Groundwater recharge dynamics by snowmelt

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Mots clés en anglais : Groundwater recharge, karst, climate change, vadose zone, snow hydrology, model uncertainty, snowmelt

Abstract

Understanding the recharge mechanisms for karst aquifers is of great importance because these aquifers are relied upon by a quarter of Earth's population, and because their recharge and discharge dynamics are often strongly coupled. Given this coupling, karst groundwater resources are inherently susceptible to surface processes such as climate, where the timing and volume of water infiltration is a direct consequence of precipitation and temperature. Climate change is anticipated to shift the ratio of snow to liquid precipitation, particularly in regions that currently receive a substantial portion of precipitation as snow.

Thus, we have dedicated this body of research to better understand 1. the mechanisms controlling karst aquifer recharge from snowmelt; 2. how well snow process models actually predict infiltration of snowmelt and what are the predictive uncertainties surrounding these models; and 3. how karst aquifer recharge patterns will shift in a warming climate. We collected three years of data from a unique field site where recharge rates can be tracked in a shallow cave, and which can be considered as an oversized, real-world lysimeter. The collected data embodied spatially integrated behaviors across the lysimeter's catchment area, and allowed us a rare opportunity to depart from system study using point-data.

Through these studies we found that a substantial amount of infiltrating water was stored in the vadose zone (predominantly in soils versus the epikarst), which led to temporal redistribution of water from melt events to cold periods lacking snowmelt infiltration. Vadose zone storage and flow have a strong control on aquifer discharge at the scale of weeks, while phreatic storage becomes dominant during prolonged periods without input. Further, we observed that snow process model predictive uncertainty is reduced with increased parameterization of melt processes. Rigorous snow process model calibration, which allows for model optimization, should become standard practice for water resource managers in cold regions. Lastly, we found that increased air temperature reduces both a snowpack's snow water equivalent at a given time and also its duration of emplacement and that recharge distribution throughout the winter can have significant impacts on

groundwater availability, rendering karst aquifers particularly susceptible to climate change.

Résumé

Comprendre les mécanismes de recharge des aquifères karstiques revêt une grande importance car ces aquifères sont utilisés par un quart de la population mondiale et que leurs dynamiques de recharge et de décharge sont souvent fortement couplées. Compte tenu de ce couplage, les ressources en eaux souterraines karstiques sont intrinsèquement sensibles aux processus de surface tels que le climat, où le moment et le volume de l'infiltration d'eau sont une conséquence directe des précipitations et de la température. Les changements climatiques devraient modifier le ratio précipitations de neige / liquide, en particulier dans les régions qui reçoivent actuellement une part importante des précipitations sous forme de neige.

Ainsi, nous avons consacré ce corpus de recherches à mieux comprendre 1. les mécanismes contrôlant la recharge des aquifères karstiques à partir de la fonte des neiges; 2. dans quelle mesure les modèles de traitement de la neige prédisent réellement l'infiltration de la fonte des neiges et quelles sont les incertitudes prévisionnelles entourant ces modèles; et 3. comment les schémas de recharge de l'aquifère karstique vont changer dans un climat qui se réchauffe. Nous avons collecté trois années de données sur un site unique où les taux de recharge peuvent être suivis dans une grotte peu profonde et qui peuvent être considérés comme un lysimètre surdimensionné et réel. Les données collectées reflétaient des comportements intégrés spatialement dans la zone de captage du lysimètre et nous donnaient une rare occasion de nous écarter de l'étude du système en utilisant des données ponctuelles.

À travers ces études, nous avons constaté qu'une quantité substantielle d'eau infiltrante était stockée dans la zone vadose (principalement dans les sols par rapport à l'epikarst), ce qui a conduit à une redistribution temporelle de l'eau des temps de fonte aux périodes froides sans infiltration de la fonte des neiges. Le stockage et le débit de la zone vadose exercent un contrôle puissant sur le débit de l'aquifère à l'échelle de la semaine, tandis que le stockage de la nappe phréatique devient dominant pendant des périodes prolongées sans apport. De plus, nous avons observé que l'incertitude prédictive du modèle de processus de neige est réduite avec une paramétrisation accrue des processus de fonte. Un calibrage rigoureux des modèles de processus sur neige, permettant l'optimisation du

modèle, devrait devenir une pratique courante pour les gestionnaires de ressources en eau dans les régions froides. Enfin, nous avons constaté que l'augmentation de la température de l'air réduisait l'équivalent en eau de la neige d'un manteau neigeux à un moment donné, ainsi que sa durée de mise en place et que la répartition de la recharge tout au long de l'hiver pouvait avoir des effets importants sur la disponibilité des eaux souterraines, rendant les aquifères karstiques particulièrement vulnérables au changement climatique.

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I would like to thank Daniel Hunkeler, Philip Brunner and Michelle Pronk, who collectively supervised this thesis, for their support, mentorship and openness to discussion. Thank you for prompting new lines of thought and for supporting the extenuating circumstances of me finishing the dissertation remotely. Thank you to Prof. Bjorn Klove and Prof. Nico Goldscheider who have accepted to join the jury of this thesis and for taking the time to read this research. A special thanks to Christian Moeck for his guidance, comradery and patience throughout our collaborations. His time and insights were instrumental in the advancement of my research. The enormous amount of fieldwork completed would not have been possible without Roberto Costa. Thank you for solving innumerable technical problems and for helping to design and install the VCB field station. I want to thank Guillaume Bertrand and Alice Badin, with whom I shared an office. The hours discussing life inside and out of the science world were laughter filled and I count myself incredibly lucky to have been paired with the two of you. I would like to thank my colleagues at the CHYN for a good balance between work and extracuriculars, and for making my PhD years pleasurable ones.

Thank you to the GENESIS project on groundwater systems 513 (financed by the European Commission) 7FP large-scale project contract 514226536, who financially supported my research. The GENESIS colleagues were instrumental in broadening my understanding of hydrogeology and of what it means to be a part of the global scientific community.

I am beyond grateful to my husband Tim for his love and support and for his innumerable hours spent snowshoeing out to the VCB in a meter of fresh snow while carrying 15kilos of equipment. Finally, I would like to thank the rest of my family, especially my mother Lynne and mother-in-law Tina, for their support while I worked to finish writing this document.

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Chapter 1 - Introduction

Thesis objectives

Groundwater resources and climate are inextricably linked, as the timing and volume of water infiltrating into the ground is a direct consequence of factors such as precipitation and temperature. Significant impacts on the hydrologic cycle, including global alterations to groundwater recharge patterns (Barker T. et al., 2007), are anticipated to accompany a warming climate. In karst environments, aquifer recharge and discharge are often strongly coupled. Thus, it is particularly important to understand how recharge might change in these geologic settings. While karstic aquifers are relied upon directly or indirectly by approximately a quarter of Earth's population (Ford and Williams, 2007b; Hartmann et al., 2014), scant scientific attention has been given to understanding the potential impacts of climate change on karst systems' recharge patterns, particularly in cold-climate regions, compared to other aquifer types. Therefore, in this body of research we address the following questions: (i) what are the mechanisms controlling winter-season groundwater renewal in a karst setting; (ii) how do these mechanisms impact karst aquifer storage and discharge; (iii) how well do snowmelt models actually simulate winter season infiltration; (iv) what is the predictive uncertainty surrounding those models; and lastly, (v) how might karstic groundwater recharge change with a warming climate.

Assessment of recharge to karst aquifers: methods and challenges

Karst aquifers are typified by dissolution cavities, such as caves and conduits (Figure 1), which may allow for rapid transmission of water through their vadose and phreatic zones (Ford and Williams, 2007a). Karstic features such as sinkholes and dolines can allow surface water, and any entrained nutrients and contaminants, to quickly enter the subsurface.

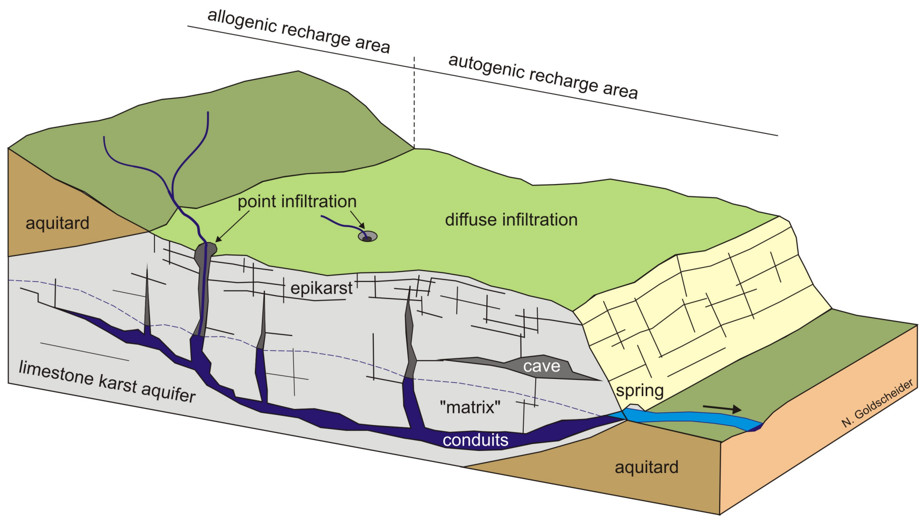

Figure 1: Depiction of the hydrogeologic functioning of a karst aquifer.

Source: (Goldscheider and Drew, 2007)

Direct connection to the aquifer renders subsurface water resources particularly susceptible to surficial conditions such as climate. Furthermore, karst features are heterogeneous across a watershed, which can lead to spatially varying recharge. The spatially inconsistent, yet potentially rapid recharge in a karst watershed results in the heightened need to understand infiltration processes. One way to establish the connection between an aquifer's infiltration and discharge is by employing tracer technologies. Tracer tests are often used to estimate the travel time distribution within an aquifer all the way to its spring, thereby providing information about a site's recharge. Traditionally, tracers such as stable isotopes (Allison et al., 1985; Aquilina et al., 2005; Kohfahl et al., 2008; Lee and Krothe, 2001), radioactive isotope (Savoy et al., 2011), chloride (Allison et al., 1985), ionic and chemical composition (Katz et al., 1997; Lee and Krothe, 2001), fluorometric dies (Flury and Wai, 2003; Göppert and Goldscheider, 2008; Mull, 2001; Smart and Smith, 1976), temperature and chlorofluorocarbons (Long et al., 2008; Long and Putnam, 2006; Ozyurt and Bayari, 2005) have been used to assess recharge to karst aquifers. A comprehensive review of tracers in karst hydrology was presented by Goldscheider et al. (2008). Tracer tests may commence with the introduction a known volume of water (i.e. potential recharge or infiltration) and mass of associated tracer to a groundwater body, either naturally or through injection. The tracer is then monitored for and quantified somewhere hydraulically down gradient in the aquifer. Alternately, one can monitor the physical and chemical properties of a karst spring over time to evaluate infiltration and transmission times of recharge waters. For example, a pulse of infiltrating precipitation may present itself as a breakthrough curve of decreased temperature in a spring's hydrograph (Figure 2).

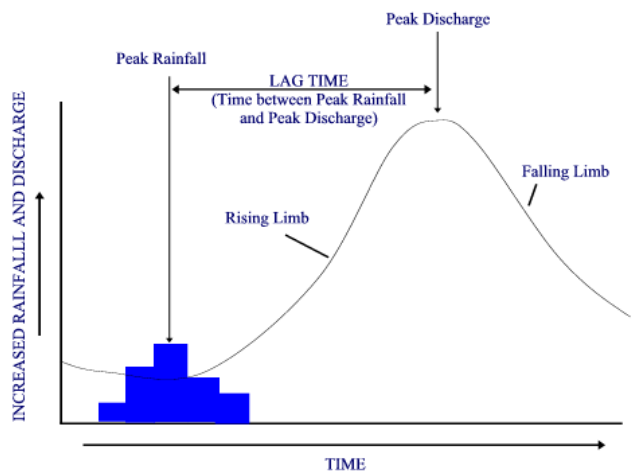

Figure 2: Idealized hydrograph depicting a pulse of rainfall followed by an increase in discharge followed by a decrease in discharge. The lag time between the peak rainfall and peak discharge speaks to the hydrogeologic properties of the geologic media, such a porosity and hydraulic conductivity.
Source: http://vudeevudeewiki.blogspot.com/2012/01/hydrographs-and-river-discharge.html

In locations without snow, a karst system's precipitation input (including the time, volume and chemical makeup) can be directly monitored. However, when a snowpack is emplaced, the volume and signature of liquid precipitation is moderated by the surficial storage of that water in the snowpack, prior to entering the subsurface. Monitoring water volume and signature draining from the base of a snowpack, prior to entering the subsurface, is inherently difficult due to lack of access to the base of a snowpack, without disruption of the snowpack itself. Despite its difficulty, researchers have made efforts to better constrain winter season recharge within karstic settings, typically through analysis of karst aquifer discharge data. While indirect, Geyer et al. (2008) identified daily pulses of recharge from snowmelt based on diurnal fluxes in conduit air temperature, pulses which were not easily identified via spring hydrograph analysis. Reisch and Toran (2014) monitored electrical conductivity, water levels, air temperature, and depth of snowpack at a karst spring to differentiate internal runoff along preferential flow paths from diffuse infiltration. In this study, road-salt served as a tracer to track infiltrating meltwater. These researchers observed patterns indicating that internal runoff, i.e. vadose zone drainage, dominated during frozen periods. Diffuse infiltration became predominate during warmer periods because subsurface thawing allowed the snowmelt to penetrate the epikarst, the layer of enhanced fracturing and solutional voids in the upper part of a karstified bedrock body (Figure 3).

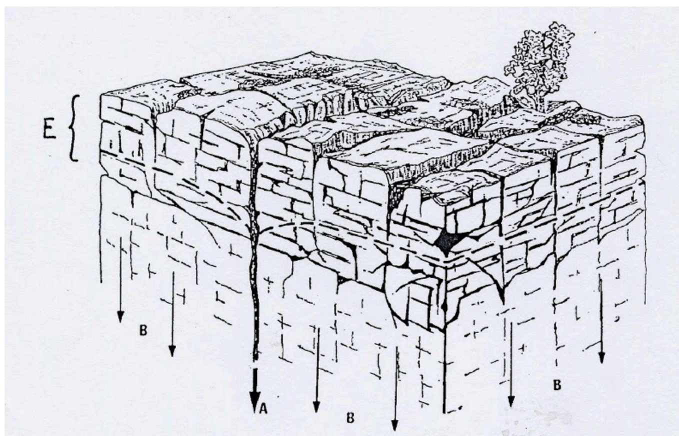

Figure 3: The epikarst (E) is the upper layer of the bedrock that is heavily fractured and riddled with solutional voids.
Source: https://www.geocaching.com/geocache/GC3GFXD_hubelj?guid=c2a747a7-9f7b-46a7-aa1d-867a8f76e639

Penna et al. (2015) used stable isotopes of water (^{18}O and ^{2}H) and electrical conductivity to quantify the role of snowmelt in spring water in the lower part of a small Dolomitic catchment. These researchers concluded that: 1. rainfall, snowmelt, shallow riparian groundwater and near-surface soil water were the main contributors to stream runoff; 2. springs and surface runoff in the upper part of the catchment reflected a more direct influence of rainfall and snowmelt than springs and surface runoff in the lower part; and 3. relatively constant isotopic composition and EC in streams and springs in the lower part, and evaporation signal suggest slower subsurface flow and longer residence time. Audra and Nobecourt (2013) assessed discharge and temperature data at the Coulomp Spring in France and derived snowmelt to contribute 30–50% toward discharge with the remainder attributed to baseflow. Through hydrograph analysis of peak air temperature and peak discharge from snowmelt, these researchers determined the relative contributions of vertical transfer (7hrs) through the vadose zone and horizontal transfer (3hrs) through the karst conduits. Though somewhat tangential, Gremaud and Goldscheider (2010a) and Gremaud (2009) used tracer tests to understand the influences of glacial meltwater on karst aquifer recharge.

Only a few studies have investigated karstic recharge by snowmelt in more detail directly in the recharge zone. Penna et al. (2015) collected samples from a snow lysimeter, piezometers, snow courses, and soil-water suction cups. With snow courses, the snow water equivalent (SWE) is determined repeatedly with a snow sampler along well-defined transects. These water samples however are discrete and only represent the point from which they were sourced. Uncertainty is introduced when extrapolating watershed-scale processes from point data, a consistent difficulty in snow hydrology studies, as is discussed in depth in the Introduction of Chapter 2. Regardless of the geologic setting the rate of snowmelt water entering the ground is usually monitored using point-data sources such as snow lysimeters (Haupt, 1969; Kattelmann, 2000; Martinec, 1989; Tekeli et al., 2003; Tekeli et al., 2005) and snow pillows (Archer and Stewart, 1995; Butcher and McManamon, 2011; Trujillo and Molotch, 2011). Snow courses, such as those collected by Marks et al. (2001) and Rice and Bales (2010), provide a distributed understanding of changes in snow water equivalent. However, snow courses are typically done at course time resolutions due to the intensive labor involved. As stated by Kattelmann (2000), snowmelt runoff from a larger "natural lysimeter" would provide a conceptually better basis for evaluating winter recharge processes and output from snowmelt models, than the somewhat artificial sampling of snowpack outflow by lysimeters (Figure 4) and snow pillows.

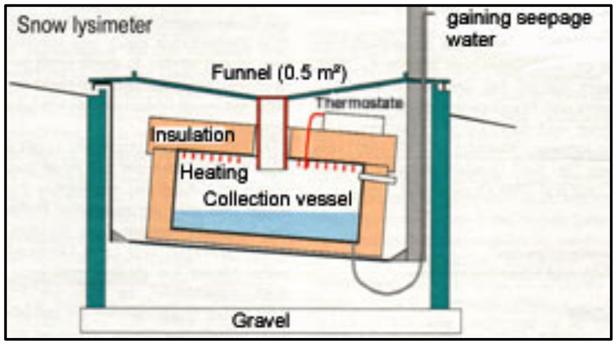

Figure 4: A cross-sectional schematic of a snow lysimeter used in northwest Austria.
Source: http://www.lysimeter.at/HP_EuLP/web/austria/at18b.html

Therefore, we used the rather unique karstified geologic configuration of the Vers Chez le Bandt, a natural lysimeter located in northwest Switzerland, to address the above delineated research questions.

Mechanisms controlling recharge

Due to karstified aquifers being highly heterogeneous and tending to maintain individualized, site-specific relationships between soils, epikarst, rock matrix and conduit network, which can vary significantly from other karstic settings, the mechanisms controlling karst aquifer recharge have been debated (Chapter 1, Introduction). Karstic settings tend to systematically present vadose zone drainage even during periods of drought, implicating the presence of perched water stores. That later feature is seen around the world in karstified location and is typically attributed to water storage in the epikarst (Aquilina et al., 2005; Arbel et al., 2010; Bakalowicz, 2004; Clemens et al., 1999; Jacob et al., 2009; Perrin et al., 2003; Trçek, 2002; Williams, 2008). An in-depth discussion of the respective roles of soil and epikarst in surficial water storage, as surmised by various researchers, is presented in the Introduction of Chapter 1, along with our findings regarding how these stratified zones influence infiltration and storage of infiltrating water.

Modeling recharge from snowmelt in a karstic setting

While many efforts to model the volume and temporal distribution of snowmelt rates have been conducted at both the point and distributed scales using a plethora of modeling algorithms (as detailed on the Snow Modelers Internet Platform and in Chapter 2's Introduction), few have published studies that modeled winter season groundwater renewal in karst settings. Rimmer and Hartmann (2010) modeled recharge from snowmelt to karst. These researchers applied a degree-day based snow routine to HYdrological Model for Karst Environment (HYMKE) to estimate the effect of snow accumulation and snowmelt on the timing of stream flow in the Jordan River. They showed that at higher elevations there is a clear difference between the timing of effective precipitation (precipitation plus snowmelt) compared to just precipitation. However, according to their model, these changes had almost no effect on the final results of the daily stream flow, most likely due to snowmelt's low contribution compared with rain at lower evaluations. This study failed to collect data, such as evolution in SWE, which could have been used to calibrate the degree-day portion of their model.

Predictive uncertainty surrounding snowmelt models

As stated, there have been many efforts to constrain snow processes and associated drainage from snow packs using a spectrum of model complexities which range from the simple temperature index method to the intricate and physically-based energy balance method (as discussed in Chapter 2's Introduction). With each snowmelt model simulation, regardless of complexity, there results a unique outcome of snowmelt rates. These outcomes are based on the model's initial conditions, configuration, and used calibration data (Hill and Tiedeman, 2006). To understand how reliable and in turn powerful these model-forecasts may be, a model user must understand the uncertainty surrounding both a model's input (such as accuracy of air temperature data) and output (estimated snowmelt). Predictive uncertainty of the output may arise from inaccuracies in the model's structure, i.e. representation of a system's functioning (Moore and Doherty, 2005). For example, simplified models that lump many processes via one index, may be more prone to produce predictive uncertainty as they are not physically based and do not consider the explicit nuances of a particular system in space and time. However, parameter uncertainty may be small as there are fewer parameters to constrain. Conversely, using a physically based model that considers a highly parameterized system may reduce uncertainty of the model's predictions. However, with an increase in parameterization, parameter uncertainty may also increase (Brunner, 2015; Moeck et al., 2016). Also, highly parameterized models may not be practical, as they require greater volumes of input data, which can be expensive with regard to time, cost and computational requirements. Prior to the study presented in Chapter 2, a systematic assessment of the predictive uncertainty surrounding various grades of snowmelt model complexity had yet to be completed.

Effect of climate change in a karst setting

According to Hartmann et al. (2014), climate simulations project a strong increase in temperature and a decrease of precipitation in many karst regions in the world over the next decades. Specifically, projections across Switzerland include an increase in mean temperature and decrease in summer mean precipitation (CH, 2011). Hartmann et al. (2014) then noted that despite this potentially changes, few studies have specifically quantified the impacts of climate change on karst water resources. Loaiciga et al. (2000) were among the few. These researchers applied climate change scenarios, derived from general circulation models, to assess the potential shifts in water resources in the Edwards Balcones Fault Zone (BFZ) aquifer in Texas, USA, a region that does not receive snow. In this study, historical climatic time series in periods of extreme water shortage, near-average recharge, and above-average recharge were scaled to $2\times CO_2$ conditions to create aquifer recharge scenarios in a warmer climate. These scenarios were then combined with several pumping states. This was done to assess the sensitivity of water resources to human-induced stresses. Their simulations indicated that $2\times CO_2$ climatic conditions could exacerbate negative impacts and water shortages in the Edwards BFZ aquifer even if pumping does not increase above its present average level.

As stated by Moeck et al. (2016), sustainable management of water resources requires the evaluation of potential climate change impacts. If one is interested in how groundwater systems react to climate change, the first thing to do is evaluate potential changes in water input to the aquifer, i.e. groundwater recharge. Increasing temperatures and changing precipitation patterns could either increase or decrease the rate of groundwater recharge throughout a year, altering the quantity and possibly the quality of water available for human use. This is due to the coupling between precipitation, air temperature and groundwater recharge rates. The last two decades has seen a surge in research focused on modeling the interplay between climatic parameters and groundwater recharge rates at everything from the point to global scales (Adam et al., 2009; Brouyère et al., 2004; Calanca and Semenov, 2013; Dibike and Coulibaly, 2005; Eckhardt and

Ulbrich, 2003; Herrera-Pantoja and Hiscock, 2008; Loáiciga et al., 2000; Moeck et al., 2016), research that has improved our understanding of how a warmer climate might impact water resources. This area of research has been in large part motivated and facilitated by the Intergovernmental Panel on Climate Change (IPCC), a group tasked to assess the scientific, technical and socio-economic information relevant for the understanding of the risk of human-induced climate change (Barker T. et al., 2007). The IPCC has now released three generations of emission scenarios that have been used to drive global circulation models (GCM), which in turn have been used to develop climate change scenarios. GCMs have been down-scaled and climate scenarios tailored for application at smaller scales that allow for region-specific predictions. However, only few studies have specifically evaluated potential changes of recharge to karst aquifer in regions with a seasonal snowpack. Thus, we modeled how karstic aquifer winter-season recharge might respond to increasing air temperatures, to better understand potential shifts in future groundwater availability.

Thesis outline

The subsequent chapters of this dissertation consist of two primary authorships in peer-reviewed journals (Chapters 1 and 2), an unpublished study and thesis conclusions (Chapter 3), followed by two appendices that contain two co-authorships in peer-reviewed journals (Appendixes A and B).

Chapter 1 identifies the mechanisms behind winter season recharge to a karstic aquifer. This was assessed through analysis of surface and vadose zone time series data collected from a unique catchment with a known recharge area (~1600m^2) that drains to a point 53m below ground surface in a cave, a configuration akin to a large natural lysimeter. Implications of shallow vadose zone processes on karst aquifer recharge, storage and spring discharge are then evaluated. This study was undertaken to understand not only the aforementioned processes, but to establish baseline conditions of karst aquifer response to the current climate for use in future climate change predictions as presented in Chapter 3. Chapter 1 was published in the Journal of Hydrology in 2015.

The unique experimental configuration and data set provided an opportunity for the evaluation of the predictive uncertainty surrounding snowmelt models. For karst systems, snowmelt model predictive uncertainty is particularly important as discharge and recharge are tightly coupled, and the latter strong depends on the melt rate dynamics. Due to this tight coupling you need a good melt-model even if you are mainly interested in the spring discharge. This study was undertaken to constrain how well snowmelt models of different complexity actually predict infiltration from snowmelt. The results from this study were published in the Journal of Hydrology in January of 2017 and are presented in Chapter 2.

Chapter 3 presents the results from the application of simplistic climate change scenarios to the energy balance snowmelt model used in Chapter 2 as well as some conclusions and implications of these studies. These projections were made to evaluate how groundwater recharge rates in a karst system may shift with a warming climate.

As a contributor to the GENESIS project, of consortium scientists from across the European Union tasked to evaluate groundwater and dependent ecosystems and funded by the European Union's 7th Generation Groundwater Directive, I participated in research somewhat tangential to the principal thread of my dissertation. Byproducts of my involvement in GENESIS work-packages 2 and 4 were the two publications attached as Appendices A and B. Appendix A presents a study wherein stable isotopes (^{18}O and ^{2}H) were used to identify the sources of water used by alluvial tress. Appendix B presents a critical review of groundwater vulnerability assessment methods.

Chapter 2

Snowmelt infiltration and storage within a karstic environment, Vers Chez le Brandt, Switzerland

Jessica Meeks, Daniel Hunkeler

Accepted by Journal of Hydrology

Abstract: Even though karstic aquifers are important freshwater resources and frequently occur in mountainous areas, recharge processes related to snowmelt have received little attention thus far. Given the context of climate change, where alterations to seasonal snow patterns are anticipated, and the often-strong coupling between recharge and discharge in karst aquifers, this research area is of great importance. Therefore, we investigated how snowmelt water transits through the vadose and phreatic zone of a karst aquifer. This was accomplished by evaluating the relationships between meteorological data, soil–water content, vadose zone flow in a cave 53 m below ground and aquifer discharge. Time series data indicate that the quantity and duration of meltwater input at the soil surface influences flow and storage within the soil and epikarst. Prolonged periods of snowmelt promote perched storage in surficial soils and encourage surficial, lateral flow to preferential flow paths. Thus, in karstic watersheds overlain by crystalline loess, a typical pedologic and lithologic pairing in central Europe and parts of North America, soils can serve as the dominant mechanism impeding infiltration and promoting shallow lateral flow. Further, hydrograph analysis of vadose zone flow and aquifer discharge, suggests that storage associated with shallow soils is the dominant source of discharge at time scales of up to several weeks after melt events, while phreatic storage becomes import during prolonged periods without input. Soils can moderate karst aquifer dynamics and play a more governing role on karst aquifer storage and discharge than previously credited. Overall, this signifies that a fundamental understanding of soil structure and distribution is critical when assessing recharge to karstic aquifers, particularly in cold regions.

Keywords: snowmelt, recharge, storage, karst, vadose zone

Introduction

With increased global temperature, the hydrologic cycle could undergo significant alteration including possible reductions in seasonal snow cover (Beniston et al., 2003) and shifts in amount and type of precipitation (Arnell, 2001). Alterations in these parameters would invariable affect the volumetric and temporal distribution of groundwater recharge, particularly in cold-regions (Eckhardt and Ulbrich, 2003). Given that seasonal snowpacks play a significant role in the storage and redistribution of water resources (Bayard et al., 2005), several studies have addressed recharge and runoff processes attributed to spring onset snowmelt (Barnett et al., 2005; Buttle, 1989; Flerchinger, 1992; Nabi et al., 2011). However, with proposed temperature shifts possibly leading to reductions in seasonal snowpack duration and volume, the classic paradigm of winter snowpack water storage and spring on-set melt of snow may transition to multiple, ephemeral accumulation and melt cycles of snow throughout a winter/spring cycle. Therefore, more attention must be given to inter-winter infiltration processes and the mechanisms that control them, enlarging the historic focus of recharge studies beyond spring onset snowmelt. Thus we aim to expand on the few previous studies that have investigated inter-winter recharge (Iwata et al., 2010) with our study which takes place in a karstified watershed. In such aquifers, recharge is often tightly coupled to discharge due to the presence of conduit networks (Moore et al., 2009). Hence, changes in recharge patterns, due to increasing temperatures, might have a particularly strong effect on discharge trends in karstic watersheds. Additionally, caves present the opportunity to physically enter the vadose zone of a study area, a convenient advantage to other aquifer types. As such, observation of temporal trends in recharge rates can be observed directly within conduits, thereby more effectively elucidating the actual hydrological processes involved (Buttle, 1989). Karstic aquifers are broadly relied upon by an estimated 25% of earth's population for drinking agricultural and industrial purposes and are thus an area of great concern (Ford and Williams, 2007b; Hartmann et al., 2014).

The epikarst, a spatially variable (Hartmann et al., 2012) layer of enhanced porosity that can encrust soluble bedrock, is thought to influence the temporal distribution of groundwater recharge (Klimchouk, 2004; White, 2004; Williams, 2008). This conceptual understanding is based on a breadth of studies that investigated how and why cave drip water continued to appear within karstic vadose zones even during extended periods of drought (Bakalowicz et al., 1974; Friederich and Smart, 1982; Mangin, 1973; Williams, 1983). The epikarst was identified as a layer in which perched storage and lateral flow can occur (Friederich and Smart, 1982; Smart and Friederich, 1986; Williams, 1983). Observed rapid reactions to rain events at stalactite drip points were explained by Williams (1983) as the result of shallow lateral flow in the epikarst to vertical drains, allowing for rapid infiltration. Alternately, Klimchouk and Jablokova (1989) proposed that rising hydraulic head in the epikarst induces rapid infiltration of storm event water. Trcek (2002) built upon this latter theory by proposing "the piston effect", in which "old" water stored in the soil and epikarst must be flushed out first followed by "new" water from a storm. Water storage was hypothesized to occur due to decreases in permeability through an epikarst's vertical profile (Perrin et al., 2003) and differences in hydraulic conductivity between the epikarst and the lower unsaturated zone (Trçek, 2007). The significance of the epikarst storage was postulated by Trcek (2007), who concluded that karst aquifer flow largely depends on the hydraulic behavior of the epikarst zone and by Aquilina et al. (2006) who asserted the major role of the epikarst reservoir in the karst recharge functioning. And in a broad assertion, Perrin et al. (2003) posited that storage in the epikarst could be more significant than storage in the underlying phreatic zone.

The interactions between and respective function of the epikarst and overlying soils have been debated, complicating the identification of recharge mechanisms in karst settings. White (2004) asserted soil cover to be unrelated to water storage in the epikarst, while Jones (2003) believed much of the apparent epikarst storage of storm water to be held in soil-filled fissures of the epikarst. White (2004) considered that while the A and O soil horizons (the American soils classification system) should be excluded from the epikarst, normally the B horizon that fills the solutional voids, should be included. Celico et al. (2010) concluded that epikarst formation can be reliant on soil thickness. Williams (2004) conceded that where soil is present, it would most likely moderate infiltration and provide further storage of water. Lee and Krothe (2001) found epikarst, rather than soils, to be the dominant contributor to river recharge following a storm event. In contrast to these works, Tooth and Fairchild (2003) saw soil matrix flow as the dominant karst water source during dry periods, rather than the epikarst. Perrin et al. (2003) assessed soil and epikarst storage as a cohesive unit and deduced that while significant soil moisture storage did moderate mixing and infiltration velocities, dynamic storage could only occur in the epikarst. Therefore, ambiguity still exists regarding where exactly in the vadose zone modifications to recharge are actually taking place.

The need to resolve this ambiguity is further heightened when considering recharge from snowmelt water, where surficially stored precipitation is temporally redistributed, complicating the groundwater recharge process. While infiltration from glacial melt has been studied (Gremaud and Goldscheider, 2010a; Gremaud and Goldscheider, 2010b; Zeng et al., 2012), only Reisch and Toran (2014) have assessed the transient nature of recharge from seasonal snowmelt in karstic aquifers. By assessing hydrochemographs at a karst spring, these researchers related signatures in overall spring discharge to variations in internal runoff and diffuse infiltration. While Reisch and Toran (2014) considered soils separate from the epikarst, not much consideration was given to the role in which soils may influence the epikarst and underlying aquifer.

The unique configuration of the Vers Chez le Brandt (VCB) study location, where this study takes place, allowed us to build upon these recharge studies and also take into consideration the array of methodologies for assessing snowmelt infiltration used in other lithologic settings (Bayard et al., 2005; Buttle, 1989; Flerchinger, 1992; Sutinen et al., 2008). While varied in approach, all these studies sought to relate snowpack basal outflow to an increase in recharge, via surficially accessed data. Our analysis builds upon these surficial study configurations by assessing for snowmelt infiltration within the karst conduits in addition to the soils, upper epikarst and aquifer's spring.

The objectives of this study were to assess how and when snowmelt waters transit and store within a karstic aquifer's vadose zone during winter and spring snowmelt, and how groundwater recharge and discharge are related. We investigated recharge processes at two different temporal scales: firstly, to investigate how daily melt water pulses are attenuated by the soil/epikarst system and aquifer; and secondly how soil/epikarst storage can provide water during prolonged cold periods and how quickly this reserve is replenished again during periods of snowmelt. The study was carried out at the VCB in the karstified Areuse aquifer in the Swiss Jura Mountains. Here, inter-winter melt events frequently occur and the vadose zone can be accessed via a cave. The recharge area and discharge for a cave drainage point (ie. vadose zone outflow, VCB1), 53m below the ground surface, are known. Consequently, the VCB site resembles a large, real-world lysimeter and presents an ideal situation to evaluate recharge dynamics, as infiltration rates can be directly quantified.

Site Description

Meteorological winter conditions in the Jura range are characterized by cold temperatures associated with significant snow accumulation (Bouoncristiani, 2004). The VCB sites receives approximately 1550mm of precipitation annually, 30 to 40% of which falls as snow between the months of December and March (www.meteoswiss.ch). A proximal Swiss Agrometeo meteostation in Les Verriers (525500, 199175 universal polar stereographic (UPS) coordinate system; Campbell-CR10x) shows that average summer and winter temperatures for the area are +14^{o}C and -1^{o}C respectively. The site is primarily vegetated by cocksfoot and ryegrass meadow and flanked by forest composed of fir and spruce species.

The Areuse karst aquifer, with a catchment area of 130 km^2, is located in the Swiss Jura Range's western edge. It discharges to a single spring with an average discharge rate of 7.15 m3/sec located at an elevation of 793 m. The watershed's groundwater resources are directly or indirectly relied upon by thousands of people via pumping wells within the karstic aquifer or for water supply and/or hydroelectric production needs generated hydraulically down gradient.

The VCB (526450/199010 UPS) site is situated within a flat area 1160 m above sea level (Figure 1). The single-chamber VCB karst cavity, which underlies the study site (average orientation N145^{o}), inclines at an angle of 13° and parallels the imbricated bedding planes (Kiraly and Simeoni, 1971). From the VCB entrance shaft, the cave drops approximately 55 vertical meters over a distance of 260 m. The chamber is underlain by a marl sequence that is thought to confine the karst's basal development (Goldscheider et al., 2008) and serve as an aquiclude within the watershed. Waters entering the VCB cave via drips from stalagmite straws and from the two or three (depending on degree of overburden saturation) vadose zone drainage points contribute to a perennial stream that traverses the cave length (Figure 1). The primary vadose-zone exfiltration point (VCB1) is located approximately 175 m from the cave entrance (Savoy, 2007).

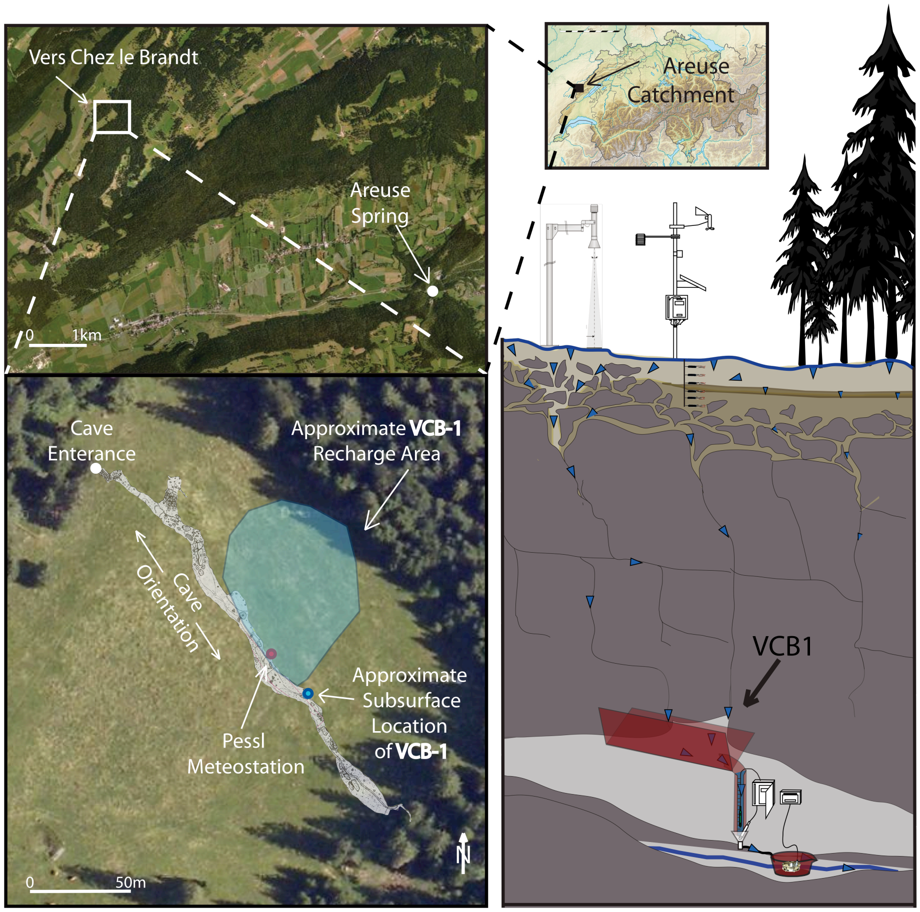

Figure 1: Counter-clockwise: a regional site map; the Vers Chez le Brandt (VCB) site with approximate recharge area, cave roof drip point (VCB1) and cave orientation denoted; and profile (not to scale) of site instrumentation with conceptual flow indicated by the blue arrows.

Previous studies at the site concluded that the soil zone (Perrin, 2003) and epikarst (Perrin et al., 2003; Pronk, 2009; Savoy, 2007) serve as a collective buffering reservoir to diffuse recharge. While not validated through detailed geophysics or soil borings, water was thought to percolate through the soil zone and enter the epikarst where it was stored prior to reaching the karst conduits that drain to VCB1. A series of unpublished VCB tracer tests revealed the approximate VCB1 recharge location to be north and adjacent to the cave's orientation (Figure 1). During this series, tracers applied above the southern side of the cave axis were never observed at VCB1. A 1979 tracer test proved hydraulic connection between the VCB cave and the Areuse Spring (Müller, 1982).

VCB bedrock is composed of Upper Jurassic (Portlandian, Kimmeridgian and Sequanian) aged marl and fossilferous limestone (Sommaruga, 1997; Valley, 2002). Two separate

Quaternary glaciers acted upon the VCB region leaving the post-ablation landscape completely denuded of soil (Campy, 1982; Campy, 1992). Mineralogical comparison of Jura soils with underlying calcareous bedrock showed that Aeolian silts of crystalline origin (plagioclase, chlorite, feldspar and an abundance of quartz) blanketed the range since glacial ablation (Pochon, 1978). These sediments accumulated to form the Neoluvisol loess soils currently in place. During VCB instrument installation, three soil pits were dug for emplacement of soil moisture sensors and for concurrent soil classification. Additionally, 7 soil cores were taken within the recharge area in the summer of 2011 to verify soil type (using the French soils classification scheme) and distribution. In all investigated locations, the upper 10 cm of VCB soil were found to be an organic-rich type A soil, dominated by shallow root structures and with a pH of 4.5. A silty type E loess, with a pH of 4.5, extends from 10cm to approximately 50cm b.g.s.. This is in turn underlain by a 10cm thickness of brown clayey BT soil (pH of 5.5). Beneath this is an undefined thickness of grey, clayey IC soils, with a pH of 8 (Figure 2). According to Baize and Girard (2009) Neoluvisol (aka. sols bruns) soils correspond with the Luvic Cambisols as described by the FAO World Reference Bank for Soils (Micheli et al., 2006). Table 1 presents further detail pertaining to each soil horizon. Hand auger borings terminated at 70 to 130cm b.g.s. due to increased fractions of limestone cobbles. Field observations were confirmed by a seismic study (Müller, 1978), which revealed a layer of low permeability at 0.5 to 3m, and by Elouardi's (1998) seismic refraction study, which implied a joint soil and epikarst thickness of approximately 2m.

Soil Classification (French System)	Description	pH
A	Organic-rich (> 0.5 g / 100 g) surficial horizon with incorporated rhizome layer	4.5
E	Clay-depleated (<10%) silty brown soil	4
BT	Clay-enriched (>10%) silty brown soil	5.5
IC	Chemically degraded limestone bedrock	8

Table 1: Presents descriptions of the observed VCB soil horizons and their corresponding French classification.

Observed VCB soils can be subdivided into two mineral systems. When thick accumulations of loess deposit directly over denuded limestone bedrock, acidic rainwater is able to mobilize and redistribute clay particles within the loess (Gobat, 2011). This only occurs in locations with shallow rooting species, such as the ryegrass found at the site. These initially, vertically-distributed loess clays are put into solution with the infiltrating acidified rainwater and migrate downwards within the soil column until they come into contact with the basic limestone. Once in contact with the higher pH, the clay precipitates out to form an accumulation horizon. This secondary clay deposit and the overlying silt layer (now clay-poor) make up the first (allochthonous) mineral system (Figure 2). The underlying IC layer consists of clay-sized particles of chemically weather limestone that grade with depth to include an increasing fraction of limestone gravel and cobbles. The IC layer is the second mineral system, autochthonous, and is also considered the epikarst's upper boundary. If loess accumulation is less than 40cm or if the location is vegetated by deeper rooting plant species, the limestone derived basicity is vertically redistributed within the soil profile by plant uptake and degradation, deterring remobilization of clays and consequently preventing the formation of the secondary clay accumulation horizon (Havlicek, 1999). If loess is calcareous in origin, this evolutionary soil paradigm is not applicable. The VCB soil configuration is one of four soil structures seen throughout the Jura range and should be expected in a karstified region blanketed by crystalline-sourced aeolian loess, an arrangement not uncommon throughout central Europe.

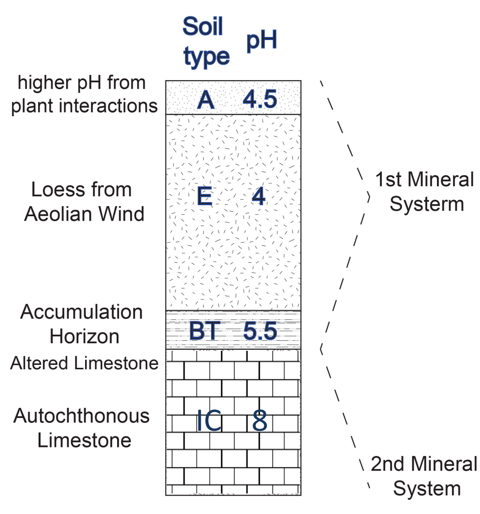

Figure 2: VCB soil structure and composition. The left letter column (A, E, BT, and IC) indicates the type of soil while the right number column (4.5, 4, 5.5, and 8) corresponds to pH values.

Methods

To discern the storage and transfer mechanisms of recharge from snowmelt in a karstic setting, approximately two years of field data, collected throughout a vertical profile within the VCB site, were qualitatively and quantitatively assessed for trends in water volume flux. Continuous measurement of snow height and time discrete sampling of snow density tracked water storage in the snow layer. Water storage in the soil was characterized based on soil moisture measurements up to a depth of 90cm across the two soil layers. Recharge was quantified by measuring discharge at the cave roof drainage point VCB1. By calculating a water balance, chances in soil water content were related to recharge, which provides some indirect insight into storage at deeper locations not accessible by measurements. Finally storage in the phreactic zones were evaluated by comparing discharge at VCB1 with discharge at the spring, especially under low flow conditions. Through these measures, we assessed for transmission and storage mechanisms for a karstic aquifer, and how these mechanisms may influence winter recharge in a changing climate.

A meteostation and soil moisture sensors were installed within the recharge area of VCB1 (Figure 1). The Pessel iMETOS Pro meteostation recorded air temperature (range of -40°C to +60°C, accuracy of +/- 0.1°C), relative humidity (range of – to 100%, accuracy of 1%), and radiation (range of 0 to 2000 W/m2). Snow height was recorded adjacent to the VCB meteostation by a Sommer USH-8 Ultrasonic Snow-depth sensor (range of 0 to 8 m, accuracy of +/- 1 cm). Decagon 5TE sensors, installed in a semi-vertical profile at 10, 25, 40, 55, 70 and 90 cm below ground surface (b.g.s.) within virgin soils recorded soil temperature (range of -40 to 50°C, accuracy of +/- 1°C) and volumetric water content (accuracy of +/- 3%). Data were recorded hourly between November 16, 2011 and May 16, 2013.

In the cave, a V-shaped 5.5m long PVC collection device mounted to the roof funneled VCB1 discharge water to a 1.4m vertical PVC pipe. Water stage within the pipe was recorded with a pressure transduce and correlated to the discharge rate via manual discharge measurements collected bi-weekly between 2010 to 2012. A WTW TertaCon 96A electrical conductivity (range of 0 to 199.9 µS/cm, accuracy of ≤ 0.5 %) and temperature (range of -5°C to +50°C, accuracy of ≤0.1°C) sensor measured exfiltrating cave water hourly between November 16, 2011 and May 16, 2013.

For the Areuse spring (755 m a.s.l., 532980, 195880 UPS), discharge values, recorded by the Swiss Federal Office for the Environment, were used.

Randomly throughout the VCB1 recharge area, weekly snow cores were collected during the 2011/12 and 2012/13 winters using a steel snow-tube and proximal snow heights were measured. Snow core volumes and corresponding masses were used to derive snow-water equivalent (SWE). The snow height to SWE relationship was then used to approximate whether or not snowpack outflow occurred. Snowpack outflow was then directly related to groundwater recharge.

The size of the VCB1 recharge area was identified using a series of isolated summer rain events of varying intensity and duration, as observed in VCB1 hydrograph records. The integrated area (m^3) under each summer-storm event hydrograph was divided by its corresponding total-event precipitation (m), resulting in a recharge area (m^2). Base-flow was subtracted from each event hydrograph prior to calculation and all events were preceded by periods of drought to ensure minimal effects of storage.

Results

Temporal evolution of meteorological and hydrological parameters

Observed parameters at the VCB surface and within its vadose zone, at VCB1 and the Areuse Spring changed considerably throughout both the 2011/12 and 2012/13 winters, respectively identified as the 2012 and 2013 winter seasons (Figure 3). To simplify the data presentation and discussion, the time series data are segregated into periods during which data showed distinctive trends and are annotation as Period A, B, etc. followed by either a 12 or 13 to indicate the study year (Figure 3). As data summary for these periods is presented in Table 2.

	2012		2013	
Snow Accumulation Length (days)	119		146	
Total Snow Accumulation (cm)	97		103	
	Precipitation (cm)	**Total VCB1 Discharge (cm)**	**Precipitation (cm)**	**Total VCB1 Discharge (cm)**
Period A	0	0.2	0	1
Period B	43	33	55	47
Period C	5	17	21	27
Period D	22	27	24	23

Table 2: Presents the snow cover duration and maximum snow depth for 2012 and 2013 in addition to total accumulative precipitation and VCB1 discharge for each period.

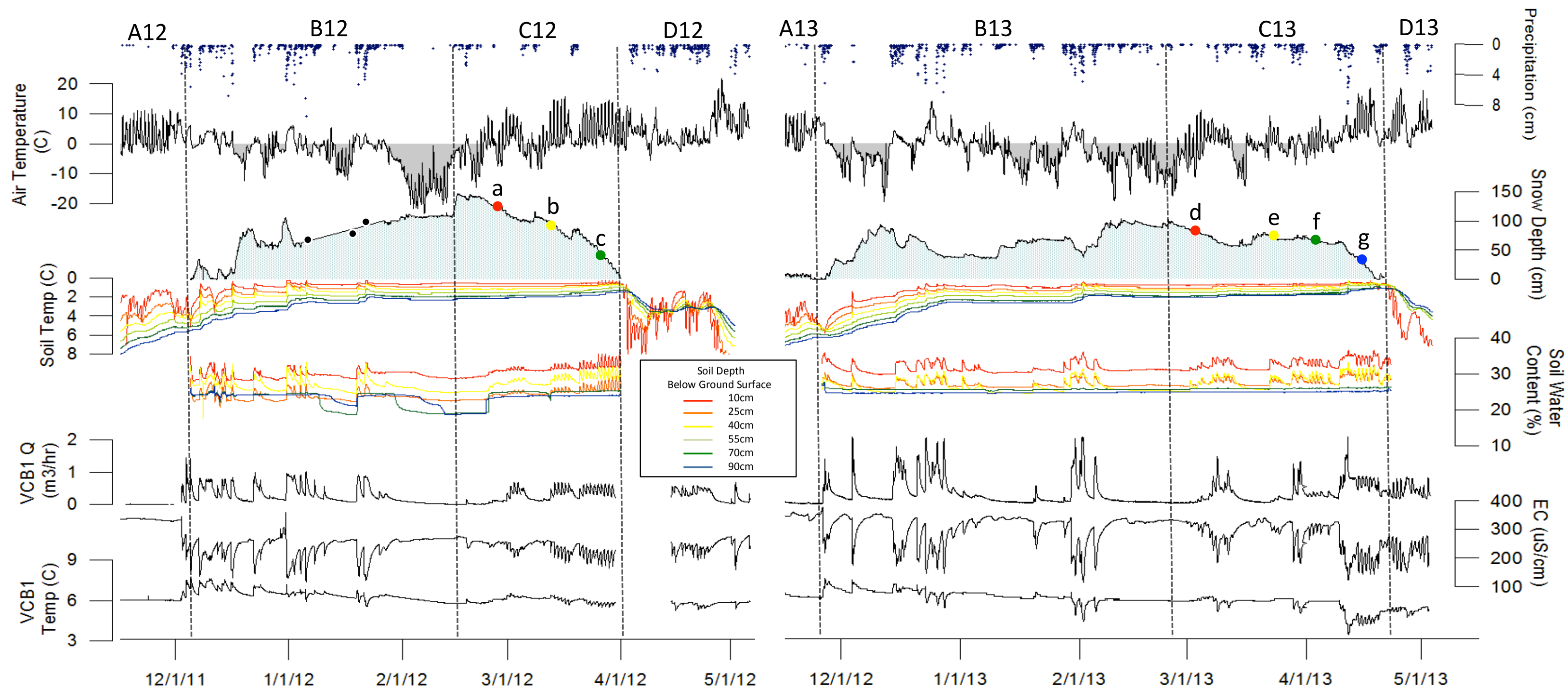

Figure 3: Time series data for precipitation, air temperature, snow height, soil temperature, soil water content, and VCB1's discharge, electrical conductivity and temperature for the winter seasons of 2011/12 and 2012/13. Time series data is segregated into periods during which data showed distinctive trends and are annotated as Period A, B, etc. followed by either a 12 or 13 to indicate the study year.

The snow accumulation period (Figure 3) was approximately 20% shorter in 2012 (119 days) than in 2013 (146 days), but only slightly less snow accumulated in 2012 (97cm) compared to 2013 (103cm). Winter 2012 was characterized by an intense and dry cold period of three weeks at the end of the snow accumulation phase (B12). In winter 2013, cold periods tended to be shorter and less intense. During both winter seasons, several thaw events occurred during the snow accumulation phases B12 and B13 often accompanied by rain (Figure 3). The 2012 winter's melt season (C12) had 3 distinct phases (a, b and c) during which an additional 5cm of precipitation fell and 16cm drained from VCB1. C13 was a wet period with an additional 30cm of precipitation and was broken up into 4 snowmelt phases (d, e, f and g), which occurred more gradually and over a longer timeframe than C12's.

In both years, soil temperatures were above zero when snow accumulation started and remained positive throughout the snow accumulation and melt periods (Figure 3). Frost tubes confirmed the absence of soil frost. Throughout the accumulation and melt period, soil temperatures increased with depth. After the disappearance of the snow cover, the temperature gradient inverted to increasing temperatures with depth. The soil moisture content showed a distinctly different pattern in the upper and lower mineral system. Probes in the upper system (10, 25 and 40cm) reacted to rain events during phases B12 and B13 and showed diurnal fluctuations during the snowmelt phases (C12 and C13). In contrast probes in the lower system (70 and 90cm) showed stable values except for the intense cold periods in 2012. During these periods the moisture content dropped first in 70cm and subsequently also in 90cm. As indicated by flow rates at VCB1 (Figure 3), recharge events occurred not only during snow-melt (B13 and C13) but also during the snow accumulation phase (B12 and C12). During the latter, recharge events were usually associated with rain-on-snow events or days with average positive air temperatures. In this period, the total outflow amounted to 74% (2012) and 86% (2013), respectively of the total precipitation.

VCB1 event hydrograph shapes depended on whether or not the event occurred during an accumulation or melt stage (Figure 3). During snow accumulation periods, infiltration events presented as sharp rising limbs, a distinct peak discharge followed by rapidly declined recessional limbs with a substantial tailing. Even during the longest period with sub-zero air temperatures (end of B12), VCB1 outflow never ceased, demonstrating that storage in soil and/or epikarst is sufficiently high to sustain outflow over several weeks. Hydrographs during the snowmelt period (C12 and C13) were somewhat different, reflecting much drier conditions in C12 (5cm precipitation) compared to C13 (30cm precipitation). Hydrograph events associate purely with melt water have a more Gaussian shape, with recessional and rising limbs having similar aspects and durations. Period C12 is characterized by diurnal discharge variations typical for snowmelt while in C13, sharp discharge peaks associated with rain fall are superimposed on the snowmelt pattern. During both the snow accumulation and melt phases, electrical conductivity (EC) at VCB1 dropped during each outflow event, suggesting that during all flow events freshly infiltrated water reached VCB1, not only "old" stored water from previous events. During high flow events, early in the snow accumulation phase, water temperatures at VCB1 increased likely as a result of the higher temperatures in deeper soil zones while in later periods temperatures drops reflecting colder soil temperatures. Much of the discussion below focuses on the 2012 winter with its pronounced drought period and a snow-melt period with little perturbation by rain. This period is particularly well suited to evaluate storage, because during periods of dry, freezing temperatures, discharge can be equated to changes in storage in following with the fundamental water budget equation.

Dynamics of water content in soil and epikarst

In order to assess in more detail how infiltrating snowmelt is transmitted and stored in the soils and upper epikarst, soil moisture profiles for the upper 90cm of the vadose zone were constructed for selected days of the snowmelt periods (Figure 4). The selected dates cover different stages of the melt phase of the snowpack as indicated in Figure 3 (a, b and c for 2012; d, e, f and g for 2013). The profiles represent the amplitude of soil moisture variation within a 24-hour period, with the left-most line indicating the minimum and the right-most line denoting the maximum soil water content. For both the 2012 and 2013 winters, the daily minimum and maximum moisture lines became increasingly divergent in the upper 55cm of soil as the stages of melt progressed, indicating an increase in water input from the overlying melting snowpack as melt advanced. Additionally, the average daily soil moisture content increased with each stage of melt, particularly between the 25 and 55cm depths. Saturation increased with time at the base of the upper mineral system, just above the pedological contact (55cm) with the lower mineral system, suggesting the formation of a perched water lens. Soil moisture content remained stable below 55cm, indicating that the daily influx of melt water observed in the overlying layers had minimal influence on the saturation of the lower mineral system (Figure 4). Moisture content in the upper epikarst was essentially consistent with each successive melt phase. The overall constancy of soil moisture in the calcareous clays implies that saturation in the lower mineral system (upper epikarst) had been reached over winter's first half, which favored the formation of an overlying perched lens of melt water.

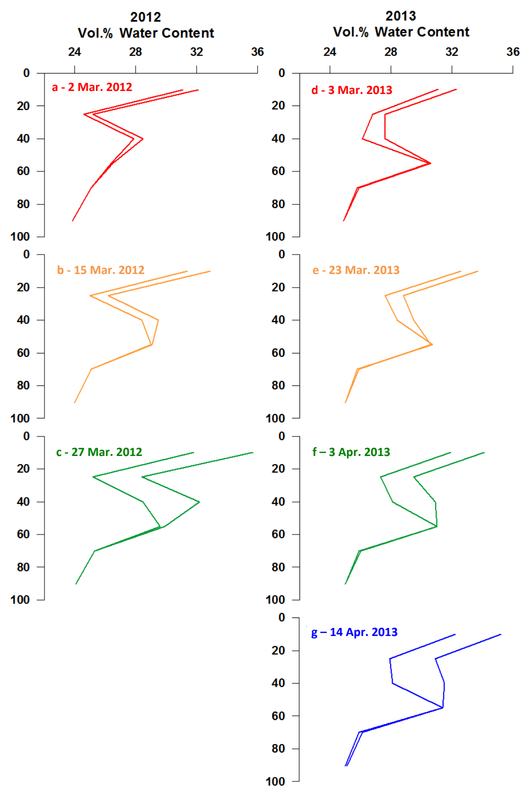

Figure 4: Soil moisture profiles for the upper 90cm of the VCB vadose zone, during the 3 melt stages of 2012 and the 4 melt stages of 2013.

To evaluate if the changes in soil moisture corresponded to significant water volumes, the temporal changes in soil water amount (cm) in the different layers were quantified and related to the SWE and outflow rates (Figure 5). This 2012 time segment (Period B12 and C12, Figure 3) was selected because it best represents the overall storage dynamics due to two prominent phenomena; a inter-winter mixed precipitation event followed by a dry spell, where VCB1 reaches winter low flow conditions (B12); and a multi-stepped melt phase (C12). Figure 5 displays the temporal evolution of dynamic storage of water (cm) throughout the snow and soil profile in addition to the system's input (precipitation) and output (VCB1 discharge). Between 1/25/12 and 2/17/12, the cold spell which lacked precipitation, the 0-40cm layer's water content decreased by 2cm, the 40 to 70cm layer's water content decreased by 2.7cm, and the 70-90cm layer's water content decreased by 1.1cm, totaling 5.8cm of water collectively drained from the soils. 5.5cm of water arrived at VCB1 during this same timeframe. Hence, VCB1 outflow roughly balanced water draining from the soils in approximately 21 days, which verifies our conceptual model of the site as an oversized real world lysimeter. During this same period the water content in the clay-dominant lower mineral system showed very little reactivity to diurnal infiltration fronts, with limited drainage occurring only after two weeks without precipitation (Figure 5). Due to their high water hold capacity, the clayey soils of the lower mineral system stored 1.5cm of water for 21 days, implicating the lower mineral system as a source for base flow waters. During the three phases of melt, the 0 to 40cm segment revealed itself to be a layer of flow-through with pronounced diurnal water-fluxes. Daily water storage changes of 0.5cm during the first stage of melt in early march, increased to 1.5cm during the final stage of melt in late march (Figure 5). Also the perched lens of snowmelt in the upper mineral system drained over a 6-day period after snowmelt had finished.

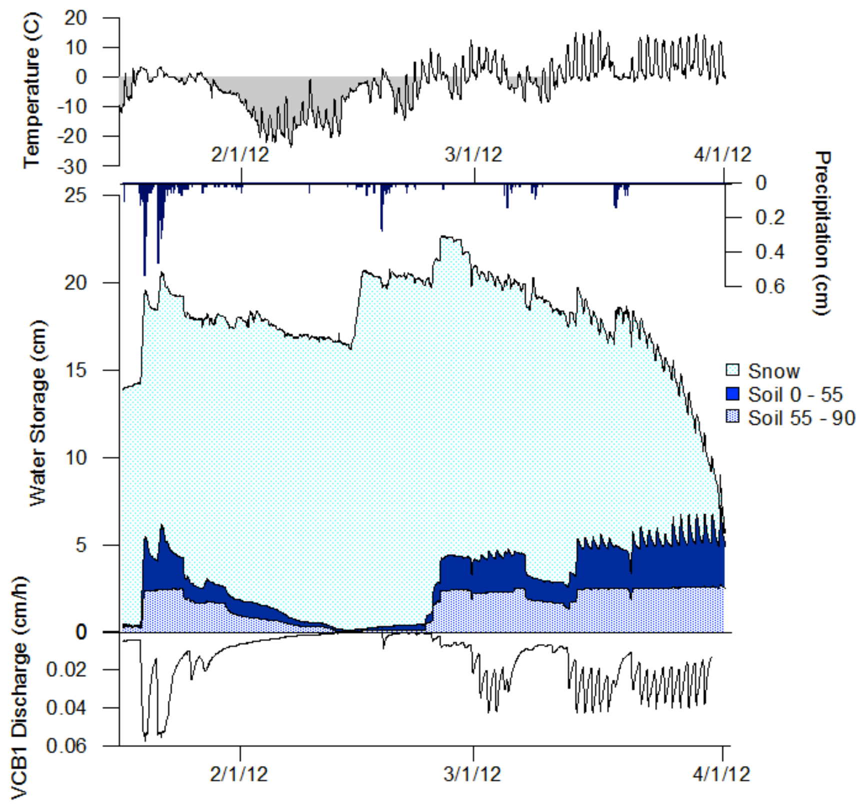

Figure 5: VCB water balance showing precipitation (system input), snow water equivalent, soil water storage and VCB1 discharge (system output).

Relationship between cave and spring discharge

The significance of these small-scale, shallow hydraulic processes is evaluated by comparing the VCB1 discharge with that of the Areuse Spring, the watershed's discharge point (Figure 6). The 2012 winter is particularly well suited to evaluate storage relationships between the VCB1 and Areuse Spring, because as mentioned, discharge can be equated to changes in storage due to subfreezing temperatures and lack of precipitation. Discharge at VCB1 and Areuse Spring were made comparable by normalizing them to their respective catchment areas (Figure 6). The normalized VCB1 and Areuse spring discharges agree surprisingly well despite the large difference in catchments size and despite relying on measurements from "opposite ends" of the groundwater flow systems. Both hydrographs show sharper peaks for events before January, likely due to rain-on-snow, which releases larger quantities of water quickly compared to the more rounded peaks observed during the snowmelt periods. However, some differences in flow between the two scales can be observed.

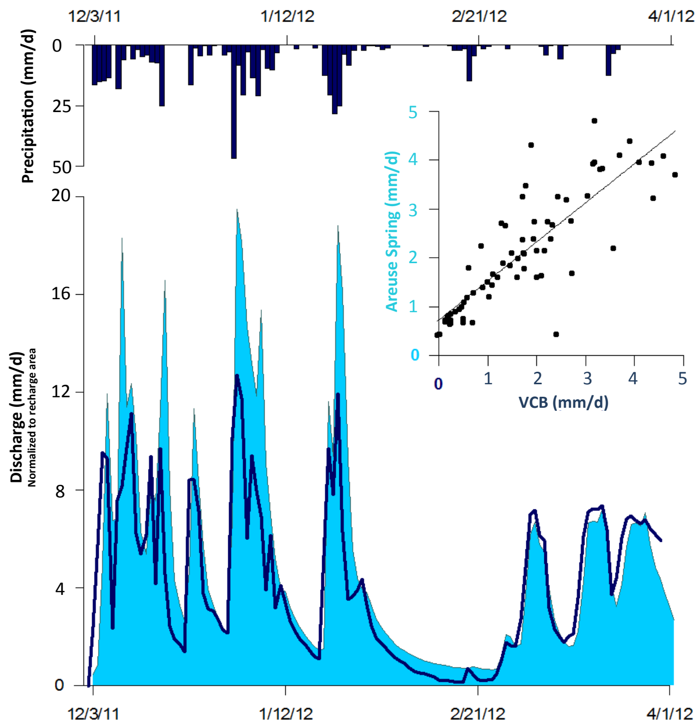

Figure 6: The bottom graph depicts precipitation along with the synchronous evolution of VCB1 (dark blue line) and Areuse Spring (light blue fill) discharge (normalized to respective recharge areas) during 2012. The enclosed bivariate graph shows the approximately linear relationship between the two monitoring points.

Discussion

The mechanisms of infiltration were identified via critical study of observed vadose zone water fluxes (soils, upper epikarst and VCB1) in their peodological and geological context. Water fluxes shown in the soil moisture profiles (Figure 4) and the 2012 water balance (Figure 5) clearly indicate that melting snow infiltrates into the vadose zone throughout most of the winter and passes through the coarser upper soils to the underlying clays. Once the clay layer is saturated, when infiltration exceeds drainage capacity, they serve as a temporary aquitard, promoting the formation of an overlying water lens, the thickness of which relies on the duration of water influx from the melting snow. Said in another way: the greater the number of consecutive days with snowpack outflow, the thicker the perched water lens. The presence of the observed clay layer presents a complication to previous conceptual models of karst aquifer recharge. Infiltrating precipitation was assumed to percolated vertically though the soil to then enter the epikarst where it would store or move laterally, prior to draining into the underlying karstic network. Conceptually, the barrier to downward flow was thought to have been created by the decrease in epikarst permeability (Perrin et al., 2003) or the differences in hydraulic conductivity between the epikarst and the lower unsaturated zone (Trçek, 2007). Findings indicate that in the case of the VCB, it is the soil's clay layer that serves as the impediment to infiltration and not the epikarst. If observed soil stratigraphy were ubiquitous throughout the VCB1's recharge zone, event waters would not pass through the vadose zone for a minimum of 42 hrs; a conservative calculation that presupposes all vadose water must be renewed before event water discharges from VCB1. This time estimation is based on the assumption of a saturated 50cm thickness of silty loam with a hydraulic conductivity (K) of 12.5cm/hr, and a saturated 20cm clayey loam layer with a K of .9cm/hr (Clapp and Hornberger, 1978). That being the case, waters still transits the 53 vertical meters to arrive at VCB1 within 2 hrs, much more swiftly than the calculation would indicate. It could be argued that piston flow (Trçek, 2002) induced the rapid arrival of event water at VCB1. However, this rapid water arrival occurs even at the end of the mid-winter cold spell (2/20/12, Figure 5), a time in which the perched lens was very thin and a relatively low hydraulic head existed. Hence, some portion of event water must flow around the soil's clay layer to quickly arrive deeper in the karstic system, a theory further supported by the close and inverse relationship between VCB1's discharge and electrical conductivity. Thus, while direct recharge points, such as swallow hole, were not identified within the recharge area, regions of thinner soil lacking a clay accumulation horizon must exist, allowing rapid infiltration. If distributed recharge were uniform, an infiltration front would be represented by peak water content arriving at 10cm b.g.s., followed by peak water content arriving at 25cm b.g.s., followed by peak water content arriving at 40cm, etc. until finally peak discharge at VCB1 at the end of the temporal succession. However, peak VCB1 discharge repeatedly occurred prior to peak water content at the 25, 40, and 55cm soil horizons, imply that preferential flow paths supplying water to VCB1 probably originate at a depth around 25 b.g.s., This supports the hypothesis that shallower soils, lacking a clay accumulation horizon, exist within the VCB1 source area. While previous investigators (Friederich and Smart, 1982; Perrin et al., 2003; Williams, 1983) implicated the epikarst as the medium in which shallow lateral flow occurs, it appears that lateral flow within soils overlying the epikarst could, in some cases, be more important.

Some degree of lateral flow within the silty loam of the upper mineral system, toward thinner soils lacking an accumulation horizon, most likely occurs year round. However, as the perched lens of snowmelt water increase during the spring snowmelt periods (C12 and C13, Figure 4), the hydraulic gradient towards these infiltration points would increase. Thus, perched lateral flow may become more relevant to vadose zone hydraulics during extended periods of infiltration associated with the spring melt of a snowpack, increasing the relative importance of the concentrated flow at isolated locations.

The interpretations of the VCB and Areuse Spring relationship results are discussed here. During rain-influenced events, peak flow at the spring tends to be higher possibly due to activation of rapid, preferential flow-paths during extensive rain-on-snow events e.g. via dolines present throughout the watershed. The overall good agreement between the two discharge patterns suggests that the soil/epikarst infiltration system has a strong influence on the spring discharge pattern.

As mechanisms of water storage are of main interest in this study, the discharge rates during recession periods were compared in more detail for low flow periods (<5 mm/d) not influenced by recent infiltration at a daily time step (Figure 6). A linear relationship, nearly 1:1, between discharge fluxes was observed indicating that drainage of water from soil/epikarst provides a significant contribution to spring discharge. When extrapolated towards zero, the linear relationship does not pass through the origin of the plot, i.e. as discharge rates approach zero at VCB1, the discharge rate at the spring approaches a value of 0.54mm/d. This suggests that the deep phreatic zone provides a steady base flow component on which recharge from the soil/epikarst zone is superimposed.

Indeed, the average Q347 (discharge exceeded during 95% of the days of the year) and the average NM7Q (lowest 7-day flow average for a year) for a 52 year period (1959-2010) having respective values of 0.57mm/d and 0.49mm/d (data from Swiss low flow database), correspond very well to the base flow component estimated from Figure 6. Compared to this base flow value, the dynamic water storage in the soil/epikarst of about

50mm is significant, corresponding to about 90 days of phreatic zone base flow. This storage volume is mainly relevant in providing water to the spring at the time scale of up to several weeks, while the storage time scale of the phreatic zone is likely rather months or years. Overall, this signifies that vadose hydraulics play a governing role in karst aquifer behavior as posited by Trcek (2007).

A critical evaluation of the potential sources of uncertainly associated with the selected methods was made. As is typical in karstic regions, catchment area can be reliant on volume and intensity of a given precipitation event. Thus it was not surprising that VCB1 recharge area was shown to increase exponentially between 700m^2 and 1600m^2, with total water volume of a given summer precipitation event. The complexity of time-variant recharge area, such as that of the VCB, has been well studied by the likes of Ravbar et al. (2011) and Hartmann et al. (2013). The former researchers identified that anomalous specific electrical conductivity at karst springs can result from variable catchment boundaries, while the latter group developed a calibration approach that incorporates identification of variable recharge area for predictive modeling. While the recharge area for the VCB system does fluctuate, 80% of studied summer precipitation events indicate a recharge area above 1000m^2. Also, for smaller precipitation events preceded by drought, a larger proportion of a given events infiltration may have gone toward satiating soil moisture deficits, thereby implying the recharge area to be smaller than it actually is. Further, variable recharge area has not been studied in the context of snowmelt and as such it seemed conservative to use the maximum recharge area 1600m^2 for water balance calculations given that the volume of infiltration associated with snowmelt is in the same order of infiltrating water associated with large summer precipitation events. Had a smaller recharge are been used for water balance calculations, the agreement between the normalized discharges for the VCB and Areuse Spring would not have agreed as well, even though the similarities in trend would have maintained. Advanced modeling of the VCB's recharge area was not in the scope of study and may be considered in future modeling efforts associated with this system.

A second source of uncertainty may have arisen from the method selected to approximate the temporal evolution of snowmelt infiltration. As indicated, snow depth data collected and remotely transmitted by the Sommer Sensor were used to estimate SWE, based on the seasonally averaged relationships between field measured snow depth and snow density. Monthly snow courses during the 2013 winter, showed snow depth ranged between 13% and 16% across the recharge area at a given time. Given the remote location of the site, it was not feasible to conduct snow courses at a finer time resolution. In contrast to alpine regions (Jonas, 2009), the relationship between snow density and height did not show a systematic seasonal trend likely due to many episodic accumulation/melt cycles at this lower altitude. While snow compaction and blowing snow can certainly result in snow-height reduction, it was reasonably assumed that reductions in snow height, averaged across the season, related to a loss of snowpack SWE. This broad assertion was validated through analysis of our soil moisture data, where reductions in snow height corresponded to increases in soil water content. Snowpack loss due to sublimation was assumed insignificant due to high ambient air humidity averaged across the winter seasons. Reductions in snow height were equated to snowpack out-flow, and in turn equated directly to infiltrating recharge, which was appropriate for two reasons. Firstly, the recharge zone was flat, with no surface runoff ever observed during summer months. Secondly, methylene-blue frost tubes, which extended 30cm b.g.s. at the VCB, did not at any point indicate the presence of soil frost during snow cover for both studies winters.

Since soil moisture can be variable in space, a confirmatory profile of soil moisture sensors was emplaced at the site within the upper 90cm of soil and epikarst approximately 5m from the metoostation. Confirmatory data showed synchronous, volumetrically comparative trends in soil moisture as the primary 5TE sensors previously discussed. That said, deviations from the measured soil moisture data should be anticipated as natural soils are rarely homogeneous throughout a watershed.

Summary and Conclusion

We investigated the transit and storage of snowmelt water through the vadose and phreatic zones of a karst aquifer. Although vadose zone flow showed diurnal patterns during snowmelt, suggesting a tight coupling between melt events and recharge, a substantial amount of water was stored in the vadose zone. Such storage led to a temporal redistribution of water from melt events to cold periods lacking snowmelt infiltration. As suggested by soil moisture time series, water storage probably occurred in the soil rather than the epikarst. The soil structure, consisting of a permeable layer overlying clay, likely favored the formation of superficial perched lenses. The importance of soil water storage was confirmed by water balance calculations that showed a good agreement between soil water storage loss and vadose zone outflow during a cold period lacking melt infiltration. Such superficial water storage is likely more relevant for vadose zone flow in winter as no evapotranspiration occurs due to the snow cover and cold temperatures. While the soil layer seems to have a buffering effect on meltwater inputs, surprisingly little attenuation of water flow occurred between the vadose zone and the spring. Normalized hydrographs for these two discharge points agreed well, expect that shorter-term variations observed at the spring were superimposed on a baseflow component that likely originated from the aquifer's phreatic zone. Hence vadose zone storage and flow has a strong control on aquifer discharge at the scale of weeks, while phreatic storage becomes dominant during prolonged periods without input. The strong coupling of recharge and discharge underlines the importance of understanding recharge mechanisms when attempting to predict future groundwater availability from karst aquifers under changing climatic conditions. Soil water storage might have a larger influence on discharge at karst springs than previously assumed, especially during winter when evapotranspiration is absent. At our site, the storage mechanism is strongly associated with genesis of the soil i.e. the deposition of siliceous loess on top of the calcareous bedrock. Such a configuration might have been commonly formed in Europe and North America after the last glaciation. As such, identification of soil type and distribution should be an integral part of karst aquifer assessments, particularly in regions that receive seasonal snowfall. This study further identifies the importance of maintaining soil health in karst watershed, as extended storage of perched water in soils may allow for extended chemical exchange with soil constituents, altering water quality. Henceforth, the influence of such soils types on the behavior of karst aquifers deserves further attention.

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Chapter 3

Infiltration under snow cover: Modeling approaches and predictive uncertainty

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Abstract: Groundwater recharge from snowmelt represents a temporal redistribution of precipitation. This is extremely important because the rate and timing of snowpack drainage has substantial consequences to aquifer recharge patterns, which in turn affect groundwater availability throughout the rest of the year. The modeling methods developed to estimate drainage from a snowpack, which typically rely on temporally dense point-measurements or temporally-limited spatially-dispersed calibration data, range in complexity from the simple degree-day method to more complex and physically-based energy balance approaches. While the gamut of snowmelt models are routinely used to aid in water resource management, a comparison of snowmelt models' predictive uncertainties had previously not been done. Therefore, we established a snowmelt model calibration dataset that is both temporally dense and represents the integrated snowmelt infiltration signal for the Vers Chez le Brandt research catchment, which functions as a rather unique natural lysimeter. We then evaluated the uncertainty associated with the degree-day, a modified degree-day and energy balance snowmelt model predictions using the null space Monte Carlo approach. All three melt models underestimate total snowpack drainage, underestimate the rate of early and midwinter drainage and overestimate spring snowmelt rates. The actual rate of snowpack water loss is more constant over the course of the entire winter season than the snowmelt models would imply, indicating that mid-winter melt can contribute as significantly as springtime snowmelt to groundwater recharge in low alpine settings. Further, actual groundwater recharge could be between 2 and 31% greater than snowmelt models suggest, over the total winter season. This study shows that snowmelt model predictions can have considerable uncertainty, which may be reduced by the inclusion of more data that allows for the use of more complex approaches such as the energy balance method. Further, our study demonstrated that an uncertainty analysis of model predictions is easily accomplished due to the low computational demand of the models and efficient calibration software and is absolutely worth the additional investment. Lastly, development of a systematic instrumentation that evaluates the distributed, temporal evolution of snowpack drainage is vital for optimal understanding and management of cold-climate hydrologic systems.

Key Words: uncertainty, snowmelt, energy balance, day degree, recharge, karst, groundwater

Introduction

Infiltration resulting from snowmelt represents the temporal redistribution of liquid precipitation. This is extremely important because the rate and timing of snowpack drainage has substantial consequences to aquifer recharge patterns, which in turn affect groundwater availability throughout the rest of the year. In spite of its significance, direct measurement and modeling of snowpack outflow remains challenging due to the inherent limitations of monitoring instrumentation.

A number of field methods have been used to measure water drainage from snow packs (loss of snow water equivalence, SWE) including snow pillows (Archer and Stewart, 1995; Butcher and McManamon, 2011; Trujillo and Molotch, 2011), and snowmelt lysimeters (Jost et al., 2012; Kattelmann, 1989, 2000; Tekeli et al., 2005), both of which can render temporally dense point data. Extrapolation throughout a watershed of point measurements such as these is difficult due to the considerable spatial variability that exists in both snow depth and corresponding SWE and heterogeneous infiltration processes resulting from different soil types and structures across a watershed. Further, snow lysimeters have structural configurations that impose bias to the output data, such as sidewalls which are used to mitigate gains or losses from lateral flow within a snowpack (Haupt, 1969; Martinec, 1986). Snow pillow data can also be skewed due to snow bridging. Snow courses (Marks et al., 2001; Rice and Bales, 2010) produce a more distributed understanding of SWE, however they are highly laborious and are typically done at a coarse time resolution. Assessment of SWE evolution is further complicated when considering that spatial variability in recharge from snowmelt also results from irregularity in the amount of water released from the base of the snowpack. This ensues from complicated, preferential pathways in which melt water travels through a snowpack before percolating to the base (Kattelmann, 1989). Ultimately, snow hydrologists still must rely on limited and possibly biased field data to obtain basic liquid inputs for snowmelt modeling (DeWalle and Rango, 2008).

Numerous modeling methods have been developed to evaluate snow processes, with complexities ranging between simple index models and physically based multi-layer models which simulate a snowpack's energy balance (Etchevers et al., 2004). The ongoing debate regarding the relative merits of these modeling end members (Franz et al.

2008) has manifested in several model inter-comparisons (Feng et al., 2008; Magnusson et al., 2011; Rutter et al., 2009). In its simplest form, the degree-day (DD) method of modeling snowmelt is based on the assumption that snowmelt during a time interval is proportional to positive air temperature, with the proportionality factor being the degree-day factor C (Hock 1999), an association first presented in 1943 (Linsley). The relative contributions of the different energy balance components can shift in space and time affecting the parameter C. These changes include cloud cover, snowpack conditions, shift in season or progression of day, aspect, slope and vegetation cover (Hock, 2003). That withstanding, Ohmara (2001) was able to show the computational validity of melt rate parameterization using air temperature, and that the degree-day method "works" because temperature information is transferred to earth's surface mainly through long wave atmospheric radiation, which is by far the most important heat source for melt. Several studies have demonstrated improvements to the DD method via incorporation of solar radiation (Hock, 1999, 2003; Jost et al., 2012) and progression of day (Tobin et al., 2013). Overall though, the efficacy of this index method is usually attributed to the way in which air temperature effectively integrates the influence of a range of meteorological variables, or energy fluxes (Hodgkins et al., 2012). Acquiring air temperature data is relatively easy and inexpensive. In contrast, more rigorous energy balance models are data intensive and usually require expensive instrumentation. At a minimum, physically based assessments take into account air temperature, relative humidity, wind speed, precipitation, global and incoming long wave radiation. With this breadth of information researchers can explicitly model changes in heat storage of a snowpack and solve for snow surface temperature using a heat budget formula (Jost et al., 2012), thereby more concisely modeling accumulation and ablation. The physics behind the energy balance method has been well documented (Anderson, 1968; Cline, 1995; Herrero et al., 2009; Male and Gray, 1981; Marks and Dozier, 1992). An exhaustive overview of snow models is presented by Yang (2008) and updated regularly on the Snow Modelers Internet Platform.

Choice of modeling method is in part dictated by data and computational availability. The empirical degree-day method requires little data and is easily applied in distributed modeling efforts, but does not explicitly take into consideration climatic forcing functions operating during snow accumulation and ablation. In contrast, the computationally intensive physically-based energy balance methods offers more insight into the processes controlling the energy balance (Hodgkins et al., 2012) but requires vast amounts of data, which in consequence hinders distributed application, needed for up-scaling of point-processes. Further, uncertainty may be introduced when adopted model parameters are unknown. Thus, to some degree, these modeling end members serve different needs within the modeling community.

Most numerical models are employed to aid in environmental management, and as such the uncertainty associated with predictions made by such models must be assessed (Gallagher and Doherty, 2007; Jost et al., 2012). However, given the issues with the above-discussed field methods for collection of calibration data and the lack of data for comparison, it has been difficult to quantify 1. to what extent these branches of snowmelt models provide robust estimates of snowpack outflow and 2. how well these models perform at different time scales. That said initial attempts on this front have been made. Seibert (1997) examined parameter uncertainty within the HBV model using a Monte Carlo approach. Since ranges in parameters can provide an almost equally good model fit, Seibert concluded that model predictions should be given a probability distribution rather than a single value, which is in keeping with assertions made by Melching et al. (1990) and Beven and Bingley (1992). Franz et al. (2010) applied the Bayesian Model Averaging (BMA) method to an ensemble of twelve snow models, that varied in their heat and melt algorithms, parameterization, and/or albedo estimation method, to quantify the uncertainty associated with these sources of error in the stream flow forecasting process associated with snowmelt. Here the individual models BMA predictive mean, and BMA predictive variance were evaluated. An individual snow model would often outperform the BMA predictive mean. However, observed snow water equivalent was captured within the 95% confidence intervals of the BMA variance on average 80% of the time. Franz et al. concluded that consideration of multiple snow structures would provide useful uncertainty information for probabilistic hydrologic prediction. Slater et al. (2013) investigated uncertainty surrounding SWE reconstruction, when using remote sensing, and found that errors in model forcing data were at least as important, if not more so, than image availability when reconstructing SWE. Even though a few isolated studies have look at uncertainties surround snow processes models, uncertainty assessment of model performance is not routinely quantified for recharge estimates associated with snowmelt. So far, there have been no systematic comparisons of the uncertainties arising from different snowmelt modeling approaches at either the parametric or structural levels.

This paper presents a comparison of three snow-process models' ability to predict recharge from snowmelt and a short discussion pertaining to the application of these results at different temporal scales. This study was not intended to be an exhaustive analysis of either parameters nor model structure uncertainty but rather help shed light on how well snow process models are able to predict recharge, either at the event or seasonal scale. To generate snowmelt model calibration data, we used a large and natural lysimeter as proposed by Kattelmann (2000). This researcher stated that snowmelt runoff from a larger "natural lysimeter", a well defined catchment with an easily-monitored drainage point, would provide a conceptually better basis for evaluating output from snowmelt models than the somewhat artificial sampling of snowpack outflow by lysimeters and snow pillows. In following, the karstified Vers Chez les Brant (VCB), which can be viewed as an oversized, real-world lysimeter, consists of a $1600m^2$ watershed that drains infiltrating water to a cave discharge point (VCB1) 53m below the ground surface (Meeks and Hunkeler, 2015). We used this rather unique natural lysimeter to evaluate the uncertainty surrounding modeled snowmelt predictions. We used a simple, albeit physically based vadose zone model, to back calculate snowmelt from the observed cave drainage. The back-calculated snowmelt does not retain any of the aforementioned data biases imposed by traditional lysimeters or snow pillows, has a fine time resolution, and represents the integrated behavior of snowmelt across the VCB recharge zone. This back-calculated snowmelt data was then used as a point of comparison for the snow process models' predictions of time-series snowmelt data. The uncertainty associated with each model's recharge prediction was then systematically quantified through a rigorous calibration process, something that had yet to be undertaken by the snow modeling community. A visual workflow of these modeling efforts is included as Figure 1.

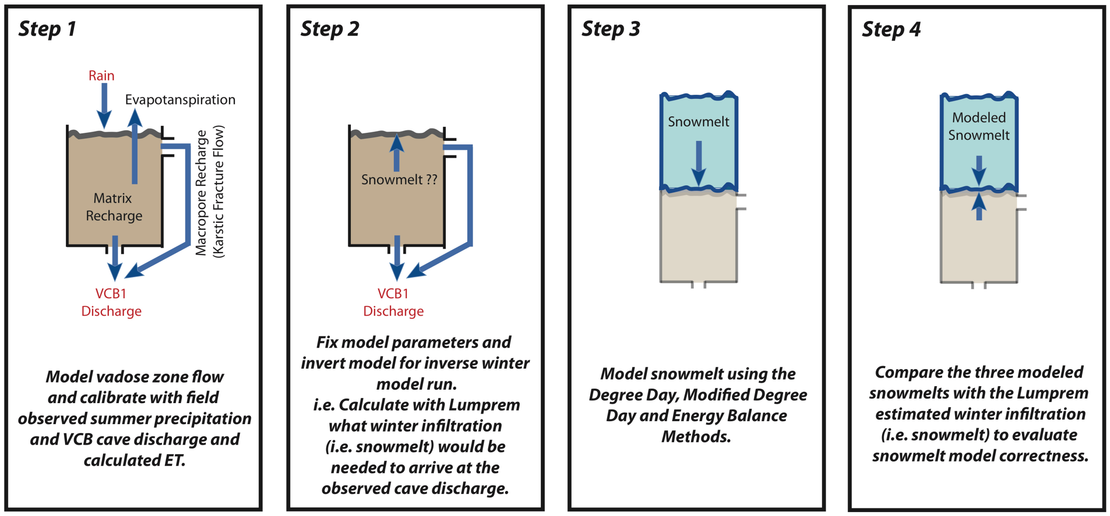

Figure 1. Workflow of the modeling completed using the Vers Chez le Brandt (VCB) calibration data. Each of the four models (Lumprem, degree-day, modified degree-day and energy balance) was subject to a null-space Monte-Carlo calibration.

Study Area

The Vers Chez le Brandt (526450/199010 UPS) research site is situated within the 130km^2 karstified Areuse Catchment in the Swiss Jura Range's western edge (Figure 2). Upper Jurassic (Portlandian, Kimmeridgian and Sequanian) aged marl and fossilferous limestone (Sommaruga, 1997; Valley, 2002) make up the region's bedrock and house the site's single chamber karst cavity. Up to 70cm of Neolovisol loess soils blanket the autochthonous, solutionally altered bedrock (Havlicek, 1999) and are composed of two mineral systems that collectively make up the soil and epikarst, which are approximately 2m in thickness (Elouardi, 1998; Müller, 1978). Based on field observations of VCB pedology and lithology, in conjunction with analysis of precipitation, soil moisture and VCB1 discharge time series data, subsurface flow at the site is routed and or stored through a combination of three pathways. The three flow routing components include the site's upper mineral system composed of silty soils and a clay accumulation horizon, the underlying mineral system made up of clayey soils and the epikarst and lastly a shallow karstic drainage pathway/network originating in the silty-soil horizon. An in depth discussion on the study site's pedology, lithology, karstification, and recharge zone determination has been presented by Meeks and Hunkeler (2015). As demonstrated by Meeks and Hunkeler (2015), shallow vadose zone processes can have a governing role in karst aquifer dynamics, as is the case at the VCB and Areuse Catchment, and that assessment of infiltration in sub-catchments can shine light on aquifer-scale infiltration, storage and drainage patterns. Further the VCB site aligns with Dewall and Rango's (2008) suggested optimal conditions for direct percolation of melt water into the subsurface: negligible slope, lack of soil frost and permeable soils and strata. With these conditions, runoff was insignificant and the balance of melt water was assumed to infiltrate into the soils and groundwater (Mullem et al., 2004).

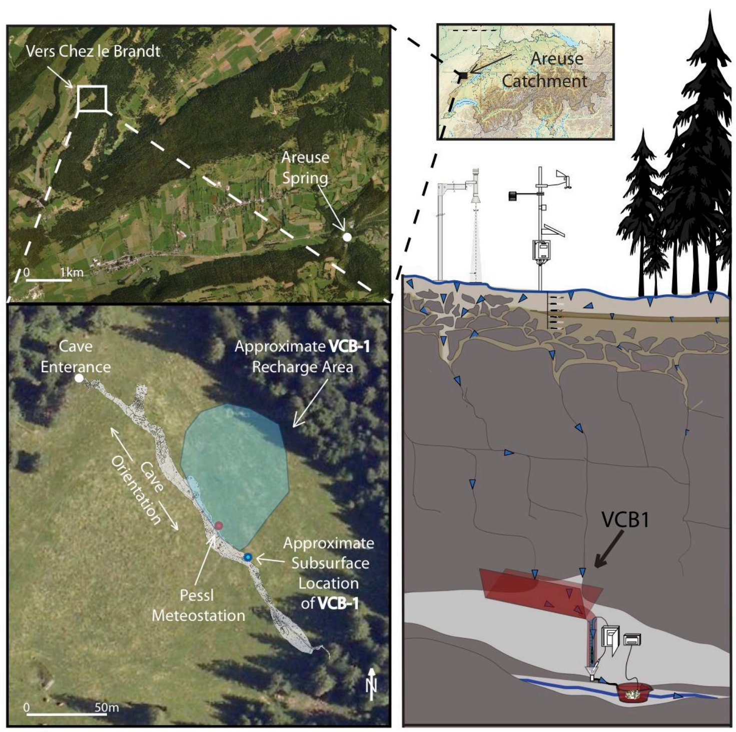

Figure 2. Counterclockwise is the Areuse Catchment location, Vers chez le Brandt (VCB) and Areuse Spring locations, the VCB areal configuration showing the site's underlying cave, and a not-to-scale conceptual model of site cross section with surficial and cave instrumentation.

The VCB receives approximately 1550mm of precipitation annually, 30 to 40% of which falls as snow between the months of December and March (www.meteoswiss.ch). A proximal Swiss Agrometeo weather station in Les Verriers (525500, 199175 UPS; Campbell-CR10x) shows that average summer and winter temperatures for the area are +14^{o}C and -1^{o}C respectively. The VCB catchment is primarily vegetated by cocksfoot and ryegrass species.

The size of the VCB1 catchment area was identified using a series of isolated summer rain events of varying intensity and duration, as observed in VCB1 hydrograph records (Meeks and Hunkeler, 2015). The integrated area (m^3) under each summer-storm event hydrograph was divided by its corresponding total-event precipitation (m), resulting in a recharge area (m^2). It should be noted that karst aquifers are known to have time variant recharge areas (Hartmann, 2013; Hartmann et al., 2012). This effect is not considered here. An in-depth discussion on the validity of using a catchment area of 1600m^2, despite the influences of a time-variant catchment size, is presented in Meeks and Hunkeler (2015).

A series of unpublished VCB tracer tests revealed the approximate VCB1 recharge location to be north and adjacent to the cave's orientation. A 1979 tracer test proved hydraulic connection between the VCB cave and the Areuse Spring, the drainage point of the Areuse watershed (Müller et al., 1982). Meeks and Hunkeler (2015) normalized the VCB1 and Areuse Spring discharges to one another according to their respective catchment areas. These normalized discharges agree surprisingly well despite the large difference in catchments size and despite relying on measurements from ''opposite ends'' of the groundwater flow systems. This hydrograph agreement indicates the relevance of shallow vadose zone processes on karst aquifer dynamics and that the inverse modeling of snowmelt from cave drainage could be extrapolated to the larger watershed scale (Meeks and Hunkeler, 2015).

Data

Between November 16, 2011 and May 16, 2013, VCB air temperature (range of -40°C to 60°C, accuracy of +/- 0.1°C), relative humidity (range of 0 to 100%, accuracy of 1%), global radiation (range of 0 to 2000 W/m^2), wind speed and direction and precipitation were recorded hourly by a Pessl iMETOS Pro meteostation. An attached Sommer USH-8 Ultrasonic Snow-depth sensor (range of 0 to 8 m, accuracy of +/- 1 cm) simultaneously measured snow depth. Methylene-blue frost tubes, which extended 30cm into the VCB soil substrate, did not indicate the presence of soil frost at any point during snow cover. A funneling device mounted to the cave roof, directed VCB1 discharge water to a vertical PVC pipe, within which suspended a pressure transducer measuring water stage. Water

stage was then correlated to discharge volume via manual discharge measurements collected weekly throughout the 2010 to 2013 test period. VCB1 stage was monitored hourly between November 16, 2011 and May 16, 2013 to generate discharge data representing the drainage for the entire VCB catchment. This data was then divided by the recharge area to arrive at point discharge data used in the calibration of the vadose zone model. A detailed depiction of cave instrumentation and overall site configuration can be found in Meeks and Hunkeler (2015). Additionally, weekly snow cores were randomly collected throughout the VCB1 recharge area for later establishment of SWE.
Weekly snow cores were randomly collected throughout the VCB1 recharge area for later establishment of SWE. Manual measurement of SWE, via snow samples collected in snowtubes, served as calibration data for the snowmelt models. While the sampling devices and methods used were "industry standard", immaculately collection of a snow sample is impossible as small amounts of snow were lodged in the uneven and grass-covered ground surface and therefor left unaccounted. More importantly, collected SWE samples may not have reflected the distributed average SWE within the capture zone, as snow packs have a high anisotropy and by very nature snow cores are discreet samples. To overcome this inherent limitation, we collected SWE samples randomly throughout the recharge area and performed a few snow courses to establish the range of SWE values at a given time (on average ~15cm). The later exercise was only performed three times throughout the winter, as the capture zone is only 1600m^2 and repeated snow courses would have greatly disturbed the snowpack in the relatively small recharge zone and possible alter infiltration rates.

Modeling

In the following section, we define the Lumprem model, which serves to generate the dataset of snowmelt infiltration and the point of comparison for the snowmelt models' output. Then we describe the three applied snowmelt models, of varying degrees of mathematical complexity and physical basis, used to simulate snowpack accumulation and ablation and consequent snowpack drainage.

Lumprem

The 1-D unsaturated zone model LUMPREM (LUMPed Parameter REcharge Model) was used to back-calculate infiltration from VCB1 discharge. This semi-physically based, lumped parameter model is capable of providing basic simulation of major unsaturated zone water balance components including rainfall, evapotranspiration, recharge, macro-pore recharge and runoff (Watson et al., 2013). This latter process in the model is turned off for this study given the lack of observed runoff at the VCB site. Lumprem serves as the user and data interface to the subroutine RECHMOD (RECHarge MODel), where all model calculations pertaining to water movement and storage within the vadose zone take place. The model receives field-measured precipitation (cm) and calculated evapotranspiration (cm, ETo) as hourly inputs. Daily ETo was calculated using the FAO56 Penman-Monteith method (Allen and Pruitt, 1991). Input data for ETo calculations included field-measured average solar radiation, minimum and maximum air temperature, minimum and maximum relative humidity, dew point and wind speed for each day. Given that the evapotranspiration rate of plants is dependent on the volume of available water in the unsaturated zone, water loss through evapotranspiration was calculated in Lumprem using the following equation:

$$E = fE_p \frac{1 - e^{-\gamma v'}}{1 - 2e^{-\gamma} + e^{-\gamma v'}} \tag{1}$$

Where E is water loss through evapotranspiration, E_p is potential evapotranspiration (ETo is used in exchange for E_p within this model), f is a crop factor, v' is the relative volume of water in the container (ie. V/V_{max} where V is the current amount of water in the container and V_{max} is the total container volume), and γ is a parameter determining the shape of the evapotranspiration rate versus the stored water relationship.

Recharge, as calculated by Lumprem, is the water lost from the model bucket as a continuous unsaturated vertical flow to the subsurface underlying the root zone. Since hydraulic conductivity of unsaturated material decreases with decreasing saturation, the rate at which water is lost from the bucket depends on the volume of moisture currently stored. In accordance with van Genuchten's (1980) equation, rate of water lost as recharge is expressed by the following equation:

$$R = K_s[v']^l\left[1 - \left(1 - [v']^{l/m}\right)^m\right]^2 \tag{2}$$

Where R is rate of drainage, K_s is saturated hydraulic conductivity, l is the pore-connectivity parameter (estimated by Mualem (1976) to be about 0.5 for many soils), and m is a parameter determining the shape of the drainage rate versus stored water relationship.

Macro-pore recharge (R_m), a phenomena quite common in karstic aquifers, occurs when temporary saturated conditions in the upper subsurface allow water to migrate laterally to zones of preferential downward flow. Lumprem allows for this type of drainage when saturated conditions occur in the soil store. Given that runoff is not present in this system, all overflow of the bucket become macro-pore recharge.

Degree-Day Snowmelt Model

SWE evolution was simulated based on the melt equations governing HBV's (Bergstrom and Singh, 1995; Sten, 1975) snow routine.

$$M = \left(\frac{C}{24}\right) * (T_a - T_t) \tag{3}$$

Where M is melt (mm h^{-1}), C is the degree-day factor (mm $^{o}C^{-1}h^{-1}$), T_a is the mean hourly air temperature (^{o}C), and T_t is the threshold temperature (^{o}C). The model accounts for the inputs and outputs for both the liquid and solid stores along with water retention capacity of the snowpack and refreezing of meltwater.

$$R = C_{fr} * C(T_t - T_a) \tag{4}$$

Where R is refreezing (mm h^{-1}) and C_{fr} is the refreezing coefficient, the latter of which assumed a default value of 0.05. Measured hourly precipitation and temperature served as input parameters, while hourly snowpack drainage was model output.

The degree-day model can have limited ability to appropriately assessing sub-daily time increments due to the variations in the correlation between air temperature and snowpack energy supply between day and night periods (DeWalle and Rango, 2008). That withstanding, hourly simulations were completed in this study to: 1. provide a more accurate comparison to the physically based model, which is capable of accurate hourly melt simulation; and 2. to stay consistent with the high reactivity of the vadose zone to recharging water, where infiltrating waters transit the unsaturated zone in under an hour. Further, the daily aggregate of hourly simulations, rendered the same simulation results as discussed below, just in a courser time step.

All three snow process models use a constant but calibrated threshold temperature at which rain converts to snow or vice versa. In reality, a temperature range exists in which the percentage of rain decreases as the percentage of fallen snow increases, when air temperature cools.

In total, the degree-day model has three parameters, which have to be calculated (T_t, C, and SCF, the latter a snowfall correction factor) and two time series inputs (T_a and precipitation).

Modified Degree-Day Snowmelt Model

While many adaptations to the degree-day approach (Brubaker et al., 1996; Cazorzi and Dalla Fontana, 1996; Hock, 2003; Kuusisto, 1980; Pellicciotti et al., 2005) have been fabricated to overcome the model's simplicity and consequent limitations, the recently developed Tobin method (Tobin et al., 2013) was selected as a midpoint in assessed model complexity (i.e. physical basis). The Tobin method uses measured daily air temperature extremes to impose a diurnal cycle on the melt rate. The resulting time variant degree factor (A_s) accounts for the actual distribution of snowmelt rates in time, which peak at hours of maximum incident radiation and fall to a minimum during the night (Tobin et al., 2013). The association between C and A_s is as follows:

$$A_s = \begin{cases} C + \beta\ \Delta_T \sin\left(\pi \dfrac{t_d - t_o}{t_1 - t_0}\right) & t_0 \leq t_d < t_1 \\ C - \beta\ \Delta_T\, Z & otherwise \end{cases} \tag{5}$$

Where β is a factor to convert the temperature amplitude into a degree-day factor amplitude, Δ_T is the difference between the maximum and the minimum daily temperature on day *d* (h), t_d is the hour of the day *d* (h), t_0 is the start time of daylight on day *d*, t_1 is the end time of daylight on day *d* and Z is a factor to ensure that the daily mean value of A_s equals C.

$$Z = 2\frac{t_1 - t_o}{\pi l_n} \tag{6}$$

Where l_n (h) is the duration of the night.

$$l_n = 24 - t_1 + t_o \tag{7}$$

Similarly to the classic degree-day method, the Tobin approach accounts for the solid (snow) and liquid (rain) water inputs to the system and the solid and liquid water content of the snowpack (refreezing and drainage). Snowpack drainage incorporates both rain and melting snow.

The modified degree-day model had four parameters that have to be calibrated (T_t, C, SCF, and β) and five time series inputs (T_a, t_1, t_0, Δ_T and precipitation).

Energy Balance Snowmelt Model

The point energy balance model used to simulate the energy and mass balance and melt rates of the VCB's snow surface was based on melt equations of ESCIMO.spread (Strasser and Marke, 2010). The 1-D, single-layer process model considers short and long wave radiation, sensible and latent heat fluxes, energy conducted by solid and liquid precipitation as well as sublimation/re-sublimation and a constant soil heat flux. This physically based model assumes a homogeneous, isotropic snowpack and does not consider lateral process in the mass balance calculation. For snowmelt to occur, a snow cover must be present at a given time step and the surface energy balance must be positive, indicating that energy is available for snowmelt. In keeping with ESCIMO.spread, the used energy balance model had eight parameters that required calibration (A_{min}, A_{add}, DP_p, DP_n, SS, S_e, SHF, and T_p). Input included VCB field measured air temperature (K), relative humidity (%), wind speed (m/s), precipitation (mm/h) and global radiation (W/m^2) along with calculated incoming longwave radiation ($Q_{l\downarrow}$, W/m^2). This latter parameter was derived using the Stefan-Boltzmann constant:

$$Q_{l\downarrow} = \sigma \cdot \varepsilon \cdot T_a^4 \tag{8}$$

Where T_a is the air temperature (K) at 2m and ε is atmospheric emissivity. This latter parameter was established using Prata's (1996) approach:

$$\varepsilon = 1 - (1 + W_p) \cdot e^{-(1.2+3W_p)^{0.5}} \tag{9}$$

W_p is the precipitable water as an empirical function of actual vapor pressure and air temperature.

$$W_p = 46.5 \cdot \frac{e_o}{T_a} \tag{10}$$

Where e_o is the actual vapor pressure (kPa) and was calculated in following with the FAO56 method (Allen et al., 1998).

Calibration

All four models were calibrated using the automatic parameter estimation software PEST (Doherty et al., 2016). The suitable parameters sets were found using a null space Monte Carlo (NSMC) approach, wherein 10,000 random parameter values were generated. NSMC allows the computational demand of a standard Monte-Carlo approach. NSMC is readily available through two powerful tools of the PEST-suite, SVD-assist (Tonkin and Doherty, 2005) and pre-calibration null space projection (Doherty and Hunt, 2009). In the former, paramenter space was subdivided into solution and null spaces. The subdivision took place only once and the number of estimated parameters was equal to the number of chosen dimensions of the solution space. The latter PEST tool was used to modify random parameter sets in the null space (NSMC). A variation of a parameter in the null space will not affect the target objective function (OF), defined in this study as 6000mm^2, wherein the squared difference between the observed and calculated snow water equivalent was minimized.

Through this calibration approach, it was possible to efficiently obtain a set of very different parameter fields, which respect both the stochastic variability of the model parameters as well as the historical measurements of system state with only a handful of runs per parameter field (Moeck et al., 2015). For more details the interested reader is refereed to (Dausman et al., 2010; Doherty, 2003; Fienen et al., 2009; Tonkin and Doherty, 2009).

Lumprem

Lumprem calibration occurred in two phases. In the initial phase, herein referred to as the "forward" Lumprem calibration, model parameters for soil and vegetation were calibrated based on observed hourly precipitation (rain) and VCB1 discharge for the snow-free period (04/01/12 00:00 to 6/22/2012 23:00) following the above detailed process. The initial 311 hours (until 4/13/2012 23:00) served as a model warm up phase to consider antecedent soil moisture prior to the calibration of the modeled vadose zone system. Following the calibration, the period between 6/23/2012 00:00 and 11/21/2012 23:00 was used to validate the calibration. The best-estimated parameter set for Lumprem's forward calibration had the smallest differences between observed and simulated discharge (mean-error for the test period) for the calibration and validation period.

In Lumprem's second phase of use, herein referred to as the "inverse" Lumprem calibration, the model was fixed using the 30 best-estimated parameter sets from the forward Lumprem calibration. Lumprem was then used to estimate the amount of snowmelt infiltration needed to arrive at the observed VCB1 discharge, between the winter period of 11/1/12 00:00 and 6/1/13 00:00. Lumprem's 30 inverse optimizations are presented in the results and discussion as the point of comparison for the snow process models' output.

Snowmelt Models

The degree-day, modified degree-day and energy balance snow process models were calibrated to manual SWE measurements for the period between 11/1/12 00:00 and 6/1/13 00:00 following the above detailed process. SWE values were log transferred and given equal weights during the PEST calibration. We assumed that all observations are equally relevant and therefore the weights are not changing between the observations. We calibrated T_t, C, and SCF for the degree-day model, T_t, C, SCF, and β for the modified degree-day model, and A_{min}, A_{add}, DP_p, DP_n, SS, S_e, SHF, and T_p for the energy balance model. Success of snow process simulation (for the best-fit parameter set for each snowmelt model) was evaluated via coefficient of determination (R^2), index of agreement (IA), (Willmott, 1981) and Nash-Sutcliffe model efficiency (NSME), (Nash and Sutcliffe, 1970).

Results

Degree of snow process model success was defined by the models' ability to meet the objective function of 6000mm^2. One iteration of the modified degree-day, 15 iterations of the energy balance and none of the classic degree-day iterations met the objective function. Given this range of ability to meet the OF, snow model results are compared in two ways; firstly according to the 30 best-fit iterations for each respective model; and secondly according to the iterations that actually met the objective function.

Over all, Lumprem was able to reproduce observed VCB1 discharge during the summer forward calibration and validation periods very well (the time series of observed VCB1 discharge and Lumprem estimated discharge, using the optimal parameter set for the forward calibration and validation are presented in Figure 3).

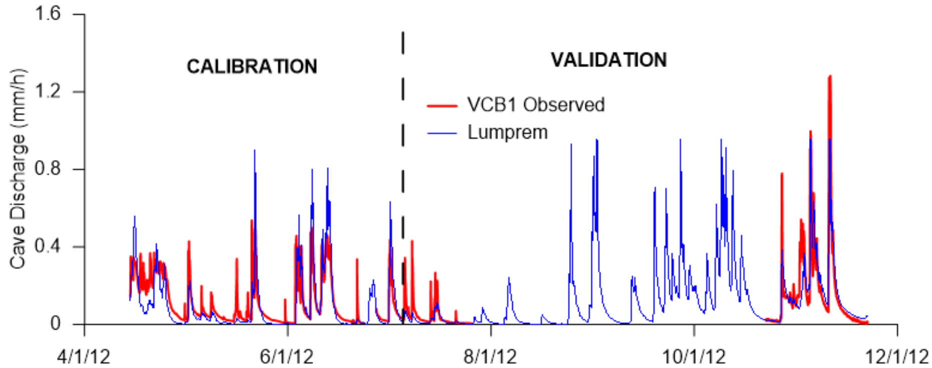

Figure 3: Best-fit Lumprem calibration and validation during snow-free summer and fall

However, the validation period underestimated peak discharge events and was not able to fit the tailing of individual hydrograph events with great accuracy. During the validation phase, peak discharge was overestimated by Lumprem and modeled event hydrograph recessions more closely matched observed decreases in flow within the cave. The gap in observed VCB1 discharge resulted from equipment failure in the cave. Lumprem inversely estimated a total accumulated infiltration from snowmelt to have been between 493 to 837mm, with the mean cumulative infiltration of 703mm. Figure 4 shows the cumulative, inversely-estimated snowpack drainage from Lumprem (30 best-fit iterations and ensemble mean (EM, the average of the model results from the 30 best-fit parameter sets)) along with the simulated cumulative drainage by the snowmelt models. Only the iterations that met the objective function are presented for the modified degree-day and the energy balance models. In contrast, the 30 best-fit iterations for the classic degree-day (none of which met the objective function) are presented in this figure. Increases in cumulative Lumprem modeled infiltration correspond with either rain (or mixed precipitation) on snow events (end of December, beginning of February and April) or positive air temperatures (mid March). Periods of snow accrual with consistently negative air temperatures (second half of February) did not correlate to an inversely modeled increase in snowmelt infiltration.

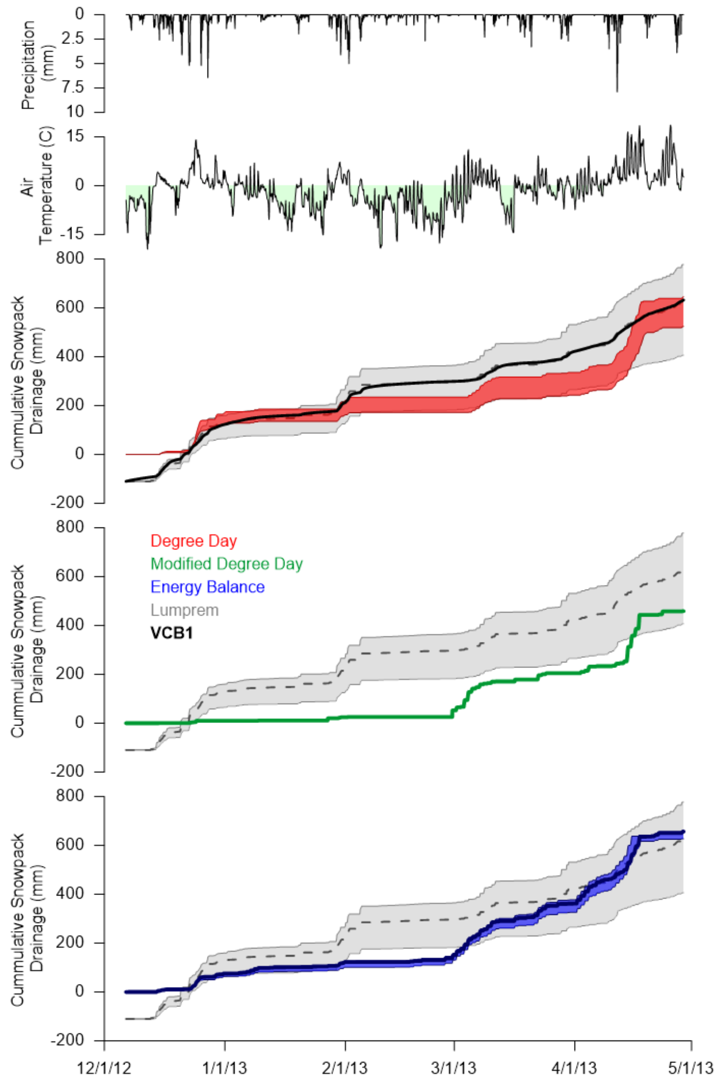

Figure 4. Comparison of accumulative snowpack drainage for each snowmelt model with Lumprem's inversely estimated snowpack drainage. Depicted are the; thirty best-fit Lumprem iterations for inversely estimated snowmelt (gray); the ensemble mean of the thirty best-fit Lumprem iterations (dashed line); the discharge observed at the VCB (black line); the thirty best-fit degree-day iterations (none having met the objective function, red); the one modified degree-day iteration that met the objective function (green); and the 15 energy balance iterations that met the objective function (blue). Lumprem results are offset by 110mm to account for a melt event at the beginning of winter that was not observed by the snow models.

The classic degree-day method over estimated the early-winter snow accumulation in December, underestimated snow accumulation in January, and initially underestimated and then overestimated SWE during the melt phase in March and April. None of the classic degree-day iterations reached the objective function of 6000mm^2. The range in cumulative drainage from the snowpack is relatively small, between 518 and 638mm, and lies in the lower bounds of the range of inversely estimated snowpack drainage from Lumprem (493 to 837mm). The EM for degree-day simulation of snowpack drainage is approximately 110 mm less than the ensemble mean generated by Lumprem. A boxplot of the calibrated parameters representing the 30 best-fit NSMC simulations is presented in Figure 5. The model's SCF maintained the highest parameter variability while the C maintained the least. This calibration resulted in a NSME of 0.16, a R^2 of 0.31 and an IA of 0.75.

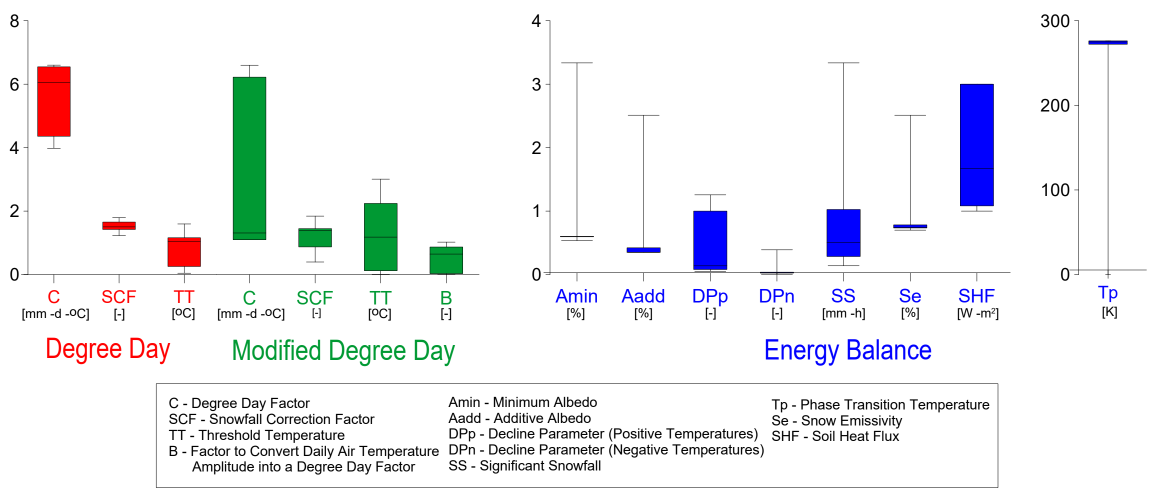

Figure 5: Boxplot of calibrated parameters representing the 30 best-fit NSMC simulations for each of the three snowmelt models.

While able to reproduce SWE values, on average, the modified degree-day modeling approach slightly overestimated SWE during the accumulation phase of early winter and then slightly underestimated SWE during the slow accumulation phase in January (Figure 6). The modified degree-day approach then went on to initially underestimate and then overestimate SWE during the spring snowmelt phase. When considering the 30 best-fit iterations, the modified degree-day model had a larger variability of estimated cumulative snowpack drainage, ranging between 448 and 650mm, with an ensemble mean of 524mm, approximately 167mm less than what was estimated by Lumprem (EM). Of these 30 iterations, only one was able to achieve the objective function, which presented a cumulative drainage of 458.4mm. This iteration had a C of 1.1, a SCF of 0.73, a T_t of 2.7, and a β of 0.96 and resulted in a NSME of 0.72, a R^2 of 0.73 and IA of 0.92. Similarly to its more simplified cousin, the model's SCF maintained the highest parameter variability while the C maintained the least. However, with the introduction of the temperature amplitude conversion factor, the three calibrated parameters that were also used in the classic degree-day approach had an increase in parameter variability of approximately 100% in relation to that observed in the simplified degree-day version.

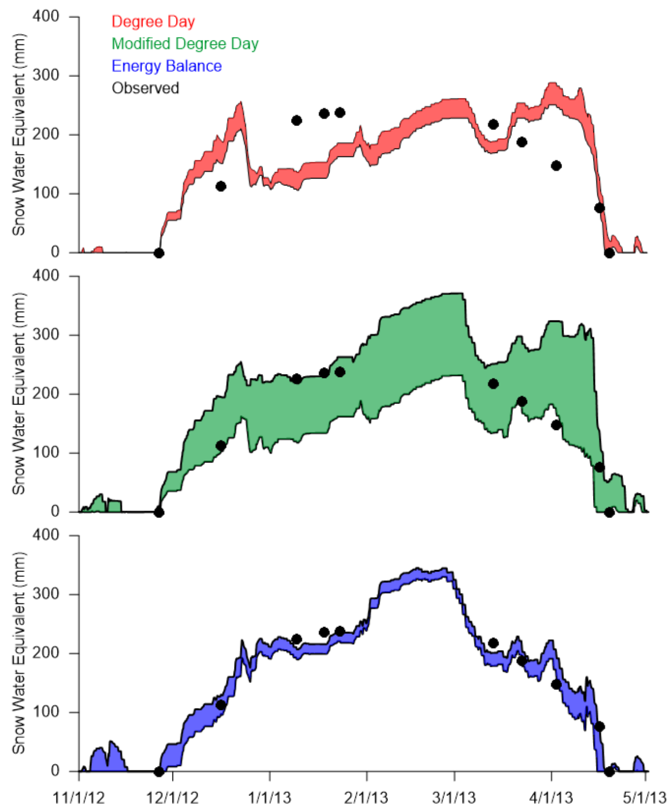

Figure 6: Shows the range of the 30 best-fit null-space Monte Carlo iterations for each snowmelt model. The black dots represent the field-measured SWE.

The energy balance model reproduced observed SWE measurements very well throughout the entire winter, with only nominal underestimates of mid-winter observations (Figure 6). The cumulative snowpack drainage, as indicated by the EM, was only 49mm less than that estimated by Lumprem (EM) and presented a potential cumulative range of 625 and 656mm. 15 iterations reached the objective function with the calibrated parameters ranging accordingly: A_{min} (0.57 to 0.6), A_{add} (0.35 to 0.45), DP_p (0.05 to 1), DP_n (0.02 to 0.06), SS (0.18 to 1.91), S_e (0.7 to 0.78), SHF (1 to 3), and T_p (272.16 to 276.16). The cumulative drainage for these 15 iterations ranged between 653mm and 662mm (Figure 4). The iteration with the lowest objective function presented a cumulative drainage of 656mm (only 40mm less than that estimated by Lumprem (EM)) and resulted in an IA of .98, a R^2 of .91 and a NSME of .91. Albedo (A_{min} and A_{add}), decline parameter (positive temperatures) and snow emissivity presented the highest parameter variability while soil heat flux and phase transition temperature presented the lowest (Figure 5).

Discussion

This is the first study to systematically compare the uncertainties surrounding modeled predictions for recharge from snowmelt. It is important to distinguish the contributing sources of uncertainty. The first category resulted from the input data used to calibrate the models, while the second category resulted from the modeling approaches themselves. While snow process model predictions are the aim of this study, we will begin the discussion by evaluating the uncertainty arising from the models' input.

As depicted in Figure 4, cumulative snowpack drainage estimated by the snowmelt models and Lumprem do not coincide perfectly, principally so in the first half of the winter. During this time, Lumprem simulated two infiltration events (12/15/12 to 12/27/14 and 1/28/13 to 2/4/13) that were only marginally detected by the three snowmelt models. The synchronous underestimation of snowpack drainage in the first half of winter by all three snowmelt models, which range between physically based and index, implies input data error. In the first half of winter, precipitation data was derived from an insulated and heated tipping bucket attached to the remotely-located weather station. A solar panel sourced electricity to the tipping bucket's heating unit. The small heating unit, limited by its electrical supply, was unable to generate enough heat to melt all fallen snow and prevent ice bridging. Therefor on 2/22/13, a more rigorous precipitation gauge was installed approximately 3km away at the closest source of electricity. An increased electrical supply allowed for a more robust heating unit in a larger weighting precipitation gauge. Snow bridging was never observed at this latter instrument. The weighting precipitation gauge was not installed earlier due to lack of observed snow bridging in the previous winter and a consequent lack of knowledge of the need for a hardier device. As a reminder, Lumprem's inversely estimated infiltration during the winter is based solely on discharge observed in the cave. Therefor, the two early-winter infiltration events are believed to be "real" as they are not biased by concurrent precipitation measurements. The lack of observed snowpack drainage by all three models during these two periods implicates precipitation under-catch by the original tipping bucket as the source of off-set shown in the cumulative plots of Figure 4. Tipping bucket gauge under-catch relative to weighting gauges, particularity during periods of snowfall has been estimated to be upwards of 22% (Hanson et al., 1999). Furthermore, Gurtz et al. (2003) showed that liquid and solid precipitation measurements with conventional gauge are associated with large errors. They applied time depending monthly correction factors which were high in late fall, over the entire winter and early spring, starting from +5% in October for rainfall and rising up to +62% in March for snow. The discrepancy between the snowmelt models and Lumprem in the first half of winter re-affirms that selection of precipitation measurement device is absolutely paramount in snowmelt modeling studies.

The second source of uncertainty, and principal motivation for this study, resides in the snow process models' structural ability to predict loss of SWE, i.e. available infiltration for recharge. Lower reproducibility of SWE by the index model is logical, given that physical properties such as vapor pressure, albedo, and wind are known to influence the energy balance and snowmelt processes (Tobin et al., 2013) and are not explicitly accounted for. The elevated parameter variability for the degree-day factor (C, Figure 5) relative to the other variable, most likely resulted from air temperature serving as a proxy for the all the physical forces acting on and within snowpack.

The modified degree-day model was able to more closely approximate the accumulation and ablation of VCB SWE, however only using certain parameter sets resulting from the calibration. This modified index method used a factor (b) to account for the semi-sinusoidal fluctuations in daily air temperature, which indirectly accounts for environmental aspects like cloud cover, length of day, and seasonality. The robustness of this approach is called into question though, when considering the doubling of parameter variability for the degree-day factor, snowfall correction factor and the threshold temperature that arises with the addition of β in the calibration. While parameter variability is significantly greater than in the degree-day approach, the model does allow for the opportunity to more closely reproduce observed SWE data, emphasizing the importance of a NSMC simulation coupled with parameter estimation software in modified index models.

The energy balance model presented the smallest snowpack drainage range of the three models when considering the 30 best-fit iterations, with the 15 best-fit NSMC runs resulting in only mild deviations of modeled accumulation and ablation (Figure 4). As the energy balance equations used in this study were in keeping with those of ESCIMO.spread, latent and sensible heat fluxes, sublimation and re-sublimation, snow age and albedo and the constancy of soil heat flux may serve as sources of model error (Strasser and Marke, 2010). For example, parameter variability may have developed from the simplified calculation of turbulent fluxes, which assumes a medium snow surface roughness and stable snow-surface temperature. In areas where the contribution of the turbulent fluxes to the energy balance of the snowpack is small, the induced loss of model accuracy is negligible (Strasser and Marke, 2010), however stable conditions never truly exist and snow roughness certainly varied throughout the winter. Also, in our model, advective energy supplied by precipitation depends on its phase. Since the percentage of precipitation attributed to each phase was not measured, an adjustable threshold temperature was assumed for the distinction between snow and rain, thereby inducing potential error into the advective component of the energy balance equation. As noted, this latter source of error resulting from a course consideration of the rain/snow phase transition, was uniform throughout all the snowmelt models. All models used the same amount of precipitation in the same form at a given time. Given this consistency across models, this source of uncertainty is not overly important in this study, as the induced uncertainty would be consistent across all snow models. More precise energy balance assessment of accumulation and melt parameters is possible, such as surface roughness, wind speed and snow surface temperature, but with the added cost of high precision instrumentation that was beyond the scope of this study.

While the NSMC parameter sets used for Lumprem's calibration resulted in a large variability of cumulative infiltration (383 to 727mm, Figure 4) at the end of the snow period, the ensemble mean of the 30 best-fit iterations overlies almost perfectly with the cumulative discharge from VCB1 within the cave. This implies Lumprem's successful performance at inverse estimation of snowpack drainage and also reveals that ensemble means should be applied to produce credible predictions of snow drainage and the associated uncertainties. If only one realization of Lumprem, and each snowmelt model, had been completed, our results and associated conclusions would have been greatly altered. As depicted in Figure 4, all three snowmelt models underestimate cumulative snowpack drainage. When considering the potential precipitation under-catch of 22% in the first part of winter, the degree-day model would have underestimated (based on the EM) cumulative snowpack drainage by 65mm; the modified degree-day model would have underestimated (based on the iteration with the lowest objective function) drainage by 215mm; and the energy balance would have underestimated (based on the iteration with the lowest objective function) drainage by 14mm when compared to the Lumprem's estimated cumulative drainage (EM). These underestimates respectively account for 9%, 31% or 2% of the total Lumprem estimated (EM) drainage by the end of the snow period. When comparing the trends in cumulative increases in snowpack drainage over time (EM for each model), some interesting observations can be made. With increased complexity of snowmelt model, there is an increased underestimation of snowpack drainage during the first half of winter. While the underestimation of snowmelt rates in the early part of the snow season is probably related to input error associated with precipitation under-catch, this error is consistent across all models. Through the second half of winter, the degree-day model initially tracks Lumprem drainage well, but then grossly overestimated rate of snowpack drainage. This latter period speaks of the degree-day model's oversensitivity to air temperature. The modified degree-day model tracks similarly to the classic degree-day approach. The energy balance model maintains a more persistent overestimation throughout the second half of winter, issuing to the models more parametrically distributed sources of uncertainty tying into the computation of latent and sensible heat fluxes, sublimation and re-sublimation, snow age and albedo.

The finding from this study must be considered in light of how snow process models are typically used. Water management agencies may apply snow process models to understand such things as melt pulses resulting from a rain on snow events (as might be the case for flood protection or water quality monitoring), or to determine a snowmelt's contribution to an annual water budget (such as contributions of snowmelt to base flow in the summer). The results show that none of our models are well suited for very short time scales, but rather the longer seasonal scale. Our study reinforces the need to carefully choose the model based on not only in terms of uncertainty surrounding predications but also temporal scale associated with intended application of model results.

Overall, the results indicate that with increased physical basis to the model and increased parameterization of the factors influencing melt and accumulation, there is a higher ability of the model to reproduce the observed SWE values. Further these results show that will rigorous calibration, one can achieve model results that have low uncertainty surrounding predictions and produce results that are within the realm of reality as is the case for the energy balance model.

Conclusions

As water demands outstrip water supply, more in depth knowledge of recharge processes will be needed. The rate and timing of infiltration from snowmelt is frequently devised via snowmelt models, with varying degrees of complexity and physical basis. Given the importance of snow process model results for water management agencies, we presented here the results from the first systematic inter-comparison of uncertainties associated with modeled snowmelt predictions. Back-calculated snowmelt data was used as a point of comparison for three snowmelt algorithms; a simple degree-day model, a modified degree-day model and an energy balance snowmelt model. Through null space Monte Carlo calibration, we evaluated the performance and uncertainty arising from each snowmelt model's predictions. When compared with the back-calculated snowmelt, all three melt models underestimate total snowpack drainage, underestimate the rate of early and midwinter drainage (possibly related to precipitation gauge under-catch) and overestimate spring snowmelt rates. Further, the rate of snowpack water loss is more constant over the course of the entire winter season than the snowmelt models would imply when compared to the back-calculated snowmelt. These results indicate that mid-winter melt can contribute as significantly as springtime snowmelt to groundwater recharge. Further, these groundwater management agencies bodies should be aware that actual groundwater renewal could be between 2 and 31% greater than snowmelt models suggest.

The choice of the snowmelt model used might be dictated by the available data. For example, in many cases only temperature data is available and thus only the day-degree method can be use. Our study however shows that the uncertainties are considerable for snowmelt modeling and can be reduced by inclusion of more data that is integrated into a more complex approach such as the energy balance method, and that in any case quantifying the uncertainties should be done. This study demonstrated that an uncertainty analysis of model predictions is easily accomplished due to the low computational demand of the models and efficient calibration software. Further, this analysis is absolutely worth the additional investment as it allows model users to arrive at an optimal model realization with a minimized objective function.

Also, this study was only possible due to the solid data for the infiltration under the snowpack, generated via the unique, oversized, natural lysimeter of our field site. The importance of real distributed calibration data such as this cannot be underestimated or understated. Given that the geology allowing for a study such as this is quite uncommon, development of a systematic instrumentation that evaluates the distributed, temporal evolution of snowpack drainage is paramount for optimal understanding and management of cold-climate hydrologic systems.

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Chapter 4 - Conclusions and Implications

Findings

As presented in Chapter 1, we studied the transit and storage of snowmelt water through the vadose and phreatic zones of a karst aquifer to identify firstly the mechanisms controlling winter-season groundwater renewal in a karst setting, and secondly how these mechanisms impact karst aquifer storage and discharge. The findings can be distilled down to: 1. vadose zone storage modifies the meltwater signal, even during extended periods with subzero temperature, outflow takes place; 2. perched storage is occurring dominantly in the soil horizons rather than the epikarst as had been previously postulated by the scientific community; 3. this perching may be more relevant in winter and early spring when evaotranspiration is a nominal aspect of the water budget, enhancing the temporal redistribution of water from melt events to cold periods lacking snowmelt infiltration; 4. little attenuation of water flow occurred between the vadose zone and the aquifer's primary spring, indicating that vadose zone storage and flow has a strong control on aquifer discharge at the scale of weeks, while phreatic storage becomes dominant during prolonged periods without input.

These deductions implicate a strong coupling of recharge and discharge in karst aquifers and highlight the importance of understanding recharge mechanisms when attempting to predict future groundwater availability from karst aquifers. Thus, in our second study, we sought to assess how well snowmelt models actually simulate winter season infiltration and the predictive uncertainty associated with these models' predictions. This study was novel for two reasons: firstly, we completed the first systematic inter-comparison of uncertainties associated with modeled snowmelt predictions and secondly, this study was carried out using data collected from an oversized real-world lysimeter. The spatially integrated catchment data allowed us to complete our assessment not at the usual point scale, but at the catchment scale, which was unique in and of itself. Through null space

Monte Carlo calibration, we evaluated the performance and uncertainty arising from three snowmelt models' predictions (a simple degree-day model, a modified degree-day model and an energy balance snowmelt model). Results indicated that mid-winter melt can contribute as significantly as springtime snowmelt to groundwater recharge. Snow process model predictive uncertainty, while significant for all three model approaches, is reduced with increased parameterization, rendering the energy balance model as superior in its ability to reproduce observed snow water equivalent values. Further, the results show that none of our melt models are well suited for very short time scales, but rather the longer seasonal scale. This lack of temporally-refined modeling ability might pose particular difficulty in well-developed karst settings where individual melt events can have direct impacts on aquifer discharge. For other types of aquifers that buffer the input signal very strongly and do not react to individual events, this may not be as important.

We then expanded on these findings to obtain a better understanding of the sensitivity of winter season groundwater recharge to systematically increasing air temperatures. This was not a full-blown climate change impact study, but rather an initial evaluation of temperature impacts on winter season recharge to a karst aquifer. The energy balance snowmelt model iteration, from the Null Space Monte Carlo calibration with the lowest phi from the study presented in Chapter 2, was used to for this purpose.

The temperature shifts were selected based on Swiss Climate Change Scenarios CH2011; an easily used tool that enables modeling of climate change impacts specifically within Switzerland. The CH2011 were informed by the IPCC's work and results from the ENSEMBLE Project (Van der Linden and Mitchell, 2009), new statistical methods developed to better quantification of uncertainties in climate projections and improved downscaling of climate variables at specific sites (CH, 2011). The three CH2011 emission scenarios, based on IPCC's two nonintervention emission scenarios (A2 and A1B) that anticipate increases in emissions, and one climate stabilization scenario (RCP3PD) that supposes emissions are cut by about 50% by 2050, approximately correspond to increased seasonal mean air temperatures of 1, 3, and 5 degrees Celsius. Air temperature from a brief reference period of 11/21/12 to 4/28/13 was increased by these amounts and then used as input to the snowmelt energy balance model discussed in Chapter 2. All other input parameters such as precipitation, incoming longwave radiation, wind speed, etc. remained the same as in the reference period. While projected changes in precipitation from CH2011 could have been applied as well, this parameter was not altered to better identify temperature effects, as temperature is the key driver for change in snow-covered regions. The calculated snowmelt rates were then applied as input to the calibrated VCB vadose zone Lumprem model (Chapter 2) to generate time series recharge data. In Chapter 2, Lumprem was calibrated using the null space Monte Carlo approach. The optimal iteration with the lowest phi was used in this study. The changes in recharge distribution associated with the warmer air temperatures were then evaluated.

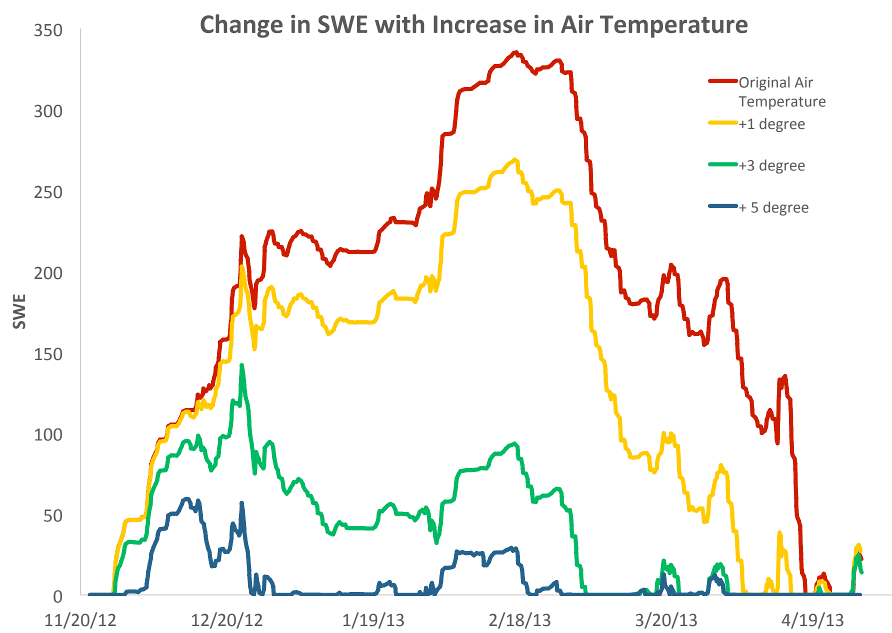

Figure 1: The snow water equivalent time series were modeled using the energy balance snow model, previously calibrated with a null space Monte Carlo approach. Air temperature was increased respectively by 1°C, 3°C and 5°C, while all other input parameters remained consistent between model runs.

The evolution of SWE for each of the temperature-increase scenarios is presented in Figure 1. As would be expected, increased air temperature reduces both a snowpack's SWE at a given time and also its duration of emplacement. A 1°C air temperature increase reduced the period with continuous snow cover by 13 days, while a 3°C increase reduces continuous snow cover by 46 days. A 5°C increase in air temperature results in a bifurcation of the seasonal snowpack emplacement. Snowpack became more ephemeral and isolated to individual storm events. Maximum SWE was reduced from 331mm to 265mm with a 1°C increase, to 138mm with a 3°C increase, and to 59mm with a 5°C increase in air temperature. Additionally, peak SWE appeared earlier and earlier in the winter: 2/18/13 for the original air temperature, 2/27/13 for a 1°C increase, 12/23/12 for a 3°C increase, 12/12/12 for a 5°C increase.

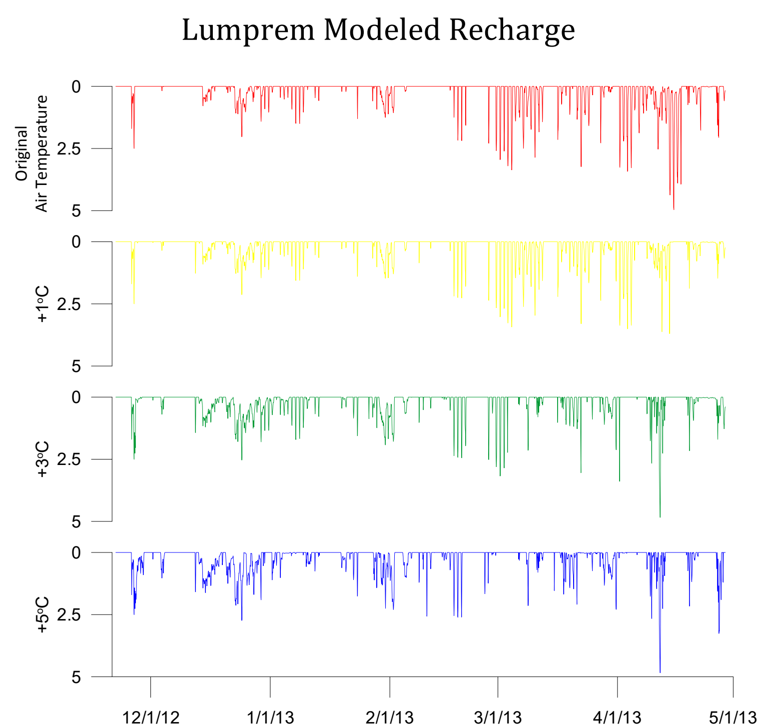

Figure 2: Lumprem modeled recharge intensity (mm/hr)

Figure 2 shows the recharge intensity over the 2012/13 winter period. An increase in air temperature results in an increased rate of recharge in the first half of winter and a reduced rate in the second half of winter. By the first week of January, a 5°C increase in air temperature resulted in an approximate infiltration increase of 2.5 times compared with recharge associated with current temperatures. The amount of change in recharge rate shifts significantly between a 1°C increase and a 3°C increase for the middle part of the winter, in comparison to the changes between current air temperatures and 1°C increase and the changes between a 3°C increase and a 5°C increase. This implies that a 3°C increase has more significant implications for temporal distribution of infiltration than a 1°C increase. We can anticipate that with an additional degree or two of warming, snow will no longer accumulate and all infiltration will result from rain and enter the subsurface directly.

These results imply that even slight increases in air temperature significantly impact the temporal distribution of recharge to an aquifer. Warming air temperatures result in reductions in surficial storage of precipitation, increases in early and mid-winter recharge and reductions in late winter recharge. This change may have impacts on agriculture as water demands are highest during the summer growing season, which has historically been preceded by spring snowmelt pulses. With a warming climate, late spring pulses of recharge will no longer occur with great intensity. For aquifers that can store and retain water well, it might not make much of a difference how recharge is distributed in winter as long as the overall quantity remains fixed. For a karst aquifer however, recharge distribution throughout the winter can have significant impacts on groundwater availability, rendering these aquifer types particularly susceptible to climate change.

Limitations and Further research needs

The first study highlighted the importance of knowing site soil structure for appropriate karst aquifer management. The storage mechanism at the Vers Chez le Brandt is strongly associated with genesis of the soil i.e. the deposition of siliceous loess on top of the calcareous bedrock, a configuration not uncommon in Europe and North America. As such, identification of soil type and distribution should be an integral part of karst aquifer assessments, particularly in regions that receive seasonal snowfall. This study further identifies the importance of maintaining soil health in karst watershed, as extended storage of perched water in soils may allow for extended chemical exchange with soil constituents and any present contaminants, altering water quality. Given these conclusions, the influences of different soil types on the behavior of karst aquifers deserves further research at the point and distributed scales.

The second study implicated the importance of a thorough calibration process in snowmelt modeling and the need for the snow process modeling community to adopt rigorous calibration as standard practice. We demonstrated that an uncertainty analysis of model predictions can be easily accomplished due to the low computational demand of the models and efficient calibration software. This additional analysis is certainly worth the investment as it allows model users to arrive at an optimal model realization with a minimized objective function.

Further, our study was only possible due to the solid data for the infiltration under the snowpack, generated via the unique, oversized, natural lysimeter of our field site. The importance of real distributed calibration data such as this cannot be underestimated or understated. Given that the geology allowing for a study such as this is quite uncommon, development of a systematic instrumentation that evaluates the distributed, temporal evolution of snowpack drainage is paramount for optimal understanding and management of cold-climate hydrologic systems. More broadly, the study underscored the importance of input data being as accurate as possible. The snow process modeling results were certainly influenced by the source of the precipitation data, ie. the traditional heated tipping bucket versus the more robustly heated weighted precipitation gauge. In short, the ubiquitously used heated tipping buckets are ill advised for anything other than a cursory understanding of local snowfall patterns. Finally, this study showed that the uncertainties are considerable for snowmelt modeling and can be reduced by inclusion of more data that is integrated into a more complex approach such as the energy balance method. A future study could build upon this work by doing a similar uncertainty analysis of snow process models used more commonly in the private, public and academic sectors, that incorporate a distributed, larger watershed-scale component to the work.

Further study assessing the potential impacts of warmer air temperature on groundwater recharge to karst aquifers should be completed with more realistic climate scenarios that consider region-specific, expected changes in precipitation and evapotranspiration. This work should then be extended to the catchment scale, where altitude gradients and sucessive snow melts are evaluated to see how temperature effects manifest. Lastly, this work should be caried out for an array of geographic regions, given the inherant site specific nature of karst.

In final conclusion, this body of work shed light on the mechanisms controlling winter recharge to karst aquifers, the true ability of snow process models to predict recharge from snowmelt, and the impacts of a warming planet on karst aquifer resources, yet it raises many other questions that I, and my academic peers, can work toward resolving.

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Appendix A

Determination of spatiotemporal variability of tree water uptake using stable isotopes (δ18O, δ2H) in an alluvial system supplied by a high-altitude watershed, Pfyn forest, Switzerland

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Accepted by *Ecohydrology*

Abstract: Sources of water use by 10 alluvial trees in various hydrogeological and ecological situations at the Pfyn forest (Wallis canton, Switzerland) were assessed by analyzing 18O and 2 H isotopes of precipitation, soil water at different depths, surface water, groundwater and xylem sap. The soil water line in a $d^{18}O$ versus d2 H diagram shows evidence of kinetic fractionation related to evaporation. The tree water line is close to the soil trend; however, an additional enrichment may occur and could be related to xylem–phloem communication under water stress. At sites where the water table was at least 2 m below the ground surface, isotopic temporal variability of soils and trees was strongly linked with seasonal variation of soil water content. When soil water content was low and water table shallow, trees used both soil water and groundwater. When soil water content was high, this source was usually the dominant source for transpiration. In addition, some ecological strategies, reproduction or growth competition, could explain shifts in the utilization of different water sources, for example, from soil water to a mix of soil water and groundwater. At one site where soil water and groundwater were abundant throughout the year (next to the river course), neighboring trees permanently used distinct water sources. This is consistent with a strategy of competition limitation, which would favor bank colonization. These results provide insight into the ecohydrological functioning of this system and will aid future management responses to both local and climate changes.

Key Words: water uptake; deuterium; oxygen-18; trees; alluvial ecosystem; Switzerland

Contributions toward publication:
I served as a reviewer of the paper and completed two iteration of rigorous edits and re-writes of the publication prior to revision by technical contributors.

Appendix B

Towards Operational Methods for the Assessment of Intrinsic Groundwater Vulnerability: a Review

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Abstract: Assessing vulnerability of groundwater to adverse effects of human impacts is one of the most important problems in applied hydrogeology. At the same time, many of the widespread vulnerability assessment methods do not provide physically meaningful and operational indicators of vulnerability. Therefore, this review summarizes (i) different methods used for intrinsic vulnerability assessment, (ii) methods for different groundwater systems. It particularly focuses on (iii) time scale methods of water flow as appropriate tool and (iv) provides discussion on the challenges in applying these methods. The use of such physically meaningful indices based on time scales is indispensable for groundwater resources management.

Key Words: intrinsic vulnerability, groundwater, pathway, objective methods, time scales, mean residence time, transit time distribution

Contributions toward publication: I served as the lead author on all sections pertaining to groundwater vulnerability assessment methodologies for karst systems. I also participated in the assimilation of subsections by other co-authors and multiple iterations of edits and paper re-writes.