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Hydrology and Earth System Sciences

Threshold behaviour in hydrological systems as (human) geo-ecosystems: manifestations, controls, implications

E. Zehe[1] **and M. Sivapalan**[2,3]

[1]Institute of Water and Environment, Department of Hydrology and River Basin Management, Technische Universität München, Arcisstrasse 21, 80330 Munich, Germany
[2]University of Illinois at Urbana-Champaign, Departments of Geography and Civil and Environmental Engineering, 220 Davenport Hall, 607 South Mathews Avenue, Urbana, IL 61801, USA
[3]Department of Water Management, Faculty of Civil Engineering, Delft University of Technology, P.O. Box 5048, 2600 GA Delft, The Netherlands

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Abstract. In this paper we review threshold behaviour in environmental systems, which are often associated with the onset of floods, contamination and erosion events, and other degenerative processes. Key objectives of this review are to a) suggest indicators for detecting threshold behavior, b) discuss their implications for predictability, c) distinguish different forms of threshold behavior and their underlying controls, and d) hypothesise on possible reasons for why threshold behaviour might occur. Threshold behaviour involves a fast qualitative change of either a single process or the response of a system. For elementary phenomena this switch occurs when boundary conditions (e.g., energy inputs) or system states as expressed by dimensionless quantities (e.g. the Reynolds number) exceed threshold values. Mixing, water movement or depletion of thermodynamic gradients becomes much more efficient as a result. Intermittency is a very good indicator for detecting event scale threshold behavior in hydrological systems. Predictability of intermittent processes/system responses is inherently low for combinations of systems states and/or boundary conditions that push the system close to a threshold. Post hoc identification of "cause-effect relations" to explain when the system became critical is inherently difficult because of our limited ability to perform observations under controlled identical experimental conditions. In this review, we distinguish three forms of threshold behavior. The first one is threshold behavior at the process level that is controlled by the interplay of local soil characteristics and states, vegetation and the rainfall forcing. Overland flow formation, particle detachment and preferential flow are examples of this. The second form of threshold behaviour is the response of systems of intermediate complexity – e.g., catchment runoff response and sediment yield – governed by the redistribution of water and sediments in space and time. These are controlled by the topological architecture of the catchments that interacts with system states and the boundary conditions. Crossing the response thresholds means to establish connectedness of surface or subsurface flow paths to the catchment outlet. Subsurface stormflow in humid areas, overland flow and erosion in semi-arid and arid areas are examples, and explain that crossing local process thresholds is necessary but not sufficient to trigger a system response threshold. The third form of threshold behaviour involves changes in the "architecture" of human geo-ecosystems, which experience various disturbances. As a result substantial change in hydrological functioning of a system is induced, when the disturbances exceed the resilience of the geo-ecosystem. We present examples from savannah ecosystems, humid agricultural systems, mining activities affecting rainfall runoff in forested areas, badlands formation in Spain, and the restoration of the Upper Rhine river basin as examples of this phenomenon. This functional threshold behaviour is most difficult to predict, since it requires extrapolations far away from our usual experience and the accounting of bidirectional feedbacks. However, it does not require the development of more complicated model, but on the contrary, only models with the right level of simplification,

Correspondence to: E. Zehe
(e.zehe@bv.tum.de)

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which we illustrate with an instructive example. Following Prigogine, who studied structure formation in open thermodynamic systems, we hypothesise that topological structures which control response thresholds in the landscape might be seen as dissipative structures, and the onset of threshold processes/response as a switch to more efficient ways of depleting strong gradients that develop in the case of extreme boundary conditions.

1 Introduction

We recognise threshold behaviour from our common experiences boiling water in a kettle. Thermal energy enters the water from across the kettle's bottom and sets up a vertical temperature gradient. In the early phase of heating molecular diffusion is sufficient for transporting the heat upwards, dissipating it and depleting the temperature gradient in the kettle. When the stove temperature and consequently the temperature gradient increase further, convection cells begin to form (Nicolis and Prigogine, 1977; Prigogine and Stengers, 1984), as these allow for a more efficient energy dissipation at the higher temperature gradient than does molecular diffusion. When the vertical temperature gradient increases even further, above a second threshold, turbulent eddies begin to form (Nicolis and Prigogine, 1977; Haken, 1983) and water begins to "boil".

Although simple, this example sheds light on many important aspects of threshold behaviour: there are apparently different modes of dynamic behaviour, which are qualitatively different at the "macroscale", including the more efficient mixing/depletion of gradients/dissipation of energy. These dynamic modes are not stable; when the stove's temperature is reduced the system switches back from turbulent mixing to convective mixing. Hence, the occurrence and stability of these dynamic modes depend upon a combination of (a) the boundary conditions (here energy inputs), (b) an internal threshold determined by system properties (here the molecular coefficient for heat diffusion and viscosity) and (c) the initial system state. Strikingly, the individual water molecules, at the microscopic level, do not "know" anything, neither about turbulence nor about eddies, they simply "move" under the prevailing temperature gradient and the corresponding dynamic mode. Speaking about the different modes of the dynamics is only appropriate at the macroscopic level. As suggested by Haken (1983) and Prigogine and Stengler (1984) the transition from laminar flow (in this case molecular diffusion) to turbulence (in this case turbulent eddies) is a process that involves self-organisation and positive feedbacks. It is difficult to predict the threshold values that determine these qualitative changes in system behaviour from analyzing just the behaviour of the individual water molecules at the microscopic level – although it is still possible in some unique cases, such as in the case of a laser (Haken, 1983) or chemical clocks (Prigogine and Stengler, 1984). In many cases they have to be derived empirically at the macroscopic level.

Threshold behaviour can be deemed as an extreme form of nonlinear dynamics when phenomena are intermittent and the related state variables/fluxes switch from zero to non-zero values (Blöschl and Zehe, 2005; McGrath et al., 2007) over a short time or space increment. Alley et al. (Alley et al., 2003) suggested that threshold behaviour can be defined as a rapid and sudden change in macroscopic dynamics, which occurs much faster than the typical time scales of the system and of the external forcing/boundary conditions. This sudden change in behaviour manifests itself either in the form of an activity/triggering event as in the case of earthquakes (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 McClain et al., 2003), or as a qualitative change in the macroscopic dynamics as in the case of a laser (Haken, 1983). When the threshold is crossed, and associated processes or responses become considerably faster or slower, and/or the thermodynamic mixing is much more or much less efficient (Kleidon and Schymanski, 2008; Zehe et al., 2009). Examples of elementary threshold phenomena include phase transitions between the liquid, solid and gas phases of an element, or from laminar to turbulent flow, or from emission of normal light to laser light (Haken, 1983). Generally, a threshold is crossed when a macroscopic state variable, or a ratio of macroscopic state variables, or the external forcing/boundary conditions rise above or drop below an empirical threshold value (with some uncertainty). For instance, turbulence in open channel flow occurs when the Reynolds number (i.e., the ratio of inertial forces to viscous forces) increases above a certain threshold value that depends on the fluid viscosity and a characteristic length scale. Or laser light, which is a new, highly coherent quality of light that results from stimulated light emission from atoms and molecules, and occurs when the rate of energy input (pumping rate) into a gas laser exceeds a threshold value.

Threshold behaviour manifests itself in many of the things we experience in day to day life, including many environmental phenomena. Several authors have suggested that threshold behaviour is of key importance for understanding and predicting the dynamics and stability of our climate system (Claussen, 1999; Pitman and Stouffer, 2006), the dynamics and resilience of geo-ecosystems (With and Crist, 1995; Wilcox et al., 2003a; Schröder, 2006; Emanuel et al., 2007; Saco et al., 2007), and the dynamics of fluxes in hydrologic systems (Beven and Germann, 1981; Woods and Sivapalan, 1999; Uhlenbrook and Leibundgut, 2002; Weiler and Naef, 2003a; Zehe et al., 2007). As suggested by Thom (Thom, 1977, 1989) in the context of catastrophe theory, and as will be discussed below, threshold behaviour drastically reduces our ability to make predictions at the level of (a) an individual process, (b) the response of larger units (e.g., hillslopes or catchments) that involve interactions of many processes, and (c) the long-term hydrologic functioning of complete geo-ecosystems. Detecting and/or understanding threshold behaviour in each of these cases is, therefore, a significant and important challenge to hydrological science, especially for predictions in the context of global change. The objective of this paper is to review and discuss many aspects of threshold behaviour in hydrological systems – viewed as closely coupled human geo-ecosystems – and related earth system sciences.

The main difficulty in investigating or predicting threshold phenomena in hydrological systems, and environmental systems in general, is that the underlying controls are often hidden and complex due to (a) possible multiple feedbacks between biotic and abiotic components (Saco et al., 2007), and (b) the fact that observations of internal states and boundary conditions are highly uncertain, especially at larger scales (Rundle et al., 2006; Zehe et al., 2007; Beven, 2006). Identification of cause-effect relations in the context of threshold behaviour is made difficult as a result (Zehe et al., 2007), especially when these are accompanied by structural or morphological changes as well (Newson, 1980; Newson and Newson, 2000; Phillips, 2004, 2006; Faulkner, 2008). Hence, threshold behaviour in hydrological and geo-ecosystems becomes much more difficult to *detect*, *understand* and *predict* in comparison to elementary threshold phenomena, which can be predicted on the basis of well observable dimensionless variables such as the Reynolds number. In response, this paper is organized around the following key questions:

1. How can we detect threshold behaviour and how does it manifest in hydrologic systems, viewed as coupled human geo-ecosystems?

2. What are the implications of threshold behaviour on the predictability of individual hydrological processes, and the hydrological responses of hillslopes and catchments?

3. Can we conceptualize different forms of threshold behaviour and understand their first order controls?

4. How can we model threshold behaviour and how accurately can we observe the patterns of the controlling state variables and boundary conditions?

5. Are there possible explanations for why hydrological systems exhibit threshold behaviour?

Section 2 addresses questions one, two and three by providing multiple examples of threshold behaviour in hydrology and related earth system sciences and through discussing common implications of threshold behaviour. Sects. 3–5 provide evidence to show that it is useful to distinguish threshold behaviour at the process level, response level and the "functional level", explain differences in the underlying process controls, and provide illustrative experimental and modelling studies. Questions four and five are discussed in the Conclusions section (Sect. 6).

2 Threshold behaviour: evidences, implications and manifestations

2.1 Examples of threshold behaviour in hydrology and earth system sciences

Threshold behaviour is ubiquitous in the environment around us in many forms. It is frequently discussed as influencing surface and subsurface runoff generation processes at the local (Horton, 1933; Dunne and Black, 1970; Dunne et al., 1991; Weiler and Naef, 2003a; Zehe and Blöschl, 2004), hillslope (Mosley, 1982; Bonell et al., 1990; Elsenbeer et al., 1994; Tani, 1997; Tromp-van-Meerveld and McDonnell, 2006a; Weiler and McDonnell, 2007), and catchment scales (Sklash et al., 1996; Buttle and Peters, 1997; McGlynn et al., 2002; Uhlenbrook et al., 2002; Deeks et al., 2004; Wenninger et al., 2004; McGuire et al., 2005; Zehe et al., 2005; Blume et al., 2008a, b). The interactions of antecedent wetness, rainfall intensity and depth with surface and subsurface structures are deemed as first order controls in each of these cases. Likewise, particle detachment and soil erosion is clearly a threshold process (Hicks et al., 2000; Salles et al., 2000; Hairsine et al., 2002; Shao et al., 2005; Maerker et al., 2008; Scherer, 2008; Ternat et al., 2008), which is controlled by rainfall intensity, shear stress due to overland flow and soil stability. Infiltration, vertical flow and transport of contaminants in field soils may be observed in two qualitatively different modes, namely in preferential pathways or in a slow form in the soil matrix continuum (Bouma, 1981; Beven and Germann, 1982; Edwards et al., 1989; Flury et al., 1994, 1996; Stamm et al., 1998; Zehe and Flühler, 2001a, b; Vogel et al., 2005; McGrath et al., 2007). Lateral preferential flow in soil pipe systems, which crucially determines hillslope scale subsurface runoff generation, is intermittent and is likely to be a threshold phenomenon as well (Adams and Parkin, 2002; Buttle and McDonald, 2002; Negishi et al., 2007; Weiler and McDonnell, 2007).

Threshold behaviour is also discussed in the context of the long-term development of soil structures and landforms (Newson and Newson, 2000; Phillips, 2004, 2006), fluvial morphology (Willgoose et al., 1991a, 1991b; Grant, 1997; Piégay et al., 2000; Fryirs et al., 2007; Harnischmacher, 2007), rill and gully erosion (Bull and Kirkby, 1997; Kirkby et al., 2003; Poesen et al., 2003; Shao et al., 2005; Parkner et al., 2006, 2007), badlands formation (CalvoCases and Harvey, 1996; Howard, 1997; Faulkner et al., 2000, 2003; Boardman et al., 2003; Faulkner, 2008), and formation and growth of channel networks (Willgoose et al., 1991a, 1991b; Rinaldo et al., 1996; Rodriguez-Iturbe et al., 1998; Rodriguez-Iturbe and Rinaldo, 2001). Threshold behaviour and feedbacks between vegetation patterns and redistribution of abiotic resources determine the resilience of geo-ecosystems in semi-arid regions (Grant, 1997; Piégay et al., 2000; Wilcox et al., 2003a; Wilcox and Newman, 2005; Fryirs et al., 2007; Saco et al., 2007; Tietjen and Jeltsch, 2007), grassland establishment (Baer et al., 2005; Baer and Blair, 2008), regeneration of forests in floodplain areas (Dufour and Piegay, 2008), plant diversity in wetlands (Ellison and Bedford, 1995; Bedford et al., 1999; Gusewell et al., 2005; Loheide and Gorelick, 2007), plant water stress (Emanuel et al., 2007), and species response to landscape structures (With and Crist, 1995; Schröder, 2006). Dynamics of the "climate system", which includes the atmosphere and ocean, marine and terrestrial biosphere, cryosphere, lithosphere and hydrosphere (Claussen, 1999), is also strongly affected by threshold behaviour at various spatio-temporal scales, as recently discussed by Pitman and Stouffer (2006).

To summarize, many types of threshold behaviour occur in our earth system at a wide range of spatio-temporal scales. However, the underlying controls are different for the different forms of threshold behaviour. Before we further elaborate on the different forms of threshold behaviour in hydrological systems and geo-ecosystems, and the underlying process controls, in the following section we will first discuss common implications and manifestations of threshold behaviour.

2.2 Implications of threshold behaviour for predictability

Rundle et al. (2006) and Blöschl and Zehe (2005) have suggested that *intermittence* of phenomena/processes in space and/or time is a characteristic feature of threshold behaviour in environmental systems, and poses problems for predictability. In general, predictions of intermittent phenomena are two-level mixed discrete-continuous problems: 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. Ever since the introduction of catastrophe theory (see for instance Thom, 1989) we have known that the accuracy of level A predictions of threshold phenomena (does it occur or not?) depends on the current state of the system and the expected forcing/boundary condition. We therefore suggest that we might detect the existence of threshold behaviour based on the intermittence of phenomena and the fact that predictability of system behaviour drastically decreases for certain combinations of states and expected boundary conditions (Fig. 1 upper panel), or in a range of states when a certain forcing is expected (Fig. 1 lower panel).

This is nicely explained using our tea kettle example. At the beginning, when the water from the water tap is at 12°C, say, we are sure that increasing the water temperature by just 1°C will not induce formation of convection cells nor turbulent eddies. The dynamic "mode" of energy mixing is stable. When we increase the water temperature to a state

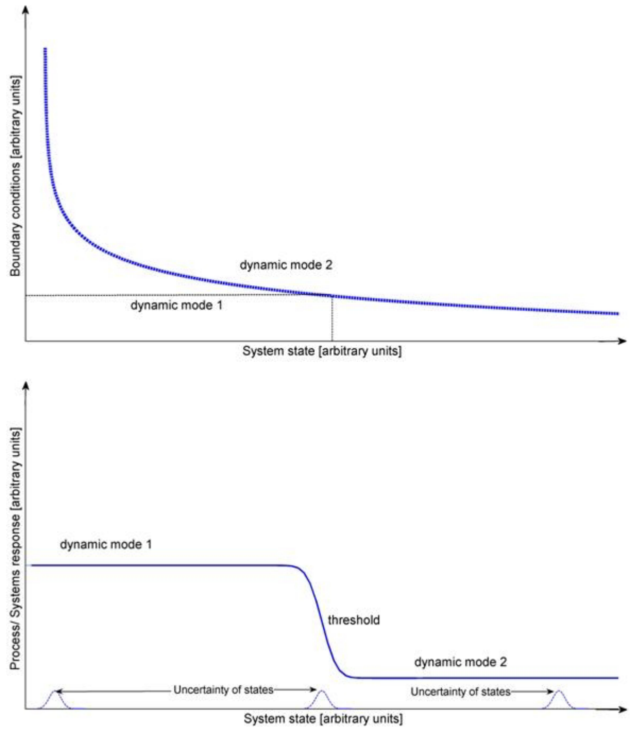

Fig. 1. "Sketch of a stability diagram" of a system with threshold dynamics and two dynamic modes (upper panel); the solid line marks combinations of states and expected boundary conditions for which the threshold is crossed (with uncertainty). The lower panel shows how the dynamic mode will change with different states for the case of fixed boundary conditions (marked in the upper panel). The distribution functions illustrate the uncertainty of state observations. If the state the uncertainty range of the state observations and the transition range start to overlap, predictability of the system is lowest (Zehe et al., 2007).

which is, for example, 1.05°C below the threshold needed for the formation of 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°C, 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 will happen. Small, non-observable differences within the error range of our measurements of the macroscopic systems state can determine whether convection cells will form or not and therefore we cannot accurately predict whether or not the energy dissipation will switch from the molecular regime to the convective regime (Thom, 1989; Haken 1983). Repeated, identical trials of the experiment will, therefore, lead to significant scatter in the observed macroscopic dynamics. We then say that predictability of the system is low at this critical state (Fig. 1b). If we increase the temperature to a value which is much larger than the threshold for eddy formation, the mode of energy dissipation becomes stable again, unless and until we come close to the next threshold value with respect to the uncertainty of our measurements of states and boundary conditions.

Thus, level A predictions – whether a threshold phenomenon/ process occurs or not – are most difficult and uncertain if the "system state" is in the vicinity of a threshold. Whether a range of states is "unstable" in respect of the expected dynamic mode depends on the expected forcing/boundary condition (Nicolis and Prigogine, 1977; Thom, 1977; Haken, 1983; Zehe and Blöschl, 2004; Zehe et al., 2007) and the accuracy that system states and boundary fluxes can be measured (Fig. 1 lower panel). In hydrology and related earth system sciences we often struggle with level A predictions: Will an expected rainfall event cause preferential transport of a pesticide into the subsoil where degradation processes are significantly slower (Bolduan and Zehe, 2006) or not? Will an expected rainfall event trigger lateral preferential flow/subsurface storm flow so that upslope parts of the landscape or groundwater will contribute to flood formation (McDonnell, 1990; Kirnbauer et al., 2005; Wenninger et al., 2004; Graeff et al., 2009), or not? Will an expected rainfall event cause erosion rills to connect to the main channel which in turn triggers gully formation (Faulkner, 2008), or not? Will an expected rainfall event trigger a landslide (van Asch et al., 1996, 1999; Boogard et al., 2002; Lindenmaier et al., 2005), or not? Will a higher disturbance due to enhanced grazing exceed the resilience of a savannah ecosystem (Rietkerk and van de Koppel, 1997; Kefi et al., 2007; Tietjen and Jeltsch, 2007), or not? To address such questions we must better understand whether the system of interest is in a critical state, which means whether the expected forcing/boundary conditions/ disturbance could be strong enough to push the system over the threshold (Fig. 1 upper panel).

Judgement of whether a system is in a critical state requires a detailed understanding of the first order controls of threshold behaviour. In hydrology and related earth systems sciences this is rather difficult to obtain because:

– Threshold processes deplete/work against the initial and boundary conditions that make up their cause (for instance they deplete gradients). Post-hoc identification of cause-effect-relations to explain when the "state" became critical is therefore very difficult, especially when morphological changes are also involved (Beven, 1996b).

– Our ability to perform observations under controlled identical experimental conditions – a crucial requirement to establish cause-effect relations – is poor because observations hydrological states are non exhaustive even at small scales (Zehe et al., 2007; Hills and Reynolds, 1969).

– Complexity of underlying controls increases when moving from simple hydrological systems to hydrological systems that include higher levels of complexity, including interactions and feedbacks (Dooge, 1986).

These issues are discussed through the use of examples in the next few sections.

2.3 Manifestations of threshold behaviour at different levels of complexity

Threshold behaviour has broad implications, including for predictability, as discussed in the previous section, and it occurs in hydrological systems at different scales, across different phenomena as well as in different contexts. Hydrologists speak about threshold behaviour in the context of investigating individual processes such as infiltration, but also in the context of overall system responses, such as the runoff response or sediment yields of whole hydrological systems such as hillslopes, catchments or larger river basins. We want to highlight the differences between these different forms of threshold behaviour, along with the complexity of the underlying controls. In this respect, we suggest that threshold behaviour in hydrological systems occurs at three different *levels of complexity*, as schematically illustrated in Fig. 2.

Threshold behaviour at the process level, i.e. when individual hydrological processes are intermittent. They either emerge/vanish as, for instance, overland flow (Horton, 1933; Dunne and Black, 1970; Beven, 2004), or show qualitative changes – when infiltration switches from slow water flow through the soil matrix continuum to faster flow through preferential pathways. The threshold character of these processes is strongly determined by nonlinear interactions of local (soil) structures/ properties with soil moisture as well as rainfall intensity and depth. In the following we will use the term "threshold process" as synonymous with threshold behaviour at the process level. Illustrative examples are provided in Section 3.

Threshold behaviour in the response of hydrological systems of intermediate complexity; according to Dooge (1986), such systems are too large to be treated in a fully deterministic manner and too small to be treated with first order statistical methods. The hydrological response of a system is, in general, dependent on a specific environmental problem context such as flooding, land degradation or contamination and is often characterised at the systems boundary. A hydrological system has many more degrees of freedom to react as compared to a (threshold) process because its response is composed of several processes that interact in *space and time*. Instructive examples are the runoff response of a hillslope or sediment export from a catchment. Crossing the threshold for local overland flow generation/ particle detachment is a necessary but not sufficient condition for the system to respond because upslope runoff may re-infiltrate and upslope eroded material may settle on the way downslope.

Whether and how the system will respond in each case is crucially dependent on *multivariate or even topological*

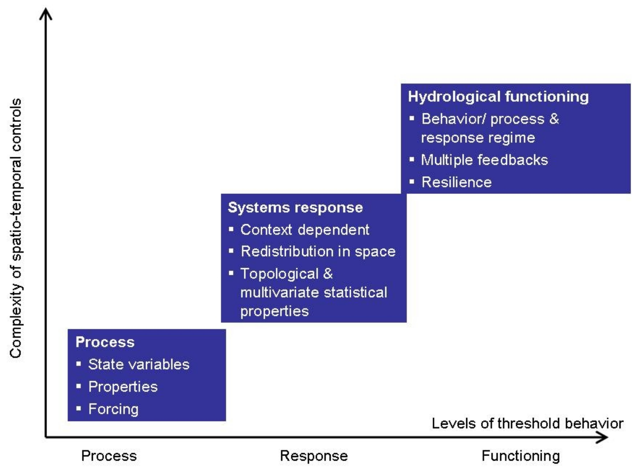

Fig. 2. Different levels of threshold behaviour and complexity of the underlying controls.

characteristics of several patterns (Schulz et al., 2006) – for instance topography, vegetation, soil properties, soil moisture – that determine the redistribution of overland flow and substances in space or the redistribution of mechanical stress in the case of an earthquake (Rundle et al., 2006). The response, and therefore the underlying controls, is more complex than at the process level, since they involve more degrees of freedom and the satisfaction of stronger conditions. As discussed in Sect. 4, the connectivity of surface (Hearman, 2008) or subsurface flow paths (Tromp-van-Meerveld and McDonnell, 2006a, b; Lehmann et al., 2007) appears to be very important in the case of the runoff responses of hillslopes and catchments. In the following we will use the term "threshold response" to be synonymous with threshold behaviour at the response level.

Drastic/threshold-driven changes in the functioning of hydrological systems as human geo-ecosystems i.e. the way they respond to rainfall and other climatic boundary conditions from a general or long-term perspective. Hydrological functioning can be characterized using common statistical indicators as, for instance, the flow duration and flood frequency curves, the monthly runoff regime, inter-annual variability of the water balance and – if available – patterns of sediment yield and hydro-chemistry. These indices encapsulate threshold behaviour at the process and response levels but in a statistical manner, including how patterns of topographic variables, soil structures, and vegetation patterns transform the energy and water/mass flows into the system into hydrological responses. Change of hydrological functioning – that may manifest through changes in any one of these indices – can either be in response to external climate change, or due to persistent, substantial often human-induced changes in geo-ecosystem properties that determine either local process thresholds or the topology/structures in the system and thus response thresholds. Illustrative examples are a die-back of earthworms and related substantial changes in infiltration properties and local bioturbation (Edwards et al., 1990b; Lavelle et al., 1997, 2004; Shipitalo and Butt, 1999; Milcu, 2005) the onset of gully erosion and badlands formation (Howard, 1997; Boardman et al., 2003; Boardman and Foster, 2008; Faulkner, 2008; Maerker et al., 2008), leaky vegetation patterns due to overgrazing in arid runoff–runon systems (Saco et al., 2007), or as a classic example, the results of past regulation of the upper Rhine river (Dister et al., 1990; Hofius, 1991). Again, the underlying control for this form of threshold behaviour is more complex when compared to the response level, since it involves long-term feedbacks between biotic and abiotic geo-ecosystem components or morphological processes, as well as substantial disturbances that go beyond the *resilience* of the system. In the following we will use the term "functional threshold" to be synonymous with the drastic changes that occur in the functioning of hydrological systems.

3 Threshold behaviour at the process level

3.1 The classics: overland flow formation and infiltration

3.1.1 Overland flow generation

Overland flow is the most prominent example of an intermittent hydrological process. 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. Saturation excess runoff, on the other hand, is a capacity (i.e., volume) controlled threshold process (McGrath et al., 2007). It occurs when the entire pore volume within a soil column is saturated with water from incoming precipitation added to previously stored water (Dunne and Black, 1970). It dominates surface 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; Beven 2004). The latter is strongly governed by the unsaturated hydraulic conductivity and thus soil water content. Hence, both mechanisms of overland flow generation are soil moisture controlled threshold processes. 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.

3.1.2 Infiltration and vertical preferential flow

Infiltration and soil water flow occur in qualitatively different forms, such as preferential flow and matrix flow. The term preferential flow was coined upon realising that water flow and solute transport in non-capillary soil structures were much faster than would be expected from classical theory of flow and transport in porous media (Beven and Germann, 1982; Germann, 1990; Roth et al., 1991). In coarse grained soils wetting front instability may lead to fingered flow, especially during conditions when water repellence is involved (Ritsema et al., 1998; Blume et al., 2008a). Or (Or, 2008) suggests that fingers start to form when the Bond number – which relates gravity, capillary forces and viscous forces, becomes smaller than 0.2.

In fine grained soils the existence of connective soil structures such as root channels, shrinkage cracks (Vogel et al., 2005a, b) and worm burrows (Edwards et al., 1990b; Edwards et al., 1992; Smettem, 1992; Shipitalo and Butt, 1999) are key pre-conditions for the occurrence of vertical preferential flow. These allow for, locally, up to 100–1000 times faster water fluxes than compared to matrix flow (Beven and Germann, 1982) because capillarity may be neglected in those structures and flow resistances in the direction of the driving potential gradient are much smaller. Flury et al. (1994, 1995) were among the first to develop effective methods to visualize infiltration and flow patterns in heterogeneous field soils by using dye tracer techniques. Today we know of numerous studies that have provided further evidence, demonstrating that that preferential flow in structured soils is indeed the rule rather than the exception (Fig. 3) (Zachmann et al., 1987; Kladivko et al., 1991; Roth et al., 1991; Flury, 1996; Mohanty et al., 1998; Stamm et al., 1998; Stamm et al., 2002). Earthworms are as classical ecosystem engineers of special interest in this context because they build long-lasting soil structures that exert a significant influence on water flows (Edwards et al., 1990b, 1992; Smettem, 1992; Shipitalo and Butt, 1999) and solute transport (Edwards et al., 1993; Zehe and Flühler, 2001a, b; Domínguez et al., 2004; Alekseeva et al., 2006; Le Bayon and Binet, 2006) but also on organic matter dynamics, pedogenetic processes and plant growth (Lavelle et al., 1997, 2004; Hedde et al., 2005; Milcu, 2005; Milcu et al., 2006).

However, whether or not connected structures are activated during a given rainfall event depends on the interplay of initial soil moisture and rainfall forcing. Weiler and Naef (2003b) suggested that the supply of rainfall is crucial for initiating and maintaining vertical preferential flow. They also highlighted that only a small number of macropores becomes active during most rainfall events. Zehe and Flühler (2001b) showed that after slope position as a proxy for soil type, initial soil moisture was the second most important variable to explain differences in dye tracer patterns observed in the Weiherbach catchment. Van Schaik (2009) found in a similar study that slope position, stoniness and texture were the most important explanatory variables. For reasons of brevity we omit a review of the approaches to model preferential flow; an excellent overview of this is presented in Simunek et al. (2003).

3.2 Threshold behaviour of process thresholds

3.2.1 Shrinking and swelling soils, overland flow and infiltration

In landscapes with cracking clay soils the opening/closing of cracks can cause an abrupt increase/decrease of soil infiltration capacity. Internal shrinkage 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 climate regions normal shrinkage is the most relevant mechanism, where volume reduction is approximately proportional to the water loss (Chertkov, 2000). When soil moisture drops below a threshold value cracks begin to develop and expand, which might enhance infiltration due to preferential flow (Bronswijk, 1988; Wells et al., 2003; Vogel et al., 2005a, b). 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 has been observed in vertisols in a catchment of northern Mexico (Navar et al., 2002), the Tannhausen catchment in Germany (Lindenmaier et al., 2006) (compare Sect. 4), and the Riesel Y-2 catchment in Texas (Allen et al., 2005).

3.2.2 Water repellency, overland flow and infiltration

Overland flow generation and infiltration in water-repellent soils may exhibit an even stronger threshold character, because the process threshold itself exhibits soil moisture dependent threshold behaviour. Water repellency of soils is a phenomenon that is caused by the presence of degraded organic matter (Krammes and Debano, 1965), e.g., by forest fires (DeBano 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). DeBano and Rice (1973) suggested that water repellency occurs after soil moisture drops below a certain threshold. Water repellency has been 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.

The potential for hydrophobicity can be determined by the Water Drop Penetration Time test, or 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

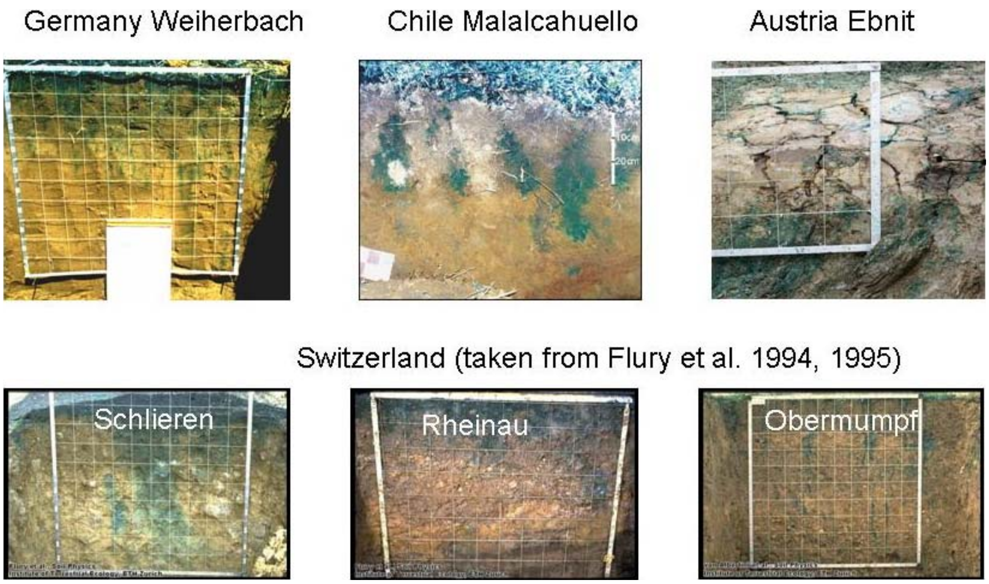

Fig. 3. Preferential flow patterns observed in Switzerland (taken from Flury et al., 1994), Austria (Wienhöfer et al., 2009), Chile (Blume et al., 2008) and the Weiherbach catchment.

scale for hydrophobicity that begins at zero, which indicates no hydrophobicity. Thus, hydrophobicity is a threshold phenomenon, which can display varying levels of intensity when it does occur.

Water repellency may favour formation of finger flow as suggested by (Dekker and Ritsema, 1994, 2000; Ritsema et al., 1998) for soils in the Netherlands, and by Blume et al. (2008a) for volcanic as soils in a pristine catchment in Chile. Zehe et al. (2007) performed 53 sprinkling experiments with 50 mm in one hour to shed light on the control of antecedent soil moisture on overland flow generation in a landscape comprising hydrophobic soils in southern Switzerland. They found significant differences in observed overland flow response depending on initial soil moisture state measured with TDR at each plot: strong overland flow of, on average, 45 mm for initial soil moisture states less than $0.11\,m^3\,m^{-3}$, weak overland flow response of, on average, 9.5 mm when initial soil moisture is greater than $0.21\,m^3\,m^{-3}$ and a transition region in between. In this "unstable range" the system appears to gradually change 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 (Fig. 4a, note the similarity with Fig. 1 lower panel).

Zehe et al. (2007) simulated repeated trials of these sprinkling experiments using a statistical model that describes threshold behaviour by assuming that the system response is characterised by two response functions, with constant moments in the stable ranges (Fig. 1b). Switching between the two response functions is achieved by assuming a fast nonlinear transition of the first and second moment of the response function. With this simple model they were able to reproduce the observed scatter in these sprinkling experiments (Fig. 4b) as well as the state dependent reproducibility of these experiments for an asymptotical high number of trials. They found that for a given forcing, reproducibility and predictability were lowest when the difference between the initial state and the values in the transition region was of the order of measurement uncertainty.

4 Threshold behaviour at the response level: hydrological systems of intermediate complexity

Dynamic responses of hydrological systems of intermediate complexity are controlled by multivariate statistical or topological/structural features of patterns that determine the surface and subsurface redistribution of water and substances within the system and exchanges with neighbouring systems (Schulz et al., 2006). Even if the process threshold of, for example, overland flow formation is crossed a substantial part of a hillslope overland flow may not reach the hillslope toe and thus the stream, because it may re-infiltrate on its way. These key topological properties, such as the presence of connected flow paths that link internal areas/volumes to the system boundaries, may change with overall average system states as, for instance, the top soil water content, as suggested for the Tarrawara catchment (Grayson et al., 1997; Western et al., 2001) or subsurface storage and thus the rainfall depth for subsurface stormflow (Mosley, 1982; Tromp-van-Meerveld and McDonnell, 2006a; Lehmann et al., 2007). Response thresholds such as the connectedness of flow paths to the

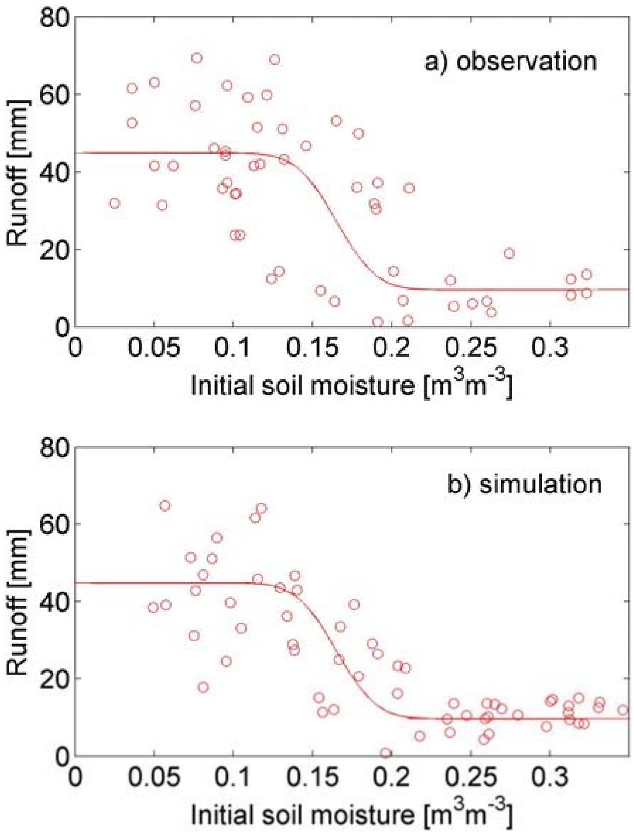

Fig. 4. Observed overland flow response from 53 irrigated field plots plotted against the initial soil water content (upper panel **a**). The lower panel **(b)** show the corresponding graphs simulated by the statistical model that assumes to stationary pdfs and non stationary transition range as inferred from the observations.

system's "outlet" or the mobilization of a high amount of pre-event water are controlled by key characteristics that depend crucially on the dominant process and typical landscape characteristics. The following sections will further elaborate these ideas based on illustrative field and model studies.

4.1 Threshold controls in lateral subsurface flows

Hillslope and catchment streamflow responses to rainfall events in forested mid-mountain reaches are often dominated not by Hortonian overland flow but by a combination of lateral subsurface flows and saturated overland flow (Buttle and Peters, 1997; Hoeg et al., 2000; Uhlenbrook et al., 2000; Buttle and McDonald, 2002; Freer et al., 2002; Uhlenbrook et al., 2002; McGuire et al., 2005; Weiler and McDonnell, 2007; Zillgens et al., 2007). There is considerable evidence that subsurface storm flow is an intermittent phenomenon that occurs after a threshold of rainfall depth is exceeded: for instance 35 mm of rainfall for a hillslope in Ohio as reported by Whipkey (1965), 20 mm of rainfall for field sites in New Zealand (Mosley, 1979) and Japan (1997) and 55 mm for a hillslope trench site in Georgia, USA (Tromp-van-Meerveld and McDonnell, 2006a). Many authors suggest that lateral subsurface flow is controlled by lateral subsurface structures. Beven and Kirkby (1979) found that in the case of shallow permeable soils, when bedrock topography might be assumed to be parallel to the surface, first order controls for saturated overland flow and subsurface flow are transmissivity and surface topography. This fundamental insight was condensed into the famous TOPMODEL concept. Networks of lateral preferential pathways/pipes are another likely explanation for subsurface storm flow, as suggested by several previous studies (Bonell et al., 1990; Elsenbeer et al., 1994; Sklash et al., 1996; Uchida et al., 2001; Buttle and McDonald, 2002; Weiler et al., 2003; Weiler and McDonnell, 2004; Negishi et al., 2007; Weiler and McDonnell, 2007).

Various authors suggest bedrock topography as a first order control for subsurface stormflow (Noguchi et al., 2001; Freer et al., 2002; McGlynn et al., 2002; Güntner et al., 2004; Weiler and McDonnell, 2004; Tromp-van-Meerveld and McDonnell, 2006b; Uchida et al., 2006; Negishi et al., 2007). Tromp-van-Meerveld et al. (2006a) introduced the *fill and spill* mechanism to explain bedrock control on observed threshold behaviour of subsurface storm flow at the Panola hillslope. They argue that during storms with rainfall depths smaller than the threshold value free subsurface water table may form within isolated areas of the hillslopes, but subsurface flow downslope is impeded by local barriers in the bedrock micro-topography. For precipitation events larger than the threshold, pits in the bedrock relief are filled, excess water spills over the micro-relief, subsurface saturated areas get connected and subsurface flow is established rather abruptly and delivers water to the stream channel. Lehmann et al. (2007) used percolation theory to set up a hillslope model that is capable of reproducing the observed threshold behaviour and to support this "fill and spill" hypothesis. The hillslope was discretized into a two dimensional lattice where each cell has only two states, denoted as occupied or non-occupied. The state of a site is regarded as dependent upon the soil depth, porosity, hydraulic conductivity, and an occupied site corresponded to a local transient water table at the soil bedrock interface. So-called bonds connect the sites and the average number of connections per site is named coordination number. Occupied and connected sites are deemed conducting with respect to subsurface flow and form a cluster. The fraction of occupied clusters – corresponding to the connected areas with a free water table – is called the occupation probability p. When p reaches the so called percolation threshold there is a cluster that spans the whole system i.e. the system is globally connected and subsurface storm flow response is initiated. Lehmann et al. (2007) was able to reproduce the observed threshold of 55 mm rainfall for onset of subsurface flow response as well as to capture the magnitude of the observed subsurface storm flow events (Fig. 5).

When displacement of relatively "old" groundwater largely contributes to subsurface runoff production, as found by Sklash and Farvolden (1979), Cloke et al. (2006), Wenninger et al. (2004) and Graeff et al. (2009), first order controls are not that obvious any more (Uhlenbrook et al., 2002; Wenninger et al., 2004). Transmission of pressure signals could be a possible explanation for the mobilization of a high amount of pre-event water (Buttle and Peters, 1997; Uhlenbrook et al., 2002; Wenninger et al., 2004). First order controls in this case would be the specific storage coefficients and transmission properties of confined aquifers (Wenninger et al., 2004) and the trigger could be rapid vertical flow either in preferential pathways or in gravel rich peri- and post-glacial sediments (Hoeg et al., 2000; Uhlenbrook et al., 2002). This appears to be effective in establishing connectivity between internal subsurface stores and the stream channel fairly rapidly.

Using rather simple models for both intensity and capacity controlled threshold preferential flow events, McGrath et al. (2007) investigated how intermittence of rainfall translates into intermittence of preferential flow events, since not every rainfall event will trigger a preferential flow event. For capacity controlled systems, which seem to correspond to some of the systems discussed above, they found an enhanced probability for a second event to occur shortly after a preferential flow event. The reason is that such a system captures the carry-over of storage from one rainfall event to the next, which is irrelevant in the case of intensity controlled preferential flow systems. Consistent with this result, Graeff et al. (2009) found that a successful reproduction of observed bimodal runoff events with a ground water model was only possible for a very high initial ground water table. Furthermore, they found that the pre-event discharge, as a proxy for the filling of the deep groundwater store, was the most important variable needed to predict the occurrence of bimodal flood events in the Schäfertal. Reliable information on subsurface storage therefore seems to be crucial for predicting flow responses of capacity controlled hydrological systems.

4.2 Bio-geomorphological thresholds for Hortonian overland flow response

Semi-arid areas cover over 30% of the world's land surface (Saco et al., 2007). Vegetation, when abundant, is often arranged into patterns that consist of patches with strong plant cover that alternate with low-cover or bare soil patches (Tietjen and Jeltsch, 2007; Tietjen et al., 2009b). Typical patterns are either characterised as spotted or stippled, consisting of dense irregularly shaped vegetation clusters that are surrounded by bare soil (Ludwig et al., 1999). Other examples are banded patterns such as the "tiger bush" in Africa and the "mogotes" in Mexico (Fig. 6), in which the dense biomass patches form bands, stripes or arcs (Saco et al., 2007; Ludwig et al., 1999) aligned along contour lines. These patterns exert a strong influence on the re-distribution of water and nutrients within the system (Saco et al., 2007; Tietjen et al., 2009a). The underlying reason is that infiltration in bare soil patches is impeded due to surface soil crusting but very effective in vegetated patches due to the presence of inter-connected root channels that act as preferential pathways (Thiery et al., 1995; Ludwig et al., 2005; Saco et al., 2007).

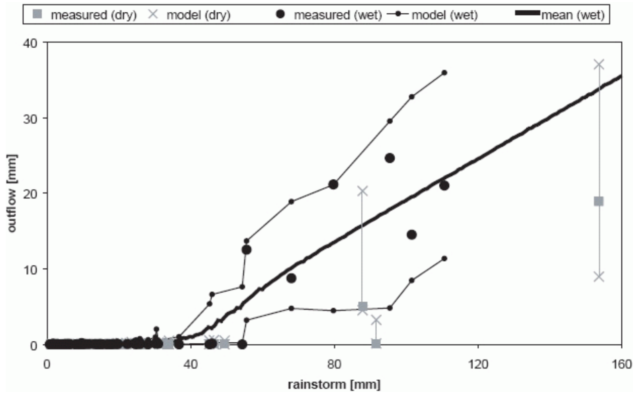

Fig. 5. Subsurface storm flow response plotted against the rainfall forcing (Figure is taken from Lehmann et al., 2007) as well as hillslope response predicted with the percolation model. The solid black line is the average response of 100 model realizations, the thin lines mark the uncertainty ranges.

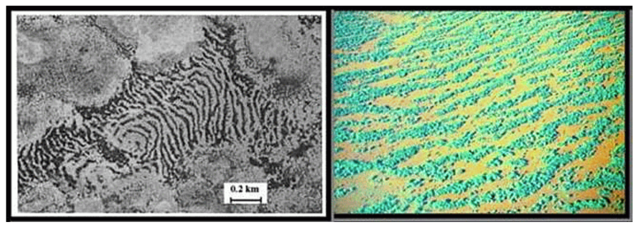

Fig. 6. Observed (optimal) vegetation patterns in semi-arid run-off-run-on ecosystem: Tiger bush in Niger, Africa (left panel Valentin et al., 1999), and banded pattern in Northern Territory, Australia (right panel www.nt.gov.au/ipe/pwcnt).

Using a model of Hortonian overland flow based on percolation theory (compare former section for a brief explanation of a percolation model), Hearman (2008) investigated how different vegetation patterns – stippled, labyrinth or banded forms – determine the overland flow response of hillslopes in semi-arid areas. She suggested that the global connectivity of bare soil patches to the hillslope toe is a first order control for the rainfall threshold that has to be crossed to initiate overland flow response at the hillslope toe. She showed that this threshold is, as would be expected, a functioning of the total vegetation cover and the type of vegetation pattern. She further showed that striped vegetation was most efficient in impeding overland flow from reaching the toes of the hillslope at a minimum of total vegetation cover.

Zehe et al. (2005a) showed that the typical hillslope scale pattern of worm burrows, with a higher amount of deeper worm burrows in the lower hillslope sectors, exerts a crucial

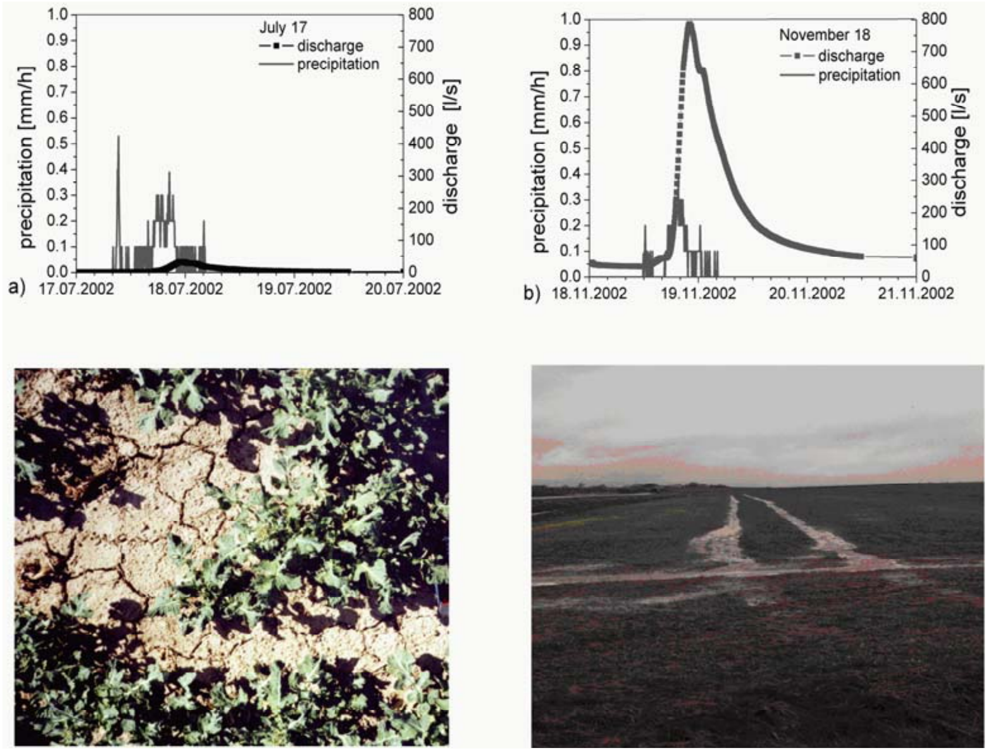

Fig. 7. Rainfall runoff response observed in the Tannhausen catchment (2.3 km^2) for two events in 2002. Event precipitation and antecedent precipitation of the summer event (left) were 18.3 and 20.3 mm, respectively; the corresponding parameters of the winter event (right) were 14.1 and 23.3 mm. Although the precipitation totals of the two events are similar, the runoff response was clearly 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.

control on hillslope overland flow and thus flood response of the Weiherbach catchment in Germany. Both changes in the worm burrow population, while preserving the typical spatial pattern as well as disturbing the spatial pattern at a preserved mean, did strongly affect hillslope scale runoff response to extreme rainfall events (Zehe et al., 2006). Flipping the typical pattern of worm burrows at the hillslope up side down reduces worm burrow/ macropore density and thus infiltration capacity downslope, leading to increased connectivity of surface flow paths.

Temporal changes in the density of macropores and thus in the infiltration capacity commonly occur in landscapes with shrinking and swelling soils. Lindemaier et al. (Lindenmaier et al., 2006) reported a strong seasonality of both flood volumes and runoff coefficients for 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) of low hydraulic conductivities (1.5×10^{-6} to 2×10^{-7} m/s) have developed. However, 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. Figure 7 provides two examples where two almost identical rainfall events (in July and November) caused almost no overland flow responses in July – due to open cracks in the valley floors – but produced strong overland flow responses when the cracks were closed. Zehe et al. (2007) found a sharp increase from low to high observed runoff coefficients when the surface soil water content increased from 0.29 to 0.35 m^3 m^{-3}. Catchment scale overland flow response during dry conditions is impeded when open cracks disconnect the surface flow paths. The closing of cracks increases the connectivity of flow paths, which increases overland flow response for a given forcing.

5 Threshold behaviour at the functional level and geo-ecosystem resilience

Thinking about the functioning of hydrological systems requires changing perspectives from short-term, event-based thinking to a long-term, holistic view of hydrological systems as closely coupled complex, human geo-ecosystems. In the last two sections we discussed how the internal "geo-ecological architecture", i.e. vegetation patterns, surface and subsurface structures, control process and response thresholds, i.e. topologies that determine redistribution of water and substances within the system and exchange with neighbouring systems. Pristine geo-ecosystems are characterised by typical patterns of topography, soil and vegetation and other biota that have co-evolved over long time scales (Phillips et al., 1999; Phillips, 2004, 2006; Dietrich and Perron, 2006). They represent a system architecture that is well adapted to the climate and water flow regimes (including not too extreme extremes; Newson, 1980), the geological setting and availability of resources for species growth (Watt, 1947; With and Crist, 1995; Schröder, 2006). Resilience of such a geo-ecosystem configuration and thus of its hydrological functioning – i.e. the extent of disturbance the system may tolerate without responding with drastic qualitative changes in dynamics – is determined by the multiple feedbacks between abiotic and biotic components that comprise the system. However, most places in the world are not pristine but are under substantial human influence and face considerable disturbance and changes including climate change. What is a substantial disturbance of this internal geo-ecological architecture that leads to (irreversibly) altered process and response thresholds, and thus to a substantially altered hydrological functioning? The following sections will elaborate on this question using simple examples that allow a clear identification of the cause and effect relations in the context of functional thresholds.

5.1 Biological controls of hydrological functioning in pristine and rural areas

As discussed above, ecosystems in semi-arid regions are highly coupled morphological-ecological-hydrological systems and water is the key factor in determining plant growth (Walker et al., 1981; Ehleringer et al., 1991; Kemp et al., 1997; Rodriguez-Iturbe et al., 1999). The arrangement of vegetation in stippled, labyrinth or banded patterns is obviously an efficient way to 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 the system configuration, as soil moisture in vegetated patches is higher, which reinforces biomass production, which in turn reinforces development/persistence of soil structures (Thiery et al., 1995; Wilcox et al., 2003a, b; Ludwig et al., 2005; Wilcox and Newman, 2005). Such soil-vegetation patterns control, along with topography, the hydrologic functioning of catchment systems in semi-arid and arid regions, in particular the redistribution of water, nutrients and soil material. Water losses due to evaporation 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 minimizes erosion and soil losses (Saco et al., 2007).

However, a strong and persistent disturbance might seriously endanger the architecture of these systems (Tietjen and Jeltsch, 2007) and their ability to minimize the export of water, sediments and nutrients. Leaky vegetation structures, often caused by grazing pressures (Rietkerk and van de Koppel, 1997; Rietkerk et al., 2000, 2002; Kefi et al., 2007), are less efficient in trapping runoff and sediments (Walker et al., 1981, 1999; Ludwig et al., 2004, 2005). 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). The hydrological functioning, as characterized by patterns of sediment export, and the flow duration curve, will start to change irreversibly.

Semi-arid savannah ecosystems contain over 50% of the global livestock (Tietjen et al., 2009b). These ecosystems thus provide a service that is most crucial for global food production, especially in developing countries; the provision of ecosystem service depends crucially on the fragile equilibrium between woody and herbaceous vegetation. By applying a recently developed eco-hydrological model to a Namibian thornbush savannah, Tietjen et al. (2009a) evaluated the separate and combined effects of decreased annual precipitation, increased temperature, more variable precipitation, and elevated atmospheric CO_2 on soil moisture and on vegetation cover. They suggest that expected climate change tends to promote shrub growth in Namibian thornbush savannah. As a consequence infiltration into the deeper subsoil and transpiration losses would increase and redistribution of water resources due to runoff would be impeded (Tietjen et al., 2009a, b). The drier surface soil moisture regime implies worse conditions for grass growth. As grasses and shrubs compete for water resources this means a positive feedback for shrub growth and further change of the hydrological functioning. Once a degraded state is reached, regeneration is hardly feasible, and a reduction in biodiversity or in productivity for livestock farming is likely (Roques et al., 2001).

Earthworms are classic ecosystem engineers. Long-lasting soil structures generated by earthworms exert significant influence on process thresholds for overland flow and the initiation of rapid solute transport (Edwards et al., 1990a; Shipitalo and Butt, 1999; Shipitalo et al., 2000; Bastardie et al., 2002, 2003) and their spatial distribution affects connectivity of flow paths that determine response thresholds for overland flow at the catchment scales (Zehe et al., 2005). Earthworm behaviour and their metabolic activity are also crucial for organic matter dynamics, pedogenetic processes, plant growth and degradation of herbicides (Lavelle et al., 1997; Hedde et al., 2005; Milcu, 2005; Bolduan and Zehe, 2006; Milcu et al., 2006). Abundance and spatial distribution of earthworm populations depend on spatio-temporal patterns of organic matter content (Rossi et al., 1997, 2003), soil hydrological properties and soil water content (Cannavacciuolo et al., 1998; Nuutinen et al., 1998, 2001), biotic interactions (Nuutinen, 1997; Rossi, 2004), and disturbance regimes related to agricultural practice (Whalen et al., 1998, 2004; Whalen and Costa, 2003; Joschko et al., 2006; Whalen and Fox, 2007).

In a modelling study Zehe et al. (2005, 2006) showed that a reduced population of earthworms that goes along with a reduction of apparent connected preferential pathways would thus reduce the intensity threshold for overland flow production and increase connectivity of surface flow paths that control overland and thus flood response. A positive effect of an apparently lower number of earthworms and thus worm burrows would be a lower susceptibility for rapid transport of agrochemicals and thus a possible reduction of related environmental risk. However, this might be compensated by increasing transport of agrochemicals in overland flow into surface water bodies (Flury, 1996).

5.2 Morphological triggers for drastic changes in hydrological functioning

Morphological processes, though in most cases they may be regarded as rather slow, may cause drastic changes in hydrological functioning that go along with degradation of the entire geo-ecosystem, such as the formation of badlands. This is well known from many places around the world such as Spain (Bull and Kirkby, 1997; Kirkby et al., 2003), Chile (Maerker et al., 2008), France (Maquaire et al., 2003; Rey et al., 2005) New Zealand (Hicks et al., 2000; Parkner et al., 2007), and South Africa (Boardman et al., 2003; Achten et al., 2008; Boardman and Foster, 2008). Initiation of erosion rills/pipes and their possible evolution into gullies is a crucial factor (Bull and Kirkby, 1997) that is strongly controlled by connectivity of surface flow paths in combination with the occurrence of extreme rainfall events (Fryirs et al., 2007; Faulkner, 2008). Formation of erosion gullies is commonly associated with agricultural landscapes where vegetation clearance increases runoff, leading to fluvial incision (Bull and Kirkby, 1997). However, gully erosion can also occur in pristine forests, as recently found by Parkner et al. (2007).

Faulkner et al. (2008) recently suggested, for the case of the Mocatan badlands catchment that after pipes achieve coupling to the main channel the pipe network develops in a systematic, sequential way. They suggest that crossing this threshold causes a "wave of incision" to positively feed back on the development of deep cut rills that collapse to form an extensive, partially coupled steep-sided gully network. An interesting point in the context of erosion is that we humans can and often do impede the system from crossing such a threshold through appropriate land use planning. For instance Scherer (2008) demonstrated that a simple rearrangement of the land use pattern by co-locating crops that hamper particle detachment at locations of high erosion risk can substantially reduce sediment export from the Weiherbach catchment.

5.3 Impacts of land use/land cover changes

Human activities can also significantly impact hydrological functioning when lateral subsurface flows dominate runoff production, such as in the case of the Schäfertal, a small headwater catchment located in the Harz Mountains of Germany. Up to the mid-seventies rainfall-runoff behaviour was frequently affected by bi-modal rainfall-runoff events (Becker and McDonnell, 1998; Becker, 2005) that consist of a fast flood peak that occurs shortly after the onset of the rainfall event as well as a second peak that occurs with a lag of approximately one day and shows a pronounced recession lasting up to several days. In a combined field and modelling study Graeff et al. (2009) explained the second delayed peak by fast displacement of groundwater. In 1974 nearby mining activities started to exploit fluorite deposits at a depth of 190 m and a distance of nearly 1 km and bimodal rainfall-runoff events vanished almost completely (Wenk, 2000). Graeff et al. (2009) explained this change in hydrological functioning by groundwater pumping activity that significantly lowered groundwater levels in this area. It is interesting to note that although mining has stopped for several years, the former regime has not re-established yet.

Regulated rivers are amongst the most prominent examples for human induced changes in the hydrological functioning of large and complex systems. The rectification of the upper Rhine is, at least in Europe, among the most prominent examples of both the great success and also negative impacts of river regulation (Dister et al., 1990). Initiated by Johann Gottfried Tulla in 1817, the objective was to increase habitability of the upper Rhine valley by lowering groundwater levels, changing local climate, reducing periods of flooding and thus achieve favourable conditions for agriculture. This was achieved by cutting off river meanders and increasing discharge and flow velocities substantially. Since the early days of Tulla many measures up to the "Grand Canal d'Alsace" were established that completely reshaped the upper Rhine and strongly amplified river bed erosion and the lowering of groundwater levels far beyond the original expectations (Dister et al., 1990; Hofius, l99l). Just to name a few of the unexpected changes in hydrological functioning of this river system: 1) travel times of flood waves from Basel to Cologne were reduced to almost a third of the initial values (Pinter et al., 2006), 2) erosion of the river bed has to be impeded by application of debris to the upper Rhine that is taken from the lower Rhine, and 3) habitat quality for important fish species was seriously degraded (Aarts et al., 2004). Maintaining such a river architecture, far from the original minimum energy expenditure state (Rodriguez-Iturbe et al., 1992; Rodriguez-Iturbe and Rinaldo, 2001), requires a continuous investment of energy and thus money. Even if this is invested a river might partly restore its old architecture during severe, morphological flood events, as was seen in the case of the Weißeritz River in the famous flood in the city of Dresden in August 2002 (Petrow et al., 2006).

6 Implications and conclusions

6.1 Threshold behaviour and hydrological models

As elaborated above, threshold behaviour implies in general a fast qualitative change of dynamics either of a process, the response of a hydrological system in a given context, or the hydrological functioning of a system. In hydrology and related earth system sciences we are highly interested in predicting these qualitative changes in dynamics as they often go along with hazards such as flooding, environmental problems such as erosion and contamination or even geo-ecosystem degradation. Predicting threshold behaviour requires, as a first step, identification of first order controls and as a second step implementation of these controls into hydrological models. The first step becomes, as discussed in the previous sections, increasingly difficult when moving from the process, to the response, and then to the functional level of threshold behaviour. Also because necessary identification of first order controls and cause-and-effect relations are impeded by poor quality of observations and the limited reproducibility of controlled experiments, especially when morphological processes, ecological feedbacks or systems with long memories are involved (as in the case of the Rhine regulation). Up to now hydrology has only partly been successful in achieving the second step, and we think that current models are either capable:

- To predict threshold behaviour at the process level which is controlled by local state variables, boundary conditions and system properties;

- To predict threshold behaviour at the response level by implicitly conceptualizing the (hidden) multivariate statistical or topological controls based on effective states, parameters and fluxes.

Models of the first category have traditionally been called "physically based". They describe soil water flow using the Darcy-Richards approach – including different approaches to preferential flow – solute transport using the convection dispersion approach and overland flow based on the 1- or 2-D hydraulic approaches (MIKE SHE: Refsgaard and Storm, 1995; HYDRUS Šimunek et al., 1999, 2005; CATFLOW: Zehe et al., 2001; HILLFLOW, Bronstert and Plate, 1997; InHM, VanderKwaak and Loague, 2001). These models allow a successful prediction of overland flow generation, infiltration and preferential flow/transport at the plot and small catchment scale (with uncertainty, Beven, 1989; Beven and Binley, 1992). The charm of these models is that we might check their predictions using distributed observations of soil states and soil parameters within the system and, thus, at least get a quantitative assessment as to how local changes of system properties affect process thresholds. However, predictions of (threshold) responses of hydrological systems with this type of models requires a representation of the *spatial fields* of soil and surface properties at a spatial resolution that is sufficiently fine to capture their key multivariate statistical and topological properties. Except for a few microscale research catchments this information will never be at hand. Thus, we cannot use these models to predict how local (threshold) dynamics translates into the response of hydrological systems at useful scales (Beven, 2006) – not to mention deficiencies of REV based process descriptions at larger scales.

Models of the second category implicitly conceptualise the control of multivariate statistical and topological system properties usually on runoff and stream flow response of hydrological systems by means of effective states, effective parameters and fluxes. The TOPMODEL concept is one example of such a conceptualisation (TOPMODEL: Beven and Kirkby, 1979; THALES: Grayson et al., 1992, TOPOG: VERTESSY et al., 1993; WASIM-ETH: Jasper et al., 2002) that is well suited for humid climates and shallow permeable soils (with uncertainty Beven and Binley, 1989). The HBV model concept (Bergström, 1995; Hundecha and Bardossy, 2004), including further refinements such as TAC^d (Uhlenbrook and Leibundgut, 2002), is another example that works well as long the runoff-storage relation is monotonically increasing. The charm of these models is that they allow a reproduction of hydrological responses including the effects of threshold behaviour, for example, of subsurface stormflow, at useful scales (with uncertainty, Uhlenbrook et al., 1999; Beven and Freer, 2001). However, reproducing the response of the system, in most cases, does not mean that simulated dynamics or the local model structure is consistent with distributed observations of hydrological states and local catchment properties and our process knowledge. Thus, we might not infer from these models which combination of local (threshold) processes and redistribution mechanisms caused the observed response and therefore cannot predict how changes of the system will translate into altered response thresholds (Sivapalan et al., 2003) and altered hydrological functioning.

Understanding and predicting (drastic) changes in hydrological functioning are among the greatest challenges in hydrological science as it requires extrapolations far beyond our range of experience in closely coupled human geo-ecosystems (with feedbacks). This requires models that work for the right reasons (Sivapalan et al., 2003; Kirchner, 2006). They must capture how threshold behaviour at the process level and the redistribution of local dynamics translate "forward" to the response of a hydrological system and allow a "backward" estimate on the pattern of local dynamics and structures that control redistribution after they have been shown to reproduce system behaviour.

6.2 Way forward?

Though predicting change in hydrological functioning requires that we have to deal with systems of intermediate *complexity* this does not automatically mean that we have to develop more and more *complicated* models. On the contrary, it is crucial to find the right level of simplification or complexity, as discussed by Savenije (2009). This can only be achieved by combining bottom-up thinking to avoid oversimplification with top-down thinking that sharpens our view on what is important and helps us to avoid losing ourselves in the details, i.e., "missing the forest for the trees".

6.2.1 Need for better "observables"

As subsurface storage and topology appear to be first order controls of threshold behaviour at the response level, we need better techniques to assess dynamics and topological properties in the subsurface at useful scales (Beven, 2006). Geophysical methods such as ground penetrating radar (Binley et al., 2002; Roth et al., 2004), electrical resistivity soundings (ERT) (Kemna et al., 2002; Wenninger et al., 2008; Graeff et al., 2009b); or seismic soundings (Herbst et al., 1998; Schmelzbach et al., 2007) are promising in this respect. The former yields the pattern of the dielectric permittivity, the second the pattern of the apparent specific resistivity – both are related to subsurface wetness – the latter yields velocity profiles that are mainly related to bulk density and thus porosity. As the inversion of geophysical observations is not a well defined problem several authors have suggested joint inversion of several data sources to reduce equifinality (Binley and Beven, 2003; Paasche et al., 2006; Paasche and Tronicke, 2007; Looms et al., 2008). Another promising idea is to derive multivariate characteristics of subsurface parameters, for example, the correlation length of lateral structures by analysing the "raw" radar data. This approach avoids the usual smoothing effect of regularisation during inversion (Tronicke et al., 2004). Combining distributed geophysical methods with artificial and natural tracer observations seems even more promising (Kemna et al., 2002; Cassiani et al., 2006) to understand how structures/topology translate "forward" into signatures of integrated tracer responses, which is the first step towards understanding how we may backward-infer subsurface topological properties from tracer and geophysical observations at multiple scales. Reductionist physically based models may play a supporting role here. Through simulations of flow and tracer transport in non-trivial, i.e. non-Gaussian structured heterogeneous, systems one may generate a virtual reality for simulating tracer and geophysical observations, inverting these virtual measurements and exploring how much of the known structure we can ideally get back from these observations (Weiler and McDonnell, 2004; Zehe et al., 2005a, b, 2006). However, such a research strategy requires sustaining joint experimental work in already established well documented research catchments as long term hydrological observatories (Sivapalan et al., 2003).

6.2.2 Future models based on observables and landscape structures

To make progress in the future in terms of predictions, we must reunify the different fields of hydrological research. Today our community is stuck, i.e., "lost in translation", between mesoscale modellers, process modellers and experimentalists/field people. A fruitful cooperation appears difficult, often impossible (Fenicia et al., 2008a, b). We need to build our theoretical concepts and models around those "observables" we can assess today and hope to assess in the *near* future rather than around effective concepts that scale up REV scale approaches by means of effective parameters, which is what we have done so far. This means establishing a modelling framework that provides *common concepts* and a *common language* for field hydrologists and modellers – to avoid Babylonian confusion – and that allows integration of distributed process knowledge, geophysical data and available geo-ecological data. In the sense of Vogel and Roth (2003) these models have to

1. explicitly represent key topologies and structures that determine response thresholds in the landscape, and
2. effectively represent subscale dynamics – controlled by subscale structures – in a way that is though simplified, at least in principle, and compatible with subscale process understanding and observations.

We suggest that the REW approach (Reggiani et al., 1998, 1999, 2000; Reggiani and Rientjes, 2005) already meets the second requirement (Lee et al., 2005, 2007; Zehe et al., 2005b, 2006; Zhang et al., 2005, 2006; Zhang and Savenije, 2005; Beven, 2006), however, it fails to meet the first. Furthermore, it is a suitable framework to reunify the different fields of hydrological research as it postulates the existence of *functional units* with homogeneous *hydrological response/behaviour* in the landscape (Beven, 2006; Zehe et al., 2006). Therefore the REW approach can be interpreted as a mathematically rigorous implementation of the concept of hydrological response units (HRU). The latter is also well suited to guide hydrological field work.

However, all applications of the REW approach up to now (Reggiani and Rientjes, 2005; Zhang et al., 2005, 2006; Varado et al., 2006; Mou et al., 2008) treat subcatchments and REWs as synonymous and thus compromise the first requirement by averaging across key topologies inside the catchments. This destroys the large scale architecture of the catchment by smoothing out key topologies that should be explicitly resolved. Future refinements of the REW approach should therefore arrange REW/HRU's along lead topologies in the landscape, for instance, along the catena, and allow for exchange processes between REWs/HRUs. This would preserve the rigorous mathematical formulation and internal structure of the approach and also allow for a distributed process representation that can capture the connectivity in the landscape – without ending up in the numerical overkill of partial differential equations. It would, furthermore, facilitate the representation of available observables and vegetation patterns in the model structure as well as the use of present and near future distributed data to assess closure relations (Beven 2006; Zehe et al., 2006).

6.2.3 Explore first order controls of threshold behaviour, predictability and feedbacks

The use of (simplified) models as "diagnostic tools" can, as elaborated, provide valuable insights into first order controls on threshold behaviour such as connectivity of bare soil patches in the case of overland flow response and global connectivity of the subsurface water tables in case of subsurface storm flow (Lehmann, 2007; Hearman, 2008).

Furthermore, such diagnostic model studies may be used to explore patterns of reproducibility of field observations in the presence of threshold processes (Zehe et al., 2007). From such studies we might learn about which patterns in our observations might hint on threshold behaviour and how to explain observations that appear inexplicable at first sight. A nice example is that of Lischeid et al. (2000). They observed tracer velocities that ranged between 30.6 and 10.6 m/d during three identical steady-state field scale breakthrough experiments at the Gårdsjön test catchment, which could not be explained by any *measurable* difference in the experimental conditions and can be easily explained with repeated trials close to a threshold to establish preferential flow.

We thus learn that repeatability of observations and, therefore, our understanding of cause-and-effect relations is limited in threshold systems as also suggested by Beven (1996a). This means that "we have to accept limits of predictability as we cannot expect hydrological models to predict more accurately than the repeatability of nature herself" (Beven, 1996a). This also means that we might learn which observations are crucially needed to improve predictability and narrow down the unstable range sketched in Fig. 1 (lower panel). Furthermore, diagnostic models allow us to shed light on how the intermittency and statistical properties of the rainfall forcing translate to intermittency of threshold responses that are either capacity or intensity controlled (McGrath et al., 2007). From these studies we might, for example, learn how to adapt our temporal sampling design when observing capacity controlled threshold processes, as the carry over of storage leads to a temporal clustering of events.

However, to avoid interpreting artefacts produced by oversimplified models any kind of diagnostic model study should be based on models that have been shown to portray system behaviour sufficiently well. This is even more crucial when studying bidirectional feedbacks between biotic and abiotic ecosystem components, for instance, vegetation and hydrological processes. This is a crucial task for understanding geo-ecosystem resilience and thus the stability of hydrological functioning; however, it is also most delicate even when the focus is on ecosystems where feedbacks and cause-and-effect relations appear to be rather obvious, as pointed out by Tietjen and Jeltsch (2007). These authors compared 41 different models to simulate coupled vegetation and water dynamics of savannah ecosystems and found a series of deficiencies concerning the representation of water dynamics and feedbacks of vegetation on key hydrological processes such as the lateral redistribution due to Hortonian overland flow and infiltration. Tietjen et al. (2009a, b) developed their own coupled eco-hydrological model that accounts for the complete water cycle in a simplified form, sound vegetation dynamics of shrubs and grasses and key feedbacks such as the effects of shrub cover on infiltration, runoff redistribution and transpiration. With their model they were able to reproduce observed soil water dynamics at several different sites in Israel and observed vegetation patterns at several locations. Thus, water-, vegetation dynamics and feedbacks are at least consistent with hydrological and ecological observations and the model structure is acceptable for both hydrologists and ecologists. Thus, one may be cautiously optimistic that extrapolations to possible future climate conditions will not point in a totally wrong direction. This underpins the notion that finding the right level of simplification (or complexity) is crucial, even when we deal with systems that appear "nice and neat" at first sight.

6.2.4 Ask the "why- questions"

Why does threshold behaviour occur in our tea kettle? Why do structures in the landscape evolve the way they do? Why do they evolve and persist in the light of the second law of thermodynamics, which postulates that the universe approaches maximum entropy and thus maximum disorder in thermodynamic equilibrium? Fundamental issues that have been addressed by Haken (1983) in his theory of synergetics and also by Prigogine (Nicolis and Prigogine, 1977), who explained that energy inflow into open systems far from thermodynamic equilibrium may create "ordered structures". He called these structures dissipative structures and was awarded the Nobel Prize for chemistry in 1979 for his theory. Convection cells in our tea kettle example are dissipative structures, which increase the efficiency of dissipating the energy inputs against the thermodynamic gradient through turbulent eddies, which are highly organised at the microscopic scale according to Prigogine and Stengler (1984).

Since our landscape and structures have been formed by dissipative processes (Leopold and Langbein, 1962; Phillips et al., 1999, 2006), may we then explain them as slowly evolving dissipative structures? Rodriguez-Iturbe et al. (Rodriguez-Iturbe et al., 1992; Rodriguez-Iturbe and Rinaldo, 2001) as well as Rinaldo et al. (1996) employed thermodynamics to explain the organisation of river networks as "least energy structures". Recently, Kleidon and Schymanski (2008) proposed that most processes in the hydrological cycle, including soil wetting, are irreversible and produce entropy. They suggest that an optimal/steady state architecture ensures that exchange fluxes of mass and energy maximise entropy production (MEP). The MEP principle emerges from the trade-off between thermodynamic "forces"/gradients and fluxes, because the latter depletes the former (Kleidon et al., 2006; Kleidon and Schymanski, 2008). Zehe et al. (2009) recently showed that preferential pathways, when activated, are more efficient dissipators of Helmholtz free energy in cohesive soils, especially during dry summer conditions. Apparent preferential pathways offer additional degrees of freedom to the flow process and Zehe et al. (2009) suggested that water flow in soils organises in such a way that dissipation of Helmhotz free energy becomes maximum (MED) for a given rainfall input, for a given soil and given soil structures. Maybe connective structures in hydrological systems are in general favourable from a thermodynamic perspective because they allow formation of dissipative structures during rainfall events, faster dissipation and faster relaxation towards equilibrium? One important implication of this MEP/MED hypothesis is that we may infer how an optimal hillslope structure should look like, if the univariate distributions of key properties of hillslope are known: simply by re-arranging pixels with different properties as long as the water and mass flows across the systems boundary maximise entropy production/energy dissipation. In soils this means to assure a fast drainage of excess water and a fast redistribution of water to dry parts within the soil. As strong gradients in the matric potential are equivalent to strong gradients in chemical potential, depleting these gradients means to reduce Helmholtz free energy of the system (Kleidon and Schymanski, 2008; Zehe et al., 2009). Such an optimal architecture could then be compared with true architecture of the hillslope to test the MEP/MED hypothesis.

Understanding such "why" type questions in a real quantitative manner might, in the long term, be one of the cardinal challenges to hydrology and related earth systems sciences to better understand the resilience of the human geo-ecosystems and to make judgements about the extent of the human interferences.

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