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Threshold behavior in hydrological systems and geo-ecosystems: manifestations, controls and implications for predictability

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Abstract

The aim of this paper is to provide evidence that the dynamics of hydrological systems and geo-ecosystems is often influenced by threshold behavior at a variety of space and time scales. Based on well known characteristics of elementary threshold phe-
nomena we suggest criteria for detecting threshold behavior in hydrological systems. The most important one is intermittence of phenomena, i.e. the rapid switching of related state variables/fluxes from zero to finite values, or existence of behavior regimes where the same process/response appears qualitatively differently at the macroscopic level. From the literature we present several examples for intermittent hydrological phe-
nomena, ranging from overland flow generation in different landscapes, including the effects of hydrophobicity, to soil water flow occurring in the matrix continuum or via preferential pathways, including the case of cracking soils, nonlinear subsurface stormflow response of hillslopes during severe rainfall events, and long-term catchment flood-
ing responses. Since threshold phenomena are often associated with environmental hazards such as floods, soil erosion, and contamination of shallow groundwater re-
sources, we discuss common difficulties that complicate predictions of whether or not they might even occur. Predicting the onset of threshold phenomena requires a thor-
ough understanding of the underlying controls. Through examples we illustrate that threshold behavior in hydrological systems can manifest at (a) the process level, (b) the response level, and (c) the functional level, and explain that the complexity of the underlying controls and of the interacting phenomena that determine threshold behav-
ior become increasingly complex at the higher levels. Finally we provide evidence from field observations and model predictions that show that within an "unstable range" of system states "close" to a threshold, it is difficult to predict whether or not the system will switch behavior, for instance, as a result of the expected rainfall forcing. The term close, in this respect, depends on the expected (rainfall) forcing and the accuracy of our data/knowledge on the macroscopic state of the system.

1 Introduction

We all know threshold behavior from our experience in preparing our morning tea. Thermal energy enters the water across the bottom of the tea pot. In the early phase molecular heat diffusion is sufficient for transporting the energy upwards and to dissipate the energy against the vertical temperature gradient. When the stove temperature, and therefore the vertical temperature gradient increase, suddenly convection cells form, because these permit much more efficient energy transport and dissipation than molecular diffusion. If the vertical temperature gradient increases further above a second threshold, suddenly turbulence occurs (Haken, 1983). The water starts boiling and tea is ready.

Although simple, this example sheds light on many important aspects of threshold behavior: there are different dynamic modes/ regimes for energy mixing and they differ according to their efficiency in dissipating the incoming energy. These regimes are not stable, they are intermittent, i.e., when temperature of the stove is reduced the system switches back from turbulent mixing to convective mixing with convection cells. The occurrence of these dynamic regimes depends, therefore, on the strength of the external forcing (here energy flow into the system), internal system properties (here the molecular coefficient for heat diffusion) and the current system state. Occurrence of these regimes may be predicted on the basis of a macroscopic state variable (here temperature). Strikingly, the individual water molecules at the microscopic level do not know or control anything, neither about turbulence nor about eddies, they simply move. The different quality of the dynamics is only appropriate at the macroscopic level. As suggested by Haken (1983) this is a result of self-organization. Consequently, it is very difficult to estimate the magnitude of the threshold values from analyzing the behavior of the individual water molecules at the microscopic level (Haken, 1983; it is possible for a laser, though). They have to be derived empirically at the macroscopic level.

Generally threshold behavior can be deemed as an extreme form of nonlinear dynamics. The response of a nonlinear system is in general no longer linearly propor-

tional to the external forcing (e.g., Sivapalan et al., 2002). When doubling the forcing the system response can, depending on the degree of nonlinearity and the current system state, grow more rapidly in the case of an amplifying system or grow less rapidly in the case of an inhibiting system (Haken, 1983; Rundle et al., 2006). Hence, the nature

5 of the system response depends crucially on both the magnitude of the forcing and the actual system state. One suitable definition for threshold behavior that includes our example of the tea pot was suggested by Alley et al. (2003). Threshold behavior manifests as a rapid and sudden change in macroscopic dynamics to another dynamic regime or dynamic equilibrium. Fast, in this respect, means much faster than the typi-

10 cal time scales of the internal system dynamics and of the external forcing. As shown in the tea pot example, a threshold is triggered when a macroscopic state variable or a ratio of macroscopic of state variables or the energy flow into the system rises above or drops below an empirical threshold value. We might observe this sudden change in behavior either in the form of an activity/triggering event as in the case of earthquakes

15 (a hot spot in space as suggested by Rundle et al., 2006), or in the form of strongly increased reaction rates in biogeochemical systems (a hot moment in time as suggested by McCain et al., 2003), or as a fast qualitative change in macroscopic dynamics as in the case of a laser (Haken, 1983). In the other regime, i.e., once the threshold has been crossed, processes/responses are often considerably faster/slower and the en-

20 ergy dissipation is often much more/less efficient. As a consequence the system often behaves, at the macroscopic level, totally differently, as in the case of turbulence (e.g., as we saw in the tea pot example).

Elementary threshold phenomena can be deemed as phase transitions between the liquid, solid and gas phases of an element. Further examples of elementary threshold

25 phenomena are the switching from laminar to turbulent flow (Arya, 2003), or the switch from normal light to laser light (Haken, 1983). Turbulence in open channel flow, for example, occurs when the Reynolds number (the ratio of acceleration forces to viscous forces) increases above a certain threshold, which depends on the fluid and the geometry of the channel. Turbulent flow represents a different quality of open channel flow

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when compared to laminar flow, just as turbulent mixing is a different quality of mixing when compared to molecular diffusion. Laser light is a new, highly coherent quality of light resulting from stimulated light emission from atoms and molecules. The latter is in contrast to the normal instantaneous emission of light and occurs when the rate of energy input (pumping rate) into a gas laser exceeds a threshold value.

The present study deals with threshold behavior in hydrological systems or geo-ecosystems. Various authors suggest that threshold behavior is also of key importance for understanding and predicting either the dynamics and stability of our climate system (Pitman and Stouffer, 2006, Claussen, 1999), or the dynamics and resilience of geo-ecosystems (Saco et al., 2006; Wilcox et al., 2003), or the dynamics of environmental flows in hydrologic systems (Woods and Sivapalan, 1999; Zehe et al., 2007). As discussed in these studies and will be shown in our study the presence of thresholds can drastically impact our ability to make predictions either at the level of an individual process, or at the level of the response of larger units such as hillslopes or catchments, or even at the level of the long-term hydrologic functioning a complete geo-ecosystem. Detecting and understanding threshold behavior in complex hydrological systems or geo-ecosystems is, therefore, a significant and important challenge to hydrological science, especially for predictions in the context of global change. The main difficulty, when compared to elementary threshold phenomena, is that the underlying physical controls in geo-ecosystems are far more complex due to possible multiple feedbacks between biotic and abiotic components (Saco et al., 2006), and due to the fact that internal states and dynamics are often very difficult to observe (Rundle et al., 2006, Zehe et al., 2007). Hence, threshold behavior in geo-ecosystems becomes much more difficult to detect and to predict in comparison to elementary thresholds which can be predicted on the basis of dimensionless quantities such as the Reynolds number, or even more fundamentally on the basis of physical-chemistry/ thermodynamics or quantum statistics in the case of phase transitions. In response, this paper is organized around the following key questions:

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- how can we detect threshold behavior and how does it manifest in hydrologic systems scales?

- can we conceptualize different forms of threshold behavior and understand their first order controls?

5 - is there any reason/"advantage" for the system to switch from one regime to another dynamic regime?

- how can we model threshold behavior and how accurately can we observe the patterns of the controlling state variables and forcing?

- what are the implications of threshold behavior on the predictability of individual
10 hydrological processes, or hydrological responses of hillslopes and catchments?

Following Rundle et al. (2006) and Zehe and Blöschl (2004) we suggest that intermittence of phenomena/processes in space and/or time is a good indicator for threshold behavior in hydrological systems (although it might not be a sufficient condition). In this respect threshold behavior is frequently discussed as influencing local, hillslope and
15 catchment scale runoff generation processes (Deeks et al., 2004; Zehe et al., 2005; Tromp van Meerveld and McDonnell, 2006; Tani, 1997). Soil moisture and/or rainfall intensity and depth are deemed as first order controls in this case. Subsurface flow and transport of contaminants in field soils may be observed at least in two different regimes: as a slow form in the soil matrix continuum and as a faster form in pref-
20 erential flow pathways when the necessary conditions have been created (Vogel et al., 2005; Zehe and Blöschl, 2004; Flury, 1996; McGrath et al., 2007; Ritsema et al., 1998). Threshold behavior is also discussed in the context of the long-term development of soil structures, soil patterns and landforms (Phillips, 2006; Blöschl and Zehe, 2005). Saco et al. (2006) suggested that threshold behavior and feedbacks between
25 vegetation patterns and redistribution of abiotic resources determine the resilience and even stability of geo-ecosystems in semi-arid regions. Taking another example, the "climate system" consists of the atmosphere, ocean, marine and terrestrial biosphere,

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cryosphere and lithosphere. The interactions and feedbacks between these components may trigger threshold behavior, as discussed through several examples by Pitman and Stouffer (2006). The most prominent example discussed by these authors is the enhanced freshwater input into the North Sea as a result of increased precipitation in the mid-latitudes, and their net effect on the Thermo-Haline Circulation (THC) in the North Atlantic ocean and consequently on the global climate system.

To summarize, threshold behavior occurs in our earth system at various spatio-temporal scales. However, the forms and possibly the process controls might be very different in each case. In Section 2 we will further elaborate on different forms of threshold behavior in hydrological systems and geo-ecosystems, and will propose a hierarchy with respect to the complexity of the underlying controls and also in respect of scale (Fig. 1).

In the above discussion we have suggested intermittence of processes/phenomena in hydrologic systems and geo-ecosystems as a potential indicator for detecting threshold behavior. Predictions of intermittent phenomena represent, in general, a two-level mixed discrete-continuous problem: level A is to predict whether the phenomenon/process will occur or not, level B is then to predict the strength of the phenomenon/process, if it does indeed occur. For elementary threshold systems it is well known that the difficulty of level A predictions (does it occur or not?) depends on the current state of the system. This is nicely explained by our tea pot example. At the beginning, when the water from the water tap is at 12°C, say, then we are sure that increasing the water temperature by just 1°C will neither induce formation of convections cells nor of turbulence. The dynamic regime of energy mixing is "stable". When we increase the water temperature to a state which is, say for example, 1.05°C below the threshold needed for formatting convection cells, a level A prediction is not that simple any more. It depends largely on the accuracy/error of our temperature measurement. If the error is 0.01, we can safely state that increasing the temperature by 1°C will surely not induce formation of convection cells. However, if the error of our temperature observation is higher at 0.1°C, say, we are not sure anymore about what

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will happen. Small, non-observable differences within the error range of our measurements can determine whether convection cells will form or not. Hence, we may not be able to predict with confidence whether or not the energy dissipation will switch from the molecular regime to the convective regime. Repeated trials of the experiment will,

5   therefore, lead to a strong scatter in the observed macroscopic dynamics. We then say that we have a low predictability. If we increase the temperature to a value which is clearly and unambiguously above the threshold for eddy formation, the regime of energy dissipation remain becomes stable again, unless and until we come close to the next threshold value. "Close" in this context is in respect of the error range of our

10 temperature measurements. In general, level A predictions – whether a phenomenon/ process occurs or not, are most difficult and uncertain if the "system state" is in the vicinity of a threshold. This range of states is also called the "unstable range" (Haken, 1983; Zehe and Blöschl, 2004; Zehe et al., 2007).

   Coming back to hydrological science, the difficulties of level A predictions are often
15 discussed in the context of intermittent hydrological phenomena. Does subsurface storm flow during a rainfall event occur or not, and what are the underlying controls (Weiler et al., 2005, Tromp van Meerveld and McDonnell, 2006)? Does preferential transport of an agrochemical into the subsoil occur or not, and what are the underlying controls (Vogel et al., 2005; Flury, 1996; McGrath et al., 2007; Ritsema et al., 1998)?

20 Does a hillslope respond to a moderate rainfall event with surface runoff or not, and if so, will this trigger erosion or not (Scherer, 2008)? Therefore, as in the tea pot example, hydrological processes/phenomena may also be deemed to be threshold driven, if they exhibit intermittence characteristics, and their occurrence is found to be harder to predict over a certain unstable range of system states (depending on the

25 forcing) and quite easy to predict in other system states. This discussion pertains to intermittent behaviors and low predictability of level A predictions, and are known as elementary threshold phenomena (Haken, 1983). However, this statement does not by any means imply that thresholds are behind the lack of predictability in all hydrological systems or behaviors. Indeed, the difficulties with respect to thresholds are in addition

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to more common predictability problems in hydrological and environmental systems.

The width of the unstable range of system states where level A predictions are difficult is – as discussed for the tea pot example – strongly dependent on the question of how accurately we can observe the local values and spatio-temporal patterns of critical state variables – such as soil moisture – and of the external rainfall forcing (Zehe et al., 2007; Rundle et al., 2006). Observations of hydrological state variables, processes and rainfall forcing continue to rely strongly on point measurements. These measurements are therefore non-exhaustive due to the heterogeneity of natural soils and spatio-temporal variability of rainfall, and to the fact that measurement and process scales are quite often incompatible (Blöschl and Sivapalan, 1995). Consequently, patterns of state variables may only be insufficiently characterized, especially at larger scales. In Sect. 3 we will elaborate on how the limited accuracy of observations hampers level A predictions in hydrological systems whenever threshold behavior is involved. In Section 4 we will introduce alternative potential modeling approaches that can more appropriately predict threshold behavior, taking into consideration the issues of observability and predictability associated with threshold driven systems. In Sect. 5 we will draw several conclusions about the role of threshold behavior in hydrologic systems and will close with remarks about further research required in this area.

2 Different levels of threshold behavior and their manifestations in hydrological systems

The emphasis of the previous section was on presenting indicators for detecting threshold behavior in hydrological systems and geo-ecosystems. We proposed that intermittence as well as the lack of predictability of their occurrence over a range of system states (e.g., unstable range) as possible criteria for this purpose. In the context of elementary threshold phenomena, we showed further that the switch between different dynamic regimes can lead to more efficient mixing/dissipation of energy, provided the incoming energy fluxes exceed a certain threshold level.

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However, threshold behavior in hydrological systems can occur over a range of scales, across different phenomena, as well as at different levels of complexity. Threshold behavior can occur locally at the level of an individual process, for instance infiltration, but also in respect of the overall response of larger hydrological unit, such as the runoff response of hillslopes and catchments. The latter can be composed of several interacting processes. In this section we want to highlight the differences between these different forms of threshold behavior, along with the complexity of the underlying controls that trigger these and the spatio-temporal scales of their manifestation. In this respect, we suggest that threshold behavior in hydrological systems occurs at different levels of complexity, as schematically illustrated in Fig. 1:

– Threshold behavior at the process level, which often manifests at the point scale of 0.1–1 m, and short temporal scale (e.g., event sale): the simplest form of threshold behavior is when individual hydrological processes are intermittent. They either emerge/vanish, such as overland flow, or show drastic changes, as when infiltration switches from slow water flow in the matrix continuum to faster flow through preferential pathways (Vogel et al., 2005; Zehe and Flühler, 2001b; Flury, 1996; McGrath et al., 2007; Ritsema et al., 1998). The threshold character of these processes is strongly determined by the nonlinear interaction of local (soil) structures/ properties with the soil moisture as well as rainfall intensity and depth. In the following we will use the term "threshold process" as synonymous with threshold behavior at the process level.

– Threshold behavior at the response level, such as at the hillslope and catchment scale, and the temporal scale is the event scale: the fact that a process threshold has been triggered at the point scale, for instance that overland flow is locally generated during a rainfall event, is necessary but not sufficient for a larger hydrological unit (such as a hillslope or a whole catchment) to respond in the same way. The response of a hillslope or even catchment is governed by the interplay of several processes in space and time: local runoff generation is only one

part of this mosaic, and runoff concentration and re-infiltration are of equal importance. Whether a hillslope responds with surface or subsurface runoff, or whether a catchment responds with a flood, is determined by an interplay of local threshold processes with the hillslope/catchment scale pattern of soil properties and structures, related patterns of state variables (i.e., soil moisture), along with patterns of precipitation intensity and depth (Lindenmaier et al., 2005; Zehe et al., 2005; Tromp-van Meerveld and McDonnell, 2006; Tani, 1997). The underlying controls are therefore more complex than at the process level, since they involve satisfying stronger conditions. One of the conditions, as will be explained in Sect. 2.2, is connectivity of patterns, which describes how various parts of the landscape are connected to each other and to the "catchment outlet". Another important condition is spatial correlation or covariation of patterns of initial states, as they relate to forcings and key landscape parameters: for instance if the rainfall pattern is such that areas of high rainfall intensities coincide with areas of high macroporosity, fast preferential infiltration will occur in the entire area and the topsoil will respond with leaching of chemical contaminants (for example see: Zehe and Blöschl, 2004). In the following we will use the term "threshold response" to be synonymous with threshold behavior at the response level.

– Threshold behavior at the functional level, which may occur at the hillslope and catchment scales, whereas the temporal scale is long-term (e.g., multi-year): by hydrological functioning we refer here to the way larger units such as hillslopes and catchments respond to the rainfall and other climate forcing (e.g., population of events) in a statistical sense. Common statistical indicators that can quantify hydrological functioning are the flow duration curve, the flood frequency curve, the monthly runoff regime, the water balance and – if available – patterns of sediment yield and hydro-chemistry. These indices encapsulate threshold behavior at the process and response levels but in a statistical manner, including how patterns of topographic variables, soil structure, and vegetation – much of it of biological origin or biologically mediated – translate the energy input (precipitation and ra-

diation) into hydrological responses. We argue that hydrological functioning of hillslopes/catchments is well defined and can be considered stationary if these indices are also stationary. Stationarity of hydrological functioning of a catchment – defined here at the human time scale – is probably determined by the same multiple, and possibly biologically mediated, feedbacks between abiotic and biotic components that also determine resilience/ stability of the geo-ecosystem present within a catchment (Phillips 2006; Saco et al., 2006). Non-stationarity of these indices and therefore of hydrological functioning can either be in response to external climate change, or due to persistent, substantial changes in those geo-ecosystem properties (e.g., human induced land use changes) that determine the process and response spectrum in this catchment. Examples could be a substantial and persistent change in vegetation dynamics due to a change in habitat conditions and related feedbacks on local process thresholds and hydrological responses (Schröder, 2006), or changes in biologically formed soil structures and related feedbacks on the thresholds for runoff generation and preferential flow, or changes in topography due to landslides and related feedbacks on sediment yields. Hence, a functional threshold is either triggered by persistent change of climate, changes in internal process and response thresholds, or a combination thereof. Either way, it could manifest in the non-stationarity of certain associated functional indices and may manifest itself only on a long time scale. Once again, the underlying control is more complex compared to the behavior at the process and also response levels, since it involves complex, long-term feedbacks between biotic and abiotic geo-ecosystem components, as well as substantial disturbances that go beyond the resilience characteristics of the geo-ecosystem.

Sects. 2.1 to 2.3 will further elaborate these ideas and will provide instructive examples from field observations to illustrate these concepts.

2.1 Threshold behavior at the process level

2.1.1 Threshold behavior of overland flow generation process

Threshold behavior at the process level occurs when phenomena are either intermittent and the related state variables/ fluxes switch from zero to non-zero values (Blöschl and Zehe, 2005; McGrath et al., 2007) or if the process happens to cross two different macroscopic regimes. Overland flow is the most prominent example of intermittent phenomena, whereas infiltration is a good example of the second type. Overland flow initiation is determined by the interaction of the soil hydraulic properties, soil moisture and the rainfall forcing, arising from the fact that both the water retention curve and the hydraulic conductivity of the soil are strongly nonlinear functions of the soil water content (Dingman, 1994). Hence, infiltration and overland flow generation exhibit highly nonlinear behavior even in homogeneous soils. Depending on the landscape, the saturation excess or infiltration excess runoff generation mechanism may dominate overland flow generation. Saturation excess runoff is a capacity controlled threshold process. It occurs when the entire pore volume within a soil column is saturated with water from the incoming precipitation (Dunne, 1970; Dunne and Black, 1978), and dominates runoff generation when either shallow impermeable bedrock or impermeable soil layer (e.g. gley-soils) underlies a top soil layer of high infiltrability, or when there is a water table at shallow depth. Infiltration excess runoff generation is an intensity controlled threshold process, and occurs when precipitation intensity exceeds the local infiltrability of the soil (Horton, 1933). The latter is strongly governed by the unsaturated hydraulic conductivity of the soil which, as mentioned above, is a nonlinear function of the soil water content (Dingman, 1994). Hence, both mechanisms of overland flow generation are soil moisture controlled threshold processes (Zehe and Blöschl, 2004).

The threshold associated with the onset of Hortonian overland flow is strongly increased by the presence of soil structures/ preferential pathways that can be either semi-permanent, or temporary as in the case of cracking soils. In the case of swelling

and cracking soils soil structures exhibit threshold behavior themselves since they emerge and then vanish (Návar et al., 2002; Lindenmaier et al., 2006). Hydrophobicity is another soil property that exhibits threshold behavior as it can be activated, and increases in strength, during dry conditions and but disappears when the soil is
5 wetted sufficiently (e.g. DeJonge et al., 1999; Doerr and Thomas, 2000). As shown in Sects. 2.1.2 and 2.2.3 this threshold behavior of soil properties themselves may lead to an even more distinct threshold behavior at the process level – for overland flow generation – as well as at the response level for catchment scale rainfall-runoff response.

2.1.2   Threshold behavior of the infiltration process: preferential flow and transport in
10   soils

Most natural cohesive soils offer a richness of connected structures such as root channels and earthworm burrows. In hydrology, such structures are traditionally called macropores or "preferential" flow paths, because water may flow along these larger pathways without being affected by capillary forces (Beven and Germann, 1982). In
15 the case of high precipitation amounts or wet soil moisture such macropores may become active and allow much higher infiltration rates, e.g., 10–100 times faster flow and transport of solutes into the subsoil than when compared to flow in the continuum of the soil micro-pore matrix.
   Flury et al. (1994, 1995) were the first to develop effective methods to observe and
20 visualize infiltration and flow patterns in heterogeneous field soils by using dye tracer techniques. Within the last 12 years numerous studies have followed on from this pioneering work and have provided further evidence that macropore flow in structured soils is indeed the rule rather than the exception (Zachmann et al., 1987; Roth et al., 1991; Kladivko et al., 1991; Richter et al., 1996; Mohanty et al., 1998; Flury et al., 1995;
25 Flury, 1996; Zehe and Flühler, 2001a; Roulier and Jarvis, 2003). A prime precondition for the occurrence of macropore flows in cohesive soils is the presence of connected soil structures or biopores that link the soil surface with the subsoil (Vogel et al., 2005; Zehe and Flühler, 2001b; Flury et al., 1994). However, whether or not these structures

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are activated at any given time (i.e., during a given rainfall event) depends on the interplay of initial soil moisture and rainfall forcing (Zehe and Flühler, 2001b; McGrath et al., 2007). Figure 2 shows, for instance, two dye tracer distributions observed in the German Weiherbach catchment, both having a similar surface density of connected macropores, shortly after irrigation of these field plots with 25 mm of tracer solution (Zehe and Flühler, 2001b). In the case of a constant rainfall forcing, the initial soil moisture determined whether indeed matrix flow dominated (left panel, $0.20\,\mathrm{m^3\,m^{-3}}$ initial soil water content) or if the system switched to macropore flow and transport (right panel, $0.25\,\mathrm{m^3\,m^{-3}}$ initial soil water content). Maximum tracer depth in the latter case was 1 m, which was ten times larger than for the matrix flow dominated case. The question of whether or not the soil switches from matrix to preferential flow regime is critical for flood generation (Zehe et al., 2005) and even more in the context of the fate of agrochemicals (Flury, 1996). Although a typical transport depth of 1 m appears to be "not much" at first sight it makes an enormous difference to the life-time of a pesticide in the soil. For example, the life-time in the topsoil is typically 6–30 days, and at a depth of 80 cm it is already longer than 150 days (Bolduan and Zehe, 2006; Issa and Wood, 1999).

An interesting question is now whether a switch from matrix flow to preferential flow does allow more efficient dissipation of the energy input that accompanies heavy rainfall, similar to our tea pot example? We think so. Zehe and Blöschl (2007) showed that fast infiltration through preferential pathways allows for a faster reduction of potential energy during rainfall events than during flow in the matrix continuum of cohesive soils. Furthermore, and most importantly in cohesive soils, the redistribution of water from the macropores into the soil matrix is a much more efficient way to reduce capillary energy than simple matrix flow from the soil surface. Hence, it is worth emphasizing that the switch to the preferential flow regime also allows a more efficient dissipation of energy in the case of strong energy flow/heavy precipitation events, similar to many elementary threshold phenomena that we have covered so far (e.g., the tea pot example).

2.1.3 Threshold behavior of soil properties, I: water repellency and overland flow generation

Overland flow generation in water-repellent soils may exhibit an even stronger threshold character, because the threshold control for overland flow generation itself exhibits
5 soil moisture dependent threshold behavior. Water repellency of soils is a phenomenon that is caused by the presence of degraded organic matter (Krammes and DeBano, 1965) and sometimes by forest fires (de Bano and Rice, 1973), and is often associated with sandy soils and only occasionally with clay soils. The main controls are the type of organic matter, the occurrence of dry spells and soil moisture (DeJonge et al., 1999).
10 DeBano and Rice (1973) suggested that water repellency occurs after soil moisture drops below a certain threshold. Water repellency was reported as early as 1910 by Schreiner and Shorey (1910) for soils in California that could be wetted neither by infiltration nor by the rise of ground water tables. A recent bibliography on water repellency (Dekker et al., 2005) highlights the global occurrence of this phenomenon.
15 The potential for hydrophobicity can be determined by the Water Drop Penetration Time test (WDPT, Dekker and Ritsema, 1994), which is based on the time a water drop needs to penetrate the soil after it has been applied. Another method is the molarity-of-ethanol-droplet test on air-dried soil samples (Letey et al., 2000). Both tests lead to an ordinal scale for hydrophobicity that begins at zero, which indicates no hydrophobicity.
20 Hence, hydrophobicity is a threshold phenomenon, which however can display varying levels of intensity when it does occur.

To shed light on the control of antecedent soil moisture on overland flow generation in a landscape with hydrophobic soils, Zehe et al. (2007) performed 53 sprinkling experiments in southern Switzerland, a landscape prone to occasional droughts and wildfires.
25 They applied artificial rainfall with an intensity of 50 mm/h, which occurs with a return period of 5 years. Runoff from the plots was measured manually, and initial soil moisture was measured at two points in each plot using Time Domain Reflectometry (TDR). Water repellency was estimated before each sprinkling experiment by measuring the

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water drop penetration time and also by performing the molarity-of-ethanol-droplet test on air-dried soil samples (Letey et al., 2000).

Figure 3 shows examples of observed runoff at the same field plot for wet and dry antecedent soil moisture conditions (0.30 and 0.07 m^3 m^{-3}, respectively). Due to the high saturated hydraulic conductivities of about 5×10^{-5} m/s, which is equivalent to a rainfall intensity of 180 mm/h, a large part of the water infiltrates when the soil is wet. In contrast, when the soil is dry surface runoff is increased threefold at this plot as a result of the strong water repellency. Figure 4 (right panel) presents a plot of the total runoff depths that were observed during the 53 irrigation experiments against the average initial soil moisture. It is quite apparent that the rainfall intensity threshold for initiating overland flow strongly depends on the average wetness state of the plot. There are two "regimes" observable in the data, one associated with strong overland flow response of, on average, 45 mm for an initial soil moisture of less than 0.11 m^3 m^{-3}, and the other with a much weaker overland flow response of, on average, 9.5 mm for soil moisture greater than 0.21 m^3 m^{-3}. For wetness states between 0.11 and 0.21 m^3 m^{-3} the overland flow response exhibits much larger variability than in either of the stable ranges. In this "unstable range" the system seems to gradually switch from strong hydrophobic conditions where the threshold for Hortonian surface runoff generation is much lower and runoff generation is strong, to hydrophilic conditions where the threshold for Hortonian surface runoff generation is higher and runoff generation is much weaker. This behavior is similar to that discussed earlier in our tea pot example.

In Sect. 4, which deals with level A predictions, i.e. the question of whether or when a threshold is triggered, and the system switches to another regime, we will revisit this problem.

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2.2 Threshold behavior at the response level

2.2.1 Interactions of spatial patterns and local threshold processes

Understanding local scale process thresholds, whilst important, is nevertheless insufficient for predicting the hydrological responses of larger entities such as catchments or
5　even hillslopes. This is because the overall hillslope or catchment response is determined by not only the local process responses, which may of course have a threshold
character, but also by the redistribution of these responses in space through additional
processes that interact across different parts of the landscape (Rundle et al., 2006).
For instance, the overland flow response of hillslopes is determined by local runoff
10　generation as well as by runoff concentration (i.e., routing), which may or may not possibly re-infiltrate on its way down slope. What is necessary for the hillslope as a whole
to respond with overland flow is a connective path down slope so that overland flow
can reach the foot of the hillslope and possibly a river.

　　This idea of global connectivity (Lehmann et al., 2007; Schulz et al., 2006) is
15　very nicely illustrated for overland flow in the water repellent landscape introduced in
Sect. 2.1.3. If an upslope patch has a dry soil surface and is strongly hydrophobic it will
respond with strong overland flow generation during a rainfall event (compare Figs. 3
and 4). When the next patch on the down-flow path – which is determined by local
topography – is in a wet and hydrophilic state, the upslope run-on, together with local
20　precipitation, may yet fully infiltrate due to the high infiltration capacity of soil in this
case. Hence, the spatial distribution of the local runoff responses has the net effect of
inhibiting the overland flow signal coming down from upslope. If on the other hand, the
down-slope patch is also in a dry, highly water repellent state, run-on from the up-slope
patch will not infiltrate and, in combination with the effects of local precipitation, the
25　down-slope patch will generate additional surface runoff. Thus, in this case the spatial
variability of the local runoff response adds to the downstream effects of the upslope
response and the redistribution leads to an amplification of the signal. In the presence
of spatial heterogeneity of soils, the hillslope responds as a whole with overland flow

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whenever or wherever there is at least one globally connected path of wet patches that allows down slope runoff concentration without (total) re-infiltration. Since soil moisture is the macroscopic state variable to determine hydrophobicity at the local level, the hillslope scale soil moisture pattern then has to be connective (Lehmann et al., 2007; Schulz et al., 2006) for there to hillslope scale overland flow response. There must be a well connected path of patches across the hillslope along which the surface soil moisture content remains below the value of $0.11\,\mathrm{m^3\,m^{-3}}$ (see Fig. 4, left panel) at the same instant in time. This is quite a strong condition, because it refers not only to the number of dry patches but also to their spatial arrangement, and it involves a process interaction such that a positive feedback in space is generated (Rundle et al., 2006). The underlying control of this threshold phenomenon is clearly more complex than at the process level.

A second important control of response thresholds is the spatial correlation or spatial coherence between the spatio-temporal patterns of the forcing and spatial patterns of key parameters or state variables within a catchment. Think of the preferential flow phenomenon in the context of groundwater contamination. Zehe and Blöschl (2004) showed that co-variation of the spatial patterns of initial soil moisture and macroporosity is a first order control in this case. A co-location of wet soil patches with areas of high macroporosity favored preferential infiltration and transport of solutes into the subsoil and shallow groundwater at the plot as well as at the catchment scale. In contrast, if dry patches are co-located with areas of high macroporosity, the solute did not leach into the subsoil but remained within the plough horizon.

Spatio-temporal coherence of forcings and landscape structure or patterns is also a precondition for the occurrence of extreme or severe floods (Naden, 1992; Woods and Sivapalan, 1999). In this case, the major sub-catchments have to all respond with floods and the sequence of the sub-scale flood responses has to be such that the consequent flood hydrographs arrive together at a downstream location leading to the generation of much larger flood peaks than otherwise (Rundle et al., 2006; Blöschl and Zehe 2005).

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Sediment export from hillslopes into the river network is another classical example of threshold behavior at the response level. It is determined by the triad of local particle detachment, down slope particle transport and sedimentation (Scherer, 2008; Meyer and Wischmeier, 1969). Both detachment and sedimentation are threshold processes. The first occurs when momentum input during rainfall exceeds a critical value or when the shear stress of overland flow increases above a critical value, which is determined by land use, root- and soil cohesion (Scherer, 2008; Huan et al., 1998). The latter occurs when sediment concentration in overland flow increases above the transport capacity. Following Scherer (2008) and Hairsine et al. (1992, 2002), the latter increases with overland flow depths and velocity, and these two variables are determined by surface roughness, which is determined by land use and also by micro- and macro-topography, in a sense, the form of the hillslope. Sediment export from hillslopes therefore occurs, depending on the land use pattern and hillslope form, only if rainfall intensity and depth are large enough to establish a connected flow path of sufficiently high flow depths and speed that particles can be transported all the way to the foot of the hillslopes. The latter often exhibits a concave shape and is therefore the sink region for the deposition of sediment particles from the upslope source areas (Scherer, 2008). Hence, the land use pattern, soil erodibility, hillslope form as well as the precipitation pattern, determine whether or not the hillslope flow paths become sufficiently connected to allow sediment export.

In the following sections, we will provide two examples on how the connectivity of patterns controls subsurface stormflow and how correlated interactions of various patterns control the rainfall-runoff response in a catchment with cracking soils.

2.2.2 Subsurface storage control on hillslope runoff response

Hillslope and catchment streamflow responses to rainfall events in forested midmountain reaches are often dominated not by overland flow but by lateral subsurface stormflow (Weiler et al., 2005), which also happens to be an intermittent phenomenon. The generation of subsurface stormflow is strongly governed by the interplay of the

precipitation input, and the subsurface drainage and storage characteristics, which in turn depend on the patterns of soil depth and soil hydraulic conductivity, the permeability of the underlying bedrock as well as the presences of soil layers and macropores. There is considerable evidence that these complex interactions of patterns and nonlinear subsurface flow processes result in different types of threshold behavior of hillslope subsurface flow responses. Whipkey (1965) observed, during a field experiment, that a precipitation threshold of 35 mm had to be exceeded for the hillslope to respond with subsurface flow. Mosley (1979) and Tani (1997) found similar threshold behavior but with a precipitation threshold of 20 mm for a field site in New Zealand and at a trench site near Okayama in Japan.

At a hillslope trench site in Georgia, USA, Tromp-van Meerveld and McDonnell (2006a) observed a threshold value of 55 mm of rainfall (Fig. 5) for the hillslope to respond with subsurface stormflow. Based on observed spatial and temporal patterns of saturation at the soil bedrock interface, Tromp-van Meerveld and McDonnell (2006b) suggested the "fill and spill" mechanism to explain this observed threshold behavior. Their idea basically was that subsurface topography controls water flows from upslope positions to lower reaches of the hillslope. During storms with rainfall depths smaller than a threshold value a free subsurface water table forms within isolated areas of the hillslopes. However, local barriers in the bedrock micro-topography hinder flow down-slope. For precipitation events larger than this threshold, according to Tromp-van Meerveld and McDonnell (2006b), pits in the bedrock relief are filled and excess water then spills over the micro-relief of the bedrock depressions. The subsurface saturated areas thus become connected and subsurface flow is established rather abruptly and delivers water to the stream channel down below. In this case, the threshold value refers to the volume of water necessary to establish a globally connected water table that connects otherwise isolated saturated parts of the bedrock topography.

2.2.3 Threshold behavior of soil properties, II: soil cracking and catchment scale flood response

In landscapes with cracking clay soils the opening/closing of cracks can cause an abrupt increase/decrease of soil infiltration capacity at the local scale. Internal shrink-
age and swelling of clay minerals such as smectites or vermiculites can cause soil shrinkage (Kariuki and van der Meer, 2004), which is driven by soil moisture changes. In temperate climatic regions normal shrinkage is the most relevant mechanism, where volume loss is approximately proportional to the water loss (Bronswijk, 1988; Chertkov, 2000). When soil moisture drops below a threshold value cracks begin to develop and
expand, which might enhance infiltration due to macropore flow (Bronswijk, 1988; Wells et al., 2003). Likewise, if the soil again wets up above a certain threshold value, the cracks begin to shrink and eventually close, and the local infiltration capacity is then substantially reduced. This, for example, is the case of the vertisols of northern Mexico (Návar et al., 2002) or in the Tannhausen catchment in Germany.

The interaction of this local threshold process with the catchment scale pattern of soil moisture can cause threshold behavior in the catchment scale overland flow response, as observed by Zehe et al. (2007) in the 2.3 km^2 Tannhausen catchment in Germany. The geological setting of this catchment consists of clayey and marly sediments of lower Jurassic age, where Luvisols, stagnic Gleysols and Regosols (ISSS-ISRIC-FAO,
1998) have developed. Saturated hydraulic conductivities are generally quite low, lying between 1.5×10^{-6} and 2×10^{-7} m/s, which correspond to rainfall intensities of 5.4 and 0.7 mm/h. Hence, one should expect the catchment to react, in general, with strong Hortonian overland flow generation, with high event runoff coefficients, especially in summer time, which is the season of thunderstorms and extreme rainfall events. How-
ever, the Regosols, which are located in the valley floors close to the river, exhibit considerable shrinkage and crack formation during the dry spell and swelling/ closing of cracks during wet conditions. The latter occurs within several hours to 1–2 days (Fig. 6, Lindenmaier et al., 2006). Hence, catchment scale runoff response exhibits a strong

seasonality, somewhat counter intuitively, with high runoff coefficients in winter and fall and small runoff coefficients in summer (Fig. 7, Lindenmaier et al., 2006), which are mainly controlled by the interaction of the catchment scale soil moisture pattern with the soil structure pattern. Figure 6 shows, for instance, the rainfall-runoff response observed in the Tannhausen catchment ($2.3 \, km^2$) for a summer event (upper panel) and a winter event (lower panel). Although precipitation depths were very similar between these two events, runoff responses were very different due to different antecedent conditions. In summer, cracks in the Regosol soils were open due to the very dry soil conditions and hence most of the overland flow from upslope regions re-infiltrated in the valley floors (see upper panels). Onset of overland flow response lags by more than half a day behind the onset of rainfall. During the winter season, no cracks were present. The catchment reacts much faster, with strong overland flow runoff generated in the upslope area that does not re-infiltrate in the valley floors, and can therefore reach the river without loss to infiltration, and thus produce a strong flood response.

Figure 4 (left panel) shows observed runoff coefficients for 260 rainfall runoff events plotted against the estimated average initial soil moisture of the catchment (for details refer to Zehe et al., 2007). As no soil moisture observations were available in this catchment, we used the model CATFLOW (Zehe and Blöschl, 2004; Lindenmaier et al., 2006, see Sect. 4) to estimate soil moisture in the upper 30 cm of the soil. Based on a digital elevation model, detailed soil hydraulic properties, as well as observed meteorological data, the water balance of the catchment was simulated for the period of 1994 to 2004. For each time step, the simulated soil moisture field was averaged over the catchment, yielding a time series of catchment average soil moisture as an estimator for the initial states.

Similar to the case of overland flow production in water repellent soils (compare Sect. 2.1.3, Fig. 4, right panel), two stable ranges are clearly apparent. One is associated with weak flood responses and a runoff coefficient of, on average 0.06, for average initial soil moisture less than $0.29 \, m^3 \, m^{-3}$. The other is a strong runoff response with an average runoff coefficient of 0.66 for an average soil moisture value larger than

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$0.35\,\mathrm{m}^3\,\mathrm{m}^{-3}$. Clearly, in the drier regime soil cracks are likely to be open (compare Fig. 6), so most of the rain water infiltrates into the Regosol soils and catchment scale flood response is weak. In contrast, in the wet regime, the cracks are closed, and so the transformation of rainfall to runoff is much stronger. Closing of these cracks seems
5 to take several hours to days, as may be estimated from the hydrographs in Fig. 6.

In the unstable range between 0.29 and $0.35\,\mathrm{m}^3\,\mathrm{m}^{-3}$ the scatter of the observed runoff coefficients is – similar to the case of water repellent soils – very large. The spatial average initial soil moisture of the catchment does not of course give any information on how the catchment scale soil moisture pattern and the soil pattern interact.
10 Soil moisture patterns in catchments where the dry regions are co-located with the Regosol soils and soil cracks are open are indistinguishable from those patterns where the Regosols are wet and the cracks are closed, even if the spatial average soil moisture is the same for both cases. However, this information is essential for constraining the catchment scale flood response: in the first case Hortonian overland flow from up-
15 slope areas infiltrates into the open cracks, in the latter case the upslope areas can be connected to the river reach. Hence, we may state that:

– correlation of spatial patterns of soil moisture and soil properties determines whether upslope areas are connected to the river and whether catchment scale response is weak or strong.

20 – information on the average wetness state is not sufficient for predicting catchment scale runoff response when the catchment is, on average, in the vicinity of a catchment-scale process threshold.

It is interesting, and also important, to stress that both scatter plots in Fig. 4 are similar, although they are based on observations at totally different scales. As argued for our
25 tea pot example we observe that in both cases stable ranges are apparent where a Level A prediction should be straightforward. In both cases an unstable range is apparent where a prediction about whether the system switches from one regime to

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another regime appears to be much more difficult because information about the initial state seems to be insufficient. We will come back to this interesting phenomenon in Section 3 as we discuss the nature and extent of information that may be usefully extracted from hydrological observations to characterize spatial patterns in the context of predicting threshold behavior.

2.3 Threshold behavior at the functional level

2.4 Functional thresholds and stability of geo-ecosystem

We introduce this third level because vegetation patterns, soil patterns and structures that determine threshold behavior at the process and response levels are not necessarily stationary (Phillips, 2006). There are indeed typical patterns and structures in a landscape, depending on climate and bio-geophysical setting; otherwise the term landscape would not even exist. Indeed, Watts (1947) and Turner (1989) have argued that similar patterns have been formed by similar processes. As these patterns of soil and vegetation and other biota have co-evolved over long time scales (Dietrich and Perron, 2006) we think they represent a configuration that is well adapted to the climate forcing and the geological setting. This configuration determines the hydrological functioning of a landscape, i.e. how the energy inputs due to both precipitation and radiation are redistributed in the landscape and translate into well known characteristic hydrological responses and into biomass production. Common statistical indices such as the flow duration curve, the flood frequency curve, the monthly runoff regime, average annual water balance and – if available – similar statistics on sediment yields and hydro-chemistry are some of the signatures that can be used to define hydrological functioning (Wagener et al., 2007).

Resilience of such a geo-ecosystem configuration and related hydrological functioning – defined as the extent of disturbance the system may tolerate without responding with qualitative (i.e., drastic) changes – is determined by the multiple, often biologically mediated feedbacks between abiotic and biotic components that comprise the system

(Saco et al., 2006). Stationarity of these hydrological indices signifies stationarity of hydrological functioning as well. Non-stationarity of these indices and therefore hydrological functioning can either be the effect of climate change, or persistent or substantial changes in those geo-ecosystem properties that determine the process and response

5 spectra associated with the catchment. Examples of these could be substantial and persistent changes in vegetation dynamics due to change in habitat conditions and related feedbacks on local process thresholds and hydrological responses (Schröder, 2006), and likewise changes in biologically formed soil structures and related feedbacks on process thresholds associated with runoff generation and preferential flow.

10 This can be seen as another form of threshold behavior that manifests on the long time scale through significant changes in, for example, sediment yields, flood frequency curve etc. The underlying controls are far more complex than those at the process and response levels since they involve substantial disturbances that go beyond the process or response levels to the resilience of the geo-ecosystem as a whole, and

15 involve interactions and feedbacks between the biotic and abiotic components of the geo-ecosystem (Dietrich and Perron, 2006), which have co-evolved over long periods of time.

In the following we will discuss two instructive examples of threshold behavior at the functional level, both from the field of eco-hydrology. We leave out the field of climate

20 change impacts, which we think is well covered in the literature. The first example deals with semi-arid runoff-run-on systems, their functioning and what happens when they are subjected to substantial disturbances. The second example deals with spatial patterns of earthworm burrows in humid landscapes, their key role in Hortonian overland flow generation, and the associated flood response.

25 2.4.1 Semi arid geo-ecosystems, resilience of soil-vegetation patterns and runoff-run-on mechanisms

The question of resilience of the geo-ecosystem and related hydrological functioning is most instructively illustrated for semi-arid geo-ecosystems which, according to Saco

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et al. (2006), cover over 30% of the world's land surface. Geo-ecosystems in semi-arid regions are highly coupled geological-ecological-hydrological systems with strong feedbacks that occur across multiple scales. Vegetation is often arranged into a mo-saic, consisting of patches with strong plant cover that alternate with low-cover or bare

5 soil patches. This pattern leads to spatial organization of infiltration/overland flow pro-cesses (Saco et al., 2006; Schröder, 2006), an efficient re-distribution of water and nutrients from bare to vegetated patches, which has a strong positive feedback on plant habitat conditions (Seyfried and Wilcox, 1996). Infiltration capacities in bare soil patches are generally low due to surface soil crusting. In contrast, vegetated patches

10 allow high infiltration rates and also preferential flow due to the presence of intercon-nected preferential pathways (root channels and earthworm burrows (Ludwig et al., 2005; Thiery et al., 1995). Hence, these areas trap most of the local precipitation and the water that runs off from the upslope bare soil patches on to the vegetated patches. This so called runoff-run-on mechanism has a strong positive feedback on

15 the system configuration, as soil moisture in vegetated patches is higher, which re-inforces biomass production, which in turn reinforces development/persistence of soil structures (Wilcox et al., 2003). This efficient redistribution of water and nutrients is a first order control on the resilience and geo-ecosystem stability within drylands (Saco et al., 2006). The macroscopic result of these feedbacks are the typical vegetation

20 patterns which may be either spotted or stippled, consisting of dense vegetation clus-ters that are irregular in shape, surrounded by bare soil (Saco et al., 2006). Another common pattern is banded vegetation, also known as "tiger bush" in Africa (compare Fig. 8) and "mogotes" in Mexico, where the dense biomass patches are seen to form coherent and long-lasting bands, stripes or arcs.

25 These typical soil-vegetation patterns control, along with topography, the hydrologic functioning of such semi-arid and arid systems, in particular the redistribution of water, nutrients and soil material (compare also Sect. 2.2.1). Water losses due to evapo-ration and overland flow velocities are minimized as a consequence, as the overland flow re-infiltrates in the next vegetation cluster located down-slope. This in turn mini-

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mizes erosion and soil losses. However, a severe disturbance of this geo-ecosystem, e.g., due to overgrazing, might seriously endanger the stability and the hydrological functioning and ability of this landscape to control the export of water, sediments and nutrients. Leaky vegetation structures caused by over-grazing are less efficient in trap-
5 ping runoff and sediments (Walker et al., 1981; Thiery et al., 1995). Consequently, the geo-ecosystem starts to lose water, soil and nutrient resources, which might then lead to further reduced vegetation cover, which reinforces the water and nutrient losses, which in turn reinforce further degradation of the landscape (Lavee et al., 1998). Hydrological functioning, as characterized by patterns of sediment export, and the flow
10 duration curve, start to change irreversibly, and we can then say that a threshold at the functional level has been triggered.

2.4.2 Humid geo-ecosystems, stability of earthworm population and biological controls of Hortonian flood response

Such strong geo-eco-hydrological interactions with positive feedbacks do of course
15 also occur in regions within temperate climates, although they are often not that obvious. Zehe and Flühler (2001b) found a spatially organized pattern of anecic earthworm burrows in the Weiherbach catchment, located in the humid climate of southern Germany. This landscape, characterized by gentle slopes, exhibits typical loess soil catenas formed by erosion: moist but drained Colluvisols at the foot of the hillslopes and
20 valley floors and drier Calcarig Regosol soil in the upper hillslope sectors. A mapping of anecic earthworm burrows showed a higher spatial density of larger and deeper worm burrows within the Colluvisol soils, which can be easily explained by the fact that the earthworms are likely to find better habitat conditions in the more even moisture regime and the higher amount of soil organic material within the Colluviosols. This typical spa-
25 tial pattern of soils and macroporosity in this landscape did, in turn, lead to a spatially organized pattern of flow and transport (Zehe and Flühler, 2001b) and exerted a key influence on the catchment scale runoff response to extreme rainfall events (Zehe et al., 2005).

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Once again, this arrangement of biologically mediated patterns of soils and soil structures is stabilized by positive feedbacks: the enhanced infiltration rates in the valley floors reduce overland flow velocities, and favor sedimentation (i.e., deposition) of fine eroded material and organic matter that emanates from upslope regions (Scherer,
5 2008, compare Sect. 2.2.1). This creates moist but drained conditions that are favorable to anecic earthworm populations (Joschko et al., 2006; Whalen and Fox, 2007), which in turn assures a high density of connected earthworm burrows at the foothill sector. Figure 9 presents the response to a large flood event that occurred in June 1994, which was simulated with a physically based model for different areal densities
10 of macropores in this catchment (cf. Sect. 3.2.2 for model setup). On the other hand, simulations without the presence of macropores over-estimated the observed flood response by a factor of three.

However, anecic earthworm populations are very sensitive to disturbances from extensive agriculture, especially to applications of agrochemicals, or specific crops such
15 as mustard, and to surface preparation. These can even lead to a collapse of the earthworm populations (Joschko et al., 2006; Whalen and Fox, 2007). As long as the earthworm population is stable there is seasonality to the number of anecic earthworm burrows that connect the surface to the subsoil. Earthworm burrows are disrupted and re-distributed due to plowing in spring, which reduces their net effect on infiltration
20 (Flury, 1996; Zehe and Blöschl, 2004). However, during summer they get reconnected to the surface, the system is dynamically stable, contributing to the seasonal variation in the areal density of connected earthworm burrows. Modeling rainfall-runoff response in this catchment needs to capture this seasonality of soil structure, requiring a higher density of macropores in late summer than in early summer, for example (Fig. 9).
25 In the event of a collapse of the earthworm population, the disrupted earthworm burrows are no longer reconnected to the surface. As shown in Fig. 9, this could have a drastic influence on the hydrologic functioning of this rural geo-ecosystem, as shown by simulations with the physically based hydrological model, CATFLOW. If the infiltration is limited to only the soil matrix continuum, the flood responses become four

times larger than in the present equilibrium state (i.e., in the presence of earthworm burrows), sediment yields and phosphorus export during strong flood events would increase by even one order of magnitude (Scherer, 2008). The stronger erosion could, in the very long term, also affect the geomorphology of this landscape. A collapse of the anecic earthworm populations in such a Loess landscape catchment could therefore substantially change the flood frequency curve as well as patterns of sediment export: this, according to our definition, is another example of a threshold on the functional level.

An interesting point in the context of erosion is that we ourselves can trigger more positive functional thresholds through appropriate land use planning. Scherer (2008) demonstrated that a simple re-arrangement of land use pattern by co-locating crops in locations of high erosion risk can substantially reduce sediment export from the Weiherbach catchment. A thorough understanding of threshold behavior at the functional level is therefore pivotal for understanding the resilience of geo-ecosystems and for sustainable land management. Appropriate models, which incorporate the various process interactions and feedbacks, can provide valuable assistance in this respect.

3 Observability of state variables and level A predictions of threshold behavior

3.1 Uncertainty and limited observability of hydrological state variables

As argued in the introduction, predictions are a mixed discrete-continuous problem whenever or wherever threshold behavior is involved: level A is to predict whether or not the system switches to another regime (e.g., see the last section), and level B is to predict the strength/intensity of the new process/regime, in the case where the switch does happen. We already argued in the case of our tea pot example and provided detailed simulation evidence in Sect. 3.2 that level A predictions are themselves very difficult, if the system gets "close" to the threshold needed to switch to another regime. Even if detailed information on the initial soil moisture patterns is at hand, small dif-

ferences in the initial states, which cannot be resolved with current observation technology, may yet cause substantially different flood responses. "Close" in this respect means close with respect to the error/ uncertainty scale of our observations. Hence, the unstable range over which level A predictions are difficult to make is strongly related
5 to the accuracy/error range of our observations. However, from the experimental point of view we should also be aware that our observations may be highly uncertain when the observed system is close to a threshold necessary to switch to another regime. The question of how well one may be able to repeat observations of cause and effect is therefore fundamental to the understanding of how well we might predict the onset
10 of a threshold phenomenon.

A system one hopes to predict with perfect accuracy needs to give exactly the same response if the experiment is repeated under identical conditions, i.e., when the forcings and the initial states of repeated trials are exactly the same. However, distributed observations of state variables, especially at larger scales, are non-exhaustive and
15 therefore inherently uncertain. This is because the support, spacing and extent scales of our distributed measurements tend to be incompatible with the spatial scale of state variables and processes in natural systems (Blöschl and Sivapalan, 1995). As shown in Sect. 3.2.2 even 61 point measurements of soil moisture did not suffice to fully characterize the spatial (or probability) distribution of soil moisture at a catchment of only
20 $3.2\,km^2$ size. Even if the probability density function (pdf) was known, the detailed spatial pattern will still elude full quantification (Deeks et al., 2004; Western et al., 2004), as discussed in the previous section. Observations of hydrological state variables, in particular with respect to their spatial patterns, are not sharp and are highly uncertain.

Even when we repeat experiments in the laboratory, which is the classical physical
25 approach to understanding cause and effect, we can determine "identical" conditions only with respect to the errors/uncertainty of our measurements. Identical conditions in the case of hydrological field experiments, due to limited observation technology, are of course much more uncertain. We cannot repeat observations under truly identical conditions but can do so only under apparently identical conditions. Consequently

observations of hydrological system response, such as tracer transport depths, residence times and runoff volumes, often exhibit considerable scatter, even if the initial and boundary conditions are apparently identical (Lischeid et al., 2000; Zehe et al., 2005; Zehe and Blöschl, 2004). In the context of threshold behavior this implies that thresholds at the process level are even harder to identify in the field than in the laboratory. What we can do is to identify a range for a macroscopic state, within which the process might switch to the other dynamic regime, rather than to identify a sharp threshold as in the case of a typical laboratory system.

In the following we will provide two examples to further elaborate these ideas for the case of overland flow generation over water repellent soils and for catchment scale flood response.

3.2 Level A predictions and uncertainty of apparently identical observations

Zehe et al. (2007) suggested a simple conceptual model to assist in unraveling how well we might detect threshold behavior in hydrological systems and to understand its implications for predictability of hydrological systems. The idea is to artificially simulate hydrological observations explicitly involving threshold behavior. As suggested by Fig. 4, and discussed in Sects. 1 and 2, they assumed that threshold behavior in a hydrological system can manifest through the existence of at least two regimes where a process/response behaves macroscopically differently. These regimes are stable when within finite ranges of a macroscopic state variable, in this case the initial soil moisture. Within each of the two regimes, Zehe et al. (2007) represented the system response/observation process by a stationary probability density function, assumed to be a Gaussian as a first guess. The response functions differ with respect to their expected values, and also their standard deviations. As shown in Figure 4, the threshold for the switch from one regime to the other can only be identified within an uncertain range of states. Within this unstable range of the macroscopic variable, the system switches between the two response functions. The transition between the two regimes is conceptualized as an error function that relates the expectation of the response func-

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tion to the control variable.

This simple 7-parameter model was employed to simulate repeated trials of observations of point scale overland flow on a water repellent soil in the Swiss Alps (cf. Sect. 2.1.3) as well as of observations of floods in the Tannhausen catchment, which is located in a cracking soil scape (cf. Sect. 2.2.3). The 7 model parameters i.e. the mean and standard deviation of the response functions in the two stable regimes as well as the width of the unstable range, were determined based on the available experimental data (cf. Zehe et al., 2007). The next step was to estimate the inherent uncertainty of the initial soil moisture observations, both for the irrigation plots in the Swiss Alps as well as for the entire Tannhausen catchment. In this case, Zehe et al. (2007) assumed that the inherent uncertainty of the initial states may, as a first order approximation, be represented by a Gaussian distribution, which was then parameterized on the basis of available data.

Zehe et al. (2007) then simulated repeated trials of the observations for apparently identical initial conditions, for sample sizes that were equal to the number of observations at each study area. Figure 10 (upper and lower left panels) presents the scatter plots of simulated observations for both systems plotted against the average initial soil moisture states. Both graphs correspond very well to the patterns of the true observations of plot scale overland flow in the hydrophobic soil-scape and the flood response of the Tannhausen catchment (cf. Fig. 4). Hence, we can say that the simple model is able to reproduce the scatter of the true observation process, although the scale and the underlying mechanisms are different in these two cases. However, both the process and the response exhibit threshold behavior. Thus, the nature of the process /response is the same.

In a next step the authors simulated the observations for a very large sample size of more than 10 000 observations (of course this can never be achieved in the real world). Uncertainty in the simulations was quantified using the scaled width of 95% confidence intervals of all observations that belong to an apparently identical initial state. The upper right panel of Figure 10 shows the width of the scaled 95% confi-

dence intervals plotted against the average initial soil moisture for plot scale overland flow observations. The peak value of 3.5 at $0.17 \, \text{m}^3 \, \text{m}^{-3}$ initial soil moisture indicates that the uncertainty range of predicted plot scale overland flow rates is 350% of the average value. For initial states far away from the threshold the error ranges of the predicted overland flow rates drop to values around 0.7. As predictability is low when uncertainty is high, a level A prediction, i.e., whether the system switches from strong overland flow production during hydrophobic conditions to weak overland flow production in the hydrophilic regime, is most simple in the stable ranges but highly uncertain in the vicinity of the threshold.

The simulation of the catchment scale flood response in the Tannhausen exhibits, in principle, vey similar pattern as in the above example (Fig. 10, lower right panel). Prediction uncertainty is clearly higher, and a level A prediction is very difficult, for average initial soil moisture conditions ranging between 0.28 and $0.34 \, \text{m}^3 \, \text{m}^{-3}$. The maximum predictive uncertainty is around 2.5, which indicates a scatter of 250% in the predictions compared to the average runoff coefficient. We already explained that the information about the average initial soil moisture state lacks key information about the spatial correlation of the soil pattern in the Tannhausen catchment and the soil moisture pattern (cf. Sect. 2.2.3), i.e., whether locations of Regosol soils coincide with wetter than average patches and cracks are closed or it is the opposite case and the cracks are open. It turns out therefore that close to a threshold, the response of the hydrological system depends very strongly on details that may not be resolvable with current observation technology.

4 Modeling approaches for threshold behavior

4.1 Hydrological models in general

Modeling approaches for simulating hydrological response as a whole or key individual processes are dependent on the scale and modeling target and are as rich and

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diverse as hydrology itself. In the field of hydrological modeling we find many physically based, process oriented, numerical models at the catchment scale (MIKE SHE: Refsgard and Storm, 1995; HYDRUS: Simunek et al., 1994; CATFLOW: Zehe et al., 2001, HILLFLOW, Bronstert and Plate, 1997; InHM, VanderKwaak and Loague, 2001).
5 Their application at present is restricted to scales ranging from the plot to the small catchment. On the other hand, we also find conceptual models that are based on simpler conceptualizations of hydrological processes, such as the HBV model (Bergstöm, 1996; Hundecha and Bárdossy, 2004), the Sacramento model (Crawford and Linsley, 1966), and the LASCAM model (Sivapalan et al., 1996). Such models are commonly
10 employed to simulate hydrological responses of catchments larger than $500\,km^2$ in size. In between these two extremes there is a lare a variety of hybrid models that combine conceptual and process oriented approaches, such as the TAC model (Uhlenbrook and Leibundgut, 2002), LARSIM (Bremicker, 1998), WASIM-ETH (Gurtz et al., 2003), TOPMODEL (Beven and Kirkby, 1979), THALES (Grayson et al., 1995), and
15 TOPOG (Vertessy et al., 1993). Since most of the relevant hydrological processes are nonlinear, many of these models have to simulate/mimic nonlinear behavior and could therefore, in principle, be easily extended to simulate threshold behavior as well, and some already do (e.g., InHM, VanderKwaak and Loague, 2001; CATFLOW: Zehe et al., 2001). An exhaustive review of all such models and modeling approaches is beyond
20 the scope of this paper. For a review of modeling preferential flow, see Simunek et al. (2003).

Here we will focus on modeling approaches that were either explicitly used for simulating forms of threshold behavior presented in the last section, or are explicitly developed to simulate threshold behavior or threshold events. We will leave out the large
25 field of dynamic models for the climate and earth-system, as well as eco-hydrological models. Interested readers can refer to Pitman and Stouffer (2006) and Saco et al. (2006), respectively, for more details of these modeling approaches.

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4.2 Modeling of threshold behavior at different scales and different levels of complexity

4.2.1 Modeling Hortonian overland flow response in a Loess landscape using the process model CATFLOW

As explained in Sect. 2.1.2 macropore flow is a threshold process, which may locally increase infiltration rates significantly, and its occurrence is controlled by the interplay of initial soil moisture, soil structures and precipitation. However, the effect of this local threshold process on the overland flow response of an entire catchment depends, as elaborated in Sect. 2.2.1, on the interaction between the spatial pattern of connected macropores within the catchment with the patterns of the rainfall forcing and initial states. As already explained in Sect. 2.3.3 the hillslope scale patterns of soils and preferential pathways are spatially organized, with a smaller density of connected macropores in the upslope areas but a clearly higher density of earthworm burrows in the Colluvisols in the valley floors. In modeling studies, Zehe et al. (2005) and Zehe et al. (2006) showed that this pattern has a distinct, state dependent influence on the overland flow response of this catchment.

To this end they employed the physically based numerical model CATFLOW. The model subdivides a catchment into a large number of hillslopes and the associated drainage network. Each hillslope is discretized into a two-dimensional finite difference grid along the main slope line. Hillslope soil water dynamics, solute transport, overland flow and evapo-transpiration as well as discharge in the stream are simulated using widely accepted approaches based on a set of coupled partial differential equations (Richards equation, convection-dispersion equation in two dimensions, Saint-Venant equation, Penman-Monteith equation: refer to Dingman, 1994 for the governing equations and mathematical expressions). Preferential flow is represented by a simplified, effective approach similar to the 1-D approach of Zurmühl and Durner (1996). However, while these authors used a bimodal function to account for high unsaturated conductivities at high water saturation values, Zehe et al. (2005, 2006) used the field capacity

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of the soil as a threshold value to initiate preferential flow. If connected macropores are present and soil water saturation exceeds field capacity total hydraulic conductivity of the elements increase linearly with increasing saturation up to a maximum that is determined by the macroporosity factor. The latter is the ratio of the water flow rate in all
5 the macropores present in the cross section to water flow in the soil matrix continuum. Zehe and Blöschl (2004) showed that this effective approach to model macropore flow is appropriate for yielding simulation results (a) in good accordance with tracer experiments conducted at the plot and hillslope scales, and (b) showing a good reproduction of the observed rainfall-runoff behavior at the event and long time scales in the Wei-
10 herbach catchment.

The model was selected based on a comprehensive field data set described in Zehe et al. (2001), including soil hydraulic data, surface roughness data, land use and plant morphological data, and observations of precipitation and meteorological forcings that were gathered for more than 15 years. The basic model element was a typical hillslope
15 structure with Calcaric Regosol soils in the upper reaches and Colluvisol in the lower 20%, which is a reasonable average representation of the soil catena observed in the Weiherbach valley. The catchment was represented by 169 of these hillslopes and the inter-connected river network. The widths of surface hillslope raster elements were dependent upon the land use and soil pattern between 50 and 100 m. Guided by a
20 survey of earthworm burrow densities at 15 distributed field sites, the macroporosity factor was increased from hilltop to the valley floor by a constant proportion. The latter was derived based on the total volume of macropores observed at these sites. The average density of connected macropores of the hillslope had, however, to be estimated from test simulations, as this parameter was not observable at this scale. As
25 shown in Fig. 9, an average macroporosity factor of 2.1 was necessary for matching the largest flood event observed in this catchment. Hence, the maximum value of modeled infiltration rates is 2.1 times the saturated hydraulic conductivity. A model structure without macropores overestimates flood response by a factor of almost four. But even if we keep the average macroporosity as it is, and flip the spatial pattern

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upside down, i.e. arrange a high number of macropores in upper hillslope reaches and a low macropore density at the foothill sector, catchment scale runoff response is completely different (graph not shown here, cf. Zehe et al., 2006, for more details). Hence, the spatial pattern of macropore density is indeed a first order control.

5   The next step was to use this model to shed light on the question of how well we may predict the response of the Weiherbach catchment to an extreme rainfall event, if very detailed information about the spatial pattern of initial soil moisture is at hand. To this end Zehe et al. (2005) estimated the initial soil moisture state before a large rainfall event observed in August 1995 (compare Fig. 11 upper right panel) based on 61 point
10   observations using geo-statistical methods. In the next step they generated initial soil moisture patterns with identical estimates of mean and variance, and the covariance structure for the Turning Bands synthetic generation method for spatial random fields, and conditioned them to exactly reproduce the initial soil moisture values observed at 61 TDR stations. Hence, all these realizations were apparently identical as they are not
15   distinguishable with respect to the amount information that could be extracted from the 61 point observations. Nevertheless, the various realizations of catchment responses were not at all identical due to the local statistical fluctuations at those locations where no soil moisture observations were at hand.

    Zehe et al. (2005) showed that this inherent uncertainty causes a strong scatter in the
20   simulated flood response when the spatial average initial soil moisture of the catchment is in the vicinity of the "local" process threshold (in this case, the field capacity) for the soil to switch on enhanced infiltration rate due to macropore flow. The maximum scatter in the predicted flood peaks was larger than $1\,m^3/s$, which corresponds to an uncertainty range of more than 30% (Fig. 11, lower left panel). This is 6 times larger
25   than the uncertainty of the initial soil moisture pattern, which was around 5% only. During the wet and dry initial conditions "far away" from the process threshold, the uncertainty of the initial soil moisture pattern did not have such a significant influence on the simulated flood response (Fig. 11, upper left and lower right panel). The scatter of the simulated flood peaks was smaller than 5% of the average response.

The reason for the much stronger scatter of flood responses in the vicinity of the local process threshold is that the correlation between the different realizations of the initial soil moisture pattern and the spatial pattern of macroporosity fields is slightly different. Weaker flood responses were caused by soil moisture patterns that had a
5 weakly positive correlation of 0.3 with the spatial pattern of macroporosity, whereas in the case of stronger flood response the patterns of macroporosity and initial soil moisture were negatively correlated. Note that all realizations were identical at the locations of the 61 TDR and still un-resolved differences in the pattern of initial states did have a drastic influence on the flood response of the catchment, when the system
10 was on average close to the local process threshold.

4.2.2 Modeling the temporal occurrence of threshold events

In the present study we have highlighted several times the key importance of preferential flow as a point scale threshold process as well as in the context of micro- and meso-scale threshold responses of catchments and hillslopes. The modeling approach
15 introduced in Sect. 3.2.2 is aimed at understanding how the spatial pattern of the onset of preferential flow interacts with the spatial patterns of initial soil moisture and precipitation when the catchment is responding to a single extreme rainfall event. The modeling approaches discussed in this section are aimed at explaining the temporal occurrence of these threshold responses to a long term rainfall forcing, i.e. it deals with
20 level A predictions, as outlined in the introduction. Rainfall is itself intermittent; however, not every rainfall event will trigger a preferential flow event, hence preferential flow events are themselves highly intermittent. The question is now whether the temporal pattern of the occurrence of preferential flow events will follow the pattern of rainfall forcings or not. This is of highest importance, for example, in the context of hazard
25 assessment of pesticide contamination of groundwater resources.

To answer this and other related questions McGrath et al. (2007) and Struthers et al. (2007a, 2007b) employed much simpler models of threshold behavior (as compared to those introduced in the last two sections), to explore the triggering of threshold

events subjected to synthetic rainfall time series. McGrath et al. (2007) introduced a spatially lumped model to gain further insights into the temporal occurrence of threshold events, as a function of the type of threshold, i.e. whether it is capacity controlled or intensity controlled, and as a function of the statistics of storm properties and climate
5 settings. They examined two simple conceptualizations of runoff triggering: an infiltration excess mechanism based on a rainfall intensity threshold while neglecting any soil moisture storage controls, and the opposite extreme, a saturation excess mechanism based on the existence of a storage threshold that depends only on the rainfall event depth and not its intensity. The latter approach is therefore able to capture the carry-
10 over of storage from one rainfall event to the next. As a result there is an enhanced probability that a second preferential flow event will occur shortly after a preferential flow event had just occurred because a higher value of soil moisture storage is more likely. McGrath et al. (2007) found that the rainfall intensity threshold produces an inter-event time distribution which is a filtered version of the inter-event time distribu-
15 tion, conditioned by the probability of the intensity threshold being exceeded during any single event. In contrast, the soil moisture storage threshold leads to temporal clustering of threshold events, for example, in the case of leaching of pesticides, which can have deleterious consequences for groundwater contamination. Furthermore, they also found that the mean and the variance of inter-event times increase with increas-
20 ing climate aridity. However, for systems with a deep storage capacity the coefficient of variation of inter-event times, which is a measure of the temporal clustering, was a maximum for an aridity index of approximately 1 (unity).

4.2.3 Percolation theory to model threshold behavior of hillslope subsurface stormflow response

25 Lehmann et al. (2007) used percolation theory to set up a hillslope model that is capable of reproducing threshold behavior observed in the subsurface stormflow response of the Panola hillslope (cf. Sect. 2.2.2). The idea of percolation theory is to discretize the hillslope into a two dimensional lattice where each cell has only two states, de-

noted as occupied or non-occupied. So-called bonds connect the sites of the lattice. Lehmann et al. (2007) generated the average number of connections per site (termed the coordination number) by a random process. An occupied site is a location with a transient water table at the soil bedrock interface and a bond corresponds to a flow path
5 of high conductivity, such as a lateral preferential pathway or a pipe. Two occupied and connected sites are deemed conducting with respect to subsurface flow. The state of a site is regarded as dependent upon the soil depth, porosity, hydraulic conductivity, and the initial state. Occupied sites connected by a bond are said to form a cluster. The fraction of occupied clusters (corresponding to the connected areas with a free water
10 table) is called the occupation probability p.

If p increases to a certain threshold, p_c, named the percolation threshold, there is a connected cluster that spans the whole hillslope system and subsurface flow response is then deemed to be initiated. This state corresponds to the state of the real hillslope where subsurface flow paths in the hillslope are connected by a free water table (cf.
15 Sect. 2.2.2 or Tromp-van Meerveld and McDonnell, 2006b) and the hillslope responds to rainfall with shallow subsurface flow response. Further details on the statistical assignment of hydraulic properties to the sites and the description of the flow processes can be obtained from Lehmann et al. (2006). These authors showed that their percolation model could match the observed threshold behavior observed at Panola, with
20 subsurface stormflow commencing, on average, when rainfall depth exceeds 55 mm (Fig. 5). Also the magnitude of the observed storm events triggering subsurface flow response fell within the uncertainty range of their Monte Carlo simulations. As subsurface stormflow response in the model starts when the percolation threshold is exceeded, i.e., there is a connected cluster that spans the whole system, the model
25 results lend support to the hypothesis that global connectivity of the free water table is the first order control of the onset of subsurface stormflow response of the hillslope.

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4.2.4 Summary of modeling approaches

In this paper we have presented three alternative approaches to modeling of threshold driven processes, responses and hydrological system functioning. The modeling approach of Zehe et al. (2005, 2006) is of the bottom-up type, where a detailed distributed model based on the Richards equation was extended to include the effects of macropores at the process level. This modeling was supported by the availability of extensive data on the number, density and distribution of macropores in the study catchment. Nevertheless, the authors found that there remained considerable uncertainty due to the presence of considerable unmapped heterogeneity of soil properties and soil structure, including macropores.

It needs to be realized that the wealth of detailed information on soil structure and other forms of heterogeneity will normally not be available for most catchments of the world, even if it is clear that threshold nonlinearities abound, and have a significant impact on hydrologic processes or responses, and will lead to uncertainty of predictions. The approach of McGrath et al. (2007) and Struthers et al. (2007a, 2007b) is built on an explicit acknowledgement of this fact, whereby they resort to the use of simple, parsimonious models that only purport to make level A predictions, i.e., the occurrence and timing of threshold exceedances, and do not yet aspire to make level B predictions, i.e., responses. This apparently simple-minded analysis has nevertheless led to profound conclusions; in the case of pesticide contamination of groundwater through preferential flow pathways, McGrath (2007) showed conclusively that the first order control on risk of contamination is in fact no more than fine time-scale precipitation data! This "pattern dynamics" approach has been previously successfully adopted in seismology for the prediction of the timing and location of earthquakes (see Rundle et al., 2006).

The percolation threshold approach of Lehmann et al. (2007) goes one step further, and does away with any pretence to physically based hydrologic predictions. Instead, this modeling approach is a purely patterns based approach, and aims to mimic the physical behavior through a statistical representation of the underlying patterns, in this

case patterns of connectivity, leading to simple conceptual models that nevertheless mimic some key aspects of observed behavior.

There could of course be other modeling approaches that may be adopted to deal with threshold behaviors, including a combination of the approaches that were presented here. One approach that was not highlighted here, but is promising, is to focus the possible ecosystem function of the threshold behaviors, and in this way avoiding the explicit characterization of the associated heterogeneities, but by replacing them by an an organizing principle underpinning the ecosystem function (Christoph Hinz, personal communication). This is left for further research.

5   Conclusions

The aim of this paper has been to illustrate through examples dynamics of hydrological systems and geo-ecosystems which are influenced by threshold behavior occurring at a variety of space and time scales. Based on well known characteristics of elementary threshold phenomena we suggested suitable criteria for detecting threshold behavior in hydrological systems. The most important one is intermittence of phenomena, i.e., the rapid switching of related state variables/fluxes from zero to finite values, or the existence of characteristic behavior regimes where the same process/response appears qualitatively differently at the microscopic and macroscopic levels. From the literature we provided several examples for intermittent hydrological phenomena, ranging from overland flow production in different landscapes, which includes the effects of hydrophobicity, to soil water flow that may occur in the matrix continuum or in preferential pathways, including cracking soils, highly nonlinear subsurface storm flow response of hillslopes during severe rainfall events, and finally to long-term catchment flood responses and behavior. From our elementary example of threshold behavior in the tea pot we learned that different dynamic regimes occur when the external forcing exceeds a threshold value and that in the case of the tea pot, energy dissipation is more efficient in the new regime. Fast overland flow generation, the switch from matrix to preferential

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flow, and the onset of subsurface stormflow can all be seen as switches of the system to a more efficient forms of dissipating the energy in the incoming rainfall as flow processes to areas of low potential energy and redistribution processes in the subsurface get much faster.

5 As threshold phenomena are quite often associated with environmental hazards such as floods, soil erosion and contamination of shallow groundwater resources, it is of key importance to predict whether they might occur at all, and if so, then their intensity. The first will be difficult in an "unstable range" of states as it depends on both the expected rainfall forcing and the accuracy of our data/knowledge on the macro-
10 scopic state of the system. When, with respect to the accuracy of observations, the system is close to the threshold, it is then difficult to predict whether or not it will switch, for instance within the predicted rainfall forcing. In contrast, when the system is sufficiently far away we may more confidently neglect the question of whether or not the system will switch to the other regime. Predicting the onset of threshold phenomena
15 requires, therefore, a thorough understanding of the underlying controls. In this context we showed that threshold behavior in hydrological systems may manifest at range of levels, with the complexity of the underlying controls and of the interacting phenomena that determine threshold behavior becoming more and more complex (Fig. 1). In this respect we distinguish:

20 – Threshold behavior at the process level, which manifests at the point scale during events. Good examples are overland flow production with and without the influence of hydrophobicity, or the switch from simple to preferential infiltration, that is, as argued in Sect. 2, controlled by threshold values for soil moisture and rainfall intensity/amount.

25 – Threshold behavior at the response level, which manifests at the hillslope and catchment scales, also during events. Good examples are subsurface storm flow response, overland flow response or sediment export from hillslopes and catchments. These phenomena are more complex as they are composed of interac-

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tions of more than one process. Furthermore, these phenomena occur if patterns of rainfall forcing, state variables and also static properties such as land use fulfill strict conditions such as global connectivity of flow paths or a positive spatial correlation between different patterns (Sect. 2.2).

5 – Threshold behavior at the functional level, which manifests through non- stationarity of indices that characterize hydrological behavior of a geo-ecosystem, in a statistical sense. Examples are trends or even sudden breaks in the flood frequency curve or the sediment export rates, as discussed in Sect. 2.3 for semi-arid runoff-run-on systems. Threshold behavior at the functional level is strongly related to
10 the resilience of geo-ecosystems, which is determined by multiple biologically mediated feedbacks between biotic and abiotic system properties. Examples are soil structures/ preferential flow paths that are due to plant roots in the context of vegetation patterns in semi-arid areas, or patterns of anecic earthworm burrows in humid areas.

15 What are the common implications for hydrological science but also for hydrological practice? The onset of intermittent processes or responses, in the vicinity of a threshold, is difficult to predict, as already discussed above. Uncertainties in observed initial and boundary conditions will be "amplified" in the sense that we cannot confidently predict whether the threshold will be triggered or not. The second implication is the
20 uncertainty of hydrological observations themselves. In the vicinity of a threshold a single observation is not appropriate to characterize how a hydrological system will react under given observed initial and boundary conditions. For example, Lischeid et al. (2000) observed tracer velocities of between 30.6 and 10.6 m/d during three identical steady-state field scale breakthrough experiments at the Gårdsjön test catchment.
25 The differences could not be explained by any measurable difference in the experimental conditions. Because of the limited observability of states the system can give a different response when the experiment is repeated under apparently identical conditions. Thus, the repeatability of observations and, therefore, our understanding of

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cause and effect, are limited in threshold systems. In the same vein, Beven (1996, p. 260) states that we cannot expect hydrological models to predict more accurately than the repeatability of nature herself!

This "unstable range" of system states can be reduced by increasing the accuracy of measurement technology. Promising technologies to observe proxies of soil moisture patterns are, the ground penetrating radar (GPR) (Binley et al., 2002; Roth et al., 2004), electrical resistivity soundings (ERT) (Kemna et al., 2002). The former yields the pattern of the dielectric permittivity, the latter the pattern of the apparent specific resistivity in the subsurface. A better characterization of the soil moisture pattern, especially connectivity, could be very useful for improved predictions of subsurface stormflow. The big drawback of both methods is that they are restricted to field campaigns and therefore provide only a coarse temporal resolution. Another promising development is the combination of weather radar and rain gauge data for a better characterization of the precipitation pattern (Ehret, 2002).

Predictions of threshold behavior at the functional level seem much more difficult (Groffmann et al., 2006; Saco et al., 2006) as they are strongly related to resilience of geo-ecosystems and only show up within long-term changes in functional indices. One important message to pick up again from the teapot example is that a switch to a different regime seems to occur when a different/most efficient way of energy dissipation is needed in case of high energy input into the system. Geo-ecosystems are certainly not that simple. However, we think that the architecture of a geo-ecosystem, the patterns of soils and vegetation and other biota collectively represent a configuration that is well adapted to the climate forcing and the geological settings. This configuration determines the hydrological functioning of a landscape, i.e., how energy input due to precipitation and radiation is redistributed in the landscape and translates into hydrological responses and into biomass production, and is stabilized by multiple feedbacks (Saco et al., 2006). However, we cannot answer the question whether the triggering of a functional threshold leads to yet another functional architecture, with an even better adapted way of redistributing and dissipating energy input. Only further and sustained

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research on the role of threshold behavior in hydrological systems will provide the answers to such questions.

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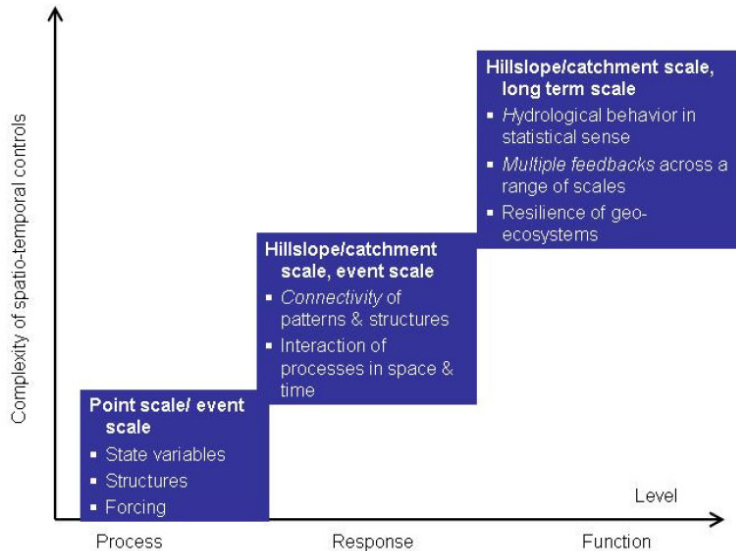| Abstract | Introduction |
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Fig. 1. Different levels of threshold behavior and complexity of the underlying controls.
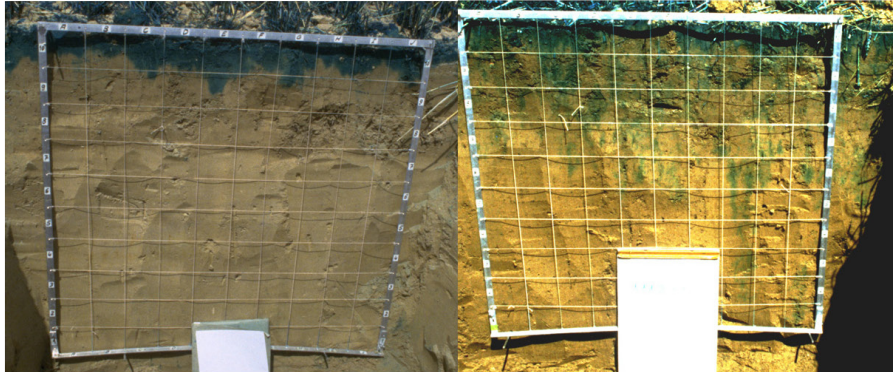
Fig. 2. Dye tracer distributions in soil profiles observed during sprinkling experiments in the Weiherbach catchment (Zehe and Flühler, 2001). Both plots were irrigated with approximately 25 mm in 2 h and at both sites connected earth worm burrows were present. However initial soil moisture at the site shown in the right panel was with 0.25 m^3 m^{-3} clearly higher than at the site of the left panel (0.20 m^{3}m^{-3}).
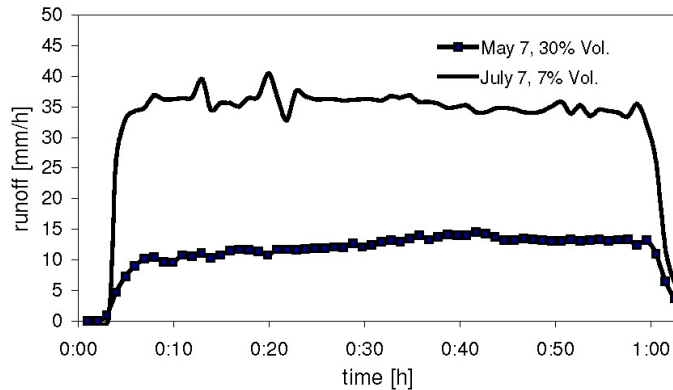
Fig. 3. Surface runoff in response to irrigation of the same plot during wet antecedent soil moisture conditions of 30% Vol. in the upper soil (7 May) and dry antecedent soil moisture conditions (7 July) of 7% Vol.
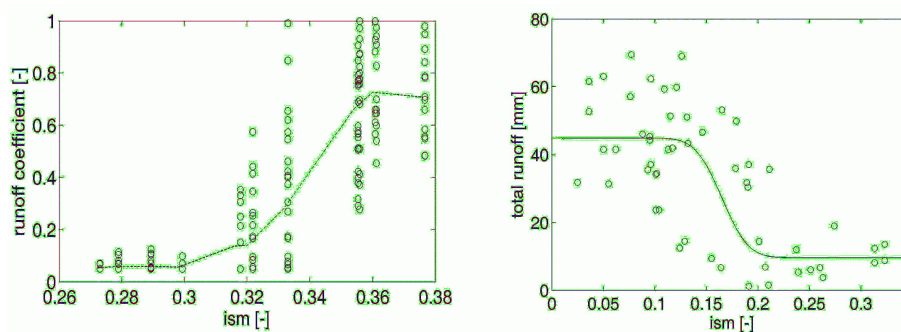
Fig. 4. Observed overland flow response from 53 irrigated field plots plotted against the initial soil water content (right panel). Observed catchment scale runoff response plotted against the average initial soil moisture of the catchment estimated based on process model simulations (left panel). Please note that the scales and the processes are different, however, the principal patterns of the scatter plots are the same. This highlights the similarity of threshold phenomena with respect to the existence of different regimes where runoff generation/response is clearly different and an unstable range where the system switches between both regimes.
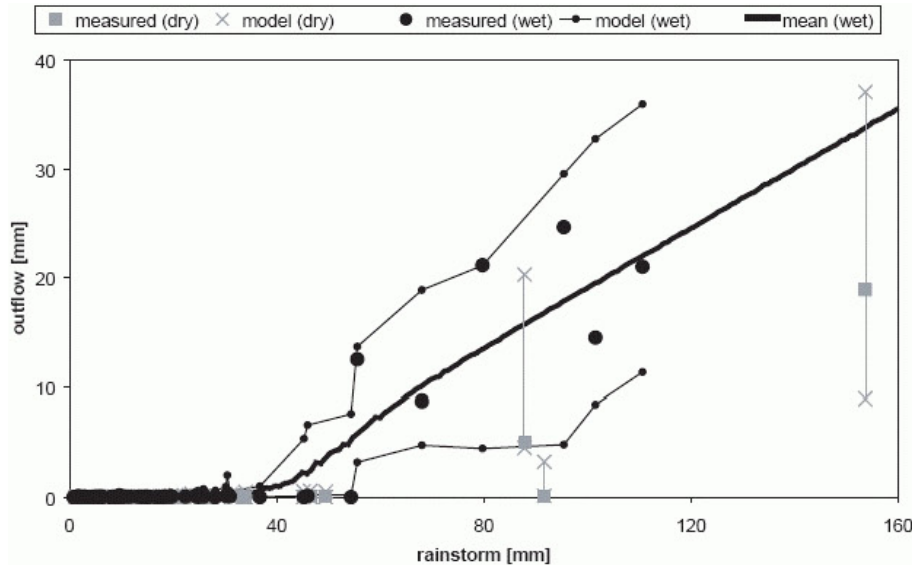
Fig. 5. Subsurface storm flow response plotted against the rainfall forcing (Figure is taken from Lehmann et al., 2007) as well as hillslope response predicted with the percolation model. The solid black line is the average response of 100 model realizations, the thin lines mark the uncertainty ranges.
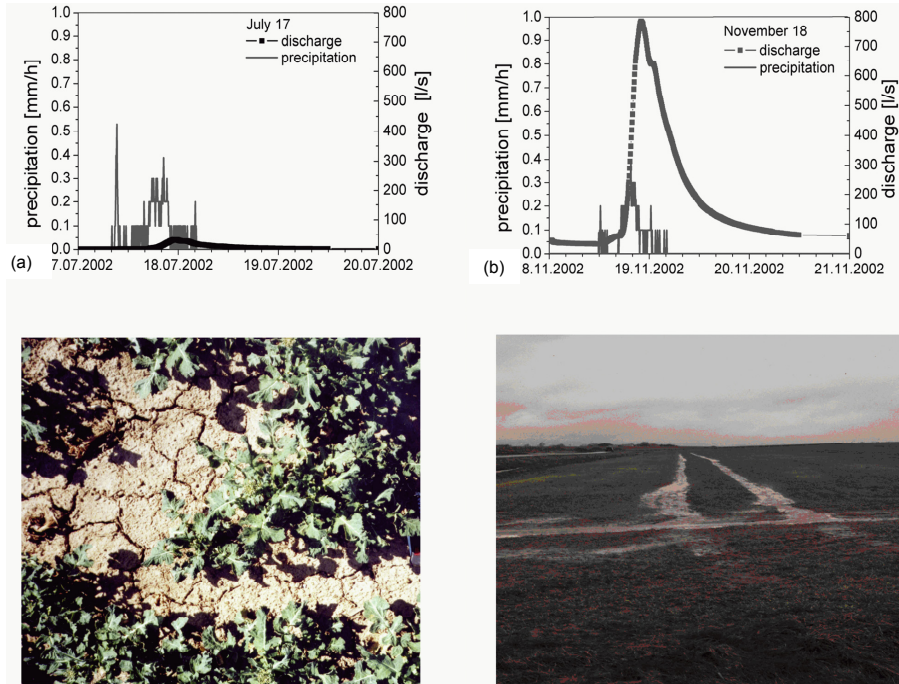
Fig. 6. Rainfall runoff response observed in the Tannhausen catchment (2.3 km^2) for two events in 2002. Event precipitation and antecedent precipitation of the summer event (left) were 18.3 and 20.3 mm, respectively; the corresponding parameters of the winter event (right) were 14.1 and 23.3 mm. Although the precipitation totals of the two events were similar, the runoff response was vastly different. The lower left and right panels show cracked soil in the Tannhausen catchment during summer in July, and overland flow during a flood event in November at the same site in July 2002.
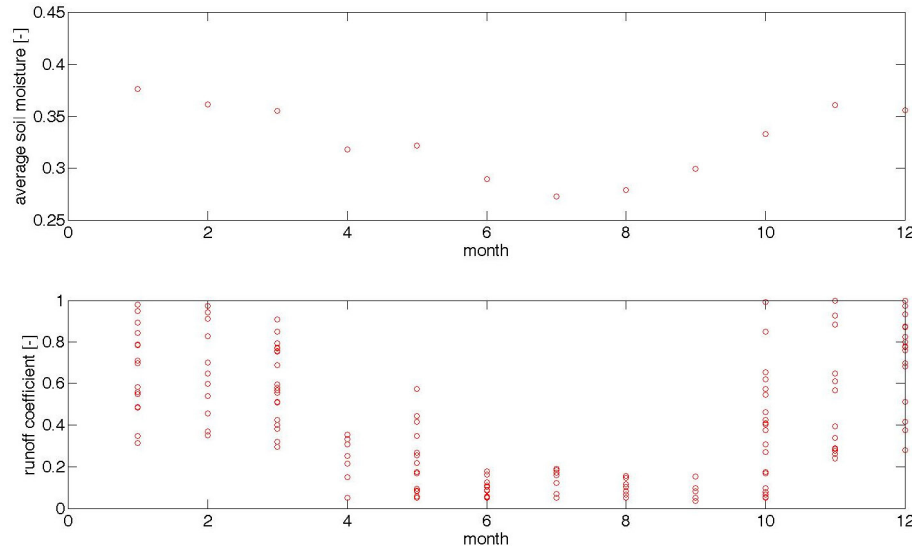
Fig. 7. The upper panel shows the annual regime of catchment average initial soil moisture simulated with CATFLOW as described in Zehe et al. (2007). The lower panel shows the annual regime of observed runoff coefficients in the Tannhausen catchment.

(a)　　　　　　　　　(b)
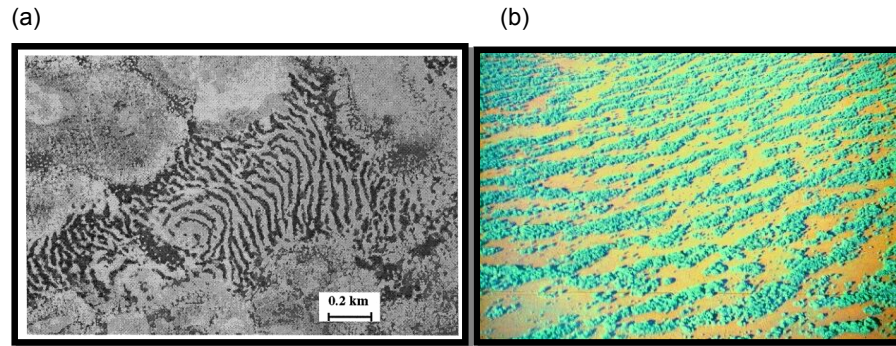
Fig. 8. Observed (optimal) vegetation patterns in semi-arid run-off-run-on ecosystem: **(a)** Tiger bush in Niger, Africa (Valentin et al., 1999), and **(b)** banded pattern in Northern Territory, Australia (www.nt.gov.au/ipe/pwcnt).
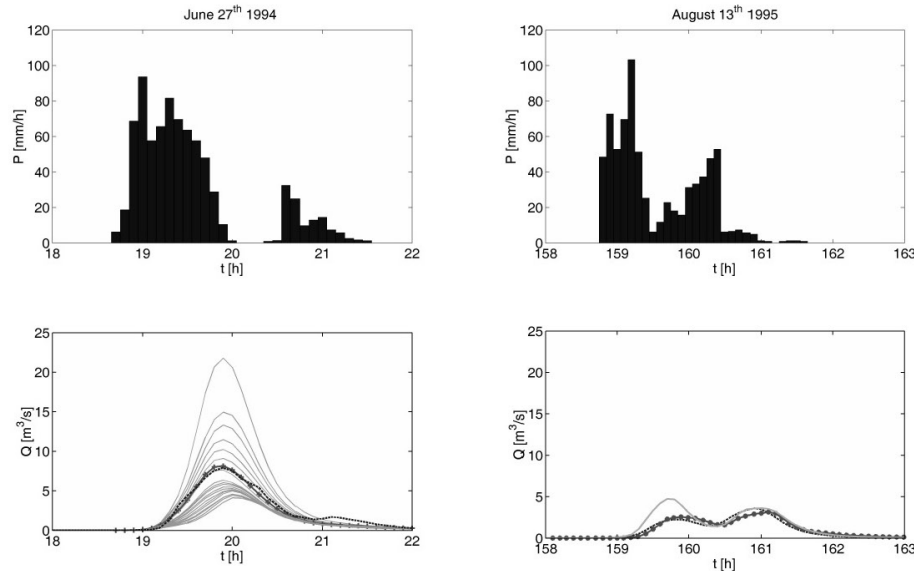
Fig. 9. Simulated flood response for different average macroporosity values (lower left panel) for the largest rainfall runoff event observed in June 1994. The dashed line is the observed hydrograph, solid gray lines simulations for different areal densities of connected macropores the solid line with crosses the best fit for a macroporosity of 2.1. The right panel shows the simulation of the second largest flood event observed in August 1995. Dashed line is the observation, solid gray line the simulation with a macroporosity of 2.1, the solid line with crosses a simulation with increased macroporosity of 3.2. Please note that the best fit and the observed hydrograph match too well so that they cannot be separated at that scale.

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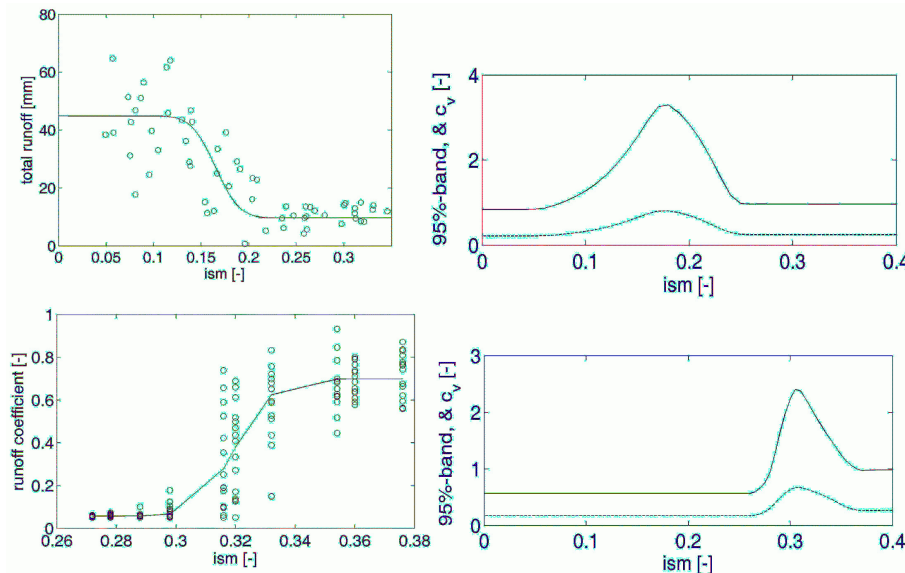
Fig. 10. Simulation results obtained with the statistical model for threshold behavior (taken from Zehe et al., 2007). The upper left and right panels show overland depths and their scaled 95% confidence interval as a function of the initial soil moisture ism simulated for the 53 sprinkling experiments on water repellent soils in the southern Alps (cf. Sect. 2.1.3 and Fig. 4). The lower left and right panels present event runoff coefficients of the flood response and their scaled 95% confidence interval plotted against the initial soil moisture that were simulated for the Tannhausen catchment with cracking soils (cf. Sect. 2.2.3 and Fig. 4.).
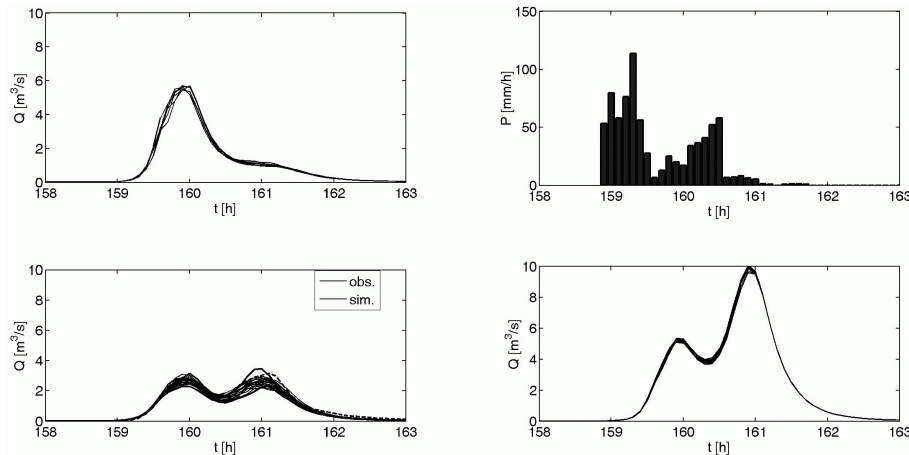
Fig. 11. Catchment scale flood response to an extreme rainfall event of 80 mm in 4 h (upper right panel) simulated for apparently identical initial soil moisture states at dry (upper left panel), intermediate (lower left panel) and wet (lower right panel) average soil moisture conditions of the Weiherbach catchment (taken from Zehe et al., 2005). In the intermediate soil moisture state that is close to the threshold that could trigger enhanced infiltration, the inherent uncertainty of the observation causes a clearly higher scatter in predicted hydrographs. Catchment scale flood response is therefore not able to be predicted well.