Master Thesis

Hydrological simulations of a hillslope prone to shallow landslides

Author(s):
Gilgen, Marc

Publication Date:
2008

Permanent Link:
https://doi.org/10.3929/ethz-a-005676916

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DIPLOMA THESIS

Marc GILGEN

Hydrological Simulations of a Hillslope Prone to Shallow Landslides

DEPARTMENT OF ENVIRONMENTAL SCIENCES
ETH ZURICH

authored at the

SWISS FEDERAL INSTITUTE FOR FOREST, SNOW AND LANDSCAPE RESEARCH WSL

Supervisors:
Prof. Dr. Sarah SPRINGMAN
Dr. Manfred STÄHLI
Dr. Peter KIENZLER

January 2008
Contents

Vorwort iii
Summary v
Zusammenfassung vii
List of Figures x
List of Tables xi
List of symbols and abbreviations xiii

1 Introduction 1
  1.1 Context ................................................................. 1
  1.2 Goal and Objectives .................................................. 2
  1.3 Outline ................................................................. 3

2 Theory 5
  2.1 Hillslope Hydrology .................................................... 5
    2.1.1 Introduction ....................................................... 5
    2.1.2 Flow processes at the hillslope scale .......................... 6
  2.2 Pedology ............................................................... 10
    2.2.1 Soil configuration ............................................... 10
    2.2.2 Soil water ......................................................... 10
  2.3 Shallow landslides and slope stability ............................ 14

3 Field Site 17
  3.1 Tössegg, Canton Zürich ............................................. 17
    3.1.1 Climate and Hillslope Description ............................ 17
    3.1.2 Field Data and Experiments ................................... 19
## Contents

4 Methods .................................................. 25
   4.1 HillVi ................................................. 25
      4.1.1 Introduction .................................... 25
      4.1.2 Lateral flow .................................... 27
      4.1.3 Pipe flow ....................................... 29
      4.1.4 Unsaturated zone and recharge ................. 29
      4.1.5 Vertical bypass flow ............................ 30
      4.1.6 Overland flow ................................... 31
      4.1.7 Water balance and storage ..................... 31
      4.1.8 Bedrock seepage ................................ 32
      4.1.9 Other routines ................................... 32
      4.1.10 Model output ................................... 32
   4.2 Parameter sampling ................................... 33
      4.2.1 Monte Carlo and Latin hypercube sampling .... 33
   4.3 Sensitivity analysis and parameters ................. 35
      4.3.1 Parameters .................................... 37

5 Results .................................................. 41
   5.1 Event analysis ....................................... 41
      5.1.1 July 2005 ....................................... 41
      5.1.2 April 2006 .................................... 42
   5.2 Sensitivity Analysis .................................. 45
      5.2.1 Parameter discussion ......................... 46
   5.3 Comparing simulation and field data ................. 52
      5.3.1 Influence of soil porosity on model processes 54
   5.4 Spatial patterns of maximum saturation ............ 57
   5.5 Influence of pipe flow on saturation patterns .... 61

6 Discussion .............................................. 67
   6.1 Discussion of objectives ............................ 67
   6.2 Synthesis and recommendations ..................... 70
   6.3 Final remarks and personal statement ............... 71

Bibliography .............................................. 73

A Graphs and Tables .................................... 1
Meinem Betreuer, Dr. Manfred Stähli, Leiter der Forschungseinheit Gebirgshydrologie und Wildbäche der WSL, danke ich für die Möglichkeit, eine sehr interessante Diplomarbeit unter seiner Betreuung zu schreiben. Die Involvierung der Diplomarbeit in das Gross-Projekt TRAMM hat mir zahlreiche Einblicke in ein sehr aktuelles, spannendes und interdisziplinäres Forschungsprojekt gegeben. Manfred Stähli stand mir zu jeder Zeit für die Beantwortung von Fragen zur Verfügung, gab mir etliche wichtige fachliche Anstösse und verstand es, mich mit innovativen und weitsichtigen Forschungs-Ansätzen vertraut zu machen.

Meiner Diplomprofessorin Dr. Sarah Springman vom Geotechnischen Institut der ETH Zürich möchte ich für die Möglichkeit danken, in Zusammenarbeit mit ihrem Institut diese Arbeit zu verfassen. Sie hat mich mit ihrem Wissen in der Geotechnik und Hangstabilitätsforschung in zahlreichen ergiebigen Diskussionen unterstützt, damit ich bei der hydrologischen Modellierung des rutschgefährdeten Hanges stets auch die mechanischen Aspekte im Auge behalten konnte.

Bei Dr. Peter Kienzler vom Geotechnischen Institut der ETH Zürich möchte ich mich für die informativen Diskussionen, seinen kritischen Umgang mit der Thematik und die Möglichkeit, den Infiltrationsexperimenten in Tössegg/ZH beizuwohnen bedanken.

Dr. Peter Lehmann, wissenschaftlicher Mitarbeiter in der Forschungseinheit Gebirgshydrologie und Wildbäche der WSL, hat mir mit seinem breiten Wissen und seiner grossen Erfahrung über bodenphysikalische und hydrologische Prozesse immer wieder interessante Anregungen und Stossrichtungen für die Diplomarbeit gegeben.

Professor Markus Weiler von der University of British Columbia, Canada, danke ich für das Überlassen des Quellencodes und für die Einführung in das von ihm programmierte hydrologische Modell HillVi und für das Beantworten meiner Fragen betreffend der Modellierung.
Bei folgenden Personen möchte ich mich für ihren Beitrag zu dieser Diplomarbeit bedanken:

Dr. Massimiliano Zappa von der Forschungseinheit Gebirgshydrologie und Wildbäche für die Beantwortung zahlreicher hydrologischer Fachfragen.

Dr. Christian Hoffmann von der Forschungseinheit Wald-Ökosystemprozesse der WSL für die Unterstützung in statistischen Fragen.

Jürg Schellendorfer vom Seminar für Statistik der ETH Zürich für seine Informationen bezüglich Regressionsmodellen.

Dr. Andrea Thielen vom Geotechnischen Institut der ETH Zürich für das zur Verfügung stellen und das Beantworten meiner Fragen betreffend den Felddaten.

Allen Mitarbeitern der Forschungseinheit Gebirgshydrologie und Wildbäche für die sehr angenehme und kameradschaftliche Zeit an der WSL.

Für das Korrekturlesen der Arbeit danke ich ganz herzlich Annemarie Schneider, Kelly Hess, Peter Kienzler und Benjamin Loretz.

Summary

During the major storm events of the last two decades hundreds of shallow landslides occurred all over Switzerland. Although the general factors promoting the occurrence of shallow landslides are known, there is a lack of knowledge about the single triggering event that finally results in slope failure. Shallow landslides seem to be quasi random in their nature, occurring in a wide range of slope inclinations, precipitation intensities and soil characteristics, making their prediction difficult. To improve the assessment of risks arising from such natural hazards and provide sustainable protection for society, it is necessary to understand processes leading to such natural disasters. The Competence Center Environment and Sustainability (CCES) of the Swiss Federal Institute of Technology (ETH) runs a major four year research project (2006-2010) entitled "Triggering of Rapid Mass Movements" (TRAMM) that investigates rapid mass movements with a main focus on the triggering and initiation mechanisms, propagation from slow to fast and on flow characteristics. The approach of the CCES-TRAMM project is truly interdisciplinary, as research fields ranging from soil physics, soil mechanics, fluid dynamics, geotechnics, hydrology and others meet on a platform for cooperation research. Within this framework, the diploma thesis at hand aimed to improve and refine the hydrological modelling of water flow and distribution of a hillslope prone to shallow landslide in Tösseg, canton Zurich, Switzerland.

The simulations were performed with the spatially distributed, physically based conceptual hydrological model HillVi which particularly considers preferential flow through soil channels.

In a first step, a regional sensitivity analysis has been conducted to study the influence of the parametrization on model outputs. To reduce the number of model runs necessary to perform a sensitivity analysis, the "Latin Hypercube Sampling Method" has been implemented into HillVi. This method allows an entire sampling of the parameter space with a drastically reduced number of model runs compared to standard random sampling methods (e.g. McKay et al. (1979)). Eleven model parameters have been assessed in the sensitivity analysis by means of two storm events in 2005 and 2006 and were ranked according to their influence on saturation and runoff.

Measured soil moisture content data from April 2 to 15 2006 served as a basis to compare the simulated
to the local field data. The modelled results show a high "goodness-of-fit" for this particular event regarding the temporal change of water content.

Combinations of two total and five drainable soil porosities were investigated with the base-calibrated model. It is shown, that the combination of the lower total porosity with a higher drainable porosity produce the highest local water levels.

The model runs of the sensitivity analysis were then analysed regarding their spatial distribution of maximum water levels. As a result, three slope areas showing frequent high water table depths were identified. Because these three locations emerged from the model runs for the sensitivity analysis with randomly sampled parameters, this shows a certain parameter independence of the occurrence of such areas and further underlines the influence of bedrock and surface topography to cause the spatial distribution of such saturation patterns.

In a final step, the influence of preferential flow on the spatial distribution of areas with high water levels was assessed. The existence of pipes showed two main effects: (1) local maximum water tables are efficiently drained by preferential flow leading to their dissipation, (2) soil pipes ending in a dead-end passageway increase local water availability. The latter effect is assumed to be of major importance regarding triggering events of shallow landslides because the local increase in water levels may increase pore pressure in the surrounding matrix which may then lead to the final triggering event and slope failure as shown in other research studies. However, these particular locations were not identifiable because of the random implementation algorithm for the preferential flow network in the model.

The thesis concludes by mentioning strengths and weaknesses and constraints of the modelling approach described here and makes suggestions for a continued model development for hillslope modelling with HillVi and recommends a project procedure for further studies assessing the triggering problematic of shallow landslides.

**Keywords**: hillslope hydrology, natural hazards, shallow landslides, pipe flow, virtual experiments, HillVi
Zusammenfassung


Im Rahmen dieser Diplomarbeit wurden Wasserfluss und -verteilung in einem rutschgefährdeten Hang in Tössegg, Kanton Zürich (CH), mit einem hydrologischen Modell simuliert. Die Simulationen sind mit dem räumlich verteilten, physikalisch basierten, konzeptuellen hydrologischen Modell HillVi durchgeführt worden, welches im Speziellen präferentielles Fliessen im Boden simulieren kann.


\[1\text{ englisch: Triggering of Rapid Mass Movements / TRAMM}\]
und Sättigung untersucht worden.

Gemessene Daten zur Bodenfeuchtigkeit vom April 2005 wurden als Basis herangezogen, um Simulationsdaten und Feldmessungen zu vergleichen. Das Modell zeigte eine hohe Güte in der Simulation der Änderung des Wassergehaltes.

Untersuchungen von Modellläufen mit verschiedenen Werten, für die totale sowie drainierbare Porosität, zeigten, dass die Kombination einer kleinen totalen mit einer hohen drainierbaren Porosität zu den höchsten Wasserspiegeln im Hang führt.


Die Arbeit schliesst mit dem Aufzeigen der Vorteilen, aber auch der Grenzen und Schwächen der in dieser Arbeit angewandten Modellierung, zeigt Möglichkeiten für die weitere Entwicklung von HillVi auf und schlägt einen Ablauf für eine künftige Modellierung von rutschgefährdeten Hängen vor.
## List of Figures

1.1 Pictures of two shallow landslides ........................................ 2

2.1 Spatio-temporal classification of different branches in hydrology .......................... 5

2.2 Flow routes followed by subsurface runoff on hillslopes .................................. 6

2.3 Schematic diagram describing the processes which explain effects of pipeflow on landslide initiation ......................................................... 9

2.4 Schematic illustration of stresses acting on a block and a soil element .................. 15

3.1 Location Tössegg, Kanton Zürich ............................................. 17

3.2 Climate diagram, Station: Zürich-Kloten ...................................... 17

3.3 Tössegg slope with locations of instrumentation ....................................... 18

3.4 DEM of the Tössegg hillslope .................................................... 18

3.5 Soil profile of left and right boarding transects ....................................... 20

3.6 Sprinkling installation (a) and excavated profiles (b) and (c) ......................... 21

3.7 Evidence of preferential flow structures in an expanded view ......................... 22

4.1 Model concept of HillVi ............................................................. 26

4.2 Grid cell by grid cell approach .................................................... 27

4.3 Graphical output window .......................................................... 33

5.1 MP, TDR and tensiometer curves for all instrumented soil depth profiles from 17 to 29 July, 2005 ................................................................. 42

5.2 TDR and tensiometer curves for all instrumented soil depth profiles from 2 to 14 April, 2006 ................................................................. 43

5.3 Sensitivity analysis model outputs (top) and corresponding SRC’s (bottom) July 2005 event ................................................................. 50

5.4 Sensitivity analysis model outputs (top) and corresponding SRC’s (bottom) April 2006 event ................................................................. 51

5.5 Model validation period April 2.-15. 2005 ...................................... 53

5.6 Influence of soil porosities on model outputs ......................................... 55
<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.7</td>
<td>Percentage of model runs whose maximum water level at the soil-bedrock interface exceeded the datum given above the Figure (in cm), July 2005</td>
<td>57</td>
</tr>
<tr>
<td>5.8</td>
<td>Spatial patterns of percentage of model runs whose maximum water level at the soil-bedrock interface exceeded the datum given above the Figure (in cm), April 2006</td>
<td>59</td>
</tr>
<tr>
<td>5.9</td>
<td>Influence of pipe flow on the spatial distribution of maximum water levels (cm) along the hillslope for the April 2006 event</td>
<td>62</td>
</tr>
<tr>
<td>5.10</td>
<td>Cumulative frequency distribution of water levels above bedrock (%) for different pipe densities</td>
<td>63</td>
</tr>
<tr>
<td>5.11</td>
<td>Graphical HillVi output for the July 2005 event with pipes</td>
<td>64</td>
</tr>
<tr>
<td>5.12</td>
<td>Graphical HillVi output for the July 2005 event without pipes</td>
<td>65</td>
</tr>
<tr>
<td>A.1</td>
<td>3D model of the Tössegg hillslope, own illustration (IDL)</td>
<td>II</td>
</tr>
<tr>
<td>A.2</td>
<td>Layout of instrumentation</td>
<td>II</td>
</tr>
<tr>
<td>A.3</td>
<td>Time series data of all Moisture Point instruments</td>
<td>III</td>
</tr>
<tr>
<td>A.4</td>
<td>Times series data of all TDR instruments</td>
<td>IV</td>
</tr>
<tr>
<td>A.5</td>
<td>Time series data of all Tensiometer instruments</td>
<td>V</td>
</tr>
<tr>
<td>A.6</td>
<td>Decline of $ko$ and $no$ with depth for the values shown in the legend</td>
<td>VI</td>
</tr>
<tr>
<td>A.7</td>
<td>Meta model statistics for sensitivity analysis</td>
<td>VII</td>
</tr>
</tbody>
</table>
List of Tables

3.1 Soil characteristics. Source: Thielen (2007), modified ........................................ 19
3.2 Instrumentation ........................................................................................................... 20

4.1 Information contents of graphical output windows .................................................... 33
4.2 Parametrization for the sensitivity analysis ............................................................... 39

5.1 Rainfall characteristics for the selected periods ....................................................... 44
5.2 Sensitivity analysis parameter ranking based on absolute values (mean SRC over time), 17 to 29 July, 2005 ................................................................. 48
5.3 Sensitivity analysis parameter ranking based on absolute values (mean SRC over time), 2 to 14 April, 2006 ........................................................... 48
5.4 Calibrated parameter set ......................................................................................... 54
5.5 Summary Table of flow amount, pipe-matrix flow ratio and mean water table elevation for 10 different pipe densities ......................................................... 64

List of symbols and abbreviations

ΔS  Storage change
εs  dielectric constant of soil [-]
εw  dielectric constant of water [-]
βs  regression coefficients
es  regression residuals
ψ  (total) water potential [J kg⁻¹]
ψg  gas potential
ψH  hydraulic potential [J kg⁻¹]
ψm  matrix potential [J kg⁻¹]
ψo  osmotic potential
ψz  gravity potential [J kg⁻¹]
σ  normal stress
τf  shear strength
θw  volumetric water content
θr  residual water content
θs  saturation water content
c  speed of light [ms⁻¹]
D  soil depth
e  regression residual
g  gravity constant
H  piezometric head
J  slope gradient
K₀  saturated hydraulic conductivity at soil surface
kbed  Saturated hydraulic conductivity of bedrock
k_sat  saturated hydraulic conductivity
m  mass
nd  drainable porosity
List of symbols and abbreviations

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
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</thead>
<tbody>
<tr>
<td>$n_{tot}$</td>
<td>total porosity</td>
</tr>
<tr>
<td>$p$</td>
<td>precipitation</td>
</tr>
<tr>
<td>$q$</td>
<td>flow [$LT^{-1}$]</td>
</tr>
<tr>
<td>$q_p$</td>
<td>pipe flow [$LT^{-1}$]</td>
</tr>
<tr>
<td>$R$</td>
<td>hydraulic radius</td>
</tr>
<tr>
<td>$T$</td>
<td>transmissivity</td>
</tr>
<tr>
<td>$V$</td>
<td>total volume (of soil)</td>
</tr>
<tr>
<td>$V_p$</td>
<td>pore volume</td>
</tr>
<tr>
<td>$V_w$</td>
<td>water volume</td>
</tr>
</tbody>
</table>

CCES Competence Center Environment and Sustainability

cmwc cm water column

DEM Digital Elevation Model

DF Degrees of freedom

E Evapotranspiration

ERT Electrical resistivity tomography

HOF Hortonian Overland Flow

IGT Institute for Geotechnical Engineering, ETH Zurich

IQ(R) Inter quartile (range)

LHS Latin hypercube sampling

ll lower left plot

lr lower right plot

MC Monte Carlo

MCLHS Monte Carlo Latin hypercube sampling

MP Moisture Point

OLS Ordinary least squares

ORC ordinary regression coefficient

P Precipitation

$pF$ pF-Value [-], $pF$=log-$h_m$ if $h_m$ in [hPa]

Q Runoff

SA sensitivity analysis

SOF Saturation Overland Flow

SRC standardized regression coefficient

SSF Subsurface Flow

TDR Time Domain Reflectometry

TRAMM Triggering of Rapid Mass Movements

u pore water pressure

ul upper left plot
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Abbreviation</th>
</tr>
</thead>
<tbody>
<tr>
<td>ur</td>
<td>upper right plot</td>
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<tr>
<td>WRC</td>
<td>water retention curve</td>
</tr>
</tbody>
</table>
Chapter 1

Introduction

1.1 Context

During the major storm events of the last two decades hundreds of shallow landslides occurred all over Switzerland. A main problem of shallow landslides is their quasi random nature of occurrence in a wide range of slope inclinations, precipitation intensities and soil characteristics, making their prediction difficult. Decades of research on triggering mechanisms have so far not revealed a common understanding of this type of mass movement. Affected areas often exhibit dramatic spatial and temporal variations of vegetation, hydrological, lithological and hydro geological conditions. Shallow landslides often appear regionally clustered during extreme precipitation events. Two events in Switzerland are representatively mentioned here: a series of 42 shallow landslides occurred in May 2002 close to Rüdlingen (canton Schaffhausen) and the hillslope investigated in this diploma thesis at the river Rhein due to extreme precipitation events. In summer 2002 several hundred landslides occurred in the Napf and Appenzell region (see Rickli and Bucher (2002)).

The Competence Center Environment and Sustainability (CCES) of the Swiss Federal Institute of Technology (ETH) runs a major four year research project (2006-2010) entitled "Triggering of Rapid Mass Movements" (TRAMM) that investigates rapid mass movements with a main focus on the triggering and initiation mechanisms, propagation from slow to fast and on flow characteristics. Within this project, shallow landslides are a main type of natural hazards addressed. Research will focus on the role of heterogeneity and criticality of hydro-mechanics on triggering mechanisms.

A major constraint of the research work done so far was the isolated treatment of the topic by single science disciplines. In contrast to this, the approach of the CCES-TRAMM project is truly interdisciplinary, as research fields ranging from soil physics, soil mechanics, fluid dynamics, geotechnics, hydrology and others meet on platform for cooperational research.

A common perception is that heavy rainstorms or intense water inputs may trigger shallow landslides,
1 Introduction

(a) Landslide near Oberiberg (SZ) occurred during a storm event on 20.06.07. Photo: C. Rickli
(b) Landslide near Rüdlingen / SH in May 2002. Photo: Archive IGT ETH Zurich

Figure 1.1: Pictures of two shallow landslides

particularly by contribution of preferential flow through soil channels (a more precise description is provided in Chapter 2). 90% of 133 shallow landslides analysed by Rickli and Bucher (2002) were affected by preferential flow. In 70% of all cases, channels extended to the shear plane, giving evidence for fast infiltration and accumulation of water in these zones. According to Rickli and Bucher (2002) plant root channels and faunal burrows are among the most influential preferential flow channels in soils. At numerous locations root channels were found, where only the outermost primary cell layer of the root, the rhizodermis was still present, but the inner wooden core was degraded. Such root channels are like tubes in the soil, supplying certain slope locations quickly with high amounts of water. This effect may result in saturation of certain hillslope areas and being therefore a critical factor regarding triggering of shallow landslides.

The Institute for Geotechnical Engineering (IGT) at ETH Zurich investigated in the TRAMM framework the influence of soil saturation on the stability of a hillslope in Tössegg (canton Zürich), Switzerland from mid 2004 to late 2007 (Thielen, 2007). Calculations showed that even during the major rainstorm events of the investigated period slope stability did not reach a critical value. For future research, Thielen (2007) recommended to assess the influence of preferential water flow in the hillslope.

1.2 Goal and Objectives

Goal of this diploma work is the refinement of hydrological modelling at the hillslope scale to provide a more detailed spatial simulation of water flow and storage. Especially the role of preferential flow on hillslope hydrology and the water distribution in the soil should be assessed.

The spatial distributed, physically based conceptual hillslope model HillVi serves as basis for the modelling exercises (Weiler and McDonnell, 2004). HillVi is a simple hydrological model, based on a digital
elevation model (DEM) of bedrock and surface topography. A special feature of HillVi is its ability to generate a random preferential flow network.

Within the described framework the following objectives are addressed in this thesis:

(i) Identification of sensitive model parameters and quantification of their influence on model output

(ii) Modelling of single rainstorm events and comparison to field data

(iii) Investigation of maximum saturation patterns across the hillslope for selected rainstorm events and isolation of frequent saturated areas

(iv) Assessment of the influence of preferential flow on saturation patterns and and water balance of the hillslope

The findings of this thesis should improve knowledge about HillVi by the outcomes from the sensitivity analysis and strengthen understanding of water flow and distribution across the hillslope. Within the TRAMM research project such findings provide valuable information about possible critical triggering locations in hillslopes prone to shallow landslides.

1.3 Outline

Chapter 2 provides the theoretical background on hillslope hydrology, pedology and slope stability. The investigated hillslope in Tössegg is described in Chapter 3 based on a PhD project (Thielen, 2007) and from field experiments carried out during this thesis. The methodical approach is presented in Chapter 4 focusing on the hydrological model HillVi, with which the modelling exercises were conducted and presents the multiple linear regression model applied for the sensitivity analysis. Chapter 5 is geared to the above enumerated objectives and initially interprets soil and precipitation data for two selected rainstorm and one validation event before the sensitivity analysis is discussed. The second objective leads to the model validation through comparison of simulation runs and field data before Sections 5.4 and 5.5 address the spatial distribution of saturated areas in the hillslope and the influence on pipe flow on this process. Chapter 6 discusses the outcomes of each objective addressed, combines them to an overall conclusion and gives an outlook with recommendations for future hydrological modelling exercises.
2.1 Hillslope Hydrology

2.1.1 Introduction

Hillslope Hydrology is a branch of hydrology that studies the spatial and temporal distribution of water and its movement in a hillslope. Horton (1933) was the first to describe a classical model of hillslope hydrology based on a theory referred to as *Hortonian overland flow (HOF)*; a type of flow that is produced when rainfall or snowmelt rates exceed the infiltration capacity of the soil. Kirkby (1988) defined hillslope hydrology as the science of the partition of precipitation as it passes through the vegetation and soil between overland flow and subsurface flow. A hillslope as a hydrological unit is characterized by a spatial extent with magnitudes ranging from square meters to hectares at a corresponding temporal resolution within a range of minutes to days. Figure 2.1 visually classifies hillslope hydrology among the different branches of hydrology (adapted version of a figure in Becker and Nemec (1987)). The fol-

![Spatio-temporal classification of activities in hydrology](image)

Figure 2.1: Spatio-temporal classification of activities in hydrology. Source: Zappa (2003)
Following introduction to flow processes is illustrated by Figure 2.2. Every watershed may be considered to be composed of a mosaic of hillslopes. Therefore understanding of processes at hillslope scale will strengthen the knowledge about behaviour of whole watersheds (Bronstert and Plate, 1997). However, hillslope hydrology and its applications are by far not limited to watershed understanding. As an example they can be directly used for questions arising at the hillslope itself, for example by investigating pollution dynamics in environmental chemistry or in conjunction with other related scientific disciplines (Bronstert and Plate, 1997) as demonstrated within the thesis at hand.

### 2.1.2 Flow processes at the hillslope scale

This section briefly summarizes flow processes at the hillslope scale being important for the hydrological modelling exercises in later chapters. Refer to Figure 2.2 for an illustration of these processes.

**Hortonian Overland Flow** (HOF) occurs when the infiltration capacity of the soil is exceeded. Although several studies observed other flow processes and pointed to the deficiencies of the HOF concept (see e.g. Kirkby (1978) and Scherrer and Naef (2003)), HOF dominated the perception of runoff generation at the hillslope scale. Research showed that HOF occurs only on a minor part of hillslopes because its initiation was found to be hindered by higher infiltration rates than precipitation intensities.
mainly due to the effects of plant roots and burrows from faunal activity. Additionally, HOF is rarely observed on vegetated slopes (Kirkby and Chorley, 1967).

**Saturation Overland Flow** (SOF) was suggested as a flow process by Kirkby and Chorley (1967). They found that a non-Hortonian overland flow was produced by saturated soils (partly by lateral flow) even though the local infiltration capacity has not been exceeded by the rainfall intensity.

**Subsurface Flow** (SSF) Surprisingly, the concept of lateral water movement in the soil is almost as old as the concept of HOF but won recognition decades later (works of Lowdermilk, Hursh, Brater and others, see Kirkby (1978)). These studies are the origin of a process called subsurface flow or through flow. In this thesis, the abbreviation of subsurface flow is explicitly SSF, as elsewhere in literature other definitions can be found, referring SSF to subsurface storm flow. The latter term is written out here if used. Despite the fact that the origins of the SSF concept existed since the 1940's, research only intensified 15 years later.

**Bypass Flow** Bypass flow is the general term for fast preferential flow bypassing the matrix. Bypass flow is substantially different from matrix flow, as large portions of flow occur in a small percentage of the total soil pore volume (Faeh et al., 1997). He indicates, that no clear distinction between macropores and soil pipes is evident and proposes that a differentiation should not rely on cross-sectional size, as suggested by some other studies (Beven, 1982), but on length and orientation of these flow channels. If bypass flow is orientated vertically in soil channels, focusing on the infiltration process, it is referred to as macropore flow (as e.g. described in Beven and Germann (1982)) and as pipe flow, if flow channels are orientated parallel to the slope. The thesis at hand accepts this definition, especially because Weiler and McDonnell (2004) used this definition and terminology for developing the here applied hydrological software HillVi.

Chains of connected soil pipes, developed nearly parallel to the soil surface, are commonly found on vegetated hillslopes (e.g. in Uchida et al. (2001)). Soil pipes can either be formed by soil fauna (e.g. mole, mouse and earthworm burrows) or more frequently by dead root channels in forest soils (Weiler et al. (2003) or Rickli and Bucher (2002)). Diameters of pipes are within a narrow range of a few centimetres. In clayey soils, they are often uniformly spaced and were found to lay either in the top soil layer with a high faunal activity, or within a narrow range of about 20 cm above an impermeable layer as the soil-bedrock interface (Uchida, 2004). Beven and Germann (1982) discussed the importance of bypass flow on total water flow in the soil and reviewed the research done so far on macropore flow. They mentioned that the idea of water movement through soil channels dates back to at least 1864. In the 1980s Mosley (1982) initiated studies describing the role of lateral preferential flow on fast subsurface runoff. Bypass flow was found to be an important control at sites where the subsoil had a closed structure or even impermeable bedrock. At sites where the parent rock was shattered and permeable, formation of a saturated zone is hindered and bypass flow plays a negligible role. Mosley (1982) con-
cluded that “for rapid flow through macropores to have significant influence on streamflow, saturation of the soil is necessary before water can enter the macropores”. Even though many studies on the effect of channelled flow and its contribution to runoff formation have been conducted, a general theory on combined subsurface and bypass flow at hillslope scale is still not available.

Pipe flow has been considered by many studies to have significant influence on landslide initiation (e.g. Blong and Dunkerley (1976), Brand et al. (1986), Pierson (1983), Rickli and Bucher (2002), Sidle and Kitahara (1995)). Under normal rainfall conditions, a pipe network may drain a hillslope distributing water downslope and perched water tables may dissipate quickly (McDonnell, 1990). Under heavy rainfall conditions, when pipe flow transmissivity is lower than concentration of water into the soil pipes (McDonnell, 1990) or if soil pipes are blocked or are a dead-end passageway, thus increasing pore water pressure of the surrounding matrix (Pierson (1983) and Brand et al. (1986)), susceptibility for slope failure increases. Uchida and Mizuyama (2002) showed with a numerical approach, that once a collapse of the pipe walls causes clogging of a soil pipe, the safety factor (see Section 2.3 of the slope soils promptly decreased. Rickli and Bucher (2002) found by investigating 133 shallow landslides in Napf and Appenzell (CH), that macropores were present in 90% of all landslides. In 70% of all locations, macropores penetrated into the shear plane. Macropores (mainly root channels and worm holes) were washed out by infiltrating water. The schematic diagram in Figure 2.3 illustrates the effects of pipeflow on landslide initiation (taken from a review study on pipe flow effects on hydrological processes and its relation to landslides (Uchida et al., 2001)).

More recent works by Faeh (1997) and Faeh et al. (1997), Scherrer and Naef (2003), Scherrer et al. (2007) or Kienzler and Naef (2008) attempt to describe discharge formation at the hillslope scale by conducting a series of comparable sprinkling experiments at different sites in Switzerland. As an outcome of such investigations, Scherrer and Naef (2003) proposed a decision scheme to indicate dominant runoff processes of hillslopes (hillslope-plots), with a more sophisticated classification of the dominant flow processes.

**Deep Percolation (DP)** Under certain conditions infiltrating water reaches the soil-bedrock interface. Depending on the bedrock characteristics, water may either be blocked and forced to flow laterally (and downslope) along a soil-bedrock interface or it percolates deeper into the soil (Anderson et al. (1997), Scherrer et al. (2007)). In past studies only minor attention has been paid to bedrock-permeability and its effects on water balance and flow due to severe difficulties in observing and quantifying such processes (Tromp-van Meerveld and McDonnell, 2006).

**Perched water table** Depending on the soil configuration, an impeding layer or a impermeable layer may block water and result in a perched water table (perched aquifer). This is an aquifer in the vadose zone and occurs above the regional ground water table. In this thesis the term water table is used, it refers always to a perched water table within a slope and not to the ground water table. Synonyms for this term are depth of saturation (because zones above an impermeable layer are saturated to a certain
2.1 Hillslope Hydrology

Figure 2.3: Schematic diagram describing the processes which explain effects of pipeflow on landslide initiation. Source: Uchida et al. (2001)

To conclude this chapter the state of knowledge of the processes contributing to discharge/runoff formation, as observed and described by numerous studies is summarized based on Kirkby (1978), Kirkby (1988) and Scherrer and Naef (2003). These processes are:

- **Hortonian Overland Flow (HOF):** occurs if the precipitation intensity or snow melt rates exceed the infiltration capacity of the (top)soil layer and water flows on the soil surface.

- **Saturation Overland Flow (SOF):** occurs when the storage capacity of the soil is exceeded (saturation of the soil profile), so that all additional water is forced to flow on the surface. Thereby, the infiltration rate of the soil is never exceeded.

- **(Fast) Subsurface Flow (SSF):** is water flow in the soil between the surface and the bedrock.

- **Deep Percolation (DP) or seepage:** occurs when the infiltrating water percolates to the groundwater zone or into the soil bedrock.
2.2 Pedology

2.2.1 Soil configuration

**Texture**  Soil is composed of three components: solids, liquids and gases. The solid phase is a mix of humus particles, rocks and mineral fragments of different sizes and forms. Together, they constitute the soil texture, which is a key factor in soil characterization (Scheffer and Schachtschabel, 1992). The soil components are classified by grain size, which allows soils to be classified according to their grain size distribution. Grain sizes < 2 mm constitute the fine soil, grain sizes > 2 mm the soil skeleton. The fine soil is further divided into: sand (2000-63 \( \mu \)m), silt (62-2 \( \mu \)m) and clay (<2 \( \mu \)m) (this classification is according to DIN4220. US-Soil Taxonomy indicates other classes).

**Porosity**  Grains constituting soils do never entirely fill the given volume by their mass. The ratio of void space to solid material is either described by the amount of hollows (pores) or by soil density. The pore volume, often referred to as porosity \( n_{\text{tot}} \), is defined by:

\[
 n_{\text{tot}} = \frac{V_p}{V} \tag{2.1}
\]

where \( V_p \) represents the pore volume, \( V \) the total volume of the soil.

**Drainable Porosity**  Drainable porosity \( n_{DP} \), in some cases also referred to as specific yield, was introduced in hydraulic groundwater equations to model the relation between water outflow volumes and groundwater dynamics. From a hydrological point of view, drainable porosity is important to estimate water-yielding capacities of soils. According to Johnson (1967) specific yield is defined as the ratio of the volume of water that a saturated rock or soil will yield by gravity to the total volume of the rock or soil, usually expressed as a percentage. McGuire et al. (2006) accordingly define drainable porosity by the difference in volumetric water content between saturation and field capacity (e.g. approximately from 0 to 100 cm WC). For this thesis, the latter definition is applied, as in literature slightly different definitions of drainable porosity or specific yield can be found related to groundwater hydraulics.

2.2.2 Soil water

**Soil water content**  The soil water content or moisture content \( \theta_w \) is, not only from the hydrological perspective, a dominant soil state quantity. Processes such as heat transport, transport of soluble substances and gas transport depend highly on the soil water content. Further, soil mechanical properties such as the shear strength or biological processes such as water uptake by plant roots depend on the soil water content. The volumetric water content is defined as:

\[
 \theta_w = \frac{V_w}{V} \tag{2.2}
\]

where \( V_w \) is the water volume and \( V \) the total (soil) volume \([m^3/m^3]\). In practice, measurements of the soil water content pose a problem. The simplest and most efficient way to measure soil water content
2.2 Pedology

is the collection of soil samples, weighing them directly after excavation before drying them in an oven. The loss of weight of water during the drying process equals the water content at excavation. However, this method destroys the soil plot and makes continuous temporal measurements impossible, which is in most cases of main interest. For the purpose of continuous temporal measurements indirect methods are chosen: so called proxy quantities are measured, which are closely linked to the water content. According to Flühler and Roth (2004), measurements relying on this method need to pursue at least three steps: "1. Measurement of the proxy quantity, 2. calculation of water content out of this quantity and 3. calibration of the measurements with simultaneously excavated soil samples". A common method to measure soil water content is *Time Domain Reflectometry* (TDR). This method is based on the difference of dielectrical constants of water ($\epsilon_w$) which is approximately 81.5 and soil ($\epsilon_s$) which is about 3 to 5. Along two electrical conductors with a length of 10 to 30 cm and a distance of a few centimetres between the conductors, electric current induces a electromagnetic field. Soil media with their corresponding dielectrical constants induce a change of impedancy and reflect a part of the pulse energy. By analysing the magnitude, direction and shape of the reflected waveform the soil water content can be calculated:

\[
v = \frac{c}{\sqrt{\epsilon_s}}
\]

where $v$ represents the velocity of the electromagnetic wave, $c$ the light speed and $\epsilon_s$ the dielectrical constant of the soil. The propagation distance along the probe is twice the probe length (2L). Consequently, propagation velocity is:

\[
v = \frac{2L}{t_s}
\]

Combining (2.4) and (2.3) leads to:

\[
\epsilon_s = \left[\frac{ct_s}{2L}\right]^2
\]

As mentioned initially, $\epsilon_s$ needs to be transformed to water content. Different approaches have been developed, but most commonly the following approach is applied (Topp, 1980):

\[
\theta_w = -5.3 \times 10^{-2} + 2.92 \times 10^{-2} \epsilon_s - 5.5 \times 10^{-4} \epsilon_s^2 + 4.3 \times 10^{-6} \epsilon_s^3
\]

Additionally and/or alternatively, the probes can and should be calibrated in the laboratory. Another instrument to measure soil water content is called *Moisture Point (MP)* (from "Environmental Sensors Inc. ESI"). MP devices base on the same technology as TDR. The main difference is that each probe consists of several conducting elements, making continuous profile measurements possible.

**Soil water potential** Forces driving the soil water movement are difficult to define, therefore it is common in soil science to consider their potential for work (Hillel, 1980). The potential is defined as "the mechanical work that is needed to transport a unit of water (volume, mass or weight) from a given point in a field of force to a reference point" (Scheffer and Schachtschabel, 1992).

The soil water potential ($\psi$) is defined as:

\[
\psi = mgh
\]
where \( m \) is the mass of water, \( g \) the gravity and \( h \) the height over a free water table. If the weight is the reference, equation (2.7) reduces to \( \psi = h \) (cm water column). The total potential equals the sum of different partial potentials:

\[
\psi = \psi_z + \psi_m + \psi_g + \psi_o
\] (2.8)

\( \psi_z \) is the gravity potential and equals the energy necessary to move a unit of water from a reference level to a certain height. If weight is the reference, the gravity potential is equal to the height \( h \).

\( \psi_m \) is the matrix or capillary potential (or also suction). This potential describes all the effects of the soil matrix on soil water and acts in most cases in the opposite direction than \( \psi_z \). The matrix potential can be measured with tensiometers. Tensiometers are composed of a porous ceramic candle connected to a manometer by a fluid filled tube. The manometer is calibrated to zero if the candle lies exactly on the soil water surface. The dryer the soil around the candle is, the more water will be drawn out of the system and the manometer shows an increase in suction. This pressure is given in Pa or sometimes as a height in cm in a water column (cmwc), where 1 cmwc = 100 Pa. The sum of \( \psi_m \) and \( \psi_z \) is the hydraulic potential \( \psi_H \).

The effects of salts on the soil water status is called osmotic potential (\( \psi_o \)) and is measured as the potential for water to move across a semipermeable membrane. Finally, the gas potential (\( \psi_g \)) has to be considered, if the air pressure in the soil is not equal to the atmospheric pressure.

**Water Retention Curves** If water in the top layer of a soil profile is evapo(transpi-)rated, \( \psi_m \) decreases toward higher negative values which is equal to an increase in suction, but \( \psi_z \) remains constant, forcing the water to flow in the direction of the higher negative potential \( \psi_m \), thus toward the soil surface. The opposite occurs, if for example rain infiltrates in the upper soil layers, forcing the water to flow further down into the soil (driven by \( \psi_m \)). This phenomenon demonstrates the relation between suction (\( \psi_m \)) and water content: TDR and Tensiometer data can be combined to provide information about soil characteristics and water balance. Such curves are called water suction curves, pF-curves or water retention curves (WRC) and give valuable information about:

- soil characteristics as the shape of the curve is influenced mainly by the grain size distribution (granulation, texture)
- stresses and stress history.

Besides the structural influence on the shape of a WRC, also the direction of change of water content influences the curve. This phenomenon is called *hysteresis*.

**Saturated hydraulic conductivity** By investigating water infiltration into an artificially saturated soil column, the French engineer H. Darcy (1803-1858) found, that water flow in the column is proportional to the hydraulic height (*Darcy’s Law*):

\[
j_w = -k_{sat} \frac{\partial}{\partial z} h_w
\] (2.9)
2.2 Pedology

where $j_w$ is the water flux, $h_w$ the hydraulic height and $z$ the depth of infiltration. $k_{sat}$ is a factor which varies with depth and defines the rate of movement of water.
2.3 Shallow landslides and slope stability

Shallow landslides  Landslides are geological phenomena which include wide ranges of mass movements like rock and block fall, failure of slopes with different depths and shallow debris flows. Over-steepened slopes are the primary reason for landslide occurrence. Although other factors as precipitation intensity and total amount, soil materials, snow melt, vegetation or even fire may affect the original slope stability. Shallow landslides in particular are characterized by their sliding surface located within the soil mantle or weathered bedrock. A typical depth is from a few centimetres to a few meters. Due to the quasi random nature of occurrence of shallow landslides and difficulties in modelling, Rickli and Bucher (2002) conducted case studies by investigating several hundreds landslides occurred in Switzerland in 2002. Shallow landslides were found in an inclination range between 23-50° and more than 90% had a volume smaller than 300 m². Most of the investigated landslides showed instabilities already before the event. For a significant number of events, the sliding surface was the bedrock.

Slope stability  As described earlier, soils are an aggregation of different solids (minerals) with voids, filled with water and/or air. The shearing resistance of soils is commonly described as a combination of cohesive forces and friction. Cohesion is due to soil particles, dominantly minerals, cemented mainly by chemical (molecular and ionic) bonds or by electrostatic and electromagnetic forces. Frictional strength originates from the frictional resistance of minerals against each other. Figure 2.4 (a) serves as basis to illustrate forces acting on a block with help of the triangle of forces. As long as a soil element is stationary, the normal force $N$, the tangential force $H$, the reactionary force $R$ and the angle $\alpha$ are related by:

$$\tan\alpha = \frac{H}{N}$$  \hspace{1cm} (2.10)

Increasing $H$ results in increasing $\alpha$ until the soil element starts to slide. $\alpha$ then reaches its maximum value and is defined as $\phi$. For several purposes in soil mechanics, the triangle of forces is converted into stresses. Equation (2.10) can then be rewritten as $\tan\phi = \tau/\sigma_n$ where $\tau$ is the shear stress and $\sigma_n$ the normal stress. In combination with cohesion $c$, the mobilised shear strength at failure $\tau_f$ can then be rewritten (Coulomb, 1773):

$$\tau_f = c + \sigma_n\tan\phi.$$  \hspace{1cm} (2.11)

Above expressions only apply when the soil is dry. Soil water content influences the shear strength by apparent cohesion. Apparent cohesion is produced by capillary stresses occurring as surface tension in water films between particles (Selby, 1982). In a moist soil, particles are temporary cemented by apparent cohesion and thus pore water pressure $u_w$ is negative. Full saturation results in total loss of apparent cohesion making soils particularly susceptible to shallow landslides. The following modified version of equation (2.11) takes the effect of pore water pressure into account:

$$\tau_f = c' + (\sigma_n - u_w)\tan\phi'.$$  \hspace{1cm} (2.12)
2.3 Shallow landslides and slope stability

(a) Forces acting on a block and frictional angle

(b) Stresses acting on a soil element on a shear plane

Figure 2.4: Schematic illustration of forces acting on a block on a plane (a) and stresses acting on a soil element in a hillside, adapted from Selby (1982)

(‘) stands for effective stresses and indicates their modification by pore water pressure.

The stability of a hillslope is usually quantified as factor of safety (FoS) defined by:

\[
\text{FoS} = \frac{\text{sum of resisting stresses}}{\text{sum of driving stresses}} \equiv \frac{\tau_f}{\tau} \tag{2.13}
\]

Thus, a FoS of less or equal 1 indicates a condition for slope failure. Figure 2.4 (b) illustrates stresses acting on a soil element in a hillside. The weight of the soil element is:

\[
W = \gamma z \cos \beta \tag{2.14}
\]

The normal stress on the shear plane is then:

\[
\sigma_n = \frac{W \cos \beta}{l} = \gamma z \cos \beta \cos \beta \tag{2.15}
\]

and the shearing stress on the shear plane is

\[
\tau = \frac{W \sin \beta}{l} = \gamma z \cos \beta \sin \beta \tag{2.16}
\]

where \( \gamma \) its unit weight at a given soil water content, \( z \) the vertical distance from surface to the shear plane (vertical) and \( \beta \) the inclination of the slope. Combining equations 2.12 - 2.16 results in:

\[
\text{FoS} = \frac{\epsilon' + (\gamma z \cos^2 \beta - u_w) \tan \phi'}{\gamma z \sin \beta \cos \beta} \tag{2.17}
\]

The pore water pressure \( u_w \) can be derived either from the piezometric head \( h \) multiplied by the unit weight of water \( \gamma_w \) or from tensiometer measurements. Equation (2.17) illustrates the contribution of water table height to the stability of a hillslope: an increasing water table results in an increase of \( u_w \).
(\(\gamma_w h\)), thus the numerator decreases and (2.13) converges to the value 1 or below. However, the stability criterion described in by equation (2.11) does not apply for the conditions in a unsaturated environment. It has been shown that slope instabilities may occur even before full saturation of the soil is reached. Equation (2.13) (respectively equation (2.17)) can be adapted to apply in these conditions (Springman et al., 2003).
Chapter 3

Field Site

3.1 Tössegg, Canton Zürich

3.1.1 Climate and Hillslope Description

Tössegg is situated at the northern border of canton Zurich at the river Rheine (Figure 3.1). The investigated hillslope is located some 20 meters away from the river at an elevation of about 400 m above sea level (47° 33’ 13.14” N, 8° 33’ 40.87” E). The slope exposition is north west. Slopes in this region are susceptible to failure due to heavy rainfalls as a series of 42 shallow landslides in May 2002 in close proximity to here investigated hillslope showed (Thielen, 2007). A dissertation project in the CCES-TRAMM framework investigated the influence of soil saturation on slope stability from 2004 to 2007 (Thielen, 2007). According to field and laboratory experiments and modelling with the finite-element

Figure 3.1: Location Tössegg, Kanton Zürich. Source: Swisstopo

Figure 3.2: Climate diagram, Zürich-Kloten, Source: Norm period 1961-1990, MeteoSchweiz
software VADOSE/W (GEO-SLOPE Int. Ltd.) the factor of safety did never fall below a critical value. The slope inclination was found to be too low to allow for a triggering event. Thielen (2007) concluded, that future research should focus on preferential flow in both modelling and field experiments - a main justifications for the modelling work done in the thesis at hand.

**Climate** Figure 3.2 shows the climate diagram of the nearest neighbouring meteo station Zürich-Kloten at a distance of 5 kilometres (436 m above sea level, 47°28’ 59.59” N/ 8°32’ 10.16” E). The climate norm period 1961-1990 showed an average yearly total precipitation amount of 1017 mm. Mean yearly temperature was 7.8°C with a maximal amplitude of 18.1°C. For the hydrological simulations of extreme events presented in Chapter 5, heavy precipitation events have been chosen according to MeteoSchweiz (2005a) and (2006b) or analysis of the station data of Zürich-Kloten.

**Hillslope characteristics** The meadow slope is characterized by shallow quaternary sediments overlying a sandstone basement (Thielen, 2007). At 17 meters form the slope base level, an edge divides the slope into a lower part, with an average inclination of 27° and an upper part of about 20° inclination. The depth distribution of soil layers and bedrock were measured with electrical resistivity tomography (ERT) (Friedel et al., 2006). The soil mantle is constituted of three layers: below surface, an organic

---

Figure 3.3: Tössegg slope with locations of the four instrumented depth profiles (red circles) and the location of the meteorological station. Source: IGT ETH Zurich

Figure 3.4: DEM of the Tössegg hillslope. The figure shows the soil depth in meters from surface to the bedrock topography. The left and right transect from Figure 3.5 represent two horizontal lines through x=2 and x=13. Source: ERT measurements Friedel et al. (2006)
### 3.1 Tössegg, Canton Zürich

<table>
<thead>
<tr>
<th>Grain Size Diameter [mm]</th>
<th>Clayey Sand Mass %</th>
<th>Silty Sand Mass %</th>
</tr>
</thead>
<tbody>
<tr>
<td>&lt; 31.5</td>
<td>100.0</td>
<td>100.0</td>
</tr>
<tr>
<td>&lt; 0.5</td>
<td>97.0</td>
<td>98.3</td>
</tr>
<tr>
<td>&lt; 0.06</td>
<td>41.2</td>
<td>26.8</td>
</tr>
<tr>
<td>&lt; 0.02</td>
<td>23.3</td>
<td>15.8</td>
</tr>
<tr>
<td>&lt; 0.002</td>
<td>7.1</td>
<td>5.6</td>
</tr>
</tbody>
</table>

| Mean Pore Size [µm]      | 5.85                |
| Dry Weight [kN/m³]       | 12.9-14.9           |

| Angle of Internal Friction [°] (φ') | 31 | 37.5 |
| Apparent Cohesion [kPa] (c')        | 0  | 0    |

Table 3.1: Soil characteristics. Source: Thielen (2007), modified

Highly rooted top layer extends to a depth of 30 cm; the second layer is of clayey sand with a thickness of up to 1.5 meter; the base layer above the sandstone bedrock is of silty sand with a thickness of up to half a meter in the lower part of the hillslope. Mean soil thickness from surface to bedrock is around 2.5 meters and 1 meter from surface to the interlayer. This interlayer represents the boundary between clayey and silty sand. At the upper end of the slope, the soil is generally thicker with a maximum depth of approximately 5 meters to the bedrock and 1.8 meters to the interlayer. Minimum heights are 1 meter respectively 0.3 meters for bedrock and interlayer. In the center of the hillslope, a bedrock depression was predicted by the ERT.

Figure 3.3 shows a picture of the hillslope and Figure A.1 illustrates the 3D model of the bedrock topography underlying the surface topography. Figure 3.4 presents the digital elevation model (DEM) with bedrock to surface depth information derived from electrical resistivity tomography measurements. Several soil and soil core samples of differing depths and locations across the hillslope were analysed by Thielen (2007) to define mineralogy, grain size distribution, plasticity, shear strength and hydraulic conductivity (Table 3.1). Soil hydraulic conductivity was determined from 8 soil core samples from 4 different depths in triaxial cells, excluding the influence of macropores. Geophysical data and geotechnical investigations, are in good agreement with ERT measurements according to Friedel et al. (2006) and Thielen (2007).

#### 3.1.2 Field Data and Experiments

**Instrumentation** At four selected locations of the hillslope (according to Figure 3.3) depth profiles were measured continuously using TDR, MP and tensiometer devices. Depth profiles and layer configuration for the instrumented locations are according to the red dashed lines in Figure 3.5, which shows the left and right outermost ERT transects looking up slope. Due to the shallow soil depth of about 1 meter in the upper left soil profile as predicted by the ERT, the MP was equipped with shorter segments.

All installed instruments are summarized in Table 3.2, where Figure A.2 and Table A provide further information about place and depth of single TDR, MP and tensiometer probes. Three piezometers were
Applied soil instrumentation | Data | Unit
---|---|---
Temperature probes (Rotronic) | air and soil temperature | °C
Rain gauge, Markasub (CH) | precipitation | mm/time
Barometer (SENSYM) | air pressure | mbar
Humidity sensor (Rotronic) | air humidity | %
TDR100 (Campbell/USA, Markasub/CH) | volumetric water content | m³/m³
MP917 (E.S.I./Canada) | volumetric water content | m³/m³
Tensiometer (ETH Zurich/ITÖ) | suction (tensiometer) | cmwc
Channel with box (construction by Keller (CH)) | surface runoff | l/s
Piezometer (Keller/CH) | ground water level (piezometer) | m below surface

Table 3.2: Instrumentation

Figure 3.5: Soil profile of left and right boarding transects. Red lines indicate positions of instrumented soil depth profiles

installed at the top flat part of the slope due to accessibility reasons with the core drill machine. They lay either outside the DEM of HillVi or at the top border, giving therefore no directly comparable data for the simulations and are not used in this study. A complete meteorological station was set up in the centre of the hillslope measuring precipitation, humidity, air pressure, air and soil temperature. Unfortunately, technical problems of the rain gauge caused the on-site precipitation data to be erroneous. Thus precipitation data from the MeteoSchweiz station Zürich-Kloten were used for the simulations. Whole time series of MP, TDR and tensiometers are plotted in Appendix A.3, A.4 and A.5.
Sprinkling Experiments  Before disassembling the field site in autumn 2007, dye sprinkling experiments have been conducted by Peter Kienzler (Geotechnical Institute, ETH Zurich) at three of the four instrumented soil depth profiles (upper left, upper right and lower left). The dye applied in this experiment was Brilliant Blue FCF. The sprinkling intensity was set to 60 mm/h for 2 hours irrigation (120 mm total precipitation amount with a Brilliant Blue concentration of 4 g/l). The day after irrigation, the soil plots were excavated by digging long trenches of about 1.5 m depth and subsequently cutting off profile slices of about 10 cm width. On the basis of the sprinkling experiments pictures in Figures 3.6, 3.7, soil characteristics have been discussed and conclusions for the modelling part with HillVi have been drawn.

Figure 3.6: Sprinkling installation (a) and excavated profiles (b) and (c)

All profile pictures show that the Brilliant Blue distribution is quite homogeneous in the upper 50 cm for all three plots. This is in agreement to the ERT measurements indicating a highly rooted and permeable 30 cm thick top soil mantle. Below this depth, the infiltration is highly preferential. The existence of preferential flow channels was confirmed in all of the three excavated plots in different forms. Figure 3.7 (b) and (c) show two such macropores which were coloured by the through flow of Brilliant Blue. As a consequence of such preferential flow, Figure 3.6 (b) shows a Brilliant Blue concentration spot in 100-150 cm depth. This spot lies in the higher permeable silty sand layer. Here, the interaction of soil matrix and macropores is higher than in the clayey sand layer above which results in diffuse colouring of the soil matrix. This shows that already in a micro environment as in the upper left plot, measurements of water content and suction may differ significantly between neighbouring instruments depending on the preferential flow behaviour. Figure 3.7 (a) shows a TDR probe lying directly in a water supplied zone, whereas the TDR in Figure 3.6 (b) to the right is not affected by preferential flow. These findings show, that interpretation of soil moisture and tensiometer data has to be done with care because their measurements depend highly on the particular micro environment in the soil. The upper left and the
lower left excavation sites did not indicate the existence of an impeding soil-bedrock layer and lateral flow due to a impeding layer was not observed in these profiles.

The upper right soil depth profile has to be discussed separately because a differing infiltration and flow behaviour compared to the upper left and lower left plot was found. The ERT measurements (refer to Figure 3.5) predicted a layer change between clayey and silty sand at a depth of about 40 cm. From this datum a silty sand layer of more than 1 meter thickness should extend to the bedrock. However, as picture 3.6 (c) demonstrates, the lower layer had a compact constitution comparable to sandstone. A possible interpretation is that the clayey and silty sand layers are very shallow and the impeding layer at 40 cm depth is already the sandstone basement underneath the organic A-horizon of 40 cm thickness. The water perched at the top end of the impeding layer in 40 cm depth and a saturated layer developed. Lateral preferential flow was observed at this location occurring along soil strata and contact veins, but not in soil pipes. At several locations, the water percolated further into this layer, and flew preferentially along contact veins of other constitution (on the right side of the picture). At the bottom of the soil profile a crack was found spreading 80 cm horizontally to the right and 20 to 30 cm further vertically into the soil. The crack transmitted several litres of water without impoundment. These findings from the upper right soil depth profile demonstrate clearly the local heterogeneity of soil constitution (within a few meters to the other depth profiles), the existence of saturated zones if the conditions of impeding layers within the soil profile are given and the occurrence of soil cracks and veins wherein water flows preferentially. Information regarding parametrization of HillVi was difficult to derive from the sprinkling experiments. As described in Section 2.1.1 in several studies soil pipes were found to be mostly located either at the soil surface or at the soil-bedrock interface. Even though the existence of such soil pipes could be verified from these field experiments, no clear height
or density emerged. Further the preferential flow occurred also along other preferential flow structures than soil pipes. If they significantly influence fast subsurface runoff remains doubtful. The cracks in the sandstone enhance the bedrock permeability and large amounts of water may leak deep into the bedrock.
Chapter 4

Methods

4.1 HillVi

4.1.1 Introduction

The hydrological model applied in this work is HillVi (Version 44_MC) developed by Markus Weiler (see Weiler and McDonnell (2004)). HillVi's name is derived from so-called virtual hillslope experiments and was programmed as a basis for a "novel approach" to investigate hydrological processes at the hillslope scale. Traditional modelling of hillslope hydrology focused on analytical solutions of soil physical and fluid dynamical process description (namely Richard's, Darcy's and kinematic wave equations). Such models usually need a large number of input parameters. However, since the processes of water and solute fluxes in a hillslope are very complex, difficult to quantify and very heterogeneous, an appropriate parametrization of numerically explicit models is rarely possible and differ between each investigated hillslope. Out of this statement follows the lack of a common conceptualization of hillslope hydrology. Furthermore, Weiler and McDonnell (2004) note, that there is a poor dialogue between field experimentalists, who often have a conceptual understanding and knowledge of particular hillslopes based on their field experiences and modellers, who focus mostly on physical description of processes and do not incorporate field knowledge in the model structure. The mentioned "novel approach" attempts to improve conceptualization and process understandings in hillslope hydrology based on a virtual experiment framework. Virtual experiments are defined as "numerical experiments with a model driven by collective field intelligence" (Weiler and McDonnell, 2004). The intent is to investigate and isolate so-called "first order controls" (the main and essential process in water (and solute) flux at the hillslope scale), where modeller and experimentalist work together, analysing the modelling results collectively (Weiler and McDonnell (2004); Seibert and McDonnell (2002)). HillVi should allow modellers and experimentalists to avoid above mentioned constraints and strengthen the knowledge of first order controls. HillVi is a simple model, with few "tunable" parameters and provides a sophisticated graphical quasi 4-dimensional output. Figure 4.1 shows the model concept with the main processes described in
4 Methods
detail later in this chapter. HillVi is entirely written in the Interactive Data Language (IDL Ed. V5.6).

![Figure 4.1: Model concept of HillVi (own illustration). Refer to Section 4.1 for a detailed mathematical description of the model processes.](image)

The basic data requirements to run HillVi are:

- Digital Elevation Model (DEM) of surface and bedrock (subsurface) topography
- Precipitation input
- Optional: evapotranspiration data, measured runoff (or other experimental data)

HillVi belongs to the category of spatially distributed, physically based conceptual models. A two layer DEM provides detailed information about surface and bedrock topography of the hillslope. Since Dunne
4.1 HillVi

(1978) and others found that lateral subsurface flow in hillslopes is mainly driven by a perched water table, most models explicitly represent the unsaturated and the saturated zone and the coupling between them. HillVi is programmed on this basis too. As initially defined in Section 2.1.2, the terms related to the perched water table depth of saturation, water table depth and water level are equivalent here.

4.1.2 Lateral flow

Wigmosta et al. (1994) described the grid cell by grid cell method to route topographically driven subsurface flow, which is implemented in HillVi. This method allows to calculate quasi three dimensional saturated subsurface flow by approximation of transient conditions by a series of steady state solutions based on local hydraulic gradients. Each hillslope cell is surrounded by eight neighbouring cells as shown in Figure 4.2. The centre cell can exchange water with each of its neighbouring cells. The centre cell node is indexed by \(i\) and \(j\) (as its centre point). The possible eight routing directions “\(k\)” are indexed from 0 to 7 clockwise (in a programming context, 0 is the first number). As an example, \(k=4\) stands for the direction from the centre cell node \(i,j\) to the lower right cell node \(i+1, j+1\). Hydraulic gradients, calculated as water table slopes for each cell, define the flow direction for each time step. The time dependent saturated subsurface flow rate from a cell \(i,j\) to its down-gradient neighbours is calculated under Dupuit-Forchheimer assumptions (Freeze and Cherry, 1979):

\[
q(t)_{i,j,k} = \begin{cases} 
-T(t)_{i,j,k} \tan \beta_{i,j,k} w_{i,j,k} & \beta_{i,j,k} < 0, \\
0 & \beta_{i,j,k} \geq 0.
\end{cases} \tag{4.1}
\]

where \(q(t)\) is the rate of SSF from cell \(i, j\) in \(k\) direction, \(T(t)\) the transmissivity from cell \(i, j\) in \(k\) direction, \(\beta\) the water table slope in \(k\) direction and \(w\) the width of flow.

**Exponential transmissivity function** For some soils Beven (1982) found that an exponential function accounts for the decreasing saturated hydraulic conductivity with depth from soil surface originating

![Figure 4.2: Grid cell by grid cell approach. Source: Wigmosta et al. (1994)](image)
mainly from changing void ratios.

\[ k_s(z) = k_0 \exp \left(-\frac{z}{m}\right) \]  \hspace{1cm} (4.2)

Transmissivity from 4.1 can now be expressed as:

\[ T_{i,j,k}(z) = k_0 m \left[ \exp \left(-\frac{z_{i,j}}{m}\right) - \exp \left(-\frac{D_{i,j}}{m}\right) \right] \]  \hspace{1cm} (4.3)

where \( T(z) \) is the depth dependent transmissivity, \( k_0 \) the saturated conductivity at soil surface, \( D \) the soil depth, \( m \) the decline factor and \( z \) the depth into the soil profile.

If the hydraulic gradient, respectively the water table slope is smaller than 0 (case 1 in (4.1), substitution of (4.3) into (4.1) leads to:

\[ q_{i,j,k}(t) = \gamma_{i,j,k} h_{i,j}(t) \]  \hspace{1cm} (4.4)

with

\[ \gamma_{i,j,k} = - \left( w_{i,j,k} k_0 m \right) \tan \beta_{i,j,k} \]  \hspace{1cm} (4.5)

\[ h_{i,j}(t) = \exp \left(-\frac{z_{i,j}}{m}\right) - \exp \left(-\frac{D_{i,j}}{m}\right) \]  \hspace{1cm} (4.6)

**Power law transmissivity function** The behavior of declining saturated hydraulic conductivity with depth is not the best approximation for certain landscapes, such as mountainous catchments with thin surface layers. Iorgulescu and Musy (2006) therefore presented a power law transmissivity function, which allows transmissivity to reach zero at a prescribed depth:

\[ T_{i,j,k}(z) = \int_{z(t)}^{D} k_{s,i,j,k}(z)dz = k_0 m \left( 1 - \frac{z_{i,j}}{D_{i,j}} \right)^m \]  \hspace{1cm} (4.7)

where \( T(z) \) is the depth dependent transmissivity, \( k_0 \) the saturated conductivity at soil surface, \( D \) the soil depth, \( m \) the exponent and \( z \) the depth into the soil profile. Analogous to (4.4) flow can be calculated with

\[ \gamma_{i,j,k} = - \left( \frac{w_{i,j,k} k_0 D_{i,j}}{m} \right) \tan \beta_{i,j,k} \]  \hspace{1cm} (4.8)

and:

\[ h_{i,j} = \left( 1 - \frac{z_{i,j}}{D_{i,j}} \right)^m \]  \hspace{1cm} (4.9)

As mentioned above, \( q(t)_{i,j,k} \) is the water flux from cell \( i,j \) to the downward neighboring cell in k direction in m³ / timestep. \( \tan \beta_{i,j,k} \) is calculated for each cell \( i,j \) by subtracting the water table height in cell \( i,j \) minus the water table height of the neighbouring cells in k direction divided by the horizontal distance of the cell nodes. Because \( \gamma \) is dependent on k, it has to be calculated for all eight k’s for each slope cell, whereas \( h \) is only calculated once for each slope cell. For each cell the amount of matrix flow is calculated for all k-directions and summed up to the total flux.
4.1.3 Pipe flow

Several studies show the importance of pipe flow contribution to subsurface flow and subsurface storm flow (see Chapter 2.1.2). Mosley (1982) stated, that a significant contribution of pipe flow to total subsurface flow is only given when soils are saturated. Based on this suggestion pipe flow in HillVi is only allowed to occur within the saturated zone. Since the exact location and density of the pipe network is unknown, HillVi generates a random pipe network. Main input parameters for the random generation of the pipe network in HillVi are:

- density $d_p$ of pipes (as percentage of hillslope cells contributing to the pipe network)
- height of pipe above bedrock $h_p$ (if positive as distance from bedrock, if negative from soil surface)
- standard deviation of pipe height above bedrock

With these parameters, HillVi computes a random pipe network based on two statistical distributions. The sum of $h_p$ and a randomly sampled value of a normal distribution with the given standard deviation is added to the height of each bedrock cell. This information is passed to an array and subsequently cells where pipes start are chosen by a uniform distribution of the same array dimensions multiplied by $d_p$. The algorithm to calculate pipe flow along the formerly generated pipe network is empirical and was suggested by Sidle and Kitahara (1995). This study found the following relation by realizing bench-scale experiments of water flow in PVC-tubes embedded in sand filled boxes:

$$q_p = \alpha (H - 0.03) \beta$$

where $q_p$ is the pipe flow rate, $H$ the average piezometric heads at several depths, $\alpha$ the coefficient including hydraulic conductivity, roughness coefficient, hydraulic gradient, pipe and box dimensions and $\beta$ representing the slope of the log-linear regression of the curve fitted to the experimental data. The factor 0.03 represents the height of pipes above bedrock in meters. According to (4.10), HillVi calculates pipe flow using:

$$q_p(t) = p_{const} k_p \sqrt{A} (w(t) - z_p)^\alpha$$

where $q_p(t)$ is the time dependent pipe flow, $p_{const}$ the multiplicative constant for determining inflow into pipe $k_p$ the hydraulic conductivity, $A$ the grid cell area, $w(t)$ the time dependent water table height, $z_p$ the height of pipe start and $\alpha$ the power exponent Pipe flow only occurs if the local water table is higher than pipe start height. As follows, pipe flow is never larger than the storage above pipe start height.

4.1.4 Unsaturated zone and recharge

The amount of water in both the saturated and unsaturated zone available for fluid flow depends on the soil porosity $n$. Total porosity $n_{tot}$ is the total pore volume in the soil matrix and an input parameter for HillVi. However, this total volume is not available for fluid flow, due to capillary forces and increase
in suction retaining the water in the soil matrix (see Chapter 2.2.1). The amount of water available for fluid flow equals the drainable porosity of the soil. Studies carried out by Gillham and McGuire et al. (2006) and others suggest that drainable porosity is a key parameter in modelling water balance in the unsaturated zone and varies in space and time. Older versions of HillVi (Weiler and McDonnell, 2004) defined drainable porosity as constant throughout the soil profile, whereas the present version accounts for an exponential decline of \( n_d \) with depth:

\[
n_d(z) = n_0 \exp\left(- \frac{z}{b}\right)
\]  

(4.12)

where \( n_0 \) is the drainable porosity at soil surface, \( z \) the depth into the soil and \( b \) a shape parameter. The influence of the parameter \( b \) on the model is visualized in Figure A.6.

Given the two distinct porosities for the soil profiles, HillVi calculates the water storage compartments. Water storage in the unsaturated zone starts with an initial condition of volumetric water content and is calculated by:

\[
V_{\text{unsat}}(z) = z(n_{\text{tot}} - n_d(z))A
\]  

(4.13)

Water exchange from the unsaturated to the saturated zone (recharge) is calculated with a recharge function based on relative saturation. While depending on the saturated hydraulic conductivity at soil surface and its exponential- or power law- decline with depth, recharge is either:

\[
r(t) = \left(\frac{\theta(t)}{\theta_s}\right)^c k_0 \exp\left(- \frac{z(t)}{m}\right)
\]  

(4.14)

if the model is exponential or

\[
r(t) = \left(\frac{\theta(t)}{\theta_s}\right)^c k_0 \left(1 - \frac{z(t)}{D}\right)
\]  

(4.15)

if the model is power law function, where \( \theta(t) \) is the water content in unsaturated zone, \( \theta_s \) the maximum water content in unsaturated zone, \( c \) the power law exponent, \( k_0 \) the hydraulic conductivity at soil surface, \( z(t) \) the depth into the soil profile, \( m \) the shape factor and \( D \) the soil depth.

### 4.1.5 Vertical bypass flow

Vertical bypass flow represents the amount of water bypassing the matrix in macropores and feeding the saturated zone directly. The infiltration capacity of the soil is set as an input parameter. If infiltration excess occurs, water is directed to the overland flow algorithm and represents the HOF process. Remaining water, or the whole precipitation amount if no HOF occurs, infiltrates and is directed either to the unsaturated zone storage or the saturated zone storage by three different bypass flow models. All of them rely on a power law function with exponent \( \text{bypass} \). The first model

\[
bypass_{\text{flow}} = \text{infiltration} \left(\frac{\theta}{\theta_s}\right)^{\text{bypass}}
\]  

(4.16)

depends on the water content in the unsaturated zone. A second model depends on a rainfall intensity threshold by

\[
bypass_{\text{flow}} = \text{infiltration} \left(\frac{\text{infiltration \ threshold} \times \Delta t}{\Delta t}\right)^{\text{bypass}}
\]  

(4.17)
The third model calculates the amount of bypass flow by a constant fraction of rainfall input.

\[
\text{bypass}_\text{flow} = \text{infiltration} \times \text{fraction} \left( \frac{\theta}{\theta_s} \right)^{\text{bypass}}. \tag{4.18}
\]

### 4.1.6 Overland flow

If the infiltration rate is exceeded (HOF situation) or if the whole cell is saturated (SOF situation) the overland flow routine is initiated if a roughness coefficient is defined in the parameter file of HillVi. According to the Manning-Strickler equation:

\[
v = k J^{1/2} R^{2/3} \tag{4.19}
\]

where \(v\) is the flow velocity, \(k\) the roughness coefficient, \(J\) the slope gradient and \(R\) the hydraulic radius multiplied by the flow cross section, HillVi calculates the amount of overland flow and routes the flow to the cell with the highest gradient.

### 4.1.7 Water balance and storage

All the above described processes contribute in some way to the model’s water balance. The balance for the saturated zone is formulated as follows:

\[
\text{balance}_{\text{sat}} = \text{recharge} + \text{bypass} + \frac{\text{SSF}}{A} + \frac{\text{pipeflow}}{A} - \text{seepage} \tag{4.20}
\]

Accordingly, balance of the unsaturated zone is calculated:

\[
\text{balance}_{\text{unsat}} = \theta_{\text{temp}} + (\text{infiltration} - \text{bypass}) - \text{recharge} - \text{evapotranspiration} - \Delta(n_{\text{tot}} - n_d)(\exp \left(\frac{z'}{b}\right)) \tag{4.21}
\]

where \(\Delta\) is the change of water table and \(\theta_{\text{temp}}\) the unsaturated water content at former time step. To calculate the change of water table height per time step, equation (4.12) is integrated from height \(z_1\), which is the water table height at the former time step \((z(t-\Delta t))\) to \(z_2\), which is the actual water table height,

\[
\Delta B = \int_{z_1}^{z_2} n_d(z) dz = n_0 b \left[ \exp \left( -\frac{z_1}{b} \right) - \exp \left( -\frac{z_2}{b} \right) \right] \tag{4.22}
\]

and subsequently solved for \(z_2\) (the actual time step \(z(t)\)), which is

\[
z(t) = -bln \left[ \frac{\Delta B}{n_0 b} + \exp \left( -\frac{z(t-\Delta t)}{b} \right) \right] \tag{4.23}
\]

with

\[
\Delta B = \left[ \frac{(Q_{\text{out}} - Q_{\text{in}})}{A} - R \right] \Delta t \tag{4.24}
\]

where \(Q_{\text{out}}\) is the outflow from each cell, \(Q_{\text{in}}\) the inflow into each cell, \(A\) the cell area and \(\Delta t\) the time step.
4 Methods

4.1.8 Bedrock seepage

Deep percolation or bedrock seepage may play a dominant role in runoff generation at the hillslope scale as discussed in Chapter 2.1.1 (Anderson et al. (1997), Scherrer et al. (2007), Lehmann et al. (2006)). HillVi calculates seepage into the bedrock based on the total head above bedrock according to

\[ B(t) = k_{s_{bed}}(1 + w(t)) \]  

(4.25)

where \( k_{s_{bed}} \) is the saturated hydraulic conductivity of the bedrock (\( \text{bedrock}_{-}k_{sat} \)) and \( w(t) \): time dependent water table height. If no saturated zone is present, seepage is limited by the recharge flux from the unsaturated zone (equation (4.14) or (4.15)). Bedrock exfiltration as described by Anderson et al. (1997) is a process that is not included in HillVi. Water seeping into the bedrock is assumed to be lost from the system.

4.1.9 Other routines

Weiler and McDonnell (2004) and McGuire et al. (2006) conducted transport tracer experiments and modelled solute mass flux with a mass flux routine included in HillVi. This functionality is not applied in this thesis. The model further includes a simple evapotranspiration algorithm, which was not used either due to the focus on modelling of short rainstorm events, where evapotranspiration has negligible influence on water distribution and runoff.

4.1.10 Model output

**Graphical output** An important feature of HillVi is the graphical output. Four dynamic windows show the bedrock topography with superposed information of water table elevation (saturation depth), unsaturated water content, relative flow and pipe flow. Figure 4.3 shows an example output which explains how to interpret the graphs later in Chapter 5. The basic structure shown is the same for all four animated windows, only the layer information on bedrock topography and the x-y plots in the upper left corner of the animated windows are different (see Chapter 5). Table 4.1 specifies the information content displayed in the different graphical outputs.

**Numerical Output** HillVi generates ASCII files of all calculations done by the model. Due to the availability of the source code of HillVi, the numerical output has been customized for the modelling exercises in this thesis. Examples of data outputs are total matrix or pipe runoff and matrix or pipe runoff at each trench cell, water table depths, water content for the unsaturated storage, seepage (all per cell and time step) and maximum daily water table depths of each cell.
4.2 Parameter sampling

4.2.1 Monte Carlo and Latin hypercube sampling

Monte Carlo (MC) sampling is a computational algorithm which relies on repeated random sampling of input values to compute model outputs. Physicist developed this sampling method in the early 1940's. Physical or mathematical systems never describe "truth", therefore several sources of uncertainty are inherent. MC sampling offers a way to assess this uncertainty in modelling by considering not one deterministic model run but several with randomly sampled input parameters. HillVi includes a MC sampling method, where lower and upper limits of a parameter value and the number of simulation runs have to be defined. Subsequently, for each simulation run, a parameter set is sampled, based on uniform distributions of all input parameters within a given range. A shortcoming of the random MC method is that in a multiple parameter environment many MC simulations are required to cover the whole parameter space. Sieber and Uhlenbrook (2005) state that the "classical" MC approach is impos-
sible to apply to complex models with a large number of model parameters due to the high necessary number of model runs for representative results. Even if HillVi is, as stated earlier, a simple model with a "few tunable parameters", the complexity is too high for MC analysis, if all parameters are randomly sampled (about 15 parameters that are meaningful). To face this problem McKay et al. (1979) developed the "Latin hypercube sampling" method (LHS). LHS has been widely used in environmental modeling as basis for sensitivity analysis (Sieber and Uhlenbrook (2005) according to Hofer (1999) and Helton (1999)). However, there are only few examples in hydrological modelling, where LHS has been used so far (Sieber and Uhlenbrook (2005), Christiaens and Feyen (2002)).

LHS divides the range of each parameter into $n$ intervals and randomly samples a parameter value within this interval. The size of these intervals represents the underlying probability of a chosen distribution. Each parameter value is then combined randomly with other parameters to provide the input parameter set for the simulation, where it is indispensable that all the parameter ranges are divided into the same number of intervals $n$. After one parameter interval being sampled, it is discarded and the remaining will be sampled in further simulation runs. This results in $n$ parameter sets and the same number of model runs. The method ensures, that each parameter range is considered. To determine the necessary numbers of model runs Christiaens and Feyen (2002) suggested that $n$ should be at least between two and five times the number of varied parameters $p$. In their study Sieber and Uhlenbrook (2005) compared several thousand random MC with Latin hypercube sampled model runs and found that both sampling strategies yield similar statistical measurements if $n$ is about ten times the number of varied parameters; the number of intervals $n$ equals $10 \times p$ with the iteration step of $1/n$. $n$ is defined in the main steering file of HillVi as number of Monte Carlo simulation runs. As all the parameter ranges in HillVi are sampled based on a uniform distribution, all LHS intervals have equal lengths. For this diploma study, the LHS algorithm has been therefore implemented into HillVi (HillVi V44.1_MC_LHS).
4.3 Sensitivity analysis and parameters

According to Saltelli et al. (2004), sensitivity analysis (SA) is “the study of how the uncertainty in the output of a model (numerical or other) can be apportioned to different sources of uncertainty in the model input”. First of all, the response sensitivity of model output to the diverse inputs is of interest. Then questions about magnitude, direction and correlation of influence of these parameters arise. The achieved knowledge about model behaviour is then used to pursue as follows:

- when important factors are identified, the model structure may be simplified by removing parts that seem to be irrelevant
- if factors are shown to be influential which were initially thought to be relatively irrelevant (or vice versa) the model structure may be revised
- estimates of the factors may be further improved and optimized to increase the accuracy of model predictions

(after Saltelli et al. (2004))

As an outcome of the sensitivity analysis a list of ranked factors (parameters) can be generated according to their relative importance. Sensitive parameters can be distinguished from insensitive ones. The latter may be then held constant to reduce model dimensionality and to proceed with the calibration or the main modelling exercise. It is crucial to note, that sensitivity analysis is uniquely an investigation of the model behaviour in itself and no measured data are necessary for this task (Reichert, 2005). A variety of sensitivity analysis techniques exist, but so far no common classification is present in literature that we know of. The following paragraphs summarize briefly two distinct method families.

**Local methods** Local methods investigate the effect on model output by variation of single parameters of a reference parameter set, which represents a single point in the parameter space. This method is appropriate, when knowledge about parameter values are available. Saltelli et al. (2004) formulate the local method as an equation $S_j = \frac{\delta Y}{\delta X_j}$, where $S$ is the local sensitivity, $Y$ is the output of interest and $X_j$ an input parameter. The sensitivity measure is usually normalized by the reference value, multiplying the right side of this equation with $X_j/Y$. The shortcoming of this method is that sensitivity can only be analysed at one point in the parameter space but lacks information about the interaction of other parameters and the significance of parameter changes (Blasone, 2007).

**Regional or global methods** Global or regional methods consider the effect of the whole input factor space on model output. Some authors as Saltelli et al. (2004) name this type sampling methods global, because the whole possible parameter space should be sampled. However, in reality estimates of parameters or at least a range of parametrization is given, therefore an other group of authors (see Reichert
Methods

(2005) underline the regional character of such sampling methods. This thesis agrees on the latter definition.

Regional methods are often based on MC sampling of the factor space, the reason why these sampling methods are described in Chapter 4.2.1 in detail. Again, several methods exist, as summarized briefly in Blasone (2007). For this work, namely scatter plots and the regression technique are applied using standardized regression coefficients (SRC) described later. So far, regression techniques have not been widely applied in hydrology (Sieber and Uhlenbrook, 2005). Sieber and Uhlenbrook (2005) and Christiaens and Feyen (2002) presented a possible approach to perform a sensitivity analysis with a multiple linear regression meta model. This meta model is applied also in this thesis and is of the form:

\[ y_v(s) = \hat{\beta}_{0,v} + \sum_{i=1}^{p} \hat{\beta}_{i,v}x_i(s) + \hat{\epsilon}_v(s) = \tilde{y}_v(s) + \tilde{\epsilon}_v(s) \]  

(4.26)

where \( s \) is the number of simulation runs, \( v \) the number of studied output variables, \( p \) the number of parameters, \( \hat{\beta}_{0,v} \) and \( \hat{\beta}_{i,v} \) the regression coefficients and \( \hat{\epsilon}_v(s) \) regression residual. \( \beta 's \) in equation (4.26) are ordinary regression coefficients (ORC) and are absolute sensitivity measures that indicate how much \( y \) will change if \( x_i \) changes. To allow for comparison between input parameters standardized regression coefficients have to be calculated according to:

\[ SRC = \hat{\beta}_{i}^{(s)} = \hat{\beta}_{i} \frac{s_{x_i}}{s_{y}} \]  

(4.27)

SRCs quantify how much the output of the meta model \( Y \) will change relative to its standard deviation \( (s_{y}) \), when \( x_i \) is changing with its standard deviation \( (s_{x_i}) \). The meta model for the multiple linear regression analysis in this thesis was set up as follows:

\[
\begin{bmatrix}
Y_1(t) \\
\vdots \\
Y_s(t)
\end{bmatrix} =
\begin{bmatrix}
1 & x_1^1 & \cdots & x_p^1 \\
1 & \vdots & \ddots & \vdots \\
1 & x_1^s & \cdots & x_p^s
\end{bmatrix} \begin{bmatrix}
\hat{\beta}_0 \\
\vdots \\
\hat{\beta}_p
\end{bmatrix} + 
\begin{bmatrix}
\hat{\epsilon}_1 \\
\vdots \\
\hat{\epsilon}_s
\end{bmatrix}
\]  

(4.28)

where \( Y \) is the model output, \( x \) the parameter, \( \hat{\beta} \) the regression coefficient and \( \hat{\epsilon} \) the regression residual, \( s \) the number of simulation runs and \( p \) the number of investigated parameters. As the left side of equation (4.28) shows, \( Y \) is a function of time and therefore the meta model has to be calculated for every time step, applying the ordinary least square (OLS) method.

When applying regression analysis, model accuracy and requirements should always be tested and fulfilled beforehand. These requirements are:

1. Residuals should scatter around the regression line without indicating major trends
2. Standardized residuals should be normally distributed
3. Residual variance should be equal over the whole prediction range
4. Regression coefficients should not be mainly influenced by single influential cases
4.3 Sensitivity analysis and parameters

To follow this procedure, six summary graphs presented later in Chapter 5 served as basis to check if the requirements 1-4 from the list above are fulfilled.

1. **Tukey-Anscomb (residuals vs fitted):** input residuals are plotted against the predicted (fitted) residuals. No systematic trends should emerge from this plot.

2. **Normal Q-Q:** standardized input residuals are plotted against the residuals theoretical quantiles of the regression model outputs. All scatter points should lie close or on a straight line to fulfil the requirements of a normal distribution.

3. **Scale Location:** the square root of the standardized input residuals are plotted against the predicted (fitted) residuals. This plot demonstrates, if the condition of variance homogeneity of the residuals is given.

4. **Cook’s distance:** informs how much one case (residuum) influences the regression coefficients of the meta model. The equation is:

   \[ D_i = \frac{\sum (\hat{y} - \hat{y}_i)^2}{(k + 1)\sigma_{\text{residual}}^2} \]  

   (4.29)

   Cases are classified as influential if they exceed the value of the F-statistic at \( \alpha = 0.5 \) with \( (k+1) \) and \( (n-k-1) \) degrees of freedom (DF).

5. **Residuals vs Leverage:** measures the extremity of cases regarding the independent variables. Large values are classified if leverage exceeds \( 2(k+1)/n \).

6. **Cook’s distance vs Leverage:** plots the Cook’s distance as a measure of influence on the regression coefficients versus the leverage, which is measure for the extremity of cases regarding the independent variables.

### 4.3.1 Parameters

The input parameters for all the equations described in Section 4.1 are collected in Table 4.2. The Table shows upper and lower limits of the parameters and parameter estimates from field data. The value ranges represent the regional character of this sensitivity analysis because they cover not the entire possible parameter space, but relying on ranges derived from field measurements or that are indicated in literature related to HillVi and sample model runs. Parametrizations can be found e.g. in McGuire et al. (2006), Weiler (2004, 2005, 2006, 2007, 2007a). Eleven parameters have been sampled for this sensitivity analysis. Refer to Table 4.2 for a description of these parameters. Parameters investigated in the sensitivity analysis are gray coloured in Table 4.2. These are namely \( c, b, m, ko, wc\_init, wattab\_init, thres, p\_const, infk, \text{bypass} \) and \text{bedrock\_ksat}. Other parameters such as \( p\_density, p\_height, p\_stdv, n \) and \( n_0 \) are kept constant. The reason for this selection is the loss of "goodness" of the multiple regression model the more parameters are included in the meta model. Therefore, "internal" parameters that are more difficult to measure or interpret from field or laboratory data have been evaluated in the regression model.
Total ($n$) and drainable ($n_0$) porosities are indisputably important input factors for the model because they largely determine the water amount available in the soil. If ever possible, these values should be defined from field or laboratory measurements. Compared to other parameters sampled in the sensitivity analysis, $n$ and $n_0$ are relatively easy to quantify. To account also for the sensitivity of drainable and total porosity, Section 5.3 presents a local sensitivity analysis of these two parameters, by holding the others constant. The ranges of values of parameters $b$ and $m$ are within the ranges cited in literature (see above for complete list of references). Values for the recharge parameter ($c$) were indicated higher in literature with ranges up to 120 in some papers. However, for the modelling setup of this thesis, no meaningful simulations could be produced with values greater than 10. $w_{c\text{ init}}$ and $w_{\text{attab init}}$ have been chosen to study in the sensitivity analysis to assess the influence on model runs particularly at the beginning of simulations and the temporal development. Test model runs showed that these parameters have a big influence especially in the first simulation time steps. Otherwise, these parameters, as in the case of $n$ and $n_0$, are relatively easy to quantify.

Laboratory definition of $k_o$ by Thielen (2007) excluded the influence of preferential flow. The measured values are therefore very low and could not be used for the simulations in the following chapter. Based on literature data on $k_o$, a wide range in-between two order of magnitudes (0.005-0.5 m/h) has been specified for sampling based model parametrization.

$infk$ data were not available. The chosen range lies between 0.01-0.1 m/h. Sprinkling experiments at Tössegg with an intensity of 60 mm/h did not produce overland flow, therefore the infiltration capacity was not exceeded. This rate lies in the middle of the defined sampling range. In this particular modelling context with HillVi, the upper limit of the range is of minor importance, because if the rate is not exceeded by the rainfall, the parameter has no influence on the water flow in the model.

The range of $bedrock_{ksat}$ was defined from literature data and single model runs with HillVi. Higher values as the indicated produced no subsurface flow any more.

Parameters related to pipe geometry are highly empirical as described in Section 4.1.3 and were not per se quantifiable from the sprinkling experiments. As discussed in Section 3.1.2 pipes were found in the whole profile. Therefore, a mean height above bedrock of 1 meter was defined for the modelling exercises and a large standard deviation of 0.5 meter was assumed to generate soil pipes within the whole soil profile. In Section 5.5 the influence of pipe flow based on pipe height above bedrock and pipe density will be further discussed.
<table>
<thead>
<tr>
<th>Parameter (Unit)</th>
<th>Name</th>
<th>Value</th>
<th>Sampling method</th>
<th>Field estimates</th>
</tr>
</thead>
<tbody>
<tr>
<td>$n$ (-)</td>
<td>total porosity</td>
<td>0.4 - 0.4</td>
<td>uniform</td>
<td>0.3</td>
</tr>
<tr>
<td>$n_o$ (-)</td>
<td>drainable porosity</td>
<td>0.1 - 0.1</td>
<td>uniform</td>
<td>0.07</td>
</tr>
<tr>
<td>$b$ (-)</td>
<td>shape parameter for exponential depth function of drainable porosity</td>
<td>0 - 10</td>
<td>uniform</td>
<td></td>
</tr>
<tr>
<td>$k_o$ (m/h)</td>
<td>saturated hydraulic conductivity at soil surface</td>
<td>0.005 - 0.5</td>
<td>uniform</td>
<td>0.02</td>
</tr>
<tr>
<td>$m$ (-)</td>
<td>shape parameter for conductivity depth function</td>
<td>0 - 10</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$c$ (-)</td>
<td>shape parameter for recharge function</td>
<td>0 - 7</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>bypass (−)</td>
<td>bypass flow exponent</td>
<td>0 - 10</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$thres$ (-)</td>
<td>constant fraction for bypass flow (fraction model equation (4.18))</td>
<td>0.1 - 0.3</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$wu_{\text{init}}$ (θ)</td>
<td>initial water content</td>
<td>0.1 - 0.2</td>
<td>uniform</td>
<td>0.3</td>
</tr>
<tr>
<td>$wat_tab_init$ (m)</td>
<td>initial water table height</td>
<td>0.02 - 0.2</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$p_density$ (m/m$^2$)</td>
<td>pipe density</td>
<td>0.7 - 0.7</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$p_const$</td>
<td>empirical conductivity factor for pipe flow initiation</td>
<td>0 - 7</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$p_stdv$ (m)</td>
<td>standard deviation of height of pipe start above bedrock</td>
<td>0.5 - 0.5</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$p_exp$</td>
<td>exponential factor for the pipe flow equation (4.11). Should be close to 0.4 (Sidle and Kitahara, 1995)</td>
<td>0.4 - 0.4</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$infk$ (m/h)</td>
<td>max infiltration rate into soil matrix</td>
<td>0.01 - 0.1</td>
<td>uniform</td>
<td>no obs</td>
</tr>
<tr>
<td>$bedrock_ksat$ (m/h)</td>
<td>hydraulic conductivity of bedrock</td>
<td>0.0 - 0.0006</td>
<td>uniform</td>
<td>no obs</td>
</tr>
</tbody>
</table>

Table 4.2: Parametrization for the sensitivity analysis
Chapter 5

Results

As the focus of this thesis lies on hydrological extreme events, a time series analysis of MP, TDR, tensiometer and precipitation data was performed in order to find storm events for the following modelling exercises. The criteria for selection of the in Section 5.1 discussed storm events were data availability (time series of some soil instruments were not continuous), either high precipitation amounts or intensities or both and clearly visible response of the soil instruments to the rainfall input. The temporal resolution is 10 minutes for all data. The modelling time step was set accordingly.

5.1 Event analysis

5.1.1 July 2005

According to MeteoSchweiz (2005b), July 2005 was "capricious - with locally heavy rainstorms". Total rain amount was 166.4 mm with a maximum intensity of 19.8 mm/h (July 18). Between cold air from the north sea and warm air from south eastern Europe, a heavy convective rainstorm occurred from July 24 to 25 (72h). On July 25, intensities of up to 10.4 mm/10 min were registered. The total rain amount of this event was 36.1 mm. The precipitation patterns are plotted in Figures 5.1 from the top of the graph. Precipitation data from the Zürich-Kloten station are plotted in black, field precipitation data is green coloured. Stratiform rainstorm events may only produce minor differences in precipitation patterns between the data from station Zürich-Kloten and the field site, but local convective (summer) storms may only occur at one of the sites. Therefore, even though the amounts of the field measurements were lower due to mentioned measurement problems, they are plotted in Figures 5.1 and 5.2 in green to account for local precipitation patterns not present at Zürich-Kloten.

At first glance, the graphs in Figure 5.1 show only minor overall variations over time in either water content measured by TDR and MP or soil suction measured by tensiometers ("T" in figure legends) except the top most MP devices. The increase in volumetric water content for MP device in the upper left plot during the rain event on 25 July is about 15%. Consider that this MP device segment lies
5 Results

Figure 5.1: MP, TDR and tensiometer curves for all instrumented soil depth profiles from 17 to 29 July, 2005

only 7.5 cm below soil surface. Other top situated MP devices (15 cm below soil surface) also register an increase in volumetric water content, but only about 5 to 10%. This is in agreement to rough calculations of the increase in volumetric water content by precipitation input. Water content should increase by 7-13% (taking soil depth of the four plots into account). All other devices show only minor response to rainfall events. This leads to a possible conclusion, that only the uppermost 10 to 20 cm of the soils have been partially saturated during this summer event and water did not percolate further. Water was then lost to the atmosphere or through biological activity by evapotraspiration. This could be in agreement to the ERT measurements, which reveal a highly rooted top soil layer zone. Another possible explanation could be, that water percolating further than some 20 cm into the soil flows highly preferential, as seen in Figure 3.6 (b) and (c) and Figure 3.7 (a) bypassing the instruments.

5.1.2 April 2006

According to MeteoSchweiz (2006a), April 2006 was "capricious and mild - wet in the north and dry in the south". Total monthly rain amount was 176.9 mm with a maximum intensity of 4 mm/h. On 9 April,
5.1 Event analysis

Figure 5.2: TDR and tensiometer curves for all instrumented soil depth profiles from 2 to 14 April, 2006

station Zurich-Kloten measured a 24 hour sum of 51 mm. Main precipitation event was 9 to 11 April, 2006 with a total duration of 72 hours and a total rain amount of 81.5 mm and a maximum intensity of 1.2 mm/10 min. Precipitation, TDR and tensiometer (T) data are plotted in Figures 5.2 for April 2. to 14 2006. For this event, no MP data are available. Analysis of the TDR and tensiometer (“T”) curves in Figure 5.2 clearly distinguish between the July 2005 and April 2006 event. Periodic daily oscillation, as also present in Figure 5.1 was corrected manually in order to obtain a smooth curve. (Oscillations started from a clearly identifiable base level. Points, where the oscillation started and where it reached again this base level have been connected). All the four plots show an adequate response to rainfall input. Increasing temporal delay in response to the precipitation input with depth can be found. Instruments at the top soil show a decrease in suction and an increase in volumetric water content very quickly, whereas instruments deeper in the soil show a delayed response. Tensiometers and TDRs installed at the same soil depth react mostly simultaneous to a change in water content. The delayed response of the lower instrument depends on multiple factors and is spatially very heterogeneous. Tensiometer 7 (upper left plot, 90 cm) reacts about 15 hours after the precipitation input, whereas TDR 10 (lower right plot, 150 cm) signals an increase in volumetric water content already after 12 hours. Tensiometer 3 (lower left plot, 90 cm) at the same depth as tensiometer 7, shows a decrease in suction 8 hours after
5 Results

precipitation input. This exemplary event shows, that under wet winter half term conditions, water may percolate to the very bottom of the soil profile (e.g. TDR5 (upper right), TDR10 (lower right)). The roughly calculated increase in volumetric water content due to the precipitation should be around 10 to 25%. Again, as compared to Section 5.1.2, the calculated increase is much higher than the observed increase which lies in between 5 to 8%. The possibility of bypassing the probe by preferential flow or other losses or distribution of soil water may explain the difference.

Conclusions  Under dry summer conditions, rainstorms result in a saturation of only the upper parts of the soil and water does not percolate further into the soil (according to the data). Suction is therefore increased and contributes additionally to slope stability (see Section 2.3). Thielen (2007) found, that during the summer period, when suction was highest the factor of safety was highest too.

Rainfall amounts and intensities for the investigated months, the period for simulation with HillVi and the main precipitation events are indicated in Table 5.1. The April 2005 event in Table 5.1 served as model validation storm. Precipitation patterns and average soil water content can be seen in Figure 5.5.

The origin of the periodic oscillations of some instruments as seen in Figure 5.1 (smoothed for the April 2006 event) could not be explained on a physical basis. Pressure conditions in the tensiometer tubes may be influenced by solar radiation an result in a diurnal variation. However, electrically driven MP and TDR devices were also affected. This behaviour is not explainable on the basis of a diurnal physical based variation, because instruments in a soil depth of 1.5 m were affected. In this depth, diurnal variations are relatively small. The reason for this periodic oscillations should be further investigated.

<table>
<thead>
<tr>
<th>Event Name</th>
<th>Month</th>
<th>Simulation Period</th>
<th>Main Events</th>
</tr>
</thead>
<tbody>
<tr>
<td>April 2005</td>
<td>01.-30.4</td>
<td>02.-14.4</td>
<td>07.-09.4</td>
</tr>
<tr>
<td>Amount</td>
<td>117.7 mm</td>
<td>50.7 mm</td>
<td>50.7 mm</td>
</tr>
<tr>
<td>Max. Intensity</td>
<td>1.2 mm/h</td>
<td>0.6 mm/10min</td>
<td>0.6 mm/10min</td>
</tr>
</tbody>
</table>

| July 2005 | 01.-31.7 | 17.-29.7 | 18.-19.7 | 24.-25.7 |
| Amount | 166.4 mm | 95.4 mm | 47.7 | 36.1 mm |
| Max. Intensity | 19.8 mm/h | 10.4 mm/10 min | 8.7 mm/10 min | 10.4 mm/10 min |

| April 2006 | 01.-30.4 | 02.-14.4 | 09.-11.4 |
| Amount | 176.9 mm | 114.6 mm | 81.5 mm |
| Max. Intensity | 4 mm/h | 1.2 mm/10 min | 1.2 mm/10 min |

Table 5.1: Rainfall characteristics for the selected periods
5.2 Sensitivity Analysis

The here presented sensitivity analysis is based on the two storm events described in Section 5.1. 100 Latin hypercube sampled model runs have been performed for each of the two events. The whole simulation period was 336 hours. The first 24 hours serve as warm up period for model initialization and have not been evaluated and plotted. Figures 5.3 and 5.4 present the graphical results of the analysis. Each sub-figure (b), (c) and (d) of Figures 5.3 and 5.4 display ensemble plots of 100 sampled model runs at the top and the process corresponding SRC graphs at the bottom. Sub-figures (b) show the development of the unsaturated water storage [mm], (c) the saturated water content [mm] and (d) the subsurface flow hydrograph with median in red and the interquartile range highlighted in colour.

A summary graph of different model outputs was plotted for each of the 100 model runs. Based on these graphs, numerical unstable runs have been rejected. The remaining runs have then been processed by the regression analysis. 44 runs of the July 2005 event were unstable and 19 for the April 2006 event.

As mentioned in Section 4.3, regression diagnostics of the meta model have been conducted before proceeding to the SRC evaluation. This is done by analysis of the residuals to fulfil the four points indicated on page 36. It is hardly possible to evaluate all regression models (for every time step), therefore the diagnostics have been performed for several randomly selected time steps (for each model output a randomly selected time step is indicated in Tables 5.2 and 5.3). This procedure is justifiable, because the SRC curves are continuously differentiable (showing no jumps). The corresponding diagnostic plots can be found in Figures 5.3 (a), 5.4 (a) for the unsaturated storage and in A.7 for the saturated storage and the subsurface runoff. Description of the information content of each of these six plots is according to the list in Section 4.3 on page 36. The regression diagnostics for the unsaturated storage Figures 5.3 (a) and 5.4 (a) show, that in plot 1 no systematic trend is present. Plot 2 shows, that the residuals lie approximately on the given line for both events, which indicates normally distributed data and fulfils the requirement to proceed with the regression analysis. Model run 4, 52 and 56 (July 2005) and 73 (April 2006) have the biggest distance to the line. These four observations show also the biggest variance in plot 3 and their behaviour has to be considered in plots 4 to 6. For the other observations variance homogeneity is given.

The mentioned observation are also the ones with highest Cook’s distances (threshold values for Cook’s distance (see 4.3) have been calculated and indicated in Table 5.2 and 5.3 for the selected time step). Accordingly both plots 5 and 6 indicate the influence of the mentioned observations. However, this influence does not exceed the calculated thresholds. Visual analysis of the graphical outputs of these model runs does not show instabilities or abnormal results. On the basis of this diagnostics, also the discussed influential observations have been kept. The regression meta model fulfils the requirements and further analyses are justified. Regarding the saturated storage and runoff, the Tukey-Anscomb plots showed a small systematic trend which could not have been improved by adaptation measures to the regression models done within this thesis (see Figure A.7 Tukey-Anscomb plot). Also the normal Q-Q plots show, that the data is not optimally normally distributed. Further improvement by applying non-linear models for the sensitivity analysis are suggested here.
5 Results

5.2.1 Parameter discussion

c  The exponent \( c \) of the recharge function (equation (4.14)) is the most influential factor of all the sampled parameters in both the July 2005 and April 2006 event and on all three investigated model outputs (water storage in the saturated and unsaturated zone [mm] and subsurface runoff [mm/10 min]). If equation (4.14) is considered, we see that an increase of \( c \) results in a decrease of its quotient \( \theta / \theta_s \) and therefore the recharge rate too. SRC of \( c \) is highly positive for the water content in the unsaturated zone, allocating water to this storage with each increase relative to its standard deviation and highly negative for the "loss" model processes "saturated storage" and "subsurface runoff". SRC time series of \( c \) show, that its importance is higher immediately after a rainfall event and decreases in the hydrograph recession phase.

bedrock_ksat  The second highest ranking parameter is bedrock_ksat. bedrock_ksat is a parameter in equation (4.25) (saturated hydraulic conductivity of bedrock). An increase in this parameter results in a faster loss of water in the saturated domain and influences therefore the saturated water storage and the subsurface flow negatively. Consequently, this parameter has a priori no influence on the unsaturated water content and ranks therefore 10th, respectively 11th with values very close to 0 (a minor influence is given by the fact, that the saturated storage defines the water table (which is strongly influenced by bedrock_ksat) and therefore the extent of the unsaturated zone). SRC time series of this factor show a decrease of the absolute SRC value during a rain storm event and an increase in the recession phase of the hydrograph. This behaviour represents the fact, that during a storm phase, recharge, runoff and other directly storm related processes are dominant and that bedrock_ksat influences and dominates mainly the recession or inter-storm phases.

ko  Saturated hydraulic conductivity at soil surface (ko) is among a group of parameters (with m, b, bypass, thres, wc_init and wattab_init), which significantly influence the meta model, but are not strong enough to show the same behaviour for all the investigated events and processes. Standardized regression coefficients of ko are negative for the unsaturated storage, but only significant for the July 2005 event (on the 95% level), which is indeed discussible, if the straight run of the graph around -0.1 is considered. According to equation (4.14) recharge increases with increasing ko, explaining the negative value for the July 2005 event. For the saturated storage, ko ranks third in both events and on a 95% significance level for the meta model. With a negative value of around -0.2, its influence is quite strong and indicates parametrization of a "loss" process, which is the lateral flow discussed next for the subsurface runoff process. ko further parametrizes the lateral flow equation (4.3) resulting in a linear increase of flow, if its value is raised. This is particularly true during a rain event, where the water table rises and more water is available for lateral flow. This is reflected by the similarity of the SRC curve progression and the (runoff) hydrograph. For the meta model no significance for ko is given for the July 2005 event and for April 2006 only on the (very low) 80% level. Although the sensitivity of this parameter
5.2 Sensitivity Analysis

Concerning the runoff process seems to be reasonable, the influence should be further evaluated.

\textbf{m} The shape factor for the exponential depth function of saturated hydraulic conductivity parametrizes equations (4.3) and (4.14). Low values of \( m \) lead to a sharp decline with depth, \( k(z) \) values quickly reduce from \( k_0 \) to zero. With increasing \( m \) the curve becomes linear, approximating the defined \( k_0 \) as a constant with depth (\( k(z)=k_0 \)) (refer to Figure A.6 to see graphs visualizing this behaviour). \( m \)'s influence on the unsaturated storage is about the same for both events with -0.14 respectively -0.19. An increase in \( m \) results, as discussed above, in higher \( k(z) \)'s for each depth, therefore more water will recharge to the saturated zone. Against expectation, SRC values do not support this behaviour (with values of -0.06 and 0.1 and no significance for the meta model). As \( m \) influences the transmissivity for subsurface runoff, higher values should increase the lateral runoff. However, \( m \) is not significant for the July 2005 event, but for April 2006 with a value of 0.16. Once again, SRCs are higher during the storm event due to water availability.

\textbf{b} Parameter \( b \), shape factor for the drainable porosity depth function, has to be discussed analogous to \( m \) due to the same mathematical origin (see equation (4.12) and A.6). Significance is given for all the outputs in April 2006 but only for the unsaturated storage for the July 2005 event. SRC values are all positive. This is not in agreement with Weiler and McDonnell (2004), where for example a higher drainable porosity (resulting e.g. from an increase in \( b \)) results in lower depths of saturation.

\textbf{bypass and thres} The parameters \( \text{bypass} \) and \( \text{thres} \) have to be discussed together because of their involvement in the same equation (4.18). In this sensitivity analysis, \( \text{thres} \) accounts for a fraction of precipitation directly allocated to the saturated zone, representing a fast vertical macropore flow process (refer to Table 4.2). Together with the power exponent \( \text{bypass} \) they define the bypass flow rate. Both parameters rank mostly in the same region of the output, being significant only together (except for April 2006 unsaturated storage, where only \( \text{bypass} \) is significant). For July 2005, significance is given only for the subsurface flow output with SRC values of -0.09 and 0.06 for \( \text{bypass} \) resp. \( \text{thres} \). The combination of negative and positive SRC, supports the fact, that an increase in \( \text{bypass} \) reduces the bypass flow rate but an increase in \( \text{thres} \) augments it. For the April 2006 event, this relation is not given. A possible reason could emerge, if we consider equation (4.18). \( \text{bypass} \), as the exponent of the bypass function, is the dominating parameter here, superposing the linear parameter \( \text{thres} \).

\textbf{wc_init} and \textbf{wattab_init} The influence on the model runs of the initial condition parameters \( \text{wc_init} \) and \( \text{wattab_init} \) are heterogeneous. For both the July 2005 and April 2006 event, \( \text{wc_init} \) is a significant and influential factor regarding the unsaturated storage. The same should be true for the saturated storage, where \( \text{wattab_init} \) is the pendant to \( \text{wc_init} \) in the unsaturated zone regarding the initial water content. The values indicate the positive correlation between SRC and saturation, but Table 5.2 and 5.3 do not indicate a significance. This is due to the randomly selected time step which is 1372 (ca. 230h)
## 5 Results

<table>
<thead>
<tr>
<th>Rank</th>
<th>Parameter</th>
<th>Storage unsaturated SRC</th>
<th>Storage saturated SRC</th>
<th>Subsurface runoff SRC</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>$c \bullet$</td>
<td>0.79</td>
<td>-0.58</td>
<td>-0.54</td>
</tr>
<tr>
<td>2</td>
<td>$b \bullet$</td>
<td>0.26</td>
<td>$bedrock_{ksat} \bullet$ -0.25</td>
<td>$bedrock_{ksat} \bullet$ -0.44</td>
</tr>
<tr>
<td>3</td>
<td>$wc_{init} \bullet$</td>
<td>0.23</td>
<td>$ko \bullet$ -0.21</td>
<td>$b \bullet$ 0.21</td>
</tr>
<tr>
<td>4</td>
<td>$m \bullet$</td>
<td>-0.14</td>
<td>$bypass \bullet$ 0.16</td>
<td>$ko \bullet$ 0.12</td>
</tr>
<tr>
<td>5</td>
<td>$ko \bullet$</td>
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<td>$wattab_{init} \bullet$ 0.13</td>
<td>$m \bullet$ 0.13</td>
</tr>
<tr>
<td>6</td>
<td>$thres$</td>
<td>-0.08</td>
<td>$bypass \bullet$ -0.09</td>
<td>$ko \bullet$ 0.12</td>
</tr>
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<td>7</td>
<td>$wattab_{init}$</td>
<td>0.08</td>
<td>$wc_{init} \bullet$ 0.08</td>
<td>bypass $\bullet$ -0.09</td>
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<tr>
<td>8</td>
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<td>infk $\bullet$ -0.08</td>
<td>$thres \bullet$ 0.06</td>
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<td>9</td>
<td>bypass</td>
<td>-0.07</td>
<td>$p_{const} \bullet$ 0.08</td>
<td>$p_{const} \bullet$ -0.04</td>
</tr>
<tr>
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<td>$bedrock_{ksat}$</td>
<td>-0.02</td>
<td>$m \bullet$ -0.06</td>
<td>$wc_{init} \bullet$ 0.02</td>
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<tr>
<td>11</td>
<td>infk</td>
<td>-0.0</td>
<td>$thres \bullet$ 0.2</td>
<td>infk $\bullet$ -0.01</td>
</tr>
</tbody>
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### Statistics for the variable selected regression models

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Random time step for diagnostics</th>
<th>$R^2$</th>
<th>F-statistic</th>
<th>Threshold for Cook's distance</th>
<th>Threshold for Leverage</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>1672</td>
<td>0.71</td>
<td>24.03 on 5 and 50 DF</td>
<td>0.91</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1372</td>
<td>0.43</td>
<td>13.07 on 3 and 52 DF</td>
<td>0.85</td>
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<tr>
<td></td>
<td></td>
<td>924</td>
<td>0.57</td>
<td>17.17 on 4 and 51 DF</td>
<td>0.88</td>
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Parameters on 95% significance level ($\bullet$) in meta model

---

Table 5.2: Sensitivity analysis parameter ranking based on absolute values (mean SRC over time), 17 to 29 July, 2005

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<tr>
<th>Rank</th>
<th>Parameter</th>
<th>Storage unsaturated SRC</th>
<th>Storage saturated SRC</th>
<th>Subsurface runoff SRC</th>
</tr>
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<tr>
<td>1</td>
<td>$c \bullet$</td>
<td>0.73</td>
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<td>-0.55</td>
</tr>
<tr>
<td>2</td>
<td>$b \bullet$</td>
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<td>$bedrock_{ksat} \bullet$ -0.32</td>
</tr>
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<td>3</td>
<td>$wattab_{init} \bullet$</td>
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<td>$ko \bullet$ 0.18</td>
</tr>
<tr>
<td>4</td>
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<td>$bypass \bullet$ -0.14</td>
<td>$m \bullet$ 0.16</td>
</tr>
<tr>
<td>5</td>
<td>$thres$</td>
<td>0.14</td>
<td>$thres \bullet$ -0.14</td>
<td>$wc_{init} \bullet$ 0.16</td>
</tr>
<tr>
<td>6</td>
<td>$wc_{init} \bullet$</td>
<td>0.1</td>
<td>$p_{const} \bullet$ 0.11</td>
<td>bypass $\bullet$ -0.15</td>
</tr>
<tr>
<td>7</td>
<td>bypass $\bullet$</td>
<td>0.09</td>
<td>$m \bullet$ 0.1</td>
<td>$thres \bullet$ -0.15</td>
</tr>
<tr>
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<td>$b \bullet$ 0.1</td>
<td>$b \bullet$ 0.11</td>
</tr>
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<td>9</td>
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</tr>
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<td>11</td>
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<td>0.0</td>
<td>infk $\bullet$ -0.0</td>
<td>infk $\bullet$ 0.03</td>
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### Statistics for the variable selected regression models

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<td></td>
<td>567</td>
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<td>9.256 on 9 and 71 DF</td>
<td>0.94</td>
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Parameters on 95% significance level ($\bullet$) in meta model
Parameters on 80% significance level (*) in meta model

---

Table 5.3: Sensitivity analysis parameter ranking based on absolute values (mean SRC over time), 2 to 14 April, 2006
5.2 Sensitivity Analysis

resp. 243 (ca. 41h): with progressing time, the influence of this parameter decreases and indeed the statistical analysis of the regression model for an early timestep reveals highly significant values for the parameter. The ensemble plots at the top of Figures 5.3 and 5.4 show how important parametrization of initial conditions is. All six plots depend highly on their initial conditions: the unsaturated storage amount seems to remain stable over time (Figure 5.3 and 5.4 (b)); the saturated storage and the subsurface flow are initially highly affected by the initial water table height ($w_{atb \_init}$) but are overlaid by other processes with progressing time.

**infk** The infiltration capacity has no significance in the variable selected meta model and ranks low in all processes. This could be due to too high sampled values (between 0.01 and 0.1 (m/h)). If $infk$ is higher than rainfall input, no overland flow is produced and all water infiltrates. Consequently and in agreement with the sprinkling experiments, where no HOF was observed (Kienzler 2007, personal communication), no overland flow is produced and justifies the process parametrization. SRC time series of $infk$ are approximately straight lines close to zero, showing minor amplitudes in the July 2005 event.

**p\_const** Parameter for equation (4.11) is not significant for all outputs and events, which is in agreement with the SRC time series plot. $p\_const$ represents a nearby straight line around 0 in all SRC graphs. If we consider that pipe height was defined 1 meter above bedrock with a standard deviation of 0.5 meter no pipe flow occurred because pipes did not lie in the saturated zone.
5 Results

(a) Statistics of the meta model for unsaturated storage at time step 1672 (ca. 279h)

(b) Unsaturated storage (with IQ-range and median) and SRCs

(c) Saturated storage (with IQ-range and median) and SRCs

(d) Subsurface runoff (with IQ-range and median) and SRCs

Figure 5.3: Sensitivity analysis model outputs (top) and corresponding SRC’s (bottom) July 2005 event
5.2 Sensitivity Analysis

(a) Statistics of the meta model for unsaturated storage at time step 719 (ca. 120h)

(b) Unsaturated storage (with IQ-range and median) and SRCs

(c) Saturated storage (with IQ-range and median) and SRCs

(d) Subsurface runoff (with IQ-range and median) and SRCs

Figure 5.4: Sensitivity analysis model outputs (top) and corresponding SRC’s (bottom) April 2006 event
5 Results

5.3 Comparing simulation and field data

To describe natural systems with numerical models, an adequate parametrization needs to be found to represent measured data in a first step. For a hydrological model, this means, that simulated and observed data of a particular time period, mostly discharge data of a catchment, are compared using statistical or visual methods. Step by step, model performance is optimized by adjusting the parameter input set to fit the simulated results optimally to the observed. A second modelling task then leads to model validation, where a different time period is modelled and compared to the measured data to check whether the "calibrated" parameter set produces similar "goodness-of-fit" measures for other storm events. As no subsurface runoff data was available for the hillslope investigated in this thesis, only soil moisture data can be used to compare modelling results with real field measurements. Thielen (2007) remarks, that field moisture data from MP and TDR probes were not quantitatively comparable, but qualitatively. Model output performance is therefore not explicitly judged based on absolute soil moisture values, but on temporal variation and change in water content. These shortcomings, the non-availability of subsurface runoff data and the only qualitative comparable data, lead to the conclusion, that a classical "calibration-validation" approach is not suitable in this context. Instead the term "comparison" of simulated and measured data was used and the "calibrated" parameter set is referred to be a "base-calibration" set. To compare modelled and measured soil moisture content, the mean field data value for each time step of all four instrumented soil depth profiles at the depths of 45 and 60 cm has been calculated. Because of the discussed spatial heterogeneity and uncertainty in observed soil moisture, this approach seems to be the most justifiable as a basis for a model comparison: the depth range between 45 and 60 cm seems to be a good approach because at this depth, the soil is in most cases not saturated and the depth is deep enough to not be mainly influenced by superficial evapotranspiration.

As a starting point for comparison of simulated and measured soil moisture content, model runs of the sensitivity analysis have been analysed. Parameter sets of runs that fit the observed temporal change in moisture content well have been collected and optimized in further model runs using the Nash-Sutcliffe efficiency measure (NSE):

\[
NSE = 1 - \frac{\sum_{i=1}^{N} (Y_i - \hat{Y}_i)^2}{\sum_{i=1}^{N} (Y_i - \bar{Y})^2}
\]

(5.1)

where \(Y_i\) is a observed data point, \(\hat{Y}_i\) a simulated data point and \(\bar{Y}\) the mean of observed data. Because HillVi does not explicitly calculate the water content in the unsaturated zone, the unsaturated water content [m] has been divided by the height of the unsaturated zone for each cell and time step [m] in order to calculate volumetric water content [m/m] (grid cell area is 1 m\(^2\)). To compare this curve to the field average moisture content, base water content of 0.285 [m/m] was assumed. This was done only to get a similar base water content as the field data and has no physical justification. Therefore the NSE value has to be interpreted as a performance indicator for the shape fit of simulated and measured water content for the temporal change and peak moisture representation.

According to the sensitivity analysis, high ranking parameters proved to be very influential for the pro-
5.3 Comparing simulation and field data

procedure to find a "base-calibration" set by fitting the moisture curve to the observed data. This underlines the importance of performing a sensitivity analysis before starting simulation tasks with the model. *bedrock_ksat* especially influences the recession limb of the hydrograph and the soil moisture curve, whereas *c* allocates the available water to the saturated and unsaturated zone.

**Model "validation" period April 2005**  As "validation" period, 2 to 15 April, 2005 has been chosen. TDR time series of this period showed clear responses, making this event particularly suitable for validation purposes. Total rainfall amount was 50.7 mm with a maximum intensity of 0.6 mm/10 min. The best fitting porosity values from the calibration phase have been taken for this validation. This "base-calibration" set is presented in Table 5.4 and the resulting graph is presented in Figure 5.5. The acquired NSE value is 0.78 (with the same base water content of 0.285 [m/m]). The results show, that for a rainstorm event with different character (both total amount and intensity are only half of the April 2006 event), the model performs equally well. The peak hydrograph is even performed better for the validation period.

![Figure 5.5: Model validation period April 2005](image)

This "base-calibration" parameter set results, as stated earlier, from visual analysis of the sensitivity analysis model runs successively optimized by adaptation of single parameters with the NSE performance measure. However, the restrictions concerning the parameter ranges discussed in Section 4.3.1 apply here too. To judge the appropriateness of single parameters, field measurements are indispensable, as will be discussed later in Chapter 6. This concerns mainly the two parameters *c* and *ko* as their influence on model runs is high (however, the ranges given for the sensitivity analysis are within the field estimates from the sprinkling experiments too). Other parameters are within ranges according to field estimates and literature.
Note Limiting the calibration exercise of the model to the measured water content curve in the unsaturated zone would result in the equifinality problem in this particular case with HillVi: as described earlier, the unsaturated zone is only linked to the saturated zone by recharge and is one of the first storages in the models storage chain (here, a "storage chain" is defined as a series of storages where water flows from one storage to another). Several parameter sets would produce "good fits" for water content curves, but may also influence other processes, e.g. related to the saturated zone, that are then not represented well. This behaviour would not be identifiable by evaluating only water content curves. Therefore it is urgent to calibrate the model at the end of a storage chain, as e.g. to runoff data and evaluate and validate the here discussed water content as so called internal model process.

5.3.1 Influence of soil porosity on model processes

Both input parameters total \( n \) and drainable porosity \( no \) have not been analysed in the sensitivity analysis for the reasons given in Section 4.3.1. With the "base-calibration" parameter set resulting from the task above, the influence of \( n \) and \( no \) can be studied by varying the values while holding the other parameters constant. The results of these model runs are presented in Figure 5.6. The porosity values are listed in the legend box. Plot 1 in Figure 5.6 shows ten modelled water content curves (mean over the the investigated area, delineated by the four measuring plots), plot 2 combined matrix and pipe flow [mm/10min] and plot 3 mean water table depth in cm above bedrock. Plot 4 displays the cumulative frequency distribution of water table height above bedrock for 240 to 264 simulation hours (day 11 to 12), where the highest water tables after the rainstorm developed. The shape of all water content curves in plot 1 are similar. Hydrograph recession is modelled well in terms of temporal decrease. However, the modelled increase in water content during the main event is too low. The main difference of the different parametrization lies in a vertical shift of the curves: curves with \( n=0.45 \) lie generally above curves with \( n=0.35 \) and increasing drainable porosity leads to a decrease in water content in the unsaturated zone. This behaviour becomes apparent by referring to equation (4.13) on page 30 and its term \( (n - no) \).

Combined matrix and pipe flow (plot 2) and saturation depth (plot 3) are inversely influenced by total and drainable porosity as is water content in the unsaturated zone (plot 1). Increase in total porosity results in lower water tables because more void space is available in the same unit soil element. Since Dupuit-Forchheimer assumptions are applied to calculate lateral flow, a lower water table results in

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
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<td>( n )</td>
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<td>( wc_{init} (\theta) )</td>
<td>0.12</td>
</tr>
<tr>
<td>( no )</td>
<td>0.1</td>
<td>( wattab_{init} (m) )</td>
<td>0.02</td>
</tr>
<tr>
<td>( b )</td>
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<td>( p_{density} (m/m^2) )</td>
<td>0.7</td>
</tr>
<tr>
<td>( ko (m/h) )</td>
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<td>( p_{height} (m) )</td>
<td>0.2</td>
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<tr>
<td>( m )</td>
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<td>( p_{const} )</td>
<td>2.6</td>
</tr>
<tr>
<td>( c )</td>
<td>3</td>
<td>( p_{stdv} (m) )</td>
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<td>( bypass )</td>
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<td>( infk (m/h) )</td>
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</tr>
<tr>
<td>( thres )</td>
<td>0.2</td>
<td>( bedrock_{ksat} (m/h) )</td>
<td>0.00016</td>
</tr>
</tbody>
</table>

Table 5.4: Calibrated parameter set
5.3 Comparing simulation and field data

Figure 5.6: Influence of total and drainable porosity on model outputs. Plot 1: water content curves, plot 2: combined matrix and pipe flow [mm/10min], plot 3: maximum saturation depth above bedrock [cm] and plot 4: cumulative frequency distribution of saturation depth above bedrock [%]
5 Results

decreased matrix and pipe flow. Counter wise, higher drainable porosities alter saturation depths and combined matrix and pipe flow. A lower total porosity produces wider ranges of observed flow especially at hydrograph peaks than soils with a higher total porosity. High drainable porosities but limited total porosities produce maximum water table depths and combined flow. This finding is in agreement to other studies, as e.g. Muklisina et al. (2006). Plot 4 shows that the differences in cumulative frequencies of maximum depths are quite small with a maximum range of about 2 cm. 50% of all cells have a water table depth of about 4 cm. Only 5% of all cells have a higher water table depth than 6 cm.

Note Especially from plot four emerges, that the differences between the different parametrization are extremely small regarding the depth scale of the whole soil profile. By referring to the sprinkling experiment, it becomes clear, that especially the outputs of the latter performed analysis are meaningless if we consider "real" conditions, because local heterogeneity is very high and both the lower and upper left plots did not even show a perched water table. However, analyses in this sense may reveal controls, how a natural system with high complexity responds to certain factors.
5.4 Spatial patterns of maximum saturation

A main objective of the diploma thesis is the identification of areas in the hillslope that are susceptible to high water table levels. The sensitivity analysis above revealed parameters that highly influence the amount of water in the saturated zone, therefore also the water table height above the soil-bedrock interface. Consequently, the question arises, if and how the different model parametrizations affect the spatial patterns of maximum water table depths. To approach this question, the simulation results from both the July 2005 and the April 2006 events with their 56 and 81 numerically stable simulation runs, respectively have been further analysed. The modelled storm period has been subdivided into daily intervals and maximum water table heights for each cell and each interval have been recorded. With these data, graphs showing the spatial patterns of maximum water table depths at given benchmark depths of 5, 10, 15 and 20 cm (and 25 cm for the April 2006 event) have been produced.

Figure 5.7: Percentage of total 56 model runs whose maximum water level at the soil-bedrock interface exceeded the datum given above the Figure (in cm) for the July 2005 event

**July 2005** The July 2005 event reached its maximum water table elevations between 120 and 144 simulation hours (Wednesday, 20th July). For this interval, spatial patterns of the percentage of model runs exceeding the given benchmark depth in Figure 5.7 are displayed. The benchmark depth is indicated on
the top of each Figure and the legend bar indicates percentages of all model runs exceeding the given level. The horizontal axis represents the hillslope base and the vertical the upslope direction in meters. As the DEM is not rectangular in shape, the boundaries of the hillslope are indicated. As discussed in Section 3.1.1 and visualized in Figure 3.4, the bedrock topography shows a depression approximately in the middle of the hillslope at (6/14) (coordinates x and y in meters). Not surprisingly, this depression shows frequent high water tables up to more than 20 cm, because water accumulates here. The upper part of the hillslope has generally low water tables because it is relatively flat compared to the lower part and has a high soil thickness (see Figure 3.5). This leads to limited water table heights due to lower recharge rates. Furthermore, bedrock seepage is elevated, because water flows generally slow in this flat part. The concentration of patterns at the upslope diagonal right boundary is a modelling artefact: HillVi adds cells with increased height to the DEM boundary - this prevents water from flowing sideways out of the model. The water is blocked at the boundary cell row and causing an increase water table at that location.

Regarding potential triggering locations (if the height of water table at particular locations is accepted to be a first order control for a triggering event), the main hillslope part of interest lies in the steeper downward zone. Interestingly, two areas at (11/6) and an area similar in size at (6/6) show frequent high water table levels. These patterns are not uniquely explainable on the basis of prevailing shallow soil depth at the corresponding areas. Moreover the lower left pattern at the 15 cm benchmark plot lies in a diagonal depression channel expanding from (1/2) to approximately (8/10), where water flows preferentially. Around (3/10), a bedrock elevation additionally supplies this channel with water.

**April 2006** The April 2006 event reached its maximum water table elevations at simulation day 11 (Monday, 11th April). The graph with the saturation patterns are plotted in Figure 5.8 according to the information content in Figure 5.7. For the same reason as for the July 2005 event, the upper part has only low water table levels for all parametrizations. The modelling artefact at the diagonal boundary is again present as is the water accumulation in the bedrock depression. The lower hillslope part shows very similar cumulative frequency patterns as during the July 2005 event. Especially the area around (11/6) is modelled as a critical zone for high water table levels, which was also found for the July 2005 event. From the third graph in Figure 5.8 on (> 15 cm benchmark) a grey homogeneous zone is identified as being particularly dominant in the > 25 cm Figure. Investigations showed, that a group of simulation runs (thus, a certain parametrization) produced this uniformly frequent high water tables. These runs are characterized by high initial water tables and very low values for the recharge parameter ($c$) close to 1, which results in high recharge rates. Further $ko$ had a low value and hindered the subsurface flow, which then results in a sustained water table development.

**Comparison of the two events** Comparing the spatial maximum saturation patterns for the two events reveals a generally high correlation. Thus, according to HillVi simulations, stratiform, long lasting rainstorm events do not produce significantly different saturation patterns than convective summer
Figure 5.8: Spatial patterns of percentage of total 82 model runs whose maximum water level at the soil-bedrock interface exceeded the datum given above the Figure (in cm) for the April 2006 event storm events. However, the water table heights are generally higher during the April 2006 event, which is due to antecedent wetness and higher rainfall amount. In addition to the bedrock depression zone, which shows frequent high water tables, two zones around (11/6) and (6/6) were identified with frequent high water table depths.

Conclusions The analysis of maximum saturation patterns is based on the same MCLHS simulation runs as the sensitivity analysis in Section 5.2, with parametrization given in Table 4.2. Analysis of the saturation patterns interestingly pointed to local areas susceptible to high maximum water tables. This shows a certain degree of parameter independency, where saturated areas frequently occur and leads to the hypothesis that first order controls concerning local saturation patterns are shallow soil depth and existence of bedrock "channels" combined with elevated supply zones. The latter characteristics are entirely defined, if a surface and a bedrock topography is available.

For both investigated events, July 2005 and April 2006, no parameter set produced water table depths of more than 30 cm. To get order of magnitude estimates if the simulated changes in water tables are realistic, piezometer data from outside the investigated hillslope area have been checked (for informa-
tion about the installed piezometer see Thielen (2007)). Change in water table depth measured by piezometer 1 was 0.16 m for the July 2005 and 0.58 m for the April 2006 event (respectively 0.03 and 0.59 for piezometer 2). The piezometer field data show changes in water table depths twice as high as the modelled ones. As discussed under Section 5.3, further model runs with an extended parametrization regarding the parameter $c$ and other influential parameters could be performed. Hypothetical storm events with higher intensities and total precipitation amounts could be modelled to further check the model behaviour.
5.5 Influence of pipe flow on saturation patterns

All results presented so far were simulated with a pipe density of 0.7. This value defines the percentage of hillslope cells contributing to the pipe network (containing pipes). The sprinkling experiments, demonstrated clearly the occurrence of pipes and macropores but their exact location, density and connectivity is a priori unknown. This Chapter therefore assesses the influence and importance of pipe flow on total subsurface flow and saturation processes in the hillslope. Pipe height above bedrock was set to the reference height of 0.2 m with a standard deviation of 0.5 m because soil pipes were found to be located within a narrow range of about 20 cm above the soil-bedrock interface (Uchida and Mizuyama, 2002) and because the water tables with the given parametrization have shown to be restricted to a depth of about 20 cm in former chapters. (The excavation experiments at Tössegg showed, that such a parametrization is not justified, because no clear soil bedrock was definable, nor a clear height, distribution and density of pipes was found. To study the influence of pipe flow this parametrization was necessary).

Figure 5.9 shows ten graphs of maximum water table depths for ten different pipe densities of the April 2006 event. Graphs are similar to the ones presented in 5.7 and 5.8 with the difference, that not patterns of frequency distributions of certain depths are shown, but the average saturation depth per cell (in cm) of ten Monte Carlo simulations at the same pipe density. The average value, instead of the median, has been chosen to better account for extreme values. The evaluated time period is again day 11 of the simulated event (Tuesday, April 11, 2006).

The bedrock depression in the middle of the hillslope is apparently important for water accumulation for all simulated pipe densities. For lower pipe densities (particularly for 0 and 0.1) the maximum water table depths in this depression is allowed to reach high values of about 15 cm. If the pipe density is increased, this zone seems to be quickly drained by pipe flow and water table depths may even fall to half of the level at low pipe densities. This is in agreement with McDonnell (1990), who found that water tables quickly dissipated by the effect of pipe flow (see Section 2.1.2, bypass flow). Especially at low pipe densities, the same saturation areas were identified as in Figures 5.7 and 5.8. As already stated, the bedrock and surface topography mainly influences these patterns.

The likelihood of restricted water table development by high drainage capacities is further supported by Table 5.5. Total lateral runoff at the lower end of the hillslope is augmented by each increase in pipe density. Accordingly, the pipe-matrix flow ratio increases until a pipe flow contribution of nearly 40% is achieved. Higher combined matrix and pipe flow consequently leads to lower water tables. Average water table values are according to Table 5.5 and strictly decreasing with increasing pipe densities.

Figure 5.10 shows the cumulative frequency of maximum saturation depth for all the pipe density simulation runs presented in Figure 5.9. One curve shows the average distribution of water table height of ten Monte Carlo pipe network implementations for the same pipe density. The shape and positions of the curves are consistent with the finding of restricted water table levels by increased pipe densities. High pipe densities result in generally lower water tables. Therefore the curves are shifted stepwise to
Figure 5.9: Influence of pipe flow on the spatial distribution of maximum water levels (cm) along the hillslope for the April 2006 event
5.5 Influence of pipe flow on saturation patterns

Figure 5.10: Cumulative frequency distribution of water levels above bedrock (%) for different pipe densities. A value of 0 means no pipes, a value of 0.9 that 90 percent of all hillslope cells contribute to the pipe network.

Figure 5.11 and 5.12 further illustrate the influence of pipe flow on hydrological processes in a hillslope. (Description of the graphical HillVi outputs can be found in Section 4.1.10 and are summarized in Table 4.1 on 33). Both Figures show a random deterministic simulation run of the July 2005 event (July 19, 2005). In Figure 5.11 a pipe network at a density of 0.7 is included and in 5.12 mo preferential flow network. Despite the single deterministic model run with a random pipe network, the critical saturation locations appear both in Figures 5.11 (c) and 5.12 (c) at (11/6) and in the channel reaching from (1/2) to (8/10), as discussed in Section 5.4.
These two Figures illustrate clearly the draining effect of pipe flow. Where in Figure 5.12 (c) a large part of the bedrock depression in the middle of the hillslope is simulated to be saturated, the pipe network in Figure 5.11 drains the depression, leading to a lower depth of saturation but distributing the water downslope, where it may drain into the trench or enter the soil matrix elsewhere in the hillslope, if the pipe is closed ended. The relative flow along the hillslope ((b) in the same Figures), is homogeneous for the hillslope with no pipe network, showing generally higher flow rates in the lower, steeper part of

<table>
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<th>Flow type</th>
<th>Pipe density</th>
</tr>
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<td>0.0 0.1 0.2 0.3 0.4 0.5 0.6 0.7 0.8 0.9</td>
</tr>
<tr>
<td>Matrix flow [mm]</td>
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</tr>
<tr>
<td>Pipe flow [mm]</td>
<td>0 1.8 5.4 7.9 13.8 9.7 12 22.4 21.8 25.9</td>
</tr>
<tr>
<td>Total amount [mm]</td>
<td>76.7 78.2 80.8 82.9 85.6 87.1 89.4 90.8 92.9 95.1</td>
</tr>
<tr>
<td>Pipe-matrix flow ratio [-]</td>
<td>2.3E-2 7.1E-2 0.11 0.19 0.13 0.16 0.33 0.3 0.37</td>
</tr>
<tr>
<td>Mean water table over whole slope [cm]</td>
<td>5.1 5 4.8 4.6 4.3 4.2 3.9 3.7 3.6 3.4</td>
</tr>
</tbody>
</table>

Table 5.5: Summary Table of flow amount, pipe-matrix flow ratio and mean water table elevation for 10 different pipe densities (April 2006 event)
5.5 Influence of pipe flow on saturation patterns

(a) No pipe network implemented, x-y showing matrix and pipe flow

(b) Relative (combined matrix and pipe) flow

(c) Saturation depth

(d) Water content in unsaturated zone

Figure 5.12: Graphical HillVi output for the July 2005 event without pipe network

the hillslope. Implementation of a pipe network increases flow rates at the particular location, resulting in a heterogeneous shaded layer in Figure 5.11 (b). The bar plots in (a) shows the pipe flow rates per trench cell (where in Figure 5.12 no pipes are included, no pipe flow occurs). Both bar plots from (c) and (d) only indicate matrix flow, where (b) is the combination of (a) and (c/d) (pipe and matrix flow). Combined flow in Figure 5.11 shows peak flow rates into the trench cells, into which a pipe directly drains.

Note  The modelling done within this Section assumes again a distinct soil-bedrock interface, which was, as discussed earlier, not present in the field. All the values concerning the lateral runoff Table 5.5 are based on the assumption of a perched water table above bedrock. As subsurface runoff data were not available, the runoff data can not be related to reality. However, the runoff amounts are very high. At the highest pipe densities, more than 80% of the precipitation amount appears as runoff (which is at least doubtful for this soil type). Increasing the parameter $\text{bedrock}_\text{ksat}$ or $c$ would drastically reduce this amount. Nevertheless, as the intention in this Section was the investigation of the influence of pipe flow on saturation patterns, this parametrization is justifiable.
Discussion

Section 6.1 briefly summarizes the results of each objective addressed in this thesis, mentions strengths and weaknesses and suggests solutions addressing particular emerging issues. Section 6.2 then takes one step back and discusses a holistic approach focusing on the goal of this thesis, the refinement of hydrological modelling of a hillslope prone to shallow landslides. Section 6.3 concludes the thesis by some final remarks.

6.1 Discussion of objectives

Objective 1: "Identification of sensitive model parameters and quantification of their influence on model output and relationship"

Performing a sensitivity analysis in combination with a "Latin Hypercube Sampling" approach has proven to be a very useful tool to assess model behaviour and to investigate the tuning importance of each parameter. An advantage of the regional method applied here, a multiple linear regression approach, is the ability of varying several parameters at the same time and rank their relative influence on model output. Prior to the interpretation of the sensitivity analysis results, regression diagnostics have to be performed to assess the quality of the regression model. The influence of single model runs on the regression can thus be evaluated and the performance can be increased stepwise. In particular, this procedure insures that influential but meaningless or even numerically unstable model runs originating from parameter sampling are eliminated.

The quality of the regression model applied could be improved further regarding the saturated storage and runoff. Tukey-Anscomb plots showed a small systematic trend which could not have been improved by adaptation measures done within this thesis. A possibility would be to extend the sensitivity analysis to non-linear models.

The results of the sensitivity analysis based on the interpretation of standardized regression coefficients enhanced model knowledge and seemed to be very comprehensible. The model reacted very sensitively to the following parameters: \( c, \text{bedrock}_\text{ksat}, k_0, b \) and \( m \) (see Tables 4.2, 5.2 and 5.3).
For future modelling tasks with HillVi, it is recommended to specify or at least constrain these parameters by field experiments and laboratory analysis. For \( b \) this could be done by analysing the drainable porosity values of different soil samples from different depths and fit the exponential model form HillVi to the data. \( k_0 \) could be defined in the field by infiltration experiments. The parameter \( c \) could be defined by analysing water retention curves (according to Weiler and McDonnell (2006)). Measuring \( k_{sat} \) in different depths to define the shape parameter \( m \) would have been impossible in Tössegg because the soil texture did not allow taking samples of a size, large enough to include the influence of bypass flow on the saturated hydraulic conductivity. Quantification of \( \text{bedrock}_{-}k_{sat} \) is particularly difficult. A suggestion to assess the seepage problem is to calculate the water balances with help of soil instruments (MP, TDR, Tensiometer), runoff data and particularly piezometers and/or crest-stage gauges. On this basis a quantification of bedrock seepage would be possible (e.g. Tromp-van Meerveld (submitted)). \( \text{bypass} \) and \( \text{thres} \) are two parameters whose implementation into HillVi is highly empirical and parametrization delicate. An algorithm that generates random macropores for vertical bypass flow could be implemented analogical to the soil pipe network (this is especially justifiable if a multiple layer model is considered as described later). This implementation could conceptually be based on Weiler (2001), who investigated the mechanism controlling macropore flow during infiltration.

**Objective 2:**  "Modelling of single rainstorm events and comparison to field data"

Based on the results of the sensitivity analysis, a "base-calibration" parameter set was acquired and further optimized by varying single sensitive parameters. The model was able to describe the temporal change in mean water content of the unsaturated zone for the two events in April 2005 and 2006 (Nash-Sutcliffe Efficiency of 0.78, respectively 0.81) adequately. However, limiting the calibration exercise of the model to the measured water content curve in the unsaturated zone would result in the equifinality problem in this particular case with HillVi: as described earlier, the unsaturated zone is only linked to the saturated zone by recharge and is one of the first storages in the models storage chain ("storage chain" is defined as a series of storages where water flows from one storage to another). Several parameter sets would produce "good fits" for water content curves, but may also influence other processes situated at the end of the storage chain, e.g. related to the saturated zone, that are then not represented well. This behaviour would not be identifiable by evaluating only water content curves. Therefore it is urgent to calibrate the model at the end of a storage chain, as e.g. to the runoff data and evaluate and validate the water content as so called internal model process. Due to the lack of runoff data, the model could not be calibrated in a traditional sense. For future hydrological modelling it is suggested to measure surface and subsurface runoff separately, most suitably with appropriately sized tipping buckets, as the triangular spillway method applied in Tössegg has shown to be accident-sensitive (Thielen, 2007).

**Objective 3:**  "Investigation of maximum saturation patterns across the hillslope and isolation of frequent saturated areas"

The model runs generated for the sensitivity analysis served as basis for the spatial analysis of satu-
6.1 Discussion of objectives

Discussion of objectives. Although these simulations generated large differences in model outputs of runoff or saturated and unsaturated water content, overall three hillslope areas emerged to be susceptible for frequent high water depths. One location is a bedrock depression in the middle of the hillslope and the other two are situated downslope at 6 m distant from the lower end of the hillslope. With the sensitivity analysis approach a certain parameter independence regarding locations with frequent high water levels at the soil-bedrock contact has been demonstrated. Bedrock and surface topography seem to be the first order control for the appearance of these two locations. This is in agreement with studies of Weiler and McDonnell (2004) and Tromp-van Meerveld and McDonnell (2006) and others. Thus, a detailed DEM of bedrock and surface topography optimally derived from ERT measurements are indispensable for adequate hillslope modelling (van Meerveld and McDonnell, 2006). Pipe flow had a minor influence on these saturation patterns, because pipe density was set to 1 m above bedrock with a standard deviation of 0.5 m and modelled water tables rarely rose up to this height. If this range of modelled water table change is realistic, the simulation data was compared to values from field piezometer measurements and studies by Tromp-van Meerveld (2006). These values were approximately twice as high, as the modelled water tables. Finally, to discuss the modelled water table values, the hillslope should be densely instrumented with piezometers or crest-stage gauges on the soil-bedrock contact zone; such an instrumentation set-up could contribute to validate model predictions of water levels and help to quantify the amount of bedrock seepage through water balance calculations. Such studies have been done at the Panola hillslope (Tromp-van Meerveld and McDonnell, 2006).

**Objective 4:** "Assessment of the influence of preferential flow on saturation patterns and water balance"

Dye tracer infiltration experiments clearly demonstrated the occurrence of soil pipes and macropores. However, it was not possible to define a specific density or height of pipes for model parametrization. This is not in agreement with some studies, which found that soil pipes mainly occur in the topsoil (A-horizon of the soil) or a the soil-bedrock interface (C-horizon) (see Uchida et al. (2001)). Since the distribution and density of pipes was largely random at Tössegg, mean soil depth and a large standard deviation have been specified for general modelling tasks in order to represent pipe occurrence over the whole soil profile. Pipe flow strongly controls the appearance of saturation patterns. Mainly two effects have been isolated from the simulations:

1. Drainage efficiency is elevated by pipe flow, leading to restricted water levels.

2. Pipes ending in a dead-end passageway cause local water accumulations and result in a very local increase of water levels.

The pressure head in this part of a pipe is elevated and therefore the pore water pressure in the surrounding matrix, which then may become an endangered location for the triggering of shallow landslides. These two pipe flow effects are in agreement with the review study of Uchida et al. (2001) (be referred to Section 2.1.2 for an overview and possible effects of pipe flow).
Due to the random nature of the pipe network generation in HillVi, only minor spatial predictions of the two effects described can be made. No information is won regarding the spatial distribution from this analysis: the areas that are influenced by pipe flow are delineated by the areas where already the analysis of the latter objective showed high water levels. Under these particular circumstances, the implementation of a preferential flow network provides not more information, than if the model is run without pipes and seems to be unsuitable for predicting possible triggering locations. However, the general effects of pipe flow (point 1 and 2 above), regarding the terminology of "first order controls" emerged from this investigation.

6.2 Synthesis and recommendations

The conclusions drawn from the objectives discussed above give partial information about model characteristics when addressing a certain task. This Section takes one step back and discusses the whole modelling approach and experiences while modelling with HillVi. As mentioned, the bedrock topography is a first order control for the characteristics of the saturation patterns. One could argue that it’s enough to know the bedrock topography of the hillslope to characterize these locations of maximum water levels. However, as Tromp-van Meerveld and McDonnell (2006) and others showed, the bedrock topography for modelling is only meaningful combined with a surface topography because of variable soil depth: before the soil-bedrock interface can influence the water flow, water has to flow there. Hence, all processes involved in water transmission in reality as in modelling, are important factors and influence largely the water flow behaviour (as soil porosity, saturated hydraulic conductivity of the flow, the preferential flow behaviour, infiltration capacity among others). Successful modelling can only be done, if knowledge about such parameters are available. Therefore a process procedure for future modelling tasks with HillVi is provided here:

1. Definition of measurable parameters
   - Geotechnical analysis of the hillslope (including ERT) as described by Thielen (2007)
   - Measuring infiltration capacity on large soil samples and with ring infiltrometer
   - Measuring total and drainable porosity at soil surface and at different soil depths and locations across the hillslope to define related model parameters as $n$, $n_o$ and $b$

2. Perform first modelling of storm events based on a sensitivity analysis approach within limited parameter ranges derived from field measurements to familiarize oneself with the hillslope’s characteristics

3. Install the measuring devices based on the spatial information obtained from the results from the simulations performed under step 2:
   - Collecting channels for surface flow (top 10 cm)
6.3 Final remarks and personal statement

- Collecting channel for subsurface flow right at the soil-bedrock contact
- Placing crest-stage gauges and/or Piezometers on a regular grid across the hillslope and at critical locations as predicted by the first simulations from point 2
- Install common measuring instruments as a meteorological station (precipitation, temperature, humidity) and MP, TDR and tensiometers
- Measure soil deformations in the hillslope (as suggested by Springman 2008, personal communication)

4. Field data interpretation

5. Modelling tasks based on a "calibration-validation" approach

While working with HillVi and the Tössegg data, two ideas for further model development arose:

- The first recommendation emerged from analysis of the ERT and geotechnical data and from a specific finding of the sprinkling experiments: the Tössegg hillslope is constituted of two different soil layers, clayey and silty sand. Soil characteristics between or along these two layers may be significantly different and have an influence on the flow behaviour and the soil water balance. However, such a two layered soil model cannot be supported by HillVi. HillVi could be extended to support multi layer soil configurations in combination with a probabilistic macropore infiltration network extending vertically into the soil as recommended above.

- As a strong relation between increased pore water pressure and the triggering of shallow landslides seems to exist (e.g. Blong and Dunkerley (1976), Brand et al. (1986), Pierson (1983), Rickli and Bucher (2002), Sidle and Kitahara (1995)), the model could be extended to provide slope stability information based on a saturated or unsaturated factor of safety value approach as described in Section 2.3. The data basis to solve the stability equations is available though adequate instrumentation and supporting laboratory testing. A comparable numerical modelling approach is described in Uchida and Mizuyama (2002).

6.3 Final remarks and personal statement

HillVi was developed as a platform to discover first order controls in hillslope hydrology (Weiler and McDonnell, 2004) based on a simple model concept. This approach is criticized by many scientists as being too far away from physical reality. My task was to investigate the influence of preferential flow on the hydrological processes and to understand the model concept of HillVi and its philosophy. Beside this theoretical modelling task, I got to know the field site Tössegg during the sprinkling experiments. I saw, that requirements for HillVi, namely the existence of a distinct soil-bedrock interface, were not given or at least only under very heterogeneous conditions. In this way, I knew that my modelling task could never reproduce the reality, also due to the lack of data, namely subsurface runoff and piezometer data.
and due to not having access to a complete calibrated parameter set. However, I think, that the concept behind HillVi allows for investigations of complex natural systems and the main factors that possibly control them by including random components as the pipe network. The continuous development of research techniques will lead to improved knowledge about natural systems. Perhaps, for example, soil pipe networks and soil porosities in the whole soil profile will be definable in future by remote sensing technologies or radar investigations and provide the necessary data for successful modelling. Until then, researchers are urged to model with large uncertainties. These uncertainties are increased with each increase in model complexity. From my point of view, a model should focus on the first order controls discussed, while keeping the mathematical model as simple as possible. The thesis at hand therefore tried to investigate the influence of preferential flow on the hydrology of a hillslope prone to shallow landslides, while keeping in mind, that the results have to be interpreted not in a quantitative, but in a conceptual way of thinking.


Horton, R., 1933. The role of infiltration in the hydrologic cycle. Transactions, American Geophysical Union, 446–460.


### Table A.1: Depth of soil probes in Toessegg. Source: Thielen (2007)

<table>
<thead>
<tr>
<th>Mean depth of probe [cm]</th>
<th>TDR</th>
<th>Moisture Point Segment</th>
<th>Tensiometer</th>
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</thead>
<tbody>
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<td>3.1</td>
<td></td>
<td></td>
</tr>
<tr>
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<td>1, 11, 15, 16</td>
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</tr>
<tr>
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<td>3.2</td>
<td></td>
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</tr>
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</table>
Figure A.1: 3D model of the Tössegg hillslope, own illustration (IDL)

Figure A.2: Layout of instrumentation. Source: Thielen (2007)
Figure A.3: Time series data of all Moisture Point instruments
Figure A.4: Times series data of all TDR instruments
Figure A.5: Time series data of all Tensiometer instruments
Figure A.6: Decline of $ko$ and $no$ with depth for the values shown in the legend
(a) Statistics for saturated storage at time step 1372 (ca. 229h), July 2005

(b) Statistics for subsurface flow at time step 924 (154h), July 2005

(c) Statistics for saturated storage at time step 243 (40.5h), April 2006

(d) Statistics for subsurface flow at time step 567 (94h), April 2006

Figure A.7: Meta model statistics for sensitivity analysis (top: July 2005, bottom: April 2006)