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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 watershed storage changes during a rainy season (dV). 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 watersheds 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 the maximum volume of 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. The amount of water required to activate rapid rainfall–runoff response is larger for steeper watersheds where the more restricted expansion of seepage from bedrock to the soil limits the connectivity between stored water and stream runoff.

KEY WORDS catchment; storage; rainfall-runoff; recession; water balance; bedrock

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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 good conceptual framework for overland flow type and occurrence based on aridity indices and precipitation intensity is now well defined in the literature (Kirkby, 2005; Reaney *et al.*, 2007). Of course, many landscapes do 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). However, unlike overland flow shedding processes, 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 the question: How much water can a watershed store? Watershed storage is the key function of a watershed (Black, 1997) and a fundamental descriptor for catchment classification (Wagener *et al.*, 2007). It is also important as a primary variable of rainfall–runoff models (e.g. Sugawara and Maruyama, 1956; Brutsaert, 2005; Kirchner, 2009), a controlling factor for hydrogeochemical evolution (e.g. Burns *et al.*, 2003) and directly related to water resource and watershed resilience under climate change (Tague *et al.*, 2008). Despite the importance of watershed storage, few attempts have been made to estimate the volume of subsurface water storage at the headwater watershed scale (McDonnell, 2003; McDonnell, 2009). Attempts to measure storage, especially in the subsurface, 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). There have been a number of studies in groundwater hydrogeology and hillslope hydrology using ground-based geophysical approaches to characterize the subsurface (e.g. Collins *et al.*, 1989) and, recently, using gravity-based satellite measures for large river basins (Rodell *et al.*, 2006; Troch and Durcik,

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2007; Strassberg *et al.*, 2009). Nevertheless, we have not been able to answer the fundamental question for headwaters where most watershed runoff is generated (Soulsby *et al.*, 2009). Answering such a question would help with understanding better and vexing questions of subsurface stormflow delivery mechanisms.

Of course, determining total water storage *senso stricto* is an impossible task, given the ill-defined bottom boundary condition. Here, we focus on the dynamic component of total watershed storage – the amount of storage change in a system over the course of a rainy season. 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 by Spence (2007) and Spence *et al.* (2010) provided a useful model of the large-scale storage and discharge relations at catchments with lakes and wetlands. Here, we build upon this earlier work and 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 ten 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 dynamic 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 hydrological 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 also has suggested that storage elements in the hillslope need to be filled before releasing water from the slope base (see Graham and McDonnell, 2010; Graham *et al.*, 2010; 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 as follows:

1. How much subsurface water can a watershed store?
2. How does dynamic storage differ between sites and scales? and
3. How does topography and geology influence dynamic 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

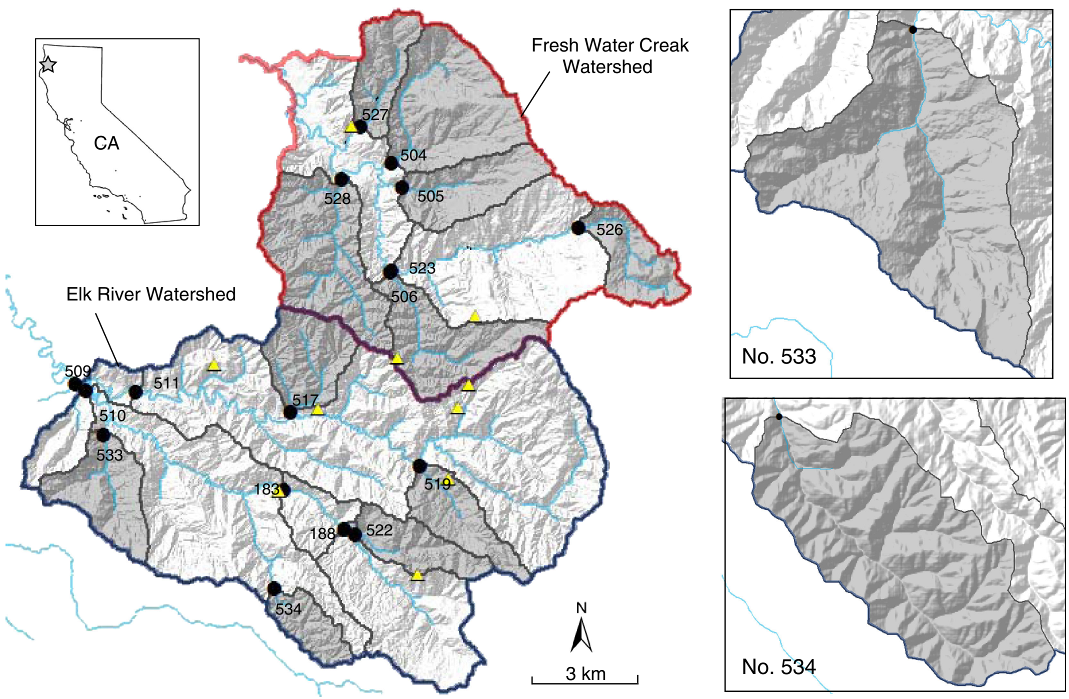

Figure 1. Map of Elk River watershed (110 km^2) and Freshwater Creek watershed (76 km^2). The black dots represent the 17 discharge gauging stations, and the triangles represent the ten rain gauges in the two adjacent watersheds

watershed (76 km^2), also is used for 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 composed mostly of a coniferous lowland forest community (stand age $\approx$ 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, which is fine grained, clay-rich thick sedimentary rocks. These rocks are predominantly marine sandstone, mudstone, and siltstone deposited as the sequence of the transgressive-regressive movement in the late Miocene to Middle Quaternary (Reid, 1999). The Wildcat Group deposits weather readily into loam to clay loam soils, typically named 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 (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 massive sandstone is cracked and fissured to create deep gravelly soils, whereas the argillite is prone to slaking and deep weathering and often is easily sheared. Because of the different erosion rates, slopes underlain by the Yager Formation often are 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).

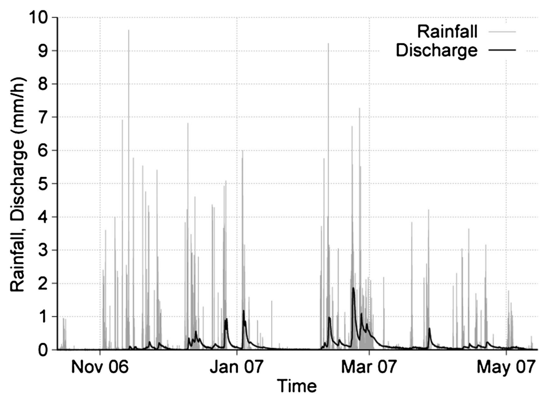

Figure 2. Watershed average rainfall and observed discharge at the outlet of the Elk River watershed (no. 509, 112 km^2) during a wet season (from 13 October 2006 to 15 May 2007)

METHODS

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 follows:

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

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 15 May 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 (nos. 500 and 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 1 h 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 the dynamic storage increase or decrease from $t = t_0$ to $t = t$. Because the absolute volume of the watersheds' total storage cannot be quantified using the water balance method, we focused exclusively on how their dynamic storage changed over time from the beginning to the end of the rainy season. Errors in these estimates could be caused by discharge observations (our approach was based on the U.S. Geological Survey gauging protocol), watershed-average rainfall estimates (using our methods described above), and evapotranspiration estimates. In terms of the spatial averaging of gauged rainfall, the interpolation method we used, or the Thiessen polygon method, does not account for the orographic effect of rainfall. We used this simple interpolation method to avoid any subjective error into the interpolation algorithm, given that the ten rain gauges were considerably well distributed throughout the watershed including at valley bottoms and ridges along forest roads.

In addition, because the standard deviation of the total rainfall among the ten gauging stations was only 73 mm (6% of total rainfall: 1187 mm), the errors induced by the interpolation was thought to be negligible. In Northwest Californian forest watershed, fog water condensation by leaves also may be important and allow augmented transpiration, especially during summer months (Burgess and Dawson, 2004). However, in terms of annual water balance, Keppeler (2007) reported that the effect of fog-drip is relatively small compared with the annual rainfall (≈3%) based on the field measurement at the Caspar Creek Experiment Watershed also located at the Northern Californian coast. For a further detailed water balance analysis, interception by foliage, bark, and litters also should be explicitly treated because the total interception would account for as much as 25% of annual rainfall, and the difference between potential evapotranspiration and actual evapotranspiration reaches about 70 mm, corresponding to about 5% of the annual rainfall (Reid and Lewis, 2009). Thus, we should realize that the similar degree of uncertainty in our $E(t)$ estimate is included, which generally causes the underestimation of $dV(t)$. Nevertheless, given our focus on a rainy season, during which evapotranspiration is estimated to be about 230 mm – this error appears to be relatively small compared with the 1187 mm of rainfall and 594 mm (from the whole Elk River or no. 509 watershed) of runoff during the same period. Another potential error is from trans-boundary groundwater flux. 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). Quantifying this flux is very difficult if not impossible. Nevertheless, by focusing on relatively large watersheds ($> \sim 5\,km^2$), we argue that the influence of such a flux should be negligible compared with analysis at smaller headwater scales.

Recession analysis

Streamflow recession analysis is another powerful tool to investigate the characteristics of storage feeding streams (Tallaksen, 1995; Rupp and Selker, 2005; Brutsaert, 2008; Rupp and Woods, 2008). A recession curve contains valuable information concerning storage properties and aquifer characteristics (Tague and Grant, 2004; 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:

$$-dQ/dt = f(Q) \tag{2}$$

where f denotes an arbitrary function. We considered recessions only during nighttime periods to avoid errors associated with evapotranspiration (Kirchner, 2009). In addition, to avoid measurement noise in individual hourly measurements, we computed first average discharge for 4 h during the following period; (Q_1) 19:00–22:59, (Q_2) 23:00–02:59, and (Q_3) 03:00–06:59. 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. Data were excluded from the plot if rainfall during the periods of 19:00–02:59 and 23:00–06:59 exceeded 0.1 mm to avoid the impact of rainfall.

RESULTS

Total storage changes estimated by water balance analysis

Figure 3 illustrates the relative temporal changes in dynamic storage (dV) estimated by the water balance approach described in Equation (1), showing the storage in each of the Elk River 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 dynamic storage increased by about 400 mm during the rainy season. The increase was almost linear throughout November and December and then reached a peak 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 peak value because of rainfall events in February. It is interesting to note that, although a rainfall event in the end of February (20 February–4 March) was the largest of the measured rainfall events (total of 237 mm as averaged over the eight 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

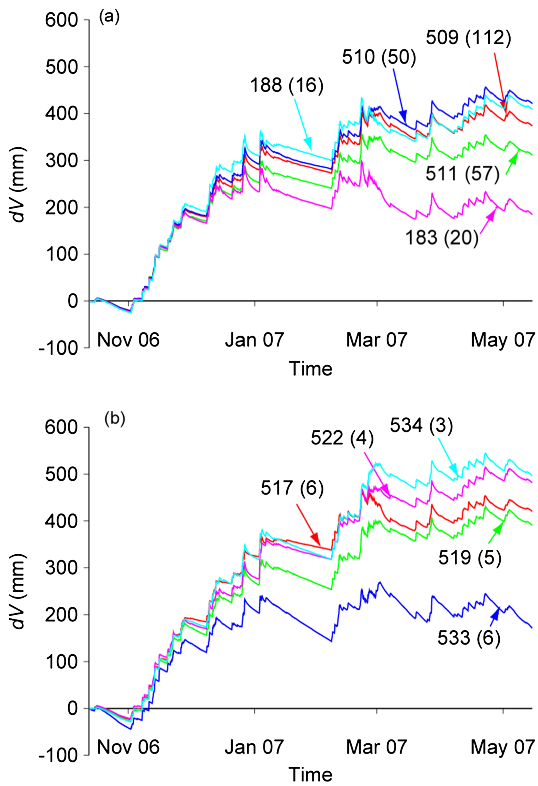

Figure 3. Temporal trends of total storage changes (dV) during the wet season for the ten gauged watersheds. The numbers in the legend represent watershed ID number with their sizes in square kilometers in the parentheses

Table I. Various topographic indices and maximum total storage change (dV_{max}) at each watershed are listed

Watershed	No.	*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.
Elk	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
	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
	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
	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
Fresh	523	22.8	1.01	16.6	2678	0.509	F	286.7
	528	12.0	1.39	24.4	924	0.501	W	514.1
	504	11.9	0.97	16.0	1961	0.449	F	294.3
	506	8.2	1.41	22.5	2198	0.358	W	651.7
	505	6.2	1.04	17.5	2111	0.441	F	392.4
	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

COR represents correlation between each topographic index and dV_{max}. *Area* is a watershed area (km^2). *G* is a median gradient [A 10-m resolution DEM is used to compute all the topographic indices including the median gradient (*G*), which is the median value of slopes for all grid cells in a sub-watershed. The slope value for each pixel is estimated as the maximum rate of elevation change between the cell and its eight-direction neighbors]. D_d is a drainage density. *R* is a relief (elevation difference between basin summit and basin outlet). *HYP* is a hypsometric integral [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]. *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 ($p < 0.05$).

more constant behavior (Figure 3b). However, the peak storages and the time required to reach the peaks 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 changes are best illustrated in *dV versus* discharge (*Q*) plots shown in Figure 4. These patterns shows that discharge in nos. 533 and 534

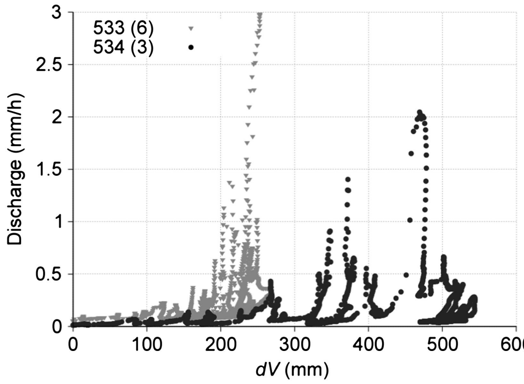

Figure 4. The relationship between change in total storage *dV* and discharge *Q* from two sub-watersheds. Both watersheds have almost no runoff response when the *dV* values are below 200 mm at no. 533 watershed (6 km^2) and 350 mm at no. 534 (3 km^2), respectively. At the no. 533 watershed, the *dV*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

watersheds was not activated until their *dV* reached 200 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 about an additional 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.

Topographic controls on total storage change

For each sub-watershed, we calculated a variety of topographic indices listed in Table I with our available 10-m resolution digital elevation model (DEM). We calculated also the maximum dynamic storage changes for each sub-watershed during this study period; hereafter, we denote this maximum dynamic storage change during this period as dV_{max}. Then, we computed the correlation coefficients between the topographic indices and dV_{max} using the data from all the sub-watersheds in both Elk River and Freshwater Creek watersheds. Table I summarizes the correlation coefficients between each topographic index and the storage. Among these indices, median gradient (*G*)

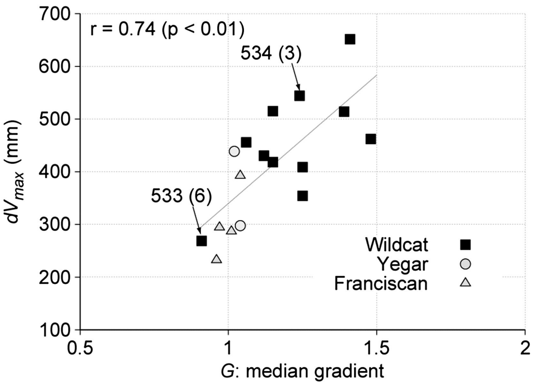

Figure 5. The relationship between median gradient G for each sub-watershed and its maximum total storage change (dV_{max}) during the rainy season. The symbols represent the three basic geologic units that comprise the overall watershed area

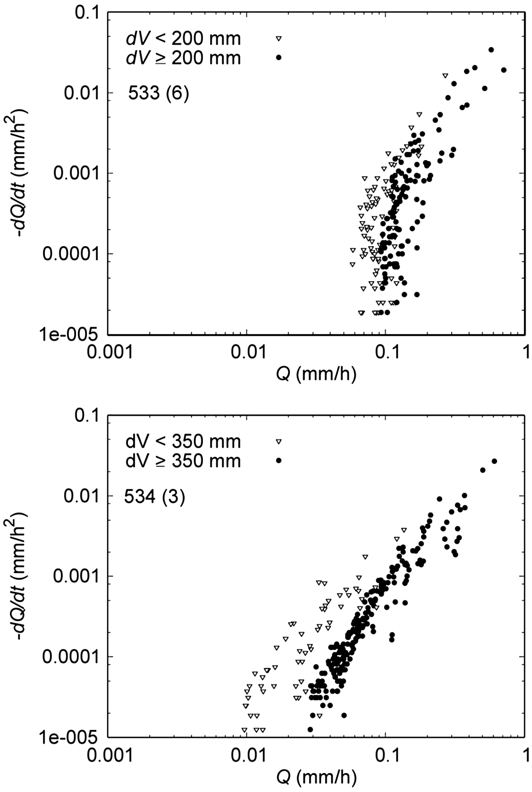

Figure 6. The relationship between recession rates ($-dQ/dt$) and runoff Q from two sub-watersheds (nos. 533 and 534). The plots are classified into two groups based on the dV values ($dV = 200$ mm and $dV = 350$ mm were used as the thresholds to distinguish before and after wet-up)

showed statistically significant positive correlation with dV_{max}. This positive correlation indicates that a watershed with steep slopes shows a larger dynamic storage increase during a rainy season than a watershed with milder slopes.

Although the median gradient metric (G) is objective and readily quantifiable, we acknowledge that there is undoubtedly a co-relation and co-evolution of local geology 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, whereas the Yager and Franciscan groups are a greater mixture of geologic conditions. Notwithstanding these complexities, the geologic variation within the sub-watersheds was overall relatively small with all the geologic groups within a class of marine-derived sedimentary rock.

Table I 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. However, the computed correlation coefficients shown in Table I did not show clear correlations between the volumetric indices and the watershed storage and storage change.

Recession analysis

Recession analysis was conducted for each sub-watershed, and the results are summarized in the form of Q *versus* $-dQ/dt$ plots in Figure 6. These analyses show contrasting results from nos. 533 and 534 watersheds. Recall that the no. 533 watershed is a gentler slope watershed with smaller dV_{max}, whereas 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 a 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 and one where dV is less than 350 mm, which was the amount of water required at watershed no. 534 to start generating rapid storm runoff, as described above.

DISCUSSION

So how much water can 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 macroscale hydrological laws. The method presented here of watershed intercomparison capitalizes on the extremely intensive gauging network – the densest of its kind that we are aware – rather than relying on mapped storage volumes (e.g. Spence *et al.*, 2010). 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. Our water balance approach was motivated by the visual observation of increasing baseflow levels through the wetting up season, onto which the wet season hydrographs are superimposed. 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. Of course, this represents the dynamic storage and not the total water storage in the watershed (because of the ill-defined bottom boundary problem). The simple water balance analysis showed how a watershed increases its dynamic storage in the beginning of a rainy season and then remains almost constant after reaching a peak 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 on the order of a few hundred millimeters with the individual watershed values depending on the local topographic and geologic properties.

Although each watershed showed distinct differences in its dynamic storage limit, each watershed did indeed reach a storage limit during the wetting up cycle – varying in timing by approximately 60 days. 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 work of Hewlett and Hibbert (1967) who viewed 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 also are 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 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.

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 because it suggests that catchments with steeper slopes tend to store more water. All things being equal, one might expect that catchments with gentle 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 with 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 caused by 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 watersheds in the California and Oregon Coast Ranges (see Montgomery and Dietrich, 2002, for review), revealed a very different sort of flow response, conditioned by permeable bedrock. If one considers permeable bedrock groundwater involvement in streamflow, as evidenced in the region by Anderson *et al.* (1997); Torres *et al.* (1998), and Anderson and Dietrich (2001), the positive relation between storage and topographic gradient immediately 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 groundwater 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 dynamic 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 with the milder slope. This expansion of exfiltration zones drastically changes the runoff generation response because this controls the connectivity between the stored soil water and stream flow (Fiori *et al.*, 2007).

Uchida *et al.* (2008) called this type of catchment system – with a permeable bedrock zone that stores and releases precipitation – a '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 and Dietrich (2002) explained their runoff generation mechanisms via deep permeable groundwater 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.

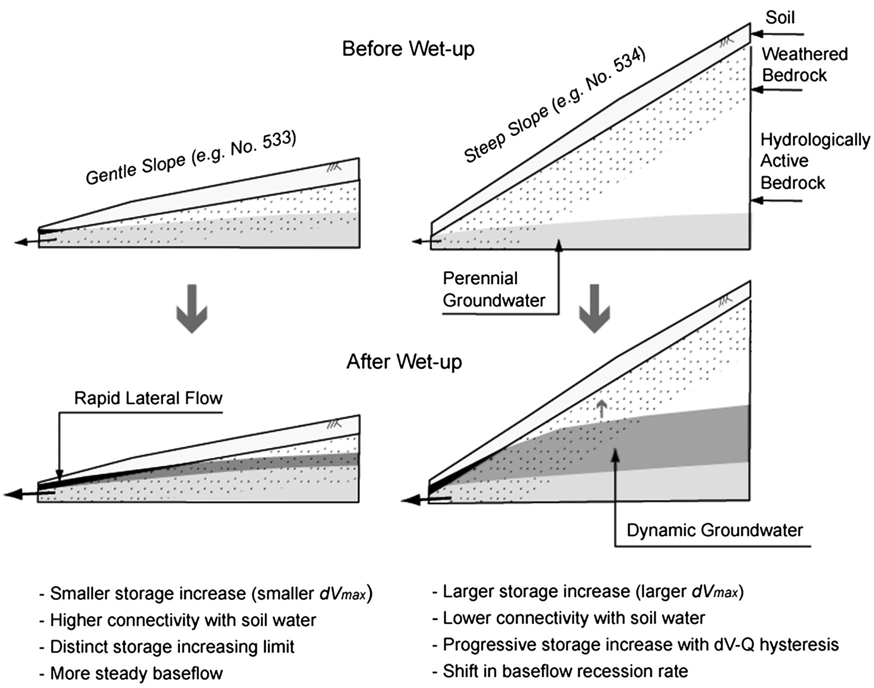

Figure 7. A conceptual diagram of hydrologically active bedrock hypothesis. A steeper watershed (e.g. no. 534, right side) requires more water to fill the weathered bedrock zone even if the depths of the soil and bedrock layers are the same as the gentler sloping watershed (e.g. no. 533, left side). 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

The results shown in Figures 4 and 6 also support the hydrologically active bedrock zone hypothesis. The gentle slope watershed, such as the no. 533 watershed, increased its dynamic storage up to about 200 mm and maintained almost the same level regardless more precipitation input. Alternatively, steeper watersheds, e.g. 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 groundwater starts seeping to the soil layer, creating saturated near stream zone, in which additional storm rainfall creates quick lateral saturated subsurface flow through better connection between the soil water and stream flow. This is when the storage rate increase becomes slower compared with the beginning of a wet season. At the same time, part of the slope can still 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 with the recession rate after the wet-up period. Our hypothesis is that when the groundwater 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 become 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. Linked to this active bedrock hypothesis is the difference in hydrological connectivity within catchments. It may be that the gentler no. 533 watershed has a better connected riparian zone; its *HYP* value shows that it is more concave than the no. 534 watershed and has flatter valleys (albeit within a generally incised topography overall). Because discharge will only react to hydrologically connected storage, the results obtained using a coarse value, such as dV_{max}, which includes both connected and disconnected storage, may need to be interpreted through this filter. Exploring these reductionist process details is a logical next step to the top-down analysis of data presented in this paper.

CONCLUSIONS

This work has explored watershed storage dynamics and function associated with collection and release of water across multiple nested watersheds in Northern California. 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 hydrological data. Our water balance analysis from the 17 nested macroscale watersheds revealed that each watershed stores different amounts (varying between 200 and 500 mm of precipitation) before actively generating storm runoff. The regression analysis between the maximum dynamic storage increase dV_{max}, and topographic indices showed that watersheds with steeper slopes store more water than watersheds with gentler slopes. 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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