Measurements of the initiation of post-wildfire runoff during rainstorms using \textit{in situ} overland flow detectors

John A. Moody\textsuperscript{1*} and Richard G. Martin\textsuperscript{2}

\textsuperscript{1} US Geological Survey, National Research Program, Boulder, CO, USA
\textsuperscript{2} Martin Enterprise, Wheat Ridge, CO, USA

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*Correspondence to: John A. Moody, US Geological Survey, National Research Program, Boulder, CO, USA. E-mail: jamoody@usgs.gov

\textbf{ABSTRACT:} Overland flow detectors (OFDs) were deployed in 2012 on a hillslope burned by the 2010 Fourmile Canyon fire near Boulder, Colorado, USA. These detectors were simple, electrical resistor-type instruments that output a voltage (0–2.5 V) and were designed to measure and record the time of runoff initiation, a signal proportional to water depth, and the runoff hydrograph during natural convective rainstorms.

Initiation of runoff was found to be spatially complex and began at different times in different locations on the hillslope. Runoff started first at upstream detectors 56\% of the time, at the mid-stream detectors 6\%, and at the downstream detectors 38\% of the time. Initiation of post-wildfire runoff depended on the time-to-ponding, travel time between points, and the time to fill surface depression storage. These times ranged from 0.5–54, 0.4–1.1, and 0.2–14 minutes, respectively, indicating the importance of the ponding process in controlling the initiation of runoff at this site. Time-to-ponding was modeled as a function of the rainfall acceleration (i.e. the rate of change of rainfall intensity) and either the cumulative rainfall at the start of runoff or the soil–water deficit.

Measurements made by the OFDs provided physical insight into the spatial and temporal initiation of post-wildfire runoff during unsteady flow in response to time varying natural rainfall. They also provided data that can be telemetered and used to determine critical input parameters for hydrologic–runoff models.

\textbf{KEYWORDS:} unsteady overland flow; time-to-start of runoff; initial abstraction; wildfire; infiltrability; convective rainstorms; threshold; hydrologic model parameter

\section*{Introduction}

Knowing the time from the start of rainfall to the peak runoff discharge is central to forecasting floods. This is especially true in burned areas in steep mountainous terrain where land and emergency managers need advance warning and where the time is frequently shortened substantially by the effects of wildfire. Before runoff can begin, rainfall must first satisfy 'initial losses' related to interception, infiltration, surface depression storage, and any travel time from a source area. These initial losses are often referred to as the initial abstraction, $I_a$ (in millimeters), (Natural Resources Conservation Service, 2004; Springer and Hawkins, 2005), and $I_a$ represents a critical threshold for runoff generation. The rainfall minus $I_a$ is the excess rainfall (Figure 1), which has been used extensively in rain-runoff models (e.g. Henderson and Wooding, 1964; Woolhiser, 1975; Dunne and Dietrich, 1980; Julien and Moglen, 1990).

Historically, time-of-concentration, $T_c$ (in minutes), (Figure 1) rather than lag time, $T_l$ (in minutes), has been used to predict runoff volumes based on the unit hydrograph (Natural Resources Conservation Service, 2004, 2010). These two times are difficult to measure and uncertain because one must know either the centroid of mass of the excess rainfall or when the excess rainfall ends, and consequently many empirical equations have been proposed to estimate $T_c$. However, the time-to-start of runoff, $T_s$ (in minutes), (also referred to as the 'response lag', Carey and DeBeer, 2008) is precise, easy to measure, and is a quasi, real-time parameter because one does not need to know the entire hyetograph or hydrograph as is the case for $T_c$ and $T_l$ (Figure 1). The $T_s$ value is the time from the start of rainfall until the start of runoff, and is given by:

$$T_s = t_i + t_p + t_r + t_h,$$ (1)

where $t_i$ (in minutes) is the time for the rainfall to satisfy interception storage by vegetation and by surface litter and duff layers, $t_p$ (in minutes) is the time-to-ponding equal to the time for the initial infiltration to saturate the soil at the surface, $t_r$ (in minutes) is the time to fill surface depression storage on the irregular soil surface (Dunne, 1978; Darboux et al., 2002), and $t_h$ (in minutes) is the travel time from a source point of runoff to a downstream point. Much effort has gone into developing theoretical expressions for $t_i$ (Parlange and Smith 1976; Clothier and White, 1981; Diskin and Nazimov, 1996; Kunke and Mullins, 1997; Chu and Mariño, 2005; Assouline et al., 2007;
Figure 1. Conceptual diagram of rainfall-runoff parameters. $T_p$ is the time-to-peak measured from the start of the runoff. Time-to-start of runoff, $T_o$, equals the sum of the interception time, $t_i$, the ponding time, $t_p$, the time to fill surface depressions, $t_d$, and the travel time, $t_t$. For sources areas of runoff $t_i = 0$, and for burned areas $t_i$ can be assumed zero. The cumulative amount of rainfall, $R_o$ (gray-shaded rectangle in the upper figure) at $T_t$ is equal to the initial abstraction, $I_a$. At $T_r$ the rainfall intensity equals the infiltrability, $I_r$. Infiltration rate decreases with time shown by the white line in the upper figure. $T_t$ is the lag time from the centroid of the excess rainfall to the time of the peak discharge. $T_c$ is the time to concentration equal to the travel time from the hydrologically most distant point in a basin to the basin outlet. Modified after National Resources Conservation Service, 2004, 2010.

Xue and Gavin, 2008), but all require a priori knowledge of various soil hydraulic properties (such as minimum and maximum infiltrability, sorptivity, and saturated hydraulic conductivity). Additionally, the rainfall intensity often is constrained to be constant or slowly varying near the ponding time. Time to fill depression storage is a function of the surface roughness and the connectivity of the micro-topography (Antoine et al., 2011). Wildfires can change surface roughness, connectivity, and thus surface depression storage (Moody et al., 2013), and this can have a large impact on runoff (Stone et al., 1995). At runoff sources, $t_i = 0$, and in burned areas $t_i$ is reduced substantially by combustion (but is not zero, i.e. Mitsudera et al., 1984). Thus, for these conditions $T_r$ is essentially the sum of $t_p$ and $t_d$ (Figure 1). By measuring or predicting $T_r$, one can determine the critical runoff threshold, i.e. $I_a$, because $I_a$ equals the cumulative rainfall, $R_o$ (in millimeters), at the time, $T_c$.

Runoff models need to predict the time varying infiltration and the initial abstraction. Infiltration and excess rainfall in most models (e.g. Julien et al., 1995; Hydrologic Engineering Center Hydrologic Modeling System [HEC-HMS], 2000; CASC2D, 2014; KINEROS2, 2014; Water Erosion Prediction Project [WEPP], 2014; WRF-Hydro, 2014) are calculated using either Green and Ampt (1911), Mein-Larson (1973), or Parlangue et al. (1982) infiltration models. Most infiltration models were developed for agriculture settings and values for critical input variables such as the effective saturated hydraulic conductivity, $K_s$ (in mm h$^{-1}$) and wetting front suction, $S_f$ (in millimeters), are often unknown for mountainous soils affected by wildfire. The initial abstraction is usually assumed to be 20% of the maximum basin water storage, $S$ (in millimeters), when the curve number method is used (National Resources Conservation Service, 2004). However, publications list a wide range of values for $I_o$ from 5 to 99.6% of $S$ (Ponce and Sletty, 1995; Woodward et al., 2003; National Resources Conservation Service, 2004; Hawkins et al., 2010; Yuan et al., 2014). Appropriate curve numbers for burned areas are still uncertain (Springer and Hawkins, 2005; Foltz et al., 2009) and may, like $I_o$, also vary with time (Cydzik and Hogue, 2009). Another method is to estimate $I_o$ by calibrating the model using optimization and measured runoff (Cydzik and Hogue, 2009; Alonistioti et al., 2011; Yuan et al., 2014). For low-relief, unburned basins typical values of $I_o$ range from 2.5 to 50 mm (Carey and DeBeer, 2008; Huizinga, 2014) depending upon the environment. For an unburned, forested basin in the southern California mountains, typical calibrated values of $I_o$ ranged from 50.5 to 210 mm (Cydzik and Hogue, 2009).

Under ideal conditions of spatially uniform soil-hydraulic properties, the time-to-start of infiltration-excess overland flow at each point on a hillslope is simultaneous. This begins when the rainfall intensity exceeds a threshold equal to the maximum infiltrability, $I_r$ (in mm h$^{-1}$), which depends on initial soil–water content, $θ_i$ (in cm$^3$ cm$^{-3}$), and soil hydraulic properties (Smith, 2002; Liu et al., 2011). However, soil-hydraulic properties are rarely uniform, and the partial-area conceptual model introduced spatially variable infiltration properties such that runoff could start first down slope near stream channels (where the soil–water content might be higher) and then start later farther upslope (Betson, 1964). Field observations have indicated that under some conditions runoff started upslope, then at mid-slope but ‘was not initially continuous’ (Bryan et al., 1978, p. 156) or infiltrated before reaching ‘a belt extending at the base of the slope area’ (e.g. Yair et al., 1980, p. 243) adjacent to the channel. The complexity of runoff in semi-arid landscapes has been described qualitatively as being ‘generated patchily’ (Kirby, 2011, p. 3), as ‘potentially chaotic’ (Phillips, 1992, p. 191), or ‘scattered across the hillslope’ (Srinivasan et al., 2002, p. 649). Later, numerical models addressed some of this spatial complexity in infiltration patterns and in the resulting overland flow (e.g. Woolhiser et al., 1996). Overland flow on hillslopes has been described as a braided or anastomosing flow superimposed on a much slower moving thin film of sheet flow (Emmett, 1970; Dunne and Dietrich, 1980). The deeper, faster braided flow is between surface obstructions and within subtle topographic depressions, which have been referred as ‘micro-valleys’, ‘lateral concentrations’ (Emmett, 1970, pp. A13 and A27) or ‘microchannels’ (Smith and Goodrich, 2005, p. 1715). We referred to these as micro-drainages because they have no distinct banks but widths (between inflection points of profiles orthogonal to the flow) of 5 to 20 cm and depths of 1 to 10 cm. Sheet flow is confined to the interfluvial surfaces between micro-drainages and these surfaces will be referred to as hillslope facets.

In the past, measurements of overland flow have been made during constant rainfall simulations, steady-state equilibrium flow, or overland flow simulations. These have been made in laboratory flumes (e.g. Emmett, 1970; Raews, 1988; Bunte and Poesen, 1994; Abrahams et al., 2001); in the field using either rainfall simulation (e.g. Emmett, 1970; Dunne and Dietrich, 1980; Abrahams and Parsons, 1991; Gilley et al., 1992; Kinner and Moody, 2010) or overland flow simulation (e.g. Abrahams et al., 1986; Abrahams and Parsons, 1991; Parson et al., 1996; Sheridan et al., 2007; Robichaud et al., 2010; Nyman et al., 2013), and a few have used both types of simulations (e.g. Dunne and Aubry, 1986). Fewer measurements have been made during natural rain storms, which create unsteady flow (e.g. Esteves et al., 2000; Bartley et al., 2006; Bautista et al., 2007; Sen et al., 2010; Orchard, 2013), and no studies to our knowledge have been made on burned hillslopes during unsteady flow.

Few studies have focused specifically on measuring and predicting time-to-start of runoff but none in post-wildfire environments. Thus, the purpose of our research was to understand the spatial and temporal controls on the time-to-start of runoff and hence the initial abstraction for fire-affected soils during
temporally variable rainfall. To achieve this we: (1) developed an overland flow detector (OFD) to measure the time-to-start of runoff and initial runoff velocities, (2) investigated the spatial pattern of the initiation of runoff, and (3) used field observations to test two hypotheses that relate the time-to-start of runoff from fire-affected soils with rainfall characteristics and soil–water properties.

Methods

Field site

The research site was within an area burned by the 2010 Fourmile Canyon fire in the Front Range Mountains near Boulder, Colorado, USA. The site was a north-facing hillslope, which had burned at high intensity (Ebel et al., 2012; Moody and Ebel, 2012a, 2012b, 2013, Figure 1), and contained a small basin (8440 m²) (Figures 2 and 3) at an elevation of about 2400 m dominated by lodgepole pine (Pinus contorta). Two reasons for using this site were: (1) to ensure that a runoff response to unsteady rainfall could be measured, which would probably not be possible in an unburned site; (2) to take advantage of previous research results (Ebel et al., 2012; Moody and Ebel, 2012a, 2012b, 2013).

Gravely sand soils in the site are derived from a bedrock geology that is primarily Boulder granodiorite. They are within the Allens Park member of the Fern Cliff–Allens Park–Rock outcrop complex (Moreland and Moreland, 1975), which are frigid Lamellic and typic Haplustalfs (USDA, 2010; Ebel et al., 2012). In 2011, K₆ was estimated using an inverse method and a one-dimensional (1D) infiltration model (Moody and Ebel, 2013). The geometric mean value of K₆ for these soils after the wildfire was 1.8 and 0.44 mm h⁻¹ for the upper layer (4–9 mm) and lower layer (>9 mm), respectively (Moody and Ebel, 2013).

The area surrounding the research site has a Continental climate (Pepin, 2000) where the precipitation is primarily a mix of cyclonic storms in the spring and fall, convective storms in the summer, and snowstorms in late fall through early spring. Convective storms during the summer are typically high-intensity (>25 mm h⁻¹), short-duration rainfall (10–60 minutes) from storms with monsoon moisture originating from the Gulf of California (Ebel et al., 2012; Douglas...
et al., 2004). These storms often consist of a series of cells defined as continuous rainfall with intervals of ‘no rain’ lasting less than one hour (Moody and Ebel, 2013), and are designated with an A, B, etc. after the date.

Overland flow detection

The OFDs were electrical resistor-type instruments that were relatively inexpensive (c. ~$150 in 2011) and simple to construct. They output a voltage (0–2.5 V) and were designed to measure the time of runoff initiation and the runoff hydrograph during natural rainstorms. Rainstorm detectors have been used to measure the timing or onset (i.e. change from ‘no flow’ to ‘flow’) of streamflow (Blasch et al., 2002; Srinivasan et al., 2002; Goulsbra et al., 2009) in ephemeral channels. The current model of the OFD represents a modification of an earlier model designed for and used by Schmidt et al. (2011) to measure water depths. These detectors differ from the ‘flow–no flow’ type and the type that capture a small sample volume (Kirkby et al., 1976; Zimmermann et al., 2014) by measuring a time series signal that is proportional to the runoff depths.

Technical description

OFDs recorded the time-to-start of runoff as a rapid voltage drop. Detectors had three common electrodes (165-mm long and threaded at one end) that also served as support ‘legs’ and one sensing electrode (127-mm long). All electrodes were 3.2-mm diameter stainless steel rods. Common electrodes were mounted in a flat, ‘3-inch’, plastic (PVC) bottom cap at the vertices of an equilateral triangle (Figure 4B, 71.1-mm diameter bolt circle) and electrically connected together. The sensing electrode was mounted in the center of the triangle and connected to a voltage pulse generated by the four-channel data logger (model U12-006, Onset Corp., Bourne, MA). This pulse (2.5 V) was impressed across a half-bridge circuit (upper part was a fixed ~10 kΩ resistor) and the measured voltage across the electrodes generated an electrical signal related to the water depth. When the sensing electrode was not in contact with water, the base voltage was 2.5 V, and when it was in contact the voltage drop was proportional to the water depth. Laboratory calibration indicated that the water depth was a linear function of the voltage drop ($R^2$ values ranged from 0.93 to 0.96 for water depths < 10 mm and from 0.97 to 0.99 for depths < 5 mm). The proportionality constant depends on the specific conductance of the water. We did not measure the specific conductance of the runoff water for each storm so we could not compute absolute water depths, but illustrate the hydrograph as a voltage-drop ‘hydrograph’, for which relative changes during each storm represent relative changes in water depth.

Field deployment

Six OFDs where deployed in two groups of three detectors in micro-drainages on the burned hillslope. The upper group (OFDs 1, 2, and 3) was located in micro-drainages with slopes ranging from 0.25 to 0.29 (Figure 3A, Table I). These micro-drainages were relatively wide (100–120 cm) and shallow (4–5 cm) at the location of the OFDs, and the bed had a relatively uniform surface texture of fine to coarse sand with only occasionally roots crossing the micro-drainage. Originally, the detectors were placed on two different micro-drainages whose flow lines joined and contributed flow to a ‘1-inch’ Parshall flume (1–3 in Figures 2 (upper) and 3). Later OFD#1 was relocated to the same micro-drainage as OFD#2 and OFD#3 to provide an additional pair of points for measuring velocity. The lower group (OFDs 4, 5, and 6) was located along a single micro-drainage (slopes ranging from 0.26 to 0.51) that contributed flow to a ‘1-inch’ Parshall flume (1–1 in Figure 3). This micro-drainage tended to be narrower (60–160 cm) and deeper (2–12 cm) than the upper micro-drainages. The bed was rougher with outcrops of cobbles and roots creating a ‘step-pool’ system on the order of 2 to 5 m

Figure 4. Overland flow detector-model 3. This model is self contained with the data logger (U12-006 Onset corp.) inside a 3-inch PVC housing and therefore avoids having to run wires across the hillslope. It has three (3-mm diameter, stainless steel rods) ‘legs’ for support, which penetrate the ground and one 3-2-mm diameter stainless steel sensing electrode that is deployed to be about 1 mm above the ground surface. A 76-mm piece of PVC (or ABS) 3 inch diameter pipe was inserted and sealed into the bottom cap to house the data logger (not water proof or water resistant), and a domed, 3-inch PVC cap completed the water proof overland flow detector. This figure is available in colour online at wileyonlinelibrary.com/journal/esp

Table 1. Downhill location of overland flow detectors (OFDs) on the hillslope

<table>
<thead>
<tr>
<th>OFD#</th>
<th>Distance above flume (m)</th>
<th>Elevation (m)</th>
<th>OFD pair separation distance (m)</th>
<th>Slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>1*</td>
<td>4.9</td>
<td>2394.54</td>
<td>—</td>
<td>0.27</td>
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<tr>
<td>flume</td>
<td>0.0</td>
<td>2393.22</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>1*</td>
<td>18.8</td>
<td>2398.28</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>3</td>
<td>10.4</td>
<td>2396.18</td>
<td>1.3</td>
<td>0.25</td>
</tr>
<tr>
<td>2</td>
<td>5.2</td>
<td>2394.67</td>
<td>3.2</td>
<td>0.29</td>
</tr>
<tr>
<td>flume</td>
<td>0.0</td>
<td>2393.22</td>
<td>—</td>
<td>0.28</td>
</tr>
<tr>
<td>6</td>
<td>35.5</td>
<td>2381.43</td>
<td>6.4</td>
<td>21.0</td>
</tr>
<tr>
<td>5</td>
<td>24.6</td>
<td>2375.86</td>
<td>6.5</td>
<td>10.9</td>
</tr>
<tr>
<td>4</td>
<td>14.5</td>
<td>2373.24</td>
<td>5.4</td>
<td>10.1</td>
</tr>
<tr>
<td>flume</td>
<td>0.0</td>
<td>2366.47</td>
<td>—</td>
<td>0.47</td>
</tr>
</tbody>
</table>

Note: OFDs deployed in the area burned by the 2010 Fourmile Canyon fire were checked, downloaded, and re-launched every day; the sampling interval was one second.

*Initially this detector was by itself in a separate micro-drainage 4.9 m uphill from the flume 1–3, but was moved 7.4 m uphill from OFD#3 on 8 July 2012.
between steps. OFD#5 was located at the confluence with another micro-drainage. These micro-drainages contributed to a first-order hillslope channel gaged by a ‘3-inch’ modified Parshall flume (3–1 in Figure 3).

Sampling interval was set to one second to adequately resolve the initiation of flow and flow velocities. The clock for each data logger was reset each day to the same identical time (within one second using a global positioning system [GPS] satellite signal). Typical hillslope velocities over bare soil could be on the order of 5 to 50 cm s$^{-1}$ and travel times between pairs of OFDs could range from 10 to 200 seconds (Table 1) so that by using this sampling interval the uncertainty of the velocity was $\pm 5$–10%. The sampling interval for the flumes was 10 seconds such that the uncertainty of the velocity between the last detector and the flume was $\pm 5$–100% and these velocities were not calculated. For this reason OFD#1 (initially alone in a single micro-drainage) only provided information on the initiation of runoff and no estimates of overland flow velocities. At a one second sampling interval the memory of the data logger (U12-006, Onset Corp.) was filled in about 12 hours. Therefore, the OFDs were downloaded and re-launched each day in the late morning so that they could record the afternoon and evening runoff from summer convective storms. Daily downloading and re-launching served to determine the quality of the data by inspecting the detectors, cleaning off trash that may have collected during runoff, and re-positioning the sensing electrode (to be within 1 mm of the bed of the micro-drainage) to compensate for any soil erosion under the electrode or soil deposition around the electrode.

Ancillary data

Soil–water content

Surface soil–water content was measured daily by collecting soil cores. Four soil cores (4.7-cm diameter, 1.5-cm long) were collected each morning from a small area (0.1 m $\times$ 0.1 m) within a larger area (2 m $\times$ 2 m) at two locations near rain gages 669624 and 10236 (Figure 3), placed in soil cans, sealed, and the volumetric soil–water content, $\theta$ (in cm$^3$ cm$^{-3}$) was determined thermogravimetrically in the laboratory (Topp and Ferré, 2002). Subsurface arrays measured temperature and $\theta$ (model STE, Decagon Devices, Pullman, WA) every minute at four depths (0.05, 0.10, 0.15, and 0.20 m), and had been deployed since 2010. When these sensors were deployed, soil samples were collected to determine saturated soil–water content, $\theta_r$ (in cm$^3$ cm$^{-3}$) which was 0.49 cm$^3$ cm$^{-3}$ and the van Genuchten soil hydraulic parameters (van Genuchten et al., 1991; Ebel et al., 2012; Moody and Ebel, 2013). These soil cores provide an estimate of the initial soil–water content, $\theta_i$ (in cm$^3$ cm$^{-3}$), for the first rain cell A (Table II). For succeeding storm cells (B and C) during a convective storm, $\theta_i$ was estimated by using the Hydrus-1D numerical infiltration model (Simunek et al., 2008; Moody and Ebel, 2013). This model requires the initial subsurface soil–water profile at the start of rainfall, $\theta_{r_0}$ van Genuchten soil-hydraulic parameters, and the observed hydrograph near the sub-surface sensor array (see Moody and Ebel, 2013, for details). The observed hydrograph was recorded by the runoff gages associated with the rain gages at 669624 and 10236 (Figure 3; Moody and Ebel, 2012b).

Rainfall characteristics

Three rainfall characteristics (cumulative rainfall, rainfall intensity, and rainfall acceleration) were computed from data collected by three rain gages. These were 15-cm diameter, tipping-bucket gages (see locations in Figure 3) with each tip equal to 0.254 mm, a data logger that recorded the time of each tip (HOBO, Onset Corp.), and installed ~1 m above the ground. They were inspected, downloaded, and re-launched about every 14 days to ensure that there were no obstructions and that the total clock drift (~2–3 seconds day$^{-1}$) was minimal. One-minute rainfall intensities, $I_1$ (in mm h$^{-1}$), were computed by first interpolating the irregular time series of ‘tip times’ to a regular interval of 0.1 minute, and then computing a one-minute backwards difference from the interpolated cumulative rainfall values. The rainfall acceleration, $a$ (in mm h$^{-2}$), was the slope of the least-squares standard linear regression line ($I_i(t)$ versus $t$) from the start of rainfall at $t = 0$ when $I_1(0) = 0$ to the time of peak rainfall intensity (Moody and Ebel, 2013).

Time-to-start of runoff

Time-to-start of runoff in the micro-drainages was measured by the OFDs and on hillslope facets it was measured at the two different infiltrometers. Each difference infiltrometer (669624 and 10326 in Figure 3) consisted of a tipping-bucket rain gage and a runoff plot (radius ~ 0.25 m) with a runoff gage. Because the runoff plots were small, the travel time, $t_p$ and the time-to-fill the surface depressions, $t_d$ were negligible compared to the time-to-ponding, $t_3$ (Moody and Ebel, 2012b, 2013). Thus, measurements of $t_3$ provide estimates of $T_r$ on the hillslope for comparison with $T_r$ values in the micro-drainages.

Overland flow velocities

Overland flow velocities were simulated (during periods with no rain) in micro-drainages on the same hillslope where the OFDs were deployed. Water was released from a 20-l jug at a maximum rate of 0.05 l s$^{-1}$ and velocity was computed from the travel time of dye or particles in the water between two points separated by a known distance. These velocity measurements were used for comparison with the OFD flow velocities and with flow velocities measured using the same method in the area burned by the 2005 Harvard Fire near Burbank, California (Table III).

Analysis

Analysis was limited by the number of the convective rainstorms during the summer of 2012. Most storms were during two narrow windows from 5 to 12 July and from 27 to 30 July. The dataset consisted of hydrographs generated by 10 separate storm cells on 5, 6, 7, 8, 12, 27, 29 and 30 July 2012 (Table II).

Observed time-to-start of runoff

Observed time-of-start of runoff, $T_r$, was the elapsed time from the start of rainfall to the start of runoff. Clock time for the start of runoff was measured by the OFDs to within one second; however, clock time for the start of rainfall is not known when tipping-bucket rain gages are used – only the time of the first tip. These start times were estimated by linearly extrapolating back in time, using data for the first three tips. Median value of the 90 and 95% uncertainty limits for the 20 estimated times for the start of rainfall were $\pm 1.0$ and $\pm 2.0$ minutes, respectively. Cumulative rainfall was digitized at 0.1 minute interval. Cumulative rainfall at the start of runoff was determined by interpolation between the 0.1-minute intervals if necessary, and one-minute rainfall intensities, $I_1$, were computed using these 0.1-minute values.
## Table II. Rain and soil characteristics and runoff characteristics measured by overland flow detectors (OFDs) during eight rain storms

<table>
<thead>
<tr>
<th>OFD#</th>
<th>Start of rain</th>
<th>Duration of rain</th>
<th>Maximum one-minute intensity</th>
<th>Intensity at start of runoff, $i_0$</th>
<th>Initial water content at start of runoff, $θ_i$</th>
<th>Time-to-ponding, $t_p$</th>
<th>Cumulative rainfall at start of runoff, $R_p$</th>
<th>Observed, $T_r$</th>
<th>Travel time, $t_t$</th>
<th>Predicted, $t_p$</th>
<th>Time of first peak OFD pair</th>
<th>Leading edge of water</th>
<th>First peak</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>(minutes)</td>
<td>(mm h⁻¹)</td>
<td>(mm h⁻¹)</td>
<td>(mm h⁻¹)</td>
<td>(cm³ cm⁻³)</td>
<td>(minutes)</td>
<td>(mm)</td>
<td>(minutes)</td>
<td>(minutes)</td>
<td>(minutes)</td>
<td>(minutes)</td>
<td>(minutes)</td>
<td>(minutes)</td>
</tr>
<tr>
<td>5A July 2012</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>16:01:10</td>
<td>43</td>
<td>101</td>
<td>31·3</td>
<td>290</td>
<td>0·015</td>
<td>3·5</td>
<td>16:02:11</td>
<td>0·41</td>
<td>1·0</td>
<td>2·2</td>
<td>16:16:53</td>
<td>—</td>
</tr>
<tr>
<td>2</td>
<td>16:01:10</td>
<td>43</td>
<td>101</td>
<td>30·8</td>
<td>290</td>
<td>0·015</td>
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12B July 2012

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<th>Intensity at start of runoff</th>
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<th>Initial water content at start of rain, ( \theta_i )</th>
<th>Time-to-ponding, ( t_p )</th>
<th>Start of Runoff</th>
<th>Cumulative rainfall at start of runoff, ( R_p )</th>
<th>Observed, ( T_r )</th>
<th>Travel time, ( t )</th>
<th>Predicted, ( t_p )</th>
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<td>7.2</td>
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<td>5.3</td>
<td>0.4</td>
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<td>6.5</td>
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<td>1.5</td>
<td>16:06:56</td>
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Note: saturated for 12B July 2012 indicates that the soil was saturated from rain cell A and enough soil moisture was still present to indicate runoff had started before the rain cell B started; soil-water content at saturation was 0.49 cm$^3$ cm$^{-3}$, Moody and Ebel, 2013.
Predicted time-to-start of runoff

The \( T_p \) value depends partly on time-to-ponding, \( t_p \). Time-to-ponding at the point scale (~1 m²) on burned hillslopes facets depends on rainfall characteristics more than on fire-affected soil hydraulic properties (Moody and Ebel, 2013). For convective rainfall, an empirical relation was found to predict \( t_p \) on burned hillslopes (Moody and Ebel, 2013):

\[
T_p = \left( \frac{R_p}{a} \right)^{1/2}
\]

Substituting Equation (2) into Equation (1) provides one relation between \( T_p \) and rainfall characteristics with the initial assumptions that \( t_p \) and \( t_i \) are zero. Thus \( R_p \) is a posteriori variable known only after runoff starts. If it is assumed to depend on the soil-saturation deficit, \( (\theta_s - \theta_i) \) in cm³ cm⁻³, then site-specific relations can be determined to predict the cumulative rainfall, \( R_p \) based on a priori variables. Infiltration and runoff are usually non-linear functions of \( \theta_s - \theta_i \) (often exponential e.g. Horton, 1939; Smith, 2002), thus data collected from the two difference infiltrometers in 2011 (Moody and Ebel, 2012b, 2013) were used to determine the following empirical soil-saturation deficit equations to predict \( R_p \):

\[\begin{align*}
2011 \text{ at } 669824 & : R_p = 0.93e^{0.68(\theta_s-\theta_i)}, \quad R^2 = 0.76 \quad (3a) \\
2011 \text{ at } 10236 & : R_p = 0.59e^{2.96(\theta_s-\theta_i)}, \quad R^2 = 0.72 \quad (3b)
\end{align*}\]

Equation (3a) would apply to OFDs 1, 2, and 3 and equation (3b) to OFDs 4, 5, and 6. Substituting the appropriate soil-saturation deficit equation into Equation (1) provides a second relation between \( T_p \) and rainfall characteristics and soil properties.

Flow velocities

Flow velocities in the micro-drainages were calculated from the separation distance between detectors (Table I) and the time difference in arrival of one of two distinct features of the hydrograph. One feature was the relatively rapid initial rise and the second was the first peak (Peak, Table II) in the hydrograph. Several OFD hydrographs had multiple peaks, but peaks after the first peak were not used to determine the velocity because of the possibility of deposition of debris on or erosion beneath the sensing electrode after the first peak. Velocities were only computed between consecutive downhill pairs of detectors (i.e. 1–3, 3–2, 6–5, and 5–4).

Table III. Simulated overland depth and flow velocities

<table>
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<tr>
<th>Source</th>
<th>Conditions</th>
<th>Slope</th>
<th>Depth (mm)</th>
<th>Velocities (cm s⁻¹)</th>
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<td>Abrahams et al., 1986</td>
<td>Desert soils</td>
<td>0.092–0.687</td>
<td>1.1–7.7</td>
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<td>Bunte and Posen, 1994</td>
<td>Bare soils in flume</td>
<td>0.014</td>
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<td>Emmett, 1970</td>
<td>Semi-arid soils</td>
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<td>0.2–46</td>
<td>0.18–12</td>
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<td>Nyman et al., 2013</td>
<td>Burned soils</td>
<td>0.36–0.40</td>
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<td>62–114</td>
</tr>
<tr>
<td>Robichaud et al., 2010</td>
<td>Low burn severity</td>
<td>0.24–0.64</td>
<td>6.3 ± 2.2</td>
<td>7.3 ± 5.8</td>
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<td>Robichaud et al., 2010</td>
<td>High burn severity</td>
<td>0.23–0.75</td>
<td>6.5 ± 2.0</td>
<td>31 ± 12</td>
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Results

Rain and soil–water content

The summer monsoon season was brief in 2012 lasting only from 5 July through 30 July 2012 with essentially no rainfall greater than 5 mm day⁻¹ in August. Eight days in July had at least 5 mm of rainfall and average rainfall intensities > 5 mm h⁻¹. These eight days (5, 6, 7, 8, 12, 27, 29, and 30 July) had 10 storm cells with median duration of 25 minutes; a median, maximum one-minute rainfall intensity of 48 mm h⁻¹; and a median rainfall acceleration of 195 mm h⁻² (Table II). Each storm cell had multiple one-minute intensity peaks (Figure 5) with 7B July having five distinct peaks, but the rainfall was spatially uniform over the research area (Figure 3). Near surface (~1–2 cm) soil–water content, \( \theta_i \) at the beginning of the monsoon rain season on 5 July 2012 was quite dry (0.015–0.018 cm³ cm⁻³; Table II), increased during the first week of rain to a maximum of (0.459 cm³ cm⁻³) before rain cell 12B July, and then decreased during the dry period from 13 July to 27 July 2012 down to 0.032–0.037 cm³ cm⁻³.

Hydrographs

All OFDs recorded a hydrograph for each storm cell, and the shape of the hydrograph resembled the hyetograph. The rising limb of the hydrographs was relatively gradual and not abrupt as would be expected for the sudden arrival of a flood wave, and there was no obvious dispersion of the peak at the hydrograph downhill. There were exceptions like the sudden abrupt peak for OFD#3 during the 12A July 2012 rain cell (Figure 5A). The falling limb sometimes did not return to zero water depth because small debris caught on the electrode or soil moisture at the surface continued for a while to provide an electrical pathway, which produced a voltage drop. This was one of the reasons for inspecting the OFD detectors daily in order to remove the debris and re-deploy the detectors.

Time-to-start of runoff

The spatial distribution of runoff was complex. Runoff did not start simultaneously on the hillslope and in the micro-drainages. Linear regression between the observed \( T_p \) on hillslope facets (\( T_i = t_p \), Table II) and the observed \( T_i \) in the micro-drainages (Table II) indicated the \( T_i \) on the hillslope was about 20% shorter (\( R^2 = 0.76 \)) than in the micro-drainages. Observed \( T_i \) in the micro-drainages ranged over two orders of magnitude from 0.3 to 79.4 minutes (Table II). In general, \( T_i \) in the micro-drainages were inversely correlated (power law, \( R^2 = 0.54 \)) with rainfall acceleration, \( a \), reflecting the
explained 96% of the variance in observed values of $T_r$ (one-second voltage drop expressed as approximate water depth).

The cumulative-rainfall relation (Equations (1) and (2)) explained 80% of the variance in $I_a$ for values of $t > 2$ minutes (Figure 6):

$$\text{Observed } T_r = 1.33 \text{ Predicted } T_r - 0.58; \quad R^2 = 0.96 \quad (4)$$

The soil-saturation deficit relation (Equations (1)–(3)) explained 80% of the variance in $T_r$:

$$\text{Observed } T_r = 1.91 \text{ Predicted } T_r - 5.29; \quad R^2 = 0.80 \quad (5)$$

and on average it also under predicted $T_r$. The 95% confidence limits for the slope and intercepts for these equations were $\pm 0.06$ and $\pm 0.22$ and $\pm 1.02$ and $\pm 2.67$ minutes, respectively, so that the intercept for the cumulative-rainfall relation (Equation (4)) is essentially zero, whereas the intercept for the soil–water deficit relation (Equation (5)) is not. This bias in the soil–water deficit relation is reflected in the predicted values for $T_r < 2$ minutes, which were over predicted by a factor of 4 to 10 (Figure 6). The largest over predictions correspond to two different extreme conditions (data points are circled in Figure 6) when $\theta_i$ was the driest (5A July 2012, Table II) and when $a$ was the largest (30A July 2012, Table II). The 95% confidence limits for the start of the rainfall are $\pm 2$ minutes, and one could justify using only the data for $T_r > 2$ minutes to determine the empirical relations for Equations (4) and (5). This was done and Equations (4) and (5) were essentially unchanged with nearly identical slopes, intercepts, and $R^2$ values.

**Figure 6.** Comparison of predicted and observed time-to-start of runoff, $T_r$. The dotted line is the line of perfect agreement. Predicted values are based on the cumulative rain at start of runoff (solid line) and on an empirical soil–saturation deficit approximation of the cumulative rainfall (dashed line) based on 2011 data (Equation (3a)). The encircled data represents two extremes when the initial soil–water content, $\theta_i$, was the driest (0.015–0.018 cm$^{-3}$), 5A July 2012 and when the rainfall acceleration, $a$, was the largest (1680–1970 mm h$^{-2}$, 30A July 2012).

**Initial abstraction, $I_a$, and maximum infiltrability, $I_c$**

When travel time is zero, $I_a$ equals measured values of $R_a$ at $T_r$. The median value of $I_a$ was 0.7 mm for the upper group of OFDs (1, 2, and 3), 1.0 mm for the lower group of OFDs (4, 5, and 6), and the range for the combined groups was from 0.1 to 5.8 mm. Differences in $I_a$ between the upper and lower groups of OFDs for each storm cell were not significantly different ($p = 0.33$). So the mean value of $I_a$ was 1.2 mm (coefficient of variation of 0.94) and is an estimate of the threshold.
for runoff initiation at this burned site. These values are usually much lower than typical threshold values published for unburned low-relief areas (2·5 to 50 mm; Carey and DeBeer, 2008; Huizinga, 2014) and for mountainous areas (9·6 mm; Cydzik and Hogue, 2009). Few measurements of $I_p$ exist for burned areas, but for comparison, a ‘conservative’ lower limit of 1·0 mm was assumed by Elliott et al. (2005) to model runoff using the curve number method for the 2002 Hayman fire (also in the Front Range Mountains of Colorado). This was based on a minimum rainfall value observed to generate runoff after wildfires. In contrast, a post-wildfire model calibration value of 19·6 mm was reported by Cydzik and Hogue (2009) for the first year after the 2003 Old fire in the southern California mountains.

The maximum infiltrability can also be determined by knowing $T_r$ when $t_i$ = 0, because the rainfall intensity $I_i(t)$ at $T_r$ equals $I_c$. Median values of $I_c$ were 17·1 and 17·3 mm h$^{-1}$ for the upper and lower group of OFDs, respectively. These, like $I_p$, were not significantly different (two-tail, $p = 0·88$) between the upper and lower sites, so the mean value of $I_c$ was 18·9 mm h$^{-1}$ with a coefficient of variation of 0·60. This infiltrability threshold is often expressed as ‘rainfall-intensity threshold’ and published values for burned areas range from 8·5 to 20 mm h$^{-1}$ (Moody and Martin, 2009).

Flow velocities

Overland flow velocities measured during simulations on hillslope facies and in micro-drainages on the hillslope were comparable. Those on the hillslope facies ranged from 8 to 21 cm s$^{-1}$ and those in the micro-drainages ranged from 7 to 14 cm s$^{-1}$ (Table III). Velocities in micro-drainages were not detected by the OFDs for all storm cells. For 50% of the storm cells the leading edge of the water had velocities that ranged from 1 to 93 cm s$^{-1}$ (Table II) with median values of 5·8 and 8·9 cm s$^{-1}$ for the upper and lower group of detectors. For 78% of the storm cells the hydrograph peak had velocities ranging from 3 to 52 cm s$^{-1}$ (Table II) with median values of 12·8 and 9·4 cm s$^{-1}$ for the upper and lower group, respectively.

Discussion

Initiation of runoff

Based on these few OFD, the initiation of runoff appears to be complex, starting at different times at different locations during different storm cells. Runoff from storm cells started first in micro-drainages (69%) before it started on hillslope facies (31%). Runoff was initiated first at an upstream detector 56%, at the mid-stream detector 6%, and at the downstream detector 38% of the time. As an example of this complex response, flow began first ($T_r$ = 2·1 minutes) during the 5A July 2012 storm cell at OFD#6, then on the lower hillslope facet ($T_r$ = 2·8 minutes), and finally ($T_r$ = 5·6 minutes) at the most downstream OFD#4, which included 3·5 minutes of travel time from OFD#6. A similar pattern of complex response was observed by Kinner and Moody (2010) during rainfall simulations on 1-m$^2$ plots on hillslopes burned by the 2003 Overland fire in Colorado. They noted ponding first in topographic lows at different locations along the micro-drainage network, which often expanded in the uphill and downhill directions and eventually became connected to form a single body of water flowing downstream.

Whether or not runoff started first on hillslope facets or in micro-drainages did not depend on $\theta_i$ ($R^2 < 0·1$). However, $\theta_i$ was measured at only two sites (669824 and 10236, Figure 3) on the hillslope and not at OFD sites in the micro-drainages. Thus, one cannot rule out $\theta_i$ at the OFD sites from affecting $T_r$. This suggests that even at the relatively small scale of this hillslope (~100 m × 200 m), $\theta_i$ cannot be characterized by two values. Differences in $\theta_i$ between hillslope and micro-drainage may represent first-order effects, whereas differences on a hillslope and differences within a micro-drainage may represent second-order effects on $T_r$. This is partially supported by the similarity of the two values of $\theta_i$ at the start of rain on hillslope facets (Table II) for which the average difference is only 0·046 cm$^3$ cm$^{-3}$. For $t_i > 0$ (27%), the travel time correlated ($R^2 = 0·39$) inversely with a suggesting that the travel time component of $T_r$ may be more important for storms having high rainfall acceleration.

Soil–saturation deficit relation

The soil–saturation deficit relation explained 80% of the variance in the observed values of $T_r$ when travel times are included (Equation (5)). If only data for $t_i = 0$ is considered, then the corresponding predictive relation for $T_r$ could be used, with appropriate values of $\theta_i$ and $\theta_s$, to determine the initial abstraction from a hyetograph as an alternative to deploying OFDs. This relation is similar to Equation (5), under predicts for $T_r > ~5$ minutes, and over predicts for $T_r < ~5$ minutes:

$$\text{Observed } T_r = 2·03 \text{ Predicted } T_r - 5·47, \quad R^2 = 0·79, \quad \text{for } t_i = 0 \tag{6}$$

where the 95% confidence limits for the slope and intercept are ± 0·28 and ± 3·30 minutes, respectively. Saturated $\theta_i$ was assumed to be 0·49 cm$^3$ cm$^{-3}$, but varying $\theta_s$ by ± 0·05 cm$^3$ cm$^{-3}$ only changed the slope in Equation (6) by about ± 6%.

Over prediction is reflected by the non-zero intercept, and there are several possible explanations. First, Equations (3a) and (3b) relating $R_p$ to $\theta_i - \theta_s$ were developed from data collected at the two difference infiltrometer sites and not at each of the six OFD sites. Not surprisingly, these equations predict $R_p$ at the difference infiltrometer sites (669824 and 10236) better ($R^2 = 0·61$ and 0·26, respectively) in 2012 than they predict $R_p$ at the OFD sites ($R^2 < 0·01$). Additionally, Equation (3a) always predicts the same value of $R_p$ (i.e. the value associated with the single $\theta_i - \theta_s$ value at the difference infiltrometer site) for each of the three upper OFD sites, whereas given the spatial variability of soil properties, the values probably differ. Same is true for Equation (3b) and the three lower OFD sites.

Two unusual conditions may also explain why the soil–water deficit equation (Equations (5) and (6)) over predicts $T_r$. One was the unusually dry soil conditions (0·015–0·018 cm$^3$ cm$^{-3}$) on 5A July 2012 (Figure 6) that may have caused the reappearance of water repellency (Doerr and Thomas, 2000; Huffman et al., 2001). This would reduce infiltration and explain the reduction in $I_c$ from the predicted value of 3·7 mm to the observed values of 0·41, 0·28 and 0·74 mm (Table III). The other condition was the extreme values of rainfall acceleration (1680 and 1970 mm h$^{-1}$) on 30A July 2012 (Figure 6). The soil-deficit equations were developed for rainfall–runoff conditions for which 90% of the rainfall accelerations were less than 1000 mm h$^{-1}$ and thus may not be applicable for these extreme rainfall conditions.

Depression storage

Initial abstraction must satisfy the initial infiltration until ponding and the depression storage. A subset of the OFD
measurements, used to determine Equation (4), can be used to
determine another empirical relation to predict $T_i$ when $t_i = 0$,
and thus, determine the value of $R_p$ that equals $I_c$. The relation
is nearly the same as Equation (4):

$\text{Observed } T_i = 1.46 \text{ Predicted } T_i = 0.38, R^2 = 0.98 \text{ for } t_i = 0$

(7)

with 95% confidence limits for the slope and intercept of
$\pm 0.05$ and $\pm 0.75$ minutes. This relation also, on average,
under predicts the observed values of $T_i$, but still over pre-
dicts values for $T_i < 2$ minutes. However, time to fill depres-
sion storage, $t_c$ was assumed to be zero in determining
Equation (7), but it may account for some of the time differ-
ence between observed and predicted values. While we did
not measure $t_c$ directly, an order of magnitude estimate can
be made to assess whether it is potentially a significant
process in determining $T_i$ and thus $I_c$. To do this, high-
resolution longitudinal profiles of the micro-drainages were
extracted from a data set collected using a tripod laser scan-
ning system (Rengers et al., 2012). These profiles showed
that about 10% of the micro-drainages consisted of depres-
sions with typical depths, $d$ (in centimeters), on the order
of 3 cm, lengths ranging from 15 to 45 cm, and local bed
slopes that were either flat or negative. Estimates of the time
($t'_c = \sqrt{(2d/g)}$) for rainfall to fill these depressions were on
the order of 10 to 200 minutes, and much longer than the
time difference. Whereas, estimates of the time, $t'_c$ for flow
in the micro-drainages (~0.5 cm deep) to fill these depres-
sions gave 0.4 to 1.1 minutes assuming a velocity of ~6.9
$\text{cm s}^{-1}$. Thus, filling surface depressions appears to account
for most of the time difference between observed and pre-
dicted values, but in general, it is an order of magnitude less
than the ponding times, $t_c$ (Table II). A more rigorous inves-
tigation would require high-resolution digital elevation
models of hillslopes and micro-drainages to determine the
spatial distribution and volumes of surface depressions.

Flow velocities

OFD measurements of unsteady flow velocities during rainstorms
are all within the range of steady flow velocities published in the
literature for burned or bare soils (Table III). A few maximum ve-
celocities of the leading edge of the flow (93 and 50 $\text{cm s}^{-1}$, upper
group, 12A July 2012; and 73 $\text{cm s}^{-1}$, lower group, 8A July 2012,
Table III) appear to be outliers. They may represent the situation
when ponding at the upstream detector occurred just before
ponding at the downstream detector rather than representing
flow in the micro-drainage between detectors.

The maximum velocity of the hydrographic peak was 52 $\text{cm s}^{-1}$
(upper group) and the relatively high peak velocities of 24 and
34 $\text{cm s}^{-1}$ (lower group) were all associated with one of the
most intense storm cells (5A July storm cell, maximum
one-minute intensity of 101 to 104 $\text{mm h}^{-1}$, Table III) during the
entire summer. These velocities are comparable with those
reported by Nyman et al. (2013) and Robichaud et al. (2010)
for burned soils. Thus, these OFD measurements of unsteady
flow velocities during convective rain storms are physically re-
alistic. Moreover, these and future OFD travel time measure-
ments can be used with confidence to verify hydrologic
models of unsteady runoff, and thus improve predictions of
post-wildfire runoff.

Applications

Uncertainty exists in selecting values of model parameters to
account for the effects of wildfire. OFD measurements made
in burned areas can provide estimates of critical input parame-
ters for hydrologic models and can be used to verify these
models. In addition to the measurements of initial abstraction
and travel times, OFD measurements can be used to deter-
mine the critical rainfall intensity, $i_c$ (in $\text{mm h}^{-1}$), at the start
of runoff (Table II, column 6). This intensity can be used to es-
timate critical soil variables for infiltration models such as $K_s$
(given the wetting-front suction, $S_f$) or $S_f$ (given $K_s$). The value
of $S_f$ is probably the more important variable because it is more
difficult to measure than $K_s$, which has been reported for fire-
affected soils in the literature (Robichaud, 2000; Kinner and
Moody, 2010; Nyman et al., 2010; Beatty and Smith, 2013;
Moody and Ebel, 2013). Normally, infiltration models must
solve for time of ponding or runoff, but OFD records this critical
time. Thus, assuming that momentarily just after the start of run-
off the infiltrability, $I_c$ (in $\text{mm h}^{-1}$), will equal $i_c$, then the
Green–Ampt equation (HEC-HMS, 2000) can be written as:

$I_c = i_c = K_s \left[ 1 + \frac{\theta_i - \theta_f}{R_p} \right]$

(8)

This can be solved for the wetting-front suction, giving

$S_f = \frac{R_p}{(\theta_i - \theta_f)} \left[ i_c - \frac{1}{K_s} \right]$

(9)

To illustrate this application, we used the value of $K_s$ (1.8 $\text{mm h}^{-1}$)
given earlier for the upper layer of soil, and assumed that
$\theta_i$ at the OFDs was equal to that at the difference infiltrometers.
The computed geometric mean value of $S_f$ was 32 mm, which is similar to that reported for sand (49.5 mm) by Rawls et al.
(1993) and physically reasonable given the gravelly nature of
the soils.

Knowledge of the excess rainfall, and hence the runoff,
hinges on the accurate prediction of infiltration, which is a
non-linear function of the soil–water deficit. This deficit also
controls soil–water repellency. Therefore, we advocate that
more in situ measurements of $\theta_i$ be made in burned areas,
and specifically that new instrumentation be developed to
measure $\theta_i$ at shallow depths (i.e. ~1–2 cm) because post-
wildfire infiltration and runoff are controlled within these
depths (Moody and Ebel, 2013). Soil–water deficit is a critical
variable and initial values before a storm cell are needed if hy-
drological models are to be run in forecast mode. It is now cer-
tainly possible to telemeter these data as quasi real-time inputs
to hydrologic models.

These OFD measurements coupled with $\theta_i$ measurements at
shallow depths can also help to understand the redistribution of
soil–water content between storm cells. Models often make
crude approximation or ignore it (KINEROS2, 2014), yet one
storm cell may ‘prime the pump’ for latter cells that may pro-
duce substantial floods even though the storm may be smaller in
magnitude.

Finally, the time-to-start of runoff can be output from a hydro-
logic model of a burned area and compared to measurements
made using OFDs. It was found in this study that time for de-
pression storage was small relative to the time for ponding,
but this might not be the case for different burned sites with dif-
f erent surface roughness and connectivity characteristics.
These comparisons would serve as a way to verify post-wildfire
hydrologic models.
Summary and Conclusions

Hydrographs of unsteady overland flow in micro-drainages on hillslopes during actual rainstorms were measured in situ at a sampling frequency of one second using inexpensive OFDs. These remote measurements of the time-to-start of runoff, $T_0$, can be telemetered and can be used to determine important quasi, real-time parameters for hydrologic models. Measurements made at multiple sites can be used to compute unsteady overland flow velocities for model verification. If $T_0$ measurements are combined with a hyetograph, actual values of the critical runoff thresholds (initial abstraction) and critical infiltrability threshold can be determined. With additional measurements of soil–water content and saturated hydraulic conductivity, the wetting-front suction parameter can be estimated and used in the infiltration component of hydrologic models.

For the particular field site in this study, overland flow did not start simultaneously on hillslope facets and in micro-drainages as is often assumed in the development of most runoff models. Flow from upstream accounted for 27% of the initiation of runoff at OFDs. Runoff was initiated first at upstream detectors 56%, at mid-stream detectors 6%, and at downstream detectors 38% of the time. Thus, initiation of runoff exhibited a complex spatial and temporal response. It started at different times at different locations during different storm cells.

Temporally, $T_0$ was shown to depend on the time-to-ponding, time to fill surface depression storage, and travel time between points in the micro-drainages. At this particular field site, these times were 0.5–5 s, 0.4–1.1 s, and 0–2–14 minutes, respectively and indicate the importance of the ponding process in controlling $T_0$. To understand the temporal initiation of runoff from burned areas, two relations were used to model the dependence of $T_0$ on travel time, cumulative rainfall, rainfall acceleration, and the soil–water deficit. The cumulative-rainfall relation predicted time-to-ponding, and it, combined with the travel time explained 96% of the variance in the observed values of $T_0$. The soil–water deficit relation explained 80% of the variance, which was probably lower than the cumulative-rainfall relation because soil–water content was only measured daily at the surface (0–1.5 cm) at two locations separated from the six OFD locations. This highlights the need to develop remote sensing instruments that are capable of measuring soil–water content at shallow depths with a time resolution compatible with that of the OFDs to improve predictions of flow initiation from burned areas.

These results describing the initiation of post-wildfire runoff are an initial attempt to understand the process during unsteady flow produced by actual rainstorms rather than simulations. They highlight the importance of the rainfall acceleration characteristic in determining the time-to-start of runoff, and the value of OFDs in determining critical threshold parameters such as initial abstraction and maximum infiltrability for hydrologic modeling. However, additional work is needed to develop instrumentation and similar relations for other sites before a more general relation can be used as a forecast tool or adopted into existing hydrologic rainfall–runoff models.

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