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Storage-discharge relationships at different catchment scales based on local high-precision gravimetry

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Abstract

In hydrology, the storage-discharge relationship is a fundamental catchment property. Understanding what controls this relationship is at the core of catchment science. To date, there are no direct methods to measure water storage at catchment scales (10^1-10^3 km^2). In this study, we use direct measurements of terrestrial water storage dynamics by means of superconducting gravimetry in a small headwater catchment of the Regen River, Germany, to derive empirical storage-discharge relationships in nested catchments of increasing scale. Our results show that the local storage measurements are strongly related to streamflow dynamics at larger scales (> 100 km^2; correlation coefficient = 0.78-0.81), but at small scale no such relationship exists (~ 1 km^2; correlation coefficients = -0.11). The geologic setting in the region can explain both the disconnection between local water storage and headwater runoff, and the connectivity between headwater storage and streams draining larger catchment areas. More research is required to understand what controls the form of the observed storage-discharge relationships at the catchment scale. This study demonstrates that high-precision gravimetry can provide new insights into the complex relationship between state and response of hydrological systems.

Introduction

Total terrestrial water storage is the state variable of a hydrological system. Discharge is directly dependent on the amount of water stored in the catchment. Understanding how a catchment stores and releases water is key to understanding the discharge generation processes that are active in the catchment. Different concepts of the role of storage in the context of discharge generation have been proposed and the relationship between water storage and discharge has received increasing attention in recent years (e.g., Tetzlaff *et al.*, 2011). In general, a catchment consists of different storage compartments, such as riparian and hillslope zones (e.g., McGlynn *et al.*, 2004), upslope and downslope areas (e.g., Haught and van Meerveld, 2011; Seibert *et al.*, 2003) or active and passive storage components (e.g., Birkel *et al.*, 2011; Dunn *et al.*, 2010). These storage components are dynamically connected by a multitude of different processes including preferential flow (e.g., Flury *et al.*, 1994; Weiler and Naef, 2003), bedrock-soil interface flow (e.g., McDonnell, 1990; Tromp-van Meerveld and McDonnell, 2006a), or pressure waves (e.g., Torres *et al.*, 1998). Therefore, the relationship of storage and discharge is complex and can be hysteric (e.g., McGlynn *et al.*, 2004; Spence *et al.*, 2010) or threshold controlled (e.g., Tromp-van Meerveld and McDonnell, 2006b). The spatial heterogeneity and high process complexity at all temporal and spatial scales make it difficult to describe and predict the state and, hence, the response of the catchment. Different approaches have been developed to investigate the storage-discharge relationship: nested observation approaches (e.g., Ali *et al.*, 2011; Haught and van Meerveld, 2011), natural tracers in combination with models (e.g., Birkel *et al.*, 2011; Soulsby *et al.*, 2011) or theoretical studies of stochastic influences (e.g., Suweis *et al.*, 2010). However, more simplifications are needed in order to establish a single storage-discharge relationship for a catchment (e.g., Brutsaert, 2008; Kirchner, 2009; Teuling *et al.*, 2010). The form of the storage-discharge function could serve as a means to compare and classify catchments across climate and geologic gradients (McNamara *et al.*, 2011; Wagener *et al.*, 2007).

Since water storage is highly variable in space and time and most of the water is stored in the subsurface, it is notoriously difficult to measure or estimate water storage and to investigate the relationship between precipitation, storage and discharge. Hydrological science has developed several methods to track water storage or the time variation of water storage at different spatial scales (Troch *et al.*, 2007). In hydrology, fluxes are frequently used to infer storage change at the catchment scale and storage is estimated based on the water balance approach (e.g., Sayama *et al.*, 2011)

$$\frac{dS}{dt} = P - E - Q \qquad (1)$$

where S [L] is the storage change during a time step t [T], P [L/T] is the precipitation, E [L/T] is the evapotranspiration and Q is the discharge [L/T]. One main problem with the water balance approach is that by integrating fluxes one accumulates errors in measurements/estimations in the storage term.

In most catchments, discharge is directly related to the amount of water stored in the catchment (e.g., Brutsaert and Nieber, 1977; Troch *et al.*, 1993; Wittenberg and Sivapalan, 1999). Therefore, catchment storage can be derived from discharge directly based on a storage-discharge function

$$S = aQ^{b} \qquad (2)$$

where Q [L/T] is discharge and a [$L^{1-b}T^{b}$] and b [-] are parameters that can be estimated by recession analysis of discharge data. With known values for E and P (usually set to zero), the rate of decrease of discharge can reveal the nature of the storage-discharge relationship and can be used to derive the catchment storage. This method suffers from practical problems like the selection of the recession periods and the need of high quality discharge measurements (Rupp and Selker, 2006). In addition, making assumptions on the usually unknown relationship between the output and system state to derive storage dynamics is especially critical in view of non-linear discharge generation processes. If independent and direct measurements of terrestrial water storage were available at the catchment scale, hydrologists would be able to directly investigate the relationship of water storage and discharge generation to advance the understanding and prediction of catchment hydrology.

Direct measurements of the water storage and its temporal variations are traditionally limited to point or plot scales (~10^{-2} m^2 to ~10^{1} m^2) and to single components of total water storage (snow, soil moisture, or groundwater). But water storage varies in space and the variability is controlled by the underlying processes and system properties (e.g. boundary fluxes, internal system parameters), which have very different spatial characteristics (e.g. correlation lengths). Hence, the spatial variability of water storages is a function of the spatial scale (e.g., Blöschl and Sivapalan, 1995; Western *et al.*, 2002; Woods, 2005). Theories about the spatio-temporal variability of near-surface water storages have been developed and tested (e.g., Brocca *et al.*, 2010; Famiglietti *et al.*, 2008; Western *et al.*, 2004), but only a few studies have used data from soil moisture measurements in the deeper subsurface (e.g., Kachanoski and de Jong, 1988; Pachepsky *et al.*, 2005; Seyfried *et al.*, 2009). At the point scale, water storage can vary at short horizontal distances and can often be assumed to be randomly distributed (nugget in a variogram; e.g., Western *et al.*, 2004). At the plot scale (~10^1 m^2), vegetation and soil hydraulic properties largely control water storage (e.g., Teuling and Troch, 2005; Vereecken *et al.*, 2007a). At the next larger scale (the field/hillslope scale; ~10^4 m^2), distribution of water storage is primarily linked to the topography, land use, or soil types (e.g., Merz and Plate, 1997; Western *et al.*, 2002). Geostatistical techniques can help to inter- and extrapolate these measurements to estimate water storage with depth and for larger lateral spatial scales (e.g., Bardossy and Lehmann, 1998; Entin *et al.*, 2000; Vereecken *et al.*, 2007b). New methods such as cosmic-ray soil moisture probes (e.g., Rivera Villarreyes *et al.*, 2011; Zreda *et al.*, 2008), electro-magnetic methods (electrical resistivity (e.g., Samouelian *et al.*, 2005), ground penetrating radar (e.g., Huisman *et al.*, 2003), magnetic resonance tomography (e.g. Lubczynski and Roy, 2004)), GPS receivers (e.g., Larson *et al.*, 2008), distributed temperature sensing (e.g., Steele-Dunne *et al.*, 2010) or high-precision gravimeter measurements, present exciting opportunities to explore the water storage, including deeper zones, directly at plot (~10^1 m^2) to field scale (~10^4 m^2).

Superconducting gravimeters provide a measure of the dynamics of water storage with sufficient precision and temporal resolution to be useful in hydrology (e.g., Hasan *et al.*, 2008;

Hokkanen *et al.*, 2007; Kroner and Jahr, 2006; Lampitelli and Francis, 2010). The gravity signal integrates all mass variations over the entire soil column within its footprint, including snow, soil moisture or groundwater changes (e.g., Creutzfeldt *et al.*, 2010a; Jacob *et al.*, 2009; Mäkinen and Tattari, 1988; Van Camp *et al.*, 2006b). In this context, the gravity signal corresponds to the integral character of discharge measurements (Creutzfeldt *et al.*, 2010b; Hasan *et al.*, 2008). The downside of gravimeter measurements (which similarly applies to discharge measurements) is that it is difficult to unambiguously identify the signal source and that the sampling volume changes over time (Creutzfeldt *et al.*, 2010a). A general rule for the measurement support of a gravimeter states that around 90 % of the gravity signal is generated within a radius of 10 times the vertical distance between the sensor and an assumed flat and thin layer where the water storage changes occur (Leirião *et al.*, 2009). This implies that the measurement support is a function of the vertical distribution of mass change below the sensor. As also topography determines the distribution of hydrological masses in space, the relationship of WSC and gravity response is site specific (e.g., Creutzfeldt *et al.*, 2008; Hokkanen *et al.*, 2006; Kazama and Okubo, 2009). Thus, gravity residuals primarily reflect WSC on the field scale (10^2-10^5 m^2), but the exact measurement support is difficult to define and depends on the mass distribution in space.

The measurement support of gravimeters is still small compared to the catchment scale (~10^1-10^3 km^2). For high-precision gravimeters to be useful tools to determine the storage-discharge relationships of catchments, this scale gap needs to be bridged. Some studies used time-lapse measurements to estimate spatial gravity variation to derive water storage changes or the hydraulic properties of the subsurface in space (e.g., Chapman *et al.*, 2008; Ferguson *et al.*, 2007; Gettings *et al.*, 2008; Jacob *et al.*, 2010). At the large river basin to global scale (~10^5-10^8 km^2), time-variable gravity observations of the GRACE satellites provide valuable information on terrestrial water storage dynamics (e.g., Ramillien *et al.*, 2008). Different studies focus on the relationship of in-situ and satellite gravity measurements (Crossley *et al.*, 2003; Hinderer *et al.*, 2006; Neumeyer *et al.*, 2006). However, only few studies examined the relationship between local gravity variations and discharge (Jacob *et al.*, 2008; Kroner and Weise, 2011; Lampitelli and Francis, 2010; Van Camp *et al.*, 2006a). For example, Lampitelli and Francis (2010) concluded for the Alzette River in Luxembourg that temporal gravity measurements do not improve the prediction of discharge. At catchment scales, the usefulness of gravity measurements still has to be evaluated.

In this study, we focus on the relationship between total water storage change and discharge at a range of catchment scales. In absence of total storage measurements at the catchment scale, we compare local water storage dynamics measured by a superconducting gravimeter with streamflow measurements. In other words, we ask: do local total water storage measurements provide information about larger scale discharge dynamics? If so, what catchment characteristics might explain such correlation?

Study area

The study area includes the watersheds surrounding the superconducting gravimeter (SG) CD029 of the Geodetic Observatory Wettzell in the Bavarian Forest, Germany, operated by the Federal Agency for Cartography and Geodesy (BKG; Schlüter *et al.*, 2007). Mean annual precipitation is 995 mm and the mean annual potential evapotranspiration was estimated to be 577 mm according to the Thornthwaite equation (Thornthwaite, 1948). The observatory is located in the Regen watershed on a mountain ridge, which divides the watershed of the river Regen into the tributaries Weißer Regen and Schwarzer Regen (Figure 1). The observatory lies within the catchment Schwarzer Regen but is also close to the drainage divide (~300 m). The superconducting gravimeter is located in the headwaters of the Höllensteinbach but the distance to the drainage divide with Augraben is only 50 m. It is located 480 m and 530 m away from the first-order rivers Höllensteinbach and Augraben, respectively. The altitude drop to the next stream is about 50 m. Hence, the measurement site of the superconducting gravimeter can be characterized as an upslope headwater away from the river and with a distinct altitude difference to the nearest river (Figure 1). The mean annual runoff and the mean annual precipitation of the catchment Weißer Regen are 617 mm and 1141 mm, respectively (gage Kötzting, catchment size: 224 km^2). For the catchment Schwarzer Regen the mean annual runoff and the mean annual precipitation amounts to 747 mm and 1334 mm, respectively (gage Höllenstein; catchment size: 981 km^2).

The geomorphology of the region is characterized by flat ridges and plateaus (mainly used for agriculture and grassland), steep long slopes (dominated by forest), and valleys. In the Schwarzer Regen and Weißer Regen catchments, valleys make up 37 %, slopes 32 % and ridges 31 %, based on the Slope Position Classification and the Topographic Position Index (Jenness, 2010; Weiss, 2001). The geology of this area is mainly magmatic and metamorphic rocks (Granite and Gneiss) but also sedimentary rocks are present. Above the basement, a loosened and highly fractured rock zone can be found. This zone merges into a deeply weathered saprolite zone of up to 10 m or more (Raum, 2002). Typical soils are sandy-loamy Cambisols and their genesis is closely associated with periglacial weathering covers (Völkel, 1995).

In the immediate vicinity of the gravimeter station, the soil is made up of gravelly sandy loamy brown soils (Cambisols), and the basement of intact gneiss can be found at a depth of 19 m. Electrical resistivity tomography measurements in combination with cores of boreholes and soil pits show that the basement declines towards the rivers Höllensteinbach and is covered by loess deposits. At the weir location (Figure 1), the bedrock basement is at a depth of approximately 40 m. This impermeable layer is covered by a highly fractured zone, a saprolite layer, colluvial layers, and a fen (Grams, 2010; Heim, 2010).

Data and methods

We use data from the superconducting gravimeter (SG) at the Geodetic Observatory Wettzell gathered between January 2000 and July 2009 (Creutzfeldt *et al.*, 2012). SGs record the vertical change of gravitational acceleration by levitating a superconducting Niobium sphere in an extremely stable magnetic field of two superconducting coils. The current in a feedback coil is adjusted to keep the sphere at one position, so that the current is proportional to the gravity change. Observatory SGs are high-precision instruments with sub-μGal resolution, recording gravity changes at 1 Hz frequency (Goodkind, 1999). The noise level is 0.02 μGal during a period of 100 s (Van Camp *et al.*, 2005). For geophysical applications such as hydrology, it is critical to correctly estimate the drift and scale of the SG as well as non-hydrological gravity effects. Approximately two times per year, absolute gravimeter measurements were used to estimate the scale factor with a relative precision of better than 0.1 % and the almost linear drift with an accuracy of better than 0.5 μGal/year (Wziontek *et al.*, 2009). The gravity effect of tides (Hartmann and Wenzel, 1995), polar motion (Wahr, 1985) and atmospheric mass changes (Klügel and Wziontek, 2009) were modeled and removed from the gravimeter signal (Hinderer *et al.*, 2007; Neumeyer, 2010). The accuracy of the remaining signal – the gravity residuals – can be estimated roughly with 0.1 μGal for short-term variations (1–30 days) and with 0.5 μGal for interannual variations. For the Geodetic Observatory Wettzell, the remaining signal, denominated the gravity residual in the following, is considered to be largely influenced by local hydrological mass variations (Creutzfeldt *et al.*, 2010c).

We calculate water storage from the SG residuals by coupling a hydrological and geodetic model during the inversion process (Creutzfeldt *et al.*, 2010b). Water storage was calculated with a conceptual hydrological model. The model was set up with the proviso that it accounts for both parameter parsimony and adequate representation of hydrological processes. The model is based on the HBV model (Bergström, 1992; Seibert, 2005) but has been adopted and modified to reflect storages and fundamental mechanisms of the study area. Based on the subsurface conditions, the model calculates water storage in the snow, top soil, soil, saprolite, and groundwater storage. As input data, the model uses precipitation, reference evapotranspiration and snow height. The modeled water storage is translated to gravity response using a spatially explicit geodetic model by distributing the mass variations along the topography (Creutzfeldt *et al.*, 2008). The gravity response is calculated for a square area with a side length of 4 km and the SG located in its center. We developed a spatially nested discretization domain and used the DEM to distribute the estimated water storage along the topography. The z component of gravity effect due to spatial homogenous water storage variation is calculated for each cell and time step using the approximation of the MacMillan (1958) equation presented by Leirião *et al.* (2009). The total gravity response is the sum of the gravity effect of each elementary body. Finally, model parameters were calibrated to match the measured SG residuals. The calibration was based on the GLUE method developed by Beven and Binley (1992), where 50 000 Monte Carlo runs were performed with different parameter sets and the top 0.1 % of the model runs were defined as behavioural model runs. This model framework could then be used to model water storage for the study period. However, to explore the SG data directly, the results of the coupled hydro-geodetic model are used only to estimate the regression factor to calculate water storage from gravity residuals. Using the regression coefficient to estimate water storage from SG residuals has the advantage that model uncertainties only influence the absolute values but not the temporal characteristics of the time series. In this study, we use the 10-year time series (2000-2009) of water storage estimated directly from the SG. Water storage is expressed using the minimum observed water storage of the study period as the baseline, so that the estimated water storage can be considered as the *water storage change* (Christiansen *et al.*, 2011; Creutzfeldt *et al.*, 2012; Jacob *et al.*, 2010) or *active water storage* (McNamara *et al.*, 2011). In this study, we will refer to *water storage* taking into account that this term only refers to relative water storage since total water storage cannot be measured using temporal gravity observations.

Water storage and discharge were compared assuming a nonlinear relationship between water storage S [mm] and baseflow QB [mm/d] data

$$QB = cS^d + QB_0 \tag{3}$$

where QB_0 is the minimum baseflow [mm/d], c [mm^{1-d}/d] and d [-] are fitting parameters. Here, we focus on the baseflow component since it is directly dependent on the water stored in the catchment whereas total streamflow is also influenced by surface runoff and quick interflow components. Using the lower envelope of discharge values, we established the storage-baseflow relationship rather than the storage-discharge relationship, which is in line with the approach presented in the study of Brutsaert and Nieber (1977). The minimum baseflow can be interpreted as a constant outflow from the aquifer. In this study, we examine the storage-baseflow relationship for the whole study period, avoiding the subjective determination of recession periods when the streamflow is assumed to be the only function of storage, e.g., periods without precipitation or evapotranspiration. To reduce the effects of observation errors on the relationship, the data were binned and the mean of three adjacent data pairs was used for the analysis.

Streamflow measurements where available for the entire 10 year study period for the Kötzting (Weißer Regen) drainage area, Sägmühle (Schwarzer Regen) drainage area, Höllenstein (Schwarzer Regen) drainage area, and Pulling (Regen) drainage area with catchment sizes of 224.4 km², 839.3 km², 981.0 km², and 1236.5 km², respectively (Figure 1). Additionally, a V-notch weir was installed in December 2007 to measure streamflow in the first-order stream Höllensteinbach (catchment area 0.5 km²). The baseflow was separated from the streamflow using a recursive digital filter (Lyne and Hollick, 1979; Nathan and McMahon, 1990) implemented in the Web based Hydrograph Analysis Tool WHAT (Lim *et al.*, 2005). The direct runoff QD_t [mm/d] at time t is calculated from the total streamflow Q_t [mm/d] and the total streamflow and the direct runoff at the previous time (t-Δt) from

$$QD_t = \beta QD_{t-\Delta t} + \frac{(1+\beta)}{2}(Q_t - Q_{t-\Delta t}) \tag{4}$$

where β is the filter parameter and was set to 0.925 according to the study of Nathan and McMahon (1990) and Arnold *et al.* (1995). The baseflow is determined with the equation

$$QB_t = Q_t - QD_t \qquad (5).$$

This recursive digital filter was chosen for the baseflow separation to avoid making assumptions on which form the relationship of water storage and streamflow takes (Eckhardt, 2005).

Results

Figure 2 shows the water storage for the study period (Creutzfeldt *et al.*, 2012). As we can see, there are distinct seasonal variations. The average seasonal storage change is 215 mm with maxima reached between January and March and minima occurring between July and December.

When comparing water storage to streamflow of the small headwater catchment Höllensteinbach for the period of December 2007 to October 2009 (where measurements are available for the weir at the Höllensteinbach), the correlation between storage and streamflow is negative and weak (Spearman rank correlation coefficients = -0.11; p-value = 0.01; Figure 3). In contrast, for the same time period, we observe significant Spearman rank correlation coefficients between total water storage and streamflow of 0.81 (p-value < 0.01), 0.78 (p-value < 0.01), 0.80 (p-value < 0.01), and 0.81 (p-value < 0.01) for the Kötzting drainage area, the Sägmühle drainage area, the Höllenstein drainage area, and the Pulling drainage area, respectively (Figure 3).

For the whole study period, the Spearman rank correlation coefficients between total water storage and streamflow are 0.72 (p-value < 0.01), 0.72 (p-value < 0.01), 0.71 (p-value < 0.01), and 0.68 (p-value < 0.01) for the gages at Kötzting, Sägmühle, Höllenstein, and Pulling, respectively. Figure 4 shows storage versus discharge in combination with the estimated storage-baseflow curve. We see that storage and baseflow show a strong non-linear relationship. The non-linear character of the storage-baseflow relationship was also observed in other studies where streamflow was used to derive the storage of a catchment (e.g., Brutsaert and Nieber, 1977; Troch *et al.*, 1993; Wittenberg, 1994). For the four catchments, we can observe a similar shape of the storage-baseflow curves. The fitted parameter of Equation 3 agrees well for the different catchments with the largest difference for the Pulling catchment (Table 1). This might be related to the fact that the discharge regime is altered due to the confluence of the rivers Schwarzer and Weißer Regen and also due to the flow regulation of the Schwarzer Regen by two dams just before the confluence. However, the similarity of the storage-baseflow curves suggests a universal relationship in the study area assuming that the measured storage is representative for the whole catchment area.

The baseflow can be modeled using the superconducting gravimeter and the derived storage-baseflow curves. Figure 5 displays the baseflow estimated from superconducting gravimeter measurements (SG flow) together with the measured discharge and baseflow (derived from the low-pass filter algorithm) at the four different gages. Baseflow and SG flow exhibit a distinct seasonal variation. We can see an overall good agreement between measured baseflow and SG flow in time, taking into account that both measurements have a support scale that differs by several orders of magnitude. On a monthly time scale, a strong linear correlation between baseflow and the SG flow can be observed (Figure 6). In general, however, the SG flow is smaller than the baseflow, except for the years 2008 and 2009.

Discussion

In this study, we compare local water storage dynamics measured by a superconducting gravimeter with streamflow measurements at increasing spatial scales. Depth-integrating gravity measurements of storage dynamics revealed a strong correlation between total water storage dynamics at the field scale and baseflow measured at increasing spatial scales at four gages, and suggest that water fluxes between the storage reservoirs and the river network have similar dynamics at all scales. At the same time, the correlation between water storage and discharge of the closest measured headwater is remarkably weak. This discrepancy can be explained by the geological settings and the dominant runoff generation processes. At the weir location of the headwater catchment, the impermeable geologic basement occurs at a depth of approximately 40 m so that the channel of the small-scale headwater does not intersect the aquifer. As a result, it is likely that water from the upslope headwater passes the gage un-quantified as subsurface flow and only near-surface flow processes are measured at the weir. Hence, the streamflow of the first-order catchment is characterized by runoff generation processes associated to the (near-sub)surface like surface saturation overland flow or interflow on an impeding layer of the Stagnosol or the periglacial weathering cover. Gravimeters integrate over different storage components and connecting processes within the response unit instead of providing a signal on individual hydrological processes. At larger scales of the four sub-basins, the same integrated hydro-meteorological processes control state and response dynamics and, hence, a strong correlation between intermediate storage and catchment discharge can be observed. We do not know if this correlation can be observed throughout the basin. However, this study indicates that increasing the measurement scale above point scale and integrating over the whole hydrological system, the high variability of water storage collapses and common features become visible at the field scale, which is also recognizable at the catchment scale. This 'step back' in perspective provides a unique and simplified view of the overall hydrologic system and helped us to reveal similarities at different spatial measurement scales.

The fact that the gravimeter-derived field-scale storage signal correlates strongly with the overall system response at large scales suggests that the channel system is connected to the aquifers at all times at these scales so that the relationship between upslope headwater storage and streamflow is observable. Earlier studies have shown that the vadose zone storage is an important component of total terrestrial water storage in the study area (Creutzfeldt *et al.*, 2010a). Therefore, the baseflow signal at the catchment scale must contain information about the vadose zone and recharge fluxes between vadose and saturated storage. This is in close agreement with the study of Hewlett and Hibbert (1963), which proved that the vadose zone can significantly contribute to the baseflow in steep watersheds and that therefore unsaturated flow has to be taken into account in hydrograph analysis.

The strong correlation of water storage and baseflow poses the question if and how the upslope headwaters are linked to the streams draining large catchment areas? A direct connection of upslope headwaters and first-order stream catchments via the fractured zone was demonstrated by several studies for steep hillslopes with mainly shallow soils or weathering cover (Burns *et al.*, 1998; Onda *et al.*, 2001; Uchida *et al.*, 2008). For the Bavarian Forest, with the focus on the river Regen watershed, a high connectivity of the underlying fractured zone of upslope headwaters and receiving streams was demonstrated in the study of Raum (2002). Flow pathways with a distance of up to 1150 m and flow velocities of up to 6500 m/d were observed indicating a “hydraulic short-circuit” through the fractured zone and providing a hydrogeological basis for connectivity. These regional geological features, in combination with the high correlation of water storages and discharge, strongly suggest that upslope headwaters are directly connected to the receiving stream and that most water bypasses the first-order stream through the subsurface. The soil and thick saprolite zone act as the water storage of the headwater system, whereby the underlying fractured zone enables the rapid communication between upslope headwater and receiving stream. Fractured rocks overlain by a think saprolite zone in combination with periglacial weathering covers is a common feature of the Bavarian forest (Raum, 2002; Völkel, 1995; Vornehm, 2004). Whether the above mentioned mechanism can be found throughout the Bavarian forest and also in other mid-mountain ranges with similar geological settings needs to be evaluated. It was, however, also observed for the rain-fed Meuse catchment. Here, plateaus and hillslopes directly discharge into the stream via the groundwater system so that the landscape classes like wetlands, hillslopes and plateaus are not connected in series but in parallel (Savenije, 2010).

In the context of the old water paradigm, it is well recognized that stormflow is composed of pre-event water stored in the catchment before it is quickly released during an event (e.g., Buttle, 1994; McDonnell, 2003). Furthermore, the chemical composition of this pre-event water varies over time. This double paradox of rapid mobilization and varying chemistry of pre-event water suggests that "catchments have several different water stores … (which; note from the author) are mobilized in different proportions at high and low flows" (Kirchner, 2003). This study shows that upslope headwaters can play a vital role in baseflow generating processes. It also provides evidence that pre-event upslope water can rapidly reach the stream while the displacement process of pre-event water by event water and the rapid transport through the vadose zone remain unexplained.

In line with other studies (e.g., Kirchner, 2009; Wittenberg, 1994), we assume a non-linear one-to-one relationship of storage and baseflow. For this area, this simplification seems to be justified but has to be elaborated for other scales and areas, because the storage-discharge function might be non-unique and/or discontinuous (e.g., Spence, 2010). Gravimeters might be especially suitable because of the integral character similar in nature to discharge measurements. Contrary to streamflow measurements, high-precision gravimeters can characterize the catchment above the outlet point since they are not limited to the stream network but can be employed anywhere in the landscape. The observed scale gap can be bridged by spatially distributed gravity measurements (e.g., Jacob *et al.*, 2010) or the combination with other methods. In this context, the combination of conservative tracer inverted storage (e.g., Soulsby *et al.*, 2009) and storage estimated with high-precision gravimetry might be especially fruitful because both methods provide an integrative measure on storage and aim at the estimation of the active storage part of the catchment. This study demonstrates the potential use of high-precision gravimetry in the context of hillslope and catchment hydrology. Using gravimeters to estimate water storage dynamics, it is possible to directly evaluate the relationship of storage and discharge, which further fosters the catchment understanding and the catchment classification.

Conclusion

In this study, high-precision gravimeter measurements are interpreted in the context of catchment hydrology to improve our understanding of the storage-discharge relationship. Independent storage measurements in combination with streamflow measurements lead to an improved understanding of the hydrological processes that dynamically connect different compartments and scales. Contrary to streamflow measurements, high-precision gravimeter measurements can be realized anywhere in the landscape and are not limited to the stream network. Thus, they can provide new insights into the large-scale structure of hydrological systems and can help to characterize hydrological processes throughout a basin. It requires advances in gravimeter technology to develop portable high-precision instruments that will allow for such spatially distributed field investigations.

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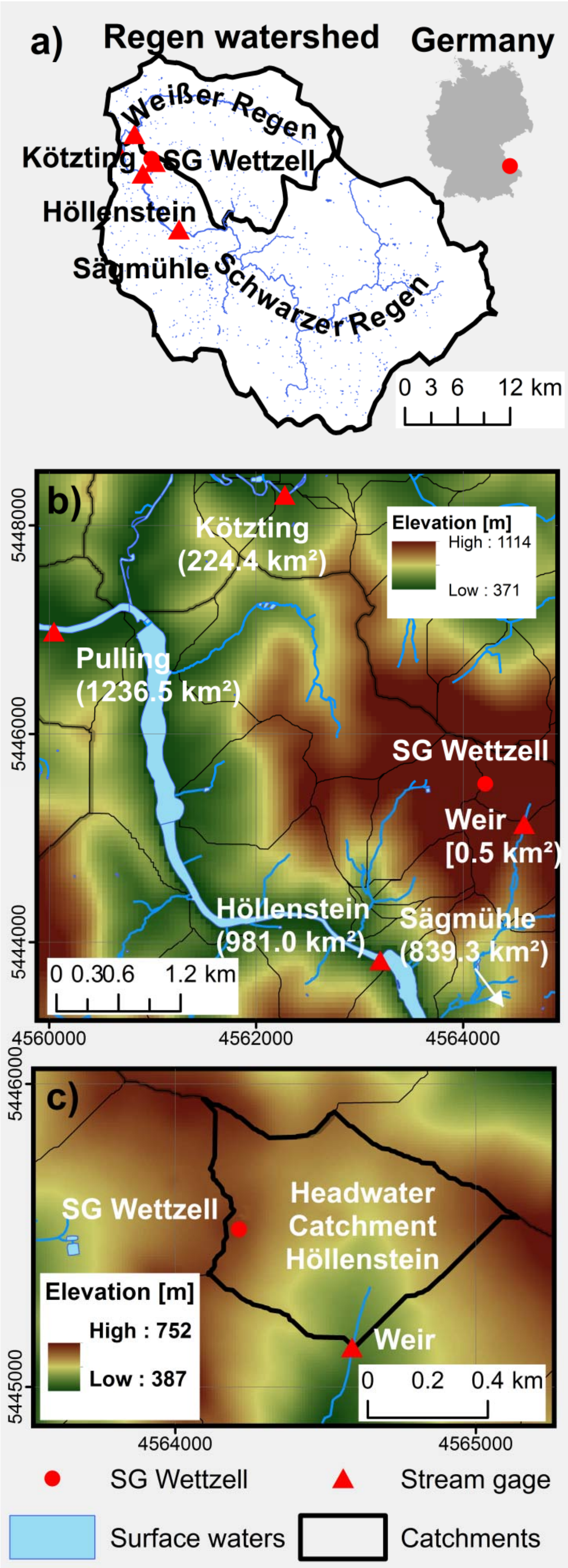
Tables

Table 1: Estimated parameters of the non-linear relationship between storage and baseflow (Equation 3) for the different catchments (see Figure 1), where *c* and *d* are fitting parameters and QB_0 is the estimated minimum baseflow.

	Kötzting	Sägmühle	Höllenstein	Pulling
Catchment	Weißer Regen	Schwarzer Regen	Schwarzer Regen	Regen
Catchment size [km²]	224.4	839.3	981.0	1236.5
C [mm^{1-d}/d]	1.253e-010	2.314e-010	1.058e-010	2.208e-008
d [-]	3.850	3.777	3.899	3.020
QB_0 [mm/d]	0.560	0.563	0.400	0.543

Figures



a)
Regen watershed
Germany
Weißer Regen
Kötzting
SG Wettzell
Höllenstein
Sägmühle
Schwarzer Regen
0 3 6 12 km
b)
Kötzting
(224.4 km²)
Elevation [m]
High : 1114
Low : 371
Pulling
(1236.5 km²)
SG Wettzell
Weir
[0.5 km²)
Höllenstein
(981.0 km²)
Sägmühle
(839.3 km²)
0 0.30.6 1.2 km
5448000
5446000
5444000
4560000
4562000
4564000
c)
SG Wettzell
Headwater
Catchment
Höllenstein
Elevation [m]
High : 752
Low : 387
Weir
0 0.2 0.4 km
5446000
5445000
4564000
4565000
SG Wettzell
Stream gage
Surface waters
Catchments

Figure 1: Map of the study area. (a) the watershed Regen divided into the catchments Weißer Regen and Schwarzer Regen, the corresponding stream gages and the superconducting gravimeter (SG) of the Geodetic Observatory Wettzell and the location of the SG Wettzell in Germany. (b) location of the stream gages and the SG within the catchment underlain by the topography represented as a Digital Elevation Model (DEM; BKG, 2004). (c) the headwater catchment Höllensteinbach, the location of the weir and the SG underlain by a DEM (DEM; LVG, 2007).

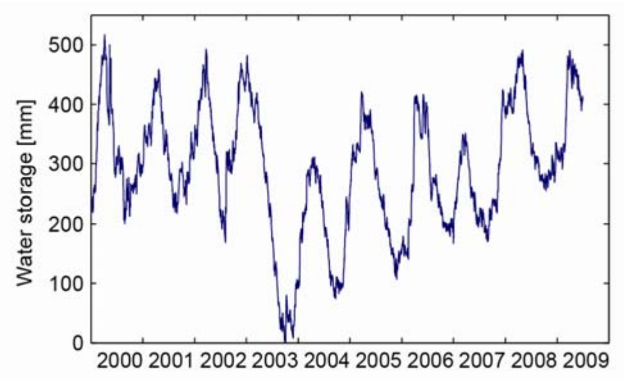


Figure 2: Water storages estimated from superconducting gravimeter. Water storage is expressed using the minimum observed water storage of the study period as the baseline.

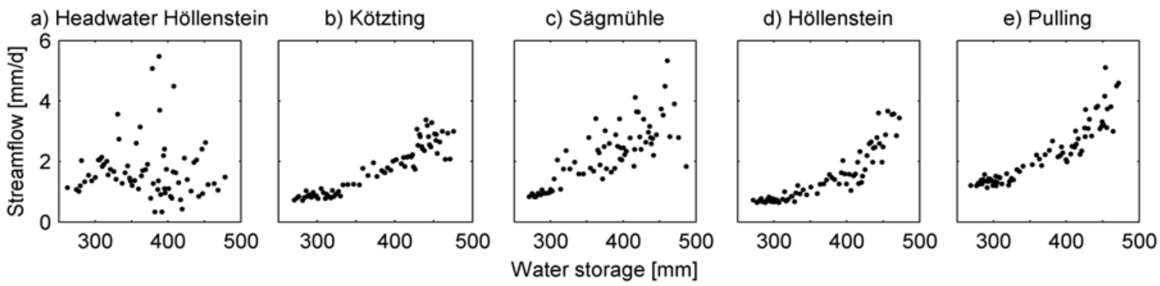


Figure 3: Relation between water storage and streamflow for the five different stream gages: (a) headwater Höllenstein drainage area, (b) Kötzting drainage area, (c) Sägmühle drainage area, (d) Höllenstein drainage area, and (e) Pulling drainage area for the period December 2007 to October 2009. Black dots are the bin averages.

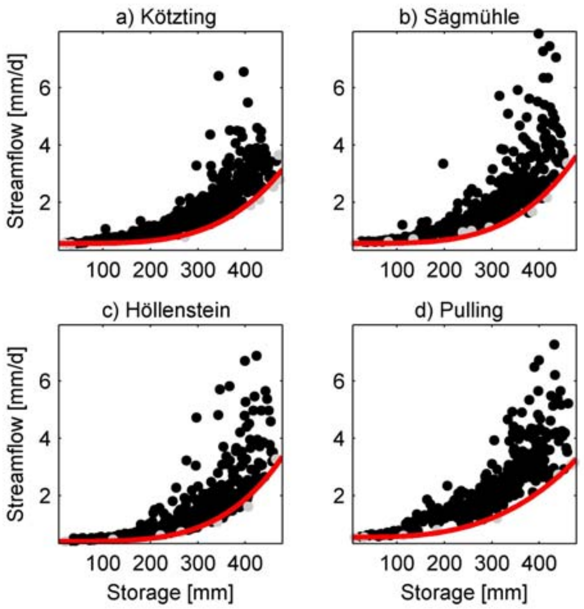


Figure 4: Relation between water storage and streamflow for the four different stream gages (a) Kötzting drainage area, (b) Sägmühle drainage area, (c) Höllenstein drainage area, and (e) Pulling drainage area for the whole study period from January 2000 to October 2009. Red lines are the lower envelop curve and black dots are the bin averages.

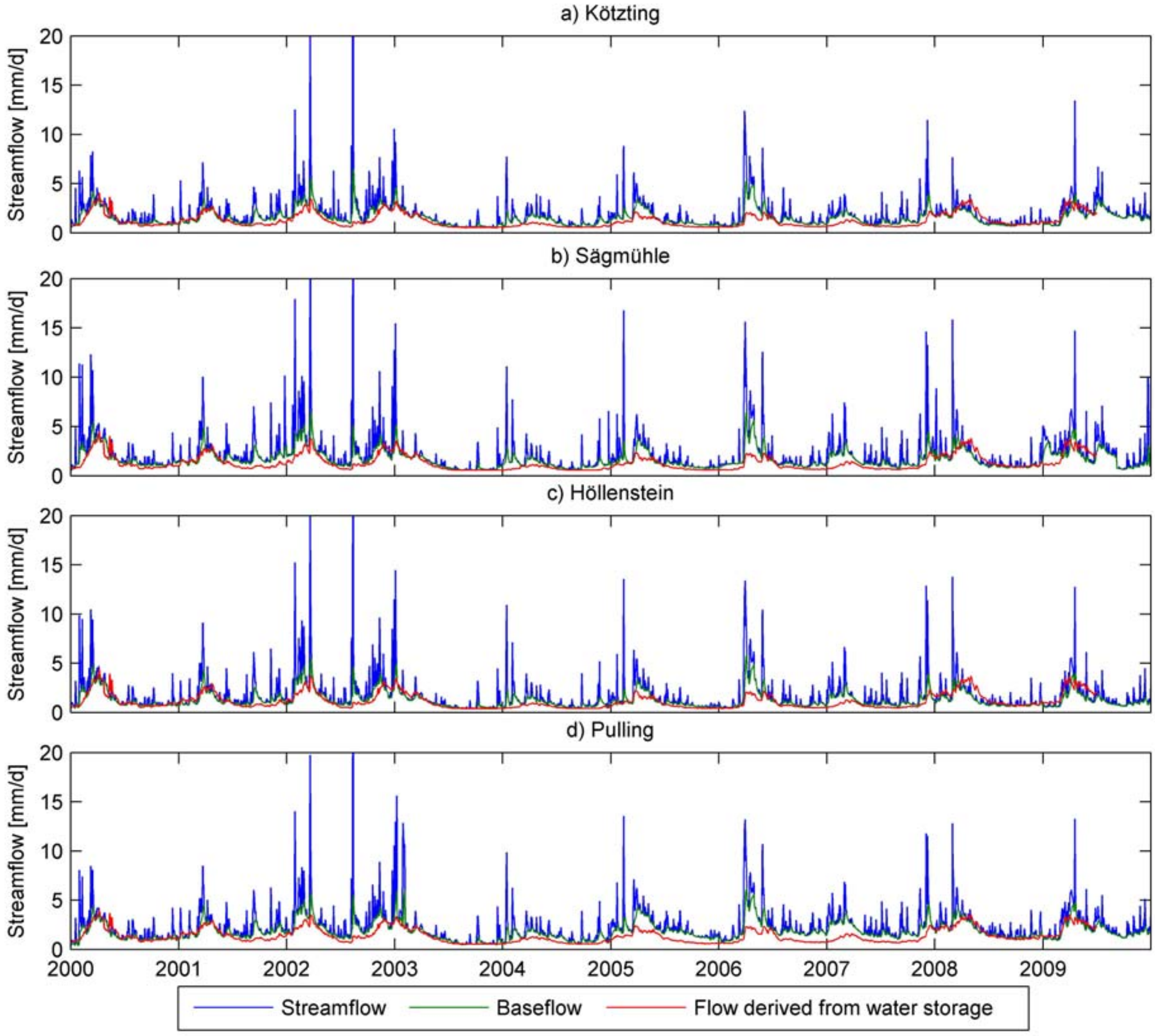


Figure 5: Measured streamflow and estimated baseflow measured at the four different stream gages: (a) Kötzting drainage area, (b) Sägmühle drainage area, (c) Höllenstein drainage area, and (d) Pulling drainage area in combination with the flow derived from SG measurements, using the lower envelop curve of Figure 4 for the storage-baseflow-relationship.

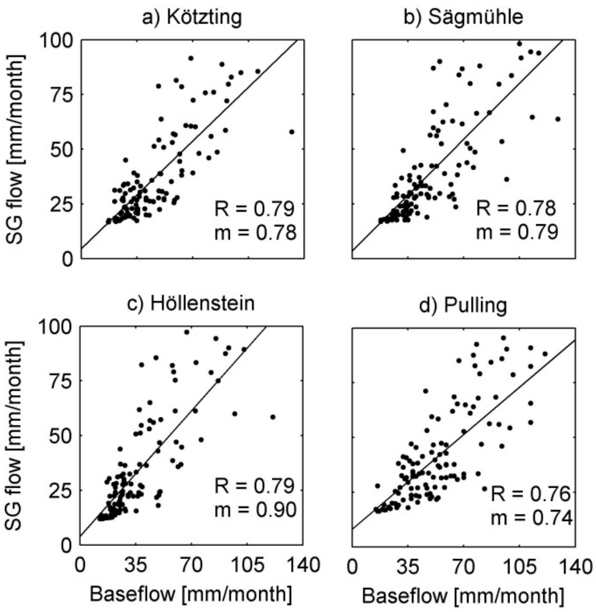


Figure 6: Monthly baseflow in comparison to flow derived from SG measurements (SG flow) for four different stream gages: (a) Kötzting drainage area, (b) Sägmühle drainage area, (c) Höllenstein drainage area, and (d) Pulling drainage area in combination with the regression line (black line), the Pearson correlation (R) and the slope of the regression line (m). All correlation coefficients are below the significance level of p=0.01.