Professur für Hydrologie

Albert-Ludwigs-Universität Freiburg im Breisgau

Matthias Ritter

Combined field- and model-based intercomparison of hillslope hydrological dynamics

Masterarbeit unter Leitung von Prof. Dr. Markus Weiler

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Ehrenwörtliche Erklärung

Hiermit erkläre ich, dass die Arbeit selbständig und nur unter Verwendung der angegebenen Hilfsmittel angefertigt wurde.

Ort, Datum Unterschrift

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Summary

Investigating first-order controls of hydrological dynamics at the hillslope scale is also crucial for a fundamental understanding at catchment scale. Especially the role of vegetation on hillslope runoff processes must to be examined. Hence, the intention of this study was to explore the influence of contrasting vegetation cover on hillslope hydrological dynamics in terms of flow pathways and solute transport. It was built on an existing large-scale hillslope intercomparison setup of three adjacent well-instrumented hillslopes with similar topography and parent material, but different vegetation cover (grassland, coniferous forest, mixed forest).

Firstly, sodium chloride was used as hydrological tracer and injected into wells, 30 m upslope to a monitoring trench at each of the three hillslopes. Tracer breakthrough was measured as electrical conductivity in near trench wells and the trench outflow. At the grassland and coniferous hillslopes a breakthrough could be observed in the downslope wells and trench flow during the weeks after tracer application. Additionally, one-third of the total applied tracer mass was exported with streamflow but not captured in the trenches. This proved hillslope-stream connectivity via deep (below 2 m) flow pathways. Differences in the tracer response could be attributed to local microtopography but not to vegetation effects.

Secondly, the hydrological model HillVi was set up for each hillslope and calibrated against observed trench flow, water table and tracer transport data. The model could approximately reproduce the trench flow dynamics but failed for water table heights and tracer transport. Consequently the model structure was regarded as inadequate for the hillslope subsurface flow processes. In addition parameter identifiability was low and optimized parameter ranges unrelated to vegetation cover.

Based on the study results a revised perceptual model of hillslope runoff mechanisms for the study site was hypothesized and should be tested with numerical and field studies in the future.

Keywords: Hillslope hydrology; Runoff generation; Subsurface Flow; Vegetation cover effect; Intercomparison study; Tracer; Numerical model

Zusammenfassung

Die Untersuchung der dominanten Einflussgrößen auf das hydrologische Verhalten der Hangskale ist ebenfalls für das Verständnis auf der Einzugsgebietsskale von großer Bedeutung. Insbesondere der Vegetationseinfluss auf Hangabflussprozesse sollte erforscht werden. Deshalb war das Ziel dieser Arbeit zu untersuchen, wie unterschiedliche Vegetationsbedeckung die Hanghydrologie bezüglich der Fließwege und des Stofftransports beeinflusst. Die Arbeit knüpft dabei an eine bestehende Vergleichsstudie zwischen drei benachbarten und stark instrumentierten Hängen an. Die Hänge ähneln sich in Bezug auf Topografie und Ausgangsgestein, unterscheiden sich aber in der Vegetation (Wiese, Nadelwald, Mischwald).

Zunächst wurde auf jedem der drei Hänge Natriumchlorid als Tracer in Bohrlöcher 30 m oberhalb von Drainagegräben eingespeist. Die Durchgangskurven der elektrischen Leitfähigkeit wurden in Bohrlöchern nahe der Gräben und im Grabenabfluss aufgezeichnet. Auf dem Wiesen- und Nadelwaldhang wurden während der Wochen nach der Einspeisung Durchgänge in den unteren Bohrlöchern und im Grabenabfluss beobachtet. Außerdem wurde ein Drittel der gesamten Tracermasse über den Bach ausgetragen, was aber nicht in den Drainagegräben gemessen wurde. Somit konnten tiefe Fließwege (unterhalb 2 m) zwischen Hang und Bach nachgewiesen werden. Unterschiedliche Tracerdurchgänge konnten der Mikrotopografie, nicht aber Vegetationsunterschieden zugerechnet werden.

Danach wurde das hydrologische Modell HillVi für jeden der Hänge mit Hilfe der gemessenen Grabenabflüsse, Grundwasserspiegel und Tracertransportdaten kalibriert. Das Model konnte näherungsweise die Grabenabflüsse abbilden, versagte jedoch für die Wasserspiegelhöhen und Tracertransport. Daher wurde die Modellstruktur als ungeeignete Repräsentation der unterirdischen Hangabflussprozesse verworfen. Außerdem war die Parameteridentifizierbarkeit gering und die Parameterwerte waren unabhängig von der Vegetation.

Auf Grundlage der Ergebnisse wurde eine überarbeitete Hypothese über die Hangabflussmechanismen für den Untersuchungsraum aufgestellt, die zukünftig mit Model- und Feldstudien überprüft werden sollte.

Stichworte: Hanghydrologie; Abflussbildung; Zwischenabfluss; Vegetationseffekte; Vergleichsstudie; Tracer; Numerisches Model

1 Introduction

1.1 Literature review

Hillslopes are critical zones that shape the hydrological behavior of catchments in steep and humid environments (Tromp-van Meerveld and Weiler, 2008). As such basic landscape building blocks they control the dynamics of water storage, its vertical and lateral redistribution and release to the streams during and between rainfall or snowmelt events. Thus, they fundamentally affect flood generation, mass transport, slope stability and ecohydrological cycling (Bachmair and Weiler, 2011; Wagener et al., 2007). Identifying and understanding the dominant driving processes is crucial to robust predictions of runoff generation and water quality in space and time, especially under shifting land use and climatic conditions (Bachmair and Weiler, 2011). For accurate projections water source, flow path and age must be integrated (Bonell, 1998; Hewlett and Hibbert, 1967; Hewlett and Troendle, 1975). Current hydrological models purport a high degree of realism and sophisticated theoretical background, but they widely prove inconsistency when confronted with field–based process descriptions and observed scaling relationships (McDonnell et al., 2007). Although progress was made particularly at the hillslope scale, our ability to classify and conceptualize these hydrological dynamics (e.g. for modeling) based on our process understanding is still limited (McDonnell, 2003; McDonnell et al., 2007, 2010; Sivapalan, 2003; Tetzlaff et al., 2008).

Recently the quest for an unifying hydrological theory and thus a more holistic depiction of runoff response was initiated (Sivapalan, 2005; Troch et al., 2009). Instead of mapping omnipresent landscape and process complexity, the study of underlying organizing principles would derive generalizable knowledge for different scales and climates (McDonnell et al., 2007). Therefore, first-order controls of hillslope hydrological processes must be systematically explored through field surveys and numerical experiments. Frequently field studies were carried out at intensively monitored individual hillslopes and their hydrological behavior was characterized (e.g. Brooks et al., 2004; Graham et al., 2010a; Harr, 1977; McDonnell, 1990; Ohara et al., 2011; Sklash et al., 1986; Tromp-van Meerveld and McDonnell, 2006b, 2006c; Woods and Rowe, 1996). However, the result was the documentation of hillslope idiosyncrasies rather than a systematic evaluation of relevant hillslope hydrological controls across the studied sites (Uchida et al., 2006; Weiler and McDonnell, 2004). Better transferable insights can be drawn from functional intercomparisons of well-instrumented hillslopes. By this means, site characteristics similarity (e.g. climate, vegetation, topography) can be used to isolate dissimilarity (e.g. geology, soil properties) effects and thus identify first-order controls on hydrometric, isotopic and chemical dynamics (Uchida et al., 2006). During the past decade several studies designed as hillslope intercomparisons were conducted (e.g. Asano et al., 2002; Bachmair et al., 2012; Gabrielli et al., 2012; Kienzler and Naef, 2008a, 2008b; Nordmann et al., 2009; Scherrer and Naef, 2003; Scherrer et al., 2007; Uchida et al., 2004, 2005, 2006). However, field-based intercomparison is constrained by the trade-off between slope size and number (Bachmair et al., 2012), different monitoring setups (Uchida et al., 2006) and the difficulty to find suitable sites due to the uniqueness and heterogeneity at even seemingly homogeneous natural hillslopes (Bachmair and Weiler, 2011; Bachmair et al., 2012). Alternatively, artificially physical (Hopp et al., 2009) or completely virtually constructed hillslopes can be compared. Weiler and McDonnell (2004) introduced virtual experiments as "numerical experiments with a model driven by collective field intelligence". Virtual experiments provide an effective framework to test the role of hypothesized first-order controls from field observations and transfer process understanding between different sites (e.g. Dunn et al., 2007; Hopp and McDonnell, 2009, 2011; Keim et al., 2006; Sayama and McDonnell, 2009; Tromp-van Meerveld and Weiler, 2008; Weiler and McDonnell, 2004, 2006).

One consistent outcome of many hillslope studies seems to be a nonlinear rainfall-runoff relationship (McDonnell, 2003; Weiler et al., 2005). Often rainfall amount must exceed a threshold until hillslope discharge increases significantly or is triggered at all (e.g. Graham et al., 2010b; McGuire and McDonnell, 2010; Tromp-van Meerveld and McDonnell, 2006b; Uchida et al., 2005). This system state dependent ability to transfer water and solutes or matter from hillslopes to streams is unified in the hydrological connectivity concept (McDonnell et al., 2007; Michaelides and Chappell, 2009). In this context Bachmair and Weiler (2011) proposed a "connect-and-react" mechanism via hydrologically-active areas (surface flow, saturated soil patches, high-permeability features). These areas develop during rainfall on and within the slope due to local heterogeneity. Once the active areas interconnect, they contribute significant hillslope runoff. Surface runoff areas can be initiated as Hortonian or infiltration excess (Horton, 1933), saturation overland flow and return flow (Dunne and Black, 1970; Hewlett and Hibbert, 1967). On forested hillslopes real Hortonian overland flow (rainfall intensity exceeds soil infiltration capacity) has been rarely observed, yet near-surface flow can also occur within the litter layer (Gomi et al., 2008; McDonnell et al., 1991; Sidle et al., 2007; Weiler and McDonnell, 2004). Saturation overland flow is induced by upward saturation within the soil profile to the surface. It is dependent on the wetness state and transmissive soils. Topography or impeding soil layers may route subsurface water back to the soil surface as return flow (Dunne and Black, 1970). In general, water redistribution within the soil as subsurface flow (also termed subsurface stormflow or interflow) is the main mechanism of lateral water movement in humid environment and steep terrain (Hewlett and Hibbert, 1967; Weiler et al., 2005). It is the fast lateral movement of water through the soil or weathered bedrock above a layer of reduced permeability. Although subsurface flow initiation is not fully understood, it seems that transient saturation of soil parts are a prerequisite (Weiler et al., 2005). Lateral preferential flow paths in form of highly permeable soil matrix parts or soil pipes (Uchida et al., 2001; Weiler and McDonnell, 2007) are assumed to prevail homogeneous matrix flow (Bachmair and Weiler, 2011).

Several interacting key factors control the dominance of hillslope runoff processes and the build-up and persistence of connected subsurface flow pathways. According to time scales, geology, surface and subsurface topography can be categorized as static (in the sense of years) and precipitation, soil moisture and vegetation as dynamic factors (Bachmair and Weiler, 2011).

Surface and bedrock topography is obviously a fundamental first-order control for flow routing and also influences the distribution of soil properties and moisture, atmospheric fluxes and vegetation (e.g. Anderson and Burt, 1978; Beven and Kirkby, 1979; Troch et al., 2003; Wagener et al., 2007). Irregular hillslope bedrock microtopography and permeability were found to be closely linked to the observed nonlinear runoff response. Bedrock depressions have to fill up during a rainfall event before water spills over and a connected flow path network can contribute to hillslope outflow (the “fill-and-spill” hypothesis) (Graham et al., 2010b; Tromp-van Meerveld and McDonnell, 2006b, 2006c). Parent material and weathering rate define bedrock permeability and hence vertical losses can occur during flow transmission along the soil-bedrock interface at the event and longer time scales. It was recently noted that deep percolation into and also exfiltration from the underlying bedrock can be important hillslope water balance terms (e.g. Anderson et al., 1997; Gabrielli et al., 2012; Graham et al., 2010a; Kosugi et al., 2006; Salve et al., 2012; Tromp-van Meerveld et al., 2007; Tromp-van Meerveld and Weiler, 2008). Surface and subsurface topography are linked via variable soil depths. Deeper soils can delay and reduce subsurface flow. In shallower areas transient saturation above bedrock and subsurface flow start earlier caused by the smaller total storage volume. Changes in drainable porosity show a similar effect on subsurface flow initiation (Hopp and McDonnell, 2009; Weiler and McDonnell, 2004). Often the hydrological response is linked to the spatial distribution of soil types because of feedbacks between pedogenetic processes and the dynamics of water movement through soils (e.g. Tetzlaff et al., 2007).

Significant preferential pathways may develop as pipe networks from decayed root channels, animal burrows and subsurface erosion. Despite their importance for hillslope runoff hydrographs, the mechanisms of pipe formation and pipe flow initiation are not fully understood (Anderson et al., 2009a, 2009b; Jones, 1997; Tromp-van Meerveld and McDonnell, 2006b; Weiler and McDonnell, 2007). Additionally, preferential flow in vertical direction through macropores enhances the infiltration capacity. Water can quickly move into deeper soil layers and bypass parts of the soil matrix (Jarvis, 2007).

Many surface and subsurface soil properties and consequently runoff processes are affected by the presence of vegetation. Aboveground partitioning of incoming rainfall in interception, stemflow and throughfall (Crockford and Richardson, 2000) causes an evaporative loss term in the water balance and smoothes the intensity of rainfall reaching the soil surface (Keim et al., 2006). Canopy heterogeneity results in spatial patterns of throughfall amount und soil moisture (André et al., 2011; Gerrits et al., 2010; Keim et al., 2005; Zimmermann et al., 2008). Virtual experiments suggest an effect of interception losses on the timing, peak flow rates and total volume of subsurface flow (Keim et al., 2006). In contrast, spatial throughfall patterns have only small influence on subsurface flow dynamics at steep hillslopes with variable bedrock (Coenders-Gerrits et al., 2012; Hopp and McDonnell, 2011). However, there is some field evidence that concentrated water flow via stemflow locally increases soil water content and triggers the downslope onset of subsurface flow above bedrock (Liang et al., 2007, 2011). Ground vegetation and litter layers can be of additional importance for evaporative losses (Gerrits et al., 2010) and may act as a barrier to or enhance infiltration (Sato et al., 2004; Schume et al., 2004).

By root water uptake and hence transpiration, vegetation alters the spatio-temporal soil water distribution and subsurface flow dynamics (Barnard et al., 2010; Tromp-van Meerveld and McDonnell, 2006a). Root-derived preferential flow pathways are often visualized in vegetated soils via dye tracing experiments (e.g. Alaoui et al., 2011; Anderson et al., 2009a; Bachmair et al., 2009; Blume et al., 2008; Bundt et al., 2001; Lange et al., 2009).

Most studies of vegetation influence on hydrological dynamics are designed as either field experiments at the plot scale or as virtual experiments at the hillslope scale (Bachmair and Weiler, 2012a). In contrast, field studies at the hillslope scale that explore the influence of vegetation are rare. Notable exceptions are comparative hillslope studies of Nordmann et al. (2009), Burke and Kasahara (2011), Jost et al. (2012) and Bachmair et al. (2012). These studies showed effects of different tree species and grassland on soil moisture and subsurface flow dynamics. Still, it proved difficult to clearly isolate vegetation factors from others, such as soil properties (Bachmair and Weiler, 2012a; Jost et al., 2012).

As pointed out, vegetation interacts with many other controls of hydrological dynamics and affects the connectivity of flow pathways across scales. Nevertheless, effects of vegetation presence for above, on and below ground water fluxes are often disregarded in hydrological studies at all scales (Bachmair and Weiler, 2011). This is even more obvious for surveys with regard to water transit time and event vs. pre-event water contributions to runoff, although vegetation-soil interactions may be of key importance (Bachmair et al., 2012). Recent literature on water age and event water dynamics is mainly focused on landscape structure, topography, soil and storage distribution at the catchment scale (e.g. Asano and Uchida, 2012; Broxton et al., 2009; Buttle, 1994; Dunn et al., 2007; Hrachowitz et al., 2009; McGuire et al., 2005; Segura et al., 2012; Soulsby and Tetzlaff, 2008; Tetzlaff et al., 2011).

It can be hypothesized that vegetation controls water age through atmospheric water fluxes (interception, transpiration), thus decreases event water inputs and available soil water storage. Transpiration can occur spatially (soil depth, hillslope position) and temporally (events, seasons) preferential and influence water age signatures. Additionally, changed infiltration patterns and lateral preferential flow features at vegetated sites alter runoff generation mechanisms and subsurface flow velocities. Dunn et al. (2007) and Sayama and McDonnell (2009) identified unsaturated zone dynamics as first-order control of water transit times at the catchment scale. Asano et al. (2002) mention that root water uptake may decrease soil water storage volume and hence shorten hillslope soil- and groundwater residence times. Canopy cover was found to be a descriptor of mean water water age, with longer transit times for higher canopy cover at north facing slopes (Broxton et al., 2009). Shanley et al. (2002) reported increasing pre-event water contributions to streamflow with increasing forest cover, however also linked to catchment size and ground frost. An only weak and non significant relation between baseflow age and shrub cover resulted from a recent study (Mueller et al., 2012). Roa-Garcia and Weiler (2010) found similar event water contributions for grassland and forest dominated catchments, but shorter transit times for forest due to higher evapotranspiration. These partly contrasting results, in particular between hillslope and catchment scale studies, support the presence of complex interactions of vegetation, flow pathways and water age dynamics.

1.2 Objectives and hypotheses

The aim of this work is to explore the influence of contrasting vegetation cover on hillslope hydrological dynamics in terms of flow pathways and solute transport. It is built on an existing large-scale hillslope intercomparison setup of three adjacent well-instrumented hillslopes with the objective to characterize interrelations between subsurface flow processes and vegetation type. The hillslopes exhibit similar topography and parent material, but differ in vegetation cover (grassland, coniferous forest, mixed forest) (Bachmair and Weiler, 2012a, in review; Bachmair et al., 2012).

The following research hypotheses and questions will be explored in a stepwise, but combined field- and model-based approach to test the general hypothesis that **vegetation is a first-order control of hydrological dynamics at the hillslope scale (H1)**:

Field study (chapter 2):

(H2) Vegetation is a first-order control of subsurface flow and solute transport.

- Can solute transport give additional information about the connectivity, velocity and pathways of subsurface flow?

Model study (chapter 3):

(H3) Vegetation is a first-order control for parameterization of a hillslope model.

- Can the model replicate observed hydrometric and solute transport dynamics?
- Do the model parameters isolate the vegetation effect from e.g. topography?
- What is the water balance of the hillslopes?

2 Field study

2.1 Data and methods

2.1.1 Site description

The studied hillslopes are situated within a small v-shaped, zero-order catchment at the foot of the Black Forest in southwestern Germany (Figure 2.1). A small creek drains an area of 0.21 km² (outlet location: 47.957° N, 7.838° E) and and the catchment covers an elevation range of 340 - 585 m above sea level. Also during dry summer periods creek baseflow does not totally cease (flow rate < 1 l/s). The three instrumented hilsllopes (Figure 2.2) are located adjacent to each other at the northwest facing side slopes and hence have similar planar topography, parent material, aspect (~ 330°) and slope (~ 26°), but differ in vegetation cover (Bachmair et al., 2012).

The climate is warm temperate (“Cfb” Koeppen classification) with mean annual precipitation 970 mm and air temperature 11°C (years 2007-2011). Evapotranspiration is highest during the summer months. Also mean monthly precipitation peaks in summer due to frequent high-intensity convective storms (Bachmair et al., 2012).

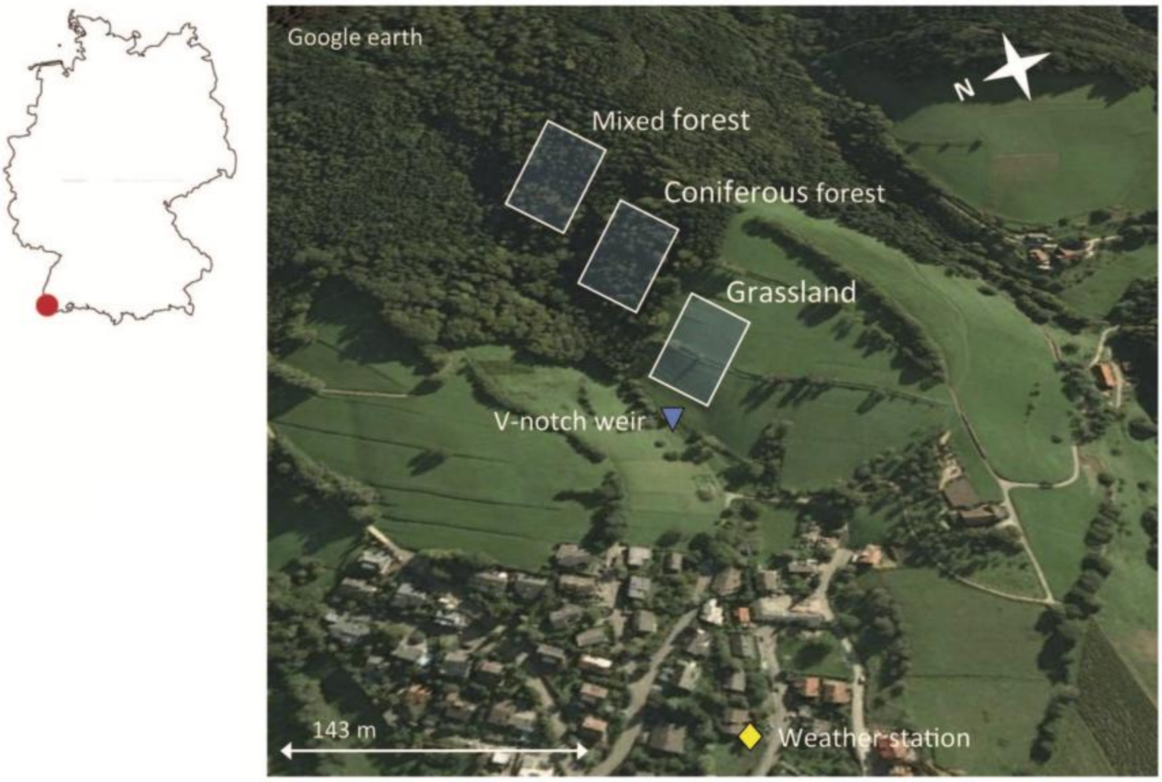

Figure 2.1: Location and aerial photograph of the hillslopes (from Bachmair, 2012).

Geology is composed of crystalline bedrock and overlain by periglacial drift cover which is mainly formed by solifluction, cryoturbation and loess mixed into the local-borne material (Bachmair et al., 2012; Kleber et al., 1998). Three lithologic units can be distinguished in the drift cover, a basal layer, an intermediate layer and an upper layer (Kleber et al., 1998; Völkel et al., 2001). The basal layer exclusively consists of local bedrock and regolith with high bulk density and compacted, slope-parallel aligned clasts. In the intermediate layer the coarse fraction is characterized by finer-sized clasts, varying clast orientation and bulk density; it often features high loess content. The upper layer is the predominant unit of root growth with the finest texture and lowest bulk density (Kleber et al., 1998; Völkel et al., 2001). Both, the intermediate-basal and upper-intermediate layer interfaces are documented zones of perched water table development and subsurface flow pathways (Chifflard et al., 2008; Kleber et al., 1998; Nordmann et al., 2009; Völkel et al., 2001). In the drift cover cambisols have developed at all three hillslopes with loamy texture. Soil texture differences are greater profile-wise than between hillslopes (for detailed soil properties see Bachmair et al., 2012).

The three hillslopes are covered by grassland, coniferous forest and mixed forest (Figure 2.3). The grassland is used for sporadic sheep grazing during summer. Forest to grassland conversion took place around 200 - 300 years ago. The coniferous forest is dominated by spruce (*Picea abies*), silver fir (*Abies alba*) and douglas-fir (*Pseudotsuga menziesii*). At the lower near creek hillslope part few sycamore maple (*Acer pseudoplatanus*) and European ash (*Fraxinus excelsior*) are found. Understory vegetation is dense and deadwood frequent, a needle layer covers the forest floor. The mixed forest is dominated by European beech (*Fagus sylvatica*) and silver fir, with some sycamore maple, European ash and spruce interspersed. Understory vegetation is rare and only little deadwood is found. A thick layer of beech leaves covers the soil surface. At both forested hillslopes tree age is 70-100 years. (Bachmair and Weiler, 2012a; Bachmair et al., 2012).

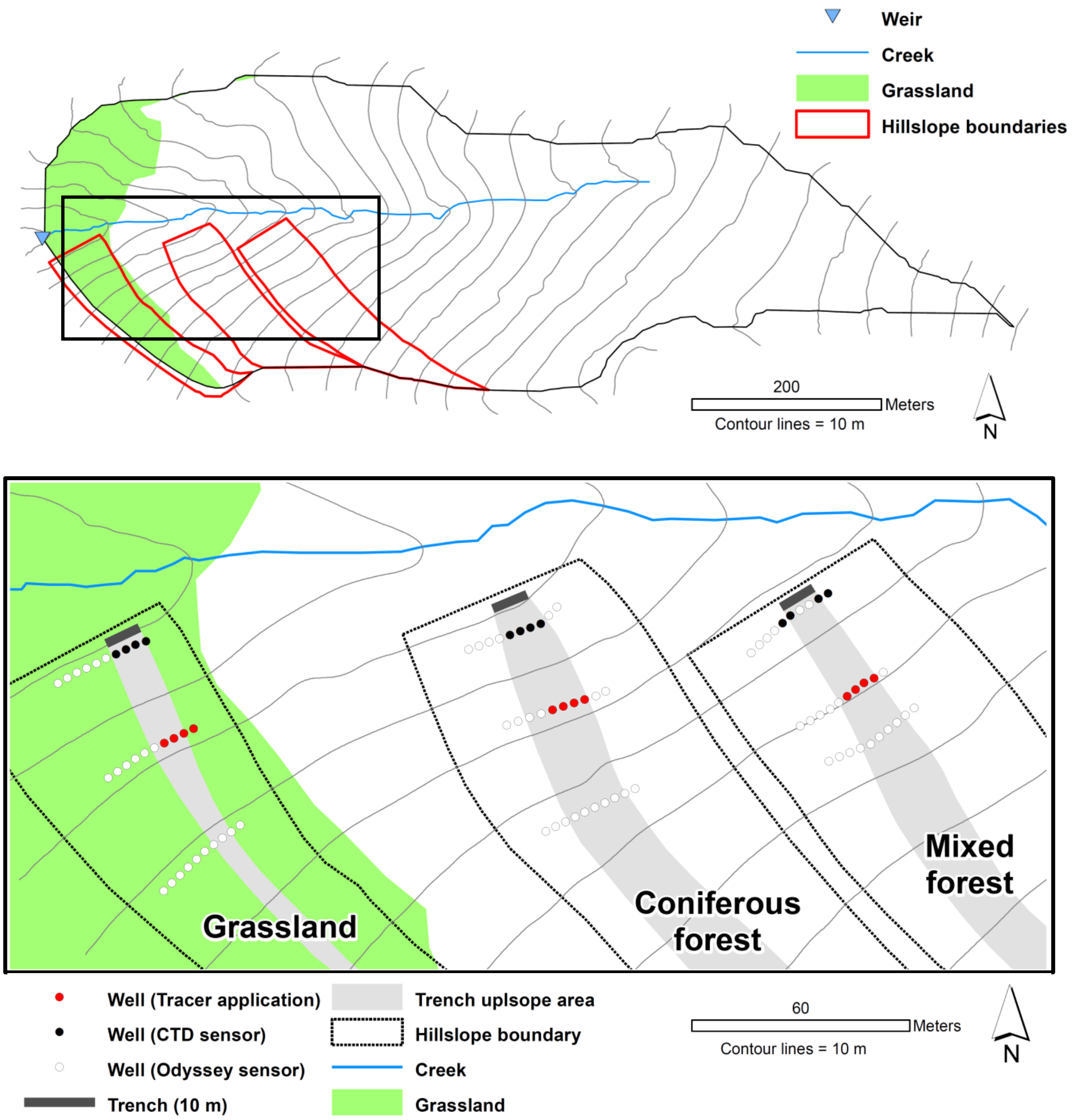

Figure 2.2: Overview of the catchment and location of hillslopes (top), locations of hillslope instrumentation (bottom).

Grassland **Coniferous forest** **Mixed forest**

Figure 2.3: Photographs of the three hillslope vegetation types at the middle well transects (October 2012).

2.1.2 Monitoring design

At the three hillslopes 90 wells were installed (Figure 2.2) to monitor the internal subsurface response to precipitation in high spatio-temporal resolution (Bachmair and Weiler, 2012a, 2012b, in review; Bachmair et al., 2012). At each hillslope 30 wells are arranged in three transects of always 10 wells along the contour lines. Generally, the transects are spaced 30 m apart, except for the upper mixed forest transect (15 m due to an intersecting forest road) and distance between wells per transect is 3 m. A hand-held gasoline-propelled breaker (Cobra Standard) was used for well drilling. The aimed drilling depth was 2 m, but actual well depths depend on subsoil conditions. Many wells are shallower as the resistance of the periglacial drift cover or single boulders prevented deeper drilling. Mostly the wells end in dense drift cover layers far above the actual bedrock. PVC pipes (diameter 40 mm) were perforated over the entire length, wrapped in geotextile, inserted into the wells and sealed around with bentonite clay at the soil surface.

The majority of wells was equipped with Odyssey Capacitance Water Level Recorders (Data Flow Systems, New Zealand) with length between 1.0 and 2.0 m according to well depths (Figure 2.2). The integrated data loggers were set to 2 min measuring interval but only water level changes > ±5 mm were recorded (compressed logging mode). For this study, data of the nearly 90 capacitance probes were downloaded in April and October 2012, the probes were cleaned and water table heights were measured via an electronic contact gauge for water level validation. Accuracy of calibrated probes is 5 mm according to the manufacturer's specifications. However, field validation regularly showed deviations of up to 25 cm for completely dry wells and several cm for higher water tables. Hence, probe readings lower than 30 cm were omitted to reduce data uncertainty. During the study period May - October 2012 at each hillslope four capacitance probes were replaced with CTD sensors (Decagon Devices, USA) to monitor not only water level but also electrical conductivity in the wells (Figure 2.2). At each hillslope the CTD sensors were connected to a Campbell data logger (models CR800 and CR1000, Campbell Scientific, USA). The measurement were taken in 1 min intervals and stored as 5 min averages. Manual water level validations yielded satisfactory deviations < 1 cm (manufacturer reported accuracy: 0.20%). Electrical conductivity accuracy was tested in the laboratory prior to field installation with a calibrated (1413 µS/cm, 25°C) hand-held EC meter (LF-325, WTW Wissenschaftlich-Technische Werkstätten GmbH, Germany). Most sensors deviated less than 5% and hence fall within the specificated range (manufacturer reported accuracy: 10%). However, in the beginning some sensors failed after field installation and had to be replaced. Thereof several data gaps and inconsistent time series length resulted between the wells and hillslopes.

In addition to water tables also subsurface flow dynamics were monitored via hillslope trenches. At each hillslope a 10 m long trench was excavated close (some meters) to the lowest wells transect (Figure 2.2). After installation of a drainage system the trenches were back-filled to ensure hydraulic continuity. For the grassland and coniferous hillslopes the trench depth is 2 m, whereas at the mixed forest the irregular bedrock surface was hit at depths between 0.6 and 1.4 m. Trench outflow is separately measured for the left and right 5 m wide trench sections via tipping buckets (RainWise, USA). For the left trench section at the coniferous hillslope a larger tipping bucket was constructed (~100 ml bucket volume) because the flow rate often exceeded the capacity of the smaller RainWise model (max. 15 ml volume). Tipping bucket counts were also stored as 5 min sums with the Campbell data loggers. The tipping bucket calibrations were regularly checked in the field (weekly to biweekly) and yielded volumetric deviations of less than 5%. Frequent data gaps resulted from failures of the reed switches and electrical contact problems.

A small v-notch weir is installed in the creek at the catchment outlet (Figure 2.2). It was equipped with a CTD-Diver to measure water level and electrical conductivity and with a Mini-Diver (both Schlumberger Water Services; 5 min measurement interval) for atmospheric pressure correction. Water levels and electrical conductivity were regularly (weekly to biweekly) validated. Additionally, manual volumetric discharge measurements (bucket method) at 10 dates between May and August 2012 at various flow conditions validated the stage-discharge relationship. Rainfall and standard meteorological parameters were measured with a Davis Vantage Pro 2 weather station (15 min measurement interval). It is located 260 m northwest of the catchment outlet at an elevation of 316 m above sea level (Figure 2.1). A short rainfall data gap (15/08 - 27/08/2012) was filled via linear regression with daily precipitation information from a neighboring weather station located approximately 2 km north-northwest (regression based on the period 19/06 - 14/08/2012; R^2 = 0.93). More technical details about the construction and installation of the monitoring infrastructure can be found in Bachmair and Weiler (2012b).

As pointed out before, the study is built on on the earlier PhD work of Bachmair (2012). Based on the same experimental setup the following main findings about the spatio-temporal variability of subsurface flow processes, the explainability of shallow water table dynamics, effect of vegetation cover and the scaling of hydrological processes are:

Subsurface flow response at the plot scale (well observations) and at the hillslope scale (flow from different trench sections) is highly variable in space and time. Adjacent wells showed different responses to rainfall (dry vs. water table rise in 3 m distance) and a homogenous transient water table along the transects was never observed. In contrast, during certain events distinct spatial patterns of subsurface saturation emerge and suggest upslope expanding flow paths. The spatial patterns were found to vary seasonally. During summer well water tables of all transects rose quickly and strong but with high spatial variability. At wetter conditions (fall/winter/spring) mainly the lower transect wells were saturated with generally weaker and slower water table rise. The pattern of water table response was explainable only to a low degree based on measured hillslope characteristics, particularly for high rainfall intensities and dry antecedent conditions. Also vegetation seemed to be only of minor importance. Therefore, other not mapped factors may be of additional importance, e.g. preferential pathways, bedrock topography or hydrophobicity. Trench flow dynamics were found to be unrelated to vegetation cover, however the internal response of the grassland hillslope differs from the forested hillslopes. Obviously the grassland flow paths could not be captured by the 3 m well spacing and the water table response was less predictable compared to the forest hillslopes. Trench flow response was related to shallow water table dynamics, whereas water table dynamics reflect hillslope outflow dynamics only when they capture the spatially limited flow paths. Hillslope runoff was identified as strong driver of catchment peak runoff, but other processes and flow pathways (overland flow, riparian zone, deeper subsurface flow) must be additional contributors to total catchment runoff.

2.1.3 Tracer application

Subsurface flow processes are difficult to observe but natural and artificial tracers can yield important insights into the involved flow pathways and their connectivity, flow velocities and subsurface heterogeneities (e.g. Wienhofer et al., 2009). The intention of this tracer experiment is to move beyond the existing hydrometric observations at the three hillslopes. Observations of the tracer transport dynamics may give additional information about similarities and dissimilarities about the hillslopes' general functioning and the effect of different vegetation cover (H2). Furthermore, the results will be integrated in the numerical simulation of the hillslopes.

Chloride (in form of sodium chloride, NaCl) was selected as non-sorptive and conservative artificial tracer for the experiment. The rationale for this choice is the challenging task of monitoring tracer breakthrough at several locations (6 trench sections and weir) in high temporal resolution (minutes). When applied in large amounts, the breaktrough of NaCl can be detected and quantified via measurements of electrical conductivity (EC) and a site-dependent concentration-EC relationship at the hillslope scale (Anderson et al., 2009b; Wienhofer et al., 2009).

Based on the reported tracer studies (Anderson et al., 2009b; Wienhofer et al., 2009) a total tracer amount of 60 kg NaCl was applied at the hillslopes mid-June (mixed forest: 14+15/06/2012, coniferous forest: 16+17/06/2012, grassland: 18/06/2012). The large tracer mass was used because high dilution, retention in the unsaturated zone and deep percolation losses were expected. At each hillslope 20 kg of NaCl tracer (supermarket table salt) was injected into each of the four wells (5 kg each, equivalent to a pure chloride mass of 3 kg) of the middle transect located around 30 m from the trench in its upslope area (Figure 2.2). For each well the NaCl tracer was dissolved in 15 l of water from the creek. This is the smallest possible amount of water as solvent and should preserve natural flow conditions. The tracer solution was poured into the well pipe with a hose, a bit shorter than well depth. In this way the retention of tracer in the unsaturated zone was intended to be small. However, despite wet antecedent conditions, water tables in the wells were generally low (mixed forest 10 - 70 cm, coniferous forest 0 - 15 cm, grassland all wells dry). On the other hand, a hydraulic head of approximately 1 m within the wells was necessary to infiltrate the 15 l of tracer solution in reasonable time (< 3 h). Therefore, large amounts of tracer can assumed to be stored in unsaturated soil parts at the beginning of the tracer experiment. After the tracer solution additional 7.5 - 15 l of pure creek water were poured on top of the well water table to push the tracer solution out of the well pipe.

During the next ~ 5 months tracer breakthrough was monitored as EC in trench flow and additionally in the four near trench wells with CTD sensors (Figure 2.2). The CTD setup was slightly modified for the mixed forest hillslope due to the known bedrock ridge at the trench location. Hence, subsurface flow is probably diverted around the trench there. Trench outflow EC was separately measured for the left and right trench sections. Therefore, at each trench two flow-through cells were installed at the drainage system outlets and also equipped with Decagon CTD sensors and connected to the Campbell Scientific data loggers. The flow-through cells were constructed from 100 ml plastic sample bottles with the top cut off and ensured a small mixing volume and proper flow along the sensor. After passing the CTD sensor, the water spilled over into the tipping bucket. Weekly to biweekly the trench flow EC readings from the CTD sensors were checked with the calibrated hand-held EC meter (WTW LF-325). For all trench sections individual linear regressions were established ($R^2 > 0.98$) to correct for systematic measurement errors and ensure comparability between the trench sections. Grab samples of trench flow were taken before and after the tracer injections from all trench section in order to establish a functional chloride-EC relationship. Most samples were taken one day after rainfall events during flow recession or when residual water in the flow-through cells yielded at least 30 ml. Creek discharge was grab sampled at the weir location. The samples were stored cool and dark until analysis. Ion chromatography (Dionex DX 500, accuracy 5%) was applied to determine anion concentrations (chloride, nitrate, sulfate).

2.2 Results

2.2.1 Hydrometric dynamics

Cumulative rainfall for the monitoring period 01/05 - 22/10/2012 is 715 mm and 443 mm after the start of the tracer experiment (15/06/2012), respectively (Figure 2.4(a)). After a series of intense convective storms in June and at the beginning of July, frontal systems contributed to rainfall from September onwards. Three distinct hydrological sub-periods can be identified from the creek hydrograph (Figure 2.4(b)): A considerably wet May until mid-August, causing persistent high flow rates; mid-July until mid-September with high evaporative demands, infrequent small rainfall events and hence mostly baseflow conditions; a successive rewetting period starting mid-September and peaking with a large runoff event at the end of the study period. High antecedent wetness and the series of large rainfall events in June caused the largest observed creek discharge (actual hydrograph peak missing due to weir blocking by sediment and deadwood) since the initiation of the hillslope and catchment monitoring three years ago. Return flow was prominent at several concave footslope locations, however strongly limited in its spatial extent (Figure 2.5 left). Opposite to the grassland at the footslope an active soil pipe (diameter 4 - 5 cm) could be excavated 15 cm below the soil surface. It originated from a mousehole and could be tracked over a total distance of 5 m, but was actually longer (Figure 2.5 right).

Hillslope shallow water tables showed high spatio-temporal variability as already observed earlier at this site (Bachmair et al., 2012). Water table dynamics (hillslope means and standard deviations) are calculated as water table depth below surface, shown in Figure 2.4(c)-(e). The actual number of wells included in the statistics varies temporally and between the hillslopes, due the omission of uncertain low water tables and probe failures. Data from the 12 injection wells also were omitted because the high salt concentrations interfered with the capacitance probes after tracer solution was added. Generally, the water table dynamics show distinct differences between the grassland and both forested hillslopes. The grassland wells were activated less frequent, display lower water tables and only weak event responses even in June. In contrast, at the forested hillslopes water tables roughly resemble the creek discharge dynamics. This behavior is in agreement with the findings of Bachmair et al. (2012).

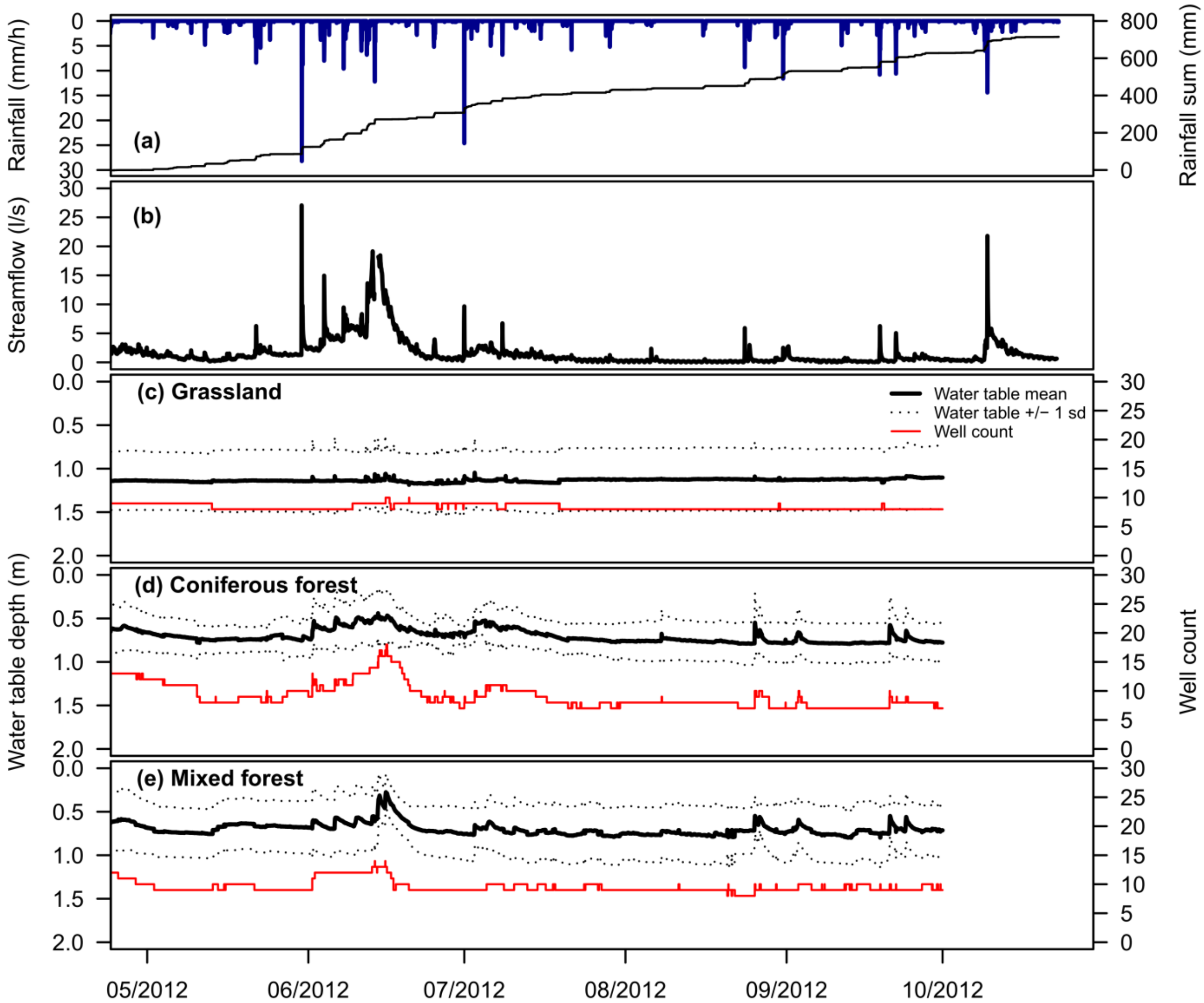

Figure 2.4: Observed (a) precipitation, (b) creek runoff, (c)-(e) spatially water table dynamics. Well count gives the number of wells included in water table statistics calculations.

Figure 2.5: Return flow at the grassland footslope, channeled in a small dug ditch (left), pipe flow originating from a mousehole (right) (June 2012).

As shown in Figure 2.6(c)-(e) trench flow dynamics strongly fluctuated. In addition, the correlations between the left and right sections of the same hillslope or among the three hillslopes were found to be complex and vary between events (Bachmair and Weiler, in review). Flow rates and peaks can be very different for the individual sections of a trench which is obvious for the grassland and coniferous forest hillslopes. Outflow from these hillslopes tailed off after rainfall events and ceased only during the dry summer period, whereas the mixed forest trench showed a different response pattern. There, trench flow was restricted to times of rainfall and flow ceases immediately after events. Also maximum flow rates from the mixed forest trench were considerably lower compared to the other hillslopes. This can be attributed to the shallow trench depths and local bedrock ridge. Similar to flow rates, also the total flow volumes during the study period vary over orders of magnitude between the hillslopes (Table 2.1), despite significant data gaps in all time series. The highest flow volume at the coniferous forest hillslope may be attributable to the local surface microtopography expressed in the larger upslope area. However, compared to creek runoff at the weir, contributions from subsurface flow measured as trench flow are low for the whole period. Bachmair et al. (2012) described similar results for several rainfall events during the year 2011.

Table 2.1: Upslope areas, flow and missing data statistics for the catchment and hillslope trench sections (period 01/05 - 22/10/2012).

	Creek	Grassland			Coniferous forest			Mixed forest		
		left	right	both[a]	left	right	both[a]	left	right	both[a]
Upslope area (m²)	214497	-	-	1340	-	-	3138	-	-	2571
Total flow (l)	-	3216	589	1820	18644	3662	20084	39	10	49
Total flow (mm)	104	-	-	1.36	-	-	6.40	-	-	0.02
Runoff coefficient (%)[b]	14.58	-	-	0.19	-	-	0.90	-	-	0.003
Missing data (%)	0.7	14.2	31.9	34.7	11.7	6.1	11.8	14.0	17.1	17.1

[a] calculated only from periods of coincident data availability

[b] based on 715 mm total precipitation

2.2.2 Tracer response

The EC dynamics in trench flow and the near trench wells are depicted in Figure 2.6. Data from the grassland and coniferous trenches show similar EC ranges and strong fluctuations (Figure 2.6(c)-(e)). During rainfall events EC rapidly decreases, indicating a contribution of event water and hence dilution of trench flow. For the coniferous forest both trench sections have similar dynamics, while the grassland trench sections exhibit divergent EC behaviors. This is most obvious during July when left section EC slightly decreases and right section EC strongly increases. However, it must be noted that during that period trench flow was small and dominated by the left trench section, as can be seen from the flow weighted mean EC.

Generally, a tracer breakthrough of chloride should elevate EC compared to the background value before the tracer application. Though, there is no such discernible difference in EC for the pre- and post-application period at the coniferous hillslope. At the grassland hillslope an increase in EC is visible despite high scatter and missing right section trench flow data before the tracer application. Because of the EC drops during rainfall events and an anticipated tracer mobilization mainly during events, also dampened dilution of trench flow can indicate a tracer breakthrough. This might be the case at the coniferous forest hillslope. There the trench outflow (mainly from the left section) EC remains on a near constant level for several weeks after the salt injection. Particularly during the large rainfall event in the beginning of July, EC dynamics show no dilution and two distinct small spikes linked to flow peaks.

These findings are supported by the near trench wells (Figure 2.6(f)-(h)). At the grassland hillslope the highest mean EC values correspond to an increase of EC in the two left wells (well 1 and 2). Interestingly though, the right trench section dominated the trench flow rate. The coniferous forest wells reacted strongly to the rainfall events during June and July due to the surface microtopography (relatively flat and large upslope area). Water levels rose up close to the surface and return flow was observed near the wells. During six weeks after the tracer application EC increased continuously in three wells, one well fell dry quickly. This corresponds well with the dampened EC dynamics observed for trench flow. The high water tables additionally may act as a buffer through mixing of pre-event and event water and thus reduce EC dynamics in the wells and trench flow. As soon as the wells dry out in August, subsequent rainfall events again imprint their dilution signal in the trench flow EC.

The mixed forest results again stand out. Saturation in the wells developed only during a few events for three wells, well 1 was nearly permanently saturated. However, all water tables were limited to a few cm above the well bottom. Because of the very intermittent observed trench flow and well dynamics no tracer breakthrough can be identified. Additionally, EC in the wells and trenches display very wide ranges (up to 2500 µS/cm) even before tracer was applied.

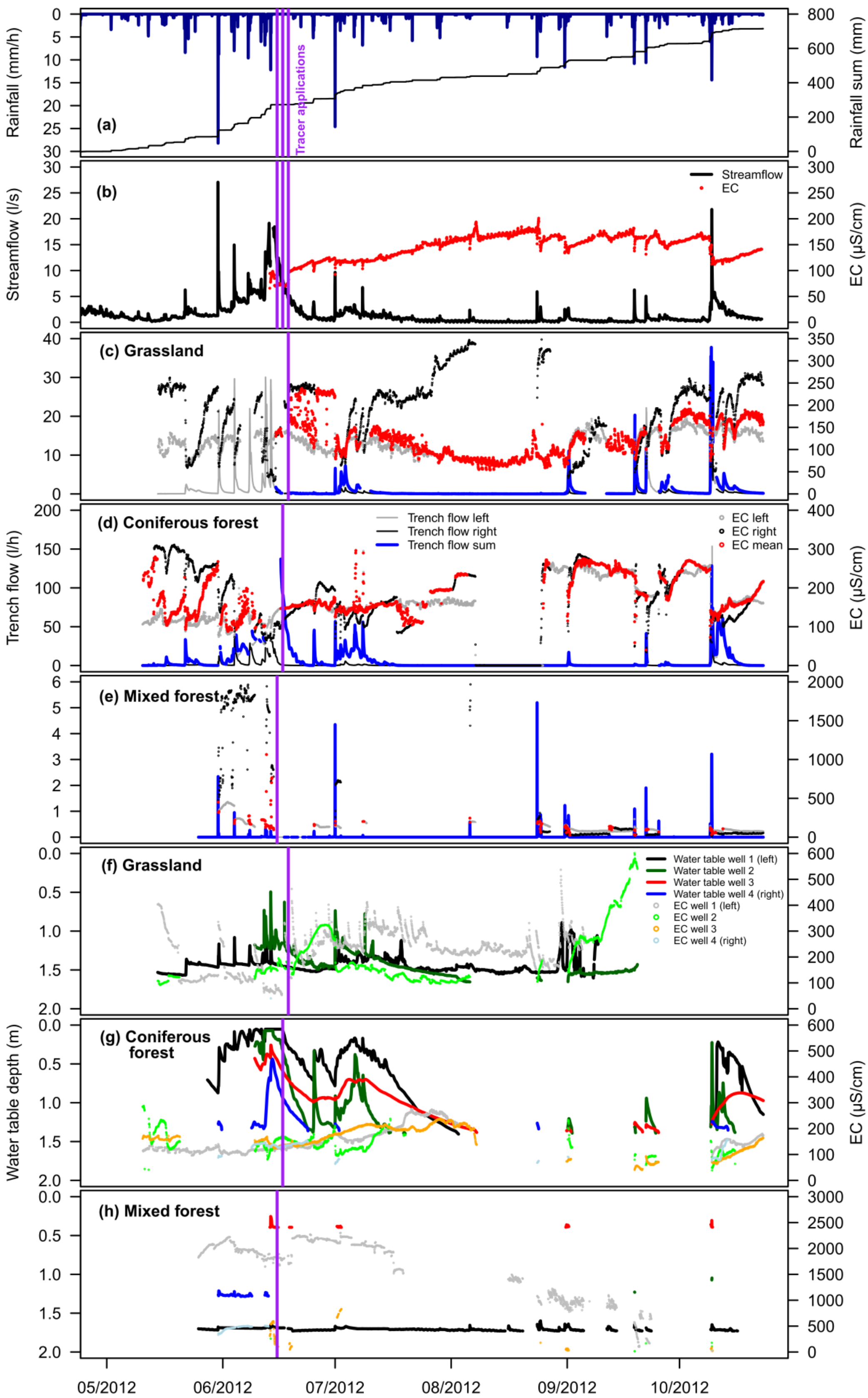

Figure 2.6: Observed (a) precipitation, (b) creek runoff and electrical conductivity (EC), (c)-(e) trench flow and EC, (f)-(h) near trench water table and EC dynamics. The vertical purple lines indicate tracer applications.

To quantify the mass of transported chloride, water samples from the trenches and the creek were analyzed (Table 2.2). Sample sizes at each location are relatively small ($n \leq 8$) because trench flow was intermittent. Also it was waited for an expected stronger increase of EC during the first weeks after the tracer injections. Thus, the (in hindsight) observed tracer breakthroughs were missed with the sampling procedure.

The anion concentrations are not uniform for left and right sections of the individual hillslopes and also differ between the hillslopes. Surprisingly high nitrate concentrations were found at all trenches (>> 50 mg/l), though the creek samples had usual background concentrations (~ 10 mg/l). Sulfate was higher in the creek and coniferous forest trench flow (~ 20 mg/l) compared to the grassland and mixed forest (~ 4 mg/l). Chloride concentrations were mostly low before the tracer application (< 4 mg/l), only two pre-tracer samples of the grassland trench displayed higher values (> 6 mg/l). After the tracer application concentrations increased slightly, but only three samples contained > 10 mg/l chloride.

Linear regression between chloride content and EC of the samples resulted in a variety of slope and intercept parameters for the different sample locations (Table 2.3). For most hillslope sites the regression yields either a poor fit ($R^2 < 0.3$) or a slope close to zero. Only for creek samples a reasonable fit and slope was achieved ($R^2 = 0.91$). It seems that EC dynamics at the hillslope scale is strongly driven by other solutes, such as nitrate and sulfate, at least at the sampling times. The regression results may also indicate a variable hydrochemical composition of trench flows. Due to this and the unknown, probably variable background concentrations of chloride no continuous chloride time series were calculated. Yet, a rough estimation of chloride mass recovery in the trenches until the end of the study period was done with the derived regression parameters and flow rate data. Background chloride was assumed as 0 mg/l to get the maximum theoretically possible recovery mass. Despite this unrealistic assumption, due to the small total flow volumes the estimations yielded very small masses, 30 g (grassland), 1 g (mixed forest) and 65 g (coniferous forest). Although the values are the maximum possible masses based on the available samples, it can be speculated that sampling during the days and weeks after the tracer applications would have lead to stronger chloride-EC relationships. Thus the calculated values may also underestimate real chloride recovery in the trenches.

Table 2.2: Anion concentrations (mg/l) of sampled trench flow and streamflow. Negative times indicate sampling before tracer applications.

Location	Time since tracer application (d)	Chloride left /right	Nitrate left /right	Sulfate left /right
Creek[a]	-32	3.6	8.4	20.1
	-28	3.7	8.5	19.7
	-1	2.9	9.4	15.0
	96	4.6	13.5	19.8
	102	4.2	9.8	21.0
	104	4.3	10.1	19.7
Grassland	-36	2.9 / n.s.[b]	29.3 / n.s.[b]	4.3 / n.s.[b]
	-32	2.8 / 7.5	19.3 / 74.8	3.9 / 2.4
	-4	6.5 / 2.2	34.9 / 3.4	2.7 / 4.3
	50	3.8 / 20.0	13.1 / n.d.[c]	3.9 / 0.7
	70	4.2 / 31.2	13.5 / 1.0	3.5 / 7.9
	93	8.7 / 6.3	19.3 / 4.9	2.2 / 2.1
	99	6.6 / 6.1	42.8 / 61.1	3.0 / 2.3
	101	7.7 / 5.2	28.6 / 22.3	2.2 / 2.2
Coniferous forest	-34	3.2 / 3.7	18.4 / 93.1	17.7 / 15.7
	-31	3.1 / 3.6	16.1 / 82.9	15.9 / 14.4
	-2	3.9 / 2.0	27.6 / 18.2	20.4 / 11.1
	30	4.4 / 3.2	17.9 / 32.0	25.0 / 15.6
	71	4.7 / 3.2	28.4 / 47.6	18.5 / 19.3
	94	3.6 / 2.6	39.6 / 82.0	21.3 / 16.3
	100	3.7 / 3.4	19.1 / 66.9	18.9 / 18.3
	102	3.1 / 2.9	26.2 / 49.3	21.8 / 13.6
Mixed forest	0	1.7 / 3.3	0.9 / 0.0	5.6 / 25.3
	53	3.5 / 10.1	44.53 / > 100	4.8 / 32.8
	73	3.8 / 2.0	4.8 / 0.9	3.9 / 4.2
	96	3.2 / 1.9	33.3 / 10.9	4.3 / 3.8
	102	2.7 / 1.8	5.5 / n.d.[c]	4.5 / 3.0

[a] Time since tracer injection based on mixed forest, [b] n.s. = not sampled, [c] n.d. = below detection limit

Table 2.3: Results of linear regressions between anion concentration (independent variable) and EC (dependent variable).

Location	Chloride			Nitrate			Sulfate		
	R^2	slope	intercept	R^2	slope	intercept	R^2	slope	intercept
Creek	0.91	0.03	0.28	0.22	0.04	4.61	0.81	0.08	8.13
Grassland left	0.13	0.04	1.59	0.92	0.44	-21.74	0.09	-0.01	4.34
Grassland right	0.71	0.07	-2.14	0.00	0.01	22.38	0.03	0.00	2.48
Coniferous forest left	0.35	0.01	2.53	0.20	0.07	12.18	0.03	0.01	18.29
Coniferous forest right	0.63	0.00	1.58	0.76	0.36	-19.44	0.55	0.03	9.05
Mixed forest left	0.23	0.01	1.77	0.43	0.31	-21.79	0.01	0.00	4.47
Mixed forest right	0.99	0.00	1.57	0.13	0.00	3.46	0.80	0.01	5.82

In comparison to the injected pure chloride mass at each hillslope (~ 12 kg), the EC dynamics, measured concentrations and calculated recovered masses are very small. Thus, in October 2012 a great part of the applied tracer was still retained in and below the soil zone or had left the hillslopes through deeper unmonitored flow pathways. The latter assumption was tested via analysis of the creek EC time series (Figure 2.6) and chloride-EC functional relationship (Table 2.3). As can be seen after applying the tracers, EC increased continuously at the weir location. Cumulative chloride mass and runoff volume were plotted as double mass curve (Figure 2.7). Breakpoints in the slope were identified and linear regression lines fitted segment-wise (R package “segmented”; Muggeo, 2003). A first rise of slope occurred two to five days after the tracer injections, a second after an additional month; in October the slope decreased again. Interestingly the timing of breakpoints matches the observations made at the grassland and coniferous forest hillslopes (Figure 2.6). Hence, the first break may be interrelated to the increasing EC in the grassland wells, only two days after the start of the tracer experiment at this hillslope. The second break coincides with the high EC values observed in the coniferous forest wells in mid-July (especially well 1). The last break towards a lower slope is clearly related to the intense rainfall event, which probably changed the hydrochemical composition and clearly diluted the discharge. The recovered chloride mass at the catchment outlet can be estimated by extension of the segmented slopes. The difference at the y-axis gives an additionally exported mass of ~ 12 kg. This is one-third of the total applied tracer mass at all three hillslopes. Of course, this mass also includes potential chloride contributions from the mixed forest, although the transport could not be observed at this hillslope. However, also no third positive break of slope was identified, suggesting only minor or more diffuse tracer export from the mixed forest site. Clearly breakpoints do not prove causality and the EC monitoring in the creek started only immediately before the onset of the tracer experiment adding uncertainty of the initial slope. Yet, it gives strong and plausible evidence of a significant tracer recovery at the catchment scale.

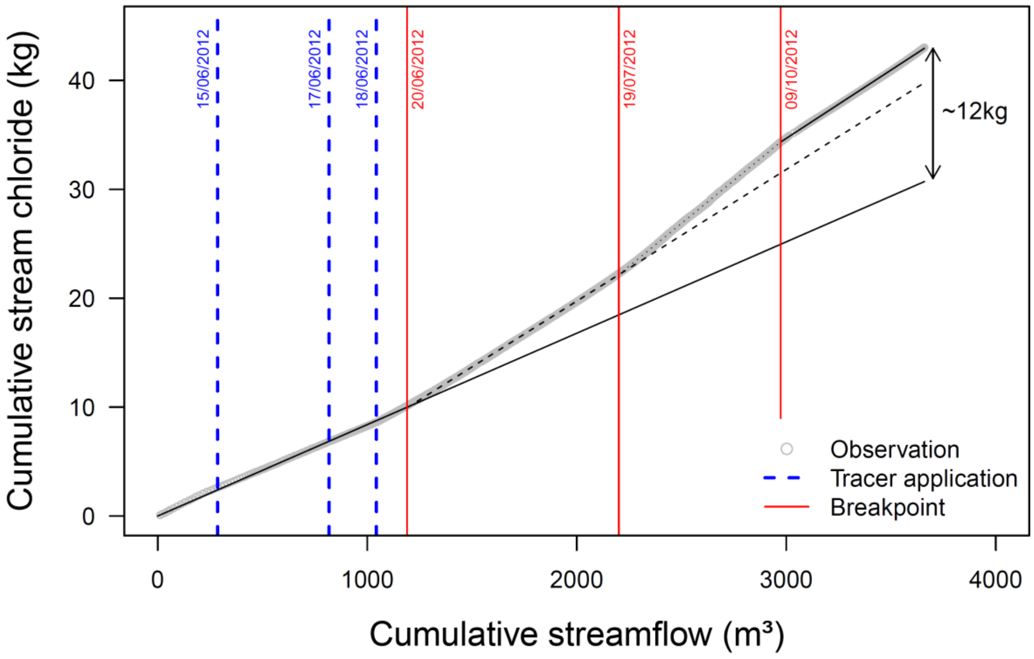

Figure 2.7: Double mass curve of chloride mass and runoff at the weir.

3 Model study

3.1 Data and methods

3.1.1 Model description

Bachmair and Weiler (2012a) identified topography and soil properties as rather important predictors of water table response. The role of vegetation cover, however, was difficult to assess and played only a minor role. If a numerical model described both topography and soil parameters spatially explicit, it can be hypothesized (H3) that other parameters must vary between different vegetation covers. Hence, the hydrological hillslope model HillVi was used to test this hypothesis. Futhermore, the model should quantify previously not observed (or not observable) terms of the hillslope water balances.

HillVi is a physically based and spatially distributed model of water flow and solute transport. It was first introduced as hypothesis testing tool for virtual experiments at the hillslope scale (Tromp-van Meerveld and Weiler, 2008; Weiler and McDonnell, 2004, 2006, 2007). Subsequently it was also used in combination with a field experiment (McGuire et al., 2007), at the catchment scale (Gascuel-Odoux et al., 2010; Stoll and Weiler, 2010) and in the context of model intercomparison (Hollander et al., 2009). The model and underlying equations are described in great detail in McGuire et al. (2007), Tromp-van Meerveld and Weiler (2008) and Weiler and McDonnell (2004). HillVi is based on spatial representations of surface and subsurface topography defined by DEM and soil depth information. Each grid cell represents a coupled unsaturated-saturated zone and allows water and solute transfer between both zones under changing water table conditions. Depth from the soil surface to the water table and time-variable water content define the unsaturated zone. Precipitation input, vertical recharge into the saturated zone, evapotranspiration and change in water table determine the water balance of the unsaturated zone. The saturated zone is defined by the thickness of the water table above permeable or impermeable bedrock (or more general, the lower horizontal boundary) and the porosity. The water balance in the saturated zone is calculated from vertical unsaturated zone recharge, bypass flow during rainfall, lateral subsurface in- and outflow, lateral pipe flow (not included here), deep percolation to the lower boundary and the water table change. The Dupuit-Forchheimer assumption is used to calculate lateral subsurface flow with an explicit grid cell by grid cell approach (Wigmosta and Lettenmaier, 1999). For each time step the flow direction and outflow partitioning to the neighboring grid cells is calculated based on the local water table gradients. Deep percolation is calculated from the water table position and the hydraulic saturated conductivity of the bedrock and is treated as loss from the model domain (no exfiltration back into the soil). Descriptions of the integrated solute transport routine can be found in McGuire et al. (2007)

and Weiler and McDonnell (2004, 2005). For this study it is used to replicate the field-based tracer experiment and serve as additional constraint for model evaluation. Complete mixing is assumed in the unsaturated and saturated zone of each grid cell. Solute mass is only transported advectively in and between grid cells. The effective porosity for solute transport was assumed to be 80% of the total porosity (Weiler and McDonnell, 2007).

For this thesis the HillVi source code was transferred from IDL (Interactive Data Language, ITT-VIS) to the open source programming language R (R Development Core Team, 2011). The model structure is consistent with McGuire et al. (2007) with the exception of additional seepage flux (Tromp-van Meerveld and Weiler, 2008).

3.1.2 Input data

Individual HillVi model domains were created for the three hillslopes based on a DEM with a grid size of 3 x 3 m (= distance of installed wells). The domain boundaries covered the whole hillslopes, approximately from the contour line at the trench location to the catchment boundary. They correspond to the hillslope boundaries depicted in Figure 2.2. The downslope outflow boundary measured ~ 60 m to reduce edge effects from the no flow boundaries at both sides. For model calibration subsurface flow was read out every time step at the four grid cells that correspond to the trench dimensions (a multiplicative correction factor was applied to account for the 3 m grid length vs. the trench length of 10 m). Depth to water table was monitored at the grid cells closest to the real well locations at each time step.

Depth to bedrock is not known for most parts of the hillslopes, hence the depth to the basal layer defined the soil depth which was assumed equal to well depths. Based on the geostatistical parameters of the well depths, for each hillslope and model run an unconditional Gaussian simulation defined a random, but spatially correlated pattern of soil depths above the basal layer. At the well locations the actual known well depths were retained to guarantee comparability of the field observed with the simulated water tables positions.

To force the HillVi model a simple interception model was fitted for the forested hillslopes, based on observed throughfall amounts for 8 events during the growing season 2010/2011 (Bachmair and Weiler, 2012a). The model allows canopy interception and evaporation, direct throughfall and slow dripping from the interception storage after events. The fitted models reduced open area rainfall by 18% (coniferous forest) and 29% (mixed forest) and performed well (Nash-Sutcliffe-Efficiencies 0.94 and 0.95).

Potential evapotranspiration was also taken from the weather station. It automatically calculates grass reference evapotranspiration (ET_0) from the meteorological measurements. To correct for the station location outside the catchment close to buildings and allow vegetation differences, a crop factor was introduced. The crop factor is a multiplicative constant which increases or reduces ET_0 for all time steps.

The tracer experiments were replicated by adding the same amounts of tracer mass (3 kg chloride per well) to the corresponding grid cells at the same dates (mid-June). At each tracer location the injected mass was partitioned between the unsaturated and saturated zone depending on the relative water table position to mimic the conditions that were met during the tracer injection in the field. If the simulated water table height in the grid cell was e.g. at 30% of the total soil depth, 30% of the tracer mass was added to the saturated zone and 70% to the unsaturated zone.

Table 3.1 shows the parameters used in this study. For all three hillslopes the same initial parameter ranges were allowed. Total porosity, drainable porosity, the parameter for the decline of the drainable porosity with depth and the unsaturated recharge exponent were estimated from the measured soil properties reported in Bachmair et al. (2012) and an assumed variable stone content of 10 - 50% (Hangen et al., 2001; Uhlenbrook et al., 2008). The saturated hydraulic conductivity was simulated as a spatially uncorrelated random field with a log-normal distribution. Slug test data from all wells were available (Bachmair and Weiler, 2012a). However, a direct interpretation of the observed soil hydraulic properties to derive the related model parameters was not possible. Hence, wide ranges for spatial mean and standard deviation of the saturated hydraulic conductivity at the soil surface and the parameter for exponential decline with depth were chosen. The vertical saturated hydraulic conductivity of the basal layer was described as a reduction factor in relation to the spatially varying lateral (soil) hydraulic conductivity. This ensured a realistic anisotropic behavior of the basal layer with higher lateral than vertical hydraulic conductivity (Kleber et al., 1998). The bypass flow exponent controls which fraction of rainfall is directly added to the saturated zone without storage in the unsaturated zone. The range of the exponent was selected to allow high fractions of bypass flow, a process that has been observed at the hillslopes (Bachmair et al., 2012)

Table 3.1: Model parameters and ranges.

Parameter	Parameter limit lower	Parameter limit upper
Total porosity (-)	0.2	0.35
Drainable porosity at the soil surface (-)	0.08	0.15
Drainable porosity shape factor of exponential depth function (-)	0.1	1
Mean of saturated hydraulic conductivity at the soil surface (log_{10} m/h)	-3	1
Saturated hydraulic conductivity shape factor of exponential depth function (-)	0.01	1
Standard deviation of saturated hydraulic conductivity at the soil surface (log_{10} m/h)	-3	1
Bypass flow exponent (-)	1	15
Unsaturated recharge exponent (-)	8	12
Ratio of vertical to lateral saturated hydraulic conductivity at the basal layer (log_{10})	-3	0
Crop factor (-)	0.7	1.2

3.1.3 Model calibration

HillVi was run and calibrated for the time period 01/05 – 22/10/2012 with one additional month for warm up. Random parameter sets were created using Latin Hypercube Sampling (Helton and Davis, 2003). Input and calibration data time step was 1 h. The internal time step for water and solute flux calculations was shorter during periods of intense flow to minimize water balance errors. Observed trench flow rates (sums of left and right trench sections; Figure 2.6) and the water table spatial statistics (mean, standard deviation) were used as calibration data (Figure 2.4).

The objective function should allow a combined calibration of modeled trench flow, water table and tracer transport dynamics in comparison to the observations. Therefore three separate efficiencies were calculated and merged into a single performance measure.

Trench flow efficiency (E_{SSF}) was calculated as the Nash-Sutcliffe-Efficiency from the simulated and observed trench flow time series (Nash and Sutcliffe, 1970). To regard possible measurement uncertainties, relative error bands of ± 5% were calculated around the observed time series. For time steps with a model deviation inside the band the deviation was set to zero. For time steps with a model deviation outside of the bands the deviation from the band was calculated.

Water table efficiency (E_{GW}) was the average of the two mean relative errors (MRE) for the time series of spatial water table mean and spatial water table standard deviation. Similar to E_{SSF} absolute error bands of ± 5 cm were calculated around the observed time series to account for measurement uncertainty. Note that for every time step only exactly the wells were compared for which observations were available (Figure 2.4). The resulting MRE was rescaled as $E_{GW} = 1 – MRE$ to achieve a perfect fit for $E_{GW} = 1$.

Because no continuous time series of chloride tracer flux from the trenches could be calculated, instead the estimated maximum possible masses (see chapter 2.2.2 Tracer response) were used to assess the efficiency of chloride transport (E_{Cl}). Therefore a trapezoidal function (Seibert and McDonnell, 2002) was defined; E_{Cl} = 1 if the simulated chloride recovery at the trench is smaller than 50% of the maximum possible mass ("fully acceptable"); E_{Cl} = 0 if the simulated recovery exceeds the maximum possible mass ("not acceptable"); linear interpolation of E_{Cl} between both extremes.

The combined model efficiency E_{Total} was calculated as the average of E_{SSF}, E_{GW} and E_{Cl}. The best possible fit is achieved for E_{Total} = 1.

3.2 Results

3.2.1 Model evaluation

More than 3000 parameter combinations were evaluated for each of the three hillslopes. The best 100 parameter sets for E_{Total} are analyzed in further detail below to assess the model fit, identifiability of the parameters, parameter differences among the hillslopes and the hillslope water balances.

Interestingly, for all evaluated random parameter sets and all hillslopes the recovered chloride mass in the trenches and over the total outflow boundary was zero, despite wide parameter ranges. Hence, the efficiency for tracer transport resulted always in a perfect fit and did not contribute to the identification of feasible parameter combinations. E_{Cl} was therefore not used in the calculation of E_{Total}, shown in Table 3.2.

Table 3.2: Model performance ranges of the best 100 parameter sets.

Hillslope	Model efficiency		
	E_{SSF}	E_{GW}	E_{Total}
Grassland	0.16 - 0.49	0.71 - 0.88	0.49 - 0.63
Coniferous forest	0.42 - 0.76	0.55 - 0.90	0.65 - 0.79
Mixed forest	0.02 - 0.19	0.56 - 0.70	0.33 - 0.39

The best overall model performance was found for the coniferous forest hillslope, the worst for the mixed forest hillslope. Time series of observed and simulated trench flow dynamics are depicted in Figure 3.1. Trench flow dynamics are reproduced approximately well at the coniferous forest hillslope with an E_{SSF} range of 0.42 – 0.76. However, for both the grassland and mixed forest hillslopes trench flow efficiency is smaller than 0.50 for all parameter sets, indicating a weak agreement of modeled and observed trench flow rates.

Observed and simulated time series of the spatial water table mean and standard deviation are shown in Figure 3.2.

For all hillslopes the mean water table position is modeled too low. At the grassland hillslope the simulated water tables respond strongly to every rainfall event and mostly decline close to the basal layer between events. This is contrary to the observed nearly constant water table depths. For the coniferous forest the model performed best also in terms of the water table efficiency. During most of the time and for all parameter sets the mean water table position is simulated ~ 0.4 m lower than observed. Alternatively, it is simulated too high during the wet period in June. Also at the mixed forest hillslope the water table is always simulated far below the measured height. The time series of observed and simulated spatial standard deviation of the well water tables agree better for all hillslopes. However, this can be expected, since the water table depth in a grid cell is constrained by the soil depth which is equal to the measured well depth for the simulations. It seems that HillVi cannot reproduce the spatial pattern of observed water tables. Yet, the simulated trench flows approximately mimic the observed flows and are in the same order of magnitude. Obviously, there is a mismatch in the internal subsurface flow dynamics of the model and the real hillslopes. Hence the model cannot replicate the observed hydrometric dynamics adequately.

The distributions of calibrated parameter values for the three hillslopes are shown in Figure 3.3 as boxplots. The parameter values are normalized to a range between 0 and 1 for better comparability between different parameters. Generally, wide interquartile ranges and long whiskers indicate poorly constrained parameters and may indicate model structural inadequacy or overparameterization. For many parameters the values of the best 100 sets still spread relatively symmetrically over the complete initial range (porosity, drainable porosity, unsaturated recharge exponent, crop factor). Exceptions are the parameters for the mean saturated hydraulic conductivity at the soil surface (k_{sat}) and the shape factor of the depth function. At the mixed forest hillslope k_{sat} is lowest with the weakest reduction with depth; the highest k_{sat} with a rapid reduction over depth was calibrated for the coniferous forest hillslope. It seems that the range of the optimized k_{sat} parameter is strongly constrained by the trench flow rate and cumulative volume (high k_{sat} = high flow). However, this implies a dominant control of surface and subsurface microtopography around the trench for this parameter. The only parameter that is obviously linked to vegetation is the crop factor, which is not identifiable for all hillslopes.

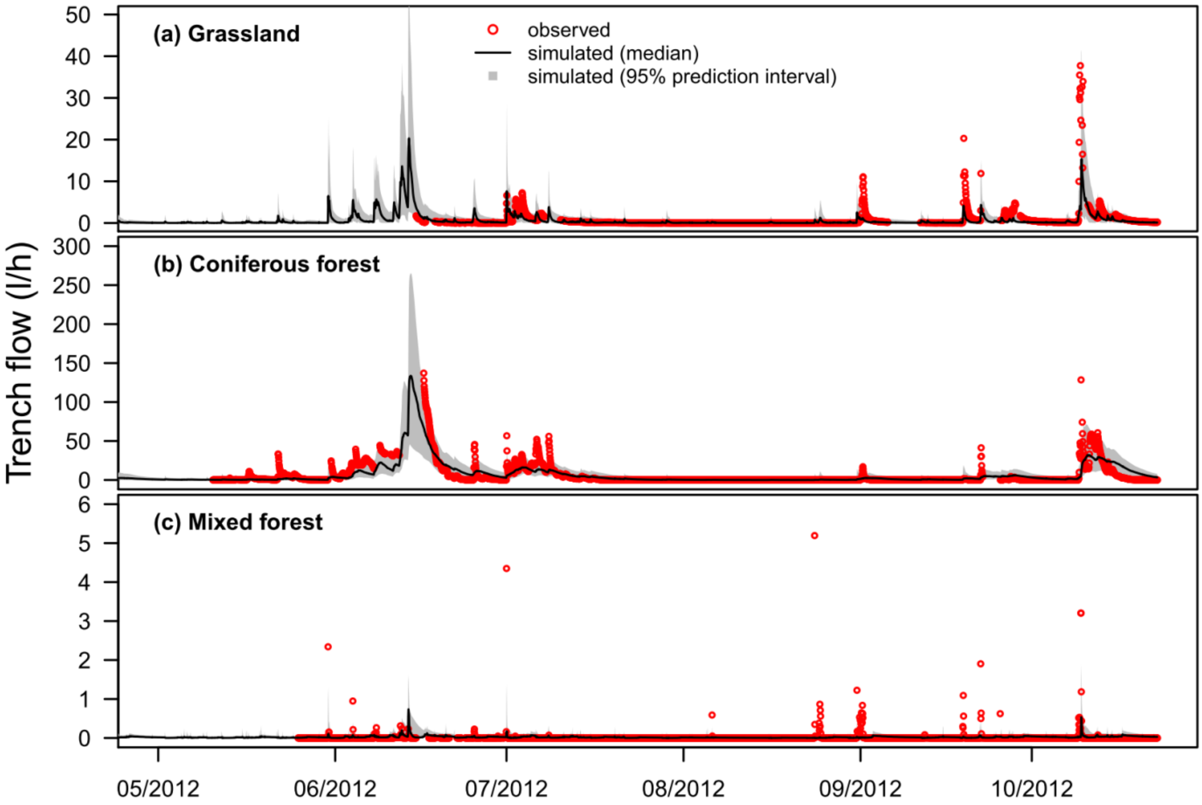

Figure 3.1: Model results trench flow (best 100 parameter sets).

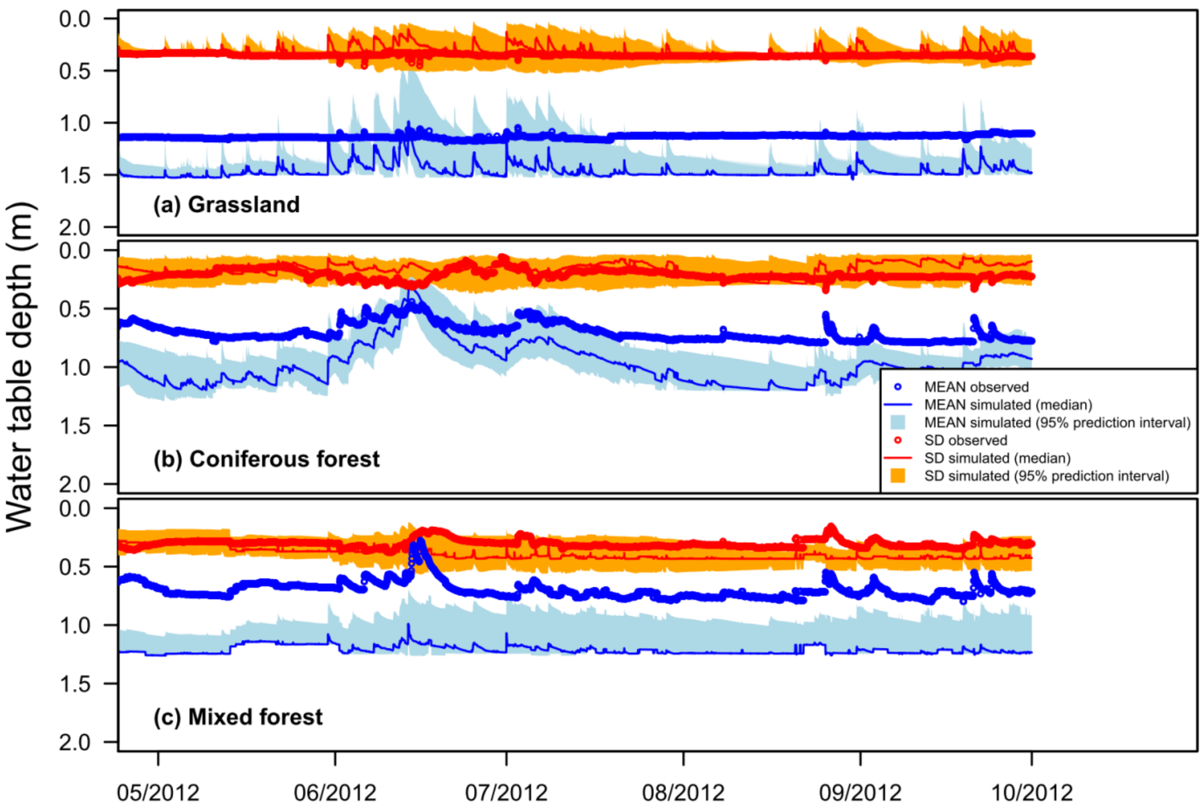

Figure 3.2: Model results water table depth (best 100 parameter sets).

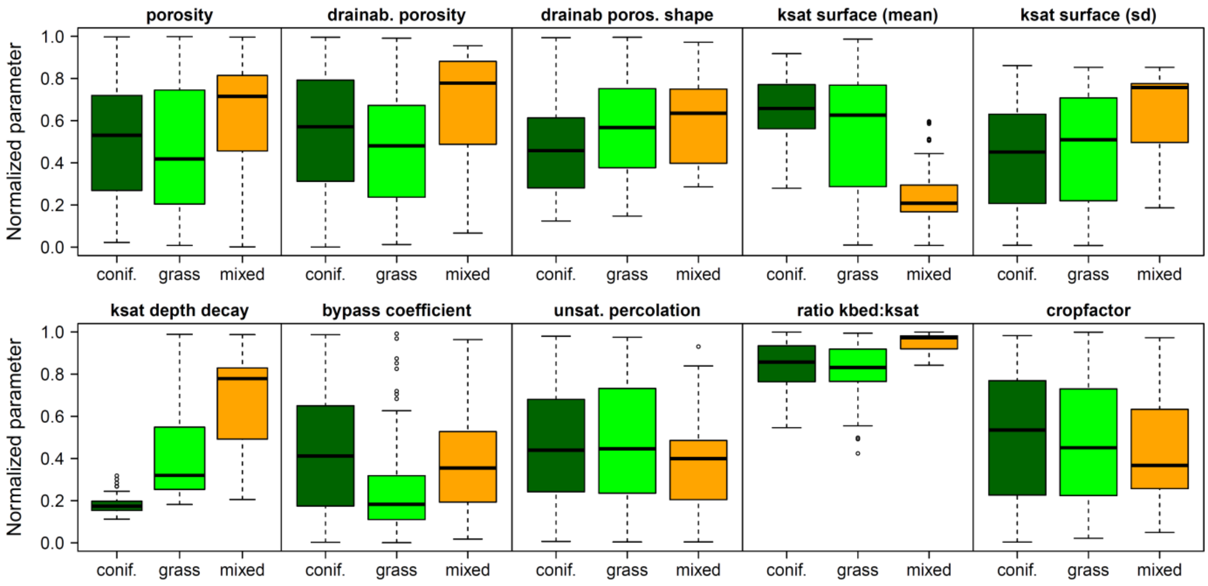

Figure 3.3: Boxplots of normalized parameter ranges (best 100 parameter sets).

3.2.2 *Hillslope water balances*

Achieving a closed water balance at the hillslope scale is very difficult (Graham et al., 2010a). In a model environment all flux and storage terms can be quantified. Yet, the simulated fluxes are closely linked to the assumptions about hydrological behavior inherent to the choice of a model structure and its parametrerization. Due to the low identifiability of most HillVi parameters at the three hillslopes, parametrical uncertainty is high. Consequently, also the calculated hillslope water balances for each hillslope are widely scattered for the selected best 100 parameter sets. Despite this and the unsatisfactory representation of the internal water table dynamics, the median water balance fluxes and storage changes for the total HillVi model domains during the simulated period (May – October 2012) are illustrated in Figure 3.4. Note that the balances are not closed due the choice of the median.

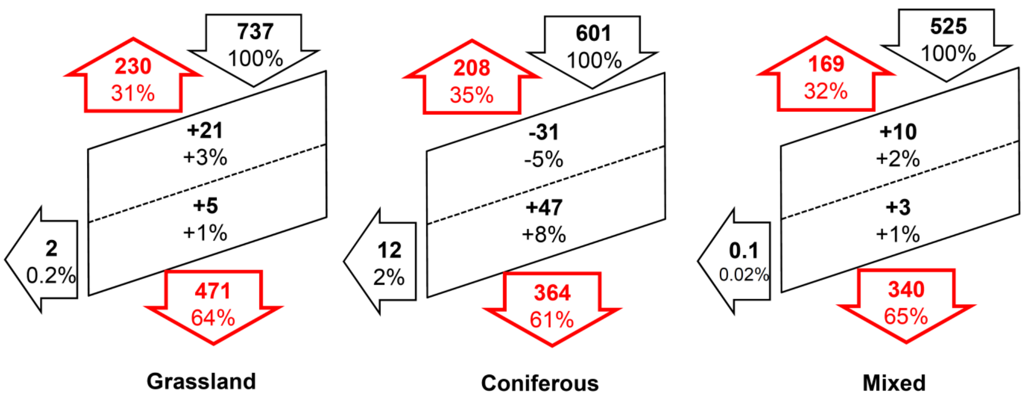

Figure 3.4: Hillslope water balances (median of 100 best parameter sets), flows and storages are given in mm and fraction of precipitation. Major fluxes are colored in red.

Despite uncertainties, the general partitioning of incoming precipitation (evaporative losses from canopy interception are already subtracted from the input) is very similar among the three hillslopes. Approximately one-third of the hydrological input left the hillslope via evapotranspiration, and nearly two-thirds as deep percolation. This suggests that subsurface flow of the upper 2 m (trench depths) seems to be of minor importance for the long-term water balance. However it was found to be an essential contribution during storms also at the catchment scale (Bachmair and Weiler, in review). Due to the model setup, the deep percolation losses may still participate as lateral subsurface flow and not necessarily infiltrate vertically at the real hillslopes. Interestingly, a prominent effect of vegetation cover is not obvious from these results.

3.2.3 Tracer balances

As mentioned before, not a single parameter combination for any of the three hillslopes resulted in any chloride recovery at the trenches. In contrast, at the grassland and coniferous forest hillslopes tracer breakthroughs were observed during the first weeks after the application, although they cannot exactly be quantified. Figure 3.5 shows the simulated travel distances from the injections wells towards the trench cells. It is obvious that the chloride transport distance is vastly underestimated, even with some inevitable numerical dispersion. Hence, in addition to water table dynamics also the simulated flow pathways and velocity distributions of subsurface flow seem to be not representative for the observations at the field site.

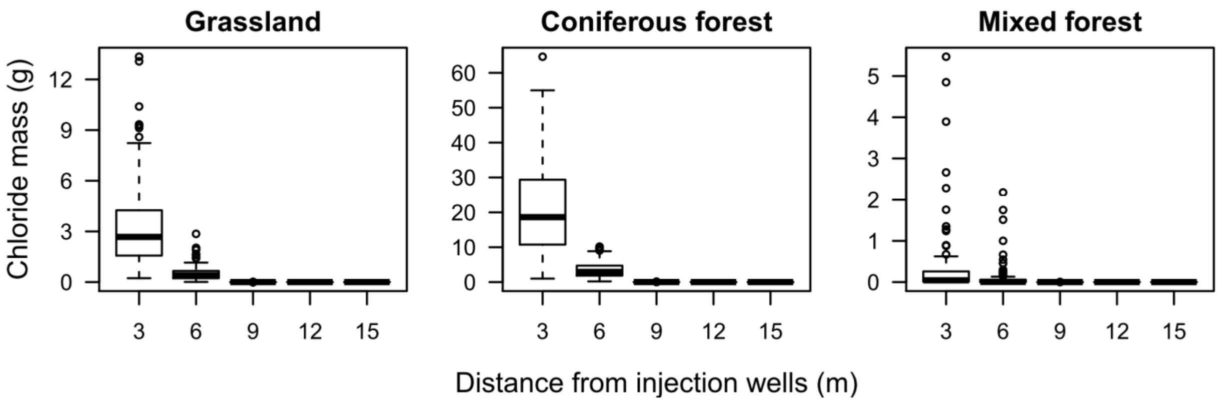

Figure 3.5: Chloride masses (unsaturated and saturated zone) at the end of the simulation period (100 best parameter sets).

From Figure 3.6 it can be seen that also large amounts of the tracer mass were retained in the unsaturated zone or transported with the deep percolation losses. The large stored tracer mass in the saturated zone of the coniferous forest hillslope in combination with the underestimated travel distance again straightens out the underestimated flow velocities. However, higher hydraulic conductivities (or a less pronounced decay with depth) would imply even lower water table heights in order to match the observed trench flow rates.

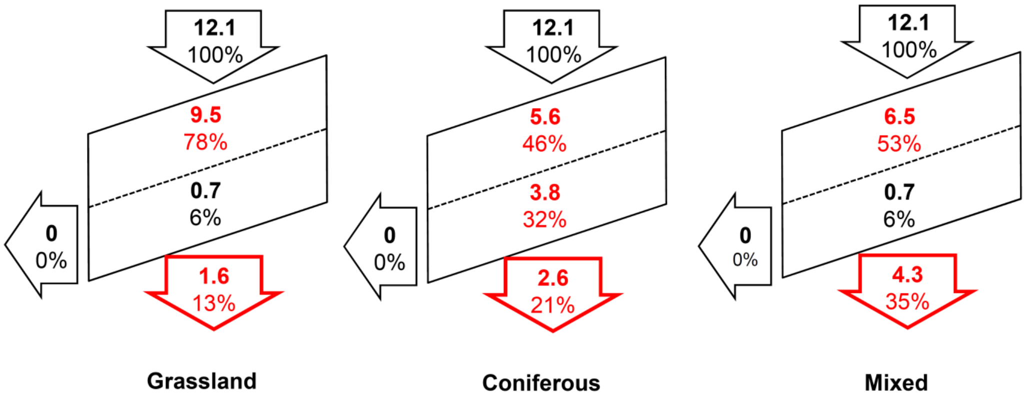

Figure 3.6: Chloride mass (kg) balances (median of 100 best parameter sets). Major fluxes and storages are colored in red.

4 Discussion

4.1 Insights gained from the tracer experiments

At the study site high spatial and temporal variability of subsurface flow processes with a marked difference between the grassland and the forested hillslopes has been observed (Bachmair, 2012). Several different flow pathways were suggested as explanations for the interactions between responses of water tables, trench dynamics and catchment runoff. In this respect some additional information could be picked with the aid of the tracer experiments.

At the grassland and coniferous forest hillslopes tracer breakthroughs were observed in the trench flow and near trench wells indicating some active flow path between the locations of injection and monitoring. Even without a well water table present during tracer application at the grassland hillslope an increase of well EC was observed 30 m down slope after two days. Some preferential flow pathways in the unsaturated zone were fed with the tracer solution, which was obvious from a very limited water table rise during the injection. However, it is unlikely that the applied water volume was able to activate a preferential flow network over such distance. Hence, most of the tracer solution was retained in the unsaturated zone around the wells or percolated deeper into the soil. There it was probably transported laterally via the basal layer towards the trench. The same can be assumed for the coniferous forest. However, due to a present water table across the hillslope greater mixing and thus dampening of the tracer signal occurred there. The presence of a flow path that is located deeper than the wells and trenches is also suggested from the chloride breakthrough in the creek. Additionally, during periods where no trench flow was observed for the coniferous forest and grassland a continuous tracer export from the hillslopes can be found.

Hydrochemical analyses of sampled trench flow revealed high spatial and temporal variability of the ionic water composition which can be a characteristic of varying flow pathways. The higher sulfate concentrations in the coniferous forest trench flow and creek samples may be interpreted as higher contributions from deeper soil layers (Hangen et al., 2001). What causes the high nitrate concentrations is not clear. It could indicate near surface flow contributions, such as preferential flow along channels of decayed roots.

Again, the role of vegetation for the observed tracer transport dynamics cannot be clearly assessed. Topographic characteristics of the hillslopes seem to be a dominant control. From the mixed forest hillslope no tracer breakthrough could be identified, suggesting higher chloride retention in the unsaturated zone or maybe also percolation into the shallow bedrock. However, the role of local bedrock topography for re-routing a possible tracer transport via subsurface flow is not clear.

4.2 Does the model adequately represent observed hillslope dynamics?

The development of HillVi was guided by the process understanding gained from field studies (Weiler and McDonnell, 2004). The fundamental assumed runoff mechanism is lateral subsurface flow initiated by a transient saturation above a single impeding layer.

However, at all three simulated hillslopes the dynamics of water tables, trench flow and tracer transport could not be reproduced satisfactorily. Although trench flow mimicked the observed behavior, the model structural representation of the flow pathways is unrealistic for the hillslopes. This shows the importance of internal model validation in addition to flow data.

The inadequacy of the model suggests that at the real hillslopes the response of wells and trench flow is partially decoupled. That is, trench flow or generally subsurface flow is there also initiated without a related response in the well water tables. Consequently flow path variability and activation operates on a smaller scale than the well distance. This was also suggested by Bachmair and Weiler (in review).

A way of conceptualizing preferential pathways is the soil pipe routine in HillVi. However, the activation of pipe flow also depends on a rising water table. At the moment no pipe activation is possible via partial saturation of the topsoil due to the simple conceptualization of the unsaturated zone. Possibly the routine could be modified to test the hypothesis of a preferential flow network without a rising water table from the soil-bedrock interface (or basal layer).

The assumption of the basal layer as horizontal boundary seems to be an unsuitable choice, as there may be fluxes into, from and within the layer at the natural hillslopes. The depth to bedrock could be simulated as a random field with a modified depth function of the saturated hydraulic conductivity. Within the basal layer water flow can be faster than in the soil above and the weathered bedrock below.

Generally, the model structure is rejected for the studied hillslopes and a vegetation cover effect on the model parameters was not found. The only identifiable parameters were closely linked to the trench flow dynamics, which however are mainly a result of local surface and subsurface microtopography.

4.3 A revised view on the runoff generation at the studied hillslopes

Following the insights gained from hydrometric observations, the tracer application and model rejection a revised perceptual model of runoff processes at the study site is proposed (Figure 4.1). Depending on the wetness state and rainfall intensity different processes of flow pathway activation and partitioning can be distinguished:

Baseflow: No significant lateral flow above the basal layer, recharge into the weathered bedrock, baseflow is sustained from weathered and fractured bedrock.

Dry / low: Infiltration into soil, storage in soil matrix, partially percolation to the basal layer and little lateral flow.

Wet / low: Infiltration into the soil, subsurface initiation of bypass flow, increasing lateral flow at basal layer, rise of footslope water table due to flow accumulation, increasing bedrock recharge and baseflow (small double peak).

Dry / high: Infiltration into soil, surface and subsurface initiation of bypass flow, local saturation of topsoil and connecting via preferential flow pathways, lateral flow from topsoil and basal layer.

Wet / high: Surface initiation of preferential flow, water table upward saturation into the topsoil and merging with local topsoil saturation, high lateral flow, increasing bedrock recharge and pronounced double peak.

Furthermore, a high spatial variability of soil properties affects the initiation of transient topsoil saturation. Water percolates down until the percolation rate exceeds the vertical hydraulic conductivity (percolation excess) and saturation starts. This may be due to changes in the texture, bulk density, macroporosity or other factors. Saturated patches interconnect via coarse texture heterogeneities, soil pipes, root channels, mouseholes etc.

This perceptual model is slightly modified for the presence of forest. In that case the roots enhance vertical fluxes and the probability of percolation excess decreases. Upward saturation gets more important, causing a higher representativity of water table positions for subsurface flow rate. However, the exact effect depends on the tree species and root architecture (Jost et al., 2012; Nordmann et al., 2009).

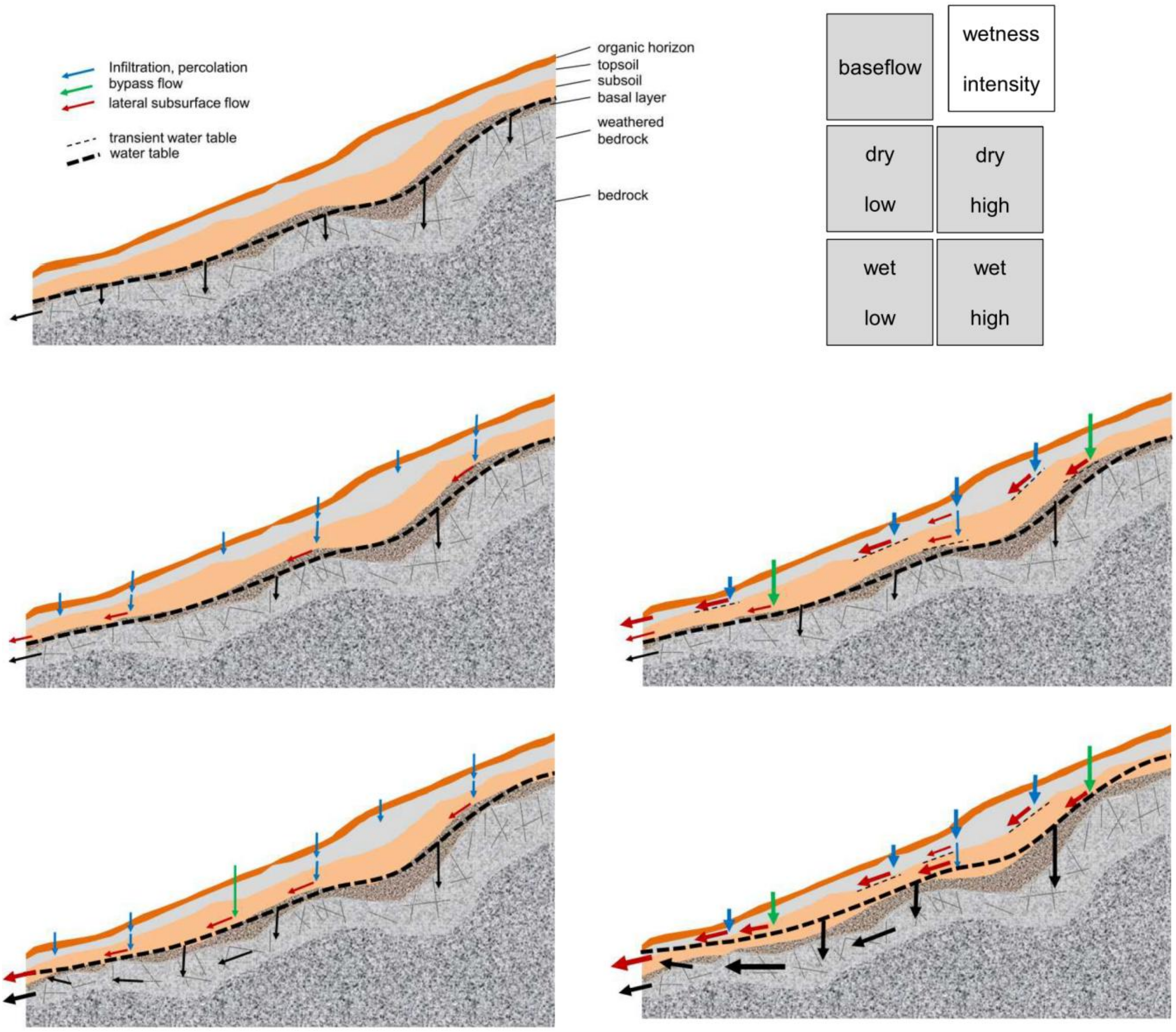

Figure 4.1: Perceptual model of runoff generation at the study site for different settings of antecedent wetness and precipitation intensity.

5 Conclusions and outlook

The combined field- and model-based hillslope study was used to test a hypothesized dominant control of vegetation cover on hydrological dynamics in general, in terms of tracer transport and for model parameterization. All three hypotheses must be rejected based on the results. Local surface and subsurface microtopography was identified as important control of subsurface flow and transport dynamics. The tracer experiment could prove hillslope to creek connectivity via deep flow pathways. Lateral subsurface flows within or below the basal layer may be important flux terms for the hillslope and catchment water balance, but are not captured with the current field monitoring setup. Model calibration with combined trench flow, water table and tracer transport data did not reduce parameter uncertainty but proved the model structure inadequate in terms of flow path representation. Upward saturation over a single impeding layer was thus rejected as dominant runoff process at the study site. However, the current monitoring design of subsurface flow dynamics was partly based on this assumption.

The revised perceptual model of runoff generation can be seen as a testable new hypothesis about hillslope hydrological dynamics. It should be tested in the near future both with numerical and field tests. The HillVi pipe flow routine could be adapted in order to conceptualize a connect-and-react mechanism induced by infiltration and partial soil saturation instead of a rising water table. The adapted model structure can then be tested again with the observed trench flow and water table data. The hypothesized shallow flow paths that contribute also to trench flow should be explored in the field. Sprinkling experiments and dye staining for different antecedent conditions and intensities would give insights into vertical water fluxes and zones of lateral subsurface flow initiation at the different hillslopes. Also long-term and event hydrochemical sampling seems to be a potential tool for characterizing time-variant flow pathways. New findings then can guide an efficient extension of the instrumentation to monitor subsurface flow dynamics at the hillslope scale.

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