Rangeland Processes: Hydrology and Soil Erosion

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Editors: Mark Weltz and Gary Frasier

Contributors:

Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Mark A. Weltz, Research Leader, Rangeland Hydrologist, USDA-ARS, Reno, NV.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ, Tucson, AZ.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX
Kenneth McGwire, Physical Scientist, Desert Research Institute, Reno, NV.
Jason Nesbit, IT Specialist, USDA-ARS, Reno, NV.

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Abstract: The purpose of this series of Rangeland Hydrology Handbooks is to improve the understanding of hydrologic processes and sources and transport mechanisms of sediment in rangeland catchments. Soil loss rates on rangelands are considered one of the few quantitative indicators for assessing soil quality, rangeland health, and conservation practice effectiveness. Concentrated flow erosion processes are distinguished from splash and sheetflow processes in their enhanced ability to mobilize and transport large amounts of soil, water, and dissolved elements. On rangelands, soil, nutrients, and water are scarce, and only narrow margins of resource losses are tolerable before crossing the sustainability threshold. In these ecosystems, rill flow processes are indicators of degradation and often warrant the implementation of mitigation strategies. Vegetation lifeform type, distribution, and amount influence both infiltration and soil erosion processes. Vegetation and ground cover are the principal environmental attributes that can be manipulated by managers on rangelands and have a direct impact on raindrop splash and concentrated flow erosion processes and rates. At the ecohydrologic level, vegetation and concentrated flow pathways are engaged in a feedback relationship, the understanding of which can help improve rangeland management and restoration strategies. In the USDA-NRCS National Range and Pasture Handbook, Part 645, Subpart G, rangeland ecohydrologic principals and plant/soil water interactions are discussed in detail (to be released in the near future). A compete review of fire and how it can dramatically influence rangeland hydrology and erosion by altering ecohydrologic relationships is presented in Pierson, F.B., Williams, C.J. (2016). Ecohydrologic impacts of rangeland fire on runoff and erosion: A literature synthesis, U. S. Dept of Agriculture, Forest Service, Rocky Mountain Research Station, Fort Collins. In this handbook, Part 646, we review published literature pertaining to rangeland hillslope scale hydrologic and soil erosion processes to (1) present the fundamental science underpinning ecohydrologic processes, (2) present raindrop splash and concentrated flow erosion processes, (3) discuss the influence of vegetation on these erosion processes, (4) discuss the influence of management on these erosion processes, and (5) discuss the history and current state of modeling runoff and soil erosion processes at the hillslope scale on rangelands. The tools to predict soil erosion and detailed user guides are provided in the second handbook, Part 647: "Rangeland Hydrology and Soil Erosion Processes: A guide for Conservation Planning with the Rangeland Hydrology and Erosion Model (RHEM)." The RHEM assessment tool provides information that can be combined with ecological state and transition models and enhance Ecological Site Descriptions (ESDs). The RHEM assessment tool has also been incorporated into the Automated Geospatial Watershed Assessment (AGWA) tool for understanding and predicting hydrologic and soil erosion processes at the watershed scale. The AGWA tool is a GIS-based hydrologic modeling tool that uses commonly available GIS data layers to fully parameterize, execute, and spatially visualize results for the RHEM, KINEROS2, KINEROS-OPUS, SWAT2000, and SWAT2005 watershed runoff and erosion models. The tool and documentation can be accessed at https://www.tucson.ars.ag.gov/agwa/. With these handbooks and tools, the user can understand causes and

consequences of soil erosion and design management plans to prevent or correct issues of concern on rangelands at scales ranging from hillslope to watersheds.

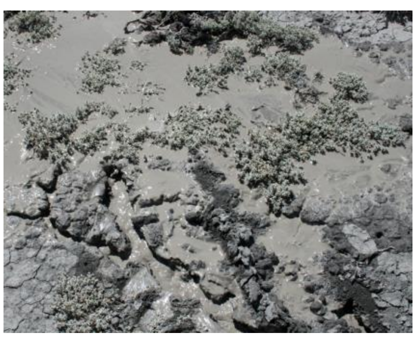
Keywords: soil erosion; rangelands; rill erosion; concentrated flow; interrill erosion; soil erodibility; slope length, steepness, and shape; runoff; infiltration; risk assessment; foliar and ground cover; soil texture; precipitation intensity; duration and frequency; Ecological Site Description; conservation practice; grazing management, brush management, and fire.

Figure i. Soil erosion processes on rangelands: (A) Convective thundershower, (B) Raindrop splash erosion, (C) Sheetflow erosion, (D) Concentrated flow erosion, (E) Channel erosion during a flash flow, and (F) Gully erosion.

(A) Convective thundershower.

(B) Raindrop splash erosion.

(C) Concentrated flow erosion



(D) Sheet flow erosion

(E) Channel flow erosion

(F) Gully erosion.

Part 646 – Rangeland Processes Handbook: Hydrology and Soil Erosion

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Figure ii. Walnut Gulch (left) and Rotating boom (right) rainfall simulators used to collect field data to develop the Rangeland Hydrology and Erosion Model (RHEM) and further understanding of hydrologic and hillslope scale soil erosion processes on rangelands.

Introduction

A. Soil erosion on rangelands was recognized as a serious problem at both local and national scales in the United States in the 1920s. By 1935 soil erosion was considered a national menace on an area covering over half of the country (Bennett and Chapline, 1928; Chapline, 1929; Sampson and Weyl, 1918; Weaver, 1935). "Soil erosion" is an all-inclusive term describing the deflation of the landscape by wind and water. Specific terms like "interrill" and "rill" erosion are used to define the detachment of soil particles by raindrop impact and by flowing water, respectively (table i).

Table i. Definition of soil erosion terms.

Term	Definition
Interrill erosion	Detachment of soil particles by raindrop impact and their transport by broad sheet flow to concentrated flow areas.
Rill erosion	Detachment and transport of soil particles by concentrated (rill) flow.
Sediment transport	Movement of detached soil particles (sediment).
Deposition	Settlement of detached soil particles.
Sediment yield	Total sediment outflow per unit area measured at a point of reference and for a specific time period (including deposition)
Soil loss	Quantity per unit area and time of soil detached and transported from an area, without significant deposition.
Sediment discharge	The rate of movement of a mass of sediment past a point or through a cross section, related to velocity.

B. In natural plant communities, the erosion potential of a site is the result of complex interactions between soil, vegetation, topographic position, land use and management, and climate. Soil erosion is a natural process, but the quantity and rate of surface runoff and sediment yield may be altered and accelerated through land use and management practices. Many abiotic and biotic factors affect soil erosion and sediment yield on rangelands. Plant and ground surface cover variables influence runoff and the basic erosion processes of soil detachment by raindrops and concentrated flow, sediment transport, and sediment deposition through the amount and distribution of exposed bare soil, the tortuosity and connectivity of the concentrated flow paths, hydraulic roughness, and soil properties of the site (i.e., interrill and rill erodibility). Soil erosion is a function of:

(1) total standing biomass
(2) biomass by lifeform class (i.e., grass vs. shrub)
(3) distance between plants
(4) canopy cover
(5) components of ground cover (rock, litter, plant basal area, biological soil crust)
(6) bare soil
(7) soil bulk density
(8) soil texture
(9) soil organic carbon
(10) soil aggregate stability
(11) amount and connectivity of interspace or coppice dune areas
(12) number or size of surface depressions
(13) rainfall intensity
(14) rainfall duration
(15) antecedent soil moisture

C. The complex interaction of these and other abiotic and biotic variables determines how much, when, and where soil erosion will occur across upland rangeland hillslopes.

Figure iii. Soil erosion control using straw-filled wattles in Pelekane Bay, Hawaii.

RHEM Guides for specific Ecological Sites

RHEM Guides for specific Ecological Sites, State-and-Transition models, and disturbances are available as Appendices on the Title 190, Part 647 eDirectives web page. They have been developed to quantify how changes in state within an ecological site influences hydrologic and soil erosion process at the scale of a hillslope. The Guides provide documented case studies that discuss how to define sustainability in the context of risk of soil erosion. The Guides discuss when sites may cross an ecological or environmental threshold where restoration may not be physically possible. By altering model inputs for vegetation community attributes, the RHEM tools allow for evaluation of hydrologic and soil loss response in relation to defined user management action (i.e., percent increase in bunchgrass and decrease in bare soil derived from a seeding practice). This allows quantification of conservation benefits. By comparing runoff from various states within an ecological site one can enhance ecological sites descriptions by incorporating this information, how hydrologic properties change as plant communities change, into ecological site descriptions.

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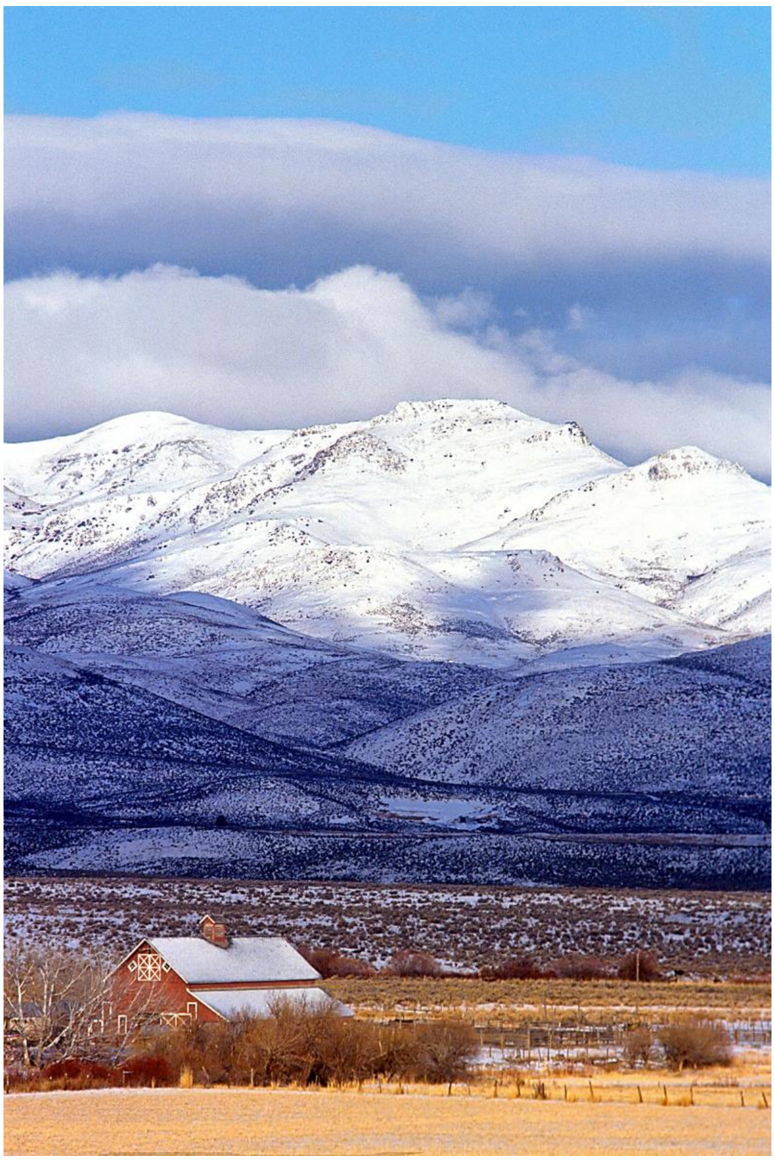
Subpart A – Rangeland Hydrologic Cycle

646.01 Contributors:

C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.

Figure A-1. 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).



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646.02 Hydrologic Cycle

A. The hydrologic cycle (figure A-2) 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 (Brutsaert, 2005; Dingman, 2002; Hornberger et al., 1998; Maidment, 1993; Weltz and Blackburn, 1995).

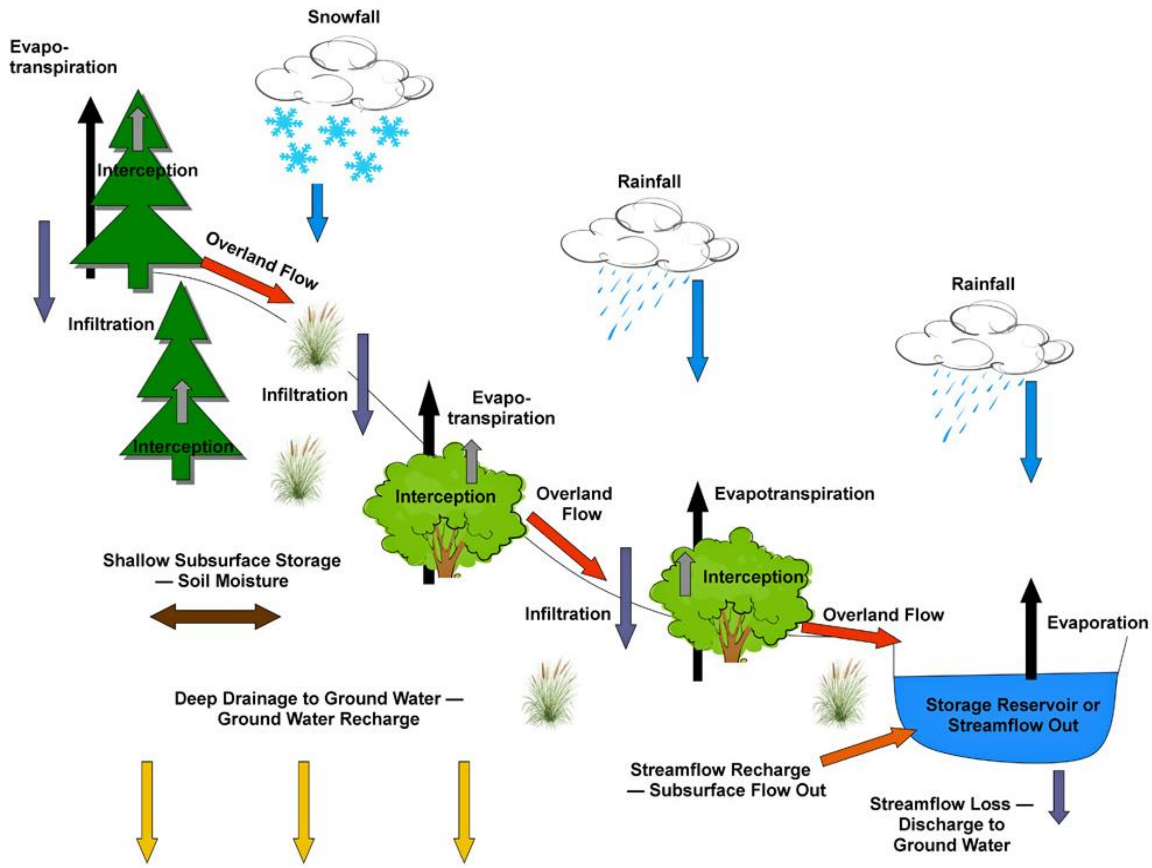
B. Knowledge of the hydrologic cycle provides a useful framework from which to conceptualize and understand vegetation-soil-climate-hydrology interactions and to understand how fire and other disturbances influence runoff and erosion behavior. For a particular hillslope or watershed, the cycle consists of water inflow, transit and storage, and outflow. Inflow primarily occurs as precipitation, overland flow, streamflow from upslope areas, and ground water returns from springs and into streambeds and lakebeds.

C. Transit and storage components of water arriving at the land-atmosphere interface include precipitation (rain and snow), interception, infiltration, and water storage on and underneath the land surface. Outflows include gaseous phase losses to the atmosphere through evaporation and transpiration (evapotranspiration), watershed runoff, and deep drainage to aquifers.

646.03 Precipitation

A. The primary water input to the land-atmosphere interface on rangelands 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 below 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.

Figure A-2. Illustration of the 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). (Caption and figure from Pierson and Williams, 2016.)



B. Storm 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 of warm air over the cool air at the surface generates precipitation, commonly at an extremely high intensity (Dingman, 2002).

C. 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, cause flash flood events.

D. 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 (e.g., Great Basin Desert) form on the leeward side of mountains as the air crests and warms, and clouds dissipate. Orographic events are often associated with frontal or convective events that encounter a topographic barrier.

E. 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) (table A-1).

Table A-1. Locations of national and international climate data.

Climate Data Source	Web Site
National Oceanic and Atmospheric Administration	http://hdsc.nws.noaa.gov/hdsc/pfds/pfds_map_cont.html
National Weather Service	https://www.weather.gov/help-past-weather
Natural Resources Conservation Service	https://www.wcc.nrcs.usda.gov/scan/
World Meteorological Organization	http://worldweather.wmo.int/en/home.html
Food and Agricultural Organization of the United Nations	http://geonetwork3.fao.org/climpag/agroclimdb_en.php

F. Rainfall depth is the quantity of accumulated rainfall (expressed as a length measurement such as mm or cm) at a point on the landscape (e.g., 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 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 (e.g., 48 mm during a 45-min interval) or an intensity-duration relationship (e.g., 64 mm h^{-1} for 45 min) and are related to specified recurrence intervals (frequency) or return periods (e.g., a 100-year event).

G. 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. The 100-year precipitation event has a one in 100 chance, or one percent probability, of occurring each year. Precipitation frequency estimates for average recurrence intervals from one to 1,000-year durations can be found at the following NOAA site: http://hdsc.nws.noaa.gov/hdsc/pfds/pfds_map_cont.html. The various types of precipitation gauges, surveys (figure A-3), and methods for extrapolating and applying precipitation data from available sources can be found in most hydrology textbooks (Brutsaert, 2005; Dunne and Leopold, 1978; Haan, 2002; McCuen, 1989).

Figure A-3. Snow measurement techniques used by the USDA Natural Resources Conservation Service (photo A: Agricultural Research Service; photo B: Ron Nichols, Natural Resources Conservation Service).

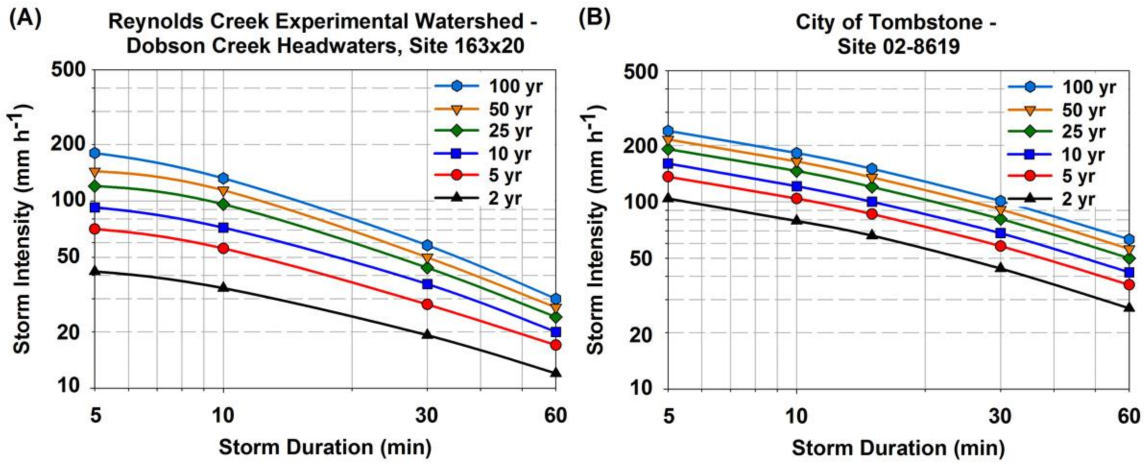
(A) Examples of shielded (right) and unshielded (left) precipitation gages.

(B) Snow water equivalent measurement along a snow course transect.

H. 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) (figure A-4). Average annual rainfall is calculated from gauges with long-term records. The total rainfall depth in a given year may be reported as a percentage of the average annual rainfall (e.g., 120 percent) for a number of complete years of record divided by the number of years used. Snowfall is most often measured as a storm-specific depth (accumulation) of newly fallen snow at a point, as a snowpack accumulation at the land surface over some duration (daily, monthly, seasonally, or annually), or as 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 (Dingman, 2002).

I. Snow water equivalent (SWE) 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. Some of the snow is lost directly into the atmosphere by sublimation.

Figure A-4. 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); (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). (Caption and figure from Pierson and Williams, 2016.)



J. 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).

K. 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 (figure A-5). In addition to annual quantity differences, the timing or seasonality of precipitation inputs varies significantly across U.S. rangelands (figure A-6).

Figure A-5. (A) Annual precipitation (PRISM, 2011) and (B) landcover (U.S. Geological Survey 2011) for the western United States. (Caption and figure from Pierson and Williams, 2016.)

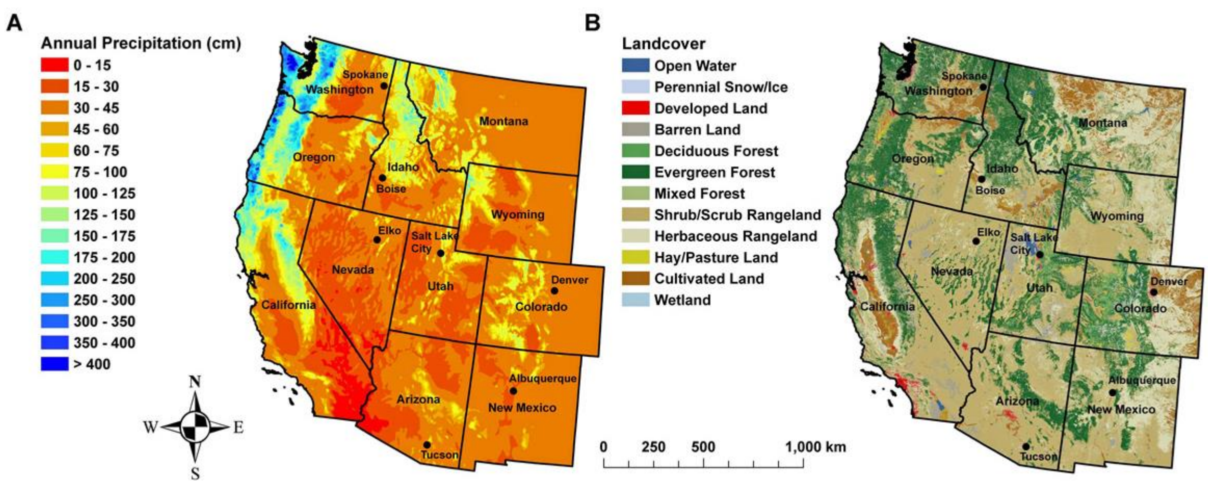
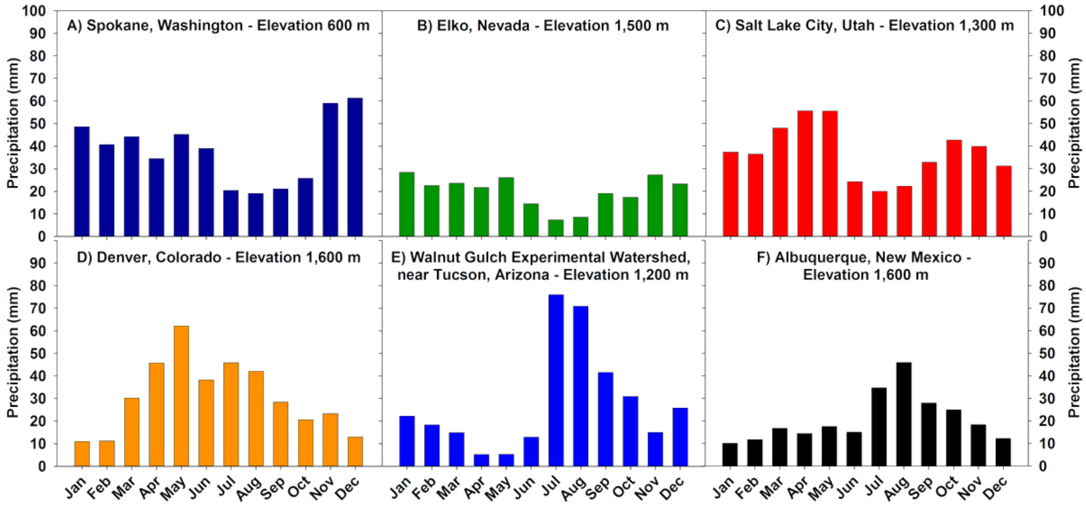
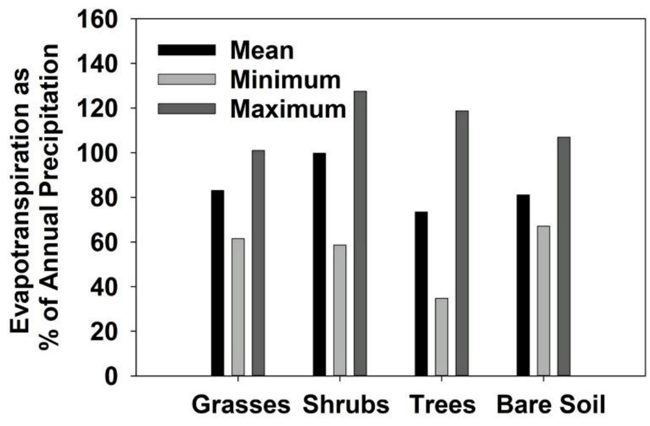


Figure A-6. Average monthly precipitation for western U.S. urban centers immediately adjacent to rangelands (PRISM, 2011). Panels A–C are indicative of annual precipitation in the Inland Northwest. Panel D is indicative of annual precipitation trends on central U.S. rangelands. Panels E–F demonstrate the influence of the monsoonal season (July–August) on the rainfall in the desert Southwest. (Caption and figure from Pierson and Williams, 2016.)



L. Precipitation inputs on southwestern U.S. rangelands (figure A-6E and F) occur mostly during the summer monsoonal season as intense convective thunderstorms (up to 60 to 70 percent of annual precipitation can occur during July and August) (Goodrich et al., 2008; Mendez et al., 2003; Osborn, 1983a; Osborn, 1983b; Wainwright, 2006). Most of the annual precipitation in the northwest and north-central United States falls from November through May and is snowfall-dominated in mountain locations and rainfall-dominated in valley locations (figure A-7 and figure A-8).

Figure A-7. 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 (Brandes and Wilcox, 2000; Carlson and Thurow, 1996; Carlson et al., 1990; Flerchinger and Cooley, 2000; Flerchinger et al., 1998; Joffre and Rambal, 1993; Weltz and Blackburn, 1995). (Caption and figure from Pierson and Williams, 2016.)



M. Rainfall patterns at these locations usually occur as low-intensity, long-duration events, in late autumn or winter, and during a four to eight-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 (Pierson et al., 2001; Wilcox et al., 1989). Mountain and valley locations experience high-intensity convective storms during the dry summer months. Annual precipitation is usually greater at higher elevations than on valley floors due to adiabatic processes (figure A-9). Precipitation trends are bimodal for some south-central U.S. rangelands (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; Bowen, 1996). Elevational precipitation trends for the central United States are similar to those occurring on northwest and north-central U.S. rangelands. Precipitation and streamflow shown are from a convective thunderstorm even (~22 mm rainfall during two hours) occurring July 21, 1971, in the USDA Agricultural Research Service Reynolds Creek Experimental Watershed, as recorded at the Tollgate Weir Site 116 x 83. Event and base flow contribution to streamflow are separated by the dashed line. Arrows and callouts indicate rising limb, peak discharge, and falling limb of the hydrograph as responses to the storm event. (Caption and figure from Pierson and Williams, 2016.)

Figure A-8. Example streamflow hydrograph and precipitation hyetograph.

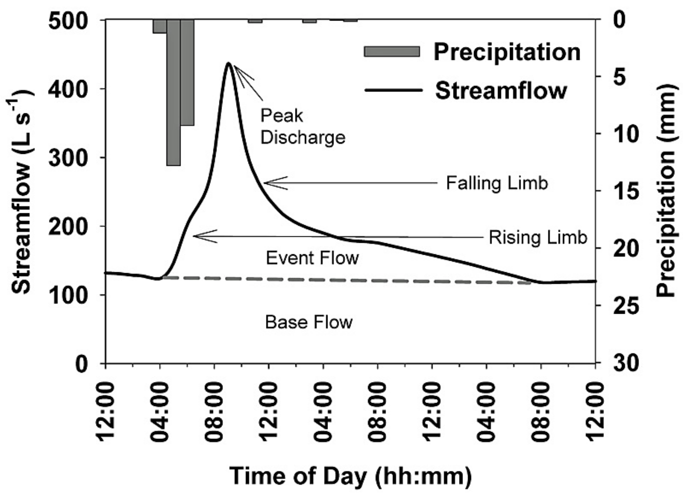
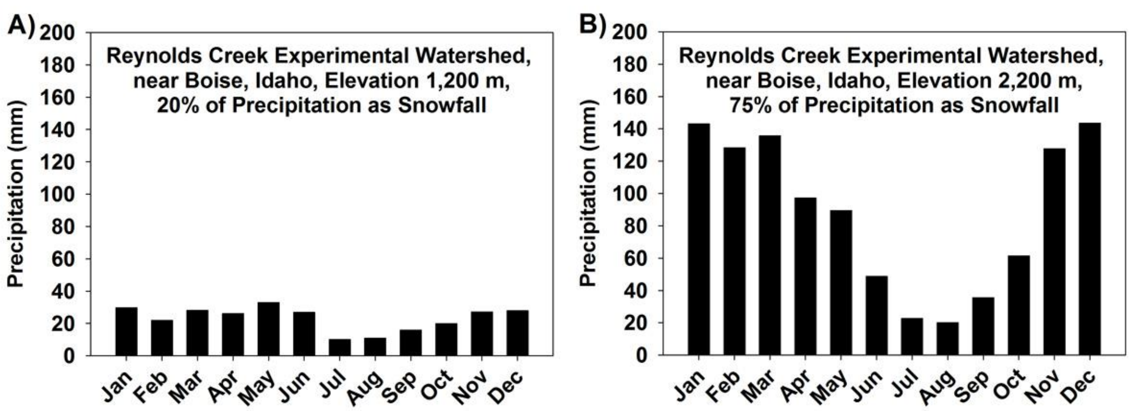


Figure A-9. Average monthly precipitation for two climate stations (A: Site 076x59, low elevation; B: 163x20, high elevation) located in the Reynolds Creek Experimental Watershed, Idaho (Hanson and Pierson, 2001). Precipitation differences between the two sites are primarily driven by the difference in elevation. (Caption and figure from Pierson and Williams, 2016.)



B. Interception is strongly influenced by the precipitation frequency and intensity and by the type and structure of the vegetative community (Branson and Miller, 1981; Dunkerley, 2008; Owens et al., 2006). Interception losses from a series of small storms and clearing events are proportionately greater than those from large, prolonged rainfall events (Hamilton and Row, 1949; Owens et al., 2006; Rowe, 1948). 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 conifers than broad-leaved tree species and is greater from tree species than from shrubs and grasses.

646.04 Interception

A. Precipitation arriving at the land-atmosphere interface either is intercepted by vegetation, rocks, or litter debris or falls unimpeded to the soil surface (figure A-10). Intercepted precipitation evaporates into the atmosphere (interception loss) or is transferred as liquid water to the soil surface as throughfall or stemflow (Abrahams et al., 2003; Bhark and Small, 2003; Carlyle-Moses, 2004; Martinez-Meza and Whitford, 1996; Navar and Bryan, 1990; Owens et al., 2006; Thurow et al., 1987; Wainwright et al., 1999; Whitford et al., 1997). 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 to the soil surface. 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).

B. Interception is strongly influenced by the precipitation frequency and intensity and by the type and structure of the vegetative community (Branson and Miller, 1981; Dunkerley, 2008; Owens et al., 2006). Interception losses from a series of small storms and clearing events are proportionately greater than those from large, prolonged rainfall events (Hamilton and Row, 1949; Owens et al., 2006; Rowe, 1948). 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 conifers than broad-leaved tree species and is greater from tree species than from shrubs and grasses.

C. Variations in the canopy density of individual plants and the vegetative community complicate quantification of interception losses over large scales (table A-2). Eddleman (1994) reported that canopy interception in western Juniper can exceed 12 percent of annual precipitation. 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).

Table A-2. Event and annual interception rates (as percent of gross rainfall) reported in literature for various individual rangeland plant and community types (Branson et al., 1981).

Cover type	Event interception (%)	Annual interception (%)
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

Figure A-10. Measuring interception, stemflow, and throughfall on a Juniper tree to determine components of the water balance on Wyoming Big sagebrush plant community that has been encroached by Juniper in Desatoya Mountains, central Nevada.

D. Gross rainfall interception by individual rangeland shrubs (Branson et al., 1981; Tromble, 1983) and conifer trees (Owens et al., 2006; Slaughter, 1997; 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 multi-storm to annual timescales ranges from 5 to 46 percent (Domingo et al., 1998; Martinez-Meza and Whitford, 1996; Navar and Bryan, 1990; Serrato and Diaz, 1998; Thurow et al., 1987; Tromble, 1988), 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 (Carlyle-Moses, 2004; Dunkerley and Booth, 1999; Pressland, 1973; Thurow et al., 1987; Tromble, 1988; Tromble, 1983). 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. They estimated that interception of gross annual rainfall at two Texas grassland sites with 56 percent cover (shortgrass) and 62 percent cover (midgrass) was 11 percent and 18 percent, respectively. Dunkerley and Booth (1999) reported a 32 percent interception of gross annual rainfall by grass in Australia. Pierson and Williams (2016) provide a thorough review of canopy interception on rangelands.

646.05 Infiltration

A. 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 (Branson et al., 1981; Brutsaert, 2005; Dingman, 2002; Dunne and Leopold, 1978; Hillel, 1998). 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 a 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).

B. 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 and may result in significant downstream watershed flooding (Freudiger et al., 2014).

C. Water infiltrates soil mainly due to a negative pressure gradient or suction (matrix suction) into the soil matrix and secondarily due to gravity (Brutsaert, 2005; Dingman, 2002; Hillel, 1998; Hornberger et al., 1998; Selby, 1982). 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 is caused 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.

D. Porous or rough surfaces are usually more conductive than uniform surfaces. However, high-intensity rainfall can break apart highly conductive porous structures or aggregates of surface soils, facilitating infilling of soil pores with fine soil particles (Selby, 1982; Thornes, 1980). Infilling of pores creates a hydrologic barrier by sealing the surface of the material. 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 et al., 1990; Blackburn and Wood, 1990).

E. 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, 1998), 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 (DeBano and Rice, 1973; Doerr et al., 2000; Doerr et al., 2009; Meeuwig, 1971). The formation of water-repellent soils and the effects on infiltration are discussed in more detail in Pierson and Williams (2016).

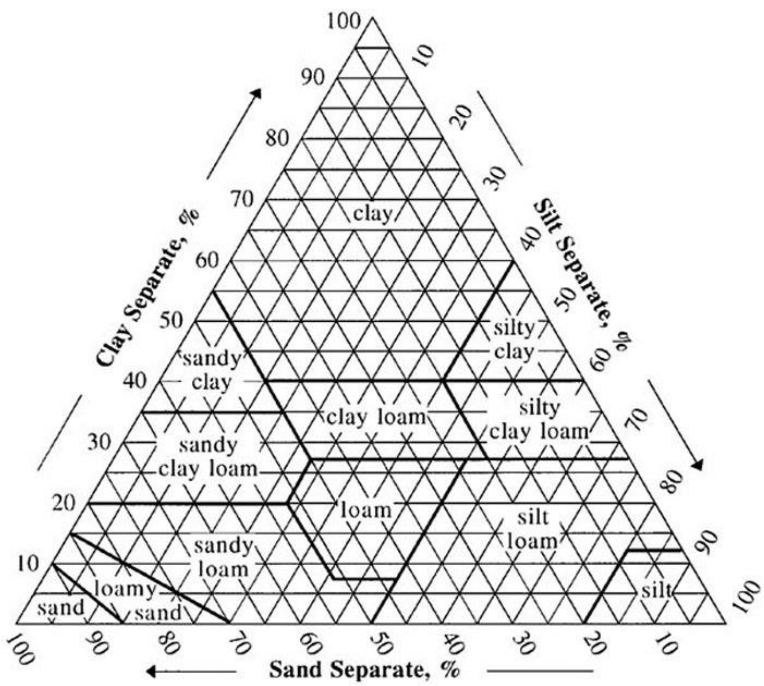
F. Infiltration into wet soils is significantly influenced by the saturated hydraulic conductivity of the soil profile (Dingman, 2002; Hillel, 1998; Selby, 1982). Once the soil profile becomes wet, any further water input is partially dependent on the redistribution or downward transmittance (percolation) of existing soil water. Hydraulic conductivity (measured as length per unit of 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. 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.

G. Finally, redistribution and, hence, infiltration is also influenced by the structure of the soil profile. The presence of an impeding or 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 flow (macropore or preferential). The processes and measurement methods for water infiltration and redistribution in unsaturated and saturated soils are described in detail by Hillel (1998).

646.06 Soil Water Storage and Ground Water Recharge

A. Soil texture is an important soil characteristic that influences infiltration, redistribution of water, and storage of water in the soil profile. Soil texture is classified based on the percentage of sand, silt, and clay. The clay fraction is made up of particles < 0.002 mm in diameter, silt ranges from 0.002–0.05 mm, and sand particles range in size from 0.05–2.0 mm. Rocks > 2 mm in diameter are not considered soil and are not used when determining soil texture. USDA uses 12 different classes to represent soil texture. Once the percent sand, silt, and clay are known, then soil texture can be identified (figure A-11). Calculations of soil texture can be simply done using a web application developed by the Natural Resources Conservation Service at: https://www.nrcs.usda.gov/wps/portal/nrcs/detail/soils/survey/?cid=nrcs142p2_054167

Figure A-11. USDA soil textural triangle used to calculate soil texture from the percentage of sand, silt, and clay in the soil.



(1) Soil texture influences the rate at which water drains through soil. Water moves faster through sandy soils than it does through clayey soils. 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.

(2) The unsaturated zone encompasses the soil water or rooting zone, an intermediate zone, and the capillary fringe immediately overlying the saturated zone (Dingman, 2002; Hornberger et al., 1998).

(i) 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.

(ii) 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.

(iii) Pore spaces in the capillary fringe are saturated or are near saturation and pore water is held there by capillary forces (Dingman, 2002; Hornberger et al., 1998). Generally, less than one percent of water entering the soil profile passes through the soil water and intermediate zones into deep storage below the capillary fringe.

B. The discussion on soil water movement and storage is restricted to the unsaturated zone because most surface responses are associated with soil water content in the rooting zone of the soil profile that affect upland rangeland soil erosion processes. Explanations of the soil physics that dictate soil water movement and retention in the unsaturated and saturated zones are provided in Dingman (2002), Hillel (1998), and Kramer and Boyer (1995). 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. 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 (the product of volumetric water content and the thickness of the layer) results from the resolution of the matrix and gravitation forces, and plant water use.

C. Water content available for plants is the difference between the field capacity content and permanent wilting point water content (Dingman, 2002; Kramer and Boyer, 1995). 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 (Dingman, 2002; Kramer and Boyer, 1995) is the soil water content at which water availability is too low to support plant transpiration demands and the point at which plants wilt. 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 retain limited water storage capacity. The presence of organic matter in soils usually increases the water storage capacity.

646.07 Evapotranspiration and Transpiration

A. 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 on arid rangelands (Branson et al., 1981; Brutsaert, 2005; Dingman, 2002; Hornberger et al., 1998; Kramer and Boyer, 1995). 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) (Branson et al., 1981; Dingman, 2002).

B. Transpiration is the direct evaporation of liquid water pulled 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, photosynthetic pathways (plant water use strategies), soil microclimate, water availability, and plant influences on wind and the aboveground microclimate (Branson et al., 1981; Dingman, 2002; Hornberger et al., 1998; Kramer and Boyer, 1995).

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

D. 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 of annual precipitation (but can be as high as 50 percent), depending on the type and structure of the plant community, meteorological patterns, and soils and geology (Carlson and Thurow, 1996; Flerchinger and Cooley, 2000; Nearing et al., 2007; Weltz and Blackburn, 1995 and 1996; Wilcox et al., 2006, 2008, and 1997). Runoff generation from rainfall-dominated rangelands primarily occurs as infiltration-excess overland flow and is minimally influenced by evapotranspiration demands (Pierson et al., 2001; Stone et al., 2008; Wilcox et al., 2003b). Exceptions occur following multi-storm events or prolonged low-intensity storms that wet up the near-surface environment, shifting the runoff process to saturation-excess (Castillo et al., 2003; Wilcox, 1994). Evapotranspiration demands at snow-dominated sites strongly influence the seasonal duration of ephemeral streamflow, along with seasonal water input (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. Estimates from literature indicate evapotranspiration as a percentage of precipitation (figure A-7).

E. 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 between plant communities (7 to 80 percent) 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 (Huxman et al., 2005; Moran et al., 2009; Scott et al., 2006; Stannard and Weltz, 2006; Wilcox et al., 2003a). 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; Moran et al., 2009; Scott et al., 2006).

646.08 Upland Surface Runoff

A. 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 (Dingman, 2002; Dunne, 1978; Hornberger et al., 1998; Horton, 1933; Selby, 1982). Hortonian overland flow (Horton, 1933) or rainfall excess is generated when water input on land exceeds the rate at which water can infiltrate the soil. Hortonian 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” (Dingman, 2002; Hornberger et al., 1998; Selby, 1982).

B. Rainfall-excess and saturation-excess overland flows may create sheetflow (interrill) or concentrated flow (rills) or a combination of the two on sloping terrain.

(1) Sheetflow refers to overland flow as a thin, relatively spatially connected film or sheet on the land surface.

(2) Concentrated flow is runoff that accumulates or converges into well-defined microchannels or rills.

(3) Direct precipitation into stream channels occurs during all precipitation events and can contribute significantly to peak and total event streamflows. Subsurfaces return 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 (Dingman, 2002; Wilcox et al., 1997).

(4) 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 flow or event flow (Dingman, 2002; Dunne and Leopold, 1978; McCuen, 1989).

(i) 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.

(ii) 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.

(5) 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.

C. Watershed or hillslope response to precipitation events is commonly quantified graphically and analyzed by using streamflow hydrographs and precipitation hyetographs (figure A-8) (Dingman, 2002; Hornberger et al., 1998; McCuen, 1989). 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-axis and y-axis of a streamflow hydrograph to view respective stream responses to water-input events.

(1) Interception is strongly influenced by precipitation frequency and intensity and by the type and structure of the vegetative community (Branson and Miller, 1981; Dunkerley, 2008; Owens et al., 2006). Interception losses from a series of small storms and clearing events are proportionately greater than those from large, prolonged rainfall events (Hamilton and Row, 1949; Owens et al., 2006; Rowe, 1948). 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 conifers than broad-leaved tree species and is greater from trees species than from shrubs and grasses.

(2) The shape of the streamflow hydrograph provides qualitative and quantitative evidence 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 recessional limb) (figure A-8).

(3) 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.

D. The overall response to a particular event and the resultant hydrograph shape depend on the size of the drainage area, soils and geology, slope, and land use and vegetation patterns (Dingman, 2002; McCuen, 1989). Large watersheds commonly exhibit delayed responses to precipitation events, unless the event is near the stream outlet or occurs over a large, contiguous area. Shorter event and peak response times, higher peak discharges, and steepened rising limbs with respect to similar rainfall events post-disturbance, indicate degraded surface conditions.

E. 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 and hillslope data are available. Infiltration responses to water-input events may also be quantified in hydrograph form, as the inverse of the runoff relationships (Dingman, 2002; Dunne, 1978; Hillel, 1998; Hornberger et al., 1998; Selby, 1982).

646.09 Water Balance

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

$$P + G^{in} - G^{out} - E - T - Q - I = Q\Delta S$$

where P is precipitation, G^{in} is incoming ground water, G^{out} is groundwater outflow, E is evaporation, T is transpiration, Q is streamflow or surface runoff, I is interception, and ΔS is the change in water storage over the period of interest.

B. Evaporation and transpiration are often combined as a single term referred to as evapotranspiration due to difficulty in direct measurement of the term. 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., 2003a). The net change in water storage is generally considered to be zero over long time periods (such as several years). 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., 2003a). Nearly all of the remainder of precipitation falling on rangelands is lost to evapotranspiration.

Figure A-12. Camel grazing in Kazakhstan, with open connected bare interspaces which increase risk of accelerated soil erosion.

646.10 References

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Subpart B – Rangeland Soil Erosion Processes

646.21 Contributors:

C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.

Figure B-1. Remnant vegetation mound anchored by deep tree roots where streamflow has eroded the bank, leaving an island of vegetation.

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646.22 Introduction

A. Soil erosion refers to the detachment and transport of soil by raindrop impact, running water, wind, ice flows, and other mass-movement processes (Selby, 1982). Soil detachment is a function of the erosive energy acting on the soil surface and the resistance of the surface to erosion (Selby, 1982; Thornes, 1980; 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 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; Kinnell, 2005; Nearing et al., 1999).

B. The removal of detached sediment requires entrainment by a transport mechanism. Transport may occur by displacement from raindrop impact (splash) or entrainment into flowing water, or both (Kinnell, 1988, 1990, and 2005). The 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 (water velocity), and efficiency of transport (Kinnell, 2005; Thornes, 1980; Toy et al., 2002). In this section we summarize these fundamental sediment detachment and transport mechanisms. Emphasis in the handbook is on hillslope processes. Julien (2010), Knighton (2014), and McCuen (1989) provide detailed explanations of channel or fluvial sediment entrainment, transport, and routing processes beyond the scope of this handbook. Soil erosion through wind, creep, weathering, and other mechanisms not directly involving rainfall and flowing water are discussed in Field et al. (2012); Li et al. (2013); Okin et al. (2009); Ravi et al. (2007); Sankey et al. (2009); Selby (1982); and Toy et al. (2002).

646.23 Raindrop Splash and Sheetflow Processes

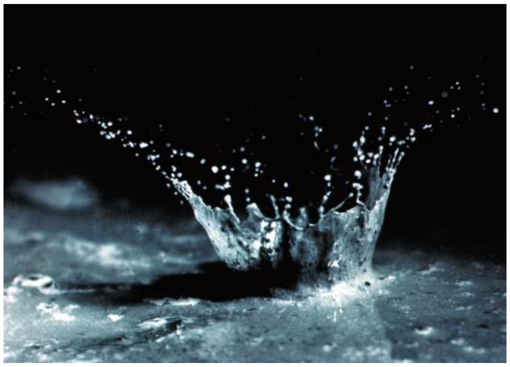
A. The combined effects of raindrop splash and sheetflow/sheetwash processes are termed “rainwash” or “interrill” processes (Selby, 1982; Toy et al., 2002). Interrill processes often represent the dominant erosion processes where microchannel formation is substantially limited by soil aggregation, soil cohesion, and surface protection. The two components of interrill (raindrop splash) and sheetflow are differentiated because their co-occurrence is dependent on overland flow generation. Interrill erosion is the transfer of sediment resulting from raindrop impact. The effects of raindrop impact include detachment and displacement of soil (Kinnell, 2005), disaggregation of soil aggregates, and reduced infiltration due to surface sealing (Assouline, 2004; Kinnell, 2005; Moss, 1991). Sediment detachment and transport by interrill erosion occur in a “splash- crown” (figure B-2A), with sediment mass declining exponentially outward from the point of impact on flat surfaces (Kinnell, 2005; Thornes, 1980; Toy et al., 2002). On sloping terrain, downslope transport may exceed three times the mass of upslope transport where slopes are greater than 10 percent (Thornes, 1980).

B. 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 maybe highest at rainfall onset if the supply of detachable sediment is low (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 (Bryan, 2000; Gilley and Finkner, 1985; Salles and Poesen, 2000; Salles et al., 2002; Sharma et al., 1991). 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). The impact energy of rainfall and sediment detachment generally increases with increasing rainfall intensity. Cumulative sediment detachment may be significant from low-intensity storms over long durations.

C. A thin layer of water on the soil surface can often increase the effect of the raindrop impact on soil detachment. The decrease in detachment by raindrops during rain events is often associated with an increase in the depth of sheetflow (Ferreira 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 (figure B-2C) (Kinnell, 1990, 1991, and 1993; Moss and Green, 1983). 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 (Assouline, 2004; Bradford et al., 1987), may create a surface crust or compact the surface soil, resulting in reduced infiltration (by a factor of one 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.

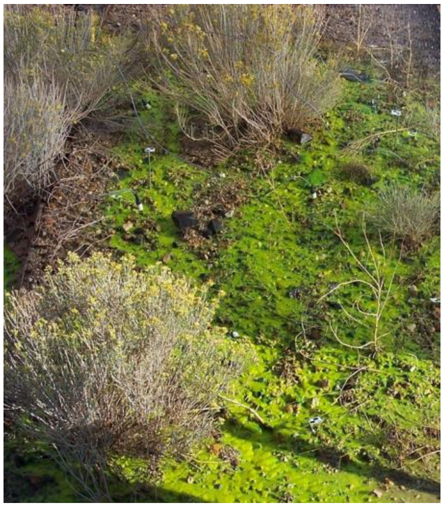
D. Sheetflow generates shear stress that is minor compared to raindrop impact (about 100-fold less), but it serves as an additive detachment and transport mechanism for interrill sediment yield (Kinnell, 2005; Selby, 1982; Thornes, 1980; Toy et al., 2002). Sheetflow does not usually occur as a broad flow across a hillslope; rather, it occurs in isolated irregular-flow patches of several millimeters in depth separated by flow obstacles (figure B-2B) (Emmett, 1978; Emmett, 1970). 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 (Kinnell, 2005; Thornes, 1980). 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. Entrained particles remain in suspension until a deposition velocity occurs (< 0.015 m s^{-1}).

Figure B-2. Soil erosion by (A) raindrop splash, (B) sheetflow, and (C) concentrated flow processes (photo A: USDA Natural Resources Conservation Service; photos B and C: USDA Agricultural Research Service, Great Basin Rangelands Research Center).



(A) Raindrop splash erosion.

(B) Concentrated flow erosion. Flow channels are color-enhanced for research using fluorescein dye.



(C) Sheet flow erosion. The pattern of flow is enhanced for research with application of fluorescein dye.

E. 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 raindrop splash and sheetflow processes (Blackburn et al., 1990; Blackburn and Wood, 1990). Ice formation in the soil can result in frost heaving (figure B-3). Frost heaving is the process where the soil is lifted upwards by the formation of the ice. This results in a pedestalled soil surface with reduced soil erodibility, bulk density, and aggregate stability of the soil surface (figure B-3). After the first several rainfall events in the spring, the pedestals are broken down and the soil is reconsolidated through wetting and drying processes. Soil erodibility is returned to a baseline condition that is more resistant to raindrop slash and concentrated flow processes.

Figure B-3. Frost heaving lifts the soil, reduces the bulk density and aggregate stability. This increases susceptibility to both raindrop splash and concentrated flow soil erosion processes.

646.24 Concentrated Flow Processes

A. Concentrated flow paths or rills are microchannels of several to tens of centimeters in width and several to 300 millimeters in depth that may be removed between storm events by wetting and drying cycles or growth of new plants. These microchannels form when surface roughness elements (microtopography) concentrate sheetflow into narrow, deeper flow paths, increasing the velocity and erosive energy of runoff (figure B-4) (Emmett, 1978; Emmett, 1970). Sediment yield from concentrated flow (rill) processes is several orders of magnitude greater than that of sheetflow and raindrop splash (Pierson et al., 2008b; Thornes, 1980; Wainwright et al., 2000). Concentrated flow processes may account for 50 to 90 percent of total sediment yield on slopes with little vegetation. 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 the soil surface critical shear stress or erodibility (Al-Hamdan et al., 2012b; Nearing et al., 1989 and 1999). 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 (Al-Hamdan et al., 2015, 2012a, 2012b, and 2013; Toy et al., 2002).

B. Deposition occurs when the water flow sediment 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 flow path incision or where sediment supply is limited (Al-Hamdan et al., 2012b; Nearing et al., 1997; Pierson et al., 2008b; Wainwright et al., 2000). On freshly exposed surfaces, parallel concentrated flow paths may cross grade or merge by breaking down the divides between microchannels (micropiracy). This process diverts the flow into the deeper, more dominant flow paths and generally increases the spacing of concentrated flow paths across the hillslope.

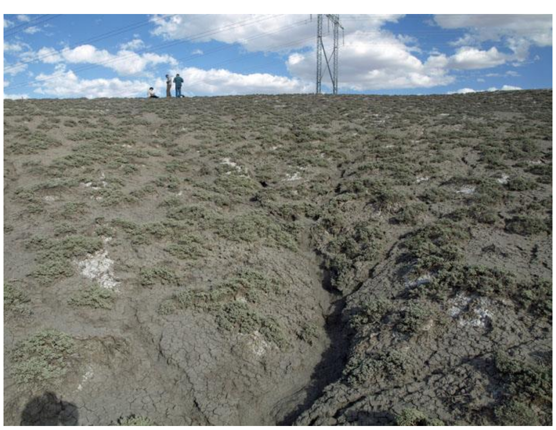
Figure B-4. Concentrated flow paths (rills) in a burned Pinyon site following wildfire (left) and gully formation and deposition where the slope gradient changes in the foreground (right).


646.25 Gully Erosion Processes

A. Gullies are recently channelized drainage features that transmit ephemeral flow, usually have steep sides and a head scarp (leading upslope to area of exposed soil and rock) and are more than 30 cm wide and 60 cm deep (Selby, 1982; 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 (figure B-5) (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 on the hillslope.

B. 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, 1982). Gully erosion can be controlled with various control structures using wooden or rock structures; however, these structures require frequent maintenance or the erosion within the gully can be exacerbated (Pederson et al., 2006).

Figure B-5. Gully erosion on an unprotected soil in central Utah.



646.26 Mass Movement Processes

A. 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 (figure B-6) (Selby, 1982; Sidle et al., 1985). 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 when shifting of soils alters the soil structure, bedding planes of soil horizons, and disrupts root anchoring points within the soil. Additionally, shear strength is reduced through increased buoyancy and capillary tension associated with pore water changes, 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. Sidle et al. (1985) and Selby (1982) provide extensive descriptions and explanations of the types, causes, and occurrences of soil mass movements and present approaches to slope stability assessment and analysis.

B. 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 *enmasse* (slope creep). Particle creep occurs due to gravity, particle expansion and contraction with heating and cooling, wetting and 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 year^{-1} on well-vegetated slopes, but may be as high as 500 mm year^{-1} on exposed slopes and areas with frequent freeze-thaw cycles (Selby, 1982). Falls result from the undercutting of slope faces or toe slopes by flowing water or from cliff-top sloughing after freeze-thaw or wetting and drying periods. Falls may contribute significantly to the downslope transport of rocks and sediment contributions until the system comes back into equilibrium with the transport capacity of this recently deposited unconsolidated material.

Figure B-6. Debris flow following a thunderstorm event occurring on burned forest land (photo: USDA Forest Service).

C. Slides occur on failure planes that are either straight (translational) or curved (slumps) and are the most common form of landslide (Selby, 1982; Sidle et al., 1985). 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.

D. Overburdening, slope steepening, and ground vibrations also cause 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 over-consolidated 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.

E. Debris (soil) flows are gravity-induced mass movements that are intermediate between sliding and water flows. They occur as debris, earth, or mud flows resulting from wet or dry liquefaction of coarse debris, fine-grained soil, or clay soil, respectively (Selby, 1982). 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. Debris 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.

(1) Debris flows (figure B-6) 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." The discussion that follows is similar for debris torrents and flows and refers 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. Flow paths can extend for many kilometers and commonly cease in low-gradient alluvial fans with boulder levees (Selby, 1982).

(2) 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 (five 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 (Ellen and Wieczorek, 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 (1982) provides a brief review of this and other catastrophic debris flow events of similar magnitude. Cannon et al. (1998, 2001a, 2008, and 2001b) 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.

F. 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 a greater threat to life, property, and resources than earthflows. 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 (Scott, 1988).

646.27 Water Repellant Soils: Hydrophobic Solis

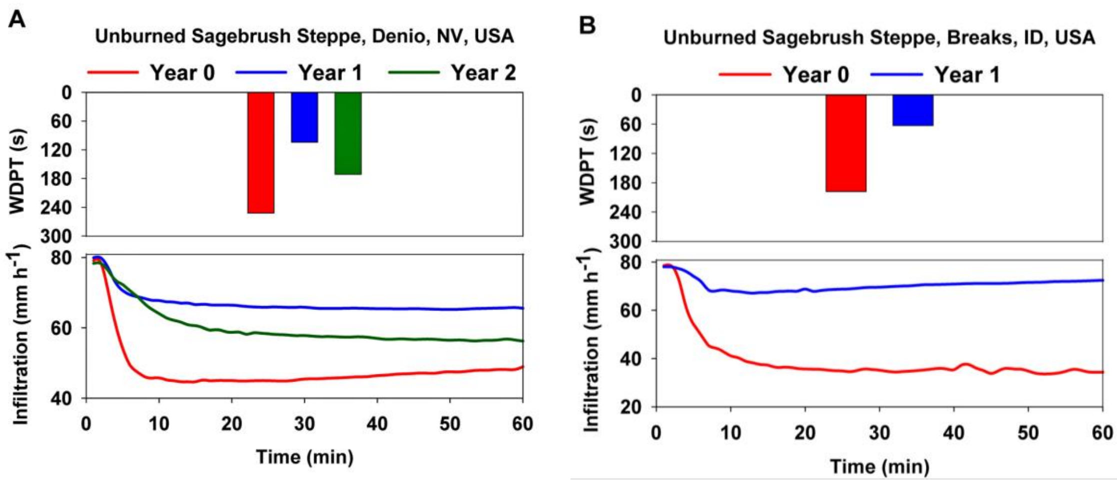
A. Soil water repellency is a naturally occurring soil condition that inhibits infiltration. Its occurrence has been well documented on shrubland, chaparral, woodland, and semi-arid forest ecosystems (Bodí et al., 2013; DeBano, 2000a; Hubbert et al., 2006; Pierson et al., 2014; Robinson et al., 2010; Williams et al., 2014). These compounds are long carbon chain molecules with an affinity on one end to attach to a soil particle, and a water repellent chemical structure on the other end that repels water molecules. They extend outward from the soil particles as a protective coating one molecule thick. 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 (Bisdom et al., 1993; Doerr et al., 2000; Imeson et al., 1992; Savage et al., 1972). 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; Doerr and Thomas, 2000; Pierson et al., 2008a, 2009, and 2010; Verheijen and Cammeraat, 2007). The type and quantity of vegetation dictate the amount and type of hydrophobic compounds potentially available. Clay soils have more particle surface area per unit of soil than sand. Coarse-textured soils generally are more susceptible to soil water repellency than fine-textured soils due to their greater particle surface area (Bisdom et al., 1993; DeBano, 1971; Huffman et al., 2001). Recent research has demonstrated that strong soil water repellency can occur in fine-textured soils. Doerr et al. (2006), Doerr et al. (2009), and Doerr and Thomas (2000) provide a review of occurrence of soil water repellency, its causes, hydrologic and erosional effects, and measurement methods. Water repellency is frequently a result of fires burning the vegetation and depositing the organic repellency from the burning residue onto the soil particles (DeBano, 2000 a, b).

B. The strength and persistence of soil water repellency is highly variable in time and space (Dekker et al., 2001; MacDonald and Huffman, 2004; Madsen et al., 2008; Witter et al., 1991; Woods et al., 2007). Soil water repellency for a particular soil may be present under dry conditions, decrease with soil wetting, and reappear with soil drying (Doerr et al., 2000; Shakesby et al., 1993). 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 five 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. (2008a and 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 (figure B-7).

C. 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; Verheijen and Cammeraat, 2007; Woods et al., 2007). 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., 2008a; Pierson et al., 2009; Pierson et al., 2010).

D. Soil water repellency facilitates runoff initiation either by inhibiting infiltration at the surface (infiltration-excess runoff) or by 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 soils become wet (Burch et al., 1989; DeBano, 2000b; Pierson et al., 2008a and 2009; Robichaud, 2000). Increasing infiltration rates over time (minutes to hours) occurs due to a gradual decrease in repellency with wetting, or due to lateral and vertical water transfer through preferential infiltration and flow via macropores or breaks in the repellent layer (Dekker and Ritsema, 1994, 1995, and 1996; Doerr et al., 2000; Jaramillo et al., 2000; Leighton-Boyce et al., 2007; Ritsema and Dekker, 1994, 1995, and 1996; Ritsema et al., 1998a, 1997, and 1998b). Vegetation and ground cover store intercepted rainfall, as well as snow, and retain overland flow, allowing more time for infiltration and soil wetting. 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 (Jaramillo et al., 2000; Leighton-Boyce et al., 2007; Ritsema and Dekker, 1994; Ritsema and Dekker, 1995; 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 (Burch et al., 1989; Imeson et al., 1992; Meeuwig, 1971). The overall effect of preferential flow depends on the extensiveness of the macropore network and the strength of soil water repellency.

Figure B-7. Temporal variability of soil water repellency effects (measured by using water drop penetration time, WDPT) 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., 2008b); 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). (Caption and figure from Pierson and Williams, 2016.)



E. Vertical preferential flow along wet spots has been referred to as fingered flow (Ritsema et al., 1997). 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 (Dekker and Ritsema, 1996). In multiyear rainfall simulation studies of two steeply sloping mountain big sagebrush sites in Nevada and Idaho, Pierson et al. (2008a, 2009, 2008b, and 2001) found that minimum- and steady-state infiltration rates on unburned shrub coppices (0.5 m^2 rainfall simulation plots) 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. (2008a and 2001) 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. (2008a, 2009, 2008b, and 2001) 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 flow paths most likely subdues the effects at hillslope and catchment scales (Doerr and Moody, 2004; Pierson et al., 2009; Shakesby et al., 2000).

F. 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 (Doerr and Moody, 2004; Imeson et al., 1992; 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 rainfall simulations (0.36 m^2) 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 two needle Pinyon litter yielded hydraulic conductivities (as measured by an infiltrometer) six to more than 30 times greater than under water-repellent conditions. They observed (without taking specific measurements) that tree coppices retained surface water and routed it laterally toward preferential wet spots under the tree canopy.

G. Hydraulic conductivity and infiltration rates can be as much as 25- to 30-fold lower for water repellent versus wettable soils. The litter layer in vegetated areas buffers repellency effects on infiltration by trapping water input and allowing it to slowly infiltrate via macropores and breaks in the water repellent layer or slow wetting of the soil profile (Pierson et al., 2008b; Williams et al., 2014). Collectively, interception and enhanced infiltration in vegetated areas commonly result in two- to more than 20-fold less event runoff relative to bare or sparsely vegetated areas across the point to patch scales (Pierson and Williams, 2016).

H. Soil water repellency may provide water conservation and increased plant productivity for some woody species and may indirectly mitigate runoff generation (Jaramillo et al., 2000; 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 two needle 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 flow paths and deep soil recharge (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).

I. 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 (Huxman et al., 2005; Ludwig and Tongway, 1997; Ludwig et al., 2005; Wilcox et al., 2003). Surface flow routing by soil water repellency may function similarly to the lateral surface transfers of overland flow (run-on) in maintaining shrub, grass, and tree islands of higher biological activity and water retention (Breshears et al., 1997; Ludwig and

Tongway, 1995; Ludwig et al., 2005; Puigdefabregas et al., 1999; Reid et al., 1999; Robinson et al., 2010; Schlesinger et al., 1990; Wilcox and Breshears, 1994; Wilcox et al., 2003).

Figure B-8. Flume 1 at the USDA Agricultural Research Service Walnut Gulch Experimental Watershed, Tombstone, Arizona measuring suspended sediment during a summer monsoon convective thundershower.

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Figure B-9. Tollgate weir at USDA Agricultural Research Services Reynolds Creek Watershed near Boise, Idaho measuring runoff and suspended sediment. Tollgate weir defines an area of about a quarter (55 km²) of Reynolds Creek Experimental Watershed, but that area is almost entirely comprised of the upper elevations that receive the greatest annual precipitation which predominately falls as snow. Elevations range from 1410 m to 2241 m. Vegetation is primarily sagebrush rangelands with small stands of Douglas fir (*Pseudotsuga menziesii*), aspen (*Populus spp*.), and Alpine Fir (*Abies lasiocarpa*) at the higher elevations.

Subpart C – Rangeland Vegetation Effects on Soil Erosion Processes

646.31 Contributors:

C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ, Tucson, AZ.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX.

Figure C-1. Embedded rocks and perennial vegetation alter the direction of concentrated flow paths of water on rangelands, resulting in braided channels (left) and detention ponds (right) above the obstruction resulting in reduced runoff and sediment yield.

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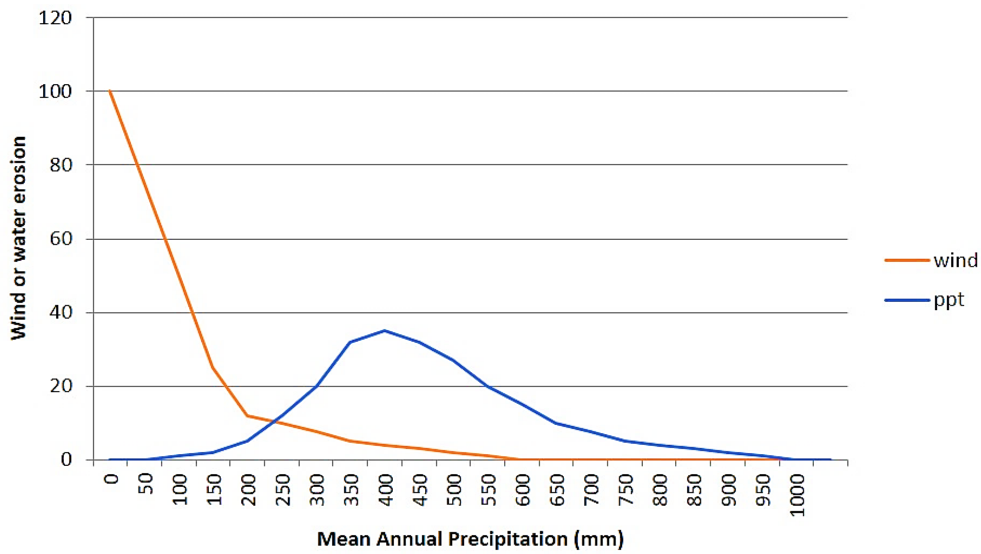
646.32 Vegetation Influence on Soil Erosion Processes

A. The most vulnerable rangeland areas for soil movement are where annual precipitation is between 100 and 400 mm yr^{-1} (figure C-2), which limits soil moisture available to sustain plant growth. With low plant density and minimal plant and ground cover, arid and semi-arid areas are prone to both wind and water erosion. Arid and semi-arid regions have low plant density, which often results in open and connected bare interspaces where aerodynamic roughness is low and fetch length is sufficient to allow for wind and water erosion (Okin et al., 2009). In addition, there is insufficient vegetative canopy and ground cover to prevent soil movement from raindrop splash, sheetflow, and rill erosion in the bare connected interspaces (Puigdefabregas, 2005). The relatively low vegetation cover, combined with high intensity convective rainfall events, makes the Upper Colorado River Basin one of the most erosive areas of the United States. Average sediment yield frequently exceeds 3 t ha^{-1} yr^{-1} on the Colorado Plateau (Langbein and Schumm, 1958). As water erosion is exponentially related to rainfall intensity, most of the soil erosion occurs during rare convective storm events that have high rainfall intensities. Consequently, rilling and arroyo formation are very pronounced in the Colorado Plateau where convective storms annually occur (West, 1983). Interaction between wind erosion and deposition–and water erosion, transport, and deposition–is poorly understood, but linkages do exist, and total erosion may be maximized in arid and semi-arid regions because of limited cover and the steep, highly dissected slopes of poorly weathered marine shales that are highly erosive in the Upper Colorado River Basin.

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., 2015). Raindrop splash and sheetflow processes dominate at the small-plot scale (one to two m^2), and erosion rates highly depend on the amount of vegetation and litter that controls 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 (figure C-1) and the connectivity of these processes (Williams et al., 2015).

Figure C-2. Conceptual diagram of wind and water erosion with mean annual precipitation for naturally vegetated arid and semi-arid range lands (Modified from Marshall 1973).



B. 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 (Abrahams et al., 1995; Puigdefabregas, 2005; Puigdefabregas et al., 1998; Reid et al., 1999; Wilcox et al., 2003b; Williams et al., 2014b). For example, small perturbations on a hillslope may create small patches of exposed bare soil highly susceptible to raindrop splash 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.

C. The same landscape with uniform surface roughness may undergo substantially more soil erosion from a similar storm due to an increase in the spatial connectivity of well-organized sheetflow or concentrated flow paths (Williams et al., 2015). 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 and driving forces (Pierson et al., 1994a and 1994b; Wilcox et al., 2003b). 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.

D. Vegetation regulates sediment availability for concentrated flow erosion through protection of the soil surface from raindrop impact and the erosive energy of overland flow. Surface protection and soil stabilization by cover elements are paramount in minimizing erosion, given that raindrop impact is the primary sediment contributor to shallow overland flow and ultimately a source for concentrated flow (Kinnell, 2005; Wainwright et al., 2000; Williams et al., 2015). Vegetation and ground cover can reduce rainfall erosivity by nearly 50 percent (Wainwright et al., 1999). In addition to reducing raindrop impact, vegetation and ground cover are roughness elements that trap and slow runoff and promote sediment deposition (Emmett, 1970; Pierson et al., 2007a; Pierson et al., 2009; Wainwright et al., 2000). Plants and associated organic material also contribute to the soil shear strength by anchoring soils and promoting aggregate stability (Al-Hamdan et al., 2014; Al-Hamdan et al., 2013;

Blackburn, 1975; Cammeraat and Imeson, 1998; Cerdà, 1998; Pierson et al., 2013; Pierson et al., 2014; Pierson et al., 2010).

E. Parsons et al. (1992 and 1994) evaluated the effect of cover elements on raindrop splash erosion during high intensity rainfall simulations. They found that raindrop splash erosion rates on arid, well-vegetated grassland were 0.01–0.04 g m^{-2} min^{-1} for 73–86 mm h^{-1} rainfall intensities. The same studies measured a 0.34 g m^{-2} min^{-1} erosion rate on a degraded arid shrubland for a simulated event with 145 mm h^{-1} intensity (Wainwright et al., 2000). Raindrop splash during the shrubland experiments eroded about 1.6-fold more sediment from areas between plant canopies than from areas underneath plant canopies (Parsons et al., 1992). Results from numerous other studies indicate that erosion rates from raindrop splash and sheetflow at the point scale can be two-fold to more than three orders of magnitude greater for bare areas than areas underneath vegetation or with litter cover (Pierson and Williams, 2016). Actual differences vary with cover, soil, rainfall, and topographic characteristics. Erosion from combined raindrop splash, sheetflow, and concentrated flow processes is typically negligible where ground cover exceeds 50 percent (Gifford, 1985; Pierson et al., 2009, 2013, and 2011; Weltz et al., 1998).

F. 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 run-on underneath and immediately adjacent to the canopy area. Litter amassment and decomposition underneath shrub and tree canopies (figure C-3) and differential raindrop splash contribute to soil, organic matter, and nutrient accumulation (Belnap et al., 2005; Blackburn et al., 1992; Ludwig and Tongway, 1995; Ludwig et al., 2005; Schlesinger et al., 1999). 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 (Beven and Germann, 1982; Blackburn and Skau, 1974; Cerdà, 1998; Parsons et al., 1996; Pierson et al., 2013; Pierson et al., 2014; Pierson et al., 2010; Thurow et al., 1986; Williams et al., 2014a; Wood and Blackburn, 1981).

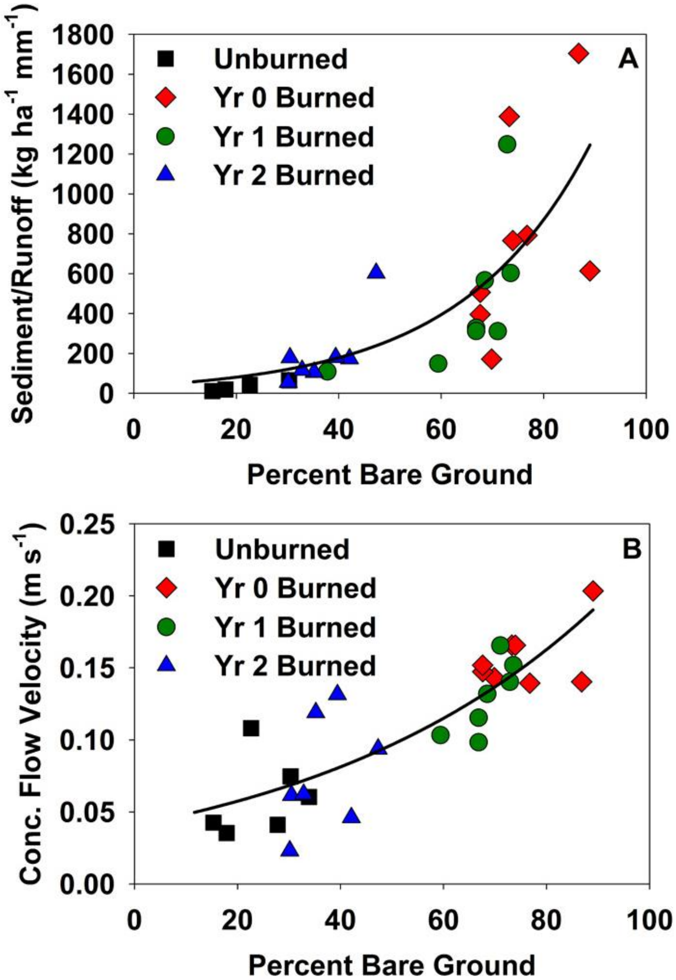
G. Litter and slash from fallen woody plants can result in formation of debris dams (figure C-3). Debris dams form when concentrated flow pushes loose unconsolidated organic material down slope. When the material contacts restrictions like rocks or the basal area of a plant, the material can bridge the gap, forming a dam. As more litter is pushed down slope, the dam can build to several centimeters high. This results in a temporary detention pond being formed, downslope water velocity is retarded, and sediment is deposited behind the dam. These dams are not stable and can be breached if the velocity or volume of the incoming water increases (increase in rainfall intensity). This breach can then result in a sudden release of water and the formation of an incised rill with accelerated soil erosion. 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 (Beven and Germann, 1982; Blackburn and Skau, 1974; Cerdà, 1998; Parsons et al., 1996; Pierson et al., 2013; Pierson et al., 2014; Pierson et al., 2010; Thurow et al., 1986; Williams et al., 2014a; Wood and Blackburn, 1981).

Figure C-3. Debris dam formed from loose Pinyon and Juniper slash, litter, and duff.

H. 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 infiltration (Belnap et al., 2005; Cammeraat and Imeson, 1998; Dunkerley, 2002; Ludwig et al., 2005; Puigdefabregas et al., 1999). Stemflow concentrates water input at plant bases, allowing rapid vertical recharge of the soil profile via preferential flow along root channels (Bhark and Small, 2003; Devitt and Smith, 2002; Lebron et al., 2007; Martinez-Meza and Whitford, 1996; Navar and Bryan, 1990; Thurow et al., 1987). Plant growth form also influences infiltration processes. Infiltration rates are generally higher for bunchgrasses than sod-forming grasses (Blackburn et al., 1992; Knight et al., 1984; Pierson et al., 2002; Thurow et al., 1986; Thurow et al., 1988; Wood and Blackburn, 1981). 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; Thurow et al., 1988). Additionally, biomass and organic matter accumulations under bunchgrasses most likely favor infiltration-increasing microbial activity (Blackburn et al., 1992). Infiltration under shrub canopies is usually greater than under grass canopies (Schlesinger et al., 1999; Wood and Blackburn, 1981), 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 (Belnap et al., 2005; D'Odorico et al., 2007; Schlesinger et al., 1990).

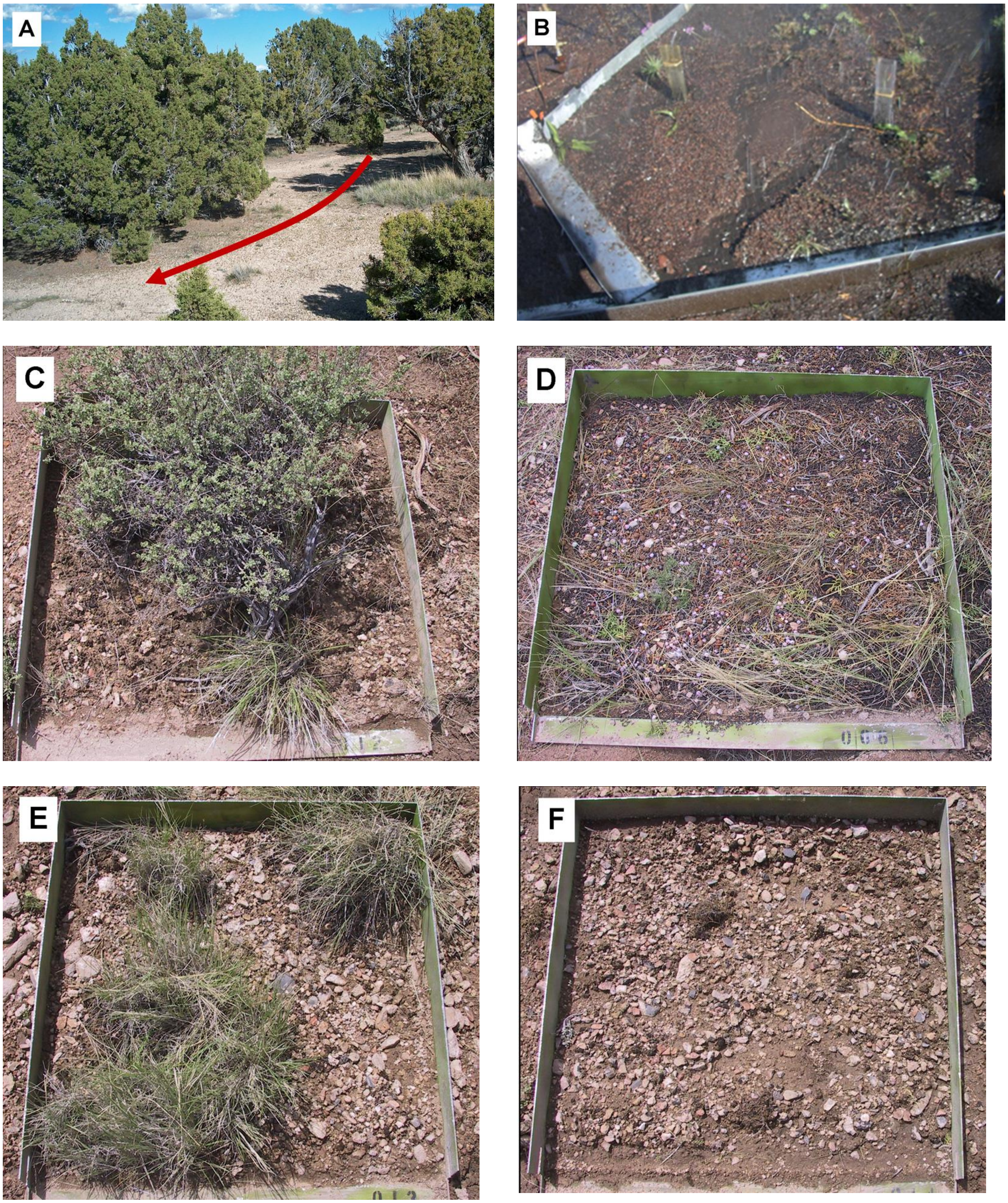
I. In general, surface characteristics of interspace areas are consistently different from coppices throughout the year, but the magnitude of the differences and respective influences on infiltration exhibit some seasonality. The spatial differences in vegetation cover and surface characteristics exert 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). See figure C-4.

Figure C-4. Sediment per unit of runoff (A) and concentrated flow velocity (B) versus percent bare ground measured on rainfall simulation plots (32.5 m^2, 85 mm h^{-1} rainfall intensity) and concentrated flow experiments (12 L min^{-1} flow release) on unburned and burned shrublands. Data from Pierson et al. (2009). (Caption and figure from Williams et al., 2014; Nouwakpo et al., 2016.)



J. 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 and runoff initiation. Interspaces occur with various amounts of herbaceous cover or exist as contiguous bare patches (figure C-5) (Abrahams et al., 1995; Blackburn et al., 1992; Parsons et al., 1996; Pierson et al., 1994a; Pierson et al., 1994b; Wilcox and Breshears, 1994). Well-vegetated interspaces may exhibit similar surface characteristics as canopy areas to some degree, but usually generate more surface runoff (Bhark and Small, 2003; Reid et al., 1999; Wilcox et al., 2003b). On more water-limited or degraded sites, interspaces have low plant biomass and organic matter (figure C-5F) and thin surface soil horizons (Abrahams and Parsons, 1991; Blackburn and Skau, 1974; Parsons et al., 1996; Pierson et al., 2010; Wilcox et al., 1996).

Figure C-5. Photographs from a Utah Juniper woodland site in Utah, showing (A) general direction for overland flow over contiguous bare interspace, (B) interspace concentrated flow path during artificial rainfall simulation, (C) hydrologically stable shrub, (D, E, F) and vegetated and bare interspaces (photos: USDA Agricultural Research Service, Northwest Watershed Research Center). (Caption and figure from Pierson and Williams, 2016.)



These characteristics result in poor aggregate stability, soil structure, and high bulk densities relative to coppices. They also facilitate low infiltration rates (Blackburn et al., 1992; Blackburn and Skau, 1974; Johnson and Gordon, 1988; Roundy et al., 1978; Wilcox and Wood, 1988).

646.33 Rock Cover

A. Rock cover on the soil surface has a complex relationship with infiltration and soil erosion processes (figure C-6). The effects of rock cover (> 2 mm) depend on the size, amount, and embeddedness of the rocks (Poesen and Ingelmo-Sanchez, 1992; Poesen et al., 1990; Wilcox and Wood, 1988). 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 (Cerdà, 2001; Poesen et al., 1994; Valentin, 1994).

Figure C-6. Accelerated upland rill erosion showing exposed rock fragments in Pinyon-Juniper interspace following intense thunderstorm.

B. Infiltration is negatively correlated with embedded rock cover due to a decrease in nonabsorbing area. Numerous authors reported negative correlations between rock cover and infiltration in interspace areas but did not explicitly evaluate embeddedness (Abrahams and Parsons, 1991 and 1994; Pierson et al., 2010 and 2013; Wilcox and Wood, 1988). The studies by Wilcox et al. (1988) and Abrahams and Parsons (1991) indicate that interspace areas occurred in swales and were more compacted and crusted than coppice areas. These authors 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 that 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 that rock cover can facilitate infiltration, that negative effects of rock cover on infiltration most likely occur when smaller rocks dominate, and that the rock cover is embedded rather than freely lying atop the soil surface (Brakensiek and Rawls, 1994). Rock fragments may provide protection from raindrop impact but do not substantially reduce hydraulic shear stress or rilling in semi-arid shrub-dominated landscapes. For large rainfall events, the depth and shear stress of flow in the rills exceed the resistance offered by the rock fragments, and substantial rilling does occur between the shrub-dominated coppice dunes in the desert southwest (Tiscareno-Lopez et al., 1993).

646.34 Biological Soil Crust

A. Infiltration in interspace locations is strongly influenced by the expanse of bare ground, rock cover, or vesicular crusts (Blackburn and Skau, 1974; Johnson and Gordon, 1988; Parsons et al., 1996; Pierson et al., 1994a and 2010; Reid et al., 1999; Wood et al., 1978). Exposure of bare ground to raindrop impact increases the potential for surface sealing or development of infiltration-inhibiting surface crusts (Branson et al., 1981; Puigdefabregas et al., 1999). Decreasing infiltration and increasing runoff with increasing expanse of bare or vesicular surfaces are well documented in literature (Blackburn et al., 1992; Pierson et al., 2007a, 1994a, 2002, 2014, and 2010; Schlesinger et al., 1999; Schlesinger and Andrews, 2000; Williams et al., 2014a and 2014b). Of particular significance in rangelands are biological soil crusts (cryptogams), which have soil erosion resistance-conferring properties and have extreme susceptibility to disturbance (figure C-7). Biological soil crusts are a term used to define a collection of nonvascular plants: mosses, algae, lichens, liverworts, and cyanobacteria.

Figure C-7. Biological soil crust.

B. The impact of biological soil crusts on infiltration rates and soil erosion is poorly understood and often contradictory. Biological soil crusts can reduce infiltration rates and increase soil erosion by blocking flow through macropores, or they may enhance porosity and infiltration rates by increasing water-stable aggregates and surface roughness (Eldridge, 1993; Loope and Gifford, 1972; West, 1991). Disturbance of the soil surface can disrupt biological soil crusts and result in enhanced wind erosion and may or may not affect water erosion processes (Barger et al., 2006; Belnap and Gillette, 1998; Belnap et al., 2009; Eldridge and Koen, 1998; Li et al., 2008). Li et al. (2008) evaluated the interactions between biological soil crusts and runoff on a hillslope with patchy shrub vegetation. They reported that in undisturbed areas 53 percent of the simulated rainfall became runoff from the crust patches, and 55 percent of this was redistributed and absorbed by the shrub patches. In addition, approximately 75 percent of the sediments, 63 percent of soil carbon, 74 percent of nitrogen, and 45 percent to 73 percent of the dissolved nutrients transported in runoff from the crust patches were delivered to shrub patches. The disturbance of crust patches tended to result in the uniform distribution of water over the whole slope, with a corresponding reduction in the transport of runoff and nutrients from the crust patches to the shrub patches.

C. The exact response on runoff and soil erosion is a function of site disturbance and level of development of the biological soil crusts (Belnap et al., 2013). When studies are evaluated based on biological crust type, and considering naturally occurring differences between crust types, results indicate that biological crusts in hyper-arid regions reduce infiltration and increase runoff. Biological soil crusts have mixed effects in arid regions and increase infiltration and reduce runoff in semi-arid cool regions. Most research has shown that intact biological soil crusts are effective in reducing soil erosion and transport of soils and associated contaminants (Belnap, 2006). Additional research is required before the role that biological soil crusts play in altering transport of salts (dissolution of salt crusts by efflorescence) is fully understood. Also, while mechanisms of concentrated flow detachment are well understood, prediction of sediment delivery is often complicated by the less studied sediment deposition processes.

D. Numerous attempts have been made to establish cover guidelines required for site protection from soil erosion. Various cover types offer varying degrees of soil protection (i.e., rock, biological soil crusts, litter, and vegetation). The amount and effectiveness of cover necessary for site protection depends on other factors such as slope, soil type, time of year, and rainfall intensity and duration. Wilcox (1994) found that most erosion was produced by large convective summer thunderstorms within the bare interspace areas of Pinyon-Juniper woodland, and erosion was slight during the winter due to the absence of raindrop detachment, even with high runoff rates from snow melt. Generally, the greater the bare soil area, the greater the erosion rate. Reported levels of cover necessary for site protection range from 20 percent in Kenya (Moore et al., 1979) to 100 percent for some Australian conditions (Costin et al., 1959). Most studies indicate that cover of 50 to 75 percent is probably sufficient to provide erosion control (Gifford, 1984; Orr, 1970; Packer, 1951).

E. The percentage of event rainfall captured by vegetation and associated ground cover generally decreases as rainfall intensity increases (Carlyle-Moses, 2004; Owens et al., 2006). For low-intensity, short-duration rainfall events, most of the precipitation is captured by plant canopies, litter, and other ground cover and is lost to evaporation (Dunkerley, 2008; Owens et al., 2006). Water input during high-intensity or prolonged rainfall events usually exceeds interception storage capacity, resulting in delivery of water to the ground surface via throughflow and stemflow (Carlyle-Moses, 2004; Dunkerley, 2008; Martinez-Meza and Whitford, 1996). Interception by individual shrubs and conifers commonly averages 50–60 percent of water input for low-intensity rainfall events and 5–35 percent for high-intensity or prolonged rainfall events (Hamilton and Row, 1949; Owens et al., 2006; Skau, 1964; Taucer et al., 2008). Water arriving at the ground surface during an event either ponds at the soil surface, is stored in the litter layer, infiltrates into the soil, or is transferred downslope as runoff. Organic matter contributions and soil fauna activity are typically greater in vegetated and litter covered areas relative to bare areas and facilitate macropore development and soil properties associated with enhanced infiltration (Belnap et al., 2005; Blackburn, 1975; Cammeraat and Imeson, 1998; Dunkerley, 2002; Imeson et al., 1998; Ludwig et al., 2005; Puigdefabregas et al., 1999). Litter layers underneath vegetation also trap water input behind debris dams and thereby delay runoff generation. Prolonged storage at the ground surface allows water to slowly infiltrate, even in the presence of water repellent soils (Leighton-Boyce et al., 2007; Pierson et al., 2010 and 2013).

646.35 Spatial Distribution of Vegetation’s influence: Soil Erosion Process

A. The vegetation- and soils-driven hydrologic heterogeneity (VDSH) (Puigdefabregas, 2005) of rangeland ecosystems creates a mosaic of runoff source and sink areas at the hillslope scale (figure C-8) (Ludwig et al., 2005; Puigdefabregas et al., 1999; Schlesinger et al., 1996; Turnbull et al., 2012; Wilcox and Breshears, 1994; Williams et al., 2014a). The timing and quantity of overland flow generation on rangeland sites is strongly correlated to the density and distribution of canopy, ground cover, and bare interspace (Branson et al., 1981). Runoff occurs more rapidly after the onset of rainfall in bare 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; Puigdefabregas, 2005). The hydrologic connectivity and downslope surface hydraulic conductivity dictate the progression or decay of surface runoff with increased slope length (Abrahams et al., 1995; Bergkamp, 1998; Cerdà, 1997; Puigdefabregas, 2005).

Well-connected flow paths develop in consecutive source areas where overland flow is routed around topographically elevated coppice mounds, grass clumps, or roughness elements (figure C-8, C.9A, and C.9B) (Emmett, 1970 and 1978; Parsons et al., 1996; Schlesinger et al., 1996 and 1999; Thornes, 1994; Wilcox and Breshears, 1994). Where concentrated, these flow paths transfer large volumes of water laterally at greater overland flow depths and velocities than occur in sheetflow processes. These effects are amplified on steep slopes (Al-Hamdan et al., 2013), where infiltration and slope steepness have been shown to have a positive effect on runoff from rangelands (Wilcox et al., 1988). The interception of flow paths (mostly due to ponding behind coppices or topographic features) and subsequent re-infiltration (run-on) in coppice or vegetated hydrologic sinks are thought to stimulate biological productivity and further facilitate coppice-interspace structure (Bhark and Small, 2003; Joffre and Rambal, 1993; Ludwig et al., 2005; Schlesinger et al., 1990; Wilcox and Breshears, 1994). Fine-scale vegetative heterogeneity of a hydrologically stable grassland facilitate run-on processes. Coarseness or increased spatial distance between shrubs in degraded rangelands amplifies runoff with increasing slope length (Abrahams et al., 1995; Parsons et al., 1996). On disturbed rangelands, increased connectedness of bare soil patches allows the formation of concentrated flow paths, which initiates gully formation and increases runoff and soil loss (Pierson et al., 2009; Urgeghe et al., 2010; Wilcox et al., 2003a, 2003b, and 2003c). Once the ecosystem's flow paths have been altered, subsequent storms reinforce gully formation, further accelerating soil loss and decreasing water infiltration rates.

B. In arid and semiarid rangelands, where vegetation is typically sparse, a synergistic relationship has traditionally been observed between spatial distribution of vegetation and runoff structuring. The VDSH system stems from differential soil development and evolution processes between areas under canopies and bare ground (Caldwell et al., 2012; De Ploey, 1984; Nulsen et al., 1986), resulting in feedback mechanisms perpetuating or further accentuating the bare ground and under-canopy soil dichotomy. Observations in semiarid rangelands (figure C-9A) suggest that deposition mounds form upstream of plant clumps as a result of energy losses and changes in transport capacity that accompany overland flow diversion by plant stems (Meire et al., 2014; Rominger and Nepf, 2011). Entrapment of nutrients along with sediments in these mounds creates areas of nutrient concentration where plants thrive–spatially alternated by bare or poorly vegetated zones of water and nutrient depletion–forming the premise of the "resource islands" or "vegetation island" concept (Li et al., 2007; Ridolfi et al., 2008). From a hydraulic standpoint, these "vegetation islands" can further exacerbate the concentrated flow process (figure C-9B).

C. Examples of this negative feedback loop are seen most often in shrub-dominated landscapes in the United States, which have formed coppice dunes such as sagebrush (*Artemisia* spp.), saltbush (*Atriplex* spp.), creasotebush (*Larrea tridentate*, DC. Coville), mesquite (*Prosopis glandulosa* Torr.), greasewood (*Sarcobatus vermiculatus*, Hook. Torr.) and in Pinyon (*Pinus* spp.) and Juniper (*Juniperus* spp.) woodland dominated areas in arid and semi-arid rangelands (Davenport et al., 1998; Eldridge and Rosentreter, 2004; Li et al., 2013; Pierson et al., 1994a and 1994b; Spaeth et al., 1994).

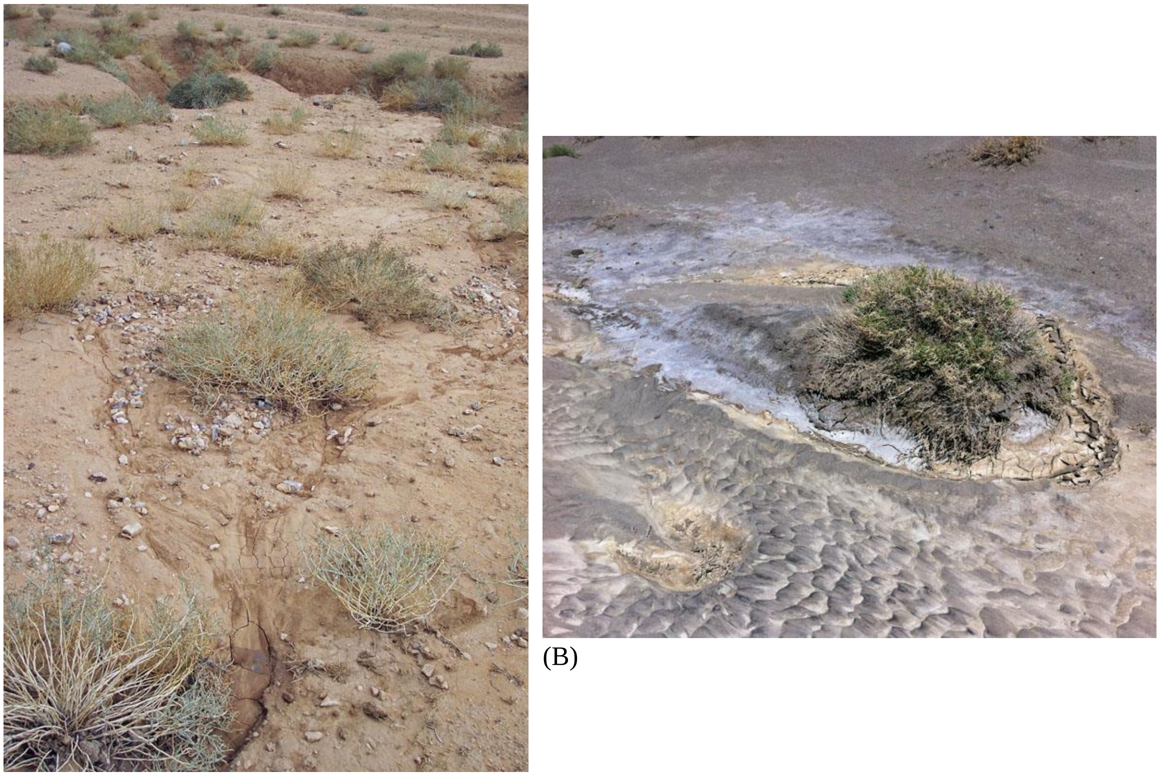
D. These studies illustrate that a coarsely arranged source-sink structure, as observed on degraded sites, potentially generates and releases more surface runoff than a finely structured source-sink community (Abrahams et al., 1995; Bhark and Small, 2003; Davenport et al., 1998; Parsons et al., 1996; Schlesinger et al., 1990; Wilcox et al., 1996). Numerous studies have been conducted that provide comparative examples of these relationships for fine (grassland) versus coarsely arranged (shrubland) rangeland communities in southern Arizona (Parsons et al., 1996; Turnbull et al., 2010; Turnbull et al., 2012; Wainwright et al., 2000). Pierson et al. (2013, 2010) and Williams et al. (2014a) present examples of similar relationships following conifer encroachment into Great Basin shrub steppe.

Figure C-8. 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).

E. Experimental research at the Walnut Gulch Experimental Watershed in southern Arizona revealed that coarsening of the spatial structure of vegetation in shrublands led to an increase in flow concentration and erosion rates (Abrahams et al., 1995; Parsons et al., 1996; Wainwright et al., 2000). VDSH influences not only runoff partitioning into sheet and concentrated flow processes but also seems to control flow characteristics in hillslope rills and channels. The same landscape with uniform disturbance may experience significantly more runoff and soil loss from a similar runoff event due to increased connectivity of bare soils and formation of well-organized concentrated flow paths. These organized flow paths rapidly accelerate runoff velocity and the ability of water to erode and transport sediment downslope (Davenport et al., 1998; Urgeghe et al., 2010; Wilcox et al., 1996). Tongway and Ludwig (1997) found, for example, that on degraded tussock grasslands, overland flow was concentrated in long straight paths between the grasses. In the good condition grassland, overland flow was tortuous, uniformly distributed, and produced less soil loss. Plant community physiognomy affects concentrated flow by controlling the connectivity of runoff and sediment sources and the energy of overland flow where it does occur (Williams et al., 2014a, 2015, and 2016). On well-vegetated rangelands, downslope transmission of runoff and erosion generated by raindrop splash and sheetflow in isolated bare or sparsely vegetated patches is limited by ground cover or roughness elements that promote infiltration and deposition (Pierson et al., 1994a, 1994b, and 2009; Reid et al., 1999; Wilcox et al., 2003a and 2003b). Soil detachment by concentrated flow is well-correlated with flow velocity and discharge (Al-Hamdan et al., 2012a and 2012b; Govers et al., 2007; Nearing et al., 1997 and 1999; Pierson et al., 2008a and 2009). Flow velocity is strongly related to discharge (Al-Hamdan et al., 2012a and 2012b; Gimenez and Goves, 2001; Goves et al., 2007; and Nearing et al., 1997 and 1999).

F. Grass clumps, plant bases, root mounds, and litter dams create topographic highs that may concentrate overland flow where runoff occurs, but the transport and erosive energy of concentrated flow are greatly reduced when flow intersects these roughness elements (Abrahams and Parsons, 1994; Abrahams et al., 1991; Al-Hamdan et al., 2012b and 2013; Giménez and Govers, 2001; Govers et al., 2007; Nearing et al., 1997 and 1999).

Figure C-9. (A) Vegetation islands redirecting concentrated flow around, resulting in deposition on the down slope side of the hillslope. (B) Greasewood coppice dunes illustrating interconnected bare spaces and linked concentrated flow paths that readily transport salt in surface runoff or are vulnerable to wind transport of soils and salt.



(B)

(A)

G. Reduced flow velocities and energy limit detachment and transport and allow surface runoff to disperse and for sediment to fall out of suspension. Rangeland studies from the Great Basin Region, USA, have reported two-fold higher concentrated flow velocities for experiments on bare plots (80 percent bare ground) relative to well-vegetated plots with 20–60 percent bare ground, (Pierson et al., 2007a, 2007b, and 2009). In those studies, erosion from concentrated overland flow was four-fold to eight-fold greater for bare plots than well-vegetated plots. Sediment transported by concentrated flow, where it does occur on well-vegetated sites, often forms miniature alluvial fans adjacent to vegetative clumps (Emmett, 1970; Meire et al., 2014; Rominger and Nepf, 2011). These features indicate that concentrated flow does redistribute surface soil from bare areas to vegetated zones on hydrologically stable rangelands, but hillslope soil loss from this process is minor under such conditions (Pierson et al., 2007a, 2007b, and 2009). Al-Hamdan et al. (2013) infer that the existence of a channel network is dictated not by hydraulic stresses exerted by runoff on bare soil but rather by the spatial distribution and structure of vegetation at which this network is in equilibrium. Concentrated flow becomes the dominant erosion mechanism on degraded rangelands where ground cover is sparse (Pierson et al., 2008a, 2009, 2008b, 2013, and 2011; Williams et al., 2014a, 2015, and 2014b).

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Figure C-10. Teaching principles of rangeland soil erosion in Jordan (right) and Kazakhstan (left) and how to use the Rangeland Hydrology and Soil Erosion Model to assess sustainability.

Subpart D – Vegetation and Management Interaction on Soil Erosion Processes

646.41 Contributors:

Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.
C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ, Tucson, AZ.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX.

Figure D-1. Grazing by domestic livestock and wildlife influence rangeland hydrology and soil erosion processes primarily by altering the type, quantity, and distribution of vegetation across the landscape.

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Figure D-2. Great Basin sagebrush steppe with high resistance to soil erosion due to dense shrub cover and minimal connected interspaces.

Figure D-3. Great Basin salt desert bottom lands. Source: ARS.

646.42 Grazing

A. A governing principal of land management is that changes in land cover result in changes in watershed condition and response (Hernandez et al., 2000; Miller et al., 2002). Land management practices influence runoff and soil erosion on rangelands because they affect plant distribution, biological diversity, canopy, ground cover, and soil properties (Spaeth et al., 1996). Vegetation is the primary factor controllable by human activity that influences the spatial and temporal variability of surface runoff and soil erosion on rangelands (figure D-3). Hydrologic response (runoff volume, peak discharge, and sediment yield) at either the hillslope or watershed scale can be used as an integrated index to quantify impacts of conservation practices, such as prescribed grazing (Thurow et al., 1986; 1988; Weltz and Wood, 1986a and 1986b), invasive species control (Afinowicz et al., 2005; Pierson et al., 2007a and 2007b), or impacts of fire (Knight et al., 1983; Reed and Schaffner, 2007). Rapid advances in hydrologic assessments at hillslope and watershed scales have been facilitated by the revolution in computer science and information technology. Furthermore, remote sensing and analytical tools to display the results through geographic information systems (GIS) now make it possible to quantify the spatial and temporal changes in land cover at both national and regional scales to understand and predict soil erosion.

B. This handbook focuses on how management can influence upland soil erosion processes on rangelands. The Natural Resources Conservation Service commissioned a review of impacts and benefits of numerous conservation practices (grazing, fire, etc.) as part of the Conservation Effects Assessment Project program that documented the benefits and impacts of conservation on rangelands. For insights on how other management practices alter rangeland ecosystem goods and services, other than soil erosion, see Briske (2011).

C. Vegetation amount, distribution, and lifeform are the primary factors controllable by human activity that influence the spatial and temporal variability of surface runoff and soil erosion on rangelands. Blackburn (1975) found that shrub coppice dunes have significantly different erosion rates than the associated interspace areas. Erosion decreases significantly as plant lifeform changes from short grass to midgrass to tall grass (Thurow et al., 1986 and 1988). Grazing management practices impact soil erosion on rangelands through their influence on the type, amount, and distribution of cover (Gifford and Hawkins, 1978). By reducing both canopy and ground cover and increasing the number and size of bare soil patches, improperly applied grazing management practices increase the risk that a site will have significantly different erosion rates than the associated interspace areas and be eroded by both raindrop splash and concentrated flow path processes.

D. Grazing as a range management practice implies a host of activities promoting the consumption of standing forage by domestic animals. The effect of grazing on rangeland hydrology has been well documented, and inferential relationships to salinity have also been proposed. Grazing may influence hydrology through vegetation alteration and direct soil surface property modification as a result of animal trampling (figure D-3).

E. In general, the effects of livestock grazing on hydrologic resilience are associated with the degree to which grazing pressure affects surface susceptibility or the fire regime (Gifford and Hawkins, 1978; Thurow, 1991, 1986, and 1988; Trimble and Mendel, 1995; Wood and Blackburn, 1981). Grazing pressure that substantially reduces vegetation and ground cover or compacts and disturbs surface soils will likely increase losses of water and soil resources through water and wind erosion processes (Field et al., 2011; Greene et al., 1994; Trimble and Mendel, 1995). Soil compaction (increased bulk density) and disturbance associated with intense or repetitive grazing will increase runoff and erosion, and these effects are strongly influenced by the season of use (Branson and Miller, 1981; Daniel et al., 2002; Gifford et al., 1978; Greenwood and McKenzie, 2001; Teague et al., 2011; Warren et al., 1986a). High intensity grazing, particularly over multiple years, can alter plant composition such that the biotic structure triggers long-term site degradation through abiotic-driven losses of water and soil resources (Greene et al., 1994; Ludwig et al., 2007; Rietkerk and Van de Koppel, 1997; Schlesinger et al., 1990; Turnbull et al., 2008 and 2012; van de Koppel et al., 1997; Warren et al., 1986b and 1986c).

F. Two production scale grazing treatments were sampled in north Texas to evaluate their impact on hydrologic processes. Treatments were year-long continuous grazing, stocked at a moderate rate (MC); 16-paddock rotational grazing treatment, stocked at a heavy rate (RG); moderately stocked pasture; 3-herd deferred rotation treatment (DR); and ungrazed exclosure (EX). As above-ground biomass and cover increased, there was a corresponding increase in infiltration rates and decrease in sediment production. The RG treatment had higher sediment yield and lower infiltration rates than the MC treatment. Infiltration rates and sediment production in the RG and DR treatments before grazing were not significantly different from those in the MC treatment. Grazing caused a significant decline in infiltration rates and a significant increase in sediment production in both treatments as a function of removal of above-ground biomass, cover, and proportion of the area in midgrasses. Midgrasses had higher infiltration rates and lower sediment production than shortgrasses. Sediment production was lowest in the exclosure (Pluhar et al., 1987). This is similar to results that Weltz and Wood found (1986a and 1986b), that high intensity rotational grazing removes excessive amounts of above-ground biomass and increases the vulnerability of the site to accelerated soil erosion.

G. West et al. (1984) reported that after 13 years of no livestock grazing in west central Utah that desirable perennial vegetation had not been reestablished despite a trend of increased precipitation over the length of the study. They concluded that the site had transitioned to a stable shrub-dominated site. The concept that removing livestock would return the plant community to the original sagebrush-native shrub-grass assemblage was unlikely. Therefore, direct manipulation of the site is mandatory if rapid return to the desired plant community is desired. Belnap et al. (2009) reported that grazed watersheds in southeast Utah had significantly more soil loss from wind than ungrazed watersheds. When comparing soil losses among the sites, they determined that biological soil crusts were the most important in predicting site stability, followed by perennial plant cover.

H. Extensive woody plant encroachment and subsequent amplified runoff and soil erosion across much of the western USA have been partially attributed to intensive grazing and an associated decrease in wildfire activity during the 20th Century (Archer et al., 1995; Bahre and Shelton, 1993; Buffington and Herbel, 1965; Grover and Musick, 1990; Miller, 2005; Miller and Wigand, 1994; Pierson et al., 2010; Romme et al., 2009; Turnbull et al., 2010a, 2010b, and 2012; Van Auken, 2000 and 2009). Overall, vegetation, soil properties, and the associated hydrologic and erosion responses to grazing can vary tremendously depending on inherent characteristics of the respective ecological site, pre-grazing rangeland condition, and the grazing prescription (Branson and Miller, 1981; Burke et al., 1999; Castellanos et al., 2005; Emmerich and Heitschmidt, 2002; Field et al., 2011; Milchunas and

Lauenroth, 1993; Teague et al., 2011; Thurow, 1991, 1986, and 1988; Trimble and Mendel, 1995; Vermeire et al., 2005). In many cases, proper grazing can be used to augment restoration of rangeland ecosystems or to reduce fuel accumulations and potential fire severity without negative ecohydrologic impacts (Briske, 2011).

I. Bentley (1978) published an extensive review examining the influence of grazing on rangeland vegetation. The recurring theme from this review relates to the threat of overgrazing on ecosystem health suggested by many studies of temporary or permanent loss of vegetative cover, especially during droughts, decline of desirable species through selective browsing, dominance of less desirable species, etc. He observed that during the severe droughts of 1933–1939 and 1952–1955 that prevailed in the Great Plains, loss of vegetative cover on heavily grazed grassland ranges was nearly double that on moderately grazed ranges and more than double that on ungrazed ranges. Erosion decreases significantly as plant lifeform changes from short grass to midgrass to tall grass (Thurow et al., 1986 and 1988) as a function of grazing intensity. Grazing management practices impact soil erosion and salinity transport on rangelands through their influence on the type, amount, and distribution of cover (Gifford et al., 1978). By reducing both canopy and ground cover and increasing the number and size of bare soil patches, improperly applied grazing management practices increase the risk that a site will be eroded by both raindrop and concentrated flow processes.

J. In the northern, central, and southern plains grasslands, the runoff and erosion potential of a site are closely related to management activity. Prolonged heavy continuous grazing results in significant change in plant community structure in which the more productive tall- and mid-grasses are replaced with less productive short-grasses, resulting in increased surface runoff and soil erosion (Rauzi et al., 1968; Thurow et al., 1988). Other studies have concluded that proper grazing and brush management practices result in infiltration, surface runoff, and soil loss characteristics similar to those of ungrazed landscapes (Blackburn, 1983; Blackburn et al., 1982; McCalla et al., 1984; Weltz and Wood, 1986b; Warren et al., 1986b; Thurow et al., 1986 and 1988).

K. The conventionally accepted wisdom is that ground cover is negatively related to runoff production and erosion (e.g., Hanson et al., 1970; Moore et al., 1979; Thurow et al., 1986; Wine et al., 2012) and presumably to surface water salinity (e.g., Bentley et al., 1980; Moore et al., 1979). Nevertheless, no consensus exists on the magnitude of change in hydrologic properties imputed to grazing, probably due to the diversity of research methodologies published in the literature. Numerous authors have evaluated the impacts of grazing and its impact on runoff and soil loss on the Badger Wash Watershed near Grand Junction, Colorado. They reported that very heavy grazing did increase runoff and sediment yield (Lusby, 1970, 1979, and 1964). However, Thompson (1968) reported that after 10 years with exclusion of grazing, infiltration rates were lower in both the grazed and ungrazed watershed. They found that the time of year the sample was collected had more influence on infiltration rates than grazing. Branson and Owen (1970) reported that runoff from all 17 watersheds was directly related to percent of bare soil. Sediment yield was not significantly related to bare soil. They also reported that the time of year and when vegetation was sampled (spring vs. fall) had significant impact on the ability to predict runoff.

L. Using runoff and sediment monitoring data (1956–1966) from the Badger Wash Basin, CO, Lusby and Reid (1964) found that grazing increased total watershed runoff 1.3- to 1.45-fold and sediment yield 1.8-fold. Local soil properties have also been used to quantify the impact of grazing on hydrology. Bentley (1978) reviewed an extensive list of studies demonstrating a positive correlation between vegetation health indicators (ground cover and plant successional development) and infiltration rate. Intensive grazing, for example, may cause reduction in ground cover which would in turn expose a larger area of the soil to raindrop impact, causing surface sealing, increased bulk density, and decrease in infiltration rate (e.g., Thurow et al., 1986; Wilcox and Wood, 1988).

M. Vegetation composition has a direct influence on water balance and ground cover (Moore et al., 1979) and has been proposed as a potential factor controlling runoff. Hanson et al. (1970) observed a predominance of short grasses and sedges on heavily grazed watersheds, while lightly grazed watersheds showed a mixture of grasses with a large proportion of western wheatgrass. Thurow et al. (1986) demonstrated that the presence of trees in pastures (oak in the study) promotes increased infiltration and lower erosion. In an analysis aiming at proposing salt reduction alternatives in Upper Colorado River Basin rangelands, Bentley et al. (1980) recognized the dependence of watershed hydrology on vegetation composition by assuming that one percent of bare soil would result in a 17.3 mm (0.68 inch) increase in runoff in most studied vegetation communities and 60.2 mm (2.37 inches) in Shadescale (*Atriplex confertifolia* (Torr. & Frém.) S. Watson)-Galleta grass (*Pleuraphis jamesii* Torr.) and Big sagebrush (*Artemisia tridentata* Nutt.)-Shadescale communities. However, the paucity of observed data linking vegetation state and composition to salt production in runoff limits any knowledge in this domain to be inferred from the known effect of grazing on hydrology. Besides the lack of observed data, understanding the impact of grazing on salinity is further complicated by the confounding effect of animal trampling on runoff generation and salinity.

N. One of the most referenced soil property changes resulting from animal trampling is bulk density increase by compaction, leading to decrease in infiltration rate and potential increase in erosion (e.g., Bentley, 1978; Hiernaux et al., 1999; Warren et al., 1986c). The negative effect of trampling on soil properties is exacerbated when grazing occurs in riparian zones (e.g., Belsky et al., 1999; Flenniken et al., 2001). In fact, a review of literature on grazing influence on stream and riparian ecosystems by Belsky et al. (1999) revealed many studies reporting negative impacts, but none reporting potential benefits on these ecosystems. Some studies have, however, shown in some cases (e.g., George et al., 2002) no apparent effect of grazing on stream morphology.

O. Gilley et al. (1996) evaluated grazing and haying effects on runoff and erosion on a former Conservation Reserve Program (CRP) site near Streeter, North Dakota. Treatments evaluated undisturbed CRP, twice-over rotational grazing, season-long grazing, haying, and burning. No significant difference in runoff or erosion was found between the season-long grazing and burned treatments after initial period for treatment was established. Grazing and haying of the CRP lands resulted in a significant increase in runoff. Similar amounts of erosion were measured from the grazing and hayed treatments. Use of CRP sites for grazing or haying would not be expected to result in excessive erosion if adequate canopy and basal cover are maintained.

P. Because inappropriate grazing management may lead to rangeland degradation and potentially increase surface water salinity, prescribed grazing management has been proposed as a key control in reducing salinity transport and soil erosion. Generally, the greater the bare soil amount, the greater the erosion rate. Levels of cover necessary for site protection against accelerated soil loss range from 20 percent in Kenya (Moore et al., 1979) to 100 percent for some Australian conditions (Costin, 1959). Most studies indicate that cover of 50 to 75 percent is probably sufficient (Gifford, 1984; Gifford et al., 1978; Orr, 1970; Packer, 1951). Bentley et al. (1980) suggested that moderate grazing (40 percent–60 percent utilization of forage plants) during winter when the soil is frozen would result in less compaction and disturbance by trampling, while periodic rest from grazing (rest-rotation) would also ensure healthy plant communities and the buildup of litter to protect soil from erosion. Bentley et al. (1980) estimated a potential salt reduction of 15 percent in the Upper Colorado River Basin and similar reduction in soil erosion from careful grazing management.

Q. Hydrologic response to grazing largely parallels those of other ecological variables in that stocking rate and weather are the dominant variables that must be addressed to achieve desired results. In many cases, prescribed grazing management can be used to augment restoration of rangeland ecosystems or to reduce fuel accumulations and potential fire severity, without negatively impacting hydrologic processes (Briske et al., 2011).

646.43 Fire

A. The ecology of North American rangelands depends strongly on fire (Fuhlendorf et al., 2011). The consensus about fire within the rangeland community is that native fire regimes have been altered by various human induced factors (figure D-4). Two significant trends have been reported in relation to these shifts in fire regimes: (1) intentional suppression of fire or reduction in fuel load through grazing has led to invasion of woody plants, and (2) increased fire frequency as a result of invasion of exotic herbaceous species (*Bromus tectorum* L, Cheatgrass) (Fuhlendorf et al., 2011). In some cases, fire has been used or prescribed by various rangeland management agencies for restoration of historical rangeland conditions or promotion of specific rangeland services.

B. The consequences of fire on rangeland hydrologic processes include loss of vegetative cover, increasing vulnerability to wind and water erosion, and physiochemical changes in the soil surface layer depending on burn severity, resulting in temporary water repellency and increased runoff (e.g., Glenn and Finley, 2010; Pierson et al., 2011; Wright et al., 1976). Long term effects of fire on hydrologic processes in general follow the trend of vegetation and ground cover recovery to pre-fire amounts and distributions. Most published long-term effects of fire on hydrology suggest that the short-term detrimental effects of fire on runoff and erosion wane as vegetation is progressively reestablished. Prescribed fire can be used successfully by various rangeland management agencies for restoration or promotion of specific rangeland services successfully (e.g., Garza Jr. and Blackburn, 1985; Knight et al., 1983; Wright et al., 1976). Wright et al. (1976) has, however, found that recovery from the initial detrimental effect of fire is lengthened on steeper slopes to 15 to 30 months or more, compared to 9 to 15 months of recovery time on moderate slopes. Time of year of the fire will also influence the recovery and soil erosion rates due its impact on vegetation response after the fire (Pierson and Williams, 2016).

Figure D-4. Wildfire in Pinyon-Juniper woodland. Source: BLM.

C. Fire alters hydrologic processes by altering the geospatial structure of plants and removal of biomass, thereby increasing surface susceptibility to transport of salt by increasing runoff and soil loss by wind and water on bare soils immediately following the fire (figure D-5). The destruction of organic matter in soils during a wildfire can alter soil structure and aggregate stability, increase bulk density and pH, and decrease porosity and infiltration capacity and rates (Giovannini and Lucchesi, 1997; Hester et al., 1997; Mataix-Solera et al., 2011; Stoof et al., 2010). Fire can also damage or kill invertebrates, microorganisms, and mycorrhizae fungi in the surface soil with extremely hot fires. These organisms facilitate soil aggregation, nutrient cycling, and infiltration rates and capacities (DeBano, 2000; Mataix-Solera et al., 2011; Shakesby and Doerr, 2006). Physical soil crusting may occur after fire in soils that are high in clay. This can reduce water infiltration rates and water availability and slow recovery after fire (Mills and Fey, 2004; Snyman, 2002). Soil health and quality are also degraded by increased rates of soil erosion and surface runoff following fire (Emmerich and Cox, 1994; Pierson et al., 2011; Shakesby, 2011; Williams et al., 2013). If not mitigated, accelerated soil erosion can degrade an ecological site to such a state that it can permanently alter its form and function (Herrick et al., 1999). A detailed and comprehensive review of fire effects on vegetation and soils of the western United States is provided by Pierson et al. (2011), Miller et al. (2013), and Pierson and Williams (2016).

D. The impact of fire on hydrologic and erosion processes largely depends on the spatial arrangement of burn severity, bare soil exposure, degree of water-repellency created, rainfall intensity, and storm patterns (Al-Hamdan et al., 2011). Rare, often unexpected, rainfall event(s) may trigger a nick-point along the hillslope and facilitate the formation of a concentrated flow path (rill). On rangelands, these concentrated flow paths facilitate water accumulation and accelerate soil erosion (Al-Hamdan et al., 2012). If left unchecked, these concentrated flow paths can remove enough soil to result in the site crossing an ecological or hydrologic threshold and becoming permanently degraded (Pierson et al., 2013; Urgeghe et al., 2010). Intense rainstorms after wildfires may cause flooding and mudflows and result in extensive damage to property and infrastructure (Klade, 2007; Pierson et al., 2002). In rangeland and woodland dominated watersheds, Reed and Schaffner (2007) found a 10-fold increase in peak flow rates and soil erosion following fire. The magnitude of these changes was so great that they developed a new procedure to estimate potential post-burn flash floods for use by NOAA.

E. The relative post-fire hydrologic recovery of rangeland plant communities is primarily influenced by the pre-fire ecological state, fire severity, and post-fire climate and land use that relate to vegetation recovery (Kinoshita and Hogue, 2011; Miller et al., 2013). The pre-fire ecological state influences spatial variability in burn severity and post-fire plant recruitment. High severity burns on productive sites may consume nearly 100 percent of canopy and ground cover, but runoff and erosion rates can return to pre-fire levels within one to three years, respectively, depending on post-fire plant recovery (Pierson et al., 2009, 2008, and 2011). Relative hydrologic recovery appears to occur within one to three years post-fire. Burned rangelands will remain susceptible to runoff and erosion during extreme events until overall site characteristics (live plant and litter biomass) are consistent with pre-fire conditions (Pierson et al., 2011). Arid rangelands with warm/dry soil temperature/moisture regimes may require longer periods to recover hydrologically than cool/moist sites and may be vulnerable to annual grass invasion and subsequent re-burning (Brooks and Chambers, 2011; Chambers et al., 2007; Davies et al., 2012).

Figure D-5. Concentrated flow paths formed and actively eroding as result of loss of protective vegetation and formation of hydrophobic surface soil layer following wildfire in Central Nevada.

646.44 Mechanical

A. Gully Plug

(1) In the 1930s the US Government, through the Civilian Conservation Corps, established numerous earthen, rock, and brush structures (gully plugs) to reduce soil erosion in northern New Mexico. When the sites were reevaluated in the 1990s, it was reported that 60 percent of the 47 structures had breached, and 65 percent of the structures were more than 50 percent full of sediment. Reasons for breaching of all structural types was related to piping, scour immediately below the structures, under-sized design for the size of the drainage area, and poor maintenance (Gellis et al., 1995). A gully plug (figure D-6) is a small earthen dam constructed at one or more locations along the gully to provide grade control and retain sediment (Schaffrath, 2012). Gully plugs and rock structures have been used for centuries to mitigate soil erosion problems (Castillo et al., 2007; Lenzi and Comiti, 2003; Xiang-zhou et al., 2004) by controlling the geomorphic grade of the area and the velocity of the runoff water. A side benefit of these structures is in capturing water and its use in subsistence agriculture (Norton et al., 2002) or in revegetating the area with natural rangeland vegetation (Nichols et al., 2012). Hessary and Gifford (1979) conducted a series of experiments to study the effect of various range improvement practices, including gully plugs, on salt loading to surface waters in the Upper Colorado River Basin. They compared salt accumulation in different layers of soil along channel bottoms upstream and downstream of a series of gully plugs to those measured in upland areas in the vicinity of the gullies. Even though some significant differences were found among sampling locations, no consistent trend in salt accumulation with sampling position was found as the result of experiments to study the effect of various range improvement practices, including gully plugs, on salt loading to surface waters in the Upper Colorado River Basin.

Figure D-6. Rock gully plug in southern Arizona used to retard overland flow and trap sediment behind, reducing futher downstream erosion through reduction in runoff velocity.

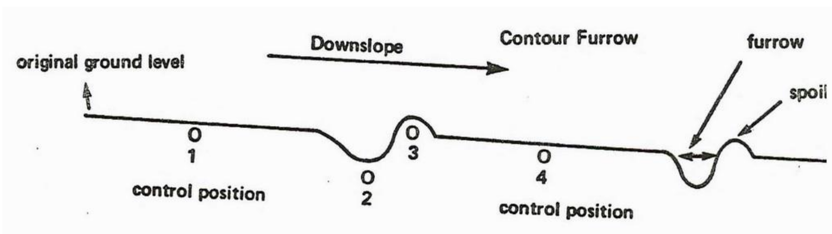
(2) Gully plugs can be effective in reducing runoff and sediment delivery out of the watershed. However, they should be viewed as an engineering solution of last resort as they are costly to install, require frequent maintenance of the structure, and frequently require the sediment be removed and placed in an appropriate place. Failure to provide this maintenance will result in failure of the structure, and all trapped sediment and salt will then be remobilized and

transferred downstream. The need to install a gully plug structure indicates that the uplands are unstable and unsustainable. In general, it should be less expensive to stabilize the uplands through vegetation management than through costly engineering solutions.

(3) Contour furrowing (figure D-7) involves converting a level soil surface into a series of ridges and furrows along the contours of the landscape (Bates et al., 2005), allowing soil to retain water and retard runoff. Since contour furrows were found to be an effective land treatment practice to reduce runoff and sediment production in rangelands (Branson et al., 1966; Gifford et al., 1977), it is conceivable to hypothesize that this practice may have an effect on salt loading and redistribution along the furrowed hillslope.

(4) Riley et al. (1982) measured surface runoff quality from the Coal Creek (a sub-watershed of the Price River basin) 15 years post implementation of contour furrowing and found lower total dissolved solids (TDS) compared to other sub-watersheds where this practice was not implemented (Soldier, Wattis, and Grassy trail sub-watersheds). However, due to the absence of prior treatment data and the spotty nature of the treatment data, Riley et al. (1982) were unable to unequivocally attribute the lower TDS to contour furrowing.

Figure D-7. Cross section of furrowed hillslope showing relative position of sampling points (Hawkins et al., 1977).



(5) Hawkins et al. (1977) looked at the effect of contour furrowing on the spatial redistribution along the furrowed hillslope. Soil samples were taken at different positions along upslope, downslope, and inside the furrows. On three out of 11 study sites they found a significant furrowing effect on spatial distribution of salt. Two of the significant results indicated a higher concentration inside the furrows, while the third significant case suggested the opposite trend. On eight of the 11 sites they found a higher salt concentration in the top layers of non-saline soils compared to the bottom layers, suggesting that contour furrowing may promote salt entrapment in non-saline soils.

(6) The Vallerani plow creates a deviate and pushes up soil to form a berm (i.e., bund) that traps water from the uphill slope (Gammoh and Oweis, 2011a; Gammoh and Oweis, 2011b). The Vallerani plow (figures D-8 and D-9) mimics the natural process of spatial distribution of vegetation (VDSH) and provides the necessary water for plant establishment and initiation of the restoration process, if properly located to prevent overtopping and breaching of the bunds. This mechanized water harvesting system provides additional water to the shrubs transplanted into the depression (*Atriplex* spp. and *Salsola* spp.) that is necessary for their survival. In addition, a ripping blade is part of the system that is pulled through the soil to a depth of 60 cm. This fractures the cemented petrocalcic subsoil and improves water storage capacity in the soil profile. Without this treatment, the cemented petrocalcic horizon impedes downward percolation of water and limits the volume of stored soil moisture to the upper 50 cm of the soil profile. This limits the ability of the soil to store the water derived from water harvesting that is critical for plant growth and survival in these dry areas. This technique has proven successful in watershed restoration in Jordan and other countries in the Middle East and North Africa.

Figure D-8. Restored watershed using Vallerani plow, with reduced runoff and soil erosion and increased forage production.

Figure D-9. Vallerani plow.

B. Chaining and Cabling

(1) Schott and Pieper (1987) examined the recovery of the vegetation community following cabling a Pinyon-Juniper rangeland in New Mexico. The pattern of vegetation recovery and the types of vegetation and biomass produced following Juniper removal had an important influence on both the short- and long-term hydrology of a site. The rate of recovery and extent of biomass produced was largely dependent on soil depth. Because the cover on sites that were dominated by annual plants varied seasonally, the soil was well covered during portions of the year but bare during other times, resulting in seasonal changes in soil erosion potential. Runoff and erosion will increase directly after cabling because of exposure of bare soil on these annual plant-dominated communities. As perennial vegetation began to dominate the site, vegetation cover increased and became more stable over time, and erosion potential decreased. See figure D-10.

Figure D-10. Chaining to remove trees killed in a wildfire. Source: BLM.

(2) In Utah, several studies evaluated the effects of chaining and debris management on site hydrology (Gifford and Shaw, 1973; Gifford et al., 1970; Williams et al., 1969 and 1972). The three treatments included chaining with windrowed debris, chaining with debris left in place, or untreated. Sites were seeded with perennial grasses following chaining. The chained and windrowed sites produced 1.2 to 5 times more runoff and 1.6 to 6 times more sediment than the undisturbed sites. The chained sites with debris left in place produced equal or less runoff and sediment than the undisturbed sites. Windrowing the debris required additional use of mechanical equipment (greater soil disturbance) and also exposed large areas of bare soil. This caused infiltration to be lower on windrowed sites.

(3) Recent studies in Utah evaluating chaining after burning have reported that chaining reduced the effect of the hydrophobic layer of the surface soil, increased infiltration rates, and increased desired seeded grasses (Madsen et al., 2015; Ott et al., 2003). Roundy et al. (2016) and Farmer et al. (1999) evaluated two different chainings in semi-arid sagebrush steppe sites in Utah that were in an advanced phase of encroachment by Pinyon (*Pinus* spp.) and Juniper (*Juniperus* spp.) trees. They measured runoff and sediment for five years following treatment in each of the studies. Chaining increased cover by 5 times, reduced runoff by 5 to 6 times and sediment yield by 9 to10 fold. Both studies attributed the effectiveness of the treatment, in part, to the above average precipitation following the treatment. This allowed the herbaceous vegetation to establish and reduce bare soil. On rangeland sites in Nevada, there was no difference between chained and seeded sites and undisturbed sites 5 to 11 years after treatment (Blackburn and Skau, 1974). The effectiveness of chaining or cabling to remove trees depends on site disturbance, the amount of residue bare soil after the treatment, the precipitation following treatment, and the management of the site to appropriate grazing levels. If sufficient precipitation follows the chaining treatment and allows for establishment of herbaceous vegetation, then recovery of the site is possible and reversible, assuming that the abiotic threshold of advanced accelerated soil erosion to a condition of sustainability can be achieved (figure D-11).

Figure D-11. Hand cutting of Pinyon and Juniper trees with slash laid perpendicular to slope to reduce runoff and soil erosion.

C. Cutting

(1) The concern over the expansion of Pinyon-Juniper woodlands and methods of controlling them (chaining, cutting, and bulldozing) have been evaluated for decades (figure D-11). Most of the early research focused on increasing forage production (Williams et al., 1969). Recent studies have focused on watershed value and soil protection aspects of such land management techniques (Roundy et al., 2016). On steeply sloped landscapes, a common technique is to cut the trees and lay the slash on the soil surface to reduce soil erosion (Hessing and Johnson, 1982; Jacobs and Gatewood, 1999; Wood and Javed, 2001). Woodland management practices designed to cut and thin Pinyon-Juniper trees can significantly reduce water and soil loss by altering the type, amount, and distribution of ground cover through the addition of slash (Jacobs and Gatewood, 1999; Stoddard et al., 2008).

(2) The spatial arrangement of the slash piles is critical because it influences erosion behavior (Reid et al., 1999). The slash should be placed perpendicular to the slope to reduce slope length and should be in direct contact with the soil surface to increase ground cover and minimize soil erosion processes.

(3) Pierson et al. (2007a; 2007b) compared uncut Juniper stands with stands that had been cut and allowed to recover for ten years. It was found that the uncut areas rapidly produce significant amounts of runoff, with cumulative sediment yield being two orders of magnitude higher for the woodlands compared to the areas where the understory had successfully reestablished itself. Additionally, areas of concentrated flow were more frequent in the woodlands and produced higher sediment concentrations.

(4) Bare intercanopy zones can have up to 24 times more sediment loss than in the canopy zones (Reid et al., 1999). These large areas of interconnected interspaces with minimal ground cover provide opportunities for water to form concentrated flow paths and accelerate the soil erosion process. Erosion from these interspace zones increases with decreasing ground cover

(Wilcox, 1994). One of the largest factors in limiting sediment movement on Pinyon-Juniper woodlands is the addition of vegetation cover in the form of slash (Hastings et al., 2003; Stoddard et al., 2008; Wood et al., 1987). Vegetation cover increases roughness which can reduce sheet flow and concentrated flow, as well as diminish sediment transport capacity. Reduction in soil movement is likely to be attributed to the physical barriers associated with litter and branches from the addition of slash (Jacobs and Gatewood, 1999; Stoddard et al., 2008; Wood and Javed, 2001). These physical barriers aid in reducing the effective length of the slope, which limits the ability of concentrated flow paths to erode the soil (Dunne and Foster 1978).

D. Off Road Vehicles

(1) Access control entails a host of measures regulating animals, people, and vehicular traffic in an area. In the context of this handbook, access control refers specifically to the exclusion of vehicles from rangelands (Off Highway Vehicles, OHV) (figure D-12). In most consulted references, the effect of OHV exclusion was assessed by comparing disturbed to undisturbed areas**.**

Figure D-12. Off highway vehicles. Source: BLM.

(2) Scientific consensus concluded that OHV use on rangelands often results in land degradation and accelerated erosion when vehicles are used indiscreetly across the landscape (figure D-12). Nevertheless, the magnitude of damage incurred by OHV-disturbed rangelands compared to undisturbed sites varied substantially between studies. Dohrenwend (2003) reported erosion rates five times higher on OHV-disturbed Mancos shale hillslopes compared to native erosion rates, while Goodloe (2006) suggested erosion rates 26 times soil loss tolerance levels due to OHV activity in the Panoche Hills area of California.

(3) Impacts of motorcycle and 4-wheel drive truck traffic on infiltration rates and sediment production were evaluated on two desert soils in southern Nevada (Eckert Jr. et al., 1979). Infiltration was similar for both soils. More sediment was produced from a surface with exposed mineral soil than from gravel-mulched surfaces. Infiltration was 3 to 13 times greater on the coppice dunes beneath shrubs than on interspace soil between shrubs. Sediment rates were10 to 20 times greater on interspace soil. After soil was disturbed by vehicular traffic, infiltration was decreased, and sediment yield was greater. Protection of shrub-dominated areas will minimize overall soil erosion on these sites. Shrub coppice dunes had higher infiltration rates, more organic matter, and greater aggregate stability in comparison to the interspaces–which resulted in reduced soil erosion rates.

(4) Off-road military vehicle traffic is a major consideration in the management of military lands. Fuchs et al. (2003) evaluated the impacts of military tracked M1A1 heavy combat tank vehicles on sediment loss from runoff, surface plant cover, and surface microtopography in a desert environment. In general, sediment losses from tank treatments did not differ from natural sediment losses under nominal rainfall events. Intense rainfall events did generate significantly greater sediment losses from the tank triple pass treatments. When disturbances were imposed under dry seasonal conditions, triple pass M1A1 tank impacts had detrimental effects. With current precipitation regimes, a minimum of three years for most triple pass

M1A1 tank impacts is suggested for suitable vegetation recovery and soil stability. Increase in erosion rates due to OHV have been related to dramatic changes in soil surface hydraulic characteristics (Iverson, 1980). In fact, Iverson (1980) estimated a 13-fold decrease in Darcy-Weisbach hydraulic roughness friction factor and a 5.5-fold increase in Reynolds numbers–two trends suggesting increase in runoff velocity, erosivity, and soil erosion.

646.45 Herbicide

A. Herbicide treatment removes undesired vegetation to increase desired vegetation, reducing soil erosion. Increasing desired vegetation will decrease bare soil and the interconnectedness of bare soil concentrated flow paths through interspace between plants. The reduction of connected interspaces is accomplished by increasing tortuosity of flow paths, as undesirable shrubs or woody plants are removed and replaced with herbaceous vegetation in the interspaces. This results in reduced velocity and reduced transport capacity, and deposition will occur upstream of the basal area of the new vegetation. Furthermore, with reduced flow velocity, the residence time is increased, providing more time for water to infiltrate into the soil, further reducing flow velocity and volume. If the amount of bare soil is decreased, then raindrop splash erosion will decrease, further reducing sediment available for transport off site. The theory is explained using Pinyon and Juniper woodlands by Davenport et al. (1998) and Williams et al. (2016).

B. The long-term effectiveness of herbicide applications as a range management tool to reduce upland soil erosion at the watershed scale on arid and semi-arid rangelands is not universal and depends on the type and density of vegetation removed, the type and density of vegetation that is increased, and the precipitation regime. Work in northern Arizona in the 1950s and 1960s was designed to augment water availability through removal of Pinyon and Juniper trees. When herbicides were aerially applied to kill Pinyon and Juniper trees, no change in sediment yield was reported on watershed WS3 at the U.S. Forest Service Beaver Creek Experimental Watershed in northern Arizona (Lopes et al., 1999). The reason for this was that the soil surface was not disturbed, and the trees were left in place.

C. Herbicides have been widely used for removing honey mesquite and enhancing forage in Texas. The highest infiltration rate and lowest sediment production were closely associated with honey mesquite canopy zone. In shortgrass interspace areas, infiltration and erosion rates were about one-half of the canopy zone rate. Midgrass-dominated areas had the greatest infiltration rate improvement due to treatment. Sediment production on the shortgrass interspace was double that of the mesquite canopy zone or midgrass interspace areas. As above-ground biomass and cover increased, bare soil decreased, and soil erosion decreased (Brock et al., 1982). In contrast, Bedunah and Sosebee (1985) reported that use of herbicide to kill honey mesquite did not alter infiltration rates in north Texas. They did find that various mechanical treatments increased infiltration rates, primarily through the interaction of increased litter and above-ground biomass.

D. Balliette et al. (1986) reported that total sediment production was about 29 percent to 41 percent higher under the canopy of tebuthiuron-treated sagebrush, compared to the canopy zone of untreated rangeland. These differences were variable and were significant at only one site. Total sediment production was related primarily to a combination of soil texture, sagebrush canopy cover, and total vegetation production. Overall infiltration rates and sediment concentration in runoff within the canopy zone and interspace areas were not affected by chemical control treatments to reduce sagebrush.

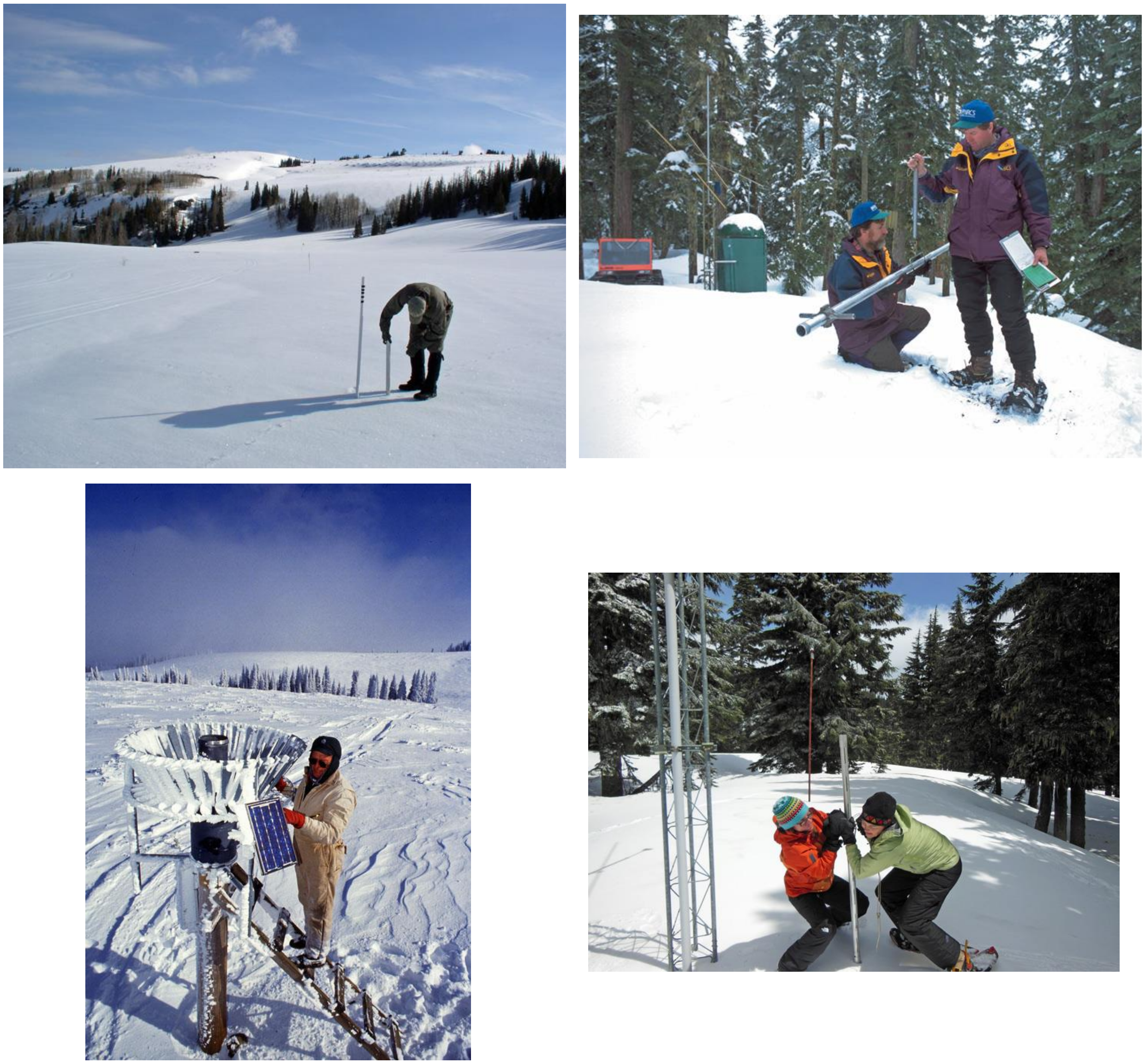
646.46 Water Augmentation

A. Water yield increase (augmentation) by removal of Pinyon-Juniper trees by herbicides did produce an increase in stream flow, but only in years with above average precipitation on the Beaver Creek Watershed in northern Arizona (Baker, 1984). The increase in water yield will decease as the woody vegetation reestablishes itself across the watershed, making water yield augmentation usually economically unfeasible. Research across the Colorado River Basin has found that little if any sustained increase in water yield should be expected by removal of Pinyon-Juniper trees, no matter what treatment is used to remove Pinyon and Juniper trees (Baker and Ffolliott, 2000; Ffolliott and Brooks, 1988; Roundy and Vernon, 1999).

B. The area where vegetation control (e.g., mechanical) has been demonstrated as effective is in areas where Pinyon and Juniper trees have been removed, and shrubs and grass have successfully established in the historic bare interspaces as discussed above in the previous sections (Buckhouse and Mattison, 1980; Bybee et al., 2016; Gaither and Buckhouse, 1983; Pierson et al., 2007a and 2007b). In northern Texas, there is an indication that removing shrubs can increase infiltration and reduce soil erosion rates (Bedunah and Sosebee, 1985; Brock et al., 1982). In the desert Southwest, conversion of desert grasslands to creosote brush has resulted in increased runoff and soil erosion (Archer et al., 2011). Successful conversion of these degraded sites back to grasslands has not generally been possible, and erosion rates are still high in the Sonoran and Chihuahuan deserts (Lane et al., 2000).

C. Numerous studies have evaluated the water budget of forest and scrub woodlands, and excellent reviews of the literature are provided by Bosch and Hewlett (1982) and Hibbert (1983). Relatively few studies have evaluated components of the water budget for mesquite-dominated rangelands (Carlson et al., 1990; Richardson et al., 1979). Heitschmidt and Dowhower (1991) and Carlson et al. (1990) evaluated the effect of honey mesquite removal on herbage response and water balance in Texas. They reported that annual above-ground net primary productivity increased significantly following removal of honey mesquite. The increase was the result of increased production of the species present at the time of control rather than a shift in species composition. Evapotranspiration accounted for 95 percent of rainfall from both sites. They reported no net change in evapotranspiration, runoff, or drainage associated with removal of honey mesquite. Increased annual above-ground net primary productivity of the treated site offset any water yield benefit that accrued through removal of honey mesquite trees. Dugas and Mayeux Jr. (1991) reported that both percentage and absolute difference in evapotranspiration between treated and untreated mesquite dominated rangelands were greatest under dry conditions and were essentially zero immediately after rainfall. While mesquite used substantial amounts of water, evapotranspiration from rangelands with mesquite that was killed was essentially the same due to the increased annual above-ground net primary productivity of other species following mesquite control. Two honey mesquite-dominated watersheds were evaluated in the Blackland Prairie region of Texas. Mesquite trees on one watershed were killed by hand application of one liter of diesel oil to the base of each tree. Removal of mesquite trees reduced evapotranspiration by 244 mm over a three-year period and increased surface water runoff by 10 percent compared to an untreated watershed. Removal of honey mesquite had minimal effect on soil water in the surface soil profile during the growing season. The mesquite-dominated community used considerably more water from the subsurface than did the herbaceous vegetation (Richardson et al., 1979). Weltz and Blackburn (1995) reported that no net change in evapotranspiration, runoff, or drainage would occur if mesquite-dominated shrub clusters were replaced by deep-rooted perennial grasses in south Texas. No change in soil erosion was reported when grassland interspaces were compared to the mesquite-dominated areas.

Figure D-13. Measuring snow fall and soil water equivalent at the USDA Agricultural Research Service Reynolds Creek Experimental Watershed near Boise, Idaho.



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Subpart E – Historical Erosion Prediction Technology

646.51 Contributors:

Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.
C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ, Tucson, AZ.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX.

Figure E-1. Concentrated flow paths in a greasewood plant community. Site is near Fallon, Nevada and illustrates vulnerability of rangelands to accelerated rill erosion that puts sites at risk of unsustainability if proper conservation is not applied.

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Figure E-2. Catastrophic soil erosion following wildfire in Pinyon and Juniper woodland in western Nevada.

646.52 Erosion Prediction Technology Over Time

A. The basis of mathematical equations used to estimate soil erosion can be traced to the work of Cook (1937), who identified three major variables: (1) the susceptibility of soil to erosion, (2) the potential erosivity of rainfall and runoff, and (3) the protection offered by vegetation. Zingg (1940) evaluated the effects of slope length and steepness on soil erosion and is often cited as the developer of the first erosion prediction equation. Smith (1941) included the influences of vegetation (C-factor) and supporting farming practices (i.e., P-factor representing the type, depth, frequency, and direction of mechanical disturbance). During the 1960s, factors for crop rotation, management, and rainfall for areas of the United States east of the 104th meridian were added to the existing "rational" equation for estimating soil loss from upland areas. More than 10,000 plot years of data, representing a variety of soil, crop, and management practices from multiple research locations across the Midwest were used to statistically derive a model called the Universal Soil Loss Equation (USLE). The USLE groups the physical and land management variables that influence soil erosion into six factors. Conversion factors for A, R, and K between U.S. customary units and SI units are given by Foster et al. (1981). The USLE is defined as:

$$A = R\ K\ L\ S\ C\ P$$

where:

A is a computed soil loss per unit area (metric tons • hectare^{-1})

R is a rainfall and runoff factor based on 22 years of climate records (megajoule • millimeter)/(hectare • hour •year)$^{-1}$

K is a soil erodibility factor based on a slope length of 22.1 m and a uniformly sloping 9 percent surface in continuously clean-tilled fallow ((metric tons • hectare • hour)/(hectare • megajoule • millimeter)$^{-1)}$;is a slope length factor determined as the ratio of soil loss from the field slope (unitless)

L is 1 when length is 22.1 m

S is a slope steepness factor determined as the ratio of soil loss from the field slope to that from a 9 percent slope under otherwise identical conditions (unitless)

C is a cover and management factor determined as the ratio of soil loss from an area with specified cover and management practices to that of continuous fallow (unitless)

P is a support practice factor determined as the ratio of soil loss with conservation practices to straight-row tillage parallel with the slope (unitless)

B. The USLE is a lumped empirical model that does not separate factors that influence soil erosion, such as plant growth, decomposition, infiltration, runoff, soil detachment, or soil transport. The USLE was designed to estimate sheet and rill erosion from hillslope areas. Davenport et al. (1998) developed a concept of soil erosion potential for Pinyon-Juniper woodlands based on the Universal Soil Loss Equation. They concluded that soil erosion was an interaction of site factors (soil texture, slope steepness, climate, etc.) and cover factors, with management being able to influence only canopy and ground cover–but the model had limitations. It was not designed to address soil deposition and channel or gully erosion within watersheds. The applicability, accuracy, and precision of the USLE on rangelands has been debated (Foster, 1982; Trieste and Gifford, 1980). In general, the USLE has been found to poorly estimate actual soil erosion on rangelands (Blackburn, 1980; Hart, 1984; Johnson et al., 1980 and 1984). The potential for improving rangeland estimates of soil erosion with the USLE is limited because of its restrictive structure, reliance on an empirical database rather than physical processes, and lack of temporal adjustments for factors of soil erodibility (K), cover (C), and management practice (P).

C. Advancements in erosion science since the release of the USLE in 1978 were incorporated into the Revised Universal Soil Loss Equation (RUSLE) (Renard et al., 1997). These refinements included techniques to address slopes over 20 percent, compound slopes, and time varying adjustments for soil erodibilities and cover for cropland. The RUSLE model maintains the simple linear form of the USLE as a product of six factors, but subfactors that reflect current knowledge of erosion science are used to calculate each factor.

D. The RUSLE model was compared to the USLE for three different soil-vegetation assemblages using a large rotating boom rainfall simulator on the USDA-ARS Walnut Gulch Experimental Watershed near Tombstone, AZ (Weltz et al., 1987). Three surface conditions were evaluated: natural vegetation, clipped plots where all standing vegetation was removed, and bare plots where all above-ground biomass and surface cover were removed. Both dry and wet soil moisture conditions were evaluated twice a year (spring and fall) over a four-year period. The regression coefficients of predicted versus observed erosion for the different model comparisons were used to evaluate the different models and indicate that the models were similar in predicting soil loss. In each instance, the slope of the line is less than unity, demonstrating that the predicted values of soil loss were substantially less than the measured values.

E. In a similar comparison, Renard and Simanton (1990) evaluated the USLE and RUSLE models at 17 sites in seven western states using the procedures described above. The differences in the comparisons between the two models involve the K, LS, and C factors. They concluded that RUSLE did a better job of estimating soil loss than USLE for naturally-vegetated and clipped plots, although both models were poorly correlated with actual soil loss. Both RUSLE and USLE gave improved soil loss estimates when the bare soil treatments were included in the analysis with the vegetated and clipped treatments. However, as in the previous study, the slope of the line was less than unity for the RUSLE model, demonstrating that the predicted values of soil loss were substantially less than the measured values.

F. Benkobi et al. (1994), working with rainfall simulation from 1 m^2 plots on a sagebrush-grassland area in Idaho, reported that RUSLE significantly underestimated soil erosion; and the slope of the line was near zero indicating a very poor relationship between measured and predicted soil loss. They replaced the surface cover subfactor (SC) with a multiple regression equation based on litter and rock cover in an attempt to improve prediction of soil loss. This new equation did not substantially improve the estimate of soil erosion, and both versions of RUSLE significantly underestimated soil erosion. Soil loss was most sensitive to changes in values of the slope steepness and slope length factors.

G. Spaeth et al. (2003) reported that USLE over-predicted soil loss and that RUSLE underestimated soil erosion, when compared to rainfall simulation plots from a diverse set of rangeland vegetation types (8 states, 22 sites, 132 plots). They recommended, given the structure of both the USLE and RUSLE, that a new system is necessary to estimate soil loss on rangelands, that is developed using rangeland soil and a modeling structure that reflected the differences between cropland and rangelands.

H. In 1985, the USDA-ARS initiated the Water Erosion Prediction Project (WEPP). The WEPP model was released in 1995, representing the assemblage of state-of-the-art process-based erosion modeling technologies (Flanagan and Nearing, 1995; Tiscareno-Lopez et al., 1993 and 1994). This model is based on fundamentals of infiltration, hydrology, plant science, hydraulics, and erosion mechanics (Nearing et al., 1989). As a process-based model, WEPP has the advantages over empirical models for its capabilities to estimate spatial and temporal distributions of net soil loss and to extrapolate to a broad range of conditions (Nearing et al., 1999). During 1987 to 1988, the WEPP team collected a large set of erosion data from rangelands across the western U.S. for parameterization of erosion and hydrology factors.

I. Current erosion models like WEPP predict that erosion takes place on hillslopes where areas of concentrated flow paths or rills are present, and that these areas do not evolve during a rainfall event (Favis-Mortlock, 1998). It is assumed in WEPP that areas of concentrated flow are uniformly spaced at one per meter and have a uniform rectangular cross-sectional width (Gilley et al., 1990). The model calculates the number of potential areas of concentrated flow, based on the distance between plants. Concentrated flow equations used in WEPP are based on the detachment-transport coupling concept that was developed by Foster and Meyer (1972). Huang et al. (1996) conducted a field study focused on how concentrated flow areas function and concluded that the existing coupling concept was inadequate. Various studies have found that erosion processes in areas of concentrated flow are more complex than that described in WEPP (Al-Hamdan et al., 2011, 2012a, and 2012b; Lei et al., 1998).

J. Watters et al. (1996) developed a Site Conservation Threshold concept using WEPP. They found that estimated basal cover was the best single indicator of site stability of indicators evaluated in their study. Basal cover varied with season and weather conditions but was much less sensitive to these short term seasonal climatic variations and grazing pressures than was standing biomass or litter.

K. WEPP is the basis for the Rangeland Hydrology and Erosion Model (RHEM) (Wei et al., 2009). Contrary to the WEPP paradigm, areas of concentrated flow are dynamic and do evolve during a rainfall event and do not have a uniform cross-sectional depth/width ratio. When rainfall intensity exceeds infiltration capacity and runoff occurs, it may concentrate in flow paths where accelerated soil erosion may occur. If runoff velocity exceeds the soil’s shear strength, then down cutting and widening of the flow path may occur, resulting in accelerated soil erosion. When and where this occurs is poorly understood, currently. The WEPP model is limited in application to rangelands because many of the model’s concepts and erosion equations were developed from experiments on croplands. It has not been widely accepted and utilized by rangeland managers for addressing conservation planning.

Figure E-3. Saline and sodic badlands in central Utah illustrating natural geologic soil erosion processes. Uniformly spaced rills and concentrated flow channels are developed as runoff is routed down the steep side slope from the mesa top to the valley floor.

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Subpart F – Current Rangeland Erosion Prediction Technology

646.61 Contributors:

Mark A. Weltz, Research Leader, Rangeland Hydrologist; USDA-ARS, Reno, NV.
S. Kossi, Nouwakpo, Soil Scientist, University of Nevada Reno, Reno, NV.
C. Jason Williams, Research Hydrologist, USDA-ARS, Tucson, AZ.
Fred B. Pierson, Research Leader, USDA-ARS, Boise, ID.
Ken E. Spaeth, Rangeland Management Specialist, USDA-NRCS, Fort Worth, TX.
Mark A. Nearing, Research Agricultural Engineer, USDA-ARS, Tucson, AZ.
Mariano Hernandez, Hydrologist, USDA-ARS, Tucson, AZ, Tucson, AZ.
Osama Z. Al-Hamdan, Assistant Professor, Texas A&M Kingsville, Kingsville, TX.

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646.62 Current Rangeland Erosion Prediction Technology

A. Rangeland Hydrology and Erosion Model (RHEM) V1.0 is based on state-of-the-art technology from the Water Erosion Prediction Project (WEPP) (Flanagan and Nearing, 1995). However, the basic equations in the WEPP model are based on experimental data from croplands. While many of the fundamental hydrologic and erosion processes can be expressed in a common way on both crop and rangelands, there were several aspects of the WEPP model that are not optimum for rangeland application and hence were modified, dropped, or replaced in RHEM (Nearing et al., 2011b).

B. The RHEM V1.0 model was initially developed for undisturbed rangelands, where the impact of concentrated flow erosion is limited, and most soil loss occurs by rain splash and sheet erosion processes. This model included a new splash and sheet erosion equation developed by Wei et al. (2009), based on rainfall simulation data collected on rangeland plots from the WEPP and Interagency Rangeland Water Erosion Team and National Rangeland Study Team (IRWET and NRST, 1998) projects. These studies covered 49 rangeland sites distributed across 15 western states. Two methods of computing the peak discharge were used in the WEPP model: a semi-analytical solution of the kinematic wave model (Stone et al., 1995), and an approximation of the kinematic wave model that was developed to reduce run times on computers of two decades ago. The first method is used when WEPP is run in a single event mode, while the second is used when WEPP is run in a continuous simulation mode. In RHEM V1.0, the semi-analytical solution of the kinematic wave model was used instead of the approximate method for calculating peak runoff.

C. The steady state cropland-based shear stress approach from WEPP for modeling concentrated flow erosion was adapted and used in RHEM V1.0. Consequently, it was not possible to quantify within-storm sediment dynamics (Bulygina et al., 2007). That is, a steady state model does not allow the user to represent changing parameter values, such as soil erodibility, during the rainfall event. It also does not provide information on peak sediment discharge or the sediment load pattern within the storm, both of which can be useful for assessing potential pollution loadings from sediment fluxes into water courses and identifying sediment sources for designing appropriate management alternatives that reduce sediment losses (Kalin et al., 2004). The Yalin sediment transport capacity equation (Yalin, 1963) and the shear stress partitioning detachment and deposition concepts developed by Foster (1982), which distributes the transport capacity among various particle types, were used in RHEM V1.0. This configuration of RHEM V1.0 provided reasonable erosion rates on undisturbed rangeland soils (Nearing et al., 2011a).

D. Weltz and Spaeth (2012) used RHEM to assess the impact of ecological sites invaded with Ash Juniper (*Juniperus ashei* J. Buchholz) on the Edwards Plateau near Johnson City, Texas, USA. They determined that applying conservation to return the invaded site to reference conditions could reduce soil loss by up to six-fold, depending on the runoff return period being evaluated. Weltz et al. (2014) used RHEM V1.0 to estimate the impact of transitioning from one state to another for two different plant community types. For a typical Wyoming sagebrush site near Austin, NV, water-induced soil loss was 2.4 to 3 times lower than it was on a burned site previously dominated by cheatgrass. In addition to greater soil loss, the burned cheatgrass site had 1.2 to 1.6 times more runoff during intense summer thunderstorms. Runoff and soil loss from the cheatgrass-dominated site was estimated to be slightly elevated over the Wyoming sagebrush site at current potential. In a mountain sagebrush site that had been encroached by Pinyon and Juniper trees, the type and distribution of canopy and ground cover are altered relative to the Current Potential State. In the Current Potential State, more uniformly distributed vegetation makes concentrated flows unlikely and minimizes soil loss and runoff. When Pinyon and Juniper trees invade, and canopy closure advances, the understory cover (grasses and forbs) declines (Miller et al., 2000). This further increases the probability of concentrated flows in the connected bare spaces and results in accelerated runoff and soil erosion (Pierson et al., 2011). After a wildfire, runoff may increase on the order of 4 to 10 times, and soil loss can increase 4 times, increasing the probability of downstream floods. These results are consistent with those reported by others that sites encroached by Pinyon and Juniper trees, both pre- and post-fire, have increased potential for accelerated soil erosion (Pierson et al., 2013; Pierson et al., 2011).

E. The RHEM V1.0 tool was used to estimate runoff and erosion at the hillslope scale for over 10,000 NRI sample points in 17 western states on non-Federal rangelands (RCA, 2011; USDA-NRCS, 2011). The national average annual erosion rate on non-Federal rangeland was estimated to be 1.4 ton ha^{-1} $year^{-1}$. Nationally, 20 percent of non-Federal rangelands generate more than 50 percent of the average annual soil loss. Over 29.2 x 10^6 ha (18 percent) of the non-Federal rangelands might benefit from treatment to reduce soil loss to below 2.2 ton ha^{-1} $year^{-1}$. National average annual erosion rates combine areas with low and accelerated soil erosion. Evaluating data in this manner can misrepresent the magnitude of the soil erosion problem on rangelands. Between 23 percent and 29 percent of the Nation's rangelands are vulnerable to accelerated soil loss (soil erosion > 2.2 ton ha^{-1} $event^{-1}$) if assessed as a function of vulnerability to a runoff event greater than a 25-year storm intensity. Hernandez et al. (2013) reported that RHEM could effectively assess the influence of foliar, ground cover, plant life form, soils, and topography on current soil erosion rates using data from USDA National Resources Inventory (NRI on-site data collection) in southern Arizona. Results suggested that the model could be further improved with additional measured experimental data on infiltration, runoff, and soil erosion within key ecological sites. These data could better quantify model parameters to reflect ecosystem changes and the risk of crossing interdependent biotic and abiotic thresholds. The results of Hernandez et al. (2013) from southern Arizona and national assessment of soil erosion (USDA-NRCS, 2011) indicated that RHEM can be used to assess the relative erosion rates on rangelands and can be used to assess the potential benefits of conservation and land management practices.

F. Belnap et al. (2013) evaluated the RHEM model for its effectiveness in estimating runoff and soil loss on sites in Utah dominated by biological soil crusts. They reported that RHEM, once calibrated, predicted that sites with the least development of biological soil crusts had the highest amount of soil loss, and that erosion potential increases by a factor of 10 as slope gradients increase from 0 percent to 10 percent. The model results indicated that as biological soil crusts increased, soil loss decreased. RHEM results also illustrated how soil erosion potential rises as antecedent soil moisture levels rise.

G. RHEM V2.0 is fundamentally a different model than previous versions because of changes to critical hydrologic and soil erosion prediction equations. This includes the integration of the dynamic formulation of a stream-powered sediment continuity equation (Bennett, 1974), instead of the steady-state shear stress approach for modeling concentrated flow erosion. Nearing et al. (1997) and others found that the best predictor for sediment load was stream power instead of shear stress.

H. Al-Hamdan et al. (2012) used concentrated flow simulations on disturbed and undisturbed rangeland to estimate soil erodibility, as well as to evaluate the performance of linear and power law equations that describe the relationship between erosion rate and several hydraulic parameters. They showed that stream power provided the best linear function to describe the detachment rate on disturbed rangeland sites. Based on these findings, stream power was integrated into the empirical equation developed by Nearing et al. (1997) to calculate the sediment transport capacity into RHEM V2.0. In subsequent versions of RHEM leading to V2.3, major changes have been implemented to the parameter estimation equations for describing and quantifying disturbed rangelands, testing the model efficiency in predicting larger soil loss events, and not to the simulation model itself.

I. The enhanced RHEM V2.3 model discussed herein provides major advantages over existing erosion model prediction technology, including RHEM V1.0. The updated model is capable of capturing the influence of different plant types, disturbances such as fire and climate change, and rangeland management practices on important hydrological and erosion processes acting on rangelands. The model has undergone a critical review and expansion of capabilities. The most significant differences between this model and the original include:

(1) The model uses a dynamic solution of the sediment continuity equation based on kinematic wave routing of runoff, and the integration of the newly developed splash and sheet source term equation, and stream power for predicting sediment transport of concentrated flow erosion (Nearing et al., 1997).

(2) It integrates the approach for estimating the splash and sheet erodibility coefficients formulated by Al-Hamdan et al. (2015), who developed equations to predict the differences of erodibility before and after disturbance across a wide range of soil texture classes and vegetation cover types.

(3) The model integrates the method for predicting concentrated flow erosion based on the work by Al-Hamdan et al. (2013), who developed a dynamic erodibility approach for modeling concentrated flow erosion (for sites with relatively immediate disturbance, such as fire).

(4) The model includes a user-friendly web-based interface to allow users to simplify the use of RHEM, manage scenarios, centralize scenario results, compare scenario results, and provide tabular and graphical results (Hernandez et al., 2015).

J. Cadaret et al. (2016a and 2016b) evaluated the highly erosive saline soils of the Mancos Shale formation in the Price-San Rafael River Basin in Utah, USA using a Walnut Gulch rainfall simulator. The data were used to evaluate the RHEM's ability to provide unbiased estimates of runoff, total dissolved solids (TDS), and sediment load. Simulated inter-plot variability was highest at sites with the greatest slope angles, vegetation cover, and sediment loads. The calibrated surface erosion parameters in RHEM (K_{ss}, K_ω) were substantially greater than any published in prior studies from non-saline environments. The spatial distribution of vegetated canopy cover was quantified using photogrammetric modeling and landscape pattern metrics. Plot-averaged sediment and total dissolved solids (TDS) had a positive linear relationship. Total dissolved solids in runoff derived from these upland rangelands in central Utah was successfully predicted with RHEM. As the patches of vegetation became more contiguous and the tortuosity of the bare soil area increased, RHEM over-predicted sediment output, suggesting that vegetation-driven spatial heterogeneity influenced erosion in a way that is not captured by the model.

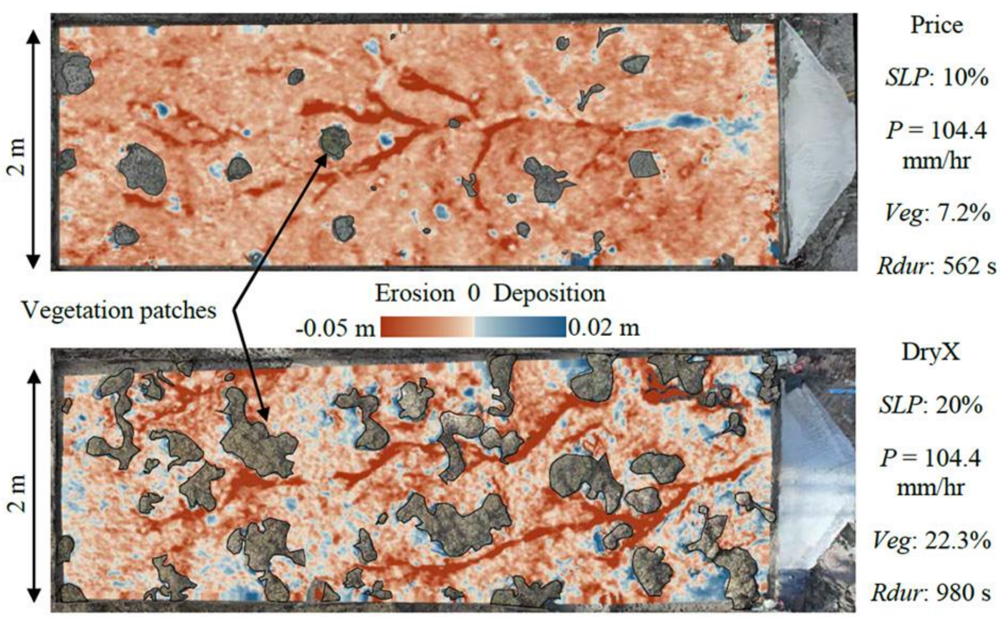
K. The ability to assess the impact of management actions on soil and water resources is crucial to sustainable land management. Hillslope scale hydrologic processes predicted with RHEM have been applied successfully to illustrate the influence of plant and soil characteristics on soil erosion and hydrologic function and to predict runoff and erosion rates for refinement and development of Ecological Site Descriptions (Williams et al., 2016). This model has been used to characterize rangeland conditions based on a probabilistic approach, subject to the presence of a set of soil erosion thresholds (Hernandez et al., 2015). Runoff, sediment, and microtopography data were collected during a series of rainfall simulation experiments aimed at evaluating the effectiveness of RHEM to evaluate the success of revegetation on a shrub-dominated post-construction hillslope (Nouwakpo et al., 2016). Experimental results showed that a mix of the shrub species rabbitbrush (*Ericameria nauseosa* a (pallas ex Pursh) G.L. Nesom & Baird) and the invasive annual grass cheatgrass (*Bromus tectorum L.*) were more effective at reducing runoff and soil loss than rabbitbrush alone. Plots with more than 45 percent of litter cover produced as much as 2.1 times less runoff and 16 times less sediments than those with less than 15 percent of litter. Analysis of these 3D data highlighted the central role of concentrated flow erosion in sediment delivery on rangeland hillslopes (see figures F-1 and F-2). These studies demonstrate that RHEM can be effectively used by land managers and project managers to estimate soil erosion as a function of precipitation events on construction sites that have been revegetated to rangeland plant communities. The hydrologic and soil erosion impact of land management practices can be estimated with RHEM, such as grazing, prescribed fire, or mechanical treatments that can alter the species present, density, and spatial arrangement of vegetation on rangelands.

L. Title 190 Part 646, "Rangeland Hydrology and Soil Erosion Processes: A guide for Conservation Planning with the Rangeland Hydrology and Erosion Model (RHEM)," is available for download. This is The RHEM Guide and describes the science and equations for RHEM and provides a tutorial to use the model. In addition, it has guides on use and interpretation of model results. The model can be used to inform and enhance ecological state and transition models and enhance Ecological Site Descriptions (ESDs). In the new NRCS National Range and Pasture Handbook – Title 190, Part 645 RHEM – is now approved for use by NRCS when conducting grazing management plans and resource inventory.

> *Rangeland: When planning on rangeland, conduct interpreting indicators of RHA on each field in the grazing plan. If the preponderance of evidence summary indicates any category except "None to Slight" for Soil/Site Stability and/or Hydrologic Function, then RHEM will be used to further determine the extent and risks of potential erosion based on long-term average and for 2-, 5-, 10-, 25-, 50-, and 100-year rainfall events. If the planner identifies that the RHA "Biotic Integrity" may be affecting hydrologic function and/or soil and site stability, then RHEM can be used to evaluate the effects of plant foliar cover dynamics on hydrology and erosion.*

Figure F-1. Naturally occurring erosion in Badlands of South Dakota on marine-derived soils.

Figure F-2. Examples of maps of elevation differences showing plot-wide differences (top) and differences in delineated channel network (bottom). Gray polygons mark the outlines of plant canopies demonstrating how plant coppice dunes alter flow path trajectory and reduce flow velocity and sediment transport capacity. For this example, plot, slope = 16.8 percent, plant canopy cover = 5.0 percent.



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