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How much water can a watershed store?

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Abstract

Subsurface runoff dominates the hydrology of many steep humid regions, and yet the basic elements of water collection, storage, and discharge are still poorly understood at the watershed scale. Here we use exceptionally dense rainfall and runoff records from two Northern California watersheds (~100 km^2) with distinct wet and dry seasons to ask the simple question: how much water can a watershed store? Stream hydrographs from 17 sub-watersheds through the wet season are used to answer this question where we use a simple water balance analysis to estimate total storage changes (dV) during a rainy season. Our findings suggest a pronounced storage limit and then "storage excess" pattern; i.e. the watersheds store significant amounts of rainfall with little corresponding runoff in the beginning of the wet season and then release considerably more water to the streams after they reach and exceed their storage capacities. The amount of rainfall required to fill the storages at our study sites is the order of a few hundred millimeters (200-500 mm). For each sub-watershed, we calculated a variety of topographic indices and regressed these against maximum dV. Among various indices, median gradient showed the strongest control on dV where watershed median slope angle was positively related to total storage change. We explain this using a hydrologically active bedrock hypothesis whereby the amount of water a watershed can store is influenced by filling of unrequited storage in bedrock—itself linked to topography and slope angle, which increase the required amount of water to activate rapid rainfall-runoff response at the watershed scale.

> The secret to 'doing better hydrological science': change the question!,
> Sivapalan, M. (2009)

Introduction

Much of the focus of watershed hydrology has been aimed at how much water a watershed can shed (Tetzlaff *et al.*, 2009). Such shedding mechanisms in humid regions have focused on combinations of infiltration excess overland flow and saturation excess overland flow (Easton *et al.*, 2008). Surface water shedding is readily observed and as a result a conceptual framework for overland flow type and occurrence based on aridity indices and precipitation intensity is now well defined in the literature (Kirkby, 2005; Reany *et al.*, 2007). Of course, much of the landscape does not "surface saturate" and in upland humid catchments, subsurface stormflow may dominate the "shedding" of water, with rainfall-runoff ratios that sometimes rival overland flow rates (Beckers and Alila, 2004). Unlike overland flow shedding processes, however, subsurface stormflow mechanisms are seemingly endless and a multitude of subsurface stormflow mechanisms have been put forward in the literature (see McDonnell *et al.*, 2007 for review).

Here we explore the age-old subsurface runoff issue, but change the question—from one aimed at watershed water shedding, to one aimed at answering a question: How much water a watershed can store? Despite the importance of watershed storage as the key watershed function (Black, 1997, Wagener *et al.*, 2007) and also as a fundamental variable for many rainfall-runoff models (e.g., Sugawara and Maruyama, 1956; Brutsaert, 2005; Kirchner, 2009), few attempts have been made to estimate the volume of subsurface water storage (McDonnell, 2003; McDonnell, 2009). Whilst there are notable groundwater hydrogeology, for hillslopes using ground-based geophysical approaches (e.g. Collins *et al.*, 1989) and for some large river basins using gravity-based satellite measures (Troch and Durcik, 2007; Strassberg *et al.*, 2009; Rodell *et al.*, 2006), for headwaters where most watershed runoff is generated, we have not been able to answer this basic question (Soulsby *et al.*, 2009). Answering such a question would help with understanding better, the more transient, and vexing questions of subsurface stormflow delivery mechanisms.

The variable source area concept of Hewlett and Hibbert (1967) and the hydrogeomorphic concept of Sidle *et al.* (1995, 2000) are useful foundational elements for considering subsurface storage and release. Recent work on lake storage by Spence *et al.* (2007, 2010) has provided a useful model for how the subsurface store and storage-discharge process may be conceptualized and understood. Unlike these surface water storage phenomena, however, subsurface storage attempts are hindered by boundary conditions that are difficult or impossible to define. In addition subsurface heterogeneity makes the storage-discharge relationship even more complicated (Beven, 2006). Here we explore the links between subsurface water collection, storage and discharge within a set of diverse nested catchments in Northern California USA. To our knowledge, this is the most intensive continuous rainfall-runoff installation ever collected: 17 stream gauging stations (covering a wide range of scales) and 10 rainfall recorders distributed throughout two neighboring ~100 km^2 watersheds. We leverage this unique dataset against an extremely sharp wet-dry season transition that allows us to explore the limits of subsurface storage across each of the catchments and at different scales. We deliberately avoid any plot or hillslope scale process analysis and instead, work with watershed rainfall runoff data. Our work is motivated by recent calls for creative analysis of the available runoff data to gain insights into the functioning of catchments, including the underlying climate and landscape controls (Sivapalan, 2009) and early pleas for macroscale hydrologic laws (Dooge, 1986).

We build upon the work of Sidle *et al.* (2000) who noted the importance of threshold-like activation of different geomorphic positions at a steep, humid catchment in Japan. They observed that as antecedent wetness increased, zero-order basin activation began after an accumulation of shallow groundwater. Recent work at the hillslope scale has also suggested that storage elements in the hillslope need to be filled before releasing water from the slope base (see Graham et al., 2010; Graham and McDonnell, 2010 for recent review, McGuire and McDonnell 2010). Seibert and McDonnell (2002) used a similar approach to define a series of cryptic units within a watershed that were then translated into a predictive rainfall-runoff model structure. Furthermore, Sayama and McDonnell (2009) showed how subsurface storage in the soil mantle influences the source, flowpath and residence time of water flux in the headwaters. Our method is simple and straightforward: water balance analysis from the sites, regression with available topographic data and hydrogeomorphological interpretation. Our specific research questions are:

1. How much subsurface water can a watershed store?
2. How does storage amount differ between sites and scales? and
3. How does topography and geology influence on storage at the watershed scale?

Study site

Our study site is the Elk River watershed (110 km^2), which drains into Humboldt Bay just south of Eureka, California (Figure 1). A neighboring watershed, the Freshwater Creek watershed (76 km^2), is used for validation of our analysis. The climate in the area is temperate and Mediterranean: dry summers followed by wet winters. The area's average annual rainfall is about 1100 mm; about 90 % of which occurs between November and May (Figure 2). The rainfall intensity is typically moderate with maximum hourly rainfall reaching up to 20 mm/h. The strong contrasts between summer and winter precipitation amounts result in a gradual wet-up period from about November to December, and thereafter, very high soil wetness is maintained until late spring. The average slopes are short (~75 m) and very steep (~45 degrees) with large variations in topography at the sub-watershed scale (<~5 km^2). The forest is comprised mostly of a coniferous lowland forest community (stand age ≈ 60 years), which includes second and third growth redwood (*Sequoia sempervirens*) and Douglas-fir (*Pseudotsuga menziensii*).

Approximately 86% of the Elk River watershed (65% of the Freshwater Creek watershed) is underlain by the Wildcat Group geology; thick sedimentary rocks as the sequence of the transgressive-regressive movement in the late Miocene to Middle Quaternary (Reid, 1999). These rocks are predominantly marine sandstone, mudstone, and siltstone. The Wildcat Group deposits weather readily into loam to clay loam soils, classified typically as Larabee soils, which are the deepest soils among the major three soil types present in the area. The combination of the Wildcat Group geology with the Larabee soils occurs mostly in the lower reaches (west part of the two watersheds), covered with comparatively deeper soil layers (100-180 cm). The upper reaches of the Elk River watershed are underlain by the Yager Formation, which covers approximately 14% of the watershed. This Cretaceous formation consists typically of well-indurated and highly folded arkosic sandstone and argillite. The sandstone is typically very strong and often forms cliffs, whereas the argillite is prone to slaking and deep weathering and is often easily sheared. Because of the different erosion rates, slopes underlain by the Yager Formation are often irregular and have a higher surface relief. The typical soil type on the formation is the Hugo soil, which is the shallowest of the three major soil types, averaging about 75 - 100 cm in depth. The upper reaches of the Freshwater Creek watershed are underlain by the Franciscan Formation, the oldest formation in the Humboldt Bay area, consisting of a heterogeneous mix of sedimentary, igneous and metamorphic rocks. Soils developed from these rocks are Atwell soils, which are typically plastic sandy clays and clayey sands (Reid, 1999).

Methods

1. Water balance analysis for total storage change

We used water balance analysis to estimate total storage changes for each sub-watershed. The total storage changes were estimated as:

$$dV(t) = \sum_{t=1}^{T} \left(R(t) - Q(t) - E(t) \right) \quad (1)$$

where t : elapsed hours from the beginning of the data record (in this study $t = 0$ at 0:00 on 13 October 2006 and $t = T$ at 23:00 on May 15 2007), $dV(t)$: total storage change from $t = 0$ to t, $R(t)$: average rainfall, $Q(t)$: discharge and $E(t)$: evapotranspiration.

We used streamflow records from ten gauging stations in the Elk River watershed and seven gauging stations in the Freshwater Creek watershed. Two gauging stations (No. 500 and No. 502) were excluded from the analysis because we found some data quality issues after careful data screenings. The data period covered the 2007 rainy season from 13 October 2006 to 15 May 2007. In terms of rainfall records, we used data from ten rain gauges distributed in the two watersheds. We applied the Thiessen polygon method to estimate average rainfall for each sub-watershed. Both discharge and rainfall data were originally recorded at 15 min intervals but aggregated to one hour for further analysis. We computed potential evapotranspiration using the Penman equation applied to the climate data at "Gasquet California site", the nearest site to our study watersheds, archived by Western Regional Climate Center (http://www.calclim.dri.edu/).

The $dV(t)$ term in equation (1) represents storage increase or decrease from $t = t_0$ to $t = t$. Since the absolute volume of the watersheds' total storage cannot be quantified using the water balance method, we focused exclusively on how their total storage changed over time from the beginning to the end of the rainy season. Errors in these estimates could be due to discharge observations (our approach was based on the USGS gauging protocol), watershed-average rainfall estimates (using our methods described above) and evapotranspiration estimates. Errors from evapotranspiration estimates can be large if the water balance calculation is implemented in a dry season when evapotranspiration contributes greatly to the watershed water balance. Given our focus on a rainy season, during which evapotranspiration is estimated to be about 200 mm—this error appears to be relatively small compared to 1200 mm of rainfall and 600 mm of runoff during the same period. Another potential error is from trans-boundary groundwater fluxes. The loss of water from one watershed to another through deep groundwater systems can potentially be important in this coastal mountain, marine-derived uplifted sedimentary geologic environment (*Reid*, 1999), and to quantify this flux is very difficult if not impossible. Nevertheless, by focusing on relatively large watersheds (> ~5km^2) we argue that the influence of such a flux should be negligible compared with analysis at smaller headwater scales.

2. Recession analysis

Streamflow recession analysis is another powerful tool to investigate the characteristics of storages feeding streams (Tallaksen, 1995; Rupp and Selker, 2005; Rupp and Selker 2008; Brutsaert, 2008). A recession curve contains valuable information concerning storage properties and aquifer characteristics (Tague and Grant, 2004l; Clark *et al.*, 2008). Brutsaert and Nieber (1977) proposed plotting an observed recession slope of hydrograph $-dQ/dt$ versus discharge Q in log-log space by eliminating time as a reference:

$$-dQ/dt = f(Q) \quad (2)$$

where f denotes an arbitrary function. We consider recessions only during night period to avoid errors associated with evapotranspiration (Kirchner, 2009). In addition, to avoid measurement nose in individual hourly measurements, we computed first average discharge for four hours during the following period; (Q_1) 19:00 – 22:00, (Q_2) 23:00 – 2:00 and (Q_3) 3:00 – 6:00. Then we calculated $-dQ/dt$ and Q as $(Q_1 - Q_2)/4$, $(Q_1 + Q_2)/2$ and $(Q_2 - Q_3)/4$, $(Q_2 + Q_3)/2$ for each day. We excluded from the plot if rainfall during the periods of 19:00 – 2:00 and 23:00 – 6:00 exceeded 0.1 mm to avoid the impact of rainfall.

Results

1. Total storage changes estimated by water balance analysis

Figure 3 illustrates the relative temporal changes in total storage (dV) estimated by the water balance approach described in equation (1), showing the storage in each of the watersheds initialized at the beginning of the data record (13 October 2006) and the relative changes during the rainy season. In the entire Elk River watershed (No. 509), the total storage increased by about 400 mm during the rainy season. The increase was almost linear throughout November and December and then plateaued at approximately 350 mm in January. After a month of relatively dry weather in January, the storage reduced by about 30 mm but then increased back to its plateau value due to rainfall events in February. It is interesting to note that even though a rainfall event in the end of February (Feb. 20 – Mar. 4) was the largest of the measured rainfall events (total 237 mm as averaged over the 8 rain gauges of the Elk River watershed), the storage increase in the watershed was only about 50 mm during that event.

The large and small sub-watersheds of the Elk River watershed showed similar temporal patterns of the parent watershed with progressive storage filling followed by plateau behavior (Figure 3(b)). However, the storage plateaus and the time required to reach the plateaus varied considerably from sub-watershed to sub-watershed. For example, the No. 533 watershed (6 km^2) reached its maximum storage of 200 mm in the beginning of January and remained almost at the same level for the rest of the rainy season. Alternatively, the No. 534 watershed (3 km^2) was characterized by the storage increases more progressively until the beginning of March.

The dynamics storage change are best illustrated dV vs discharge (Q) plots shown in Figure 4. These patterns shows that discharge in the No. 533 and the No. 534 watersheds was not activated until their dV reached 200 mm and 350 mm, respectively. At the No. 533 watershed, storage filling did not increase during the subsequent rainfall events and the *dV-Q* plot showed a large increase in discharge with minimal storage increase. On the other hand, at the No. 534 watershed, even after the *dV* reached 350 mm when the watershed started generating storm runoff, the storage progressively increase until it reached more than 500 mm. Furthermore, during the largest storm event in February, when the peak specific discharge was more than 2 mm/h, the watershed still stored additional about 20 mm of rainfall. The *dV-Q* plot during this event showed a hysteretic clockwise storage relation. This pattern was not observed at the No. 533 watershed; i.e. no storage change was observed before and after the largest storm event in February.

2. Topographic controls on total storage change

For each sub-watershed, we calculated a variety of topographic indices listed in Table 1 with our available 10 m resolution digital elevation model. We calculated also the maximum total storage changes for each sub-watershed during this study period; hereafter we denote this maximum storage change during this period as dV_{max}. Then we computed correlation coefficients between the topographic indices and the total storage changes (dV_{max}) using the data from all the sub-watersheds. Table 1 summarizes the correlation coefficients between each topographic index and the storage. Among these indices, median gradient (*G)* showed statistically significant positive correlation with dV_{max}. This positive correlation indicates that a watershed with steep slopes shows a larger total storage increase during a rainy season than a watershed with milder slopes.

While the median gradient metric (*G*) is objective and readily quantifiable, we acknowledge there is undoubtedly a co-relation and co-evolution of localgeology topography and consequently storage characteristics (Onda, 1992; Onda *et al.*, 2006). As described earlier, our watersheds are formed on three sedimentary rock groups. Figure 5 presents the relationship between *G* and dV_{max} for all sub-watersheds with the notation of their dominant geologic settings. The plot indicates that the watersheds on the Wildcat group are categorized into higher *G* with larger dV_{max} , whereas ones on the Yager and Franciscan groups are categorized into smaller *G* with less dV_{max}. The Wildcat group is the thick sedimentary rocks, which weather readily into loam to clay loam soils, while the Yager and Franciscn groups are comparatively old formations with a greater mixture geologic conditions. Notwithstanding these complexities, the geologic variation within the sub-watersheds were overall relatively small withall the geologic groups within a class of marine-derived sedimentary rock..

Table 1 shows correlations between dV_{max} and other computed topographic indices. For relief (*H)* and hypsometric integral (*HYP)*, we expected that a larger three dimensional control volume (as indicated by H and HYP) would result in larger water storage volumes. The computed correlation coefficients shown in Table 1, however, did not show clear correlations between the volumetric indices and the watershed storage and storage change.

3. Recession analysis

Recession analysis was conducted for each sub-watershed and the results are summarized in the form of Q vs $-dQ/dt$ plots in Figure 6. These analyses show contrasting results from the No. 533 and the No. 534 watersheds. Recall that the No. 533 watershed is a gentler slope watershed with smaller dV_{max}, while the No. 534 watershed has steeper slopes with higher dV_{max}. Comparing the recession analysis results from the two sub-watersheds shows that the recession rates are similar to each other when the Q is greater than 0.1 mm/h. When Q is smaller than 0.1 mm/h, the values of $-dQ/dt$ vary greatly between the two sub-watersheds. For the No. 533 watershed, Q did not drop below 0.05 mm/h, suggesting that the watershed has more stable baseflow source. At the No. 534 watershed, the variability of $-dQ/dt$ is more systematic. If we differentiate the $-dQ/dt$ plots based on the corresponding dV values, the recession plots separate into two groups: one where dV is greater than 350 mm/h and one where dV is less than 350 mm/h, which was is the amount of water required at Watershed No. 534 to start generating rapid storm runoff, as described above.

Discussion

1. So how much water does a watershed store?

The question of how much water a watershed requires is, in some ways, the type of analysis of the available runoff data advocated by Dooge (1986) and Sivapalan (2009) to gain insights into the functioning of catchments, the underlying landscape controls on water flux and the search for macro-scale hydrologic laws. Without mappable volumes like surface storage phenomena (e.g. Spence *et al.*, 2010), our method of watershed intercomparison capitalizes on the extremely intensive gauging network—the densest of its kind that we are aware of. Our approach goes beyond variable source area (Hewlett and Hibbert, 1967) and hydrogeomorphic (Sidle *et al.*, 2000) concepts by focusing on the quantitative assessment of subsurface collection, storage and discharge. The water balance approach was motivated by the visual observation of increasing baseflow levels through the wetting up season, onto which are superimposed the wet season hydrographs. Like some of our early observations of storage filling from simple hydrograph analysis (McDonnell and Taylor, 1987), the sites in California displayed clear "limits" to their wet season baseflow level attainment.

The amount of water a watershed can store varied from 200 to 500 mm. The simple water balance analysis showed how watershed total storage increases in the beginning of a rainy season and then remains almost constant after reaching a plateau value. Such observations have been made in other regions where a series of wet-up events follow an extended dry period (Sidle *et al.*, 2000). Our analyses suggest that the amount of rainfall required to fill the storage at our study sites was the order of a few hundred millimeters with the individual watershed values depending on the local watershed properties.

While each watershed showed distinct differences in its storage limit, each watershed did indeed reach a storage limit during the wetting up cycle—varying in timing by approximately 60 days. The extremely sharp wet-dry season transition allowed for this analysis. Our storage estimates are in the range of other studies that have explored soil mantle storage estimates (Sayama and McDonnell, 2009) and in many ways this is very consistent with early view of Hewlett and Hibbert (1967) who view the watershed as a "topographic pattern of soil water storage". Of course our storage estimates include an unknown blend of soil water and groundwater storage and represent the dynamics of total storage.

Our findings are also analogous to the hillslope-scale fill and spill mechanism outlined by Tromp van Meerveld and McDonnell (2006) now writ large over the watershed. In fact others observing fill and spill have observed such behavior at intermediate scales of soil-filled valleys (Spence and Woo, 2003). How much water a watershed can store seems to be a function of how much water a watershed can hold until it spills—i.e. when the wet season hydrograph response is superimposed on a steady state pre-event water background. Indeed, such analysis could be very helpful in modeling studies, where cryptic reservoirs in a lumped rainfall-runoff model (Seibert and McDonnell, 2002) could be potentially defined by such a storage-based view of the watershed.

2. Steeper watersheds store more water: An active bedrock zone hypothesis

Our watershed topographic analysis revealed a positive relation between median slope gradient of a watershed and total storage change ($dV_{\max}$) through the wet-up. This may seem a somewhat counter-intuitive relation since it suggests that catchments with steeper slopes tend to store more water. All things being equal, one might expect that catchments with lower slopes should store more water. Indeed some previous studies have shown that this is the case. For example, Troch et al (2003) used a storage based Boussinesq model and compared two idealized slopes with different gradients. Their analysis showed that flow rates from the steeper slope were more responsive and as a result, the dynamic storage change was limited compared to milder gradient slope sections. Similarly, Hopp *et al.* (2009) used a three-dimensional Darcy-Richards equation solver to show that as slope angle increases, the layer of transient saturation driving lateral flow decreases.

These previous negative correlations between dV_{max} and G are opposite to our findings. We hypothesize that this is due to bedrock permeability. In the Troch *et al.*, 2003 and Hopp *et al.*, 2009 analyses, the boundary between soil and bedrock was sharp and the bedrock was poorly permeable. On the other hand, in our watershed, like other watershed in the California and Oregon Coast Range, numerous studies have revealed a very different sort of flow response, conditioned by permeable bedrock (see Montgomery and Dietrich 2003 for review). If one considers permeable bedrock groundwater involvement in streamflow, as evidenced in the region by Anderson *et al.* (1997), Torres *et al.* (1998), Anderson and Dietrich (2001), the positive relation between storage and topographic gradient makes sense.

Figure 7 compares two idealized slopes with a porous soil underlain by a permeable bedrock layer. The conceptual diagram assumes that the depths of the soil and bedrock layers are the same for the gentle and steep slopes. The positions of the ground water tables are shown in the permeable bedrock layers at the beginning of a rainy season, as linked to our observed continuous baseflow even after the long dry season (Figure 6). Precipitation at the beginning of the rainy season infiltrates the soil and then the permeable bedrock. The water table rise represents the increase of catchment water storage and indicates the expansion of seepage area through the soil-bedrock interface. Comparing the gentle and steep slopes, the amount of precipitation water required to fill the permeable bedrock layer is greater at the steeper slope given the same gradient of water table at the beginning of the rainy season. In addition, the area of groundwater seepage, or exfiltration zone, is smaller at the steeper slope; i.e. the steeper slope needs more water to expand the same area of the seepage compared to the milder slope. This expansion of exfiltration zones drastically changes the runoff generation response also noted by Fiori *et al. (*2007)—one of the reasons we believe that we observe the storage excess patterns at the watershed scale. Invoking the hydrologically active bedrock zone hypothesis, therefore, explains the positive correlation between dV_{max} and watershed steepness.

Uchida *et al.* (2008) called this type of catchment system—with a permeable bedrock zone that stores and releases precipitation— "hydrologically active bedrock zone". At their biotite granite and granodiorite bedrock study site, Uchida *et al.* (2008) used tracer and hydrometric data to show how hydrologically active bedrock zones influence channel stormflow. We use a similar logic to Uchida *et al.* (2008) and also the Coos Bay body of work, a site less than 200 km North of ours and where the Montgomery *et al.* (2002) group explained their runoff generation mechanisms via deep permeable ground water involvement. This same runoff generation mechanism is highly likely at our study site because the geographic location and geologic setting are very similar to the Coos Bay catchments

The results shown in Figure 4 and Figure 6 also support the hydrologically active bedrock zone hypothesis. The gentle slope watershed, such as the No. 533 watershed, increased its storage up to about 200 mm and plateaued regardless more precipitation input. Alternatively, steeper watersheds, for example of the No. 534 watershed, increased its storage amount up to about 350 mm and then commenced rapid rainfall-runoff response. It is notable that even after the watershed began releasing more runoff, the watershed still stored additional water, with dV finally reaching about 500 mm. Our conceptual model with a hydrologically active bedrock zone would explain that once the groundwater table rises up to a certain level, the ground water starts seeping to the soil layer, on which additional storm rainfall creates quick lateral saturated subsurface flow. This is when the storage rate increase becomes slower compared to the beginning of a wet season. At the same time, part of the slope still can store some water gradually, particularly at the steeper watershed. This behavior influences also the streamflow recession characteristics as shown in Figure 6. At the No. 534 watershed, the recession rate is faster during the wet up period compared to the recession rate after the wet-up period. We would hypothesize that this is due to the fact that when the ground water table is low enough and rainfall infiltrates into the active bedrock zone through the soil layer, the storm runoff is created only from a limited zone (e.g. the near stream riparian zone) (Sidle *et al.* 1995). Alternatively, as the groundwater table rises and starts exfiltrating water to the above soil layer, the baseflow becomes more stable and therefore the recession rates becomes smaller. The No. 533 watershed showed generally low recession rates without dropping its discharge below 0.5 mm/h, which again supports the hydrologically active bedrock zone hypothesis as the gentle gradient watershed tends to have more steady baseflow even early in the wet season as shown in Figure 7.

Conclusions

In many ways, the work presented in this paper is a response to Dooge's (1986) call for looking for macroscale laws and more recently Sivapalan's (2009) call for more creative analysis of standard hydrologic data. We have explored watershed storage dynamics and function associated with collection and release of water across multiple nested watersheds in Northern California. These storage dynamics play an important role in determining runoff response as well as water quality (Soulsby *et al.*, 2009). Our water balance analysis from the 17 nested macro-scale watersheds revealed that each watershed stores different amounts (varying between 200 – 500 mm of precipitation) before actively generating storm runoff. The regression analysis between the maximum storage increase dV_{max} and topographic indices showed that watersheds with steeper slopes store more water than watersheds with gentler slopes. This relation could not be explained with the assumption of impermeable bedrock layers. We explained this via the hydrologically active bedrock layer hypothesis—a response type reported in similar geologic and geographic settings and our own further evidence that steeper watersheds in our study increased their storage amount gradually even after activation of storm runoff generation. Conversely, our study watersheds with gentler topography exhibited more distinct storage limits. This spatial and temporal pattern of storage plays an important role for stream flow as evidenced by distinctly different hydrograph recession rates before and after the watershed storage filling.

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Table 1 Various topographic indices and maximum total storage change (dV_{max}) at each watershed are listed. COR represents correlation between each topographic index and dV_{max}. *Area* is a watershed area (km^2). G is a median gradient. D_d is a drainage density. R is a relief (elevation difference between basin summit and basin outlet). *HYP* is a hypsometric integral[1)]. *Geol* is a dominant geologic type (W: Wildcat formation, Y: Yeger formation, F: Franciscan formation). An asterisk (*) indicates a correlation coefficient that is statistically significant ($p<0.05$).

	Area (km^2)	G	D_d	R (m)	*HYP*	*Geol*	dV_{max} (mm)
COR	-0.06	0.74*	0.32	-0.23	-0.12	N.A.	N.A.
509	111.7	1.15	18.7	2338	0.372	W	418.3
511	56.9	1.25	20.8	2328	0.353	W	354.3
517	5.7	1.48	28.1	821	0.458	W	462.2
519	4.9	1.12	15.3	1641	0.493	W	430.5
510	50.3	1.06	16.2	2092	0.453	W	455.9
183	19.5	1.04	16.6	1853	0.529	Y	297.7
188	16.2	1.02	15.6	1621	0.511	Y	438.7
533	6.3	0.91	16.6	1179	0.407	W	268.7
522	4.3	1.15	13.8	1197	0.621	W	514.9
534	3.0	1.24	13.9	815	0.568	W	544.4
504	11.9	0.97	16.0	1961	0.449	F	294.3
505	6.2	1.04	17.5	2111	0.441	F	392.4
506	8.2	1.41	22.5	2198	0.358	W	651.7
523	22.8	1.01	16.6	2678	0.509	F	286.7
526	5.1	0.96	14.8	1371	0.636	F	232.3
527	4.6	1.25	19.5	1297	0.440	W	408.7
528	12.0	1.39	24.4	924	0.501	W	514.1

[1)] A hypsometric distribution (e.g. Luo, 1998; Vivoni *et al.*, 2008) is depicted as the relative height (h/H) versus the relative area (a/A), where a is the area of watershed above height h, A is the total watershed area, h is the height above the watershed outlet, and H is the total relief of the basin. Hypsometric integral (*HYP*) is an index calculated by the integral of the hypsometric distribution. *HYP* becomes large for a watershed with convex surface, whereas *HYP* becomes small for a watershed with concave surface.

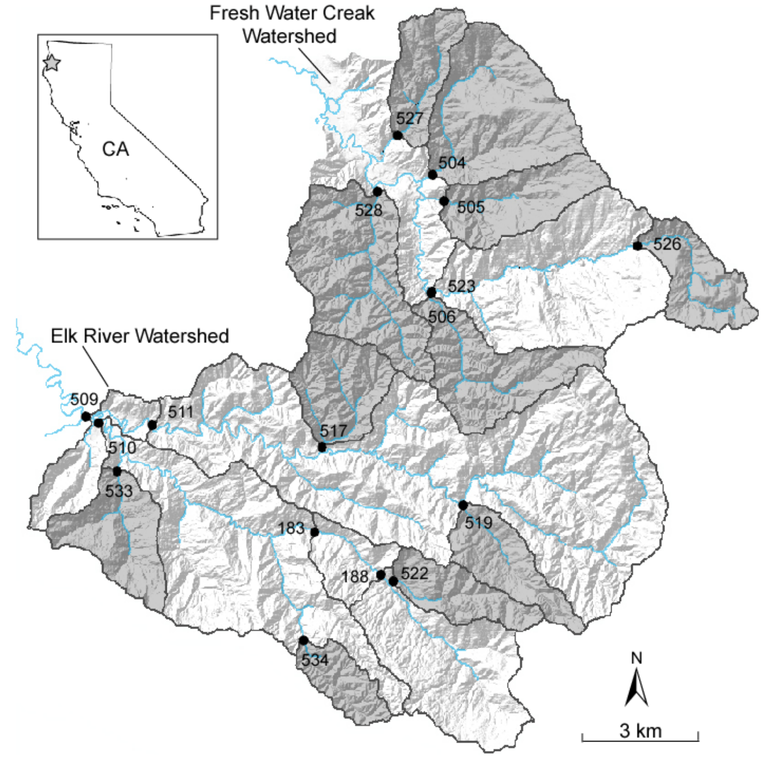

Figure 1 Map of Elk River watershed (110 km^2) and Freshwater Creek watershed (76 km^2). The black dots represent the 17 discharge gauging stations in the two adjacent watersheds.

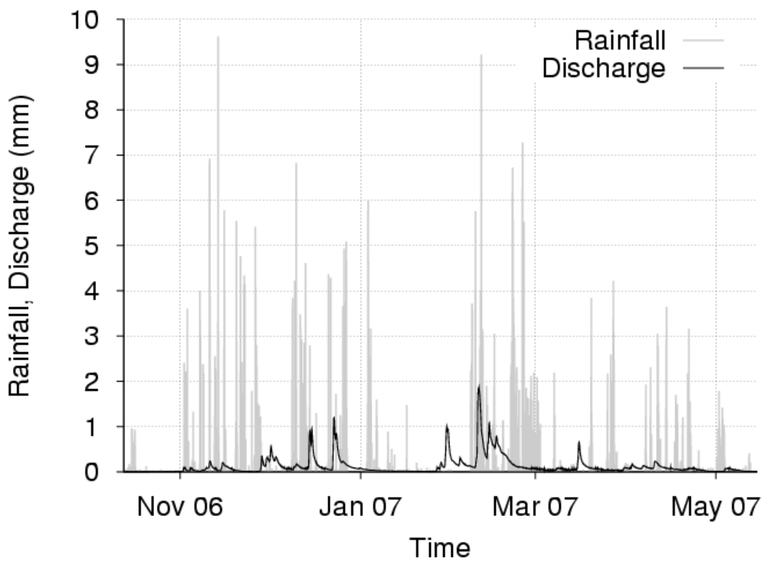

Figure 2 Watershed average rainfall and observed discharge at the outlet of the Elk River basin (No. 509, 112 km^2) during a wet season (from Oct. 13, 2006 to May 15, 2007).

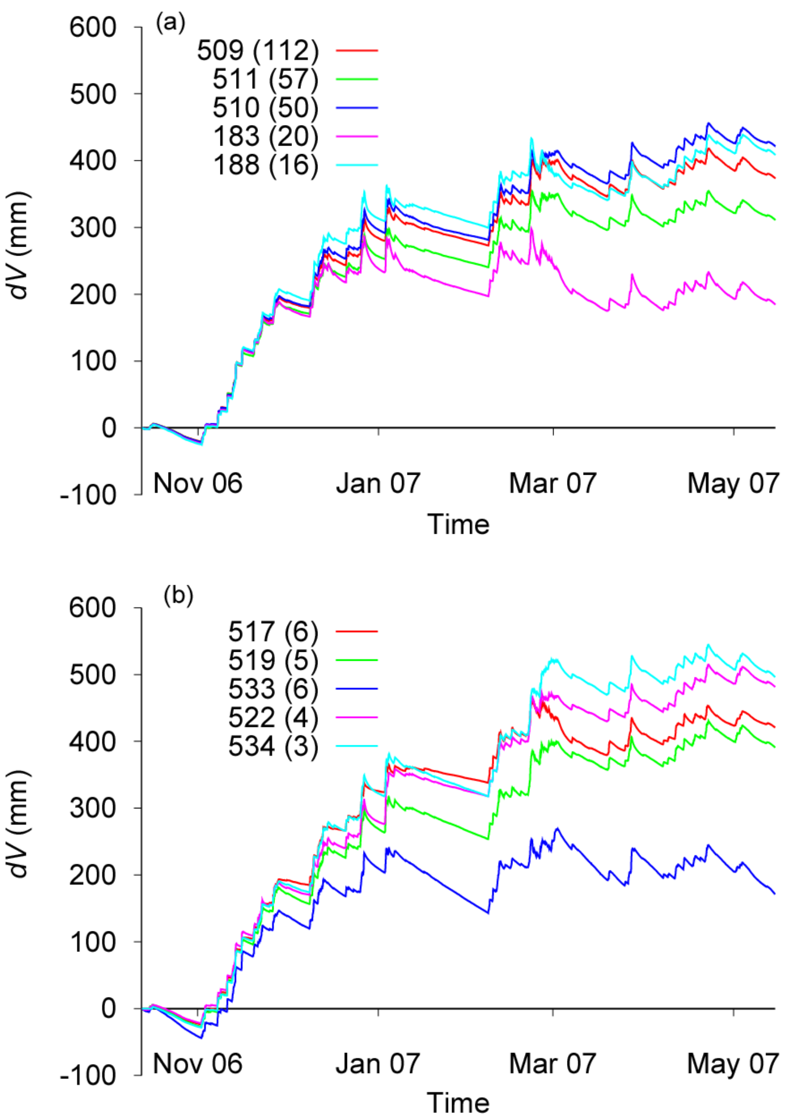

Figure 3 Temporal trends of total storage changes (*dV*) during the wet season for the 10 gauged watersheds. The numbers in the legend represent watershed ID number with their sizes in km^2 in the parentheses.

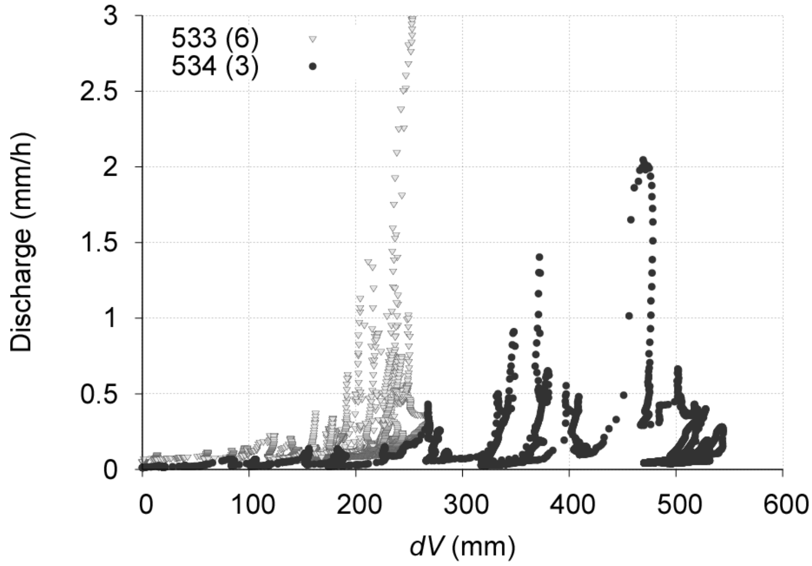

Figure 4 The relationship between change in total storage dV and discharge Q from two sub-watersheds. Both watersheds have almost no runoff response when the dV values are below 200 mm at No. 533 watershed (6 km^2) and 350 mm at No. 534 (3 km^2), respectively. At the No. 533 watershed, the dV plateaus around the 200 to 250 mm level, whereas at the No. 534 watershed, the dV increases gradually even after active runoff activation and finally it exceeds 500 mm.

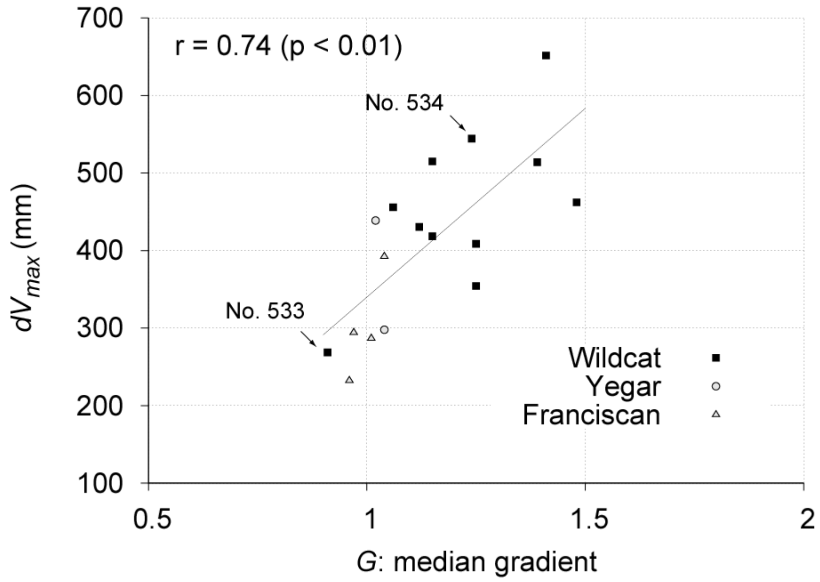

Figure 5 The relationship between median gradient G for each sub-watershed and its maximum total storage change (dV_{max}) during the rainy season. The dot colors represent the three basic geologic units that comprise the overall watershed area.

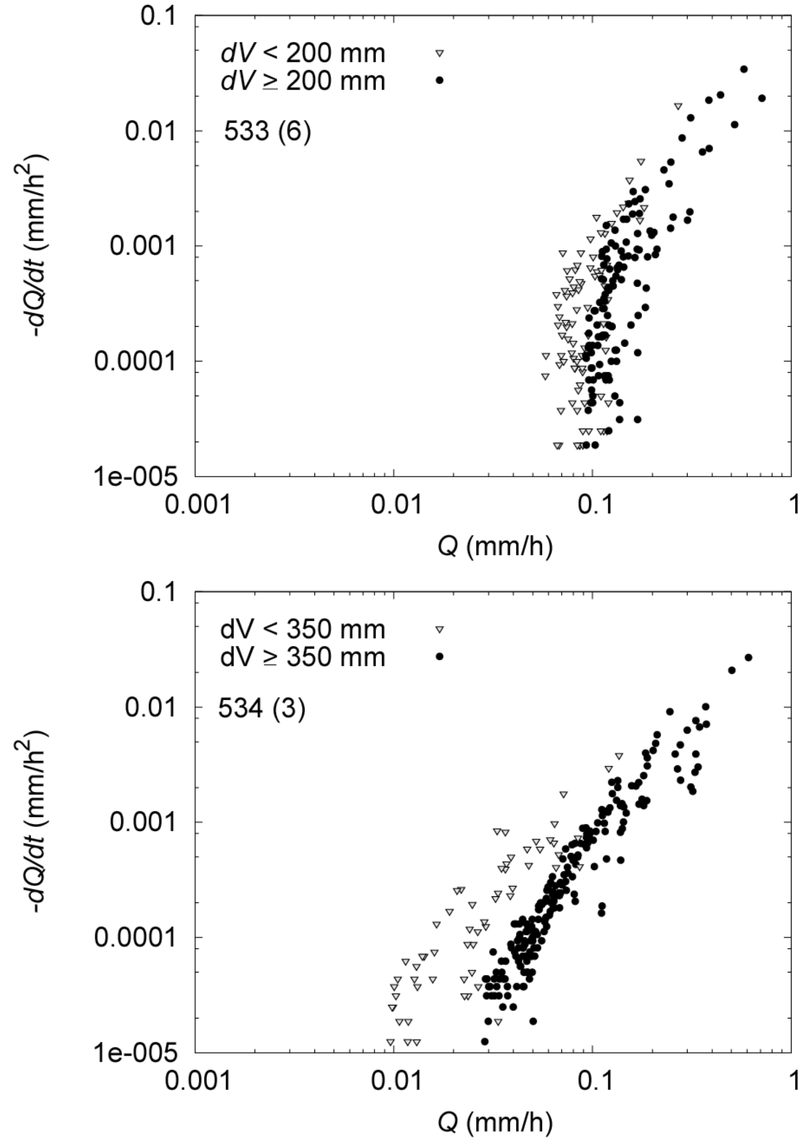

Figure 6 The relationship between recession rates (*-dQ/dt*) and runoff *Q* from two sub-watersheds (No. 533 and No. 534). The plots are classified into two groups based on the *dV* values (*dV* = 200 mm and *dV* = 350 mm were used as the thresholds to distinguish before and after wet-up).

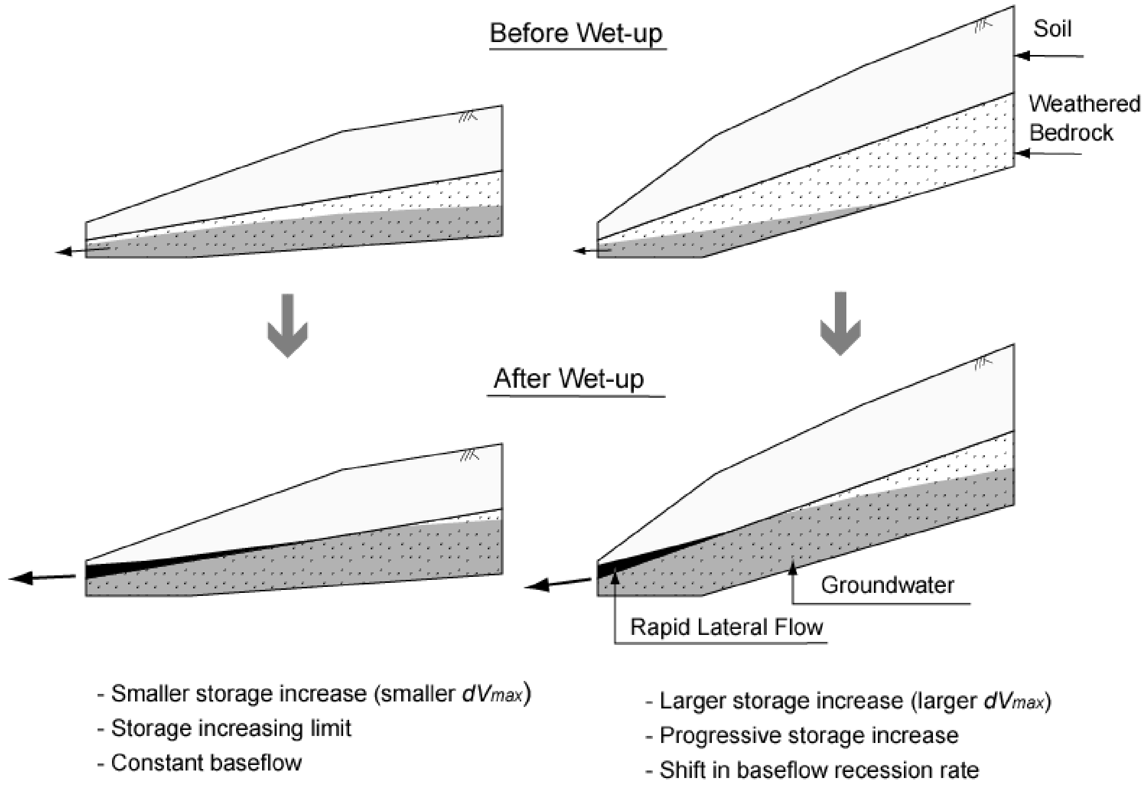

Figure 7 A conceptual diagram of hydrologically active bedrock hypothesis. A steeper watershed (right side) requires more water to fill the weathered bedrock zone even if the depths of the soil and bedrock layers are the same as the gentler sloping watershed. In addition, the area of bedrock groundwater exfiltration to the soil layers tends to be smaller at the steeper watershed; as a result it still stores some additional water even after the commencement of rapid runoff response.