Stormflow response and 'effective' hydraulic conductivity of a degraded tropical *Imperata* grassland catchment as evaluated with two infiltration models

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Key points:

- The Spatially Variable Infiltration model outperformed the Green–Ampt model when simulating hydrographs, especially for multi-peak events.
- SVI-model parameter values varied markedly, but were correlated with antecedent topsoil moisture content
- Model-derived infiltration capacities were much higher than field-measured K_{sat}, regardless of the model or field method used

Abstract

Predicting catchment stormflow responses after tropical deforestation remains difficult. We used five-minute rainfall and storm runoff data for 30 events to calibrate the Green–Ampt (GA) and the Spatially Variable Infiltration (SVI) model and predict runoff responses for a small, degraded grassland catchment on Leyte Island (the Philippines), where infiltration-excess overland flow is considered the dominant storm runoff generating process. SVI replicated individual stormflow hydrographs better than GA, particularly for events with a small runoff response or multiple peaks. Calibrated parameter values of the SVI model (*i.e.*, spatially averaged maximum infiltration capacity, I_m and initial abstraction, F_0) varied markedly between events, but exhibited significant negative linear correlations with (mid-slope) soil water content at 10 cm (SWC_{10}) – as did the 'catchment effective' hydraulic conductivity (K_e) of the GA model. SWC_{10}-based values of F_0 and I_m in SVI resulted in satisfactory to good predictions (NSE > 0.50) for 18 out of 26 storms for which data on SWC_{10} were available, but failed to reproduce the hydrographs for six events (23%) with mostly small runoff responses. Median values of field-measured near-surface K_{sat} (~2–3 mm h^{-1}, depending on method) were distinctly lower than the median I_m (32 mm h^{-1}) and, to a lesser extent, K_e (~8 mm h^{-1}), confirming previously suspected under-estimation of field-measured K_{sat}. Using pre-storm topsoil moisture content and 5-min rainfall intensities as the driving variables to model infiltration with SVI gave more realistic results than the classic GA approach or the comparison of rainfall intensities with field-measured K_{sat}.

Plain Language Summary

It is important for flood management to be able to predict the volume and peak value of streamflow during intense rainfall (so-called 'stormflow'). We used rainfall and streamflow data for a small, degraded tropical grassland catchment on Leyte Island (the Philippines) to calibrate two rainfall infiltration models of different complexity: the simple Green–Ampt model (GA) and the Spatially Variable Infiltration (SVI) model that describes rainfall infiltration into the soil as a function of the intensity of the rain. SVI generally performed better than GA in simulating observed stormflow responses, especially for events with multiple rainfall peaks. Values for the two main parameters of SVI (the amount of rainfall required to initiate stormflow, and the maximum infiltration capacity of the soil) varied with the amount of moisture in the top 10 cm of the soil prior to the rain. Using the measured topsoil moisture contents for 26 rainfall events to estimate the SVI parameter values and predict the stormflow response from the measured rainfall intensity produced satisfactory to good results for ~70% of the examined storms. However, it failed to reproduce the stormflow patterns for six events with mostly small to very small runoff responses.

1 Introduction

Large areas in the humid and seasonal tropics suffer moderate to severe soil degradation (Bai et al., 2008; Gibbs & Salmon, 2015). Repeated cycles of slash-and-burn cultivation, as well as more intensive forms of agricultural cropping and grazing, have resulted in reductions in topsoil organic matter content, soil faunal activity and macroporosity, and an increase in bulk density (Martinez & Zinck, 2004; Shougrakpam et al., 2010; Recha et al., 2012; Zwartendijk et al., 2017; Toohey et al., 2018). The associated decline in soil infiltration capacity typically leads to increased occurrence and amounts of infiltration-excess overland flow (IOF) in regions and/or periods with high rainfall intensities (Chandler & Walter, 1998; Ziegler et al., 2004; Molina et al., 2007; Ghimire et al., 2013; Bush et al., 2020). IOF, in turn, causes accelerated erosion, as well as higher runoff peaks at the headwater catchment scale (Ziegler et al., 2009; Liu et al., 2011; Recha et al., 2012; Ribolzi et al., 2017; Birch et al., 2021a, 2021b), which exacerbates flooding and sedimentation problems downstream (Bruijnzeel, 2004; Sidle et al., 2006; Valentin et al., 2008; Yin et al., 2019).

Despite the extent of tropical land degradation and associated environmental problems, comparatively little progress has been made with the quantitative prediction of storm runoff for degraded tropical catchments (Yu, 2005; Sidle et al., 2006; Ribolzi et al., 2017; Yamamoto et al., 2021; Birch et al., 2021a). Some of the more frequently used approaches include the Green–Ampt infiltration model (GA; Mein & Larsen, 1973; Chu, 1978) and the US Soil Conservation Service curve number (SCS–CN) method (Ponce & Hawkins, 1996). GA partitions rainfall between infiltration and IOF. The SCS–CN method estimates 'direct runoff' (*i.e.*, a fast runoff component that is assumed to be linearly related to rainfall) from hillside plots or small catchments using a dimensionless 'curve number' (CN-value) that is assumed to capture catchment-wide water retention as a function of soil texture, drainage conditions and land cover/use (Ponce & Hawkins, 1996). Both approaches have their limitations. Although GA takes short-term changes in infiltration rate as the soil wets up during rainfall into account (Koorevaar et al., 1983), the method applies only to individual points. The notoriously high spatial variability of near-surface saturated soil hydraulic conductivity (K_{sat}) makes it difficult to obtain 'representative' estimates at the hillslope- to catchment scale (Sharma et al., 1987; Dunne et al., 1991; Chappell et al., 1998; Zehe & Flühler, 2001; Campos Pinto et al., 2018). Hence, a spatially uniform 'effective' final infiltration rate (K_e) is usually assumed in catchment-scale applications of GA (Aston & Dunin, 1979; James et al., 1992; Nearing et al., 1996; Leemhuis et al., 2007; Yira et al., 2016). More importantly, once infiltration reaches steady-state condition, infiltration rates as predicted by GA do not respond to changes in rainfall input anymore (Yu, 1999), despite ample evidence to the contrary (Hawkins, 1982; Dubrueil, 1985; Dunne et al., 1991; Yu et al., 1997a; Stone et al., 2008). On the other hand, the SCS–CN method is incapable of providing information on the spatio-temporal variation in storm runoff (Garen & Moore, 2005; Ogden et al., 2017). Nevertheless, GA or SCS–CN constitute a core element of widely used erosion and hydrological models, such as WEPP (Flanagan et al., 2001; Nearing et al., 1996) and SWAT (Neitsch et al., 2011; Arnold et al., 2012). Therefore, Ogden et al. (2017) called for the identification of '*more appropriate dynamic hydrological formulations for different hydro-geographic regions*' (such as the tropics) to replace the static and spatially lumped SCS–CN method, as did Yu (1999) in relation to GA (*cf*. Yamamoto et al., 2020).

Arguably, in areas with significant surface degradation, where IOF is likely to be the dominant storm runoff generation mechanism (Sutherland & Bryan, 1990; Mathys et al., 1996; Chandler & Walter, 1998; Molina et al., 2007), a dynamic model of infiltration that takes the spatial variability of surface K_{sat} into account, *as well as* the positive impact of rainfall intensity on infiltration rates (Hawkins, 1982; Dunne et al., 1991), would go some way towards the improved process description called for by Ogden et al. (2017). Building upon earlier work by Hawkins and Cundy (1987), Yu et al. (1997a) developed a spatially variable infiltration model (SVI) that relates actual infiltration rates at the plot scale (as determined by subtracting measured IOF from rainfall over short consecutive periods) to rainfall intensity and a spatially averaged infiltration parameter. SVI proved to be consistently superior to GA with regard to predicting IOF from (mostly large) storms on (mostly bare) hillside plots at various tropical sites (Yu, 1999). Fentie et al. (2002) considered SVI the best choice amongst eight different methods to predict IOF from grazed plots in Queensland, whereas Van Dijk and Bruijnzeel (2004) concluded that SVI provided a '*robust and accurate method for predicting runoff*' from terraced fields on volcanic substrate in Indonesia. Recently, Z. Cheng et al. (2018) compared the performance of GA and SVI under much drier conditions on the Chinese Loess Plateau, and concluded that the amount of simulated IOF was less sensitive to changes in model parameter values for SVI than for GA. Despite SVI's superior performance at the plot scale in a range of tropical settings (Yu, 1999; Fentie et al., 2002; Van Dijk & Bruijnzeel, 2004; *cf.* Patin et al., 2012), the model has so far not been used to predict stormflow at the *catchment scale*. Conversely, GA has been used extensively for this purpose (*e.g.*, Aston & Dunin, 1979; Van Mullem, 1991; James et al., 1992; Obiero, 1996; Conolly et al., 1997; Leemhuis et al., 2007; Yira et al., 2016; Yamamoto et al., 2020).

This paper marks the first attempt to evaluate SVI's ability to predict stormflow hydrographs and peak discharge, using detailed rainfall and streamflow data for the 3.2 ha Basper catchment on Leyte Island (the Philippines). After decades of slash-and-burn, much of the catchment is covered by *Imperata* and *Saccharum* grasses. Fire-climax grasslands constitute a widespread form of degraded land, occupying an estimated area of up to 57 million ha across South and Southeast Asia in the early 1990s (Garrity et al., 1997). More than two-thirds of the estimated 6.5 million ha under *Imperata* in the Philippines (17% of the national land base) were classified as experiencing moderate to severe surface erosion (Concepcion & Samar, 1995). Despite its widespread existence, quantitative hydrological information for this type of grassland is scant (Jasmin, 1976; Lim Suan, 1995; *cf.* Sirimarco et al., 2018). Earlier work in the Basper catchment revealed very low (near-) surface values of K_{sat}, suggesting the likelihood of frequent IOF occurrence, even though K_{sat} may have been under-estimated (Zhang et al., 2019a). Nearly two-thirds of the annual streamflow at

Basper consists of stormflow (here defined as the component of the hydrograph above the Hewlett and Hibbert (1967) separation line), rendering the catchment one of the hydrologically most responsive humid tropical sites described to date (Zhang et al., 2018a; *cf.* Chappell et al., 2012; Birkel et al., 2021). Although no explicit measurements of hillslope IOF were made at Basper, the extreme dilution of streamflow during rainfall events (Zhang et al., 2018a; Van Meerveld et al., 2019) and isotope hydrography separation results (Van Meerveld et al., 2019) all suggest a major contribution of low electrical conductivity 'new water' to stormflow. Hence, our objectives were to: (*i*) test the appropriateness and relative performance of GA and SVI for describing storm runoff for a small catchment in a state of advanced surface degradation; (*ii*) examine the temporal variability of the calibrated model parameters, and their relationships with antecedent soil water content and rainfall characteristics; and (*iii*) compare calibrated model infiltration parameter values with the previous field measurements of K_{sat} by Zhang et al. (2019a) to assess the degree of possible under-estimation of the latter at the catchment scale.

2 Materials and methods

2.1 Study area

The south-facing 3.2-ha Basper catchment (11°15'28" N; 124°57'22" E) is located 14 km west of Tacloban, the capital of Leyte Island. Elevations range from 50–135 m a.s.l. The climate is tropical ever-wet (Köppen-type Af) with a mean annual rainfall at Tacloban Airport (1977–2012) of 2,660 mm (range: 1,435–4,790 mm), distributed over 195 rain days (with ⩾0.5 mm of rain each) on average per year. There is no clear dry season, but average monthly rainfall totals are distinctly higher (>350 mm mo^{-1}) between November and January than for April–May (>100 mm mo^{-1}). Typhoons and tropical storms can bring large amounts of rain and supply roughly one-third of the annual rainfall in the region (Cinco et al., 2016). Between 1977 and 2011, ~50% of all rain days at Tacloban Airport received less than 5 mm of rain. Considering only events with ⩾5 mm of rain, 64% of storms were 5–20 mm in size, whereas 10% and 2.5% of events were larger than 50 and 100 mm, respectively. The median 5-, 15-, 30-, and 60-min rainfall intensities measured at Basper during 99 events with at least 5 mm of rain between June 2013 and May 2014 were 3.2, 2.1, 1.5 and 1.0 mm h^{-1}, respectively. Corresponding 95^{th}-percentile intensities were 34, 22, 18, and 12 mm h^{-1}.

The upper slopes are straight to slightly concave, while foot-slopes generally steepen towards the stream. Landslides are a prominent feature and made up 3.4% of the catchment area at the time of the investigation (Zhang et al., 2018a; Figure 1). The vegetation consists of *cogon* grass (*Imperata cylindrica*) on the ridges and upper slopes, with additional sedge (*Cyperus* sp.) in less well-drained parts. The mid-slope parts have a mixture of *Saccharum spontaneum* grass and low shrub (<1.5 m, mostly *Melastoma* and *Chromolaena*), while shrubs and young trees (<3 m, mostly *Neonauclea* and *Leukosyke*) are common on the lower slopes. Although regularly burned in the past, the area did not experience fire after 2003 and young regenerating forest occupied an estimated 4,500 m^2 (~14%) in the central portion of the catchment at the time of the study (Figure 1).

Eutric Cambisols of predominantly clay loam texture, grading to a sandy clay loam below 90 cm depth, overlay the gabbro bedrock. Soil organic carbon content, porosity and drainable pore space decline with depth, while median bulk densities increase with depth in the top 40 cm (Zhang et al., 2019a). The median (± median absolute deviation, MAD) steady-state surface infiltration rate (determined using a portable double-ring infiltrometer with inner and outer ring diameters of 15 and 21 cm) was 2.1 ± 0.7 mm h^{-1} (n = 13). The median near-surface K_{sat} (<10 cm depth) obtained from small cores (laboratory permeameter) was 1.7 ± 1.6 mm h^{-1} (n = 27). The median K_{sat} at ~20 cm depth as derived with a constant-head well permeameter was 2.7 ± 2.2 mm h^{-1} (Amoozegar, 1989; n = 20; see Zhang et al. (2019a) for details).

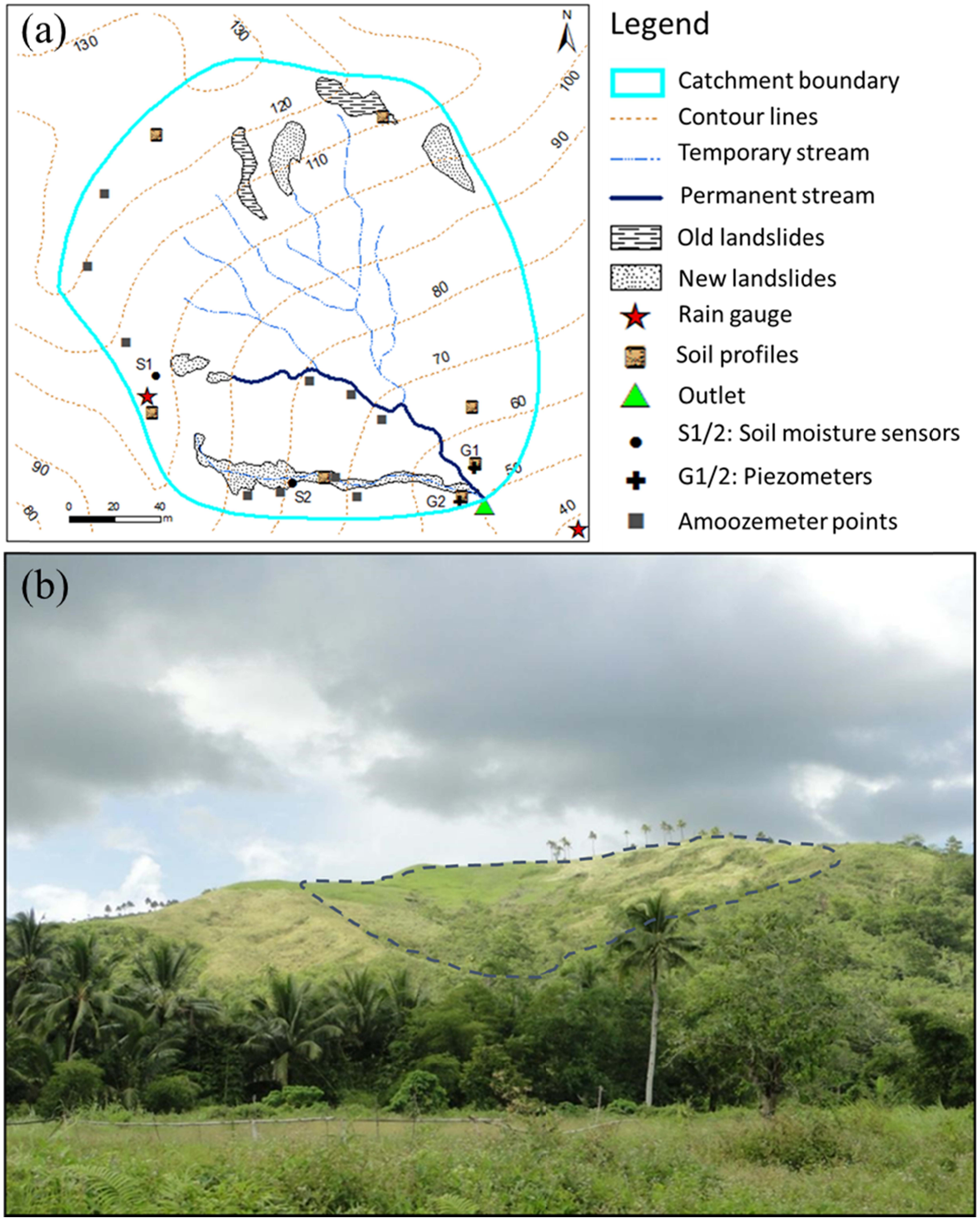

Figure 1. Basper micro-catchment. **(a)** Map showing the drainage network, and locations of landslides, hydrological instrumentation, soil profiles (core sampling and double-ring infiltration sites), and soil hydraulic conductivity measurements using well permeametry. **(b)** Photo showing the land cover. The broken line indicates the catchment boundary. Photo credit: Jun Zhang.

2.2 Methods

2.2.1 Hydrological monitoring

For this study, we used measurements of rainfall, streamflow, soil water content and foot-slope groundwater levels taken between 3 June and 7 November 2013. These measurements represent the conditions prior to the major disturbance to vegetation and soils by Typhoon Haiyan on 8 November 2013 (Zhang et al., 2018a).

Rainfall (P) was measured using two Onset Computer Corporation RG3 tipping-bucket rain gauges (0.25 mm per tip, confirmed by manual calibration) connected to a HOBO Pendant event data-logger. One gauge was located in the open near the catchment outlet and the other on the upper western ridge (Figure 1a). A standard manual rain gauge (100 cm^2 orifice) was placed next to each of the recording gauges and read every morning as a check.

Streamflow (Q) was measured using a sharp-crested compound weir consisting of a 0.55 m high 90° V-notch and a horizontal beam extending 0.5 m to each side from the edge of the V-notch (Zhang et al., 2018a). Water pressure was measured at five-minute intervals using a HOBO U20L04 logger and corrected for atmospheric pressure, which was measured by a similar device in a hut located ~100 m from the weir. The standard V-notch weir equation (Bos, 1989) was checked through volumetric discharge measurements below 4.4 $l\ s^{-1}$ (staff heights < 0.3 m), and Price Type-AA current-meter measurements at stages up to 0.55 m. Water levels exceeded the shoulder of the V-notch for ~1.3% of the total duration of the 30 selected storm events (Section 2.2.2), representing ~33% of the corresponding total storm runoff amount. For these conditions the Bergmann compound weir equation as given by USBR (1997) was used to calculate the streamflow.

Volumetric soil moisture content (θ) *and shallow groundwater levels* were monitored at different sites within the catchment (Figure 1a; Zhang et al., 2018a). The present analysis only used soil moisture data from site S2 (*Saccharum* grassland at mid-slope position) and shallow groundwater levels as measured at piezometer site G1 (left bank, 0.9 m deep; Figure 1a). Soil moisture at S2 was measured at five-minute intervals using simplified Time Domain Reflectometry (TDR) sensors (MP-306, ICT

International, Australia) installed at 0.1, 0.2, 0.4, 0.6, 0.8, and 1.1 m below the surface, and connected to an ICT International Microvolt data-logger. Water levels in piezometer G1 were also measured at five-minute intervals using a HOBO U20L04 logger.

2.2.2 Stormflow separation and event selection

To separate stormflow (Q_q) from baseflow (Q_b), the constant-slope method of Hewlett and Hibbert (1967) was applied to the streamflow record for each event prior to Typhoon Haiyan (3 June–7 November 2013). The following criteria were used to define the start of each 'stormflow event': (*i*) total rainfall $\geqslant$ 5 mm; and (*ii*) the event was preceded by a rain-free period $\geqslant$ 6 h. For the end of an event, a threshold value of 0.005 mm per five minutes (0.06 mm h^{-1} equivalent) was used. Furthermore, we only included events for which the five-minute rainfall- and streamflow measurements were complete (*i.e.*, no data gaps). Lastly, events for which > 10% of the stormflow could have been generated by precipitation falling directly onto the perennial stretch of the stream (~90 m^2 or 0.28% of the total catchment area; Figure 1a) were excluded. Thus, only events with a minimum hillslope runoff contribution > 90% were considered. Application of the above criteria yielded 30 stormflow events for comparative testing of the Spatially Variable and Green–Ampt infiltration models.

2.2.3 Likelihood of an overland flow dominated system as inferred from the transit time of the rainfall-to-streamflow wave propagation

The extremely low sub-soil K_{sat}-values determined in the field (Zhang et al., 2019a), the high electric conductivity of foot-slope groundwater and pipe flow (~270 μS cm^{-1}) but strong dilution of streamflow during times of stormflow (Zhang et al., 2018a; Van Meerveld et al., 2019), and high event-water contributions to stormflow (Van Meerveld et al., 2019) all suggest that runoff generation in the Basper catchment is dominated by IOF. But before comparing the performance of the GA and SVI models for the prediction of catchment-wide overland flow generation, we investigated whether the selected storm runoff events were more likely to be generated primarily by IOF than return flow and saturation overland flow (SOF) (*cf*. Dunne & Black, 1970; Lapides et al., 2022) in more detail. Using a data-based mechanistic modeling approach, the rainfall-generated streamflow response time was compared to that of the groundwater level in piezometer G1 in the riparian zone that is influenced by lateral subsurface flow. To facilitate the comparison, the observed piezometer water levels were converted to pore-water depth equivalents by multiplying the water levels times the measured soil porosity (Zhang et al., 2019a). The response times (strictly speaking, of the celerities, not of the velocities of the water particles), were identified from optimal Transfer Function (TF) models using the Nash-Sutcliffe model efficiency (NSE; Nash & Sutcliffe, 1970) and a heuristic measure that helps avoid selection of over-parameterised models (the Young Information Criterion; Young, 2001) as selection criteria. A discrete-time, rather than continuous-time, transfer function identification algorithm was used (Chappell et al., 1999) to account for the presence of occasional short breaks in the observed streamflow record. This algorithm, RIVID, is part of the CAPTAIN Toolbox for Matlab (Taylor et al., 2007). A wide range of model structures were evaluated covering first- to third-order models, with pure time delays ranging from zero to 30x the five-minute time-steps, and various non-linearity transformations, including the established Store–Surrogate (Chappell et al., 1999) and Bedford–Ouse approaches (Chappell et al., 2006).

The modeling identified time constants of first-order (*i.e.*, single pathway) models of the rainfall-streamflow response that varied between 9–15 min, depending on the event. These times are much smaller (*i.e.*, the response is much faster) than for the cyclone-affected South Creek basin in Queensland, where hillside SOF is important (Chappell et al., 2012), but comparable to those derived for overland flow plots (*e.g.*, Chappell et al., 2006). This suggests a dominance of IOF at Basper. Further, the subsurface response to rainfall in the riparian zone at Basper (*i.e.*, foot-slope groundwater levels) was 9–70 h and thus much slower than the streamflow response to rainfall (9–15 min), again pointing to IOF as the main mechanism for stormflow generation (see examples in Supporting Information S1).

2.2.4 Infiltration models

Two models of contrasting complexity were used to quantify the infiltration process and to derive the associated amounts of excess rainfall (r_e). However, an identical runoff routing algorithm was employed in both cases for subsequent comparison with the observed storm runoff hydrographs at the catchment outlet.

The first model is based on the Green–Ampt (GA) equation, in which the infiltration capacity (i_c) is expressed as a function of the cumulative infiltration amount, F (in mm) as:

$$i_c = K_e \left(1 + \frac{\psi_m}{F}\right) \tag{1}$$

where K_e (mm h^{-1}) can be regarded as the 'effective' saturated hydraulic conductivity of the surface soil, and ψ_m (mm) as the 'effective' matric potential at the wetting front across the catchment. An application of the GA equation for a rainfall event of constant intensity was developed by Mein and Larsen (1973) and for an event of varying intensity by Chu (1978). Computational procedures are described in detail by Chow et al. (1988). Briefly, for each time interval j, given a rainfall intensity p, and a cumulative infiltration F at the beginning of the interval, there are three possible scenarios for the actual rate of excess rainfall (r_e): (*i*) $p < i_c$, and $r_e = 0$ throughout the interval; (*ii*) ponding condition, *i.e.* $i_c = p$, is met at some point during the time interval; or (*iii*) ponding has occurred and p exceeds i_c throughout the interval, hence $r_e = p - i_c$. In each of the three cases, F is updated to the end of the time interval.

The second model was SVI (Yu et al., 1997a), which conceptualizes overland flow generation during two distinct phases. At the start of an event, i_c is typically much larger than p, and an initial abstraction, F_0 (in mm) is used to represent the amount of infiltration prior to the commencement of excess rainfall. In other words, r_e is zero at this stage – irrespective of rainfall intensity, as long as cumulative rainfall is less than F_0:

$$r_j = 0, when\ \sum_{i=1}^{j} p_i \leq F_0 \tag{2}$$

Once cumulative rainfall has exceeded F_0, the actual rate of infiltration, i_a is modeled as a function of the rainfall intensity and a spatially averaged maximum infiltration rate, I_m (both in mm h^{-1}). The main assumption behind the SVI model is that i_c varies in space according to an exponential distribution that involves I_m as a single parameter (Yu et al., 1997a; *cf.* Hawkins & Cundy, 1987; Supplementary Figure S1).

It can be shown (Yu et al., 1997a) that:

$$i_a = I_m(1 - e^{-p/I_m}) \tag{3}$$

Application of either SVI or GA leads to a time series of excess rainfall on hillslopes as the difference between rainfall intensity and the modeled rate of infiltration:

$$r_e = p - i_a \tag{4}$$

To take the rain falling directly on the surface of the perennial stream (*i.e.*, channel precipitation; see Section 2.2.2 above for rationale) into account, the total excess rainfall, r^* was expressed as the area-weighted sum of rainfall excess over the stream channel and that over the hillslopes:

$$r^* = (1 - f_w)r_e + f_w p \tag{5}$$

where f_w is the fractional area of the perennial stream channel (in this case: 0.28%).

Regardless of the infiltration model used, for each time interval, j, with excess rainfall computed using equations (4) and (5), r^* is routed to the catchment outlet using a simple kinematic wave approximation:

$$Q_j = \alpha Q_{j-1} + (1 - \alpha) r_j \tag{6}$$

where Q_j is the stormflow rate at the catchment outlet for time interval j (in mm h^{-1}). The routing parameter, α, is related to the catchment lag time, T (in hours), and the adopted time interval for the rainfall and storm runoff observations, Δt as follows (Yu et al., 1997a):

$$\alpha = \begin{cases} T/(T + \Delta t) & T \leq \Delta t/2 \\ (2T - \Delta t)/(2T + \Delta t) & T > \Delta t/2 \end{cases} \tag{7}$$

The advantage of using Equation (6) for routing is the guaranteed numerical stability, irrespective of the magnitude of T relative to Δt.

2.2.5 Model calibration and evaluation

The parameters for the two models were optimized by minimizing the sum of squared errors (SSE) between the observed and modeled stormflow using the Levenberg-Marquardt algorithm (Marquardt, 1963):

$$\min SSE = \sum_{j=1}^{N}\left(Q_j - \hat{Q}_j\right)^2 \quad (8)$$

where $\hat{Q}_j$ and Q_j are the modeled and observed stormflow rates, respectively (in mm h^{-1}), and N is the total number of time intervals for the event. Model parameters were calibrated for each individual event to account for temporally varying infiltration rates, resulting in 30 parameter sets (one for each event) for each of the two infiltration models. The two infiltration models were fully integrated with the Parameter ESTimation Software (PEST++) for efficient parameter estimation (White et al., 2020).

To evaluate model performance, the Nash-Sutcliffe efficiency was calculated for each of the 30 individual storm runoff hydrographs. Further, we computed the Sum of Squared Errors, percent bias (PBIAS; Gupta et al., 1999), and the ratio between the RMSE of the observations and their standard deviation (RSR; Legates & McCabe, 1999) for each event. Although the two infiltration models were applied primarily to test their ability to predict storm hydrographs at five-minute intervals, model performance was also examined in terms of stormflow amount (Q_q) and peak runoff rate (Q_p) for individual events.

2.2.6 Relations between infiltration model parameters and event characteristics

The calibrated infiltration model parameters were related to event rainfall characteristics to examine whether – and to what extent – the model parameters were affected by rainfall characteristics and antecedent conditions. The main event characteristics used in the Spearman rank correlation analysis were the peak intensity and the maximum rainfall intensities during 15 and 30 min. The main indicators of antecedent wetness conditions were the three-day antecedent precipitation index (API_3) and the volumetric water content in the top 10 cm of the soil at mid-slope position (SWC_{10}). The Antecedent Precipitation Index (API) is a measure of catchment wetness based on the rainfall that occurred over preceding days and was calculated as:

$$API = \sum_{n=1}^{N} P_n k^n \quad (9)$$

where P_n is the precipitation during the nth day preceding the day for which the API is calculated, and k is a decay constant. Given the small size and comparatively shallow soils of the study catchment, we decided to use a three-day antecedent precipitation index (API_3) using a k value of 0.80 (Shaw et al., 2010).

2.2.7 Comparison of hydraulic model parameters with field measurements

The point-measured K_{sat} data from Zhang et al. (2019a) were compared directly with the model-calibrated values of near-surface K_{sat} (*i.e.*, K_e in GA and I_m in SVI). In addition, the *distributions* of the two data series were compared, noting that the underlying idea of the SVI model is that the spatial variation in infiltration capacity i_c can be described by an exponential distribution of the maximum infiltration capacity I_m according to Equation (3) (Yu et al., 1997a). To approximate an overall distribution of i_c for the Basper catchment, the 30 event-based values of I_m were each inserted separately into Equation (3) to derive the corresponding distributions of i_c. The average distribution for all 30 events was regarded as representing the overall spatial distribution of i_c across the catchment. Because differences in mean field K_{sat} based on portable-ring infiltrometry (n = 13), near-surface well permeametry (n = 20), and laboratory permeametry on small cores (n = 27) were not statistically significant (p-value > 0.35), all data were bulked (n = 60).

3 Results

3.1 Characteristics of selected storm events

Rainfall amounts for the 30 events ranged from 6.6 to 149 mm, with a mean of 26 mm (median 18.5 mm; Figure 2). Event total stormflow at the catchment outlet varied from 0.3 mm to 76 mm, averaging 7.2 mm (median 3.5 mm), while stormflow runoff coefficients (Q_q/P) ranged from 3–56%, averaging 21% (median 18%). Collectively, these events represented ~66% of the total rainfall during the 3 June–7 November 2013 study period (1,187 mm) and ~92% of the total storm runoff (235 mm; Zhang et al., 2018a). Event duration (defined as the time between the initial rise in discharge and the stormflow cut-off point; Section 2.2.2) varied from 0.8 to 40.8 h, averaging 10.4 h (median 6.0 h). Event-averaged rainfall intensity (4.3 mm h^{-1}; median 3.3 mm h^{-1}) was approximately an order of magnitude smaller than the five-minute peak rainfall intensity (average: 58 mm h^{-1}, median 55 mm h^{-1}; Figure 2). Based on their Q3/Q1-ratios (*i.e.*, between the third and first quantiles), rainfall amounts and peak rainfall intensities varied less between events than stormflow amounts and peak runoff rates (Figure 2).

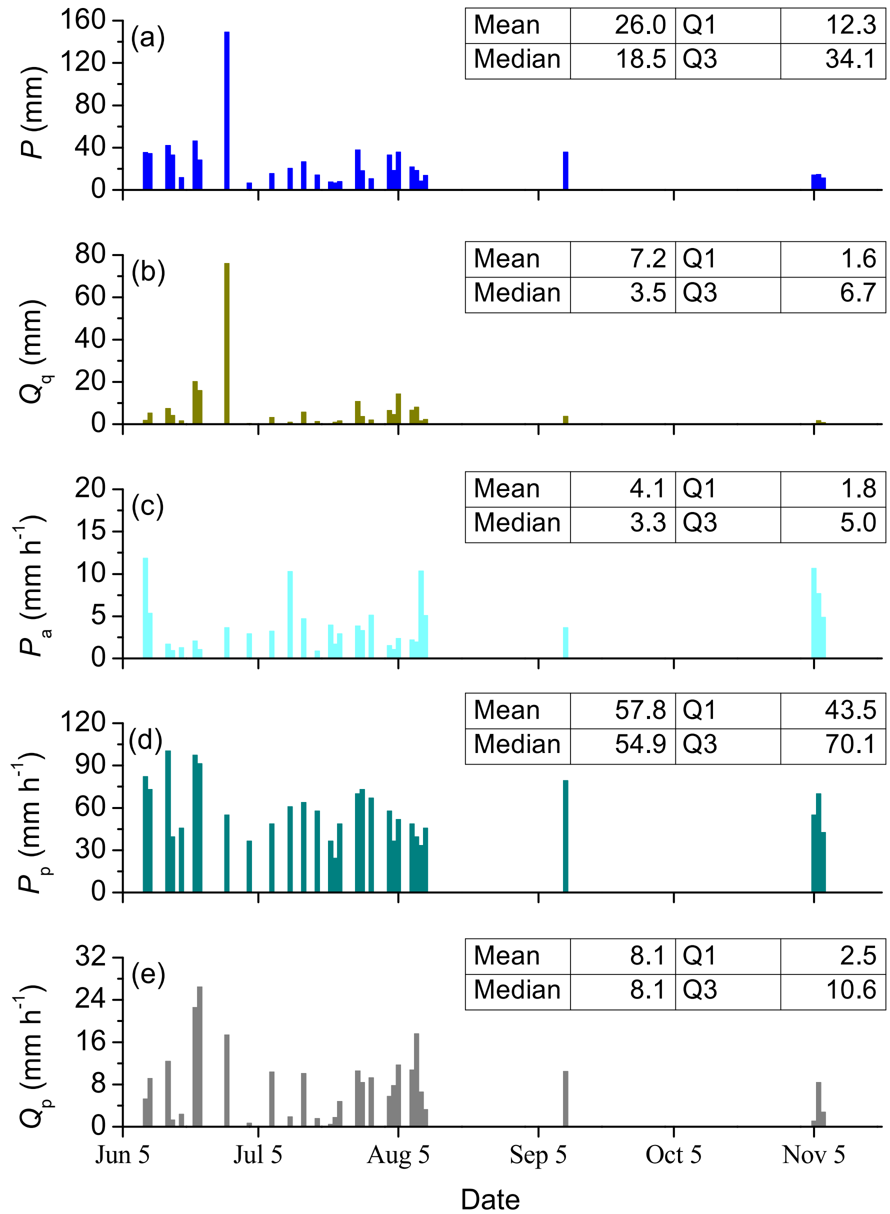

Figure 2. Time series showing the basic characteristics of the 30 examined runoff events at Basper catchment between 6 June and 7 November 2013: **(a)** rainfall (P, mm), **(b)** total stormflow (Q_q, mm), **(c)** average rainfall intensity (P_a, mm h^{-1}), **(d)** peak rainfall intensity (P_p, mm h^{-1}) and **(e)** peak stormflow rate (Q_p, mm h^{-1}). Insets list the means, medians, as well as the first (Q1) and third (Q3) quantiles for the respective variables.

3.2 Comparative infiltration model performance

GA and SVI could be calibrated equally well to simulate event-based Q_q and Q_p, with simulated Q_q and Q_p being in good agreement with observed values (R^2-values of 0.98 and 0.99, respectively, regardless of the model used; Figure 3). Nevertheless, the median Q_q tended to be under-estimated by about 10% (GA) to 14% (SVI; Table 1) and by 21–22% for larger events (based on the slopes of the regression lines in Figure 3a). Peak runoff rates were under-estimated by about 8–12% (Figure 3b; Table 1). However, based on the higher NSE- and lower RSR-values, SVI performed slightly better than GA in terms of simulating event-based stormflow (Table 1).

Table 1. Model performance of SVI and GA for the prediction of storm runoff totals (Q_q) and peak runoff rates (Q_p) for the 30 examined storm events.

Model	Evaluation	Min	Q1	Median	Q3	Max
SVI	SSE[1] (Calibration)	0.04	1.6	4.6	23	418
	NSE[2]	0.57	0.84	0.92	0.95	0.99
	PBIAS[3]_Q_q	-2	6	14	26	54
	PBIAS[3]_Q_p	-12	2.5	8	18	39
	RSR[4]	0.10	0.23	0.28	0.40	0.66
GA	SSE[1] (Calibration)	0.03	2.1	6.8	29.5	589
	NSE[2]	0.12	0.74	0.88	0.94	0.99
	PBIAS[3]_Q_q	-10	-0.8	10	28	60
	PBIAS[3]_Q_p	-6	5	11.5	26	63
	RSR[4]	0.10	0.25	0.35	0.51	0.94

[1]Sum of squared errors; [2]Nash-Sutcliffe efficiency; [3]Per cent bias; [4]Ratio between the RMSE and the standard deviation of the observations

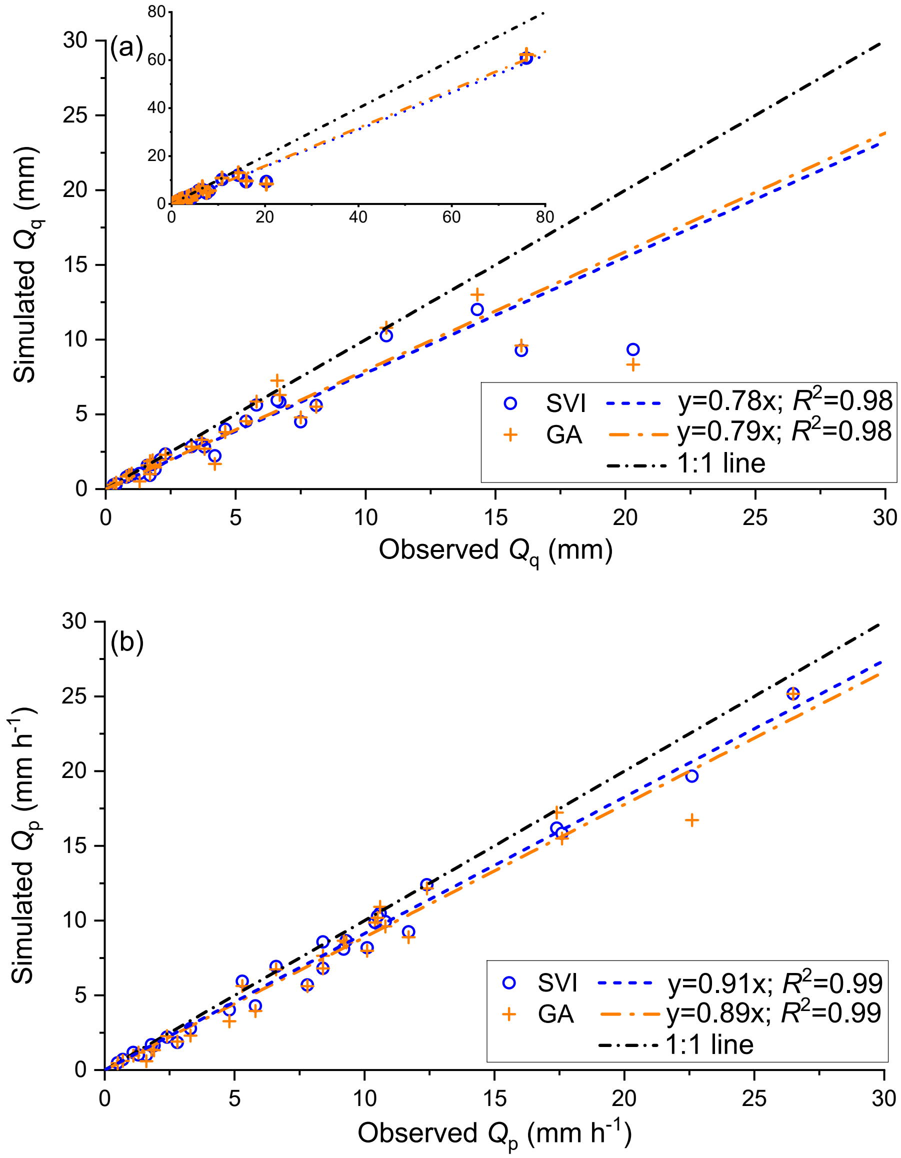

Figure 3. Comparison of the observed and modeled **(a)** stormflow totals (Q_q, mm) and **(b)** peak runoff rates (Q_p, mm h^{-1}) for the 30 examined runoff events. The models were calibrated for each individual event by minimizing the sum of squared errors.

Figure 4 shows the model performance in terms of the NSE-values derived for individual events *versus* corresponding stormflow runoff coefficients ($R_c = Q_q/P$). As indicated by the enveloping line, both models captured events with higher runoff coefficients better than events with lower R_c, for which low NSE-values suggested a poor model fit (Figure 4). Overall, SVI outperformed GA in terms of its ability to reproduce event-based hydrographs, with average NSE-values for all 30 events of 0.88 for SVI *versus* 0.81 for GA (difference significant at a p-value < 0.05). Out of 13 events with $R_c \leq 0.16$, three were captured poorly by GA (*i.e.,* NSE ≤ 0.50) *versus* none for SVI (Figure 4). Simulations for two specific events with multiple runoff peaks are presented in Figure 5 to illustrate the difference in model performance for complex events. GA missed the second peak of the hydrograph entirely for both events, whereas SVI was capable of simulating all peaks despite a certain degree of under-estimation. A similar pattern was noted for the events with a particularly high rainfall intensity at the beginning of the storm, which caused GA-modeled stormflow to occur earlier than observed (see Supplementary Figures S2a and S2b).

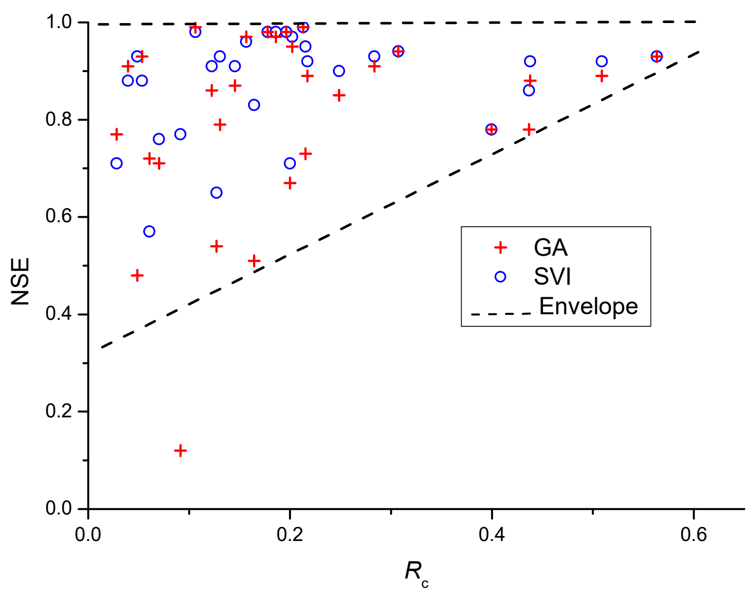

Figure 4. Relationship between stormflow runoff coefficient (R_c) and the Nash–Sutcliffe model efficiency as a measure of model performance for the GA and SVI models for the 30 examined events.

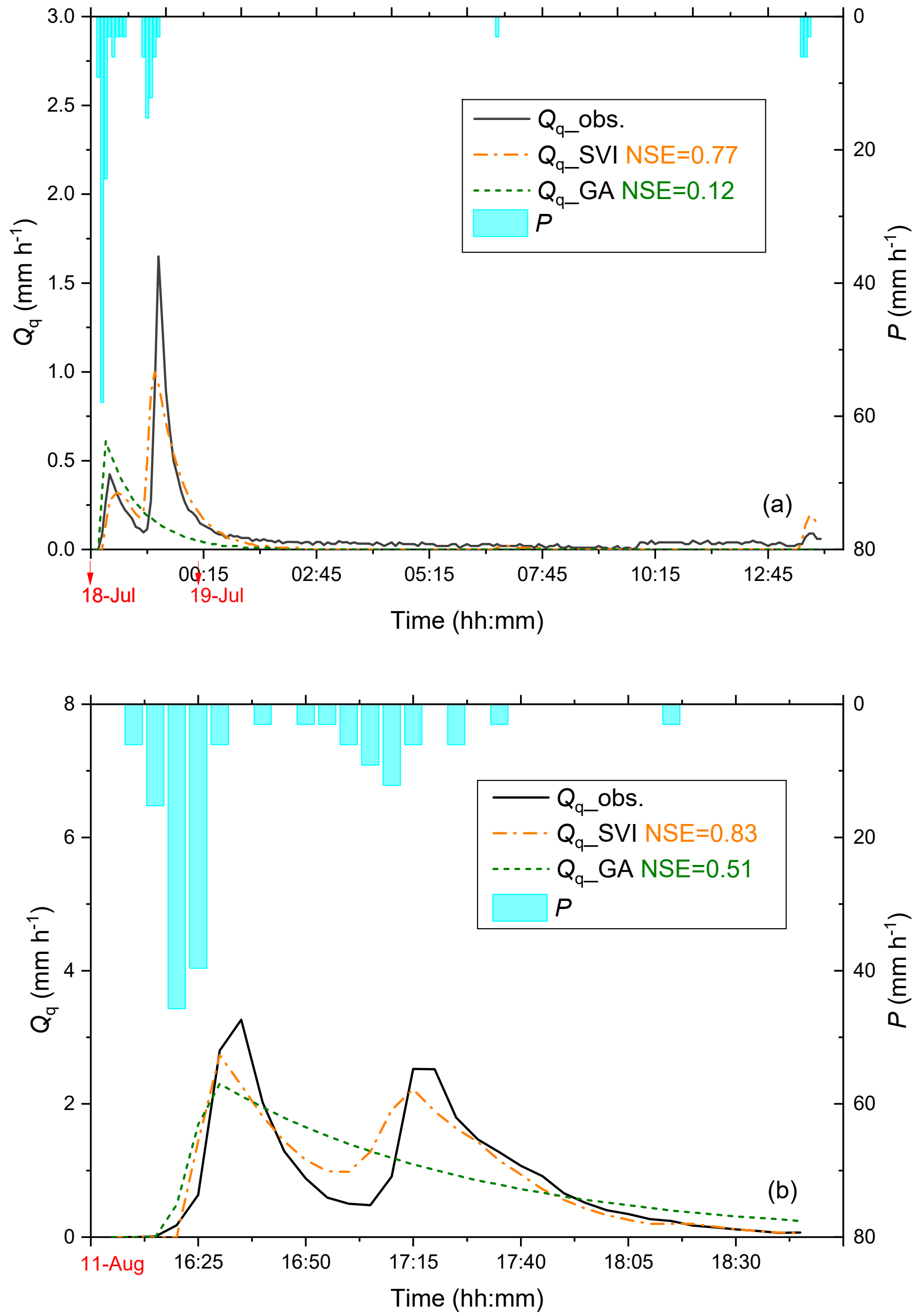

Figure 5. Observed and simulated stormflow hydrographs for two example events with two runoff peaks for which SVI outperformed GA due to the latter's failure to simulate the consecutive peaks: **(a)** the 14 mm event of 18–19 July 2013, with a stormflow runoff coefficient (R_c) of 10%, and **(b)** the 14 mm event of 11 August 2013, with R_c of 17%.

3.3 Infiltration model parameter variability

The optimized values for the three parameters for each infiltration model (F_0 and I_m for SVI; K_e and ψ_m for GA, plus lag time T in both models) are summarized in Table 2. Coefficients of variation (CV) were larger for I_m and K_e compared to the other parameters. The comparison of the ratio of the third and first quantiles (Q3/Q1) suggests that K_e and lag time T in GA varied more from event to event than I_m and T in SVI. Mean lag times for the two infiltration models did not differ significantly (p-value = 0.19).

Table 2. Variability of the optimized infiltration model parameters for the 30 examined events: F_0 = initial abstraction (mm), I_m = spatially averaged maximum infiltration capacity (mm h^{-1}), T = lag time (min), K_e = 'effective' final infiltration rate (mm h^{-1}), and ψ_m = matric potential at the wetting front (mm). Q1, Q2 and Q3 indicate the 1st, 2nd and 3rd quantiles of the respective parameter values. CV denotes the coefficient of variation and C_s the skewness.

Model	Model parameter	Q1	Q2	Q3	Mean	CV	Q3/Q1	C_s
SVI	F_0	5.1	7.0	9.4	7.9	0.5	1.8	1.2
	I_m	22.9	31.6	48.7	47.8	1.1	2.1	2.5
	T	8.6	14.0	19.9	16.9	0.7	2.3	1.7
GA	K_e	3.1	7.5	11.7	9.4	0.9	3.8	1.3
	ψ_m	21.9	24.6	27.4	27.8	0.7	1.2	1.4
	T	12.0	22.3	32.0	25.9	0.6	2.7	0.8

The infiltration-related parameters F_0, I_m and K_e (but not ψ_m) were all positively affected by rainfall intensity (regardless whether represented by the five-minute peak intensity P_p, or maximum intensities over 15 or 30 min, P_{15} or P_{30}), whereas the lag time for either infiltration model was inversely related to rainfall intensity (Table 3). Furthermore, both F_0 and I_m exhibited significant, negative correlations with SWC_{10} (Figure 6), but not with API_3. So did K_e to a lesser extent, but not ψ_m (Table 3).

Table 3. Spearman rank correlation coefficients between the infiltration model parameters and selected rainfall and catchment wetness characteristics: P = event precipitation (mm), P_a= average rainfall intensity (mm h^{-1}), P_p, P_{15} and P_{30} = maximum 5-min, 15-min and 30-min rainfall intensities (mm h^{-1} equivalents), API_3 = three-day antecedent precipitation index (mm), SWC_{10}, SWC_{30} and SWC_{60} = mid-slope soil water contents (%) down to 10 cm, 30 cm, and 60 cm depth, respectively.*** indicates p-value < 0.001, ** p-value < 0.05, * p-value < 0.1.

	P	P_a	P_p	P_{15}	P_{30}	API_3	SWC_{10}	SWC_{30}	SWC_{60}
F_0	0.23	0.34*	0.51***	0.52***	0.46**	-0.13	-0.59***	0.09	0.14
I_m	0.47***	0.22	0.63***	0.63***	0.63***	0.01	-0.57***	0.10	0.21
T_SVI	0.03	-0.34*	-0.38**	-0.48***	-0.35*	-0.27	0.11	-0.17	-0.05
K_e	0.53***	0.06	0.74***	0.72***	0.68***	-0.03	-0.48**	-0.16	-0.20
ψ_m	-0.25	0.288	-0.13	-0.11	-0.14	-0.06	-0.01	0.21	0.33
T_GA	-0.08	-0.23	-0.36**	-0.45**	-0.37**	-0.39**	0.05	-0.37*	-0.30

3.4 Stormflow prediction using SVI

Because the optimized values of the infiltration-related parameters in SVI (*i.e.*, F_0 and I_m) varied considerably between events (Table 2), predictions of individual stormflow hydrographs using average or median parameter values might not be very satisfying (see example events with wet and dry antecedent conditions in Supplementary Figure S3). However, both F_0 and I_m were clearly related to the near-surface wetness condition of the catchment as represented by the moisture content of the top 10 cm of the soil as measured in mid-slope position (though not by that down to 30 or 60 cm, nor by API_3; Table 3). Hence, the linear relationships between SWC_{10} and F_0 or I_m shown in Figure 6 were used to estimate the values of F_0 and I_m for each of the 26 events for which SWC_{10}-data were available. For each event, we used the median value of the lag time (T = 14.0 min).

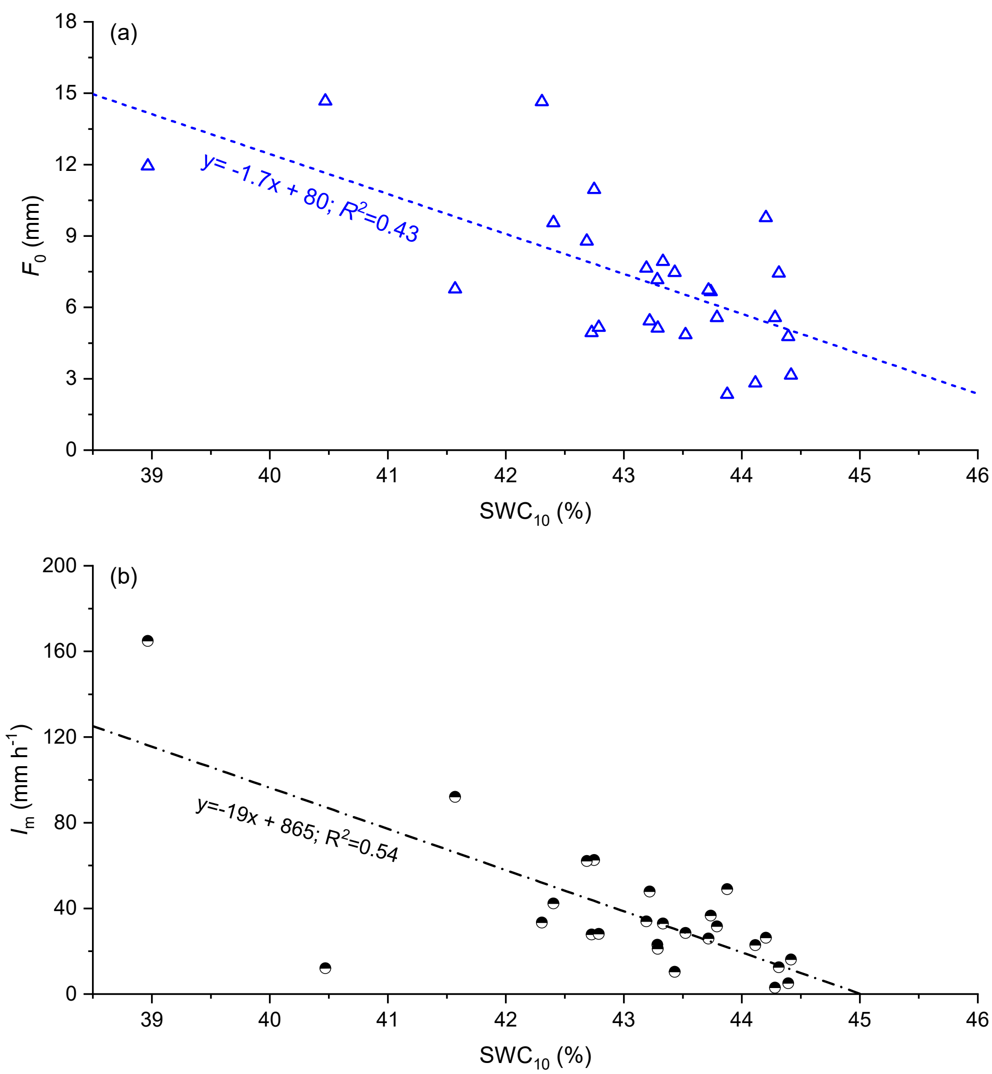

Figure 6. Linear relationships between mid-slope soil water content at 10 cm depth (SWC_{10}) and the optimized parameter values for **(a)** spatially average maximum infiltration rate, I_m, and **(b)** initial abstraction, F_0 for all 26 runoff events for which SWC_{10} data were available.

Satisfactory to good (NSE > 0.5) results were obtained for ~70% of the 26 events with group-based average stormflow runoff coefficients (R_c = Q_q/P) larger than ~0.15 (Figure 7). However, SVI was less successful at capturing the stormflow hydrographs of two events with contrasting runoff responses (NSE 0.18–0.28; Figure 7).

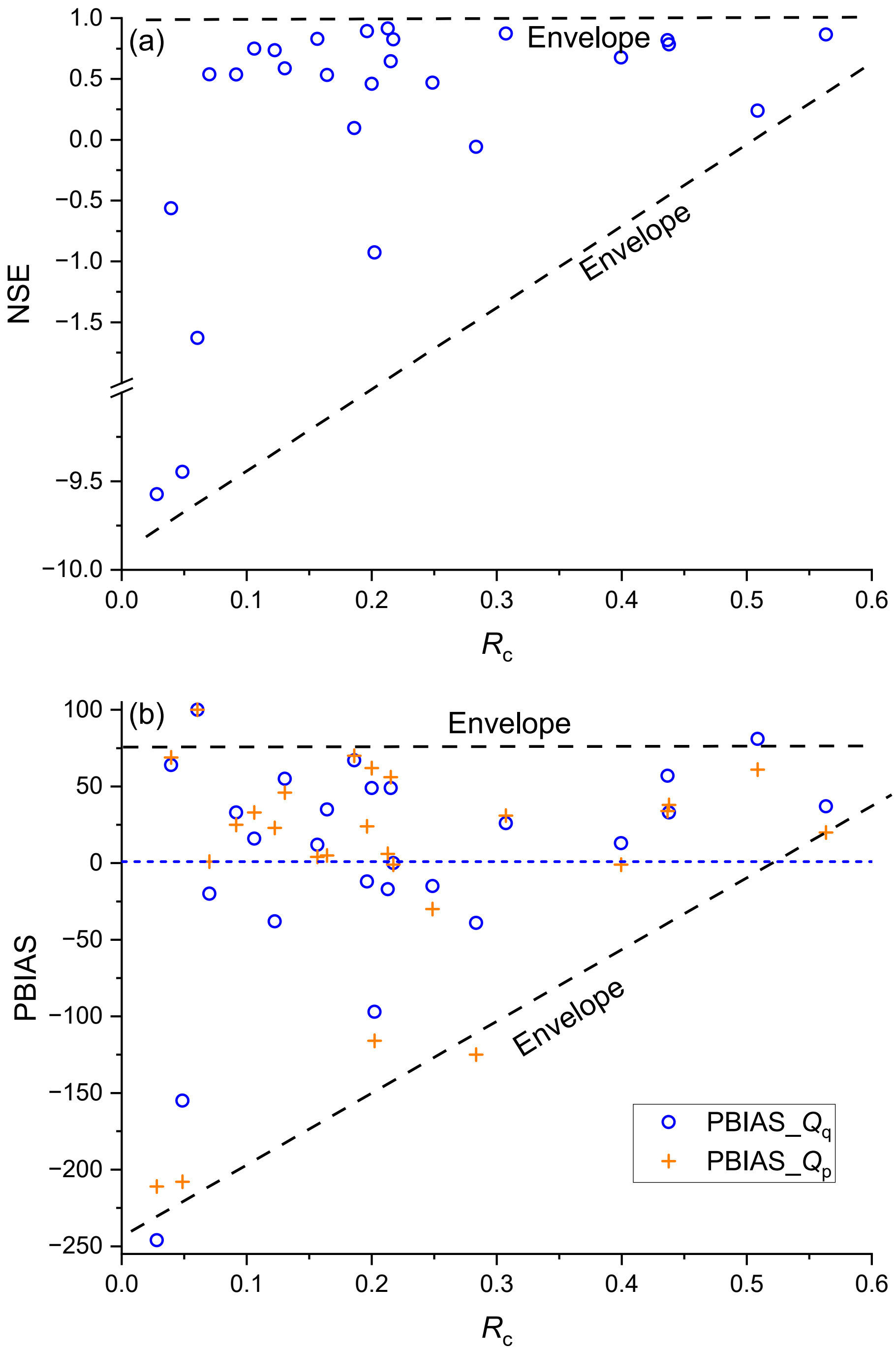

Figure 7. Relationship between stormflow runoff coefficient (R_c) and **(a)** the Nash–Sutcliffe model efficiency (NSE) when the SVI model parameters are based on the correlation with soil moisture at 10 cm (F_0 = -1.7·SWC$_{10}$+80; I_m = -19·SWC$_{10}$+865, according to Figure 6); and **(b)** PBIAS for stormflows (PBIAS_Q_q) and peak flow rates ((PBIAS_Q_p).

Negative NSE-values were obtained for another six events (23%) representing mostly (but not exclusively) low stormflow runoff coefficients (Figure 7). Two of these six events had comparatively low rainfall amounts (6.6–7.6 mm) and stormflow totals were severely under-estimated by the model. The remaining four events received more substantial amounts of rain (14–38 mm), but SVI over-estimated the amounts of stormflow considerably. Therefore, a comparison was made between calibrated and estimated values of F_0 and I_m (n = 26) for different classes of NSE and PBIAS; Supplementary Figure S4). Discrepancies between predicted and calibrated values of F_0 had a significant impact on the model performance (*i.e.*, lower NSE), whereas discrepancies in I_m had a smaller effect (Supplementary Figure S4a). Discrepancies in both F_0 and I_m had an important and significant effect on the simulated amount of stormflow (Supplementary Figure S4b). Higher values of F_0 and I_m led to underestimation of stormflow and *vice versa* (Supplementary Figure S4b).

3.5 Comparison of modeled infiltration parameters and measured K_{sat}

The average value for the highly skewed distribution of measured near-surface K_{sat} (22 ± 94 mm h^{-1} *versus* a median of 2 mm h^{-1}; skewness: 6.4) was roughly half the ~42 mm h^{-1} derived from the averaged exponential distribution of infiltration capacities for the 30 events (skewness: 2; Figure 8). In addition, the shape of the two distributions differed in that comparatively low infiltration capacities (< 20 mm h^{-1}) were encountered far more frequently during the field measurements than implied by the modeling, whereas the reverse applied for intermediate (20–50 mm h^{-1}) and higher infiltration capacities (50–500 mm h^{-1}; Figure 8).

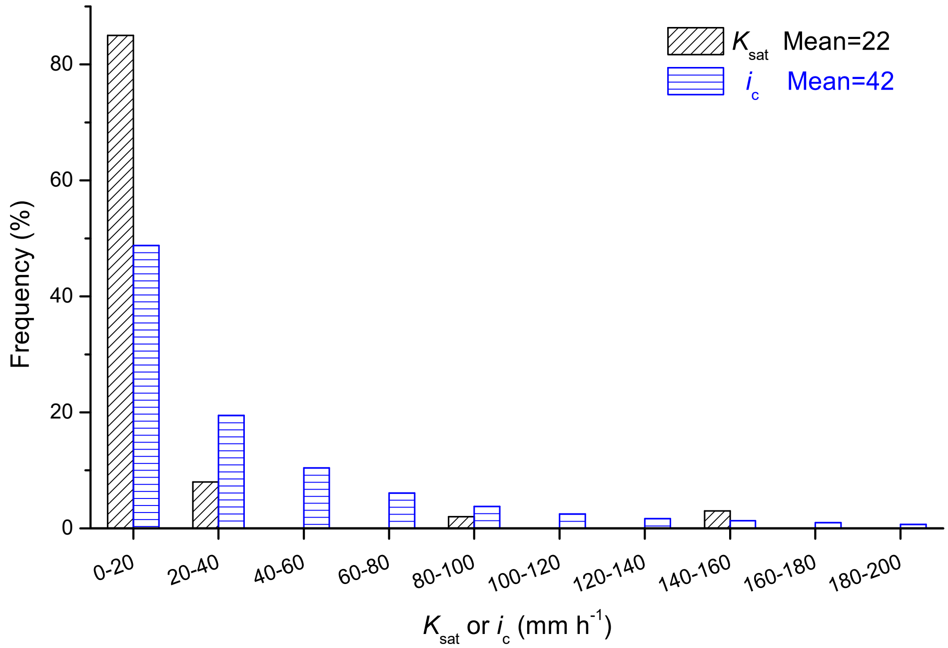

Figure 8. Comparison of the spatial distributions of measured K_{sat} (n = 60; data from Zhang et al., 2019a) and the modeled infiltration capacity (i_c) based on individual values of I_{m} for all 30 examined events.

4 Discussion

4.1 Infiltration model performance

With median NSE-values of 0.88 and 0.92, respectively, both GA and SVI performed well for the 30 examined events, with a few notable exceptions. In comparison to SVI, GA is inherently not responsive to changes in rainfall intensity, especially after infiltration reaches steady-state conditions (Yu, 1999). This is likely the main reason why GA was not able to reproduce events with consecutive peaks as well as SVI (Figure 5). Further, high-intensity rain falling on an initially dry soil causes the infiltration capacity to decrease rapidly to values approaching an ‘effective’ K_{e} (*cf.* Supplementary Figure S1). If subsequent rainfall intensities are less than K_{e}, this leads to the simulation of low stormflow rates (Figure 5a). Similarly, for events with a particularly high rainfall intensity at the beginning of the storm, the simulation led to large decreases in infiltration capacity within a short period of time, causing GA-modeled stormflow to occur earlier than observed (Supplementary Figure S2).

However, SVI did not perform perfectly for events with multiple bursts of rain either. The main reason for this discrepancy lies in the use of constant values for the model parameters for a given event. This assumption is likely to be violated during events with multiple rainfall peaks. An example of this occurred on 3–4 August 2013, when three successive bursts occurred within the event (Supplementary Figure S2c). The first burst occurred on 3 August between 15:20–17:45, the second on 4 August between 00:40–02:45, and the third between 03:20–12:35. The modest runoff peak for the second burst was greatly over-estimated, whereas the larger, third peak was substantially under-estimated (Supplementary Figure S2c). This is likely because the time gap between the first and second bursts (~7 h) was large enough to allow the soil to drain somewhat, thereby re-creating some additional storage opportunity. As a result, part of the rainfall of the second burst was used to fill this additional capacity, causing predicted stormflow rates to be over-estimated. For the third burst, which followed soon after (Supplementary Figure S2c), a lower value of I_m than the applied constant value would have been more appropriate to reflect the wetter soil conditions during this part of the event. Instead, applying a higher, constant I_m throughout the event led to under-estimated stormflow rates for the third burst. A similar under-estimation was also noted for the latter part of the event occurring the following day (Supplementary Figure S2d), where a lower I_m would again have given better results.

Both models performed fairly for several events with low stormflow runoff coefficients, even though they were calibrated for these events (*i.e.,* $Q_q/P < 0.10$; Supporting Figures S2e and S2f; *cf.* Figure 4). When applied in predictive mode with F_0 and I_m estimated from mid-slope SWC_{10} (Figure 6), SVI behaved less than satisfactorily for several other events with (mostly) low runoff coefficients (Figure 7).

4.2 Infiltration model parameters: variability and influences

Calibrated values for initial abstraction loss (F_0) and spatially averaged maximum infiltration capacity (I_m) in SVI, as well as for the 'catchment effective' infiltration capacity (K_e) in GA, varied substantially between the 30 examined events, with overall Q3/Q1-ratios of 1.8, 2.1, and 3.8, respectively (Table 2). As expected on the basis of general infiltration theory (Brutsaert, 2005), all three parameters were negatively correlated with topsoil moisture content (SWC_{10}), albeit not with moisture contents down to greater depths, nor with the three-day antecedent precipitation index (Table 3). Patin et al. (2012) did not find clear relationships between I_m and API per land cover for numerous 1-m^2 microplots under various land covers in Lao PDR either, but low values were derived at the height of the rainy season and maximum values late in the dry season. In addition, temporal variability in I_m of soils under young fallow vegetation after slash-and-burn cropping (as practiced in the past at Basper; Zhang et al., 2019a) was markedly greater than that for bare soil, upland rice or *Imperata* grassland. Patin et al. (2012) concluded that variations in the water use of (taller) vegetation types between rainfall events affected I_m through modification of soil water contents in more subtle ways than could be captured by a proxy like API with a stationary (*i.e.*, fixed) recession constant (*cf.* Eq. (9)) that does not capture variations in wetness conditions due to differences in evapotranspiration rates. As such, linking infiltration model parameter values to measured topsoil moisture contents is to be preferred (*cf.* Figure 6). In line with the findings of Patin et al. (2012), I_m at Basper also varied seasonally. Calibrated values less than 50 mm h^{-1} were obtained for events during the rainy June–August period, increasing to 75–175 mm h^{-1} during the drier September–November period (Supplementary Figure S5).

The presence of a well-developed vegetation cover affects the magnitude of I_m and F_0 also in other, indirect ways. Vegetation provides protection of the soil surface against rain drop impact, slaking and crust formation (Wiersum, 1985; Rose et al., 1997; Durán-Zuazo & Rodríguez-Pleguezelo, 2008; Miyata et al., 2009; Lacombe et al., 2018), and promotes soil faunal activity and macropore formation, thereby enhancing infiltration (Blanchart et al., 2004; Shougrakpam et al., 2010; Zwartendijk et al., 2017; Toohey et al., 2018). Indeed, the strongest correlation between I_m and any particular soil characteristic in the Laotian study by Patin et al. (2012) was that with the extent of surface crusting. Hence, comparative median values of I_m for different land-cover types effectively reflected their capacity to prevent crust formation (low for bare soil, high for fallows). Crusting was not studied explicitly at Basper, but the low infiltration capacities recorded by Zhang et al. (2019a) were attributed primarily to erosion during former slash-and-burn cropping phases that exposed the denser sub-soil to the impact of rain drops, as well as a general absence of soil biotic activity and macropores (Quiñones, 2014), and inherent limitations of the K_{sat} measurements (see also discussion below). Repeated cycles of slash-and-burn agriculture can effectively destroy the macropore systems formed during fallow periods (Shougrakpam et al., 2010; Zwartendijk et al., 2017). Pertinently, soil moisture contents at 60 cm depth in the Basper grassland hardly responded to fluctuations in rainfall (Zhang et al., 2018a). Conversely, soil moisture at the same depth beneath a nearby forest responded rapidly to rainfall (Zhang et al., 2018b), suggesting the presence of preferential flow pathways that allowed rapid percolation to deeper layers (Van Meerveld et al., 2019; Zhang et al., 2019a; *cf.* Y. Cheng et al., 2018).

In line with the trend noted above for I_m, F_0 can also be expected to be higher for well-vegetated or mulched surfaces than for bare soils (Yu et al., 1997b; Van Dijk & Bruijnzeel, 2004). The limited data available for tropical sites do not suggest that soil texture has a notable influence on the magnitude of F_0 or I_m (in contrast to findings for K_e by Nearing et al., 1996). Increases in soil organic matter content (SOM) tend to have a positive effect, whereas increases in bulk density tend to have a negative effect (Coughlan, 1997; Yu et al., 1997b; Van Dijk & Bruijnzeel, 2004). However, with the possible exception of the relationship between I_m and bulk density ($R^2 = 0.923$, $n = 7$), the predictive capacity of such tentative equations is still low (Supplementary Figure S6) and many more empirical data are required.

The currently derived median F_0 (7.6 mm, Table 2) exceeded most of the values reported by Yu et al. (1997b) for various bare agricultural plots in Southeast Asia and Queensland (2.3–6.0 mm), which generally had higher bulk densities and lower SOM than the Basper grass- and shrubland (Coughlan, 1997; Zhang et al., 2019a; Supplementary Figure S6). Higher values of F_0 were obtained at the same sites after application of a surface mulch (~13 mm; Yu et al., 1997b). As such, the interception storage capacity afforded by the tall grasses and shrubs at Basper (and their litter) may well have raised the effective value of F_0 somewhat (*cf.* Leopoldo et al., 1981; Waterloo et al., 1999; Bruijnzeel, 1988). In addition, it cannot be excluded that variations in rainfall intensity at Basper further affected the magnitude of F_0 indirectly through variations in wet canopy evaporation rates between successive storms as observed in a nearby forest by Zhang et al. (2018b). This would not only go some way towards explaining the positive correlations between F_0 and short-term rainfall intensities (P_{15} and P_{30}; Table 3), but possibly also the discrepancies between SWC_{10}-based estimates of F_0 and calibrated values for certain poorly predicted events (negative NSE; Supplementary Figure S4a).

4.3 Difficulty of estimating effective hydraulic conductivity and infiltration capacity from point measurements

The median values of the model-based estimates of catchment-wide 'effective' (K_e) and 'maximum' (I_m) infiltration (7.5 mm h^{-1} for GA and 31.6 mm h^{-1} for SVI) were distinctly higher than the field-based measurements of K_{sat} (1.7–2.7 mm h^{-1}, depending on the method used; Zhang et al., 2019a). Also, the SVI-inferred *distribution* of infiltration capacities suggested generally higher values compared to the results obtained by the measurements (Figure 8). However, the measured values were also much lower than the median value reported for similarly textured, non-grazed *Imperata* grassland soils elsewhere in the Palaeo-tropics (35 mm h^{-1}, n = 8; range: 15–95 mm h^{-1}; Zhang et al., 2019a; Ghimire et al., 2021). The methods used for measuring near-surface K_{sat} at Basper may have under-estimated actual hydraulic conductivities to some extent – either because of under-sampling of macropores in the case of small cores and small-diameter ring infiltrometry (Davis et al., 1996; Lai &

Ren, 2007) or due to smearing of boreholes during augering in the case of well permeametry (Sherlock et al., 2000; Bonell et al., 2010). In addition, it cannot be excluded that somewhat higher values of K_{sat} may have been associated with the denser (less penetrable) parts of the regenerating vegetation in the central part of the catchment, where only a few K_{sat} measurements were conducted (Figure 1a). As such, overall mean catchment-wide K_{sat} may also be higher than inferred from the measurements by Zhang et al. (2019a) due to the spatial bias in field sampling. Furthermore, point-measured K_{sat}-values typically under-estimate the 'block permeabilities' of whole hillslopes (Wen & Gomez-Hernandez, 1996; Chappell et al., 1998; Brooks et al., 2004; Pirastru et al., 2017). This under-estimation of block permeability is also seen where statistical distributions of point-measured K_{sat} values are compared directly with 'effective' parameter values derived from inversion of catchment models (*e.g.*, Beven, 1989; Blöschl & Sivapalan, 1995; Mertens et al., 2005).

Both I_{m} and K_{e} are commonly applied to characterize soil infiltration capacity (Yu, 2000; Nearing et al., 1996). The relationship between the two is of interest because it allows derivation of modeled I_{m} (the spatially averaged *maximum* infiltration capacity) from K_{e} (the 'effective' infiltration rate after reaching steady-state conditions; *cf.* Supporting Figure S1) obtained by inverse means from either IOF (plots) or stormflow (catchments) measurements and GA (*e.g.*, Nearing et al., 1996). In agreement with these definitions, derived values for I_{m} at Basper (3–259 mm h^{-1}) were higher than those for K_{e} (1–31 mm h^{-1}). As also reported by Yu (1999) for six different locations in Australia and Southeast Asia, I_{m} at Basper was positively correlated with K_{e}. As shown in Supporting Figure S7, the second-order polynomial describing the relation between K_{e} and I_{m} for the Basper grassland had an R^2 of 0.45 (n = 30) compared to R^2 = 0.80 (n = 60) for the equation derived by Yu (1999). Additional empirical data for different tropical locations are desirable to complement these tentative equations.

5 Conclusions

Five-minute rainfall and runoff data collected during 30 events (6.6–149 mm of rain) were used to calibrate two infiltration models of different complexity for the prediction of stormflow responses for a 3.2 ha fire-climax grassland catchment at Basper, Leyte Island (the Philippines). The catchment has soils with very low hydraulic conductivity (K_{sat}) and infiltration-excess overland flow is inferred to be the dominant storm runoff generation mechanism. Landslide scars with low-infiltrability slip surfaces are prominent, covering 3.4% of the area.

In the Green–Ampt model (GA), the infiltration rates decline steadily after the start of infiltration, whereas the Spatially Variable Infiltration model (SVI) describes infiltration as a function of short-term fluctuations in rainfall intensity. SVI systematically reproduced the observed stormflow hydrographs better than GA, especially for events with multiple peaks. Calibrated values of the parameters for SVI (notably, spatially averaged maximum infiltration capacity, I_m and initial abstraction, F_0) varied markedly between events, and showed significant negative linear correlations with mid-slope topsoil water content (SWC_{10}) – as did the 'effective' hydraulic conductivity (K_e) in GA. Using SWC_{10}-based values of I_m and F_0 in SVI produced satisfactory to good (NSE > 0.5) predictive results for ~70% of the examined storms, but failed to reproduce hydrographs for six events (23%) with variable runoff responses, possibly because F_0 was also affected by variations in rainfall interception losses between storms. Deviations between calibrated and SWC_{10}-predicted values of F_0 had a greater impact on predicted stormflow amounts than corresponding deviations in I_m.

The median I_m and, to a lesser extent K_e, inferred for the 30 examined events (31.6 and 7.5 mm h^{-1}, respectively) were much higher than the median values of near-surface K_{sat} measurements (2–3 mm h^{-1}, depending on method), confirming the previously suspected under-estimation of field-measured K_{sat} in the study catchment.

Summarizing, using pre-storm topsoil moisture content and 5-min rainfall intensities as the driving variables to model infiltration with a spatially variable infiltration model resulted in more realistic simulated stormflow responses than the classic

Green–Ampt approach or the comparison of rainfall intensities with field-measured K_{sat} to predict stormflow responses at the small catchment scale.

Acknowledgements

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Open Research

The data used for visualization of all figures, the model input data for the 30 examined storm events, and the Python codes employed in the infiltration modeling using GA- and SVI can be accessed via HydroShare: Cheng, Z., J. Zhang (2022). Data_resource_of_figures; Model code_and_input, HydroShare, http://www.hydroshare.org/resource/6a63073f0361493f81e4e48c93fae299

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*Note:*references from the supporting information have been listed in the main reference list.*