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

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S U M M A R Y

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.

1. 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

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and are thus an area of great concern (Ford and Williams, 2007; 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,b; 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), 53 m 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.

2. 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 1550 mm of precipitation annually, 30–40% 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 °C and −1 °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 m^3/s 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 (Fig. 1). The single-chamber VCB karst cavity, which underlies the study site (average orientation N145°), 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, 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

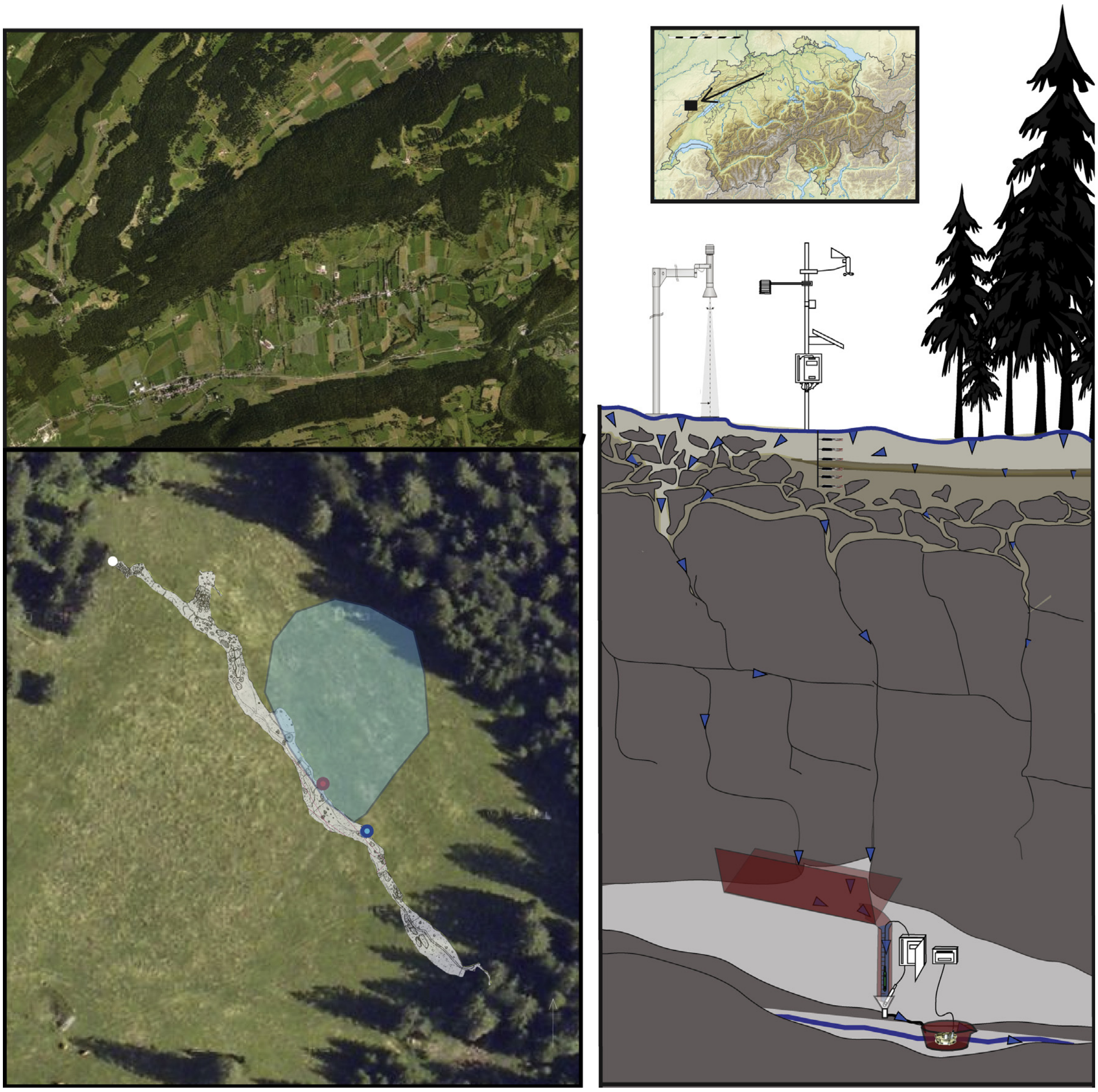

Fig. 1. Counter-clockwise: a regional site map; the Vers Chez le Brandt (VCB) site with approximate recharge area, cave roof drip point (VCB1) and cave orientation denoted; and profile (not to scale) of site instrumentation with conceptual flow indicated by the blue arrows. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

traverses the cave length (Fig. 1). The primary vadose-zone exfiltration point (VCB1) is located approximately 175 m from the cave entrance (Savoy, 2007).

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 (Fig. 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 fossiliferous 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, 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 10 cm to approximately 50 cm b.g.s. This is in turn underlain by a 10 cm thickness of brown clayey BT soil (pH of 5.5). Beneath this is an undefined thickness of gray, clayey IC soils, with a pH of 8 (Fig. 2). According to Baize and Girard (2009) Neoluvisol (aka. sols bruns) soils correspond with the Luvic Cambisols as described by the

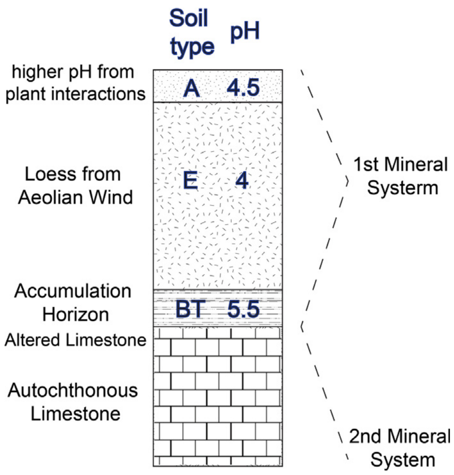

Fig. 2. VCB soil structure and composition. The left letter column (A, E, BT, and IC) indicates the type of soil while the right number column (4.5, 4, 5.5, and 8) corresponds to pH values.

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 70–130 cm 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 0.5–3 m, and by Elouardi's (1998) seismic refraction study, which implied a joint soil and epikarst thickness of approximately 2 m.

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 rye-grass 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 (Fig. 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 40 cm 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.

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

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

3. 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 90 cm 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 (Fig. 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 0–2000 W/m^2). Snow height was recorded adjacent to the VCB meteostation by a Sommer USH-8 Ultrasonic Snow-depth sensor (range of 0–8 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 5.5 m long PVC collection device mounted to the roof funneled VCB1 discharge water to a 1.4 m 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 and 2012. A WTW TertaCon 96A electrical conductivity (range of 0–199.9 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.

4. Results

4.1. 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 (Fig. 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 (Fig. 3). As data summary for these periods is presented in Table 2.

The snow accumulation period (Fig. 3) was approximately 20% shorter in 2012 (119 days) than in 2013 (146 days), but only slightly less snow accumulated in 2012 (97 cm) compared to 2013 (103 cm). 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 (Fig. 3). The 2012 winter's melt season (C12) had 3 distinct phases (a, b and c) during which an additional 5 cm of precipitation fell and 16 cm drained from VCB1. C13 was a wet period with an additional 30 cm 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 (Fig. 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 40 cm) 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 90 cm) showed stable values except for the intense cold periods in 2012. During these periods the moisture content dropped first in 70 cm and subsequently also in 90 cm. As indicated by flow rates at VCB1 (Fig. 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 (Fig. 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 (5 cm precipitation) compared to C13 (30 cm 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.

4.2. 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 90 cm of the vadose zone were constructed for selected days of the snowmelt periods (Fig. 4). The selected dates cover different stages of the melt phase of the snowpack as indicated in Fig. 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-h 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 55 cm 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 55 cm depths. Saturation increased with time at the base of the upper mineral system, just above the pedological contact (55 cm) with the lower mineral system, suggesting the formation of a perched water lens. Soil moisture content remained stable below 55 cm, indicating that the daily influx of melt water observed in the overlying layers had minimal influence on the saturation of the lower mineral system (Fig. 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.

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 (Fig. 5). This 2012 time segment (Period B12 and C12, Fig. 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). Fig. 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 0–40 cm layer's water content decreased by 2 cm, the 40 to 70 cm layer's water content decreased by 2.7 cm, and the 70–90 cm layer's water content decreased by 1.1 cm, totaling 5.8 cm of water collectively drained from the soils. 5.5 cm 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

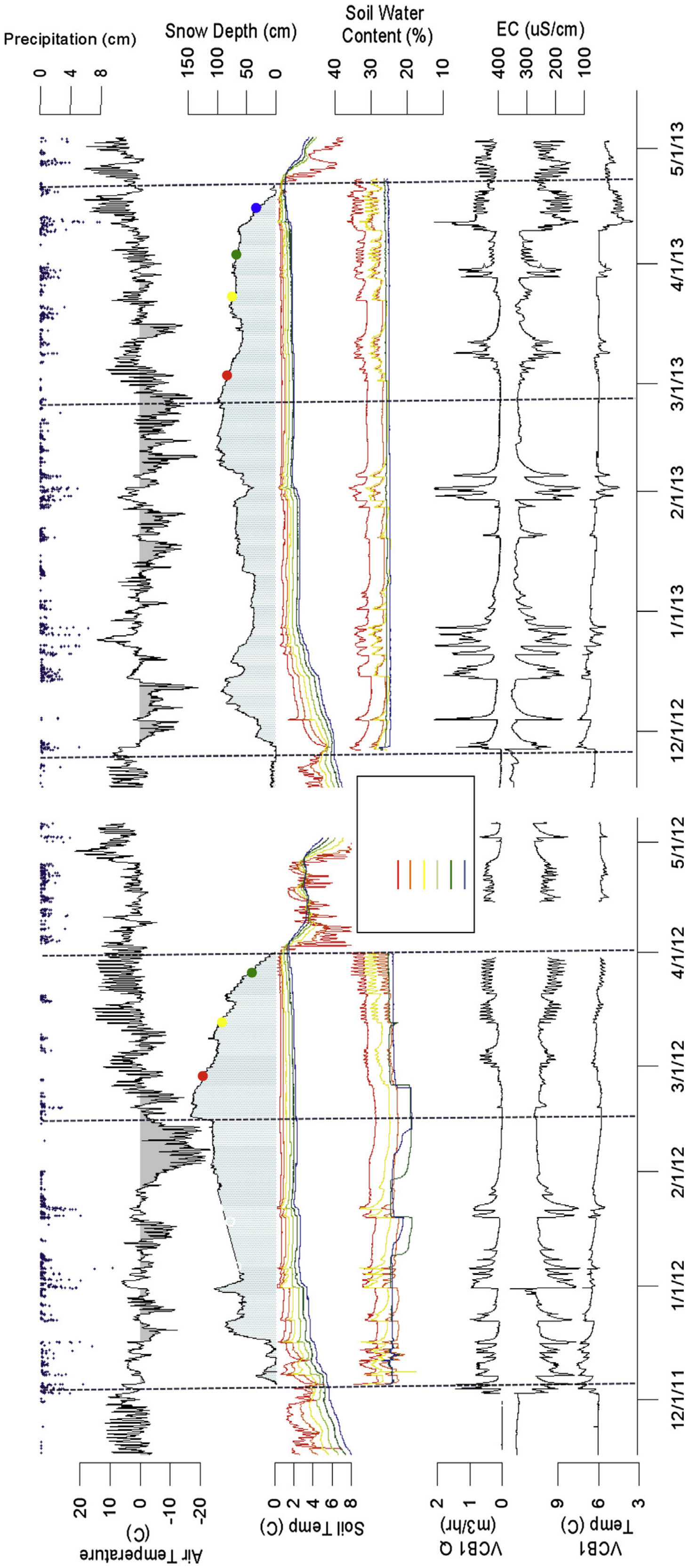

Fig. 3. Time series data for precipitation, air temperature, snow height, soil temperature, soil water content, and VCB1's discharge, electrical conductivity and temperature for the winter seasons of 2011/12 and 2012/13. Time series data is segregated into periods during which data showed distinctive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

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.

	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

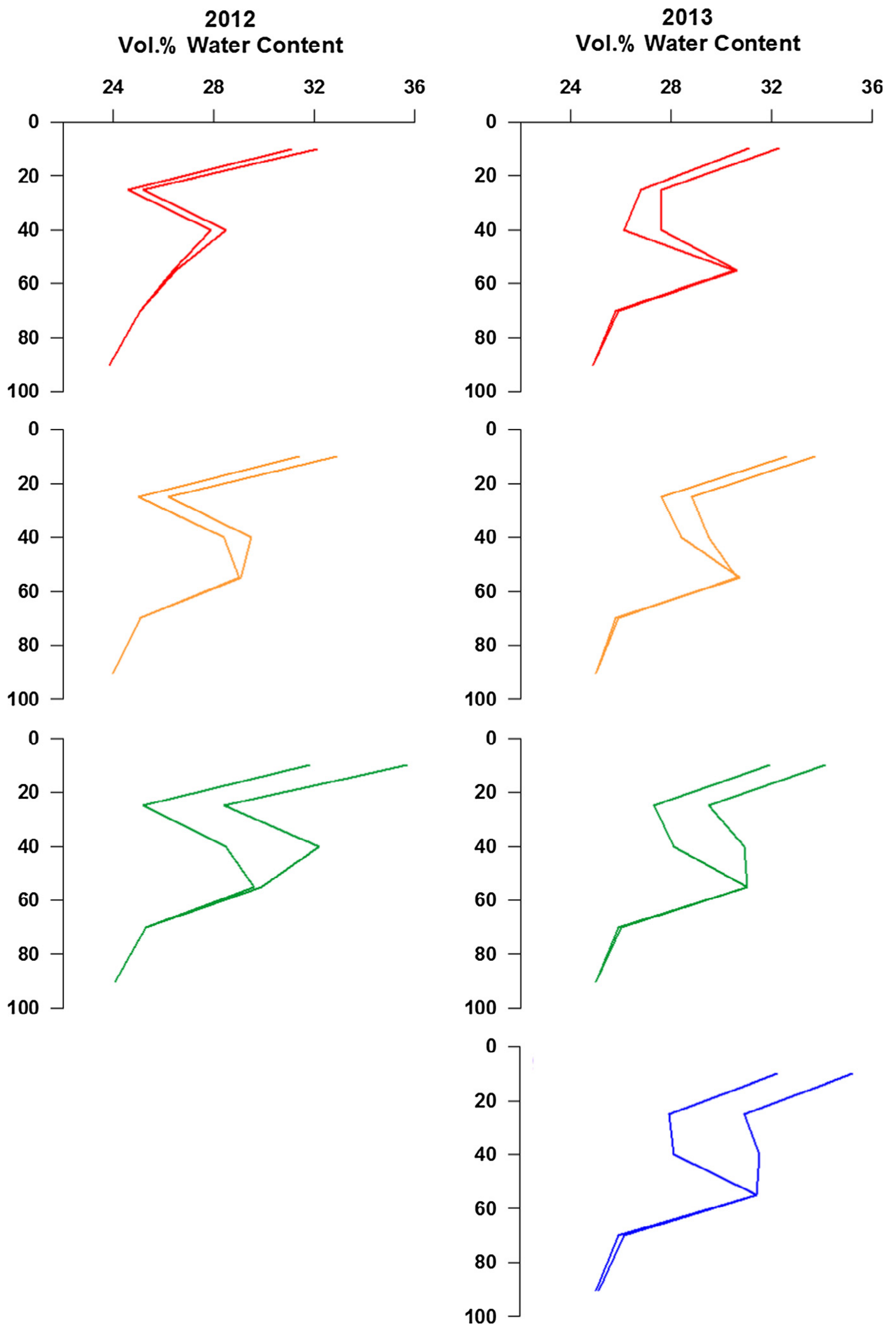

Fig. 4. Soil moisture profiles for the upper 90 cm of the VCB vadose zone, during the 3 melt stages of 2012 and the 4 melt stages of 2013.

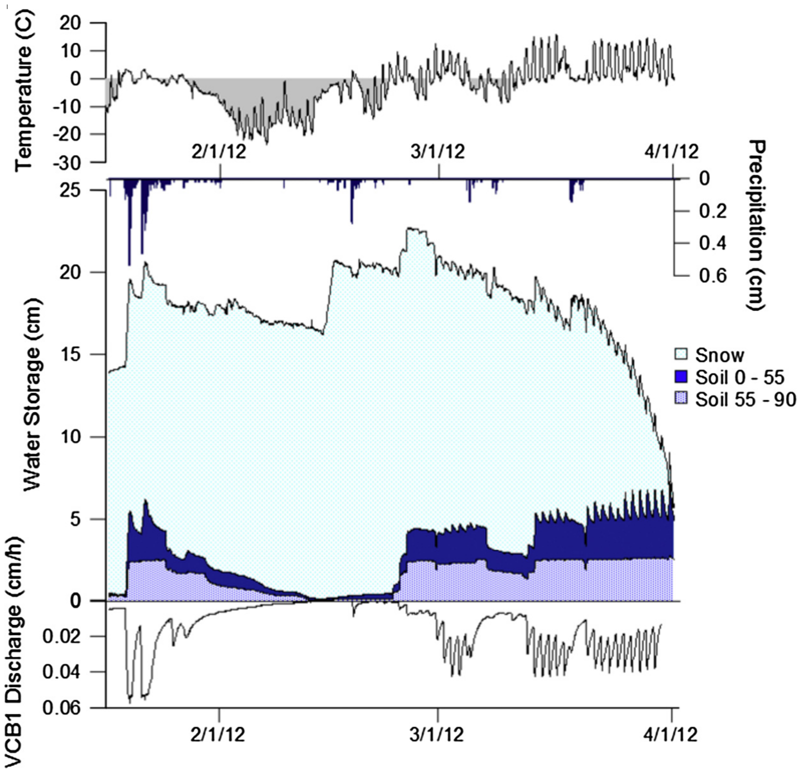

Fig. 5. VCB water balance showing precipitation (system input), snow water equivalent, soil water storage and VCB1 discharge (system output).

system showed very little reactivity to diurnal infiltration fronts, with limited drainage occurring only after two weeks without precipitation (Fig. 5). Due to their high water hold capacity, the clayey soils of the lower mineral system stored 1.5 cm of water for 21 days, implicating the lower mineral system as a source for base flow waters. During the three phases of melt, the 0–40 cm segment revealed itself to be a layer of flow-through with pronounced diurnal water-fluxes. Daily water storage changes of 0.5 cm during the first stage of melt in early March, increased to 1.5 cm during the final stage of melt in late March (Fig. 5). Also the perched lens of snowmelt in the upper mineral system drained over a 6-day period after snowmelt had finished.

4.3. 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 (Fig. 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 (Fig. 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.

5. 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 (Fig. 4) and the 2012 water balance (Fig. 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 h; 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 50 cm thickness of silty loam with a hydraulic conductivity (K) of 12.5 cm/hr, and a saturated 20 cm clayey loam layer with a K of .9 cm/hr (Clapp and

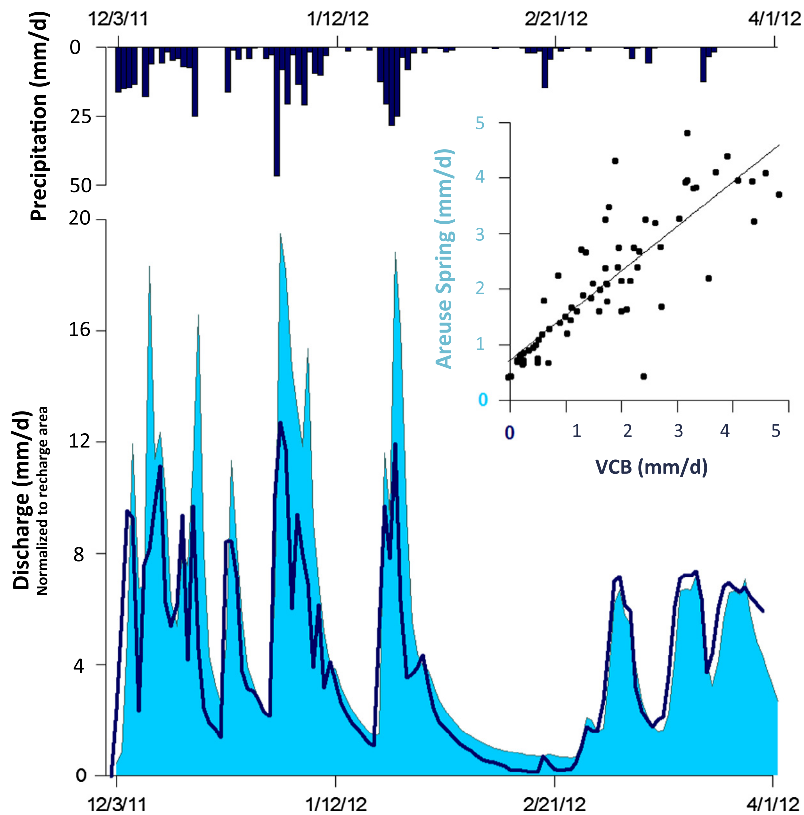

Fig. 6. The bottom graph depicts precipitation along with the synchronous evolution of VCB1 (dark blue line) and Areuse Spring (light blue fill) discharge (normalized to respective recharge areas) during 2012. The enclosed bivariate graph shows the approximately linear relationship between the two monitoring points. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Hornberger, 1978). That being the case, waters still transits the 53 vertical meters to arrive at VCB1 within 2 h, 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, Fig. 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 accumula-tion horizon must exist, allowing rapid infiltration. If distributed recharge were uniform, an infiltration front would be represented by peak water content arriving at 10 cm b.g.s., followed by peak water content arriving at 25 cm b.g.s., followed by peak water content arriving at 40 cm, 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 55 cm 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, Fig. 4), the hydraulic gradient toward 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 over-all 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 (Fig. 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 toward 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 0.54 mm/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 0.57 mm/d and 0.49 mm/d (data from Swiss low flow database), correspond very well to the base flow component estimated from Fig. 6. Compared to this base flow value, the dynamic water storage in the soil/epikarst of about 50 mm 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 700 m^2 and 1600 m^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 1000 m^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 1600 m^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 30 cm 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 90 cm of soil and epikarst approximately 5 m 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.

6. 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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