Quantifying effects of catchments storage thresholds on step changes in the flood frequency curve

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[1] Previous work has shown that non linear catchment response related to a storage threshold may translate into a step change in the flood frequency curve. The aim of this paper is to understand the controls of this step change for catchments where runoff is generated by the saturation excess mechanism and a clear separation between a permanently saturated region and a variably saturated region with spatially uniform storage deficits exists. The magnitude of the step change is quantified by the maximum of the second derivative (curvature) of the flood peaks with respect to their return period. Sensitivity analyses with a stochastic rainfall model and a simple rainfall runoff model show that the magnitude of the step change decreases with increasing temporal variability of antecedent soil storage, and increases with increasing area of the variably saturated region. The return period where the step change occurs is very similar to the return period of the rainfall volume that is needed to exceed the storage threshold. Diagrams are presented that show the joint effects of spatial and temporal storage variability on the magnitude and return period of the step change. The diagrams are useful for assessing whether step changes in the flood frequency curve are likely to occur in catchments where the runoff generation characteristics are as examined here and the flood records are too short to indicate a step change.

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1. Introduction

[2] Flood frequency curves are typically obtained by fitting a distribution function to a record of measured flood peaks in order to estimate floods of a given return period. They are important for many engineering tasks including water resources planning and design, and risk management in floodplains. If the return period of interest is large relative to the record length, the flood peaks so estimated are associated with significant uncertainty. Understanding of the flood generation processes may assist in reducing this uncertainty [*Merz and Blöschl*, 2008a, 2008b; *Viglione et al*., 2013].

[3] Flood generation is controlled by a number of nonlinear, threshold-driven processes [*Kusumastuti et al*., 2007; *Blöschl and Zehe*, 2005; *Zehe and Sivapalan*, 2009] that depend on the catchment setting. Infiltration excess [*Horton*, 1933] refers to a sudden increase in surface runoff when the rainfall intensity exceeds the infiltration capacity of the soil. Saturation excess runoff [*Dunne and Black*, 1970] occurs when soils get saturated and any additional precipitation transforms into surface runoff. Macropore flow may lead to threshold behavior as a function of soil moisture [*Zehe and Blöschl*, 2004; *Zehe et al*., 2007]. Threshold processes may also occur in the subsurface when bedrock depressions get hydraulically connected during rainfall events as their storage capacity is exceeded causing a sudden increase in subsurface stormflow [*Tromp-van Meerveld and McDonnell*, 2006] or in shallow groundwater systems near rivers [*Derx et al*., 2010].

[4] A number of authors argue that threshold processes in runoff generation may affect the shape of the flood frequency curve. *Sivapalan et al*. [1990] suggested that, in some catchments, a transition from saturation excess runoff at low return periods to the infiltration excess runoff at high return periods may occur, resulting in a sudden increase of flood magnitudes at the return period where the transition takes place. Other authors have linked threshold processes at the catchment scale to the exceedance of storage thresholds. *Blöschl and Sivapalan* [1997] showed that a step change in the flood frequency curve may occur if rainfall exceeds a storage threshold in a large part of the catchment. For their catchment settings, *Kusumastuti et al*. [2007] and *Struthers and Sivapalan* [2007] found that, at low return periods, the flood frequency curve is controlled by tension storage depending on the field capacity of the soil which reduces the occurrence of small flood events, since rain must first bring soil moisture to a basic level before excess water for runoff generation becomes available. They suggested that at large return periods a threshold in the gravity storage or total storage capacity of the soil can cause an

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inflection point or step change in the flood frequency curve. While the above studies were all performed for hypothetical catchments, *Rogger et al*. [2012a] showed the occurrence of a step change in the flood frequency curve for two real catchments. They showed that, during heavy rainfall events, the storage threshold will be exceeded resulting in fast surface runoff in large parts of the catchment. Besides threshold processes, *Fiorentino et al*. [1985] attributed the presence of step changes in observed data to the variability in meteorological forcing and they used the two-component TCEV distribution [*Rossi et al*., 1984] to describe such step changes.

[5] There are a number of factors that influence the return period at which the step change occurs. *Kusumastuti et al*. [2007] and *Struthers and Sivapalan* [2007] varied the total storage thresholds and found that an increase in the catchments storage capacity (i.e., deeper soils) causes the step change to move from lower to higher return periods. *Struthers and Sivapalan* [2007] also showed that a variable soil depth can mask the impacts of the storage thresholds. *Blöschl and Sivapalan* [1997] showed that, with increasing catchment size, the step change may move toward larger return periods since catchment rainfall intensities may decrease with catchment scale and the soil storage deficit does not change much with scale.

[6] So far, the studies in the literature have assessed the presence and magnitudes of step changes in the flood frequency curve in a qualitative way. However, for objectively analyzing the controls, a more quantitative treatment would be useful. The aim of this paper is to examine the effects of catchment storage thresholds on step changes in the flood frequency curve in a quantitative way. We propose a new measure for the magnitude of the step change and analyze the runoff generation controls on the magnitude and return period of the step change. Runoff is assumed to be generated by the saturation excess mechanism since we are interested in catchment storage and this is the main mechanism responsible for step changes identified by *Rogger et al*. [2012b]. Furthermore, we assume a clear separation between a permanently saturated region and a variably saturated region that has spatially uniform storage deficits.

2. Methodology

2.1. Derived Flood Frequency Curve

[7] In this study, we use the derived distribution model of *Viglione et al*. [2009] and *Viglione and Blöschl* [2009] to estimate the flood frequency curve from rainfall and catchment characteristics. The model combines a statistical rainfall model with a simple, deterministic rainfall runoff model.

[8] The statistical rainfall model defines the distribution of rainfall events with intensities depending on the duration. It is a simplified version of the model of *Sivapalan et al*. [2005]. The storms are assumed to be independent. The number of storms in a year is assumed to be Poisson distributed [*Kottegoda and Rosso*, 1997] with mean m of 40. The distribution $f_{Tr}(t_r)$ of the storm duration t_r is assumed to follow a Weibull distribution with a mean of 6 h and coefficient of variation of 1.46. The rainfall intensity within a storm is assumed to be constant (block rainfall). Given that we are only interested in the saturation excess mechanism where the total storm depth controls runoff generation rather than the intensities within the storm, the assumption of block rainfall is considered to be appropriate. The rainfall intensity is assumed to be gamma distributed with moments depending on the storm duration as follows:

$$E[i|t_r] = 1.05 \cdot t_r^{0.01} \quad \text{and} \quad CV^2[i|t_r] = 1.5 \cdot t_r^{-0.55} \tag{1}$$

[9] The rainfall runoff model used in this study is a standard linear reservoir that convolves the rainfall time series. For a single storm the transformation of rainfall to runoff can be expressed through the convolution integral of the exponential unit hydrograph:

$$q(t) = \frac{r_c}{t_c} \int_0^t i(t')\exp\left(-\frac{t-t'}{t_c}\right)\mathrm{d}t' \tag{2}$$

where $i(t)$ is the rainfall input time series, $q(t)$ is the resulting runoff time series, r_c is the runoff coefficient, and t_c is the response time of the catchment. Other components, as base flow and seasonality, are not considered. Since, we assume that the rainfall intensity within the storm is constant, the flood peak simply becomes:

$$q_p = r_c \cdot i \cdot \left[1 - \exp\left(\frac{t_r}{t_c}\right)\right] \tag{3}$$

[10] The runoff coefficient is assumed to follow a beta distribution with mean δ_c and standard deviation σ_c both depending on the rainfall volume $V = i \cdot t_r$, while t_c is assumed constant and always equal to 6 h. The threshold process is represented by a switch in the mean runoff coefficient from a lower value δ_{c1} and to a higher value δ_{c2}. This switch depends on the rainfall volume $V = i \cdot t_r$ and occurs either by a sudden increase of the parameters of the probability distribution function of runoff coefficients from δ_{c1} and σ_{c1} to δ_{c2} and σ_{c2} when the rainfall volume $V = i \cdot t_r$ exceeds the threshold V^* or by a linear increase of both moments between a lower storage threshold (V_l) and a higher storage threshold (V_h). In the second case, the parameters are δ_{c1} and σ_{c1} if the rainfall volume V is smaller than V_l, δ_{c2} and σ_{c2} if the rainfall volume V is larger than V_h, with a linear transition in between. The values of the thresholds are chosen in terms of the rainfall volumes associated with given return periods.

[11] The controls on the step change of the flood frequency curve are examined in a sensitivity analysis by varying the mean runoff coefficients δ_{c1} and δ_{c2}, the standard deviations of the runoff coefficients σ_{c1} and σ_{c2} and the storage threshold V^* (or V_l and V_h). These parameters are used to represent runoff processes in catchments where runoff is generated by a saturation excess mechanism and a clear separation of two runoff contributing regions with zero and nonzero storage deficits exists. The region with zero storage deficit is assumed to be permanently saturated and contributes to all flood events. The lower mean runoff coefficient δ_{c1} can be interpreted as the spatial extent of this region relative to the total catchment area. The region

with the nonzero storage deficit is variably saturated and starts contributing to the flood events after the storage threshold is exceeded. The higher mean runoff coefficient δ_{c2} can be interpreted as the sum of the areas of these two regions relative to the total catchment area. A step increase in the runoff coefficient may occur if the storage deficit in the latter region is spatially uniform, i.e., the groundwater table is parallel to the soil surface. The storage thresholds V^* (or V_l and V_h) represent the rainfall volume that is needed to exceed the storage deficit in the variably saturated region.

[12] Antecedent storage conditions will vary between events. In the sensitivity analysis, we represent this variability by the standard deviation of the runoff coefficients σ_{c1} and σ_{c2}. This variability can be interpreted as the expansion and contraction of the permanently and variably saturated regions. This would occur if fast lateral movement of groundwater recharge takes place.

[13] Given the dependency of the storm intensity on the storm duration, and the dependency of the runoff coefficient on the rainfall volume, the probability for a given flood peak discharge Y to be less than or equal to q_p is:

$$F_Y(q_p) = \int_0^\infty \int_0^\infty F_{Rc|I,Tr}\left(\frac{q_p}{i}\left[1-\exp\left(\frac{t_r}{t_c}\right)\right]^{-1}\middle| i, tr\right) \cdot f_{I|Tr}(i|t_r) \cdot f_{Tr}(t_r)\,\mathrm{d}i\,\mathrm{d}t_r \tag{4}$$

where $F_{Rc|I,Tr}$ is the cumulative distribution of runoff coefficients r_c conditioned on the rainfall volume ($V = i \cdot t_r$). Finally, the cumulative distribution function of the annual flood peaks is given by $F_Q(q_p) = \exp\{-m[1-F_Y(q_p)]\}$ which can also be expressed in terms of return period as $T_Q = \{1-F_Q(q_p)\}^{-1}$. The curve relating q_p to T_Q is the flood frequency curve. All flood frequency curves shown in this paper have been normalized by their median $Q = q_p/\tilde{q}_p$ (Figure 1a).

2.2. Characterization of the Step Change

[14] We propose a new measure for the magnitude of the step change. The first and second derivatives of the flood frequency curve are calculated as follows (assuming n equispaced values of normalized discharge Q_i with $i = 1, \ldots, n$):

$$\left.\frac{\mathrm{d}Q}{\mathrm{dlog}_{10}T}\right|_i = \frac{Q_{i+1}-Q_{i-1}}{\log_{10}T(Q_{i+1}) - \log_{10}T(Q_{i-1})} = \mathrm{slope}_i \tag{5}$$

$$\left.\frac{\mathrm{d}^2Q}{\mathrm{d}(\log_{10}T)^2}\right|_i = \mathrm{slope}_i \cdot \frac{\mathrm{slope}_{i+1} - \mathrm{slope}_{i-1}}{Q_{i+1}-Q_{i-1}} = \mathrm{curvature}_i \tag{6}$$

[15] The first derivative is the slope of the flood frequency curve and the second derivative is denoted here as curvature for simplicity. Figure 1 illustrates the slope and curvature for a flood frequency curve with a step change. The value of slope (Figure 1b) represents the increase of the flood peak magnitude for an order of magnitude increase in return period. The value of curvature (Figure 1c) represents the rate of change of the slope. Since the step change can be thought of as a sudden increase in the slope of the flood frequency curve, we propose to quantify

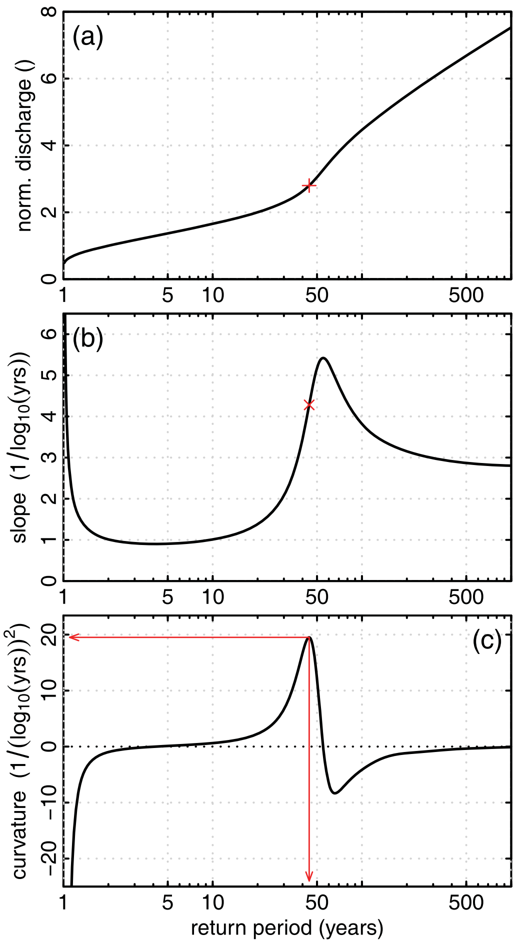

Figure 1. Characterization of a step change: (a) flood frequency curve with step change (growth curve, i.e., discharge normalized by its median); (b) slope of the curve dQ = dlog$_{10}T$; and (c) curvature of the curve d^2Q = d(log$_{10}T$)^2. The magnitude and return period of the step change are indicated by the red arrows.

it by the maximum positive value of the curvature. The value of max(curvature$_i$) calculated through equation (6) will be referred to as the magnitude of the step change, while the return period associated with it will be referred to as the return period of the step change. As indicated in Figure 1c, the step change of the flood frequency curve has magnitude 20 log$_{10}$(y)$^{-2}$, i.e., the change in slope is so strong that it increases by a factor of 20 for an order of magnitude increase in the return period, and the return period of the step change is 44 years in this example.

3. Results

[16] The following controls on the step change of the flood frequency curve are examined for catchments where a clear separation of two regions with zero and nonzero, spatially uniform, storage deficits exists:

[17] 1. Temporal variability of antecedent soil storage

[18] 2. Average size of the variably saturated region within the catchment

[19] 3. Shape of the spatial distribution of storage deficits

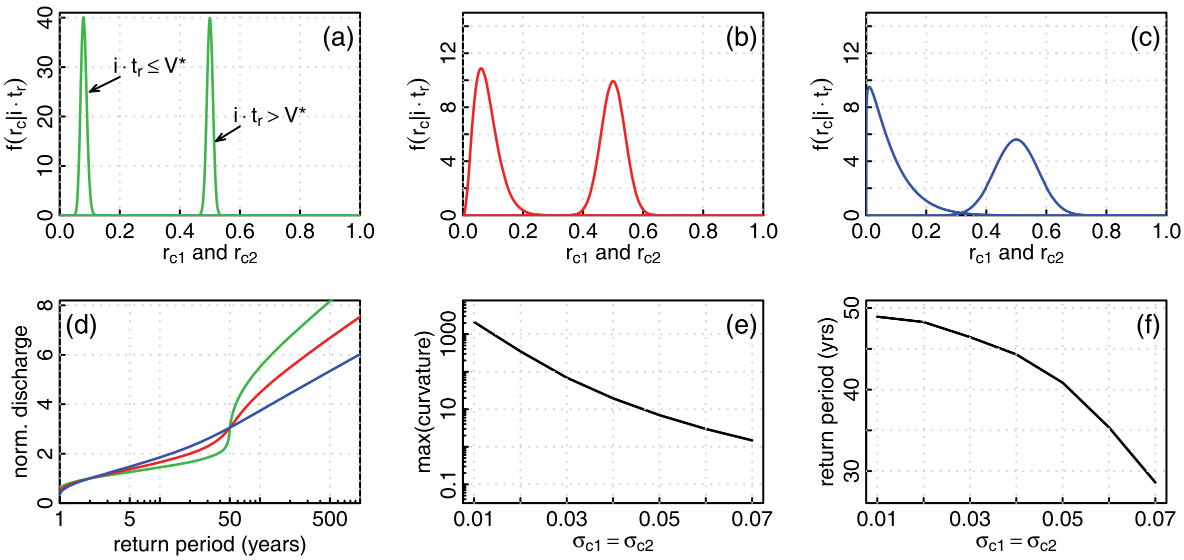

Figure 2. Step changes for different cases of temporal variability of antecedent soil storage as represented by $\sigma_{c1}$and σ_{c2}. (top) Density functions of runoff coefficients (r_{c1} and r_{c2}) for $\delta_{c1} = 0.08$, $\delta_{c2} = 0.5$, and (a) $\sigma_{c1} = \sigma_{c2} = 0.01$, (b) $\sigma_{c1} = \sigma_{c2} = 0.04$, and (c) $\sigma_{c1} = \sigma_{c2} = 0.07$. (d) Flood frequency curves for cases a, b, and c. (e) Magnitude of the step change for varying $\sigma_{c1} = \sigma_{c2}$. (f) Return period of the step change for varying $\sigma_{c1} = \sigma_{c2}$.

[20] 4. Magnitude of the soil storage deficits
[21] 5. Combined controls

3.1. Temporal Variability of Antecedent Soil Storage

[22] The extent of the saturated areas in a catchment depends on antecedent rainfall, evaporation and drainage. For some catchments, variations of the antecedent conditions between flood events may be large, e.g., because of strong seasonal cycles, while for other catchments variations are small. In the sensitivity analysis, we investigate the effect of changes in the standard deviations σ_{c1} and σ_{c2} of the runoff coefficients which can be interpreted as the degree to which the permanently saturated and the variably saturated regions expand or contract. The permanently saturated region is the region that contributes to runoff even during the smallest events (i.e., is independent of the magnitude of the event rainfall), but it may vary between events dependent on the antecedent rainfall. The contribution of the variably saturated region depends both on the antecedent rainfall and the event rainfall as the latter controls the switch when it starts to contribute.

[23] It should be noticed here that we are assuming an independence of the antecedent soil moisture conditions from the hydro-climatic conditions. In real catchments, rainfall characteristics and antecedent soil moisture conditions are interlinked, but this relationship is not straightforward. In this study we keep the simplifying assumption of independence that is used in other derived flood frequency studies such as *Sivapalan et al*. [2005], *Viglione et al*. [2009], and *Viglione and Blöschl* [2009].

[24] For the analysis, the mean runoff coefficient for the whole catchment was chosen as $\delta_{c1} = 0.08$ when only the permanently saturated region contributes to runoff and as $\delta_{c2} = 0.5$ when the variably saturated region contributes as well (assuming that the soil storage deficit is spatially uniform). Three cases for the (temporal) standard deviation are analyzed: (a) $\sigma_{c1} = \sigma_{c2} = 0.01$, (b) $\sigma_{c1} = \sigma_{c2} = 0.04$, and (c) $\sigma_{c1} = \sigma_{c2} = 0.07$. The distributions of the runoff coefficients for the three cases and for rainfall volumes smaller and larger than the threshold volume V^* are shown in Figures 2a, 2b, and 2c, respectively. In these cases, a clear separation between the permanently saturated and the variably saturated regions is assumed to exist with a threshold volume (V^*) that is chosen as the rainfall volume with a return period of 50 years ($V^* = 118.50$ mm). We define the case shown in Figure 2b as the reference scenario for the further analysis.

[25] Figure 2d shows the effect of the temporal variability of antecedent soil storage. As would be expected, with an increasing variability, the step change in the flood frequency curve becomes less pronounced due to the grater variability in the extent of saturated areas. For the very small variability case ($\sigma_{c1} = \sigma_{c2} = 0.01$, green line), the maximum curvature that represents the magnitude of the step change is as large as 2100, indicating a strong step change in the flood frequency curve. For a variability of $\sigma_{c1} = \sigma\ _{2} = 0.07$ (blue line) the maximum curvature decreases by three orders of magnitude to 1.5, indicating a very small step change. The change in variability also causes a change in the return period of the step change (Figure 2f). With increasing variability the step change moves from a return period of 49 years to a return period of 29 years, although for the latter case the step change of the flood frequency curve is hardly identifiable by visual inspection.

3.2. Average Size of the Variably Saturated Region Within the Catchment

[26] In some catchment there may be extensive regions that never contribute to flood runoff, such as deep debris fans or highly fractured rocks dominated by deep groundwater flow, while in other catchments this region may be small. There will therefore be differences between

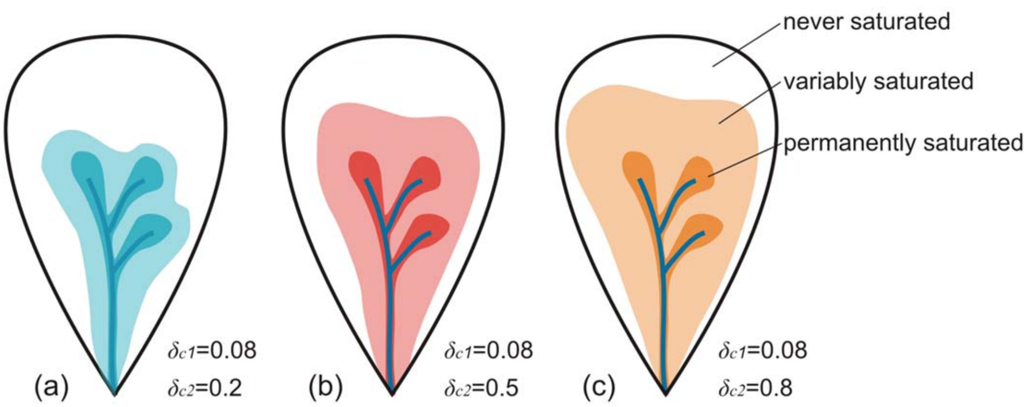

Figure 3. Schematic illustration of an increasing average size of the variably saturated region as represented by an increase in the higher mean runoff coefficient δ_{c2} for the cases where $\delta_{c1} = 0.08$ and (a) $\delta_{c2} = 0.2$, (b) $\delta_{c2} = 0.5$, and (c) $\delta_{c2} = 0.8$. Never saturated areas refer to areas with very large storage capacities that never contribute to the flood runoff, such as debris fans.

catchments in the total area that may contribute to flood runoff during extreme events. In the previous case, the average size of the two runoff contributing regions was assumed to be constant. Here, we investigate the effect of changing the average size of the variably saturated region on the step change.

[27] The average size of the variably saturated region (plus that of the permanently saturated region) is represented in the model by the mean of the larger runoff coefficient δ_{c2}. Three cases are examined, (a) $\delta_{c2} = 0.2$, (b) $\delta_{c2} = 0.5$, and (c) $\delta_{c2} = 0.8$, and the mean of the smaller runoff coefficient is set to $\delta_{c1} = 0.08$. The three cases are schematically illustrated in Figures 3a–3c. Regions denoted as never saturated have a very large storage capacity and never contribute to flood runoff (such as debris areas). The distributions of the runoff coefficients for rainfall volumes smaller and larger than V^* are shown in Figures 4a, 4b, and 4c. The case presented in Figures 3b and 4b shows the reference scenario. The standard deviations are set to $\sigma_{c1} = \sigma_{c2} = 0.04$ in all cases. We assume, again, a clear separation of the two runoff contributing regions and use the same threshold volume of $V^* = 118.50$ mm as in the previous case.

[28] The step changes are presented in Figure 4d and show that, with increasing average size of the variably saturated region, the step change becomes much more pronounced. For a small average size ($\delta_{c2} = 0.2$, pale blue line) the maximum curvature is very small and no step change is apparent from the graph, while for a large average size of $\delta_{c2} = 0.8$ (orange line) the maximum curvature is three orders of magnitude larger with a value of 930, and the step change is very clear in the graph. Figure 4f shows that the return period of the step change increases from 35 to 50

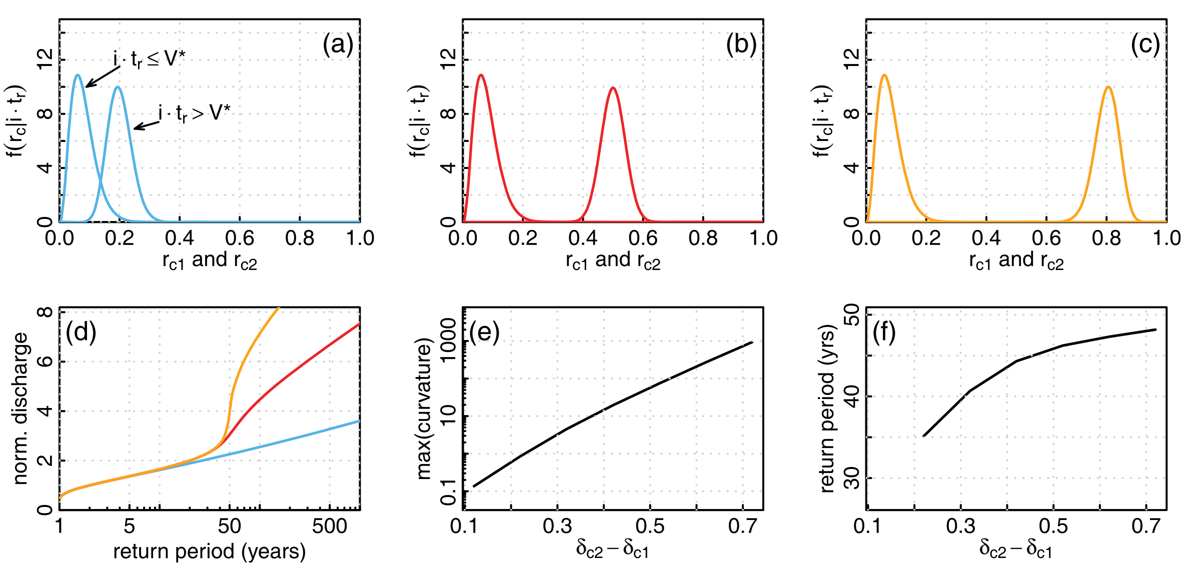

Figure 4. Step changes for different cases of average size of the variably saturated region as represented by the mean of the higher runoff coefficient δ_{c2}. (top) Density functions of runoff coefficients (r_{c1} and r_{c2}) for $\delta_{c1} = 0.08$ and (a) $\delta_{c2} = 0.2$, (b) $\delta_{c2} = 0.5$, and (c) $\delta_{c2} = 0.8$. (d) Flood frequency curves for cases a, b, and c. (e) Magnitude of the step change for varying $\delta_{c2} - \delta_{c1}$. (f) Return period of the step change for varying $\delta_{c2} - \delta_{c1}$.

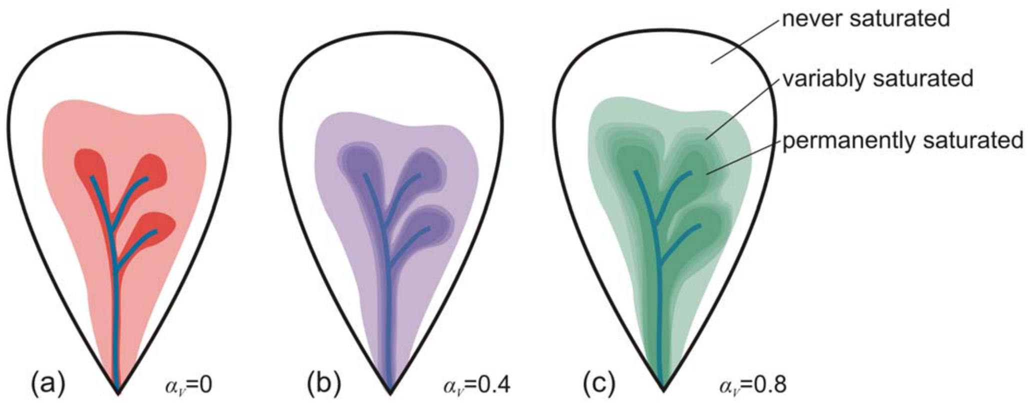

Figure 5. Schematic illustration of a gradual increase in storage deficit or flood contributing area from the permanently saturated region to the variably saturated region as represented by α_V for the cases (a) $\alpha_V = 0$ (reference scenario), (b) $\alpha_V = 0.4$, and (c) $\alpha_V = 0.8$. Never saturated areas refer to areas with very large storage capacities that never contribute to the flood runoff, such as debris fans.

years with increasing average size of the variably saturated region.

3.3. Shape of the Spatial Distribution of Storage Deficits

[29] So far we assumed that the storage deficit in the variably saturation region is spatially uniform so, once event rainfall exceeds the storage deficit, the entire region starts to contribute to runoff at the same time. In real catchments the storage deficit may not be exactly uniform and a transition between the permanently saturated areas (zero storage deficit) and the variably saturated region may occur.

[30] In the sensitivity analysis, we therefore investigate the effect of a gradual increase in storage deficit from the permanently saturated region to the variably saturated region which implies a gradual increase of the contributing area as the rainfall depth increases. This situation is schematically illustrated in Figure 5. Figure 5a shows the reference scenario, where the two regions are clearly separated, while there is a gradual increase of flood contributing area in the cases shown in Figures 5b and 5c. A gradual increase in the storage deficit is represented in the model by a gradual increase in the mean runoff coefficients with rainfall volume between a lower storage threshold (V_l) and a higher storage threshold (V_h). We assume a symmetric transition around $V^* = 118.5$ mm with $V_l = V^*(1 - \alpha_V/2)$ and $V_h = V^*(1 + \alpha_V/2)$. Three cases are analyzed: (a) $V_l = V_h = 118.5$ mm (Figure 6a, reference scenario with

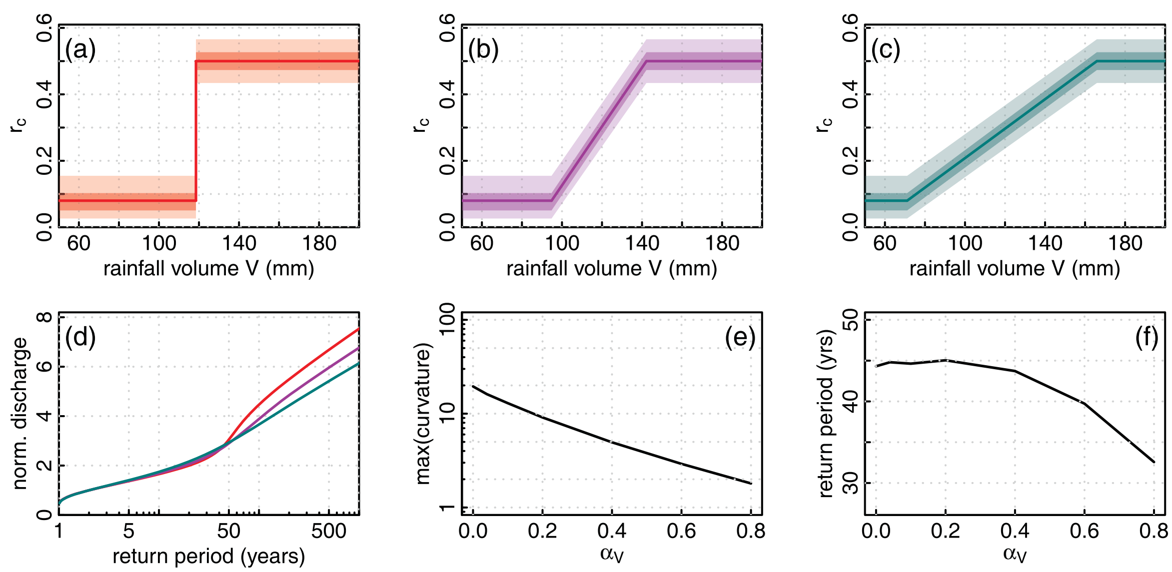

Figure 6. Step changes for different cases of a gradual increase in storage deficit or flood contributing area from the permanently saturated region to the variably saturated region as represented by α_V. The shape of the spatial distribution of storage capacities is given by the relationship between the mean runoff coefficient δ_c and the rainfall volume V for (a) $\alpha_V = 0$ (reference scenario), (b) $\alpha_V = 0.4$, and (c) $\alpha_V = 0.8$. (Darker and brighter shaded areas refer to the 50 and 90% confidence intervals, respectively.) (d) Flood frequency curves for cases a, b, and c. (e) Magnitude of the step change for varying α_V. (f) Return period of the step change for varying α_V.

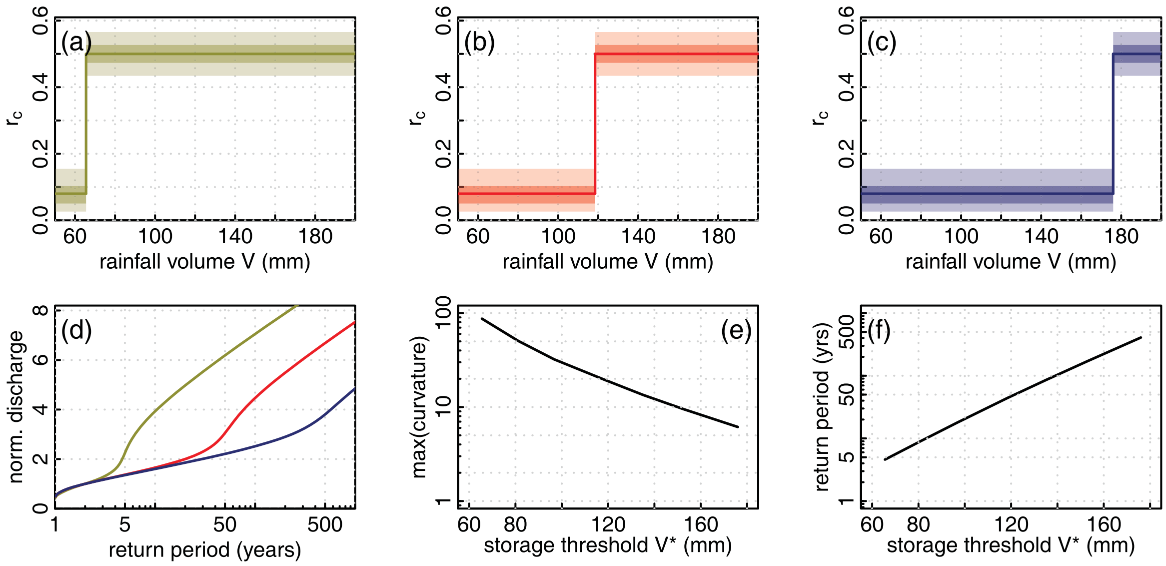

Figure 7. Step changes for different cases of magnitude of storage capacity of the variably saturated region as represented by the storage threshold V^*. The shape of the spatial distribution of storage capacities is given by the relationship between the mean runoff coefficient δ_c and the rainfall volume V for (a) threshold $V^* = 65.54$ mm, (b) $V^* = 118.5$ mm, and (c) $V^* = 176.0$ mm, respectively, corresponding to rainfall volumes with return periods of 5, 50, and 500 years. (Darker and brighter shaded areas refer to the 50 and 90% confidence intervals, respectively.) (d) Flood frequency curves for cases a, b, and c. (e) Magnitude of the step change for varying V^*. (f) Return period of the step change for varying V^*.

$\alpha_V = 0$); (b) $V_l = 94.8$ mm and $V_h = 142.2$ mm ($\alpha_V = 0.4$, Figure 6b), and (c) $V_l = 71.1$ mm and $V_h = 165.9$ mm ($\alpha_V = 0.8$, Figure 6c). The standard deviation of the runoff coefficients is assumed to be equal in all cases with $\sigma_{c1} = \sigma_{c2} = 0.04$. The mean values of the lower and higher runoff coefficients are chosen as $\delta_{c1} = 0.08$ and $\delta_{c2} = 0.5$ in all cases.

[31] In the reference scenario (red line in Figure 6d) with a uniform storage deficit ($\alpha_V = 0$) and an abrupt change in storage deficit, the maximum curvature is 20, indicating a gradual increase step change in the flood frequency curve. For the case with a strong gradual increase in storage deficit from the permanently saturated region to the variably saturated region ($\alpha_V = 0.8$, dark green line), the maximum curvature decreases to 1.8 and the step change is almost smoothened out. There is also a slight decrease in the return period of the step change from 44 years in the reference scenario to 33 years for the smoothest case (Figure 6f).

3.4. Magnitude of the Soil Storage Deficit

[32] The magnitude of the soil storage deficit represents the volume of water per area that can be stored during an event before runoff occurs. It depends on the soil depth and the depth to the groundwater level, whatever is smaller, and is therefore related to both soil evolution processes and climate through recharge and subsurface flow processes at the seasonal scale. If the soil storage deficit of the variably saturated region is small, it will be exceeded by small rainfall volumes and the region will contribute to small flood events. If the soil storage deficit is large, the region will only contribute to big flood events.

[33] The magnitude of the soil storage deficit is represented in the model by the storage threshold V^*, again assuming a clear separation between two runoff contributing regions. Three storage thresholds are examined: (a) $V^* = 65.54$ mm, (b) $V^* = 118.5$ mm and $V^* = 176.0$ mm which correspond to rainfall volumes with return periods of 5, 50, and 500 years, respectively. The dependence of the mean runoff coefficient (δ_c) on storage deficit volume is shown in Figures 7a, 7b, and 7c. Figure 7b refers to the reference scenario. The standard deviation of the runoff coefficients is set to $\sigma_{c1} = \sigma_{c2} = 0.04$ in all cases. The lower and higher mean runoff coefficients are set to $\delta_{c1} = 0.08$ and $\delta_{c2} = 0.5$, respectively, as in the previous cases.

[34] Figure 7d indicates that the step change moves from lower to higher return periods with increasing soil storage deficit of the variably saturated region, as would be expected. While for a small soil storage deficit or storage threshold ($V^* = 65.54$ mm, green line) the step change occurs at a return period of 4.6 years, a threshold of $V^* = 176.0$ mm (dark blue line) causes the step change to shift to a return period of 410 years (Figure 7f). These return periods approximately correspond to the return periods of the rainfall volumes of 5 and 500 years used to define these thresholds. With increasing storage thresholds, the step change becomes less pronounced, as reflected in the decreasing maximum curvatures (Figure 7e) as there are fewer events and longer duration events that actually exceed the threshold.

3.5. Combined Controls

[35] So far the effects of changes of a single control on the step changes in the flood frequency curve were analyzed but joint effects of a number of controls are also of interest. By combining cases (3.1) and (3.2) the joint effects of changing the temporal variability of antecedent soil storage (represented by $\sigma_{c1} = \sigma_{c2}$) and the average size of the variably saturated region (represented by δ_{c2}-δ_{c1}) are

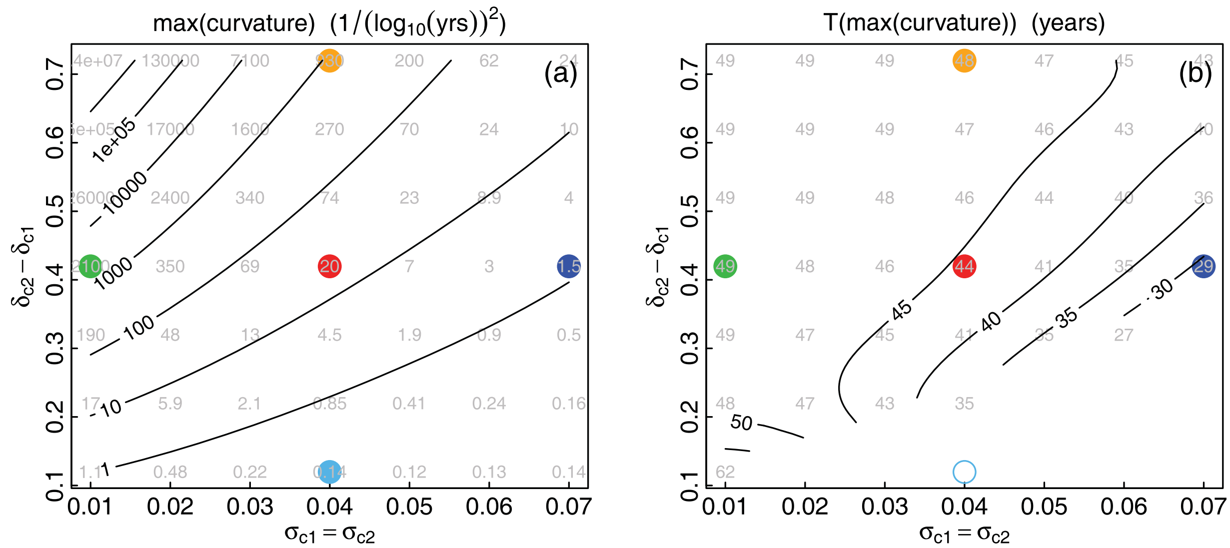

Figure 8. Joint effects of the temporal variability of antecedent soil storage (represented by $\sigma_{c1} = \sigma_{c2}$) and the average size of the variably saturated region (represented by $\delta_{c2} - \delta_{c1}$) for a storage threshold $V^* = 118.5$ mm corresponding to a rainfall volume of 50 years return period. (a) Magnitude of the step change (measured as the maximum curvature of the flood frequency curve). (b) Return period of the step change. Colored points refer to the cases shown in Figures 2 and 4.

examined in Figure 8a. The lines in Figure 8a indicate constant values of maximum curvature. The colored points indicate cases (3.1) and (3.2). The magnitude of the step changes increases from bottom right to the top left of the diagram indicating that it increases with increasing average size of the variably saturated region and decreases with increasing temporal variability of antecedent soil storage. The former has a stronger effect than the latter. For a small variability in antecedent soil storage and a small average size of the variably saturated region the values of maximum curvature are 1 and indicate no step change, while for a large variability in antecedent soil storage conditions and a large average size of the variably saturated region, maximum curvatures are around 20 and similar to the reference scenario with a clear step change.

[36] Figure 8b shows the return period at which the step change occurs. It is more strongly influenced by the variability of antecedent soil storage than by the average size of the variably saturated region. For a small variability, the return period stays almost the same with values around 50 years regardless of the average size of the variably saturated region, while for increasing variability the return period slightly decreases to approximately 30 years. The blank areas in the diagram (bottom right) refer to parameters where the step change is so small that the return period cannot be identified.

[37] In a next step the joint effects of cases (3.3) and (3.4) are examined that refer to the shape of the spatial distribution of storage deficit (represented by α_V) and the magnitude of the soil storage deficit in the variably saturated region, V^*. Storage deficit is expressed in terms of the return period of the associated rainfall volume. In this case (Figure 9a), the largest maximum curvatures (and therefore step changes) occur in the bottom left of the diagram which represents a combination of a sudden change in the shape of the spatial distribution of storage deficit ($\alpha_V = 0$) with a small soil storage deficit in the variably saturated region. No step change is to be expected if there is a gradual transition of the soil storage deficit between the two regions and the storage deficit in the variably saturated region is large (top right of the figure). The diagram also shows that the magnitude of the step change depends on both controls to similar degrees.

[38] The return period at which the step change occurs, on the other hand, is solely dependent on the soil storage deficit in the variably saturated region (expressed here as the return period of the associated rainfall volumes, Figure 9b). An increase in the return period of the threshold volume causes a similar increase in the return period of the step change (see also Figure 7f), while the shape of the spatial distribution of storage deficit has almost no influence on the return period. The return period of the threshold volume thus basically determines the return period of the step change.

4. Discussion

[39] This paper presents the effects of temporal and spatial storage variability on the magnitude and return period of a step change in the flood frequency curve. While step changes can occur due to a number of different processes such as the switch from saturation excess to infiltration excess [*Sivapalan et al*., 1990], the exceedance of storage thresholds [*Blöschl and Sivapalan*, 1997; *Kusumastuti et al*., 2007; and *Struthers and Sivapalan*, 2007] or different meteorological forcing [*Fiorentino et al*., 1985], this study only focuses on step changes caused by a saturation excess mechanism [*Dunne and Black*, 1970]. The results consequently apply to catchments where runoff is generated by the saturation excess mechanism and a clear separation between a permanently saturated region and a variably

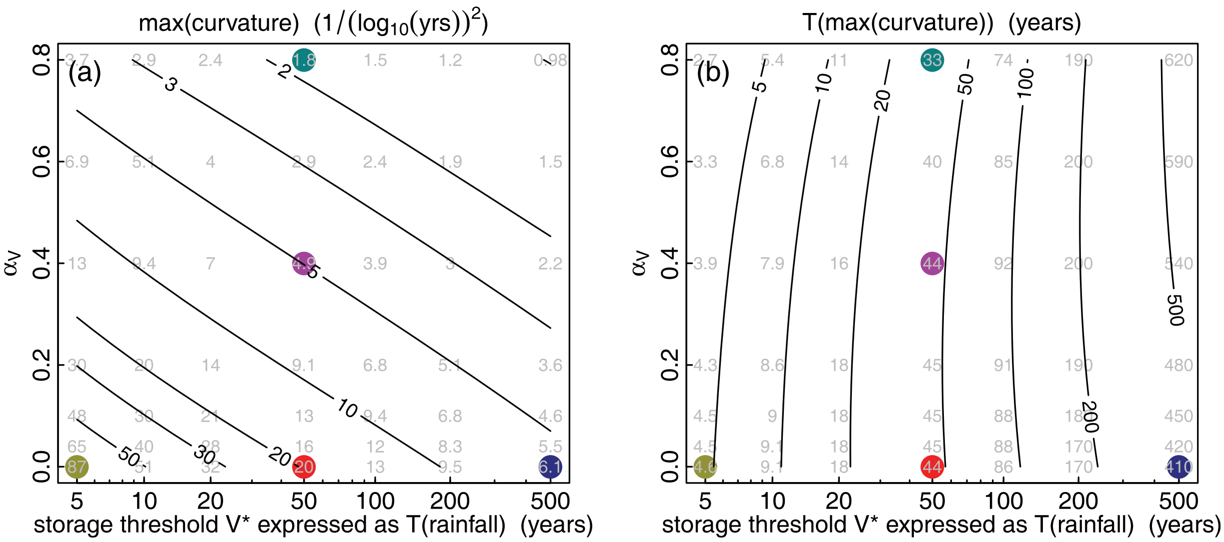

Figure 9. Joint effects of the shape of the spatial distribution of storage deficits (represented by α_V) and the magnitude of the soil storage deficit (represented by V^* and expressed as return periods of rainfall volumes) for mean runoff coefficient $\delta_{c1} = 0.08$ and $\delta_{c2} = 0.5$ and standard deviations $\sigma_{c1} = \sigma_{c2} = 0.04$. (a) Magnitude of the step change (measured as the maximum curvature of the flood frequency curve). (b) Return period of the step change. Colored points refer to the cases shown in Figures 6 and 7.

saturated region with spatially uniform storage deficits exists.

[40] The temporal controls examined are the temporal variability of antecedent soil storage in the two regions related to the expansion and contraction of the regions between events. As the variability increases, the magnitude of the step change tends to zero, while a small variability gives rise to a pronounced step change. For real catchments, the impact of this control on the step change depends on the seasonal variability of soil moisture and on the seasonality of floods which are both controlled by the climatic conditions of that catchment. If the antecedent soil storage conditions are similar throughout the year, i.e., the saturated areas do not change much, a pronounced step change would be expected regardless of the seasonality of floods. However, the more common case is that the soil storage conditions do vary because of summer evaporation and/or rainfall seasonality. Often, large saturated areas occur in the winter, sometimes enhanced by the presence of snow, and small (or no) saturated areas occur in the summer [*Western et al*., 1998]. In this case, the occurrence of a step change may depend on the seasonality of the floods [*Parajka et al*., 2010]. If the floods occur throughout the year, no step change may occur even though a clear separation between a permanently saturated region and a variably saturated region exists in the catchment. However, if floods mainly occur in one season (either summer or winter), a pronounced step change may occur as the antecedent soil storage is similar for all floods. More generally speaking, when the soil moisture and rainfall seasonalities are in phase [*Sivapalan et al*., 2005], the variability of antecedent storage conditions will be small and a clear step change in the flood frequency curve is more likely to occur.

[41] The spatial controls examined here include the average size of the variably saturated region. The size will depend both on the climate and the subsurface of the catchment. In the sensitivity analysis presented here we represent the average size of the variably saturated region by the average runoff coefficient once a threshold rainfall volume is exceeded. As the size of that region increases, the magnitude of the step change (expressed in terms of the maximum curvatures) increases by a few orders of magnitude. Such an abrupt transition has, for example, been observed by *Rogger et al*. [2012a, 2012b] in a small catchment in western Austria where the runoff coefficients jumped from 0.08 to 0.50 as the event magnitudes increased, causing a marked step change in the flood frequency curve. However, if there is a more gradual transition of the storage deficit from the permanently saturated region to the variably saturated region, the magnitude of the step change decreases. While *Struthers and Sivapalan* [2007] showed, in a similar but qualitative way, that the spatial variability of soil depths will reduce the magnitude of the step change, we quantify this effect for spatially variable soil deficits assuming a linear transition of the mean runoff coefficients between a lower and a higher threshold. For a rather large spatial variability in storage deficits ($\alpha_V = 0.8$), the step change in the flood frequency curve essentially disappears (maximum curvature of 1). Besides the spatial variability of the soil depths, the size of the permanently saturated region may also change depending on the groundwater-surface water interactions [*Derx et al*., 2010, J. Derx et al., Effects of river bank restoration during floods on the removal of dissolved organic carbon by soil passage: A scenario analysis, submitted to *Journal of Hydrology*, 2013].

[42] The spatial distribution of storage deficits therefore plays a crucial role in the occurrence of step changes in the flood frequency curve. This distribution depends on the soil depths in the catchment and the depths to the groundwater table. A rather uniform distribution of storage deficit is

conducive to the occurrence of a step change, but one might argue that this is unlikely to occur in real catchments due to the heterogeneity of processes that influence soil evolution and groundwater flow [*Fleckenstein et al*., 2006; *Derx et al*., 2010]. The distribution of soil depths in a catchment depends on the balance between soil production and soil erosion processes [*Heimsath et al*, 1997]. Soil production and soil depth have been shown to be inversely related to hillslope curvature [*Heimsath et al*., 1999; *Dietrich et al*., 1995] so that accumulation processes dominate in concave valleys where soils are deep, while convex ridges serve as sources for the sediment and are covered, if at all, with very thin soil layers [*Dietrich et al*., 1995]. The depth to the groundwater table, on the other hand, tends to exhibit inverse patterns. It is exactly in the concave valleys that the depth to the groundwater table tends to be smallest or saturation areas occur when it intersects the land surface, e.g., in the lowland areas of the river Danube [*Derx et al*., 2010]. On the hillslopes, which are essentially the domain considered as variably saturated areas here, the shallow water tables are often considered to be parallel to the land surface in modeling catchment response, e.g., in Topmodel [e.g., *Lamb et al*., 1998]. The rationale of this assumption are the similarities between landscape evolution and groundwater flow as both are driven by gravity. While the exact shape of the groundwater table will depend on the local characteristics of the subsurface, there are examples where the shallow groundwater table has indeed been shown to be parallel to the land surface [e.g., *Lamb et al*., 1998]. In such cases, the storage deficit will be spatially uniform within a spatially constrained region, as assumed in this paper. There is also evidence of the existence of clearly distinguishable hydrological zones within a catchment determined by their topographical characteristics [*Savenije*, 2010], which are related to the coevolution of the landscape, soils and vegetation [*Thompson et al*., 2011]. These may lend additional credence to the existence of regions with spatially uniform storage deficit.

[43] Another interesting aspect analyzed in this paper is the return period at which the step change occurs. *Kusumastuti et al*. [2007] showed qualitatively that increasing soil depths will cause the step change to shift from lower to higher return periods. We quantify this effect in this paper. The results suggest that the return period of the step change is very similar to the return period of the rainfall volume that is needed to exceed the storage deficit in the variably saturated region. Other controls, such as the exact shape of the spatial distribution of the storage deficit, are less important (Figure 9). This finding has important implications for the interpretation of flood frequency curves from short flood records, in particular if outliers are present [*Rogger et al*., 2012a, 2012b]. If a step change occurs at return periods larger than that covered by the flood data, fitting a smooth distribution function will underestimate extreme floods [*Rogger et al*., 2012a, 2012b]. It is therefore essential to know (a) whether a step change is to be expected and (b) at what return period it will occur.

[44] As discussed in this paper, the occurrence of a step change is related to whether storage deficit is spatially uniform or not, as well as the seasonality of soil moisture and the floods. The return period where the step change may occur is essentially a function of the magnitude of the storage deficit in the variably saturated region. Estimates of the soil storage deficit can be obtained by field mapping methods [e.g., *Markart et al*., 2004; *Rogger et al*., 2012a, 2012b] using soil observations, vegetation indicators and other storage indicators. These may be supported by sprinkling experiments in the field. The concept of 'reading the landscape' [*Blöschl et al*., 2013] may assist in the assessments. However, the return period at which a step change occurs does not only depend on the storage deficit but also on the rainfall volume. One would therefore expect major differences in the return periods between climate regions. As one moves from wet to dry climates, the rainfall volumes tend to decrease, so the return period of the step change will increase. In wet regions, the step change may occur at return periods of a few years, so will be fully covered by the flood data. In contrast, in dry regions this type of step change may occur at return periods of hundreds of years, so will never appear in the flood data with the exception of possible outliers. This is consistent with typically higher coefficients of variation and skewnesses of annual floods in dry parts of the world than in wet parts [*Merz and Blöschl*, 2009; *Blöschl et al*., 2013]. Also, as the catchment size decreases, spatially uniform storage deficits and therefore step changes are more likely to occur.

[45] The analysis presented in this paper is based on the simplified assumption that the rainfall intensities within a rainfall event are constant. Whether this assumption holds, depends on the response time t_c of a catchment. For small t_c the impact of rainfall patterns might be large, while it decreases for larger t_c. To understand the influence of the response time on the step change, additional simulations for the reference scenario ($\delta_{c1} = 0.08$, $\delta_{c2} = 0.5$, $\sigma_{c1} = \sigma_{c2} = 0.04$ and $V^* = 118.5$ mm) with t_c varying from 1 to 48 h were performed. The results of the analysis (Figure10) show that an increasing t_c results in an increase in the magnitude of the step change (Figures 10a and 10d) and in the related maximum curvature (Figure 10b), while almost no effect on the return period of the step change (Figure 10c) can be observed. The reason for this behavior is that for large t_c the highest floods peaks occur for long rainfall events (Figure 10e) with low intensities and low variability (Figure 10f). Hence the flood peak volume is mainly dependent on the distribution of the runoff coefficient r_c that exhibits an abrupt jump at the threshold volume V^* (see Figure 6a). For smaller t_c, shorter rainfall events (Figure 10e) with higher intensities and a higher variability (Figure 10f) are relevant. In this case the flood peak volume is controlled by the distribution of the runoff coefficient r_c, but also by the intensity and duration of the rainfall events. The larger variability of the rainfall events results in a less pronounced step change (Figures 10a and 10d). This is especially the case for the rainfall events below the threshold that have small volumes. An increase in the response time of a catchment results, therefore, in a more pronounced step change. For smaller t_c it should be noted that rainfall time patterns and runoff processes such as infiltration excess should also be taken into account to obtain representative results.

[46] An important aspect is the applicability of the results of this study to real catchments. In this study, we are analyzing step changes related to a saturation excess mechanism. Which runoff process is dominating during a flood

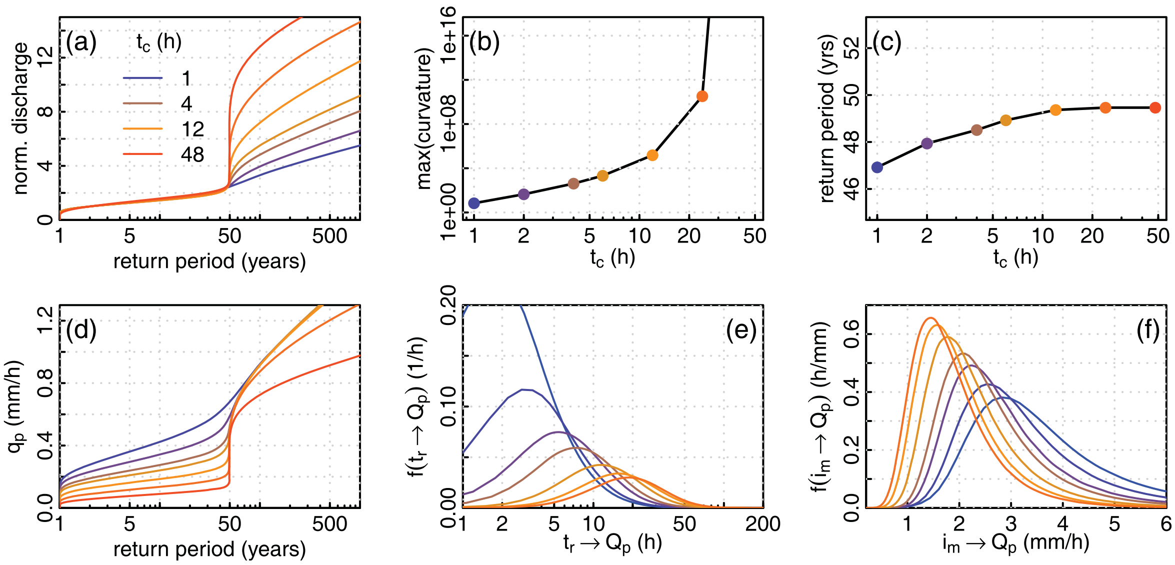

Figure 10. Step changes for varying response times t_c from 1 to 48 h: (a) related flood frequency curves, (b) magnitudes of the step change, (c) return periods of the step change, and (d) flood peak volumes. The distributions of the durations of rainfall events t_r that provoke the largest floods (q_p) for varying t_c is given in (e) and the related distributions of rainfall intensities im in (f).

event, strongly depends, besides catchment properties, on the nature of the rainfall event that triggers the flood. Frontal rainfall events with long durations and low intensities are more likely to result in saturation excess flow compared to convective storms with short durations and strong intensities that may cause infiltration excess flow. Frontal events are the trigger of synoptic flood events which are an important flood type in many regions of Austria [*Merz and Blöschl*, 2003] and have a strong impact on catchments with large response times. In the two catchments with a step change examined by *Rogger et al*. [2012], the largest flood events were synoptic floods and the dominating runoff process was saturation excess flow. One would therefore expect the findings of this study to be applicable to regions that are dominated by synoptic flood events. The impact of dominating runoff processes are also related to the question of catchment scale. *Rogger et al*. [2012b] state that they expect the step change behavior to be important for rather small catchments, i.e., for catchments in the size of tenths to a few hundreds of square kilometers as analyzed in their study. In larger catchments more averaging of the spatial hydrological variability occurs [*Sivapalan*, 2003] which is likely to mask the presence of a step change. In very small catchments, i.e., in the size of a few tenths square kilometers and less, convective storms may have a stronger impact. If such catchments are dominated by flash floods triggered by localized convective storms the infiltration excess mechanism is likely to be the dominating runoff mechanism. Step changes related to an infiltration excess flow have not been analyzed in this study. The findings of this study therefore apply to medium and small sized catchments that are dominated by synoptic flood events and have a large response time.

[47] From the perspective of design flood estimation the step changes are critically important. Design flood estimation is typically performed by fitting a distribution function to a flood record to estimate the design flood of interest. Whether a step change is taken into account strongly depends on the length of the flood record. In case the flood record is much longer than the return period of the step change, the step change will be clearly reproduced by the data and a mixed distribution function (e.g., the TCEV distribution) [*Rossi et al*., 1984] can be used to describe the data and estimate the design flood. If the length of the flood record is in the same order of magnitude as the return period of step change or shorter, the step change will only be indicated by a few data points or not at all. Resorting to purely statistical extrapolation may then lead to a strong underestimation of the design flood. In these cases it is important to expand the information beyond the flood peak sample as proposed by *Merz and Blöschl* [2008a, 2008b] in their flood frequency hydrology framework. If the step change is caused by a saturation excess mechanism due to the local hydrogeologic conditions pooling data from a regional analysis is not an option, but field visits during dry and wet catchment conditions may be conducted to map the extent of saturated areas. Available hydrogeologic information can be used to understand the catchments storage properties as shown by *Rogger et al*. [2012b] who provided a road map of how storage capacities can be estimated from field surveys. By taking the topography and vegetation into account clearly distinguishable hydrological zones may be identifiable which help characterize dominant runoff processes in the catchment. Along with the results from this study, these additional data may help ascertain whether a step change is to be expected or not and, if so, at what return period.

5. Conclusions

[48] The aim of this paper was to examine the effects of catchment storage thresholds on step changes in the flood frequency curve in a quantitative way. Runoff was assumed to be generated by the saturation excess mechanism, and a

clear separation between a permanently saturated region and a variably saturated region with spatially uniform storage deficits was assumed to exist. We proposed the maximum of the second derivative (maximum curvature) of the flood peaks with respect to their return periods as a new measure for the magnitude of the step change. A sensitivity analysis with a stochastic rainfall model and a simple rainfall runoff model gives the following results:

[49] 1. The magnitude of the step change depends on the temporal and spatial soil storage variability.

[50] 2. The magnitude of the step change decreases with increasing temporal variability of the antecedent soil storage. The step change vanishes for large variabilities.

[51] 3. The magnitude of the step change increases with increasing average size of the variably saturated region. The increase is two orders of magnitude as the average size of the variably saturated region changes from 10 to 70% of the catchment area.

[52] 4. The magnitude of the step change decreases as the spatial distribution of the storage deficit in the variably saturated region becomes more variable.

[53] 5. The return period at which the step change occurs is very similar to the return period of the rainfall volume that is needed to exceed the storage threshold. Other controls, such as the exact shape of the spatial distribution of the storage deficit, are less important.

[54] The occurrence of a step change in the flood frequency curve has important implications for the estimation of extreme floods. If flood records are short and the step change is not represented by the data, fitting a smooth distribution function to the flood data will underestimate the flood discharges. It is therefore suggested to ascertain whether a step change in the flood frequency curve is to be expected or not. The results of this study provide guidance on assessing the likely occurrence and the return period of such as step change.

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References

Blöschl, G., and M. Sivapalan (1997), Process controls on regional flood frequency: Coefficient of variation and basin scale, *Water Resour. Res.*, *33*(12), 2967–2980.

Blöschl, G., and E. Zehe (2005), On hydrological predictability, *Hydrol. Processes*, *19*, 3923–3929.

Blöschl, G., M. Sivapalan, T. Wagener, A. Viglione, and H. Savenije (2013), *Runoff Prediction in Ungauged Basins: Synthesis Across Processes, Places and Scales*, Cambridge University Press, Cambridge, UK, isbn: 9781107028180.

Derx, J., A. P. Blaschke, and G. Blöschl (2010), Three dimensional flow patterns at the stream-aquifer interface: A case study at the Danube, *Adv. Water Resour.*, *33*(11), 1375–1387.

Dietrich, W. E., R. Reiss, M.-L. Hsu, and D. R. Montgomery (1995), A process-based model for colluvial soil depth and shallow landsliding using digital elevation data, *Hydrol. Processes*, *9*, 383–400.

Dunne, T., and R. D. Black (1970), An experimental investigation of runoff production in permeable soils, *Water Resour. Res.*, *6*(2), 478–490, doi:10.1029/WR006i002p00478.

Fiorentino, M., P. Versace, and F. Rossi (1985), Regional flood frequency estimation using the two-component extreme value distribution, *Hydrol. Sci. J.*, *30*, 51–64.

Fleckenstein, J., R. Niswonger, and G. Fogg (2006), River-aquifer interactions, geologic heterogeneity, and low-flow management, *Ground Water*, *44*, 837–852.

Heimsath, A. M., W. E. Dietrich, K. Nishiizumi, and R. C. Finkel (1997), The soil production function and landscape equilibrium, *Nature*, *388* (6640), 358–361.

Heimsath, A. M., W. E. Dietrich, K. Nishiizumi, and R. C. Finkel (1999), Cosmogenic nuclides, topography, and the spatial variation of soil depth, *Geomorphology*, *271*, 2151–2172.

Horton, R. E. (1933), The role of infiltration in the hydrological cycle, *Eos Trans. AGU*, *14*, 446.

Kottegoda, N. T., and R. Rosso (1997), *Statistics, Probability, and Reliability for Civil and Environmental Engineers*, 735 pp., McGraw-Hill.

Kusumastuti, D. I., I. Struthers, M. Sivapalan, and D. A. Reynolds (2007), Threshold effects in catchment storm response and the occurrence and magnitude of flood events: Implications for flood frequency, *Hydrol. Earth Syst. Sci.*, *11*, 1515–1528, doi:10.5194/hess-11-1515-2007.

Lamb, R., K. Beven, and S. Myrabø (1998), Use of spatially distributed water table observations to constrain uncertainty in a rainfall–runoff model, *Adv. Water Resour.*, *22*, 305–317.

Markart, G., B. Kohl, B. Sotier, T. Schauer, G. Bunza, and R. Stern (2004), Provisorische Geländeanleitung zur Abschätzung des Oberflächenabflussbeiwertes auf alpinen Boden-/Vegetationseinheiten bei konvektiven Starkregen (Version 1.0) (A simple code of practice for assessment of surface runoff coefficients for alpine soil-/vegetation units in torrential rain (Version 1.0)), *Bundesforschungszentrum für Wald, BFW-Dokumentation, 3*, Austria.

Merz, R., and G. Blöschl (2003), A process typology of regional floods, *Water Resour. Res.*, *39*(12), 1340, doi:10.1029/2002WR001952.

Merz, R., and G. Blöschl (2008a), Flood frequency hydrology: 1. Temporal, spatial, and causal expansion of information, *Water Resour. Res.*, *44*, W08432, doi:10.1029/2007WR006744.

Merz, R., and G. Blöschl (2008b), Flood frequency hydrology: 2. Combining data evidence, *Water Resour. Res.*, *44*, W08433, doi:10.1029/2007WR006745.

Merz, R., and G. Blöschl (2009), Process controls on the statistical flood moments: A data based analysis, *Hydrol. Processes*, *23*(5) 675–696.

Parajka, J., et al. (2010), Seasonal characteristics of flood regimes across the Alpine–Carpathian range, *J. Hydrol.*, *394*(1–2), 78–89.

Rogger, M., B. Kohl, H. Pirkl, A. Viglione, J. Komma, R. Kirnbauer, R. Merz, and G. Blöschl (2012a), Runoff models and flood frequency statistics for design flood estimation in Austria: Do they tell a consistent story?, *J. Hydrol.*, *456–457*, 30–43, doi:10.1016/j.jhydrol.2012.05.068.

Rogger, M., H. Pirkl, A. Viglione, J. Komma, B. Kohl, R. Kirnbauer, R. Merz, and G. Blöschl (2012b), Step changes in the flood frequency curve: Process controls, *Water Resour. Res.*, *48*, W05544, doi:10.1029/2011WR011187.

Rossi, F., M. Fiorentino, and P. Versace (1984), Two-component extreme value distribution for flood frequency analysis, *Water Resour. Res.*, *20*(7), 847-856, doi:10.1029/WR020i007p00847.

Savenije, H. H. G. (2010), HESS opinions ''Topography driven conceptual modelling (FLEX-Topo),'' *Hydrol. Earth Syst. Sci.*, *14*, 2681–2692, doi:10.5194/hess-14-2681-2010.

Sivapalan, M. (2003), Process complexity at hillslope scale, process simplicity at the watershed scale: Is there a connection?, *Hydrol. Processes*, *17*(5), 1037–1041.

Sivapalan, M., E. F. Wood, and K. Beven (1990), On hydrologic similarity: 3. A dimensionless flood frequency model using a generalized geomorphologic unit hydrograph and partial area runoff generation, *Water Resour. Res.*, *26*(1), 43–58, doi:10.1029/89WR01579.

Sivapalan, M., G. Blöschl, R. Merz, and D. Gutknecht (2005), Linking flood frequency to long-term water balance incorporating effects of seasonality, *Water Resour. Res.*, *41*, W06012, doi:10.1029/2004WR003439.

Struthers, I., and M. Sivapalan (2007), A conceptual investigation of process controls upon flood frequency: Role of thresholds, *Hydrol. Earth Syst. Sci.*, *11*, 1405–1416, doi:10.5194/hess-11-1405-2007.

Thompson, S. E., C. J. Harman, A. G. Konings, M. Sivapalan, A. Neal, and P. A. Troch (2011), Comparative hydrology across AmeriFlux sites: The variable roles of climate, vegetation, and groundwater, *Water Resour. Res.*, *47*, W00J07, doi:10.1029/2010WR009797.

Tromp-van Meerveld, H. J., and J. J. McDonnell (2006), Threshold relations in subsurface stormflow: 2. The fill and spill hypothesis, *Water Resour. Res.*, *42*, W02411, doi:10.1029/2004WR003800.

Viglione, A. and G. Blöschl (2009), On the role of storm duration in the mapping of rainfall to flood return periods. *Hydrol. Earth Syst. Sci.*, *13*(2), 205–216.

Viglione, A., R. Merz, and G. Blöschl (2009), On the role of the runoff coefficient in the mapping of rainfall to flood return periods. *Hydrol. Earth Syst. Sci.*, *13*(5), 577–593.
Viglione, A., R. Merz, J. S. Salinas, and G. Blöschl (2013), Flood frequency hydrology: 3. A Bayesian analysis, *Water Resour. Res.*, *49*, 675–692, doi:10.1029/2011WR010782.
Western, A. W., G. Blöschl, and R. B. Grayson (1998), How well do indicator variograms capture the spatial connectivity of soil moisture?, *Hydrol. Processes*, *12*, 1851–1868.
Zehe, E., and G. Blöschl (2004), Predictability of hydrologic response at the plot and catchment scale: Role of initial conditions, *Water Resour. Res.*, *40*, W10202, doi:10.1029/2003WR002869.
Zehe, E., and M. Sivapalan (2009), Threshold behaviour in hydrological systems as (human) geo-ecosystems: Manifestations, controls, implications, *Hydrol. Earth Syst. Sci.*, *13*(7), 1273–1297.
Zehe, E., H. Elsenbeer, F. Lindenmaier, K. Schulz, and G. Blöschl (2007), Patterns of predictability in hydrological threshold systems, *Water Resour. Res.*, *43*, W07434, doi:10.1029/2006WR005589.