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RESEARCH ARTICLE

Regional patterns of postwildfire streamflow response in the Western United States: The importance of scale-specific connectivity

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Abstract

Wildfires can impact streamflow by modifying net precipitation, infiltration, evapotranspiration, snowmelt, and hillslope run-off pathways. Regional differences in fire trends and postwildfire streamflow responses across the conterminous United States have spurred concerns about the impact on streamflow in forests that serve as water resource areas. This is notably the case for the Western United States, where fire activity and burn severity have increased in conjunction with climate change and increased forest density due to human fire suppression. In this review, we discuss the effects of wildfire on hydrological processes with a special focus on regional differences in postwildfire streamflow responses in forests. Postwildfire peak flows and annual water yields are generally higher in regions with a Mediterranean or semi-arid climate (Southern California and the Southwest) compared to the highlands (Rocky Mountains and the Pacific Northwest), where fire-induced changes in hydraulic connectivity along the hillslope results in the delivery of more water, more rapidly to streams. No clear streamflow response patterns have been identified in the humid subtropical Southeastern United States, where most fires are prescribed fires with a low burn severity, and more research is needed in that region. Improved assessment of postwildfire streamflow relies on quantitative spatial knowledge of landscape variables such as prestorm soil moisture, burn severity and correlations with soil surface sealing, water repellency, and ash deposition. The latest studies furthermore emphasize that understanding the effects of hydrological processes on postwildfire dynamic hydraulic connectivity, notably at the hillslope and watershed scales, and the relationship between overlapping disturbances including those other than wildfire is necessary for the development of risk assessment tools.

KEYWORDS
climate change, hydraulic connectivity, streamflow, watershed, wildfire, wildland–urban interface

1 | INTRODUCTION

Wildfires are natural disturbances vital to the health of many terrestrial and aquatic ecosystems (Brown & Smith, 2000; Conway, Nadeau, & Piest, 2010; Flitcroft et al., 2015). However, forest fires in the United States and elsewhere increasingly represent a threat to water supplies due to longer wildfire seasons, increasing annual area burned, and higher fire severity associated with forest densification (Dennison, Brewer, Arnold, & Moritz, 2014), persistent drought (Borsa, Agnew, & Cayan, 2014; Diffenbaugh, Swain, & Touma, 2015), climate change (Calder, Parker, Stopka, Jiménez-Moreno, & Shuman, 2015; Rocca, Brown, MacDonald, & Carrico, 2014; Stavros, Abatzoglou, Larkin, McKenzie, & Steel, 2014), and a progressively populated wildland–urban interface (Radeloff et al., 2005). Forest fires often lead to increased peak flows and erosion, with subsequent impacts on water quality (Moody, Shakesby, Robichaud, Cannon, & Martin, 2013) and elevated risk of mass movements (Wondzell & King, 2003). This is particularly concerning given that approximately one half of the surface freshwater supply in the conterminous United States (CONUS) originates in forested watersheds (Brown et al., 2008; Sun et al., 2015). In the Western U.S., forests provide ~65% of the water supply (Furniss et al., 2010), while in the Southeastern U.S. they provide ~35% of the water supply (Caldwell et al., 2014). National Forests alone account for ~18% of the total U.S. surface freshwater supply and for more than 50% of the supply in the Western U.S. (Brown et al., 2008).

As wildfire severity increases, the impacts on surface water supply may become increasingly significant and longer lasting. Additionally, effects may be transmitted longer distances downstream from burned headwater catchments, creating challenges for drinking water treatment in downstream communities (Emelko et al., 2016; Martin, 2016; Shakesby & Doerr, 2006; Silins et al., 2014; Smith, Sheridan, Lane, Nyman, & Haydon, 2011). In particular, community drinking water suppliers must understand the impacts of wildfire on annual water yields (cumulative discharge), low flows, high (peak) flows, and timing of water availability to continue to meet the demand for adequate quantities of potable water (Emelko, Silins, Bladon, & Stone, 2011). However, the impacts on these aspects of water supply are difficult to predict because hydrologic responses are highly variable, depending on climate, topography, geology, vegetation, and characteristics of the wildfire (Campbell, Baker, Ffolliott, Larson, & Avery, 1977; Moody et al., 2013; Wagenbrenner, 2013). This can create substantial challenges for managing water supplies for downstream communities and for aquatic ecosystem health (Bladon, Emelko, Silins, & Stone, 2014; Emelko et al., 2011; Jung, Hogue, Rademacher, & Meixner, 2009).

The magnitude and longevity of hydrological impacts of wildfire vary by region, with a clear distinction between the Eastern and Western CONUS, principally due to differences in occurrence and severity of wildfire, climate, and vegetation composition. Even within the Western United States, there may be substantial variability in the postfire streamflow response, depending on prewildfire conditions, fire severity, postwildfire climate, and local catchment characteristics (Holden, Luce, Crimmins, & Morgan, 2012; Hurteau, Bradford, Fule, Taylor, & Martin, 2014; Moody & Martin, 2009; Neary, Ryan, & DeBano, 2005). Postwildfire floods with severe erosion have been observed in semi-arid Arizona (Wagenbrenner, 2013) and New Mexico (Moody & Martin, 2001), and to a lesser degree in Mediterranean California and other parts of the West (Hallema, Sun, et al., 2016; Jung et al., 2009), but such occurrences are not common in the humid subtropical Southeast (Hallema, Sun, et al., 2016). Regional variability in fire trends results in a patchwork of areas with a potential to either exclude or promote fire (Parks et al., 2015). Corresponding regional differences in fire impacts demonstrate that the recent challenge in comparative risk assessment is to link fire processes, hydrological processes, and process interactions to the corresponding environmental characteristics (Bladon et al., 2014). Notwithstanding progress in quantitative analysis reported in review papers by Swanson (1981), Neary et al. (2005), and Moody et al. (2013), there are important limitations that complicate the assessment and prediction of postwildfire run-off (Moody et al., 2013). Data on transient burned area responses is scarce, and assessment is further complicated by the episodic nature and destructive power of floods, the non-linear nature of hydrological processes related to run-off response thresholds (Germer & Braun, 2011), and the variety of methods employed to measure hydrologic impacts (Shakesby & Doerr, 2006).

To facilitate improved management of water supplies, it is critical understand regional differences in wildfire impacts on streamflow responses. One objective of this paper is to provide an overview of the effects of wildfire on hydrologic responses (e.g., interception, evapotranspiration, infiltration, and run-off generation), in particular, the impact of surface processes on hydraulic connectivity along the hillslope, which influences the delivery of water to the stream network (as opposed to hydrologic connectivity, which relates to the catchment scale and greater; Pringle, 2003). Second, we review known regional differences in streamflow responses at the watershed scale, such as water yields (cumulative discharge), low flows, and high (peak) flows. Focus is on the Western CONUS where streamflow responses to wildfire are more substantial and well documented. Where earlier review papers assume an understanding of regional forest ecology, the intent of our review is to introduce concepts of postwildfire hydrology and hydraulic connectivity to those who are not necessarily familiar with wildfire responses in the Western CONUS.

2 | WILDFIRE EFFECTS ON HYDROLOGICAL PROCESSES

Wildfires affect hydrological processes near the soil surface and alter streamflow output from watersheds (Figure 1). Therefore, a characterization of postfire hydrology relies on understanding the combined effect of hydrological processes on flow distribution within the watershed. General responses to severe wildfires include alteration of the composition and structure of vegetation, leading to reduced interception rates and more water reaching the forest floor (Winkler et al., 2010). In cases where infiltration rates are unaffected by the fire, greater net precipitation leads to more soil moisture, greater water availability for transpiration, and more run-off, ultimately resulting in more streamflow (Moody & Martin, 2001; Moody et al., 2013). When infiltration rates are reduced due to soil surface sealing or water repellency, the rate of overland flow can increase and more water is delivered more rapidly to the stream (DeBano, 2000a; Doerr, Woods, Martin, and Casimiro, 2009). In this section, we discuss wildfire effects on hydrological processes and attempt to define the connection with surface hydraulic connectivity.

2.1 | Effects on interception and evapotranspiration

2.1.1 | Interception

Severe wildfire generally results in a loss of canopy cover, which decreases canopy interception storage capacity and increases the net precipitation (throughfall) reaching the soil surface (Figure 1; Williams, Pierson, Robichaud, & Boll, 2014). The amount of leaf loss depends on the duration of heat exposure, fire temperature, and bark thickness and varies depending on weather, fuel moisture, dominant vegetative species, and succession stage (Zwolinski, 1990). The amount of foliar

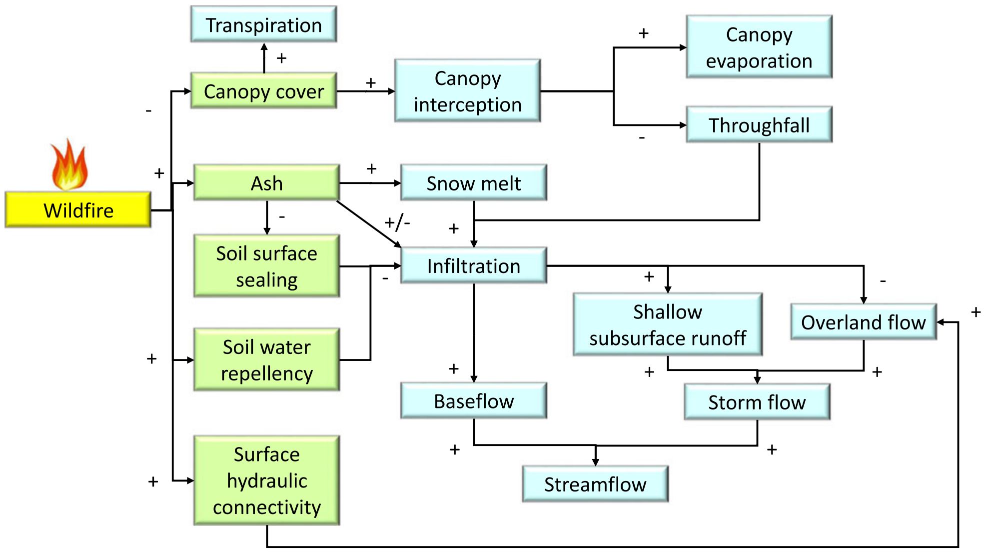


FIGURE 1 Conceptual model of postwildfire hydrology in forest-covered watersheds. The + and − symbols indicate positive and negative forcing, respectively, in the direction of each arrow

volume consumed by fire furthermore affects the portion of the soil surface exposed to rainfall impact (Winkler et al., 2010), and therefore, a decrease in interception (increase in net precipitation) can be observed immediately after a fire (Helvey & Patric, 1965; National Research Council, 2008). An increase in water availability at the soil surface, combined with wildfire impacts on infiltration and run-off generation, can modify the hydrologic response of a watershed such that postfire storms with recurrence intervals as short as 2 years can produce floods normally observed for less frequent storms (Kunze & Stednick, 2006; Moody & Martin, 2001). Therefore, interception, postfire rainfall distribution, and changes in net infiltration are essential factors affecting the hydrologic response at the watershed scale (Moody et al., 2013).

2.1.2 | Evapotranspiration

Evapotranspiration also influences the catchment-scale water balance and streamflow with variable responses depending on dominant tree species, leaf cover, and basal area (Bosch and Hewlett, 1982; McLaughlin, Kaplan, & Cohen, 2013; Sun et al., 2011). Changes in forest cover due to wildfire can influence radiant energy partitioning between latent and sensible heat fluxes, affecting the amount of soil water content and water available for groundwater recharge and runoff (Obrist, DeLucia, & Arnone, 2003). Postwildfire change in evapotranspiration (latent heat flux) has been shown to be a function of burn severity, which also impacts the amount of leaf loss (Figure 1), surface shading (Driscoll, Carter, & Ohlen, 2004), and albedo (Dore et al., 2012). For example, Montes-Helu et al. (2009) reported a 30% higher albedo at a burned site versus an unburned site in a forest in warm temperate Northern Arizona, which they associated with a 30% decrease in net efficiency (net radiation/total radiation). Increased albedo was the result of reduced leaf area and increased bare soil and light-coloured woody debris (Dore et al., 2008). The loss of leaf area and increase in albedo diminished the net radiation and caused a 20% reduction in annual evapotranspiration following this high-severity fire (Dore et al., 2012). In areas with regular snowfall in the West, wildfire can have a differential effect on albedo in summer and in winter. Gleason, Nolin, and Roth (2013) observed a two times higher snow ablation rate in a burned mixed conifer forest in the Oregon High Cascades associated with a 40% lower albedo of the hillslope surface, caused by the deposition of pyrogenic carbon particles and burned woody debris shed from charred trees. This deposition darkened the snow surface, and the combination with more solar radiation resulted in accelerated snowmelt in the spring. Many studies do not mention the effect of charred trees on albedo but attribute the albedo entirely to the colour and amount of ash on the surface. Black ash resulting from incomplete combustion (Badía & Martí, 2003) increases heat absorption and soil temperature by decreasing the overall albedo of the soil surface, enhancing evapotranspiration (Bodí et al., 2014; Massman, Frank, & Reisch, 2008). Surface evaporation from the blackened soil may increase temporarily during the period before canopy leafout and understorey development in the spring (Boerner, 2006; Iverson & Hutchinson, 2002).

2.2 | Effects on infiltration and run-off generation

During a fire, litter and duff are removed by the process of combustion, leaving behind ash and char at the surface. The ash is often wettable (Kinner & Moody, 2010), and the downward heat gradient can cause a part of the predominantly mineral soil to become (more) water repellent than before the fire (Krammes and Osborn, 1969; Finley & Glenn, 2010). Similar effects were observed for char, which results from incomplete combustion and is often present in deeper fuel beds and wetter sites (Ice, Neary, & Adams, 2004). Wildfire-induced changes that trigger infiltration excess run-off and saturation excess run-off are enhanced by soil water repellency, increased exposure of the soil surface, reduced surface water storage, soil surface sealing, and decreased canopy interception (DeBano, 2000a; Larsen et al., 2009; Moody et al., 2013). An overview of changes in hydraulic properties associated with ash and water repellency is in Table 1, with the corresponding study locations in Figure 2.

TABLE 1 Infiltration characteristics of burned sites

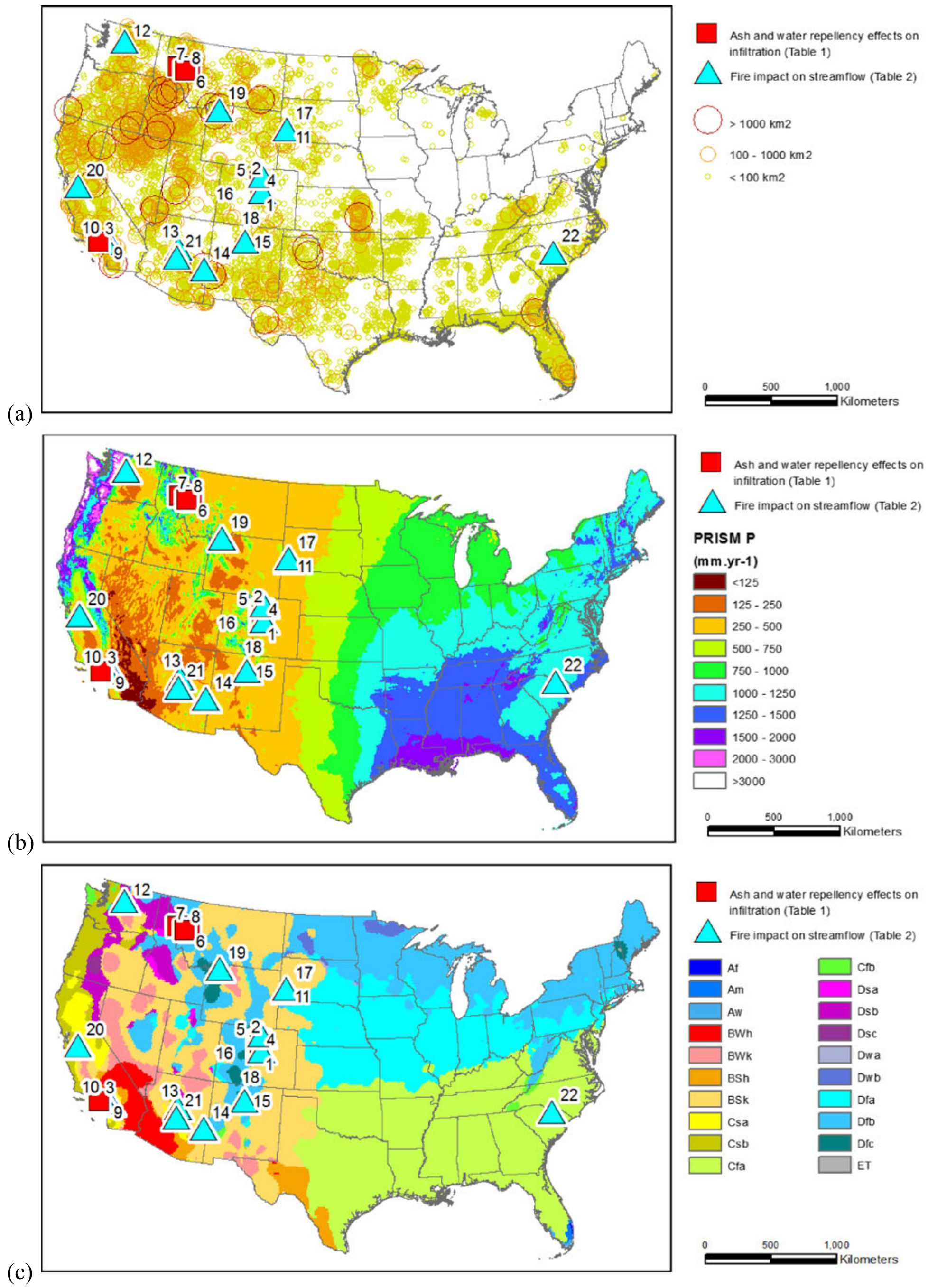
Location	Vegetative species	Experiment	Soil type and D_{50} (mm)	Fire name	Burn severity	Ash thickness (mm)	Ash colour, D_{50} (μm)	Run-off ratio (%)	Unburned soil K_f (mm/hr)	Postfire K_f (mm/hr)	Postfire S (cm/s$^{0.5}$)	Postfire WDPT (s)	Reference
Front Range, C. Colorado	Ponderosa pine	In situ and laboratory MDI test	Sand 0.76–3.7	2000 High Meadows	High[a]	–	–, 88	–	33–108[b]	33–4680[b] (soil)	0.18–0.20[b] (soil)	2.7–4.9[b] (soil)	Moody et al. (2009)
Front Range, C. Colorado			Sand 0.088	2003 Overland	High[a]					162–1908[b] (ash)	0.18–0.20[b] (ash)	0.6–3.6[b] (ash) 4.9[b] (soil)	
Verdugo Mountains, S. California	Chaparral		Sand 0.55	2005 Harvard	Mixed low to moderate[a]					47–277[b] (soil)	0.11–0.078[b] (soil)	5.6–72[b] (soil)	
Front Range, C. Colorado	Ponderosa pine, Douglas fir	Top soil reconstruction, laboratory rainfall simulation	Granular sandy loam with 3–4% OM	2002 Hayman and 2002 Schoonover	High	0 (control)	Mixed black and white, –	35	–	–	–	–	Larsen et al. (2009)
						5		17					
						12		7.2					
Fourmile Canyon, Front Range, C. Colorado	Ponderosa pine, juniper, Douglas fir	In situ falling head infiltration test	Sandy loam with 6% OM	2010 Fourmile Canyon	Mixed low to high	18	Black, –	–	124[b]	8.6[b] (ash) 274[b] (soil)	–	–	Ebel et al. (2012)
Lubrecht EF, W. Montana	Douglas fir, ponderosa pine	Ash addition, in situ rainfall simulation	Sandy loam 0.054	2006 Rx	High	5	Mixed black and white, 49	49	40	24 (soil + ash)	–	–	Woods and Balfour (2010)
						25		28	58	63 (soil + ash)			
						50		17	62	79 (soil + ash)			
Lolo NF, W. Montana	Ponderosa pine, Douglas fir, lodgepole pine	In situ constant head infiltration test	Reconstructed with 7 mm pebbles, sorted	2005 I-90 Complex	High	30	White, 14–136	–	–	165 (ash)	–	–	Gabet and Bookter (2011)
W. Montana	Ponderosa pine, Lodgepole pine, subalpine fir, white bark pine	In situ MDI test	Silt loam with gravel	2011 West Riverside, 2011 Avalanche Butte	High	38–53	Light grey to grey, –	–	–	12–14 (ash)	–	296–317	Balfour et al. (2014)

Note. NF = national forest; EF = experimental forest; D_{50} = median particle diameter; OM = organic matter; Rx = prescribed burn; K_f = near-saturated hydraulic conductivity; S = sorptivity; MDI = mini-disk infiltrometer; WDPT = water drop penetration time.

Note. Burn severity refers to the dominant burn severity class. Unreported variables are marked with a dash.

[a]Estimated a posteriori (not provided in reference).

[b]Unit conversion applied.



	Reference	Location	Climate type	Fire	Table
1	Moody *et al.* (2009)	Front Range, CO	Dfb	2000 High Meadows	I
2	Moody *et al.* (2009)	Front Range, CO	Dfb	2003 Overland	I
3	Moody *et al.* (2009)	Verdugo Mountains, CA	Csa	2005 Harvard	I
4	Larsen *et al.* (2009)	Front Range, CO	BSk	2002 Hayman, 2002 Schoonover	I
5	Ebel *et al.* (2012)	Front Range, CO	Dfb	2010 Foumile	I
6	Woods and Balfour (2010)	Lubrecht EF, MT	Dfb	2006 Rx	I
7	Gabet and Bookter (2011)	Lolo National Forest, MT	Dfb	2005 I-90	I
8	Balfour *et al.* (2014)	W Montana, MT	Dfb	2011 West Riverside, 2011 Avalanche Butte	I
9	Chen *et al.* (2013); Wohlgemuth (2016)	San Dimas EF, CA	Csa	1953 Barrett, 1960 Unnamed	II
10	Jung *et al.* (2013); Kinoshita and Hogue (2015)	San Bernardino NF, CA	Csa	2003 Old	II
11	Driscoll *et al.* (2004)	Black Hills, SD	Dfa	1988 Galena	II
12	Helvey (1980); Seibert *et al.* (2010)	Entiat EF, WA	Dsb	1970 Entiat	II
13	Campbell (1977)	Coconino NF, AZ	Csa	1972 Rattle	II
14	Wagenbrenner (2013)	Apache-Sitgreaves NF, AZ	Csb	2011 Wallow	II
15	Moody and Martin (2001b)	Santa Fe NF, NM	Dfb	2000 Cerro Grande	II
16	Moody and Martin (2001b)	Front Range, CO	BSk	1996 Buffalo Creek	II
17	Moody and Martin (2001b)	Black Hills, SD	Dfa	1988 Galena	II
18	Kunze and Stednick (2006)	Front Range, CO	Cfd	2000 Bobcat	II
19	Troendel (1996)	Shoshone NF, WY	Dfc	1988 Clover-Mist	II
20	Hallema *et al.* (2016b)	Diablo Range, CA	Csa	2003 Deer Park	II
21	Hallema *et al.* (2016b)	Tonto NF, AZ	Csa	2004 Willow	II
22	Hallema *et al.* (2016b)	Carolina Sandhills, SC	Cfa	Rx (prescribed fire)	II

FIGURE 2 Locations of studies related to infiltration on burned sites (corresponding with Table 1) and postfire streamflow (Table 2) plotted against (a) wildland fire occurrence and fire size between 1980 and 2013 (Monitoring Trends in Burn Severity; Eidenshink et al., 2007), (b) annual precipitation between 1981 and 2010 (PRISM Climate Group, retrieved 01/11/2014), and (c) climate type (Peel et al., 2007). NF = national forest; EF = experimental forest

2.2.1 | Soil water repellency

Soil water repellency is influenced by prefire soil texture, litter cover, soil moisture, soil organic matter, fire temperature, and wildfire residence time (DeBano, 2000a; Doerr, Shakesby, & Walsh, 2000). It is also often ascribed to organic materials that accumulate during fire-free intervals that volatilize during wildfire and migrate downward along temperature gradients in the soil and precipitate as a water repellant coating (DeBano, 1981, 2000b). This leads to vertical and horizontal variations in hydraulic characteristics, with water repellency usually being strongest at shallow soil depths and declining downwards through the soil profile (DeBano, 1981; Huffman, MacDonald, & Stednick, 2001).

Effects on infiltration

Postfire soils often have a wettable layer near the surface with a water repellent layer underneath—combined with postfire changes in soil structural and hydraulic properties, this affects the infiltration process, influencing the partitioning of run-off between the infiltration excess and saturation excess mechanisms (Figure 1). For example, near-saturated hydraulic conductivity (K_f) and sorptivity (S) of burned soils has been shown to relate inversely to the degree of water repellency (Moody, Kinner, & Úbeda, 2009). In turn, reductions in K_f and S can lead to enhanced infiltration-excess run-off generation with elevated risks for flash floods and debris flows (Ebel & Moody, 2017). Furthermore, unburned duff, consisting of partially decomposed litter with humus underneath the ash layer (Robichaud & Miller, 1999), can create dispersed zones along the hillslope that are water repellent when dry (Ebel, Moody, & Martin, 2012). These zones become water absorbent when moist, which not only affects vertical preferential infiltration into the soil but also creates spatial discontinuities in the run-off-generating area and adds to the spatial complexity of the hydraulic connectivity along the hillslope (Ebel et al., 2012; Moody et al., 2016).

Persistence with time

The infiltration mechanism of coarse-grained soils, with an initially high infiltration rate, can be fundamentally changed by hyper-dry or extremely dry conditions (water content <0.02 cm^3/cm^3; Moody et al., 2009) during a fire and have been associated with areas that have near-zero infiltration in the Colorado Rockies (Ebel & Moody, 2013). Nevertheless, even highly water repellent soils often return to a wettable state after prolonged rainfall (Shakesby, Doerr, & Walsh, 2000). Any link between hyper-dry conditions and reduced infiltration is usually temporary, and postfire water repellency declines gradually within 1 to 2 years after burning (Hubbert, Wohlgemuth, Beyers, Narog, & Gerrard, 2012; Larsen et al., 2009; MacDonald & Huffman, 2004).

Natural background water repellency

Increased postfire run-off has also been observed on plots without any increase in soil water repellency, suggesting that soil water repellency is not required for generating floods (Cannon, Gartner, Rupert, Michael, Rea, & Parrett, 2010). Conversely, high levels of water repellency have been observed in unburned plots, notably during drought (Goebel, Bachmann, Reichstein, Janssens, & Guggenberger, 2011). Doerr, Woods, et al. (2009) analysed topsoil samples from mixed-conifer stands in the highlands of Colorado, Idaho, Montana, Utah, Wyoming, and Oregon and detected variable degrees of water repellency already present in the soil (water drop penetration time > 5 s) for 75% of the sites, regardless of dominant conifer species. Although limited to soils found on top of sedimentary deposits and soils with clay content >4%, water repellency could not be accurately predicted from terrain or soil variables (Doerr, Woods, et al., 2009).

2.2.2 | Ash and soil surface sealing

Ash contains mineral and charred organic components that remain after the combustion of fuels and varies in depth and composition. Ash is wettable in most cases, characterized by a water holding capacity and hydraulic conductivity that is distinct from that of the soil (Kinner & Moody, 2010), and can affect infiltration directly by absorbing water, and indirectly by reducing soil surface sealing in the immediate postwildfire period (Figure 1).

Effects on infiltration

During the first few precipitation events following wildfire, the infiltration process is often influenced by a hypothetical two-layer system with ash on top of soil (Bodí et al., 2014); however, at a catchment scale, ash is often concentrated around vegetation and rarely forms a continuous homogeneous layer. Surface ash initially stores the rainfall it receives (Martin & Moody, 2001). A 60–65% surface cover of ash was shown to substantially reduce postfire sediment yields in the Colorado Rockies (Larsen et al., 2009), but once the storage capacity of the ash layer is exceeded, burned sites tend to generate more run-off than unburned sites (Martin & Moody, 2001). Ash features high rates of sorptivity (S = 0.18–0.20 cm/s) and near-saturation vertical hydraulic conductivity of ash (K_f = 33–4680 mm/hr) in Colorado and Southern California (Moody et al., 2009). Gabet and Bookter (2011) suggested that the high K_f for ponderosa pine ash (K_f = 165 mm/hr) in the Rocky Mountains of western Montana, comparable with that of fine sand, results from the particle size distribution. In Colorado, total storage capacity of the ash layer was limited and short-lasted, with a time to run-off of 8–12 min versus 6 min on an unburned control plot (Larsen et al., 2009). Ash covered plots produced only 21–49% of the amount of run-off on the control plot, considerably less than observed by Martin and Moody (2001). This may be explained by soil properties, as supported by evidence for variable hydrological response after low-intensity prescribed fire in Montana depending on local changes in soil texture and ash thickness (Woods & Balfour, 2010), where a thicker ash layer (>2 cm) favoured storage and increased infiltration by 26–30% (Woods & Balfour, 2010). There is presently no evidence of a widespread reduction of infiltration capacity due to pore clogging by ash (Bodí et al., 2014), especially not in sandy soil because ash and sand both have a negative surface charge (Stoof et al., 2016).

Relation to soil surface sealing

In contrast with the buffer effect of the ash layer, postfire soil surface sealing caused by direct raindrop impact can promote run-off generation and responds rapidly to postfire rainfall (DeBano, 2000b; Martin & Moody, 2001). Martin and Moody (2001) suggested that soil surface sealing can be attributed to precipitation and run-off, the former contributing to the development of structural crusts (Assouline, 2004) and the latter to depositional crusts (Onda, Dietrich, & Booker, 2008). Increased postfire run-off and sediment yields in the Colorado Front Range were mainly caused by soil surface sealing rather than by changes in soil water repellency (Larsen et al., 2009). An ash layer on the other hand can protect the soil from compaction and self-crusting and largely preserves the infiltration capacity of the burned soils (Ebel, 2012; Ebel et al., 2012; Moody et al., 2009). The ash layer itself can also develop a crust that will delay the inevitable erosion, as observed in western Montana (Balfour, Doerr, & Robichaud, 2014).

Persistence with time

Ash cover remains in place until it is washed out (Larsen et al., 2009) or removed by wind erosion shortly after wildfire (e.g., ~10 to 30 days), depending on the postfire weather and presence of an ash crust (Balfour et al., 2014; Woods & Balfour, 2010). Rainfall simulations in Colorado demonstrated that ash cover decreased from 77% to 45% after a second rainfall, suggesting that ash effects on run-off are temporary and most significant during the first postfire storms, depending on rainfall intensity and duration (Larsen et al., 2009).

2.3 | Relation to surface hydraulic connectivity

Surface hydraulic connectivity is an important factor controlling the amount of run-off entering the stream network (Figure 1; Hallema & Moussa, 2014, Hallema, Moussa, Sun, and McNulty, 2016) and may increase given the enhanced infiltration-excess run-off on burned soil reported by Ebel and Moody (2017). For example, field-saturated hydraulic conductivity and sorptivity may decrease in severely burned soils, acting to reduce infiltration and intensify surface run-off generation (Ebel & Moody, 2017). This effect may be amplified with increasing burn severity (Moody et al., 2016). Moody, Martin, Haire, and Kinner (2008) expressed the mean functional hydraulic connectivity in terms of the magnitude of burn severity and spatial sequence of burn severity patterns along hillslope flow paths. Spatial burn severity patterns are reflected in the fire impact on soil hydraulic parameters, and therefore, the concept of a continuous ash or water repellent layer may not be accurate for most field situations (Moody et al., 2013; Woods, Birkas, & Ahl, 2007). A mosaic of water repellent patches of variable thickness is more accurate given that fire-induced water repellency occurs mostly in the area beneath vegetation canopies and possibly the zones with unburned duff consisting of litter and humus (Robichaud & Miller, 1999). Run-off generated on these patches can infiltrate in downhill zones of wettable ash and soil (Shakesby et al., 2000).

Although soil water repellency may indeed affect postfire run-off, this contribution is inherently difficult to quantify because there are currently no techniques available to measure water repellency across the landscape (Larsen et al., 2009). Moreover, the phenomenon varies greatly in time and space as a result of changes in soil moisture and breakdown of water-repellent compounds (Doerr, Shakesby, and MacDonald, 2009; Larsen et al., 2009; Woods et al., 2007). Broader research shows that once run-off is generated, factors such as slope length, gradient, flow path convergence, infiltration rates, vegetation, and soil surface roughness play an important role in determining pathways of concentrated flow (Reaney, Bracken, and Kirkby, 2014). The organization of these pathways largely affects the timing of flow delivery at the base of the hillslope and, eventually, to the stream (Hallema, Moussa, Andrieux, & Voltz, 2013; Hallema & Moussa, 2014).

3 | EFFECTS ON STREAMFLOW BY CLIMATE REGION

Streamflow is affected by climate (e.g., rainfall seasonality and extreme events and snow melt), physical characteristics of the watershed (e.g., area, shape, relief, geology, and soils), and vegetative canopy (e.g., seasonal leaf area). The prediction of postfire streamflow is complicated by postfire climate effects, which vary regionally and seasonally, and magnitude of impact on surface characteristics (e.g., ash effect, soil sealing, soil water repellency, and infiltration capacity). Soil properties are often unknown, and postwildfire hydrographs are difficult to measure as a result of changes in streambed elevation, floods (Moody et al., 2013), debris flows (Cannon, Bigio, and Mine, 2001, Cannon, Kirkham, and Parise, 2001), or severe erosion (Benda, Miller, Bigelow, & Andras, 2003) and diminish the ability to measure flow. Here, we describe postwildfire streamflow responses per Köppen climate region (Godfrey, 1999; Peel, Finlayson, & McMahon, 2007). Special attention is given to the Mediterranean, semi-arid and highland climate types, which suffer from longer fire seasons associated with water deficit and drought resulting from the El Nino-Southern Oscillation and Pacific Decadal Oscillation (Morton et al., 2013; Swetnam & Betancourt, 1990). Changes in streamflow are reported in Table 2, with the corresponding locations in Figure 2. We are not aware of any evidence for prescribed fire-induced floods within the humid subtropical or other climate regions of the Eastern United States; however, an 8% increase in the 5-year annual water yield of a watershed in South Carolina was likely not caused by prescribed fire but by a combination of exceptionally high winter precipitation, beaver activity, and storm damage (Hallema, Sun, et al., 2016).

3.1 | Mediterranean (Csa and Csb)

Postwildfire run-off increased significantly in regions with a Mediterranean climate, in particular in Southern California (Bart, 2016; Jung et al., 2009) where the shrub vegetation (chaparral) recovered slowly and transpiration decreased for extended periods of time (Kinoshita & Hogue, 2015). Annual precipitation amounts in Southern California are very low (<500 mm) compared to Northern California (700–2500 mm) and the Sierra Nevada mountains (600–1500 mm). The Sierra Nevada and Northern California experience frequent low severity fires, but in Southern California, the drier Mediterranean climate (hot summer Csa and warm summer Csb) combined with chaparral vegetation results in frequent high-severity fires with a peak season in autumn driven by the offshore winds (Lin et al., 2014). Depending on the dominant tree species, vegetation recovery is further delayed by postfire drought or increased postfire temperatures (Meng, Dennison, Huang, Moritz, & D'Antonio, 2015). Therefore, species composition also affects the recovery rate of forest transpiration and annual water yields, which can take as long as 10 years in Southern California (Kinoshita & Hogue, 2015).

TABLE 2 Change in streamflow from CONUS watersheds in the years following wildland fire

Watershed	Location	Drainage area (km^2)	Elevation (m)	Canopy species	Fire name	Predominant burn severity	Postfire management	Postfire streamflow (attributed to fire disturbance)	Method	Reference
Wolfskill Canyon	San Dimas EF, S. California	6.16	450–1675[a]	Chaparral and mixed hardwood (oak, maple, Douglas fir)	1953 Barrett	Moderate to high	Likely sowing of annual grasses	+1,080%[a] peak flow in year 1	Run-off simulation (HEC-HMS, KINEROS2)	Chen et al. (2013)
Volfe Canyon	San Dimas EF, S. California	3.0	450–1675	Chaparral and mixed hardwood (oak, maple, Douglas fir)	1960 Unnamed	Moderate to high	–	Peak flow increased from $6.8.10^{-5}$–$3.0.10^{-4}$ m^3/s to 0.37–9.3 m^3/s	Single watershed, discharge	Wohlgemuth (2016)
Devil Creek	San Bernardino NF, S. California	14.1	500–1700	Chaparral, woodland, mixed conifer	2003 Old	Moderate to high	–	Overland flow contribution to storm run-off from 9% to 11%, interstorm baseflow unchanged 77% in year 1	Single watershed, chemical end-member analysis	Jung et al. (2009)
Devil Creek West Fork (subcatchment)		6.94						Overland flow contribution to storm run-off from 9% to 88%, interstorm baseflow from 52% to 70% in months 1–4		
Devil Creek	San Bernardino NF, S. California	14.1	500–1700	Chaparral, woodland, mixed conifer	2003 Old	Moderate to high	–	+1090% low flow in years 1–9, +7%[a] annual RC, becomes perennial	Single watershed, discharge	Kinoshita and Hogue (2015)
City Creek		51	300–2100					+118% low flow in years 1–9, +86%[a] annual RC		
12 watersheds	Coast and Transverse Ranges, S. California	7.2–625	–	Chaparral, grassland, coastal sage scrub, oak woodlands	Multiple	Variable	–	+82% to +200% annual water yield year 1	Paired watershed, mixed model discharge	Bart (2016)
Burns Creek	Entiat EF, Cascade Range, C. Washington	5	842–2156	Mixed conifer (ponderosa pine, Douglas fir)	1970 Entiat	Moderate to high	Salvage logging, areal seeding of grasses and yellow sweetclover	+50% annual RC in year 1 relative to prediction based on precipitation, +150%[a] median run-off years 1–6	Paired watershed, discharge	Helvey (1980)

(*Continues*)

TABLE 2 (Continued)

Watershed	Location	Drainage area (km^2)	Elevation (m)	Canopy species	Fire name	Predominant burn severity	Postfire management	Postfire streamflow (attributed to fire disturbance)	Method	Reference
Burns Creek	Entiat EF, Cascade Range, C. Washington	5	842–2156	Mixed conifer (ponderosa pine, Douglas fir and other)	1970 Entiat	Moderate to high	Salvage logging, areal seeding of grasses and yellow sweetclover	+120% peak flow, +150% to +200% median run-off in years 1–6	Run-off simulation (HBV)	Seibert et al. (2010)
Oak Creek West Fork	Coconino NF, N. Arizona	0.04 and 0.081	1500–1900[a]	Mixed conifer (ponderosa pine, Douglas fir, spruce)	1972 Rattle	Moderate and high	Salvage logging	+250%[a] and +357% annual RC relative to control watershed in years 1–3	Paired watershed, discharge	Campbell et al. (1977)
Willow Creek West Fork	Apache-Sitgreaves NF, E. Arizona	1.17	2682–2835	Mixed conifer (quaking aspen, Engelmann spruce, Douglas fir)	2011 Wallow	Low to high	Seeding	+210% storm peak flow, no effect on winter and snowmelt run-off in year 1	Single watershed, discharge	Wagenbrenner (2013)
Rendija Canyon	Santa Fe NF, N. New Mexico	24.8	1920	Mixed conifer (Ponderosa pine, Douglas fir, white fir)	2000 Cerro Grande	Moderate to high	–	Max. peak flow 50 $m^3{\cdot}s^{-1}{\cdot}km^{-2}$ (2-year storm)	Single watershed, discharge	Moody and Martin (2001)
Spring Creek	Front range, C. Colorado	26.8	1880	Ponderosa pine dominated	1996 Buffalo Creek	Moderate to high		Max. peak flow 24 $m^3{\cdot}s^{-1}{\cdot}km^{-2}$ (100-year storm)		
Bear Gulch	Black Hills, South Dakota	17	1100	Ponderosa pine dominated	1988 Galena	Moderate to high		Max. peak flow 3.2 $m^3{\cdot}s^{-1}{\cdot}km^{-2}$ (10-year storm)		
Bobcat Gulch	Front range, C. Colorado	2.2	1960–2575	Ponderosa pine, lodgepole pine, Douglas fir	2000 Bobcat	Moderate to high	Aerial seeding, contour log felling, mulching	Max. peak flow 3.9 and 1.1 $m^3{\cdot}s^{-1}{\cdot}km^{-2}$ in month 2 and year 2, resp.	Paired watershed, discharge	Kunze and Stednick (2006)
Jug Gulch		3.9	2020–2470				Partial seeding	Max. peak flow 0.005 and 1.7 $m^3{\cdot}s^{-1}{\cdot}km^{-2}$ in month 2 and year 2, resp.		
Jones Creek	Shoshone NF, NW. Wyoming	66.8	2087–3281[a]	Subalpine fir dominated	1988 Clover-Mist[b]	Moderate to high	–	+120%[a] annual RC than control watershed in years 3–4	Paired watershed, discharge	Troendle and Bevenger (1996)

(*Continues*)

TABLE 2 (Continued)

Watershed	Location	Drainage area (km^2)	Elevation (m)	Canopy species	Fire name	Predominant burn severity	Postfire management	Postfire streamflow (attributed to fire disturbance)	Method	Reference
Grace Coolidge Creek	Custer SP, Black Hills, W. South Dakota	69.4	1250–1750[a]	Ponderosa pine dominated	1988 Galena	Moderate to high	Salvage logging, sediment traps, slash spreading, areal seeding	−66%[a] annual RC in year 1, return to prefire conditions after year 2	Single watershed, discharge	Driscoll et al. (2004)
Del Puerto Creek	Diablo Range, C. California	187	75–1113	Sagebrush, chaparral	2003 Deer Park	Moderate to high	Sediment removal from stream bed near outlet	−64% (+38%) in 5-year annual water yield	Single watershed, CEM filter	Hallema et al. (2016)
Wet Bottom Creek	Tonto NF, N. Arizona	93	715–2157	Pinyon juniper, ponderosa pine	2004 Willow	Low to moderate	-	+266% (+219%) in 5-year annual water yield		
Black Creek	Carolina Sandhills, South Carolina	295	79–219	Longleaf pine	Rx	Low	(salvage logging after severe storms)	−39% (+8%) in 5-year annual water yield		

Note. NF = national forest; EF = experimental forest; SP = state park; Rx = prescribed burning; RC = run-off coefficient (water yield efficiency); CEM = climate elasticity model.

Note. Percentages indicate the percentage change with regard to the prefire value. Unreported variables are marked with a dash.

[a]Approximated a posteriori based on information provided in reference.

[b]Associated with the 1988 Yellowstone Fires.

High flows

Chen, Berli, and Chief (2013) observed a substantial increase from 2.3 m^3/s to 24.6 m^3/s or 1,080% in the Wolfskill Canyon headwaters (6.16 km^2) in Southern California. The adjacent Volfe Canyon had similar vegetation and soil texture, and although smaller than Wolfskill Canyon (3.0 km^2), postwildfire run-off was four orders of magnitude higher and resulted in flood damage downstream (Wohlgemuth, 2016). The differences in postwildfire streamflow response were caused by differences in prestorm soil conditions prior, notably soil moisture levels. Chen et al. (2013) simulated postwildfire streamflow response in Wolfskill Canyon and observed a temporary shift from a saturation excess, subsurface flow-driven mechanism to an infiltration-excess mechanism in the postfire period. A similar shift observed in the Colorado Rockies has been associated with water repellency (Ebel & Moody, 2013). Water repellency may have produced the same effects in Southern California, given previous and more recent observations of water repellency (Hubbert et al., 2012; Krammes & Rice, 1963) and reduced infiltration rates (Jung et al., 2009). Increased surface run-off after wildfires has caused major issues in this area and has been cited as the main cause of postwildfire debris flows (Cannon et al., 2010; Kean, Staley, & Cannon, 2011).

Low flows

Kinoshita and Hogue (2015) observed a significant increase in low flows, mostly during interstorm periods, during the first 10 years following another nearby wildfire. Low flows (90% exceedance probability) increased from an order 10-3 m^3/s to 10-1 m^3/s, or 118% in the 51-km^2 City Creek watersheds and 1,090% in the 14-km^2 Devil Creek watershed. The latter became perennial after the wildfire. A key factor for baseflow contributions in this area is the geology consisting of granitic and gneissic rocks. Deformation associated with the San Andreas Fault has created fissures and fractures in the bedrock, and these fractures act as the main source or perennial groundwater discharge (Jung et al., 2009; Troxell, 1954). Consequently, an increase in postfire baseflow could possibly be ascribed to a decrease in subsurface water storage explained by the loss of vegetation (Jung et al., 2009).

Annual water yield

Annual water yield increased by 7% and 86% in the Devil Creek and City Creek watersheds (Kinoshita & Hogue, 2015) and 82% to 200% in 10 other completely burned watersheds in Southern California (Bart, 2016). End-member mixing analysis in the Devil Creek West Fork furthermore showed that the postwildfire overland contribution to storm flow increased from 9% to 88% (Jung et al., 2009), while interstorm stream water contained more groundwater than before the wildfire (70% vs. 52%). Despite a slight increase in the overland contribution to storm flow water in the Devil Creek itself (9% to 11%), no change was observed in groundwater contribution in the postwildfire year partly due to the longer residence time of water in larger watersheds and the gentler overall slope (Jung et al., 2009). Annual water yields in the Del Puerto Creek in Northern California declined by 64% in the 5 years following a wildfire, but after filtering out the impact of drought using a climate elasticity model, an increasing trend emerged that attributes a 38% higher annual water yield to the wildfire (Hallema, Sun, et al., 2016).

3.2 | Semi-arid and warm temperate (BSk, Csa, and Csb)

Semi-arid and warm temperate climate regions have the greatest increase in postwildfire peak flows, especially in the Southwest where spring snowmelt in upstream headwaters is an important source of water supply. The largest forests are found in the mountainous regions of Arizona and Southwest New Mexico where annual precipitation varies anywhere between 300 and 700 mm. Forests with the largest fires mostly have cold semi-arid conditions (BSk) at the lower mid-altitudes and hot (Csa) to warm (Csb) summer temperate at the higher mid-altitudes. The fire regime in these coniferous forests is characterized by frequent low severity fires, which have a less severe effect on canopy cover and long-term transpiration compared to Mediterranean Southern California (Schoennagel, Veblen, & Romme, 2004). Notwithstanding, this area has also known some of the most extreme hydrological responses to severe wildfire (Hallema, Sun, et al., 2016).

High flows

In eastern Arizona, Wagenbrenner (2013) noted that the 2011 Wallow Fire had no effect on run-off in the Willow Creek West Fork during the winter and snowmelt periods in a watershed with intermittent flow (21-year prefire peak flows up to 0.47 m^3/s in a 1.17-km^2 watershed). However, storm peak flow was observed to increase by as much as 210% from the prefire peak flow, which is comparable in magnitude to change observed in the Rendija Canyon in New Mexico (50 $m^3/km^2/s$; Moody & Martin, 2001) and the changes in the California watersheds discussed above. The rainfall threshold marking a critical change in watershed response corresponded with 11 mm/hr in Rendija Canyon (Moody & Martin, 2001), which was associated with the postfire infiltration rates and surface roughness observed in this watershed. Below this threshold, no additional run-off was observed.

Annual water yield

Annual water yields in some Arizona watersheds appear more severely impacted by wildfire than in watersheds in California. Campbell et al. (1977) observed a 250% and 357% higher 3-year annual water yield for two burned watersheds in northcentral Arizona compared to an unburned watershed in the area (Campbell et al., 1977). This difference can be ascribed to the combined effect of a generally lower yield corresponding with a smaller drainage area (0.04 and 0.081 km^2, respectively, vs. 6.16 km^2) despite the salvage of most killed saw timber, which may have caused soil compaction and increased overland flow. Hallema, Sun, et al. (2016) found a comparable 266% postwildfire water yield increase in the Wet Bottom Creek in Northern Arizona, of which a 219% increase was attributed to wildfire disturbance and a 47% increase to more precipitation.

3.3 | Highlands (Dfa and Dfb)

Water supplies in regions with a warm (Dfb) or hot (Dfa) summer continental climate such as the Rocky Mountains are particularly susceptible to wildfire impacts due to the high severity of these infrequent fires, and the effects on population centres are significant given their dependence on upstream forest lands. For example, water providers in Denver incurred $26 million USD in costs to manage postfire sedimentation and restore water supply after two fires in 1996 and 2002 (Gartner, Mulligan, Schmidt, & Gunn, 2013; Martin, 2016).

High flows

Significant postfire response required a similar rainfall threshold as observed in California and the Southwest, for example, 10 mm/hr for the Spring Creek in the Colorado Front Range (Moody & Martin, 2001). Once run-off was initiated however, peak flow was lower than in California and the Southwest, with a maximum of 24 $m^3/km^2/s$ for a storm with a 100-year return interval (cf. 50 $m^3/km^2/s$ for a more common storm type with 2-year return interval in California; Moody & Martin, 2001). In watersheds of the Colorado Front Range with postfire treatments such as contour log felling and mulching, this number was even lower (<3.9 $m^3/km^2/s$; Kunze & Stednick, 2006). Storm run-off on wildfire-affected soils in the Rockies and much of the West where soils are rarely water saturated is generally controlled by the infiltration capacity of the soil (Kinner & Moody, 2010; Ebel & Moody, 2013). Moody and Ebel (2012) have argued that in the Colorado Rockies, catastrophic floods after wildfire are often linked to the first substantial rainfall. Near-zero infiltration was observed immediately after wildfire even before sealing by fine sediments, which usually takes several rainfall events. This is supported by observations of fire-induced water repellency in the Colorado Rockies and subsequent hyper-dry soil conditions causing reduced infiltration rates (Moody & Ebel, 2012). Even when run-off generation is near-immediate at the soil surface, functional hydraulic connectivity across the watershed increases in response to fire disturbance (Williams et al., 2016), especially when infiltration capacity decreases in downhill direction whether or not as a result of spatial patterns of burn severity. Moody et al. (2008) suggested that storm run-off is a linear function of the mean functional hydraulic connectivity expressed in terms of the magnitude of the burn severity and the spatial sequence of burn severity patterns along hillslope flow paths and demonstrated this theory using data from four catchments (~0.28 km^2) in the southern Rockies of New Mexico. Highlands of the Pacific Northwest are characterized by temperate rainforests and infrequent occurrence of high severity wildfires (Wimberly & Liu, 2014). Seibert, McDonnell, and Woodsmith (2010) calibrated the conceptual Hydrologiska Byråns Vattenbalansavdelning (HBV) run-off model with data from the Entiat Experimental Forest in Washington and found a 120% higher peak flow compared to conditions before the 1970 fire in this area.

Annual water yield

Troendle and Bevenger (1996) observed larger percent increases in water yield for frequent, small-magnitude storms, and smaller percent increases were observed for less frequent, large-magnitude storms after the 1988 Clover–Mist fire in the Shoshone National Forest in Wyoming. Subsequently, annual water yields were up to 120% higher compared to an unburned control watershed. Similar numbers were reported for the three watersheds studied in Washington regardless of the postfire treatments, which was different for each. Median run-off was 150% to 200% higher in the first 6 years following the fire, which corroborates with an earlier study where this watershed was compared to a control watershed (Helvey, 1980). Seibert et al. (2010) have argued that the increase in run-off was caused by deeper snow packs and more rapid snowmelt in spring based on the parameter values they found after calibrating the model.

4 | DISCUSSION

4.1 | Regional patterns of postwildfire hydrologic responses

Regional patterns in postwildfire hydrologic response are characterized by a greater increase in postwildfire peak flows and annual water yields in areas with a Mediterranean (Southern California), and especially a semi-arid climate (Arizona) compared to highland regions with snow (Rocky Mountains and the Pacific Northwest). These patterns are explained by the spatial variability of postwildfire infiltration and run-off generation (impacted by variations in fire frequency and severity and resulting spatial patterns of ash, soil surface sealing, and water repellency), changes in surface hydraulic connectivity (along the hillslope and between the hillslope and watershed outlet), and climate patterns (trends and droughts, seasonality, snowmelt and precipitation intensity, and distribution).

4.1.1 | Infiltration and run-off generation

Infiltration and water retention are promoted by ash until the ash is removed from the surface. Depending on the temporal rainfall distribution and rainfall intensity, this can happen very quickly in, for example, the Rocky Mountains, which receive regular rainfall (Ebel, 2012; Moody et al., 2009) as opposed to Southern California where annual precipitation is much lower. Findings demonstrating the unlikeliness of an effect of pore clogging on infiltration in sandy soils (Stoof et al., 2016) have shifted the research focus in the Rockies to the effects of differences between hydraulic conductivity of ash and underlying layers depending on texture, delayed water release from ash as a result of capillary forces, and lateral flow through ash and soil on top of a water repellent soil (Baker & Hillel, 1990; Bodí, Doerr, Cerdà, & Mataix-Solera, 2012; Schroth, Istok, & Selker, 1998).

4.1.2 | Surface hydraulic connectivity

Evapotranspiration, infiltration, and run-off generation are altered by fire, and the linkage between these processes along the soil surface (hydraulic or hydrological connectivity) controls changes in streamflow (Moody et al., 2008). The layer concept, although practical for hydrologic response models, is neither accurate nor useful for identifying regional patterns in ash, soil surface sealing, and water repellency effects on postwildfire streamflow. Physically based formulations of surface and subsurface hydraulic connectivity operating across different scales and rainfall intensities are more accurate (Woods et al., 2007), supported by Moody et al. (2013): (a) evidence of comparable rainfall intensity thresholds for run-off generation across the West despite substantial differences in postwildfire streamflow between semi-arid and Mediterranean watersheds (Moody & Martin, 2001; Moody et al., 2008); (b) the dependence of postwildfire storm run-off on burn severity and the spatial sequence of burn severity patterns (Moody et al., 2016); (c) correlations between unit-area peak discharge and the maximum 30-min rainfall intensity (Moody & Martin, 2001; Robichaud, Lewis, & Ashmun, 2008); and (d) variable source area contributions (Hewlett & Hibbert, 1967). The propagation of individual contributions from hillslopes and the time of concentration of the combined hydrograph depends not only on the flow velocity but also on the drainage pattern (influenced by geology) and hydrologic connectivity at the watershed scale (Hallema & Moussa, 2014; Rinaldo, Rigon, & Marani, 1991). Regularly shaped watersheds with a dendritic drainage pattern typically yield a higher peak flow than oval shaped or elongated watersheds, all other factors being equal (Black, 1972).

4.1.3 | Climate effects

Fire impacts on hydrology are highly variable between climate regions and from year to year. Decreasing precipitation trends attenuate the increase in postwildfire annual water yields in Mediterranean California, but peak flows are enhanced by a combination of dry prestorm soils and canopy loss. The effects of drought-enhanced postwildfire hydrophobicity combined with slow vegetation recovery in Southern California and the warm temperate to semi-arid Southwest (Arizona and New Mexico) have a particularly strong impact in these regions. Although drought contributes to the impact, drought itself does not automatically lead to more fires because, locally, wildfire occurrence and impacts are linked to a complex system of interactions between drought, fuel moisture conditions, surface and canopy fuels, and ignitions (Littell, Peterson, Riley, Liu, & Luce, 2016). However, future climate scenarios suggest that declines in snowpack, earlier onset of spring melt, and reduced soil and fuel moisture in the summer may increase fire potential throughout the West (Gergel, Nijssen, Abatzoglou, Lettenmainer, & Stumbaugh, 2017), and additional research must show what the effects are on streamflow. Conversely, in the humid Southeast where annual prescribed burns account for the greatest number of reported wildland fires, storm damage may have a stronger impact on annual water yields (Hallema, Sun, et al., 2016). Although infiltration and hydraulic connectivity are important factors, much of the regional variability in wildfire impacts on annual water supply (as opposed to peak flows) are ultimately controlled by climate patterns, because annual streamflow mostly depends on the annual amounts of precipitation, evapotranspiration, and withdrawals (Sun, McNulty, Moore Myers, & Cohen, 2008).

4.2 | Challenges in research

With the current state of knowledge of wildfire effects on hydrological processes, the main challenges in the understanding of postwildfire streamflow response are broadly related to (a) the scale of hydrological processes impacted by wildfire and the hydraulic connectivity that defines the linkage between these processes and (b) the transient nature of postwildfire hydraulic connectivity and overlapping disturbances.

4.2.1 | Scale and postwildfire hydraulic connectivity of hydrological processes

A principal challenge in the near future is the need for more watershed scale assessments because this is the scale where postwildfire flooding, erosion, and sedimentation issues occur and where abatement practices are implemented. These assessments will furthermore help to determine whether the respective effects of soil surface sealing, ash, and change in soil water repellency can be predicted. Soil infiltration and water retention models calibrated on field data are a valuable tool in this regard (Hallema, Périard, Lafond, Gumiere, & Caron, 2015; Hallema et al., 2014), but questions like this remain difficult to answer because they require the upscaling of experimental data obtained from local samples or in situ infiltration experiments to the watershed scale. Upscaling is complicated by the large number of pre and postwildfire experiments needed to identify any existing correlations between soil surface sealing, ash, and water repellency on one hand and terrain and soil variables on the other hand. Future wildfires and suitable locations for infiltration measurements cannot be predicted with accuracy, and therefore, such experiments remain a logistical challenge. Nonetheless, an important advantage of watershed scale assessments is that they can be aggregated to the landscape scale, which will help answer ecological questions, such as whether or not postfire changes in water repellency must be viewed as a regional landscape concern. A more complete understanding of functional hydraulic connectivity in terms of burn severity, soil surface sealing, ash, and soil water repellency may prove very helpful in explaining the linkage between these processes and their effect on storm run-off (Hallema, Moussa, et al., 2016; Moody et al., 2016). It will also prove essential in determining how the spatial distribution of fire-affected areas within a watershed (headwater vs. valley, steep vs. gentle slope, and north- vs. south-facing slope aspects) impacts postwildfire run-off and, ultimately, regional patterns in postwildfire streamflow response.

4.2.2 | Transient nature of postwildfire processes and overlapping disturbances

Recent studies have scratched the surface of hydraulic connectivity as a predominantly spatial concept, with structural and functional components. The characteristics of fire-affected areas within a watershed also change over time due to postwildfire vegetation recovery, washing of ash, soil surface sealing, erosion, mud flows, and previously unobserved forest responses (McNulty, Boggs, & Sun, 2014). All of these can affect the water balance differently depending on the climate (Kinoshita & Hogue, 2011). The cause of changes in hydrologic response is not always evident due to overlap between fire disturbance and other disturbances (drought, beetle infestations, and human disturbance), which increases the challenge of identifying the normal state of a system (Glenn-Lewin, Peet, & Veblen, 1992; Hallema, Sun, et al., 2016; Temperli, Bugmann, & Elkin, 2013). The simultaneous and asynchronous occurrence of postwildfire responses and responses to other disturbances is a complex puzzle that remains to be solved. Postwildfire precipitation patterns and rainfall intensities are essential in establishing the sequence of postwildfire responses and the migration expansion of the variable source area during run-off-generating events.

4.2.3 | Prescribed fire impacts on streamflow

Within the broader context of wildland fire impacts on water yield, not much is known on the effects of annual prescribed burning in the Eastern CONUS. More postfire assessment studies are needed in that region, given the fact that 74% of the total area burned in the mid-Atlantic between 2001 and 2010 was for the account of prescribed fires (Clark, Skowronski, Renninger, & Scheller, 2014). The lack of studies on the hydrologic response to prescribed fires in the East is explained by the lower severity of these fires and the relative absence of fire-induced hydrophobicity. Eastern forests have also recovered faster from fire-induced hydrologic disturbance compared to the West, not only because of less severe impacts commonly associated with prescribed fires but also as a result of higher precipitation and growth rates. Modelling tools exist to distinguish wildland fire impacts from climate impacts (Hallema, Sun, et al., 2016). The efficiency and reliability of risk assessment tools will rely on the continued improvement of models of mesoscale climate patterns (Jones, Nyman, & Sheridan, 2014; Moody et al., 2016; Shakesby, Moody, Martin, & Robichaud, 2016). Regional climate change is expected to lead to higher precipitation variability and evapotranspiration across the United States and possibly higher fire potential after 2040 (Duan et al., 2016; Liu, Goodrick, & Stanturf, 2013). Better knowledge of the dynamic aspects of connectivity (Williams et al., 2016) and connection with postwildfire climate will help quantify the transient nature of postwildfire hydraulic connectivity and ultimately provide more accurate predictions of the postwildfire state of the watershed.

5 | CONCLUSIONS

Studies on postwildfire streamflow in the Western United States indicate that peak flows and annual water yield generally increased more in regions with a Mediterranean or semi-arid to warm temperate climate (Southern California, Arizona, and New Mexico) compared to highland regions (Rocky Mountains and parts of the Pacific Northwest). Most wildland fires in the Eastern United States are prescribed fires, and presently, there is little evidence for any significant impacts on streamflow. A complex system of spatial and temporal interactions between hydrological processes lies at the root of regional differences in postwildfire run-off generation. The general pattern shows a reduction of leaf area, resulting in an immediate increase in net precipitation and decrease in evapotranspiration depending on burn severity, leaving the soil exposed to direct impact from rainfall. Postfire run-off generation was enhanced by rain splash-induced soil surface sealing and postwildfire water repellency, notably in the Rocky Mountains, California, and the Southwest. Regional differences in climate have been cited as the main driver for regional variations in postwildfire streamflow change because they largely determine the prestorm moisture conditions and the likelihood of soil surface sealing following a wildfire. Researcher have been able to demonstrate links between scale-specific hydrological processes, surface hydraulic connectivity, and spatial patterns of burn severity along the hillslope, which control the amount of run-off reaching the headwater streams. Nevertheless, we know little at present about the interactions between transient postwildfire interactions and overlapping disturbances, and better tools are needed to identify spatial and temporal correlations between run-off generation, burn severity, water repellency, and ash deposition patterns. This will also be helpful in the assessment of prescribed fire impacts on streamflow in the East. Further research on postwildfire hydraulic connectivity within the catchment will prove to be an essential element in the development of tools that are applicable in any wildfire impact assessment.

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