Doctoral Thesis

Flood behavior in alpine catchments examined and predicted from dominant runoff processes

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FLOOD BEHAVIOR IN ALPINE CATCHMENTS
EXAMINED AND PREDICTED FROM
DOMINANT RUNOFF PROCESSES

A thesis submitted to attain the degree of
DOCTOR OF SCIENCES OF ETH ZURICH
(Dr. sc. ETH Zurich)

presented by

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Summary

Understanding of the physical processes that produce extreme floods is often limited, especially in alpine environments. Flood predictions are therefore subject to considerable uncertainty. To explore how process knowledge may improve flood prediction in alpine catchments, I developed specialized mapping and modeling tools on the basis of field studies of the storage and drainage mechanisms in different alpine landforms in the Schaechen catchment, Central Switzerland. This thesis reports the findings at the field sites and examines how the developed tools help to explain and predict the flood responses in three catchments with strongly contrasting flood behaviors.

The Schaechen catchment was chosen for the field studies because the largest floods on record are markedly different from the medium floods. This difference could be caused by threshold-like changes in runoff formation occurring in large parts of the catchment. The fact that the largest floods were caused by long-duration storms that produced over ~100 mm of precipitation, suggests that the relevant threshold-like changes occur in areas that can store much of this precipitation volume over event time scales.

To explore this hypothesis, the runoff behavior was studied in three spring catchments and three headwater catchments that are dominated by an alpine landform type with large storage capacity. Rainfall-runoff responses were monitored during three summers, recording three floods with return periods of more than 2 years. One spring catchment with a threshold-like runoff response was studied in more detail, with sprinkling experiments and measurements of the storage dynamics in the soil and the underlying fractured rock.

In each of the field sites, during every single storm event in the study period, more water was stored than discharged. Such a large storage considerably dampens the rainfall-runoff response; i.e., runoff rates are much smaller than the flood-producing rainfall intensities, and peak flows occur several hours to days after the storm. Areas with thick sediment deposits produced only little runoff during the largest events of the study period. Runoff generation in the site with threshold-like response occurred mainly in the heavily fractured bedrock that underlies the ~1 m of permeable soil cover. The soil could store much water, but not for long enough to cause the observed damped responses.

These case-studies indicate that the time scale of storage may be large for various commonly found alpine landforms. These landforms can cover large parts of an alpine catchment, such that understanding their occurrence and hydrological functioning is important for the prediction of floods in alpine terrain.
Based on the process knowledge gained at the Schaechen field sites, I developed a classification tool for mapping the dominant runoff processes (DRP). The classification specifies the runoff mechanism and response strength from a geomorphological characterization of the landscape. It requires only generally available spatial data, and complements existing methods for mapping runoff processes in the shallow subsurface.

Concurrently, I developed the model QArea for simulating flood runoff on the basis of the DRP maps. It has a different structure for each runoff mechanism; to reduce the model complexity, parameters are shared between the classes as much as possible. The parameters were mainly derived from small-scale observations during small floods in the Schaechen catchment. Simulations of other extreme events in the Schaechen, including the largest flood on record, showed that the model could predict the flood peaks and volumes quite accurately.

The DRP mapping and modeling framework was applied in two other meso-scale catchments with markedly different flood behavior from the Schaechen: the Hinterhein, with a stronger runoff response, and the Dischma, with a more damped runoff response. In each catchment, one small flood and one large flood were simulated to evaluate how differences in runoff behavior may be explained from differences in the spatial distribution of dominant runoff processes.

Without adjusting the parameterizations of the runoff processes—the DRP maps thus being the only difference in model setup for the three catchments—, QArea predicted the flood behavior in these events satisfactorily. This suggests that the DRP framework describes the physical processes reasonably well. Furthermore, the DRP maps and event simulations indicate that threshold-like flood behavior is more likely in the Schaechen catchment than in the Hinterhein and Dischma catchments.

As a control, the same small and large floods were simulated with simple lumped models that were calibrated to observations in the respective catchments. Despite being calibrated, these models could not predict the floods in the Schaechen and Dischma catchments, which indicates that the process representation by the DRP framework helps to improve the assessment of flood behavior in alpine catchments.

These results give confidence that the developed framework can be useful in other alpine catchments as well. For example, to evaluate the sensitivity to storms that are more extreme than have been observed until now, or to assess the plausibility of a threshold-like change in the flood response.
Zusammenfassung


Die Felduntersuchungen wurden im Einzugsgebiet des Schächens durchgeführt. Im Schächen sind die grössten Ereignisse wesentlich grösser als die übrigen Hochwasser. Die Tatsache, dass die grössten Hochwasser durch langanhaltende Niederschläge mit über 100 mm Regen verursacht wurden, deutet auf Schwellenwerte hin; das heisst: anfänglich wird ein grosser Teil des Niederschlags gespeichert, und wesentlicher Abfluss erst nach ergiebigen Niederschläge.


Die verschiedenen typischen alpinen Geländeformen, die nur mit grosser Verzögerung reagieren, können grosse Teile alpiner Einzugsgebiete bedecken. Um die Speicher- und Drainageprozesse in dieser Hänge zu bewerten, wurde ein Verfahren zur Kartierung dominanter Abflussprozesse (Dominant Runoff Processes; DRP) entwickelt, das auf verfügbaren Geo-Daten, sowie einer geomorphologischen Charakterisierung der Landschaft basiert. Es stellt eine Erweiterung vorhandener Verfahren zur Kartierung oberflächennaher Abflussprozesse dar.

Die entwickelte Methodik wurde danach im rasch reagierenden Hinterrhein und dem stark verzögert reagierenden Dischma evaluiert. Die Simulationen, sowohl von kleineren als auch grossen Hochwassern, basierend auf den DRP Karten führten, ohne weitere Anpassung der Parameter, zu befriedigenden Resultaten. Dies lässt darauf schliessen, dass sich das unterschiedliche Verhalten durch die DRP erklären lässt.

Diese Ereignisse wurden auch mit einfachen Konzeptmodellen simuliert, die für die jeweiligen Einzugsgebiete kalibriert wurden. Weder im Schächen noch im Dischma konnten die grosse Hochwasser zufriedenstellend simuliert werden.

Diese Tests zeigten dass sich mit Hilfe der entwickelten Methodik das Verhalten alpiner Einzugsgebiete besser einschätzen lässt. Durch Szenarien-basierte Analysen lässt sich mit dieser Vorgehensweise die Sensitivität auf unterschiedliche Niederschlagsverläufe evaluieren und die Wahrscheinlichkeit von abrupten Änderungen der Hochwasserreaktion abschätzen.
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CD  Coarse Debris (sediment texture type, see Table 3.1)
CI  Confidence Interval
CV  Coefficient of Variation
DEM Digital Elevation Model
DGSD Deep-Seated Gravitational Slope Deformation
DP  Deep Percolation
DRP Dominant Runoff Process
EC  Electrical Conductivity
ERT Electrical Resistivity Tomography
ESMA Explicit Soil Moisture Accounting
EWA Elektrizitätswerk Altdorf AG
FFA Flood Frequency Analysis
FG  Fine-Grained (sediment texture type, see Table 3.1)
FOEN Federal Office for the Environment
GA25 Geologischer Atlas der Schweiz 1:25000
GEV Generalized Extreme Value
GWL Ground Water Level
HOF Hortonian Overland Flow
JJA months of June, July, August
MAM months of March, April, May
ME Mean Error, see Eq. (4.16), p. 94
MX Mixture of CD and FG sediments
m.b.g.s. meter below ground surface
m.s.l.  meter above mean sea level

NSE  Nash-Sutcliffe Efficiency, see Eq. (4.14), p. 94

ODE  Ordinary Differential Equation

RT  Runoff Type

SN  short for the paper by Scherrer and Naef (2003)

SOF  Saturation Overland Flow

SON  months of September, October, November

SSF  Subsurface Stormflow

SWC  Soil Water Content

VE  Volumetric Efficiency, see Eq. (4.15), p. 94

VWC  Volumetric Water Content
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<th>Description</th>
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<tr>
<td>$A$</td>
<td>$L^2$</td>
<td>planar area (e.g., of a catchment or model element)</td>
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<tr>
<td>$A_{\text{contr}}$</td>
<td>$L^2$</td>
<td>total area of all QAREA$^+$ model elements with a $S_{g,s}$ reservoir</td>
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<tr>
<td>$\alpha_f$</td>
<td>$-$</td>
<td>QAREA$^+$ diversion regulator; fraction of $q_f$ flux routed to the $S_g$ reservoirs</td>
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<td>$\beta$</td>
<td>$-$</td>
<td>QAREA$^+$ model $S_u$ nonlinearity exponent</td>
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<tr>
<td>$C_{R,d}$</td>
<td>$-$</td>
<td>direct runoff coefficient</td>
</tr>
<tr>
<td>$C_{R,e}$</td>
<td>$-$</td>
<td>event runoff coefficient</td>
</tr>
<tr>
<td>$c$</td>
<td>$ML^{-3}$</td>
<td>tracer concentration</td>
</tr>
<tr>
<td>$D$</td>
<td>$L$</td>
<td>maximum sediment cover depth associated with a DRP class</td>
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<tr>
<td>$E_a$</td>
<td>$LT^{-1}$</td>
<td>actual evaporation flux</td>
</tr>
<tr>
<td>$E_p$</td>
<td>$LT^{-1}$</td>
<td>potential evaporation rate</td>
</tr>
<tr>
<td>$E_u$</td>
<td>$LT^{-1}$</td>
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<tr>
<td>$\eta$</td>
<td>$-$</td>
<td>nonlinearity parameter (exponent) of the L2 nonlinear reservoir model</td>
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<td>$f_{A,b0}$</td>
<td>$-$</td>
<td>fraction of the catchment area lying below the $0^\circ C$ line</td>
</tr>
<tr>
<td>$K$</td>
<td>$T$</td>
<td>time constant of linear reservoir</td>
</tr>
<tr>
<td>$K_{L2}$</td>
<td>$TL^{a-1}$</td>
<td>time constant of the simple nonlinear reservoir model L2</td>
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<tr>
<td>$K_d$</td>
<td>$T$</td>
<td>time constant of the QAREA$^+$ $S_d$ reservoir</td>
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<tr>
<td>$K_{g,f}$</td>
<td>$T$</td>
<td>time constant of the QAREA$^+$ $S_{g,f}$ reservoir</td>
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<tr>
<td>$K_{g,s}$</td>
<td>$T$</td>
<td>time constant of the QAREA$^+$ $S_{g,s}$ reservoir</td>
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<td>$K_{ob}$</td>
<td>$T$</td>
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<td>$K_s$</td>
<td>$T$</td>
<td>time constant of the QAREA-C model $S_s$ linear reservoir</td>
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<tr>
<td>$L_i$</td>
<td>$LT^{-1}$</td>
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<tr>
<td>$P$</td>
<td>$LT^{-1}$</td>
<td>precipitation flux</td>
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<tr>
<td>$P_{\text{eff}}$</td>
<td>$LT^{-1}$</td>
<td>effective precipitation flux into the the QAREA$^+$ model $S_u$ reservoir</td>
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<tr>
<td>$Q$</td>
<td>$L^3T^{-1}$</td>
<td>discharge</td>
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<td>$Q_T$</td>
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<td>baseflow rate (as specific discharge)</td>
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<td>q&lt;sub&gt;f&lt;/sub&gt;</td>
<td>LT&lt;sup&gt;-1&lt;/sup&gt;</td>
<td>nonlinarily responding 'fast' runoff flux from the QA/r.sc/e.sc/a.sc model</td>
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<td>Hortonian Overland Flow flux of the QA/r.sc/e.sc/a.sc model</td>
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<td>specific discharge at the moment of peak flow</td>
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<td>q&lt;sub&gt;tot&lt;/sub&gt;</td>
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<td>total outflow flux from a QA/r.sc/e.sc/a.sc model element</td>
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<td>q&lt;sub&gt;u&lt;/sub&gt;</td>
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<td>storage in simple nonlinear reservoir model L2</td>
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<td>routing delay reservoir of the QA/r.sc/e.sc/a.sc model</td>
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<td>storage 'slowly' draining reservoir of the QA/r.sc/e.sc/a.sc model</td>
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<td>fast response storage threshold upper reservoir of the QA/r.sc/e.sc/a.sc model</td>
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<td>subcatchment channel routing delay QA/r.sc/e.sc/a.sc model</td>
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<td>precipitation volume, expressed as depth</td>
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<td>event discharge volume, expressed as depth</td>
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Surely, a PhD research project is a journey, and, as I like to travel, embarking on this journey seemed like a sensible thing to do. Even more so because the project concerned hydrological research in the beautiful Swiss Alps and I happened to be curious about how water-related processes occur in nature. Keen to learn about these processes and the environment in which they occur, I set out on this project. The aim was to obtain new insights about how Alpine slopes store and release water, and use these insights to develop a mapping technique for characterizing the dominant hydrological processes in the landscape, and through this, improve the understanding and prediction of flood runoff formation in Alpine catchments.

This very useful goal has become dear to me, and I think my PhD research has managed to contribute to it a fair bit. I can thus say with confidence that this PhD project was a good thing for me to do, however, not only because of the research itself: the journey was at least as important as the destination.

Of course I should not detail the journey here, as it is mainly relevant to me. But of all the nice, interesting and rewarding experiences that the journey entailed, there is one particular insight that has shaped much of this research. I think it is fair that I here try to explain what this insight is about.

As many hydrologists before me, I had some challenging encounters with hydrological reality. For example, I learned that there are many steep areas that can store so much rainfall that they hardly contribute to the catchment-scale flood runoff, whereas in other areas, deep subsurface drainage processes may become important contributors to floods caused by long-duration storms. This is not common wisdom in the scientific community, and asks for the question: how can we distinguish these areas? Another example: today’s discharge, rain, and snow data proved insufficient for testing many of our hypotheses about the hydrological functioning of Alpine watersheds.

These encounters were demanding, because they further complicated the already challenging problem that flood prediction in Alpine terrain. I learned from these encounters that much of what I thought I knew —or was taught that ‘we’ know—, was in fact pretty much unknown. In some ways, this was a rejuvenating experience, giving me the energy to go back to the basics and explore new frontiers. In other ways, this was a disappointing lesson: goals had to be reformulated, which often felt like giving up (and is not my favorite thing to do).

Dealing with these issues required me to remain positive and focused. For this, I found it helpful to keep in mind that much of what we think of as reality, is actually only a model of it. Of course, I had learned that before. It may even seem trivial. But what are the consequences?
Well... my limited understanding of the philosophical perspectives on this matter tells me that the consequences are far from clear and trivial: knowing that we do not know reality does not tell you how to ‘deal’ with the unknown reality, let alone its prediction in the future!

The insight came when I was trying to map out hydrological reality in the landscape, the core of my PhD research. It became clear that a map is only a model of reality, that is useful for its purpose at best. A full description of reality is thus an unattainable goal, and one instead has to define what is ‘fit for purpose.’ A good thing of models is that they are easier to change and improve than realities, such that it is clearly a good idea to remain open — but at the same time critical— to new ideas. To me it seems also clear that at their cores, model and reality should be aligned as much as possible. This means that realistic relationships have to be distilled from case studies and existing data sets. These relationships, possibly together with new data, may yield an improved model.

For the development of the mapping tool and the runoff prediction model that builds upon it, this insight thus taught me to ‘see through’ the quantitative data and instead focus on qualitative understanding first. Of course, I would like to have hard data, but as this is often insufficiently available, I should find a description that is at least qualitatively meaningful and coherent. This then became the ultimate goal for the research: develop a suitable qualitative description of flood runoff formation in Alpine terrain, and find proper ways to obtain and evaluate the quantitative predictions that can be based on this description. This thesis presents the results of my best shot at meeting this goal.

My increased appreciation of ‘qualitativeness’ was probably not fully coincidental, as it occurred at the same time that I was following a course called “Readings in Environmental Thinking,” led by Professor Jaboury Ghazoul. Through some of the works we read during the course, and the texts I got inspired to read after it, I learned to respect that many things are simply impossible to quantify, but still have an important qualitative value. For example, the value of being in an environment where you can get in contact with courses like these.

This course was one of the many unexpected but rewarding things that happened along my PhD journey. I am thankful I had the opportunity to encounter these things, and for the personal enrichment they have provided, but for many of them I do not know whom to actually thank. I am glad that this is mostly different for the PhD research itself, and want to reserve a few lines to express my gratitude to the most important contributors.

I first want to thank my main supervisor, Felix Naef; thank you very much for challenging me to thoroughly study this exciting topic of flood behavior in alpine catchments and never giving up trying to keep me focused on the key issues, ... and for enduring and editing my sometimes somewhat prosaic writings. Many special thanks also go to James Kirchner, my Doktorvater, for the great support, patience, thorough proofreading, and the many inspiring and thought-provoking discussions. Further thanks go to Wolfgang Kinzelbach, for his supportive attitude, interesting discussions and for nurturing the good group atmosphere that I have enjoyed for so many years. Not in the least do I want to thank Bruno Merz, the external examiner of this thesis, for the inspiring and pleasant cooperation and interesting discussions.

I am also very grateful to Michael Margreth, Peter Kienzler and Simon Scherrer, who, as experts in researching and mapping of runoff formation processes, proved excellent sparring partners who contributed substantially, directly or indirectly, to the design of the research.
Similarly, I want to thank Nina Volze, for the fun times during our joint field campaigns, contribution to the research design, and the handling of many of the organizational matters involved.

My sincere thanks also go to the great bunch of people who helped making the field campaigns successful, for example by: providing great technical support (Thomy Keller, Ernst Bleiker, Dani Brown and his always helpful team of assistants); helping to setup the Magic Drill for installing piezometers for monitoring bedrock groundwater tables (Chris Gabrielli), helping with geophysical explorations and geodetic measurements (Lasse Rabenstein, Marian Hertrich, Sebastian Tilch); exploring the influence of karstified limestone formations in the catchments I got interested in (Arnauld Malard, Pierre-Yves Jeanin); setting up and maintaining our field sites and analyzing some of the data (Raphael Riedler, Michael Schwab, Adrian Castrischer, Jacob Anz, Matthias Castrischer); supporting the fieldwork by allowing us to use their land and facilities, always providing useful information, help and lodging (Michi and Elsbet Arnold, Fridolin Gisler and his family, Karl and Luzia Gisler, and the hospitable people of Skiclub Effretikon), as well as the many helpful people at Elektrizitätswerk Altdorf AG and Uri Cantonal Authorities. A great thank you to you all!

I further want to thank Marius Floriancic, Veronica Morales, and Alina Tyukhova for the valuable help with some parts of the writing (and the thesis title, again many thanks for that, Veronica!). Special thanks go to Florian Köck, Sebastian Stoll, Manuel Antonetti, Benjamin Fischer, Michael Rinderer, Ilaria Clemenzi, Marco Carenzo, Dominik Krug, Marc Wolf, Georgina Bennett, Nans Addor, Michael Wernli, Maxence Carrel, Dani Sidler and François Michalec for the interesting discussions and the occasional help. And of course I want to thank all the nice colleagues and staff at the Groundwater and Hydromechanics Chair for contributing to the pleasant working atmosphere.

Much of this research project was conducted within the SACflood project in the NFP61 program of the Swiss National Science foundation and I want to thank the agency and in particular the NFP61 steering committee for the support and interesting cross-disciplinary workshops, with special thanks to Petra Schmocker-Fackel for putting the SACflood project into motion.

Most of the analyses, modeling and writing were done with open source software. I want to gratefully acknowledge the people behind the exquisite Python programming language, especially those contributing to the SciPy, NumPy, pandas, Spyder, IPython and Matplotlib packages, and the communities contributing to the very praiseworthy QGIS, GDAL, LiEnX, LiX, Zotero, JabRef, and Subversion projects.

On a journey like this you also need much support from those that are close to you, so I am really grateful to my dear friends and family, who coped so well with me being so far away, both physically and mentally, and have always been great at having joyful times together. I am particularly grateful to my parents, for their seemingly infinite support and always encouraging me to do the things that are important to me.

My deepest gratitude goes to Femke, my dear wife and best mate, who is always inspiring and the right kind of critical, and our lovely son Minas, who can always make me laugh and knows best how to boost my energy. You both have been the best support I could have.

Maarten Smoorenburg
November 2015, Zürich

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Chapter 1

Introduction

1.1 Floods in alpine catchments

In alpine catchments, the most damaging floods are often caused by long-duration storms (Hilker et al., 2009). In August 2005, for example, flooding affected large parts of Switzerland and caused six casualties and a damage of 1870 million Euro, almost a quarter of the in total 8000 million Euro in damages caused by floods, debris flows and landslides in Switzerland between 1972 and 2007 (Hilker et al., 2009). This event made clear that the flood runoff generation in alpine areas is not well understood, and, as will be outlined below, it thereby triggered various questions that motivated the research presented in this thesis.

Improved management and reduction of flood risk require that probabilities of the occurrence of extreme floods are properly estimated. Yet, present-day methods for both operational forecasting and estimating design floods often exhibit substantial uncertainties. The methods often rely heavily on inductive inference techniques based on the available catchment-scale discharge observations; they define a relation for predicting flood magnitudes that is not grounded in understanding of how these floods are generated. Predictions of extremes larger than those found in the observation record may then be unreliable, and their plausibility cannot be assessed from process knowledge. These problems of inductive techniques may be addressed by incorporating knowledge obtained through techniques that can represent the underlying causal mechanisms (e.g., Taleb, 2005, 2007; Kleinhans et al., 2010).

Flood frequency analysis (FFA) is an example of an almost purely inductive technique for estimating flood probabilities. It does not consider the mechanisms that caused the floods, and estimates the probabilities of extreme floods by fitting an extreme value distribution to observed magnitudes of small- and medium-sized floods. It thereby neglects that the mechanisms that caused small floods may be entirely different from those that cause large floods, which conflicts with the underlying assumption that all floods are part of the same random sample (e.g., Klemeš, 1993, 2000).

Various techniques for estimating the probabilities of floods in catchments where no detailed flood records are available go further along this inductive pathway, by conditioning some model for transferring probabilities (e.g., a regression relation) on correlations between some FFA results and some catchment properties (see Rosbjerg et al., 2013 for a recent review in the context of prediction in ungauged basins). Again, there is little knowledge of the flood formation processes involved, leading to relatively uncertain estimates, sometimes even based
Another problem is that predictions based on the above-mentioned methods assume that the driving climatological regime is stationary. It is, however, well known that the climate is non-stationary, and that there are considerable fluctuations in flood occurrence over time (e.g., Klemes, 2000; see Schmocker-Fackel and Naef, 2010a,b, and Wirth et al., 2013) for studies of historical floods in Switzerland. Anthropogenic climate change may further challenge this stationarity assumption (e.g., Milly et al., 2008). Understanding of the flood-producing mechanisms may help to evaluate the uncertainties in flood risk caused by changes in the driving climatological regimes through scenario analysis, provided that the changes can be well characterized.

As indicated before, the limitations of above-mentioned inductive methods may be amended by incorporating the underlying causal relations. For assessing flood probabilities, this means including more knowledge of the relevant flood-generating mechanisms (e.g., Merz and Blöschl, 2008a,b; Beven, 2012; Rosbjerg et al., 2013; Merz et al., 2014). This knowledge may then be represented in a rainfall-runoff model to facilitate experimenting with scenarios of extreme storms. At present, however, knowledge of the relevant mechanisms is often too limited to follow this route, in large part also because no suitable method for characterizing the spatial heterogeneities of the processes are available (e.g., McDonnell et al., 2007). For example, preferential flow and subsurface stormflow in fractured rock can be important runoff generating mechanisms, but they are often not adequately represented in present-day rainfall-runoff models because spatial data about their occurrence and behavior are difficult to obtain.

Because of this limited representation of process knowledge, the models often depend greatly on catchment-specific calibration (e.g., Beven, 2012), which, even if the model structure entails appropriate simplifications of reality, still means that the knowledge used for prediction is largely based on inductive inference. Another consequence—that may well be related, see Klemes (1986a)—is that these models often have little predictive power outside the calibration conditions, whether these conditions are more extreme events in the same catchment (e.g., Seibert, 2003; Coron et al., 2014) or similar climatological conditions in different catchments (e.g., Blöschl, 2006; Oudin et al., 2008, 2010). For high mountainous regions like the Swiss Alps, the state-of-the-art is unlikely to be better, as “observations of the states of nature in mountainous terrain are the most difficult to make and the processes governing mountain hydrology cover the widest range thus posing the greatest demands on theoretical understanding” (Klemes (1990), titled “The modelling of mountain hydrology: the ultimate challenge”).

There is thus considerable scope and need for improving the understanding of how floods in alpine catchments are generated. Improving the understanding of these runoff generating processes and its representation in tools for prediction for extreme floods is the general aim of this thesis; the specific problems that the research has focused on are discussed in the next section.

### 1.2 Aims of the research

The research presented in this thesis focuses on the understanding and prediction of flood formation in meso-scale (~50–500 km²), high mountainous catchments of the Swiss Alps (peaks
Figure 1.1: Annual maximum floods in the Schaechen catchment and the precipitation sums in the preceding days. The four largest floods, indicated with Roman numerals, are markedly larger than the rest, and are difficult to predict. (a) Annual maximum floods, specified per season. (b) Flood frequency plot with Generalized Extreme Value (GEV) distributions fitted to the full observational record and the longest period between two extreme floods (1940–1976); the 95% confidence intervals (CIs) hardly overlap. (c) Catchment average precipitation; the grey tones indicate the contributions of the three days until the day of the flood peak (day 3). (d) The largest of the three daily precipitation sums of (c) versus the peak discharges in (a); only the two largest events received more than 100 mm d\(^{-1}\) precipitation. Methodology and details of the results are discussed in Appendix A.

These catchments are of interest because flood estimation at this intermediate scale is particularly challenging (e.g., Dooge, 1986), and there can be considerable damage potential.

The largest floods are sometimes much larger than the rest, such that extrapolations of inductively inferred relations severely underestimate the magnitudes of the extreme floods. An interesting example is the Schaechen catchment, where most of the research presented in this thesis was conducted (see Sect. 1.3). The four largest floods on record are markedly larger than the other floods (Fig. 1.1a). A flood frequency analysis based on the 37 years between the second and the third extreme floods (marked II and III, respectively) illustrates how extrapolation of an obtained relationship can seriously underestimate extreme floods (Fig. 1.1b). Appendix A presents the analysis and its findings in more detail.

This behavior is here referred to as ‘threshold-like,’ because of the apparent step-change in
the flood frequency curve (Fig. 1.1b), which could be caused by the exceedance of thresholds in the catchment’s rainfall-to-runoff transformation process (e.g., Struthers and Sivapalan, 2007). Rogger et al. (2012), for example, showed how in two meso-scale Alpine catchments a step-change could be attributed to a storage threshold being exceeded.

Such threshold-like behavior might be visible as a strong nonlinearity in the relation between precipitation and flood discharge, but no such relation could be established for the Schaechen flood response. There is little correlation between flood peak discharge and the 24-, 48-, or 72-hour precipitation sums that lead up to them, as is visible in the large differences between the flood peaks and the 72-hour precipitation sums (Figs. 1.1a and c), and the considerable scatter in the plot of flood magnitude versus the highest 24-hour precipitation sum in the three preceding days (Fig. 1.1d).

The poor correlation may be related to the inadequate precipitation measuring interval of 24 hours. For example, the two large floods of 1977 and 2005 occurred in response to 10- and 18-hour periods with moderately high intensity precipitation; considerably shorter than 24 hours. In addition, many smaller floods were produced by storms with low snow lines or substantial snowmelt, making the estimation of the catchment liquid water input subject to considerable error. These issues may often complicate the analysis of flood formation in alpine catchments (Appendix A).

The described threshold-like flood behavior is quite common. The flood records of many meso-scale alpine catchments in Switzerland show extreme floods that are considerably larger than the small floods, or have a marked step change in the flood frequency curve; for example in the Engelberger Aa at Buochs, the Berninabach at Pontresina, the Taschinasbach at Gruesch, and the Brenno at Loderio.

A detailed study by Scherrer AG after the extreme flood of August 2005 in the Schaechen suggested that the poor predictability of the most extreme floods may be caused by a large portion of the area having a large storage capacity (Scherrer AG, 2007). These areas would need to exhibit a process change during large and longer-duration events, causing them to contribute little to catchment flood runoff during small events, but substantially during larger ones. Such process change may be caused by the rainfall depth exceeding a storage threshold, or by the storm duration exceeding a critical concentration time (i.e., if an area contributes mainly to the receding limb of the flood hydrograph during small events, but already to the rising limb during more extreme events), or a combination of both mechanisms.

This seems a reasonable hypothesis. The ~4.2 mm h\(^{-1}\) peak discharge of the largest flood in the Schaechen catchment is smaller than the largest floods in the Muota and Reuss catchments to its north and south, respectively: the largest flood in the Muota (at Ingenbohl), with a catchment area of 316 km\(^2\), was 4.9 mm h\(^{-1}\), and the largest flood in the 192 km\(^2\) Reuss (at Andermatt) was 5.5 mm h\(^{-1}\). The catchments have comparable record lengths, and the same ranking is found if the magnitudes of the 2- to 10-year floods are compared. This goes against the often found relation that flood magnitudes in nearby catchments get smaller with increasing catchment area (e.g., Rosbjerg et al., 2013). These figures indicate that the Schaechen has a relatively damped flood response, which could indeed, as Scherrer AG (2007) suggested, be caused by a comparatively large storage.
Because the largest floods in the Schaechen catchment are caused by long-duration storms, i.e., storms lasting between ~8 hours and several days, the areas responsible for the hypothesized threshold-like response must either have a storage threshold between roughly 50 mm and 150 mm (Fig. 1.1d), or store such volumes for longer than a few hours, yet release a considerable fraction of the storage within a few days. The behavior of these areas with large storage capacity is relevant for the understanding of flood formation in meso-scale catchments, because here the large floods are mostly caused by long-duration storms. However, as will be discussed in Sect. 2.1 it is not well understood what kind of areas may exhibit such behavior.

The Schaechen flood behavior illustrated several issues in present-day techniques for flood runoff estimation (Sect. 1.1). These problems may be addressed by methods that can represent the runoff contribution of areas with large storage capacity. The research presented in this thesis therefore explores the following research questions:

1. What kinds of alpine landforms have a large storage capacity?
2. What are the storage and drainage time scales of these alpine landforms?
3. How can the spatial distribution of runoff generating processes be described at the scale of meso-scale catchments?
4. How can the understanding of runoff generating processes and their occurrence in alpine terrain be used for flood prediction?
5. Do these advances in the representation of process understanding improve the prediction of extreme events?

Questions 1 and 2 were explored through process monitoring and field experimentation at various hillslopes and subcatchments in the Schaechen catchment. Question 3 involved development of a classification tool for mapping dominant runoff processes in meso-scale catchments. The resulting maps form the basis of a newly developed distributed rainfall-runoff model. These mapping and modeling tools were used to qualitatively and quantitatively evaluate question 4. Finally, predictions with these tools are evaluated in different catchments (question 5).

This thesis thus entails a feasibility study on how process knowledge may lead to improved prediction of floods in alpine terrain. The five research questions thereby help to establish a proof of concept for the developed methods, that is, to show that they are useful for addressing catchment-specific questions about the probabilities of large floods.

### 1.3 Methodology and outline of the thesis

Based on a literature review, along with field and GIS explorations in various Alpine catchments, I made a shortlist of commonly occurring landform types with large storage potential. Several research sites were set up in the Schaechen catchment to study some of these landform types. These sites were spring or headwater catchments predominantly covered by one typical landform. This field research included various plot- and hillslope-scale experiments and
 explorations, combined with monitoring of precipitation and discharge at relevant locations. Much of this work was done in collaboration with fellow PhD student Nina Volze, who focused on improving the process understanding at the plot and hillslope scale. These results are presented in Volze (2015). My focus was on comparing the flood runoff reactions of these sites and their landform characteristics to learn about the storage and drainage mechanisms, quantities and time scales. The key findings are presented in Chapter 2 thereby addressing research questions 1 and 2 formulated in Section 1.2.

Based on these findings, a mapping tool was developed for classifying the runoff response of alpine landforms throughout a catchment (Chapter 3). Through the combination of field visits, measurements and interpretation of generally available data, the infiltration, storage and drainage processes were characterized. This forms the basis of the classification of dominant runoff generating mechanisms and their process intensities at the hillslope scale, in order to obtain a meaningful hydrological description of the landscape (e.g., Grayson and Blöschl, 2000; Woods, 2002; Devito et al., 2005; Schmocker-Fackel et al., 2007). The mapping method is an extension of the dominant runoff process classification schemes first outlined by Scherrer and Naef (2003), which have been successfully applied in various meso-scale catchments.

The resulting dominant runoff process maps form the basis of the concurrently developed QA\textsuperscript{Area} model, which is presented in Chapter 4. Each mapped landscape element receives a model structure that matches its dominant runoff mechanism, using established model concepts with parsimonious parameterizations. The parameterizations of the different dominant runoff process classes were derived from our hillslope- and subcatchment-scale observations in the Schaechen, and from response relations that were found to give useful predictions with the maps obtained with the classification schemes of Scherrer and Naef (2003). Model performance was evaluated for a 2-year flood event in a headwater subcatchment of the Schaechen, and for various events in the Schaechen main catchment. The model was found to represent the behavior of the dominant runoff processes and adequately predict the catchment-scale floods.

To explore questions 4 and 5, the developed mapping and modeling tools were applied in two meso-scale catchments with contrasting flood behavior: the Hinterrhein catchment with a strong runoff response, and the Dischma catchment with a strongly damped runoff response. The large differences in response require that differences in spatial distribution of runoff formation processes are considered, as these differences may not be fully explained from differences in storm characteristics alone. The selected catchments and events therefore permitted an evaluation of the prediction skill of the developed tools. The mapping and modeling tools were applied without catchment-specific adjustments (e.g., calibration), thus transferring only the small-scale knowledge that informed the parameterization of the QA\textsuperscript{Area} model. The results are presented in Chapter 5 which also includes a comparison with simple models that do not explicitly account for differences in the spatial heterogeneity.

The results of this knowledge transfer to other catchments are promising, and demonstrate the potential of the developed tools. In the concluding chapter (Chap. 6), I summarize these findings and elaborate on why the methods may be expected to work also in other alpine landscapes and what are viable strategies for further improvement and evaluation of the tools.
1.4 Definitions

Some words that other authors have used differently are defined here.

I use the word ‘Alpine’ to refer to high mountainous landscapes, with peaks well above ~2000 m.s.l, situated in the European Alps. These landscapes are characterized by relatively recent deglaciation, such that the Quaternary sediment packets deposited on the mountain slopes and higher-elevation valley floors are poorly consolidated, and soils are relatively young (more detailed discussions follow in the introductions of Chaps. 2 and 3). The word ‘alpine’—without capital ‘A’—is used to refer to these kinds of landscapes in a more general way, as similar types of terrain can be found, for example, in the Andes and the Rocky Mountains, such that our results may be compared with studies in these regions.

The word ‘event’ is often used as an adjective; for example ‘event storage.’ It here only indicates the time scale, not the origin or age of the water. This is different from the use in studies where ‘event water’ refers to the actual water parcels that have fallen as precipitation during an event, and that may be treated as different from ‘pre-event water.’ For floods research the increase in runoff at the event time scale is more relevant than the actual age of the discharged or stored water. The word ‘event’ is used here to denote the change in volume state or flux relative to the ‘pre-event’ conditions. For example, ‘event storage’ refers to the increase in water stored since the onset of a storm event (Sect. 2.3.4).
Chapter 2

Observations of damped runoff behavior in the Schaechen catchment

2.1 Introduction

2.1.1 Storage and its effect on flood runoff in steep terrain

Catchment-scale water balance evaluations often indicate that a large portion of the rainfall and snowmelt is stored in the subsurface, even in mountainous terrain (e.g., Hood et al., 2006; Sayama et al., 2011; Andermann et al., 2012; Hood and Hayashi, 2015). Event time scale storage is an important determinant of flood magnitude in alpine catchments (alongside meteorological inputs), particularly for floods caused by multi-day storms, when also hillslopes with delayed responses may contribute to the rising limb of the catchment flood hydrograph. The combination of peak runoff reduction and response delay caused by event time scale storage is here referred to as damping of the runoff response.

How much water is stored for how long is poorly understood at the hillslope scale, as are drainage flowpaths and stormflow generation in general (e.g., McDonnell, 2003; McNamara et al., 2011; McDonnell, 2013). This may be even more so for extreme floods in high mountainous terrain like the Swiss Alps (e.g., Klemeš, 1990). In steep terrain, subsurface stormflow generation is often indicated as the dominant runoff generating mechanism, particularly when soils are deep (Dunne, 1978, 1983). Subsurface stormflow is often associated with late runoff initiation and flow peaks occurring after the storm peak intensity (e.g., Dunne, 1978; Montgomery and Dietrich, 2002), and threshold-like behavior of the hillslope-scale runoff response appears common (e.g., Weiler et al., 2006; Zehe and Sivapalan, 2009; McDonnell, 2013).

Threshold-like behavior in hillslope-scale response may be reflected at the catchment scale if a large fraction of the area is governed by the same behavior (e.g., Rogger et al., 2012, 2013). Scherrer AG (2007) hypothesized that in the Schaechen catchment such threshold-like response causes the strong runoff responses during the most extreme events (Sect. 1.2). Scherrer AG (2007) pointed out various slopes with thick debris deposits or permeable bedrock that may have a relatively high storage threshold, which might be only exceeded during the most extreme events. In the Schaechen catchment, thick sheets of debris are mainly deposited by glacial processes (moraines of different ages), gravitational processes (rockfall, forming talus cones), and debris flows (recognizable as debris cones). Permeable bedrock is mainly associated
with strong fissuring in karstified limestone formations and creeping landmass slopes (German: Sackungen), a form of deep-seated gravitational slope deformation (e.g., Soldati, 2013). These landscape features are discussed in more detail below.

### 2.1.2 Landscape features with large storage potential

#### Moraine deposits

Unsorted sediments deposited through glacial activity are referred to as glacial till, the associated landforms are called moraines. There is little hydro(geo)logical knowledge about the relatively young and permeable tills in alpine terrain, especially in comparison with the typically older, more consolidated ‘continental’ tills, commonly associated with low permeability (Ronayne et al., 2012). The glacial till deposition process is complex and different advance-retreat cycles may cause mixed till deposits with markedly different properties (e.g., Roy and Hayashi, 2009). Information about the deposition processes is often not available.

Fluvial reworking of the till decreases the permeability due to sorting of the grains and increasing silt and clay contents. Fluvially reworked landscape features are sometimes still referred to as moraines, but, strictly speaking, the sediment is not glacial till and the landform would be for example an outwash plain or outwash terrace. On the other hand, some slopes are dominated by other features, masking the occurrence of glacial till. This may cause substantial underrepresentation of slopes with glacial mantle.

Alpine slopes with thick (≥ 2 m) till deposits are often reported to store water long enough to support seasonal or perennial outflow (Caballero et al., 2002; Clow et al., 2003; Roy and Hayashi, 2009; McClymont et al., 2011; Langston et al., 2013; Hood and Hayashi, 2015). Estimates of short-term runoff rates of these kinds of slopes could not be found, but occasionally a fast response is mentioned (Roy and Hayashi, 2009) or suspected (Ronayne et al., 2012).

#### Talus deposits

Talus slopes are rockfall debris accumulations, often with a cone shape. Talus and scree are sometimes used interchangeably, although talus is a more general term for the landform and material, and scree is often specifically used for gravel-sized sediment deposited by rockfall. A wide variety of talus slopes exist, from cone-shaped deposits with up to boulder-size debris extending tens of meters deep (e.g., Pierson, 1982; Clow et al., 2003; Schrott et al., 2003; Otto and Sass, 2006; Sass, 2007; Phillips et al., 2009), to shallow talus ‘sheets’ consisting mainly of gravel-sized material (Pierson, 1982). Other (re-)deposition mechanisms may occur on the same slope, such as debris flows and avalanches. Fine-grained sediments may be found at larger depth even though little is found at the surface (e.g., Pierson, 1982; Brazier, 1988; Williams et al., 1997; Clow et al., 2003; Schrott et al., 2003). Some studies report stratification of the sediments (Yair and Lavee, 1976; Pierson, 1982; Otto and Sass, 2006; Sass, 2007) and an increase in grain size towards the base (e.g., Pierson, 1982).

The wide variety of talus slopes may well imply similarly wide variety of the runoff response, as Pierson (1982) argued based on estimates of sediments’ hydraulic conductivities. A few hydrological process studies reported a fast runoff response, and attributed this to lack of fine-grained material (Muir et al., 2011) or the existence of a compacted, flow-impeding
2.1. Introduction

horizon of fine-grained sediment above which stormflow is generated (Yair and Lavee, 1976). On the other hand, studies of tracer flow velocities (e.g., Pierson, 1982; Caballero et al., 2002), hydrogeochemical analyzes (e.g., Liu et al., 2004; Baraer et al., 2014), and spring discharge observations (Clow et al., 2003) typically suggest that talus slopes can store much water over seasonal time scales. However, these studies did not research contributions to flood runoff in detail.

Debris cone deposits

No studies of runoff formation in debris cone deposits were found. Subsequent debris flows often form different channels across the cone (e.g., Brazier, 1988; Stoffel et al., 2008), suggesting deposition of interfingered sheets with substantial fine-grained sediment content. This may be associated with considerable event time scale storage. Debris flows may have occurred at some of the Schaechen sites studied here, but were not the main deposition process.

Mixed deposits

The above deposition mechanisms often co-occur and the resulting ‘mixed deposits’ may largely fill alpine valleys (e.g., Schrott et al., 2003; Otto et al., 2009). Various studies point out the importance of this ‘valley fill’ for the seasonal water balance (e.g., Suecker et al., 2000; Caballero et al., 2002; Clow et al., 2003; Liu et al., 2004; Roy and Hayashi, 2009; McClymont et al., 2010; Baraer et al., 2014; Hood and Hayashi, 2015), but the importance of event time scale storage in thick debris deposits for abating catchment flood runoff appears little researched.

Fractured rock

Fractured rock may provide additional storage capacity to sediment mantled hillslopes. However, its importance for stormflow formation is often not recognized (e.g., Wilson and Dietrich, 1987; McDonnell, 2003; Gabrielli et al., 2012; Salve et al., 2012), in spite of the many studies that attribute fast and slow runoff processes to fractured rock flows (see recent review in Salve et al., 2012). In situ weathered rock is usually fractured and may extend to many meters depth, even in steep terrain (e.g., Anderson and Dietrich, 2001; Montgomery et al., 2002; Onda et al., 2004; Kosugi et al., 2008, 2011; Salve et al., 2012; Brönnimann et al., 2013).

The zone with strongly fractured rock may be much thicker if deep-seated gravitational slope deformations occur (e.g., Madritsch and Millen, 2007; Ustaszewski and Pflüffer, 2008). Furthermore, fissures in the rock may be important for various types of slope failure: water in the fractured rock aquifer may induce shallow or deep-seated landslides (e.g., Montgomery et al., 2002; Onda et al., 2004; Rickli et al., 2008; Brönnimann et al., 2013) and even rockslides (Huber, 1992), by causing high pore water pressure in the rock and/or soil mantle.

Delayed runoff responses have often been reported for steep slopes with ‘normally’ weathered (e.g., Montgomery et al., 1997; Montgomery and Dietrich, 2002; Tromp-van Meerveld and McDonnell, 2006b) or otherwise fractured rock (e.g., Onda et al., 2004; Kosugi et al., 2008); peak flows occur hours or even days after the storm. Similar delays have been observed in activation of landslides (e.g., Montgomery et al., 2002; Rickli et al., 2008) and rockslides (Huber, 1992).
Observations of damped runoff behavior suggesting similarity in hydrological processes (i.e., bedrock fracture flow). These delayed responses indicate that steep slopes may provide substantial event time scale storage.

The bedrock underlying thick debris deposits may be weathered (e.g., McClymont et al., 2011), or even be part of a deep-seated gravitational slope deformation (e.g., Madritsch and Millen, 2007; Ustaszewski and Pfiffner, 2008; Ustaszewski et al., 2008). No studies of the storm runoff response of such slopes were found in the literature, but, as pointed out by McClymont et al. (2011), it may be anticipated that the fractured rock below such deposits is responsible for sustaining baseflow.

2.1.3 Research questions

To better understand runoff formation of slopes with large storage capacity, runoff was monitored at six sites where one or more of the aforementioned landscape features predominate. The runoff responses of these sites are analyzed to explore the following questions:

1. How do alpine landforms with large storage capacity respond to heavy precipitation?
2. How much water is stored at event time scales?
3. Which landscape characteristics determine the damping of the response?

The first question is explored by characterizing the runoff responses of the sites to the largest storm for which good data was obtained. One site, with little response to small storms but considerable response to extreme events, was researched in more detail by comparing different events and more detailed description of runoff formation processes based on soil moisture and groundwater observations.

The second question is addressed through event time scale water balance evaluation. Storage quantification is challenging (e.g., McNamara et al., 2011), and is only attempted at three sites to study drainage mechanisms. Besides, storage may aid evaluation of simulation models by guiding the selection of model structure and parameters from understood proportionality between different processes (e.g., Seibert and McDonnell, 2002).

As a key aim of this PhD research is to use improved hillslope-scale understanding of runoff formation processes for flood prediction at the catchment scale, addressing question 3 involves an attempt at identifying hillslope-scale properties that help explain observed responses, or differences in runoff response in general.

The Schaechen and the six sites are introduced in Section 2.2 followed by a description of measuring and analysis techniques in Section 2.3. A brief presentation of the results (Sect. 2.4) precedes the discussion of the research questions and implications of this study for improving process-based flood prediction (Sect. 2.5).

2.2 Site characteristics

Six sites were selected for detailed monitoring and field experimentation. The Schaechen catchment was screened for typical landscape elements that may have large storage capacity and
2.2. Site characteristics

Damped flood runoff response. The report of Scherrer AG [2007] proved useful for determining what areas to explore and detailed geo(morpho)logical maps like Brückner and Zbinden [1987] helped in getting an overview which Quaternary landforms are common. Literature on runoff formation and mapping techniques provided additional insights. The typical landforms at the selected sites are commonly found in Alpine alpine terrain and the Schaechen catchment in particular.

The exploration and first installments of monitoring equipment took much of the summer season of 2010. Some sites were unsuitable for discharge monitoring, and new sites were selected in the following year with an emphasis on accessibility and robust measurements of water level and discharge. We decided to study three hillslopes in detail, mainly for the research project of Nina Volze [Volze, 2015], and to monitor discharge at additional subcatchment-scale sites to allow at least qualitative evaluation of the storm runoff response. The monitoring network was operational from June to October in the years 2011 to 2013 and also included rainfall and temperature measurements at various locations (Fig. 2.1, Tables 2.1 and C.1). The photograph in Figure 2.2 gives an impression of the central northern flank of the Schaechen main valley where five of the sites are located.

2.2.1 Schaechen catchment

The 108 km² Schaechen catchment is located in the Swiss Central Alps. Its elevation ranges between 490 and 3295 m.s.l (mean: 1717 m.s.l.). The gauging station at the outlet in Bürglen is operated by the Swiss Federal Office for the Environment (FOEN) and has measured discharge at 10 minute intervals since 1985. Until 1985, discharge was measured 3 km further upstream, with an associated catchment area of 94 km². Streamflow in the Schaechen is not significantly affected by river training or reservoirs for hydropower.

Geology is mostly sedimentary rock of the Helvetic nappes and Infrahelvetic complex, consisting of sandstone, shale, flysch, marls, and limestone (Brückner and Zbinden, 1987; Hantke and Zbinden, 2011). Some areas with limestone rock show strong karstification and are expected to drain to neighboring catchments (discussed in more detail in Sect. 3.7). The northern flanks have topographic gradients between 30 and 60 %, and are mainly used as cow meadows, with forests on the steeper slopes. The southern area is steeper and mostly forested. Alpine grasslands cover much of the terrain above the tree line.

Glacial processes have strongly affected the landscape and have left thick glacial till deposits in valley bottoms and plateaus. U-shaped cross sections and hanging valleys are still visible in the two largest tributaries, the Hinterschaechen from the south and the Vorderschaechen from the east. A small alluvial plain has formed at their confluence (Fig. 2.1). The only other large alluvial deposits are the coalescing fan systems starting just below the old gauging station.

Other thick deposits include rockslide (German: Bergsturz) and rockfall deposits (talus cones). Furthermore, debris cones are often found where steep streams encounter flat terrain and deposit much of their bedload sediments. Creeping landmasses in the area underlain by flysch bedrock cover most of the northern flank of the main valley and the Vorderschaechen below ~1800 m elevation. These movements have driven the river towards the southern flank (Jäckli et al., 1985), resulting in the V-shaped main valley and tributaries.
Cirque valleys often form the main headwater sources of the larger tributaries. A thick sediment cover is usually found on the valley floors and slopes, mainly deposited by glacial and gravitational processes. The sediment appears permeable and surface runoff generated in upslope areas often fully infiltrates into these deposits. Such cirque valleys may provide enough long-term storage to support perennial streamflow. In some cases, water is also stored in lakes formed behind terminal moraine walls.

Cirque-shape landforms are less pronounced in the headwaters above the creeping landmass slopes, where the geological map indicates displaced rock masses instead. The thick debris cover commonly found in these areas support a line of springs along a 'shoulder ridge;' thick deposits above the line and creeping landmass areas below, typically with steeper slopes. Springs form the main sources of many small tributaries in the central northern flank, some of which are diverted to a small equalizing reservoir.

These thick deposits provide important drinking water sources for the valley municipalities and some of the streams are used for hydropower generation. This indicates that the storage in the areas above the ridge is large. Many other small springs are found throughout the northern flank and some are used locally for drinking water throughout the year, for example at our Schluecht site (Sect. 2.2.2).

A detailed soil map is not available, but the more than 50 soil profiles in the Schaechen catchment presented by Scherrer AG (2007) give some general overview, which is here briefly summarized. The soil covers of the creeping landmasses are mostly thick (~1 m), permeable cambisols, often with substantial gravel content. Thick, moderately permeable soils are sometimes found on glacial and fluviatile deposits at lower elevations. Shallow, permeable soils have developed on the parent rock and (unconsolidated) debris deposits. Occasionally, gleyed soils are found on moraines in groundwater seepage zones, or because of impermeable layers (stagnosols). Sometimes even extensive swampy peat areas are formed. Bare rock is rare and mainly found above 2000 m elevation.

The Schaechen has a typical alpine water balance with snowmelt in spring and early summer, and the lowest discharge in winter. Glaciers and perennial snow comprise only 2.3 % of the catchment and are found along the southern and southeastern borders; they however cover about 6 % of the Schaechen if their catchments are included.

Mean annual precipitation is 1198 mm y\(^{-1}\) at Altdorf and 1810 mm y\(^{-1}\) at Unterschaechen (water years; period 1 October 1985 to 30 September 2009). The Unterschaechen station data correspond well with the annual catchment average precipitation estimate of 1832 mm y\(^{-1}\) derived from the RhiresM gridded data product (MeteoSwiss, 2013b). In the same period, mean annual discharge was 1436 mm y\(^{-1}\). Monthly precipitation is generally highest from May to September, but there is little seasonality; monthly precipitation of up to twice the average summer rate have been observed in November and February.

Floods mainly occur in summer and autumn (Fig. 1.1). The largest floods are caused by storms with more than 100 mm precipitation in 24 to 48 hours (Scherrer AG, 2007). Snowmelt sometimes causes small floods (Appendix A).
### Table 2.1: Site characteristics of the Schaechen catchment and the six research sites.

<table>
<thead>
<tr>
<th>Site name</th>
<th>Area ($\times 10^3$ m$^2$)</th>
<th>Predominant landform</th>
<th>Dominant lithology</th>
<th>Soil type(s)</th>
<th>Gauge location$^a$</th>
<th>Easting (m)</th>
<th>Northing (m)</th>
<th>Elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schluecht spring</td>
<td>3.4$^b$</td>
<td>creeping landmass</td>
<td>sandstone / claystone</td>
<td>cambisol</td>
<td>701 690</td>
<td>192 162</td>
<td>1484</td>
<td></td>
</tr>
<tr>
<td>Gadenstetten spring</td>
<td>85.1$^b$</td>
<td>talus</td>
<td>–</td>
<td>regosol</td>
<td>700 942</td>
<td>193 095</td>
<td>1539</td>
<td></td>
</tr>
<tr>
<td>West</td>
<td>290$^c$</td>
<td>creeping landmass</td>
<td>limestone / marl</td>
<td>gleyso / peat / cambisol</td>
<td>–</td>
<td>–</td>
<td>1654–1982</td>
<td></td>
</tr>
<tr>
<td>East</td>
<td>1458$^c$</td>
<td>talus / moraine / creeping landmass</td>
<td>sandstone / claystone</td>
<td>regosol / rendzina / cambisol</td>
<td>–</td>
<td>–</td>
<td>1654–2491</td>
<td></td>
</tr>
<tr>
<td>West</td>
<td>1748$^c$</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>697 224</td>
<td>195 406</td>
<td>1645–2491</td>
<td></td>
</tr>
<tr>
<td>Egg springs</td>
<td>438$^b$</td>
<td>–</td>
<td>–</td>
<td>regosol / cambisol</td>
<td>700 936</td>
<td>192 458</td>
<td>1373</td>
<td></td>
</tr>
<tr>
<td>Oberbutzen</td>
<td>806$^c$</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>699 933</td>
<td>192 789</td>
<td>1387–2370</td>
<td></td>
</tr>
<tr>
<td>Lehnstutzbach</td>
<td>258$^c$</td>
<td>moraine / creeping landmass</td>
<td>sandstone / claystone</td>
<td>–</td>
<td>695 475</td>
<td>192 691</td>
<td>679–1316</td>
<td></td>
</tr>
<tr>
<td>Schaechen</td>
<td>108 266$^c$</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>692 480</td>
<td>191 810</td>
<td>490–3295</td>
<td></td>
</tr>
</tbody>
</table>

$^a$ Swiss grid coordinates (i.e.: CH1903-LV03); the catchment elevation range is given if this could be defined, otherwise only the elevation of the gauging station is listed.

$^b$ Estimated from event water balance (see Sect. 2.3.3, Volze, 2015)

$^c$ Estimated manually from 25 m resolution DEM (see Sect. 2.3.3)

#### 2.2.2 Schluecht spring catchment

The Schluecht site is a creeping landmass slope with about 1 m thick silty loam cambisol covering a fractured flysch bedrock. A line of springs drains the steep (~28) grassland slope. Below these springs lies a swamp that is fed by these springs and groundwater upwelling [B.1 in Appendix B]. Discharge, water temperature and electrical conductivity were monitored just below two springs and showed similar storm runoff behavior. The catchment of the smaller spring (discharge measured about 2 m below the confluence of two point sources lying about 3 m apart) was researched in most detail. The results are briefly summarized below, further details on the measuring techniques and geophysical methods can be found in Volze (2015).

A sprinkling experiment with dyed water on a 1 m² plot found that no infiltration or saturation excess overland occurred in response to rainfall rates up to 60 mm h$^{-1}$. Excavation indicated good matrix interaction in the top 20 cm (homogeneously stained soil). Below this, water percolates more preferentially with increasing depth. A large-scale sprinkling experiment, in which 12 mm h$^{-1}$ was applied on 130 m² for 60 hours, indicated a soil water storage increase of about 50 mm (at steady state, with wet antecedent conditions), returning to pre-event levels within days after the experiment. No signs of perched saturation in soil or fractured rock were found in either piezometer nests, tensiometers or soil moisture sensors.

A seismic refraction tomography survey indicated strong fracturing in the top 10–20 m of the flysch rock. An electrical resistivity tomography (ERT) survey along the same profile suggested a groundwater table almost parallel to the land surface at about 5 m below the surface or
Figure 2.1: Schaechen catchment and measuring locations and terrain shown by Alti3D hillshade background (© 2015 Swisstopo (JD100042); Swisstopo, 2013). Subcatchment borders and divide between thick debris deposits and creeping landmass slopes at northern flank shoulder ridge are indicated. Details on research sites and meteorological stations are presented in respectively Tables 2.1 and C.1. Subsets of this map provide more detailed information about the research sites (Figs. 2.3 and 2.2.4).
2.2. Site characteristics

Figure 2.2: Photo of the Schaechen central northern flank with five sites indicated (photo taken from vantage point V in Fig. 2.1).

Figure 2.3: Subsets of Fig. 2.1 for detailed view of sites, using the same legend and hillshade background (© 2015 Swisstopo (JD100042); Swisstopo, 2013). (a) Lehnstutzbach catchment; (b) Oberbutzen catchment and Egg, Gadenstetten, and Schluecht springs.
Observations of damped runoff behavior

Groundwater level observations in boreholes of up to 9 m depth corroborated these findings, with the groundwater table being markedly flatter towards the base of the slope (Fig. 2.4a). Obtained drill cores indicated bedding of the flysch rock is near-parallel to the slope. Furthermore, the rock gets harder with depth, suggesting that fracturing decreases with depth too. Pumping and tracer tests in boreholes 1A and 2A indicated an effective porosity of the rock of about 5\%.

Spring discharge behavior was found to correspond to groundwater level fluctuations, which can be several meters during extreme events (see the event analysis in Sect. 2.4). These fluctuations occur synchronously at the different piezometers, indicating that they are located in the same aquifer. Connectivity with the spring area was demonstrated for the lower boreholes with tracer tests (Volze, 2015). In between the two point sources, already in the swampy area, groundwater head in the deeper borehole (0B; well screen only in the fractured rock) was gen-
erally above that of the shallower borehole (0C; screened over the interface between soil and rock, 0.6 m east of 0B). This head difference across the soil-rock interface usually increased in response to storms, indicating increased groundwater upwelling across a flow impeding zone. The increasing head difference is demonstrated for the two largest events in Section 2.4.2.

Together, the observations described above indicate that runoff formation at the Schluecht slope consists only of drainage from the groundwater body found in the fractured rock. This perceptual model is sketched in Figure 2.4b.

2.2.3 Gadenstetten spring catchment

The Gadenstetten spring drains a thick debris deposit (setting detailed in Fig. 2.3). Topographic gradient is about 50% (27), becoming slightly flatter towards the base. The sediments were deposited by glacial and gravitational processes as indicated by the unsorted mix of gravels, cobbles and boulders. A shallow regosol soil supports the grassland and coniferous forest covering the deposit. A seismic refraction tomography survey indicated the debris mantle depth may locally exceed 50 m (Volze, 2015). Fine-grained sediments (clay, silt, sand) have presumably filled most of the void space between the boulders and cobbles, as could be observed in a gravel pit situated in same deposit, roughly 270 m to the northeast. Springs draining the same deposit about 100 m to the west serve as a drinking water supply throughout the year, indicating long-term storage.

The spring area is a diffuse seepage zone at the base of the slope, with the uppermost important point source situated ~20 m above the discharge monitoring station. The highest observed discharges were caused by snow melt in spring and early summer and not by rainfall. The area below the gauging station is typical creeping landmass terrain. The Gadenstetten spring is thus exemplary for the line of springs on the shoulder ridge between thick debris deposits and creeping landmass slopes (see Sect. 2.2.1). No drilling or further hydrometric monitoring was conducted at the slope. There are no indications of perched saturation in the slope, as no episodic springs were observed in or above the seepage area.

The characterization presented above supports a perceptual model that shows similar processes as at the Schluetteh spring (Fig. 2.4); all rainfall infiltrates, percolates to great depth and recharges a perennial groundwater system. The main differences with the Schluetteh slope are that the soil is shallower and that it is underlain by a thick debris deposit. No information about the underlying bedrock could be obtained, but geological maps suggest that it is the same flysch rock type as in Schluetteh.

2.2.4 Wissenboden headwater

The 1.75 km² Wissenboden headwater is a typical cirque valley, with sharp topographic water divides along the upper half-circle and glacial till covering the main valley floor (Fig. 2.5). Discharge was measured about 70 m downstream of the confluence of the Wissenboden-East (W-E; 1.46 km²) and Wissenboden-West (W-W; 0.29 km²) tributaries. The topographic water divide between the subcatchments crosses the catchment of a relict rock glacier flowing towards W-E, such that the ridge south of this feature was selected as water divide instead (Fig. 2.5).
Vegetation cover is predominantly alpine (hay) meadows, with small patches of bushes or coniferous trees.

The W-W subcatchment is dominated by shallow soils and swamps that have formed on wildflysch bedrock. Strong erosion has formed a dense channel network with strong runoff response already during small storms.

The landscape of the W-E subcatchment contrasts strongly with the W-W subcatchment. The area below 2000 m elevation, almost half the catchment, is dominated by thick debris deposited by gravitational and glacial processes. Extensive geophysical exploration revealed the debris mantle is generally thicker than 20 m and locally up to more than 50 m (Volze, 2015). A few slopes in the southwest have only the shallow, swampy soil cover that makes up most the W-W catchment. Springs with perennial flow and at least episodic connectivity to the main stream are found only in the valley bottom (Fig. 2.2.4). The perennial part of the channel starts at a spring area delivering more than ~10 L s\(^{-1}\).

Above 2000 m, mainly shallow soils are found on limestone bedrock, interspersed with bare rock and small moraine and rockfall deposits. The steep lower slopes exhibit erosion rills that indicate fast runoff formation. These rills are not connected to the main stream network, how-
ever, because the water infiltrates fully into the thick debris deposits downslope. The north-
ernmost slopes bear no sign of runoff formation, indicating that all precipitation infiltrates to a 
karstic aquifer system. The karstic system is expected to drain northwards, in accordance with 
the strata local dip and strike (see Sect. 3.7).

2.2.5 Egg spring catchment

The Egg spring catchment is situated in the creeping landmass area below the shoulder ridge discussed earlier (Fig. 2.3). Land use is predominantly meadows. The Egg spring feeds a small 
stream that the Elektrizitätswerk Altdorf AG (EWA) diverts for hydropower production. Dis-
charge was measured at the intake structure. The EWA reported the spring shows only small, 
seasonal fluctuations. The topographic catchment area is too small to support the observed 
flow, which was in the order of $10 \text{ L s}^{-1}$. I suspected that water flows in the subsurface via old 
stream channels that got sedimented. As at the Gadenstetten slope, the highest flows were 
observed during melt of the winter snowpack.

2.2.6 The Oberbutzen catchment

Streamflow was measured at the “Vorderer Mühlebach” intake structure of the EWA, just below 
the confluence of two streams, both predominantly fed by a single spring. The springs may be 
positioned on the same shoulder ridge line as the Egg and Gadenstetten springs. Roughly 60% 
of the catchment is covered by a thick debris cover. As in the Wissenboden-East catchment, 
all runoff produced in the upslope areas infiltrates into the thick debris cover. The lower parts 
of the catchment have thick permeable cambisols like in Schluecht, and are also indicated as 
creeping landmass areas in the geological map (Brückner and Zbinden, 1987). Land use consists 
of mixed forest and meadows.

The uppermost spring is relatively large with discharges in the order of $100 \text{ L s}^{-1}$. Much or 
all of its water is lost through streambed infiltration over the 480 m long channel reach above 
the confluence (elevation drop of ~120 m). The smaller spring usually discharges less than 
$10 \text{ L s}^{-1}$, and reaches the confluence without visible losses over its 60 m channel reach (30 m 
elevation drop). The catchment area of the intake structure has various private and public 
intakes for drinking water.

Because of these intakes and the re-infiltration in the streambed the water balance of the 
catchment is not well understood. However, the site may still be used to test the hypothe-
sis that runoff generation from the thick debris deposits above the shoulder ridge is strongly 
damped. These springs constitute the only surface runoff from the area and the intake struc-
ture provides a relatively narrow, fixed cross section wherein water level could be measured 
accurately enough to recognize a strong response, should it occur.

2.2.7 Lehnstutzbach subcatchment

The Lehnstutzbach subcatchment is a small tributary of the Schaechen. It is dominated by 
thick glacial tills deposited at the end of the last ice age about 12 000 years ago, when the Reuss 
glacier still protruded into the Schaechen valley (Brückner and Zbinden, 1987). Hantke and
Zbinden, 2011). The damming of the valley by that glacier caused fluvial deposition of glacial till that is still visible as moraine terraces (German: Glazial Stauschotter-Terrasse) at the lower western and eastern catchment boundaries. The Lehnstutzbach has cut a steep V-shaped valley through these terraces without reaching the underlying bedrock. The higher areas are also mapped as creeping landmasses (Brückner and Zbinden, 1987) and various small landslips are found.

Soils are generally permeable and land use is mainly meadows, with mixed forest on the steeper slopes. Due to the low elevations and deposition mechanisms, it may be suggested that the sediments in the moraine terraces are more consolidated and have higher clay content than the thick deposits at the other sites. Such a deposition mechanism may favor extensive flow-impeding layers but no such features where found in the gravel pit in the western terrace.

The river bed is permeable and only the lower ~80 m had water throughout our measuring campaigns. Given the occasional damaging floods reported by the cantonal authorities, the reduction of total runoff due to infiltration in the streambed was expected to be too small to affect large floods.

2.3 Methods

2.3.1 Hydrometrical data

All routine monitoring data was measured at 10 minute intervals and reported timestamps constitute Swiss winter time values (UTC+01:00). Discharge data were obtained from stage-discharge relationships (i.e., 'rating curve') in structures with fixed cross section. They were serviced every two to three weeks.

Discharge at the Schluecht and Gadenstetten sites was measured in sharp-crested V-notch weirs with notch angles of 19° and 60°, respectively. Both were equipped with Keller AG type DCX-22-AA pressure sensor loggers and Odyssey capacitive water level loggers for recording stage inside installed perforated pipes (Figs. B.1 and B.2 in Appendix B). The pressure sensors were found to be more robust than the capacitive sensors. They are however less accurate and were only used when the capacitive sensors failed.

In addition, water temperature and electrical conductivity (EC) were measured using Onset HOBO Conductivity Data Loggers type U24-001. Sensor drift was corrected with manual measurements using WTW Cond 340i handheld meters when sites were serviced. These data are here mainly used to qualitatively scrutinize the direct rainfall contribution to discharge.

The Wissenboden headwater discharge was measured at a check dam (Figs. B.3). A stage-discharge relationship was obtained by fitting the slope and roughness coefficients of Manning’s equation to manual discharge measurements using WTW Cond 340i handheld meters when sites were serviced. These data are here mainly used to qualitatively scrutinize the direct rainfall contribution to discharge.

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tary catchments. Its eastern and western flanks are therefore added to the respective tributaries to simplify analysis. Streamflow in the tributaries was estimated from a two-component end-member mixing model, combining water and tracer mass balance equations whereby tributary flows are the only two unknowns:

\[ Q_g = Q_W + Q_E \]  
\[ c_g Q_g = c_W Q_W + c_E Q_E \]

with discharge \( Q \) \([L^3T^{-1}]\), tracer concentration \( c \) \([ML^{-3}]\); subscripts ‘g’, ‘E’, and ‘W’ indicate the gauging station, eastern tributary, and western tributary.

Discharge at the Egg site was measured with a Nivus PCM 3 device for in-stream ultrasonic Doppler velocity measurements with a hydrostatic pressure probe for measuring stage. It was installed in an EWA water intake structure with circular cross-section (Fig. B.3). The rating curve was constructed from regular salt dilution gauging measurements. The data presented here was scrutinized with data from the additionally installed Odyssey capacitive water level logger.

Discharge of the Lehnstutzbach was also measured with a Nivus PCM 3 device, but in a rectangular culvert (~2 m wide) and without ultrasonic velocimetry (Fig. B.3). Flow depths were generally too shallow for obtaining a rating curve and the relevant high flows were estimated with Manning’s equation using an accurate description of the cross-sectional geometry and the roughness coefficient estimated at the Egg station.

Water level at the Oberbutzen station was measured with a Keller pressure probe mounted in the center of a rectangular intake structure of the EWA (Fig. B.3). Accurate streamflow measurements for constructing a rating-curve could not be obtained, and therefore Manning’s equation was used with the same roughness coefficient as at the Egg station.

The Schluecht slope above the small spring was also equipped with three profiles with Decagon soil moisture sensors at five depths (types 5TM and ECTM). Only the measured volumetric water content (VWC) measured at the profile closest to the spring is used here, because it contains only 5TM sensors, suffered least from data acquisition problems, and shows similar dynamics as the other two profiles. Total soil water content (SWC) in the upper 1 m of soil was estimated from depth-weighted averaging of the five VWC measurements.

Piezometers were installed in a transect of boreholes in the direction of steepest ascent from this spring (Fig. 2.4 Table 2.2). The piezometers consisted of Eijkelkamp 32 mm HDPE pipes with a prefabricated screen/sand filter combination at the bottom 0.5 m; the remaining depth was backfilled with bentonite. Keller AG DCX-22 pressure sensors were installed to measure groundwater levels. The shallower piezometers where no saturation was observed are not presented here. Of the analyzed piezometers, only the deepest, furthest upslope piezometer (4A) fell dry between storms.

Hydraulic connections between boreholes 1A, 2A, and the spring were demonstrated with tracer tests. Groundwater level (GWL) depth increases in the upslope direction. Saturated hydraulic conductivities between \(10^{-4}\) and \(10^{-3}\) m s\(^{-1}\) were estimated with pumping and tracer tests for the lower boreholes. Further details are provided in Volze (2015).
Table 2.2: Properties of the piezometers in the transect upslope from the Schluecht spring (distance and height refer to soil surface).

<table>
<thead>
<tr>
<th>name</th>
<th>depth (m)</th>
<th>height above spring (m)</th>
<th>distance from spring (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0B</td>
<td>2.15</td>
<td>0.60</td>
<td>0.63</td>
</tr>
<tr>
<td>0C</td>
<td>1.10</td>
<td>0.67</td>
<td>0.63</td>
</tr>
<tr>
<td>1A</td>
<td>4.90</td>
<td>5.23</td>
<td>10.84</td>
</tr>
<tr>
<td>2A</td>
<td>6.20</td>
<td>7.42</td>
<td>15.32</td>
</tr>
<tr>
<td>3A</td>
<td>7.70</td>
<td>9.58</td>
<td>20.13</td>
</tr>
<tr>
<td>4A</td>
<td>8.95</td>
<td>28.00</td>
<td>54.66</td>
</tr>
</tbody>
</table>

2.3.2 Meteorological data

Seven rain gauges were operated during the measuring campaigns to get accurate measurements at our sites and to better estimate catchment areal precipitation (Fig. 2.1 and Table C.1 provide details about locations and measured variables). Rainfall was measured with calibrated Davis Rainfall Collector II type tipping bucket gauges with a precision of about 0.2 mm and no heating see Fig. C.1 for example). Tips were logged with Onset HOBO Pendant Event loggers (type #UA-003-64) that can also measure temperature. The loggers were installed in Onset RS1 Solar Radiation Shields except for the Haldi station, and measured air temperature at 10-minute intervals. Gauges were installed about 2 m above ground, except for the Brunni station (~4 m), in well-exposed but sloping terrain. Rainfall data was aggregated to match the 10-minute resolution of the temperature data.

Permanently installed measuring stations have provided additional data (Table C.1). MeteoSwiss measures precipitation and other common meteorological variables at 10-minute intervals in Altdorf, just west of the catchment outlet, and daily precipitation at Unterschaechen station, centrally located in the catchment. MeteoSwiss also provides raster data products based on interpolation of station data on a ~2 × 2 km² grid. Particularly the daily precipitation product RhiresD (MeteoSwiss, 2013a) was found to be useful. The SLF has operated three IMIS stations in the region since 1999. These stations are located above 2000 m and measure some meteorological variables at 30-minute intervals. Their rain gauges are not heated and can therefore not measure snowfall.

2.3.3 Estimation of catchment area

Runoff response is expressed as specific discharge to allow comparison of sites and comparison with the rainfall input signal. Computation of specific discharge requires an estimate of catchment area and this was obtained differently for the sites. Resulting catchment areas are presented in Table 2.1.

The catchment areas of the Schaechen and its main tributaries, the Vorder- and Hinterschaechen (Fig. 2.1), were determined from topographic water divides in the 25 × 25 m² digital elevation model (DHM25: Swisstopo, 2005). Such estimation is generally robust at this scale, as small boundary errors hardly affect the total area. However, some water divides are found in areas with karstified rock. As a result, these catchment areas may be overestimated, with about
5% (Schaechen), 2.1% (Vorderschaechen), and 12.5% (Hinterschaechen) expected to drain to neighboring catchments (see the results of the mapping of dominant runoff processes in the next chapter, i.e., Sect. 3.7 and Table 3.2).

Catchment delineation errors may play a larger role for the smaller subcatchments Wissensboden, Oberbutzen and Lehnstutzbach, because many topographic water divides are located in areas with permeable substrate like thick debris deposits and fissured rock and may not reflect the subsurface water divide. Therefore, some of the catchment boundaries were drawn manually whereby geomorphological features and known spring locations were considered.

No sensible catchment areas could be distilled from topographic data for the Schluecht, Gadenstetten and Egg springs. The catchment areas were therefore determined from a short-term water balance evaluation (details in Volze, 2015). Here, the estimate based on the October 2012 event is taken because this event is used to characterize runoff responses at our research sites (see Sect. 2.3.5). The estimate represents a lower bound estimate; up to two times larger areas were found for smaller events and longer analysis periods. Because of the uncertainty in the catchment area estimation, the resulting specific discharge estimates are mainly analyzed in terms of temporal fluctuations and qualitative comparison of the sites.

### 2.3.4 Event runoff coefficient and event storage

The effect of storage on flood runoff is characterized with the temporal development of the event runoff coefficient $C_{R,e} [-]$ since the start of the event:

$$C_{R,e} (t) = \frac{\int_{t_0}^{t} (q (t) - q_b) \, dt}{\int_{t_0}^{t} P (t) \, dt} \tag{2.3}$$

with $q \, [LT^{-1}]$ denoting specific discharge (discharge $Q$ divided by catchment area $A \, [L^2]$), and $P \, [LT^{-1}]$ denoting precipitation. Definitions are illustrated in Figure 2.6. Event start time $t_0 \, [T]$ is the latest time with lowest flow until the first 1 mm of rain has fallen. A constant ‘baseflow’ $q_b \, (= q(t_0)) \, [LT^{-1}]$ is subtracted to get the discharge increase $\Delta q$, integration of which gives the event discharge volume $V_{Q,e} \, [L]$. The event runoff coefficient $C_{R,e}$ is thus the fraction of event cumulative precipitation $V_{P,e} \, [L]$ that is discharged since $t_0$. At some point in time, $q$ may get smaller than $q_b$ and although the computation may continue, the resulting curve for this period is here not plotted.

Ignoring actual evaporation, $E_a \, [LT^{-1}]$, as it is only a small fraction of $P$ for large storms, event storage depth $S_e \, [L]$ is defined as $V_{P,e} - V_{Q,e}$. Underestimated catchment area causes overestimated $q$ and $C_{R,e}$, and underestimated $S_e$. Omission of $E_a$ implies overestimation of $S_e$ by at most a few millimeters per day.

Various methods for calculation of the event runoff coefficient have been proposed in the literature. Typically, a baseflow separation is performed and a single value is calculated for an event. Differences in definitions of baseflow and how to separate it from the ‘event flow’ have caused much ambiguity in the interpretation of these values (see Blume et al., 2007 for a discussion of the ambiguities and a comparison of techniques). Beven (2012) even suggested that one may best avoid any such procedure altogether. Disregarding baseflow may however result
2.3.5 Characterizing the runoff behavior of the research sites from observations during the October 2012 event

Flood runoff formation at the different sites is characterized by a hydrograph analysis of the flood event of 10 October 2012, hereafter referred to as the October 2012 event. The observed Schaechen flood peak of 36 m³ s⁻¹ has a return period of about 2.3 years (as determined from a GEV distribution fitted to the full observation record; see Sect. 1.2 and Appendix A). This event was selected because it is the largest event for which discharge data was obtained at all sites. Furthermore, interpretation is relatively straightforward because the storm developed similarly throughout the catchment and the snow line was high.

Pre-event conditions did not favor a strong flood response. It was dry the 9 days before the small storm of 7 October (13–33 mm) that preceded the main event, starting on 8 October 2012, 18:00. Pre-event discharge was 3.4 m³ s⁻¹, close to October mean discharge. The storm characteristics are detailed in Section 2.4.1.
2.4. Results

2.3.6 Further characterizing the runoff behavior of the Schluecht slope from responses to 28 different storms

To better characterize the threshold-like response of the Schluecht slope, also observed during the October 2012 event (see results in Sect. 2.4.1), more events were studied. Various hydrograph metrics were compared for 28 storms with more than 10 mm total rainfall and no snowfall or snowmelt (analysis period: May 22nd, 2012, until June 5th, 2013). Storms were separated by at least 24 hours with less than 1 mm d$^{-1}$ of rainfall.

The sketch in Figure 2.6 presents the definitions used for a typical Schluecht spring hydrograph. Event end time $t_{\text{end}}$ is the first dry moment followed by at least 24 hours with less than 1 mm d$^{-1}$ of precipitation. Only the two highest peaks are considered. The second peak, $q_{p2}$, per definition should correspond to an observed groundwater level peak in the upslope boreholes. The first peak, $q_{p1}$, may also correspond to groundwater behavior, for example during long events with multiple rainfall periods, but is usually caused by fast runoff from within direct vicinity of the spring. This is visible as flashy discharge peaks that occur simultaneously with spikes in the EC and rainfall data, suggesting important contributions from quick subsurface flow and ‘direct’ rainfall on the wet area between spring and gauge. The definition of the lag-to-peak follows Montgomery and Dietrich (2002): the time elapsed between the first time stamp that more than half of the total event precipitation has fallen and a discharge peak that corresponds to a peak in groundwater table in the piezometers upslope.

To further characterize the behavior of the Schluecht slope, also the largest flood event observed was analyzed and compared to the October 2012 event. The main storm took place from May 31st to June 2nd, 2013, hereafter referred to as the June 2013 event, delivering about 140 mm of rain in 48 hours. Antecedent conditions were wet; the 222 mm measured at Unterschaechen belonged to the 25 % wettest May months in the 1982–2013 record. Below 1700 m the winter snowpack had completely melted by May 27th. The main event started in the night of May 31st, after about 25 mm of precipitation in the preceding 48 hours. Part of the precipitation may have fallen as snow, but temperatures were mostly above freezing point and our rain gauges recorded liquid precipitation.

2.4 Results

2.4.1 Responses at the six sites: flood of 10 October 2012

Storm characteristics

The main storm lasted about 48 hours and was preceded by a minor storm separated by a 24-hour dry period (Fig. 2.7a). Webcam footage near Wissenboden showed snowfall only occurred at high altitudes. This was confirmed by observations at the Aelpler Tor station (2330 m.s.l., not shown in Fig. 2.7); it registered liquid precipitation, indicating that the 0 °C line remained above 2400 m.s.l. This implies that at most 10 % of the Schaechen and none of the field sites received much snow during the event.
The Wissenboden station received the most rainfall and the Altdorf station the least. They are presented together with the data from the Schluecht station that closely resembles the Schaechen catchment average rainfall estimated at 24-hour intervals from the RhiresD gridded data product (MeteoSwiss, 2013a). All RhiresD grid-cells’ data fall between the observations at Altdorf, 24-hour (48-hour) precipitation sums of 40 mm (62 mm), and Wissenboden, with rainfall sums of 109 mm (145 mm). Despite these large differences, the temporal development of the storm was similar at all stations, with four periods (B1 to B4) with moderately intense precipitation. This suggests that the observations at the Schluecht station may well serve as estimates of average precipitation for computation of the runoff coefficient and storage in the Schaechen (Fig. 2.7e–f). Maximum 24-hour and 48-hour rainfall sums at Schluecht were 50 mm and 88 mm. Estimated catchment average precipitation was 11–14 % more than at Schluecht, but the Schluecht data is used for the computations without correcting for this difference.

**Discharge response**

The Wissenboden subcatchment has the strongest runoff response (Fig. 2.7b), with the highest peak coinciding with the highest rainfall rate (period B2). Subsequent peaks are slightly lower, in accordance with the lower rainfall intensities in periods B3 and B4. Although the Wissenboden response is the strongest of the six sites, the response may be only considered strong in terms of response time and not in terms of runoff volume, as the discharge peaks are not much higher than those observed at the other sites. This apparent paradox is explained by the contrasting behavior of the tributaries. Tributary discharges were only estimated for the first part of the event because of sensor clogging problems at the western tributary and the gauging station, i.e., where reactions were strongest and substantial bedload was transported. The Wissenboden-East tributary showed little response, whereas the Wissenboden-West subcatchment reacted strongly to precipitation, already early in the event: the maximum observed increase of 3.6 mm h\(^{-1}\) occurred after only 19 mm or rain had fallen (an instantaneous runoff coefficient of 26 %).

The Schaechen hydrograph is similar to the Wissenboden hydrograph, with a minor runoff increase during the B1 period and a rapid increase during periods B2 and B3 (Fig. 2.7c). However, specific discharge is lower and, unlike at Wissenboden, the peak in response to the B2 period is the highest of the four main discharge peaks. Also, the superimposed peaks (rapid fluctuations) are smaller. The Schaechen response may thus be regarded as more damped than the Wissenboden response.

The Schluecht spring response is even more damped, with only minor intensity peaks. Discharge increases slowly until about 55 mm of rain has fallen, then rises markedly and peaks within hours after the B4 period (Fig. 2.7c; minor superimposed peak during the B4 period not counted). Peak discharge is about 0.71 mm h\(^{-1}\) above pre-event flow, a 6-fold increase. The Gadenstetten spring discharge response is even more strongly damped. Discharge peaks at only 0.19 mm h\(^{-1}\) above pre-event flow, constituting a factor 2.8 increase, two days after the end of the storm. Both sites have a more damped response than the Schaechen, where discharge increased by a factor 10.

The responses of the Egg and Oberbutzen sites resemble the strong damping of the Gadenstetten spring, whereas no streamflow increase was observed at the Lehnstutzbach site (Fig. 2.7d).
2.4. Results

These sites have larger measurement errors due to sensor inaccuracies and the relatively wide channel cross-sections. At the Oberbutzen station a diurnal signal was observed with a quick drop in water level between 8:00 and 12:00. The origin of this diurnal fluctuation could not be identified, but the effect is found in the full time series and is at least partly caused by the Keller pressure sensor installed here; the same sensors at the Schluecht and Gadenstetten sites showed similar diurnal patterns, only rarely exceeding the 1 cm accuracy the manufacturer specifies for these sensors. As the sensors worked normally and no damages or sedimentation were found after the event, the absence of a strong response cannot be attributed to a detection problem; the obtained rating curves indicate that a five-fold increase in discharge, half the Schaechen discharge increase, requires a more than 7 cm (6 cm) rise in stage at the Egg (Oberbutzen) stations.

Discharge measurements at the Lehnstutzbach station are less accurate due to the wide cross-section of the structure. The absence of response may not be attributed to measurement inaccuracy because no complementary evidence of strong discharge variations was found after the event. Such evidence was found after the larger event of June 2013, when the sensor was damaged by flows with substantial erosive power and transport capacity. Flood marks indicated that runoff peaked at well over 1 mm h\(^{-1}\), a more than 10-fold increase. This was the only event during our campaigns for which we observed such evidence of erosion and sediment transport. A strong response was also reported by the authorities for the most extreme event of August 2005.

Event storage and event runoff coefficients

Developments of the event runoff coefficient \( C_{R,e} \) and event storage \( S_e \) are presented in Figure 2.7e–f for the Schaechen and Wissenboden catchments and the Schluecht and Gadenstetten springs. The strong runoff response of the Wissenboden-West tributary is reflected in a quick rise in \( C_{R,e} \), but the stored volume is comparable to those of the other sites. This is due to the higher rainfall rates, which also cause the comparatively large storage of 120 mm in the Wissenboden subcatchment; only up to 80 mm was stored at the other sites. The water appears to be stored for a relatively long time, as \( S_e \) decreases only about 12 mm during the two days after the storm.

The Schluecht \( C_{R,e} \) response is diametrically opposite from the Wissenboden response: enough water is stored long enough to prevent a quick rise in \( C_{R,e} \) (and discharge), but short-term drainage (~20 mm in 2 days) is strong enough to cause a higher \( C_{R,e} \) within 24 hours after the storm. Still, more than 50 mm remains in storage until more than 4 days after the storm.

Similarly strong drainage after the storm is visible in the development of the Schaechen \( C_{R,e} \). The Gadenstetten spring has the lowest \( C_{R,e} \) and the slowest drainage.

Soil moisture and groundwater responses at Schluecht

Storage in soil and groundwater measured at the Schluecht slope shows contrasting behavior (Fig. 2.7g–h). Whereas soil moisture at most depths reacted already to the small storm of 7 October, groundwater levels were in decline until the start of the main event. The responsive soil moisture behavior reflects rainfall intensity variations throughout the storm, with maximum
Figure 2.7: Responses at different sites during the October 2012 event in the Schaechen. Grey backgrounds indicate the event start, $t_0$, and four periods with more intense rainfall, B1 to B4. (a) Precipitation $P$ at Wissenboden and cumulative precipitation $V_P$ at various sites, together with daily catchment areal average. (b–d) Specific discharge $q$ at all sites, plotted at the same scale.
Figure 2.7: Continued. (e–f) Event runoff coefficient $C_{R,e}$ and event storage $S_e$ since $t_0$ at the main sites. (g) Groundwater level (GWL) in boreholes above the Schuecht spring (Fig. 2.4), expressed in meter below ground surface (m.b.g.s.). (h) Volumetric water content (VWC) at 5 depths and estimated soil water content (SWC) in soil profile next to borehole 1A.
estimated SWC during B3, an increase of 51 mm since \( t_0 \), and rapid depletion after the B1, B3 and B4 periods. The rapid drainage causes SWC and VWC at most depths to drop to less than 10 % above pre-event levels within 36 hours after the storm, indicating little long-term water storage in the soil.

Groundwater levels (GWL) rose markedly at the end of period B2, synchronous with the spring discharge. Total GWL fluctuation increases with distance from the spring for boreholes 0B–3A, with a maximum groundwater table rise of 2.76 m in 3A. Borehole 4A was dry prior to the event, but the groundwater level reached the sensor just after period B2.

### 2.4.2 Comparison of storms at the Schluecht slope

The two largest storms cause a much stronger response of the Schluecht spring than the other 26 events included in the analysis (Fig. 2.8a). The maximum discharge increases during the smaller events were often caused by fast runoff from within the vicinity of the spring \( \Delta q_{p1} \) in Fig. 2.6, see Sect. 2.3.6. These smaller events consisted of event total rainfall depths of up to 64 mm, but discharge did not exceed 0.3 mm h\(^{-1}\) (Fig. 2.8b). Associated event runoff coefficients \( C_{R,e} \) until the end of the storm were at most 0.12, and only once reached 0.2 within 48 hours after the storm (the longest storm in Fig. 2.8c). The rise of \( C_{R,e} \) for these smaller events is slower than for the two largest events.

The groundwater driven discharge peaks typically occur after the end of the storm, except for the two largest events. Groundwater lag-to-peak times range between a few hours and multiple days (Fig. 2.8d). This does not depend on the lag-to-peak definition; a similarly wide range was found for lag-to-peak values referenced to the time of the storm’s end or storm center of mass (not shown).

The largest and second-largest events showed peak increases of 1.2 mm h\(^{-1}\) (June 2013) and 0.7 mm h\(^{-1}\) (October 2012, see Fig. 2.8b). Rainfall intensities were low for these long-duration storms, with the June 2013 event being generally more intense. Maximum hourly rates were, however, similar with 8.5 mm h\(^{-1}\) in the October 2012 event and 8.9 mm h\(^{-1}\) in the June 2013 event. The wetter antecedent conditions are reflected in higher pre-event discharge, groundwater levels, and soil water content (Figs. 2.8 and 2.9). However, the differences between the events’ peak flow increases are larger than the difference between their pre-event flows, although the peaks occur after similar rainfall sums of 80–90 mm (Fig. 2.8b). This indicates that elevated basflow was not the only factor causing the stronger response to the June 2013 storm. The stronger response during the June 2013 events is also visible in the steeper rise of \( C_{R,e} \) (Figs. 2.8 and 2.9). Interestingly, this difference is less pronounced when plotted against cumulative rainfall (Fig. 2.8d), suggesting that event storage is little affected by antecedent conditions or rainfall intensity.

The second peak occurred about 24 hours after the first and is only little higher, although more than 60 mm of rain fell in between the two peaks. Maximum hourly rainfall rates were, however, much lower than during the first peak (5 vs. 9 mm h\(^{-1}\)). Discharge decreased slowly between peaks and started to rise first about 8 hours after the onset of the second period with more intense rainfall, i.e., about 2 hours after the first rainfall intensity spike. This shows that the delay between rainfall and runoff response can even occur under wet conditions. During
2.5. Discussion

2.5.1 How do alpine landforms with large storage capacity respond to heavy precipitation?

The selected sites were expected to store rainfall long enough to cause a damped runoff response. This was confirmed at all sites. The damped response means that the discharge peaks hours or even days after the storm, and that the runoff rates are substantially smaller than rainfall. The Gadenstetten peak runoff increase was less than 5% of the maximum hourly rainfall intensity, and even less at the time of peak flow in the Schaechen catchment (Fig. 2.7). Such a contribution to catchment flood runoff may be considered negligible, and was observed at all sites with large areal coverage of thick debris deposits: Gadenstetten, Wissenboden-East, Egg, and Oberbutzen. The damped response indicates that drainage rates were low, which also explains why the highest flows from the Egg and Gadenstetten springs occurred in response to long periods of snowmelt (Sect. 2.2.1).

This strongly damped response may cause inflows from upslope areas to be effectively ‘absorbed.’ Even strongly reacting upslope areas may contribute negligibly to flood runoff, as in the Wissenboden-East subcatchment where runoff from areas with little sediment cover re-infiltrated in the thick deposits downslope. This is visible as erosion rills ending on the thick deposits at large distances from any channel connected to the main stream network. The total source area disconnected from the main stream appears comparable to the rapidly reacting area of the Wissenboden-West catchment, suggesting that the total flood response would be twice as strong without the absorption effect of the thick deposits.

The Wissenboden-West subcatchment produced substantial runoff during both small and large events. The response of this strongly reacting area depended mainly on rainfall intensity (Fig. 2.7b). The most intense rainfall period (i.e., B2) caused more discharge than subsequent periods with lower intensity rainfall, illustrating that strongly reacting areas may mask storage dependent contributions from areas with a more damped response.

Such strongly reacting areas may be located close to areas with damped response, as the contrasting tributaries in the Wissenboden subcatchment illustrate. The contrasting runoff responses determine the behavior after the confluence, whereby the total runoff increase may be dominated by the contribution from the fast reacting areas. Similar aggregation of different
Figure 2.8: Discharge response of the Schluecht spring to 28 rain events of more than 10 mm, with the two largest events emphasized. (a) Discharge $q$ from $t_0$ to $t_e + 48$ h. (b) Peak specific discharge increase $\Delta q$ versus accumulated rainfall until time of the peak, specified per driving mechanism (see Fig. 2.6); markers are color-coded according to (a). (c) Event runoff coefficients $C_{R,e}$, also from $t_0$ to $t_e + 48$ h, with $t_e$ indicated. (d) $\Delta q$ versus lag-to-peak, using only the groundwater driven peaks, as indicated with the same markers as in (b).
Figure 2.9: Observed responses at the Schluecht slope to the extreme storms of October 2012 (in grey) and June 2013 (in black); the 10-minute data are referenced to the start of the event, $t_0$. (a) Rainfall $P$ measured at Schluecht. (b) Accumulated rainfall $V_p$ and event storage $S_e$ (dashed lines). (c) Specific discharge $q$. (d) Event runoff coefficient $C_{R,e}$. (e) Soil water content (SWC) in the top 1 m of soil. (f) Groundwater level (GWL) depths in boreholes 1A and 3A (dashed lines) in meter below ground surface (m.b.g.s.). (g) GWL gradients between borehole pairs 4A–3A and 1A–0B (dashed lines). (h) GWL in borehole 0B and head difference between boreholes 0B and 0C (dotted lines; 0B is deeper but has a higher groundwater table).
response types may also occur at the meso-scale, such as is visible in the Schaechen response to the October 2012 event (Fig. 2.7).

At the Schluecht and Lehnstutzbach sites, large differences were observed between small and large events. These differences may be seen as a threshold-like response; contributing little during small events but substantially during extreme events.

The Schluecht creeping landmass slope reacted substantially to moderately intense storms of more than 70 mm, after dry as well as wet antecedent conditions of respectively the October 2012 and June 2013 event. However, peak discharges occurred hours after the main rainfall periods and the maximum discharge increases were only about 10% of the maximum 10-minute rainfall intensities (Figs. 2.8 and 2.9).

The Schluecht spring runoff behavior corresponds to groundwater fluctuations in the fractured rock (Figs. 2.7 and 2.9). The groundwater system appears to be the sole driver of the runoff response as no perched saturation was observed in any of the piezometer nests, and soil moisture dynamics correlated little with spring discharge. The relatively rapid depletion of event storage indicates that drainage from the fractured rock may become an import runoff source during long-duration storms. Such storms must deliver at least moderately intense precipitation to cause a substantial contribution from a slope like Schluecht: the delayed increase in runoff in response to the second rainfall period of the June 2013 event shows that discharge only increases further if rainfall is more than the few millimeters per hour that fell at the start of the second rainfall period. As the time lag even occurs under wet conditions, it is expected that the runoff response will always be much below peak rainfall intensities.

Apart from the runoff formation occurring in the fractured rock, the storage capacity of the soil cover also exerted a relevant damping control on the runoff response. The damping results from the time required for water to pass through the soil and the associated short-term storage. The time lag is visible in the delayed response of the deeper soil moisture sensors, even after the soil had wetted up substantially (Fig. 2.7). The effect of short-term storage in the soil is visible in the comparison of the two most extreme events (Fig. 2.9), where the first 20 mm of the October 2012 storm caused a strong increase in SWC and minor groundwater and discharge responses, compared to similarly intense periods later in the storm or during the June 2013 event. This buffering capacity of the soil appears particularly effective during small events or short storms, when it limits input to the groundwater.

Antecedent conditions may affect the runoff response, since moist soil has less storage capacity left, and drains faster. This relation is confirmed by the fact that the SWC increased less during the June 2013 event than during the October 2012 event, despite higher rainfall intensity and total rainfall depth in the former. However, soil moisture decreased more rapidly than groundwater tables, suggesting that the groundwater system has a longer ‘memory’ and is therefore more sensitive to antecedent conditions. This difference in drainage time scales may also explain why the groundwater peaks can occur days after the storm, when drainage from the soil has diminished.

The soil cover thus appears to play a more secondary role in both damping of the stormflow and the slope’s sensitivity to antecedent conditions. During long-duration storms, when the permeable soil provides only a little extra storage because rainfall and drainage occur at similar
2.5. Discussion

rates (i.e., storage is exhausted), damping of the runoff response occurs predominantly in the fractured rock groundwater system.

**Stormflow generating mechanisms at the Schluecht slope**

The strong response to extreme storms of the Schluecht slope is reflected by the strong nonlinearity in the storage-discharge relation as found from recession analysis (Volze, 2015). It may be hypothesized that such a nonlinear response is caused by a physical threshold in the fractured rock system. For example, permeability of fractured rock could decrease markedly at some depth, causing a considerable transmissivity increase when groundwater tables rise beyond this permeability contrast (i.e., transmissivity feedback mechanism, e.g., Rodhe, 1989; Kendall et al., 1999).

Given that groundwater tables can rise several meters during an event, transient saturation of a more permeable fractured rock zone appears feasible. Such ‘activation’ of fissures would also explain the erratic rising of the groundwater table at borehole 4A (Fig. 2.7g). The hypothesis is also supported by the groundwater table gradient being steeper between the deep boreholes 3A and 4A than between the shallower boreholes 1A, 2A and 3A; a steeper gradient generally points towards lower permeability. Furthermore, assuming Darcy’s law can be used to describe groundwater outflow at the spring, a 2.5-fold increase in total transmissivity may be inferred; the 5-fold increase in discharge was associated with only a 2- to 3-fold increase in groundwater table gradient at the base of the slope (Fig. 2.9). Such a transmissivity increase suggests a much higher permeability at shallow depth, as the total depth of the groundwater system may be orders of magnitude larger than the ~2 m rise in borehole 2A (Sect. 2.2.2).

On the other hand, the few estimates of hydraulic conductivity at different depths did not point out any strong contrast (Volze, 2015), and no mechanistic reasons for the formation of such zone could be identified. The smooth fluctuations of discharge and driving groundwater table gradients near the spring may even argue against a quick change in permeability. In addition, the head difference between boreholes 0B and 0C (near the point sources) may increase more strongly than runoff (up to a factor 10; Fig. 2.9). This suggests that an increase in local pressure gradient could also be an important driver of the strong runoff increase. Understanding the relative importance of the pressure gradient increase and transmissivity feedback mechanisms requires further study.

It may be suggested that empirical relations may reveal useful knowledge of first-order controls on runoff formation. For example, Montgomery and Dietrich (2002) pooled observations of recession constants and lag-to-peak times to characterize the hillslope response time scale. They found that the topographic gradient has no effect on the subsurface stormflow response and suggested that vertical unsaturated flow properties predominantly determine the response time scale, as the topographic gradient must affect lateral saturated flow.

However, our observations at the Schluecht slope indicate that changes in properties of the medium where lateral flow occurs determine the hillslope response. Soil water content and its drainage (i.e., value and rate of change in SWC, Fig. 2.7) peaked earlier than spring discharge and there is little reason to expect much additional delay by the percolation through the fractured rock. Besides, vertical flow paths constitute only a minor part of the total flowpath length...
for most of the slope, such that variation in groundwater depth may hardly explain the large variability in lag-to-peak times. Unfortunately, lateral processes are much harder to describe than vertical processes, which would explain the emphasis on the latter in many process studies.

It may also be questioned whether lag-to-peak time, as used in the Montgomery and Dietrich (2002) study, is an appropriate measure of hillslope runoff response time. For instance, the lag-to-peak times at the Schluecht slope vary over more than an order of magnitude, ranging from a few hours to more than four days, the latter well in excess of the up to 40-hour time lags reported in Montgomery and Dietrich (2002). Besides, rainfall intensity spikes may affect the timing of the maximum peak (see the October 2012 event).

Alternatively, it may be suggested that the magnitude of the specific discharge peak may be a suitable measure of response strength. This can be illustrated by the difference in behavior of the Schluecht and Gadenstetten slopes for which catchment area could be determined relatively accurately: the timing of the discharge peaks may be uncertain, but the damped response of the Gadenstetten spring is consistently reflected by its smaller peak runoff increase.

Even though it remains an open question which properties of the fractured rock system determine the Schluecht runoff formation, the above research and the findings of Volze (2015) summarized in Section 2.2.2 make clear that the fractured rock at the base of the slope determines the stormflow response. This is the zone where the actual runoff formation takes place and where groundwater level fluctuations are strongest. The response can provide a substantial contribution to flood runoff formation during long-duration events: the delayed rising limb of the hydrograph may occur within the storm event and activation of faster drainage mechanisms requires a large rainfall sum.

### 2.5.2 How much water is stored at event time scales?

Calculation of event storage requires relatively accurate estimates of precipitation, discharge, and catchment area. Such estimates could be obtained for the Schaechen catchment and the Wissenboden, Gadenstetten, and Schluecht sites for the October 2012 event. The estimates are discussed and compared to other studies of storage in mountainous catchments.

**Estimates of event storage at the sites**

In spite of the markedly different runoff responses and event runoff coefficients, event storage developed similarly among the different sites (Fig. 2.7). Only the strongly reacting Wissenboden-West subcatchment may have discharged more water within the storm than it stored, but this could not be demonstrated due to clogging of the EC sensors. Despite the strong runoff response, much of the first rain is stored for at least a short time: ~32 mm of the 47 mm of rainfall that fell in the analyzable period was in storage. This appears feasible as it is comparable to storage at the Schluecht (34 mm) and Gadenstetten sites (29 mm), but is well below the 45 mm stored in the Wissenboden-East subcatchment. Even areas with fast response may thus store most of the rainfall during small events.

The large storage in the Wissenboden-East catchment can be responsible for the damped response observed after the confluence with the five time smaller Wissenboden-West catchment.
This explains the similarity in developments of $C_{R,e}$ at Wissenboden and Schaechen, in spite of the Wissenboden subcatchment receiving much more rainfall. This resulted in a storage increase at Wissenboden of 108 mm at the time the Schaechen peak flow occurred and 121 mm at the end of the storm; 88 % and 84 % of the rainfall. Event storage in the Schaechen was about 63 mm at the flood peak, comparable with the 70 mm stored in Schluecht and the 63 mm in Gadenstetten. Despite its more damped response, computed $S_e$ at Gadenstetten was less than at Schluecht, which is explained by the larger rainfall sum in Schluecht (104 mm vs. 95 mm).

Maximum event storage at the Schluecht slope for the larger storm of June 2013 was estimated at 112 mm (Fig. 2.9). This comprises 78 % of cumulative rainfall, comparable to the stored water fraction during the October 2012 event.

At all sites, maximum storage increase rates exceeded maximum depletion rates (i.e., the slopes of the $S_e$ curves presented in Fig. 2.7). This suggests that similar storm water fractions are stored during more extreme events because the drainage mechanism is not expected to differ from event to event (always deep subsurface processes) and drainage rates remain below rainfall intensity.

Storage depletion at the Wissenboden subcatchment during the first 48 hours after the storm was only 6 mm d$^{-1}$, whereas storage depletion at the Schluecht slope was about 9–10 mm d$^{-1}$ during both the October 2012 and June 2013 events. This difference is also visible in the recession limbs of the sites (Fig. 2.7b–c) and causes the $C_{R,e}$ of the Schluecht site to exceed its Wissenboden counterpart within days after the event (Fig. 2.7e). Furthermore, the similarity between the Gadenstetten and Wissenboden depletion rates may indicate that thick debris deposits release storm water slower than soil mantled (fractured) rock slopes like Schluecht. The continuing quick rise in $C_{R,e}$ of the Schaechen suggests that such slowly draining slopes do not predominate here.

The fast drainage of the permeable soil at the Schluecht slope causes the near-synchronous fluctuations of rainfall and soil moisture, and the relatively fast drop to pre-event conditions after rainfall has ceased. As a rule of thumb, permeable soils may stop draining within few days after major storms, whereby the remaining moisture (held against gravity) is referred to as field capacity (e.g., WMO, 2008). The soil cover at Schluecht appears to follow this rule. Even so, the soil provides substantial short-term storage, even when the soil is already wet and rainfall intensity is low, like during the June 2013 event (Fig. 2.9, first peak of ~30 mm increase in SWC, almost 50 % of rainfall input). This short-term storage helps to prevent drainage rates from approaching the high rainfall intensities that occur only for short periods of time (intensity spikes).

The computed SWC values may yield some overestimation of event storage in the soil as it was observed that they can exceed short-term rainfall depth, for example, at the start of the October 2012 event, when the soil was relatively dry. However, the increases in SWC were not more than the storm rainfall depth, indicating that measurement errors do not hinder qualitative interpretation. Besides, the lack of groundwater response until the second half of the B1 period (Fig. 2.7g) also indicates that there is considerable water storage in the soil.
Plausibility of the storage estimates

Above estimates indicate that more water is stored than discharged during the storm and the following 48 hours at all of our sites. This may be interpreted as storage, rather than runoff formation, being the dominant hydrological process, something that might be regarded as counterintuitive for such steep terrain. However, event runoff coefficients smaller than 50% are often found in other parts of the Alps (e.g., Merz et al., 2006; Norbiato et al., 2009), indicating that many slopes store most of their precipitation inputs. The behavior of our sites should thus not be viewed as exceptional. The sensitivity analyses discussed below indicate that the event storage estimates are robust and that our sites are typical examples of mountain slopes where event time scale storage causes substantial damping of the flood response.

Errors in calculation of the event runoff coefficient are multiplicative; a 10% higher runoff causes a 10% higher $C_{R,e}$, and the combination of 20% less rainfall with a 20% higher specific discharge causes a 50% lower $C_{R,e}$ ($1.2/0.8 = 1.5$ times the 'original' value; its reciprocal thus yields a 33% decrease). Even with errors of about 20%, none of the sites’ $C_{R,e}$ exceeds 0.3 at the end of the storm, and only the Schluecht and Schaechen values may approach 0.5 within the following 48 hours. Even with such large errors, it thus remains valid that at all sites more rainfall is stored than discharged during the event, and that this may only reverse days after the storm.

Estimates of the relatively small catchment areas of the Schluecht and Gadenstetten springs are conservative (Sect. 2.3.3) and rainfall is measured relatively accurately, such that total storage is likely underestimated in above computations. The Schaechen and Wissenboden catchment areas are likely overestimated, because of karstic systems draining parts of the topographic catchment areas to neighboring basins. About 5.1% of the Schaechen catchment was estimated to drain to other catchments, and even 23.8% (28.5%) of the Wissenboden (-West) subcatchment (mapping procedure discussed in Sect. 3.5). Associated storage overestimation at the end of the storm is however relatively small and does not affect the above conclusions: 1 mm for the Schaechen catchment and 8.4 mm for the Wissenboden catchment.

Likewise, uncertainty in the storage estimate due to the simple baseflow separation (constant baseflow since $t_0$, Sect. 2.3) is comparatively small. A further extrapolation of the receding limb until $t_0$ would typically approach zero asymptotically (e.g., exponentially), such that a linear extrapolation gives the lowest feasible baseflow and largest event runoff (coefficient). Linear fits to discharge measured in the 6 hours before $t_0$ indicated a decrease rate $\frac{dq}{dt}$ between $0.8 \times 10^{-4}$ (Schluecht) and $23 \times 10^{-4}$ mm h$^{-2}$ (Schaechen). The resulting difference in event storage, $\Delta S_e$, increases parabolically with time since $t_0$ (i.e., $\Delta S_e = \frac{1}{2} \frac{dq}{dt} (t - t_0)^2$), causing an overestimation of at most 6 mm over the first 72 hours after $t_0$ (Schaechen), less than 10% of the event rainfall depth.

Comparison with other studies

The event runoff coefficients found for the Schluecht slope are similar to those found at other soil-mantled fractured rock slopes. For example, Montgomery et al. (1997) report total event runoff coefficients (until discharge has returned to pre-event state) between 0.38 and 0.49 for the Coos Bay 1 hillslope. The 43 slope with permeable sandstone bedrock covered by 0.5 to 2 m of colluvial silty-sand soils appears similar to our Schluecht site, where (conservative)
estimates showed $C_{R,e}$ increased only towards 0.5 at the end of the analysis period. Tromp-van Meerveld and McDonnell (2006a) reported even lower runoff coefficients (max. 0.27) for the gentler Panola hillslope (a 13 slope of granitic bedrock mantled with on average $\sim 0.63$ m of permeable sandy loam soil). Neither study explicitly states how $C_{R,e}$ was computed, but these comparisons suggest that our estimates at the Schluecht slope are not extraordinary. No studies reporting event runoff coefficients of slopes with thick debris cover could be found.

Unlike $C_{R,e}$, event storage $S_e$ estimates allow a direct physical interpretation, e.g.: where can such water volume reside? Since event rainfall depths in the Alps are only in the order of few hundred millimeters, all of a storm’s rainfall may well be stored in thick deposits or fractured rock systems. For example, fractured rock with an effective porosity of 5%, like in Schluecht, may store 100 mm per 2 m saturation depth. This corresponds well with the few meters rise in the groundwater table observed during the June 2013 event (Fig. 2.9). Similar event storage is feasible in the thick debris deposits since the porosity of such sediments is generally higher (e.g., Gelhar et al., 1992; Fitts, 2002). On the other hand, 200 mm of new water may hardly be stored in 1 m of soil without fully saturating it (e.g., if 20% out of a 40% total porosity was saturated before the storm). Indeed, the soil cover at the Schluecht site did not saturate and showed only a SWC increase of 30 to 50 mm.

Estimated event storage may also be compared to published storage quantities estimated from water balance evaluations over similar as well as different time scales. In their study of storage fluctuations in 17 catchment with relatively thick soils (75–180 cm) on short, steep slopes (~75 m; ~45) of weathered and fractured sedimentary rock that constitutes the main medium where runoff generation occurs, Sayama et al. (2011) found seasonal fluctuations of up to 652 mm (mostly 200–500 mm). They also found that the maximum storage increase, as well as the “storage threshold for activation of stronger runoff response,” correlated positively with the catchment median topographic gradient. They explain this by more water being needed in steeper slopes to cause a similar spatial extent of (nearly-)saturated area where rapid lateral flow may be generated. This could also explain why they found faster recession below than above the identified storage thresholds in the steeper catchments, a difference they did not observe in catchments with gentler slopes. Their study thereby demonstrated that large storage is feasible in steep slopes underlain by fractured rock. Moreover, similar events as studied at Schluecht often caused similar increases in storage of up to about 100 mm, supporting the hypothesis that such steep slopes with fractured bedrock may substantially dampen the rainfall-to-runoff signal.

In the alpine Colorado Front Range, Clow et al. (2003) estimated that groundwater storage change of a large talus cone from spring discharge recession over the winter season may be as much as 5.6 m (11.2 m of talus aquifer with a porosity of 50%). Such a storage depth is possible, as the talus depth was estimated at more than 18 m, but is much larger than the annual rainfall of 1100 mm. As they found that talus systems are responsible for much of the streamflow throughout the season, it may be expected that the actual catchment area of the spring is larger than the talus cone itself. The cone would then receive much larger inflows than local precipitation, for example from the bare rock faces above the talus. Such ‘extra’ storage is also expected for the Wissenboden-East catchment, and might be relatively common in alpine terrain.
In a recent contribution, Hood and Hayashi (2015) estimated a peak storage variation in an alpine headwater basin of up to 95 mm, from a water balance evaluation of two melt seasons (Canadian Rocky Mountains). Such a maximum seasonal storage increase is smaller than the event-based estimates presented for our sites. However, Hood and Hayashi (2015) interpreted that most water is stored in a proglacial moraine that appears similar, albeit younger, than the thick deposits that make up much of our Wissenboden-East, Gadensstetten, Egg, and Oberbutzen sites. In their case, the moraine deposit covers less than 10% of the 4.7 km² catchment, implying that storage may locally be much greater than the reported catchment average. Their study thus suggests that event storage in thick debris deposits may be more than 100 mm.

2.5.3 Which landscape characteristics determine the damping of hillslope-scale runoff formation?

It appears that formation of lateral flow at all investigated sites occurs at large depth (> 1 m) and that damped responses may be predominantly determined by the properties of a deep groundwater system. This large depth limits the possibilities to identify the exact mechanisms in situ and thus also the possibilities for characterizing the observed differences between the sites. It also calls for hillslope-scale characterization since this is the scale where the complex interplay of processes determines the runoff formation (Sect. 2.5.1). The discussion below therefore mainly concerns descriptions of landscape characteristics at the hillslope(-segment) scale.

Further, note how our observations at the Wissenboden headwater illustrate why the subcatchment scale is too coarse for relevant characterization, the plot scale is too small, and the hillslope scale may be a suitable compromise. The subcatchment scale is too coarse because of the strong contrasts between the slopes of the eastern and western tributary catchments. On the other hand, spatial plot-scale variability may be large and hardly relevant for the total runoff contribution at the base of a slope. For example, the thickness of soil or debris cover may vary strongly, like in the talus slopes, but play only a minor role since the lateral drainage properties in the deep subsurface determine the runoff response (Sect. 2.5.1).

Infiltration and percolation

The large depths of the zone where lateral drainage occurs at our sites warrants that local rainfall cannot fully saturate the column. The fraction of available storage capacity that is used during even the most extreme event may be small, particularly at the sites with thick debris cover. The Schluecht observations indicate that even a permeable soil cover may store a substantial portion of event rainfall long enough to help dampen the runoff formation process. However, the Schluecht soil could not account for the total event storage during long-duration storms (Sect. 2.5.2). The same may be expected for the relatively shallow soils covering large parts of the thick debris deposits that dominate at the higher-elevation sites. Permeable soil is thus mainly important for the ‘smoothing’ of the rainfall-to-percolation signal that forms the input to the lateral drainage processes at larger depth. This smoothing distinguishes slopes with permeable soils from areas where infiltration or saturation excess overland flow can cause an immediate runoff response.
Further smoothing may also occur during deeper percolation processes in fractured rock or loose material. Given the large variety of percolation depths and associated flowpaths and travel times, even very permeable media may exert such smoothing. However, there may be little storage effect in the deep unsaturated zone; inter-storm moisture content may fall only slightly below what can be held against gravity because the upward ‘loss’ fluxes are limited (unlike in soils, where evaporation processes further deplete soil moisture). Yet, the limited storage may be offset by the large depth over which it may work; for example, 5 mm per meter depth may provide 50 mm of storage over 10 m. It is unknown whether as much as 5 mm per meter of short-term storage may occur (and particularly for fractured rock this may be challenged), but it is here used to illustrate how percolation depth could constitute an important damping control on runoff formation.

The percolation storage time scale is expected to be on the order of minutes to days. The lower estimate stems from the maximum film flow velocity of $\sim 120 \text{ m h}^{-1}$ reported by Tokunaga and Wan (1997). The upper estimate matches the common assumption that field capacity is reached within days after a major storm (e.g., WMO, 2008). This wide range of time scales might co-occur within a column, such that this ‘dispersion’ may provide an additional smoothing mechanism.

Damping during the percolation process will be further enhanced if sufficient fine-grained sediments are available to decrease permeability and increase retention capacity. Since glacial deposits often have substantial fine-grained sediment content it may be expected that the thick debris deposits at our sites consist of mixed coarse- and fine-grained particles. This particularly holds in the areas where soils have developed, suggesting that much of the space between coarse sediments is filled with fine-grained material.

Several studies report that debris deposits without soil cover may have zones with substantial fine-grained material at depth, for example buried soils or moraine deposits, or sediments washed in by rainfall, snowmelt, or debris flows (Yair and Lavee, 1976; Pierson, 1982; Clow et al., 2003; Schrott et al., 2003). All these processes can co-occur anywhere in a slope, but may be of most interest at the base, ie, where the actual runoff formation takes place. For example, the absence of coarse, openwork sediments at the relatively flat bases of our sites with thick deposits suggest substantial fine-grained content, either because they were deposited there directly, or because recent deposition rates exceeded erosion rates. The opposite is sometimes evident at the the bases of steep talus slopes; little fine-grained material gets deposited here directly, and erosion patterns indicate that episodic outwash of the washed-in sediments may occur.

Occurrence of fine-grained material may also affect lateral flow processes, making it hardly possible to distinguish the damping effects on the percolation and lateral drainage processes. However, even without such understanding, the hypothesized importance of fine-grained material for damping of the hillslope-scale runoff response may be well confirmed from the following comparison of published studies. For example, the few studies of alpine talus slopes (e.g., Pierson, 1982; Williams et al., 1997) and moraine deposits (e.g., McClymont et al., 2011; Hood and Hayashi, 2015) with substantial fine-grained sediment content all report relatively long drainage time scales. In contrast, Muir et al. (2011) report quick response and drainage of a talus with little fine-grained material (exponential decay coefficient of flow recession of
Observations of damped runoff behavior

~1.0 d$^{-1}$). This suggests that even a first-order, rather qualitative assessment of fine-grained material content may help to distinguish large differences in runoff response of thick debris deposits.

Lateral drainage

Assuming lateral drainage from the groundwater body follows Darcy’s Law, the outflow is determined by the product of the groundwater head gradient, the cross-sectional area (at the ‘outlet’) and a permeability term. Of these three quantities, permeability is most uncertain because it can vary over the most orders of magnitude. Still, without accurate a priori estimation of the permeability, which might be virtually impossible in any terrain, some qualitative relations can be deduced.

For example, at very high permeability, a small increase in cross-sectional area or head gradient already gives a strong response, such that groundwater fluctuations may be little. This may be expected for permeable, coarse debris, as Muir et al. (2011) found for a talus with little fine-grained material. Event storage is relatively small since the rapid drainage prevents ‘stocking up’ much water.

Such a high permeability seems unlikely for the (strongly) damped sites presented in our study. Instead, groundwater head and cross-sectional area may vary substantially if enough water is provided, as was observed for the strong response to extreme events at the Schluelscht slope. The possibility of a quick, strong groundwater rise may therefore be regarded as an important indicator of the responsiveness of a system. The strong rise at the Schluelscht slope appears to result from low porosity combined with a low drainage rate, causing a ‘stocking up’ of water at the base.

Low porosity and drainage rates may be typical for heavily fractured rock systems, and these may be common for deep-seated gravitational slope deformations (DGSDs) like the Schluelscht slope: the deformations can cause fissuring between individual rock masses as well as inside the rock masses. In the Schlaechen, most areas identified as creeping landmasses (Bruckner and Zbinden, 1987) have little incised terrain, suggesting that stormflow generation occurs predominantly in fractured rock—as at the Schluelscht slope. Occurrence of heavily fractured rock is however not limited to DGSDs, as evidenced by the many studies stressing the importance of fractured rock for stormflow generation (see Sect. 2.1.2). Understanding of landscape formation processes may help in recognizing slopes with heavily fractured rock and associated deep storage and drainage mechanisms.

The absence of strong runoff responses at the sites with thick debris deposits suggests that the groundwater tables around the drainage locations (springs) vary little. Various causal mechanisms could be identified. They may all occur together, and ranking their importance is not attempted here.

For example, the porosity in the deposits may be higher than in the rock of the Schluelscht slope, or even fractured rock in general (e.g., Gelhar et al., 1992 Fitts, 2002), which is less favorable for strong groundwater table fluctuations. Also, the slopes have a relatively flat base that appears to reflect a relatively flat underlying bedrock. This suggests a relatively moderate groundwater table gradient, and consequently, low flow velocities, which may also prevent
rapid accumulation of groundwater in the vicinity of the spring. The comparatively deep per-
colation depth and associated smoothing discussed above, as well as the slopes being relatively
long, also favor damped runoff responses.

Factors that promote a strong response are mostly missing at our sites: a) there are no indica-
tions of extensive layers with limited permeability, b) runoff generated in strongly reacting
areas re-infiltrates at large distance from the channel heads, and c) the terrain has little concav-
ity, suggesting that convergence of flowpaths may not cause rapid accumulation of water at the
base. The importance of these factors could thus not be evaluated, but they may be contribut-
ing mechanisms at other sites.

The Lehnstutzbach is dominated by creeping landmass areas and glacial till. The till may be
more consolidated and better sorted than in the thick deposits at the other sites: it is older and
fluvial processes have likely deposited layers with high fine-grained sediment fractions (silt
and clay). This suggests that the permeability of the glacial tills is lower than at the other sites,
but this hypothesis was not tested.

**Topographic gradient**

Topographic gradient alone may be of little use for characterizing areas with (strongly) damped
runoff response. The Schluecht and Gadenstetten slopes have similar topographic gradient
but contrasting sediment cover and runoff behavior. Other studies in mountainous terrain also
report little correlation between average topographic gradient and catchment flood runoff
(e.g., Pitlick, 1994 Merz and Blöschl, 2005 Merz and Blöschl, 2009) or the hillslope-scale runoff
response (Montgomery and Dietrich, 2002). This may be of little surprise, given that hydraulic
conductivity and head gradient are equally important for saturated groundwater flow (i.e.,
Darcy’s Law), but the former typically varies over many more orders of magnitude than the
topographic gradient, which is often used as proxy for the latter. Nonetheless, topographic
gradient data may be informative when combined with other data, for example to distinguish
different types of moraine or talus deposits.

**2.5.4 Implications**

The presented study may be the first to identify steep alpine slopes that contribute little or
even negligibly to flood runoff formation due to substantial (i.e., > 100 mm) event time scale
storage. This study also indicates that the soil covers at such sites, if existing at all, may play
only a secondary role during extreme events of long duration due to their limited storage
capacity. This knowledge is relevant for understanding runoff formation processes in moun-
tainous catchments, since slopes that are geomorphologically similar to our sites are common
in alpine terrain. The knowledge of where such damped responses occur may aid in estima-
tion of design floods beyond purely statistical techniques, help explain differences in the flood
behavior of catchments, and may constitute an important step towards a realistic perceptual
model of the spatial organization of runoff formation processes.

These findings challenge three common perceptual models of runoff formation is steep ter-
rain. Firstly, that steep slopes store little water and produce runoff quickly. Secondly, that the
runoff response is mainly determined by the properties of the soil cover. And thus thirdly, that
the combination of topographic and soil data alone may constitute a meaningful description of the spatial organization of hydrological processes, for example through simple relations between soil properties and slope or elevation. In fact, our steep slopes with large storage capacity and damped runoff response were found at up to 2000 m.s.l. (i.e., Wissenboden-East catchment), and similar slopes can be found at even higher elevations in other areas.

Geomorphological maps may provide valuable information about sediment cover type and depth as well as the occurrence of strongly fractured bedrock that complements the topographic, soil and land use data that mainly yield information about the surface and shallow subsurface. It is therefore suggested that geomorphological maps provide a useful basis for spatial characterization of hydrological processes in alpine terrain. Such characterization, combined with aerial imagery, may also help to identify strongly reacting areas that are separated from the channel network by areas with large storage capacity, which may locally affect flood runoff substantially (see Wissenboden-East). These types of characterization are further explored in Chapter 3.

A threshold-like runoff response was observed at the Schluecht and Lehnstutzbach sites. Such behavior could translate to the catchment scale if similar threshold levels occur in a large part of the catchment (e.g., Rogger et al., 2012, 2013). Our observations at the Schluecht slope indicate that the effect of threshold-like behavior depends not only on how much water is needed to activate the stronger response, but also on the response time and storm duration. These time scales determine whether a slope contributes to the rising or the falling limb of the catchment flood hydrograph. This appears little studied for subsurface flow processes: reports of hillslope-scale threshold-like stormflow formation often only study thresholds in the storage state (e.g., Montgomery et al., 1997; Tromp-van Meerveld and McDonnell, 2006a, b).

Understanding the time scale effects is also important for developing process-based models. For example, the delayed timing of the peak flow cannot be described by a model that considers only a static storage threshold, instead of a dynamic balance between in- and outflows of the system responsible for runoff generation. Many simple model structures may have this deficiency.

The relations between different hillslope properties and the runoff response could not be fully clarified. This is partly due to the limited detail at which subsurface properties can be characterized; direct observations of relevant features may only be possible at the point scale and some properties, like geometry of the fractured network in the rock, can simply not be measured non-destructively. Further research into relations between slope properties and response may therefore better focus on comparisons of slopes that are similar in many but not all respects, and from this infer the relative importance of different properties.

On the other hand, the large differences in mountain slopes suggest that a crude classification of runoff response strength may be possible without complete understanding of the runoff generating mechanisms. The resulting map may facilitate a useful comparison of catchment-scale runoff responses and possibly allow prediction of the flood behavior. For example, the peculiar flood behavior of the Schaechen could be explained by threshold-like responses, such as observed at the Schluecht slope, if an adequate mapping technique indicates similar re-
2.6 Conclusions

Storage and drainage processes in typical mountainous landforms with large storage capacity are not well understood, but knowledge of the streamflow contributions from these areas is important for predicting low and high flows. It was hypothesized that some of these landforms produce little runoff during small events but substantially more during extreme events. Such threshold-like behavior, if it occurs extensively, could explain the much higher peak flows of the four largest floods in the Schaechen catchment (Sect. 1.2).

To investigate this hypothesis, the runoff response was monitored at six sites in the Schaechen that are exemplary of commonly found landforms for which a damped runoff response may be expected. The responses to the most extreme event for which reliable discharge data could be obtained were analyzed. In addition, the Schluecht site, showing threshold-like behavior, was studied in more detail.

All sites showed a damped runoff response; flow peaks were considerably lower than rainfall intensities and the runoff peak occurred hours to days after the storm. Groundwater systems at many meters depth appear to be responsible for the runoff generation. The large storage capacity of these systems explains the damped runoff response. Estimates of event storage indicated that most of the rainfall remained in storage until days after the storm. These storage estimates are relatively robust and comparable to storage volumes reported for similar landscapes.

The runoff response from sites dominated by thick debris deposits showed such a strong damping that the contribution to even the most extreme events may be considered negligible. Such areas can effectively absorb the runoff generated in upslope areas and thereby determine the runoff response of an area that is much larger than the spatial extent of the debris deposit itself. Furthermore, it was demonstrated that flashy behavior caused by a small area with a strong runoff response could mask the damped response of the rest of the catchment; the combined area may produce a rapid increase in runoff, but the peak specific discharge is low (e.g., at the Wissenboden site).

Further investigations at the Schluecht site, showing threshold-like behavior, indicated that the ~1 m of permeable soil provides important storage capacity during small events that may be exhausted during extreme storms of long duration. The heavily fractured rock system below the soil cover may provide longer-term storage and the strong groundwater table rise during extreme events causes the comparatively strong runoff response. The threshold-like behavior may originate from a combination of two mechanisms: activation of faster drainage above some storage threshold and the response time falling within the storm. The strongly fractured bedrock may be typical for the creeping landmass slopes that dominate the northern flank of the Schaecheen catchment.

To our knowledge, this is the first study demonstrating such damped responses for steep alpine slopes. This finding has important consequences for the understanding of the spatial organization of runoff formation processes, particularly since the investigated landforms are ubiquitous in alpine landscapes. Hillslope-scale properties were identified that could be indica-
tive of damped runoff generation. These indications aid in the development of a mapping tool for characterizing the strong contrasts in runoff response we demonstrated to exist in alpine terrain.
Chapter 3

A tool for mapping dominant runoff processes in alpine terrain

3.1 Introduction

The aim of this chapter is to present the mapping tool developed in this PhD research project and interpret the resulting map for the Schaechen. The development was guided by improved understanding obtained through the field experimentation in the Schaechen catchment presented in Chapter 2, complemented by classification concepts based on existing approaches, discussed in this section, and accepted hydrological principles, discussed in Section 3.2.

The conclusions from our field observations (Chap. 2) that are most important for the development of the mapping scheme are:

1. Relevant runoff formation processes may occur at meters or even tens of meters depth, as was shown for the Schluecht and Gadenstetten slopes.

2. Storage potential and associated runoff response may vary greatly between mountain slopes.

3. Short-term storage mechanisms (i.e., at the event time scale) determine the damping of catchment flood runoff response.

4. Areas with thick debris cover may have a strongly damped runoff response and contribute little to extreme floods, as observed at for example the Gadenstetten slope.

5. Some slopes may contribute mainly during the most extreme events because they are not activated during small events, or react with such delay that generated runoff contributes only to the flood recession, or both, as observed at the Schluecht site.

6. Situations where quickly generated runoff re-infiltrates in areas with damped runoff response may be common in mountainous terrain.

The hypothesis was formulated that the peculiar flood behavior of the Schaechen, the four most extreme floods being much larger than would be estimated from the smaller floods only, could be explained by a large areal fraction of slopes that only contribute considerably during the most extreme events (Sect. 1.2).
Testing this hypothesis requires evaluating whether many such slopes exist in the Schaechen, i.e., some form of classification of hillslope runoff behavior. As runoff behavior cannot be measured throughout the landscape, a method was sought for mapping slopes with similar runoff behavior based on hillslope-scale properties relevant for runoff formation, such as sediment cover type, water storage potential, and drainage pathways. The underlying hypothesis is that similar hydrological behavior may be expected if similarity of hillslope properties is classified in a hydrologically sensible way.

This ‘similarity hypothesis’ sets the context for how the word ‘expected’ is used here; from clearly defined similarities and differences, some qualitative, and maybe even quantitative, similarities and differences in flood formation are ‘expected.’ The ‘similarity hypothesis’ can be tested by comparing mapped and observed runoff behavior. I do this here qualitatively for the Schaechen and its two main tributaries, and the study sites presented in Chapter 2; quantitative evaluations with a newly developed rainfall-runoff model are presented in chapter 4 for the Schaechen catchment and in chapter 5 for two catchments whose flood runoff behavior contrasts strongly with that of the Schaechen.

3.1.1 Literature

Various hydrological mapping approaches exist or are implicitly used for deriving hydrological response units in (semi-)distributed rainfall-runoff models. The few approaches that were explicitly developed for description of flood runoff generation mechanisms in mountainous terrain are discussed here. The classification scheme developed in current research tries to explicitly classify dominant runoff generating mechanisms. Such a classification of dominant mechanisms may help to develop models with a realistic structures and to better constrain important parameters (e.g., Uhlenbrook et al., 2004; Beighley et al., 2005; Devito et al., 2005). It is thereby usually assumed that the dominant runoff formation process of an area does not change over time.

Approaches recognizing the dominant runoff generating mechanisms were developed for catchments in the Ore Mountains by Peschke and co-workers (e.g., Peschke et al., 1999; Zimmerman et al., 2001) and in the Black Forest Mountains by Uhlenbrook (1999, 2003) and Waldenmeyer (2003). These methods involve rule-based classification in a GIS using detailed spatial data obtained through extensive exploration. The resulting maps distinguish areas with infiltration and saturation excess overland flow, areas with various types of subsurface storm-flow, and areas contributing mainly to baseflow. Although these methods were developed for mountain ranges in Germany that are lower than the Alps, and formed by different processes, the rule-based interpretation of typical landscape elements, like debris and glacial till deposits, helped to define relevant landforms for the mapping tool developed in this thesis.

A different approach was developed by Markart et al. (2004), based on a large collection of plot-scale sprinkling experiments in various parts of the Austrian Alps. With a focus on floods caused by intensive precipitation, instantaneous runoff coefficients are mapped from detailed knowledge of vegetation cover, the top soil, and shallow runoff processes. These values are useful for reference, but their relevance for long-duration storms is unclear. Nonetheless, these maps identify rapidly contributing areas and, after careful interpretation, form a useful basis for a spatially distributed modeling. For example, Rogger et al. (2012) combined such
3.1. Introduction

Maps with other spatial data sets in two meso-scale Alpine catchments to define eight different hydrological response units (HRUs) that could be used with the spatially distributed model presented in Blöschl et al. (2008). The model indicated that exceedance of storage thresholds may cause a “step-change” in the flood-frequency curves of the two catchments; similar behavior was also identified in the Schaechen in Section 1.2.

Tilch et al. (2002) used generally available spatial data in a purely GIS-based classification technique, trained on the dominant runoff process map of the Brugga catchment (40 km²) obtained by Uhlenbrook (1999). The mapping technique was transferred to the contiguous Dreisam basis (258 km²) that encompasses the training area. A similar technique was developed by Tilch et al. (2006) for the Loehnersbach catchment, Tyrolean Alps, whereby slopes with thick debris cover were classified according to landscape formation processes.

Here, it was decided to develop a classification scheme in line with the dominant runoff process (DRP) classification schemes of Scherrer and Naef (2003) and simplifications of the schemes by Schmocker-Fackel et al. (2007). The schemes are hereafter referred to as SN-schemes. The decision rules and concepts of the SN-schemes are based on a large data set of plot-scale sprinkling experiments conducted by Scherrer (1996). Classification rules were developed for mapping different “runoff types” without explicitly specifying the dominant mechanism (e.g., VAW, 1994; Naef et al., 1999).

The method of Scherrer and Naef (2003) distinguished four dominant runoff generating mechanisms. Hortonian Overland Flow (HOF) occurs when rainfall intensity exceeds the infiltration capacity (e.g., Horton, 1933; Beven, 2004). This mainly occurs in areas with low infiltration capacity, like soils with high clay content or rock surfaces with low permeability; it may also be transient due to, for example, surface sealing or compaction of the top soil. Saturation excess overland flow (Saturation Overland Flow, SOF; e.g., Dunne and Black, 1970a,b), occurs due to a (potentially perched) saturation of the soil. It may be caused by direct precipitation, or through seepage of groundwater, or other subsurface (return) flow processes. Subsurface Stormflow (SSF) refers to subsurface flow processes that cause runoff within the time scale of a storm event, for example flow through soil pipes, macropores or highly permeable layers (Weiler et al., 2006). Deep Percolation (DP) occurs if lateral drainage only happens at large depth and contributes mainly to baseflow and negligibly to flood runoff.

Further developments following Scherrer and Naef (2003) include adaptation of the schemes to different land use and storm types (Scherrer, 2006) and simplifications of the scheme in a GIS-based regionalization tool using detailed soil maps in the hilly Swiss Central Plateau (Schmocker-Fackel, 2004; Naef et al., 2007; Schmocker-Fackel et al., 2007). Müller et al. (2009) presented a different GIS-based classification technique that could be trained with manually delineated DRP maps in the Rhineland-Palatinate, Germany, and the Grand Duchy of Luxembourg.

3.1.2 Research focus

Existing mapping approaches were developed for estimating catchment flood response to short-duration high-intensity storms. They therefore focus on areas with quick runoff formation, requiring detailed description of soils and shallow subsurface hydrological processes. Areas
with delayed response are of comparatively little importance for such events, yet may be im-
portant in steep, meso-scale catchments in which floods are caused by long-duration storms of 
moderate intensity. Such delayed responses may be expected in slopes where runoff formation 
occur in the deep subsurface (Chap. 2).

The mapping tool developed here tries to address the problem of characterizing delayed 
runoff formation in the deep subsurface of alpine slopes. This necessitates relying on generally 
available data for mapping the dominant runoff process directly, or through interpretation of 
a combination of spatial data products. Definitions and classification rules are similar to the 
SN-schemes, however classification will have coarser spatial resolution and the definitions of 
the DRP classes will be less detailed.

The large differences in hillslope-scale runoff responses (Chap. 2), imply that even a crude 
classification may yield meaningful insights that could not be obtained from viewing the catch-
ment as a single entity. The differentiation between areas with very fast response (steep, little 
storage potential), and strongly damped response (like the Gadenstetten site) may be relatively 
simple, and knowledge of their relative extents can already be useful for a first assessment 
of the catchment-scale flood response. Besides, the runoff response may depend strongly on 
interactions of processes within a hillslope or sequence of hillslopes, such that detailed point- 
scale data may be of little use. Instead, zooming out to characterize hillslope-scale storage and 
drainage processes may be more informative, as illustrated in the discussion of site properties 
relevant for the runoff response in Sect. 2.5.3

### 3.2 Basic concepts behind classification

A few simple concepts and accompanying assumptions form the backbone of the classification 
scheme. They require little data or interpretation and provide a first set of rules to differentiate 
landform types.

Damping of the rainfall–runoff response occurs through storage and delay. These terms are 
related; precipitation that is not yet discharged is in storage, such that there is a delay in dis-
charge response and the rainfall-to-runoff signal appears damped. Much of the damping may 
occur during the vertical flow process, i.e., the combination of infiltration into the ground and 
percolation through the unsaturated zone. The part of the subsurface where this occurs is 
here termed the vertical flow zone. It follows that, all else being equal, a deeper vertical flow 
zone causes stronger damping of the vertical flow process, because flowpaths are longer and 
dispersive flow processes become more important.

The vertical flow zone is bounded at the top by the land surface and at the bottom by ei-
ther a low permeability layer or the groundwater table. Below this, water flows laterally un-
less evaporation processes drive vertical upward flow. The vertical flow zone depth is shal-
lower than the depth to bedrock if impermeable layers or groundwater tables occur above the 
bedrock. On the other hand, it may be deeper if the bedrock is permeable, for example if the 
bedrock is strongly karstified or has become heavily fractured by other processes (like the 
Schluecht creeping landmass slope; see Sect. 2.5.1). The vertical flow zone dept may thus differ 
substantially from the depth to bedrock, indicating these hillslope properties must be clearly 
distinguished in the mapping procedure.
Damping of flow processes also depends on the properties of the flow-conducting medium. For example, hydraulic conductivity influences flow velocity, and water retention capacity influences water storage. For granular material, hydraulic conductivity may be roughly estimated from correlations with grain size (e.g., Shepherd, 1989), and this is here extended to a crude assumption about the damping effect on the runoff response: throughflow is more damped in media with substantial fine-grained material content than in media consisting mainly of coarse-grained material. This relation is expected for both saturated and unsaturated conditions and affects horizontal and vertical flow processes. The relation also applies for granular media where macropore flow processes dominate, as infiltration into the surrounding smaller pores (i.e., the soil matrix), causes damping of the flow rates (e.g., Beven and Germann, 1982; Weiler and Naef, 2003).

The same considerations apply to heavily fractured rock media, where granular material is virtually unavailable. Assuming that effective porosity of fractured rock is substantially below that of granular sediments, and that there is little exchange of water between flow-conducting fractures and their surroundings, it follows that percolation is less damped in highly fractured rock than in granular media with a substantial fraction of fine-grained sediments. Besides, in the phreatic zone, a low porosity implies a large groundwater table rise per unit storage increase. This can cause strong groundwater table fluctuations, which may result in a substantial increase in transmissivity or groundwater head gradient, or both. These mechanisms were shown to be important for the relatively strong flood runoff response of the Schluecht slope (Sect. 2.5.1), and are here hypothesized to further corroborate the difference in damping strength between fractured rock systems and fine-grained sediments.

Here, landscape elements are delineated according to their geomorphological and sediment properties, without further differentiation according to spatial extent, although this could affect the runoff response: it may be expected that damping increases with upslope accumulated area, as was also found in, for example, the hillslope comparison study of Montgomery and Dietrich (2002). This scaling relationship is not used in the mapping scheme because it is not well understood and likely depends on a combination of hillslope properties that are incorporated in the scheme; either implicitly, like the permeability assessment from grain-size properties (see above), or explicitly, like sediment cover depth or topographic gradient.

Because of this delineation method, there is no fixed mapping scale, such that the scheme may be said to operate at the ‘hillslope scale.’ The hillslope scale is smaller than, or equal to, the full slope between topographic water divide and stream, here referred to as the topographic slope. The term ‘catena’ here refers to a sequence of mapped hillslope elements connected through surface or subsurface flowpaths, for example the bare rock faces draining into the talus slopes at the Wissenboden headwater (Sect. 2.5.1). The word ‘catena’ here thus has a broader meaning than in soil science where it is used to indicate “a sequence of contrasting soils formed along a topographic slope” (e.g., Thomas and Goudie, 2000). The landscape element delineation procedure is discussed in detail in Section 3.5.
3.3 Data available for classification

The generally available data products do not allow direct delineation and classification of the properties discussed above. Therefore depth and material type must be mapped manually, by interpreting a combination of field observations, generally available spatial data products, literature and maps specifically covering the region of interest.

Field observations include interpretation of vertical profiles obtained through excavation or drilling, estimation of channel flow capacities and erosion, evaluation of spring discharge behavior, and assessment of sediment cover depth from outcrops and channel incisions.

A range of relevant spatial data is available for most of the Swiss Alps. Stream network and basic land use types are available through the VECTOR25 landscape model (Swisstopo, 2007). Particularly the land use classes urban area, bare rock, swamp, and glacier are helpful because they allow direct classification (Sect. 3.6.2). Useful aerial imagery products include the SWISSIMAGE orthophotos (Swisstopo, 2010) and the 3D representation of aerial photos in the Google™ Earth platform (Google Inc., 2013).

Topographic analyzes and derivations are here based on the $25 \times 25$ m digital elevation model (DEM) of Swisstopo (DHM25, Swisstopo, 2005). For the areas where it is available, the higher resolution ($2 \times 2$ m) terrain data of the swissALTI3D DEM (Swisstopo, 2013) are used to delineate landscape features and characterize small-scale features like erosion rills and channel initiation (Sect. 3.6).

Geological and geomorphological information is available for most of Switzerland at 1:25 000 scale in the Geologischer Atlas der Schweiz (GA25) distributed by Swisstopo. It is also available in digitized vector format in the GeoCover product, whereby gaps in the GA25 coverage are filled with other geological maps, usually with coarser resolution (Swisstopo, 2011). As definitions and scope may differ strongly between geological maps, interpretation often requires field exploration and evaluation with other products, in particular aerial imagery. Additional area-specific geo(morpho)logical knowledge may be found in the geological literature or the explanatory notes (German: Erläuterungen) accompanying the GA25 maps. Sometimes, detailed geomorphological maps may have been made for the region of interest. Of particular hydro(geo)logical interest are the 1:100 000 scale hydrogeological maps for amongst others the Schaechen catchment (i.e., Jäckli et al., 1985). Furthermore, databases of springs, wells and small streams can provide valuable auxiliary information at smaller scale.

Detailed soil maps may provide valuable information. However, in alpine areas, quality is often low at higher elevations. Soil formation in alpine terrain is often limited to few decimeters, but may occur on top of thick, permeable, debris deposits as well as impermeable bedrock, such that one must be careful not to mistake soil depth for depth to bedrock.

3.4 First-order characterization of alpine landforms

I compiled a list of commonly found landform types from studying landforms and geological maps of various alpine catchments, including the Luetschine, Reuss until Andermatt, Saltina, Engelberger Aa, Hinterrhein, and Dischma. The landforms are here categorized by defining a crude classification of sediment cover thickness and texture. This first-order categorization
forms the core of the DRP classification scheme presented in Sect. 3.6. The sediment cover thickness and texture classes are presented first, followed by a characterization of the defined landform types and the resulting categorization.

3.4.1 Definitions of sediment cover thickness and texture

Sediment texture is classified according the predominant particle sizes:

- **Coarse Debris (CD)**; sediment consists mainly of gravel, cobbles, and boulders.
- **Fine-Grained (FG)**; sediment deposited, or formed in situ, like soils, with little CD material visible; sediment consists mainly of clay, silt, and sand.
- **Mixed-Grained (MX)**; mixture of ‘Coarse Debris’ and ‘Fine-Grained’ sediments in substantial proportions.

The thickness of the sediment cover, defined as the depth to bedrock (Sect. 3.2), is grouped into four classes:

- **Shallow**; on average shallower than about 0.5 m
- **Medium**; average depth to bedrock between about 0.5 m and 1 m
- **Thick**; average depth to bedrock between about 1 m and 5 m
- **Very thick**; much more than 5 m

The indicated depths in meters are listed for reference and not for formal classification. Average depth can not be measured for the landforms of interest, but I consider it unlikely that different people with reasonable geomorphological expertise, provided with the generally available data used here, assign thickness classes with more than one class difference: a ‘medium thick’ sediment cover might be wrongly classified as ‘thick’, but is unlikely to be classified as ‘very thick’.

Likewise, situations where coarse debris or fine-grained sediments dominate may be clearly recognized from inspection of the land surface and geomorphological interpretations, as further discussed in Section 3.4.3. Besides, the assumption that damping of flow processes depends strongly on the availability of fine-grained material (Sect. 3.2), means that the sediment texture classification may be reduced to two classes; landforms with substantial fine-grained material content (i.e., FG or MX texture) and landforms with little fine-grained material (i.e., CD texture). I found these strongly contrasting situations could mostly be distinguished with little ambiguity. This crude differentiation of sediment texture was therefore given much importance in the dominant runoff process mapping scheme (detail in Sect. 3.6).

3.4.2 Definitions of landform types

Landform types are here defined broadly, such that all slopes in a typical mountainous landscape can be included in the first-order categorization. Some Quaternary landform types may be specifically recognized in detailed geological maps (Sect. 3.3), such as moraine deposits,
alluvial plains and debris cones. Other landform types, like bare rock, stream incisions, and glaciers can be determined from aerial imagery, digital elevation models and land use maps. More detailed geomorphological characterization is hardly feasible without extensive field exploration.

The landform types that receive much attention in the DRP classification scheme are briefly described below. As the classification scheme is intended for catchments where glaciers, lakes and man-made structures (urban areas, reservoirs, etc.) play only minor roles in the flood runoff formation, these features receive little attention and are not discussed in detail here.

**Alluvial Plain:** Commonly found in flat valley floors where flooding has deposited a mixture of coarse- and fine-grained sediments. The landform can be wider than the present-day floodplains along the main stream or braided river system as the terraced slopes of older deposits are included. The latter is common in deglaciated mountainous landscapes, where valleys carved out by glaciers are filled with sediment that is later incised by streams. This landform type thus covers all sorts of alluvial deposits, from the small glacial outwash plains in periglacial terrain, to the thick gravel (German: *Schotter*) aquifers found in many large valleys of the Alps.

**Bare Rock:** Land surface consisting of exposed rock, i.e., areas without soil or debris cover. Commonly found at high elevations, but also occurring as walls or outcrops at lower elevations.

**Stream Incision:** Erosional landform caused by fluvial processes cutting through bedrock or consolidated sediments. It here refers to the steep flanks along erosive streams, forming V-shaped valleys.

**Moraine:** Unsorted accumulations of sediments deposited directly by glacial processes. The sediment itself is referred to as glacial till and is typically a mixture of coarse- and fine-grained material.

**Moving Landmass:** Areas with slow, deep-seated gravitational slope deformations (DGSD; Soldati, 2013), like rotational slides (slumps) or creeping rock masses (e.g., the Schluecht slope; Sect. 2.2.2). Slope deformations caused by tectonics or post-glacial rebound (e.g., Ustaszewski et al., 2008) are also included in this class. All these deformation mechanisms may cause strong fracturing to depths beyond the saprolite horizon (chemically weathered bedrock). Occurrence of DGSD may favor collateral fast mass wasting mechanisms like land- or rockslides (Soldati, 2013). Lateral drainage may occur predominantly through this heavily fractured rock, whereby the sediment cover provides additional storage potential (see literature review in Sect. 2.1.2 and discussion of the observations at the Schluecht creeping landmass slope in Sects. 2.5.1 and 2.5.3).

**Rockslide:** A large-scale deposit formed by a collapse of a mountain slope (German: *Bergsturz*). Such deposits are typically thick and a mixture of large rock masses and fine- and coarse-grained sediments.

**Talus Cone:** A sloping mass of sediments accumulated at the foot of a cliff or slope, predominantly deposited by gravitational processes (German: *Trockenschuttkegel*), typically with
upward concave shape towards the base. Depending on formation processes, the coarse debris associated with rockfall deposits may be mixed with fine-grained material, or even covered by soil.

**Debris Cone:** A sloping mass of sediment deposited mainly by debris flows, recognizable from the lateral levees along the flow channels. Debris flows typically transport fine- and coarse-grained sediments and other debris, all of which can be found back in the debris cone deposit. This difference with a talus slope is well reflected in the German definition: *Bachschatkegel*.

**Debris Slope:** All sediment mantled slopes not falling into any of the above categories are here referred to as debris slopes, including ‘textbook example’ slopes with soil cover on (impermeable) rock. Steep cone-shaped deposits should be categorized into either talus cone or debris cone, such that a debris slope by definition has a rather uniform sediment cover thickness and slope. Various terms are used to characterize these kinds of slopes in the literature, but their definitions are often too narrow or too broad in scope; regolith, colluvium, taluvium, talus sheet, and slope waste.

**Karstified Rock:** Highly permeable bedrock mass where extensive cave and fissure systems have formed through dissolution of soluble rock types like limestone, gypsum or dolomite. Such areas normally have no surface water as all rainfall and snowmelt percolates into the rock and recharges a karstic groundwater aquifer. Occurrence of karstic landforms is usually indicated on geological maps, but the level of detail may vary and the geological literature covering the region of interest must be reviewed: some rock strata are known to be strongly karstified, whereas others usually show little karstification. Identified sinkholes and karren plateaus are usually indicated, but drainage direction of the karstic aquifer may be unknown. Likewise, the main outflows of the aquifer system may be indicated if found as springs, but are missing if discharging directly into other groundwater systems. Hydrogeological research, often with the purpose of finding secure drinking water resources, may have provided good knowledge of flowpaths and catchment areas, particularly if dye-tracer studies were conducted.

The hydrological functioning of these landform types may be little researched and difficult to generalize, as illustrated by the literature review in Section 2.1 and the field observations discussed in Section 2.5. However, there is some relevant knowledge about the geomorphological properties of the landform types that may help characterizing and differentiating their runoff responses. These properties, as well as the differences between the landform types, are discussed here.

Talus cones, debris cones and alluvial fans are often suggested to form a continuum of landform types (Brazier, 1988 p. 429), making it difficult to formulate criteria for distinguishing between them, other than by their deposition mechanisms: talus cones are formed by rockfall, debris cones by debris flows and alluvial fans by fluvial processes. All three mechanisms may have shaped an individual landform, and possibly a few more mechanisms, like rockslides or glacial deposition, have contributed too. The occurrence of debris flows is here used to distinguish the three landform types. It is a debris cone if there are various debris flow channels,
recognizable from their characteristic lateral levees (e.g., Brazier, 1988; Stoffel et al., 2008), otherwise it is a talus cone. Alluvial fans are found in more gentle sloping terrain, lack debris flow levees, and are thus mapped as alluvial plains. These differentiations are important for characterizing the sediment properties of the landform types in Section 3.4.3.

Moving landmasses and rockslide deposits may (partially) block a river, thereby strongly changing the landscape. For example the creeping landmasses that have pushed the main Schaechen river towards its southern flank (Sect. 2.2.1), or the rockslides blocking the Bavarian Reintal, forming alluvial plains upstream of the blockage (e.g., Schrott et al., 2002) that can strongly damp the flood behavior (Lauber et al., 2014).

Such catastrophic rockslides are not rare, as Albert Heim noted almost a century ago (Heim, 1931, 1932); in the Swiss Alps, a few hundreds of rockslides have occurred in historical times. Von Poschinger (2002) provides a more recent overview of large rockslides in the Alps, and points out that rockslides in sedimentary rock are often larger than in crystalline rock, and that zones with many rockslides are found in the Swiss Alps around the Aarmassif and along the Rhine-Rhone line.

Although slower processes often have less dramatic effects on the landscape than rockslides, their occurrence may be profound in large parts of a catchment. Our observations showed how deep fracturing influences subsurface flows by preventing the formation of dense drainage networks on much of the Schaechen’s northern flank. The DGSD slopes in the Urseren valley (e.g., Ustaszewski et al., 2008) provide another example: the drainage network is not well developed and the springs draining the slopes show little (episodic) channel erosion. Such deformation is also found in the granite and gneiss rocks along the Alps’ main ridge between Flims and Sierre (e.g., Hantke, 1978 p. 394), possibly affecting large parts of the upper Reuss and Rhone catchments. Compiling an overview of various types of DGSD in the European Alps, Crosta et al. (2013) found more than 1000 affected slopes, covering areas of 0.03 to 108 km². They comprise 5.6 % of the investigated area.

Rockslide deposits are here referred to as sediment cover. In contrast, this definition for moving landmasses applies only to the soil and debris covering the fractured rock mass there. The fractured rock is considered as bedrock and not sediment, because the fractured rock is not granular material and can therefore have very different hydrological properties (Sect. 3.2).

### 3.4.3 Categorization of the landforms

For the different landform types, expected sediment cover type and thickness are shown in Table 3.1. Karstified rock is not listed because the sediments covering such landforms are not relevant for the classification of the dominant runoff process. The table shows the large variation of thickness and sediment types in mountainous terrain, ranging from bare rock slopes with only depression storage to very thick debris cone deposits with large storage capacity. It also shows that landform type can sometimes serve as a proxy for the sediment cover thickness or type, or even both. For example, a debris cone is usually a very thick, mixed-grained sediment deposit, whereas a stream incision has only a shallow, mixed-grained sediment mantle.

These generalizations may guide the search for criteria for further classification. For example, an alluvial plain is a thick to very thick deposit with FX or MX type texture. One might
3.4. First-order characterization of alpine landforms

Table 3.1: Occurrence of sediment cover characteristics of typical mountainous landforms. Four thickness classes are distinguished, and three texture classes: Coarse Debris (CD), Fine-Grained (FG), Mixed-Grained (MX). Listed thickness values are only indicative.

<table>
<thead>
<tr>
<th>Landform</th>
<th>Thickness of the sediment cover</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>shallow (≤0.5 m)</td>
</tr>
<tr>
<td>Bare Rock</td>
<td>–</td>
</tr>
<tr>
<td>Stream Incision</td>
<td>MX</td>
</tr>
<tr>
<td>Alluvial Plaina</td>
<td>–</td>
</tr>
<tr>
<td>Moving Landmass</td>
<td>CD, FG, MX</td>
</tr>
<tr>
<td>Debris Slope</td>
<td>CD, FG, MX</td>
</tr>
<tr>
<td>Moraine</td>
<td>MX</td>
</tr>
<tr>
<td>Rockslide</td>
<td>–</td>
</tr>
<tr>
<td>Debris Cone</td>
<td>–</td>
</tr>
<tr>
<td>Talus Cone</td>
<td>–</td>
</tr>
</tbody>
</table>

*Sediment thickness class not expected to control the runoff formation process.

thus assume that vertical flow zone occurs in a medium with a substantial fraction of fine-grained material. However, lateral flow may occur at relatively shallow depth, as typically the groundwater table occurs within the deposit (i.e., due to and impermeable layer or connection with the river). Therefore, estimation of the depth to bedrock is not important for classifying the dominant runoff process, as also indicated in Table 3.1. This kind of evaluation points out that further classification of alluvial plains should focus on characterizing how lateral drainage occurs.

By highlighting differences and similarities between landform types, the table also provides a framework for developing internally consistent classification rules. For example, if glacial till and debris slopes both consist of thick mixed-grained sediments, they should receive the same further classification steps. On the other hand is it expected that a debris slope consisting mainly of coarse-grained material has a faster runoff response than a similarly deep moraine slope, because the moraine slope should have substantial fine-grained sediment embedded.

Table 3.1 and the basic concepts of Section 3.2 form the backbone of the DRP classification scheme, which is therefore regarded as geomorphology-based. It explicitly considers how the landscape formation processes affect the flowpaths in the different landforms. It thereby goes beyond other approaches that consider geometrical measures such as drainage density or topographic index. Such approaches are based on geomorphometry and not geomorphology, but often bear the latter in their name; for example the various geomorphological unit hydrograph approaches are mainly based on some measure of channel network density.
3.5 Delineation of landscape elements

The mapping scheme represents dominant runoff processes at the hillslope scale. First, homogeneous landscape elements are delineated, and these elements are then classified using decision rules that operate on these elements’ attributes. The landscape elements are kept as large as possible and thus only subdivided further if needed to obtain more homogeneous landscape elements. Keeping the delineated elements large allows evaluating lateral drainage processes at the scale where they cause runoff. After all, lateral drainage is not a point-scale process, but works throughout a hillslope or even a sequence of hillslopes; a catena. The largest elements are defined as discussed in Section 3.4. Topographic water divides and the channel network serve as additional bounding features.

For some landform types, like rockslide deposits, talus cones, and debris cones, further subdivision is not feasible. Sediment cover texture and thickness may be spatially very variable in these landforms, but these could only be estimated with detailed field exploration. Fortunately, the core classification of Table 3.1 is almost completely resolved; these landforms constitute thick debris deposits, the fine-grained content of which only requires further assessment for talus cones. Other areas, such as bare rock faces, swamps and urban areas, are expected to have so little internal variability of hydrological processes that further subdivision is only needed if relevant for the catena flow or runoff generated from these areas (see Sect. 3.6.2).

The other landforms often have to be further subdivided, because of sharp contrasts in the attributes of interest. Texture of the sediment cover can often be directly inferred from Table 3.1 or recognized from aerial imagery. Contrasts in sediment thickness are often bounded by topographic inflections and visible as strong changes in relief or channel network density. Changes in relief are specified as terrain ruggedness and rules for its classification are defined in Section 3.6.2.

A nice example is the valley shoulder often found above a U-shaped valley, its lower end visible as a relatively sharp topographic inflection. Much glacial till has accumulated on the relatively flat shoulder plateau, whereas the steeper slopes below the inflection have much relief and only little sediment cover, also mostly of glacial origin. Typically, episodic streams originate in the headwaters upslope of the shoulder plateau and lose water to the glacial till aquifer before they reach the steeper slopes (that constitute the ‘legs of the U’). The aquifer on the plateau feeds seasonal or perennial streams that start from the springs at the shoulder’s lower end.

Another reason for further subdivision may be the occurrence of swampy areas, as these indicate a shallow groundwater table or impermeable layer. Here, delineation can be based on a combination of aerial imagery and a detailed elevation data, possibly complemented by a soil map or soil profiles obtained in the field. Such features also occur in moraine landscapes, but may be too small to receive their own landscape element (see Sect. 3.6.2).

Areas with karstified rock may be subdivided according to the catchment they are expected to drain to (if such information is available). Further subdivision might be needed if areas are found where infiltration into the karstic system is prevented by impermeable sediments or rock strata, like marl layers or lenses (see Spillman et al., 2011, p. 175, for an area at the northern
catchment boundary of the Schaechen). Also, for various reasons, karstification may be less strong in some areas; e.g., it is less developed in steep slopes, because other weathering mechanisms and surface runoff dominate, and at high elevations, where glaciers have covered the flatter areas such that less liquid water was available for dissolution. Karstification is assessed by evaluating whether surface runoff occurs sometimes, as indicated by erosion marks, and, in the case of bare rock, whether karstification is visible at the surface. Only areas with a strong karstification and no surface runoff are then mapped as karstified rock; the remaining areas are mapped as debris slopes.

3.6 Classification of landscape elements

3.6.1 Structure and definitions

The developed scheme follows the concepts of Scherrer and Naef (2003) and Schmocker-Fackel et al. (2007), here referred to as the SN-schemes. These are useful bottom-up tools for plot-scale mapping of DRP when accurate data are available. The tool presented here is more top-down; dominant runoff processes are derived from a combination of proxy-data and understanding of landscape formation processes (Sect. 3.4). The tool is intended to complement the SN-schemes rather than replace them. This particularly holds for the classification of near-surface processes in Section 3.6.2.

The same four dominant runoff process classes are used as in the SN-schemes: Hortonian Overland Flow (HOF), Saturation Overland Flow (SOF), Subsurface Stormflow (SSF), and Deep Percolation (DP). These process may be further specified by their process intensity, by adding the numbers ‘1’, ‘2’, or ‘3’ for increasingly damped behavior. The damping depends mainly on available storage potential; shallow soils where subsurface stormflow is the dominant runoff process are mapped as SSF1 because of the limited storage potential and thicker soils with similar DRP are mapped as SSF2 or even SSF3.

Deep Percolation is here defined differently than in the SN-schemes: percolation that recharges a deep groundwater body whose drainage constitutes the dominant runoff mechanism. This definition highlights the key difference in perception of SSF and DP processes: SSF is assigned if subsurface stormflow is expected to occur via processes that are episodically activated (i.e., in response to storms), whereas DP is assigned if drainage from a deep seasonally, or even perennially saturated zone, (i.e., groundwater table more than about 2 m below the surface,) constitutes the only contribution to storm runoff. This differentiation conflicts with the more general definition by Weiler et al. (2006) in the Encyclopedia of Hydrological Sciences, wherein all subsurface flow processes contributing to storm runoff are considered subsurface stormflow. The conflict originates from the common perceptual model that deep processes contribute mainly to baseflow; Weiler et al. (2006) hardly mention the kind of deep processes observed in the Schaechen (see Chap. 2).

In extreme cases, a thick deposit may be so permeable that little water is stored beyond event time scales, like coarse debris deposited on a steep slope, and a relatively strong runoff response may be expected (Sect. 2.1.2). The dominant runoff process of such a slope will be classified as SSF and not as DP, even though water percolates over large depth. On the other
hand, a slope like Schluecht, where a relatively deep groundwater body can contribute to stormflow, will be classified as DP, although in the definition of Weiler et al. (2006) it is a SSF process.

Three types of DP are distinguished, all expected to have a more damped response than SOF3/SSF3. Slopes where the deep groundwater body is found in fractured rock are mapped as DP-rock (DPr), e.g., the Schluecht creeping landmass slope. DP type slopes with groundwater bodies in granular media, are either DP-debris1 (DPd1), if draining rapidly, or DP-debris2 (DPd2) if draining slowly. The Gadenstetten site is an example of the slowly draining DPd2 type of slope with negligible contribution to storm runoff. Only the DPd2 type of slopes are expected to contribute negligibly to flood runoff.

The DPd2 has a more damped response than the DPr and DPd1 classes, which are here considered to have similar process intensity. This means the DRP classification recognizes in total five intensity classes: the first three intensity classes are specified in the HOF, SOF, and SSF class names (e.g., SSF2), the fourth intensity class consists of the DPr and DPd1 classes, and the fifth intensity class constitutes only DPd2.

The developed mapping scheme classifies dominant runoff processes in the shallow and deep subsurface separately. The shallow subsurface part considers roughly the first 1 m of all landscape elements. It is essentially a simplification of the SN-schemes’ classification of near-surface DRP that allows mapping in mountainous terrain where detailed data is hardly available. The deep subsurface part of the scheme is only used if no runoff formation mechanisms are found in the shallow subsurface classification. It evaluates landscape properties that are relevant for percolation through the unsaturated zone and the lateral drainage from the, potentially transient, saturated zone.

Glaciers and snowfields are not included in the scheme, but are classified as SSF3 if their average slope exceeds 30% and as DP-rock otherwise. The more damped DP-rock is typically assigned to flat perennial snow fields, accumulation zones of glaciers, and the parts of glaciers that are fully covered with debris. Glaciers and snowfields should therefore be further subdivided if strong inflections in slope or contrasts in sediment cover are found.

3.6.2 Runoff formation in the shallow subsurface

The classification of the shallow subsurface is presented in Figure [3.1] The different landform types are at the top row; the black lines indicate the decision paths through at most five criteria, i.e., the grey boxes with Roman numerals I to V. The decision paths indicate how combinations of criteria result in the different dominant runoff process classes indicated at the bottom (black boxes); intermediate results are indicated in white boxes.

Some dominant runoff processes can be directly derived from land use types classified in the VECTOR25 product (Swisstopo, 2007). These include urban areas, swamps and lakes. Bare rock is not that consistently indicated in this product, as it presents outcropping rock faces that may be continuous regions or ‘slivers’ separated by narrow areas with shallow soils or even thick debris deposits. Aerial imagery helps to decide whether the bare rock is contiguous, or if the area must be mapped as debris slope. The classification of the remaining areas is based on the delineated landform types (Sects. 3.4 and 3.5).
3.6. Classification of landscape elements

**Criterion I: Thickness of the sediment cover**

This criterion is only needed for the landform types for which the thickness of the sediment cover cannot be directly inferred from Table 3.1. In some cases the sediment cover thickness may be directly inferred from the sediment texture. For example, areas with large boulders are classified as ‘thick’ or even ‘very thick.’ Similarly, the sediment cover is at least ‘medium thick’ if it consists predominantly of soil and shows no rock outcrops. In effect, sediment thickness (criterion I) and sediment texture (criterion II, see below) are thus assessed simultaneously.

If no such simple inferences can be made, classification occurs through assessing the terrain ruggedness. Thereby three ruggedness classes are defined, each used as a proxy for the sediment cover thickness classes used in Table 3.1.

**Very rugged:** Many rock outcrops indicate bedrock at ‘shallow’ depth, i.e., at less than ~0.5 m. Such areas commonly occur in the debris slope class, but also moving landmasses or moraine landforms may have very rugged parts.

**Medium rugged:** If there are virtually no outcrops, but terrain is uneven with sharp edges, a ‘medium thick’ sediment cover is expected (i.e., roughly between 0.5 and 1 m). Such terrain may be found, for example, in steep areas with dense channel networks.

**Gently undulating:** The terrain lacks sharp edges and are gently undulating. Slopes show little erosional features and often a thick soil cover has developed. The smooth transitions between convex and concave forms indicates that the sediment cover is ‘thick’ or even ‘very thick,’ i.e., more than 1 m.

It proved difficult to develop a GIS algorithm for automating the ruggedness classification because ruggedness relates to a combination of explanatory variables, such as topographic gradient, thickness, sediment texture and formation processes. Furthermore, sharp edges and smooth bumps are scale-dependent, such that the results depend on the arbitrary selection of the ‘window size’ the algorithm is applied at. Topographic data alone may thus not yield robust results. Instead, the classification is done manually, from combined high-resolution aerial imagery and elevation data. This may be more reliable, but requires expert interpretation; one must ‘read’ the landscape.

**Criterion II: Sediment texture**

If sediment cover texture cannot be determined from landform type alone (i.e., Table 3.1), it has to be estimated from aerial imagery. This is particularly relevant for debris slopes and moving landmasses, because occurrence of fine-grained sediments is not guaranteed. Also, for talus cones, the sediment texture should be assessed, although it may be only used for the classification of deep subsurface processes, as no dominance of near-surface runoff formation processes is expected.

**Criterion III: Runoff formation in the shallow subsurface**

Even in areas with thick sediment cover, runoff formation in the shallow subsurface may dominate because of groundwater tables or impermeable layers. Both mechanisms can cause rapid
lateral drainage by either saturation overland flow, or subsurface stormflow mechanisms, occurrence of which must be assessed.

Shallow groundwater tables are often visible through patches of swampy areas with different vegetation. They commonly occur in alluvial plains and flat moraine deposits, and are typically caused by seepage zones above shallow impermeable layers. Shallow groundwater tables in alluvial plains can also be related to the water level in the river. In steep terrain, occurrence of shallow groundwater tables or impermeable layers can cause strong erosion by overland flow or return flow processes, and may thus already have been recognized by the ruggedness classification for criterion I.

Criterion IV: Relating runoff process to slope

Shallow SOF or SSF processes are distinguished by a topographic gradient threshold; slopes steeper than 5% are mapped as SSF and flatter slopes as SOF, similar to Schmocker-Fackel et al. (2007).

Criterion V: Connectivity to the main channel network

Connectivity of areas with very fast response (i.e., HOF1, SOF1, SSF1) is evaluated after finishing the classification of deep subsurface processes (Sect. 3.6.3), because knowledge of areas further down the catena is required. The evaluation is needed because the difference between the response of an area and its effective contribution to catchment runoff can be large when the generated runoff infiltrates into an area with damped runoff response (Sects. 2.5.1 and 2.5.4).

This classification step evaluates whether the episodic channels fed by areas with very fast response have dead ends in DP type landforms. Unconnected areas receive a preceding ‘u’ in their classification, resulting in three additional classes: uHOF1, uSOF1, and uSSF1. Assessment of connectivity is limited to areas with very fast DRP because runoff from these areas is expected to leave visible erosion marks. Connectivity of areas with a more damped response is not evaluated, and neither is the additional damping effect of partial re-infiltration of generated runoff.

Small areas of outcropping bare rock faces can be automatically extracted from the Vector25 landscape model and may be found to occur inside talus or debris cone elements, or at their upper boundaries. Such HOF elements have little catchment area and generated runoff is therefore assumed to re-infiltrate into the thick deposits, thereby considerably damping the response. These features are therefore automatically mapped as uHOF1.

3.6.3 Runoff formation in the deep subsurface

Figure 3.2 presents the classification scheme for the areas where runoff formation occurs in the deep subsurface. Criteria concerning vertical flow processes are found in light grey boxes, and those relevant for lateral drainage processes in dark grey boxes. This distinction is here used to structure the discussion, but is not always so strict, as assessment of some properties of the vertical structure depend on interpretation of lateral flow processes. The connecting lines indicate the decision paths. An overview of the landforms and their classification is shown in Figure 3.3 to highlight differences and similarities. The different trajectories lead to processes ranging
from SSF2 to DP-debris2, i.e., from areas with fast response and substantial contribution to all floods, to areas with strongly damped response and negligible contribution to even the most extreme events.

**Criterion VI: Occurrence of fine-grained material at depth**

Table 3.1 indicates occurrence of fine-grained material in thick to very thick deposits must be evaluated for the landform types ‘talus cone,’ ‘moving landmass,’ and ‘debris slope.’ It is here assumed that the sediment texture determined via criterion II occurs throughout the vertical profile. A talus cone or debris slope with visible soil cover is thus expected to have substantial fine-grained material between the coarse debris that normally make up the rockfall material. It is thus viewed as a vertical property of the medium, but in the resulting classification it also determines the drainage response: areas are mapped as SSF2 if there are no indications of substantial fine-grained sediment content. For talus cones, this criterion is only used when no other indications of fine-grained material are found for the base of the slope, i.e., the zone where runoff is produced.

**Criterion VII: Thickness and position of moraine deposits**

The criteria that only apply to moraine landforms are combined to highlight their inter-dependence. Moraine sediments presumably have substantial fine-grained particles content (Table 3.1), such that only the thickness of the sediment cover needs to be assessed. Aerial imagery and field visits help to assess whether the depth to bedrock is much more than 5 m, i.e., the class ‘very thick’ as defined in Section 3.4.1. This is taken as the vertical flow zone depth, unless groundwater seepage zones occur mid-slope (i.e., not only at the base). In these cases, the vertical flow zone depth is considered to be less than ~5 m because a shallow groundwater table is expected.

It is here assumed that topographic gradient affects the drainage time scale of moraine deposits. The position of the deposit in the landscape is used to distinguish between steep slopes with relatively quick drainage and flat slopes with relatively slow drainage. Flatter areas are found on valley shoulders (plateaus) and valley floors. Such areas have typically been sediment sinks after glaciers have retreated, with erosion often limited to fluvial processes. Very thick till deposits may have accumulated here, particularly in overdeepened valley floors. The groundwater table may be expected to be relatively flat, with a strongly damped runoff response (see examples in Sect. 2.5.3). Moraine deposits on slopes are expected to have steeper groundwater tables and a comparatively fast drainage. The position in the landscape is thus used to differentiate response strengths. Combined with the estimated sediment cover thickness, this allows classifying moraine deposits into SSF3, DP-debris1 or DP-debris2.

**Criterion VIII: Topographic gradient at the foot of a cone**

The topographic gradient of the foot of a talus or debris cone is used to directly infer information about the groundwater table gradient, flowpath lengths and sediment texture, based on the following considerations: hydrological properties at the base of the slope determine the outflow from the cone system. If this area is not so permeable, rapid percolation pathways further upslope do not translate into a corresponding fast outflow. Besides, a flat foot indicates a
Figure 3.1: Classification scheme for mapping dominant runoff processes (DRP) in the shallow subsurface.

Classification of dominant runoff processes in the shallow subsurface.

1. Classification of runoff formation mechanisms in the deep subsurface:
   - Check if shallow groundwater table or impermeable layer exists.
   - Yes → DRP
   - No → 'u' + DRP

2. Slope > 5%?
   - Yes → SSF
   - No → SOF

3. Connected to main channel network?
   - Yes → DRP
   - No → 'u' + DRP

Classification of runoff processes in the shallow subsurface:

1. Sediment cover thickness:
   - Very rugged → shallow (s)
   - Rugged terrain → medium (m)
   - Gently undulating → thick or very thick (t)

2. Sediment cover texture:
   - Coarse Debris; gravel, cobbles → CD
   - Fine-Grained; clay, silt, sand → FG
   - Mixture of CD and FG material → MX

3. Near-surface runoff formation:
   - Is there a shallow groundwater table or impermeable layer?
     - Yes → DRP
     - No → 'u' + DRP

Figure 3.1: Classification scheme for mapping dominant runoff processes (DRP) in the shallow subsurface.
3.6. Classification of landscape elements

Classification of dominant runoff processes in the deep subsurface

**Criteria**

- **DRP**
  - SSF 2
  - SSF 3
  - DP rock
  - DP debris1
  - DP debris2

**Type**

- Karstified Rock
- Alluvial Plain
- Moving Landmass
- Debris Slope
- Rock slide
- Moraine Cone
- Talus Cone
- Debris Cone

**Figure 3.2:** Classification scheme for mapping dominant runoff processes (DRP) occurring in the deep subsurface of typical alpine landforms. Criteria mainly used for assessing vertical flow processes are placed in light grey boxes and mainly used for lateral flow processes in dark grey boxes.

*Very thick sediment cover may only be assigned if there are no midslope groundwater seepage areas. Yes (y) and no (n) for the thickness and position criteria are listed first and second, respectively.*

*Criterion only used if geo(morpho)logic indications of strong fracturing of the bedrock exist.*
Figure 3.3: Landform configurations constituting the dominant runoff process classes found in mountainous hillslopes with large storage potential. As described according to the scheme in Fig. 3.2, the diagrams illustrate differences and similarities in geometry and sediment texture and are not drawn to scale. Delineation of cone shapes is indicated if not comprising the full diagram.

Legend:
- **coarse grained debris (CD)**
- **fine grained sediment (FG)**
- **mixed grained sediment (MX)**
- **competent rock**
- **fractured rock**
- **mixed material**
- **channel formation**

The flat foot may consist of mixed material, even if the upper part of the talus consists mainly of CD material. This is not distinguished in the sketches.

1. Talus may consist of either CD or MX material, even if the upper part of the talus consists mainly of CD material. This is not distinguished in the sketches.

2. Debris Cone
delineation of the cone

3. Rockslide deposit

4.混合 grained sediment (MX)

5. fine grained sediment (FG)

6. Channel formation

7. Debris slope

8. Meltwater

9. Moraine

10. Moving landmass
3.6. Classification of landscape elements

Flat groundwater table, as freely draining groundwater tables are rarely steeper than the land surface, and fine-grained sediment is expected to have accumulated in the flat foot (Sect. 2.5.3). Cones with flat feet are thus directly classified as DP-debris2.

Debris cones without flat feet are still considered to have substantial storage capacity because of their fine-grained sediment content and relatively long vertical and lateral flowpaths. They are mapped as DP-debris1. Talus cones without flat feet are further evaluated in criterion IX.

**Criterion IX: Medium through which drainage of talus cones occurs**

A talus cone without a flat foot may be deposited on another landform with substantial fine-grained sediment content, for example a flat moraine deposit or alluvial plain. It is then mapped as DP-debris1, regardless of the sediment texture of the talus cone itself. The reasoning behind this is that such talus cone is expected to drain via a flat groundwater body. Firstly, because accumulation of fine-grained sediments is expected at the talus base. This accumulation will affect damping during the percolation process and may even cause development of a groundwater body in the cone. Secondly, because drainage will partly occur via the flat landform upon which the talus cone was deposited, warranting relatively long flow paths and a relatively flat groundwater table.

All other talus cones are expected to have little long-term storage and are mapped as SSF2 or SSF3, depending on the sediment texture class determined in criterion VI. Such cones are commonly found below rock walls at higher elevations when there is too little sediment or catchment area to form debris cones.

It is noted that subsurface flow through cones may be highly diffuse, unless bedrock structures force convergence of flows, which may, ultimately, lead to a point-scale outlet; a spring. Such structure is often obscured by the thick sediment cover, but can be well visible in some larger scale landscape structures, for example in cirque valleys. Either way, this is not included in the classification scheme because it proved difficult to define an unambiguous measure of flow convergence. If such convergence causes a strong runoff response, this is expected to be visible in the landscape somehow, for example through erosion at the base of a cone and the resulting absence of a flat foot.

**Criterion X: Absence of channel initiation indicates strong fracturing of underlying bedrock**

If indications of strongly fractured bedrock exist, it must be assessed whether lateral drainage through fractured rock is the dominant runoff process (i.e., DP-rock). Landforms where fractured bedrock is expected include karstified rock, moving landmasses and rockslide deposits. However, debris slopes and moraine deposits may also cover permeable bedrock slopes, such that dominating fracture flow should also be assessed for these landform types if they are found near areas known to have strongly fissured bedrock.

Hillslope elements drained via fractured rock systems are assumed to have no channel initiation; there are no return flow processes with enough erosive power to incise through the sediment cover. Springs may occur at the base of the slope, but may be missing if subsurface
flowpaths discharge directly into a stream. In addition, occurrence of small gullies should be considered a sign of perennially flowing systems with so little fluctuation that even during large events virtually no erosion takes place. These gullies thus indicate a damped runoff response and suggest that there is a fractured rock system with much storage capacity. Besides, they may be man-made, as commonly found in the creeping landmass area of the Schaechen, where they convey outflows from small springs to the main stream in order to prevent extensive wet areas in the meadows.

Typically, areas drained through fractured rock systems have little or no stream network, but this also depends on upslope runoff formation mechanisms. For example, headwaters above the area may have incised wide V-shaped valleys. The assessment therefore only concerns channel initiation between these 'large-scale incisions.' This highlights the importance of keeping delineated landscape elements large; the catchment must be large enough to either initiate episodic flows, or sustain more perennial flows, and thus to allow assessment of lateral flow processes.

### 3.6.4 Examples

A few exemplary situations may aid understanding of the procedure:

- A moraine slope with mixed-grained sediment cover in between rock outcrops and an average gradient of 40%, is mapped as SSF1 if runoff flows directly into the main stream network. A similar area whose generated runoff re-infiltrates into a DP-debris1 type talus cone is mapped as uSSF1.

- A gently undulating area with average slope of 20% has many small erosion rills cutting through its fine-grained sediment cover. Because the strong erosion suggests that the area is very responsive, it is mapped as SSF1. It remains unclear whether the strong response must be attributed to an impermeable soil horizon, or the sediment cover not being as thick as the gently undulating terrain suggests.

- Parts of a wide alluvial plain with many wet zones are mapped as SOF2, whereas the parts without perched saturation or shallow groundwater tables are mapped as DP-debris1.

- A debris slope with more than 1 m thick cover of mainly coarse-grained debris is mapped as SSF2.

- The Gadenstetten research site has a very thick, mixed-grained debris mantle (Sect. 2.2.3). It exhibits a concave upward form and is thus mapped as a talus cone with a flat foot; DP-debris2.

- The soil-mantled moving landmass slopes at the Schaechen northern flank have gently undulating terrain, indicating thick to very thick sediment cover and thus require assessment of the deep subsurface. The slopes without signs of channel initiation, like our Schluecht site, are mapped as DP-rock.
3.7 Dominant runoff processes in the Schaechen

3.7.1 Application of the mapping scheme

Complementary data sources

The mapping of dominant runoff processes in the Schaechen benefited from a relatively rich set of data complementing the generally available data discussed in Section 3.3. Detailed hydrogeological knowledge was contained in the 1:100 000 scale Hydrogeologische Karte (Jäckli et al., 1985) and the 1:25 000 scale geological map covering most of the catchment (Brückner and Zbinden, 1987; Hantke and Zbinden, 2011). The GeoCover product (Swisstopo, 2011) provided similarly detailed information for the upper northern flank of the main valley. Further valuable geological characterization is provided in Spillman et al. (2011).

In addition, maps of the drinking water distribution network and groundwater protection zones (Kanton Uri, 1999a,b) were helpful for locating areas with many springs and springs with large catchments. Detailed soil maps are not available for the region, but the more than fifty soil profiles in typical settings throughout the catchment presented in Scherrer AG (2007) provided useful insights, for example about occurrence of shallow groundwater table fluctuations.

Contributions from areas with karstified rock

The above-mentioned geological data were used to identify areas with karstified rock. Catchment areas of karstic springs were delineated based on reports of dye tracer tests, and geological interpretation by the Swiss Institute for Speleology and Karstology (A. Malard, personal communication). The small band of karstic limestone at the northern catchment boundary likely drains towards the neighboring Muotatal catchment. The low-elevation limestone areas to the south and southwest often show strong karstification, but this could not be confirmed for the steep north faces of the Gross Windgaellen – Gross Schärhorn ridge. Existing conduit systems in the southwestern part of the Hinterschaechen are known to drain directly to the Reuss catchment (Jäckli et al., 1985), and southward drainage is suspected for the southern ridge because the limestone strata dip in that direction. Drainage directions of the karstified rock areas of the Hinterschaechen main gorge are unknown, and so are those of the karstified areas and closed basins of the Vorderschaechen. It is assumed these areas drain within the Schaechen catchment because there are no indications of drainage to other catchments.

Out of the 13% fractional cover of closed basins and limestone bedrock, about 8% drains via karstic conduit systems. This difference arises because the degree of karstification depends on rock type and exposure to different weathering mechanisms (Sect. 3.5). The karstified areas expected to drain within the Schaechen are mapped as DP-rock and those draining to neighboring catchments as ‘Loss-Karst.’

3.7.2 Results

The dominant runoff process map of the Schaechen is presented in Figure 3.4. Spatial coverage of the different DRP classes is specified for the full catchment in a pie chart and per 100 m elevation band in a histogram (Fig. 3.5). Table 3.2 compares the coverage of the DRP classes
A tool for mapping dominant runoff processes in alpine terrain

Figure 3.4: Dominant runoff process map of the Schaechen catchment, with main sites and measuring stations indicated.
3.7. Dominant runoff processes in the Schaechen

Table: Spatial coverage (%)

<table>
<thead>
<tr>
<th></th>
<th>strong (very fast)</th>
<th>fast</th>
<th>damped</th>
<th>little contributing</th>
<th>very fast, not connected</th>
</tr>
</thead>
<tbody>
<tr>
<td>HOF1</td>
<td>100.0</td>
<td></td>
<td>97.6</td>
<td>92.1</td>
<td>91.8</td>
</tr>
<tr>
<td>SOF1</td>
<td>100.0</td>
<td></td>
<td>96.0</td>
<td>87.8</td>
<td>87.4</td>
</tr>
<tr>
<td>SSF1</td>
<td>100.0</td>
<td></td>
<td>94.5</td>
<td>82.4</td>
<td>82.1</td>
</tr>
<tr>
<td>SSF2</td>
<td>99.9</td>
<td></td>
<td>92.1</td>
<td>76.4</td>
<td>76.0</td>
</tr>
<tr>
<td>SSF3</td>
<td>99.2</td>
<td></td>
<td>87.8</td>
<td>70.1</td>
<td>69.6</td>
</tr>
<tr>
<td>DP-rock</td>
<td>99.4</td>
<td></td>
<td>82.4</td>
<td>63.2</td>
<td>62.1</td>
</tr>
<tr>
<td>DP-debris1</td>
<td></td>
<td></td>
<td>76.4</td>
<td>55.0</td>
<td>54.1</td>
</tr>
<tr>
<td>DP-debris2</td>
<td></td>
<td></td>
<td>70.1</td>
<td>47.4</td>
<td>46.5</td>
</tr>
<tr>
<td>Loss-Karst</td>
<td></td>
<td></td>
<td>40.2</td>
<td>33.7</td>
<td>32.8</td>
</tr>
<tr>
<td>uHOF1</td>
<td></td>
<td></td>
<td>47.4</td>
<td>27.9</td>
<td>27.1</td>
</tr>
<tr>
<td>uSOF1</td>
<td></td>
<td></td>
<td>33.7</td>
<td>22.7</td>
<td>22.0</td>
</tr>
<tr>
<td>uSSF1</td>
<td></td>
<td></td>
<td>27.9</td>
<td>17.4</td>
<td>16.7</td>
</tr>
<tr>
<td>uSSF2</td>
<td></td>
<td></td>
<td>22.7</td>
<td>12.5</td>
<td>11.8</td>
</tr>
<tr>
<td>uSSF3</td>
<td></td>
<td></td>
<td>17.4</td>
<td>8.2</td>
<td>7.5</td>
</tr>
<tr>
<td>DP-debris3</td>
<td></td>
<td></td>
<td>12.5</td>
<td>5.4</td>
<td>4.7</td>
</tr>
<tr>
<td>DP-debris4</td>
<td></td>
<td></td>
<td>8.2</td>
<td>3.4</td>
<td>2.7</td>
</tr>
<tr>
<td>Loss-Karst</td>
<td></td>
<td></td>
<td>5.4</td>
<td>1.7</td>
<td>1.0</td>
</tr>
<tr>
<td>uHOF2</td>
<td></td>
<td></td>
<td>3.4</td>
<td>0.6</td>
<td>0.3</td>
</tr>
<tr>
<td>uSOF2</td>
<td></td>
<td></td>
<td>1.7</td>
<td>0.6</td>
<td>0.3</td>
</tr>
<tr>
<td>uSSF2</td>
<td></td>
<td></td>
<td>0.6</td>
<td>0.1</td>
<td>0.1</td>
</tr>
<tr>
<td>uSSF3</td>
<td></td>
<td></td>
<td>0.1</td>
<td>0.0</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Figure 3.5: a) Spatial coverage of dominant runoff processes in the Schaechen, with coverage of grouped classes in bold face. b) Histogram of coverage specified per 100 m elevation bands; numbers next to the bars indicate catchment areal fraction until the respective elevation bands.
of the Schaechen and the Vorder- and Hinterschaechen subcatchments. SFF and DP type processes dominate as soils are too permeable for infiltration excess runoff formation (HOF), and saturation overland flow (SOF) in steep terrain is mainly expected at the base of the slope (e.g., Dunne, 1978).

For ease of interpretation, the different classes are grouped into four runoff response types, which are throughout the rest of this thesis indicated with single quotation marks:

‘Strong’ Comprises the very ‘fast classes’ (HOF1, SOF1, SSF1) and ‘fast’ classes (SOF2, SSF2). These areas contribute substantially to both small and large floods.

‘Damped’ DRP expected to contribute little during small events and considerably during extreme floods; SSF3, DP-rock (DPr), and DP-debris1 (DPd1). Runoff may increase strongly after a critical storage level or delay time is reached; see for example the sprinkling experiments on SSF3 type slopes in Scherrer et al. (2007) and our observations at Schluecht (Sect. 2.5.1).

‘Little contributing’ DP-debris2 (DPd2) areas, expected to contribute little to even the largest floods (e.g., a strongly damped response like the Gadenstetten slope, Sect. 2.5.1). It also includes ‘Loss-Karst’ areas, which are areas expected to drain to a neighboring catchment.

‘Unconnected’ Areas with ‘very fast’ response draining fully into DP-type areas, i.e.: uHOF1, uSOF1, and uSSF1.

Only 24 % of the catchment area contributes much to small floods (group ‘Strong’), whereas and additional 45 % contributes to large events (the ‘Damped’ group). 25 % of the catchment has a negligible runoff response, even after extensive precipitation group (‘Little contributing’). A fifth of this area drains via karstic systems to neighboring basins. Almost 7 %, roughly a third of the very fast reacting areas (HOF1, SOF1, and SSF1), is not directly connected to the main channel network (group ‘Unconnected’). This shows how important it is to account for re-infiltration processes.

Almost 95 % of the catchment lies below 2500 m (Fig. 3.5b). The remaining area consists mainly of ‘unconnected’ slopes and slopes with ‘damped’ responses. This suggests little sensitivity of flood runoff to the snow line elevation during summer storms.

Also for lower snow lines, no strong changes in runoff formation may be expected: areas with ‘strong’ and ‘damped’ response are evenly distributed throughout all elevations, and about half the area between 1900 and 2500 m is ‘unconnected’ or ‘little contributing.’

Contrasting landscape structures are reflected in the dominant runoff process map. The steep southern flank of the main valley has a large fraction of areas with a ‘strong’ response. Also, most of the areas with thick debris cover are mapped as SSF3 or DPd1, rather than DPd2. On the other hand, the thick debris deposits above the valley shoulder ridge on the northern flank have a strongly damped response (DPd2), whereas many of the slopes below this ridge have thick soil cover and are mapped as SSF3 and DPr and areas with ‘very fast’ response are rare.
3.7. Dominant runoff processes in the Schaechen

The Hinterschaechen and the southern part of the Vorderschaechen have many cirque and hanging valley headwaters. Their valley floors are often covered by thick deposits of mixed origin, which account for a large fraction of the DPd1 and DPd2 area in the Schaechen. The runoff generated in upslope areas with ‘strong’ response often drains completely into these thick deposits, such that many of the uHOF1 and uSSF1 areas are found here as well. Such landscapes, with (strongly) damped runoff response, are common in high-elevation alpine terrain.

3.7.3 Agreement with small-scale observations in the Schaechen

The mapped DRP classes agree well with our observations at spring and headwater catchments (Chap. 2). This is not very surprising for the Schluecht slope, which served as exemplary case for the definition of the DP-rock class. An example of a slope where more interpretation was needed is our Gadenstetten site. The thick MX-type debris cover was partly indicated as moving landmass in the geological map of Brückner and Zbinden (1987), but it was here mapped as a talus because of the concave upward shape at the relatively flat foot of the slope, i.e., DPd2. This shows careful interpretation of all available data is needed before a landform is assigned, as the geological map uses different definitions than the DRP classification scheme. The delayed peak with negligible contribution to flood runoff (Sect. 2.5.1) is viewed as typical for DPd2 areas.

The contrast between the strong response of the western tributary and the damped response of the eastern tributary of the Wissenboden headwater (Sect. 2.5.1) is well described in the dominant runoff process map. The dense network of incised streams of the western tributary is mapped as SSF1, the surrounding slopes, mantled by medium thick soil, mapped as SSF2. This setting also covers about 6.4% of the eastern tributary, the rest is dominated by talus and debris cones with flat feet wherein all runoff from upslope areas infiltrates. Part of the upslope area is mapped as Loss-Karst because of expected northward drainage of the strongly karstified rock system (Sect. 3.7.1). The large fraction of areas with strongly damped response combined with the small fraction of ‘strongly’ reacting areas could explain the damped response of the Wissenboden catchment (Sect. 2.5.1).

The Egg and Oberbutzen subcatchments are good examples of the valley shoulder catenas found along the central part of the Schaecheen valley’s northern flank (Sect. 2.2). Here, small areas of uHOF1 and uSSF1 drain into thick cone shaped deposits of MX-type material, which are mostly mapped as DP-debris2. The few springs at the bases of the cones have large catchment

Table 3.2: Distribution of DRP classes in the Schaechen (108.3 km²), Vorderschaechen (31.8 km²), and Hinterschaechen (27.0 km²), clustered into four groups.

<table>
<thead>
<tr>
<th>catchment</th>
<th>strong (very fast / fast)</th>
<th>damped</th>
<th>little contributing</th>
<th>very fast, not contributing</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HOF</td>
<td>SOF</td>
<td>SSF</td>
<td>SOF</td>
</tr>
<tr>
<td>Schaecheen</td>
<td>1.4</td>
<td>0.7</td>
<td>13.9</td>
<td>0.6</td>
</tr>
<tr>
<td>Vorderschaechen</td>
<td>0.0</td>
<td>0.5</td>
<td>9.8</td>
<td>0.1</td>
</tr>
<tr>
<td>Hinterschaechen</td>
<td>5.8</td>
<td>0.1</td>
<td>11.8</td>
<td>0.1</td>
</tr>
</tbody>
</table>
areas, but the deposits may also drain underground into the creeping landmass slopes below. These slopes are mapped as SSF3 or DPd. The fact that we measured no strong responses at these sites is thus well recognized in the DRP map.

The Lehnstutzbach catchment has predominantly SSF3 areas, with some DPd areas in the upper region and DPd1 areas at the old moraine terrace the stream has incised through. The stream shows little response during smaller storms, such as the October 2012 event, but had substantial runoff in response to the long-duration storms of August 2005 and June 2013 (Sects. 2.4.1 and 2.5.1). This behavior is a small-scale example of the process change that may cause the extreme flood behavior in the Schaechen, i.e., the four largest floods on record being much larger than the small floods (Sect. 1.2).

3.7.4 Agreement with extreme floods in the Schaechen

The remarkably strong flood response of the Schaechen catchment during the largest floods can be well explained by the dominant runoff process map. The small fraction of areas with a ‘strong’ response (24 %) may explain the low peak discharge during small floods, when the rest of the catchment is hardly activated. 45 % of the catchment has a ‘damped’ response and is expected to contribute substantially only during extreme storms of long duration. This is a large fraction of the catchment, such that it could be responsible for the four remarkably strong floods in the Schaechen flood record (Fig. 1.1). The rest of the catchment ‘contributes little’ to flood runoff, which could explain the low flood peak compared to neighboring catchments (Sect. 1.2).

During the August 2005 flood, the Vorderschaechen and Hinterrhein tributaries received more precipitation than the catchment average, but had similar peak specific discharge (Scherrer AG, 2007). This corresponds with the DRP coverage listed in Table 3.2: the subcatchments have smaller fractions of areas with ‘strong’ response and larger fractions of areas that are ‘unconnected’ or ‘contribute little’ to flood runoff. The small fraction of ‘strongly’ reacting areas in the Vorderschaechen also explains why none of its major floods were caused by short, intense storms (Scherrer AG, 2007).

3.8 Discussion

The developed mapping scheme may only provide a rudimentary characterization of the spatial organization of runoff formation processes in alpine terrain, as it is based on a crude hydrogeomorphological interpretation of mostly coarse, but generally available, spatial information. This rudimentary characterization may be expected to provide useful insights since different alpine slopes can exhibit strongly contrasting runoff responses (Sect. 2.5).

The developed scheme is complementary to the SN-schemes; it is less detailed but mostly uses similar concepts. The main improvements with respect to the SN-schemes are the explicit evaluation of deep subsurface drainage mechanisms, which is particularly relevant for the response to long-duration storms, and the assessment of connectivity of the ‘very fast’ reacting areas. The similarity of the schemes is beneficial in situations where classification of depth and runoff processes are more critical, like in areas where shallow processes determine the
3.9. Conclusions

A classification tool was developed for mapping dominant runoff processes (DRP) in mountainous terrain. Delineation of landscape elements is performed at the hillslope-scale, whereby a geomorphological characterization of common mountainous landforms results in a preliminary classification of sediment cover texture and thickness. Further classification of the DRP is

runoff response. A field visit to obtain soil profiles for a detailed classification according to the SN-schemes then allows a more informed mapping.

The scheme was designed such that there are only few possibilities that erroneous assessment of a single criterion causes deviations of more than one response intensity class (e.g., SSF1 or SSF3 is assigned where it should be SSF2). For example, under- or overestimation by one thickness class in criterion I or criterion VII, (e.g., ‘medium thick’ or ‘very thick’ is assigned instead of ‘thick,’) does not cause deviations of more than one intensity class. The same holds for the assessment of the occurrence of fine-grained sediments in criteria II and VI.

Also, for most landform types, the general assumptions about sediment cover thickness and texture of Table [3.1] are checked against field evidence or indications from available data. For example, rockslide or debris cone deposits are typically assigned a DP type class, as they are expected to be thick deposits of mixed-grained sediments, but may receive a much faster class if indications of shallower runoff formation processes are found in criterion III or X. Criteria like these, with a potentially large influence on the resulting map, should be assessed with much care if a large part of the catchment may be affected by the decision. Often, however, situations where these assessments are very uncertain may be rare, since most catchments are not dominated by a single landform type. For example, alluvial plains, debris cones and rockslide deposits rarely cover large parts of meso-scale alpine catchments.

The obtained DRP map could well explain the damped flood response of the Schaechen and its largest tributaries (Sect. 3.7.4), as well as capture the differences in landform characteristics and stormflow contributions of our research sites (Sect. 3.7.3). This suggests that the map may be useful for qualitative assessment of flood behavior in alpine terrain.

An important advantage of this approach is that areas with ‘damped’ runoff response are recognized and classified. It thereby gives a more complete hydrological description than can be provided by methods that focus mainly on ‘strongly’ reacting areas. This may be highly relevant in alpine terrain, because it appears that even in such steep terrain, large areas may contribute little to even the most extreme floods (Sect. 2.5).

A DRP map may also provide a meaningful basis for a quantitative assessment of flood behavior with a process-based rainfall-runoff model, like the QAREA+ model presented in the next chapter. The combination of mapping and modeling tools allows further testing of the hypothesis that the large fraction of areas with threshold-like response causes the four remarkably large floods in the Schaechen (Sect. 1.2). Distilling useful conclusions from such tests requires further scrutinizing of the mapping and modeling tool in catchments with contrasting behavior to gain more confidence in the predictive power of the tools. This is pursued in Chapter 5.
based on generally available data, such as digital elevation models, aerial imagery and channel network information. Additional data can aid interpretation, and may be obtained through field exploration or from, for example, hydrogeological maps or databases of springs covering the area of interest.

The developed classification tool for mapping of dominant runoff processes is complementary to the schemes of Scherrer and Naef (2003) and Schmocker-Fackel et al. (2007). The classification of dominant runoff processes in the shallow subsurface is directly based on these schemes, whereas the drainage processes at large depth, which are particularly relevant for long-duration storms in steep terrain, receive a more detailed assessment in the new tool. The assessment of connectivity of the most strongly reacting areas is an additional advantage of the developed tool.

The DRP map obtained for the Schaechen represented our small-scale observations well and could qualitatively explain the catchment-scale flood behavior in the Schaechen and its largest tributaries. Areas that only contribute to the more extreme events constitute a large portion of the catchment (~45%), such that they could be responsible for the threshold-like flood behavior. Areas that are expected to contribute negligibly to even the most extreme floods cover another 19 to 25% of the catchment. Knowledge of the occurrence of both types of areas is thus relevant for the understanding of catchment-scale flood runoff behavior. It was further found that a substantial fraction (one-third), of the strongly reacting areas always drains fully into areas with damped runoff response and do not directly contribute to flood formation. Assessment of this kind of connectivity may thus be relevant in alpine terrain.

Further evaluation of the procedure requires a coupling with a rainfall-runoff model that can adequately describe the different dominant runoff processes. This allows a more quantitative assessment of flood runoff behavior. A model is developed and tested in the next chapter. The predictive power of these tools is evaluated in two catchments with contrasting flood behavior in Chapter 5.
Chapter 4

Development and testing of the QAREA model in the Schaechen catchment

4.1 Introduction

Our observations of runoff formation in hillslopes and subcatchments of the Schaechen catchment have guided development of the QAREA rainfall-runoff model presented and evaluated here. The model uses the dominant runoff process (DRP) maps obtained with the classification tool presented in the previous chapter. Relevant modeling paradigms and concepts are reviewed in this section, followed by a detailed description of the developed model in Section 4.2. Parameterization and testing for floods of different magnitudes in the Schaechen are discussed in Sections 4.3 and 4.4. The obtained understanding of the flood behavior of the Schaechen is discussed alongside directions for further research and development in Section 4.5. Further evaluation of the classification and modeling tools in different catchments is presented in Chapter 5.

4.1.1 Relevant modeling paradigms

A model should adequately represent the dominant runoff formation mechanisms to allow simulation of small and large floods. Development of such a process-based model generally proceeds in three steps (Beven, 2012): (1) formulation of a perceptual model, i.e., a collection of perceived relevant processes for the hydrological problem at hand; (2) formulation of a conceptual model, i.e., mathematical descriptions, typically yielding further simplification of the perceptual model; and (3) implementation in a procedural model, usually a computer code. The outcomes of these steps depend on the purpose of the model, the available (expert) knowledge of the system and the constraints posed by, for example, data availability. It remains an outstanding challenge to determine how qualitative and quantitative knowledge about a hydrological system may be used to develop or improve perceptual and conceptual models (e.g., Seibert and McDonnell, 2002; Savenije, 2009).

In this chapter, I demonstrate how a map of dominant runoff processes may guide model development by reducing much of a landscape’s complexity to a relatively simple, explicitly
Development and testing of the QA*r* model in the Schaechen catchment

formulated, perceptual model. The dominant runoff process classification is also a first step in the development of the conceptual model, as it indicates what kind of equations are relevant. For example, there may be no need for a complex infiltration equation if runoff formation occurs predominantly through subsurface stormflow. The classification may also inform the parameterization of the model. For example, one may expect a higher storage capacity in areas with deep percolation than in areas where flows are confined to only a shallow soil cover. The DRP classification may thus allow constraining the model structure and parameters, and thereby impose a certain degree of model internal consistency. Such consistency may be scrutinized by expert judgment and, potentially, increase the predictive power of the model (e.g., Hrachowitz et al., 2014).

A DRP map indicates where similar runoff generation may be expected. The DRP classes are thus a form of Hydrological Response Units (HRUs, e.g., Leavesley et al., 1983; Flügel, 1995). HRU characterizations are often used in conceptual models of the Explicit Soil Moisture Accounting (ESMA) type (Beven, 2012), commonly referred to as box, reservoir, bucket, or tank models. In a fully distributed approach, inputs are defined per delineated element or grid cell, whereas in a semi-distributed approach, inputs and HRUs are aggregated in space, for example in subcatchments or elevation bands. Widely used examples of ESMA-type HRU models include the SWAT model of the USDA (Arnold et al., 1998), the PRMS framework of the USGS (e.g., Leavesley et al., 1983), the HBV model of the SMHI (e.g., HBV-96 in Lindström et al., 1997), and the PREVAH model that has mainly been developed and used in Switzerland (e.g., Gurtz et al., 1999; Viviroli et al., 2009). The model developed here also uses ESMA-type components.

Typically, the same model structure is used for all HRUs and only parameterization is varied (Beven, 2012), the approach of Beighley et al. (2005) and some of the models discussed in the following section being notable exceptions. Most HRU models thereby do not explicitly consider variability in dominant processes and may have many parameters that have to be set through calibration.

Acknowledging that the parameters of a more parsimonious model may be better identifiable (e.g., Kirchner, 2006), the DRP approach simulates only the dominant runoff formation mechanism of an area, using a simple model structure that can adequately describe this process. It has been frequently reported that a simple model structure with a nonlinear, or even constant, partitioning of rainfall into fast and slow flow components can simulate the rainfall-runoff process well (e.g., Naef, 1981; Jakeman and Hornberger, 1993; Beven, 2012). Hence, for this thesis it was tried to develop a model that uses model structures that resemble this two-component approach, while also keeping the structures as similar as possible to allow straightforward comparison of parameter values. This may also allow setting the constraint that parameter values are to be shared between structures, thereby making the overall model more parsimonious; fewer parameters are permitted to be adjusted.

Also note another benefit of parsimony: it is generally easier for the modeler to understand the effects of parameters, both qualitatively and quantitatively. This may be particularly useful when a model is to be used outside of calibration conditions, for example for the simulation of extreme floods.
4.1.2 Examples of models using dominant runoff process maps

Some published rainfall-runoff models are discussed in order to point out weaknesses I tried to amend and strengths I tried to build upon. The discussion is limited to the models that were developed for usage based on the mapping techniques presented in Section 3.1.1.

An important example of a model that uses different structures for different processes is the semi-distributed Tracer-Aided Catchment (TAC) model that was developed for simulating runoff formation and tracer flows on a daily basis (Uhlenbrook 1999; Uhlenbrook and Leibundgut 2000). A fraction of rainfall and snowmelt is routed to a runoff formation routine, and the rest is stored in a soil routine similar to that of the HBV model (e.g., Bergström and Forsman, 1973); the fraction of routed water depends on the fraction of maximum storage and an exponent called “Beta.” The soil routine has no other outflows; soil moisture is depleted by evaporation only. The runoff formation structures range from simple linear reservoirs to combinations of reservoirs with up to four different outlets activated by threshold exceedance or smooth nonlinear storage-discharge relations. This structural complexity, combined with the large differences between structures, complicate the interpretation of how parameters determine the simulated responses.

The model was later converted into a distributed raster model that allowed simulation at shorter time steps (Uhlenbrook et al., 2004). Some model structures were simplified, but other complexities were added: outflows may move down in a catena through multiple grid cells, and the storage-discharge relations are scaled by the local topographic gradient. With some modifications the model was also used in the Austrian Alps by Johst et al. (2008), based on the mapping approach of Tilch et al. (2006), and in a Himalayan catchment using a coarser HRU delineation (Konz et al., 2007). Neither of these studies reported adequate flood simulation skill and the model is relatively complex.

The ‘Beta concept’ mentioned above is used in many ESMA-type models where it facilitates a nonlinearly increasing storage-discharge relation that partitions inflow into two components. The actual equation differs between models, as does the pre- and post-processing of the storage inputs and outputs. Examples are found in the PREVAH model (e.g., Gurtz et al., 1999; Vivirioli et al., 2009) and PDM model (Moore, 1985, 2007), as well as in modular frameworks like FUSE (Clark et al., 2008) and SUPERFLEX (Fenicia et al., 2011). I will use this concept for the generation of fast runoff from a reservoir that is otherwise depleted by evaporation or a linear storage-discharge relation (see Sect. 4.2).

Maps obtained through the classification of Markart et al. (2004) have been used with the ZEMOKOST model, developed for estimating design floods in small-scale catchments with a time of concentration approach (e.g., Markart et al., 2006). At larger scales, these maps have been combined with “hydrogeologic runoff process maps” to set up the raster-based HRU model of Blöschl et al. (2008) for continuous simulations in meso-scale catchments, for example, by Rogger et al. (2012). In their study, the HRUs have the same model structure and make use of the ‘Beta concept,’ but some or all of the groundwater reservoirs can be deactivated by setting the percolation rate to zero or setting the percolation proportioning factor to zero or unity. This strategy for defining structures is also used in the model presented here.
DRP maps obtained through Scherrer and Naef classifications (Scherrer and Naef, 2003) have been used in conjunction with the event-based runoff model QAREA (e.g., VAW, 1994; Horat, 2000). It is here discussed in detail because it was used as reference for parameterization of the ‘strongly’ reacting DRP classes in the QAREA+ model (see Sect. 4.3.3). To make the distinction with the QAREA+ model more clearly, it referred to as QAREA-C, with the ‘C’ to denote its explicit use of response curves.

A response curve relates the direct runoff coefficient $C_{R,d}$ to cumulative rainfall ($V_p$) for each runoff type (RT; see Fig. 4.1). Runoff types are groups of DRP with similar response strength. For example, the RT2 class consists of the SSF2 and SOF2 process classes. The curves were derived from $C_{R,d}$ vs. $V_p$ relations measured in plot-scale sprinkling experiments (Scherrer, 1996; Scherrer et al., 2007). The direct runoff is routed through a ‘surface delay’ linear reservoir $S_{ob}$ [L] with time constant $K_{ob}$ [T], the remainder recharges a slowly draining groundwater reservoir $S_s$ [L] with time constant $K_s$ [T]. The value of $K_{ob}$ for all runoff types is in the order of hours, whereas $K_s$ is in the order of ten to hundreds of hours, with higher values for slower runoff types. The nonlinear partitioning of rainfall over the parallel fast and slow linear reservoirs is a good example of the parsimonious two-component models discussed above.

The model is currently used at Scherrer AG for flood estimation based on DRP maps. The value of $K_{ob}$ varies only slightly among different catchments; the $S_{ob}$ reservoir usually has faster response in catchments with high relief. Because of the confidence in the prediction of fast processes with this model, it is here used to set some of the parameters of the fast processes of the QAREA+ model.

The QAREA-C model needs improvement for long-duration storms because the response time scales are either very short, or very long, and drainage processes at intermediate time scales cannot be adequately described (i.e., delayed flow peaks). The often found overestimated decline of the receding limb is also rooted in this. Besides, the curves of the more delayed DRP are based on little data, such that more research into the behavior may be useful. These issues may be dealt with when a more process-based model structure is used, particularly if it can be meaningfully parameterized.

Such a more process-based approach, called QAREA-Pro, was developed by Schmocker-Fackel (2004). It uses a different model structure for each dominant runoff process, similar to the approach of Uhlenbrook et al. (2004). The model structure of the Subsurface Stormflow (SSF) process is complex and its parameterization is not straightforward, such that this is not maintained in the QAREA+ model. Also, the model structure for the Deep Percolation (DP) process, a simple linear reservoir, may not adequately describe the moderately fast drainage component of our DP type research sites; only a combination of reservoirs or lagging functions can simulate the kind of delayed runoff peaks we observed (Sect. 2.5).

Hellebrand et al. (2011) used four different model structures for the various DRPs mapped in Luxembourg using the method of Müller et al. (2009). The structures are very simple and cannot simulate delayed flow peaks. Hellebrand et al. (2011) found the simulation skill at seasonal time scales was comparable to that of a lumped model, but did not evaluate flood prediction skill.
Figure 4.1: The QAreea-C model concept, with response curves for partitioning rainfall into fast and slow runoff components $q_{ob}$ and $q_s$. Each runoff type (RT) $i$ has a response curve for relating relating direct runoff coefficient $C_{R,d}$ to cumulative rainfall $V_P$ (solid lines). A fraction $C_{R,d}$ of rainfall fills the fast linear reservoir $S_{ob}$ with reservoir constant $K_{ob}$. The remainder is routed to the slow linear reservoir $S_{s,i}$, which has a different reservoir constant $K_{s,i}$ for each runoff type. The fractions of accumulated rainfall directed to the fast reservoir ($V_{Q,fast} / V_P$) are indicated with dashed lines in corresponding grey tones.

4.1.3 Directions for the development of the model

The following conclusions, drawn from above reviews of model paradigms and concepts, have guided the design of the QAreea’ model presented in the next section:

1. Model structure should resemble the perceptual model of the processes formulated in the mapping scheme. This means each DRP type (e.g., HOF, SOF, etc.) should have its own adequate model structure.

2. A parameter-efficient model structure has a nonlinear mechanism for determining the fraction of incoming rainfall that quickly becomes runoff. One ‘parallel’ slow flow component is often sufficient to simulate the remainder.

3. The often used ‘beta concept’ for nonlinear proportioning of rainfall into fast and slow components requires only two parameters for relating the proportioning to the storage
Development and testing of the QA/r.sc/e.sc/a.sc model in the Schaechen catchment

4. The different structures may be derived from a modular ‘parent’ model structure; the different DRP structures share model components and parameters as much as possible. This also allows comparison of parameters that are not shared.

5. Formation of a delayed runoff peak may be simply described by putting reservoirs in series, whereby the delay depends on the storage states in the respective reservoirs.

4.2 The QA/r.sc/e.sc/a.sc model

4.2.1 Model structure

The developed rainfall-runoff model, named QA/r.sc/e.sc/a.sc, is spatially distributed, whereby delineated landscape elements (see Sect. 3.5) serve as the computational units. Each landscape element $i$ contributes discharge $Q_i$ [L$^3$ T$^{-1}$], the product of its specific discharge $q_{t,i}$ [L$^1$ T$^{-1}$] and planar area $A_i$ [L$^2$], with routing delay $t_{lag,i}$ [T] to the total simulated discharge $Q_{T,sim}$ [L$^3$ T$^{-1}$] at time $t$ [T], at some point of interest in the stream network, such that:

$$Q_{T,sim}(t) = \sum_{i=0}^{k} q_{t,i} A_i (t - t_{lag,i})$$

with $k$ the number of elements contributing to the point of interest.

As snowfall and -melt are not considered, the required inputs to each modeled element consist only of precipitation $P_i$ [L$^3$ T$^{-1}$] and potential evaporation $E_{p,i}$ [L$^3$ T$^{-1}$]. Interactions between the elements are not possible, except for areas where generated runoff re-infiltrates fully into a downslope element with large storage capacity (the ‘unconnected’ areas mapped as uHOF1, uSOF1 or uSSF1, see Sect. 3.6.2). Runoff generated in an unconnected element is added directly to the precipitation input of the associated downslope elements.

Like in the approaches of Uhlenbrook and Leibundgut (2000), Schmocker-Fackel (2004) and Hellebrand et al. (2011), a different model structure is used for each of the five dominant runoff processes; HOF, SOF, SFF, DP-fast and DP-slow. For ease of writing, a structure parameterized for a certain DRP class is sometimes called ‘model,’ i.e., the ‘SSF3 model.’ Each structure consists of two or three reservoirs, out of the four reservoirs that make up the QA/r.sc/e.sc/a.sc master model structure (Fig. 4.2). The reservoirs have well-established relations for representing processes with few parameters. In this respect, the QA/r.sc/e.sc/a.sc master structure allows less complexity and freedom of choice than other flexible model frameworks like FUSE (Clark et al., 2008) and SUPERFLEX (Fenicia et al., 2011), however it permits simple application in a spatially distributed setting.

The five different model structures are derived from the master structure by switching individual components on or off with the diversion parameters ‘$\alpha$’ [-], or setting the infiltration rate limit $L_i$ [L$^1$ T$^{-1}$] (details discussed in Sect. 4.2.2). The aim of this approach is to ease comparison of parameters within and across the different structures. This is useful if the modeler pursues a coherent parameterization or wants to set constraints on the parameter relations or
model behavior. It allows a relatively straightforward incorporation of expert knowledge into the parameterization, which may be expected to improve the process consistency in the models (e.g., Seibert and McDonnell, 2002; Gharari et al., 2014; Hrachowitz et al., 2014). If the constraint is formulated that a parameter value must be shared between different process classes this results in a more parsimonious parameterization problem because the number of 'effective' parameters that may be varied is reduced. These issues are discussed in more detailed in Section 4.3 on the model parameterization.

The model structures and associated DRP classes are summarized in Table 4.1 together with their respective groupings for crude interpretation of DRP maps (Sect. 3.7) and a brief description of the mechanisms and their occurrence in alpine terrain.

Table 4.1 also shows how the Loss-Karst and DP classes are grouped together to reduce the number of model parameters and allow parameterization of DRP classes for which we have no rainfall-runoff observations. The DP-rock and DP-debris1 classes, with a damped response, form the DP-fast group, which has a model structure that allows a relatively fast groundwater response that may contribute considerably during long-duration events. Likewise, the DP-debris2 and Loss-Karst classes are merged into the DP-slow group, which has only a slowly reacting groundwater reservoir.

### 4.2.2 Flux equations

Each model structure uses the upper zone reservoir $S_u [L]$ for interacting with the atmosphere (via $P$ and $E_p$). It has has both linear and nonlinear response relations. The other reservoirs drain only with a linear storage ($S$) - discharge ($q$) function with a reservoir constant $K [T]$:

$$q_x (t) = \frac{S_x (t)}{K_x}$$  \hspace{1cm} (4.2)

where the subscript $x$ indicates the reservoir (Fig. 4.2): the routing delay reservoir storage $S_d [L]$, the fast draining groundwater reservoir storage $S_{g,f} [L]$, and the slowly draining groundwater storage $S_{g,s} [L]$, each with corresponding reservoir constant $K_d$, $K_{g,f}$ and $K_{g,s} [T]$, respectively. In absence of new inputs, such linear reservoirs have exponentially decreasing outflows with characteristic 'half-lives' of $\ln (2) K \approx 0.69K$, and about 5% of volume left after $\ln (20) \approx 3K$; i.e., 95% is discharged after 3 days if $K = 24 h$.

The first nonlinear component of the upper reservoir, $S_u$, is infiltration excess (Hortonian) overland flow $q_h [L \cdot T^{-1}]$. It is defined as the precipitation rate exceeding the infiltration rate limit $L_i$. The remainder, effective precipitation $P_{\text{eff}} [L \cdot T^{-1}]$, infiltrates into the $S_u$ reservoir:

$$q_h (t) = \max (P(t) - L_i, 0)$$  \hspace{1cm} (4.3)

$$P_{\text{eff}} (t) = \min (P(t), L_i)$$  \hspace{1cm} (4.4)

This rather simple relation for describing Hortonian overland flow was preferred over more complex equations because it is mainly applied for areas with very limited infiltration and storage capacities, like bare rock faces, where any noteworthy rainfall rate exceeds infiltration capacity.

A fast outflow from the $S_u$ reservoir occurs if the storage depth exceeds the threshold $S_{u,pf} [L]$. Subscript 'pf' stands for preferential flow, indicating runoff through a low-resistivity
Development and testing of the QAREA* model in the Schaechen catchment

Figure 4.2: Structure of the QAREA* master model (a) and derived model structures for the different dominant runoff processes (b). Diversion regulators (grey circles) determine how structures are derived from the master structure. The process classes and corresponding model components are summarized in Table 4.1.
The QA/REA model structures for the dominant runoff processes, with description of behavior and occurrence in alpine terrain (Sect. 3.6). The grouping of Sect. 3.7 is indicated in the ‘Group’ column: ‘Strong’ refers to DRP classes with substantial response already during small floods, ‘Damped’ to classes that mainly contribute during extreme floods, and ‘Little contributing’ to classes that contribute negligibly to stormflow.

<table>
<thead>
<tr>
<th>Model Structure</th>
<th>Components</th>
<th>DRP</th>
<th>Group</th>
<th>Behavior and occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hortonian Overland Flow (HOF)</td>
<td>( L_i, S_u, S_d )</td>
<td>HOF1</td>
<td>Strong</td>
<td>Very fast response due to limited infiltration capacity and depression storage; e.g., areas with bare rock, roads.</td>
</tr>
<tr>
<td>Saturation Overland Flow (SOF)</td>
<td>( S_u, S_d )</td>
<td>SOF1</td>
<td>Strong</td>
<td>Very fast response due to rapid saturation of soils with little storage capacity; e.g., swampy areas and flat terrain with little soil cover or shallow groundwater table.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SOF2</td>
<td>Strong</td>
<td>Fast response due to saturation of soil with moderate storage capacity; e.g., urban areas, areas compacted soils or moderately shallow groundwater table.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SOF3</td>
<td>–</td>
<td>Delayed response due to saturation of soils with large storage capacity; not considered in the schemes of Sect. 3.6.</td>
</tr>
<tr>
<td>Subsurface Stormflow (SSF)</td>
<td>( S_u, S_d, S_{g,s} )</td>
<td>SSF1(^a)</td>
<td>Strong</td>
<td>Very fast response due to quick activation of lateral (preferential) flowpaths; e.g., areas with shallow sediment cover (very rugged terrain).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SSF2</td>
<td>Strong</td>
<td>Fast response due to activation of lateral (preferential) flowpaths; e.g., areas with ‘moderately thick’ sediment cover (rugged terrain).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SSF3</td>
<td>Damped</td>
<td>Damped response because the activation of lateral (preferential) flowpaths requires considerable wetting up; e.g., steep slopes with ‘thick’ sediment cover and well-drained talus cones.</td>
</tr>
<tr>
<td>Deep Percolation -fast (DP-fast(^b))</td>
<td>( S_u, S_{d}, S_{g,s} )</td>
<td>DP-rock</td>
<td>Damped</td>
<td>Damped response with only considerable contribution to stormflow during long-duration events, due to highly elevated groundwater level in the fissured rock; e.g., areas where karstification or mass movements have caused heavy fissuring of the rock.</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Damped response of groundwater body in sediments with only considerable contribution to stormflow during long-duration events; e.g., slopes with ‘very thick’ sediment cover and steep cones draining via deposits with substantial fine-grained material content.</td>
</tr>
<tr>
<td>Deep Percolation -slow (DP-slow(^b))</td>
<td>( S_u, S_{g,s} )</td>
<td>DP-debris2</td>
<td>Little contributing</td>
<td>Strongly damped response of relatively flat groundwater body with long flowpaths, such that even during the most extreme events little stormflow is produced; e.g., flat moraine deposits, cones with flat feet, and rockslide deposits.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Loss-Karst</td>
<td>Little contributing</td>
<td>Not contributing to flood runoff, because of expected drainage to neighboring basins via karstic conduit systems.</td>
</tr>
</tbody>
</table>

\(^a\) Catchment average: No \( S_{g,s} \) component as no groundwater recharge is expected in the direct vicinity of the stream network.

\(^b\) Catchment average: b DRP classes in this row receive the same parameterization in the model and are thus effectively merged into a newly defined model class, i.e., DP-fast and DP-slow.
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medium, like soil pipes, weathered rock, or the soil surface (in the case of saturation overland flow; SOF). Above this threshold a fraction of effective precipitation $P_{\text{eff}}$ "runs off" rapidly as $q_f$ [LT\textsuperscript{-1}]. As in the QA\textsuperscript{R}A-C model, this fraction is referred to as the direct runoff coefficient, $C_{\text{R,d}}$ (Sect. 4.1). It is determined by the maximum storage depth $S_{\text{u,\text{max}}}$ [L], $S_{\text{u,pf}}$, and exponent $\beta$ [-]:

$$q_f (t) = C_{\text{R,d}}P_{\text{eff}} (t) = \left(\hat{S}_{\text{u,pf}} (t)\right)^\beta P_{\text{eff}} (t)$$

whereby $\hat{S}_{\text{u,pf}}$ [-] is the fraction of storage possible above $S_{\text{u,pf}}$:

$$\hat{S}_{\text{u,pf}} (t) = \max \left(\frac{S_u (t) - S_{\text{u,pf}}}{S_{\text{u,\text{max}}} - S_{\text{u,pf}}}, 0\right)$$

The exponent is called $\beta$ because many ESMA-type models use this symbol for the exponent of ratios that define flux behavior (e.g., HBV and PREVAH), though not necessarily with the same $C_{\text{R,d}}$ vs. $S_u$ relation (see Sect. 4.1.1). It must be greater than one to produce progressively more direct runoff with increasing storage.

The linear drainage of the upper zone, $q_u$ [LT\textsuperscript{-1}], occurs as long as storage depth $S_u$ exceeds $S_{\text{u,fc}}$ [L], with 'fc' denoting field capacity. This requires a slight modification of Eq. (4.2):

$$q_u (t) = \max \left(\frac{S_u (t) - S_{\text{u,fc}}}{K_u}, 0\right)$$

Evaporation from the upper zone reservoir, $E_u$ [LT\textsuperscript{-1}] occurs at potential rate, $E_p$, for storage above $S_{\text{u,fc}}$ and decreases linearly with moisture content:

$$E_u (t) = \min \left(\frac{S_u (t)}{S_{\text{u,fc}}}, 1\right) E_p (t)$$

The model structure is adapted to the dominant runoff process by switching components on or off. The Hortonian overland flow component is only used for the HOF1 type areas, and switched on by setting a low infiltration limit $L_i$. The process is switched off by setting this limit very high, to 100 mm h\textsuperscript{-1}. The $q_f$ and $q_u$ fluxes are routed with the diversion regulators $\alpha_f$, $\alpha_u$ and $\alpha_r$ [-], which have values in the range [0, 1]. A fraction $\alpha$ is routed in one direction, and a fraction $1 - \alpha$ in the other. The $S_d$ reservoir is only turned on or off, by setting $\alpha_f$ to 0 or 1, respectively, as it is intended only to delay instantaneously generated runoff fluxes $q_h$ and $q_f$. The diversion of $q_u$ may be less binary, as some fractional loss to the groundwater reservoir(s) may be expected for some processes. The bulk recharge, $q_r$ [LT\textsuperscript{-1}], is the combined recharge from the $q_f$ and $q_u$ fluxes, and it either fills only the slow groundwater reservoir $S_{\text{g,s}}$ (DP-slow model structure, i.e.: $\alpha_r = 1$) or is subdivided in fixed proportions over the fast and slow groundwater reservoirs (DP-fast model structure).

The total outflow $q_{\text{tot},i}$ of element $i$ is the sum of its output fluxes:

$$q_{\text{tot},i} = (1 - \alpha_u) q_u + q_d + q_{\text{g,f}} + q_{\text{g,s}}$$

These element total outflows are combined in Eq. (4.1) to arrive at the catchment total outflow.
4.2.3 Procedural implementation

Combining above relations with the water balance equations of the four reservoirs gives a system of ordinary differential equations (ODEs; below, the dependence on $t$ is not explicitly written out):

$$\frac{dS_u}{dt} = P_{eff} - E_u - q_t - q_u \quad (4.10)$$

$$\frac{dS_d}{dt} = q_h + (1 - \alpha_f) q_t - q_d \quad (4.11)$$

$$\frac{dS_{g,f}}{dt} = \alpha_r (\alpha_f q_t + \alpha_u q_u) - q_{g,f} \quad (4.12)$$

$$\frac{dS_{g,s}}{dt} = (1 - \alpha_r) (\alpha_f q_t + \alpha_u q_u) - q_{g,s} \quad (4.13)$$

It is important to solve such a system of ODEs with an appropriate numerical integration technique (e.g., Clark and Kavetski, 2010; Kavetski and Clark, 2010; Beven, 2012). The model is coded in the open-source, general purpose programming language Python 2.7, for which the SciPy package (Jones et al., 2001) provides advanced numerical integration techniques. The explicit Dormand and Prince method (Runge-Kutta method of order (4, 5) with adaptive step size control) gives good results, but requires about ten times more computation time than simple Explicit-Euler integration.

As I found no substantial differences in the model behavior for the events simulated in this research (small time steps, moderate precipitation intensities), the simulations presented here used Explicit-Euler integration at a temporal resolution $\Delta t$ of 10 minutes. This is not recommended for operational use of the model, but it was here used to be consistent with the Monte Carlo simulations with model structures derived from the QARea+ master model, which are presented in chapter 5 here; the reduction in computation time was really helpful.

4.2.4 Delays in the stream network

A constant routing lag $t_{lag}$ is determined for each subcatchment from the slope and stream length of the main channel. Subcatchments are defined such that no side stream has an along-stream distance to the subcatchment outlet that is more than twice as long as the subcatchment main channel reach. The side streams are often steeper than the main channel, such that the travel time in the main reach may well serve as a proxy for the average travel time in the subcatchment. Besides, the lower gradient in the main channel is assumed to be in the range where the channel average flow velocity can be estimated with the Manning-Strickler formula. This is a simplification of the lag time determination in the QARea-C model, which requires estimating the travel times in all channels.

Here, a hydraulic radius of 1 m and Manning roughness coefficient of 0.07 are used (after Barnes, 1967, who lists roughness coefficients between 0.06 and 0.08 for similar mountain streams), and a maximum flow velocity of 5 m s$^{-1}$ is set. Slope-corrected reach length and average streambed gradient are estimated from the DHM25 digital elevation model.
Figure 4.3: Subcatchments of the Schaechen catchment with average gradients of the main channel reaches indicated. The numbers represent the time lags to the Schaechen outlet in minutes.

(Swisstopo, 2005) along the stream. Estimated total lag time to any point of interest is rounded to the nearest multiple of the simulation time step $\Delta t$.

Figure 4.3 displays reach-averaged streambed gradients and the estimated travel time to the Schaechen outlet. The total time lag of the steeper tributaries depends mostly on the travel time in the flatter main valley, such that inaccuracies in the application of the Manning-Strickler formula to steep streams have little effect at the catchment scale. Further note that uncertainties in the estimation of the delays in the stream network at the scale of the Schaechen are typically smaller than the estimation of the runoff formation process (e.g., Beven, 2012). Particularly for long-duration storms in steep catchments, estimation of the delays in the stream network may thus play only a secondary role.

### 4.3 Model parameterization

Only a few parameters could be directly related to the classification. The remaining parameter values were obtained via manual adjustment to three different sets of rainfall-runoff response data: (1) the parameters of the DP-fast and DP-slow models were set to match our observations at the Schluecht and Gadenstetten sites (Chap. 2) during two extreme events; (2) the parameters of the fast-responding DRP, except for the delay reservoir constant $K_d$, were set by adapt-
4.3. Model parameterization

4.3.1 Guidelines to constrain the parameterization

The model was designed to represent processes parsimoniously, with the DRP classification as a framework to indicate qualitative differences between landscape elements. Parameterization implies quantification of the processes, but I tried to put representation of the qualitative differences first, and not simply optimize parameter values by minimizing the mismatch between observations and simulations as quantified by some goodness-of-fit metric. Besides, the right data for fitting the model to simply do not exist, making any such optimization a rather futile exercise; data that show the average behavior of a class are just not available.

The DRP classification is thus used to constrain the parameterization by formulating assumptions and relations that help to establish a coherent parameterization that is both qualitatively and quantitatively sensible:

1. As much as possible, parameter values are shared within and between DRP types. This directly promotes an internally consistent parameterization:

   a) The idea is that if the DRP classes share certain properties in the classification, like storage depth (more on this in points 4 and 5), these properties should receive the same parameterization. This makes the best use of the classification and actually forces the modeler to choose parameters such that they are consistent with the perceptual models of the dominating processes.

   b) The sharing effectively reduces the number of parameters that may be adjusted, and thereby makes the parameterization more parsimonious. This generally makes the parameterization process more straightforward and possibly makes the ‘effective’ parameters better identifiable (Sect. 4.1.1).
2. It is assumed that recharge of the slow groundwater reservoir occurs predominantly in DP and SSF areas that have enough storage available to prevent much rapid runoff (SSF2 and SSF3). This is reflected in the model structures (Fig. 4.2), whereby higher recharge ratios in DP areas than in SSF areas are expected because DP responses are more damped. Therefore, the fraction of upper zone outflows routed to the slow groundwater reservoir in the SSF models should be less or equal to the total fraction of water routed to the slow groundwater reservoir in the DP-fast model; \( \alpha_{u,SSF} \leq 1 - \alpha_{r,DPf} \).

a) These are the only diversion regulators that are not determined by the definition of the structure, whereby \( \alpha_u \) is assumed to be roughly the same for SSF2 and SSF3 models to reduce the number of effective parameters (see point 1 above).

b) The diverted fractions should be substantial, i.e., more than 0.10, because low values effectively mean the model structure is changed; a reservoir is almost disconnected. The structure should represent the perceptual model and thus the \( \alpha \) diversion parameters cannot be chosen freely.

c) It is assumed that there is no recharge of a slow groundwater component in SOF and HOF areas, because they typically have low permeability, or, in the case of SOF, a shallow groundwater table sustained by seepage. The corresponding model structures therefore have no groundwater reservoirs.

3. Following point 1 values of the reservoir constants \( K \) are shared when possible. For example, all models may have the same slow groundwater reservoir constant, \( K_{g,s} \); it may be seen to be more a catchment property than a hillslope property, and its actual value should have little effect on the simulation of the flood event such that there is little reason to tune this parameter per DRP class. Likewise, \( K_u \) may be shared between the classes of the same DRP type: for example, all SOF-type classes have the same \( K_u \), because they are expected to have a relatively slow-draining unsaturated zone. The same holds for the SSF-type classes, which have a relatively fast-draining unsaturated zone. In addition, some relational constraints about differences between \( K \) values can be formulated:

a) Drainage from the slow groundwater reservoir \( S_{g,s} \) must be slower than drainage from the fast groundwater reservoir \( S_{g,f} \), i.e., by roughly one order of magnitude or more: \( K_{g,s} \geq 10K_{g,f} \). This is to not deviate from the perceptual model, i.e., that such slopes have fast- and slow-draining groundwater components that are strongly different.

b) The fast groundwater reservoir \( S_{g,f} \) may not have a faster response than the upper reservoir \( (S_u) \) draining to it; \( K_{g,f} \geq K_u \). This corresponds to the perception that deeper processes are typically more damped.

c) SOF areas drain substantially slower than the more permeable SSF areas (see point 3), but the difference may be smaller than an order of magnitude as the unsaturated zones in both types of areas may be expected to hold little water against gravity after several days (see discussion on field capacity in Sect. 2.5.2). Following
these assumptions, the relation between the parameters can roughly formulated as
\[ 2K_{u,SSF} \leq K_{u,SOF} \leq 10K_{u,SSF}. \]

4. In the SOF model, the slow drainage \( q_u \) component can be assumed to contrast strongly with the fast flow component \( q_f \) that is activated when the storage threshold \( S_{u,pf} \) is exceeded: the low permeability of the soils associated with SOF processes drain slowly, whereas the overland flow that is activated by saturation of the top soil may reach the stream with little delay. This results in a simple two-component nonlinear model (Sects. 4.1.1). Storage capacity \( S_{u,max} \) and \( S_{u,pf} \) can thereby be related to definitions of the storage classes formulated by Schmocker-Fackel et al. (2007), i.e.: SOF1 has a storage capacity of 0 to 40 mm, SOF2 of 40 to 100 mm, and SOF3 of 100 to 200 mm.

a) A uniform distribution of storage depths within a class may be assumed, implying a \( \beta_{SOF} \) value of 1.

b) Some relations between field capacity \( S_{u,fc} \), maximum storage depth \( S_{u,max} \) and soil depth may be assumed. Literature values of available water capacity, the difference in volumetric water content between field capacity and wilting point, range between 10 and 20 % for most soils. A similar range is typically found for the available storage between field capacity and full saturation, the latter typically lying between 40 and 50 % (e.g., Blume et al., 2010). A good starting point may therefore be: \( S_{u,max} - S_{u,pf} = S_{u,fc} = 0.1D \), with \( D [L] \) the maximum available soil depth of the DRP class.

5. Similarities and differences between the SSF and SOF model parameters may be assumed:

a) For the same intensity class, SSF maximum storage and field capacity should equal those of the SOF structure, because they are also equal in the classification scheme (the intensity class is directly determined by the maximum available storage).

b) As no overland flow may occur in SSF areas, a stronger nonlinear runoff generation \( q_f \) and more effective drainage (see point [3c]) may be expected. This is achieved by setting \( \beta_{SSF} > 1 \). A first estimate of \( \beta_{SSF} \) may be 2, which in comparison to a \( \beta \) of 1 gives a slower increase of \( C_r,d \) with \( S_u \) until \( S_{u,pf} = 0.5 \) and stronger increase above (i.e., the \( C_r,d \) vs. \( \dot{S}_{u,pf} \) relation is a straight line with slope of 1 for \( \beta = 1 \), and a parabola for \( \beta = 2 \)).

c) Besides, the dominance of subsurface stormflow in SSF classes with similar storage capacity as SOF classes suggests that the threshold for activation of fast processes must be lower or equal, i.e.: \( S_{u,pf,SSF} \leq S_{u,pf,SOF} \). Note that the actual contribution of the \( q_f \) flux may still be smaller than in the corresponding SOF model, due to the higher \( \beta \) and lower \( K_u \).

6. The HOF1 areas in alpine terrain consist mostly of bare rock faces. These may have some depression storage that is evaporated or drained rapidly, such that \( S_{u,max} \) is similar to a maximum interception storage. It is here set to 10 mm, roughly the average of what is
found in the literature. For example, Kamphorst et al. [2000] computed a maximum depression storage of 13 mm for soils and Terstriep and Stall [1974] suggested 5 mm for use in a rainfall-runoff model in urban settings. The actual parameterization of $S_u$ for HOF1 is not so important as long as it is appreciated that according to the perceptual model the storage capacity is smaller than in the SOF1 and SSF1 models and the maximum infiltration rate $L_i$ is low.

Constraining the parameterization with these guidelines in mind is difficult with automatic procedures. I therefore adjusted the parameters manually by inspecting the simulated hydrographs and evaluating the behavior of the individual runoff components according to differences and similarities the guidelines indicate they should exhibit. The event runoff coefficient $C_{R,e}$ (Eq. (2.3)) was also inspected as it provides a measure of how damped the response is and allows simple comparison with observations and between different DRP classes; behavior of different DRP classes should be substantially different in terms of short-term storage and flow rates.

Three goodness-of-fit metrics were used to evaluate the model simulations and check the direction of the model behavior when parameters are adjusted. As stated earlier, parameter values were not optimized to maximize one or more goodness-of-fit metrics, but the metrics proved useful to compare effects of parameter adjustments.

The three goodness-of-fit metrics that were used are the Nash-Sutcliffe Efficiency, NSE [-] (Nash and Sutcliffe, 1970), the Volumetric Efficiency, VE [-] (Criss and Winston, 2008), and the Mean Error, ME (dimensions of the flux of interest):

$$\text{NSE} = 1 - \frac{\sum_{t=t_{\text{first}}}^{t=t_{\text{last}}} (F_{\text{calc}}(t) - F_{\text{obs}}(t))^2}{\sum_{t=t_{\text{first}}}^{t=t_{\text{last}}} (\bar{F}_{\text{obs}}(t) - F_{\text{obs}}(t))^2} \quad (4.14)$$

$$\text{VE} = 1 - \frac{\sum_{t=t_{\text{first}}}^{t=t_{\text{last}}} |F_{\text{calc}}(t) - F_{\text{obs}}(t)|}{\sum_{t=t_{\text{first}}}^{t=t_{\text{last}}} F_{\text{obs}}(t)} \quad (4.15)$$

$$\text{ME} = \frac{\sum_{t=t_{\text{first}}}^{t=t_{\text{last}}} (F_{\text{calc}}(t) - F_{\text{obs}}(t))}{n} \quad (4.16)$$

with $n$ the total number of time steps, including the first and last time step of interest ($t_{\text{first}}$ and $t_{\text{last}}$, respectively). The fluxes, $F$, are denoted with ‘calc’ and ‘obs’ subscripts to indicate calculated and observed values, and the mean flux is indicated with a bar, i.e., $\bar{F}$. .

NSE and VE increase to up to 1 for a perfect fit and may become negative, which for the NSE means that the calculated data give a worse fit to the observations than a simple mean would. Because of the squared residuals in the NSE, this goodness-of-fit metric is particularly sensitive to large differences between observed and calculated data, which makes it more sensitive to peak flow if relative deviations are roughly the same for the rest of the time series. The VE treats “every cubic metre of water the same as any other cubic metre”, and, arguably, allows better comparison between fits to different hydrographs (Criss and Winston, 2008). The Mean Error is mainly used to indicate the average deviation of the model from the observations in meaningful units, with positive values indicating overestimation by the model, and negative values indicating underestimation. The goodness-of-fit scores of the presented simulations are listed in Appendix E.
4.3. Model parameterization

4.3.2 Obtaining parameters of the DP-fast and DP-slow models from observations of large events at the Schluecht and Gadenstetten sites

We have good observations of the Schluecht and Gadenstetten responses to the long-duration storms of October 2012 and June 2013. These storms caused floods in the Schaechen with return periods of 2.3 and 6.6 years, respectively (Sect. 1.2). The sites and events are discussed in detail in Chapter 2. The Schluecht site is a typical example of a DP-rock slope and the observations are used to parameterize the DP-fast model structure (DP-fast comprises the DP-rock and DP-debris1 classes, expected to contribute little to small events, but substantially to extreme events; see Table 4.1). The Gadenstetten site is an example of a DP-debris2 slope that contributes little to flood runoff formation. Observations of the runoff response are here used to parameterize the DP-slow model.

Model setup for simulating the events

For each simulation, the model was started \( t_{\text{start}} \) about 24 hours after the previous storm and run until the next event \( t_{\text{end}} \). The inputs consist of 10 minute resolution rainfall data measured on-site and a constant potential evaporation rate \( E_p \) of 2 mm d\(^{-1}\). Simulation time steps were also 10 minutes and goodness-of-fit metrics were computed over the full simulation period. The initial conditions were set as follows: \( S_u(t_{\text{start}}) = S_{u,fc} \), \( S_{g,f}(t_{\text{start}}) = 0 \), and \( S_{g,s}(t_{\text{start}}) = q_{\text{obs}}(t_{\text{start}}) \times K_{g,s} \). The assumption that the upper zone is at field capacity is defensible, as frameworks for practical application of the field capacity concept commonly define field capacity as the moisture state of the soil 1 to 3 days after a large storm (e.g., WMO, 2008). Our soil moisture observations at the Schluecht slope also showed such behavior (Sect. 2.5.1): volumetric water content returns to almost pre-event conditions within days. This was even the case after the application of more than 800 mm during our plot-scale sprinkling experiment (Volze, 2015).

The initial state of the two groundwater reservoirs in the DP-fast model can be easily related to observed discharge from inversion of Eq. (4.2) if one of the reservoirs is assumed to be empty. The fast reservoir is assumed empty because it drains much faster than the slow reservoir. The fast groundwater reservoir will not receive inflow as long as \( S_u \) is below field capacity, such that it remains empty until the start of the first storm.

Results

Simulations of the DP-fast model at the Schluecht slope are compared to observations for the October 2012 and June 2013 events in Figures 4.4 and 4.5. Simulations of the Gadenstetten runoff responses with the DP-slow model are presented in Figures 4.6 and 4.7. The same periods were used for the two sites. Axis scales are set per event, to allow straightforward comparison of the simulated and observed behavior at the two sites. The fast runoff component of the upper reservoir, \( q_f \), was not activated in the simulations and therefore not presented in the figures.
Figure 4.4: Simulations of the DP-fast model for the October 2012 event at the Schluecht site. (a) Observed rainfall and cumulative rainfall used as model input. (b) Observed and simulated runoff, $q_{\text{obs}}$ and $q_{\text{tot}}$, contribution of $q_u$ to the bulk recharge, $q_{u-tot}$, and outflow from the slow groundwater reservoir, $q_{g,s}$. (c) Model error ($q_{\text{tot}} - q_{\text{obs}}$) and model deviation as percentage of $q_{\text{obs}}$. (d) Evolution of the event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line). (e) Storage fluctuation in the $S_u$-reservoir and observed soil water content (SWC) in the top 1 m soil cover (values relative to the state at the simulation start).
4.3. Model parameterization

Figure 4.5: Simulations of the DP-fast model for the June 2013 event at the Schluecht site. See Fig. 4.4 for descriptions of the subplots.
The contribution of the upper reservoir to the bulk recharge, $q_{u\rightarrow r}$, and the outflow from the slow groundwater reservoir, $q_{g,s}$, are displayed alongside the simulated runoff, $q_{\text{tot}}$, and observed runoff, $q_{\text{obs}}$. The difference between simulated and observed discharge is presented as simple error ($q_{\text{tot}} - q_{\text{obs}}$) and percentage error ($100 \times (q_{\text{tot}} - q_{\text{obs}}) / q_{\text{obs}}$) in subfigures (c), and observed and simulated event runoff coefficients $C_{R,e}$ (Eq. (2.3)) are presented in subfigures (d). For the Schluecht site, also the storage variations in the upper reservoir ($S_u$) could be compared to the observed soil water content (SWC) in the top 1 m soil cover (subfigures (e) of Figs. 4.4 and 4.5).

**Discussion of the DP-fast model simulations of the Schluecht site**

The DP-fast model correctly simulated the limited response to the first rainfall in the October 2012 event and shows adequate delay of the main discharge peak. The $q_u$ flux was much smaller than the rainfall input rate, but strong enough to cause the quick depletion of SWC that corresponded to the observations. As expected, the slow groundwater reservoir outflow increased only slightly and was already responsible for the flow recession within two days after the storms, whereas the timing and magnitudes of the discharge peak were determined by the fast groundwater component $q_{g,f}$ (flux not shown in the graph). These behaviors match with our understanding of the processes in the hillslope discussed in Section 2.5.1. Storm runoff volume was overestimated, as is well reflected in the $C_{R,e}$ curves, but the absolute error did not exceed 0.3 mm h$^{-1}$.

The peak magnitudes are captured well, but the shape of the hydrograph could not be represented by the model. The underestimation of the first peak of the June 2013 hydrograph may be attributed to incorrect model inputs; part of the precipitation until the flow peak fell as snow (Sect. (2.3.6)), but the model does not account for snowfall and snowmelt. Making the fast groundwater reservoir more responsive would reduce the overestimation during the period with low intensity rainfall between the first peak and the second peak of the June 2013 event, as well as the overestimated flow immediately after the second peak, but this would also cause stronger overestimation of the second peak. The slow groundwater component was not further adjusted to improve the simulation of the flow recession because it has little effect on the flood runoff peak. Instead, the responsible parameter, $K_{g,s}$, was set equal to its counterpart in the DP-slow model (see below) to maintain an internally consistent parameterization.

**Discussion of the DP-slow model simulations of the Gadenstetten site**

The DP-slow model could simulate the two events at the Gadenstetten site well (Figs. 4.6 and 4.7). The outflow from the slow groundwater reservoir determines the timing and magnitude of the discharge peak as it has a much slower response than the upper zone reservoir that feeds into it. This implies that choosing the reservoir constant $K_{g,s}$ poses a trade-off between predicting the rising and receding limbs of the hydrograph. This structural problem, also prevalent in the DP-fast model, sets limitations to how well flow recession can be simulated with the current approach. Because the rising limb is more important for simulating flood formation, I did not try to improve the simulation of flow recession. Besides, after the storm, the hydrograph is typically almost horizontal with maximum flow occurring only after several days.
This extremely flat rising limb may not be well described with a combination of linear reservoirs. Addressing these issues was beyond the scope of the research presented here, but some possible ways forward are discussed in Section 4.6.

The underestimation of the flow in the last two days of the June 2013 event may be attributed to snowmelt in the higher parts of the catchment as temperatures rose markedly those days. These snow-related problems limit the possibilities of evaluating and improving the model for this event.

### 4.3.3 Adapting the ‘strongly reacting’ QArea\(^+\) models to the corresponding QArea-C models

The HOF1, SOF1, SOF2, SSF1, and SSF2 models were adapted to the behaviors of the different runoff types (RT) of the QArea-C model, which are ultimately based on plot-scale sprinkling experiments and allowed good simulation skill in a wide variety of catchments (Sect. 4.1.2). Table 4.2 lists which runoff response curves and parameter values of the runoff types that were used for the different DRP classes. Figure 4.1 shows the development of \(C_{R,d}\) for the different curves together with the fraction of cumulative rainfall diverted to the fast reservoir (\(S_{ob}\)). This fraction is similar to the event runoff coefficient \(C_{R,e}\) (Eq. (2.3)), but does not depend on the temporal development of the rainfall input. It indicates roughly how much of the rainfall contributes to flood runoff formation.

Two different synthetic storms were used for the adaptation process: a high intensity 4-hour storm with 60 mm h\(^{-1}\) intensity, and a moderately intense storm of 36 hours of 10 mm h\(^{-1}\) intensity. The high intensity storm corresponds to the typical rainfall rates applied in the sprinkling experiments of Scherrer (1996), whereas the moderate intensity storm represents typical conditions during long-duration events. Proper adaption to these different events was expected to result in a useful transfer of the model behavior for the types of storms the QArea-C model is normally used for.

Both models, QArea-C and QArea\(^+\), were applied without potential evaporation input and with equal values for the delay reservoir constants, i.e.: \(K_{ob} = K_d\). Tests with different \(K_{ob}\) values showed that the parameterization of the adapted QArea\(^+\) model does not depend much on the choice of this delay parameter as long as it is within the range used by Scherrer AG (\(K_{ob}\) between 1 and 5 h).

The QArea\(^+\) models could be adapted well to the QArea-C models (e.g., Nash-Sutcliffe efficiencies between 0.93 and 0.99), while maintaining an internally consistent parameterization (see the parameterization in Table 4.3 presented later). For example, the SSF models perform well using the \(K_o\) value that allowed good simulation with the DP-fast model at the Schluecht slope.

<table>
<thead>
<tr>
<th>QArea-C setup</th>
<th>HOF1</th>
<th>SOF1</th>
<th>SSF1</th>
<th>SOF2</th>
<th>SSF2</th>
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<td>1</td>
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<td>2</td>
<td>3</td>
<td>3</td>
</tr>
<tr>
<td>(K_{ob}) (h)</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
<td>1</td>
</tr>
<tr>
<td>(K_s) (h)</td>
<td>20</td>
<td>40</td>
<td>40</td>
<td>80</td>
<td>80</td>
</tr>
</tbody>
</table>

**Table 4.2:** Settings of the QArea-C model used to adapt the different QArea\(^+\) models to.
Figure 4.6: Simulations of the DP-slow model for the October 2012 event at the Gadenstetten site. (a) Observed rainfall and cumulative rainfall used as model input. (b) Observed and simulated runoff, $q_{\text{obs}}$ and $q_{\text{tot}}$, contribution of $q_{\text{u}}$ to the bulk recharge, $q_{\text{u} \rightarrow \text{r}}$, and outflow from the slow groundwater reservoir $q_{\text{g,s}}$. (c) Model error ($q_{\text{tot}} - q_{\text{obs}}$) and model deviation as percentage of $q_{\text{obs}}$. (d) Evolution of the event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line).
Figure 4.7: Simulations of the DP-slow model for the June 2013 event at the Gadenstetten site. See Fig. 4.6 for descriptions of the subplots.
(see Sect. 4.3.2). Doubling this $K_u$ value for SOF processes also gave good fits for these models, while also giving more weight to the contribution of the fast responding flux $q_{f,SOF}$. The $K_u$ value has little effect on the discharge response of the HOF and DP-slow models and the same value as for the SSF and DP-fast models was used; in accordance with the perceptual model, the outflow of the HOF model is dominated by $q_f$ and the outflow of the DP-slow model by $q_{f,SOF}$. The differences between the $q_f$ and $q_u$ contributions to the total outflow $q_{tot}$ are generally larger for lower intensity rainfall, which reflects our understanding (perceptual model) that fast processes are more important during high-intensity storms.

Likewise, the $S_{u,fc}$ and $S_{u,max}$ parameters could be kept the same for SOF and SSF types with equal process intensity (e.g., SOF1 and SSF1), while also remaining close to the $a priori$ defined proportionality between these parameters proposed in parameterization guideline 4. Also the differences between $S_{u,pf}$ and $S_{u,fc}$ of the SOF models could be set according to this guideline; they yield the minimum storage associated with the different classes, i.e., 0, 40 and 100 mm for SOF1, SOF2 and SOF3, respectively. With these storage parameters set, there was no need to further adjust the $\beta$ values, such that the values proposed in guidelines 4 and 5 were maintained. The $S_u$ reservoir could thus be coherently parameterized for the different DRP.

The obtained fits of the SOF and SSF models to the same response curves illustrates that already with few tunable parameters, even if they are set within a feasible range according to the guidelines of Section 4.3.1, the combination of model structure and its parameterization may not be identified reliably through calibration to a short period of rainfall and runoff data. It may therefore be argued that models may be more meaningfully constructed and parameterized from understanding of the physical processes, like in the presented DRP framework, than from inference techniques alone. This was recently also demonstrated by Fenicia et al. (2014) for contrasting catchments for which lots of data is available; for this reason, the authors called for a more fieldwork-based perspective to model application.

### 4.3.4 Simulating the October 2012 event at headwater and catchment scales

The flood of 10 October 2012 provides interesting possibilities for evaluating the model for the Wissenboden subcatchment and the Schaechen catchment. This event was studied in detail in Chapter 2. Besides evaluation of the flood magnitude, the event also allowed a targeted evaluation of the direct runoff contribution, the sum of $q_h$ and $q_f$, and its delay (per the delay reservoir constant $K_d$), because at this small scale the timing and ‘flashiness’ of the runoff peak is little affected by routing processes in the channel network. This was done by varying the $K_d$ value between 1 and 5 hours (see Sect. 4.3.3), and visually inspecting the timing and amplitude of the rapid fluctuations of the hydrograph. Note that the time constants of the other reservoirs are much higher and cannot produce the rapid fluctuations that are visible in the Wissenboden response.

The Wissenboden and Schaechen catchments have similar coverage of ‘strongly’ reacting areas, whereby SSF2 dominates in Wissenboden (22% of the 29% total coverage of ‘strongly’ reacting areas) and SSF1 dominates in the Schaechen (14% of the 24% total coverage of ‘strongly’ reacting areas). However, the Wissenboden catchment has much more ‘unconnected’ areas
than the Schaechen catchment, 13% versus 7%, and more ‘little contributing’ areas too: 43% versus 25%. As the fraction of areas with ‘damped’ response is also small (only 5%), the contrast between ‘strongly reacting’ and ‘little contributing’ areas is much greater in the Wissenboden catchment than in the Schaechen. This great contrast may benefit evaluation of the ‘strong’ runoff formation in the Wissenboden catchment.

On the other hand, these relatively small fractions of ‘strongly’ reacting areas mean that the Wissenboden and Schaechen catchments are well suited for evaluating the parameterization of the other model components that were thus far only adapted to small-scale observations and the three fastest runoff types of the QArea-C model. Because of the large fraction of areas affected by these components, they largely determine how much rainfall becomes runoff within hours or days, or much longer time scales, and thereby control the overall flood magnitude during long-duration storms.

Rainfall measured at the Wissenboden station was used for simulation of the Wissenboden headwater. Rainfall intensities were much higher than at the Schluecht station, which was used as input for the simulations of the Schaechen, because it was found to closely resemble catchment average precipitation (Sect. 2.4.1). The same model setup was used as for the simulations of the October 2012 event at the Schluecht and Gadenstetten slopes in Section 4.3.2 except for the initial conditions of the slow groundwater reservoir $S_{g,s}$; instead of estimating the specific discharge based on the full catchment area, only the total area of the elements that have a $S_{g,s}$ component was used:

$$S_{g,s}(t_{\text{start}}) = K_{g,s}Q_{T,\text{obs}}(t_{\text{start}})/A_{\text{contr}}$$

with $A_{\text{contr}}$ $[L^2]$ representing the total area of all model elements with a $S_{g,s}$ reservoir contributing to the discharge at the observation point, $Q_{T,\text{obs}}$. The same starting time of 29 August 2012 00:00 was chosen, but the presented analysis period starts just before the first storm on 7 October 2012.

The simulation of the runoff response of the Wissenboden headwater is presented in Figure 4.8, and the simulation of the Schaechen response is presented in Figure 4.9. The Wissenboden catchment was below the snow line during the main storm, and so was most of the Schaechen (Sect. 2.4.1). For the Schaechen, the elevation of the snow line may be assumed to follow the 0 °C line. The fractional catchment area below the 0 °C line, $f_{A,b0}$ $[-]$, is therefore indicated in Figure 4.9b. It was estimated from the DHM25 digital elevation model (Swisstopo, 2005) and the air temperature, $T_{\text{air}}$, observed at the Gross Windgaellen station (3187 m), assuming a constant lapse rate of 6.5 °C km$^{-1}$ (e.g., Isaac and Hallett, 2006). The fractional area that receives snow is likely larger, because the snow line elevation, i.e., the elevation above which precipitation falls mainly as snow, usually lies a few hundred meters below the 0 °C line.

The model could well describe the timing and magnitude of the discharge peaks in both catchments. The flood volume was also simulated well until the end of the storm, as visible in the comparison of observed and simulated event runoff coefficients ($C_{R,e}$; see Eq. 2.3); the mismatch in the Schaechen is mainly caused by the relatively large error at $t_0$ and the observed discharge getting smaller than $q(t_0)$ during the first 12 hours of the storm.

After the storm, discharge was overestimated, like at the Gadenstetten and Schluecht sites. At the Wissenboden site, however, this may not be caused by overestimated outflow from the...
Figure 4.8: QAREA+ simulations of the October 2012 event at the Wissenboden headwater. (a) Rainfall, $P$, and cumulative rainfall, $V_P$, at Wissenboden, which was used as model input. (b) Observed and simulated specific discharge, $q_{obs}$ and $q_{sim}$, with the contributions of the DRP classes specified. (c) Specific discharge of the different DRP classes; colors as per (b). (d) Observed and simulated event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line).
Figure 4.9: QAREA simulations of the October 2012 event in the Schaechen catchment. (a) Observed rainfall, $P$, and cumulative rainfall, $V_P$; the Schluecht station data was used as model input. The catchment average precipitation estimated from the gridded data product RhiresD (MeteoSwiss, 2013a) is presented as ‘MG Schaechen.’ (b) Observed and simulated specific discharge, $q_{obs}$ and $q_{sim}$, with the contributions of the DRP classes, and fractional catchment area below the 0 °C line, $f_{A,b0}$, as estimated from the air temperature, $T_{air}$, observed at the Gross Windgaellen station. (c) Specific discharge of the different DRP classes; colors as per (b). (d) Observed and simulated event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line).
DP-slow areas alone, because the drainage from the SSF2 areas could also be too strong during the first hours after the storm. This behavior was not further adjusted because it was thought to require either an improved model structure that allows quicker decline of the $q_u$ flux, or a so much stronger reacting $q_f$ flux that internal consistency of the model would be at stake.

The responses of the different DRP classes (subfigures ‘c’) and their contributions to the catchment flood hydrograph (subfigures ‘b’) are commensurate with our perceptual model. The areas with ‘strong’ response can produce high discharges, and are responsible for most of the storm discharge of the Wissenboden headwater, because the rest of the catchment is mainly of the ‘little contributing’ type. On the other hand, in the Schaechen the contribution of the ‘strongly’ reacting areas is small, because they are rare and receive a catchment average rainfall with substantially lower intensity than the local rainfall at Wissenboden. The delayed response of the DP-fast areas, i.e., the response type of the Schluecht slope, appears to be feasible, as neither the rising limb nor the peak flow was underestimated.

At the lower rainfall intensities used in the Schaechen, the SSF2 and SSF3 processes show similar response because the $q_f$ is hardly activated and thus both models have similar $q_u$. This is not desired, but was not further investigated, because we have no data to evaluate the behavior against, and the simulated behavior depends greatly on how initial conditions are defined: the assumption that both models had their $S_u$ reservoir at field capacity may not hold just after a small storm with dry antecedent conditions. This would cause the model to have similar storage deficit (in millimeters, i.e.: $S_{u,fc} - S_u$), although a higher storage deficit may be expected for the thicker soils of SSF3 areas. Besides, the difference between the model responses is substantial at higher rainfall intensities, as shown in the Wissenboden simulation, indicating that the model structure and parameterization determine the sensitivity to rainfall intensity. This is in agreement with our perceptual model, i.e., the behavior of thicker, but otherwise similar, soils may well be similar for small events, but we have no data that allows to explore above which intensity the behavior of the classes should start to diverge more.

It was found that the lower estimate of $K_d = 1$ describes the quick fluctuations at the Wissenboden headwater best. This value is therefore used for all $K_d$ used with the model. Note that the parameter is less important for moderately intense storms, as the rapid contribution occurs via both $q_u$ and $q_f$, but $K_d$ only affects the $q_f$. 
Table 4.3: Parameter values of the QArea* model for the different dominant runoff process classes. Parameters and classes are presented in Fig. 4.2 and Table 4.1.

<table>
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<th>SSF2</th>
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<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>$K_{g,f}$</td>
<td>(h)</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>160.0</td>
<td>160.0</td>
<td>160.0</td>
<td>160.0</td>
</tr>
<tr>
<td>$K_{g,s}$</td>
<td>(h)</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>160.0</td>
<td>160.0</td>
<td>160.0</td>
</tr>
<tr>
<td>$\alpha_f$</td>
<td>(–)</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>$\alpha_u$</td>
<td>(–)</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.25</td>
<td>0.25</td>
<td>1.0</td>
<td>1.0</td>
</tr>
<tr>
<td>$\alpha_r$</td>
<td>(–)</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.5</td>
<td>0.0</td>
</tr>
</tbody>
</table>

Figure 4.10: Runoff responses of the different QArea* model classes to a 4-hour storm of 60 mm h$^{-1}$ (a) and a 36-hour storm of 10 mm h$^{-1}$ (b). These storms were also used for adapting the ‘strongly’ reacting QArea* models to the corresponding QArea-C behavior in Sect. 4.3.3.

4.3.5 The resulting parameterization

The final parameterization of the QArea* model is presented in Table 4.3. Figure 4.10 presents the behavior of the different DRP for the synthetic storms used for adapting the ‘strongly’ reacting QArea* models to the corresponding QArea-C behavior in Section 4.3.3.

Many of the parameters that were set a priori from the guidelines of Section 4.3.1 could be maintained in the adjustments of the DRP models to the different data sets. This suggests that these guidelines are useful. Moreover, the guidelines resulted in a coherent parameterization with strongly contrasting behavior of the different DRP classes, such that the mapped strong contrasts in the landscape are also simulated very differently for storms with intensities of...
10 mm h$^{-1}$ or more (Fig. 4.10). This is important, because it means the differences between slopes as put forward in the perceptual model are appropriately represented in the computational model.

### 4.4 Model evaluation for more extreme events

The 2000 – 2013 period of the Schaechen flood record was screened for suitable events for further testing of the developed model. This period was used because earlier flood events had no sub-daily resolution precipitation measurements in the area, except for the Altdorf station, which lies several kilometers downstream of the Schaechen gauging station and often shows considerable deviations from the measurements at the centrally located Unterschaechen station. Only the eight floods that were larger than the October 2012 flood that was used to set up the model were considered in the screening.

Two floods were selected for testing the developed model: the floods of 6 August 2010 and 22 August 2005. The other larger floods were less suitable for evaluating the model for various reasons. For example, snow effects can make the estimation of the liquid water input highly uncertain; these effects could include a low snow line, or a rapid snowmelt due to a quick increase in incoming radiation or air temperature. Besides, rain falling on snow-covered ground may further increase the uncertainty of the liquid water input. Furthermore, the more intense storms were often found to exhibit a strongly heterogeneous rainfall distribution in both time and space, which further complicates the estimation of catchment average rainfall.

The characteristics of the presented events differ from the October 2012 flood used for parameterization of the model (Sect. 4.3). The August 2010 flood had a slightly higher peak flow than the October 2012 flood, but was caused by a comparatively short, intense storm, whereas the August 2005 flood caused by a long-duration storm and showed a remarkably strong response, ultimately becoming the largest flood ever recorded (Sect. 1.2). Rainfall intensities during both events were frequently more than 10 mm h$^{-1}$, and thus higher than during the October 2012 event. The selected events thus allow testing the developed framework under different conditions than those it was developed and parameterized for; this is a rather difficult test that model often fail (e.g., Klemeš, 1986b; Kirchner, 2006; Pilkey and Pilkey-Jarvis, 2007).

The simulations were performed with the parameters obtained in Section 4.3, using the same potential evaporation boundary condition and assumptions for setting the initial conditions as used for the October 2012 event simulation (Sect. 4.3.4).

#### 4.4.1 The flood of 6 August 2010 in the Schaechen

The main peak of the flood was caused by an 8-hour period with relatively intense rainfall. Of the stations that measure precipitation at high temporal resolution, only the Seewli and Altdorf stations showed no indications of snowfall. During the 8-hour period, these stations showed similar maximum rainfall rates and temporal developments. This suggests a relatively homogeneous spatial distribution of precipitation in the catchment. The 0 ºC line dropped substantially during these hours (Fig. 4.11b), but as the unheated tipping bucket gauge at Seewli registered precipitation, it appears the snowline remained above 2000 m.s.l. This means that roughly the
upper 25% of the catchment did not receive liquid precipitation, out of which about 3% was mapped as areas with ‘very fast’ response; roughly 20% of all the ‘very fast’ reacting areas in the catchment (Fig. 3.5).

Wet antecedent conditions may have contributed to the relatively high magnitude of the flood, as baseflow was relatively high (discharge at $t_0$ was at the 85th percentile August mean flow), and the responsible rainfall (at 45 to 65 mm) belongs to the smallest of the storms that have caused a yearly maximum flood. This rainfall sum is also smaller than that of the October 2012 event, but fell in a much shorter time. The wet antecedent conditions were caused by a roughly 2-year flood of July 30th, and following snowmelt and some local intense storms. The simulation was started about 24 hours after this storm; the local storms occurred during the first two days of the simulation (Fig. 4.11a). Also, probably much of the fresh snowfall of the preceding event melted in this period as the air temperature rose markedly and the 0°C line remained well above 3500 m.s.l.

The rainfall measured at Altdorf was used as catchment average rainfall input for the simulation because its temporal development corresponds best with the discharge dynamics. The Seewli data were not suitable for this purpose as this station continued to record liquid precipitation for many hours after the discharge peak, and thereby indicated a substantially larger precipitation sum than was recorded at the Unterschaechen station (Fig. 4.11b). The Altdorf rainfall sum, however, is only 10% lower than measured at Unterschaechen (24-hour interval). During the first 8 hours of the storm that caused the flood peak, the rainfall intensity measured at Altdorf was also about 10% lower than observed at Seewli. This moderate underestimation may well be offset by the considerable fraction of the catchment receiving liquid precipitation, such that the rainfall observed at Altdorf may adequately represent the catchment average rainfall input.

The overall flood behavior was simulated well. The observed flood peak is underestimated by 0.25 mm h$^{-1}$, and then followed by an overestimated simulated peak due to a short period of more intense rainfall. As for the October 2012 event (Sect. 4.3.4), the simulated decline of the receding limb is too strong. However, here it may also be attributed to the melt of fresh snow causing a relatively flat recession. The overestimated small peaks of August 1st and 2nd are caused by the erroneous application of the local storm at Altdorf to the rest of the catchment.

The responses and contribution of the different DRP are in line with the DRP map. Due to the relatively high rainfall rates and moderate rainfall depth, the small fraction of ‘very fast’ reacting areas is responsible for much of the flood response, whereas the DP type of responses are too delayed to contribute much to the flood peak formation. Such damped response was also observed at the large spring of the Schluecht site, which showed a peak discharge that was more than a factor 2 smaller than the peak discharge of the October 2012 event (data not presented here).

As in the simulation of the October 2012 event discussed in Section 4.3.4, the response of the SSF2 and SSF3 types may be too similar. Moreover, both SSF types respond more strongly than the SOF2 model, because the latter was hardly activated. These problems, however, have little effect on the overall representation, which still reflects our perceptual understanding well; i.e.: there are areas with a ‘strong’ reaction and areas with a (strongly) ‘damped’ response and some areas that are a bit in between.
Development and testing of the QA/rSc/eSc/aSc model in the Schaechen catchment

Figure 4.11: QA/AREA simulations of the August 2010 event in the Schaechen catchment. (a) Observed rainfall, $P$, and cumulative rainfall, $V_P$; the Altdorf station data was used as model input. The catchment average precipitation estimated from the gridded data product RhiresD (MeteoSwiss, 2013a) is presented as ‘MG Schaechen.’ (b) Observed and simulated specific discharge, $q_{obs}$ and $q_{sim}$, with the contributions of the DRP classes, and the fractional catchment area below the 0 °C line, $f_{A,b0}$, as estimated from the air temperature, $T_{air}$, measured at the Aelpler Tor station. (c) Specific discharge of the different DRP classes; colors as per (b). (d) Observed and simulated event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line).
Figure 4.12: QAREA simulations of the August 2005 event in the Schaechen catchment. (a) Observed rainfall, $P$, and cumulative rainfall, $V_P$; the Seewli station data was used as model input. The other subplots are the same as in Fig. 4.11 but have different axes scales.
4.4.2 The flood of 22 August 2005 in the Schaechen

The August 2005 flood was caused by a roughly 18-hour period with more than 100 mm of rainfall. Locally, rainfall may have been a bit less, like at the Aelpler Tor and Altdorf stations, but considerably more rainfall was recorded at Seewli and Unterschaechen (188 mm in 18 hours and 184 mm in 24 hours, respectively; Fig. 4.12). Antecedent conditions were moist, with 135 mm and 172 mm of precipitation at the Altdorf and Unterschaechen stations in the period 1 to 17 August, corresponding to the mean precipitation measured in August at both sites. Another 32–98 mm (Aelpler Tor – Altdorf) fell in the three days leading up to the main storm. The 0 °C line was always above 2800 m, mostly even above 3000 m, such that snowfall and snowmelt played only minor roles.

The temporal rainfall pattern observed at the Seewli station corresponds well with the hydrograph dynamics and the daily observations at Unterschaechen, unlike the patterns observed at Aelpler Tor and Altdorf (Fig. 4.12). Therefore, the Seewli station was taken as homogeneous input for the whole catchment. Locally, this may yield strong overestimation of the rainfall inputs, and it may be seen as an upper-bound estimate of catchment average precipitation. A reduction of 20 % of the Seewli rainfall may be seen as the lower bound of catchment average rainfall, because a reduction of 23 % matches the 24-hourly RhiresD estimates, which were shown to underestimate the heavy precipitation of the August 2005 event by 5 to 40 % (Frei et al., 2008).

With the flood peak more than twice as large as the October 2012 flood used to set up the model structure and parameters, the August 2005 flood presents a challenging test. The overestimated peak flow of 4.73 mm h⁻¹ (Fig. 4.12b) is within 10 % of the upper estimate of the flood peak (i.e., Scherrer AG, 2007 estimated the flood peak at between 120 and 130 m³ s⁻¹; 4–4.33 mm h⁻¹). This is a relatively accurate prediction for such an extreme event. Moreover, simulations based on reduction of the Seewli data by 10 to 20 % also had peak flows close to this range (not shown here). Besides, the qualitative understanding that large parts of the catchment start to contribute later in the event is adequately represented, with clearly different response of the various DRP classes (Fig. 4.12bc). This leads to a good simulation of the total volume discharged until the flood peak (Fig. 4.12d).

The event runoff coefficient, $C_{R,e}$, does not reach 50 % within the presented three days after the storm, indicating that large volumes of water are stored beyond event time scales. This also holds if rainfall is reduced by 20 % or if the accounting starts on August 18th. Such a large storage term is mainly attributed to the ‘damped’ and ‘little contributing’ areas of the DRP map, and could also be observed at our sites (Sect. 2.5.2). The roughly 25 % of the catchment that is mapped to ‘contribute little’ may thus be well realistic; to arrive at the low event runoff coefficient of the Schaechen, the other ~75 % of the catchment must on average store similarly large portions of the incoming rainfall as we observed at the Schluecht slope (Sect. 2.5.1).
4.5 Discussions

4.5.1 Could threshold-like responses explain the remarkably strong response of the Schaechen during the most extreme events?

It seems plausible that the four largest floods in the Schaechen, which are considerably larger than the rest —the threshold-like behavior discussed in Section 1.2—, are caused by a large fraction of the catchment contributing more than during smaller floods. These are the SSF3 and DP-fast areas with ‘damped’ response covering roughly 45% of the catchment (Sect. 3.7). They are responsible for a smaller portion of the peak runoff increase during the short August 2010 event, than during the longer-duration October 2012 and August 2005 events. With the magnitude of the peak flows estimated well, it appears the responses of the delayed reacting DRP are quantified satisfactorily by the current parameterization.

4.5.2 Further developments

Extending the model with a snow routine may aid further application, because adequate simulation of the different effect of snowfall and --melt may be required for events with relatively low snow lines. A snow routine should include a useful spatial distribution of the meteorological forcing data for computing snowfall and snowmelt rates throughout the landscape. Simulation of high melt rates during rain-on-snow events would be of particular interest, but testing of prediction skill for such events requires reliable observations of snowfall and snowmelt over time, which are rarely available. Development of a sound snow routine is an outstanding research challenge in itself and was not attempted here. The benefit of including snow processes also extends to simulation of rainfall-runoff process on a seasonal basis, which would allow further testing of long-term storage and drainage behavior.

Further research into the long-term drainage behavior of different landforms may also improve simulation of the recession just after the storm and thus prediction of responses to long-duration storms. Besides, it may allow for better estimation of antecedent flow conditions. It could be argued that for such longer-term processes some adaptations to the mapping scheme and model representation are needed. For example, topographic gradient and hillslope element spatial extent or landscape concavity (in how far different hillslopes drain to the same point) may influence the long-term drainage. Also, subsurface flows between all elements along a catena may become more important at longer time-scales. Such per-element properties may require little additional GIS work and could be incorporated in the model as ‘scaling factors’ of the relations that are currently used; e.g., a $K$ value could be scaled by hillslope area or some quantile of slope length. This would require further research into how these scaling relations work, and how important such hillslope characteristics are in comparison with other flow-determining properties that are not so easily estimated, such as the permeabilities of bedrock or sediment cover.
4.5.3 Further testing of the methodology

The good prediction skill for the different floods suggests that the developed framework may be useful for scenario analyses of even more extreme events. However, further testing would be useful to increase confidence in the methods and identify potential under- and overestimation biases. It proved difficult to find suitable events for such testing in the Schaechen catchment, and it is therefore suggested to do this by application in other catchments. Preferably, the other catchments have strongly contrasting flood behavior to allow testing how well the DRP mapping technique can describe these contrasts. Also, selected model components may be better tested if the different behavior is caused by a different partitioning of DRP classes; a model parameterization is more sensitive if applied to a larger part of a catchment and is thus better testable. This kind of testing is presented in the next chapter.

4.6 Conclusions

A new rainfall-runoff model, called QAREA\textsuperscript{+}, was developed to work with dominant runoff process (DRP) maps. It is an improvement of the earlier QAREA models developed for simulation with DRP maps. Particularly new are the process descriptions of delayed reacting Deep Percolation (DP) classes and the simple model structures for representing faster processes like Horton Overland Flow (HOF), Saturation Overland Flow (SOF), and Subsurface Stormflow (SSF). The simple structures are derived from a parsimonious master structure, whose individual components use established concepts for runoff generation. Sharing these components between the different dominant runoff processes allows sharing parameter values, where thought appropriate, and defining qualitative and quantitative relations for expected differences in runoff response. These relational constraints facilitate the derivation of an internally consistent parameterization that is in agreement with the perceptual model as represented by the DRP classification scheme. By focusing on the dominant mechanisms, the DRP classification thus allows formulating simple model structures, with clearly identifiable parameters.

A coherent parameterization was obtained from calibrating the DP-type models to observations at the Schluecht and Gadenstetten slopes and calibrating faster responding DRP models to responses of the QAREA-C model. Evaluations of a small flood at the Wissenboden and Schaechen catchments indicated that the QAREA\textsuperscript{+} model captures delays well and predicts flood magnitudes with good skill. Two other extreme floods in the Schaechen were simulated satisfactorily with the obtained parameterization, indicating that the large fractions of rainfall stored in many of the slopes are estimated well.

Moreover, these results were obtained with satisfying qualitative descriptions of the differences between the DRP classes and mostly adequate quantitative descriptions of the different response strengths; e.g., a SSF1 process reacts more strongly than a SSF2 process and this is simulated with substantially different behavior (Fig. 4.10). Only the differences between SSF2 and SSF3 type response simulations are not so large for the moderately intense storms evaluated here, but I had no suitable data to further investigate this. In spite of these and other uncertainties in the descriptions of the quantitative behavior of the different DRP, the model simulations do support the hypothesis that the four peculiarly large floods of the Schaechen are caused by a threshold-like intensification of the runoff response, as suggested in Section 1.2.
4.6. Conclusions

The delays and large storage capacities expected for large parts of the catchment cause these areas to contribute little to the flood formation during short events, but substantially during longer ones.

The adequate qualitative and quantitative descriptions of the flood formation indicate that the framework allows extrapolation of field-scale knowledge to the catchment scale in a useful and meaningful way. The approach may be particularly useful in catchments where detailed data are not available but an estimate of the hydrological behavior is needed, which is commonly the case in meso-scale catchments.
Chapter 5

Evaluation of the DRP mapping and modeling framework in contrasting catchments

5.1 Introduction

This chapter discusses the application of the developed DRP mapping / modeling framework in two catchments with contrasting flood behavior and landscapes, namely the Hinterrhein catchment, with a strong and flashy flood response, and the Dischma catchment, with a strongly damped flood response. Their landscapes and flood behaviors differ considerably from the Schaechen catchment. The selected catchments thus allow evaluating the mapping technique in areas that are dominated by different types of landforms as well as testing the transferability of the QAREA* model concept to areas with different storm and flood characteristics without making further catchment-specific adjustments to its parameterization. Testing the transferability is here of particular interest because the differences in flood behavior of the catchments may be mainly explained from the differences in storm characteristics and coverage of thick sediment deposits, as most other catchment properties are similar.

The application in these two catchments also serves to illustrate how the DRP framework allows assessing the sensitivities of the catchment flood response to different storm characteristics, for example, the sensitivity to storm intensity, depth or duration, and the elevation of the event snow line. Furthermore, I try to address the possibility that a threshold-like change in flood response like in the Schaechen (Sects. 1.2 and 4.5.1) may occur; i.e., that the most extreme floods may be much stronger than statistical analyses of medium floods would indicate.

Simulations are presented for one low- and one high-return-period flood (~4–5- and ~30–40-year events, respectively) for each catchment. These events were selected because of their limited observational and input data uncertainty, relatively homogeneous rainfall distribution and minor snowfall and snowmelt effects. The largest flood runoff increase in the Hinterrhein is about an order of magnitude larger than in the ~5-year event selected for the Dischma catchment. The differences in flood magnitude between these events and the events simulated for the Schaechen catchment may not be fully explained by differences in rainfall depth or intensity, such that the selected events allow proper evaluation of the representation of differences in runoff generation mechanisms by the DRP framework.
To further evaluate the power of the transferability test and set a benchmark against which the prediction skill of the QAArea model can be compared, the same events were simulated with two calibrated lumped rainfall-runoff models. This was also done for two events in the Schaechen catchment: the ~2-year flood of October 2012 and the extreme flood of August 2005. The selected lumped models are a simple nonlinear reservoir model with two free parameters and an eight parameter model consisting of three reservoirs that allows a nonlinear partitioning of rainfall into fast, intermediate and slow runoff components. These lumped models may be seen as representative of the lumped ESMA-type models that are widely used in hydrological practice. The lumped models are used to predict the extreme floods after calibration to the low-return-period flood in the respective catchment, and vice versa. This allows assessing how difficult it is for a model to correctly predict the floods in the studied catchments and demonstrate the added value of the extra effort involved in application of the DRP framework.

These kinds of tests are difficult to pass for a prediction model, and appear rarely reported in the scientific literature. Passing the tests thus may increase confidence in the model’s capability to predict extreme floods in alpine catchments.

5.2 Catchment characteristics

The Dischma and Hinterrhein catchments were selected because they have similar properties but strongly contrasting flood behavior, as shown in Table 5.1. Both catchments have an elongated shape, have relatively natural alpine terrain with little forest cover, and are underlain by similar metamorphic rock geology (predominantly gneiss). They also have similar catchment area, elevation range and topographic gradients, and, like the Schaechen catchment, have only a small fraction of glaciated area. Each catchment is briefly characterized, and then some important differences between the catchments and the Schaechen catchment are discussed.

5.2.1 Dischma catchment

The Dischma catchment, with its outlet at the FOEN station Kriegsmatte, has been the subject of hydrological research for decades; from snowmelt and tracer studies (e.g., Martinec et al., 1982 reporting that the catchment has an unexpectedly large groundwater storage component) to research on boundary layer meteorology (e.g., de Jong, 2005 demonstrating that condensation constitutes a considerable portion of the yearly water balance), and from reports on comparisons of snowmelt models (e.g., Zappa et al., 2003) to establishing newly developed rainfall-runoff models (e.g., Schaeffli et al., 2014). Nevertheless, research on flood runoff formation in the catchment, which, as will be discussed, has remarkably small floods, appears limited to the catchment being a case-study in Kölla (1987), who reported a new technique for incorporating more spatially distributed physical process knowledge into the rational formula for Swiss environments.

It appears that none of the studies reporting spatially distributed modeling of hydrological processes recognize the occurrence of thick sediment deposits, although land use (Swisstopo, 2007 and geomorphological maps (e.g., Bearth et al., 1935; Vögele, 1984) clearly indicate that thick packets of sediment have been deposited by glacial, gravitational, and fluvial processes in
5.2. Catchment characteristics

Figure 5.1: Dischma catchment with discharge and meteorological stations indicated, and terrain shown by Alti3D hillshade background (© 2015 Swisstopo (JD100042); Swisstopo, 2013). Meteorological stations without precipitation measurements are marked with a center dot; further details about the meteorological stations are presented in Table C.3. The Weissfluhjoch and Puelschezza stations are just outside the presented map, located 3.7 km NW of the Davos station and 1.8 km ESE of the Sarsura Pitschen station, respectively.
large parts of the catchment. These maps also indicate that bare rock faces are rare, except for the highest regions. The soils in the catchment are mostly cambisols and podsols that are typically shallower than 50 cm (Krause and Peyer, 1986), but some peat soils found in the swampy areas are more than a meter thick (Vögele, 1984). The few small settlements located in the main valley floor and the extensive use of alpine meadows for grazing livestock are the principal anthropogenic influences. The meteorological stations in the area are presented in the overview map of Figure 5.1 with station properties listed in Table C.3 of the appendix.

### 5.2.2 Hinterhein catchment

The hydrology of the Hinterhein catchment above the FOEN station Hinterhein is little researched, but a map of runoff response types obtained with an earlier procedure for mapping dominant runoff processes was presented in Naef et al. (1999). There are only a few meteo-
### Table 5.1: Landscape and hydrological characteristics of the Schaechen, Hinterrhein and Dischma catchments.

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Schaechen</th>
<th>Hinterrhein</th>
<th>Dischma</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Area (km²)</strong></td>
<td>108.3</td>
<td>54.2</td>
<td>43.3</td>
</tr>
<tr>
<td><strong>Minimum elevation (m.s.l.)</strong></td>
<td>490</td>
<td>1584</td>
<td>1668</td>
</tr>
<tr>
<td><strong>Maximum elevation (m.s.l.)</strong></td>
<td>3295</td>
<td>3402</td>
<td>3146</td>
</tr>
<tr>
<td><strong>Average elevation (m.s.l.)</strong></td>
<td>1717</td>
<td>2356</td>
<td>2371</td>
</tr>
<tr>
<td><strong>Topographic gradient (%)</strong>&lt;sup&gt;a&lt;/sup&gt;</td>
<td>63.0 (0.74)</td>
<td>58.6 (0.66)</td>
<td>53.8 (0.49)</td>
</tr>
<tr>
<td><strong>Dominant geology</strong></td>
<td>Sedimentary rock (sandstone, shale, flysch, marl, limestone)</td>
<td>Metamorphic rock (para-/ortho-gneisses, (Bündner) schist)</td>
<td>Metamorphic rock (para-/ortho-gneisses, amphibolite)</td>
</tr>
<tr>
<td><strong>Glacierization (%)</strong>&lt;sup&gt;b&lt;/sup&gt;</td>
<td>5.9 (2.3)</td>
<td>13.3 (9.0)</td>
<td>1.6 (1.1)</td>
</tr>
<tr>
<td><strong>Annual precipitation (mm y⁻¹)</strong>&lt;sup&gt;c&lt;/sup&gt;</td>
<td>1832</td>
<td>1714</td>
<td>1015</td>
</tr>
<tr>
<td><strong>100-year 24 h precipitation (mm d⁻¹)</strong>&lt;sup&gt;d&lt;/sup&gt;</td>
<td>130–180</td>
<td>~190</td>
<td>~90</td>
</tr>
<tr>
<td><strong>Var. year. max. 24 h precipitation (–)</strong>&lt;sup&gt;e&lt;/sup&gt;</td>
<td>2.03</td>
<td>1.83</td>
<td>1.86</td>
</tr>
<tr>
<td><strong>Var. year. max. 48 h precipitation (–)</strong>&lt;sup&gt;e&lt;/sup&gt;</td>
<td>2.23</td>
<td>1.76</td>
<td>1.99</td>
</tr>
<tr>
<td><strong>Flood record length (years)</strong></td>
<td>84</td>
<td>66</td>
<td>50</td>
</tr>
<tr>
<td><strong>Largest flood (mm h⁻¹)</strong>&lt;sup&gt;f&lt;/sup&gt;</td>
<td>4.16</td>
<td>11.29</td>
<td>1.59</td>
</tr>
<tr>
<td><strong>Annual flood (mm h⁻¹)</strong>&lt;sup&gt;f&lt;/sup&gt;</td>
<td>1.29 (0.50)</td>
<td>4.67 (0.46)</td>
<td>0.95 (0.27)</td>
</tr>
</tbody>
</table>

<sup>a</sup> Catchment average gradient as computed with procedures outlined by Tarboton (1997) using a 25 m resolution digital elevation model (Swisstopo, 2005); the coefficient of variation (i.e., standard deviation divided by the mean) is listed in brackets.

<sup>b</sup> Catchment area of the glaciers and snowfields indicated in the Vector25 product (Swisstopo, 2007), total areas of which is indicated in brackets; catchment area estimates follow from the landscape element delineation of the DRP mapping tool (Sect. 3.5).

<sup>c</sup> Catchment areal average estimated from RhiresM product (MeteoSwiss, 2013b) for the period Oct. 1985 to Sept. 2008.

<sup>d</sup> Return periods estimated per station and spatially interpolated (Jensen et al., 1997); only the Schaechen catchment has a considerable range in return values.

<sup>e</sup> Variability of the yearly maximum 24 h or 48 h precipitation depth, expressed as the ratio of the 100 y over 2 y return levels as estimated from a GEV distribution fitted to the precipitation records of the Altdorf, San Bernardino and Davos stations as reported by MeteoSwiss (2015), for the Schaechen, Hinterrhein and Dischma catchments, respectively.

<sup>f</sup> Mean annual maximum flood of the full observation records with coefficients of variation listed in brackets (details in Appendix A).

Logical stations in the region (Fig. 5.2, Table C.2). The catchment has a larger glaciated area than the Dischma catchment, most of which is found in the north-facing cirque valleys. Many of these glaciers have been rapidly retreating since the little ice age (~1850 AD), when the tongue of the Paradiesgletscher in the uppermost part of the catchment reached until 2250 m.s.l. (Burga, 1981), roughly 350 m below its current position. Comparison of the 1850 AD glacial extents presented in the geomorphological maps of Vögele (1984) and Burga (1981) indicates that a considerably larger part of the Hinterrhein catchment was recently deglaciated than in the Dischma. This may be explained by the higher annual precipitation and the larger fraction of north-facing slopes in the Hinterrhein. The effects of this recent deglaciation are reflected in the more barren landscape, which includes many areas with bare rock that are separated by areas with little soil formation, for example around the San Bernadino Pass area (i.e., the area between the San Bernardino and Hinterrhein meteorological stations; Fig. 5.2). This area also exhibits important earlier glacial landscape formation effects, for example the markedly erratic channel network that is presumably rooted in glacial striation patterns caused by the southward flowing branch of the Hinterrhein glacier (until at latest 13 000 BP; Burga, 1981).
There are no detailed soil maps of the area, and the available geo(morpho)logical maps of Frischknecht et al. (1923) and Burga (1981) provide limited geomorphological information. The stronger glaciation effects in the Hinterrhein suggest that the soils are typically shallower and less developed than in the Dischma catchment. There are a few thick sediment deposits, most notably the alluvial plains and debris cones in the main valley floor and the small cirques on the south-facing flanks. Anthropogenic influences appear confined to the San Bernardino Pass area, the main valley floor up to the military training ground, and the easternmost tributary catchment on the northern flank (Fig. 5.2). This tributary catchment is markedly different from the rest of the Hinterrhein catchment; it has Bündner schist geology (German: Bünderschiefer), and many slopes have creeping landmasses.

The overview maps of the Dischma and Hinterrhein catchments (Figs. 5.1 and 5.2) also present the channel network represented in the Vector25 landscape model. It is not clear how well they represent the perennial channel network, but there is a strong contrast in channel network density of the catchments, which corresponds well with the more heavily incised terrain of the Hinterrhein catchment, as visible in the shaded relief maps that are drawn in the backgrounds of the maps.

5.2.3 Comparison of the Schaechen, Dischma and Hinterrhein catchments

The above analysis indicates that the main difference between the catchments lies in the occurrence of landforms with large storage capacity. This appears to be reflected in the contrasting flood behaviors of the catchments, which are also different from the Schaechen catchment. As measured by peak specific discharge magnitudes, the Hinterrhein catchment has a much stronger flood response than the Schaechen catchment, the largest flood in the former (Fig. 5.3a) being more than 2.5 times larger than in the latter (Fig. 1.1a, p. 3), and more than 7 times larger than in the Dischma catchment (Fig. 5.4a). The corresponding maximum event discharge increases of the Dischma and Hinterrhein catchments (i.e., peak discharge minus pre-event discharge, a more suitable measure of the response strength), even differ by an order of magnitude.

The largest floods of the Hinterrhein catchment appear to be most adequately predicted from flood frequency analysis (Fig. 5.3b). The technique even works when the largest floods are taken out of the analysis, but only if it is based on the period after the two largest floods, which also contained the third and fourth largest floods on record and has a higher mean annual flood (Appendix A).

The estimation of the largest floods in the Dischma catchment is less reliable (Fig. 5.4b). As was found for the Schaechen catchment (Sect. 2), the flood frequency curve of the Dischma exhibits a step change at a return period of about five years. This step change is however not as profound as in the Schaechen, where the associated underestimation of the most extreme floods is more dramatic and the fitted GEV distribution has no upper bound (shape parameter larger than zero). The shape parameter of the GEV distribution fitted to the Dischma flood record still indicates an upper bound, which is more physically realistic (Appendix A).

In all three catchments, the flood-producing storms occurred mainly in summer and early
5.3 Methods

5.3.1 Mapping of dominant runoff processes

Dominant runoff processes in the Hinterrhein and Dischma catchments were mapped according to the classification schemes presented in Sect. 3.6. Interpretation of the generally available spatial data (see Sect. 3.3 for overview) was aided by the complementary maps and literature discussed below.

For the Hinterrhein catchment, the springs indicated in the cantonal water protection map (Kanton Graubünden, 2012) were useful for characterizing the creeping landmass slopes in the northeastern part of the catchment. The geological map of Frischknecht et al. (1923) and the geomorphological maps in Burga (1981) were used to characterize and delineate the landforms.

For the Dischma catchment, the delineation of the landforms was based on the geomorphological map of Vögele (1984) and the geological characterizations in Leupold et al. (1935) and
Figure 5.3: (a) Yearly maximum floods in the Hinterrhein catchment, specified per three-month season. (b) Flood frequency plot with GEV distribution fitted to the full record and the period since the two largest floods on record occurred (1989–2010); the fitted distributions and 95% confidence intervals (CIs) correspond well. Details are discussed in Appendix A.

Figure 5.4: (a) Yearly maximum floods in the Dischma catchment, specified per three-month season. (b) Flood frequency plot with GEV distribution fitted to the full record and the period between the two largest floods on record (1976–2004); the fitted distribution to the shorter period does not adequately predict the largest floods and has a remarkably narrow 95% confidence interval (CI). Details are discussed in Appendix A.
Table 5.2: Characteristics of the analyzed events. Discharge increase $\Delta q_{\text{max}}$ is the difference in discharge between the start of the event, $t_0$, and the moment of peak flow, $t_p$. The event runoff coefficient, $C_{R,e}$, is also computed over this period. The rainfall sum until the moment of peak flow, $V_p(t_p)$, is computed from the beginning of the simulation, as indicated in the figures that present the simulation results.

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Event</th>
<th>Simulation period</th>
<th>$\Delta q_{\text{max}}$</th>
<th>$C_{R,e}(t_p)$</th>
<th>Return period</th>
<th>$V_p(t_p)$</th>
<th>Pre-event flow</th>
<th>Fig.</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>start end</td>
<td>(mm h$^{-1}$)</td>
<td>(-)$^a$</td>
<td>(year)$^b$</td>
<td>(mm)</td>
<td>(perc. $q_{\text{mon}}$)$^c$</td>
<td></td>
</tr>
<tr>
<td>Hinterrhein</td>
<td>Aug 2004</td>
<td>16 Aug 24 Aug</td>
<td>5.7</td>
<td>0.47</td>
<td>4 (3; 7)</td>
<td>158.5</td>
<td>4</td>
<td>5.10</td>
</tr>
<tr>
<td>Hinterrhein</td>
<td>Aug 1988</td>
<td>18 Aug 23 Aug</td>
<td>9.5</td>
<td>0.24+</td>
<td>41 (18; 946)</td>
<td>160.3</td>
<td>48</td>
<td>5.11</td>
</tr>
<tr>
<td>Dischma$^d$</td>
<td>Jul 2008</td>
<td>4 July 17 July</td>
<td>0.8</td>
<td>0.24−</td>
<td>5 (3; 9)</td>
<td>103.5</td>
<td>52</td>
<td>5.12</td>
</tr>
<tr>
<td>Dischma</td>
<td>Aug 2005</td>
<td>18 Aug 29 Aug</td>
<td>1.2</td>
<td>0.12−</td>
<td>30 (14; &gt; 1000)</td>
<td>106.8</td>
<td>83</td>
<td>5.13</td>
</tr>
<tr>
<td>Schaechen</td>
<td>Oct 2012</td>
<td>29 Sept 15 Oct</td>
<td>1.0</td>
<td>0.13</td>
<td>2.3 (1.9; 2.9)</td>
<td>87.9</td>
<td>48</td>
<td>4.9</td>
</tr>
<tr>
<td>Schaechen</td>
<td>Aug 2005</td>
<td>17 Aug 29 Aug</td>
<td>3.7</td>
<td>0.21+</td>
<td>185 (49; &gt; 1000)</td>
<td>213.4</td>
<td>87</td>
<td>4.12</td>
</tr>
</tbody>
</table>

$^a$ observed event runoff coefficient, $C_{R,e}$, as estimated from the rainfall data used for the presented simulations at the time of peak flow. A ‘+’ symbol indicates that the catchment-averaged rainfall is likely overestimated and that the actual value of $C_{R,e}$ is higher. A ‘−’ symbol indicates the opposite.

$^b$ values of the confidence interval bounds in brackets (methodology detailed in Appendix A).

$^c$ pre-event flow at $q(t_0)$ equivalents to percentiles of monthly mean flows computed over the period Oct. 1985–Sept. 2008.

$^d$ consists of two floods with a return period of more than 2 years; properties of the second, larger, flood are listed.

Bearth et al. (1935). The map of Vögele (1984) also indicates the occurrence of springs, swampy areas, and a deep-seated gravitational slope deformation in the back of the valley that was not recognized in the earlier geological maps of the area. The soil map of Krause and Peyer (1986) allowed locating areas with gleyed soil horizons, which are indicative of frequently saturated zones and near-surface runoff formation processes.

5.3.2 Events used for model evaluation

To evaluate the model performance, one small and one large flood event were selected for the Dischma and Hinterrhein catchments according to the criteria for selecting suitable events for the Schaechen catchment (Sect. 4.4). The estimated return periods, peak discharge increases, and responsible rainfall depths are summarized in Table 5.2. The table also includes a proxy for the wetness of the antecedent conditions: the percentile of mean monthly flow that equals the pre-event discharge, $q(t_0)$. A high value suggests that the antecedent conditions in the catchment were wet for that time of the year, assuming that wet conditions are reflected by a high discharge.

The small events are 4- and 5-year floods in the Hinterrhein and Dischma catchments. The large events are the second-largest floods ever recorded in the catchments; the largest events had no reliable input data and nearly equal magnitudes (see Figs. 5.3 and 5.4). As in the Schaechen catchment, strong spatial heterogeneity in liquid water inputs limited the number of events that are suitable for model evaluation. The heterogeneity originates mostly from snowfall or snowmelt effects and from convection cells producing locally intense rainfall.

Five other events were simulated for the Hinterrhein catchment but they are not presented...
Figure 5.5: Subcatchments of the Hinterrhein (a) and Dischma (b) catchments with average gradient of their main channel reaches indicated. Numbers represent the time lags to the catchment outlets in minutes.
here. They include the third-largest flood on record and two small floods caused by moderately intense storms with under 2-year return periods. The simulated discharge peaks were within 25% of the observed peaks for these events. Such deviations are comparable to the uncertainty in catchment areal precipitation, and the simulations were therefore considered to be satisfactory. The model was not further tested in the Dischma catchment because the differences between the flood events are small (Sect. 5.2) and the strongly damped runoff response makes the resulting simulated peak flow relatively insensitive to rainfall input biases.

The Hinterrhein events had relatively dry antecedent conditions, but some storms occurred a few days before the floods. Two storms delivered about 20 mm four and five days before the 4-year event of August 2004 started and about 30 mm of precipitation fell during a five-day period that ended three days before the large flood of August 1988. Pre-event air temperatures were high, indicating that the precipitation was mostly rainfall and any fresh snow had probably melted before the flood-producing storms started.

The small event in the Dischma catchment, in July 2008, consisted of two storms, each producing a flood with a return period of more than two years. The first storm produced about 50 mm and started about two days after a three-day period with 20 to 30 mm precipitation. The air temperatures suggest that snowfall may have occurred in the upper 10% of the catchment during the last hours, which possibly had not completely melted until the start of the first flood-producing storm. This event thus had moderately wet antecedent conditions. The second storm, which produced the larger of the two floods, started four days after the first and thus had particularly wet antecedent conditions.

The large event of the Dischma catchment also caused the largest flood on record in the Schaechen in August 2005 (see Sect. 4.4.2). A preceding storm had similar magnitude as the storm preceding the small event of July 2008, but occurred with a lower 0°C line (≤ 2500 m.s.l.; suggesting that precipitation was mostly snow in more than 25% of the catchment). It ended two days before the simulation period, which itself contained a small, rather local, storm that ended two days before the flood-producing storm (see Sect. 5.4.2). Snowmelt during these days may have contributed substantially to the relatively high pre-event flood (above the 80th percentile of mean August flow, see Table 5.2). In contrast to the Schaechen catchment, here the first half of August was not extraordinarily moist in terms of total precipitation.

5.3.3 Setup of the model simulations

As in the Schaechen catchment (Sects. 4.3 and 4.4), a spatially homogeneous precipitation input was used for the simulations. The input was taken as the rainfall data of the meteorological station that shows the closest agreement between the temporal developments of rainfall and discharge. This rainfall time series is then used without any further corrections. Likewise, the potential evaporation boundary condition was set to the same constant rate of 2 mm d⁻¹.

During the 4-year flood of August 2004 in the Hinterrhein, the two gauges in the area recorded similar rainfall, suggesting a relatively homogeneous storm, such that the rainfall measured at the Hinterrhein station — being within the catchment — was used as catchment-averaged rainfall input. The difference between these two stations was considerable during the large flood of August 1988, where the San Bernardino station was chosen to represent the catchment-averaged rainfall input because of its close resemblance of the hydrograph fluctu-
lations; the Hinterrhein station data might have been more representative at the start of the storm, but the intensities towards the end of the storm were too small to have caused the strong discharge increase.

The simulations of the Dischma events use the rainfall data of the Davos station as the catchment-averaged precipitation input because of the close resemblance of the event hydrograph behavior. Besides, the 24-hour precipitation sums matched well with the daily measurements at the Dischma meteorological station, the only station lying within the catchment. Precipitation amounts at Fluela-Hospiz were also quite similar. Simulations using these data as input, which are not presented here, resulted in flood peaks of similar magnitude, but showed larger deviations in the timing.

The events were simulated using the same model parameters for the different dominant runoff process classes as were used in the simulations in the Schaechen (Table 4.3). Also, the same procedures were used for setting the starting time of the simulation and the initial conditions, except for the small flood in the Dischma catchment, where the simulation was already started 4 hours after the preceding storm to allow a bit more spin-up time.

The flow times in the subcatchments’ main channels were estimated as discussed in Section 4.2.4 for the Schaechen catchment; they are presented in Figure 5.5. Two steep drops make one channel reach of the Hinterrhein relatively steep, but apart from this, the streamed gradients and flow time lags in the Dischma and Hinterrhein are similar to those of the streams in the main valleys of the Schaechen catchment (Fig. 4.3).

5.3.4 Setup of the comparison with calibrated lumped models

Two simple lumped models were selected as benchmark models to compare the results of the DRP framework with (Fig. 5.6). The first is a simple two-parameter nonlinear reservoir model

\[ \text{Figure 5.6: Model structures of the two- and eight-parameter lumped models, L2 and L8. The L2 model is a simple nonlinear reservoir and the L8 model is essentially the QArea* SSF model (Sect. 4.2), but with both the fast (q_f) and slow (q_u) components activated above the same storage threshold (i.e., S_{up,f} = S_{u,fc}).} \]
Table 5.3: Parameters spaces sampled uniformly for the L2 and L8 models.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>L2</th>
<th>L8</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$K_L$</td>
<td>$\eta$</td>
</tr>
<tr>
<td>units</td>
<td>h mm$^{\eta-1}$</td>
<td>mm h$^{-1}$</td>
</tr>
<tr>
<td>min. value</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>max. value</td>
<td>$10^3$</td>
<td>10</td>
</tr>
</tbody>
</table>

referred to as L2:

\[ q_{\text{tot}} = \frac{S_{L2}^{\eta}}{K_{L2}} \]  \hspace{1cm} (5.1)

with the exponent $\eta$ [-] determining the nonlinearity of the response of the nonlinear reservoir $S_{L2}$ [L] and the reservoir constant $K_{L2}$ [T L$^{\eta-1}$] determining its time scale. Model L2 is here used without representation of an evaporation flux.

The second approach is the eight-parameter model referred to as L8. This is a simple structure with an infiltration limit, and fast and slow runoff components, whereby a fraction of the latter recharges an even slower groundwater reservoir. It is the same as the QARea® SSF3 model (Sect. 4.2), except that the infiltration rate limit is allowed to be low enough to cause Hortonian overland flow, and except for the definitions of the thresholds for activating the responses of the upper reservoir: they are merged into one threshold for activating both the $q_f$ and $q_u$ fluxes, which is obtained by setting $S_{u,\text{pf}} = S_{u,\text{fc}}$. This model may be seen as an example of commonly used ESMA-type models, with the $\alpha_u$ parameter responsible for the partitioning between the main fast and slow runoff components. Note that neither of the models uses any function to represent delays caused by routing through the stream network.

A Monte Carlo simulation of 100 000 runs was performed with both models in all three catchments, for both the small and large events. For each model, all parameters were varied using a uniform random sampling in the ranges presented in Table 5.3. As with the QARea® simulations, the initial conditions were set by inverting the $S_{L2}$ and $S_{g,s}$ reservoirs’ flux equations to obtain the state that produces the observed discharge at the beginning of the simulation, with the $S_d$ reservoir empty and the $S_u$ reservoir equal to the field capacity storage threshold.

The same 100 000 parameterizations were used for all the events in all three catchments, such that for each catchment the best performing parameterizations in one event, the calibrated models, can be used for prediction of the contrasting event; i.e., the model is tested outside the conditions it was calibrated for in a cross-validation exercise. This allows evaluating whether the model structures can actually represent the observed behavior and what the differences between the best parameterizations in calibration and prediction are. The 100 best parameterizations in terms of the Nash-Sutcliffe Efficiency (NSE) goodness-of-fit metric (i.e., those with the 99.9 percentile highest NSE scores; Eq. 4.14) were used to represent a typical calibrated model and its parameter uncertainty (i.e., those with the 99.9 percentile highest NSE scores). Selections using the other goodness-of-fit metrics presented in Section 4.3.1 are not presented as they gave similar patterns regarding the prediction skills for the different events, but visual inspection of the simulations indicated that they are less suitable for selecting the best fitting models. The advantages of this Monte Carlo approach to calibration are...
the straightforward methodology and simple assessment of how simulations are sensitive to changes in parameter values.

5.4 Results and Discussions

5.4.1 How do the DRP maps help to understand catchment flood behavior?

The developed DRP mapping procedure was directly applied to the Dischma and Hinterrollable

In the Hinterrollable, large sections of the main channel consist of wide braided river systems. Such features were not yet considered in the mapping scheme. They are here assumed to have a relatively shallow groundwater table below the lands adjacent to the streams, because the constant reworking prevents building up of high land surfaces (compared to the stream water level). These systems are therefore mapped as ‘fast’ reacting alluvial plains; SOF2, or, for the bars and banks surfaces consisting mainly of openwork coarse-grained sediments, SSF1. The military training ground in the alluvial plain of the Hinterrollable was the only sizeable urban area in the catchment and was also mapped as SOF2. Furthermore, there was little evidence that the creeping landmass slopes in the northeasternmost tributary catchment only drain via a deeply fractured rock system (i.e., like the Schluecht slope; Sect. 2.5.1), such that it was conservatively mapped as SSF3 rather than DP-rock.

In the Dischma catchment, the divides between adjacent slopes with thick sediment mantles could not always be clearly defined because subsurface flow patterns appear to correspond little with the surface topography. Likewise, the drainage direction of the small glaciers at the southwestern catchment boundary could not be unambiguously defined. This delineation issue was sometimes also noted when mapping similar areas in the Schaechen catchment. It is of little importance for the total coverage of the different classes at the catchment scale; it may, however, introduce subjective biases at the headwater scale. Furthermore, identification of swampy areas and stream incisions benefited from the aerial imagery and high-resolution soil and digital elevation data, which indicated considerably more of these features than the Vector25 landscape model.

The resulting DRP maps are presented in Figure 5.7 with the coverage of the different DRP classes specified per 100 m elevation bands in Figure 5.8. Distributions of the dominant runoff processes in the Hinterrollable, Schaechen and Dischma catchments are shown in Figure 5.9.

The Hinterrollable catchment is dominated by ‘strongly reacting’ areas, making up more than 52 % of the catchment. These areas are found throughout the catchment and at all elevations, and are predominantly of the SSF1 type; steep slopes with bare rock patches among shallow mixed-grained debris mantles and more gently sloping areas where little soil has formed since glaciers have retreated (for example in the San Bernardino Pass area). The remaining area is mostly mapped as ‘damped’ (41 %), almost a quarter of which is glaciated. The catchment thus has only small fractions of ‘unconnected’ and ‘little contributing’ slopes. The ‘unconnected’ areas mostly drain into thick moraine and talus deposits. Debris cones often have larger catch-
ment areas where runoff is collected in sizeable streams that pass the cone and connect to the main channel network. It could not be tested whether these connections are activated during the moderately intense storms, but it seems likely that some runoff water is lost to these debris cones. As there are many such debris cones, the effect of the ‘strongly’ reacting areas in their catchments may be overestimated by the mapping procedure.

The DRP map of the Dischma catchment differs strongly from the DRP map of the Hinterrhein. Areas with a ‘strong’ response make up only 27% of the catchment, and 24% is of the ‘little contributing’ type. About half of the area with ‘very fast’ response was found to be ‘un-connected’ (corresponding to roughly 15% of the catchment). This is a sizeable portion of the catchment and considerably more than the 3% found in the Hinterrhein catchment. These ‘un-connected’ areas are mostly found in the higher half of the catchment, above 2400 m.s.l. The runoff generated in these areas re-infiltrates mostly in the DP-debris1 and DP-debris2 type moraine and talus deposits that make up about half of this area. They are found on the valley shoulders and as ‘valley fill’ in the hanging and cirque valleys.

At lower elevations, thick deposits are mostly of the DP-debris2 type, generally in the form of debris cones along the main stream. In addition, many of the alluvial plains are mapped as DP-debris1. The slopes draining into these ‘valley deposits’ are mostly mapped as SSF2 and SSF3 and often exhibit little stream initiation (as defined in Sect. 3.6.3). I therefore expect that the subsurface stormflow generated in many of these areas recharges the thick valley deposits, which may further dampen the contribution to catchment flood runoff formation. These slopes also often receive inputs from upslope SSF1 type areas, which are thus not directly connected to the main channel network. These catena issues are not considered in the mapping scheme, as I found no unambiguous criteria for assessing the connectivity of areas with ‘fast’ and ‘damped’ responses (see Sect. 3.6.2). In the Dischma catchment this may result in overestimation of the SSF2 and SSF3 contributions to flood runoff. This also means that the difference between the Schaechen and Dischma catchments may be larger than is expressed by the mapping procedure.

The DRP maps explain well why the Hinterrhein has a stronger and flashier flood response than the Schaechen and Dischma: it has a larger fraction of ‘strongly’ reacting areas and a small fraction of areas that ‘contribute little’ to flood runoff (Fig. 5.9). As the runoff response of ‘strongly’ reacting areas may closely resemble the rainfall intensity once their storage capacity is reached, the flood response is here more sensitive to rainfall intensity than in the Dischma and Schaechen. In these catchments, areas with more delayed reactions dominate, such that the combination of storm depth and duration is also important.

About half of the area of the Dischma catchment is located between 2300 and 2700 m.s.l., and only one-eight is located in the remaining 450 m above this range (Fig. 5.8). Sensitivity of the liquid precipitation input to the snow line elevation is thus modest in the uppermost areas, but raises markedly below 2700 m.s.l. Although it has a slightly wider elevation distribution, the Hinterrhein catchment appears to be more sensitive to the snow line elevation, because the fractions of ‘strong’ or ‘damped’ response types occurring per elevation band are mostly larger, particularly in the uppermost 20% of the catchment (Fig. 5.8). The responses of both catchments are more sensitive to the snow line elevation than in the Schaechen, where the elevation distribution is wider and the fractions of areas with ‘strong’ or ‘damped’ response
Figure 5.7: Dominant Runoff Process (DRP) maps of the Hinterrhein (a) and Dischma (b) catchments; the DRP map of the Schaechen is presented in Fig. 3.4 (p. 72).
Figure 5.8: Spatial coverage of dominant runoff processes in the Hinterrhein (a) and Dischma (b) catchments, specified per 100 m elevation bands and plotted at equal scales; numbers next to the bars indicate catchment areal fraction until the respective elevation bands.

Figure 5.9: Spatial coverage of dominant runoff processes in the Hinterrhein, Schaechen, and Dischma catchments.
occurring per elevation are mostly smaller; in the uppermost 20 % of the catchment, spanning almost 1000 m elevation difference, these types of areas together represent less than 3 % of the total catchment area per 100 m elevation band (Fig. 3.5, p. 73).

5.4.2 How does the QArea$^+$ model explain catchment flood behavior?

The simulations of the small and large events in the Hinterrhein and Dischma catchments are presented in Figures 5.10 to 5.13, and the corresponding goodness-of-fit scores are listed in Table E.2 of the appendix. Recall that the event runoff coefficient $C_{R,e}$ is not plotted for the periods where discharge is below pre-event $q_b$; see Sect. 2.3.4. The simulated peak discharges are within 20 % of the observed discharge peaks, which is comparable to the uncertainty in the estimation of the catchment-averaged rainfall input. The simulations were obtained without calibrating the model to observations in the catchments. Potential biases in the rainfall input are discussed before the results are further interpreted. The direction of the effects of these biases are summarized in the $C_{R,e}$ column of Table 5.2.

The small flood event in the Hinterrhein catchment (Fig. 5.10) was simulated with realistic catchment-averaged rainfall inputs because the two available stations recorded similar precipitation, and hardly any precipitation fell as snow until a few hours after the flood peak. The rainfall also lies within 20 % of the catchment-averaged precipitation estimated by the RhiresD product. The flashy runoff response and the recession were adequately represented. The recorded discharge in the first days of the recession may however not be trusted. The quick drop in discharge below the level prior to the main storm raises the suspicion that the station’s rating curve is not stable.

The large event in the Hinterrhein catchment (Fig. 5.11) was also well simulated, but the San Bernardino rainfall data used as input may have resulted in overestimated catchment precipitation, as the Hinterrhein meteorological station and RhiresD data had considerably less rainfall. These data sets, however, indicated rainfall rates that were at the time of the flood peak only slightly above 10 mm h$^{-1}$, which are unlikely to have caused the observed discharge increase. The real catchment-averaged rainfall may thus be expected to lie between the presented observations, whereby it should be noted that towards the moment of peak flow a considerable fraction of the catchment received snow instead of rainfall. Further note that the event runoff coefficient at the moment of peak flow was smaller than at the moment of peak flow during the smaller event. This may be attributed to the combination of overestimated rainfall inputs and the smaller event lasting longer (such that also the more delayed drainage processes could contribute).

The Dischma meteorological station and RhiresD data suggest that the rainfall used for the simulation of the small event in the Dischma catchment yields a moderate underestimation (Fig. 5.12). The good prediction of the flow peak may thus be seen as support of the thesis pointed out in Section 5.4.1, namely that a considerable portion of the mapped SSF areas do not directly contribute their runoff to the main stream network. The overestimated secondary peaks during the later part of the response to the second storm may also be attributed to the large fraction of the catchment receiving snow.
The Davos station data used for the simulation of the large flood in the Dischma catchment (Fig. 5.13) may underestimate the catchment-averaged rainfall, as the peak intensities are lower than those of the small event and some stations in the areas recorded considerably more precipitation. However, there was considerable spatial heterogeneity and the Dischma meteorological station, the only station that lies within the catchment, registered less precipitation. The underestimation may thus have been small. The overestimation of the discharge peak on 20 August, prior to the main event, may be attributed to the combination of snowfall and the spatially heterogeneous temporal development of the storm (see the $V_f$ graphs in Fig. 5.13).

Simulations using the other rainfall stations for both events in the Dischma showed discharge peaks that were at most 25% higher, although the applied rainfall rates are up to 50% higher (August 2005 peak of 2 mm h$^{-1}$ when using the Weissfluhjoch data, where the rainfall burst causing the flood peak had similar intensity as the Dischma station). This illustrates that the strongly damped response of the catchment makes the simulation results relatively insensitive to input errors.

The strong and flashy response of the Hinterrhein catchment was well represented in the simulations, with the ‘strongly’ reacting areas clearly dominating the response. The dominance of these areas makes the Hinterrhein catchment sensitive to the rainfall intensity. The rainfall depths that are typical for this catchment’s floods likely cause these ‘strongly’ reacting areas to contribute close to their potential rates; storage capacities are exceeded or nearly exceeded during most extreme events, such that the response rates may approach the rainfall intensity (see also Naef et al., 1999). The sensitivity to rainfall intensity is here particularly relevant as rainfall rates in this catchment are typically high.

In the Dischma catchment, where rainfall intensities are mostly lower, the strongly damped behavior could be explained by the large fraction of areas with ‘damped’ or ‘little contributing’ responses. For the catchment floods, rainfall intensity still plays a role because the small fraction of ‘strongly’ reacting areas can be responsible for about half of the flood discharge, but, as in the Schächten (Sect. 4.5), the magnitude of the flood response also depends on the rainfall depth and duration of the storm.

The differences between the catchment flood responses are thus well explained by the model, which, accordingly, may be thought to adequately represent the different responses of the DRP classes. As was also pointed out in the discussion of the model parameterizations in Section 4.3, this leads to a meaningful qualitative description of a catchment’s runoff behavior that also makes sense quantitatively. Uncertainty associated with the quantitative description of course propagates to the catchment-scale runoff prediction, but since the differences between the classes are large, the overall behavior is still mainly determined by the distribution of the DRP. The developed methodologies thereby provide a useful framework for evaluating catchment flood behavior.

The relevance of the understanding provided by the DRP framework can be further substantiated by demonstrating that alternative explanations do not hold. The notion that the observed differences in flood response may not be explained by differences in storm characteristics (Sect. 5.2) can be concretely demonstrated with some of the studied events. For example, the flood-producing rainfall of the small event in the Hinterrhein had similar intensity and total depth (Fig. 5.10) as the flood-producing part of the storm that caused the largest flood on
Figure 5.10: QA\textsc{rea}' simulations of the August 2004 event in the Hinterrhein catchment. (a) Observed rainfall, $P$, and cumulative rainfall, $V_P$; the Hinterrhein meteorological station data was used as model input. The catchment-averaged precipitation estimated from the gridded data product RhiResD (MeteoSwiss, 2013a) is presented as 'MG Hinterrhein.' (b) Observed and simulated specific discharge, $q_{\text{obs}}$ and $q_{\text{sim}}$, with the contributions of the DRP classes, and the fractional catchment area below the 0 °C line, $f_{A,b0}$, as estimated from the air temperature, $T_{\text{air}}$, observed at the Chilchalphorn station. (c) Specific discharge of the different DRP classes; colors as per (b). (d) Observed and simulated event runoff coefficient, $C_{R,e}$, since $t_0$ (vertical dotted line).
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Figure 5.11: QAREA* simulations of the August 1988 event in the Hinterrhein catchment. Subplots are the same as in Fig[5.11] but with different axes scales and station selections; the San Bernardino precipitation data was used as catchment-averaged rainfall input, and the 0 °C lineestimate bases on the air temperature measured at the Hinterrhein meteorological station.
Figure 5.12: \textit{QArea} simulations of the July 2008 event in the Dischma catchment. Subplots are the same as in Fig 5.11, but with different axes scales and station selections; the Davos precipitation data was used as catchment-averaged rainfall input, and the 0 °C line estimate bases on the air temperature measured at Sarsura Pitschen.
Figure 5.13: QAREA simulations of the August 2005 event in the Dischma catchment. Subplots are the same as in Fig 5.13.
Figure 5.14: Observed discharge for the two Schaechen flood events and simulations with the QArea* model and the best fitting parameterizations of the L2 model (note different scales). The left column shows the calibration to the low (a) and high (b) return period flood events; presented are the 10 and 100 parameterizations with highest Nash-Sutcliffe Efficiency (NSE). The right column shows the predictions of these parameterizations for the other event, i.e.: the parameterizations with best fit to the large flood are used to predict the smaller flood (c) and vice versa (d).

Note that the simulated responses of the different DRP indicated in subfigures ‘c’ are indeed similar. This difference cannot be explained from wet antecedent conditions either: the antecedent conditions in the Hinterrhein catchment were comparatively dry. Also recall that snow effects were not important during both events.

Interestingly, doing a similar comparison between the small floods in the Schaechen and the small and large floods in the Dischma suggests that similar rainfalls cause similar peak runoff increases. This is in agreement with the roughly similar coverages of DRP, and indicates that the difference in rainfall characteristics can explain much of the difference in the observed flood behavior.

5.4.3 Are predictions with the DRP framework better than those of lumped calibrated models?

The best 100 parameterizations of the L2 model in calibration and prediction are presented in Figures 5.14 (Schaechen), 5.15 (Hinterrhein), and 5.16 (Dischma). The 10 best parameterizations are specified in a darker grey tone to indicate the tendency and robustness of the calibrated parameterizations. The figures allow comparison with observed hydrographs as well as hydrographs simulated with the QArea* model. The ranges of the NSE values of the simulations
5.4. Results and Discussions

Figure 5.15: Observed discharge for the two Hinterrhein flood events and simulations with the QArea model and the best fitting parameterizations of the L2 model (note different scales). See Fig. 5.14 for details on the subplots.

Figure 5.16: Observed discharge for the two Dischma flood events and simulations with the QArea model and the best fitting parameterizations of the L2 model (note different scales). See Fig. 5.14 for details on the subplots.
with the best 100 parameterizations are listed in Table 5.4.

Appendix D presents the same analysis for the L8 model. This model fits better in the calibration event, but shows a much larger spread in the predictions. Some of the 10 best parameterizations in the calibration even showed worse performance than the 100 best parameterizations of the L2 model, indicating that the overfitting occurs more easily for the L8 model. The main conclusions are illustrated with the L2 model because it gave similar or better predictions than the L8 model.

The calibrated lumped models predicted the flood hydrographs of the Hinterrhein catchment adequately (Fig. 5.15). The calibrations to the two events share more than 60% of the 100 best parameterizations, indicating that the behavior of the catchment does not differ much for the small and large events. The L8 model described the quick recession better than the L2 model; however, it underestimated the response to the short storms preceding the main storm of the small event (Fig. D.2). This suggests that the flashy behavior of the model depends strongly on a threshold in storage capacity or rainfall intensity.

In the Schaechen catchment, both lumped models underestimated the small flood (Figs. 5.14 and D.1) and overestimated the large flood (Figs. 5.14 and D.1). The L2 model could not satisfactorily describe the hydrographs in the calibrations, as delayed runoff peaks cannot be produced by a single reservoir model (see the discussion in Sect. 4.1). The L8 model could be better fitted to the hydrographs, but it still wrongly simulated the highest peak of the small event to be in response to the last high intensity rainfall period (Fig. 5.14), which points out that the interaction between fast and delayed runoff components is not well represented. Also, the better fit in the calibration to the small event did not result in better prediction of the large event, which some of the 100 best parameterizations overestimated even more strongly than any of the 100 best parameterizations of the L2 model. The underestimation of the October 2012 event by both lumped models may be only partially attributed to the overestimated catchment-averaged rainfall input in the calibration to the large flood (Sect. 4.4.2), as this did not make the L8 model too unresponsive: the response to the pre-event storm on 7 October 2012 was

| Table 5.4: Ranges of Nash-Sutcliffe Efficiency (NSE) scores of the presented 100 best-fitting parameterizations of the L2 (Figs. 5.14-5.16) and L8 (Figs. D.1-D.3) models in calibration and prediction. Like in the figures, the evaluation of the models is presented as a ‘cross validation;’ the NSE of the simulations of the large event using the parameterizations with the best fit to the small event are presented in the row corresponding to the large event, and vice versa. Other event properties are listed in Table 5.2 which also lists the return periods (RP) of the flood peak magnitudes repeated here. The NSE scores of the QArea* simulations listed in Table E.2 are repeated for reference. |
|----------------|----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| Catchment      | Event          | RP (y)          | L2              | L8              | L2              | L8              | QArea*          |
| Schaechen      | Oct 2012       | 2               | 0.67, 0.70      | 0.67, 0.84      | 0.03, 0.01      | 0.79, 0.70      | 0.69            |
| Schaechen      | Aug 2005       | 185             | 0.41, 0.41      | 0.83, 0.90      | −1.05, 0.33     | −4.03, 0.85     | 0.73            |
| Hinterrhein    | Aug 2004       | 4               | 0.29, 0.37      | 0.77, 0.87      | 0.20, 0.37      | −0.13, 0.76     | 0.75            |
| Hinterrhein    | Aug 1988       | 41              | 0.42, 0.65      | 0.90, 0.93      | 0.29, 0.65      | −0.87, 0.85     | 0.75            |
| Dischma        | July 2008      | 5               | 0.21, 0.22      | 0.27, 0.70      | −0.41, 0.10     | −4.98, 0.27     | 0.05            |
| Dischma        | Aug 2005       | 30              | 0.83, 0.89      | 0.73, 0.92      | 0.52, 0.63      | −0.72, 0.72     | 0.82            |
predicted well.

In the Dischma catchment, the lumped models predicted too little runoff for the large flood, but showed reasonable skill in predicting the small flood (Figs. 5.16c and D.3c). This is remarkable, as the events had similar storm characteristics and peak flow magnitudes, and were both simulated with slightly underestimated catchment-averaged precipitation, such that their adjustments to input biases may also have been rather similar. As in the Schaechen catchment, the lumped model structures do not adequately describe the delayed responses; they produce the highest peaks at the end of the storm rather than in response to the earlier periods with higher intensity rainfall that actually caused the main flow peaks.

It may not be surprising that the lumped models performed poorly in prediction mode, given that also more sophisticated models, calibrated to more data, often have similar troubles when used for the arguably simpler problem of seasonal water balance prediction (e.g., Seibert, 2003; Merz et al., 2011; Coron et al., 2012, 2014). The reasons for the poor performance may therefore not be so interesting and are not discussed in detail. The discussion here focuses on the L8 model, which at least was capable of behavioral simulation in all three catchments.

The reasons for the poor performance may be mainly sought in the model structure and the strategy that was used to obtain the model parameters, because the precipitation data biases (Sect. 5.4.2) appear to be mostly smaller than the errors in the predictions; for example, the predictions of the large flood in the Dischma catchment showed underestimations of more than 50% although none of the stations measured rates that deviated more than 30% from the used precipitation inputs. Likewise, the overestimation of catchment-averaged precipitation may hardly be more than 40%, yet peak flows could be overestimated by more than 100%. The four parameters determining the response of the $S_u$ reservoir, ($K_u$, $\beta$, $S_{u,\text{max}}$, and $S_{u,\text{fc}}$) were not well identifiable. In the Dischma and Schaechen catchments, these parameters also showed large differences between the optimal values for the different events, such that the problematic predictions may have originated from this part of the model.

The main difference between the optimal parameterizations for the catchments is that the fast response time scale controlled by the $K_u$ parameter was only sensitive in the Hinterrhein catchment, whereas the models of the other catchments showed a higher sensitivity to the $\alpha_u$ diversion parameter, the main control on the fraction of precipitation that is stored over longer time periods. This is in line with the observation that the Hinterrhein has a higher event runoff coefficient and strong runoff response, and the other two catchments store more precipitation over longer time scales, suggesting that some meaningful description of characteristic catchment behavior could be obtained through calibration. However, this did not lead to useful prediction in the Dischma and Schaechen catchments. As other lumped models often have similar structures as the L8 model, the poor performance of the selected lumped model may not be deemed an artifact of the experimental setup, and similarly poor performance may be anticipated for other lumped models.

The tests with the lumped models illustrate that simulation of the selected events is not an easy task for a model. Using the DRP mapping and modeling framework for transferring small-scale knowledge obtained in other areas to the Dischma and Hinterrhein catchments resulted in more skillful and accurate predictions than those of many of the calibrated lumped models:
the hydrograph was better represented, and the NSE values of the QA\textsuperscript{Area}+ model were higher than those of many of the 100 best-fitting parameterizations with the lumped models for all events and catchments (Table 5.4). Moreover, some of the 100 best-fitting parameterizations resulted in very poor predictions (NSE < 0). This comparison of approaches thus indicates that the developed DRP framework is more useful for analyzing and predicting floods in meso-scale alpine catchments than purely inductive approaches like lumped models or the flood frequency analysis techniques discussed in Section 5.2.3.

5.4.4 Usefulness of the DRP framework

The developed DRP mapping and modeling framework allowed adequate prediction of the small and large events in the Hinterrhein and Dischma catchments. The framework’s prediction was based on a simple transfer of small-scale knowledge obtained in other areas via the DRP maps, without catchment-specific adjustment of the parameters of the QA\textsuperscript{Area}+ model used to represent the behavior of the different DRP classes. The transferred knowledge was directly based on observations at hillslope and headwater catchment scales in the Schaechen catchment, and indirectly on sprinkling experiments on various fast reacting hillslopes via the adaptation of the QA\textsuperscript{Area}+ model to the QA\textsuperscript{Area}-C model, the behavior of which was derived from these experiments (Sect. 4.3). This gives the procedure a firm basis in the understanding of runoff formation processes, and allows a meaningful description of how the spatial organization of these processes determines the catchment-scale flood behavior.

The large differences between the catchments and events, combined with the relatively reliable estimation of the rainfall input that was possible for these events, indicate that it is unlikely that data errors have caused ‘false positives,’ i.e., that the predictions were ‘right’ by accident. The good simulation skill in the contrasting catchments thus suggests that the framework allows useful prediction of extreme floods in a wide variety of alpine landscapes. The framework is therefore likely to be useful for exposing differences between catchments and their sensitivities to different storm characteristics, like rainfall intensity, total rainfall depth and snow line elevation (Sects. 5.4.1 and 5.4.2).

The strong foundation in process understanding makes the DRP framework more useful than inductive techniques like lumped models and flood frequency analyses, because the flood causing mechanism can be interpreted more explicitly. Besides, the framework may better facilitate learning about the runoff formation mechanisms through incorporation of new field data or confronting the predictions with observations in other catchments, which purely inductive techniques are also less suitable for. Catchment-specific calibrated lumped models and flood frequency analyses were also demonstrated to be unreliable for predicting the largest floods in the selected catchments, suggesting that the presented combinations of catchments and events provide a stringent test for the flood prediction skill of the developed tools. That the DRP framework passed this test may establish it as promising proof of concept.

Assessing threshold-like changes in the catchments’ flood runoff responses

The developed DRP framework can be used to evaluate of the possibility of threshold-like changes in runoff formation processes that lead to large floods being much stronger than ex-
pected based on the analysis of smaller events. The research in the Schaechen showed that the combination of long-duration and larger rainfall sums may cause such a threshold-like response (Sect. 4.5.1). This may be more likely if there are many areas with ‘damped’ response (SSF3 and DP-fast), as they are expected to produce little runoff during small events, but produce substantial runoff during large ones. The ‘little contributing’ areas (DP-slow and areas draining to neighboring catchments), on the other hand, may also contribute disproportionately more during large floods, but the total effect may be limited: even a strong increase in a small contribution may still constitute only a minor contribution to the total flood runoff, particularly in comparison with the other areas, which will also have an increased response.

The DRP map of the Dischma catchment suggests that a threshold-like response is less likely than in the Schaechen, because it has a smaller fraction of areas with ‘damped’ response (Fig. 5.9). Also, compared to the Schaechen, the storms in the Dischma catchment produce less precipitation and exhibit less variability (Table 5.1), making the conditions that could drive strong changes in behavior less likely. The interpretation that the Dischma catchment has a larger portion of SSF areas draining into DP type landforms than in the Schaechen (Sect. 5.4.1) further supports the assessment that a threshold-like response is less likely in the Dischma than in the Schaechen.

A threshold-like response is also unlikely in the Hinterrhein catchment; increased contributions from areas with ‘damped’ response may occur during longer, more voluminous events, but the dominating ‘fast’ reacting areas may produce less runoff during such events because of the typically lower rainfall intensities.

**The importance of characterizing the connectivity of fast reacting areas**

The aforementioned issues of addressing the connectivity of areas with ‘fast’ and ‘damped’ reactions (Sect. 5.4.1) highlight the importance of secondary, more difficult to quantify, insights into differences between catchments that are not directly represented in the DRP maps. They are however facilitated by the mapping procedure, as the mapper is invited to critically examine the available data and how these lead to a catchment’s DRP map. This may be subjective, but it can be very useful expert knowledge that is hardly possible to obtain in other ways than the detailed study that a mapping exercise necessarily entails. Further research may reduce this subjectivity, as it may reveal ways to concretize knowledge of within-catena connectivity into meaningful metrics, perhaps similar to the way the developed DRP framework has concretized knowledge of dominant runoff processes, i.e., by developing a classification scheme from formulated assumptions that allow interpretation of relevant spatial data.

Assessing the connectivity in the landscape is difficult, and the developed procedure of only mapping the most extreme situations, see Section 3.6.2, may be the first step that is easiest to make. The large difference in the fraction of ‘unconnected’ areas in the Dischma and Hinterrhein catchments demonstrates that already this simple procedure yields a useful quantification of the ‘connectivity problem.’ In the Dischma catchment, about 15% of the catchment, representing roughly half the fraction of areas with ‘very fast’ response, was found to be ‘unconnected’ and thus do not contribute much to even the most extreme events. Not taking this connectivity into account would make the DRP map of the Dischma have a larger fraction of ‘very fast’ reacting areas than the DRP map of the Schaechen; it would even approach the re-
sponsiveness of the Hinterrhein. This indicates that mapping the connectivity of areas with ‘very fast’ response is important, particularly as the ‘unconnected’ areas are often found in the higher parts of alpine catchments, where rainfall rates may be highest.

The difficulties of assessing the effect of more slowly reacting areas draining into areas with large storage capacity was discussed earlier in Section 3.8. The DRP maps may provide some insight into the connectivity of these areas, which may be hypothesized to be positively correlated with the fraction of areas with DP-type thick deposits; only if these deposits exist, can there be slopes that drain into them, and the considerable accumulations of sediments suggest that connected streams with enough power to erode this sediment storage are rare.

Comparison of the fractional coverage of DP-type areas in the three catchments supports this hypothesis (Fig. 5.9): they make up 35% of the Dischma catchment, 28% of the Schaechen catchment, and 21% of the Hinterrhein catchment, a ranking that is in line with the ranking of mapped ‘very fast, not connected’ areal fractions of 15%, 6.5%, and 2.6%, respectively. This example of landscape characterization highlights the ability of tools like the developed DRP mapping scheme to provide a reference framework that can structure the available knowledge about catchment hydrological processes.

5.5 Conclusions

The Dischma and Hinterrhein catchments were selected to evaluate the DRP classification tool in areas with different landscapes and flood behavior. Based on the DRP maps, the QA/AREA+ model was evaluated by applying the model without catchment-specific adjustments; the maps serve to transfer the knowledge obtained at smaller scales in different areas, as presented in Chapters 3 and 4. The simulation skill was illustrated for one low- and one high-return-period flood event, both selected for their reliable observational data, relatively homogeneous spatial distribution of precipitation, and the limited effects of snowfall and snowmelt. The results were compared to simulations with lumped rainfall-runoff models, calibrated for each catchment. The combination of catchments and events provide a useful data set for evaluating how well the developed DRP mapping and modeling framework represents the differences in runoff generating processes, because the differences between the catchment runoff responses may not be fully explained from differences in precipitation characteristics alone.

The classification tool could be applied in the Dischma and Hinterrhein catchments with only minor problems in landform delineation and definition. The obtained DRP maps explained the differences in flood behavior well. The flashy response of the Hinterrhein catchment is due to the large portion of areas (> 50%) with ‘strong’ runoff response, and negligible coverage of ‘unconnected’ and ‘little contributing’ areas. The little-reacting Dischma catchment is dominated by areas with ‘damped’ response and has large fractions of ‘unconnected’ and ‘little contributing’ areas. In addition, the landscape analysis needed for the mapping indicated that the more delayed SSF2 and SSF3 types of slopes commonly drain into DP-type areas, setting the Dischma further apart from the Schaechen and Hinterrhein catchments where these situations are not so common. Although the damping effect of these within-catena redistributions of drainage flows is difficult to quantify, it is valuable expert knowledge that aids interpretation of modeling and predictions efforts. For example, the large fraction of SSF areas draining into
DP-type areas in the Dischma catchment suggests that the difference in damping of the runoff response between the Dischma and Schaechen is larger than the DRP map indicates.

The QA/ArA’ model predicted the wide range of flood responses in the catchments quite accurately; the floods spanned more than an order of magnitude in discharge (discharge increased between 0.8 and 9.5 mm h\(^{-1}\) for the small event in the Dischma and the large event in the Hinterrhein). The model showed similar or better skill in prediction of the extreme floods than simple lumped models that were first calibrated to the low-return-period floods; the lumped models exhibited strong biases in the predictions in the Dischma and Schaechen catchments. These strong biases illustrate the uncertainties involved when applying purely inductive techniques outside the conditions for which their behavior was inferred, a problem that was also demonstrated for predictions based on flood frequency analysis (Sect. 5.2.3).

These results give confidence that the developed DRP mapping and modeling framework, based on improved process knowledge, can be used for scenario analyses that help to assess catchment flood behavior in response to unobserved extreme conditions. The procedures require only generally available spatial data and limited field exploration, and help to evaluate sensitivities to various storm characteristics like intensity, depth, duration and elevation of the snowline, as well as threshold-like behavior. The presented analyses indicate that such threshold-like behavior is less likely in the Hinterrhein and Dischma than in the Schaechen catchment.
Chapter 6

Conclusions and outlook

This thesis explored the potential of improving the understanding and prediction of flood runoff in meso-scale alpine catchments by incorporating process knowledge in tools for mapping and modeling the dominant runoff generating mechanisms. After summarizing the main findings, I briefly discuss the values of these findings and provide some recommendations for further research.

The field research in the Schaechen catchment showed that even steep alpine landforms can store much water and produce little runoff during small events. During large events, some of these landforms contribute substantially to the catchment’s flood discharge, whereas other still produce little runoff. The occurrence of these landforms may strongly influence flood magnitudes and how greatly the large floods differ from the small floods.

The thesis presents new tools to assess the storage and drainage processes of the different landforms. The first tool is a classification scheme for the mapping dominant runoff processes (DRP) on the basis of generally available spatial data and targeted field campaigns. It differentiates between areas with strong runoff response that are directly connected to the main stream network and those that drain into landforms with damped runoff response. The second tool is the spatially distributed rainfall-runoff model QArea+, which can simulate flood runoff on the basis of DRP maps.

These tools form what is here called a DRP framework. This framework was applied in the Hinterrhein catchment, where the flood response is much stronger than in the Schaechen, and in the Dischma catchment, where the flood response is more damped. The DRP maps could explain the differences in behavior and the model predicted the flood peaks and volumes in the Hinterrhein and Dischma catchments without catchment-specific parameter adjustments; the parameterization was only based on process knowledge obtained in other areas, including our field sites in the Schaechen catchment. This indicates that the framework provides an adequate description of the runoff formation processes in different catchments, and gives confidence that it can be used for meaningful prediction of extreme floods.

6.1 On the value of the developed DRP framework

The value of process knowledge, and expert knowledge in general, has become more appreciated in the hydrological literature in recent years (e.g., Seibert and McDonnell, 2002; Merz and
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Blöschl, 2008a,b; Winsemius et al., 2009; Fenicia et al., 2014; Gharari et al., 2014; Hrachowitz et al., 2014. How such knowledge can be represented in analysis tools thereby appears a viable direction for research. The developed DRP framework can be seen as an example of how process knowledge can be included in the study of extreme floods. An important benefit of the framework is the hillslope-scale classification of dominant processes, because this representation of process knowledge may be compared with field studies on runoff formation processes. In this way, the framework may facilitate the synthesis of various sorts of hydrological knowledge; from site-specific knowledge obtained through surveying and field-experimentation to more general knowledge of physical processes and observations at similar sites.

Based on the knowledge accumulated in the framework, new hypotheses can be formulated, and new insights may be added to the framework to improve the predictions that are based on it. This can be nicely illustrated with the hypothesis that triggered the research questions explored in this thesis. A map based on the classification approach of Scherrer and Naef (2003), which mainly considers the near-surface processes relevant for high-intensity storms, indicated that the Schaechen has many slopes with large storage capacity. This lead to the hypothesis that a large fraction of these slopes may store so much water that they only contribute to the largest floods and gave some directions about where such slopes may be found. Working out this hypothesis and related questions yielded the presented tool for classifying the processes occurring in the deep subsurface, which is complementary to the schemes of Scherrer and Naef (2003).

The presented field research yielded valuable insights about how alpine slopes with large storage capacity may affect flood runoff formation. These slopes may dominate in a landscape and thereby largely determine the flood runoff behavior of a catchment, like in the Dischma and Schaechen case-studies. Understanding of the occurrence and drainage time scales of such slopes is therefore important for flood prediction and the study of the possibility of threshold-like changes in flood responses.

Evaluations of the developed mapping and modeling tools in contrasting catchments indicated that they provide a good representation of the storage and drainage processes of alpine slopes with large storage capacity, and that this representation of process knowledge benefits the prediction of large floods. This makes these tools useful for assessing a catchment’s sensitivities to different storm characteristics, such as the effects of snowfall, snowmelt, and rainfall intensity, depth, and duration. The basis in understanding of the physical processes make the DRP framework a valuable addition to the existing tools for predicting the magnitudes of very large floods that hardly reflect the flood-producing mechanisms, such as flood frequency analysis and ‘blindly’ calibrated rainfall-runoff models (Sect. 5.4.4). Moreover, the framework is particularly suited for assessing flood magnitudes in alpine areas where little or no historical data about floods are available, as the mapping tool requires only spatial data that is generally available and the model may provide useful simulations without catchment-specific adjustments.

Critical areas for the runoff response to long-duration storms have been little researched, and also the presented research in the Schaechen entails only two small-scale case-studies. As a result, although a DRP map may reliably identify the areas that could have such critical response,
it will remain difficult to quantify their responses accurately. One may therefore expect the predictions in catchments with large fractions of these areas to be relatively uncertain. Still, the advantage of the approach over more inductive techniques is that one has become aware of this ‘additional’ uncertainty and the possibility of threshold-like behavior that the critical areas may cause.

There are other situations where quantitative predictions are more challenging than in the meso-scale catchments studied here. For example, in catchments dominated by karsti/fied rock or glaciated terrain, predictions depend on how well the behavior of these areas is understood from site-specific research. Furthermore, predictions may become less reliable at smaller scales, where extreme floods are mostly caused by high-intensity rainfall. Small relative biases in the estimation of the response and coverage of fast-reacting areas can then have large effects on the absolute magnitudes of the floods.

Case-studies like those presented for the Dischma and Hinterrhein catchments takes two to four working weeks. This includes data collection, field exploration, application of the mapping scheme, and simulation of some relevant events with quality-checked precipitation inputs. The actual field exploration and mapping make up about one half of that time. This is the only additional effort compared to a study based on off-the-shelf techniques. The benefit of the doubling of the workload is the additional understanding of the system and its main uncertainties. Therefore there is much scope for using the framework in flood management studies.

6.2 Outlook

Knowledge of alpine slopes with strongly damped stormflow responses is important for the understanding of the catchment-scale flood response, however, these slopes are little researched. Researching and documenting the behavior of new exemplary sites may improve (and possibly invalidate) my assumptions about how different properties determine the hillslope response (i.e., in the mapping scheme). Besides, such a catalog of well-documented case-studies can be useful empirical reference of what responses are possible, which helps assessing the uncertainties involved in using this process knowledge for predictions.

New case-studies should not be limited to hillslope-scale research, as the subsurface boundary conditions of hillslopes with thick deposits or fractured rock are very difficult to characterize. Studies of small catchments dominated by some typical landform may suffer less from this problem, such that the catchment area and the specific discharge of the response may be determined more accurately.

In addition to detailed case-studies, there is much scope for improving the hydrological characterization of alpine landscapes. The presented research shows that such characterization can be useful, and that it requires a synthesis of the information contained in different spatial data sets; single spatial data products that provide hydrologically meaningful characterization are rare and often confined to either the soil or the groundwater.

Improving the landscape characterization requires two things: 1) field surveys to map relevant parameters, and 2) techniques like the presented mapping tool for integrating the information of these field surveys with other relevant spatial data with the aim of addressing one or
more hydrological problems (e.g., flood runoff, low flow, groundwater yield). Such field surveys may need to go beyond plot-scale characterizations, although these can be helpful (e.g., Scherrer and Naef, 2003; Markart et al., 2004), to also provide useful information about responses at subcatchment scales. Examples of possible surveying methods include the reconstruction of discharges from flood marks or erosion patterns, measuring discharge during storms, geophysical explorations, and drilling of boreholes; all these techniques have provided useful insights for the development of the presented DRP framework. Studies on new measuring techniques, survey strategies, and tools for synthesizing knowledge, as well as the surveying itself, provide interesting research topics, which, as the explorations in this thesis indicate, may improve the understanding of various hydrological issues.

The developed framework may provide a useful basis for studying other problems where deep-subsurface storage and drainage mechanisms are important, for example, the assessment of landslide risk or low flow discharge. Like for flood runoff formation, some types of slopes may be particularly important, which a DRP map may help to define.

Simulations over longer time scales than presented in this thesis, as needed for the study of seasonal water balances or low flows, are feasible with the current model, but some adaptations in both the mapping and the modeling tool may be required. For example, flowpath length, connectivity, and permeability of the bedrock might be more important factors for the understanding of low flows than for the understanding of high flows. Such factors need further investigation regarding their effects on the problem of interest and how these effects can be generalized into mapping and modeling tools. Adaptations of the model should mainly concern the representations of the groundwater recharge and long-term drainage: changes in the fraction of precipitation that causes runoff at event time scales could reduce the flood prediction skill.

Finally, the framework can be useful for the analysis of scenarios of more extreme conditions than ever observed. This is important for the estimation of design floods and the development of flood management strategies in general. As it is difficult to develop realistic scenarios of low-probability long-duration storms, or estimate the upper bounds of storm rainfall intensity, depth and duration, it was here not yet attempted to do this. It is however an important subject for further research, and it will be the focus of a follow-up project wherein an advanced numerical weather prediction model is used to establish feasible scenarios of very extreme events.
Appendix A

Flood frequency analyses for the studied catchments

The flood frequency analyses (FFA) presented in this theses were based the Generalized Extreme Value (GEV) distribution fitted to the annual maximum floods as available in the instrumental records maintained by the Swiss Federal Office for the Environment (FOEN). Because the catchment area of the Schaechen changed, the floods peaks were converted to specific discharge prior to the analysis. Of the three studied catchments, the Schaechen has the longest flood record (1930–2012). There is good estimate of the maximum flood in the year 1985, when current station was installed about 3 km downstream of the old one. The associated catchment area increased from 94 km$^2$ to 108 km$^2$, an increase of about 15%. This increase in area was assumed to have little effect on the analysis, although the two tributary catchments that were not connected to the old station have a larger fraction of strongly reacting areas than the Schaechen catchment had until 1985 (Fig. 3.4). This may be visible in the time series; the annual floods are generally higher after 1985. No such structural inhomogeneity was reported for the full records of the Dischma (1964–2012) and Hinterrhein (1945–2010) catchments, which also have no years with missing data.

The GEV distributions were fitted using Maximum Likelihood Estimation as implemented in the R package “extRemes” version 2.0-3 (Gilleland and Katz, 2011). The GEV distribution has three parameters, referred to as location, scale and shape. The location and scale parameter are

<table>
<thead>
<tr>
<th>catchment</th>
<th>period (years)</th>
<th>mean (mm h$^{-1}$)</th>
<th>CV (–)</th>
<th>location (mm h$^{-1}$)</th>
<th>scale (mm h$^{-1}$)</th>
<th>shape (–)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schaechen</td>
<td>1930–2012</td>
<td>1.288</td>
<td>0.502</td>
<td>1.017 ± 0.037</td>
<td>0.307 ± 0.029</td>
<td>0.216 ± 0.072</td>
</tr>
<tr>
<td>Schaechen</td>
<td>1940–1976</td>
<td>1.083</td>
<td>0.238</td>
<td>0.968 ± 0.041</td>
<td>0.215 ± 0.030</td>
<td>−0.049 ± 0.148</td>
</tr>
<tr>
<td>Hinterrhein</td>
<td>1945–2010</td>
<td>4.665</td>
<td>0.459</td>
<td>3.701 ± 0.233</td>
<td>1.683 ± 0.168</td>
<td>−0.006 ± 0.089</td>
</tr>
<tr>
<td>Hinterrhein</td>
<td>1989–2010</td>
<td>4.828</td>
<td>0.455</td>
<td>3.897 ± 0.461</td>
<td>1.914 ± 0.328</td>
<td>−0.106 ± 0.164</td>
</tr>
<tr>
<td>Dischma</td>
<td>1964–2012</td>
<td>0.948</td>
<td>0.274</td>
<td>0.843 ± 0.040</td>
<td>0.237 ± 0.030</td>
<td>−0.171 ± 0.135</td>
</tr>
<tr>
<td>Dischma</td>
<td>1976–2004</td>
<td>0.972</td>
<td>0.218</td>
<td>0.913 ± 0.047</td>
<td>0.224 ± 0.037</td>
<td>−0.437 ± 0.164</td>
</tr>
</tbody>
</table>
related to the mean and variance of the sample. The shape parameter describes the upper tail of the distribution; negative values indicate a heavy tail and positive values indicate that the tail has an upper bound; if the shape parameter is zero, the distribution equals the Gumbel distribution, which has a relatively thin tail. The 95% confidence intervals (CI) were estimated from a parametric bootstrap procedure with 2000 draws, because this method may provide the most reliable estimates when sample sizes are small (e.g., Katz et al., 2002).

For the studied catchments, the GEV distributions fitted to the full records were compared with GEV distributions fitted to a subset of the data. The subsets span more than 20 years and do not include the two largest floods on record. This comparison allows evaluating how well a FFA analysis based on a relatively short record could have estimated the probabilities of the largest floods on record. The results were presented in Figures A.1 (Schaechen), 5.3 (Hinterrhein) and 5.4 (Dischma). The mean and coefficient of variation (CV) of the samples and the parameters of the fitted GEV distribution are presented in Table A.1. The differences in variance and shape are clearly visible when the flood records are normalized by the mean annual flood (Fig. A.1); the Schaechen catchment has a markedly different behavior and the largest variance.

The 36-year subset selected for the Schaechen only concerns data of the old station, and could not predict the four largest floods. The prediction based on the shorter subset between the two largest floods on record (1977 and 2005), which also contains data from the present station, was similarly poor: the confidence intervals of the fits to the full record and the subset overlap only a little.

An analysis of the historical floods since the 13th century by Scherrer AG (2007) reveals that the return periods of the four largest floods may be well addressed by the plotting positions based on the 1930–2012 flood record; various floods of similar magnitude were found in the

Figure A.1: Annual maximum floods scaled by the mean annual flood for the Schaechen, Hinterrhein and Dischma catchments. The inset zooms in to the lower left corner of the graph.
historical record, such that the flood of August 2005 has an estimated a return period of 50 to
150 years. The remarkable behavior may thus not be caused by a sampling problem.

For the Hinterrhein catchment, where the two largest floods occurred in subsequent years
(1987 and 1988), the shorter subset of the period after these events was presented because the
more recent data was trusted more and the flood record appears to have a positive trend. Ac-
cordingly, analysis based on the period prior to the two largest floods indicated a considerably
flatter curve. The upper bound of the 95 % confidence interval was below the GEV fitted to the
full record, indicating the probability of the two largest floods on record would be severely
underestimated by a FFA using the more than 40 years of data available until the largest floods
occurred. The good prediction based on the subset presented in Figure 5.3 is thus not robust.

The flood record of the Dischma catchment has the smallest variation (lowest CV) of the
researched catchments, but this did not lead to a better estimation of the largest floods, even
though the two largest floods are only 8 % and 18 % larger than the flood of 1999 that was
included in the subset.

Unlike for the other catchments, the tails of the GEV distributions fitted to the Dischma
flood data appear to have a clear upper bound: the fitted negative shape parameters deviate 1.3
to 2.7 times the standard error from zero (Table A.1). Such an upper bound is physically realistic,
because the largest possible flood may not be infinitely large, like an unbounded tail would
suggest. Often, however, this issue of physical realism is not seen as a limitation of the FFA
procedure because it should be clear from the large uncertainties involved that the knowledge
about the tail of the distribution is highly uncertain and that one should not extrapolate the
fitted curve curve so far. The presented flood frequency analyses, however, illustrate that even
an ‘interpolation’ to estimate a low return period flood may not be trusted; in the Schaechen
case, many floods smaller than the apparent 5-year flood (i.e., according to ranking the floods
of the full record, lie outside the 95 % confidence interval of the GEV distribution fitted to the
36-year subset (Fig. 1.1b).

In summary: the estimates of low-probability floods based on the subsets were unreliable
for all three catchments, or, in the case of the Hinterrhein, at least strongly dependent on the
available data. The uncertainties involved in the procedure imply that even ‘interpolations’ can
be unreliable.

I tried to establish a relation between the annual floods in the Schaechen catchment and the
precipitation sums that caused them. Figures 1.1c–d show the results for the catchment aver-
age 24-hour data available in the RhiresD gridded data product (MeteoSwiss, 2013a). This data
product is derived from the available rain gauges and this data basis may have changed over
time, such that some temporal inhomogeneity may be expected. The Unterschaechen station,
for example, was relocated in 1982. The results, however, were only marginally different if
this station itself was used. The Altdorf station measures at higher temporal resolution since
1981. We found the recorded precipitation often differs substantially from the measurements
at higher elevations in and around the catchment (as discussed in Sect. 2.2.1 and clearly visible
during the August 2005 event; Sect. 4.4.2). It may therefore be no surprise that the analysis
based on this station, even when using different sub-daily intervals, found little correlation
between precipitation sums or intensities and the annual flood peak magnitudes. The IMIS
stations were installed in 1999, such that the record length is too short for this kind of analysis.
One might see a threshold in the relationship between the highest 24-hour precipitation in the three days leading up to the flood event, as the 1977 flood appears to be caused by only a little more precipitation than the smaller floods (Fig. 1.1). However, the 1977 flood was caused by a ~10-hour period with relatively high-intensity rainfall. The storm may thus have been considerably more extreme in terms of intensity, depth and duration. Using data with higher temporal resolution would prevent the masking of these differences between storm events, but the only station with higher resolution data, Altdorf (see above), is not representative of the catchment precipitation input. Because of these data problems and meteorological complexities, the data do not allow inference of how threshold-like the runoff response is.

The poor correlation may not be caused by data errors and resolution problems alone, but also by the omission of corrections for snowfall and snowmelt. Such corrections would be relatively uncertain, and at the daily time scale not meaningful, such that this was not experimented with. This issue may be well illustrated with the flood event of 10 October 2011, the third largest in the analysis period (i.e., 1961–2012). The highest 24-hour sum was about 60 mm, roughly half of the largest 24-hour sums that could be associated with the smaller floods. The 72-hour precipitation sum was not exceptional either. The relatively high flood peak was caused by a less than 12-hour period with moderately intense precipitation and concurrent strong rise in air temperature, together causing a rapid melt of the fresh snow that fell in the three previous days. These kinds of storm characteristics are difficult to describe accurately and are completely masked by the 24-hour precipitation used in the analysis.

There are no reasons why these problems would be less severe in the higher elevation Dischma and Hinterrhein catchments, such that this analysis was not conducted here. There are also no indications that the data basis for such analysis is any better in other meso-scale Alpine catchments.
Appendix B

Discharge monitoring stations at the Schaechen study sites

This appendix contains photo impressions of the research sites in the Schaechen catchment where discharge monitoring stations were installed for this research project. The sites were introduced and characterized in Section 2.2.
Figure B.1: Photos of the measuring setup at the Schluecht creeping landmass slope. Pictures were taken before the 130 m² sprinkling experiment; green windshields were put up at ~2 m from the sprinkled area. (a) Overview of the slope with boreholes (bh) and weir indicated. Dashed line indicates the line of springs that form the start of the swampy base of the slope. (b) Picture of the measuring transect in direction of steepest ascent with boreholes, points sources and weir indicated. The area is swampy (wet peaty soil with reeds) until ~1 m upslope of boreholes 0B and 0C. (c) Close-up of the V-notch weir; stage was measured inside perforated pipes with Keller pressure sensor (no cap) and Odyssey capacitive sensor (blue cap).
Figure B.2: Photo of the Gadenstetten V-notch weir. Water was taken in from a culvert under the road. Rain gauge and shielded temperature sensor were set up at the gauging station. Arrow indicates the highest point source of the spring area.
Figure B.3: Photos of the Wissenboden, Oberbutzen, Egg, and Lehnstutzbach discharge stations. (a) Wissenboden check dam seen from downstream. (b) Close-up of Odyssey capacitive water level sensor mounted on the checkdam; a Keller pressure sensor was similarly mounted on the opposite bank. (c) EWA intake at Oberbutzen with stage recording logger installed at the center of the structures (an Odyssey capacitive sensor is displayed, but the data used in this thesis were recorded with a Keller pressure sensor). (d) Odyssey capacitive sensors and Nivus PCM3 in-stream sensor installed at the Egg intake structure. Creek in the background is fed by the smaller (western) spring. (e) Nivus PCM3 installed in the center of the Lehnstutzbach culvert under the Schaechen valley main road.
Appendix C

Meteorological stations inside and around the studied catchments

The properties of the meteorological stations in and around the Schaechen, Hinterrhein, and Dischma catchments are presented in Table C.1, C.2, and C.3 respectively.

<table>
<thead>
<tr>
<th>station name</th>
<th>properties</th>
<th>location (Swiss grid; CH1903-LV03)</th>
<th>network code</th>
<th>relevant variables</th>
<th>interval</th>
<th>easting (m)</th>
<th>northing (m)</th>
<th>elevation (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Schluecht</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>701755</td>
<td>192168</td>
<td>1489</td>
</tr>
<tr>
<td>Gadenstetten</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>700938</td>
<td>193092</td>
<td>1539</td>
</tr>
<tr>
<td>Wissenboden</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>697299</td>
<td>195722</td>
<td>1704</td>
</tr>
<tr>
<td>Butzli</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>699329</td>
<td>192699</td>
<td>1285</td>
</tr>
<tr>
<td>Aesch</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>704892</td>
<td>191253</td>
<td>1265</td>
</tr>
<tr>
<td>Brunni</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>10 min.</td>
<td>700587</td>
<td>187396</td>
<td>1412</td>
</tr>
<tr>
<td>Haldi</td>
<td>this study</td>
<td></td>
<td></td>
<td>$P$</td>
<td>10 min.</td>
<td>694081</td>
<td>191106</td>
<td>1100</td>
</tr>
<tr>
<td>Altorf (ALT)</td>
<td>MeteoSwiss</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$, $v_w$, $\phi$, $E_{ref}$</td>
<td>10 min.</td>
<td>690174</td>
<td>193558</td>
<td>483</td>
</tr>
<tr>
<td>Unterschaechen (UNS)</td>
<td>MeteoSwiss</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$</td>
<td>24 hr.</td>
<td>701903</td>
<td>192000</td>
<td>1470</td>
</tr>
<tr>
<td>Gross Windgaellen (SCA-1)</td>
<td>IMIS</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$, $v_w$, $\phi$</td>
<td>30 min.</td>
<td>697551</td>
<td>185499</td>
<td>3187</td>
</tr>
<tr>
<td>Seewli (SCA-2)</td>
<td>IMIS</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$, $v_w$, $\phi$</td>
<td>30 min.</td>
<td>697551</td>
<td>185499</td>
<td>2029</td>
</tr>
<tr>
<td>Aelpler Tor (SCA-3)</td>
<td>IMIS</td>
<td></td>
<td></td>
<td>$P$, $T_{air}$, $v_w$, $\phi$</td>
<td>30 min.</td>
<td>702201</td>
<td>194250</td>
<td>2330</td>
</tr>
</tbody>
</table>

$^a$ listed variables include: precipitation ($P$), air temperature ($T_{air}$), wind speed ($v_w$), relative humidity ($\phi$), and reference evaporation ($E_{ref}$).

$^b$ measurements include melted snowfall by using a heated gauge or manual melting accumulated snow at daily read-out.
Table C.2: Measured meteorological variables in and around the Hinterrhein catchment.

<table>
<thead>
<tr>
<th>station name (network code)</th>
<th>properties</th>
<th>location (Swiss grid; CH1903-LV03)</th>
<th>interval</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>network relevant variables</td>
<td>easting (m)</td>
<td>northing (m)</td>
</tr>
<tr>
<td>Hinterrhein (HIR)</td>
<td>MeteoSwiss ( P, T_{air}, v_w, \phi, E_{ref} )</td>
<td>733 900</td>
<td>153 980</td>
</tr>
<tr>
<td>San Bernardino (SBE)</td>
<td>MeteoSwiss ( P, T_{air}, v_w, \phi, E_{ref} )</td>
<td>734 112</td>
<td>147 296</td>
</tr>
<tr>
<td>Chilchalphorn (HTR-1)</td>
<td>IMIS ( T_{air}, v_w, \phi )</td>
<td>731 600</td>
<td>155 000</td>
</tr>
<tr>
<td>Alp Piaenetsch (HTR-2)</td>
<td>IMIS ( P, T_{air}, v_w, \phi )</td>
<td>735 420</td>
<td>156 300</td>
</tr>
</tbody>
</table>

- Listed variables include: precipitation \( (P) \), air temperature \( (T_{air}) \), wind speed \( (v_w) \), relative humidity \( (\phi) \), and reference evaporation \( (E_{ref}) \).
- Precipitation measurements include snowfall by using a heated gauge or manual melting of accumulated snow at daily read-out.

Table C.3: Measured meteorological variables in and around the Dischma catchment.

<table>
<thead>
<tr>
<th>station name (network code)</th>
<th>properties</th>
<th>location (Swiss grid; CH1903-LV03)</th>
<th>interval</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>network relevant variables</td>
<td>easting (m)</td>
<td>northing (m)</td>
</tr>
<tr>
<td>Davos (DAV)</td>
<td>MeteoSwiss ( P, T_{air}, v_w, \phi, E_{ref} )</td>
<td>783 514</td>
<td>187 457</td>
</tr>
<tr>
<td>Weissfluhjoch (WFJ)</td>
<td>MeteoSwiss ( P, T_{air}, v_w, \phi, E_{ref} )</td>
<td>780 615</td>
<td>189 635</td>
</tr>
<tr>
<td>Dischma (DMA)</td>
<td>MeteoSwiss ( P )</td>
<td>786 600</td>
<td>192 990</td>
</tr>
<tr>
<td>Sarsura Pitschen (ZNZ-1)</td>
<td>IMIS ( T_{air}, v_w, \phi )</td>
<td>795 660</td>
<td>175 750</td>
</tr>
<tr>
<td>Puelschezza (ZNZ-2)</td>
<td>IMIS ( P, T_{air}, v_w, \phi )</td>
<td>797 300</td>
<td>175 080</td>
</tr>
<tr>
<td>Fluela-Hospiz (FLU-2)</td>
<td>IMIS ( P, T_{air}, v_w, \phi )</td>
<td>791 600</td>
<td>180 975</td>
</tr>
</tbody>
</table>

- Listed variables include: precipitation \( (P) \), air temperature \( (T_{air}) \), wind speed \( (v_w) \), relative humidity \( (\phi) \), and reference evaporation \( (E_{ref}) \).
- Precipitation measurements include snowfall by using a heated gauge or manual melting of accumulated snow at daily read-out.
Appendix D

Monte Carlo simulations with the L8 model

The figures present the simulations with the best parameterizations of the L8 model, in terms of Nash-Sutcliffe Efficiency (NSE), for the Schaechen, Hinterrhein, and Dischma catchments. They were discussed in Section 5.4.3.

Figure D.1: Observed discharge for the two Schaechen flood events and simulations with the QArea model and the best fitting parameterizations of the L8 model (note different scales). The left column shows the calibration to the low (a) and high (b) return period flood events; presented are the 10 and 100 parameterizations with highest Nash-Sutcliffe Efficiency (NSE). The right column shows the predictions of these parameterizations for the other event, i.e.: the parameterizations with best fit to the large flood are used to predict the smaller flood (c) and vice versa (d).
Figure D.2: Observed discharge for the two Hinterrhein flood events and simulations with the QArea* model and the best fitting parameterizations of the L8 model (note different scales). See Fig. D.1 for details about the subplots.

Figure D.3: Observed discharge for the two Dischma flood events and simulations with the QArea* model and the best fitting parameterizations of the L8 model (note different scales). See Fig. D.1 for details about the subplots.
Appendix E

Goodness-of-fit scores of the QAREA\textsuperscript{+} model simulations

The tables present the scores of the goodness-of-fit metrics applied to the simulations presented in this thesis. Table E.1 presents the results for the model parameterization process discussed in Section 4.3. The scores of the (sub)catchment scale simulations of the analyzed events are presented in Table E.2. The used goodness-of-fit metrics were defined on page 94: the Nash-Sutcliffe Efficiency (NSE; Eq. (4.14)), the Volumetric Efficiency (VE; Eq. (4.15)), and the Mean Error (ME; Eq. (4.16)).

Table E.1: Goodness-of-fit scores of the QAREA\textsuperscript{+} model parameterization processes, computed over the full simulation periods. The parameterizations of the faster DRP classes were obtained through adjusting the behavior to intense and long-duration synthetic storms simulated with the different runoff types (RT) in use with the QAREA-C model. The parameterizations of the fast and slow DP models were adjusted to the behavior observed at the Schluecht and Gadenstetten sites, respectively.

<table>
<thead>
<tr>
<th>DRP</th>
<th>adjusted to:</th>
<th>NSE (–)</th>
<th>VE (–)</th>
<th>ME (mm h\textsuperscript{–1})</th>
</tr>
</thead>
<tbody>
<tr>
<td>HOF1</td>
<td>simulations with RT 1 model; long (intense) storm</td>
<td>0.98 (0.99)</td>
<td>0.92 (0.90)</td>
<td>0.00 (0.22)</td>
</tr>
<tr>
<td>SOF1</td>
<td>simulations with RT 2 model; long (intense) storm</td>
<td>0.99 (0.99)</td>
<td>0.89 (0.93)</td>
<td>0.00 (0.34)</td>
</tr>
<tr>
<td>SSF1</td>
<td>simulations with RT 2 model; long (intense) storm</td>
<td>0.99 (0.99)</td>
<td>0.87 (0.89)</td>
<td>0.00 (0.71)</td>
</tr>
<tr>
<td>SOF2</td>
<td>simulations with RT 3 model; long (intense) storm</td>
<td>0.98 (0.94)</td>
<td>0.81 (0.70)</td>
<td>0.03 (0.69)</td>
</tr>
<tr>
<td>SSF2</td>
<td>simulations with RT 3 model; long (intense) storm</td>
<td>0.95 (0.93)</td>
<td>0.76 (0.66)</td>
<td>−0.05 (0.85)</td>
</tr>
<tr>
<td>DP-fast</td>
<td>observations Schluecht events; Oct 2012 (June 2013)</td>
<td>0.63 (0.76)</td>
<td>0.72 (0.75)</td>
<td>−0.02 (0.06)</td>
</tr>
<tr>
<td>DP-slow</td>
<td>observations Gadenstetten events; Oct 2012 (June 2013)</td>
<td>0.16 (0.81)</td>
<td>0.67 (0.82)</td>
<td>−0.03 (−0.08)</td>
</tr>
</tbody>
</table>
**Table E.2:** Goodness-of-fit scores for the events simulated with QAREA+, computed over the periods presented in the corresponding figures.

<table>
<thead>
<tr>
<th>Catchment</th>
<th>Event</th>
<th>Figure</th>
<th>NSE (–)</th>
<th>VE (–)</th>
<th>ME (mm h⁻¹)</th>
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References


CURRICULUM VITAE

Personal

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Born 30 December 1982 in Maarssen, the Netherlands
Dutch citizen

Education

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Selected work experience


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