Modeling the effects of surface storage, macropore flow and water repellency on infiltration after wildfire

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\textbf{S U M M A R Y}

Wildfires can reduce infiltration capacity of hillslopes by causing (i) extreme soil drying, (ii) increased water repellency and (iii) reduced soil structure. High severity wildfire often results in a non-repellent layer of loose ash and burned soil overlaying a water repellent soil matrix. In these conditions the hydraulic parameters vary across discrete layers in the soil profile, making the infiltration process difficult to measure and model. The difficulty is often exacerbated by the discrepancy between actual infiltration processes and the assumptions that underlie commonly used infiltration models, most of which stem from controlled laboratory experiments or agricultural environments, where soils are homogeneous and less variable in space and time than forest soils. This study uses a simple two-layered infiltration model consisting of surface storage ($H$), macropore flow ($K_{mac}$) and matrix flow ($K_{mat}$) in order to identify and analyze spatial–temporal infiltration patterns in forest soils recovering from the 2009 Black Saturday wildfires in Victoria, southeast Australia. Infiltration experiments on intact soil cores showed that the soil profile contained a region of strong water repellency that was slow to take on water and inactive in the infiltration process, thus restricting flow through the matrix. The flow resistance due to water repellent soil was represented by the minimum critical surface tension ($CST_{min}$) within the top 10 cm of the soil profile. Under field conditions in small headwaters, the $CST_{min}$ remained in a water repellent domain throughout a 3-year recovery period, but the strength of water repellency diminished exponentially during wet conditions, resulting in some weather induced temporal variation in steady-state infiltration capacity ($K_p$). An increasing trend in macropore availability during recovery was the main source of temporal variability in $K_p$ during the study period, indicating (in accordance with previous studies) that macropore flow dominates infiltration processes in these forest soils. Storage in ash and burned surface soil after wildfire was initially high ($\sim 4$ mm), then declined exponentially with time since fire. Overall the study showed that the two layered soil can be represented and parameterized by partitioning the infiltration process into surface storage and flow through a partially saturated and restrictive soil layer. Ash, water repellency and macropore flow are key characteristics of burned forest soils in general, and the proposed model may therefore be a useful tool for characterizing fire impact and recovery in other systems.

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\textbf{1. Introduction}

Fire can increase overland flow by reducing interception, infiltration and surface roughness (Martin and Moody, 2001; Robichaud, 2000; Shakesby and Doerr, 2006; Sheridan et al., 2007). Increased production of overland flow can in turn lead to increased erosion rates (Cerda and Lasanta, 2005; Lane et al., 2006a; Moody and Martin, 2001; Robichaud et al., 2008a; Shin et al., 2013; Smith et al., 2011; Wagenbrenner and Robichaud, 2013) and increased frequency of threshold driven responses such as flash floods and debris flows (Cannon, 2001; Cawson et al., 2012; Kean et al., 2013; Nyman et al., 2011). High severity wildfire (i.e. crown fire) removes vegetation, burns the topsoil and deposits ash on hillslopes. Under these conditions the surface roughness...
and the rates of interception are low and relatively homogenous across hillslopes, irrespective of the catchment conditions prior to burning (Johansen et al., 2001). Infiltration however can be highly variable due to a strong dependency on pre-fire soil properties. Soil properties such as porosity, pore-size distribution, macroporosity and water repellency are therefore important controls on variation in hydrological responses across burned landscapes (Larsen et al., 2009; Nyman et al., 2011; Robichaud et al., 2007; Shakesby and Doerr, 2006).

Infiltration models use theory of flow in porous media to estimate the rate at which water enters the soil. Basically the infiltration rate is modeled as a function of (i) the pore-size distribution of the soil matrix, (ii) the initial soil moisture and (iii) the rate at which water is supplied at the surface (Green and Ampt, 1911; Philip, 1957). Hydraulic conductivity (mm h\(^{-1}\)), sorptivity (mm h\(^{-0.5}\)), or the suction at the wetting front (mm) are infiltration parameters that reflect the combined effects of these properties on flow and retention of water within the soil (Smith et al., 2002). These infiltration parameters can be obtained from laboratory studies, field experiments or pedotransfer functions (Cook, 2007; Moody et al., 2009; Rawls et al., 1983; Risse et al., 1994; Robichaud, 2000). Models of infiltration are process-based and represent the physical processes contributing to flow and storage of water in the soil. However, the models are highly idealized and there remain large gaps in their capacity to represent the actual infiltration processes in forest soils, where infiltration is characterized by preferential flow and non-uniform wetting fronts (Beven and Germann, 2013).

The effect of burning on infiltration rates is well documented in the literature. Fire impacts on infiltration parameters by (i) adding surface storage capacity as deposits of fine ash and burned soil (Bodi et al., 2012; Cerdà and Doerr, 2008; Woods and Balfour, 2008, 2010), (ii) reducing soil structure and macropore flow (Nyman et al., 2010; Onda et al., 2008) and (iii) reducing pore-space availability and wettability due to water repellent soils (Cerdà and Doerr, 2007; Doer and Moody, 2004; Moody and Ebel, 2012; Nyman et al., 2010). Soil profiles on fire affected hillslopes typically consist of heated and burned soil that is sandwiched between ash at the surface and an underlying soil matrix that is unaffected by the fire. The layered soil profile means that there is often strong variability with depth for soil properties such as porosity, particle size distribution and water repellency (Bodi et al., 2012; Ebel, 2012; Ebel et al., 2012; MacDonald and Huffman, 2004; Moody and Ebel, 2012; Moody et al., 2009; Stoof et al., 2010; Woods et al., 2007). Characterizing soil hydrological properties and their effects on the infiltration process in these systems is challenging because it requires simultaneous examination of flow processes in soil layers with different media properties (Ebel and Moody, 2013; Moody et al., 2013).

The properties that dominate infiltration change depending on the spatial and temporal scales at which processes are measured. Water repellency for instance can be quantified as a spatial distribution of a point-based measurement of water drop penetration times (Doerr et al., 1998; Robichaud et al., 2008b; Woods et al., 2007). The water repellency has strong effect on the behavior of water drops, reducing the ability of the soil to absorb water (Doerr et al., 2000). However, the strength or persistence of water repellency at points may not translate to large impacts on infiltration if water bypasses the matrix as preferential flow through wettable patches, cracks, and macropores and along roots and rocks (Burch et al., 1989; Doerr and Moody, 2004; Granged et al., 2011; Imeson et al., 1992; Nyman et al., 2010; Shakesby and Doerr, 2006; Urbanek and Shakesby, 2009). Similarly, the temporal scale of measurement is important. Moisture-induced changes to water repellency for instance is important when infiltration is modeled across different seasons, but it might be negligible within rain storms since the time scale of imbibition in water repellent soil may be in the order of several hours to days (Crockford et al., 1991; Ebel et al., 2012; Moody and Ebel, 2012).

Representing the interactions between macropore flow, matrix flow and imbibition is important for understanding and predicting fire-impacts on infiltration processes. Most infiltration models, however, are based on theory and data from systems where the dominant processes and key properties are different from what is typically observed in fire-affected soils, particularly with regards to wetting behavior (imbibition) and macropore flow (Ebel and Moody, 2013; Nyman et al., 2010). In this paper we therefore aim to develop a model for hydraulic conductivity which incorporates moisture dependent water repellency dynamics and which accounts for changes in macropore flow during recovery from wildfire. The study combines field campaigns and laboratory measurements to:

1. Model the interactions between imbibition and hydraulic conductivity of intact soil cores that were water repellent.
2. Quantify the effects of seasonal weather, water repellency, surface storage and macropore flow on storage and steady-state infiltration in catchments recovering from wildfire.

The study was conducted at sites in Victoria, southeast Australia, burned by the 2009 Black Saturday wildfires and the 2006 Great Divide Wildfires. Previous work from the region shows that macropore flow and water repellency are important controls on infiltration (Burch et al., 1989; Crockford et al., 1991; Lane et al., 2006b; Nyman et al., 2010; Prosser and Williams, 1998; Shakesby et al., 2003; Sheridan et al., 2007; Smith et al., 2011).

2. Methods

2.1. Infiltration model

A large proportion of post-fire erosion tends to occur in response to high intensity rainfall events (Ebel et al., 2012; Kean et al., 2011; Nyman et al., 2011; Smith et al., 2011). This study assumes that infiltration during these types of events is determined by storage of water in surface material and the flow of water out of this storage and into the soil matrix. Once storage is depleted, the maximum infiltration capacity (\(K_p\)) occurs under ponded conditions and is either controlled by (i) supply rate of water (\(R\)), (ii) the hydraulic conductivity of ash/burned soil mixture (\(K_{ash}\)), or (iii) the sum of steady-state infiltration through macropores (\(K_{mac}\)) and the matrix (\(K_{mat}\)) (Fig. 1).

Field- and laboratory-based infiltration measurements were used to parameterize and analyze a storage based infiltration model which represents infiltration, \(I(t)\), as a two stage process:

\[
\begin{align*}
  \{ f(t) = Ht_0/t + K_{mat} \text{ for } (t < t_p) \} & \text{ and } (K_{ash} > K_{mat}) \\
  \{ f(t) = K_{mat} + K_{mac} \text{ for } (t > t_p) \} & \text{ and } (K_{ash} > K_{mat}) \\
\end{align*}
\]

where \(t\) is time, \(t_p\) is time to ponding, \(H\) is the surface storage (mm), \(t_0\) is the effective saturation, \(K_{mat}\) is the effective hydraulic conductivity of the soil matrix (mm h\(^{-1}\)) and \(K_{mac}\) is the macropore flow (mm h\(^{-1}\)) given unlimited supply (Fig. 1a). Effective saturation \(\theta_0\) is the difference between soil moisture at saturation (or porosity, \(\theta_i\)) and the initial soil moisture (\(\theta_0\)) relative to \(\theta_i\), \(\theta_0 = (\theta_i - \theta_0)/\theta_i\). Macropore flow, \(K_{mac}\), is the difference in steady-state infiltration between \(h = -15\) mm and \(h = 5\) mm.

The surface storage volume, \(V\) (mm\(^3\)), above and within the water repellent layer was estimated from:

\[
V = \frac{RH\theta_0}{R - K_{mat}} + A
\]
where $R$ is the supply rate of water (mm h$^{-1}$) to the soil surface and $A$ is the area under infiltration. The potential supply from infiltrometers is high relative to $K_{mat}$ ($R \gg K_{mat}$), hence $V \rightarrow AH/t$ (Kirkby, 1975; Scoging, 1979).

Water repellency can vary with changes in soil moisture. A separate model was used to represent the infiltration into this layer since wetting processes are slow compared to non-repellent soil (Moody and Ebel, 2012). Water repellency was represented as a distribution of critical surface tension, CST (dyn cm$^{-1}$) at soil depths, $d_s$, between 0 and 10 cm using the function:

$$\text{CST}(d_s) = 72.7 + \nu d_s + u^{d_s}$$

(3)

The fitted parameters $\nu$ and $u$ determine the maximum strength and the distribution of water repellency (CST) in the soil profile. Eq. (3) represents the soil surface as wettable (i.e. CST = 72.7 dyn cm$^{-1}$ when $d_s = 0$), an assumption that was supported by data from field measurements on burned soil. Using Eq. (3) to minimize the error in the spatially distributed CST measurements means that both vertical and planar variability were captured in a single function. The maximum strength of water repellency within the soil profile, CST$_{min}$, (the ‘bottleneck’ if ash is not limiting flow) was obtained by setting the first derivative of CST($d_s$) in Eq. (3) to 0, that way calculating the depth of maximum repellency ($d_{max}$):

$$d_{max} = -1/\log(\nu)$$

(4)

The slow wetting process in water repellent soils means that the interaction between soil moisture and CST$_{max}$ are mostly dependent on weather conditions spanning over days to weeks. For field conditions, the empirical Keetch–Byram Drought Index (KBDI; Keetch and Byram, 1968) was used as a predictor of moisture status in the catchments and hence temporal dynamics in CST$_{max}$:

$$\text{KBDI}_t = \text{KBDI}_{t-1} - 10P$$

$$- \frac{(2000 - \text{KBDI}_{t-1}) (0.967e^{0.008T_{max}+1.556}) - 0.83}{(1 + 1088e^{-0.007P_{max}})1000}$$

(5)

$P$ is the daily precipitation (mm), $T_{max}$ is the daily maximum temperature (°C) and $P_{max}$ is the annual precipitation at the site (mm). The temporal variability in $K_{mat}$ was measured and represented as an empirical function of CST$_{min}$ and KBDI.

In the following section we describe a set of laboratory and field experiments that were designed to:

1. Evaluate the assumption that hydraulic conductivity of the soil matrix is restricted by water repellent layer (Fig. 1) and,
2. Quantify the effects of (i) regional weather conditions, (ii) soil moisture status, (iii) water repellency, and (iv) macropore availability on infiltration processes in soils recovering from wildfire.

2.2. Laboratory and field measurements: Overview of methods

Laboratory- and field-based experiments were used to measure infiltration at three fire affected hillslopes in Victoria, southeast Australia (Fig. 2 and Table 1). The primary study sites were Sunday Creek and Stony Creek, which were burned at high severity during the Black Saturday wildfires in February 2009 (Cruz et al., 2012). A third site, Ella Creek, was burned by the Great Divide wildfires December 2006, and was included to represent similar forest environments in a later stage of recovery. All sites are located in the eastern uplands of Victoria and were burned under wildfire conditions where the forest canopy was completely burned (burn severity class 1: 75–100% Crown consumption) (see definitions in Cruz et al., 2012). The southeast Australian region experiences a Mediterranean climate with hot and dry summer and cool and wet winters. There is large variability in ecosystem properties across the study area, but the three sites used in this study were all characterized by dry eucalyptus forest and were similar in terms of rainfall, aspect and solar exposure. See Nyman et al. (2011) for a general description of the geomorphology, vegetation and fire regimes in the region.

Intact soil cores were sampled from each site in April 2010 and used for laboratory-based infiltration measurements that quantify the relation between imbibition rate and hydraulic conductivity. Laboratory experiments were best suited for this type of study because soil moisture conditions could be controlled. Infiltration rates were also sampled in the field during campaigns that were aimed at identifying key controls on infiltration in soils that were recovering from wildfire. Initial soil moisture and water repellency were measured alongside measurements of surface storage, $H$ (mm), matrix flow, $K_{mat}$ (mm h$^{-1}$) and macropore flow $K_{mac}$ (mm h$^{-1}$). The field measurements were made in headwater catchments (~2 ha) during a 3-year recovery period from wildfire. Daily temperature and rainfall were measured at each site and used to calculate KBDI using Eq. (5).

2.3. Laboratory study: Flow and imbibition in water repellent soil

Intact cores were collected from Ella Creek, Sunday Creek and Stony Creek and used to measure hydraulic conductivity and
Table 1
Attributes of the three study sites.

<table>
<thead>
<tr>
<th>Site</th>
<th>Aspect &amp; elevation</th>
<th>Burn impacta</th>
<th>Annual Rainfall</th>
<th>Forest type and dominant vegetation</th>
<th>Geology</th>
<th>Soil texture</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ella Creek</td>
<td>North</td>
<td>Moderate to high severity Dec-2006</td>
<td>1200–1400</td>
<td>Dry eucalyptus, Broad-leaved peppermint (E. Radiata) Narrow-leaved peppermint (E. Dives)</td>
<td>Shale, Marine Sedimentary</td>
<td>Stony and gravelly clay loam</td>
</tr>
<tr>
<td></td>
<td>720 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sunday Creek</td>
<td>North</td>
<td>High to very high severity dNBR = 739 Feb-2009</td>
<td>1000–1200</td>
<td>Dry eucalyptus, Broad-leaved peppermint (E. Radiata)</td>
<td>Siltstone, Marine Sedimentary</td>
<td>Stony and gravelly clay loam</td>
</tr>
<tr>
<td></td>
<td>480 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stony Creek</td>
<td>North-West</td>
<td>High to very high severity dNBR = 582 Feb-2009</td>
<td>1000–1200</td>
<td>Dry eucalyptus, Broad-leaved peppermint (E. Radiata)</td>
<td>Phyllite &amp; Gneiss Metamorphic</td>
<td>Gravelly clay loam</td>
</tr>
<tr>
<td></td>
<td>470 masl</td>
<td></td>
<td></td>
<td></td>
<td></td>
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</tr>
</tbody>
</table>

a dNBR is the change in normalized burn ratio as a result of burning (Key and Benson, 2004).

absorption rate at different stages of imbibition on a tension table. Nine cores (−8 cm deep and 5.3 cm in diameter) were sampled from each site at upper, mid and lower hillslope positions in April 2010. The soil cores were left to air dry in the laboratory (4 weeks in the lab at 20°C), then weighed before the first set of infiltration measurements. The infiltration rate was measured at 15 mm tension (h = −15 mm) using Mini-disc (MD) infiltrometer (Decagon) (Moody et al., 2009; Robichaud et al., 2008b) and a lab procedure for measuring unsaturated hydraulic conductivity (Cook, 2007). Measuring infiltration under slight tension ensures that values were representative of flow processes in the soil matrix, excluding gravity driven flow in macropores, which can obscure the effect of water repellency on flow within the matrix.

The cores and the infiltrometer were held in place using clamp and stands during infiltrations. Cheesecloth and a shallow layer of pebbles (<600 μm) were placed on the soil surface to ensure good contact between the disc and the soil. In early stages of infiltration, the rate was usually high, and then declined towards some low steady-state, usually within 5 min. The water level was read from the reservoir every 30 s for 35 min (min) and only recorded if there was a change from the previous reading. The core was reweighed at 35 min and replaced in the clamp for another 15 min of infiltration measurements if the wetting front had not appeared at the bottom of the core. If the wetting front had reached the bottom of the core at 35 min, then the core was placed on a buchner funnel set to a tension of −15 mm (same as the mini-disc tension chamber). Having the same pressure at the bottom and top of the core ensures a uniform hydraulic gradient throughout the core and allowed for direct estimation of hydraulic conductivity (Cook, 2007). The last 15 min of infiltration was used as a measure of the effective hydraulic conductivity of the matrix (K_eff) for given initial moisture conditions (h).

After the first set of infiltration experiments, the cores were immediately placed onto a tension table with the tension set so that the average water pressure (h) for the cores was equal to −15 mm. The tension table was set up with large reservoir of water and constant head burette to maintain constant tension and accommodate any water that was lost due to the water uptake by the cores. Each core was reweighed after 36 h on the tension table and a new set of infiltrations were carried out. The cores were replaced onto the tension table and reweighed at 60, 100, 148, 268 and 316 h. The multiple time steps at which the mass was obtained indicated changes in moisture content with time, and helped determine a suitable point at which to run a new set of infiltrations.

A new set of infiltrations required a new initial soil moisture status of the cores. However, after 268 h of wetting at h = −15 mm, it was evident that some cores were still dry on the surface. After this time step the water level in the tension table was therefore raised to the midpoint of the core (mean h = 0 mm) in attempts to reach a higher moisture levels and ultimately soil saturation. A third and fourth set of infiltration measurements were carried out after 374 h and 681 h of wetting. After the final infiltration, all cores appeared to be saturated and there was no further increase in mass of cores from this point onwards. The final set of infiltrations was taken to represent the saturated hydraulic conductivity (K_s) of the soil matrix when h = −15 mm. The small tension means that measurements exclude the effect of macropores that contribute to flow when h > −15 mm.

The cores were finally replaced onto the tension table and drained at 5 tension intervals (−2.5 < h < −50 cm) in order to characterize the size distribution of pores in the meso-macropore domain. The porosity of the soil at each site was estimated from the saturated volumetric water content \( \theta_s \ (\text{cm}^3 \text{ cm}^{-3}) \) of the 9 cores that were used in the laboratory infiltration experiment. The pore-size distribution was estimated from the change in water content at 5 increments of h between 0 and −50 cm, and using the capillary equation to relate h to pore radius (r, mm):

\[
r = \frac{2 \rho gh}{\cos(\alpha)}
\]

where \( \rho \) is the density of water, \( g \) is gravity and \( \alpha \) is the contact angle. A cumulative distribution function was then used to quantify the proportion of pore-space occupied by macropores (r > 0.5 mm) (Nyman et al., 2010).

Fig. 2. Infiltration measurements were made on soils from three sites in the eastern uplands of Victoria, southeast Australia.
2.4. Infiltration properties of headwater catchments during recovery from wildfire

Infiltration was measured in the field under both positive (h = 5 mm) and negative (h = −15 mm) pressure, using the min-disc tension infiltrometer (Decagon) and a custom-designed ponded infiltrometer (Perroux and White, 1988) with same disc diameter as the tension infiltrometer. Measurements were made in three headwater catchments during 4 sampling campaigns, each taken to represent burned systems in different stages of recovery and in different seasons (Fig. 3). Infiltration in each headwater catchment was measured at 4 points in 3 quadrats (1 m × 1 m) along 3 transects (80 m long) running perpendicular to the contour (n = 36). Quadrats were positioned at upper- (0 m), mid- (40 m) and lower (80 m) sections of the transects. Each campaign was conducted on allocated strips, 2 m to the side of the previous campaign to avoid disturbance from previous campaigns.

A retention ring (5.3 cm in diameter) was inserted 2–3 cm into the soil in order to prevent lateral flow through non-repellent surface material which often covered the hillside. The soil surface was relatively smooth and the infiltrometer disc was small and no contact material was required to achieve contact between the soil and the disc of the tension infiltrometer. The depth of infiltrated water was measured with the tension infiltrometer at 15 s intervals in the first minute, then at 30 s intervals for the remaining 10–15 min of infiltration, after which the infiltrometer was removed and replaced with the ponded infiltrometer. The soil within the retaining cylinder was then flooded to achieve 5 mm of ponding before measuring the ponded infiltration rate at 15 s intervals for 5–10 min.

Soil was collected at 5 depth intervals (0–1, 1–3, 3–5, 5–7.5, 7.5–10 cm) at 10 m intervals along each of the three transects where infiltration was measured. The samples were a composite of 2 cores (10 cm × 5.5 cm) collected in the 1 m quadrat where infiltration was measured. The samples were placed in sealed bags and transported back to the laboratory and weighed. Water repellency was measured on samples in the laboratory using ethanol solutions of different concentrations (0–6 M; 0.4 M intervals), and calculating the critical surface tension (CST) once three drops of solution penetrated the soil in 3 s (CST test) (King, 1981; MacDonald and Huffman, 2004). The soil was mixed inside the sealed sampling bag before a subsample was placed on a petri dish. Gravel and large organics were removed manually from the petri dish before placing the drops of ethanol solution on the soil surface.

The sub-sampled soil was returned to the original sample, which was then oven dried at 105° for 48 h in order to calculate the volumetric water content. For samples collected in March 2010, the soil was left to air-dry first so that the air-dry CST of each site could be obtained, using the same methods described above. The water repellency was often high in the 0–1 cm depth interval. However, the undisturbed soil surface always seemed to be wettable and absorbed some water during infiltration measurements in the field. Measurements of CST were therefore carried out at the soil surface during each field campaign in order to determine CST at dₙ = 0. Litter was removed while ensuring that the soil was not disturbed prior to the test. The surface material was usually a mix of mineral soil, ash and organic material and was non-repellent in 95–100% of sampled locations at each site.

2.5. Linking point and plot scale infiltration processes

Infiltration was measured in March 2010 at plots (2 m × 1.5 m) during 3 replicate rainfall simulations at Sunday Creek (Fig. 2). The plots were located on steep hillslopes (28°) in dry eucalyptus forest with clay loam soil that was burned by wildfire in February 2009. The rainfall simulation procedure has been described in Sheridan et al. (2007). The rainfall rate (target 100 mm h⁻¹) was first calibrated with steady-state runoff from a plastic sheet covering the plot. The sheet was removed and discharge was measured in 0.5 l containers at regular intervals during a 30-min runoff period. An adjacent 1 m × 1 m plot was used as a test area for measuring infiltration at 3 points per rainfall simulation plot.

3. Results

3.1. Laboratory study: Flow and imbibition in water repellent soil

The soils were water repellent at all sites under air-dry conditions (Fig. 4). The critical surface tension (CST) was highly variable at each sampling depth apart from the lowest depth interval (7.5–10) were the CST was usually close to 72 (i.e. non-repellent). The spatial variability in CST within the most hydrophobic region of the soil was more strongly skewed at Ella Creek (Fig. 4c) than at the other two sites (Fig. 4a and b). The different patterns of variability essentially show that water repellency was stronger and more homogenous in the recently burned sites. The CST(dₙ) function, Eq. (3), explained 31%, 17% and 24% of variability in water repellency of air-dry soils at Sunday Creek, Stony Creek and Ella.
Creek, respectively, when fitted across all sampling points \( (n = 36) \). The parameter optimization was highly significant \( (p < 0.01) \) for all sites.

When fitted to the mean CST, the function, Eq. (3), consistently explained more than 80% of the variability in CST with depth, indicating that spatial variability across the hillslope, and not variation with depth, was the main source of residual error when the function was fitted across all sampling points. The depth of minimum CST (i.e. maximum water repellency) in air dry soils was estimated from Eq. (4) to be 1.2, 1.8 and 1.3 cm for Sunday Creek, Stony Creek and Ella Creek respectively. The corresponding minimum value for CST, the \( \text{CST}_{\text{min}} \), was 40.1, 46.5 and 52.4 dyn cm\(^{-1}\).

The porosity and pore-size distribution values were obtained from the intact cores once the infiltration experiments were completed and the cores were completely saturated \( (h = 0 \text{ mm}) \). The porosity at Sunday Creek and Stony Creek was 0.34 and 0.41, respectively \( (\text{Table } 2) \). A higher proportion of macropores \( (r > 0.5 \text{ mm} = 4.7\% \text{ of } h_s) \) at Sunday Creek resulted in higher matrix flow \( (K_{\text{mat}} \text{ at saturation}) \) at Sunday Creek \( (66 \text{ mm h}^{-1}) \) than at Stony Creek \( (52 \text{ mm h}^{-1}) \) despite Stony Creek having higher overall porosity \( (\text{Table } 2) \). The porosity at Ella was higher \( (0.55) \) and the matrix flow at saturation was \( -4 \) times higher than the two other sites \( (\text{Table } 2) \). Porosity was inversely related to gravel content. The hydraulic conductivity of repacked cores with wettable surface material \( (\text{ash, gravel and soil}) \) was 41 mm h\(^{-1}\) and 21 mm h\(^{-1}\) for Sunday Creek and Stony Creek respectively \( (\text{Table } 2) \). At saturation the mixture of ash, gravel and burned soil had a water holding capacity of 0.40 cm\(^3\) cm\(^{-3}\).

**Table 2**

<table>
<thead>
<tr>
<th>Site</th>
<th>Bulk Density ( (\text{g cm}^{-3}) )</th>
<th>Gravel D &gt; 2 mm ( (% \text{ mass}) )</th>
<th>Porosity ( \theta_i ) ( (% \text{ of } h_s) )</th>
<th>Macroporosity ( (r &gt; 0.5 \text{ mm}) ) ( (% \text{ of } h_s) )</th>
<th>Air-dry CST(_{\text{min}}) ( (\text{dyn cm}^{-1}) )</th>
<th>( K_{\text{mat}} ) ( (h = -15 \text{ mm}) ) ( (\text{mm h}^{-1}) )</th>
<th>( K_{\text{sat}} ) ( (h = -15 \text{ mm}) ) ( (\text{mm h}^{-1}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sunday Creek</td>
<td>1.29</td>
<td>62</td>
<td>0.34</td>
<td>4.7</td>
<td>40.1</td>
<td>66 (18)</td>
<td>41 (4.0)</td>
</tr>
<tr>
<td>Stony Creek</td>
<td>1.31</td>
<td>36</td>
<td>0.41</td>
<td>1.7</td>
<td>46.5</td>
<td>52 (41)</td>
<td>21 (4.9)</td>
</tr>
<tr>
<td>Ella Creek</td>
<td>0.97</td>
<td>27</td>
<td>0.55</td>
<td>2.6</td>
<td>52.4</td>
<td>240 (81)</td>
<td>n/a</td>
</tr>
</tbody>
</table>

\( a \) This is the hydraulic conductivity of the cores when completely wetted at \( h = -15 \text{ mm} \).

\( b \) Repacked cores contained a mixture of gravel, burned soil and ash.

---

**Fig. 4.** Strength of water repellency \( (\text{critical surface tension}) \) at different depths at (a) Sunday Creek, (b) Stony Creek and (c) Ella Creek in Victoria southeast Australia. Ella Creek was burned by wildfire in December 2006 and Stony Creek and Sunday Creek were burned in February 2009. Measurements were obtained on air-dry soil samples \( (n = 36) \) at multiple depths \( (0–1, 1–3, 3–5, 5–7.5 \text{ and } 7.5–10 \text{ cm}) \) which were collected in March 2010 \( (\text{Fig. } 3) \).

---

**Fig. 5.** (a) Initial soil moisture \( (\theta_i) \), (b) matrix flow \( (K_{\text{mat}}) \) and (c) change in soil moisture \( (\Delta \theta) \) during infiltration into intact soil cores from Sunday Creek, Stony Creek and Ella Creek \( (n = 9) \) at different stages of wetting on a tension table. All values are for \( h = -15 \text{ mm} \). The water uptake in (c) was obtained from the mass of soil cores measured before and after each 35 min infiltration.
The volumetric soil moisture content, \( \theta_i \) (cm\(^3\) cm\(^{-3}\)), of air dry cores from Stony Creek, Sunday Creek and Ella Creek were 0.014, 0.022 and 0.029, respectively. In subsequent measurements the soil moisture was consistently higher at Stony Creek than Sunday Creek but highest at Ella Creek (Fig. 5a), reflecting an increasing trend in porosity. The effective hydraulic conductivity of the matrix \( K_{\text{mat}} \) was low relative to the saturated conductivity of the matrix \( (K_s) \) for all sites in the three first three sets of infiltration measurements (Fig. 5b). The largest change in \( K_{\text{mat}} \) occurred between the third and the fourth set of infiltration experiments with cores from Ella Creek displaying a larger increase than the two other sites (Fig. 5b).

The volumetric water uptake \( (\Delta \theta) \) is the change between the initial and final water content \( (\theta_i \) and \( \theta_f \) respectively) during the first 35 min of infiltration, excluding the storage component, \( V \), on the surface, \( \Delta \theta = \theta_i - \theta_f - V \). Surface storage volume during early (non-steady) stage of infiltration was obtained from \( H \), a fitted infiltration parameter in Eq. (1). The change in water content of cores before and after 35 min infiltrations was relatively large \((0.08 < \Delta \theta < 0.15)\) for first set of infiltration experiments (Fig. 5c). The water uptake \( \Delta \theta \) during infiltrations was similar for the second and third set of measurements \( (0.04 < \Delta \theta < 0.07)\) (Fig. 5c). In the fourth and last set of infiltrations the matrix was completely saturated, and \( \Delta \theta \) approached zero (Fig. 5c). At the last stage of wetting the \( K_{\text{mat}} \) was considered to represent \( K_s \) of the soil matrix when \( h = -15 \text{ mm} \) (Table 2).

The relations between initial soil moisture \( (\theta_i) \), effective hydraulic conductivity \( (K_{\text{mat}}) \) and rate of water uptake \( (\Delta \theta) \) during different stages of infiltration were combined across the three study sites by normalizing \( K_{\text{mat}} \) by \( K_s \), and \( \Delta \theta \) and \( \theta_i \) by total porosity, \( \theta_0 \) (Fig. 6a). The rate of water uptake, \( (\Delta \theta/\theta_i) \) approached zero (Fig. 6b) This decline represents the effects of (i) decreasing pore-space availability due to water repellent soil and (ii) an overall decline in available pore space. The linear relation means that the water uptake is proportional to initial soil moisture \( (\Delta \theta/\theta_i) \), thus giving rise to an exponential relation between soil moisture and the duration of infiltration. However, the hydraulic conductivity remained relatively low \((K_{\text{mat}}/K_s < 0.4)\) and constant while \( \theta_i/\theta_0 < 0.6 \), indicating that some sections of the soil remained dry and hence inactive in the overall flow process. The hydraulic conductivity \( K_{\text{mat}} \) increased exponentially with increasing soil moisture when \( \theta_i/\theta_0 > 0.6 \). The change in hydraulic conductivity with \( \theta_i/\theta_0 \) could be represented by an exponential function which asymptoted at some minimum flow in dry and repellent soils.

In summary the patterns of flow and absorption shows that the hydraulic conductivity was dependent on the rate at which pore space was activated and introduced to the infiltration process. The rate of at which the soil matrix was activated was strongly dependent on initial soil moisture conditions. The data support the assumptions underlying the conceptual model of water repellency acting as a restriction on flow through the soil matrix (Fig. 1).

### 3.2. Infiltration in burned headwater catchments

Storage, \( H \), was obtained for headwaters at each measurement campaign by fitting Eq. (1) to the daily infiltration measurements \((\text{mm h}^{-1}) \) (Table 3). The matrix flow, \( K_{\text{mat}} \) in Eq. (1) was obtained from steady-state infiltration \((t > 6 \text{ min}) \) (Table 3) and treated as a fixed parameter in the fitting procedure. Soil moisture was highest for all sites in September 2010 after a relatively wet spring period (Table 3). The normalized soil moisture decreased linearly \((R^2 = 0.61)\) with increasing \( K_{\text{BDI}} \) (Fig. 7a). Overall, the moisture content remained low relative to \( \theta_i \) despite \( K_{\text{BDI}} \) approaching 0, and the range of soil moisture contents \((0.10 < \theta_i/\theta_0 < 0.43)\) (Fig. 7a) was low compared to the range achieved under laboratory conditions (Fig. 6a). There was no relation between initial soil moisture and the matrix flow \( K_{\text{mat}} \) (Fig. 7b), a result which may have been expected given that laboratory measurements (Fig. 6) showed that \( K_{\text{mat}} \) is invariant with \( \theta_i \) when \( \theta_i/\theta_0 < 0.6 \).

The \( \text{CST}_{\text{min}} \) (dyn cm\(^{-1}\)) ranged from 24.8 to 43.1, 42.3 to 63.8 and 29.2 to 58.2 at Sunday Creek, Stony Creek and Ella Creek respectively (Table 3). The relation between \( \text{CST}_{\text{min}} \) with \( K_{\text{BDI}} \) was scattered when sites were combined without adjusting for the different background repellency. The aim however was to generalize the relation between \( K_{\text{BDI}}, \text{CST}_{\text{min}} \), and hydraulic conductivity across dry eucalyptus forest recovery from wildfire. The \( \text{CST}_{\text{min}} \) was therefore normalized by the lowest \( \text{CST}_{\text{min}} \) value (i.e. strongest water repellency) measured under field conditions, that way producing a relative measure \( \text{CST}_{\text{min}} \) which could be represented as a function of \( K_{\text{BDI}} \) (Fig. 7c). Water repellency increased with increasing \( K_{\text{BDI}} \) although the soil remained repellent even when \( K_{\text{BDI}} \) was close to 0, indicating the water repellency is likely to be present during most weather conditions for these forest systems. In terms of infiltration, the persistence of water repellency meant that flow was always restricted by water repellent soil, and that the true saturated hydraulic conductivity \( (K_s) \) could not be measured under field conditions.
The strong relation between $K_{\text{mat}}$ and $CST_{\text{min}}$ indicate that water repellency was an important source of variability when soil moisture levels were lower than water repellency thresholds (Fig. 7d). The matrix flow $K_{\text{mat}}$ was most sensitive to changes in water repellency status for $CST_{\text{mat}} > 40$ dyn cm$^{-1}$. The flow under non-repellent conditions ($K_{\text{r}}$) was estimated to be $\approx 50$ mm h$^{-1}$ by extrapolating a value for $K_{\text{mat}}$ at $CST_{\text{min}} = 72.7$ dyn cm$^{-1}$ (Fig. 7d). The extrapolated value corresponds well with laboratory measurements of $K_{\text{r}}$ for Stony Creek and Sunday Creek (Table 2) but is much smaller than the $K_{\text{r}}$ measured on soil cores from Ella Creek.

Macropores can contribute to a high proportion of flow in forest soil when given abundant supply of water. Macropore flow was quantified by subtracting matrix flow ($K_{\text{mat}}$) from ponded flow ($K_{\text{p}}$; $h = 5$ mm) (Table 2). The difference in flow was then expressed relative to $K_{\text{mat}}$ (at $CST_{\text{mat}} = 72.7$ dyn cm$^{-1}$ (Fig. 7d)). The extrapolated flow values increased exponentially with $CST_{\text{mat}}$ for values corresponding well with laboratory measurements of $K_{\text{r}}$ for Stony Creek and Sunday Creek (Table 2) but is much smaller than the $K_{\text{r}}$ measured on soil cores from Ella Creek.

Table 3

<table>
<thead>
<tr>
<th>Site</th>
<th>Date</th>
<th>$t_{\text{f}}$</th>
<th>$K_{\text{BDI}}$</th>
<th>$\alpha$</th>
<th>$CST_{\text{min}}$</th>
<th>$H$</th>
<th>$K_{\text{mat}}$</th>
<th>$K_{p}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>mo</td>
<td>–</td>
<td>cm$^{-3}$</td>
<td>dyn cm$^{-1}$</td>
<td>mm h$^{-1}$</td>
<td>mm h$^{-1}$</td>
<td>mm h$^{-1}$</td>
</tr>
<tr>
<td>Sunday Creek</td>
<td>Jul-09</td>
<td>6</td>
<td>89</td>
<td>0.09</td>
<td>27.6</td>
<td>1.9</td>
<td>16.7</td>
<td>56 (93)</td>
</tr>
<tr>
<td></td>
<td>Dec-09</td>
<td>10</td>
<td>66</td>
<td>0.06</td>
<td>24.8</td>
<td>1.6</td>
<td>16.9</td>
<td>63 (134)</td>
</tr>
<tr>
<td></td>
<td>Mar-10</td>
<td>13</td>
<td>32</td>
<td>0.09</td>
<td>26.3</td>
<td>1.8</td>
<td>18.3</td>
<td>83 (188)</td>
</tr>
<tr>
<td></td>
<td>Sept-10</td>
<td>19</td>
<td>0</td>
<td>0.15</td>
<td>43.1</td>
<td>1.2</td>
<td>19.6</td>
<td>142 (161)</td>
</tr>
<tr>
<td>Stony Creek</td>
<td>Jul-09</td>
<td>6</td>
<td>26</td>
<td>0.14</td>
<td>57.7</td>
<td>3.6</td>
<td>21.7</td>
<td>32 (61)</td>
</tr>
<tr>
<td></td>
<td>Dec-09</td>
<td>10</td>
<td>62</td>
<td>0.12</td>
<td>42.3</td>
<td>2.9</td>
<td>16.0</td>
<td>40 (96)</td>
</tr>
<tr>
<td></td>
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<td>13</td>
<td>151</td>
<td>0.04</td>
<td>46.5</td>
<td>3.7</td>
<td>21.5</td>
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<td>16</td>
<td>0.12</td>
<td>63.8</td>
<td>1.0</td>
<td>31.9</td>
<td>70 (54)</td>
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<tr>
<td>Ella Creek</td>
<td>Jul-09</td>
<td>30</td>
<td>1</td>
<td>0.21</td>
<td>58.2</td>
<td>1.2</td>
<td>34.9</td>
<td>371 (419)</td>
</tr>
<tr>
<td></td>
<td>Dec-09</td>
<td>34</td>
<td>54</td>
<td>0.19</td>
<td>29.2</td>
<td>1.2</td>
<td>19.3</td>
<td>162 (174)</td>
</tr>
<tr>
<td></td>
<td>Mar-10</td>
<td>37</td>
<td>86</td>
<td>0.14</td>
<td>46.6</td>
<td>1.3</td>
<td>20.7</td>
<td>304 (446)</td>
</tr>
</tbody>
</table>

The summary of soil hydrological properties and initial conditions in four measurement campaigns to headwater catchments in southeast Australia. Sunday Creek and Stony Creek were burned by wildfire in February 2009 and Ella Creek was burned in December 2006. The standard deviation of the effective hydraulic conductivity of the soil matrix ($K_{\text{mat}}$) and steady-state infiltration capacity ($K_{\text{r}}$) are given in parentheses.

3.3. Point- to plot-scale infiltration

Infiltration parameters derived from the mini-disc infiltration measurements at Sunday Creek were evaluated against runoff data from rainfall simulation plots. The aim was to determine how point-scale infiltration measurements compare with infiltration rates measured in larger plots (3 m$^2$) under simulated rainfall. The average steady-state infiltration was higher and more variable for the mini-disc infiltrometer ($K_{\text{mat}} = 10.1$ mm h$^{-1}$ ± 10.1) than under the rainfall simulation plot (steady-state infiltration = 7.9 mm h$^{-1}$ ± 3.65 SD) (Fig. 9a). The time to ponding $t_{\text{p}}$ for a steady-state rainfall input $R$ was estimated from tension infiltrometer data as:

$$t_{\text{p}} = H/(R - K_{\text{mat}})$$

(7)

The time to ponding was then combined with steady-state infiltration model and kinematic wave equations in order to model the hydrograph from steady-state rainfall (Fig. 9b) (Brutsaert, 2005; p. 201). The infiltration excess was routed assuming a relation between water depth ($h_{\text{sw}}$, mm), friction slope ($S_f$) and average velocity ($u$, mm h$^{-1}$):

$$u = C_s h^{2/3}$$

(8)

where $C_s$ is a resistance factor (Brutsaert, 2005; p. 172). The time to equilibrium, the build-up, and the decay-phase were modeled through approximate solutions to the kinematic wave equation under steady lateral rainfall input ($R$). The hydrograph modeled using mini-disc parameters ($H$ and $K_{\text{mat}}$) corresponded well with the run-off measurements from plot-scale rainfall simulations (Fig. 9b). The ponded infiltration, $K_{\text{p}}$ (25 mm h$^{-1}$) reflects the maximum steady-state infiltration capacity of the soil, and was $\approx 2.5$ times higher than steady-state infiltration under simulated rainfall.

4. Discussion

4.1. Contribution of the soil matrix to infiltration in burned soils

The effective hydraulic conductivity of the soil matrix, $K_{\text{mat}}$, was strongly dependent on the availability of actively conducting pore-space. Laboratory experiments on intact cores showed that $K_{\text{mat}}$ remained at <40% of $K_{\text{r}}$ while the relative soil moisture ($\theta_i/\theta_i$) in the top 10 cm of the soil profile was <0.8. The magnitude of the water repellency effect on hydraulic conductivity ($K_{\text{mat}}/K_{\text{r}}$) is similar to the 60–70% reduction measured in wet eucalyptus forest (Nyman et al., 2010; Sheridan et al., 2007) and the range of impacts reported for forest soils in other places (Imeson et al., 1992; Leighton-Boyce et al., 2007). Hydraulic conductivity ($K_{\text{mat}}$) increased exponentially with $\theta$ when $\theta_i/\theta_i > 0.8$, which means that the last 20% of pore-space within the top 10 cm of the soil profile was contributing disproportionately to flow through the soil matrix. The extremely slow wetting rate for $\theta_i/\theta_i > 0.8$ and the exponential dependency between $K_{\text{mat}}$ and $\theta$ for $\theta_i/\theta_i > 0.8$ suggests...
that flow was restricted by a water repellent layer that acted as a ‘throttle’.

Water repellency at the study sites in southeast Australia displayed consistent trends with soil depth. Water repellency peaked at depths of 1–3 cm, and remained largely non-repellent on the surface and at depths >8 cm. Similar patterns of variability with depth have been observed in water repellent soils elsewhere (Benavides-Solorio and MacDonald, 2001; DeBano, 2000; Doerr et al., 1996, 2000; MacDonald and Huffman, 2004; Miyata et al., 2007; Nyman et al., 2011), indicating that the water repellency function, $CST_{ds}$ in Eq. (3), may be useful more generally for characterizing water repellency in burned soils. Overall the wildfire-affected sites in southeast Australia displayed moderate to extreme repellency, with air-dry soils on the most recently burned sites at Stony Creek and Sunday Creek (0.5 years after fire) being more repellent than air-dry soils from Ella Creek (2–3 years after fire). Under field conditions, the variability in water repellency ($CST_{min}$) was driven by meteorological conditions (or seasonality) which is consistent with previous studies of temporal variations in water repellency (Burch et al., 1989; Crockford et al., 1991; Leighton-Boyce et al., 2005; Sheridan et al., 2007). Field measurements showed that the maximum strength water repellency ($CST_{min}$) in soil profile was driving weather dependent changes in hydraulic conductivity ($K_{mat}$) of the soil matrix.

Fig. 7. Soil hydrological properties of headwater catchments in dry eucalyptus forests during recovery from wildfire. (a) Initial soil moisture, $\theta_i$, versus KBDI. (b) Matrix flow, $K_{mat}$, versus soil moisture. (c) Normalized water repellency, $CST_{min}$, versus KBDI. (d) The change in $K_{mat}$ with changing water repellency, $CST_{min}$. Each soil moisture and water repellency ($CST$) data point represents the average of 27 measurements made on composite samples collected from two points ($d_s = 0–10$ cm) at 9 locations along 3 transects (80 m) at Sunday Creek, Stony Creek and Ella Creek ($n = 27$). Each matrix flow $K_{mat}$ data point is the average of 4 mini-disc measurement points in quadrats ($1 \times 1$ m) at 3 locations along 3 sampling transects ($n = 36$). Error bars show the standard error (SE) of the mean.

Water repellency at the study sites in southeast Australia displayed consistent trends with soil depth. Water repellency peaked at depths of 1–3 cm, and remained largely non-repellent on the surface and at depths >8 cm. Similar patterns of variability with depth have been observed in water repellent soils elsewhere (Benavides-Solorio and MacDonald, 2001; DeBano, 2000; Doerr et al., 1996, 2000; MacDonald and Huffman, 2004; Miyata et al., 2007; Nyman et al., 2011), indicating that the water repellency function, $CST_{ds}$ in Eq. (3), may be useful more generally for characterizing water repellency in burned soils. Overall the wildfire-affected sites in southeast Australia displayed moderate to extreme repellency, with air-dry soils on the most recently burned sites at Stony Creek and Sunday Creek (0.5 years after fire) being more repellent than air-dry soils from Ella Creek (2–3 years after fire). Under field conditions, the variability in water repellency ($CST_{min}$) was driven by meteorological conditions (or seasonality) which is consistent with previous studies of temporal variations in water repellency (Burch et al., 1989; Crockford et al., 1991; Leighton-Boyce et al., 2005; Sheridan et al., 2007). Field measurements showed that the maximum strength water repellency ($CST_{min}$) in soil profile was driving weather dependent changes in hydraulic conductivity ($K_{mat}$) of the soil matrix.

Moisture induced changes to $CST_{min}$ took place over long timescales which meant that the soil required long exposure to wet conditions before approaching a true $K_s$. A dry and water repellent soil is therefore unlikely to reach saturation at the timescale of convective storm events that result in major runoff events after wildfire. A parameter such as $K_{mat}$ that represents infiltration in the active region of the soil is therefore more effective than $K_s$ at describing the steady-state infiltration in these systems (Hardie et al., 2011; Liu et al., 2005; Moody et al., 2013). Water uptake in dry and water repellent soil was driven by longer term meteorological conditions, and was therefore represented separately to the infiltration processes taking place at the time-scale of rain storms. Separating between the two processes is consistent with the concept presented in Moody and Ebel (2012), where infiltration into water repellent and dry soils is controlled by diffusion-adsorption processes (timescale: days to weeks) rather than capillary and gravity driven processes (timescale: minutes to hours). Distinguishing between these two infiltration processes based on their respective timescales is an important conceptual development, which contrasts with common infiltration models such as Green-Ampt that assume a uniform wetting front, where infiltration rate equals the saturated hydraulic conductivity once the capillary potential approaches zero.

The slow wetting rate of the water repellent layer meant that the matrix never reached a saturated hydraulic conductivity ($K_s$) under field conditions, and that water repellency was restricting flow even under relatively wet conditions. Temporal variability in $K_{mat}$ due to meteorological related water repellency dynamics was modeled by linking $CST_{min}$ with the Keech–Byram Drought Index (KBDI), which varies over time as function of daily rainfall and temperature. Both soil moisture and $CST_{min}$ declined with increasing KBDI. Water repellency, $CST_{min}$, displayed consistent trends with KBDI and was causing temporal variability in $K_{mat}$ with values ranging from 16 to 35 mm h$^{-1}$. There was considerable scatter caused by different background conditions at the three study sites so the model of KBDI-dependent changes in water repellency was...
improved by normalizing CST_{min} values by CST_{min} for the most severe water repellency conditions at the site (e.g. Karunarathna et al., 2010). Soil moisture (h_i/h_s) declined linearly with increasing KBDI, but the range of soil moisture conditions was too low to drive temporal variability in K_{mat}.

4.2. Surface storage during recovery from wildfire

Ash and burned soil on the surface of recently burned hillslopes was generally non-repellent and was capable of buffering subsurface soils from participating in the infiltration process in a similar way to what has been described for burned systems elsewhere (e.g. Ebel et al., 2012; Kinner and Moody, 2010; Woods and Balfour, 2008). The surface material contained ash, but was actually a mixture of ash, gravel, burned soil and charred organics that eroded or was incorporated back into the soil profile during recovery. Initially after fire, the storage, H, was ~4 mm which is in the lower range of values for recently burned hillslopes in Colorado (3.6–5.5 mm) (Ebel et al., 2012). The wettable surface material was on average 1 cm deep and had a water holding capacity of 0.40 cm\(^{-3}\) which was low compared to the range of values (0.58–0.95) reported for ash in the literature (Cerdà and Doerr, 2008; Ebel et al., 2012; Leighton-Boyce et al., 2007; Woods and Balfour, 2008). The lower storage capacity may be explained partly by the high gravel content of 36% and 62% (by mass) at Stony and Sunday Creek, respectively. In the field, the surface storage, H, declined exponentially with time since fire and asymptoted towards background H of ~1.2 mm. This decline during recovery was primarily caused by erosion of the wettable (and largely non cohesive) surface layer.

The erosion process occurring on surface material at Sunday Creek and Stony Creek is described in Nyman et al. (2013).

Hydraulic conductivity of surface material, K_{ash} at Stony Creek and Sunday Creek was high (21–41 mm h\(^{-1}\)) compared to 8.6 mm h\(^{-1}\) in Colorado (Ebel et al., 2012); lower than 51 mm h\(^{-1}\) in Montana (Woods and Balfour, 2008); similar to 36 mm h\(^{-1}\) in Wyoming (Balfour and Woods, 2013); lower than 150 mm h\(^{-1}\) in British Columbia (Balfour and Woods, 2013); and much lower than values for other laboratory measurements (138–600 mm h\(^{-1}\)) (Bodí et al., 2012; Woods and Balfour, 2010). In experiments on ash from Coniferous fuels that were burned in a muffle furnace, Balfour and Woods (2013) found that K_{ash} (12–24 mm h\(^{-1}\)) produced from moderate burn temperatures (500–700 °C) was much lower than the K_{ash} (180–720 mm h\(^{-1}\)) of ash from low (300 °C) and high (900 °C) burn temperatures. The large range of K_{ash} values across the various studies may therefore partly be attributed to the effects of burn temperature on ash composition. With gravel (D < 2 mm) removed, the K_{ash} at Stony Creek became much lower (6.8 mm h\(^{-1}\)) while it remained high at Sunday Creek (36.4 mm h\(^{-1}\)).

The large variability in hydraulic properties of ash combined with variable subsurface soil properties makes it difficult to generalize the role of ash (K_{ash}) in controlling steady-state infiltration. At Stony Creek the mean K_{mat} (16.0–21.5 mm h\(^{-1}\)) was equal to the hydraulic conductivity of surface material (K_{ash} = 21 mm h\(^{-1}\)) on two occasion during the sampling campaign within the first year of the fire (Table 2), suggesting the K_{ash} may have been restricting steady-state infiltration. In dry conditions however, the effect of
water repellency at Stony Creek was stronger than ash effects ($K_{\text{mat}} < K_{\text{ash}}$), thus resulting in subsurface controls on steady-state infiltration, which is similar to results in Kinner and Moody (2010). At Sunday Creek, the mean matrix flow ($K_{\text{mat}}$ = 16.7–19.6 mm h$^{-1}$) was always less than the saturated hydraulic conductivity of ash/surface material ($K_s$ = 41 mm h$^{-1}$), suggesting the surface layer of ash was not limiting the steady-state infiltration in the soil matrix.

4.3. Macropore availability during recovery from wildfire

Macropores were largely inactive during the plot-scale rainfall experiments and $K_{\text{mat}}$ was therefore a better predictor of steady-state infiltration than the infiltration capacity ($K_p$), which is the sum of matrix and macropore flow ($K_{\text{mat}} + K_{\text{mac}}$). However, the activation of macropore flow depends on supply of water at the soil surface (Leonard et al., 2004; Nyman et al., 2010) and therefore depends on spatial scale and rainfall intensity. The contribution of macropore flow, $K_{\text{mac}}$, to infiltration is likely to increase with ponding, which increases with rainfall intensity and flow accumulation (Karssenberg, 2006; Langhans et al., 2011, 2013). Thus, the effective infiltration rates on hillslopes, when the supply of water at the surface is greater than the infiltration capacity of the matrix ($R > K_{\text{mat}}$), is likely to be somewhere between $K_{\text{mat}}$ and $K_p$ (Smith and Goodrich, 2000).

The linear increase in macropore availability during recovery shows that wildfire causes an initial reduction in infiltration capacity of macropores. The exact mechanism underlying this effect was not established but previous studies have shown that ash can limit flow by acting as a restricting layer and/or by clogging of pores within the soil profile (Ebel et al., 2012; Mallik et al., 1984; Nyman et al., 2010; Onda et al., 2008; Woods and Balfour, 2010). This effect may not be that pronounced for flow in the soil matrix where capillary processes dominate. However, an ash layer may be much more important in reducing infiltration capacity ($K_p$) when the surface is inundated, because a layer of unstructured material (i.e. ash) on the soil surface would limit the rate at which ponding water can be delivered to macropores in the underlying soil. Hence the structural components of the soil are unlikely to infiltrate at full capacity. At Ella Creek, 2–3 years post-wildfire, macropore flow made up 88–93% of the steady-state infiltration capacity, $K_p$. This suggests that increase in macropore availability with time since fire may be a key feature resulting in increased infiltration at during recovery. This effect however may go undetected in small runoff plots or conventional rainfall simulation experiments because of the strong sensitivity of $K_p$ to the supply of water at the soil surface.

5. Conclusion

The overall objective of the study was to identify key properties contributing to variability in infiltration during recovery from wildfire and to quantify their effect. A storage-based infiltration model was used as framework for analyzing infiltration data. The model fitted the data well and could be used effectively to partition the infiltration process into its key components of storage, matrix flow and macropore flow. Infiltration on intact cores showed that a water repellent layer acted as a ‘throttle’ on flow and that the soil moisture status of the soil therefore could have large effects on flow through the soil matrix. The slow wetting process in the water repellent layer meant that only after prolonged exposure to saturated conditions did the soil become fully saturated and able to conduct water at its full potential. In headwater catchments that were recovering from wildfire ($t_w < 3$ years) the soils always displayed some degree of water repellency resulting in incomplete saturation, which meant that the saturated hydraulic conductivity of the soil ($K_s$) could not be measured in the field. The variability in the infiltration capacity ($K_p$) during recovery was driven primarily by changes in macropore availability but also by temporal variability in water repellency due to meteorological conditions.

The results from infiltration measurements in headwater catchments were used to model the spatial and temporal variability in infiltration processes as a function of time since fire ($t_f$) and catchment dryness ($K_{\text{BDI}}$), which was calculated from monthly weather patterns. The relations underlying the model can be summarized as follows: 1. Water repellency was represented by the minimum critical surface tension ($CST_{\text{min}}$) in the soil profile. 2. The $CST_{\text{min}}$ increased exponentially with decreasing $K_{\text{BDI}}$ (i.e. the strength of water repellency decreased exponentially when the monthly weather resulted in wet conditions). 3. The matrix flow ($K_{\text{mat}}$) in turn, increased exponentially with increasing $CST_{\text{min}}$. 4. The storage ($H$) declined exponentially with $t_{sf}$ due to the loss of a discrete layer of wettable surface soil. 5. At the same time, the macropore flow increased linearly with $t_{sf}$ due to increased macropore availability during recovery from the fire impact. The relations are given as empirical equations in Figs. 7 and 8. Future work will aim to model infiltration during actual storm events and test these relations against observed runoff responses from burned hillslopes.

Overall the results of the study highlight that models of infiltration into burned soils should take into account (i) the surface storage in ash and burned soil, (ii) the effect of ponding on macropore flow, and (iii) non-uniform wetting in a soil layer that displays varying levels of water repellency depending on meteorological conditions.

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