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Contribution of alluvial groundwater to the outflow of mountainous catchments

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Key Points:

- Contributions of mountainous alluvial aquifers to total catchment outflow were quantified
- During low flow periods, subsurface outflow can dominate over stream flow
- Alluvial aquifers of limited spatial extent can release substantial amounts of water

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Abstract Alluvial aquifers in mountainous regions cover typically a limited area. Their contribution to catchment storage and outflow is rarely isolated; alluvial groundwater discharge under gauging stations is generally assumed negligible; and hydrological models tend to lump alluvial storage with other units. The role of alluvial aquifers remains therefore unclear: can they contribute significantly to outflow when they cover a few percent of catchment area? Should they be considered a dynamic storage unit or merely a transmission zone? We address these issues based on the continuous monitoring of groundwater discharge, river discharge (one year), and aquifer storage (6 months) in the 6 km^2 alluvial system of a 194 km^2 catchment. River and groundwater outflow were measured jointly through ''coupled gauging stations.'' The contribution of alluvial groundwater to outflow was highest at the outlet of a subcatchment (52 km^2), where subsurface discharge amounted to 15% of mean annual outflow, and 85% of outflow during the last week of a drought. In this period, alluvial-aquifer depletion supported 75% of the subcatchment outflow and 35%of catchment outflow—thus 3% of the entire catchment supported a third of the outflow. Storage fluctua-tions occurred predominantly in the aquifer's upstream part, where heads varied over 6 m. Not only does this section act as a significant water source, but storage recovers also rapidly at the onset of precipitation. Storage dynamics were best conceptualized along the valley axis, rather than across the more conventional riparian-channel transect. Overall the contribution of alluvial aquifers to catchment outflow deserves more attention.

1. Introduction

In the context of climate change, there is renewed interest in processes that control low-flow rates and mechanisms of water storage in catchments [e.g., *McNamara et al*., 2011; *Sayama et al*., 2011]. Improving our understanding of storage processes is particularly important for mountainous regions, since these play an important role in delivering water to the lowlands, and because substantial changes in storage behavior are expected with the retreat of glaciers and earlier snowmelt [*Viviroli and Weingartner*, 2004]. In the absence of surface contribution, base flow originates from various subsurface units—essentially soils, bed-rock aquifers, and unconsolidated deposits [*Roy and Hayashi*, 2007]. The direct quantification of storage is often elusive due to the distributed nature of the processes involved and the difficulty to extrapolate from point measurements [*Birkel et al*., 2014; *Kirchner*, 2009; *McNamara et al*., 2011]. Therefore, dynamic water storage in catchments is usually inferred from river hydrographs by fitting conceptual hydrological models to discharge time series, or derived directly from discharge fluctuations without defining, a priori, a func-tional form of the discharge-storage relationship [*Kirchner*, 2009; *Young and Beven*, 1994].

Geological, geomorphological, and pedological features are major factors that explain base flow differences among mountainous catchments [*Sayama et al*., 2011; *Tague and Grant*, 2004, 2009]. Groundwater-level fluctuations in hillslopes [*Bachmair and Weiler*, 2014; *Gomi et al*., 2002; *Rodhe and Seibert*, 2011] or bedrock aquifers [*Gabrielli et al*., 2012; *Onda et al*., 2006] have been used, for example, to characterize dynamic stor-age. Recent research has hypothesized that bedrock aquifers act as dynamic reservoirs controlling base flow [*Andermann et al*., 2012; *Birkel et al*., 2014; *Sayama et al*., 2011; *Tague and Grant*, 2004, 2009; *Welch et al*., 2012]. It has been suggested that generally more water is stored in steeper catchments due to their extensive permeable zone above the stream level—the ''hydrologically active bedrock hypothesis'' [*Sayama et al*., 2011]. In some cases, it is possible to incorporate bedrock-groundwater levels in the calibration of catchment models [*Birkel et al*., 2014], or to extrapolate catchment storage and release from hillslope processes [*Broda et al*., 2012; *Broda et al*., 2014; *Uchida et al*., 2005]. Groundwater monitoring networks, however, are often too sparse or spatially too limited to infer coarse-scale storage dynamics from water-table fluctuations alone.

Alluvial aquifers are commonly acknowledged for their role in connecting hillslopes to the streams [e.g., *Jencso et al*., 2010], in transiting bank storage during floods [e.g., *Lin and Medina*, 2003] and for their influence on stream ecology [*Brunke and Gonser*, 1997]. Yet their contribution to seasonal storage in steep catchments is rarely considered. Even in recession analysis, it is rather a practical abstraction than a documented process, and the focus is mainly upon the widespread upstream riparian soils rather than aquifers as such. There are several explanations for this: (a) mountainous fluvial aquifers cover a limited area and their influence on the water balance may seem negligible; (b) they are situated in remote locations and poorly characterized; (c) they may be subject to complex interactions with surface or other subsurface flow systems; (d) the structure of most hydrological models exhibit a limited representation of vertical processes [*McDonnell et al*., 2007]; (e) continuous monitoring of subsurface alluvial discharge is uncommon; and (f) research has focused on processes ubiquitous over a range of scales, such as hillslope and bedrock storage [*Anderson et al*., 1997; *Durand et al*., 2005; *Gabrielli et al*., 2012; *Schilling*, 2009].

Nevertheless, alluvial aquifers lend themselves to measurements of water storage and fluxes. There is therefore scope to quantify experimentally their contribution to catchment outflow, and address the following questions: should alluvial aquifers in mountainous catchments be considered a dynamic storage unit in their own right, or are they essentially a transmission zone between the catchment and the streams? Despite their limited spatial extent, can they contribute substantially to total outflow? In particular, do the narrowest branches of the alluvial network play any significant role? To act as a “water pipe” or a “storage compartment,” these aquifers must exhibit high transmissivity and storage values, and be subject to large groundwater-level fluctuations. In many mountain settings, these characteristics may not be uncommon for the following reasons: (a) despite the known heterogeneity of alluvial deposits, high transmissivity values are typical of regions having undergone Pleistocene glaciations followed by Holocene periods of energetic runoff, such as the Alps, the Himalaya, or the North American Cordillera [*Woessner*, 2000; *Younger*, 2007, p. 117]; and (b) upland aquifers should be more likely to exhibit large fluctuations of groundwater levels, since they are located at higher altitudes, where recharge is typically interrupted by longer periods of freezing and snow storage.

Any landscape-scale conceptualization of upland alluvial storage remains challenging owing to the typical variability of geological settings. This variability controls an aquifer’s stratification as well as its connectivity to rivers, hillslope soils, and deeper groundwater [*Devito et al*., 1996; *Jencso et al*., 2010; *Malard et al*., 2002; *McGlynn and McDonnell*, 2003; *Vaudan et al*., 2005; *Welch et al*., 2012; *Wright et al*., 2005]. Two aspects, however, are worth emphasizing regarding such a conceptualization. First, the notion that groundwater flow paths are “normal to the channel” might be a useful idealization in some cases, but probably not a predominant feature in the field. *Larkin and Sharp* [1992], for example, found that alluvial systems with a channel gradient exceeding 0.0008, among other factors, tend to be dominated by underflow, that is, flow parallel to the valley axis [see also *Vidon and Hill*, 2004]. This underflow does not preclude the occurrence of concentrated spots of groundwater-river exchange, or nested flow systems: longitudinal flow superimposes rather than opposes such patterns [*Kasahara and Wondzell*, 2003; *Konrad*, 2006a; *Larkin and Sharp*, 1992]. Second, alluvial aquifers are typically less continuous than the river network. Subsurface heterogeneity and bedrock constraints in particular tend to produce visible and invisible points of longitudinal discontinuity along the alluvial network. In many cases, the alluvium is segmented by bedrock “highs” or “bottlenecks” that make the system resemble “beads on a string” [*Dent et al*., 2001; *Malard et al*., 2002; *Stanford and Ward*, 1993]. Thus, while such aquifers might be complex, there is potential for novel concepts to represent explicitly alluvial storage in hydrological models.

In this experimental study, we investigate whether storage depletion in a mountainous alluvial aquifer contributes significantly to total catchment outflow during low flow periods, or whether the aquifer only acts as a transit route for water stored elsewhere. Our analysis considers both surface and subsurface discharge, since the latter may dominate in low-flow conditions. If during such periods only surface flow is considered, the capacity of a catchment to yield water might be strongly underestimated. We address this question by evaluating continuously during one year the contribution of an alluvial aquifer to the total outflow of a mesoscale perialpine catchment. Since the alluvial plain covers no more than 3% of catchment area, it might be considered representative of other regions where alluvial storage and fluxes are overlooked. More specifically, this research evaluates: (a) the transience of total catchment outflow (i.e., river and groundwater discharge) at the catchment outlet and two other locations; and (b) the transience of total alluvial storage upstream of these points. This work is based on a simple yet appropriate conceptualization of the system that integrates measurements of river discharge, groundwater heads, as well as aquifer storativity and transmissivity.

2. Field Site and Methods

2.1. Field Site

The field site is located at the northern margin of the Alps, in the region of Emmental (Canton of Bern, Switzerland)—Figure 1. The study focuses on the alluvial aquifer of the upper River Emme and its main confluent the River Roethebach, which together drain a 194 km^2 area. The bedrock underlying the aquifer consists of well-cemented Tertiary sandstone and conglomerates as well as marls, altogether several hundred meters thick and referred to in Switzerland as Nagelfluh/Molasse [*Schweizerische Geologische Kommission*, 1980; *Blau*, 1984; *Blau and Muchenberger*, 1997]. The upper part of the catchment is composed of mesozoic flysch rocks and limestone. The latter is highly karstic and drains most, if not all, of the effective precipitation out of the topographic catchment (P.-Y. Jeannin, oral communication, 2015). It is, however, clearly disconnected from the aquifer of interest.

Climate characteristics in the catchment vary from the outlet (673 m a.s.l.) to the highest peak (2221 m a.s.l.): annual precipitation increases from 1300 to 2000 mm, and mean temperature decreases from 8° to 0°C, respectively [*Weingartner and Spreafico*, 1992−2010]. During the study period (2011), precipitation amounted to 1299 mm, and the longest period without precipitation lasted 42 days, from 20 October to 1 December (Figures 1 and 2—weather station C).

The alluvial plain is 22 km long and covers a 6 km^2 area (Figure 1). Most of it (80%) is 200 to 400 m wide, whereas the remaining upstream parts narrow down progressively. The plain may be divided in two areas of similar length, up and downstream of the major confluence between the Emme river and its tributary, the Roethebach (Figure 1): the main valley has a slope of 0.9% (the river enters this area after exiting deep gorges), whereas the tributary valley has a slope of 1.8%. Both rivers exhibit a coarse gravel-bed and numerous weirs ($>$60), about 1 m high, designed for erosion control. At the confluence, the Emme drains a larger catchment (116 km^2) than the Roethebach (52 km^2). Downstream of the confluence, the Emme is fed by smaller tributaries, which drain a 25 km^2 area before the catchment outlet. In 2011, the mean annual discharge of the two subcatchments was 2.96 m^3/s for the Emme, which is below the long-term average (4.37 m^3/s) and close to the minimum of the 38 year record (2.66 m^3/s), and 0.8 m^3/s for the Roethebach, which is between the long-term average (0.98 m^3/s) and the minimum of the 10 year record (0.57 m^3/s). At both locations, river discharge is typically higher during snowmelt, in April–May.

The alluvial aquifer is formed by quaternary deposits. In some places, the bedrock-sediment interface is as deep as 90 m below ground level. The bedrock outcrops locally at the edges of the valley, either in the riverbed or as cliffs about 30 m high. No springs and only minor fractures were observed on the outcrops. Springs located on the lateral hillslopes result presumably from water circulation in the superficial weathered bedrock, or unconsolidated rocks higher than the valley bottom. The valley filling consists of Pleistocene glaciofluvial sediments underlying more recent fluvial deposits. This alluvium is composed of sandy gravels and cobbles, with a variable proportion of silt, and contains lenses of coarse uniform gravel. Most geological logs show a layer of silty material, 1–3 m thick, overlying the bedrock, and thought to be lodgement till [*Schweizerische Geologische Kommission*, 1980; *Blau*, 1984; *Blau and Muchenberger*, 1997].

Hydrogeologically, the bedrock and the lower part of the alluvium are considered as an aquitard relative to the upper part of the alluvium. Saturated hydraulic conductivity (K_s) was measured in situ at 10 locations (Figure 1) through packer and conventional pumping tests prior to this study [*Blau*, 1984, 1991; *Blau and Muchenberger*, 1997; *Würsten*, 1991]. Results show that depth-averaged K_s range between $6 \cdot 10^{-4}$ and $5 \cdot 10^{-3}$ m/s. Higher values are found predominantly in the upper 30–35 m of the valley fill, whereas smaller values, around 10^{-4} m/s or lower, occur mainly deeper. This second facies is considered as the aquitard, the top of which has been mapped by the local Water Authority (*Amt für Wasser und Abfall, Kanton Bern*) based

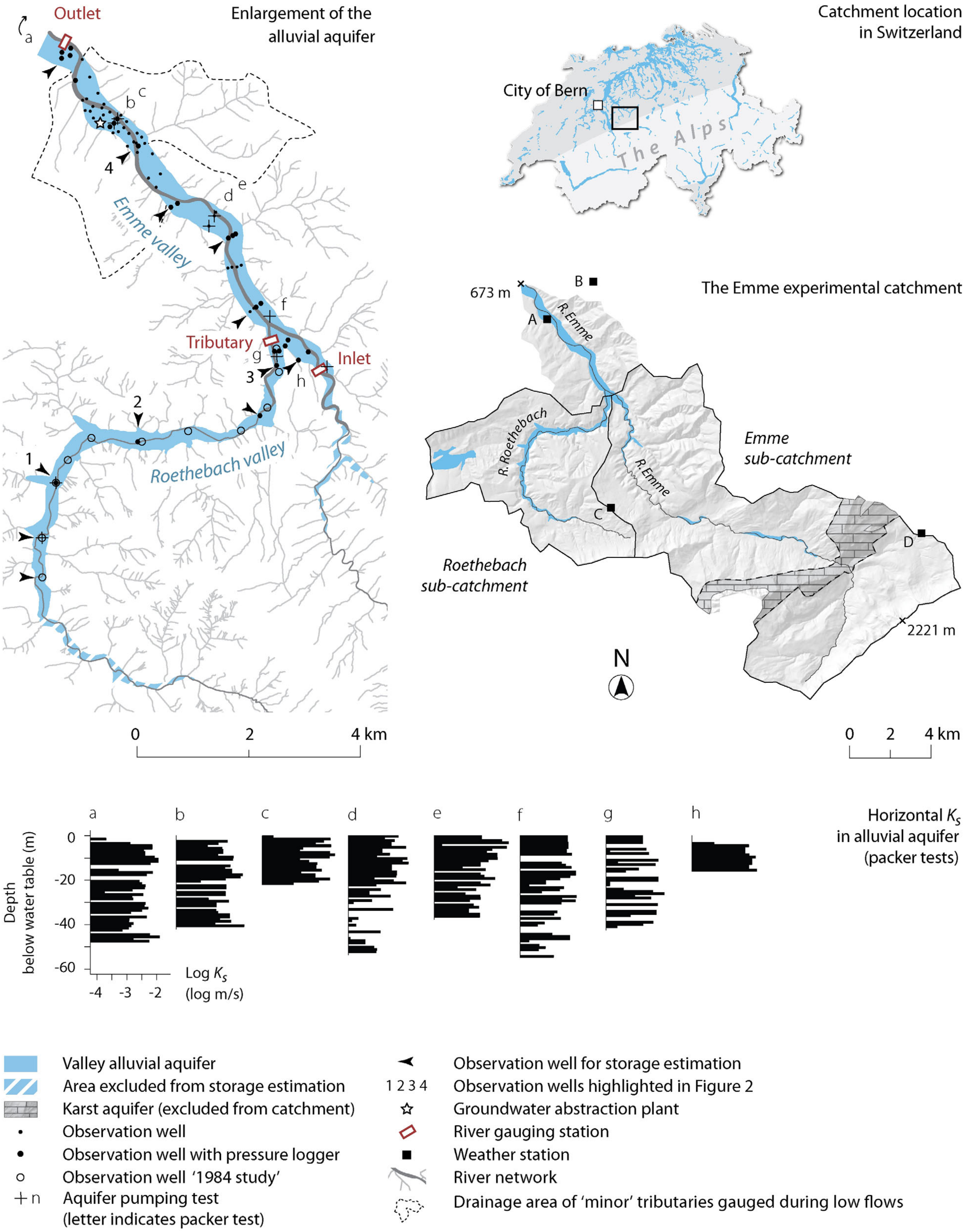

Figure 1. Location map, monitoring network, and hydraulic conductivity (K_s) measurements in the alluvial aquifer. The dashed line in the left map delineates the drainage area of "minor" tributaries gauged during the period of lowest flows (23 November to 2 December 2011). Weather stations A (755 m) and B (695 m), located in valleys, show near-identical temperature patterns, so as stations C (1170 m) and D (1383 m), situated higher in the catchment. Since the patterns of A and B are related to river discharge in a similar way, only temperatures from station B are depicted in Figure 2. Note that packer test "a" was conducted 1.2 km downstream of the outlet, in the same alluvial aquifer. Outlet coordinates: 46.924°/7.747°.

on a compilation of hydrogeological and geomorphological data [*Blau and Muchenberger*, 1997]. Values of aquifer storativity ($S = 0.1$ in average) and borehole logs indicate that the aquifer is predominantly unconfined. For several reasons, the bedrock and lateral hillslope soils in the immediate vicinity of the aquifer are not thought to contribute significantly to the alluvial groundwater system during low flows: (a) the bedrock is highly cemented, poorly fractured, and separated from the alluvium by a relatively low K_s sedimentary layer—interpreted as basal till; (b) the hillslope soils are shallow and the bedrock outcrops frequently at the edges of the floodplain; and (c) the hydraulic gradients inferred from piezometer arrays that are perpendicular to the valley axis do not indicate conspicuous lateral subsurface inflow. In any case, this assumption does not affect the study's conclusions, since the focus is upon alluvial storage fluctuations, irrespective of recharge processes.

One reason for which this aquifer is well documented is the presence of a groundwater abstraction plant (Figure 1). A number of river flow-gauging studies were conducted to assess its impact. They provide evidence of substantial river-groundwater exchanges in both the Roethebach and the Emme valleys [*Geotechnisches Institute*, 2005; *Blau*, 1984; *Gubelmann*, 1930]. A 2-D groundwater modeling study of the Emme valley suggests that 90% of recharge originates from river infiltration [*Poffet*, 2011]. The general water balance is clearly affected by pumping. In this study, however, it is the most certain of all fluxes, as abstraction rates are continuously recorded.

2.2. Methodology

2.2.1. General Approach

Our approach was devised to quantify, during 1 year, the transient contribution of alluvial groundwater to the outflow of the Emme catchment, with a focus on a major recession period. Two distinct quantities are derived from continuous measurements of surface and subsurface heads: (a) alluvial groundwater storage (6 months); and (b) total outflow, i.e., groundwater and river discharge (12 months). Additionally, the spatial pattern of exchange is examined based on vertical head differences between the river and the groundwater.

The alluvial system is characterized by a major confluence, and therefore the methodology is expanded to capture its specific contribution (Figure 1). Accordingly, the monitoring network includes three "coupled" gauging stations that measure simultaneously river and alluvial groundwater heads: one station is located at the terminal outlet, and two others immediately upstream of the confluence; these are referred to as the "Outlet," the "Inlet," and the "Tributary." Changes in groundwater storage are computed for both the entire aquifer and the tributary subsystem. The subsurface network consists of 45 observation wells, 26 of which were equipped with pressure loggers (Figure 1). It is denser in the vicinity of the pumping station, but one logger at least is placed every 2 km along the aquifer. In the Emme valley, the longitudinal spacing is shorter ($<$800 m), and the network expanded laterally with two to four wells per cross section. Heads were measured continuously from 5 January 2011 to 4 January 2012, and river discharge was measured in various flow conditions to parameterize the rating curve of the gauging stations (see section 2.2.5). Further, during the last 10 days of the driest period, all subsurface heads were measured manually; a sample of tributaries were gauged to estimate the lateral surface inputs to the Emme; and the river's elevation was measured to map its longitudinal profile. The meteorological data were provided by the network of *MeteoSwiss* (Station B) and Bern Canton (Station C)—see Figure 1.

2.2.2. Groundwater Observation Wells

We extended an existing well network [*Geotechnisches Institute*, 1991] by installing 31 wells using a Geoprobe® Direct-Push system. These wells are typically composed of a 10 m HDPE pipe (51 mm ID), with a screen extending from 4 to 9 m below ground, designed to capture the head of the phreatic surface. The cas-ing is surrounded by a coarse-sand pack sealed at the top with a layer of bentonite. The wells were surveyed with accuracies of 1 cm vertically and 5 cm horizontally. Twenty-six pressure loggers (Solinst® levelogger and Keller® DCX-22) recorded groundwater heads with a 10 min step, and four loggers (Solinst®) recorded baro-metric pressure. Several manual head measurements were carried out in 2011 to calibrate and control the logged data. Water levels were also recorded at two locations by the Water Utilities and the local Water Authority (Bern Canton). The latter in particular has a continuous record since 1984 (Well #1 in Figure 1).

2.2.3. Aquifer Storativity and Transmissivity

The aquifer's transmissivity (T) is used here to compute groundwater discharge at the gauging stations and storativity (S) to calculate changes in storage within the alluvial system. T and/or S were measured in previous studies at 10 locations, through conventional pumping tests or packer tests. Such tests "sample" a volume of the aquifer, and are therefore local but not strictly punctual. The packer tests in particular provided K_s measurements at 1 m intervals over vertical sections ranging from 16 to 55 m. These amount approximately to 350 K_s values distributed across eight locations (Figure 1).

Storativity had previously been measured in two wells located in the zone where by far the largest head changes occur (in the upper Roethebach valley)—see Figure 1 [*Blau*, 1984]. The values 0.09 and 0.11 are assumed representative of the aquifer, as the subsurface heterogeneity appears spatially stationary according to the 10 geological logs available throughout the study area [*Blau*, 1984; *Blau and Muchenberger*, 1997]—see also packer tests in Figure 1. Therefore, our estimation of storativity is based on the average $S = 0.1$, which is assumed constant in space.

For the same reason, based on packer tests carried out in eight wells along the Emme valley and the lowest part of the Roethebach valley, the saturated hydraulic conductivity (K_s) averaged over a valley section is also assumed to be spatially stationary [*Geotechnisches Institute*, 1991; *Blau*, 1984, 1991; *Blau and Muchenberger*, 1997]. For each of the three gauging stations, the valley's cross-sectional T (m^3/s) is calculated as

$$T_x = K_{s\text{-}mean} w_i e_i \tag{1}$$

where $K_{s\text{-}mean}$ ($1.8{\cdot}10^{-3}$ m/s) is the geometric mean of the profiles' K_s, this type of average being considered the best estimate for uniformly heterogeneous materials (mean $\pm$ standard deviation of Log10[K_s]: -2.74 ± 0.29); w_i is the width of the aquifer and e_i is the saturated thickness, both averaged over the area cor-responding to the groundwater gauging station (see section 2.2.5). The thickness is considered steady because the heads varied by less than 5% of the saturated thickness. The cross-sectional dimensions in meters are (width x depth): 330x26 at the inlet station, 180 $\times$ 28 at the tributary station, and 390x18 at the outlet station.

2.2.4. Groundwater Storage

Changes in alluvial groundwater storage, i.e., the net difference between discharge and recharge, are com-puted as the product of storativity, surface area, and head change in 12 adjacent zones based on one repre-sentative well per zone (Figure 1). These zones cover 88% of the aquifer's surface area: the narrow upstream parts of both the Emme and Roethebach valleys were discarded to avoid extrapolating unduly head data. The storage variations should therefore be interpreted as minimal rather than unbiased estimates.

Within each zone, when two or more wells per cross section are available, the most distant well from the river is selected, although the effect of this choice on the results is insignificant. Hourly storage variations are computed from August to December 2011, covering the main low-flow period, when all loggers were functioning. These variations are defined, irrespective of total storage, as

$$\Delta Stor = S \sum_i A_i \Delta H_i \tag{2}$$

where S (m/m) is the average storativity; A_i (m^2) is the horizontal surface area represented by well i; and H_i (m) is the hydraulic head measured in well i. The midpoints between the two neighbor wells, located upstream and downstream of well i, are used to define the area from which A_i is derived.

To compute total storage variations, the $\Delta Stor$ time series is first shifted so that its minimum value equals zero, then added to the volume of water stored in the aquifer when groundwater heads are at their lowest. This volume is defined as the product of S by the difference between the water-table (interpolated observed heads) and the aquitard elevation, available as a DEM [*Blau and Muchenberger*, 1997].

The drawdown cone at the abstraction plant, which has an insignificant effect on total storage fluctuations, and the narrow upstream ends of the aquifer, which are poorly characterized, are not represented. Last, the two uppermost wells, which were no longer operational in 2011 but had been monitored weekly in 1983/1984 [*Blau*, 1984], are included here through "virtual" head series on the following grounds: (a) the 1983/1984 study shows a distinct increase in the amplitude of head variations upstream of well #1; (b) this change in amplitude is consistent with the regional trend; (c) in all other aspects, these three time series are near-identical; (d) the heads recorded at well #1 in 1983/1984 and 2011 cover a similar range (Figure 3). Thus, first we determine the change in amplitude of the two uppermost wells relative to well #1, based on the 1983/1984 data. Second, assuming that the change in amplitude from one well to another is constant in similar flow conditions, we compute the two virtual head series by multiplying the "2011" series of well #1 by the respective factor. The historical time series are not illustrated here, but their high degree of simi-larity to the recent data makes these arguments reasonable.

2.2.5. River and Groundwater Discharge

Each coupled gauging station monitors: (a) river discharge through a conventional gauging station; and (b) groundwater discharge through measurements of transmissivity and hydraulic gradient, combined through Darcy's law. Total outflow is therefore defined as the sum of river (Q_{SW}) and alluvial groundwater (Q_{GW}) discharge:

$$Q_{total} = Q_{SW} + Q_{GW} = Q_{SW} + T_x \nabla H \tag{3}$$

where T_x is the local cross-sectional transmissivity (m^3/s), and ∇H is the local longitudinal hydraulic gradient (m/m). At each station, the gradient's magnitude and direction is derived from head observations in three wells forming a triangle, following *Devlin*'s [2003] approach. The triangles cover most of the aquifer's width, and have an average side of 220 m (Figure 4). They are located slightly upstream of the river gauging facility at the Tributary and Outlet stations, and slightly downstream at the Inlet station (Figure 1), but are considered close enough for local estimations of total discharge. This triangle-based approach is mainly justified by: the nature of the alluvium, which forms a single rather than a layered aquifer; the geometry of the aqui-fer, elongated and thin, with a predominant longitudinal flow-direction; and the documented assumption of restricted deep groundwater inflow into the alluvium. Further, the linear representation of the water table within the triangles is considered here a reasonable assumption. The triangles were designed to be large enough to cover the valley width, but sufficiently small to represent a valley cross section at the aqui-fer scale. The final well locations, however, were constrained by factors such as drilling authorizations and accessibility. The gauging station at the outlet may be considered the best setup.

In order to monitor river discharge, we installed two gauging stations (Outlet and Tributary) and used another one (Inlet) operated by the Swiss Federal Office for the Environment, FOEN (LH2409, Eggiwil, Heidb üel). All stations measure the water-surface elevation immediately upstream of a weir with a 10 min time step. The rating curve of the FOEN-station is updated several times a year since 1975, and a quality check is conducted prior to data delivery. The other stations consist of a screened PVC tube anchored to the bank and equipped with a vented pressure logger (STS® DL/N 70). The rating curves are derived from 11 discharge measurements at the tributary and 22 at the outlet-station; measurements were conducted in various flow conditions using the velocity-area method in average flow conditions (ADV Sontek® Flow-Tracker), and the instantaneous dilution method during low flows (tracer: NaCl) and high flows (tracer: Rho-damine WT).

2.2.6. Discharge of Tributaries in the Emme Valley

At the end of the driest period, in steady hydrological conditions, the contribution of tributaries between the three gauging stations was estimated by gauging a sample of streamlets draining 9.0 km^2 out of 24.6 km^2 in total (Figure 1). Several channels were dry and nine tributaries were gauged between 23 November and 2 December, using either a graduated bucket or the dilution method (tracer: NaCl). During this period, the discharge of the Emme varied only by 3%, so the data set is considered representative. The total contribution of tributaries was estimated at 71 L/s after multiplying the mean measured specific dis-charge (L/s/m^2) by the total drainage area. For this period, the water balance of the Emme valley is therefore closed, since the coupled gauging stations measure the remaining components of the balance. Despite some uncertainty, this provides a means to evaluate the consistency of the parameters S and $K_{s\text{-}mean}$.

2.2.7. Uncertainties of the Calculations

The main objective of the study is to relate alluvial storage depletion to total catchment outflow. This section discusses the associated uncertainties. Two specific aspects of the study design are thought to promote reliable estimations: (1) the optimal location of both the ''outlet'' gauging station and the points where robust S measurements were available; and (2) a water balance based on a comprehensive data set in order to verify the consistency of the estimated fluxes.

Regarding point (1), total catchment outflow was quantified in an area of low aquifer transmissivity, where groundwater discharge is relatively low (see section 3.3, Figures 5 and 7). Further, a large fraction of groundwater is abstracted upstream and the pumping rates are known. Thus, catchment outflow does not rely heavily on the estimation of K_s, which is probably the most uncertain parameter as shown with a sensitivity analysis (Figure 6). For the tributary subcatchment, the uncertainty on the outflow is higher, as groundwater discharge contributes substantially to total outflow. However, the outflow should, in principle, not be smaller than the storage depletion rate, which strongly constrains the estimated minimal groundwater discharge. Regarding alluvial storage, the most robust estimations of storativity are related to the zone where the highest head changes occur. This area accounts for 55% of total storage variations while it covers only 16% of the aquifer's extent.

Regarding point (2), we estimated the contribution of small lateral tributaries in the Emme valley, based on field measurements, in order to compute a simple a water balance for the last week of the main drought, when all fluxes were steady. The balance was calculated as $Q_{outlet} + Q_{pump} - Q_{Inlet\&Tributary} - Q_{lateral} + \Delta STOR$, where Q_{outlet} *and* $Q_{Tributary\&Inlet}$ are the total fluxes (river and groundwater) measured at the respective stations; Q_{pump} is the groundwater pumping rate; $Q_{lateral}$ is the discharge of the small tributaries located between the stations; and $\Delta STOR$ is the storage depletion rate. The error on this water balance (+28 L/s) represents only 6% of total outflow ($Q_{Outlet}+Q_{Pump}=$ 458 L/s). Further, an 11% change in T values appears sufficient to "cancel" this error, suggesting that the estimated fluxes and parameters are relatively consistent.

Although we consider these strategic aspects as a key to characterize and reduce uncertainties, we also propagated measurement errors. The estimated uncertainties, which are systematically based on the best knowledge of the site and methods but inherently subjective to some degree, are as follow:

1. Aquifer parameters: 25% for $K_{s\text{-}mean}$, 25% for S in the Emme valley, 15% for S in the Roethebach valley. These values are based on pumping tests, geological logs, and results of the initial 2-D groundwater modeling study [*Poffet*, 2011]. The uncertainty on $K_{s\text{-}mean}$ is based on the assumption that the eight vertically averaged K_s measurements represent a sample of an aquifer cross section. In this context, the uncertainty is defined as the standard uncertainty of the mean of the vertically-averaged K_s. Note that $K_{s\text{-}mean}$ differs "only" by 22% from the average K_s estimated by calibration of the initial model. For S, which is particularly costly to determine, the two values deviate by 10% from the mean.
2. Flow rates: 10% for river discharge (average and low flows), 3% for pumping rates, and 30% for the discharge of "minor" tributaries (low flows). These values are based: for river discharge on the examination of the stage/flow rating curves; for pumping rates on available information from the Water Utilities; and for tributary discharge on the authors' estimation given the methodology reported above.

2.2.8. Longitudinal River Profile

Mapping the elevation of the river's water-surface required some accuracy, as the sign of GW-SW head differences must be known to identify potential areas of groundwater recharge and discharge. In this study, the 16 km profile of the river's water-surface was obtained by linearly interpolating known water levels at weirs, which are geomorphologically the most prominent vertical features. Changes in river stage recorded at the Inlet and Tributary stations are then assigned at each time step to the entire profile of the respective rivers. Except during overbank flooding, which has not occurred during the study, the river width may be considered spatially uniform. In average conditions, the width is defined by the bank armoring structures, whereas in low flows it is shaped by relatively regular gravel bars. The profile's uncertainty was evaluated based on 14 independent water-level measurements conducted at intermediate locations leveled with 1 cm precision (April 2011, low-flow conditions). The profile's bias (+0.22 m) was then corrected and its final accuracy evaluated (standard deviation: ±0.45 m).

The initial interpolation relied on two data sets: site plans reporting weir-top elevations [*Blau*, 1984] and field measurements of the water-surface drop at the weirs. Measurements were conducted toward the end of the driest period (27 November) at 24 out of 31 weirs in the profile's lower 8 km. The water depth over the weirs was very low (ca. 1 cm); therefore the upper water level was considered equal to weir-top eleva-tion. The head drop at the seven missing weirs is estimated by linear regression (R^2: 0.95) based on a previ-ous survey of all weirs conducted in similar low flow conditions, on 2 April. In the upper 7.9 km of the Roethebach valley, less precision is required as the groundwater table becomes increasingly deeper (2–30 m below river-level). Here, the low-flow river profile was considered equal to the riverbed profile derived from the plans of 28 weirs.

2.2.9. River Temperature and Electrical Conductivity Profiles in the Emme Valley

River temperature and electrical conductivity were used to delineate finely the spatial pattern of groundwater discharge over a limited but exemplary reach (4.6 km). The primary aim is to understand the variabili-ty of groundwater discharge, and therefore evaluate to what extent river discharge measurements are sensitive to a station's location. The survey was conducted on 23/24 November in steady hydrological con-ditions, by measuring these parameters at 20 m intervals on each side of the channel, using conventional calibrated multimeters.

2.2.10. Summary of Core Assumptions

The study's main assumptions are summarized as follow: (1) the large-scale average K_s is constant in space;(2) S is constant in the zone where the highest water-table fluctuations occur, and irrelevant where the water table is relatively stable; (3) stage variations along the Emme and the Roethebach are equal to the variations recorded at the respective gauging stations, based on the assumption of a uniform river section;(4) during the low-flow period, the subsurface contribution of the underlying bedrock and lateral hillslope soils is negligible relative to the longitudinal aquifer flux or to river infiltration rates; and (5) storage varia-tions in the narrowest upstream parts of the aquifer are negligible.

3. Results and Discussion

This section focuses first on the raw time series of river discharge and groundwater heads, as well as the spatial and temporal patterns of river-groundwater exchange. Then we analyze the contribution of alluvial groundwater storage to catchment outflow, before discussing the implications of this study.

3.1. River Discharge

In 2011, river discharge at the inlet, outlet, and tributary stations show a similar pattern consisting of high flows caused by spring snowmelt and summer precipitation, and recession periods in February, April, and October/November (Figure 2b). At the end of the last and longest dry spell, however, a substantial difference appears between the three hydrographs. At the inlet, the Emme recedes following a nearly exponential decay, indicating that the Emme subcatchment behaves as a simple linear reservoir, irrespective of storage processes. The lowest daily flow, recorded on 25 November, was 68 L/s.

At the Tributary station, the Roethebach River recedes more or less steadily down to 100 L/s. The discharge remains then stable for 2 weeks, before decreasing abruptly to values ranging from 13 to 50 L/s for the last 3 weeks of the drought. The steadiness of discharge is perhaps due to groundwater discharging into the stream immediately upstream of the gauging station. As shown in Figure 3, groundwater heads in this area are very close to river stage. Following this steady period, the hydrograph is remarkably similar to the air-temperature pattern (near 0°C), suggesting that the upstream expansion and contraction of frozen reaches controlled river discharge (Figure 2a: lines a,b). This phenomenon may have similarly influenced groundwater exfiltration into the river. At this time, the upmost 4 km of the river were completely frozen, and the following 2 km partly frozen. In addition, the hourly record of river stage at the tributary station showed daily variations of 3 cm, with a minimum before sunrise (no indication of water freezing locally). This pattern was not related to instrument noise nor temperature sensitivity.

At the outlet, the Emme receded exponentially down to a plateau around 175 L/s, during the drought's last 14 days—Figure 2b. (The recession curve does not appear as a straight line on the log-scale plot, because the plateau is a nonzero value.) Previous studies in this specific area suggest that the plateau results from the thinning up of the aquifer in the vicinity of the outlet [*Geotechnisches Institute*, 1991; *Blau*, 1991]: the decrease in T_x induces groundwater discharge to the river at a rate seemingly steady over the considered time period (Figure 3a and 7). We evaluated whether this behavior might result from variations in ground-water pumping rates, and concluded that these affected possibly the value of the plateau, but not its occurrence.

3.2. Transience of the Groundwater Table

Groundwater levels vary in time following essentially one of two distinct patterns: (1) large seasonal fluctuations (2–7 m) and a damped response to surface flow events; or (2) small seasonal fluctuations ($<$2 m) and a near-immediate response to surface flow events (Figure 2c). The largest and smoothest fluctuations occur in the upstream part of the tributary valley, particularly in the uppermost well where heads varied over 6 m in 2011 (Figure 2c, series 1)—and 17 m throughout the 30 years record (Figure 3). In this well, the head variations suggest a daily to weekly lag in the response of groundwater level to river discharge (Figure 2c, series 1). This occurs despite the river's proximity (45 m) and hydrogeological properties similar to the rest of the aquifer. Thus, it is likely the result of water transfer across a thick unsaturated zone. The next well downstream is char-acterized by both seasonal and event-related fluctuations spanning 2.5 m (Figure 2c, series 2). The break-in-slope marked by line (a) is probably due to the freezing of the river that abruptly reduces surface water infil-tration into the aquifer, thus increasing the rate of storage depletion. The third well shows exclusively event-

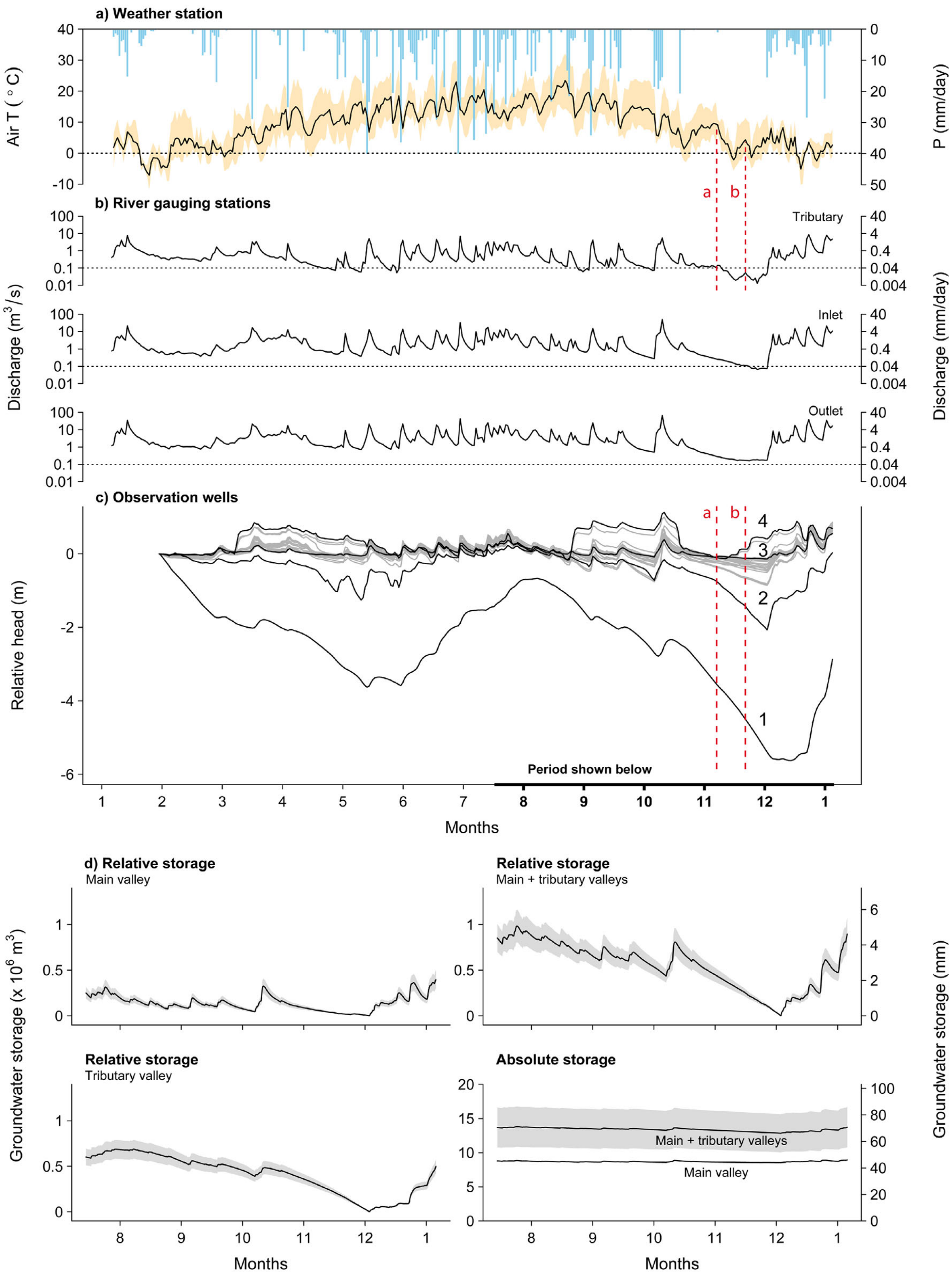

Figure 2. (a) Daily precipitation (Station C) and mean, min., max. air temperatures (Station B), (b) river discharge, (c) groundwater heads, and (d) alluvial groundwater storage. Heads are set to 0 m on 29 January 2011, and storage is computed for the period where all loggers were operational. In Figure 2c, the black lines represent head series for which the location is represented in Figure 1 (numbers 1,2,3,4). All other series are in gray. In Figure 2d, the shaded area represents the uncertainty ranges for the tributary valley ($S \pm 0.15$) and the main valley ($S \pm 0.25$). "Total storage" refers to the total amount of water stored in the aquifer. As the Y-scale is too large for a proper depiction of fluctuations, the series are also plotted in relative values. The left axis of the storage plots represents a volume, and the right axis an equivalent precipitation height over the Emme catchment (196 km^2).

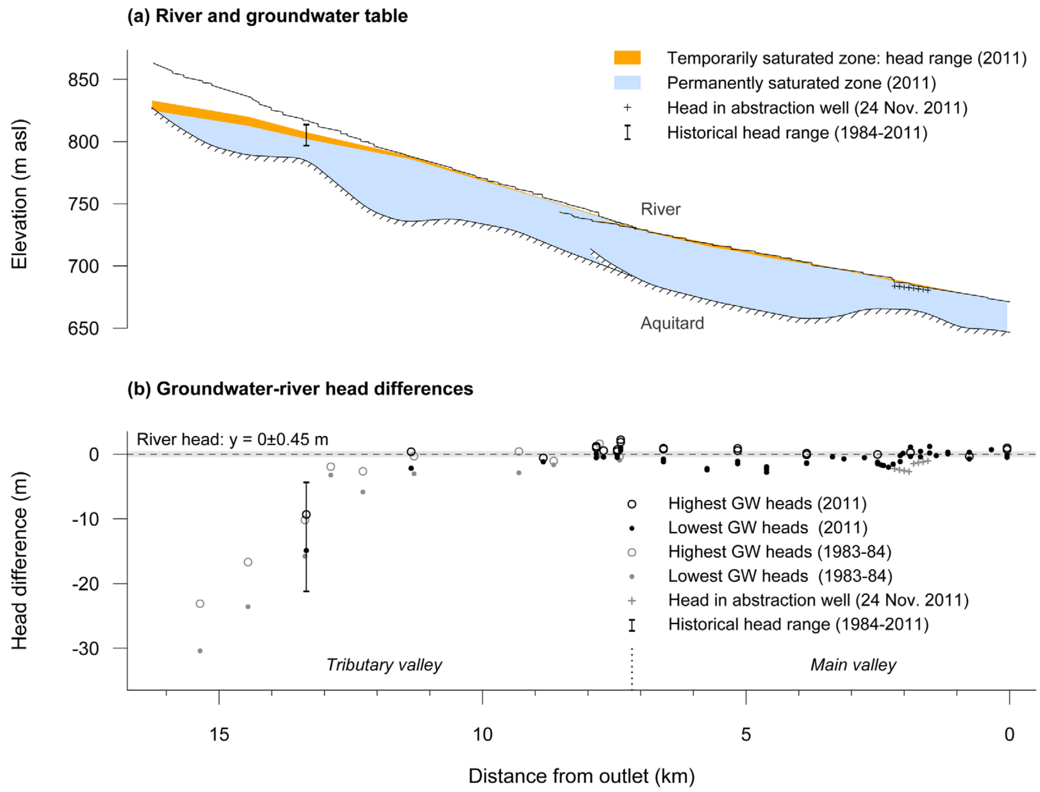

Figure 3. (a) Longitudinal profile of the Emme and Roethebach rivers and (b) groundwater heads in 2011. Figure 3a depicts the minimum and maximum groundwater levels; the thickness of the line represents river stage fluctuations. Figure 3b shows the minimum and maximum vertical head differences between the groundwater table and the river stage. The shaded area represents the standard deviation of the errors on river stage estimation. Heads in low-flow conditions are derived from manual measurements, whereas in high-flow condi-tions they were recorded by pressure loggers. The interpreted aquitard profile is defined as the top of the low-permeability sediment layer overlying the bedrock in the middle of the valley. The historical head range corresponds to well #1 (see Figure 1).

related variations spanning 1 m (Figure 2c, series 3). In the main and lower tributary valleys, head variations are smaller, more frequent, and clearly associated to individual flow events. Over this downstream segment (9 km), the groundwater table fluctuates essentially over 1 m, although three wells situated within 400 m of the abstraction plant are obviously influenced by pumping rates (Figure 2c, series 4).

3.3. Patterns of River-Groundwater Exchange

In 2011, the groundwater table is largely below river level in the uppermost 4 km, and only slightly and partly lower downstream (Figure 3). In this upstream reach, groundwater heads remain beneath river-level throughout the year, and the unsaturated zone's thickness increases to 27 m in the upstream direction. The head in well #1 never exceeded riverbed elevation throughout the 30 year record, despite variations of 17 m, thus suggesting that the upper 2 km of the aquifer are an area of recharge only. In the main and lower tributary valleys, infiltration prevails in most cases, although not in the days right after a rain event nor near the confluence or the outlet. In these last two areas, the cross-sectional transmissivity decreases longitudinally for two different reasons, thus promoting groundwater discharge to the river: (1) at the confluence, the aquifer's cross-sectional area is 30% smaller than the area of the two valleys combined; and (2) upstream of the outlet, the cross-sectional area decreases by 35%, due mainly to an increase in elevation of the aquitard-top (Figures 3a and 7). In the immediate aftermath of an event, groundwater heads may exceed surface heads at several locations (Figure 3b). A few days after peak-flow, however, the general pat-tern of exchange is restored.

3.4. Alluvial Groundwater Storage

The amplitude of variation in alluvial storage is about twice as high in the tributary valley than in the main valley despite the smaller aquifer size (Figure 2d). The tributary aquifer covers a smaller surface area, is narrower, and the mean river discharge is about 25% of the Emme discharge, but the head fluctuations are far larger than in the main valley. Note that the changes in absolute storage are small, suggesting that the aquifer is far from a state of depletion (Figures 2d and 3a).

The difference in storage amplitude between the two valleys is likely related to the unsaturated zone's thickness. A deeper groundwater table in the tributary valley allows for larger head changes and for longer residence times before surface water reaches the saturated zone; it also restricts the area of groundwater discharge to the stream. In contrast, a thinner zone favors the temporary rise of the groundwater table above river level, thus increasing the area of exfiltration, and promoting a faster equilibrium between groundwater and surface-water levels. Interestingly, the depletion curves of the two valleys show a curva-ture of opposite sign (Figure 2d). In the main valley, the depletion rate decreases with time, while it increases in the tributary valley. In the main valley and in the aftermath of high flows, groundwater heads are at or above streambed levels, thus favoring rapid storage depletion (Figure 3b). As the groundwater table drops, the depletion rate decreases. In contrast, along the upgradient half of the tributary valley, river infiltration prevails throughout the year (11.2–16 km, Figure 3b). As stream flow recedes, the infiltration rate likely decreases, and storage depletion accelerates.

In any case, whether one considers the entire aquifer or only the Roethebach subsystem, groundwater storage depletion does not proceed exponentially, as frequently assumed in conceptual hydro(geo)logical models. Previous studies have also shown that an exponential depletion can mainly be expected for long periods without recharge, which is uncommon in humid regions [*Cuthbert*, 2014].

3.5. River, Groundwater, and Total Outflows

At all three gauging stations, and despite rapid variations of river flow, groundwater discharge is remarkably steady throughout the year. The hydraulic gradients are similar to both the valley-slopes and the surrounding head gradients, in direction and magnitude (Figures 4 and 5). When considering the entire aquifer, flow appears predominantly parallel rather than normal to the river. As indicated by the initial modeling study, groundwater/river exchanges superimpose this longitudinal pattern, and are mainly controlled by the weirs, river meandering, aquifer transmissivity, and riverbed permeability [*Poffet*, 2011].

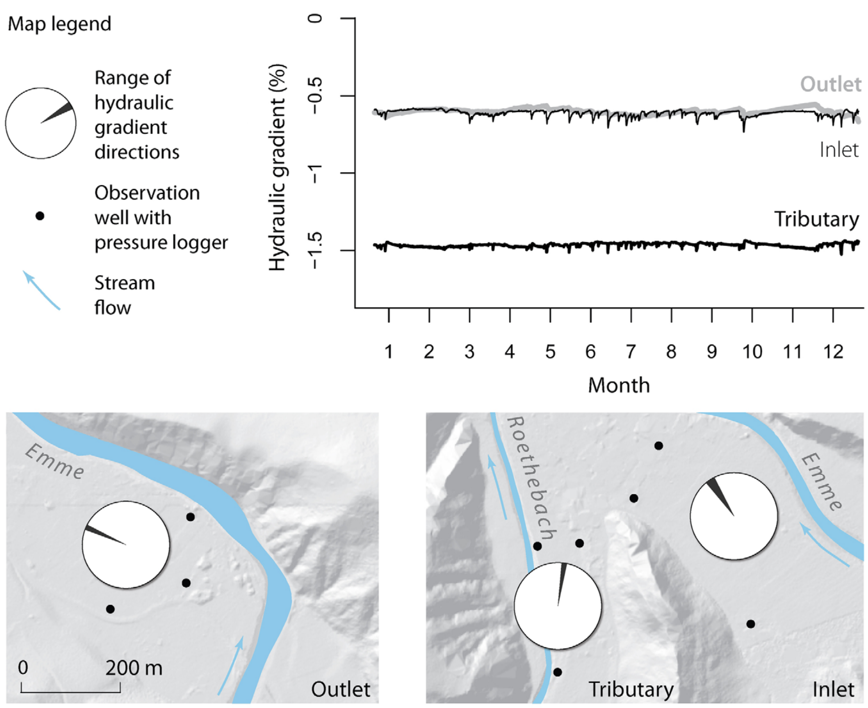

Figure 4. Groundwater hydraulic gradients and directions at the inlet, outlet, and tributary-gauging stations during 2011. The black sectors represent the range of groundwater flow directions. The errors on the time-series gradients are insignificant at plot scale.

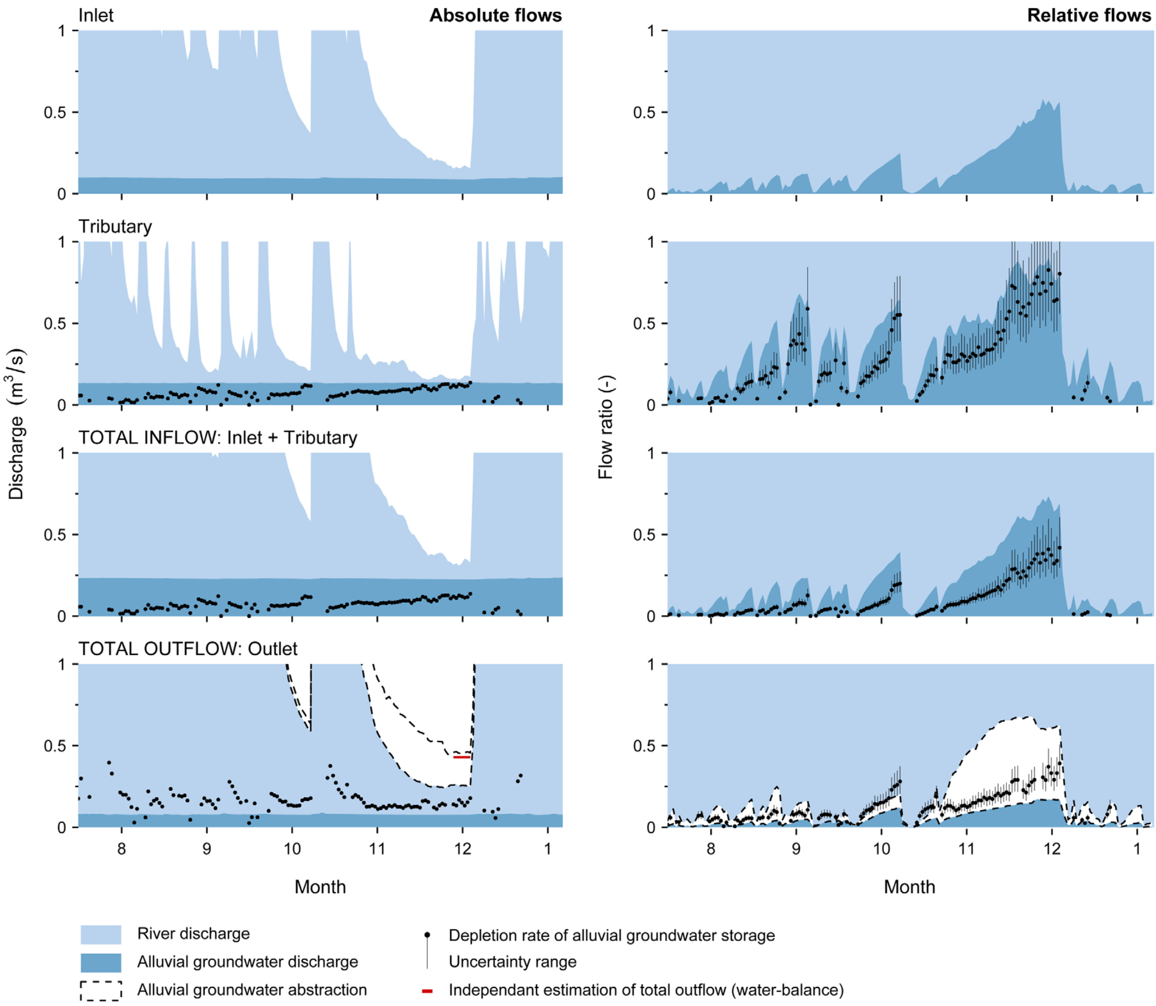

Figure 5. River and alluvial groundwater discharge measured at the coupled gauging stations (shaded areas), and alluvial storage depletion-flow estimated for the contributing area (dots), in absolute and relative terms. For clarity, error ranges have only been depicted for the estimations of relative depletion flow. These ranges, however, reflect the uncertainty of all fluxes shown in a plot. Dots are interrupted by increases in storage. The area enclosed by the dashed line represents the mean daily flow abstracted at the groundwater pumping station.

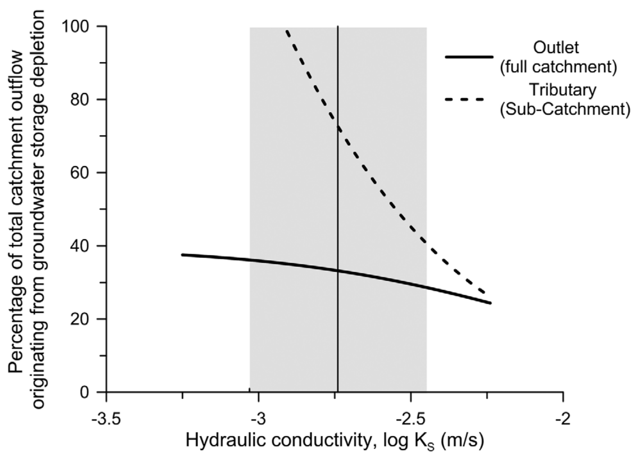

Figure 6. Sensitivity to the hydraulic conductivity K_S of the percentage of catchment outflow at tributary and outlet originating from alluvial-aquifer storage depletion. Vertical line: mean K_S. Gray area: standard uncertainty of K_S.

Groundwater contributes substantially to catchment outflow in dry conditions, although river discharge dominates the flow budget on an annual basis (Figure 5). During the last week of the drought (26 November to 2 December, referred to hereafter as the "driest week"), the groundwater flux at the inlet amounts to 88 L/s, representing a 55% contribution to the total flow at this station (160 L/s). At the tributary station, the groundwater's contribution is larger both in absolute (137 L/s) and relative terms (83%). As a result, the total fluxes at the tributary and inlet stations are near-identical (160–167 L/s) despite the significant difference in river flows. In addition, both stations together recorded a groundwater flux of 225 L/s, whereas the outlet station recorded a groundwater flux of 78 L/s. This decrease in groundwater discharge is caused by a decrease in transmissivity due to the thinning up of the aquifer near the outlet, inducing groundwater exfiltration into the river (Figures 3a and 7).

The overall consistency of the flow rates was verified by calculating a water balance for the main valley aquifer based on values averaged over the last 7 days of the drought. This period was selected because fluxes were then at their steadiest, and because it covered the gauging survey of the minor tributaries (Figure 1). In these conditions, the sum of all inflows and the storage depletion rate ($Q_{Inlet\&Tributary} + Q_{lateral} - \Delta STOR$) amounts to 430 L/s and the sum of all outflows to 458 L/s ($Q_{outlet} + Q_{pump}$); Figure 5. The error on the water balance (28 L/s or 6% of total outflow) may result from measurement uncertainty and/or minor lateral subsurface inflow. It is however quite low, suggesting that fluxes have been reasonably well quantified. The accuracy of total outflow at the outlet is relatively high, as groundwater discharge (the flux with the largest uncertainty) only accounts for 17% of the outflow (Figure 6). Thus, although the compliance of the mass balance does not imply that T and S are the exact true values, these parameters are in mutual agreement.

Coupled gauging stations provide two types of information. Quantitatively first, they capture the outflow missed by a conventional river gauging station. The best illustration is the tributary station, where groundwater discharge represents 16% of the annual outflow, and 83% of outflow over the driest week. Second, coupled stations provide better insight into the functioning of river-aquifer systems, although any interpretation requires caution as the partition of flow may change over short distances. At the tributary-station, for example, the surface drought did not affect the steadiness of groundwater flow, thus reflecting the ''buffering'' capacity of the aquifer at this location.

3.6. Contribution of Alluvial Storage to Catchment Outflow

While alluvial groundwater contributes steadily to catchment outflow, it is only a net source when storage decreases. The rate of storage depletion, or depletion-flux, when expressed as a fraction of total outflow, reflects the significance of alluvial storage versus other types of catchment storage (e.g., bedrock and hillslope)—Figure 5. Note that the ''groundwater depletion rate'' is different from the aforementioned ''groundwater discharge.'' The former integrates storage depletion *upstream* of a gauging station whereas the latter refers to flow measurements *at* the station's location. During the driest week, storage depletion sustained an important part of total outflow at both the tributary (73%) and outlet stations (33%). This is considerable given that these alluvial aquifers cover only 5% and 3% of their respective catchment areas. In terms of uncertainty, the percentage of outflow originating from storage depletion at the outlet shows a very low sensitivity to K_S (Figure 6). If K_S is varied within its standard uncertainty, the percentage only varies from 29 to 36%. This is explained by the low contribution of groundwater discharge to total outflow at the outlet. At the tributary station, the relative contribution of groundwater is higher, and therefore so is the sensitivity to K_S. During the driest week, the fraction of flow originating from alluvial storage in the 52 km^2 subcatchment may be as low as 50% and as high as 100%, as indicated by the error bars in Figure 5. It is therefore possible that the outflow is entirely supported by alluvial-storage depletion.

The temporal trend of the depletion flux differs between the lower and upper part of the aquifer. Downstream, the depletion-flux is highest in the aftermath of flow events, when the groundwater table is extensively, yet temporarily, above river levels and the exfiltration area is large. Upstream, variations of the exfiltration area are smaller, and thus the depletion-flux increases as the recharge flux decreases throughout the recession period. During the driest week, the depletion-flux in the tributary valley (121 L/s) is 4 times larger than in the main valley (31 L/s). Normalized with the surface area, the tributary valley aquifer yields 49 L/s/km^2, whereas the main valley aquifer only releases 9 L/s/km^2. Previous studies have shown that nar-row alluvial aquifers in mountainous terrain may play an important role in routing water toward valley-bottom aquifers [*Smerdon et al*., 2009]. This study reveals that they may further provide substantial dynamic storage. The tributary system is not only a reliable source of water during the drought; it recovers also rap-idly in the aftermath—it shows high ''resilience'' (Figure 2d). Within 1 month after the dry period, following 170 mm of precipitation, over half of the depleted volume was recovered again. The tributary aquifer shows a high-recovery rate compared to the main valley: the contributions of recharge events cumulate and are retained in storage due to the thicker unsaturated zone and the longer distance to drainage. In contrast, in the main valley, storage gains dissipate quickly.

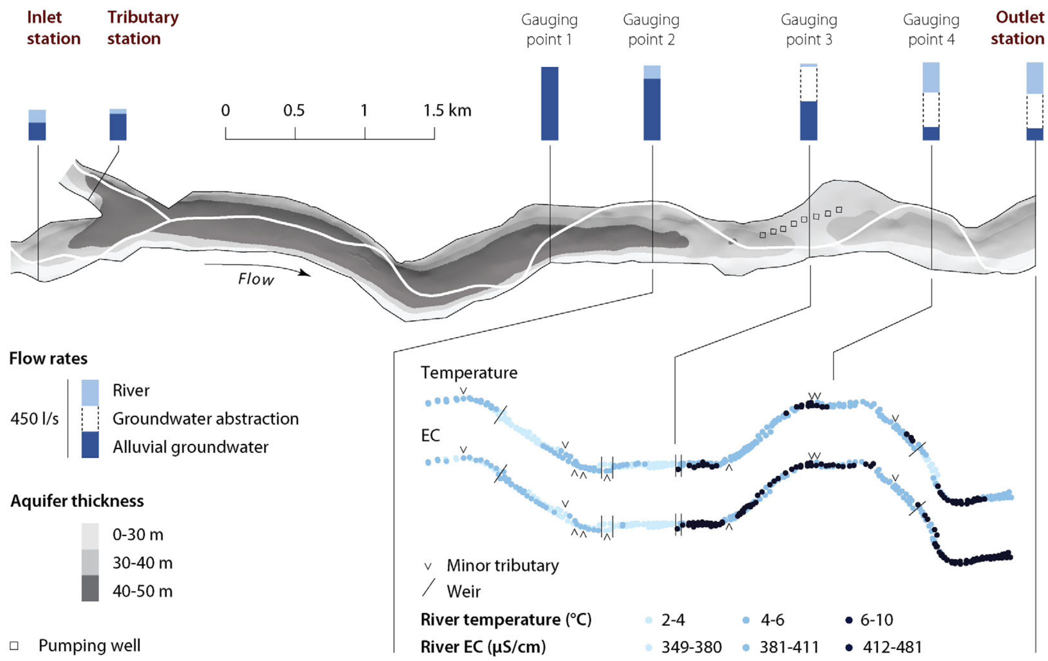

Figure 7. (top) Partition of discharge between groundwater and surface water along the Emme valley during the driest week (26 November to 2 December), based on estimated fluxes at the coupled gauging stations, and gauging surveys in-between (river and tributaries). (bottom) Temperature and electrical conductivity survey in the Emme River (23–24 November). Groundwater temperature and electrical conductivity were approximately 10°C and 450 µS/cm (corrected to 25°C) during the survey, and fluctuated seasonally between 6.6 and 10.8°C (2011), and between 400 and 500 µS/cm (December 2011 to May 2012). The aquifer thickness was derived from various data by the Canton of Bern [*Blau and Muchenberger*, 1997].

3.7. General Functioning of the River-Aquifer System

The study site is representative of mountainous alluvial aquifers that are relatively transmissive, and exhibit some potential for storage fluctuations. The river in such systems may temporarily and locally dry out, yet the subsurface flux is maintained through the depletion of upstream alluvial storage. The partition of out-flow into subsurface and surface discharge may vary strongly over short distances, as a result of hydrogeo-logical and geomorphological controls. Such exchanges are probably enhanced in regions subject to flash floods that regularly remobilize sediments and promote high streambed permeability. Thus, the fraction of total outflow captured by a river gauging station is likely to vary strongly in space, particularly in low-flow conditions.

In this aquifer, storage depletion occurs mostly in the narrower upstream part, where the unsaturated zone is thickest (Roethebach subcatchment). This area appears to supply the larger downstream region, where the groundwater-table is maintained close to and locally above river heads, and storage variations are therefore limited. In the upstream part, groundwater storage may recede continuously over a long period (e.g., 5 months in 2011, 9 months in 2003). Nevertheless, recharge is disproportionately fast: probably because groundwater exfiltration occurs only at the downstream end of this area, whereas river infiltration occurs potentially along the entire channel. This behavior is unlikely to be exceptional: significant storage fluctuations between the upstream riparian soils and the larger downstream plains might be common but relatively localized, and therefore poorly documented.

3.8. Comparison With Other Sites

In our study, we observed a down-valley groundwater flux that remained steady despite river discharge variations by nearly three orders of magnitude (Figure 4). While this pattern is known [*Younger*, 2007], it contrasts with the lateral conceptualization of base flow generation, e.g., the hillslope-riparian-stream "paradigm." According to *Larkin and Sharp* [1992], down-valley flow (underflow) is expected if channel gradients exceed 0.0008, which is the case at our site. Persistent underflow was also observed in a steep headwater valley (channel slope of 0.0113 and 0.0138) despite highly variable hydraulic conditions [*Voltz et al.*, 2013]. However, observations only covered 20–30 m long reaches, while the present study extends over 16 km.

The predominance of underflow in the studied aquifer does not necessarily reflect the behavior of the entire catchment. As frequently observed in headwater areas, alluvial deposits of limited extent and hill-slope soils may also play an important role [*Anderson and Burt*, 1978; *Gomi et al*., 2002; *Uchida et al*., 2005; *Weyman*, 1970]. Dynamic storage, for example, may occur in headwater riparian zones, modulate hillslope-valley interactions, and induce changes in subsurface flow direction [*Jencso et al*., 2009; *McGlynn and McDonnell*, 2003; *McGlynn and Seibert*, 2003; *van Meerveld et al*., 2015]. These processes, however, have received far more attention than the longitudinal conceptualization of base flow generation.

At our site, vertical exchanges between the aquifer and the river play an important role. It is most conspicuous in the Emme valley where infiltration may lead locally to the complete drying out of the river (Figure 7). Simi-lar patterns were observed in other mountain catchments [*Konrad*, 2006a]. For example at the Methow river, a well-documented site in the United States, groundwater seepage from the alluvial aquifer accounted for 37–57% of base flow and 13–14% of annual discharge [*Konrad*, 2006b]. Both this study and ours show that a zone of limited extent (2.5–5% of catchment area) may have a strong influence on base flow [*Konrad et al*., 2003]. The Methow study did not quantify storage dynamics. Therefore it is not known whether the aquifer acts as a dynamic reservoir or only transmits water stored elsewhere.

At our site, alluvial storage depletion contributes strongly to catchment outflow in dry periods. In other regions, however, storage in bedrock aquifers may be comparatively more important [e.g., *Andermann et al*., 2012]. Several studies evaluate both the contribution of bedrock-groundwater to stream flow and the inter-actions with alluvial aquifers. For example, *Tague and Grant* [2004] relate the high base flow values in the ''Higher Cascade'' catchments to the presence of volcanic deposits with an exceptionally high porosity and permeability. *Sayama et al*. [2011] obtained large dynamic groundwater storage volumes based on stream flow data of steep catchments, which led to the postulation of the ''hydrologically active bedrock layers'' hypothesis. However, these studies inferred groundwater storage from river discharge, and did not include subsurface head measurements.

Our study has also implications for recession analysis, given that both river discharge and the shape of the recession curve appear to vary over short distances. Base flow is known to vary spatially due to ground-water/surface water interactions, yet the recession's behavior is generally not expected to change dramati-cally. Thus, in addition to catchment-scale properties [*Biswal and Marani*, 2010, 2014; *Brutsaert and Nieber*, 1977; *Smakhtin*, 2001], the local conditions at the gauging station may also influence the recession behavior, as becomes evident by comparing the discharge patterns along the Emme: at the inlet, discharge follows a perfect exponential decay; at the outlet, a few kilometers downstream, it plateaus around 175 L/s, while in between the river dries out completely (Figure 7).

Last, it is worth stressing the double significance of alluvial groundwater in dry periods, to sustain catch-ment outflow on the one hand, and river flow on the other. Groundwater accretion impacts river discharge as well as chemistry (Figure 7): as a result of groundwater exfiltration mainly, the electrical conductivity of river water increased over a 4 km reach from 350 to 480 μS/cm (25°C) . Here, the higher values are related to flow paths through a relatively long and shallow alluvial aquifer, but do not indicate the arrival of deep groundwater as sometimes inferred [*Frisbee et al*., 2011].

4. Summary and Conclusions

This study shows that alluvial groundwater may contribute substantially to catchment outflow, yet at spe-cific times and locations. At all three gauging stations, the alluvial groundwater flux is remarkably steady during the year. Its relative contribution increases therefore as river discharge decreases. At the outlet of the 194 km^2 catchment, alluvial groundwater discharge accounted for only 2% of annual outflow, but amounted to 17% of daily outflow at the end of a 42 days drought. This contribution was much higher at the outlet of the upstream tributary subcatchment (52 km^2), where alluvial groundwater discharge repre-sented 15% of annual outflow, and 85% of daily outflow during the driest week. In the middle section of the Emme valley, which drains an area of 186 km^2, stream discharge completely ceased, while approxi-mately 425 L/s were transmitted in the alluvial aquifer. Thus, such aquifers may constitute a critical pathway despite their limited extent in mountainous regions. This suggests that some river gauging stations might miss a significant fraction of catchment outflow. Predictive hydrological models relying on such stations may be subject to a bias and underestimate the simulated river discharge in downstream regions, especially during low flows. In this case, “coupled gauging stations” similar to those used in this study could provide a useful solution.

Besides its function as a “transmission zone“ the alluvial aquifer also provides seasonal storage. Toward the end of the dry spell, the flux resulting from alluvial groundwater depletion supported 35% of the catchment outflow and 75% of the tributary subcatchment outflow. Storage variations occurred mostly in the uppermost and steeper area, where the unsaturated zone was thickest. As a result, steady groundwater levels were maintained in the larger downstream part of the aquifer, and the extensive drought affecting the catchment’s surface water was not concomitant, in most of the alluvial system, to a groundwater drought. These results suggest that the contribution to catchment outflow of seemingly minor tributary systems might be overlooked in mountainous regions. Yet, with the retreat of glaciers and the earlier snowmelt expected under future climatic conditions, such aquifers will likely play an increasing role in sustaining base flow.

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