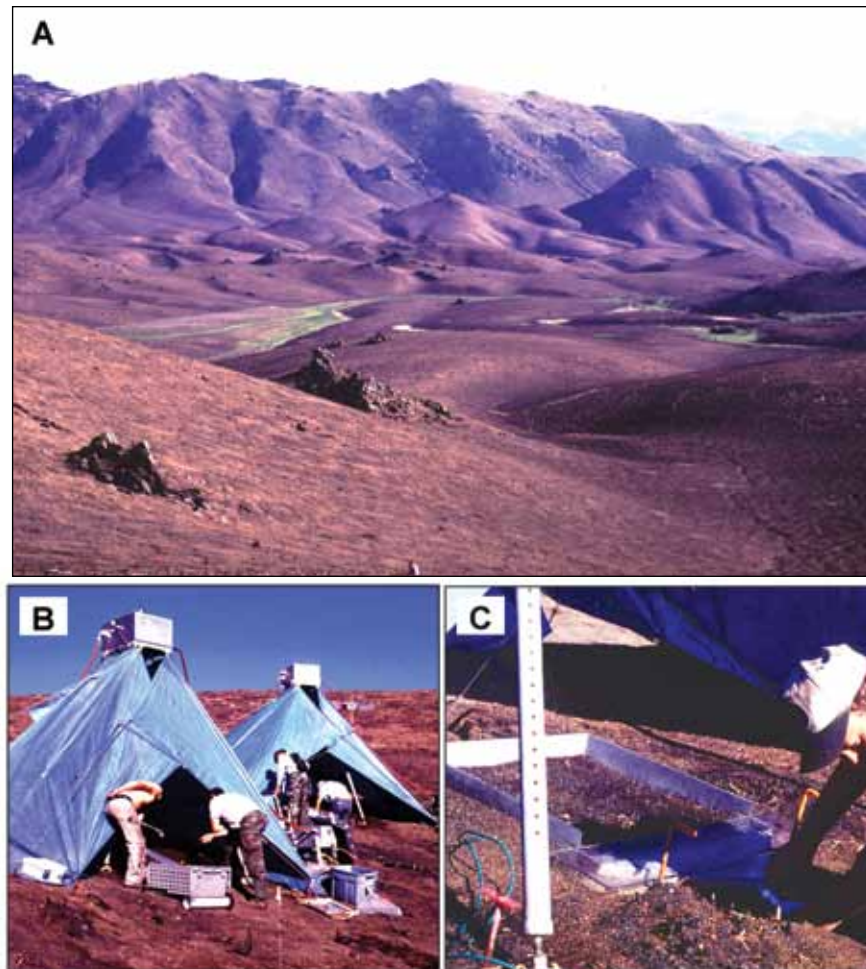
Figure 24—(A) The Denio Fire (wildfire), 1999, Pine Mountain Range, Nevada, (B) rainfall simulations, and (C) 0.5 m plots used in rainfall simulation studies at the site (Pierson et al. 2001a, 2008a,b) (photos: USDA Agricultural Research Service, Northwest Watershed Research Center).

interspace areas was associated with removal of water-shedding senescent vegetation and fire-reduced strength of soil water repellency. The influence of soil water repellency on the overall hydrologic response is shown by nearly equal runoff coefficients (10 to 20 percent) and significantly reduced strength (by 55 percent) of soil water repellency in year 1 across all plots, regardless of differences in canopy and ground cover (Pierson et al. 2008a,b). A subsequent significant increase in soil water repellency across all plots in year 2 coincided with a twofold to threefold increase in runoff coefficients for burned interspaces and all coppices and the reestablishment of differences in minimum infiltration between burned and unburned coppices (Pierson et al. 2008a,b). In summary, canopy and ground cover removal dictated water availability whereas the strength of soil water repellency exerted greater influence on infiltration processes and runoff generation (Pierson et al. 2008b).

Pierson et al. (2008b, 2009) (table 3, figs. 22B and 25) measured infiltration and runoff from rainfall simulations on small (0.5 m^2) plots on burned and unburned mountain big sagebrush rangeland (35 to 50 percent slopes) the year of, and 1 year following the Breaks Prescribed Fire in the Reynolds Creek Experimental Watershed near Boise, Idaho. A convective storm was simulated by applying 85 mm h^{-1} over a 60-min duration.

Canopy cover prefire was 84 percent on shrub coppices (almost all as shrub cover) and 31 percent on interspaces (22 percent grass). Ground cover was composed almost entirely of litter and totaled 75 percent for unburned interspaces and 99 percent for unburned shrub coppices. The fire reduced canopy cover to 10 percent on coppices and 0 percent on interspaces. Litter cover was reduced to 36 percent and 14 percent for coppices and interspaces, respectively. After one growing season (year 1), total canopy cover increased to 40 percent (29 percent forb, 3 percent grass) on burned coppices and 63 percent (40 percent forb, 22 percent grass) on burned interspaces. Total ground cover in year 1 was 22 percent (15 percent litter) on burned coppices and 32 percent (26 percent litter) on burned interspaces.

Bulk densities (0 to 2 cm) in the Pierson et al. (2009) study were equal (1.13 g cm^{-3}) on coppice and interspace microsites and were not significantly changed by burning. Soil water repellency was strong and nearly equal on coppice microsites for burned and unburned conditions the year of the fire (year 0). The strength of soil water repellency was 55 to 60 percent less on burned and unburned interspaces than on coppices in year 0, but was strong. Soil water repellency was greatly reduced (moderately strong) and nearly equal across all plots in year 1 (fig. 22B). Canopy and ground cover removal and strong soil water repellency on burned coppice plots generated significant decreases in infiltration and increases in runoff postfire (Pierson et al. 2009). Minimum infiltration decreased 60 percent and runoff doubled on coppice microsites immediately postfire (runoff coefficient of 76 percent). Runoff coefficients were greater for unburned interspaces (63 percent) than for coppices (39 percent) even though soil water repellency was twofold to threefold greater on coppices. Vegetation and ground cover mitigated soil water repellency on unburned coppices.

The low cover on unburned interspaces and the presence of strong soil water repellency facilitated runoff generation. Burning reduced cover on interspace microsites, but did not increase runoff generation relative to the unburned condition. A significant decrease (by 70 percent) in soil water repellency on burned and unburned coppices between year 0 and year 1 and the nearly uniform moderate soil water repellency across all plots resulted in a 75 to 90 percent decrease in cumulative runoff. Runoff coefficients in year 1 were 8 to 10 percent across all plots regardless of cover. Total ground cover in year 1 was significantly less on burned plots (27 percent) than on unburned plots (73 percent) and canopy cover was significantly less on burned (40 percent) than unburned (84 percent) coppices. As with the Denio Fire study (Pierson et al. 2001a, 2008a,b), cover influenced water availability, but the strength of soil water repellency exerted a greater influence on infiltration (fig. 22B) and runoff generation (table 3). That is, cover influenced the quantity of water available for runoff generation and soil water repellency, and along with other soil factors, governed the rate of infiltration for available surface water.

The collective studies by Pierson et al. (2001a, 2002b, 2008a,b, 2009) illustrate the complexity of rainfall-runoff processes on burned and unburned rangelands and are supported by other recent woodland and well-cited semiarid forested studies (Cerdà 1998a; Benavides-Solorio and MacDonald 2001, 2002; Pierson et al. 2013, 2014; Williams et al. 2014a). Recent small-plot scale investigations at a gently sloping western juniper woodland in Idaho found that severe burning had minimal effect on runoff from mostly bare interspaces and decadent shrub coppices, but fire-induced litter removal on tree coppices increased runoff of simulated rainfall by twofold to threefold 1 and 2 years postfire

(table 3; Pierson et al. 2013; Williams et al. 2014a). The wildfire reduced tree coppice litter cover by a factor of eight, from 98 to 12 percent, and soils underneath tree canopies were strongly water repellent prefire and postfire. Soils were wettable in interspaces and underneath shrubs before and after burning. Pierson et al. (2014) reported contrasting fire impacts on runoff responses to simulated high-intensity storms 1 and 2 years after low- to moderate-severity prescribed burns on degraded singleleaf pinyon (*P. monophylla* Torr. & Frém.)/Utah juniper (Marking Corral site, Nevada) and Utah juniper (Onaqui site, Utah) woodlands (table 3). Burning at both sites had minimal impact on runoff generation from degraded interspaces (table 3). The prescribed burn at the more degraded woodland, Onaqui, did not significantly reduce litter depths over the strongly water-repellent soils on tree coppices and resulted in no significant fire effect on runoff generation. In contrast, burning at the less degraded site, Marking Corral, reduced tree coppice litter depth twofold, from 40 to 23 mm, and increased runoff coefficients from 0 to 15 and 28 percent for storms of 64 mm h^{-1} and 102 mm h^{-1}, respectively (Pierson et al. 2014). Soil water repellency was strong before and after burning on tree coppices at both sites (Pierson et al. 2014). The contrast in runoff responses to burning across the two sites was attributed to greater tree litter removal at the Marking Corral site on persistent strongly water-repellent soils (Pierson et al. 2014). The lack of fire effects on runoff generation from interspaces and shrub coppices at the sites was attributed to the overall degraded conditions of interspaces before burning and the minimal impact of the moderate-severity fire on ground cover and soil properties underneath shrub canopies (Pierson et al. 2014).

Benavides-Solorio and MacDonald (2001, 2002) (table 3) measured runoff from burned and unburned areas of a ponderosa pine forest in the Colorado Front Range, Colorado. A simulated convective rainstorm applied 80 mm h^{-1} of rainfall for 60 min to 1.0 m^2 plots with unburned and varying burned severity conditions. Although the study investigated three fire sites, we focus on the primary site, the Bobcat Fire, because only two replicates existed for each burn severity at other sites (Benavides-Solorio and MacDonald 2001, 2002). Percent ground cover by burn severity class was 23 percent high-severity, 88 percent moderate-severity, and 99 percent low-severity and unburned conditions. Slopes averaged 22 to 31 percent and soil moisture averaged 1.0 to 2.0 percent. Runoff coefficients were 66 percent for the high-severity plots, 58 percent for moderate-severity plots, and 55 percent for the low-severity and unburned plots. Soils were strongly water-repellent near the surface and there was very little variation in the strength of soil water repellency with burn severity. Correlations in runoff and soil water repellency were not separated for the multiple fires studied, but when all fires were considered, runoff from high-severity plots was well correlated ($r^2 = 0.81$) with the strength of natural or fire-enhanced soil water repellency. Runoff was not well correlated with percent slope or percent bare ground. Benavides-Solorio and MacDonald (2001, 2002) concluded soil water repellency and soil moisture, as a controller of repellency strength, were the primary controls on the amount of runoff generated.

Cerdà (1998a) used rainfall simulation to study the hydrologic impacts of burning on moderately steep (25 percent) slopes within a Mediterranean scrubland plant community in southeastern Spain. A simulator applied 55 mm h^{-1} of rainfall to 0.25 m^2 plots 7, 18, 29, and 64 months after a wildfire. The site burned at high intensity, and the fire consumed the entire litter layer. Organic matter, soil porosity, and bulk density were similar across all plots. Ash cover postfire was 3 to 5 cm deep and continuous, but was removed

by autumn rains before simulation. Total ground cover was about 31 percent during the first year of simulations and was well established (74 percent) within 18 months. Time to ponding gradually increased with vegetative recovery from 2 min 17 sec the year of the fire to 15 min 35 sec 5.5 years postfire. Time to runoff increased over the same time period from 3 min 47 sec to 25 min 30 sec. The runoff coefficient was 45 percent the year of the fire, decreased to 14 percent within the first 3 years, and was 6 percent within 5.5 years postfire. The spatial variability in runoff tracked with spatial variability in vegetative recovery. Cerdà (1998a) found that the relationship between cover and runoff was negative and exponential, as reported for erosion by Johansen et al. (2001) and Pierson et al. (2008a, 2009). Runoff was greatly diminished when plant cover recovered to 50 to 60 percent, the third winter following the burn, and approximated that of other, unburned Mediterranean scrub 5.5 years postfire.

In another forested study, Woods and Balfour (2008) measured runoff from ash- and nonash-covered plots following high-severity wildfire. Rainfall was applied to 0.5 m^2 plots at 75 mm h^{-1} intensity for 1 hour. The study found that the ash layer provided 1.5 cm of water storage capacity and protected the soil surface from sealing in the immediate postfire period. Time to ponding was 12 min longer and cumulative infiltration was 2.0 cm greater on ash than on ash-free plots. Nine months after the fire, ash and ash-free plots exhibited similar runoff behavior. Similar ash cover and runoff relationships have been reported in studies by Cerdà and Doerr (2008), Larsen et al. (2009), and Ebel et al. (2012). Larsen et al. (2009) and Onda et al. (2008) indicate, however, that the positive ash and infiltration relationships are likely short lived and that soil sealing following winnowing of ash particles may promote runoff, especially where soil water repellency occurs.

Hillslope- to Watershed-Scale Runoff

Fire effects on runoff generation over large-plot (tens of meters) to hillslope scales is commonly less than observed at small-plot scales due to spatially variable cover and surface soil conditions (Shakesby and Doerr 2006; Pierson et al. 2009, 2011). This relationship is illustrated in the aforementioned Pierson et al. (2009) (table 3) study. In addition to small plots, the study measured runoff from large plots (32.5 m^2) (fig. 25B) in burned and unburned shrub steppe. Runoff coefficients from a simulated convective storm (85 mm h^{-1}, 60 min) were 27 percent for burned plots and 4 percent for unburned plots (table 3). Mean runoff coefficients from the same storm on burned and unburned small plots (0.5 m^2) were 2- and more than 10-fold greater, respectively, than on the large plots (table 3). The decreased runoff from small- to large-plot scales was attributed to greater spatial variability in infiltration at the large- versus small-plot scale. Although runoff declined with increasing scale, burned large plots still generated nearly sevenfold more runoff than unburned large plots. Greater runoff from the burned than unburned condition was attributed to the uniform threefold reduction in ground cover, 100 percent reduction in canopy cover, persistence of strong soil water repellency after burning, and formation of high-velocity concentrated flowpaths on burned plots. These conditions created more uniform surface hydrologic connectivity on burned than unburned plots (fig. 25A). Runoff coefficients from all plots were positively correlated with soil water repellency strength ($r^2 = 0.56$) and percent bare ground ($r^2 = 0.32$). Concentrated flow experiments in the same study (fig. 26) found that cumulative runoff from consecutive
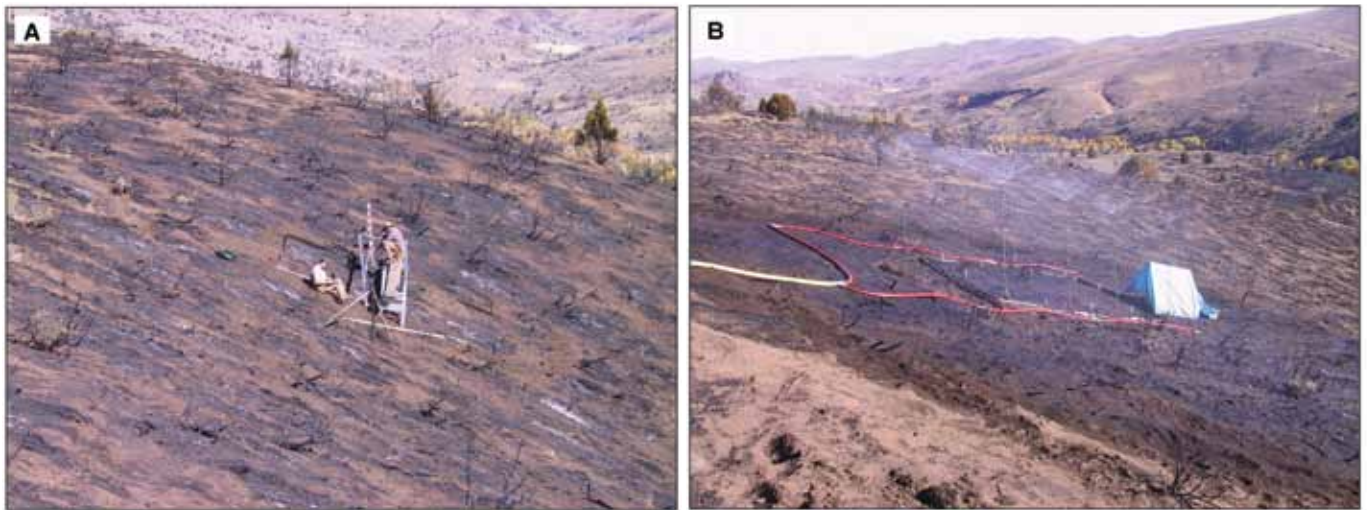
Figure 25—(A) The Breaks Prescribed Burn, Reynolds Creek Experimental Watershed, Idaho and (B) large-plot (32.5 m^2) rainfall simulations conducted at the site by Pierson et al. (2009) (photos: USDA Agricultural Research Service, Northwest Watershed Research Center).

12-min releases of 7, 12, 15, and 21 L min^{-1} of concentrated flow were 406 L on burned plots and 144 L on unburned plots. Runoff from small, large, and concentrated-flow burned plots was equivalent to that of unburned plots after ground cover recovered to 30 to 40 percent one growing season postfire. The return to prefire runoff rates was at least partially related to a 70 percent reduction in strength (from strong to moderate) of soil water repellency.
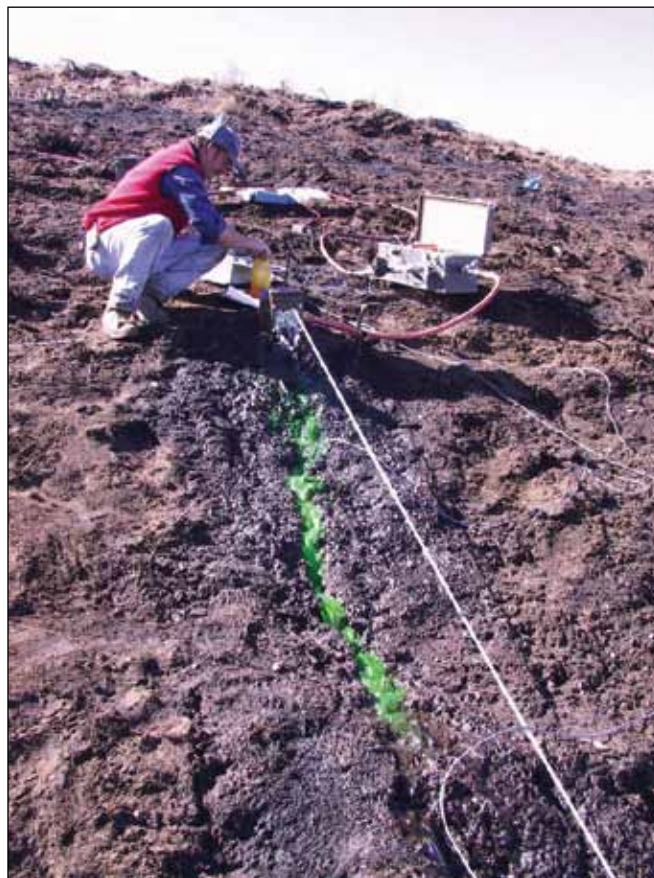
Figure 26—Concentrated flow experiments conducted by Pierson et al. (2009) immediately following the Breaks Prescribed Fire, Reynolds Creek Experimental Watershed, Idaho. Green dye in flowpath is a visual tracer used to show concentrated flowpath formation and direction. Flows were released at specified rates (7, 12, 15, and 21 L min^{-1}) for 12-min intervals each (photo: USDA Agricultural Research Service, Northwest Watershed Research Center).

In another mountain big sagebrush study, Pierson et al. (2008a) reported a 35-fold difference in runoff between high-severity burned (178 L) and unburned (5 L) plots from consecutive 12-min concentrated flow releases of 7, 12, and 15 L min^{-1}. Runoff from small plots and concentrated flow experiments in that study returned to preburn levels after ground cover increased to about 40 percent and 60 percent, respectively, and soil water repellency was reduced to moderate levels. At a western juniper woodland in Idaho, Pierson et al. (2013) and Williams et al. (2014a) found runoff ratios during simulated 64 mm h^{-1} and 102 mm h^{-1} intensity rainstorms (45 min, 13 m^2 plots) on tree coppices increased ninefold and fourfold, respectively, 1 year after a high-severity wildfire (table 3). The increased runoff from tree coppices following the fire occurred due to a threefold reduction of litter cover, a reduction of more than 4 to 5 cm in litter depth, and persistence of strongly water-repellent soils postfire (Pierson et al. 2013; Williams et al. 2014a). Burning had no effect on runoff generation from the degraded intercanopy (more than 80 percent bare soil and rock) at the site the first year after burning.

In a semiarid forest setting in New Mexico, Johansen et al. (2001) (table 3) reported that runoff from large-plot (32 m^2) rainfall simulations on burned and unburned plots was positively correlated with percent bare soil ($r = 0.76$), and that the time to runoff generation was negatively correlated with percent bare soil ($r = -0.67$). Application of 120 mm of rainfall over 2 simulation hours generated mean runoff coefficients of 45 and 23 percent for burned and unburned conditions, respectively (table 3). Mean ground cover was 26 percent for burned plots and 52 percent for unburned plots. Soil water repellency was highly variable in space and strength and had minimal effect on runoff generation.

Historical hillslope-scale studies from chaparral communities (see review by Shakesby and Doerr 2006) are consistent with the semiarid shrub steppe, woodland, and forested studies by Pierson et al. (2009, 2013), Williams et al. (2014a), and Johansen et al. (2001). The effects of fire on runoff from rangelands at the watershed scale are limited. Overall, at the watershed scale, peak discharge rather than cumulative runoff tends to be greater after burning, and is most pronounced after short-duration, high-intensity, convective thunderstorms over large expanses of severely burned landscapes (Shakesby and Doerr 2006).

Fire Effects on Soil Erosion by Water

Rainsplash and Sheetflow Processes

Higher erosion rates following fire are commonly attributed to decreased aggregate stability, increased surface exposure to raindrop impact, reduced energy dissipation of overland flow, and greater surface runoff and overland flow velocity. Greater surface vulnerability and maximized raindrop energy amplify sediment detachment by rainsplash processes. Greater water availability and flow velocity provide more efficient downslope transport and overland flow detachment. Reduced aggregate stability, where it occurs, magnifies the overall response. The previously mentioned small-plot studies in Nevada and Idaho by Pierson et al. (2001a, 2002b, 2008a, 2009, 2013, 2014) and in the Colorado Front Range by Benavides-Solorio and MacDonald (2001, 2002) provide estimates of postfire erosion increases due to rainsplash and sheetflow processes. Pierson et al. (2002b) (table 3) reported that the more than 90 percent reduction in litter biomass and 70

percent mean reduction in total live biomass 1 year after moderate and severe burning of south-facing sagebrush hillslopes increased soil erosion by factors of 7 (21 g m^{-2}) and 28 (85 g m^{-2}), respectively. Soil erosion postfire was most pronounced on severely burned interspace plots (148 g m^{-2}) due to a more than 40 percent reduction in surface roughness. Erosion from severely burned shrub coppice plots was 22 g m^{-2}. Sediment yield from all plots in the study was negatively correlated with litter biomass ($r = -0.59$), total ground cover ($r = -0.59$), and total canopy cover ($r = -0.56$).

Pierson et al. (2001a, 2008a) (table 3; fig. 24) measured a threefold increase in soil erosion from shrub coppice plots (41 g m^{-2} burned, 12 g m^{-2} unburned) after fire consumed nearly 100 percent of canopy and ground cover. Similar bare ground postfire on burned interspaces generated half the erosion rate that was measured on burned shrub coppices and produced similar erosion to prefire conditions. The differing responses were attributed to a more erodible surface, slightly greater runoff, and persistent soil water repellency on shrub coppices after burning. Erosion 1 year postfire was greatly reduced across all plots and was similar for burned and unburned conditions. Two years postfire, burned shrub coppices generated 3- to 14-fold more erosion than any other plot, burned or unburned. Soil water repellency, time to peak runoff, and minimum infiltration were the only other variables that exhibited the same temporal trend, implicating runoff generation and continued greater erodibility as the causal factors.

Pierson et al. (2009) found that 90 percent and 40 percent reductions in canopy and ground cover increased sediment yield from 17 g m^{-2} on unburned shrub coppices (80 to 90 percent cover) to 183 g m^{-2} postfire. Fire-caused reductions of 100 percent canopy and 80 percent ground cover on interspaces (31 percent and 75 percent canopy and ground cover prefire) increased sediment yield from 195 g m^{-2} to 705 g m^{-2}. Fire-induced increases in erosion on shrub coppices were attributed to greater runoff postfire whereas significantly increased erodibility, rather than runoff, explained the postfire increase in erosion from interspaces.

In woodland studies in southwestern Idaho, Pierson et al. (2013) and Williams et al. (2014a) reported that reductions of more than 80 percent of litter cover and 4 to 5 cm litter depth from tree coppices amplified erosion from small-plot rainfall simulations (102 mm h^{-1}, 45 min, 0.5 m^2) by 18- to 34-fold the first 2 years after severe wildfire (table 3). In the same studies, more than 80 percent reductions in canopy and litter cover on shrub coppices increased erosion from simulated storms more than 20-fold 1 year postfire. Two years postfire, erosion remained elevated on shrub coppice plots, but the difference in erosion between burned and unburned treatments was not significant. Slight fire-induced reductions in interspace grass canopy cover (from 15 to 7 percent) at the site 1 year postfire resulted in a fourfold increase in small-plot scale erosion from burned interspaces. As with shrub coppices, erosion from burned and unburned interspaces was not significantly different 2 years postfire.

Pierson et al. (2014) cited site-specific differences in soil erodibility as the primary driver for contrasting erosion rates from simulated rainfall at two woodland sites 1 and 2 years following prescribed fire (table 3). Simulated high-intensity rainfall (102 mm h^{-1}, 45 min, 0.5 m^2) generated amplified soil erosion from tree coppices at the less erodible site (46 to 75 g m^{-2}), but the elevated postfire erosion rates were minor relative to those of burned tree coppices at the more highly erodible site (242 to 294 g m^{-2}). Runoff rates from burned tree coppices were similar for the two sites (table 3).

Benavides-Solorio (2001, 2002; table 3) reported that sediment yield from moderate- and high-severity burned semiarid forested areas were 2- and 16-fold higher than from plots burned at low severity or unburned. Percent bare soil (table 3) explained 79 percent of the variability in erosion from all plots, and soil water repellency explained 43 percent of the variability in erosion from plots burned at high severity.

As with runoff, the postfire erosion response appears closely tied to alteration of cover. A secondary contributing factor appears to be soil water repellency, owing to its influence on infiltration and on runoff generation. In each of the studies discussed above, rainfall intensity was held constant. Runoff and erosion both exhibit some dependence on rainfall intensity as a driving force, and responses may be amplified or dampened by respective increases or decreases in the rainfall rate (Pierson et al. 2011; Williams et al. 2014b).

Concentrated Flow Processes

Concentrated flowpaths rarely occur on undisturbed rangelands, but often become the dominant conduit for overland flow and sediment transport after burning (Moffet et al. 2007; Pierson et al. 2009; Williams et al. 2014b, 2016a). Following burning, greater water availability, decreased infiltration, and reduced surface obstructions facilitate formation of concentrated flowpaths. These relationships are enhanced on steep slopes and where overland flow is promoted by soil water repellency (Shakesby and Doerr 2006; Moffet et al. 2007; Pierson et al. 2009; Williams et al. 2014a,b). Concentrated flowpaths have greater flow depth, velocity, erosive energy, and transport capacity than sheetflow (Moffet et al. 2007; Pierson et al. 2008a, 2009). The velocity and erosive energy are a function of the soil grain and form roughness, the eroded flowpath, ground cover, hillslope angle, and flow discharge (Moffet et al. 2007; Al-Hamdan et al. 2012a,b, 2013).

Pierson et al. (2008a, 2009) (figs. 24, 25, and 26) measured the velocity of simulated concentrated flowpaths under burned and unburned conditions on sagebrush rangelands in Idaho and Nevada. In both studies velocity over a range of flow rates was positively correlated (r^2 = 0.59 to 0.70) with and exponentially related to percent bare ground. Velocity decreased non-linearly with decreasing bare ground as vegetation recovered in the years after burning. Sharp increases in velocity were observed where bare ground exceeded 50 to 60 percent. Pierson et al. (2009) (table 3) measured increasing erosion with increasing plot size after burning of sagebrush hillslopes. They attributed the positive relationship to a switch in the dominant erosion process from rainsplash-sheetflow to concentrated flow following burning. In the study by Pierson et al. (2009), erosion for unburned conditions was greater from small plots (rainsplash-sheetflow) than large plots. Following burning, erosion was greater on the larger plots, and the maximum number of concentrated flowpaths increased from zero for unburned to five for burned large plots. More uniform bare soil conditions, strongly water-repellent soils, and greater hydrologic connectivity after fire contributed to rapid (3 min) runoff generation and formation of concentrated flow. Erosion increased from 8 g m^{-2} for unburned conditions to 988 g m^{-2} postburn (table 3). Large-plot sediment yield remained greater for burned conditions until vegetation recovered to nearly 60 percent two growing seasons postfire (fig. 27). No concentrated flowpaths were observed in burned plots after the second growing season.
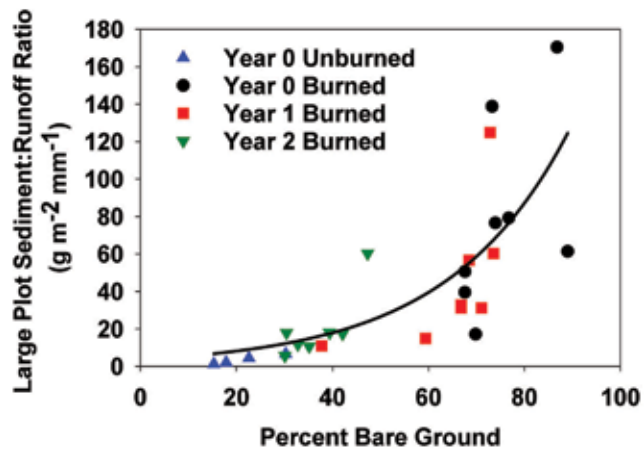
Figure 27—Erosion reported as sediment per unit of runoff from large-plot (32.5 m^2) rainfall simulations conducted on unburned and burned conditions at the Breaks Prescribed Fire, Reynolds Creek Experimental Watershed, Idaho (Pierson et al. 2009).
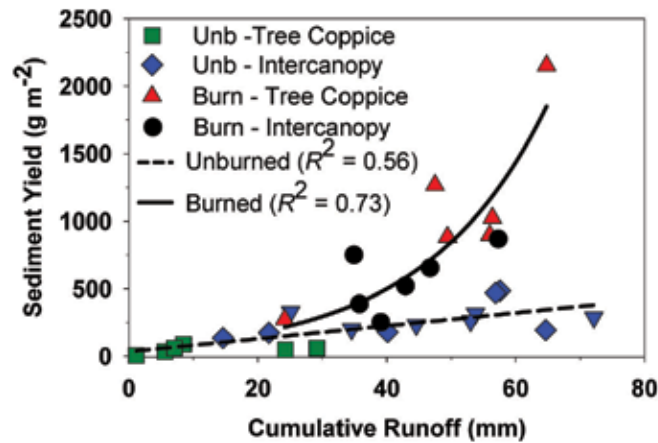

Figure 28—Cumulative sediment yield versus runoff from rainfall simulations (102 mm h^{-1}, 45 min) on unburned and burned tree coppice and intercanopy (areas between tree coppices) plots (13 m^2) in a western juniper woodland (Pierson et al. 2013; Williams et al. 2014a).

Williams et al. (2014a) and Pierson et al. (2013) also reported a shift from rainsplash and sheetflow to concentrated flow as the dominant erosion process occurring on 13 m^2 rainfall simulation plots following burning of a western juniper woodland (table 3). Erosion from burned tree coppices increased exponentially with increasing runoff postfire due to the formation of high-velocity concentrated flow paths (fig. 28). Pierson et al. (2009) used simulated consecutive 12 min concentrated flow releases of 7, 12, 15, and 21 L min^{-1} to quantify erosion solely from concentrated flow processes in unburned and burned areas of sagebrush rangeland. The cumulative flow releases generated 14,363 g of sediment immediately postfire and 2,420 g under unburned conditions prefire. Pierson et al. (2008a) conducted the same concentrated flow experiments as Pierson et al. (2009) using flow rates of 7, 12, and 15 L min^{-1} and measured 17,775 g of sediment immediately following burning and less than 10 g on unburned plots. In both studies, erosion from the respective simulated concentrated flow rates on burned plots approached that of unburned plots once ground cover recovered to 60 percent or greater. The large-plot results from Pierson et al. (2009, 2013) and Williams et al. (2014a) illustrate the profound influence that concentrated flow processes have on postfire erosion (table 3). The simulated concentrated flow experiments (Pierson et al. 2008a, 2009) demonstrate the erosive potential of high-velocity concentrated flowpaths on burned rangelands.

Hillslope- to Watershed-Scale Erosion

Erosion at hillslope to watershed scales is largely dependent on the spatial arrangement of burn severity, bare soil exposure, water-repellent soil conditions, and rainfall intensity (Shakesby and Doerr 2006). Very few large-scale erosion studies have been completed on burned rangelands. Therefore insight into large-scale fire effects on rangeland hydrology comes from studies of dry or semiarid forested sites and the limited number of hillslope-scale studies for shrub steppe, woodland, and chaparral communities. Benavides-Solorio and MacDonald (2005) used silt fences to measure (see Robichaud and Brown 2002) postfire erosion on moderate to steep, semiarid forest slopes (25 to 45 percent) in the Colorado Front Range for multiple fires of varying ages and severities. Contributing areas for the silt fences ranged from 190 to 6,600 m^2 and averaged 1,250 m^2. Over the 2-year study, percent bare soil explained about 64 percent of the variability in soil erosion across all plots ($n = 48$). Sediment yield decreased exponentially with time after burning, and was highest where bare soil was equal to or greater than 60 percent. After a high-intensity storm, however, extensive concentrated flow networks and erosion were observed on slopes with 70 to 85 percent ground cover of tree needles (15 to 30 percent bare). Benavides-Solorio and MacDonald (2005) attributed the result to the high storm erosivity and ground cover type. Soil water repellency was present at least at slight strength on most plots and was moderately strong on some of the high-severity burns. As a result, soil water repellency was weakly correlated with sediment production from all plots ($r^2 = 0.30$), but was more strongly correlated for the high-severity plots ($r^2 = 0.40$). About 90 percent of the sediment collected during the study was delivered by high-intensity convective storms whereas 10 percent resulted from frontal rainfall or snowmelt. Concentrated flow formation played an important role in postfire erosion rates, particularly in converging topographic positions. Overall sediment yield was well correlated ($r^2 = 0.77$) with a five-parameter empirical model of percent bare soil, rainfall erosivity, fire severity, soil water repellency (1 cm depth), and the 84th percentile sediment grain size. Percent bare soil and rainfall erosivity collectively explained 62 percent of sediment production variability.

A large-plot (32.5 m^2) rainfall simulation study of recently burned and unburned rangeland in southwestern Idaho by Pierson et al. (2009) found soil water repellency and sediment yield were moderately correlated ($r^2 = 0.46$) whereas runoff coefficients (table 3) were more strongly correlated with soil water repellency ($r^2 = 0.56$) and weakly correlated with bare ground ($r^2 = 0.32$). Sediment yield and repellency in the Pierson et al. (2009) study was positively correlated with runoff coefficients ($r^2 = 0.83$) and with percent bare ground ($r^2 = 0.76$). These correlations suggest that sediment yield in the study by Pierson et al. (2009) was dependent on runoff generation and bare soil whereas runoff was more dependent on water-repellent soil conditions in bare areas. The low correlation in sediment yield and repellency in the Benavides-Solorio and MacDonald (2005) study indicates that either the strength of soil water repellency was not significant enough to influence erosion or simply other factors like rainfall intensity or topography and extensive bare ground exerted more influence on erosion processes. Soil water repellency in the contrasting Pierson et al. (2009) research was strong at the soil surface immediately postfire and deteriorated to slight by the end of the 3-year study. Soil water repellency was much weaker at the sites studied by Benavides-Solorio and MacDonald (2005) and

showed a general decreasing trend through the study. In a gently sloping (5 to 7 percent) semiarid forest setting in New Mexico, Johansen et al. (2001) (table 3) found erosion from high-severity burned large plots (32.5 m^2) was 25-fold greater than measured from unburned plots of the same size. Sediment yield was strongly correlated with percent bare soil ($r = 0.84$), and soil water repellency was considered slight. The studies highlighted here illustrate the influence of ground cover and bare ground on hillslope-scale postfire erosion response over a range of settings and slopes and demonstrate how soil water repellency may magnify those effects.

Large-scale alteration of surface roughness, canopy, and ground cover on rangeland soils can trigger flash flooding and mass erosion events where superimposed on steeply sloping water-repellent hillslopes. These events have received less attention in literature for rangelands than for forested settings. One year after the Eighth Street Fire (6,070 ha) in 1996 near Boise, Idaho, a 5- to 10-year return period convective storm (~67 mm h^{-1}) of 9-min duration on steeply sloped, severely burned sagebrush caused flash flooding and mud flows in the city of Boise (Pierson et al. 2002b). Pierson et al. (2002b) (table 3) conducted rainfall simulations (67 mm h^{-1}) at the site preceding the storm and determined that elevated runoff and erosion rates from small (0.5 m^2) rainfall plots were associated with extensive bare ground, reduced surface roughness, and soil water repellency. These effects were greater for south-facing than north-facing slopes. Runoff occurred within 2 to 4 min of rainfall in the Pierson et al. (2002b) study. Flooding during the ensuing storm was driven by intense rainfall on bare (90 to 100 percent bare ground), water-repellent soils with reduced water storage capacity and low surface roughness. Most of the rainfall falling on the south-facing slopes ran off, forming concentrated flow networks (Pierson et al. 2002b).

Moody and Martin (2001) described a similar response to 100-year rainfall storm following the 4,690 ha Buffalo Creek Fire in steep, forested watersheds of the Colorado Front Range. More than 60 percent of the burn was high intensity. Within 2 months postfire, a high-intensity (90 mm h^{-1}, 1 h) rainstorm caused flash flooding that killed two people and discharged enough sediment into the Strontia Springs Reservoir to reduce storage capacity by one-third (Agnew et al. 1997; Moody and Martin 2001). Unburned hillslopes adjacent to the fire generated very little surface runoff (Elliott and Parker 2001). Hillslope erosion increased 150- to 240-fold postfire. Nearly 1,500,000 m^3 of sediment was generated from interrill, rill, and in-channel processes during the first summer after the fire, and an estimated 86 percent of erosion from the first two summers postfire was from rill and channel processes (Moody and Martin 2001). Soil water repellency was not reported for the Buffalo Creek Fire.

A torrential rainstorm 2 months after the 800 ha South Canyon Fire in Colorado generated multiple runoff-triggered debris-flow events that inundated a 13- to 14-ha area with about 70,000 m^3 of soil and engulfed 30 vehicles on an adjacent highway (Cannon et al. 1998, 2001a). The fire occurred on steeply sloping (30 to 70 percent) woodland and shrub-dominated hillslopes. Cannon et al. (1998) estimated that rainsplash, sheetflow, and concentrated flow processes during the storm removed 15 percent of the mineral soil surface to a depth of 4 cm. Amplified runoff and erosion from rainsplash and sheetflow on bare soils led to formation of concentrated flow networks and gullies with high erosive energy and sediment transport capacity. Erosion per unit of runoff generally increased downslope as flow paths incised and widened to 30 to 40 cm (Cannon et al. 2001a). The

examples presented here demonstrate the potential hydrologic responses to high-intensity rainfall after large-scale alteration of vegetation and ground surface characteristics. A more detailed chronicle of hillslope- and watershed-scale responses to fire in forested and chaparral environments can be found in Shakesby and Doerr (2006).

Assessing Postfire Hydrologic Vulnerability and Risk

The overall postfire hydrologic vulnerability of a site is a function of the site susceptibility and prevailing climatic regime (fig. 29). Susceptibility is defined by the vegetation, soil, and topographic characteristics of the site and is much greater where bare ground is extensive on steep slopes with hydrophobic surface soil conditions (Pierson et al. 2011; Williams et al. 2014b). In contrast, susceptibility to postfire runoff and erosion is low where ground cover exceeds 60 percent (Pierson et al. 2008a, 2009, 2013; Williams et al. 2014a). Susceptibility for a specified area varies in both time and space and depends on the prefire conditions and the rate of postfire recovery. Hydrologic vulnerability for a specified susceptibility can be measured as the hydrologic response (runoff, erosion, or both) to different storm events or rainfall intensities (Williams et al. 2014b). Hydrologic vulnerability for a given susceptibility then increases with increasing rainfall intensity, as shown by the different curves in figure 29. Each curve in the figure represents a hypothetical vulnerability (or hydrologic response) associated with a particular storm intensity over a range of susceptibilities, with susceptibilities defined by the site conditions. For example, hydrologic vulnerability for a low-intensity storm on highly susceptible
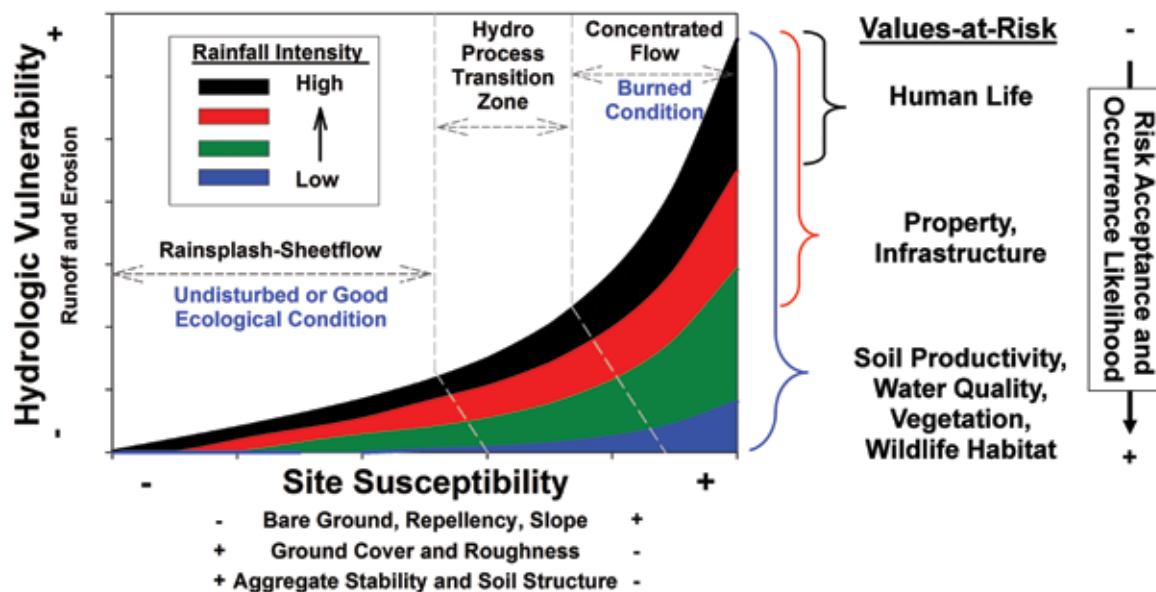

Figure 29—Conceptual model of hydrologic vulnerability and risk. Hydrologic vulnerability (measured as runoff and erosion response, Y-axis) is a function of site susceptibility (X-axis) and the prevailing rainfall intensity (indicated by colors). The level of vulnerability dictates the resources at risk. Concentrated flow processes dominate the postfire environment where site susceptibility is high (e.g., high bare ground, water-repellent soils) or rainfall is of moderate to high intensity. Rainsplash processes prevail where susceptibility is low (well-aggregated soils, land surface well protected by litter cover). Rainsplash, sheetflow, and concentrated flow processes all contribute to the runoff/erosion response where site susceptibility is indicative of disturbed conditions (reduced ground cover and aggregate stability, poor soil structure). The overall hydrologic response is amplified with increasing slope steepness. Figure modified from Pierson et al. (2011) and Williams et al. (2014b).

conditions is quite low. But hydrologic vulnerability is high for at least a moderate-intensity storm falling on a highly susceptible site (such as a site with more than 60 percent bare ground and water-repellent soils). The hydrologic vulnerability is then dictated by the storm intensity and the conditions in which the storm occurs (susceptibility).

Concentrated flow processes dictate the response when storm intensity and site susceptibility are both high postburn. Rainsplash processes dominate and overall vulnerability is minimal where site susceptibilities are low, regardless of the rainfall intensity. A combination of rainsplash, sheetflow, and concentrated flow processes may prevail where moderate rainfall intensities fall on a moderately susceptibility condition such as on a degraded site. The overall hydrologic vulnerability further defines the resource at risk (fig. 29) (Williams et al. 2014b). High-intensity rainfall on moderate to highly susceptible conditions may cause loss of life (Moody and Martin 2001) or damage to property or infrastructures (Pierson et al. 2002b; Klade 2006; Pierson et al. 2011) due to large flooding or debris flow events, or a combination thereof (Cannon et al. 1998, 2001a; Meyer et al. 2001). Low-intensity storms on highly susceptible conditions may not generate massive flooding, but may generate enough soil loss to degrade water quality or negatively affect aquatic habitat. In the framework presented (fig. 29), hydrologic vulnerability curves can be populated with quantitative data, but the mitigation of risk is value based. Land managers must consider what hydrologic vulnerability they are willing to accommodate given the potential risks with which it is associated.

Great advances have been made in postfire risk assessment in recent decades (Robichaud et al. 2007a,b; Pierson et al. 2011; Williams et al. 2014b; Al-Hamdan et al. 2015; Williams et al. 2016a,b). The Erosion Risk Management Tool (ERMiT) is one recent model developed with a probabilistic framework like the one shown in figure 29 (Robichaud et al. 2007a,b). Models like ERMiT predict the likelihood of a runoff/ erosion response associated with a given storm as the probability of that storm occurrence given a static susceptibility. The probability of a given storm can be determined from return interval or intensity/duration/frequency (fig. 3) information for the area of interest. ERMiT and similar models use climate generators to predict storm frequency and magnitude from a climate station location identified by the user. This approach allows the user to evaluate hydrologic vulnerability in a predictive way and to make mitigation decisions based on what may be at risk given a particular storm under a particular set of site conditions. For example, simulations for defined postfire conditions may show there is a 20 percent likelihood of a particular storm and an associated infrastructure-damaging flood event occurring. Managers must then decide whether they are willing to assume that type of risk, and if not, what could be used to mitigate the response if the event occurs. This, of course, becomes more complicated if a user is interested in predicting potential response for varying conditions or hydrologic recovery with time or in landscape space. Predictions of this type require the user to run multiple model scenarios, one for each set of postfire conditions and the same storm frequencies.

The Rangeland Hydrology and Erosion Model (RHEM) has been developed specifically for modeling runoff and erosion from rangelands (Nearing et al. 2011) and is applicable for predicting runoff and erosion responses on postfire conditions (Al-Hamdan et al. 2015; Williams et al. 2016a,b). The RHEM model includes optional probability-based risk assessment output. RHEM model results for different vegetation and ground cover conditions (e.g., prefire, immediately postfire, 3 years postfire) can be displayed

simultaneously to assess hydrologic vulnerability across varied surface susceptibility (Williams et al. 2016b). In summary, postfire hydrologic vulnerability assessment for large areas or over periods of time benefit from incorporating conceptual and mathematical models that consider both the probability of storm occurrences and the likelihood of respective storm occurrences over a range of postfire site conditions.

The need to address hydrologic vulnerability and risk mitigation on western rangelands is increasing due to the increased role of wildfire (Pierson et al. 2011; Williams et al. 2014b). Annual weed invasion of shrub steppe (Mack 1981; Whisenant 1990; Knapp 1996) and woodlands (Tausch 1999; Miller et al. 2008; Romme et al. 2009), woodland encroachment (Miller and Tausch 2001; Romme et al. 2009), and increasing global temperatures (Westerling et al. 2006; Keane et al. 2008) are expected to amplify the role of fire on western landscapes (Williams et al. 2014b). Cheatgrass is now the major plant constituent on 4 to 7 million of the 18 million ha of shrub steppe in the Great Basin (Mack 1981; Knapp 1996). The continuous horizontal fuel structure of cheatgrass-invaded shrubland promotes more frequent and larger-scale wildland fires than reported for historical shrub steppe (Whisenant 1990; Peters and Bunting 1994; D'Antonio 2000; Brooks and Pyke 2001; Brooks et al. 2004; Keane et al. 2008). Current fire return intervals on many cheatgrass areas are 3 to 15 years (Whisenant 1990; Brooks and Pyke 2001; Brooks et al. 2004); historical fire return intervals on Great Basin sagebrush communities range from 20 to 100 or more years depending on the productivity of the site (Wright and Bailey 1982; Keane et al. 2008). Great Basin rangelands with substantial cheatgrass coverage are 10 to 500 times more likely to burn than pristine shrub-bunchgrass communities (Hull 1965), and fire risk is near 100 percent where cheatgrass coverage approaches 50 percent (Link et al. 2006).

Recent infilling of trees in persistent woodlands and wooded shrublands of the Great Basin has also increased the risk of occurrence of large, high-severity fires (Tausch 1999; Miller and Tausch 2001; Tausch and Hood 2007; Keane et al. 2008; Romme et al. 2009). Cheatgrass invasion into persistent woodlands and wooded shrublands has further increased the horizontal fuel structure and risk of large-scale fires on many wooded sites within the Great Basin (Young and Evans 1978; Billings 1994, Tausch 1999; Miller et al. 2008). Larger, more frequent, uniform, and intense fires increase the spatial expanse and temporal exposure of these landscapes to accelerated runoff and erosion. Greater temporal exposure increases both the potential soil loss from low-return interval storms (1- to 10-year events) and from more damaging, high-intensity thunderstorm events.

The environmental and ecological implications of altered fire regimes are substantial when one considers the magnitude of cumulative soil loss associated with frequently occurring storms or infrequent but extreme events. Risks to property and human life are of particular concern at the wildland-urban interface (Craddock 1946; Cannon et al. 1998, 2001a; Meyer et al. 2001; Moody and Martin 2001; Pierson et al. 2002b; Klade 2006; Pierson et al. 2011; Williams et al. 2014b). The economic consequences are significant given current U.S. expenditures on postfire risk mitigation (General Accounting Office 2003).

Determining Site Susceptibility and Hydrologic Recovery

Hydrologic susceptibility assessments should include evaluation of those variables found to significantly affect rangeland hydrology in the postfire environment (Pierson et al. 2011; Williams et al. 2014b). Our research (Pierson et al. 2001a, 2002b, 2008a,b, 2009, 2013, 2014; Al-Hamdan et al. 2012a,b, 2013; Williams et al. 2014a,b) as well as others' research (Benavides-Solorio and MacDonald 2001; Cannon et al. 1998, 2001a,b; Johansen et al. 2001; Moody and Martin 2001; Benavides-Solorio and MacDonald 2002, 2005; Larsen et al. 2009) indicate these variables include hillslope angle and topography, percent bare soil, surface soil erodibility, and soil water repellency. Although the influence of bare soil, erodibility, and soil water repellency are important across a range of hillslope gradients, the potential magnitude of the hydrologic response may be much greater for steep hillslopes (DeBano et al. 1998). Furthermore, convergent topography tends to accentuate concentrated flow (Benavides-Solorio and MacDonald 2005). Results from numerous studies indicate both runoff and erosion, more so for the latter, are strongly influenced by the expanse of bare ground (Morris and Moses 1987; Benavides-Solorio and MacDonald 2001; Johansen et al. 2001; Pierson et al. 2001a; Benavides-Solorio and MacDonald 2002; Pierson et al. 2002b; Benavides-Solorio and MacDonald 2005; Moffet et al. 2007; Pierson et al. 2008a,b, 2009; Larsen et al. 2009; Pierson et al. 2013; Williams et al. 2014a, 2016a).

For sloping rangelands, a ground cover of 50 to 60 percent is commonly adequate to protect the soil surface from amplified runoff and erosion during frequently recurring convective thunderstorm events (Gifford 1985; Pierson et al. 2008a, 2009, 2013; Williams et al. 2014a). A greater ground cover may be necessary to protect against larger, less frequent, major events (such as a 100-year storm) capable of generating concentrated flow (Benavides-Solorio and MacDonald 2005; Pierson et al. 2008a, 2009). The presence of strong soil water repellency further inhibits infiltration, and, when combined with less than 60 percent ground cover, greatly increases the vulnerability for runoff generation, concentrated flow, and elevated erosion rates (Pierson et al. 2008a,b, 2009, 2013; Williams et al. 2014a). The influence of soil water repellency on postfire runoff is well documented, but the relationships between spatial and temporal repellency strength and the respective hydrologic responses remain largely unknown. Finally, site erodibility should be considered (Pierson et al. 2010, 2014). Sites with higher erodibility may require greater ground cover protection. The generalities presented here are based on field studies, but often one or more variables (such as slope and aspect) were held constant during investigations. Therefore, the collective interaction remains largely unevaluated.

Postfire hydrologic recovery assessments are often hampered by the difficulty in selecting the appropriate reference condition, temporal fluctuations in climate, spatial variability in postfire surface and vegetation characteristics, and the fact that conditions required for hydrologic stability differ for runoff versus erosion and for rainsplash/sheet-flow versus concentrated flow processes (Pierson et al. 2008a, 2009). Large-scale plant community transitions (D'Antonio and Vitousek 1992; Knapp 1996; DiTomaso 2000; Brooks and Pyke 2001; Brooks et al. 2004; Rice et al. 2008; Romme et al. 2009) have created prefire vegetation and surface characteristics that may differ from those that favor water and soil conservation. Therefore, determination of the appropriate reference for comparison of postburn runoff and erosion may be difficult.

Annual variations in climate influence hydrologically important variables like canopy and ground cover, soil erodibility, and soil water repellency. In a 3-year study, for example, Pierson et al. (2008a,b; figs. 22 and 24) reported that temporal controls on naturally occurring soil water repellency exerted a greater influence on runoff from burned and unburned sagebrush hillslopes than did direct fire effects. Soil erosion from burned shrub coppices in the study also exhibited significant temporal variability, but the study did not determine whether this resulted from temporal variation in infiltration/runoff or erodibility. The study highlighted the need for annual controls and suggested that results from single-year studies or those without annual unburned controls may produce unreliable evaluations of hydrologic vulnerability.

Potential spatial variability of site characteristics that influence hydrology like slope, soil properties, and vegetation should be considered in recovery assessments. For example, Pierson et al. (2002b) demonstrated that runoff and erosion were significantly greater on south-facing than north-facing slopes 1 year following wildfire in sagebrush. These differences greatly influenced hydrologic behavior after an intense, flash flood-generating storm. Analysis of the north-facing slopes solely would not have captured the potential response of the more vulnerable south-facing slopes. Other fire-influenced variables like soil water repellency may vary greatly in space depending on variations in soil properties and burn characteristics (Woods et al. 2007).

Lastly, postfire recovery periods differ for runoff and erosion and for different hydrologic processes. Pierson et al. (2008a, 2009, 2013) as well as others (Benavides-Solorio and MacDonald 2001, 2002) have shown that fire-induced increases in erosion are greater than increases in runoff (table 3). Pierson et al. (2008a, 2009) further demonstrated that postfire erosion may take longer than postfire runoff to return to unburned levels and that potential erosion from concentrated flow processes may take longer than that from rainsplash-sheetflow processes to recover. Therefore, studies that focus on one aspect of hydrologic vulnerability (such as runoff) or on one process (such as rainsplash) may not accurately evaluate recovery (Pierson et al. 2011). Studies would benefit from multiyear assessments with annual controls and evaluation of runoff and erosion over multiple scales that encompass rainsplash, sheetflow, and concentrated flow. But such all-encompassing studies are often not possible or practical. Investigations that focus on a single-scale hydrologic parameter or process should consider the potential error associated with broad inferences of hydrologic recovery.

Summary and Conclusions

Fire is a natural disturbance on western rangelands that can facilitate amplified runoff and erosion rates and place natural resources, property, and lives at risk. However, the degree to which fire increases runoff and erosion rates and the associated risks is highly variable and depends on many factors. Runoff and erosion from rangelands are dictated by interactions among climate, vegetation, soils, and topography. Fire alters the structure of vegetation and ground cover, and may physically or chemically affect soils. Therefore, the magnitude of fire-induced increases in annual or event runoff and erosion is fundamentally associated with the degree to which fire reduces vegetation and modifies

soils, and with the local topography and prevailing climate regime (annual scale) or storm input (event scale).

In general, fire-induced hydrologic vulnerability and risks are low where burning minimally alters vegetation and ground cover or when precipitation input is low. High-intensity rainfall on severely burned and steeply sloping hillslopes presents the highest risk for fire-induced increases in runoff and erosion. Plot- to hillslope-scale studies presented in this review demonstrate that burning may increase runoff, erosion, or both during high-intensity rainfall events by factors of 2 to 40 over small-plot scales and more than 100-fold over large-plot to hillslope scales. Anecdotal reports of large-scale flooding and debris-flow events from rangelands, woodlands, and semiarid forests, associated with high-intensity rainfall following burning, document the potential risk to resources (such as water quality and aquatic habitat), property and infrastructure, and human life. Such risks are of particular concern for large urban centers along urban-wildland interfaces. It is important to note the need for postfire risk assessments to consider risk in a probabilistic framework that examines the likelihood of specified storm events and the potential responses for a given set of site conditions or susceptibility. Evaluations of site susceptibility to increased runoff and erosion should focus on the key indicators described in this review: expanse of bare ground, degree of soil water repellency, slope/topography, and site-specific soil stability/erodibility. Assessments of risk over long-term intervals should also evaluate potential changes in site susceptibility associated with prevailing climate conditions and plant/cover recruitment.

Lastly, the role of fire is changing on western rangelands. Fires are increasing in frequency, duration, and size across much of the western United States. These changes potentially increase the overall hydrologic vulnerability of rangelands by spatially and temporally increasing surface exposure to runoff and erosion processes. We do not yet know the ramifications of repeated burning and the respective runoff/erosion from frequently occurring small events or less frequent large events. The materials presented in this review do, however, provide a foundation from which hydrologic vulnerability assessments can be developed and conceptual and mathematical models can be advanced to aid in the management of postfire risk evaluation and assessment.

References

Abrahams, A.D.; Anthony, J.P.; Wainwright, J. 1995. Effects of vegetation change on interrill runoff and erosion, Walnut Gulch, Southern Arizona. Geomorphology. 13: 37–48.

Abrahams, A.D.; Li, G.; Parsons, A.J. 1996. Rill hydraulics on a semiarid hillslope, southern Arizona. Earth Processes and Landforms. 21: 35–47.

Abrahams, A.D.; Parsons, A.J. 1991a. Relation between infiltration and stone cover on a semiarid hillslope, southern Arizona. Journal of Hydrology. 122: 49–59.

Abrahams, A.D.; Parsons, A.J. 1991b. Resistance to overland flow on desert pavement and its implications for sediment transport modeling. Water Resources Research. 27: 1827–1836.

Abrahams, A.D.; Parsons, A.J. 1994. Hydraulics of interrill overland flow on stone-covered desert surfaces. Catena. 23: 111–140.

Abrahams, A.D.; Parsons, A.J.; Luk, S.H. 1988. Hydrologic and sediment responses to simulated rainfall on desert hillslopes in southern Arizona. Catena. 15: 103–117.

Abrahams, A.D.; Parsons, A.J.; Luk, S.H. 1991. The effect of spatial variability in overland flow on the downslope pattern of soil loss on a semiarid hillslope, southern Arizona. Catena. 18: 255–270.

Abrahams, A.D.; Parsons, A.J.; Wainwright, J. 2003. Disposition of rainwater under creosotebush. Hydrological Processes. 17: 2555–2566.

Agnew, W.; Labn, R.E.; Harding, M.V. 1997. Buffalo Creek, Colorado, fire and flood of 1996. Land and Water. 41: 27–29.

Ahlgren, I.F.; Ahlgren, C.E. 1960. Ecological effects of forest fires. Botanical Review. 26: 483–533.

Al-Hamdan, O.Z.; Hernandez, M.; Pierson, F.B.; Nearing, M.A.; Williams, C.J.; Stone, J.J.; Boll, J.; Weltz, M.A. 2015. Rangeland Hydrology and Erosion Model (RHEM) enhancements for applications on disturbed conditions. Hydrological Processes. 29: 445–457.

Al-Hamdan, O.Z.; Pierson, F.B.; Nearing, M.A.; Stone, J.J.; Williams, C.J.; Moffet, C.A.; Kormos, P.R.; Boll, J.; Weltz, M.A. 2012a. Characteristics of concentrated flow hydraulics for rangeland ecosystems: Implications for hydrologic modeling. Earth Surface Processes and Landforms. 37: 157–168.

Al-Hamdan, O.Z.; Pierson, F.B.; Nearing, M.A.; Williams, C.J.; Stone, J.J.; Kormos, P.R.; Boll, J.; Weltz, M.A. 2012b. Concentrated flow erodibility for physically based erosion models: Temporal variability in disturbed and undisturbed rangelands. Water Resources Research. 49: art. no. W07504.

Al-Hamdan, O.Z.; Pierson, F.B.; Nearing, M.A.; Williams, C.J.; Stone, J.J.; Kormos, P.R.; Boll, J.; Weltz, M.A. 2013. Risk assessment of erosion from concentrated flow on rangelands using overland flow distribution and shear stress partitioning. Transactions of the American Society of Agricultural and Biological Engineers. 56: 539–548.

Allen, C.D.; Breshears, D.D. 1998. Drought-induced shift of a forest-woodland ecotone: Rapid landscape response to climate variation. Proceedings of the National Academy of Sciences of the United States of America. 95: 14839–14842.

Andreu, V.; Imeson, A.C.; Rubio, J.L. 2001. Temporal changes in soil aggregates and water erosion after wildfire in a Mediterranean pine forest. Catena. 44: 69–84.

Ansley, R.J.; Jacoby, P.W.; Hicks, R.A. 1991. Leaf and whole plant transpiration in honey mesquite following severing of lateral roots. Journal of Range Management. 44: 577–583.

Arcenegui, V.; Mataix-Solero, J.; Guerrero, C.; Zornoza, R.; Mataix-Beneyto, J.; Garcia-Orenes, F. 2008. Immediate effects of wildfires on water repellency and aggregate stability in Mediterranean calcareous soils. Catena. 74: 219–223.

Assouline, S. 2004. Rainfall-induced soil surface sealing: A critical review of observations, conceptual models, and solutions. Vadose Zone Journal. 3: 570–591.

Baker, W.L. 2009. Fire ecology in Rocky Mountain landscapes. Washington, DC: Island Press. 632 p.

Bauters, T.W.J.; Steenhuis, T.S.; Dicarlo, D.A.; Nieber, J.L.; Dekker, L.W.; Ritsema, C.J.; Parlange, J.Y.; Haverkamp, R. 2000. Physics of water repellent soils. Journal of Hydrology. 231–232: 233–243.

Belnap, J.; Welter, J.R.; Grimm, N.B.; Barger, N.; Ludwig, J.A. 2005. Linkages between microbial and hydrologic processes in arid and semiarid watersheds. Ecology. 86: 298–307.

Belsky, A.J.; Amundson, R.G.; Duxbury, J.M.; Riha, S.J.; Ali, A.R.; Mwonga, S.M. 1989. The effects of trees on their physical, chemical, and biological environments in a semi-arid savanna in Kenya. Journal of Applied Ecology. 26: 1005–1024.

Belsky, A.J.; Mwonga, S.M.; Amundson, R.G.; Duxbury, J.M.; Ali, A.R. 1993. Comparative effects of isolated trees on their undercanopy environments in high- and low-rainfall savannas. Journal of Applied Ecology. 30: 143–155.

Benavides-Solorio, J.D.D.; MacDonald, L.H. 2001. Post-fire runoff and erosion from simulated rainfall on small plots, Colorado Front Range. Hydrological Processes. 15: 2931–2952.

Benavides-Solorio, J.D.D.; MacDonald, L.H. 2002. Erratum: post-fire runoff and erosion from simulated rainfall on small plots, Colorado Front Range. Hydrological Processes. 16: 1131–1133.

Benavides-Solorio, J.D.D.; MacDonald, L.H. 2005. Measurement and prediction of post-fire erosion at the hillslope scale, Colorado Front Range. International Journal of Wildland Fire. 14: 457–474.

Bergkamp, G. 1998. A hierarchical view of the interactions of runoff and infiltration with vegetation and microtopography in semiarid shrublands. Catena. 33: 201–220.

Beven, K.; Germann, P. 1982. Macropores and water flows in soils. Water Resources Research. 18: 1311–1325.

Bhark, E.W.; Small, E.E. 2003. Association between plant canopies and the spatial patterns of infiltration in shrubland and grassland of the Chihuahuan Desert, New Mexico. Ecosystems. 6: 185–196.

Billings, W.D. 1994. Ecological impacts of cheatgrass and resultant fire on ecosystems in the western Great Basin. In: Monsen, S.B.; Kitchen, S.G., comps. Proceedings: Ecology and management of annual rangelands; 1992 May 18–22; Boise, Idaho, USA. Gen. Tech. Rep. INT-GTR-313. Ogden, UT: U.S. Department of Agriculture, Forest Service, Intermountain Research Station: 22–30.

Bisdom, E.B.A.; Dekker, L.W.; Schoute, J.F.T. 1993. Water repellency of sieve fractions from sandy soils and relationships with organic material and soil structure. Geoderma. 56: 105–118.

Blackburn, W.H. 1975. Factors influencing infiltration and sediment production of semiarid rangelands in Nevada. Water Resources Research. 11: 929–937.

Blackburn, W.H.; Pierson, F.B.; Hanson, C.L.; Thurow, T.L.; Hanson, A.L. 1992. The spatial and temporal influence of vegetation on surface soil factors in semiarid rangelands. Transactions of the American Society of Agricultural Engineers. 35: 479–486.

Blackburn, W.H.; Pierson, F.B.; Seyfried, M.S. 1990. Spatial and temporal influence of soil frost on infiltration and erosion of sagebrush rangelands. Water Resource Bulletin. 26: 991–997.

Blackburn, W.H.; Skau, C.M. 1974. Infiltration rates and sediment production of selected plant communities in Nevada. Journal of Range Management: 27(6): 476–480.

Blackburn, W.H.; Wood, M.K. 1990. Influence of soil frost on infiltration of shrub coppice dune and dune interspace soils in southeastern Nevada. Great Basin Naturalist. 50: 41–46.

Bodí, M.B.; Doerr, S.H.; Cerdà, A.; Mataix-Solera, J. 2012. Hydrological effects of a layer of vegetation ash on underlying wettable and water repellent soil. Geoderma. 191: 14–23.

Bodí, M.B.; Muñoz-Santa, I.; Armero, C.; Doerr, S.H.; Mataix-Solero, J.; Cerdà, A. 2013. Spatial and temporal variations of water repellency and probability of its occurrence in calcareous Mediterranean rangeland soils affected by fires. Catena. 108: 14–25.

Bonan, G. 2002. Ecological climatology: concepts and applications. Cambridge, UK: Cambridge University Press. 678 p.

Bonnin, G.M.; Martin, D.; Lin, B.; Parzybok, T.; Yekta, M.; Riley, D. 2006. Precipitation—frequency atlas of the United States, NOAA atlas 14, volume 1, version 4. Silver Spring, MD: U.S. Department of Commerce, National Oceanic and Atmospheric Administration, National Weather Service. 261 p.

Bowen, B. 1996. Rainfall and climate variation over a sloping New Mexico plateau during the North American monsoon. Journal of Climate. 9: 3432–3442.

Bradford, J.M.; Ferris, J.E.; Remley, P.A. 1987. Interrill soil erosion processes: I. Effect of surface sealing on infiltration, runoff, and soil splash detachment. Soil Science Society of America Journal. 51: 1566–1571.

Brakensiek, D.L.; Rawls, W.J. 1994. Soil containing rock fragments: Effects on infiltration. Catena. 23: 99–110.

Brandes, D.; Wilcox, B.P. 2000. Evapotranspiration and soil moisture dynamics on a semiarid ponderosa pine hillslope. Journal of the American Water Resources Association. 36: 965–974.

Branson, F.A.; Owen, J.B. 1970. Plant cover, runoff, and sediment yield relationships on Mancos Shale in western Colorado. Water Resources Research. 6: 783–790.

Branson, F.A.; Gifford, G.F.; Renard, K.G.; Hadley, R.F. 1981. 2d ed. Rangeland hydrology. Range Science Series No. 1. Dubuque, IA: Kendall-Hunt Publishing. 340 p.

Breshears, D.D.; Barnes, F.J. 1999. Interrelationships between plant functional types and soil moisture heterogeneity for semiarid landscapes within the grassland/forest continuum: A unified conceptual model. Landscape Ecology. 14: 465–478.

Breshears, D.D.; Myers, O.B.; Johnson, S.R.; Meyer, C.W.; Martens, S.N. 1997a. Differential use of spatially heterogeneous soil moisture by two semiarid woody species: *Pinus edulis* and *Juniperus monosperma*. Journal of Ecology. 85: 289–299.

Breshears, D.D.; Nyhan, J.W.; Heil, C.E.; Wilcox, B.P. 1998. Effects of woody plants on microclimate in a semiarid woodland: soil temperature and evaporation in canopy and intercanopy patches. International Journal of Plant Science. 159: 1010–1017.

Breshears, D.D.; Rich, P.M.; Barnes, F.J.; Campbell, K. 1997b. Overstory-imposed heterogeneity in solar radiation and soil moisture in a semiarid woodland. Ecological Applications. 7: 1201–1215.

Brooks, J.R.; Meinzer, F.C.; Coulombe, R.; Gregg, J. 2002. Hydraulic redistribution of soil water during summer drought in two contrasting Pacific Northwest coniferous forests. Tree Physiology. 22: 1107–1117.

Brooks, M.L. 2008. Plant invasions and fire regimes. In: Zouhar, K.; Smith, J.K.; Sutherland, S.; Brooks, M.L., eds. Wildland fire in ecosystems: Fire and nonnative invasive plants. Gen. Tech. Rep. RMRS-GTR-42-vol. 6. Ogden, UT: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: 33–45.

Brooks, M.L.; D'Antonio, C.M.; Richardson, D.M.; Grace, J.B.; Keeley, J.E.; DiTomaso, J.M.; Hobbs, R.J.; Pellant, M.; Pyke, D. 2004. Effects of invasive alien plants on fire regimes. BioScience. 54: 677–688.

Brooks, M.L.; Pyke, D.A. 2001. Invasive plants and fire in the deserts of North America. In: Galley, K.E.M.; Wilson, T.P., eds. Proceedings of the invasive species workshop: The role of fire in the control and spread of invasive species. Fire conference 2000: The first national congress on fire ecology, prevention, and management. Tall Timbers Research Station, Misc. Publ. No. 11: 1–14.

Brown, J.K. 2000. Introduction and fire regimes. In Brown, J.K.; Smith, J.K., eds. Wildland fire in ecosystems: Effects of fire on flora. Gen. Tech. Rep. RMRS-GTR-42-vol. 2. Ogden, UT: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: 1–8.

Brutsaert, W. 2005. Hydrology, an introduction. New York: Cambridge University Press. 618 p.

Bryan, R.B. 2000. Soil erodibility and processes of water erosion on hillslope. Geomorphology. 32: 385–415.

Burcar, W.W.; Tyler, S.W.; Johnson, D.W. 1994. Seasonal preferential flow in two Sierra Nevada soils under forested and meadow cover. Soil Science Society of America Journal. 58: 1555–1561.

Burch, G.J.; Moore, I.D.; Burns, J. 1989. Soil hydrophobic effects on infiltration and catchment runoff. Hydrological Processes. 3: 211–222.

Burgess, S.S.O.; Adams, M.A.; Turner, N.C.; Ong, C.K. 1998. The redistribution of soil water by tree root systems. Oecologia. 115: 306–311.

Burgess, S.S.O.; Adams, M.A.; Turner, N.C.; White, D.A.; Ong, C.K. 2001. Tree roots: Conduits for deep recharge of soil water. Oecologia. 126:158–165.

Burgess, S.S.O.; Bleby, T.M. 2006. Redistribution of soil water by lateral roots mediated by stem tissues. Journal of Experimental Botany. 57: 3283–3291.

Caldwell, M.M. 1985. Cold deserts. In: Chabot, B.F.; Mooney, H.A., eds. Physiological ecology of North American plant communities. New York: Chapman and Hall: 198–212.

Caldwell, M.M.; Dawson, T.E.; Richards, J.H. 1998. Hydraulic lift: Consequences of water efflux from roots of plants. Oecologia. 113: 151–161.

Caldwell, M.M.; Eissenstat, D.M.; Richards, J.H.; Allen, M.F. 1985. Competition for phosphorus: Differential uptake from dual-isotope-labeled soil interspaces between shrub and grasses. Science. 229: 384–386.

Caldwell, M.M.; Richards, J.H. 1989. Hydraulic lift: Water efflux from upper roots improves effectiveness of water uptake by deep roots. Oecologia. 79: 1–5.

Cammeraat, L.H.; Imeson, A.C. 1998. Deriving indicators of soil degradation from soil aggregation studies in southeastern Spain and southern France. Geomorphology. 23: 307–321.

Cammeraat, L.H.; Imeson, A.C. 1999. The evolution and significance of soil-vegetation patterns following land abandonment and fire in Spain. Catena. 37: 107–127.

Campbell, G.S.; Harris, G.A. 1977. Water relations and water use patterns for *Artemisia tridentata* Nutt. in wet and dry years. Ecology. 58: 652–659.

Campbell, R.E.; Baker, M.B.; Ffolliot, P.F.; Larson, F.R.; Avery, C.C. 1977. Wildfire effects on a ponderosa pine ecosystem: An Arizona case study. Res. Pap. RM-191. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Forest and Range Experimental Station. 12 p.

Cannon, S.H.; Bigio, E.R.; Mine, E. 2001a. A process for fire-related debris flow initiation, Cerro Grande fire, New Mexico. Hydrological Processes. 15: 3011–3023.

Cannon, S.H.; Gartner, J.E.; Wilson, R.C.; Bowers, J.C.; Laber, J.L. 2008. Storm rainfall conditions for floods and debris flows from recently burned areas in southwestern Colorado and southern California. Geomorphology. 96: 250–269.

Cannon, S.H.; Kirkham, R.M.; Parise, M. 2001b. Wildfire-related debris-flow initiation processes, Storm King Mountain, Colorado. Geomorphology. 39: 171–188.

Cannon, S.H.; Powers, P.S.; Savage, W.Z. 1998. Fire-related hyperconcentrated and debris flows on Storm King Mountain, Glenwood Springs, Colorado, USA. Environmental Geology. 35: 210–218.

Carlson, D.H.; Thurow, T.L. 1996. Comprehensive evaluation of the improved SPUR model (SPUR-91). Ecological Modelling. 85: 229–240.

Carlson, D.H.; Thurow, T.L.; Knight, R.W.; Heitschmidt, R.K. 1990. Effect of honey mesquite on the water balance of Texas rolling plains rangeland. Journal of Range Management. 43: 491–496.

Carlyle-Moses, D.E. 2004. Throughfall, stemflow, and canopy interception loss fluxes in a semi-arid Sierra Madre Oriental matorral community. Journal of Arid Environments. 58: 181–202.

Castillo, V.M.; Gómez-Plaza, A.; Martínez-Mena, M. 2003. The role of antecedent soil water content in the runoff response of semiarid catchments: A simulation approach. Journal of Hydrology. 284: 114–130.

Cerdà, A. 1997. The effect of patchy distribution of *Stipa tenacissima* L. on runoff and erosion. Journal of Arid Environments. 36: 37–51.

Cerdà, A. 1998a. Changes in overland flow and infiltration after a rangeland fire in a Mediterranean scrubland. Hydrological Processes. 12: 1031–1042.

Cerdà, A. 1998b. Soil aggregate stability under different Mediterranean vegetation types. Catena. 32: 73–86.

Cerdà, A. 2001. Effects of rock fragment cover on soil infiltration, interrill runoff and erosion. European Journal of Soil Science. 52: 59–68.

Cerdà, A.; Doerr, S.H. 2005. Influence of vegetation recovery on soil hydrology and erodibility following fire: An 11-year investigation. International Journal of Wildland Fire. 14: 423–437.

Cerdà, A.; Doerr, S.H. 2008. The effects of ash and needle cover on surface runoff and erosion in the immediate post-fire period. Catena. 74: 256–263.

Certini, G.; Nocentini, C.; Knicker, H.; Arfaioli, P.; Rumpel, C. 2011. Wildfire effects on soil organic matter quantity and quality in two fire-prone Mediterranean pine forests. Geoderma. 167–168: 148–155.

Clark, O.R. 1940. Interception of rainfall by prairie grasses, weeds, and certain crop plants. Ecological Monographs. 10: 243–277.

Craddock, G.W. 1946. Salt Lake City flood, 1945. Proceedings of the Utah Academy of Sciences, Arts, and Letters. 23: 51–61.

Craddock, G.W.; Pearse, C.K. 1938. Surface run-off and erosion on granitic mountain soils of Idaho as influenced by range cover, soil disturbance, slope, and precipitation intensity. Boise, ID: U.S. Department of Agriculture, Forest Service, Intermountain Forest and Range Experiment Station. 24 p.

D'Antonio, C.M. 2000. Fire, plant invasions, and global changes. In: Mooney, H.A.; Hobbs, R.J., eds. Invasive species in a changing world. Washington, DC: Island Press: 65–93.

D'Antonio, C.M.; Vitousek, P.M. 1992. Biological invasions by exotic grasses, the grass/fire cycle, and global change. Annual Review of Ecology and Systematics. 23: 63–87.

Davenport, D.W.; Breshears, D.D.; Wilcox, B.P.; Allen, C.D. 1998. Viewpoint: Sustainability of piñon-juniper ecosystems—a unifying perspective of soil erosion thresholds. Journal of Range Management. 51: 231–240.

Dawson, T.E. 1993. Hydraulic lift and water use by plants: implications for water balance, performance and plant-plant interactions. Oecologia. 95: 565–574.

Dawson, T.E. 1996. Determining water use by trees and forests from isotopic, energy balance and transpiration analyses: The roles of tree size and hydraulic lift. Tree Physiology. 16: 263–272.

DeBano, L.F. 1971. The effect of hydrophobic substances on water movement in soil during infiltration. Soil Science Society of America Proceedings. 35: 340–343.

DeBano, L.F. 1981. Water repellent soils: A state-of-the-art. Gen. Tech. Rep. PSW-46. Berkeley, CA: U.S. Department of Agriculture, Forest Service, Pacific Southwest Research Station. 25 p.

DeBano, L.F. 1991. The effects of fire on soil properties. In: Harvey, A.E.; Neuenchwander, L.F., comps. Proceedings—management and productivity of western-montaine forest soils; 1990 April 10–12; Boise, ID. Gen. Tech. Rep. INT-280. Ogden, UT: U.S. Department of Agriculture, Forest Service, Intermountain Research Station: 151–156.

DeBano, L.F. 2000. The role of fire and soil heating on water repellency in wildland environments: A review. Journal of Hydrology. 231–232: 195–206.

DeBano, L.F. 2001. Water repellency in soils: a historical review. Journal of Hydrology. 231–232: 4–32.

DeBano, L.F.; Krammes, J.S. 1966. Water repellent soils and their relation to wildfire temperatures. International Association of Hydrological Sciences. 2: 14–19.

DeBano, L.F.; Mann, L.D.; Hamilton, D.A. 1970. Translocation of hydrophobic substances into soil by burning organic litter. Proceedings of the Soil Science Society of America. 34: 130–133.

DeBano, L.F.; Neary, D.G.; Ffolliott, P.F. 1998. Fire's effects on ecosystems. New York: John Wiley and Sons, Inc. 352 p.

DeBano, L.F.; Rice, R.M. 1973. Water-repellent soils: Their implications in forestry. Journal of Forestry. 71: 220–223.

DeBano, L.F.; Rice, R.M.; Conrad, C.E. 1979. Soil heating in chaparral fires: Effects on soil properties, plant nutrients, erosion, and runoff. PSW-145. Berkeley, CA: U.S. Department of Agriculture, Forest Service, Pacific Southwest Research Station. 21 p.

DeBano, L.F.; Savage, S.M.; Hamilton, A.D. 1976. The transfer of heat and hydrophobic substances during burning. Proceedings of the Soil Science Society of America. 40: 779–782.

Dekker, L.W.; Doerr, S.H.; Oostindie, K.; Ziogas, A.K.; Ritsema, C.J. 2001. Water repellency and critical soil water content in a dune sand. Soil Science Society of America Journal. 65: 1667–1674.

Dekker, L.W.; Ritsema, C.J. 1994. How water moves in a water-repellent sandy soil: 1. Potential and actual water repellency. Water Resources Research. 30: 2507–2517.

Dekker, L.W.; Ritsema, C.J. 1995. Fingerlike wetting patterns in two water-repellent loam soils. Journal of Environmental Quality. 24: 324–333.

Dekker, L.W.; Ritsema, C.J. 1996. Preferential flow paths in a water repellent clay soil with grass cover. Water Resources Research. 32: 1239–1294.

Devaurs, M.; Gifford, G.F. 1984. Variability of infiltration within large runoff plots on rangelands. Journal of Range Management. 37: 523–528.

Devitt, D.A.; Smith, S.D. 2002. Root channel macropores enhance downward movement of water in a Mojave Desert ecosystem. Journal of Arid Environments. 50: 99–108.

Dingman, L. 2002. 2d ed. Physical hydrology. Upper Saddle River, NJ: Prentice Hall. 600 p.

DiTomaso, J.M. 2000. Invasive weeds in rangelands: Species, impacts, and management. Weed Science. 48: 255–265.

D'Odorico, P.; Caylor, K.; Okin, G.S.; Scanlon, T.M. 2007. On soil moisture-vegetation feedbacks and their possible effects on the dynamics of dryland ecosystems. Journal of Geophysical Research. 112: G04010.

Doerr, S.H.; Blake, W.H.; Shakesby, R.A.; Stagnitti, F.; Vuurens, S.H.; Humphreys, G.S.; Wallbrink, P. 2004. Heating effects on water repellency in Australian eucalypt forest soils and their value in estimating wildfire soil temperatures. International Journal of Wildland Fire. 13: 157–163.

Doerr, S.H.; Moody, J.A. 2004. Hydrological impacts of soil water repellency: On spatial and temporal uncertainties. Hydrological Processes. 18: 829–832.

Doerr, S.H.; Shakesby, R.A.; Blake, W.H.; Chafer, C.; Humphreys, G.S.; Wallbrink, P.J. 2006. Effects of differing wildfire severities on soil wettability and implications for hydrological response. Journal of Hydrology. 319: 295–311.

Doerr, S.H.; Shakesby, R.A.; MacDonald, L.H. 2009. Soil water repellency: A key factor in post-fire erosion. In: Cerdà, A.; Robichaud, P.R., eds. Fire effects on soils and restoration strategies. Land Reconstruction and Management Series, Volume 5. Enfield, NH: Science Publishers: 197–224.

Doerr, S.H.; Shakesby, R.A.; Walsh, R.P.D. 1996. Soil water repellency variations with depth and particle size fractions in burned and unburned *Eucalyptus globulus* and *Pinus pinaster* forest terrain in the Agueda basin, Portugal. Catena. 27: 25–47.

Doerr, S.H.; Shakesby, R.A.; Walsh, R.P.D. 2000. Soil water repellency: Its causes, characteristics and hydro-geomorphological significance. Earth-Science Reviews. 51: 33–65.

Doerr, S.H.; Thomas, A.D. 2000. The role of soil moisture in controlling water repellency: New evidence from forest soils in Portugal. Journal of Hydrology. 231–232: 134–147.

Domingo, F.; Sánchez, G.; Moro, M.J.; Brenner, A.J.; Puigdefábregas, J. 1998. Measurement and modeling of rainfall interception by three semi-arid canopies. Agricultural and Forest Meteorology. 91: 275–292.

Domingo, F.; Villagarcía, L.; Brenner, A.J.; Puigdefábregas, J. 2000. Measuring and modelling the radiation balance of a heterogeneous shrubland. Plant, Cell and Environment. 23: 27–38.

Dunkerley, D. 2000. Measuring interception loss and canopy storage in dryland vegetation: A brief review and evaluation of available research strategies. Hydrological Processes. 14: 669–678.

Dunkerley, D. 2002. Infiltration rates and soil moisture in a groved mulga community near Alice Springs, arid central Australia: Evidence for complex internal rainwater redistribution in a runoff-runon landscape. Journal of Arid Environments. 51: 199–219.

Dunkerley, D.L. 2008. Intra-storm evaporation as a component of canopy interception loss in dryland shrubs: Observations from Fowlers Gap, Australia. Hydrological Processes. 22: 1985–1995.

Dunkerley, D.L.; Booth, T.L. 1999. Plant canopy interception of rainfall and its significance in a banded landscape, arid western New South Wales, Australia. Water Resources Research. 35: 1581–1586.

Dunkerley, D.L.; Brown, K.J. 1995. Runoff and runon areas in a patterned chenopod shrubland, arid western New South Wales, Australia: Characteristics and origin. Journal of Arid Environments. 30: 41–55.

Dunne, T. 1978. Field studies of hillslope flow processes. In: Kirby, M.J., ed. Hillslope hydrology. Norwich, UK: John Wiley and Sons, Ltd.: 227–293.

Dunne, T.; Leopold, L.B. 1978. Water in environmental planning. San Francisco, CA: W.H. Freeman and Company. 818 p.

Dunne, T.; Zhang, W.; Aubry, B.F. 1991. Effects of rainfall, vegetation, and microtopography on infiltration and runoff. Water Resources Research. 27: 2271–2285.

Ebel, B.A.; Moody, J.A.; Martin, D.A. 2012. Hydrologic conditions controlling runoff generation immediately after wildfire. Water Resources Research. 48: W0352.

Ellen, S.D.; Wieczorek, G.F. 1988. Landslides, floods, and marine effects of the storm of January 3–5, 1982, in the San Francisco Bay Region, California. United States Geological Survey, Professional Paper 1434, Version 1. 319 p.

Elliott, J.G.; Parker, R.S. 2001. Developing a post-fire flood chronology and recurrence probability from alluvial stratigraphy in the Buffalo Creek watershed, Colorado, USA. Hydrological Processes. 15: 3039–3051.

Emerman, S.H.; Dawson, T.E. 1996. Hydraulic lift and its influence on the water content of the rhizosphere: An example from sugar maple, *Acer saccharum*. Oecologia. 108: 273–278.

Emmett, W.W. 1970. The hydraulics of overland flow on hillslopes. United States Geological Survey, Professional Paper 662A. 66 p.

Emmett, W.W. 1978. Overland flow. In: Kirby, M.J., ed. Hillslope hydrology. Norwich, UK: John Wiley and Sons, Ltd.: 145–176.

Ferrera, A.G.; Singer, M.J. 1985. Energy dissipation for water drop impact into shallow pools. Soil Science Society of America Journal. 49: 1537–1542.

Field, J.P.; Breshears, D.D.; Whicker, J.J.; Zou, C.B. 2012. Sediment capture by vegetation patches: Implications for desertification and increased resource redistribution. Journal of Geophysical Research G: Biogeosciences. 117. doi: 10.1029/2011jg001663.

Flerchinger, G.N.; Cooley, K.R. 2000. A ten-year water balance of a mountainous semi-arid watershed. Journal of Hydrology. 237: 86–99.

Flerchinger, G.N.; Cooley, K.R.; Hanson, C.L.; Seyfried, M.S. 1998. A uniform versus an aggregated water balance of a semi-arid watershed. Hydrological Processes. 12: 331–342.

Flerchinger, G.N.; Cooley, K.R.; Ralston, D.R. 1992. Groundwater response to snowmelt in a mountainous watershed. Journal of Hydrology. 133: 293–311.

Flerchinger, G.N.; Hanson, C.L.; Wight, J.R. 1996. Modeling evapotranspiration and surface energy budgets across a watershed. Water Resources Research. 32: 2539–2548.

Foster, G.R.; Meyer, L.D. 1972. Transport of soil particles by shallow flow. Transactions of the American Society of Agricultural Engineers. 15: 99–102.

Fox, D.M.; Darboux, F.; Carrega, P. 2007. Effects of fire-induced water repellency on soil aggregate stability, splash erosion, and saturated hydraulic conductivity for different size fractions. Hydrological Processes. 21: 2377–2384.

Gabet, E.J.; Dunne, T. 2003. Sediment detachment by rain power. Water Resources Research. 39: ESG11–ESG112.

General Accounting Office. 2003. Wildland fires: Better information needed on effectiveness of emergence stabilization and rehabilitation treatments. Rep. GAO-03-430. Washington, DC: U.S. General Accounting Office. 55 p.

Gifford, G.F. 1985. Cover allocation in rangeland watershed management (a review). In: Jones, E.B.; Ward, J.T., eds. Watershed management in the Eighties. New York: American Society of Civil Engineers: 23–31.

Gilley, J.E.; Finkner, S.C. 1985. Estimating soil detachment caused by raindrop impact. Transactions of the American Society of Agricultural Engineers. 28: 140–146.

Gilley, J.E.; Woolhiser, D.A.; McWhorter, D.B. 1985. Interrill soil erosion—Part II: testing and use of model equations. Transactions of the American Society of Agricultural Engineers. 28: 154–159.

Giménez, R.; Govers, G. 2001. Interaction between bed roughness and flow hydraulics in eroding rills. Water Resources Research. 37: 791–799.

Giovannini, G.; Lucchesi, S. 1997. Modifications induced in soil physico-chemical parameters by experimental fires at different intensities. Soil Science. 162: 479–486.

Giovannini, G.; Lucchesi, S.; Giachetti, M. 1988. Effect of heating on some physical and chemical parameters related to soil aggregation and erodibility. Soil Science. 146: 255–261.

Giovannini, G.; Vallejo, R.; Lucchesi, S.; Bautista, S.; Ciompi, S.; Llovet, J. 2001. Effects of land use and eventual fire on soil erodibility in dry Mediterranean conditions. Forest Ecology and Management. 147: 15–23.

Goodrich, D.C.; Keefer, T.O.; Unkrich, C.L.; Nichols, M.H.; Osborn, H.B.; Stone, J.J.; Smith, J.R. 2008. Long-term precipitation database, Walnut Gulch Experimental Watershed, Arizona, United States. Water Resources Research. 44: W05S04.

Govers, G. 1992. Relationship between discharge, velocity and flow area for rills eroding loose, non-layered materials. Earth Surface Processes and Landforms. 17: 515–528.

Govers, G.; Giménez, R.; Van Oost, K. 2007. Rill erosion: Exploring the relationship between experiments, modelling and field observations. Earth-Science Reviews. 84: 87–102.

Groen, A.H.; Woods, S.W. 2008. Effectiveness of aerial seeding and straw mulch for reducing post-wildfire erosion, north-western Montana, USA. International Journal of Wildland Fire. 17: 559–571.

Gunn, R.; Kinzer, G.D. 1949. The terminal velocity of fall from water droplets in stagnant air. Journal of Meteorology. 6: 243–248.

Hamilton, E.L.; Rowe, P.B. 1949. Rainfall interception by chaparral in California. Sacramento, CA: California Department of Natural Resources, Division of Forestry, with the U.S. Department of Agriculture, Forest Service, Pacific Southwest Forest and Range Experiment Station. 43 p.

Hanson, C.L.; Pierson, F.B. 2001. Characteristics of extreme precipitation and associated streamflow in the Reynolds Creek Experimental Watershed, Idaho. In: Proceedings of the 12th Symposium on Global Change Climate Variations; 2001 January 14-19; Albuquerque, NM. Boston, MA: American Meteorological Society: J2.13–J2.16.

Hendrickx, J.M.H.; Dekker, L.W.; Boersma, O.H. 1993. Unstable wetting fronts in water repellent field soils. Journal of Environmental Quality. 22: 109–118.

Hester, J.W.; Thurow, T.L.; Taylor, C.A., Jr. 1997. Hydrologic characteristics of vegetation types as affected by prescribed burning. Journal of Range Management. 50: 199–204.

Hillel, D. 1998. Environmental soil physics. New York: Academic Press. 771 p.

Hornberger, G.M.; Raffensperger, J.P.; Wiberg, P.L.; Eshleman, K.N. 1998. Elements of physical hydrology. Baltimore, MD: John Hopkins University Press. 276 p.

Horton, J.L.; Hart, S.C. 1998. Hydraulic lift: a potentially important ecosystem process. Trends in Ecology and Evolution. 13: 232–235.

Horton, R.E. 1933. The role of infiltration in the hydrologic cycle. Transactions of the American Geophysical Union. 14: 446–460.

Hubbert, K.R.; Preisler, H.K.; Wohlgemuth, P.M.; Graham, R.C.; Narog, M.G. 2006. Prescribed burning effects on soil physical properties and soil water repellency in a steep chaparral watershed, southern California, USA. Geoderma. 130: 284–298.

Huffman, E.L.; MacDonald, L.H.; Stednick, J.D. 2001. Strength and persistence of fire-induced hydrophobicity under ponderosa and lodgepole pine, Colorado Front Range. Hydrological Processes. 15: 2877–2892.

Hull, A.C., Jr. 1965. Cheatgrass—a persistent homesteader. In: Proceedings, cheatgrass symposium; 1965 July 27-30; Vale, OR. Portland, OR: U.S. Department of the Interior, Oregon Bureau of Land Management: 20–26.

Hull, A.C., Jr. 1972. Rainfall and snowfall interception of big sagebrush. Utah Academy of Science, Arts, and Letters. 49: 64.

Hull, A.C., Jr.; Klomp, G.J. 1974. Yield of crested wheatgrass under four densities of big sagebrush in southern Idaho. U.S. Department of Agriculture Tech. Bull. 1438. Washington, DC: U.S. Department of Agriculture, Agricultural Research Service. 70 p.

Hultine, K.H.; Cable, W.L.; Burgess, S.S.O.; Williams, D.B. 2003a. Hydraulic redistribution by deep roots of a Chihuahuan Desert phreatophyte. Tree Physiology. 23: 353–360.

Hultine, K.H.; Scott, R.L.; Cable, W.L.; Goodrich, D.C.; Williams, D.G. 2004. Hydraulic redistribution by a dominant, warm desert phreatophyte: Seasonal patterns and response to precipitation pulses. Functional Ecology. 18: 530–538.

Hultine, K.H.; Williams, D.G.; Burgess, S.S.O.; Keefer, T.O. 2003b. Contrasting patterns of hydraulic redistribution in three desert phreatophytes. Oecologia. 135: 167–175.

Hutchinson, B.A. 1965. Snow accumulation and disappearance influenced by big sagebrush. Res. Note RM-RN-46. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Forest and Range Experiment Station. 7 p.

Huxman, T.E.; Wilcox, B.P.; Breshears, D.D.; Scott, R.L.; Snyder, K.A.; Small, E.E.; Hultine, K.; Pockman, W.T.; Jackson, R.B. 2005. Ecohydrological implications of woody plant encroachment. Ecology. 86: 308–319.

Imeson, A.C.; Lavee, H.; Calvo, A.; Cerdà, A. 1998. The erosional response of calcareous soils along a climatological gradient in southeast Spain. Geomorphology. 24: 3–16.

Imeson, A.C.; Verstraten, J.M.; van Mulligen, E.J.; Sevink, J. 1992. The effects of fire and water repellency on infiltration and runoff under Mediterranean type forest. Catena. 19: 345–361.

Inbar, M.; Tamir, M.; Wittenberg, L. 1998. Runoff and erosion processes after a forest fire in Mount Carmel, a Mediterranean area. Geomorphology. 24: 17–33.

Inbar, M.; Wittenberg, L.; Tamir, M. 1997. Soil erosion and forestry management after wildfire in a Mediterranean woodland, Mt. Carmel, Israel. International Journal of Wildland Fire. 7: 285–294.

Ishikawa, C.M.; Bledsoe, C.S. 2000. Seasonal and diurnal patterns of soil water potential in the rhizospere of blue oaks: Evidence for hydraulic lift. Oecologia. 125: 459–465.

Jackson, R.B.; Sperry, J.S.; Dawson, T.E. 2000. Root water uptake and transport: Using physiological processes in global predictions. Trends in Plant Science. 5: 624–632.

Jaramillo, D.F.; Dekker, L.W.; Ritsema, C.J.; Hendrickx, J.M.H. 2000. Occurrence of soil water repellency in arid and humid climates. Journal of Hydrology. 231–232: 105–111.

Joffre, R.; Rambal, S. 1993. How tree cover influences the water balance of Mediterranean rangelands. Ecology. 74: 570–582.

Johansen, M.P.; Hakonson, T.E.; Breshears, D.D. 2001. Post-fire runoff and erosion from rainfall simulation: Contrasting forests with shrublands and grasslands. Hydrological Processes. 15: 2953–2965.

Johnson, C.W.; Blackburn, W.H. 1989. Factors contributing to sagebrush rangeland soil loss. Transactions of the American Society of Agricultural Engineers. 32: 155–160.

Johnson, C.W.; Gordon, N.D. 1988. Runoff and erosion from rainfall simulator plots on sagebrush rangeland. Transactions of the American Society of Agricultural Engineers. 31: 421–427.

Johnson, C.W.; Smith, J.P. 1978. Sediment characteristics and transport from Northwest Rangeland Watersheds. Transactions of the American Society of Agricultural Engineers. 21: 1157–1163.

Julien, P.Y. 1998. Erosion and sedimentation. Cambridge, UK: Cambridge University Press. 300 p.

Keane, R.E.; Agee, J.K.; Fule, P.; Keeley, J.E.; Key, C.; Kitchen, S.G.; Miller, R.; Schulte, L.A. 2008. Ecological effects of large fires on US landscapes: Benefit or catastrophe? International Journal of Wildland Fire. 17: 696–712.

Keeley, J.E. 2009. Fire intensity, fire severity and burn severity: a brief review and suggested usage. International Journal of Wildland Fire. 18: 116–126.

Kinnell, P.I.A. 1988. The influence of flow discharge on sediment concentrations in raindrop induced flow transport. Australian Journal of Soil Research. 26: 575–582.

Kinnell, P.I.A. 1990. The mechanics of raindrop-induced flow transport. Australian Journal of Soil Research. 28: 497–516.

Kinnell, P.I.A. 1991. The effect of flow depth on sediment transport induced by raindrops impacting shallow flows. Transactions of the American Society of Agricultural Engineers. 34: 161–168.

Kinnell, P.I.A. 1993. Sediment concentrations resulting from flow depth/drop size interactions in shallow overland flow. Transactions of the American Society of Agricultural Engineers. 36: 1099–1103.

Kinnell, P.I.A. 2005. Raindrop-impact-induced erosion processes and prediction: A review. Hydrological Processes. 19: 2815–2844.

Klade, R.J. 2006. Building a research legacy—the Intermountain Station 1911-1997. Gen. Tech. Rep. RMRS-GTR-184. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 259 p.

Knapp, B.J. 1978. Infiltration and storage of soil water. In: Kirby, M.J., ed. Hillslope hydrology. Norwich, UK: John Wiley and Sons, Ltd.: 43–72.

Knapp, P.A. 1996. Cheatgrass (*Bromus tectorum* L.) dominance in the Great Basin desert. Global Environmental Change. 6: 37–52.

Knight, R.W.; Blackburn, W.H.; Merrill, L.B. 1984. Characteristics of oak mottes, Edwards Plateau, Texas. Journal of Range Management. 37: 534–537.

Knighton, D. 1998. Fluvial forms and processes: A new perspective. London, UK: Arnold. 400 p.

Kramer, P.J.; Boyer, J.S. 1995. Water relations of plants and soils. London, UK: Academic Press. 495 p.

Larsen, I.J.; MacDonald, L.H.; Brown, E.; Rough, D.; Welsh, M.J. 2009. Causes of post-fire runoff and erosion: water repellency, cover, or soil sealing. Soil Science Society of America Journal. 73: 1393–1407.

Lavee, H.; Kutiel, P.; Segev, M.; Benyamini, Y. 1995. Effects of surface roughness on runoff and erosion in Mediterranean ecosystems: The role of fire. Geomorphology. 11: 227–234.

Lebron, I.; Madsen, M.D.; Chandler, D.G.; Robinson, D.A.; Wendroth, O.; Belnap, J. 2007. Ecohydrological controls on soil moisture and hydraulic conductivity within a pinyon-juniper woodland. Water Resources Research. 43: W08422.

Leffler, A.J.; Peek, M.S.; Ryel, R.J.; Ivans, C.Y.; Caldwell, M.M. 2005. Hydraulic redistribution through the root systems of senesced plants. Ecology. 86: 633–642.

Leffler, A.J.; Ryel, R.J.; Hipps, L.; Ivans, S.; Caldwell, M.M. 2002. Carbon acquisition and water use in a northern Utah *Juniperus osteosperma* (Utah juniper) population. Tree Physiology. 22: 1221–1230.

Leighton-Boyce, G.; Doerr, S.H.; Shakesby, R.A.; Walsh, R.P.D. 2007. Quantifying the impact of soil water repellency on overland flow generation and erosion: A new approach using rainfall simulation and wetting agent on in situ soil. Hydrological Processes. 21: 2337–2345.

Leighton-Boyce, G.; Doerr, S.H.; Shakesby, R.A.; Walsh, R.P.D.; Ferreira, A.J.D.; Boulet, A.K.; Coelho, C.D.A. 2005. Temporal dynamics of water repellency and soil moisture in eucalypt plantations, Portugal. Australian Journal of Soil Research. 43: 269–280.

Lentile, L.B.; Holden, Z.A.; Smith, A.M.S.; Falkowski, M.J.; Hudak, A.T.; Morgan, P.; Lewis, S.A.; Gessler, P.E.; Benson, N.C. 2006. Remote sensing techniques to assess active fire characteristics and post-fire effects. International Journal of Wildland Fire. 15: 319–345.

Letey, J.; Osborn, J.; Pelishek, R.E. 1962. The influence of the water-solid contact angle on water movement in soil. Bulletin International Association Scientific Hydrology. 3: 75–81.

Lewis, S.A.; Wu, J.Q.; Robichaud, P.R. 2006. Assessing burn severity and comparing soil water repellency, Hayman Fire, Colorado. Hydrological Processes. 20: 1–16.

Link, S.O.; Keeler, C.W.; Hill, R.W.; Hagen, E. 2006. *Bromus tectorum* cover mapping and fire risk. International Journal of Wildland Fire. 15: 113–119.

Liston, G.E.; McFadden, J.P.; Sturm, M.; Pielke, R.A., Sr. 2002. Modelled changes in arctic tundra snow, energy and moisture fluxes due to increased shrubs. Global Change Biology. 8: 17–32.

Liston, G.E.; Sturm, M. 1998. A snow-transport model for complex terrain. Journal of Glaciology. 44: 498–516.

Johnson, C.W.; Blackburn, W.H. 1989. Factors contributing to sagebrush rangeland soil loss. Transactions of the American Society of Agricultural Engineers. 32: 155–160.

Llorens, P.; Domingo, F. 2007. Rainfall partitioning by vegetation under Mediterranean conditions. A review of studies in Europe. Journal of Hydrology. 335: 37–54.

Luce, C.H.; Tarboton, D.G.; Cooley, K.R. 1998. The influence of the spatial distribution of snow basin-average snowmelt. Hydrological Processes. 12: 1671–1683.

Ludwig, F.; Dawson, T.E.; Kroon, H.; Berendse, F.; Prins, H.H.T. 2003. Hydraulic lift in *Acacia tortilis* on an East African savanna. Oecologia. 134: 293–300.

Ludwig, F.; Dawson, T.E.; Prins, H.H.T.; Berendse, F.; de Kroon, H. 2004. Below-ground competition between trees and grasses may overwhelm the facilitative effects of hydraulic lift. Ecology Letters. 7: 623–631.

Ludwig, J.A.; Tongway, D.J. 1995. Spatial organisation of landscapes and its function in semi-arid woodlands, Australia. Landscape Ecology. 10: 51–63.

Ludwig, J.A.; Tongway, D.J. 2000. Viewing rangelands as landscape systems. In: Arnalds, O.; Archer, S., eds. Rangeland desertification. Dordrecht, the Netherlands: Kluwer Academic Publishers: 39–52.

Ludwig, J.A.; Tongway, D.J.; Freudenberger, D.O.; Noble, J.C.; Hodgkinson, K.C. 1997. Landscape ecology, function, and management: Principles from Australia's rangelands. Melbourne, Australia: CSIRO Publishing. 158 p.

Ludwig, J.A.; Wiens, J.A.; Tongway, D.J. 2000. A scaling rule for landscape patches and how it applies to conserving soil resources in savannas. Ecosystems. 3: 84–97.

Ludwig, J.A.; Wilcox, B.P.; Breshears, D.D.; Tongway, D.J.; Imeson, A.C. 2005. Vegetation patches and runoff-erosion as interacting ecohydrological processes in semiarid landscapes. Ecology. 86: 288–297.

MacDonald, L.H.; Huffman, E.L. 2004. Post-fire soil water repellency: Persistence and soil moisture thresholds. Soil Science Society of America Journal. 68: 1729–1734.

Mack, R.N. 1981. Invasion of *Bromus tectorum* L. into western North America: An ecological chronicle. Agro-Ecosystems. 7: 145–165.

Madsen, M.D.; Chandler, D.G.; Belnap, J. 2008. Spatial gradients in ecohydrological properties within a pinyon-juniper ecosystem. Ecohydrology. 43(8): 349–360.

Maidment, D.R. 1993. Hydrology. In: Maidment, D.R., ed. Handbook of hydrology. New York: McGraw-Hill, Inc.: 1.1–1.3.

Marks, D.; Link, T.; Winstral, A.; Garen, D. 2001. Simulating snowmelt processes during rain-on-snow over a semi-arid mountain basin. Annals of Glaciology. 32: 195–202.

Marks, D.; Winstral, A. 2001. The effect of variable patterns of snow deposition and drifting on snowmelt, runoff, and stream discharge in a semi-arid mountain basin. In: Proceedings of the Western Snow Conference; 2001 April 1619; Sun Valley, Idaho. Western Snow Conference Proceedings 69: 127–130.

Marks, D.; Winstral, A.; Seyfried, M. 2002. Simulation of terrain and forest shelter effects on patterns of snow deposition, snowmelt, and runoff over a semi-arid mountain catchment. Hydrological Processes. 16: 3605–3626.

Martínez-Mena, M.; Alvarez Rogel, J.; Albaladejo, J.; Castillo, V.M. 1999. Influence of vegetal cover on sediment size distribution in natural rainfall conditions in a semiarid environment. Catena. 38: 175–190.

Martínez-Meza, E.; Whitford, W.G. 1996. Stemflow, throughflow, and channelization of stemflow by roots in three Chihuahuan desert shrubs. Journal of Arid Environments. 32: 271–287.

Martínez-Zavala, L.; Jordán, A. 2008. Effect of rock fragment cover on interrill soil erosion from bare soils in Western Andalusia, Spain. Soil Use and Management. 24:108–117.

Mataix-Solera, J.; Cerdà, A.; Arcenegui, V.; Jordán, A.; Zavala, L.M. 2011. Fire effects on soil aggregation: a review. Earth-Science Reviews. 109: 44–60.

Mataix-Solero, J.; Guerrero, C.; García-Orenes, F.; Bárcenas, G.M.; Pilar Torres, M. 2009. Forest fire effects on soil microbiology. In: Cerdà, A.; Robichaud, P.R., eds. Fire effects on soils and restoration strategies. Land Reconstruction and Management Series, Volume 5. Enfield, NH: Science Publishers: 133–176.

McCuen, R.H. 2004. Hydrologic analysis and design. Upper Saddle River, NJ: Prentice-Hall, Inc. 888 p.

McCulley, R.L.; Jobbagy, E.G.; Pockman, W.T.; Jackson, R.B. 2004. Nutrient uptake as a contributing explanation for deep rooting in arid and semi-arid ecosystems. Oecologia. 141: 620–628.

McNamara, J.P.; Chandler, D.; Seyfried, M.; Achet, S. 2005. Soil moisture states, lateral flow, and streamflow generation in a semi-arid, snowmelt-driven catchment. Hydrological Processes. 19: 4023–4038.

Meeuwig, R.O. 1970. Infiltration and soil erosion as influenced by vegetation and soil in northern Utah. Journal of Range Management. 23: 185–188.

Meeuwig, R.O. 1971. Infiltration and water repellency in granitic soils. Res. Pap. INT-111. Ogden, UT: U.S. Department of Agriculture, Forest Service, Intermountain Forest and Range Experiment Station. 20 p.

Meinzer, F.C.; Brooks, J.R.; Bucci, S.; Goldstein, G.; Scholz, F.G.; Warren, J.M. 2004. Converging patterns of uptake and hydraulic redistribution of soil water in contrasting woody vegetation. Tree Physiology. 24: 919–928.

Meinzer, F.C.; Clearwater, M.J.; Goldstein, G. 2001. Water transport in trees: Current perspectives, new insights and some controversies. Environmental and Experimental Botany. 45: 239–262.

Mendel, M.; Hergarten, S.; Neugebauer, H.J. 2002. On a better understanding of hydraulic lift: A numerical study. Water Resources Research. 38: 1183–1192. doi: 10.1029/2001WR000911.

Mendez, A.; Goodrich, D.D.; Osborn, H.B. 2003. Rainfall point intensities in an air mass thunderstorm environment: Walnut Gulch, Arizona. Journal of the American Water Resources Association. 39: 611–621.

Meyer, G.A.; Pierce, J.L.; Wood, S.H.; Jull, A.J.T. 2001. Fire, storms, and erosional events in the Idaho batholiths. Hydrological Processes. 15: 3025–3038.

Michaelides, K.; Lister, D.; Wainwright, J.; Parsons, A.J. 2009. Vegetation controls on small-scale runoff and erosion dynamics in a degrading dryland environment. Hydrological Processes. 23: 1617–1630.

Miller, R.F.; Chambers, J.C.; Pyke, D.A.; Pierson, F.B.; Williams, C.J. 2013. A review of fire effects on vegetation and soils in the Great Basin Region: Response and ecological site characteristics. Gen. Tech. Rep. RMRS-GTR-308. Fort Collins, CO: UT: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 126 p.

Miller, R.F.; Tausch, R.J. 2001. The role of fire in juniper and pinyon woodlands: A descriptive analysis. In: Galley, K.E.M.; Wilson, T.P., eds. Proceedings of the invasive species workshop: The role of fire in the control and spread of invasive species. Fire Conference 2000: The first national congress on fire ecology, prevention, and management; 2000 November 27–December 1; San Diego, CA. Misc. Publ. No. 11. Tallahassee, FL: Tall Timbers Research Station: 15–30.

Miller, R.F.; Tausch, R.J.; McArthur, E.D.; Johnson, D.J; Sanderson, S.C. 2008. Age structure and expansion of piñon-juniper woodlands: A regional perspective in the Intermountain West. Res. Pap. RMRS-RP-69. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 15 p.

Moffet, C.A.; Pierson, F.B.; Robichaud, P.R.; Spaeth, K.E.; Hardegree, S.P. 2007. Modeling soil erosion on steep sagebrush rangeland before and after prescribed fire. Catena. 71: 218–228.

Montaña, C.; Cavagnaro, B.; Briones, O. 1995. Soil water use by co-existing shrubs and grasses in the Southern Chihuahuan Desert, Mexico. Journal of Arid Environments. 31: 1–13.

Moody, J.A.; Martin, D.A. 2001. Initial hydrologic and geomorphic response following a wildfire in the Colorado Front Range. Earth Surface Processes and Landforms. 26: 1049–1070.

Moran, M.S.; Scott, R.L.; Keefer, T.O.; Emmerich, W.E.; Hernandez, M.; Nearing, G.S.; Paige, G.B.; Cosh, M.H.; O'Neill, P.E. 2009. Partitioning evapotranspiration in semiarid grassland and shrubland ecosystems using time series of soil surface temperature. Agricultural and Forest Meteorology. 149: 59–72.

Morris, S.E.; Moses, T.A. 1987. Forest fire and the natural soil erosion regime in the Colorado Front Range. Annals—Association of American Geographers. 77: 245–254.

Moss, A.J. 1991. Rain-impact soil crust: I. Formation on a granite-derived soil. Australian Journal of Soil Research. 29: 271–289.

Moss, A.J.; Green, P. 1983. Movement of solids in air and water by rain drop impact: Effects of drop-size and water-depth variations. Australian Journal of Soil Research. 21: 257–269.

Muñoz, M.R.; Squeo, F.A.; León, M.F.; Tracol, Y.; Gutiérrez, J.R. 2008. Hydraulic lift in three shrub species from the Chilean coastal desert. Journal of Arid Environments. 72: 624–632.

Mutchler, C.K.; Murphree, C.E.; McGregor, K.C. 1988. Laboratory and field plots for soil erosion studies. In: Lal, R., ed. Soil erosion research methods. Ankeny, IA: Soil and Water Conservation Society: 9–39.

Návar, J.; Bryan, R. 1990. Interception loss and rainfall redistribution by three semi-arid growing shrubs in northeastern Mexico. Journal of Hydrology. 115: 51–63.

Nearing, M.A.; Foster, G.R.; Lane, L.J.; Finkner, S.C. 1989. A process-based soil erosion model for USDA Water Erosion Prediction Project technology. Transactions of the American Society of Agricultural Engineers. 32: 1587–1593.

Nearing, M.A.; Nichols, N.H.; Stone, J.J.; Renard, K.G.; Simanton, J.R. 2007. Sediment yields from unit-source semiarid watersheds at Walnut Gulch. Water Resources Research. 43: W06426.

Nearing, M.A.; Norton, L.D.; Bulgakov, D.A.; Larionov, G.A.; West, L.T.; Dontsova, K.M. 1997. Water hydraulics and erosion in eroding rills. Water Resources Research. 33: 865–876.

Nearing, M.A.; Simanton, J.R.; Norton, L.D.; Bulygin, S.J.; Stone, J. 1999. Soil erosion by surface water flow on a stony, semiarid hillslope. Earth Surface Processes and Landforms. 24: 677–686.

Nearing, M.A.; Wei, H.; Stone, J.J.; Pierson, F.B.; Spaeth, K.E.; Weltz, M.A.; Flanagan, D.C.; Hernandez, M. 2011. A rangeland hydrology and erosion model. Transactions of the American Society of Agricultural and Biological Engineers. 54: 901–908.

Neary, D.G.; Klopatek, C.C.; DeBano, L.F.; Ffolliott, P.F. 1999. Fire effects on belowground sustainability: A review and synthesis. Forest Ecology and Management. 122: 51–71.

Neary, D.G.; Koestner, K.A.; Youberg, A.; Koestner, P.E. 2012. Post-fire rill and gully formation, Schultz Fire 2010, Arizona, USA. Geoderma. 191: 97–104.

Newman, B.D.; Campbell, A.R.; Wilcox, B.P. 1997. Tracer-based studies of soil water movement in semi-arid forests of New Mexico. Journal of Hydrology. 196: 251–270.

O'Dea, M.E.; Guertin, D.P. 2003. Prescribed fire effects on erosion parameters in a perennial grassland. Journal for Range Management. 56: 27–32.

Onda, Y.; Dietrich, W.E.; Booker, F. 2008. Evolution of overland flow after a severe forest fire, Point Reyes, California. Catena. 72: 13–20.

Osborn, H.B. 1964. Effect of storm duration on runoff from rangeland watersheds in the semiarid southwestern United States. International Association of Scientific Hydrology IX. 4: 40–47.

Osborn, H.B. 1983a. Precipitation characteristics affecting hydrological response of south-western rangelands. Agricultural Reviews and Manuals, Western Series 34. Oakland, CA: U.S. Department of Agriculture, Agricultural Research Service. 55 p.

Osborn, H.B. 1983b. Timing and duration of high rainfall rates in the Southwestern United States. Water Resources Research. 19: 1036–1042.

Osborn, H.B.; Lane, L.J. 1969. Precipitation-runoff relations for very small semiarid rangeland watersheds. Water Resources Research. 5: 419–425.

Osborn, H.B.; Renard, K.G. 1988. Rainfall intensities for Southeastern Arizona. Journal of Irrigation and Drainage Engineering. 114: 195–199.

Owens, M.K.; Lyons, R.K.; Alejandro, C.L. 2006. Rainfall partitioning within semiarid juniper communities: Effects of event size and canopy cover. Hydrological Processes. 20: 3179–3189.

Parsons, A.; Robichaud, P.R.; Lewis, S.A.; Napper, C.; Clark, J.T. 2010. Field guide for mapping post-fire soil burn severity. Gen. Tech. Rep. RMRS-GTR-243. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 49 p.

Parsons, A.J.; Abrahams, A.D.; Simanton, J.R. 1992. Microtopography and soil-surface materials on semi-arid piedmont hillslopes, southern Arizona. Journal of Arid Environments. 22: 107–115.

Parsons, A.J.; Abrahams, A.D.; Wainwright, J. 1994. Rainsplash and erosion rates in an interrill area on semi-arid grassland, Southern Arizona. Catena. 22: 215–226.

Parsons, A.J.; Abrahams, A.D.; Wainwright, J. 1996. Responses of interrill runoff and erosion rates to vegetation change in southern Arizona. Geomorphology. 14: 311–317.

Peláez, D.V.; Distel, R.A.; Bóo, R.M.; Elia, O.R.; Mayor, M.D. 1994. Water relations between shrubs and grasses in semi-arid Argentina. Journal of Arid Environments. 27: 71–78.

Peters, E.F.; Bunting, S.C. 1994. Fire conditions pre- and post-occurrence of annual grasses on the Snake River Plain. In: Monsen, S.B.; Kitchen, S.G., eds. Proceedings— ecology and management of annual rangelands. Gen. Tech. Rep. INT-GTR-313. Ogden, UT: U.S. Department of Agriculture, Forest Service, Intermountain Forest and Range Experiment Station: 31–36.

Pierson, F.B.; Bates, J.D.; Svejcar, T.J.; Hardegree, S.P. 2007a. Runoff and erosion after cutting western juniper. Rangeland Ecology and Management. 60: 285–292.

Pierson, F.B.; Blackburn, W.H.; Van Vactor, S.S. 2007b. Hydrologic impacts of mechanical seeding treatments on sagebrush rangelands. Rangeland Ecology and Management. 60: 666–674.

Pierson, F.B.; Blackburn, W.H.; Van Vactor, S.S.; Wood, J.C. 1994a. Partitioning small scale spatial variability of runoff and erosion on sagebrush rangelands. Water Resources Bulletin. 30: 1081–1089.

Pierson, F.B.; Carlson, D.H.; Spaeth, K.E. 2002b. Hydrologic impacts of fire on steep intermountain sagebrush-steppe rangeland. International Journal of Wildland Fire. 11: 145–151.

Pierson, F.B.; Moffet, C.A.; Williams, C.J.; Hardegree, S.P.; Clark, P.E. 2009. Prescribed-fire effects on rill and interrill runoff and erosion in a mountainous sagebrush landscape. Earth Surface Processes and Landforms. 34: 193–203.

Pierson, F.B.; Robichaud, P.R.; Moffet, C.A.; Spaeth, K.E.; Hardegree, S.P.; Clark, P.E.; Williams, C.J. 2008a. Fire effects on rangeland hydrology and erosion in a steep sagebrush-dominated landscape. Hydrological Processes. 22: 2916–2929.

Pierson, F.B.; Robichaud, P.R.; Moffet, C.A.; Spaeth, K.E.; Williams, C.J.; Hardegree, S.P.; Clark, P.E. 2008b. Soil water repellency and infiltration in coarse-textured soils of burned and unburned sagebrush ecosystems. Catena. 74: 98–108.

Pierson, F.B.; Robichaud, P.R.; Spaeth, K.E. 2001a. Spatial and temporal effects of wildfire on the hydrology of a steep rangeland watershed. Hydrological Processes. 15: 2905–2916.

Pierson, F.B.; Slaughter, C.W.; Cram, Z.K. 2001b. Long-term stream discharge and suspended-sediment database, Reynolds Creek Experimental Watershed, Idaho, United States. Water Resources Research. 37: 2857–2861.

Pierson, F.B.; Spaeth, K.E.; Weltz, M.A.; Carlson, D.H. 2002a. Hydrologic response of diverse western rangelands. Journal of Range Management. 55: 558–570.

Pierson, F.B.; Van Vactor, S.S.; Blackburn, W.H.; Wood, J.C. 1994b. Incorporating small scale spatial variability into predictions of hydrologic response on sagebrush rangelands. In: Blackburn, W.H.; Pierson, F.B.; Schuman, G.E.; Zartman, R., eds. Variability in rangeland water erosion processes. Soil Science Society of America Special Publication 38. Madison, WI: Soil Science Society of America: 23–33.

Pierson, F.B.; Wight, J.R. 1991. Variability of near-surface soil temperature on sagebrush rangeland. Journal of Range Management. 44: 491–497.

Pierson, F.B.; Williams, C.J.; Hardegree, S.P.; Clark, P.E.; Kormos, P.R.; Al-Hamdan, O.Z. 2013. Hydrologic and erosion responses of sagebrush steppe following juniper encroachment, wildfire, and tree-cutting. Rangeland Ecology and Management. 66: 274–289.

Pierson, F.B.; Williams, C.J.; Hardegree, S.P.; Weltz, M.A.; Stone, J.J.; Clark, P.E. 2011. Fire, plant invasions, and erosion events on western rangelands. Rangeland Ecology and Management. 64: 439–449.

Pierson, F.B.; Williams, C.J.; Kormos, P.R.; Al-Hamdan, O.Z. 2014. Short-term effects of tree removal on infiltration, runoff, and erosion in woodland-encroached sagebrush steppe. Rangeland Ecology and Management. 67: 522–538.

Pierson, F.B.; Williams, C.J.; Kormos, P.R.; Hardegree, S.P.; Clark, P.E.; Rau, B.M. 2010. Hydrologic vulnerability of sagebrush steppe following pinyon and juniper encroachment. Rangeland Ecology and Management. 63: 614–629.

Pierson, T.C. 1986. Flow behaviour of channelized debris flows, Mount St. Helens, Washington. In: Abrahams, A.D., ed. Hillslope processes. Boston, MA: Allen and Unwin: 269–296.

Poesen, J.; Ingelmo-Sanchez, F. 1992. Runoff and sediment yield from topsoils with different porosity as affected by rock fragment cover and position. Catena. 19: 451–474.

Poesen, J.; Ingelmo-Sanchez, F.; Mücher, H. 1990. The hydrological response of soil resources to rainfall as affected by cover and position of rock fragments in the top layer. Earth Surface Processes and Landforms. 15: 653–671.

Poesen, J.W.; Torri, D.; Bunte, K. 1994. Effects of rock fragments on soil erosion by water at different spatial scales: A review. Catena. 23: 141–166.

Pomeroy, J.W.; Gray, D.M. 1995. Snowcover: Accumulation, relocation and management. National Hydrology Research Institute Science Report No. 7. Saskatoon, Saskatchewan, Canada: National Hydrology Research Institute. 144 p.

Pressland, A.J. 1973. Rainfall partitioning by an arid woodland (*Acacia aneura* F. Muell.) in South-West Queensland. Australian Journal of Botany. 21: 235–245.

PRISM Climate Group. 2011. PRISM climate data. Corvallis, OR: Oregon State University, Northwest Alliance for Computational Science and Engineering. http://prism.oregonstate.edu [Accessed March 22, 2011].

Puigdefábregas, J. 2005. The role of vegetation patterns in structuring runoff and sediment fluxes in drylands. Earth Surface Processes and Landforms. 30: 133–147.

Puigdefábregas, J.; del Barrio, G.; Boer, M.M.; Gutiérrez, L.; Solé, A. 1998. Differential responses of hillslope and channel elements to rainfall events in a semi-arid area. Geomorphology. 23(2): 337–351.

Puigdefábregas, J.; Sole, A.; Gutierrez, L.; del Barrio, G.; Boer, M. 1999. Scales and processes of water redistribution in drylands: Results from the Rambla Honda field site in southeast Spain. Earth Science Reviews. 48: 37–70.

Pyke, D.A.; Herrick, J.E.; Shaver, P.; Pellant, M. 2002. Rangeland health attributes and indicators for qualitative assessment. Journal of Range Management. 55: 584–597.

Querejeta, J.I.; Egerton-Warburton, L.M.; Allen, M.F. 2003. Direct nocturnal water transfer from oaks to their mycorrhizal symbionts during severe soil drying. Oecologia. 134: 55–64.

Querejeta, J.I.; Egerton-Warburton, L.M.; Allen, M.F. 2007. Hydraulic lift may buffer rhizosphere hyphae against the negative effects of severe soil drying in a California oak savanna. Soil Biology and Biochemistry. 39: 409–417.

Rango, A.; Tartowski, S.L.; Laliberte, A.; Wainwright, J.; Parsons, A. 2006. Islands of hydrologically enhanced biotic productivity in natural and managed arid ecosystems. Journal of Arid Environments. 65: 235–252.

Ravi, S.; D'Odorico, P.; Zobeck, T.M.; Oliver, T.M.; Collins, S.L. 2007. Feedbacks between fires and wind erosion in heterogeneous arid lands. Journal of Geophysical Research: Biogeosciences. 112: G04007. doi: 10.1029/2007jg000474.

Reid, K.D.; Wilcox, B.P.; Breshears, D.D.; MacDonald, L. 1999. Runoff and erosion in a piñon-juniper woodland: Influence of vegetation patches. Soil Science Society of America Journal. 63: 1869–1879.

Renard, K.G. 1988. Water resources of small impoundments in dry regions. In: Thames, J.L.; Ziebell, C.D., eds. Small water impoundments in semiarid regions. Albuquerque, NM: University of New Mexico Press: 1–20.

Renard, K.G.; Lane, L.J.; Simanton, J.R.; Emmerich, W.E.; Stone, J.J.; Weltz, M.A.; Goodrich, D.C.; Yakowitz, D.S. 1993. Agricultural impacts in an arid environment: Walnut Gulch case study. Hydrological Science and Technology. 9: 145–190.

Reynolds, J.F.; Kemp, P.R.; Tenhunen, J.D. 2000. Effects of long-term rainfall variability on evapotranspiration and soil water distribution in the Chihuahuan Desert: A modeling analysis. Plant Ecology. 150: 145–159.

Reynolds, J.F.; Virginia, R.A.; Kemp, P.R.; De Soyza, A.G.; Tremmel, D.C. 1999. Impact of drought on desert shrubs: Effects of seasonality and degree of resource island development. Ecological Monographs. 69(1): 69–106.

Rice, P.M.; McPherson, G.R.; Rew, L.J. 2008. Fire and nonnative invasive plants in the Interior West Bioregion. In: Zouhar, K.; Smith, J.K.; Sutherland, S.; Brooks, M.L., eds. Wildland fire in ecosystems: Fire and nonnative invasive plants. Gen. Tech. Rep. RMRS-GTR-42-vol. 6. Ogden, UT: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: 141–171.

Richards, J.H.; Caldwell, M.M. 1987. Hydraulic lift: Substantial nocturnal water transport between soil layers by *Artemisia tridentata* roots. Oecologia. 73: 486–489.

Ritsema, C.J.; Dekker, L.W. 1994. How water moves in a water repellent sandy soil. 2. Dynamics of fingered flow. Water Resources Research. 30: 2519–2531.

Ritsema, C.J.; Dekker, L.W. 1995. Distribution flow: A general process in the top layer of water repellent soils. Water Resources Research. 31: 1187–1200.

Ritsema, C.J.; Dekker, L.W. 1996. Water repellency and its role in forming preferred flow paths in soils. Australian Journal of Soil Research. 34: 475–487.

Ritsema, C.J.; Dekker, L.W.; Hendrickx, J.M.H.; Hamminga, W. 1993. Preferential flow mechanism in a water repellent sandy soil. Water Resources Research. 29: 2183–2193.

Ritsema, C.J.; Dekker, L.W.; van den Elsen, E.G.M.; Oostindiel, K.; Steenhuis, T.S.; Nieber, J.L. 1997. Recurring fingered flow pathways in a water repellent sandy field soil. Hydrology and Earth System Sciences. 1: 777–786.

Ritsema, C.J.; Nieber, J.L.; Dekker, L.W.; Steenhuis, T.S. 1998a. Modeling and field evidence of finger formation and finger recurrence in a water repellent sandy soil. Water Resources Research. 34(4): 555–567.

Ritsema, C.J.; Nieber, J.L.; Dekker, L.W.; Steenhuis, T.S. 1998b. Stable or unstable wetting fronts in water repellent soils—effect of antecedent soil moisture content. Soil and Tillage Research. 47: 111–123.

Robichaud, P.R. 2000. Fire effects on infiltration rates after prescribed fire in northern Rocky Mountain forests, USA. Journal of Hydrology. 231–232: 220–229.

Robichaud, P.R.; Ashmun, L.E.; Sims, B.D. 2010. Post-fire treatment effectiveness for hillslope stabilization. Gen. Tech. Rep. RMRS-GTR-240. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 62 p.

Robichaud, P.R.; Brown, R.E. 2002. Silt fences: An economical technique for measuring hillslope soil erosion. Gen. Tech. Rep. RMRS-GTR-94. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 24 p.

Robichaud, P.R.; Elliot, W.J.; Pierson, F.B.; Hall, D.E.; Moffet, C.A. 2007a. Predicting postfire erosion and mitigation effectiveness with a web-based probabilistic erosion model. Catena. 71: 229–241.

Robichaud, P.R.; Elliot, W.J.; Pierson, F.B.; Hall, D.E.; Moffet, C.A.; Ashmun, L.E. 2007b. Erosion Risk Management Tool (ERMiT) user manual (version 2006.01.18). Gen. Tech. Rep. RMRS-GTR-188. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station. 24 p.

Robinson, D.A.; Lebron, I.; Ryel, R.J.; Jones, S.B. 2010. Soil water repellency: A method of soil moisture sequestration in pinyon-juniper woodland. Soil Science Society of America Journal. 74: 624–634.

Romme, W.H.; Allen, C.D.; Bailey, J.D.; Baker, W.L.; Bestelmeyer, B.T.; Brown, P.M.; Eisenhart, K.S.; Floyd, M.L.; Huffman, D.W.; Jacobs, B.F.; Miller, R.F.; Muldavin, E.H.; Swetnam, T.W.; Tausch, R.J.; Weisberg, P.J. 2009. Historical and modern disturbance regimes, stand structures, and landscape dynamics in piñon-juniper vegetation of the western United States. Rangeland Ecology and Management. 62: 203–222.

Roundy, B.A.; Blackburn, W.H.; Eckert, R.E., Jr. 1978. Influence of prescribed burning on infiltration and sediment production in the pinyon-juniper woodland, Nevada. Journal of Range Management. 31: 250–253.

Rowe, P.B. 1948. Influence of woodland chaparral on water and soil in central California. [N.p.]: State of California Department of Natural Resources, Division of Forestry, in cooperation with the U.S. Department of Agriculture, Forest Service, California Forest and Range Experiment Station. 70 p.

Ryel, R.J.; Caldwell, M.M.; Leffler, A.J.; Yoder, C.Y. 2003. Rapid soil moisture recharge to depth by roots in a stand of *Artemisia tridentata*. Ecology. 84: 757–764.

Ryel, R.J.; Caldwell, M.M.; Manwaring, J.H. 1996. Temporal dynamics of soil spatial heterogeneity in sagebrush-wheatgrass steppe during a growing season. Plant and Soil. 184: 299–309.

Ryel, R.J.; Caldwell, M.M.; Yoder, C.K.; Or, D.; Leffler, A.J. 2002. Hydraulic redistribution in a stand of *Artemisia tridentata*: Evaluation of benefits to transpiration assessed with a simulation model. Oecologia. 130: 173–184.

Ryel, R.J.; Leffler, A.J.; Peek, M.S.; Ivans, C.Y.; Caldwell, M.M. 2004. Water conservation in *Artemisia tridentata* through redistribution of precipitation. Oecologia. 141: 335–345.

Salles, C.; Poesen, J. 2000. Rain properties controlling soil splash detachment. Hydrological Processes. 14: 271–282.

Salles, C.; Poesen, J.; Govers, G. 2000. Statistical and physical analysis of soil detachment by raindrop impact: Rain erosivity indices and threshold energy. Water Resources Research. 36: 2721–2729.

Salles, C.; Poesen, J.; Sempere-Torres, D. 2002. Kinetic energy of rain and its functional relationship with intensity. Journal of Hydrology. 257: 256–270.

Sankey, J.B.; Germino, M.J.; Glenn, N.F. 2009. Aeolian sediment transport following wildfire in sagebrush steppe. Journal of Arid Environments. 73: 912–919.

Savage, S.M. 1974. Mechanism of fire-induced water repellency in soils. Proceedings of the Soil Science Society of America. 38: 652–657.

Savage, S.M.; Osborn, J.; Letey, J.; Helton, C. 1972. Substances contributing to fire induced water repellency in soils. Soil Science Society of America Proceedings. 36: 674–678.

Schenk, H.J.; Jackson, R.B. 2002. Rooting depths, lateral root spreads and below-ground/above-ground allometries of plants in water-limited ecosystems. Journal of Ecology. 90: 480–494.

Schlesinger, W.H.; Abrahams, A.D.; Parsons, A.J.; Wainwright, J. 1999. Nutrient losses in runoff from grassland and shrubland habitats in Southern New Mexico: I. Rainfall simulation experiments. Biogeochemistry. 45: 21–34.

Schlesinger, W.H.; Raikes, J.A.; Hartley, A.E.; Cross, A.F. 1996. On the spatial pattern of soil nutrients in desert ecosystems. Ecology. 77: 364–374.

Schlesinger, W.H.; Reynolds, J.F.; Cunningham, G.L.; Huenneke, L.F.; Jarrell, W.M.; Virginia, R.A.; Whitford, W.G. 1990. Biological feedbacks in global desertification. Science. 247: 1043–1048.

Schlesinger, W.H.; Ward, T.J.; Anderson, J. 2000. Nutrient losses in runoff from grassland and shrubland habitats in southern New Mexico: II. Field plots. Biogeochemistry. 49: 69–86.

Schmidt, R.A.; Gluns, D.R. 1991. Snowfall interception on branches of three conifer species. Canadian Journal of Forest Research. 21: 1262–1269.

Scholes, R.J.; Archer, S.R. 1997. Tree-grass interactions in savannas. Annual Review of Ecology and Systematics. 28: 517–544.

Scholl, D.G. 1971. Soil wettability in Utah juniper stands. Soil Science Society of America Proceedings. 35(2): 344–345.

Scholl, D.G. 1975. Soil wettability and fire in Arizona chaparral. Soil Science Society of America Proceedings. 39: 356–361.

Schulze, E.-D.; Caldwell, M.M.; Canadell, J.; Mooney, H.A.; Jackson, R.B.; Parson, D.; Scholes, R.; Sala, O.E.; Trimborn, P. 1998. Downward flux of water through roots (i.e., inverse hydraulic lift) in dry Kalahari sands. Oecologia. 115: 460–462.

Scott, D.F.; Van Wyk, D.B. 1990. The effects of wildfire on soil wettability and hydrological behaviour of an afforested catchment. Journal of Hydrology. 121: 239–256.

Scott, K.M. 1988. Origins, behavior, and sedimentology of lahars and lahar-runout flows in the Toutle-Cowlitz River system, Mount St. Helens, Washington. United States Geological Survey, Professional Paper 1447-A. 74 p.

Scott, R.L.; Cable, W.L.; Hultine, K.R. 2008. The ecohydrologic significance of hydraulic redistribution in a semiarid savanna. Water Resources Research. 44: W02440. doi: 10.1029/2007WR006149.

Scott, R.L.; Huxman, T.E.; Cable, W.L.; Emmerich, W.E. 2006. Partitioning of evapotranspiration and its relation to carbon dioxide exchange in a Chihuahuan Desert shrubland. Hydrological Processes. 20: 3227–3243.

Selby, M.J. 1993. Hillslope materials and processes. New York: Oxford University Press, Inc. 480 p.

Serrato, F.B.; Diaz, A.R. 1998. A simple technique for measuring rainfall interception by small shrub: "Interception flow collection box." Hydrological Processes. 12: 471–481.

Seyfried, M.S. 1991. Infiltration patterns from simulated rainfall on a semiarid rangeland soil. Soil Science Society of America Journal. 55: 1726–1736.

Seyfried, M.S.; Flerchinger, G.N. 1994. Influence of frozen soil on rangeland erosion. In: Variability of rangeland water and erosion processes. Special Publ. 38. Madison, WI: Soil Science Society of America: 67–82.

Seyfried, M.S.; Grant, L.E.; Marks, D.; Winstral, A.; McNamara, J. 2009. Simulated soil water storage effects on streamflow generation in a mountainous snowmelt environment, Idaho, USA. Hydrological Processes. 23: 858–873.

Seyfried, M.S.; Wilcox, B.P. 1995. Scale and the nature of spatial variability: Field examples having implications for hydrologic modeling. Water Resources Research. 31: 173–184.

Shakesby, R.A. 2011. Post-wildfire soil erosion in the Mediterranean: Review and future research directions. Earth-Science Reviews. 105: 71–100.

Shakesby, R.A.; Coelho, C.D.A.; Ferreira, A.D.; Terry, J.P.; Walsh, R.P.D. 1993. Wildfire impacts on soil erosion and hydrology in wet Mediterranean forests, Portugal. International Journal of Wildland Fire. 3: 95–110.

Shakesby, R.A.; Doerr, S.H. 2006. Wildfire as a hydrological and geomorphological agent. Earth-Science Reviews. 74: 269–307.

Shakesby, R.A.; Doerr, S.H.; Walsh, R.P.D. 2000. The erosional impact of soil hydrophobicity: Current problems and future research directions. Journal of Hydrology. 231–232: 178–191.

Sharma, P.P.; Gupta, S.C.; Rawls, W.J. 1991. Soil detachment by single raindrops of varying kinetic energy. Soil Science Society of America Journal. 55: 301–307.

Sidle, R.C.; Pearce, A.J.; O'Loughlin, C.L. 1985. Hillslope stability and land use. Water Resource Monograph Series, Vol. 11. Washington, DC: American Geophysical Union. 140 p.

Skau, C.M. 1964. Interception, throughfall, and stemflow in Utah and alligator juniper cover types in northern Arizona. Forest Science. 10: 283–287.

Slaughter, C.W.; Pierson, F.B. 2000. Sediment yield and hydrology in northwestern rangelands. International Journal of Sediment Research. 15: 253–259.

Slaughter, D.J., Jr. 1997. Throughfall, stemflow, and infiltration rates of *Juniperus ashei* on the Edwards Plateau, Texas. Austin, TX: University of Texas. Master's thesis. 120 p.

Smith, D.M.; Jackson, N.A.; Roberts, J.M.; Ong, C.K. 1999. Reverse flow in tree roots and downward siphoning of water by *Grevillea robusta*. Functional Ecology. 13: 256–264.

Spigel, K.M.; Robichaud, P.R. 2007. First-year post-fire erosion rates in Bitterroot National Forest, Montana. Hydrological Processes. 21: 998–1005.

Stannard, D.I.; Weltz, M.A. 2006. Partitioning evapotranspiration in sparsely vegetated rangeland using a portable chamber. Water Resources Research. 42: W02413.

Stone, J.J.; Nichols, M.H.; Goodrich, D.C.; Buono, J. 2008. Long-term runoff database, Walnut Gulch, Experimental Watershed, Arizona, United States. Water Resources Research. 44: W05S05.

Storck, P.; Lettenmaier, D.P.; Bolton, S.M. 2002. Measurement of snow interception and canopy effects on snow accumulation and melt in a mountainous maritime climate, Oregon, United States. Water Resources Research. 38: 5–1 to 5–16.

Sturm, M.; McFadden, J.P.; Liston, G.E.; Chapin, F.S., III; Racine, C.H.; Holmgren, J. 2001. Snow-shrub interactions in arctic tundra: A hypothesis with climatic implications. Journal of Climate. 336–344.

Taucer, P.I.; Munster, C.L.; Wilcox, B.P.; Owens, M.K.; Mohanty, B.P. 2008. Large-scale rainfall simulation experiments of juniper rangelands. Transactions of the American Society of Agricultural and Biological Engineers. 51: 1951–1961.

Tausch, R.J. 1999. Historic pinyon and juniper woodland development. In: Monsen, S.B.; Stevens, R., comps. Proceedings: Ecology and management of pinyon-juniper communities within the Interior West; 1997 September 15–18; Provo, UT. Proc. RMRS-P-9. Ogden, UT: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: 12–19.

Tausch, R.J.; Hood, S. 2007. Pinyon/juniper woodlands. In Hood, S.M.; Miller, M., eds. Fire ecology and management of the major ecosystems of southern Utah. Gen. Tech. Rep. RMRS-GTR-202. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Research Station: 57–71.

Terry, J.P.; Shakesby, R.A. 1993. Soil hydrophobicity effects on rainsplash: Simulated rainfall and photographic evidence. Earth Surface Processes and Landforms. 18: 519–525.

Thornes, J.B. 1980. Erosional processes of running water and their spatial and temporal controls: A theoretical viewpoint. In: Kirkby, M.J.; Morgan, R.P.C., eds. Soil erosion. Norwich, UK: John Wiley and Sons, Ltd.: 129–182.

Thornes, J.B. 1994. Catchment and channel hydrology. In: Abrahams, A.D.; Parsons, A.J., eds. Geomorphology of desert environments. London, UK: Chapman and Hall: 257–287.

Thurow, T.L.; Blackburn, W.H.; Taylor, C.A., Jr. 1986. Hydrologic characteristics of vegetation types as affected by livestock grazing systems, Edwards Plateau, Texas. Journal of Range Management. 39: 505–509.

Thurow, T.L.; Blackburn, W.H.; Taylor, C.A., Jr. 1988. Infiltration and interrill erosion responses to selective livestock grazing strategies, Edwards Plateau, Texas. Journal of Range Management. 41: 296–302.

Thurow, T.L.; Blackburn, W.H.; Warren, S.D.; Taylor, C.A., Jr. 1987. Rainfall interception by midgrass, shortgrass, and live oak mottes. Journal of Range Management. 40: 455–460.

Thurow, T.L.; Hester, J.W. 1997. How an increase or a reduction in juniper cover alters rangeland hydrology. In: 1997 Juniper Symposium; 1997 January 9–10; Texas Agricultural Experiment Station, San Angelo, TX. Tech. Report 97-1. College Station, TX: Texas A&M University: 9–22.

Tiedemann, A.R.; Conrad, C.E.; Dieterich, J.H. 1979. Effects of fire on water. A state-of-knowledge review. Gen. Tech. Rep. WO-GTR-10. Washington, DC: U.S. Department of Agriculture, Forest Service. 28 p.

Tongway, D.J.; Ludwig, J.A. 1990. Vegetation and soil patterning in semi-arid mulga lands of eastern Australia. Australian Journal of Ecology. 15: 23–34.

Tongway, D.J.; Ludwig, J.A. 1997. The conservation of water and nutrients within landscapes. In: Ludwig, J.A.; Tongway, D.J.; Freudenberger, D.O.; Noble, J.C.; Hodgkinson, K.C., eds. Landscape ecology, function, and management: Principles from Australia's rangelands. Melbourne, Australia: CSIRO Publishing: 13–22.

Tongway, D.J.; Ludwig, J.A.; Whitford, W.G. 1989. Mulga log mounds: Fertile patches in the semi-arid woodlands of eastern Australia. Australian Journal of Ecology. 14: 263–268.

Toy, T.J.; Foster, G.R.; Renard, K.G. 2002. Soil erosion: Processes, prediction, measurement, and control. New York: John Wiley and Sons, Inc. 352 p.

Tromble, J.M. 1983. Interception of rainfall by tarbush. Journal of Range Management. 36: 525–526.

Tromble, J.M. 1988. Water interception by two arid land shrubs. Journal of Arid Environments. 15: 65–70.

Tromble, J.M.; Renard, K.G.; Thatcher, A.P. 1974. Infiltration for three rangeland soil vegetation complexes. Journal of Range Management. 27: 318–321.

Turnbull, L.; Wainwright, J.; Brazier, R.E. 2010a. Changes in hydrology and erosion over a transition from grassland to shrubland. Hydrological Processes. 24: 393–414.

Turnbull, L.; Wainwright, J.; Brazier, R.E.; Bol, R. 2010b. Biotic and abiotic changes in ecosystem structure over a shrub-encroachment gradient in the Southwestern USA. Ecosystems. 13: 1239–1255.

Turnbull, L.; Wilcox, B.P.; Belnap, J.; Ravi, S.; D'Odorico, P.; Childers, D.; Gwenzi, W.; Okin, G.; Wainwright, J.; Caylor, K.K.; Sankey, T. 2012. Understanding the role of ecohydrologic feedbacks in ecosystem state change in drylands. Ecohydrology. 5: 174–183.

USDA Agricultural Research Service. 2011a. Long-term climate, precipitation, and stream discharge-sediment yield databases, Reynolds Creek Experimental Watershed, Idaho, United States. ftp://ftp.nwrc.ars.usda.gov/publicdatabase/reynoldscreek/ [Accessed March 23, 2011].

USDA Agricultural Research Service. 2011b. Long-term meteorological, precipitation, and runoff databases, Walnut Gulch Experimental Watershed, Arizona, United States. http://www.tucson.ars.ag.gov/dap/Default.htm [Accessed March 23, 2011].

U.S. Geological Survey. 2011. NLCD land cover 2006. Sioux Falls, SD: U.S. Department of the Interior, U.S. Geological Survey. http://www.mrlc.gov/nlcd2006_downloads.php [Accessed March 22, 2011].

Valentin, C. 1994. Surface sealing as affected by various rock fragment covers in West Africa. Catena. 23: 87–97.

Van Dijk, A.I.J.M.; Bruijnzeel, L.A.; Rosewell, C.J. 2002. Rainfall intensity-kinetic energy relationships: a critical literature appraisal. Journal of Hydrology. 261: 1–23.

Verheijen, F.G.A.; Cammeraat, L.H. 2007. The association between three dominant shrub species and water repellent soils along a range of soil moisture contents in semi-arid Spain. Hydrological Processes. 21: 2310–2316.

Wainwright, J. 2006. Climate and climatological variations. In: Havstad, K.M.; Huenneke, L.F.; Schlesinger, W.H.; eds. Structure and function of a Chihuahuan Desert ecosystem: The Jornada Basin Long-Term Ecological Research Site. Long Term Ecological Research Network Series. New York: Oxford University Press: 44–80.

Wainwright, J.; Parsons, A.J.; Abrahams, A.D. 1999. Rainfall energy under creosotebush. Journal of Arid Environments. 43: 111–120.

Wainwright, J.; Parsons, A.J.; Abrahams, A.D. 2000. Plot-scale studies of vegetation, overland flow and erosion interactions: Case studies from Arizona and New Mexico. Hydrological Processes. 14: 2921–2943.

Wainwright, J.; Parsons, A.J.; Schlesinger, W.H.; Abrahams, A.D. 2002. Hydrology-vegetation interactions in areas of discontinuous flow on a semi-arid bajada, Southern New Mexico. Journal of Arid Environments. 51: 319–338.

Walter, H. 1971. Ecology of tropical and subtropical vegetation. Edinburgh, Scotland: Oliver and Boyd. 688 p.

Wan, C.; Sosebee, R.E.; McMichael, B.L. 1993. Does hydraulic lift exist in shallow-rooted species? A quantitative examination with a half-shrub *Gutierrezia sarothrae*. Plant and Soil. 153: 11–17.

Wells, C.G.; Campbell, R.E.; DeBano, L.F.; Lewis, C.E.; Fredrickson, R.L.; Franklin, E.C.; Froelich, R.C.; Dunn, P.H.1979. Effects of fire on soil: A state-of-the-knowledge review. Gen. Tech. Rep. WO-7. Washington, DC: U.S. Department of Agriculture, Forest Service. 34 p.

Weltz, M.A.; Blackburn, W.H. 1995. Water budget for south Texas rangelands. Journal of Range Management. 48: 45–52.

West, N.E.; Gifford, G.F. 1976. Rainfall interception by cool-desert shrubs. Journal of Range Management. 29: 171–172.

Westerling, A.L.; Hidalgo, H.G.; Cayan, D.R.; Swetnam, T.W. 2006. Warming and earlier spring increase western U.S. forest wildfire activity. Science. 313: 940–943.

Whisenant, S.G. 1990. Changing fire frequencies on Idaho's Snake River plains: Ecological and management implications. In: McArthur, E.D.; Romney, E.M.; Smith, S.D.; Tueller, P.T., comps. Proceedings—symposium on cheatgrass invasion, shrub die-off, and other aspects of shrub biology and management. 1989 April 5–7; Las Vegas, NV. Gen. Tech. Rep. INT-GTR-276. Ogden, UT: U.S. Department of Agriculture, Forest Service, Intermountain Research Station: 4–10.

Whitford, W.G.; Anderson, J.; Rice, P.M. 1997. Stemflow contribution to the "fertile island" effect in creosotebush, *Larrea tridentata*. Journal of Arid Environments. 35: 451–457.

Wilcox, B.P. 1994. Runoff and erosion in intercanopy zones of pinyon-juniper woodlands. Journal of Range Management. 47: 285–295.

Wilcox, B.P.; Breshears, D.D. 1995. Hydrology and ecology of piñon-juniper woodlands: Conceptual framework and field studies. In: Shaw, D.W.; Aldon, E.J.; LoSapio, C., eds. Desired future conditions for pinyon-juniper ecosystems; 1994 August 8–12; Flagstaff, AZ. Gen. Tech. Rep. RM-258. Fort Collins, CO: U.S. Department of Agriculture, Forest Service, Rocky Mountain Forest and Range Experimental Station: 109–119.

Wilcox, B.P.; Breshears, D.D.; Allen, C.D. 2003a. Ecohydrology of a resource-conserving semiarid woodland: Effects of scale and disturbance. Ecological Monographs. 73: 223–239.

Wilcox, B.P.; Breshears, D.D.; Seyfried, M.S. 2003b. Water balance on rangelands. Encyclopedia of Water Science. doi: 10.1081/E-EWS 120010097.

Wilcox, B.P.; Breshears, D.D.; Turin, H.J. 2003c. Hydraulic conductivity in a piñon-juniper woodland: Influence of vegetation. Soil Science Society of America Journal. 67: 1243–1249.

Wilcox, B.P.; Dowhower, S.L.; Teague, W.R.; Thurow, T.L. 2006. Long-term water balance in a semiarid shrubland. Rangeland Ecology and Management. 59: 600–606.

Wilcox, B.P.; Hanson, C.L.; Wight, J.R.; Blackburn, W.H. 1989. Sagebrush rangeland hydrology and evaluation of the SPUR hydrology model. Water Resources Bulletin. 25: 653–666.

Wilcox, B.P.; Huang, Y.; Walker, J.W. 2008. Long-term trends in streamflow from semiarid rangelands: Uncovering drivers of change. Global Change Biology. 14: 1676–1689.

Wilcox, B.P.; Newman, B.D.; Brandes, D.; Davenport, D.W.; Reid, K. 1997. Runoff from a semiarid ponderosa pine hillslope in New Mexico. Water Resources Research. 33: 2301–2314.

Wilcox, B.P.; Pitlick, J.; Allen, C.D.; Davenport, D.W. 1996. Runoff and erosion from a rapidly eroding pinyon-juniper hillslope. In: Anderson, M.G.; Brooks, S.M., eds. Advances in hillslope processes, Volume 1. New York: John Wiley and Sons, Ltd.: 61–77.

Wilcox, B.P.; Seyfried, M.S.; Cooley, K.R.; Hanson, C.H. 1991. Runoff characteristics of sagebrush rangelands: Modeling implications. Journal of Soil and Water Conservation. 46: 153–158.

Wilcox, B.P.; Thurow, T.L. 2006. Preface: Emerging issues in rangeland ecohydrology. Hydrological Processes. 20: 3155–3157.

Wilcox, B.P.; Wood, M.K.; Tromble, J.M. 1988. Factors influencing infiltrability of semiarid mountain slopes. Journal of Range Management. 41: 197–206.

Williams, C.J.; McNamara, J.P.; Chandler, D.G. 2009. Controls on the temporal and spatial variability of soil moisture in a mountainous landscape: The signatures of snow and complex terrain. Hydrology and Earth System Science. 13: 1325–1336.

Williams, C.J.; Pierson, F.B.; Al-Hamdan, O.Z.; Kormos, P.R.; Hardegree, S.P.; Clark, P.E. 2014a. Can wildfire serve as an ecohydrologic threshold-reversal mechanism on juniper-encroached shrublands. Ecohydrology. 7: 453–477. doi: 10.1002/eco.1364.

Williams, C.J.; Pierson, F.B.; Robichaud, P.R.; Boll, J. 2014b. Hydrologic and erosion responses to wildfire along the rangeland-xeric forest continuum in the western US: A review and model of hydrologic vulnerability. International Journal of Wildland Fire. 23: 155–172.

Williams, C.J.; Pierson, F.B.; Robichaud, P.R.; Al-Hamdan, O.Z.; Boll, J.; Strand, E.K. 2016a. Structural and functional connectivity as a driver of hillslope erosion following disturbance. International Journal of Wildland Fire. 25(3): 306–321. doi: 10.1071/WF14114.

Williams, C.J.; Pierson, F.B.; Spaeth, K.E.; Brown, J.R.; Al-Hamdan, O.Z.; Weltz, M.A.; Nearing, M.A.; Herrick, J.E.; Boll, J.; Robichaud, P.R.; Goodrich, D.C.; Heilman, P.; Guertin, D.P.; Hernandez, M.; Wei, H.; Hardegree, S.P.; Strand, E.K.; Bates, J.D.; Metz, L.J.; Nichols, M.H. 2016b. Incorporating hydrologic data and ecohydrologic relationships into Ecological Site Descriptions. Rangeland Ecology and Management. 69(1): 4–19.

Winstral, A.; Marks, D. 2002. Simulating wind fields and snow redistribution using terrain-based parameters to model snow accumulation and melt over a semi-arid mountain catchment. Hydrological Processes. 16: 3585–3603.

Witter, J.V.; Jungerius, P.D.; ten Harkel, M.J. 1991. Modelling water erosion and the impact of water-repellency. Catena. 18: 115–124.

Wood, M.K.; Blackburn, W.H. 1981. Grazing systems: Their influence on infiltration rates in the Rolling Plains of Texas. Journal of Range Management. 34: 331–335.

Wood, M.K.; Blackburn, W.H.; Eckert, R.E., Jr.; Peterson, F.F. 1978. Interrelations of the physical properties of coppice dune and vesicular dune interspace soils with grass seedling emergence. Journal of Range Management. 31: 189–192.

Wood, M.K.; Jones, T.L.; Vera-Cruz, M.T. 1998. Rainfall interception by selected plants in the Chihuahuan Desert. Journal of Range Management. 51: 91–96.

Woods, S.W.; Balfour, V.N. 2008. The effect of ash on runoff and erosion after a severe forest wildfire, Montana, USA. International Journal of Wildland Fire. 17: 535–548.

Woods, S.W.; Birkas, A.; Ahl, R. 2007. Spatial variability of soil hydrophobicity after wildfires in Montana and Colorado. Geomorphology. 86: 465–479.

Wright, H.A.; Bailey, A.W. 1982. Fire ecology: United States and Southern Canada. New York: Wiley and Sons. 501 p.

Yoder, C.K.; Nowak, R.S. 1999a. Hydraulic lift among native plant species in the Mojave Desert. Plant and Soil. 95: 93–102.

Yoder, C.K.; Nowak, R.S. 1999b. Soil moisture extraction by evergreen and drought-deciduous shrubs in the Mojave Desert during wet and dry years. Journal of Arid Environments. 42: 81–96.

Young, J.A.; Evans, R.A. 1978. Population dynamics after wildfires in sagebrush grasslands. Journal of Range Management. 31: 283–289.

Young, R.A.; Wiersma, J.L. 1973. The role of rainfall impact in soil detachment and transport. Water Resources Research. 9: 1629–1636.

Zhang, L.; Dawes, W.R.; Walker, G.R. 2001. Response of mean annual evapotranspiration to vegetation changes at catchment scale. Water Resources Research. 37: 701–708.

Zou, C.B.; Barnes, P.W.; Archer, S.; McMurtry, C.R. 2005. Soil moisture redistribution as a mechanism of facilitation in savanna tree-shrub clusters. Oecologia. 145: 32–40.

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