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Flood risk reduction and flow buffering as ecosystem services – Part 1: Theory on flow persistence, flashiness and base flow

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Abstract. Flood damage reflects insufficient adaptation of human presence and activity to location and variability of river flow in a given climate. Flood risk increases when landscapes degrade, counteracted or aggravated by engineering solutions. Efforts to maintain and restore buffering as an ecosystem function may help adaptation to climate change, but this require quantification of effectiveness in their specific social-ecological context. However, the specific role of forests, trees, soil and drainage pathways in flow buffering, given geology, land form and climate, remains controversial. When complementing the scarce heavily instrumented catchments with reliable long-term data, especially in the tropics, there is a need for metrics for data-sparse conditions. We present and discuss a flow persistence metric that relates transmission to river flow of peak rainfall events to the base-flow component of the water balance. The dimensionless flow persistence parameter F_p is defined in a recursive flow model and can be estimated from limited time series of observed daily flow, without requiring knowledge of spatially distributed rainfall upstream. The F_p metric (or its change over time from what appears to be the local norm) matches local knowledge concepts. Inter-annual variation in the F_p metric in sample watersheds correlates with variation in the "flashiness index" used in existing watershed health monitoring programmes, but the relationship between these metrics varies with context. Inter-annual variation in F_p also correlates with common base-flow indicators, but again in a way that varies between watersheds. Further exploration of the responsiveness of F_p in watersheds with different characteristics to the interaction of land cover and the specific realisation of space–time patterns of rainfall in a limited observation period is needed to evaluate interpretation of F_p as an indicator of anthropogenic changes in watershed conditions.

1 Introduction

Floods can be the direct result of reservoir dams, log jams or protective dykes breaking, with water derived from unexpected heavy rainfall, rapid snowmelt, tsunamis or coastal storm surges. We focus here on floods that are associated, at least in the public eye, with watershed degradation. Degradation of watersheds and its consequences for river-flow regime and flooding intensity and frequency are a widespread concern (Brauman et al., 2007; Bishop and Pagiola, 2012; Winsemius et al., 2013). Engineering measures (dams, reservoirs, canalisation, dykes, and flow regulation) can significantly alter the flow regime of rivers, and reduce the direct relationship with landscape conditions in the (upper) catchment (Poff et al., 1997). The life expectancy of such structures depends, however, on the sediment load of incoming rivers and thus on upper watershed conditions (Graf et al., 2010). Where "flow regulation" has been included in efforts to assess an economic value of ecosystem services, it can emerge as a major component of overall value; the economic damage of floods to cities build on floodplains can be huge and the benefits of avoiding disasters thus large (Farber et al., 2002; Turner and Daily, 2008; Brauman et al., 2007). The "counter factual" part of any avoided damage argument, however, depends on metrics that are transparent in their basic concept and relationship with observables. Basic requirements for a metric to be used in managing issues of public concern in a complex multi-stakeholder environment are that it (i) has

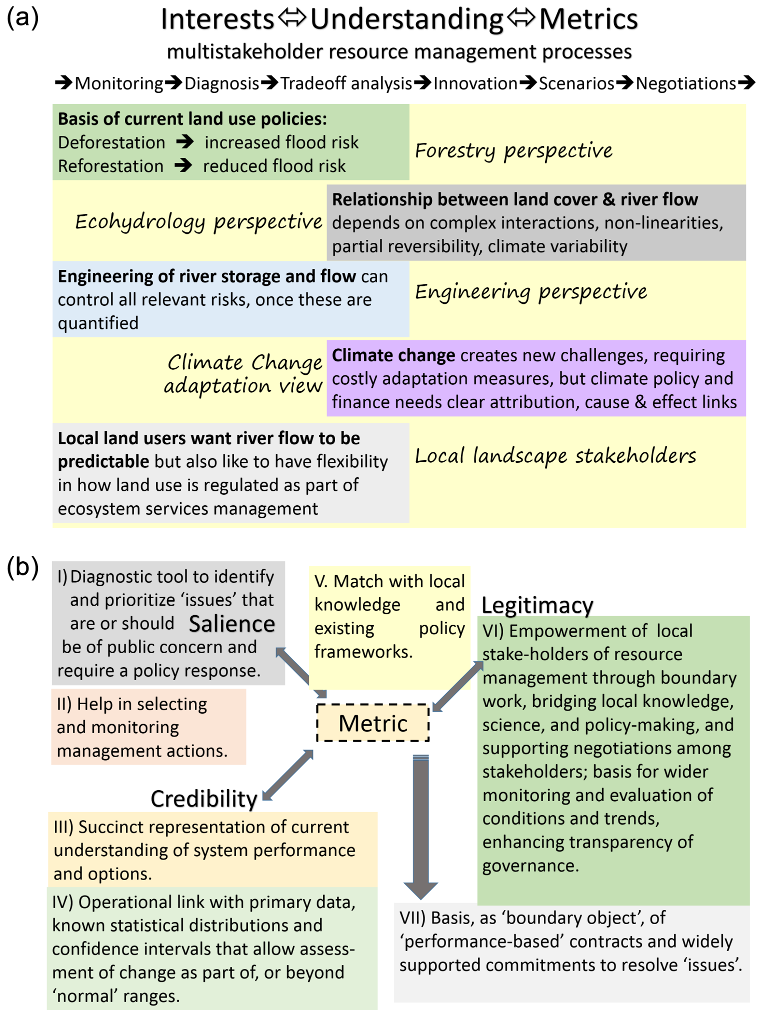

Figure 1. (a) Multiple perspectives on the way flood risk is to be understood, monitored and handled according to different knowledge systems; **(b)** basic requirements for a "metric" to be used in public discussions of natural resource management issues that deserve to be resolved and acted upon (modified from van Noordwijk et al., 2016).

a direct relationship with a problem that needs to be solved ("salience"), (ii) is aligned with current science-based understanding of how the underpinning systems function and can be managed ("credibility") and (iii) can be understood from local and public/policy perspectives ("legitimacy") (Clark et al., 2016). Figure 1 summarises these requirements, building on van Noordwijk et al. (2016).

In the popular discussion on floods, especially in the tropics, a direct relationship with deforestation and reforestation is still commonly perceived to dominate, and forest cover is seen as salient and legitimate metric of watershed quality (or of urgency of restoration where it is low). A requirement for 30 % forest cover, is for example included in the spatial planning law in Indonesia in this context (Galudra and Sirait, 2009). Yet, rivers are probably dominated by the other 70 % of the landscape. There is a problem with the credibility of assumed deforestation–flood relations (van Noordwijk et al., 2007; Verbist et al., 2010), beyond the local scales ($< 10\,\mathrm{km}^2$) of paired catchments where ample direct empirical proof exists, especially in non-tropical climate zones (Bruijnzeel, 1990, 2004). Current watershed rehabilitation programmes that focus on increasing tree cover in upper watersheds are only partly aligned with current scientific evidence of effects of large-scale tree planting on streamflow (Ghimire et al., 2014; Malmer et al., 2010; Palmer, 2009; van Noordwijk et al., 2015a). The relationship between floods and change in forest quality and quantity, and the availability of evidence for such a relationship at various scales has been widely discussed over the past decades (Andréassian, 2004; Bruijnzeel, 2004; Bradshaw et al., 2007; van Dijk et al., 2009). Measurements in Cote d'Ivoire, for example, showed strong scale dependence of runoff from 30 to 50 % of rainfall at $1\,\mathrm{m}^2$ point scale, to 4 % at 130 ha watershed scale, linked to spatial variability of soil properties plus variations in rainfall patterns (Van de Giesen et al., 2000). The ratio between peak and average flow decreases from headwater streams to main rivers in a predictable manner; while mean annual discharge scales with (area)$^{1.0}$, maximum river flow was found to scale with (area)$^{0.4}$ to (area)$^{0.7}$ on average (Rodríguez-Iturbe and Rinaldo, 2001; van Noordwijk et al., 1998; Herschy, 2002), with even lower powers for area in flash floods that are linked to an extreme rainfall event over a restricted area (Marchi et al., 2010). The determinants of peak flow are thus scale dependent, with space–time correlations in rainfall interacting with subcatchment-level flow buffering at any point along the river. Whether and where peak flows lead to flooding depends on the capacity of the rivers to pass on peak flows towards downstream lakes or the sea, assisted by riparian buffer areas with sufficient storage capacity (Di Baldassarre et al., 2013). Reducing local flooding risk by increased drainage increases flooding risk downstream, challenging the nested-scales management of watersheds to find an optimal spatial distribution, rather than minimisation, of flooding probabilities. Well-studied effects of forest conversion on peak flows in small upper stream catchments (Bruijnzeel, 2004; Alila et al., 2009) do not necessarily translate to flooding downstream. With most of the published studies still referring to the temperate zone, the situation in the tropics (generally in the absence of snow) is contested (Bonell and Bruijnzeel, 2005). As summarised by Beck et al. (2013), meso- to macroscale catchment studies (> 1 and $> 10\,000\,\mathrm{km}^2$, respectively) in the tropics, subtropics and warm temperate regions have mostly failed to demonstrate a clear relationship between river flow and change in forest area. Lack of evidence cannot be firmly interpreted as evidence for lack of effect, however. Detectability of effects depends on their relative size, the accuracy of the measurement devices, length of observation period, and background variability of the signal. A recent econometric study for Peninsular Malaysia by Tan-Soo et al. (2016) concluded that, after appropriate corrections for space–time correlates in the data set for 31 meso- and macroscale basins (554–28 643 km^2), conversion of inland rain forest to monocultural plantations of oil palm or rubber increased the number of flooding days reported, but not the number of flood events, while conversion of wetland forests to urban areas reduced downstream flood duration. This Malaysian study may be the first credible empirical evidence at this scale. The difference between results for flood duration and flood frequency and the result for draining wetland forests warrant further scrutiny. Consistency of these findings with river-flow models based on a water balance and likely pathways of water under the influence of change in land cover and land use has yet to be shown. Two recent studies for southern China confirm the conventional perspective that deforestation increases high flows, but are contrasting in effects of reforestation. Zhou et al. (2010) analysed a 50-year data set for Guangdong Province in China and concluded that forest recovery had not changed the annual water yield (or its underpinning water balance terms precipitation and evapotranspiration), but had a statistically significant positive effect on dry-season (low) flows. Liu et al. (2015), however, found for the Meijiang watershed (6983 km^2) in subtropical China that while historical deforestation had decreased the magnitudes of low flows (daily flows $\leq Q_{95\,\%}$) by 30.1 %, low flows were not significantly improved by reforestation. They concluded that recovery of low flows by reforestation may take a much longer time than expected probably because of severe soil erosion and resultant loss of soil infiltration capacity after deforestation. Changes in river-flow patterns over a limited period of time can be the combined and interactive effects of variations in the local rainfall regime, land cover effects on soil structure and engineering modifications of water flow that can be teased apart with modelling tools (Ma et al., 2014).

Lacombe et al. (2016) documented that the hydrological effects of natural regeneration differ from those of plantation forestry, while forest statistics do not normally differentiate between these different land covers. In a regression study of the high- and low-flow regimes in the Volta and Mekong river basins, Lacombe and McCartney (2016) found that in the variation among tributaries various aspects of land cover and land cover change had explanatory power. Between the two basins, however, these aspects differed. In the Mekong basin, variation in forest cover had no direct effect on flows, but extending paddy areas resulted in a decrease in downstream low flows, probably by increasing evapotranspiration in the dry season. In the Volta River basin, the conversion of forests to crops (or a reduction of tree cover in the existing parkland system) induced greater downstream flood flows. This observation is aligned with the experimental identification of an optimal, intermediate tree cover from the perspective of groundwater recharge in parklands in Burkina Faso (Ilstedt et al., 2016).

The statistical challenges of attribution of cause and effect in such data sets are considerable with land use/land cover effects interacting with spatially and temporally variable rainfall, geological configuration and the fact that land use is not changing in random fashion or following any pre-randomised design (Alila et al., 2009; Rudel et al., 2005). Hydrological analysis across 12 catchments in Puerto Rico by Beck et al. (2013) did not find significant relationships between the change in forest cover or urban area, and change in various flow characteristics, despite indications that regrowing forests increased evapotranspiration.

These observations imply that the percent of tree cover (or other forest-related indicators) is probably not a good metric for judging the ecosystem services provided by a watershed (of different levels of "health"), and that a metric more directly reflecting changes in river flow may be needed. Here we will explore a simple recursive model of river flow (van Noordwijk et al., 2011) that (i) is focused on (loss of) flow predictability, (ii) can account for the types of results obtained by the cited recent Malaysian study (Tan-Soo et al.,

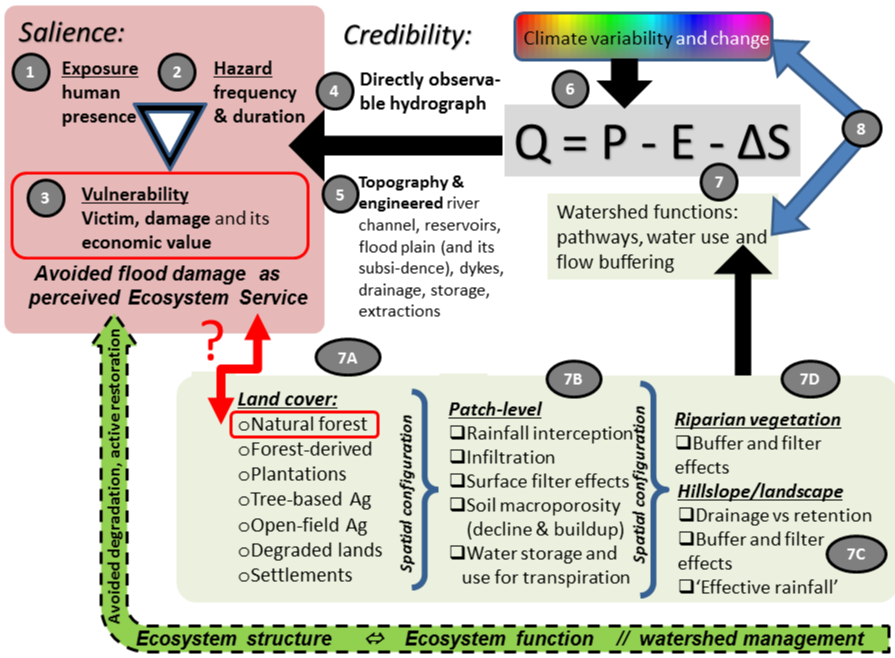

Figure 2. Steps in a causal pathway that relates the salience of "avoided flood damage as ecosystem service" to the interaction of exposure (1; being in the wrong place at critical times), hazard (2; spatially explicit flood frequency and duration) and human determinants of vulnerability (3); the hazard component depends, in common scientific analysis, on the pattern of river flow described in a hydrograph (4), which in turn is understood to be influenced by conditions along the river channel (5), precipitation and potential evapotranspiration (E_{pot}) as climatic factors (6) and the condition in the watershed (7) determining evapotranspiration (E_{act}), temporary water storage (ΔS) and water partitioning over overland flow and infiltration; these watershed functions in turn depend on the interaction of terrain (topography, soils, geology), vegetation and human land use; current understanding of a two-way interaction between vegetation and rainfall adds further complexity (8).

2016), and (iii) may constitute a suitable performance indicator to monitor watershed health through time.

Before discussing the credibility dimension of river-flow metrics, the way these relate to the salience and legitimacy issues around "flood damage" as policy issues needs attention. The salient issue of "flood damage" is compatible with a common dissection of risk as the product of exposure, hazard and vulnerability (steps 1–3 in Fig. 2). Many aspects beyond forests and tree cover play a role; in fact these factors are multiple steps away (step 7a) from the direct river-flow dynamics that determine floods. Extreme discharge events plus river-level engineering (steps 4 and 5) co-determine hazard (step 2), while exposure (step 1) depends on topographic position interacting with human presence, and vulnerability can be modified by engineering at a finer scale and be further reduced by advice to leave an area in high-risk periods. A recent study (Jongman et al., 2015) found that human fatalities and material losses between 1980 and 2010 expressed as a share of the exposed population and gross domestic product were decreasing with rising income. The planning needed to avoid extensive damage requires quantification of the risk of higher than usual discharges, especially at the upper tail end of the flow frequency distribution.

The statistical scarcity, per definition, of "extreme events" and the challenge of data collection where they do occur, make it hard to rely on site-specific empirical data as such. Inference of risks needs some trust in extrapolation methods, as is often provided by use of trusted underlying mechanisms and/or data obtained in a geographical proximity. Existing data on flood frequency and duration, as well as human and economic damage are influenced by topography, soils, human population density and economic activity, responding to engineered infrastructure (step 5 in Fig. 2), as well as the extreme rainfall events that are their proximate cause (step 6). Subsidence due to groundwater extraction in urban areas of high population density is a specific problem for a number of cities built on floodplains (such as Jakarta and Bangkok), but subsidence of drained peat areas has also been found to increase flooding risks elsewhere (Sumarga et al., 2016). Common hydrological analysis of flood frequency (called 1 in 10-, 1 in 100- and 1 in 1000-year flood events, for example) relies on direct observations at step 4 in Fig. 2, but typically requires spatial extrapolation beyond points of data collection through river-flow models that combine at least steps 5 and 6. Relatively simple ways of including the conditions in the watershed (step 7) in such models rely on the runoff curve number method (Ponce and Hawkins, 1996) and the SWAT (soil water assessment tool) model that was built on its foundation (Gassman et al., 2007). Applications on tropical soils have had mixed success (Oliveira et al., 2016). Describing peak flows as a proportion of the rainfall event that triggered them has a long history, but where the proportionality factors are estimated for ungauged catchments results may be unreliable (Efstratiadis et al., 2014). More refined descriptions of the infiltration process (step 7b) are available, using recursive models as filters on empirical data (Grimaldi et al., 2013), but data for this approach may not be generally available. According to Van den Putte et al. (2013) the Green–Ampt infiltration equation can be fitted to data for dry conditions when soil crusts limit infiltration, but not in wet winter conditions. These authors argued that simpler models may be better.

Analysis of likely change in flood frequencies in the context of climate change adaptation has been challenging (Milly et al., 2002; Ma et al., 2014). There is a lack of simple performance indicators for watershed health at its point of relating precipitation P and river flow Q (step 4 in Fig. 2) that align with local observations of river behaviour and concerns about its change and that can reconcile local, public/policy and scientific knowledge, thereby helping negotiated change in watershed management (Leimona et al., 2015). The behaviour of rivers depends on many climatic (step 6 in Fig. 2) and terrain factors (step 7a–d in Fig. 2) that make it a challenge to differentiate between human-induced ecosystem structural change and soil degradation (step 7b) on the one hand and intrinsic variability on the other. Step 8 in Fig. 2 represents not only the direct influence of climate on vegetation but also a possible reverse influence (van Noordwijk et al., 2015b). Hydrological models tend to focus on predicting hydrographs at one or more temporal scales, and are usually tested on data sets from limited locations. Despite many decades (if not centuries) of hydrological modelling, current hydrologic theory, models and empirical methods have been found to be largely inadequate for sound predictions in ungauged basins (Hrachowitz et al., 2013). Efforts to resolve this through harmonisation of modelling strategies have so far failed. Existing models differ in the number of explanatory variables and parameters they use, but are generally dependent on empirical data of rainfall that are available for specific measurement points but not at the spatial resolution that is required for a close match between measured and modelled river flow. Spatially explicit models have conceptual appeal (Ma et al., 2010) but have too many degrees of freedom and too many opportunities for getting right answers for wrong reasons if used for empirical calibration (Beven, 2011). Parsimonious, parameter-sparse models are appropriate for the level of evidence available to constrain them, but these parameters are themselves implicitly influenced by many aspects of existing and changing features of the watershed, making it hard to use such models for scenario studies of changing land use and change in climate forcing. Here we present a more direct approach deriving a metric of flow predictability that can bridge local concerns and concepts to quantified hydrologic function: the "flow persistence" parameter as directly observable characteristic (step 4 in Fig. 2), which can be logically linked to the primary points of intervention in watershed management, interacting with climate and engineering-based change.

In this contribution to the debate, we will first define the metric "flow persistence" in the context of temporal autocorrelation of river flow and then derive a way to estimate its numerical value. In Part 2 (van Noordwijk et al., 2017) we will apply the algorithm to river-flow data for a number of contrasting meso-scale watersheds. In the discussion of this paper, we will consider the new flow persistence metric in terms of three groups of criteria for usable knowledge (Fig. 1, Clark et al., 2016; Lusiana et al., 2011; Leimona et al., 2015) based on salience (i, ii), credibility (ii, iv) and legitimacy (v–vii):

i. Does flow persistence relate to important aspects of watershed behaviour, complementing existing metrics such as the "flashiness index" and "base-flow separation" techniques?

ii. Does its quantification help to select management actions?

iii. Is there consistency of numerical results?

iv. How sensitive is it to bias and random error in data sources?

v. Does it match local knowledge?

vi. Can it be used to empower local stakeholders of watershed management?

vii. Can it inform local risk management?

2 Flow persistence in water balance equations

2.1 Recursive model

One of the easiest-to-observe aspects of a river is its day-to-day fluctuation in water level, related to the volumetric flow (discharge) via rating curves (Maidment, 1992). Without knowing details of upstream rainfall and the pathways the rain takes to reach the river, observation of the daily fluctuations in water level allows for important inferences to be made. It is also of direct utility; sudden rises can lead to floods without sufficient warning, while rapid decline makes water utilisation difficult. Indeed, a common local description of watershed degradation is that rivers become more "flashy" and less predictable, having lost a buffer or "sponge" effect (Joshi et al., 2004; Ranieri et al., 2004; Rahayu et al., 2013). A simple model of river flow at time t, Q_t, is that it is similar to that of the day before (Q_{t-1}), multiplied with F_p, a dimensionless parameter called "flow persistence" (van Noordwijk et al., 2011), plus an additional stochastic term $Q_{a,t}$:

$$Q_t = F_p Q_{t-1} + Q_{a,t}. \tag{1}$$

Q_t is for this analysis expressed in mm day^{-1}, which means that measurements in m^3 s^{-1} need to be divided by the relevant catchment area, with appropriate unit conversion. If river flow were constant, it would be perfectly predictable, i.e. F_p would be 1.0 and $Q_{\mathrm{a},t}$ zero; in contrast, an F_p value equal to zero and $Q_{\mathrm{a},t}$ directly reflecting erratic rainfall represents the lowest possible level of predictability.

The F_p parameter is conceptually identical to the "recession constant" commonly used in hydrological models, typically assessed during an extended dry period when the $Q_{\mathrm{a},t}$ term is negligible and streamflow consists of base flow only (Tallaksen, 1995); empirical deviations from a straight line in a plot of the logarithm of Q against time are common and point to multiple rather than a single groundwater pool that contributes to base flow. The larger catchment area has a possibility to get additional flow from multiple independent groundwater contributions.

As we will demonstrate in a next section, it is possible to derive F_p even when $Q_{\mathrm{a},t}$ is not negligible. In climates without a distinct dry season this is essential; elsewhere it allows for a comparison of apparent F_p between wet and dry parts of the hydrologic year. A possible interpretation, to be further explored, is that decrease over the years of F_p indicates "watershed degradation" (i.e. greater contrast between high and low flows), and an increase "improvement" or "rehabilitation" (i.e. more stable flows).

If we consider the sum of river flow over a period of time (from 1 to T) we obtain

$$\sum_1^T Q_t = \sum_1^T Q_{t-1} + \sum_1^T Q_{\mathrm{a},t}. \tag{2}$$

If the period is a sufficiently long period for Q_T minus Q_0 (the values of Q_t for $t = T$ and $t = 0$, respectively) to be negligibly small relative to the sum over all t's, we may equate $\sum_1^T Q_t$ with $\sum_1^T Q_{t-1}$ and obtain the first way of estimating the F_p value:

$$F_\mathrm{p} = 1 - \sum_1^T Q_{\mathrm{a},t} / \sum_1^T Q_t. \tag{3}$$

The stochastic $Q_{\mathrm{a},t}$ can be interpreted in terms of what hydrologists call "effective rainfall" (i.e. rainfall minus on-site evapotranspiration, assessed over a preceding time period tx since previous rain event):

$$Q_t = F_\mathrm{p} Q_{t-1} + \left(1 - F_\mathrm{p}\right)\left(P_{tx} - E_{tx}\right), \tag{4}$$

where P_{tx} is the (spatially weighted) precipitation on day t (or preceding precipitation released as snowmelt on day t) in mm day^{-1}; E_{tx}, also in mm day^{-1}, is the preceding evapotranspiration that allowed for infiltration during this rainfall event (i.e. evapotranspiration since the previous soil-replenishing rainfall that induced empty pore space in the soil for infiltration and retention), or replenishment of a water film on aboveground biomass that will subsequently evaporate. More complex attributions are possible, aligning with the groundwater-replenishing bypass flow and the water isotopic fractionation involved in evaporation (Evaristo et al., 2015).

The consistency of multiplying effective rainfall with $(1 - F_\mathrm{p})$ can be checked by considering the geometric series $(1 - F_\mathrm{p})$, $(1 - F_\mathrm{p})\ F_\mathrm{p}$, $(1 - F_\mathrm{p})\ F_\mathrm{p}^2$, …, $(1 - F_\mathrm{p})\ F_\mathrm{p}^n$, which adds up to $(1 - F_\mathrm{p})(1 - F_\mathrm{p}^n)/(1 - F_\mathrm{p})$ or $1 - F_\mathrm{p}^n$. This approaches 1 for large n, suggesting that all of the water attributed to time t, i.e. $P_t - E_{tx}$, will eventually emerge as river flow. For $F_\mathrm{p} = 0$ all of $(P_t - E_{tx})$ emerges on the first day, and river flow is as unpredictable as precipitation itself. For $F_\mathrm{p} = 1$ all of $(P_t - E_{tx})$ contributes to the stable daily flow rate, and it takes an infinitely long period of time for the last drop of water to get to the river. For declining F_p, $(1 > F_\mathrm{p} > 0)$, river flow gradually becomes less predictable, because a greater part of the stochastic precipitation term contributes to variable rather than evened-out river flow.

Taking long-term summations of the right- and left-hand sides of Eq. (4) we obtain

$$\begin{aligned}\sum Q_t &= \sum \left(F_\mathrm{p} Q_{t-1} + \left(1 - F_\mathrm{p}\right)\left(P_t - E_{tx}\right)\right)\\ &= F_\mathrm{p} \sum \left(Q_{t-1} + \left(1 - F_\mathrm{p}\right)\left(\sum P_t - \sum E_{tx}\right)\right),\end{aligned} \tag{5}$$

which is consistent with the basic water budget, $\sum Q = \sum P - \sum E$, at timescales long enough for changes in soil water buffer stocks to be ignored. Therefore, the total annual and, hence, the mean daily river flow are independent of F_p. This does not preclude that processes of watershed degradation or restoration that affect the partitioning of P over Q and E also affect F_p.

2.2 Base flow

Clarifying the $Q_{\mathrm{a},t}$ contribution is equivalent in one of several ways to separate base flow from peak flows. Rearranging Eq. (3) we obtain

$$\sum_1^T Q_{\mathrm{a},t} = \left(1 - F_\mathrm{p}\right)\left(\sum_1^T Q_t E_{tx}\right). \tag{6}$$

The $\sum Q_{\mathrm{a},t}$ term reflects the sum of peak flows in millimetres. Its complement, $F_\mathrm{p} \sum Q_t$, reflects the sum of base flow, also in millimetres. For $F_\mathrm{p} = 1$ (the theoretical maximum), we conclude that all $Q_{\mathrm{a},t}$ must be zero, and all flow is "base flow".

2.3 Low flows

The lowest flow expected in an annual cycle is $Q_x\, F_\mathrm{p}^{N_\mathrm{max}}$ where Q_x is flow on the first day without rain and N_max the longest series of dry days. Taken at face value, a decrease in F_p has a strong effect on low flows, with a flow of 10 % of Q_x reached after 45, 22, 14, 10, 8 and 6 days for $F_\mathrm{p} = 0.95$, 0.9, 0.85, 0.8, 0.75

and 0.7, respectively. However, the groundwater reservoir that is drained, equalling the cumulative dry-season flow if the dry period is sufficiently long, is $Q_x/(1-F_\mathrm{p})$. If F_p decreases to F_{px} but the groundwater reservoir ($\mathrm{Res} = Q_x/(1-F_\mathrm{p})$) is not affected, initial flows in the dry period will be higher ($Q_x\, F^i_{\mathrm{p}x}(1-F_{\mathrm{p}x})\,\mathrm{Res} > Q_x\, F^i_\mathrm{p}(1-F_\mathrm{p})\,\mathrm{Res}$ for $i < \log((1-F_{\mathrm{p}x})/(1-F_\mathrm{p}))/\log(F_\mathrm{p}/F_{\mathrm{p}x})$). It thus matters how low flows are evaluated: from the perspective of the lowest level reached, or as cumulative flow. The combination of climate, geology and land form are the primary determinants of cumulative low flows, but if land cover reduces the recharge of groundwater there may be impacts on dry-season flow, that are not directly reflected in F_p.

If a single F_p value would account for both dry and wet season, the effects of changing F_p on low flows may well be more pronounced than those on flood risk. Empirical tests are needed of the dependence of F_p on Q (see below). Analysis of the way an aggregate F_p depends on the dominant flow pathways provides a basis for differentiating F_p within a hydrologic year.

2.4 Flow-pathway dependence of flow persistence

The patch-level partitioning of water between infiltration and overland flow is further modified at hillslope level, with a common distinction between three pathways that reaches streams: overland flow, interflow and groundwater flow (Band, 1993; Weiler and McDonnell, 2004). An additional interpretation of Eq. (1), potentially adding to our understanding of results but not needed for analysis of empirical data, can be that three pathways of water through a landscape contribute to river flow (Barnes, 1939): groundwater release with $F_\mathrm{p,g}$ values close to 1.0, overland flow with $F_\mathrm{p,o}$ values close to 0 and interflow with intermediate $F_\mathrm{p,i}$ values.

$$Q_t = F_\mathrm{p,g} Q_{t-1,\mathrm{g}} + F_{\mathrm{p},i} Q_{t-1,\mathrm{i}} + F_\mathrm{p,o} Q_{t-1,\mathrm{o}} + Q_{\mathrm{a},t} \tag{7}$$

$$F_\mathrm{p} = \left(F_\mathrm{p,g} Q_{t-1,\mathrm{g}} + F_\mathrm{p,i} Q_{t-1,\mathrm{i}} + F_\mathrm{p,o} Q_{t-1,\mathrm{o}}\right)/Q_{t-1} \tag{8}$$

On this basis a decline or increase in overall weighted average F_p can be interpreted as an indicator of a shift of dominant runoff pathways through time within the watershed. Dry-season flows are dominated by $F_\mathrm{p,g}$. The effective F_p in the rainy season can be interpreted as indicating the relative importance of the other two flow pathways. F_p reflects the fractions of total river flow that are based on groundwater, overland flow and interflow pathways:

$$F_\mathrm{p} = F_\mathrm{p,g}\left(\sum Q_{t,\mathrm{g}}/\sum Q_t\right) + F_\mathrm{p,o}\left(\sum Q_{t,\mathrm{o}}/\sum Q_t\right) + F_\mathrm{p,i}\left(\sum Q_{t,\mathrm{i}}/\sum Q_t\right). \tag{9}$$

Beyond the type of degradation of the watershed that, mostly through soil compaction, leads to enhanced infiltration-excess (or Hortonian) overland flow (Delfs et al., 2009), saturated conditions throughout the soil profile may also induce overland flow, especially near valley bottoms (Bonell, 1993; Bruijnzeel, 2004). Thus, the value of $F_\mathrm{p,o}$ can be substantially above zero if the rainfall has a significant temporal autocorrelation, with heavy rainfall on subsequent days being more likely than would be expected from general rainfall frequencies. If rainfall following a wet day is more likely to occur than following a dry day, as is commonly observed in Markov chain analysis of rainfall patterns (Jones and Thornton, 1997; Bardossy and Plate, 1991), the overland flow component of total flow will also have a partial temporal autocorrelation, adding to the overall predictability of river flow. In a hypothetical climate with evenly distributed rainfall, we can expect F_p to be 1.0 even if there is no infiltration and the only pathway available is overland flow. Even with rainfall that is variable at any point of observation but has low spatial correlation, it is possible to obtain F_p values of (close to) 1.0 in a situation with (mostly) overland flow (Ranieri et al., 2004).

2.5 Relationship between flow persistence and flashiness index

The Richards–Baker (R–B) "flashiness index" (Baker et al., 2004) is defined as

$$\mathrm{FI} = \sum_t |\Delta Q_t| / \sum_t Q_t = \sum_{ti} (Q_t - Q_{t-1}) + \sum_{td} (Q_{t-1} - Q_t), \tag{10}$$

where ti indicates all times t that $Q_t > Q_{t-1}$ and td indicates all times t that $Q_t \le Q_{t-1}$. Over a time frame that flow has no net trend, the sum of increments ($\sum_{ti}(Q_t - Q_{t-1})$) is equal to the sum of declines ($\sum_{td}(Q_{t-1} - Q_t)$).

Substituting Eq. (5) in Eq. (10) we obtain

$$\begin{aligned}\mathrm{FI} &= 2\left(1-F_\mathrm{p}\right)\left(0.5\Delta S + \sum_{ti}(P_t - E_{tx} - Q_t)\right)/\sum_t Q_t \\ &= 2\left(1-F_\mathrm{p}\right)\left(-0.5\Delta S + \sum_{td}(-P_t + E_{tx} + Q_t)\right)/\sum_t Q_t,\end{aligned} \tag{11}$$

where ΔS representes change in catchment storage; $\Delta S = (1-F_\mathrm{p})\ (-\sum_{ti}\ (P_t - E_{tx} - Q_t) + \sum_{td}\ (-P_t + E_{tx} + Q_t))$. This suggests that $\mathrm{FI} = 2\ (1-F_\mathrm{p})$ is a first approximation and becomes zero for $F_\mathrm{p} = 1$. These approximations require that changes in the catchment have no influence on P_t or E_{tx} values. If E_{tx} is negatively affected (either by a change in vegetation or by insufficient buffering, reducing water availability on non-rainfall days) flashiness will increase, beyond the main effects on F_p. The rainfall term, counted positive for all days with flow increase and negatively for days with declining flow, hints at one of the major reasons why the flashiness index tends to get smaller when larger catchment areas are involved; rainfall will tend to get more evenly distributed over time, unless the spatial correlation of rainfall is (close to) 1 and all rainfall derives from fronts passing over the area uniformly. Where (part of) precipitation occurs as snow, the timing of snowmelt defines

P_t as used here. Where vegetation influences timing and synchrony of snowmelt, this will be reflected in the flashiness index. It may not directly influence flow persistence, but will be accounted for in the flow description that uses flow persistence as a key parameter.

3 Methods

3.1 River-flow data for four tropical watersheds

To test the applicability of the F_p metric and explore its properties, data from four Southeast Asian watersheds were used, which will be described and further analysed in Part 2 (van Noordwijk et al., 2017). The first watershed data set is the Way Besai (414.4 km^2) in Lampung province, Sumatra, Indonesia (Verbist et al., 2010). With an elevation between 720 and 1831 m a.s.l., the Way Besai is dominated by various coffee production systems (64 %), with remaining forest (18 %), horticulture and crops (12 %) and other land uses (6 %). Daily rainfall data from 1976 to 2007, was generated by interpolation of eight rainfall stations using Thiessen polygons; data were obtained from BMKG (Agency on Meteorology, Climatology and Geophysics), PU (Public Work Agency) and PLN (National Electricity Company). The average of annual rainfall was 2474 mm, with observed values in the range 1216–3277 mm. River-flow data at the outflow of the Way Besai were also obtained from PU and PUSAIR (Centre for Research and Development on Water Resources), with an average of river flow of 16.7 $m^3\,s^{-1}$.

Data from three other watersheds were used to explore the variation of F_p across multiple years and its relationship with the flashiness index: Bialo (111.7 km^2) in South Sulawesi, Indonesia, with agroforestry as the dominant land cover type, Cidanau (241.6 km^2) in West Java, Indonesia, dominated by mixed agroforestry land uses but with a peat swamp before the final outlet and Mae Chaem (3892 km^2) in northern Thailand, part of the upper Ping Basin, and dominated by evergreen, deciduous and pine forest. Detailed information on these watersheds and the data sources is provided in Part 2 (van Noordwijk et al., 2017).

3.2 Numerical examples

For visualising the effects of stochastic rainfall on river flow according to Eq. (1), a spreadsheet model that is available from the authors on request was used in "Monte Carlo" simulations. Fixed values for F_p were used in combination with a stochastic $Q_{a,t}$ value. The latter was obtained from a random generator (rand) with two settings for a (truncated) sinus-based daily rainfall probability: (a) one for situations that have approximately 120 rainy days, and an annual Q of around 160 mm, and (b) one that leads to around 45 rainy days and an annual total around 600 mm. Maximum daily $Q_{a,t}$ was chosen as 60 mm in both cases. For the figures, realisations for various F_p values were retained that were within 10 % of this number of rainy days and annual flow total, to focus on the effects of F_p as such.

3.3 Flow persistence as a simple flood risk indicator

For numerical examples (implemented in a spreadsheet model), flow on each day can be derived as

$$Q_t = \sum_j^t F_p^{t-j}\left(1 - F_p\right) p_j P_j, \tag{12}$$

where p_j reflects the occurrence of rain on day j (reflecting a truncated sine distribution for seasonal trends) and P_j is the rain depth (drawn from a uniform distribution). From this model the effects of F_p (and hence of changes in F_p) on maximum daily flow rates, plus maximum flow totals assessed over a 2–5 days period, was obtained in a Monte Carlo process (without Markov autocorrelation of rainfall in the default case; see below). Relative flood protection was calculated as the difference between peak flows (assessed for 1–5 days duration after a 1-year warm-up period) for a given F_p vs. those for $F_p = 0$, relative to those at $F_p = 0$.

3.4 An algorithm for deriving F_p from a time series of streamflow data

Equation (3) provides a first method to derive F_p from empirical data if these cover a full hydrologic year. In situations where there is no complete hydrograph and/or in situations where we want to quantify F_p for shorter time periods (e.g. to characterise intra-seasonal flow patterns) and the change in the storage term of the water budget equation cannot be ignored, we need an algorithm for estimating F_p from a series of daily Q_t observations.

Where rainfall has clear seasonality, it is an attractive and indeed common practice to derive a groundwater recession rate from a semi-logarithmic plot of Q against time (Tallaksen, 1995). As we can assume for such periods that $Q_{a,t} = 0$, we obtain $F_p = Q_t/Q_{t-1}$, under these circumstances. We cannot be sure, however, that this $F_{p,g}$ estimate also applies in the rainy season, because overall wet-season F_p will include contributions by $F_{p,o}$ and $F_{p,i}$ as well (compare Eq. 9). In locations without a distinct dry season, we need an alternative method.

A bi-plot of Q_t against Q_{t-1} will lead to a scatter of points above a line with slope F_p, with points above the line reflecting the contributions of $Q_{a,t} > 0$, while the points that plot on the F_p line itself represent $Q_{a,t} = 0\,\mathrm{mm\,day^{-1}}$. There is no independent source of information on the frequency at which $Q_{a,t} = 0$, nor what the statistical distribution of $Q_{a,t}$ values is if it is non-zero. Calculating back from the Q_t series, we can obtain an estimate ($Q_{a,t,F_{p,\mathrm{try}}}$) of $Q_{a,t}$ for any given estimate ($F_{p,\mathrm{try}}$) of F_p, and select the most plausible F_p value. For high $F_{p,\mathrm{try}}$ estimates there will be many negative $Q_{a,t,F_{p,\mathrm{try}}}$ values, for low $F_{p,\mathrm{try}}$ estimates all $Q_{a,t,F_{p,\mathrm{try}}}$ values will be larger. An algorithm to

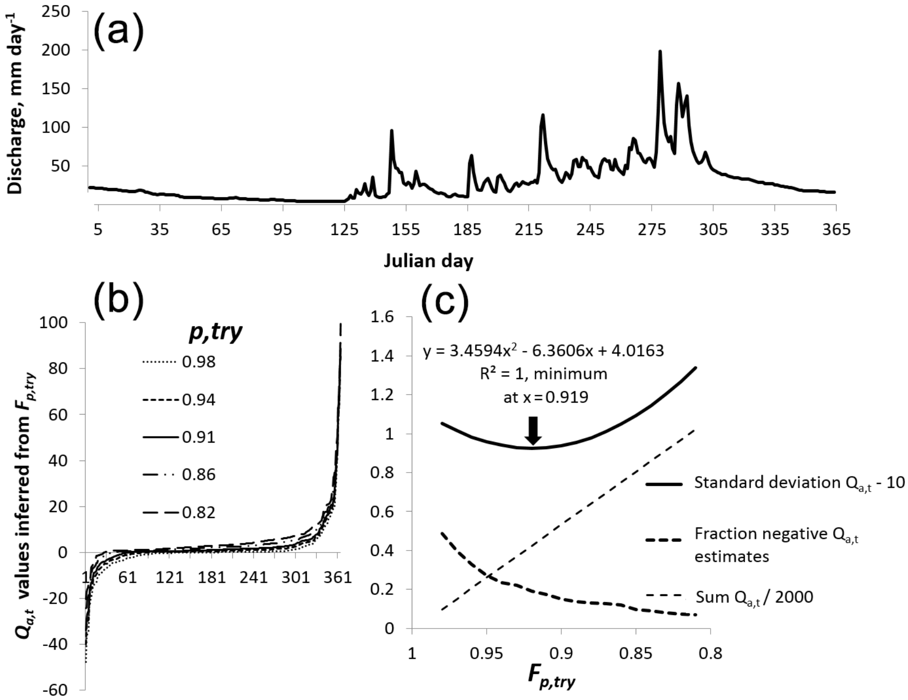

Figure 3. Example of the derivation of best-fitting $F_{\mathrm{p,try}}$ value for an example hydrograph **(a)** on the basis of the inferred Q_{a} distribution (cumulative frequency in **b**), and three properties of this distribution **(c)**: its sum, frequency of negative values and standard deviation; the $F_{\mathrm{p,try}}$ minimum of the latter is derived from the parameters of a fitted quadratic equation.

derive a plausible F_{p} estimate can thus make use of the corresponding distribution of "apparent Q_{a}" values as estimates of $F_{\mathrm{p,try}}$, calculated as $Q_{\mathrm{a},t,F_{\mathrm{p,try}}} = Q_t - F_{\mathrm{p,try}}\, Q_{t-1}$. While $Q_{\mathrm{a},t}$ cannot be negative in theory, small negative Q_{a} estimates are likely when using real-world data with their inherent errors. The FlowPer F_{p} algorithm (van Noordwijk et al., 2011) derives the distribution of $Q_{\mathrm{a},t,F_{\mathrm{p,try}}}$ estimates for a range of $F_{\mathrm{p,try}}$ values (Fig. 3b) and selects the value $F_{\mathrm{p,try}}$ that minimises the variance $\mathrm{Var}(Q_{\mathrm{a},F_{\mathrm{p,try}}})$ (or its standard deviation) (Fig. 3c). It is implemented in a spreadsheet workbook that can be downloaded from the ICRAF website (http://www.worldAgroforestry.org/output/flowper-flow-persistence-model). An R routine is being developed.

A consistency test is needed that the high-end Q_t values relate to Q_{t+1} in the same was as low or medium Q_t values. Visual inspection of Q_{t+1} vs. Q_t, with the derived F_{p} value, provides a qualitative view of the validity of this assumption. The F_{p} algorithm can be applied to any population of (Q_{t-1}, Q_t) pairs, e.g. selected from a multi-year data set on the basis of 3-month periods within the hydrological year.

3.5 Flashiness and flow separation

Hydrographs analysed for F_{p} were also used for calculating the R–B flashiness index (Baker et al., 2004) by summing the absolute values of all daily changes in flow. Two common flow separation algorithms (fixed and sliding interval methods; Furey and Gupta, 2001) were used to estimate the base-flow fraction at an annual basis. The average of the two was compared to F_{p}.

4 Result

4.1 Numerical examples

Figure 4 provides two examples, for annual river flows of around 1600 and 600 mm yr^{-1}, of the way a change in F_{p} values (based on Eq. 1) influences the pattern of river flow for a unimodal rainfall regime with a well-developed dry season. The increasing "spikiness" of the graph as F_{p} is lowered, regardless of annual flow, indicates reduced predictability of flow on any given day during the wet season on the basis of the flow on the preceding day.

A bi-plot of river flow on subsequent days for the same simulations (Fig. 5) shows two main effects of reducing the F_{p} value: the scatter increases, and the slope of the lower envelope containing the swarm of points is lowered (as it equals F_{p}). Both of these changes can provide entry points for an algorithm to estimate F_{p} from empirical time series, provided the basic assumptions of the simple model apply and the data are of acceptable quality.

For the numerical examples shown in Fig. 4, the relative increase of the maximum daily flow when the F_{p} value de-

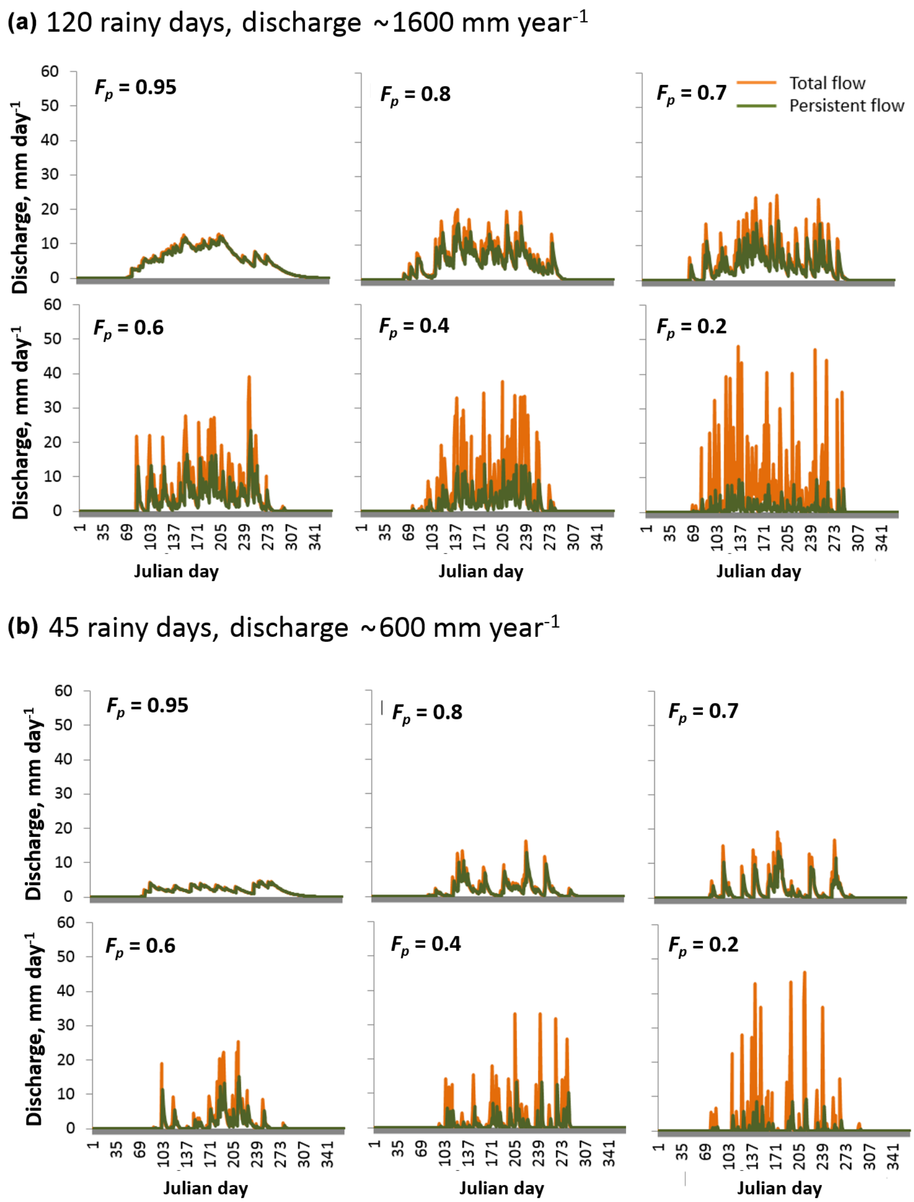

Figure 4. Effects of the F_p parameter on hydrographs of daily river flow generated by a random rainfall generator, with persistent and additional flow components indicated, for two settings with total rainfall of approximately 1600 and 600 mm yr^{-1} (NB river flow is here expressed as mm day^{-1} rather than as $m^3 s^{-1}$ as in Fig. 3).

creased from a value close to 1 (0.98) to nearly 0 depended on the rainfall regime; with lower annual rainfall but the same maximum daily rainfall, the response of peak flows to decrease in F_p became stronger.

4.2 Flood intensity and duration

Figure 6 shows the effect of F_p values in the range 0 to 1 on the maximum flows obtained with a random time series of "effective rainfall", compared to results for $F_p = 0$. Maximum flows were considered at timescales of 1 to 5 days, in a moving average routine. This way a relative flood protection, expressed as reduction of peak flow, could be related to F_p (Fig. 6a).

Relative flood protection rapidly decreased from its theoretical value of 100 % at $F_p = 1$ (when there was no variation in river flow), to less than 10 % at F_p values of around 0.5. Relative flood protection was slightly lower when the assessment period was increased from 1 to 5 days (between 1 and 3 days it decreased by 6.2 %, from 3 to 5 days by a further 1.3 %). Two counteracting effects are at play here; a lower F_p means that a larger fraction $(1 - F_p)$ of the effective rainfall contributes to river flow, but the increased flow is less persis-

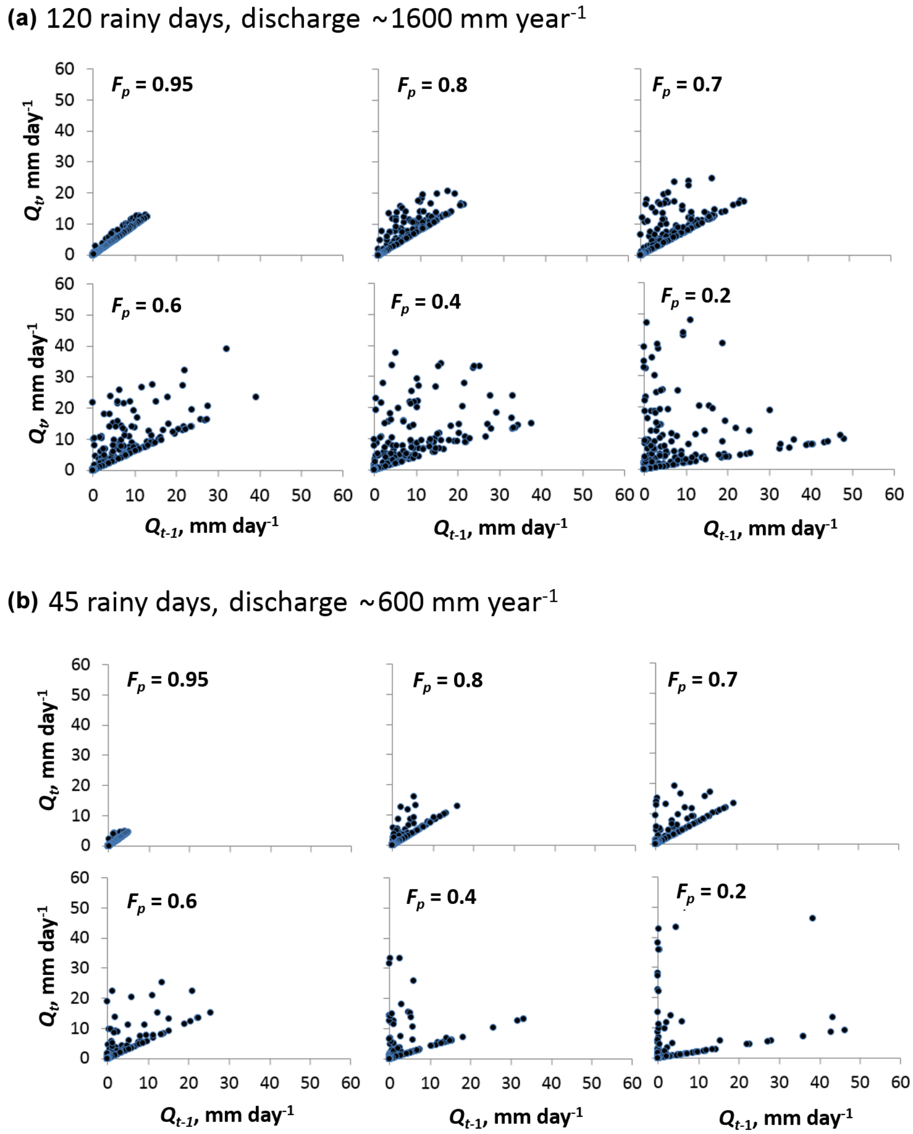

Figure 5. Panels **(a)** and **(b)** show the temporal autocorrelation of river flow for the same simulations as Fig. 4; the lower envelope of the points indicated slope F_p, the points above this line the effect of fresh additions to river flow.

tent. In the example the flood protection in situations where the rainfall during 1 or 2 days causes the peak is slightly stronger than where the cumulative rainfall over 3–5 days causes floods, as typically occurs downstream.

As we expect from Eq. (5) that peak flow is at $(1 - F_\mathrm{p})$ times peak rainfall amounts, the effect of a change in F_p depends not only on the change in F_p that we are considering, but also on its initial value. Higher initial F_p values will lead to more rapid increases in high flows for the same reduction in F_p (Fig. 6b). However, flood duration rather responds to changes in F_p in a curvilinear manner, as flow persistence implies flood persistence (once flooding occurs), but the greater the flow persistence the less likely such a flooding threshold is passed (Fig. 6c). The combined effect may be restricted to about 3 days of increase in flood duration for the parameter values used in the default example, but for different parameterisation of the stochastic ε other results might be obtained.

4.3 Algorithm for F_p estimates from river-flow time series

The algorithm has so far returned non-ambiguous F_p estimates on any modelled time series data of river flow, as well as for all empirical data set we tested (including all examples tested in Part 2, van Noordwijk et al., 2017), although there probably are data sets on which it can breakdown. Visual inspection of Q_{t-1}/Q_t bi-plots (as in Fig. 4) can provide clues to non-homogenous data sets, to potential situations where effective F_p depends on flow level Q_t and where

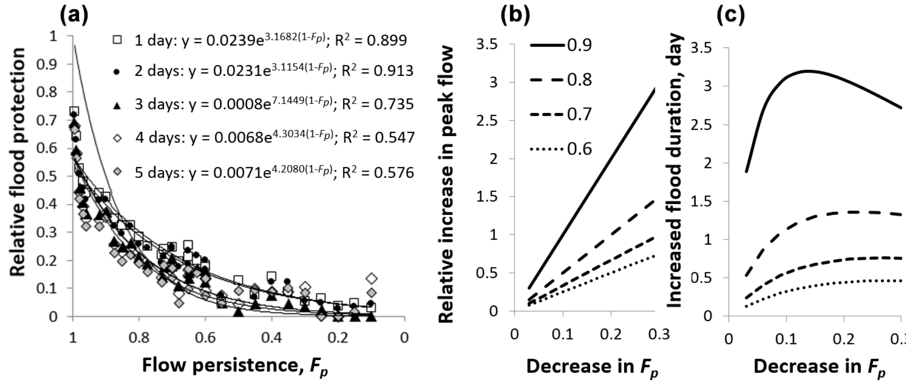

Figure 6. **(a)** Effects of flow persistence on the relative flood protection (decrease in maximum flow measured over a 1–5 day period relative to a case with $F_p = 0$ (a few small negative points were replaced by small positive values to allow for the exponential fit); **(b)** and **(c)** effects of a decrease in flow persistence on the volume of water involved in peak flows (**b**; relative to the volume at F_p is 0.6–0.9) and in the duration (in days) of floods **(c)**.

data are not consistent with a straight-line lower envelope. Where river-flow estimates were derived from a model with random elements, however, variation in F_p estimates was observed, which suggests that specific aspects of actual rainfall, beyond the basic characteristics of a watershed and its vegetation, do have at least some effect. Such effects deserve to be further explored for a set of case studies, as their strength probably depends on context.

4.4 Flow persistence compared to base flow and flashiness index

Figure 7 compares results for a hydrograph of a single year for the Way Besai catchment, described in more detail in Part 2 (van Noordwijk et al., 2017). While there is agreement on most of what is indicated as base flow, the short-term response to peaks in the flow differ, with base flow in the F_p method more rapidly increasing after peak events.

When compared across multiple years for four Southeast Asian catchments (Fig. 8a), there is partial agreement in the way inter-annual variation is described in each catchment, while numerical values are similar. However, the ratio of what is indicated as base flow according to the F_p method and according to standard hydrograph separation varies from 1.05 to 0.86.

Figure 8b and c compare numerical results for the R–B flashiness index with F_p for the four test catchments and for a number of hydrographs constructed as in Fig. 3a. The two concepts are inversely related, as expected from Eq. (11), but where F_p is constrained to the 0–1 interval the R–B flashiness index can attain values up to 2.0, with the value for $F_p = 0$ depending on properties of the local rainfall regime. Where hydrographs were generated with a simple flow model with the F_p parameter as key variable, the flashiness index is more tightly related, especially for higher F_p values, than where both flashiness index and F_p were derived

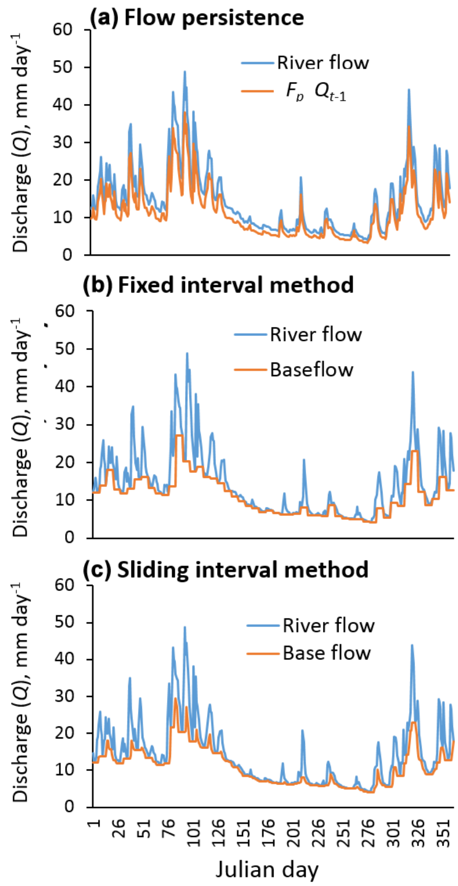

Figure 7. Comparison of base-flow separation of a hydrograph according to the flow persistence method **(a)** and two common flow separation methods, respectively, with fixed **(b)** and sliding intervals **(c)**.

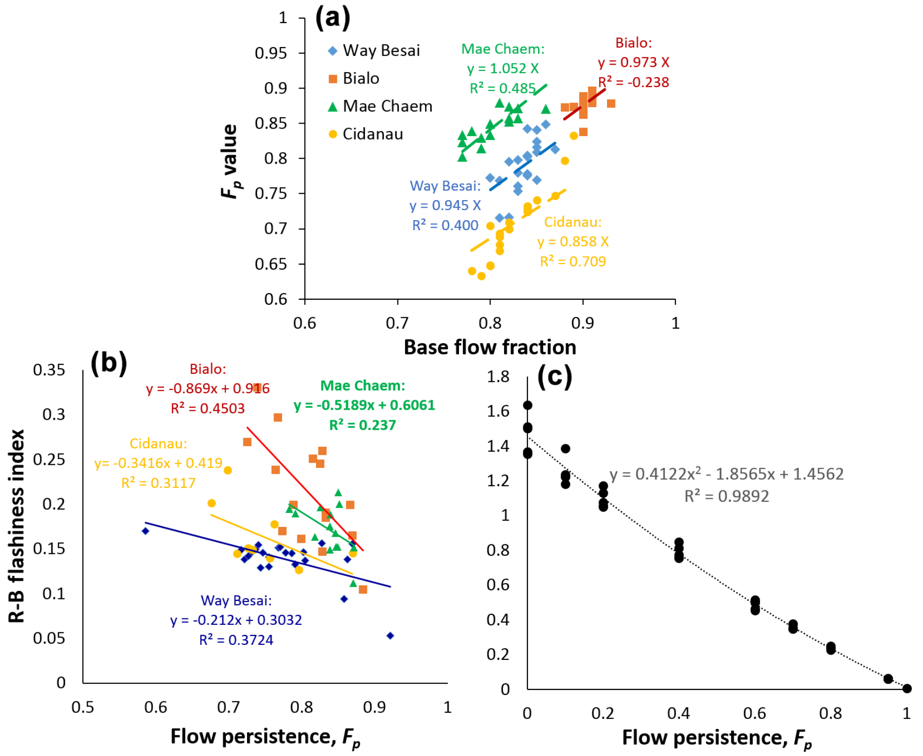

Figure 8. **(a)** Comparison of yearly data for four Southeast Asian watersheds analysed with common flow separation methods (average of results in Fig. 7) and the flow persistence method and comparison of the Richards–Baker flashiness index (Baker et al., 2004) and the flow persistence metric F_p for **(b)** four Southeast Asian watersheds, **(c)** a series of hydrographs as in Fig. 4a, with five replicates per F_p value.

from existing flow data (Fig. 8c vs. Fig. 8b). The difference in slope between the four watersheds in Fig. 8b appears to be primarily related to aspects of the local rainfall pattern that deserve further analysis in larger data sets of this nature.

5 Discussion

5.1 Salience

Key salience aspects are "does flow persistence relate to important aspects of watershed behaviour" and "does it help to select management actions?". A major finding in the derivation of F_p was that the flow persistence measured at daily timescale can be logically linked to the long-term water balance under the assumption that the watershed is defined on the basis of actual groundwater flows, and that the proportion of peak rainfall that translates to peak river flow equals the complement of flow persistence. This feature links effects on floods of changes in watershed quality, as commonly expressed in curve numbers and flashiness indices, to effects on low flows, as commonly expressed in base-flow metrics. The F_p parameter as such does not predict when and where flooding will occur, but it does help to assess to what extent another condition of the watershed, with either higher or lower F_p would translate the same rainfall into larger or smaller peak water flows. This is salient, especially if the relative contributions of (anthropogenic) land cover and the (exogenous, probabilistic) specifics of the rainfall pattern can be further teased apart (see Part 2, van Noordwijk et al., 2017). Where F_p may describe the descending branch of hydrographs at a relevant timescale, details of the ascending branch beyond the maximum daily flow reached may be relevant for reducing flood damage, and may require more detailed study at higher temporal resolution.

Figures 3 and 6 show that most of the effects of a decreasing F_p value on peak discharge (which is the basis for downstream flooding) occur between F_p values of 1 and 0.7, with the relative flood protection value reduced to 10 % when F_p reaches 0.5. As indicated in Fig. 2, peak discharge is only one of the factors contributing to flood risk in terms of human casualties and physical damage. Flood risks are themselves non-linear and in strongly topography-specific ways related to the volume of river flow after extreme rainfall events. While the expected fraction of rainfall that contributes to direct flow is linearly related to rainfall via $(1-F_p)$, flooding risk as such will have a non-linear relationship with rainfall, which depends on topography and antecedent rainfall. Catchment changes, such as increases or decreases in percentage tree cover, will generally have a non-linear relationship with F_p as well as with flooding risks. The F_p value has an inverse effect on the fraction of recent rainfall that becomes river flow, but the effect on peak flows is less, as higher F_p values imply higher base flow. The way these counteracting effects balance out depends on details of the local rainfall pattern (including its Markov chain temporal autocorrelation), as well as the downstream topography and risk of people being at the wrong time at a given place, but the F_p value is an efficient way of summarising complex land use mosaics and upstream topography in its effect on river flow. The difference between wet-season and dry-season F_p deserves further analysis. In climates with a real rainless dry season, dry-season F_p is dominated by the groundwater release fraction of the watershed, regardless of land cover, while in wet season it depends on the mix (weighted average) of flow pathways. The degree to which F_p can be influenced by land cover needs to be assessed for each landscape and land cover combination, including the locally relevant forest and forest-derived land classes, with their effects on interception, soil infiltration and time pattern of transpiration. The F_p value can summarise results of models that explore land use change scenarios in local context. To select the specific management actions that will maintain or increase F_p, a locally calibrated land use/hydrology model is needed, such as GenRiver (Part 2, van Noordwijk et al., 2017), DHV (Bergström, 1995) or SWAT (Yen et al., 2015).

The "health" wording has been used as a comprehensive concept of the way (a) climate forcing, (b) watershed vegetation and soil conditions and (c) engineering interventions interact on functional aspects of river flow. Ma et al. (2014) described a method to separate these three influences on river flow. In the four catchments we used as an example there have been no major dams or reservoirs installed upstream of the points of measurement. Where these do exist the specific operating rules of reservoirs need to be included in any model and these can have a major influence on downstream flow, depending on the primary use for power generation, dry-season irrigation or stabilising river flow for riverine transport. Although a higher F_p value will in most cases be desirable (and a decrease in F_p undesirable), we may expect that in an ecological perspective on watershed health, the change in low flows that can occur in the flow regime of degrading and intensively managed watersheds alike, depending on the management rules for reservoirs, is at least as relevant as changes in flood risks, as many aquatic organisms thrive during floods (Pahl-Wostl et al., 2013; Poff et al., 2010). Downstream biota can be expected to have adapted to the pre-human flow conditions, inherent F_p and variability. Decreased variability of flow achieved by engineering interventions (e.g. a reservoir with constant release of water to generate hydropower) may have negative consequences for fish and other biota (Richter et al., 2003; McCluney et al., 2014). In an extensive literature review Poff and Zimmerman (2010) found no general, transferable quantitative relationships between flow alteration and ecological response, but the risk of ecological change increases with increasing magnitude of flow alteration.

Various geographically defined watershed health concepts are in use (see for example https://www.epa.gov/hwp/healthy-watersheds-projects-region-5; City of Fort Collins (2015), employing a range of specific indicators, including the "R–B flashiness index" (Baker et al., 2004). The definition of watershed health, like that of human health has evolved over time. Human health was seen as a state of normal function that could be disrupted from time to time by disease. In 1948 the World Health Organization (1958) proposed a definition that aimed higher, linking health to well-being, in terms of physical, mental and social aspects, and not merely the absence of disease and infirmity. Health became seen as the ability to maintain homeostasis and recover from injury, but remained embedded in the environment in which humans function.

5.2 Credibility

Key credibility questions are "consistency of numerical results" and "how sensitive are results to bias and random error in data sources?". A key strength of our flow persistence parameter, which can be derived from a limited number of observations of river flow at a single point along the river, without knowledge of rainfall events and catchment conditions, is also its major weakness. If rainfall data exist, and especially rainfall data that apply to each subcatchment, the Q_a term does not have to be treated as a random variable and event-specific information on the flow pathways may be inferred for a more precise account of the hydrograph. But for the vast majority of rivers in the tropics, advances in remotely sensed rainfall data are needed to achieve that situation and F_p may be all that is available to inform public debates on the location-specific relation between forests and floods.

The main conclusions from the numerical examples analysed so far are that intra-annual variability of F_p values between wet and dry seasons was around 0.2, inter-annual variability in either annual or seasonal F_p was generally in the 0.1 range, while the difference between observed and simulated flow data as basis for F_p calculations was mostly less than 0.1. With current methods, it seems that effects of land cover change on flow persistence that shift the F_p value by about 0.1 are the limit of what can be asserted from empirical data (with shifts of that order in a single year a warning sign rather than a firmly established change). When derived from observed river-flow data, F_p is suitable for monitoring change (degradation, restoration) and can be a serious candidate for monitoring performance in outcome-based ecosystem service management contracts. In interpreting changes in F_p as caused by changes in the condition in the watershed, however, must be excluded with regard to changes in specific properties of the rainfall regime. At the scale of paired catchment studies this assumption may be reasonable, but in temporal change (or using specific events as starting point for analysis), it is not easy to disentangle interacting effects (Ma et al., 2014). Recent evidence that vegetation responds to, as well as influences, rainfall (arrow 10 in Fig. 2; van Noordwijk et al., 2015b) further complicates the analysis across scales.

As indicated, the F_p method is related to earlier methods used in streamflow hydrograph separation of base flow and quick flow. While textbooks (Ward and Robinson, 2000; Hornberger et al., 2014) tend to be critical of the lack of objectivity of graphical methods, algorithms are used for deriving the minimum flow in a fixed or sliding period of reference as base flow (Sloto and Crouse, 1996; Furey and Gupta, 2001). The time interval used for deriving the minimum flow depends on catchment size. Recursive models that describe flow in a next time interval on the basis of a fraction of that in the preceding time interval with a term for additional flow due to additional rainfall have been used in analysis of peak flow event before, with time intervals as short as 1 min rather than the 1 day we use here (Rose, 2004). Through reference to an overall mass balance a relationship similar to what we found here (F_p times preceding flow plus $1 - F_\mathrm{p}$ times recent inputs) was also used in such models. To our knowledge, the method we describe here at daily timescales has not been used before.

The idea that the form of the storage–discharge function can be estimated from analysis of streamflow fluctuations has been explored before for a class of catchments in which discharge is determined by the volume of water in storage (Kirchner, 2009). Such catchments behave as simple first-order non-linear dynamical systems and can be characterised in a single-equation rainfall–runoff model that predicted streamflow, in a test catchment in Wales, as accurately as other models that are much more highly parameterised. This model of the $\mathrm{d}Q/\mathrm{d}t$ vs. Q relationship can also be analytically inverted; thus, it can, according to Kirchner (2009), be used to "do hydrology backward", that is, to infer time series of whole-catchment precipitation directly from fluctuations in streamflow. The slope of the log–log relationship between flow recession ($\mathrm{d}Q/\mathrm{d}t$) and Q that Kirchner (2009) used is conceptually similar to the F_p metric we derived here, but the specific algorithm to derive the parameter from empirical data differs. Further exploration of the underlying assumptions is needed. Estimates of $\mathrm{d}Q/\mathrm{d}t$ are sensitive to noise in the measurement of Q and the possibly frequent and small increases in Q can be separated from the expected flow recession in the algorithm we presented here.

Table 1 compares a number of properties (salience and legitimacy in properties 1–4, credibility dimensions in 5–10) for the R–B flashiness index (Baker et al., 2004) and flow persistence. The main advantage of continuing with the flashiness index is that there is an empirical basis for comparisons and the index has been included in existing "watershed health" monitoring programmes, especially in the USA. The main advantage of including F_p is that it can be estimated from incomplete flow records, has a clear link to peak flow events and has a more direct relationship with underlying flow pathways, changes in rainfall (or snowmelt) and evapotranspiration, reflecting land cover change.

Seibert and Beven (2009) discussed the increase in predictive skill of models depending on the amount of location-specific data that can be used to constrain them. They found that the ensemble prediction of multiple models for a single location clearly outperformed the predictions using single parameter sets and that surprisingly little runoff data were necessary to identify model parameterisations that provided good results for "ungauged" test periods in cases where actual measurements were available. Their results indicated that a few runoff measurements can contain much of the information content of continuous runoff time series. The way these conclusions might be modified if continuous measurements for limited time periods, rather than separated single data points on river flow could be used, remains to be explored. Their study indicated that results may differ significantly between catchments and critical tests of F_p across multiple situations are obviously needed, as Part 2 (van Noordwijk et al., 2017) will provide. In discussions and models of temperate zone hydrology (Bergström, 1995; Seibert, 1999), snowmelt is a major component of river flow and effects of forest cover on spring temperatures are important to the buffering of the annual peaks in flow that tend to occur in this season. Application of the F_p method to data describing such events has yet to be done.

5.3 Legitimacy

Legitimacy aspects are "does it match local knowledge" and "can it be used to empower local stakeholders of watershed management" and "can it inform risk management?". As the F_p parameter captures the predictability of river flow that is a key aspect of degradation according to local knowledge systems, its results are much easier to convey than full hydrographs or exceedance probabilities of flood levels. By focusing on observable effects at river level, rather than prescriptive recipes for land cover ("Reforestation"), the F_p parameter can be used to more effectively compare the combined effects of land cover change, changes in the riparian wetlands and engineered water storage reservoirs, in their effect on flow buffering. It is a candidate for shifting environmental service reward contracts from input- to outcome-based monitoring (van Noordwijk et al., 2012). Therefore, it can be used as part of a negotiation support approach to natural resources management in which levelling off on knowledge and joint fact finding in blame attribution are key steps to negotiated solutions that are legitimate and seen to be so (van Noordwijk et al., 2013; Leimona et al., 2015). Quantification of F_p can help assess tactical management options (Burt et al., 2014) as in a recent suggestion to minimise negative downstream impacts of forestry operations on streamflow by avoiding land clearing and planting operations in locally wet La Niña years. But the most challenging aspect of the management of flood, as any other environmental risk, is that the frequency of dis-

Table 1. Comparison of properties of the flashiness index and flow persistence F_p.

No.	Flashiness index (Baker et al., 2004)	Flow persistence (as defined here)
1	Has direct appeal to non-technical audiences	Potentially similar
2	Where reservoir management rules imply major changes in ΔS, flashiness still describes implications for flow regimes	Is focused on the effects of changes in (upper) catchment land cover, not where reservoir management determines flow
3	Values depend on the scale of evaluating river flow; no absolute criteria for what is healthy	Similar
4	Increase generally not desirable	Decrease generally not desirable
5	Varies in range [0–2], may need normalising by division by 2	Varies in range [0–1]
6	Requires full-year flow record to be calculated	Can be estimated from any set of sequential flow observations
7	Empirical metric, no direct link to underlying process understanding	Overall F_p can be understood as weighted average of the F_p's of contributing flow pathways (overland, subsurface and groundwater-based)
8	No directly visible relationship between peak and low-flow characteristics	The F_p-term low flows and the $(1-F_p)$ term for peak flows show the water balance logic of a link between peak and low flows
9	Aggregates changes in flow regime; no directly visible link between the performance metric, rainfall (or snowmelt) and (vegetation dependent) evapotranspiration	F_p
10	Substantial empirical data bases available for comparison and meta studies	Not yet

asters is too low to intuitively influence human behaviour where short-term risk-taking benefits are attractive. Wider social pressure is needed for investment in watershed health (as a type of insurance premium) to be mainstreamed, as individuals waiting to see evidence of necessity are too late to respond. In terms of flooding risk, actions to restore or retain watershed health can be similarly justified as insurance premium. It remains to be seen whether or not the transparency of the F_p metric and its intuitive appeal are sufficient to make the case in public debate when opportunity costs of foregoing reductions in flow buffering by profitable land use are to be compensated and shared (Burt et al., 2014).

6 Conclusions

In conclusion, the F_p metric appears to allow for an efficient way of summarising complex landscape processes into a single parameter that reflects the effects of landscape management within the context of the local climate. If rainfall patterns change but the landscape does not, the resultant flow patterns may reflect a change in watershed health (van Noordwijk et al., 2016). Flow persistence is the result of rainfall persistence and the temporal delay provided by the pathway water takes through the soil and the river system. High-flow persistence indicates a reliable water supply, while minimising peak flow events. Wider tests of the F_p metric as boundary object in science–practice–policy boundary chains (Kirchhoff et al., 2015; Leimona et al., 2015) are needed. Further tests for specific case studies can clarify how changes in tree cover (deforestation, reforestation and agroforestation) in different contexts influence river-flow dynamics and F_p values. Sensitivity to specific realisations of underlying time–space rainfall patterns needs to be quantified, before changes in F_p can be attributed to changed "watershed health", rather than chance events.

Data availability. The algorithm used is freely available. Specific data used in the case studies are explained and accounted for in Part 2.

Author contributions. Meine van Noordwijk designed the method and wrote the paper, Lisa Tanika refined the empirical algorithm and handled the case study data and modelling for Part 2 (van Noordwijk et al., 2017), and Betha Lusiana contributed statistical analysis; all contributed and approved the final manuscript.

Competing interests. The authors declare that they have no conflict of interest.

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