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Key Points:

- Watershed memory effects quantifiable at varying time scales
- Variability in annual runoff largely explained by previous year's storage state
- Storage-release curve explains differences in Q for watersheds with different ET

Correspondence to:
F. Nippgen,
fabian.nippgen@duke.edu

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Watershed memory at the Coweeta Hydrologic Laboratory: The effect of past precipitation and storage on hydrologic response

Fabian Nippgen[1,2], Brian L. McGlynn[1], Ryan E. Emanuel[3], and James M. Vose[4]

[1]Department of Earth and Ocean Sciences, Nicholas School of the Environment, Duke University, Durham, North Carolina, USA, [2]Department of Land Resources and Environmental Sciences, Montana State University, Bozeman, Montana, USA, [3]Department of Forestry and Environmental Resources, North Carolina State University, Raleigh, North Carolina, USA, [4]Center for Integrated Forest Science, USDA Forest Service Southern Research Station, North Carolina State University, Raleigh, North Carolina, USA

Abstract The rainfall-runoff response of watersheds is affected by the legacy of past hydroclimatic conditions. We examined how variability in precipitation affected streamflow using 21 years of daily streamflow and precipitation data from five watersheds at the Coweeta Hydrologic Laboratory in southwestern North Carolina, USA. The gauged watersheds contained both coniferous and deciduous vegetation, dominant north and south aspects, and differing precipitation magnitudes. Lag-correlations between precipitation and runoff ratios across a range of temporal resolutions indicated strong influence of past precipitation (i.e., watershed memory). At all time-scales, runoff ratios strongly depended on the precipitation of previous time steps. At monthly time scales, the influence of past precipitation was detectable for up to 7 months. At seasonal time scales, the previous season had a greater effect on a season's runoff ratio than the same season's precipitation. At annual time scales, the previous year was equally important for a year's runoff ratio than the same year's precipitation. Estimated watershed storage through time and specifically the previous year's storage state was strongly correlated with the residuals of a regression between annual precipitation and annual runoff, partially explaining observed variability in annual runoff in watersheds with deep soils. This effect was less pronounced in the steepest watershed that also contained shallow soils. We suggest that the location of a watershed on a nonlinear watershed-scale storage-release curve can explain differences in runoff during growing and dormant season between watersheds with different annual evapotranspiration.

1. Introduction

Variability in hydrologic response within and among watersheds has long been attributed to watershed physical and biological properties [e.g., *Hewlett and Hibbert*, 1966; *Jencso and McGlynn*, 2011; *Jencso et al.*, 2009; *Nippgen et al.*, 2011; *Sidle et al.*, 1995; *Western et al.*, 1999] and climate [e.g., *Arora*, 2002; *Budyko*, 1974; *Jones et al.*, 2012]. While our understanding of the role of spatial variability on runoff has increased, understanding the influence of temporal variability has received less attention, especially the influence of past precipitation.

Studies on the effect of antecedent conditions on runoff date back to at least the first half of the 20th century [e.g., *Hamon*, 1964; *Kohler and Linsley*, 1951; *Linsley et al.*, 1949], coinciding with (or even preceding) increased attention to subsurface contributions during stormflow generation [e.g., *Hewlett and Hibbert*, 1963; *Tsukamoto*, 1963]. Some of the earliest and simplest methods to account for past precipitation are antecedent precipitation or wetness indices (API) that take into consideration the amount of precipitation before some point in time, e.g., a runoff event. APIs have been most often applied on a storm basis [e.g., *Fedora and Beschta*, 1989; *Kim et al.*, 2005; *Linsley et al.*, 1949; *Sidle et al.*, 2000; *Sittner et al.*, 1969] and rarely take into account time periods longer than a few weeks or months prior to the hydrologic event in question. *Fedora and Beschta* [1989] for example found that the effect of antecedent precipitation on runoff events vanished after 72 h in several watersheds in the north-central Oregon Coast Range. However, hydrologic models can include one or more storage terms and the modeling community devotes much effort to quantifying and parameterizing storage [e.g., *Garcia and Tague*, 2015; *Hailegeorgis et al*., 2015]. Further, many studies point out the importance of soil water storage on runoff processes, from individual events to annual water balances. *Pathiraja et al*. [2012] for example showed that incorporating antecedent moisture conditions on monthly time scales reduced underestimates of design flood peaks for watersheds in Australia. *Jothityangkoon and Sivapalan* [2009] demonstrated that incorporating carry-over storage from previous storms improved the accuracy of model-based annual water balance predictions in watersheds in Australia and New Zealand. In a modeling study using 30 day lag correlations, *Orth and Seneviratne* [2013] showed that variability in shallow soil moisture was propagated to streamflow in 100 watersheds across Europe. On an even longer time scale, *Istanbulluoglu et al*. [2012] and *Tomasella et al*. [2008] found that precipitation variability led to a carry-over of groundwater storage that affected the water balance of the following year for watersheds in Nebraska and the Amazon, respectively. This carry-over groundwater could possibly sustain base flow in years with less than average precipitation. Understanding the capacity for hydrologic systems to buffer precipitation fluctuations is especially important considering projections of increased drought severity in the coming decades (5th IPCC report) [*Hartmann et al*., 2013]. Such knowledge could have far reaching implications for policymakers and water resources managers as they anticipate and manage for increasing water scarcity. Despite the importance of this issue, quantifying the temporal dimension of past precipitation's effect on the current water balance of a watershed has received only modest attention.

The Coweeta Hydrologic Laboratory in the Appalachian Mountains of western North Carolina is one of the longest-running forest hydrology research sites in the United States. While many of the hydrologic experiments conducted at Coweeta investigated the influence of vegetation on water yield, some of the earliest research addressed the effect of soil storage on saturated and unsaturated flow [*Hursh and Brater*, 1941; *Hoover and Hursh*, 1943]. In what is now an iconic experiment, *Hewlett and Hibbert* [1963] demonstrated that $\sim$11 m^3 of saturated soil in a $\sim$14 m long, sloping concrete trough (closed at the top to eliminate evaporation from the soil surface) was able to sustain outflow for more than 140 days. *Hewlett and Hibbert* [1963] estimated that if the trough outflow after 140 days were scaled up to the size of a headwater catchment, it would be sufficient to sustain low flows observed during growing seasons in Coweeta watersheds. In the *Hewlett and Hibbert* [1963] experiment, the soil in the trough was saturated and then simply drained without being recharged again by sprinkling or rainfall. Natural watersheds, on the other hand, experience successive cycles of precipitation events and seasons. Transferring what we learned from the *Hewlett and Hibbert* [1963] experiment to the watershed scale at Coweeta and elsewhere naturally raises the question of how precipitation history, with cycles of wet and dry periods, affects soil water storage and how this precipitation legacy in return influences contemporary and future hydrologic response. We term the legacy of these influences watershed –or system– memory. Memory is a watershed characteristic which is set by a combination of climatological forcing and watershed properties (e.g., slope, soil depth, vegetation) that can impact current and future hydrologic response. Here, based on 21 years of precipitation and runoff data made available from five watersheds of the Coweeta Hydrologic Laboratory in southwestern North Carolina, we address the following questions:

1. How do variability in precipitation and system memory modulate hydrologic response across monthly, seasonal, and annual time scales?
2. What is the influence of storage state on the annual water balance and how different are storage dynamics under coniferous and deciduous vegetation?

2. Methods

2.1. Site Description

The Coweeta Hydrologic Laboratory is located in southwestern North Carolina in the Nantahala Mountain Range of the southern Appalachian Mountains (lat. 35°03′N, long. 83°26′W, Figure 1). Coweeta was established in 1934 and is a site of ongoing watershed experiments that include understanding hydrologic and ecologic changes following vegetation manipulations, such as conversion from deciduous hardwoods to evergreen conifers [*Swank and Crossley*, 1988]. The Coweeta Basin encompasses 1626 ha with watershed elevations ranging from 675 m near the outlet of Coweeta Creek in the east to 1592 m on the ridge in the west. For this study, we used runoff and precipitation data for five watersheds of the 16 currently gauged

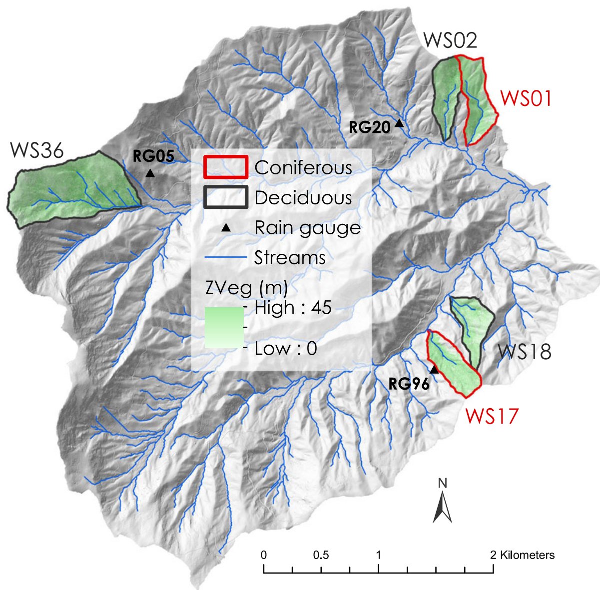

Figure 1. Map of the Coweeta Hydrologic Laboratory with the outlines of the five watersheds used in this study. The red color-coding denotes coniferous watersheds and black denotes deciduous watersheds; color scheme remains consistent for all subsequent figures.

watersheds in the basin that varied by aspect, elevation, and major vegetation type. WS01 and WS02 are south-facing paired watersheds, and WS17 and WS18 are north-facing paired watersheds. While WS02 and WS18 contain mixed hardwoods generally found in the southern Appalachians, WS01 and WS17 were converted to eastern white pine (*Pinus strobus*) trees (evergreen conifers) in the 1950s. At the beginning of the conversion process, WS01 was burned in 1942 and subsequent regrowth of hardwoods was prevented with chemicals until white pine was planted in 1957. WS17 was clear-felled in 1940 and subsequent regrowth was prevented until white pine was planted in 1956 [*Swank and Crossley*, 1988]. WS36, the fifth watershed, is a high elevation (1015–1541 m), east facing watershed consisting of mixed hardwoods (Figure 1).

Watersheds 01, 02, 17, and 18 are similar in size (12.3–15.4 ha) and are located in the lower (elevations ranging from 704 to 1051 m), eastern part of the Coweeta basin, while WS36 is 49.3 ha and located in the higher, western part of the Coweeta basin. All watersheds are generally steep with mean slopes exceeding 26°. WS36 is the steepest of the five watersheds with a mean slope exceeding 30° (Table 1).

Climate at Coweeta is classified as Marine, Humid Temperate (Köppen classification, Cfb) to Humid Subtropical (Cfa) [*Swift et al.*, 1988], especially at lower elevations. Annual precipitation at a weather station in the valley floor near the main basin outlet averages 1791 mm for the period 1937–2011. Precipitation is almost uniformly distributed over the year with slightly more (10–14%) precipitation in the winter months. There is a strong elevation effect on precipitation along the east-west axis with an increase of approximately 5% per 100 m, while this effect is almost negligible for the north-south axis slopes [*Swift et al.*, 1988]. Average annual air temperature near the main basin outlet is 12.6°C with an average monthly low of 3.3°C in January and a high of 21.6°C in July.

The soils at Coweeta are deep sandy loam inceptisols and older more developed ultisols [*Swank and Crossley*, 1988,1988b]. *Hoover and Hursh* [1943] drew a boundary at approximately 1000 m elevation above which the watersheds are typically steeper and have shallower soils. According to *Swank and Douglass* [1975], the general depth of the regolith is approximately 7 m. Shallow groundwater well installations by the authors in

Table 1. Means and Medians for Landscape Structure Metrics for the Five Watersheds[a]

Metric		WS01	WS02	WS17	WS18	WS36
Vegetation		Conif.	Decid.	Conif.	Decid.	Decid.
Size (ha)		15.4	13.0	13.5	12.3	**49.1**
Elevation (m)	Mean	834	853	894	821	**1288**
	Median	819	851	894	816	1276
Slope (°)	Mean	27.1	26.9	28.5	27.8	**30.6**
	Median	27.9	27.5	29.2	29.0	**31.7**
Gradient to creek	Mean	0.46	0.41	0.48	0.45	**0.57**
	Median	0.44	0.40	0.48	0.45	**0.57**
Dormant seas. pot. insolation (kWh/m^2)	Mean	927	982	500	568	839
	Median	975	1002	485	552	915
Growing seas. pot. insolation (kWh/m^2)	Mean	1384	1409	1187	1218	1342
	Median	1403	1419	1198	1229	1387

[a]Numbers in bold highlight differences between the low-elevation and the high-elevation watersheds.

the low-elevation watersheds 01, 02, 17, and 18 at Coweeta resulted in completion depths of up to 3.5 m to bedrock, while *Hales et al*. [2009] found soil depths in the high elevation watershed 36 to be only 1.2–1.8 m, based on the excavation of 15 soil pits in various landscape positions.

2.2. Hydrometric Measurements

Runoff data were stored as breakpoint data (a value is recorded not based on a fixed time step but only when stage changed) for the water years (WY) 1991–2011. Runoff was measured using 90° V-notch weirs at WS01, WS02, and WS17, and with 120° V-notch weirs at WS18 and WS36 and recorded using Fisher-Porter Analog to Digital punched recorders and Stevens Type A/E loggers. The breakpoint data were converted to daily runoff totals (mm/d). Runoff data were missing due to weir repairs for WY 2011 for WS02 and WYs 1991–1993 for WS36. Missing data due to weir repairs for WS17 (14 August 2003 to 27 January 2004) and WS18 (3 September 2003 to 6 June 2004) were interpolated using double mass curves with WS01 and WS02, respectively. However, only five out of a total 105 watershed years (five watersheds, 21 years each) were missing.

Precipitation data from three rain gauges associated with the watersheds (see Table 2 for rain gauge-watershed associations) were available as daily precipitation totals for the same time period (WYs 1991–2011). Precipitation was measured using Belfort Universal Recording Rain Gauges and converted to watershed precipitation with weighting factors established by Swift in 1968 [see *Swift et al*., 1988] from isohyetal maps based on a network of approximately 50 recording rain gauges operating from 1938 to 1958 [*Swift et al*., 1988].

2.3. Landscape Analysis

We calculated the means and medians of three simple topographic metrics: elevation, slope, and gradient to creek (the gradient of flow paths from a cell to the stream). The metrics were calculated based on 1 m LIDAR (LIght Detection And Ranging) data, which were resampled to 5 m resolution to obtain a resolution high enough to adequately capture the topography in the small watersheds but coarse enough to avoid oversensitivity to microtopography (e.g., fallen trees). To assess differences in available solar radiation among watersheds, we calculated potential solar insolation after *Böhner and Antonic* [2009] for all five watersheds using SAGA open source GIS software (http://www.saga-gis.org). Potential solar insolation and thus energy availability can be used as a proxy for differences in potential evapotranspiration between the watersheds, especially between the north and south-facing pairs. This is of particular interest here because the structure of the Coweeta basin contains both north and south facing watershed pairs, each with different vegetation types. We calculated potential insolation for all watersheds on an annual and seasonal basis.

2.4. Empirical Analysis of Runoff and Precipitation Data

2.4.1. Lag-Correlations

We analyzed the runoff and precipitation time series at annual, seasonal, and monthly resolutions. We calculated monthly totals and averages of precipitation (P), runoff (Q), and runoff ratios ($RR=Q/P$) and their standard errors to compare the runoff and precipitation regimes of the five watersheds. We calculated the difference in monthly Q totals between the deciduous watersheds and the coniferous watersheds to investigate potential differences in hydrologic response between the two vegetation types.

We calculated Pearson correlation coefficients between P and RR at annual, seasonal, and monthly time scales. In order to quantify the impact of past precipitation—and hence watershed memory—for each watershed, the correlations were calculated as lag-correlations: A lag of 0 means that one time step's RR was correlated with the same time step's P, at a lag of 1 a time step's RR was correlated with the previous time step's P etc. We calculated lag correlations for up to 12 time steps (i.e., 12 months, 12 seasons, 12 years). Further, we calculated lag-correlations for the P time series to test for autocorrelation in the P data.

Table 2. Rain Gauge Watershed Associations and Their Elevations

Rain Gauge	Watershed(s)	Elevation (m)
RG20	01 & 02	740
RG96	17 & 18	894
RG05	36	1144

For the annual statistics, we use the traditional water year (1 October to 30 September) as the baseline as this is the most commonly used cut-off for water budget calculations in the US. However, we also performed an analysis to test the sensitivity of our results on the start date of the year. For this we calculated annual lag-

Table 3. Studies With Published Evapotranspiration Time Series[a]

Study	Location	Vegetation	No. of Years	Mean Annual *ET* (mm)	Standard Deviation (mm)	Coefficient of Variation
Wilson et al. [2001]	Oak Ridge National Laboratory, Tennessee, USA	Mixed deciduous forest	5	571	16	0.03
Hanson et al. [2004]	Oak Ridge National Laboratory, Tennessee, USA	Oak forest	5	601	52	0.09
Lafleur et al. [2005]	Southeastern Ontario, Canada	Peatland	5	351	30	0.09
Stoy et al. [2006]	Duke Forest North Carolina, USA	Loblolly pine forest	4	657	74	0.11
Kosugi and Katsuyama [2007]	Central Japan	Cypress forest	3	735	15	0.02
Ryu et al. [2008]	Sierra Nevada foothills, California, USA	Grassland	6	319	44	0.14
Ohta et al. [2008]	Eastern Siberia, Russia	Larch forest	7	196	20	0.1
Jassal et al. [2009]	Vancouver Island, Canada	Douglas fir forest	10	404	22	0.05
Oishi et al. [2010]	Duke Forest North Carolina, USA	Deciduous Forest	4	633	26	0.04

[a]Shown here are mainly annual averages, standard deviation, and the coefficient of variation. This list is not supposed to be a comprehensive overview.

correlations with 12 varying start dates for the year, e.g., 1 January to 31 December, 1 February to 31 January, 1 March to 28/29 February, etc.

For seasonal statistics, we grouped data into growing season and dormant season, with growing season extending from 15 April to 14 October and dormant season extending from 15 October to 14 April of the following year. This is in agreement with the definitions previously used at Coweeta [e.g., *Vose and Swank*, 1994; *Jones and Post*, 2004]. While this division is appropriate for the low elevation watersheds, the seasons in WS36 are slightly different. *Hwang et al*. [2011] showed that the timing of leaf-on and leaf-off can differ by several weeks between the low and high elevations at Coweeta. However, for a consistent comparison (i.e., equal length growing and dormant seasons), we assigned all watersheds the same dates for growing and dormant seasons. As a consequence, potential seasonal effects on runoff in WS36 may be weakened as the actual growing season length in WS36 is likely shorter than the chosen 6 months period.

2.4.2. Storage Approximation

In addition to correlations between runoff ratios and precipitation at different temporal scales, we approximated changes in watershed storage in each year, ΔS, and annual watershed storage, S, from the annual 21 year P and Q time series. We developed a simple method (see equations (1)–(3) below) to estimate S and ΔS using annual P and Q from the 21 year time series, and average ET across the time series. Several studies demonstrated that average ET calculated from a long-term water balance is in good agreement with averages from–shorter–eddy covariance time series in those systems [e.g., *Kosugi and Katsuyama*, 2007; *Wilson et al*., 2001]. Note that relative storage, S, does not refer to an actual/absolute storage value, which is unknown, but rather represents the storage relative to an arbitrary storage starting value of 0 at the beginning of the time series. All uses of the term "storage" in this paper refer to the relative storage state and does not represent an absolute magnitude of watershed storage. This approach contrasts with the more typical water balance approach to calculating ET by assuming no annual changes in storage. Our approach assumes that annual evapotranspiration is relatively stable from year to year in this system where energy rather than water limits evapotranspiration. The assumption of relative constancy in evapotranspiration is not new and dates back at least 30 years. *Roberts* [1983], for example, described ET as "conservative" and noted there was little variability in ET in temperate European forest. Recently, *Fatichi and Ivanov* [2014] reported that ET at four sites was relatively insensitive to climatic variability. Relatively stable annual ET values have been reported for many other areas (both water and energy limited) in and outside the US [e.g., *Hanson et al*., 2004; *Jassal et al*., 2009; *Kosugi and Katsuyama*, 2007; *Lafleur et al*., 2005; *Nagler et al*., 2005; *Ohta et al*., 2008; *Ryu et al*., 2008; *Wilson et al*., 2001, see Table 3]. While these studies are not immediately transferable across climates, they suggest that interannual variability in ET in some climates and geographic regions may not be as sensitive to fluctuations in climate as is often assumed.

At Coweeta the annual precipitation (~1800 mm) typically exceeds estimates of potential annual evapotranspiration, suggesting that watersheds are primarily energy limited: *Rao et al*. [2011] used the *Priestley and Taylor* [1972] approach to estimate potential evapotranspiration. PET ranged from approximately 1400–1600 mm/yr for WS17 (average 1509 mm, standard deviation 63 mm, coefficient of variation 0.04) and approximately 1000 to <1200 mm yr^{-1} for WS18 (average 1079 mm, standard deviation 52 mm, coefficient of variation 0.05) for a 22 year period (1986–2007); other methods yielded much lower annual PET. It is worth noting that annual $P-Q$ in both the coniferous as well as the deciduous watersheds resulted in a much larger range (~700 mm) than the range of the PET calculations (~200 mm), suggesting that ΔS may be greater than ΔET. Further, since Priestley-Taylor is a radiation-based estimate of PET, the *Rao et al*. [2011] estimate does not reflect variability in P but rather general atmospheric conditions at Coweeta, which are relatively consistent between years.

The standard deviation of annual pan evaporation from the climate station located at the base of the Coweeta watershed over the period 1937–2011 was only 57 mm with a mean of 892 mm (coefficient of variation, $c_v = 0.06$). Estimates of actual evapotranspiration using sap-flux technology are only available for a few years: *Ford et al*. [2007] estimated ET for Coweeta WS17 be 1290 mm and 1292 mm in two consecutive years (2004–2005). These estimates are within 4% of the 1351 mm mean annual ET calculated for this study. For a 3 year period (2004–2006), *Ford et al*. [2010] found that growing season ET in the last year was ~29% lower than in the previous 2 years; however, not all components of ET (e.g., understory ET) were quantified and only a portion of the watershed was measured. It should be noted that *Oishi et al*. [2010] indicated that tree or plot level estimates of ET often suggest more variation in annual ET than is observed at larger spatial scales such as an eddy covariance footprint. In the North Carolina Piedmont, an area more prone to water-limitation than Coweeta due to much lower annual precipitation, *Stoy et al*. [2006] and *Oishi et al*. [2010] found relatively low interannual differences in ET for both a hardwood and a deciduous stand over a 4 year period. The studies summarized above provide confidence in our assumption of stable ET at Coweeta; nevertheless, we performed a sensitivity analysis to address how variable ET would impact the storage calculations and subsequent correlations with runoff ratios (please see Appendix A for this analysis).

With typical water balance applications over annual and multiyear time scales, ΔS is assumed to be negligible [e.g., *Eagleson*, 1978; *Budyko*, 1974], and differences between $P-Q$ are attributed to variability in ET. Here we suggest that annual ΔS is much greater than ΔET and therefore we use average ET for each watershed calculated from a 21 year period to estimate annual ΔS and infer the influence of storage on annual runoff dynamics. ET_{mean} for each watershed over the entire study period was calculated as

$$ET_{mean} = \frac{\sum_{i=1}^{n} (P_i - Q_i)}{n} \tag{1}$$

where n is the number of years in the study period.

The change in storage, ΔS, in each year i was calculated as

$$\Delta S_i = P_i - Q_i - ET_{mean} \tag{2}$$

where ΔS_i is the change in annual storage (mm) for each watershed, P_i is annual precipitation (mm), Q_i is annual runoff (mm), and ET_{mean} is the average annual ET for a watershed (mm) over the 21 year study period. ΔS represents the annual deviation from an undetermined storage state, whose value was arbitrarily set to set zero at the beginning of the time series to better show deviations from the starting value. Relative storage, S, for each year and watershed was subsequently calculated as

$$S_i = \sum_{i=1}^{n} \Delta S_i \tag{3}$$

and represents the cumulative deviation from the undetermined start value.

This approach assumes no net change in storage between the beginning and the end of the 21 year time year time series. Similar to the erroneous assumption of no net change in storage between individual years, assuming no net change in storage over a longer period of time would be incorrect as well, as a change in storage is equally–or more–likely over a long period of time than from one year to the next. This would be especially pronounced when calculated from the trough of a drought period to the peak of a wet period. However, if we assumed a net change in absolute storage of 300 mm over the 21 year study period, the average error in relative storage for every year would be 14.3 mm and thus negligible.

The percolation to deeper groundwater was assumed to be small at Coweeta because of impermeable bedrock [*Hatcher*, 1988]. However, if there were losses to deeper groundwater, they would not affect our analyses, as they would constitute a simple offset to the water balance equation. Further, this offset would

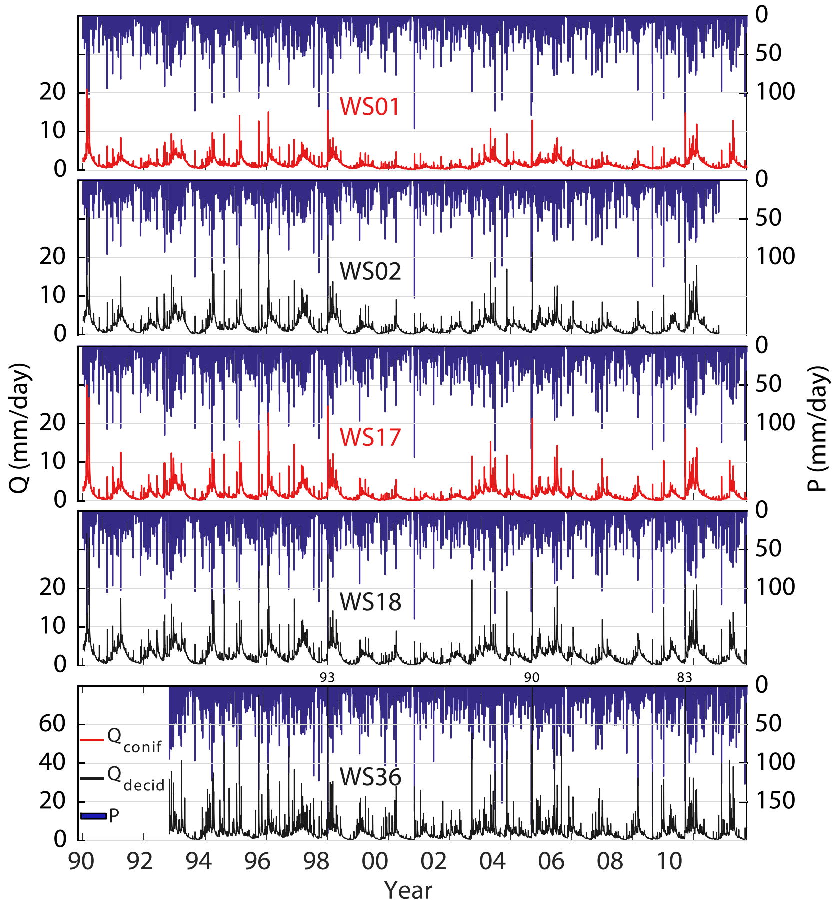

Figure 2. Precipitation and Runoff data for the 22 year study period for all five watersheds. Note the greater secondary *y* axis range for WS36. Small numbers above WS36 show runoff (mm/d) for days where runoff exceeded *y* axis maximum (80 mm/d).

effectively reduce the *ET* water balance component and thus make a potential water-limitation in the watersheds more unlikely (for annual values it would be $ET + GW_{losses} = P - Q$ instead of $ET = P - Q$).

We used linear regressions between annual *P* and *Q* and calculated the residuals of this relationship to examine factors that might explain the variability of this relationship from year to year (e.g., *S*). For example, a watershed with less storage capacity might exhibit smaller residuals than a watershed with high storage capacity. To test this idea, we calculated lag-correlations for the residuals of the regression between annual *P* and annual *Q*, and annual watershed storage, i.e., a lag of 0 is the correlation with one year's residual to the same year's storage *S*, a lag of 1 is the correlation between a year's residual and the previous year's storage term. Based on the storage calculation, one year's storage denotes the following year's storage starting point, i.e., initial conditions. All correlations were calculated as Pearson correlation coefficients.

Due to pronounced seasonal differences in *ET* between the growing and dormant seasons, especially in the deciduous watersheds, it was not possible to apply the same methodology to subannual time scales.

3. Results

3.1. Watershed Characterization

Based on their location and elevation within the greater Coweeta watershed, WSs 01, 02, 17, and 18 can be characterized as low-elevation watersheds while WS36 is a high-elevation watershed. The four low-elevation watersheds are structurally very similar (Table 1), differing primarily in aspect and vegetation. WS36, however, is steeper and larger (3–4 times larger in area) than any of the four other watersheds.

As expected, the south-facing watersheds receive more potential solar insolation over the course of a year than the north-facing ones (Table 1). The difference between the aspects is especially pronounced in the

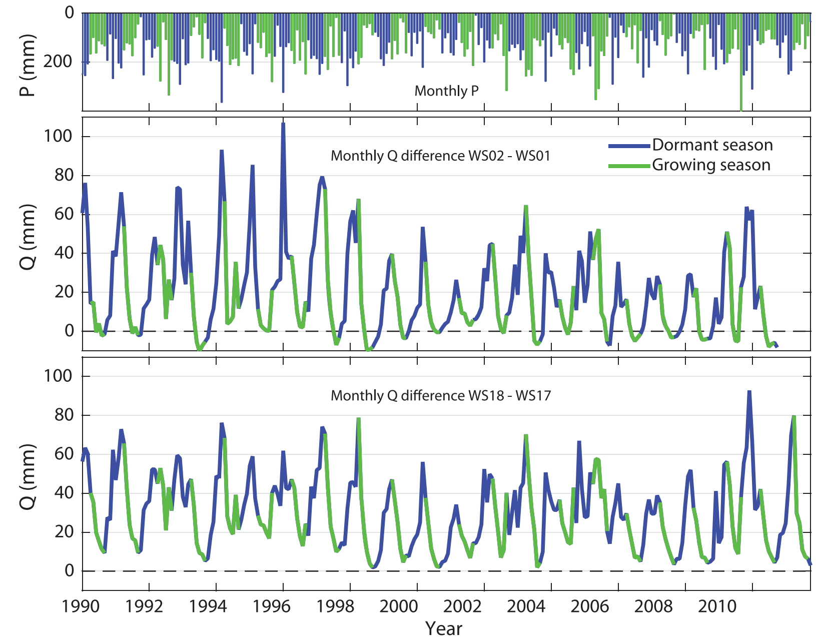

Figure 3. Monthly Q differences between the deciduous and coniferous watershed for (middle) the south-facing watersheds and (bottom) the north-facing watersheds. Monthly precipitation for WS01 is shown in the top plot for context. Blue color denotes the dormant season and green color denotes the growing season.

dormant season, when the south-facing watersheds (WS01 and WS02) receive on average 79% more direct solar radiation than the north facing ones (WS17 and WS18). This difference decreases to only about 16% more radiation during the growing season.

Mean annual precipitation was 1719 mm for WS01, 1771 mm for WS02, 1968 mm for WS17, 1923mm for WS18, and 2116 mm for WS36. There was an annual difference of approximately 200 mm precipitation between the north and south-facing watershed pairs despite having similar mean elevations.

3.2. Climate and Lag-Correlations

The daily data revealed clear seasonality with high runoff in the winter/spring and a period of decreased flows in the summer (Figure 2). The amplitude of this seasonality was variable with pronounced wet (1994–1998) and dry (1999–2002, 2006–2008) periods generally coinciding with high and low precipitation periods.

Superimposed on the seasonality in precipitation and runoff were numerous, distinct, short-duration peaks generated by individual storm events. The deciduous (WS02 and WS18) and coniferous (WS01 and WS17) watersheds exhibited similar response to precipitation in terms of timing but with dampened peaks and lower runoff magnitudes evident in the coniferous watersheds. The high elevation watershed (WS36) showed higher runoff throughout the year with a "flashier" response to precipitation, e.g., higher peak flows, than the low elevation watersheds.

Distinct differences in runoff behavior between the deciduous and coniferous watersheds were apparent in monthly Q totals (Figure 3). Monthly stream flows for coniferous and deciduous watersheds were most similar toward the end of the growing season (Figure 3, denoted with green), and most dissimilar near the end of the dormant season (Figure 3, denoted with blue), with differences often exceeding 70 mm/month, especially in the south-facing watersheds.

Average monthly Q data (over the full record length) highlight the strong seasonality evidenced in the individual daily and monthly totals. Average monthly P was highly variable between months but showed no clear seasonality (Figure 4, top). However, P was, on average, greatest in January (175–212 mm) and lowest in July (113–144 mm). Monthly Q averages (Figure 4, middle) as well as the monthly averages of RR (Figure 4, bottom) showed a clear seasonal pattern. Q was highest in March, coinciding with high P values. Q

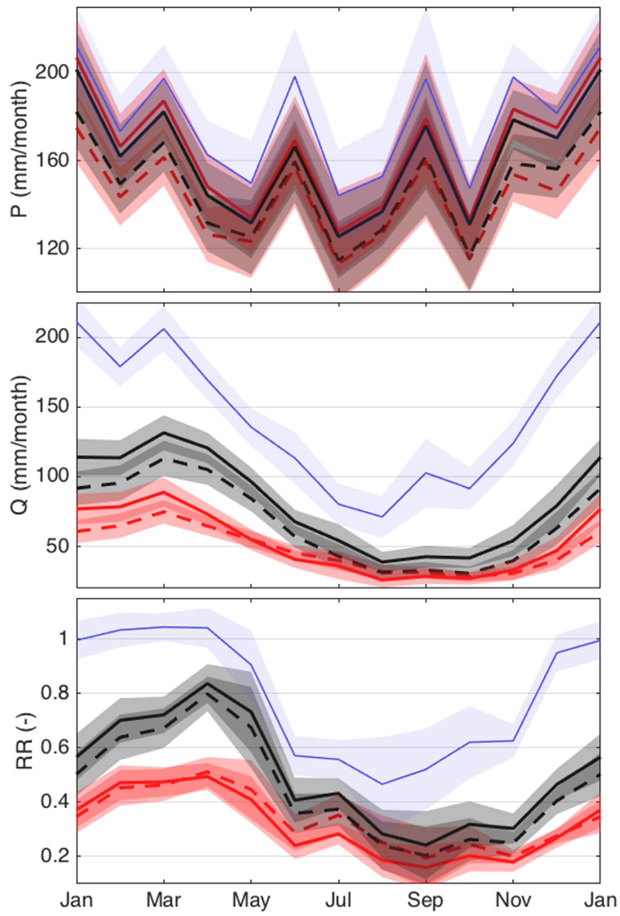

Figure 4. Monthly averages of Precipitation (*P*), Runoff (*Q*), and Runoff Ratios (*RR*) over the simulation period. Shaded areas are the standard errors of the respective time series (darker shades indicate overlap). Red denotes coniferous watersheds and black denotes deciduous watersheds. Blue denotes the high elevation watershed.

typically reached its minimum between August and October, highlighting the strong effect of both precipitation timing and *ET* on runoff.

The deciduous watersheds had greater *Q* and *RR*s than the coniferous watersheds for nearly all months. Over the course of the growing season, *Q* and *RR* became more similar, and then diverged again in the fall. Evergreen trees can transpire year round [*Ford et al*., 2007; *Swank and Miner*, 1968], whereas transpiration from deciduous trees effectively ceases with leaf senescence in the fall until leaf out in the next spring. Because of this, the differences in *Q* and *RR*s between the vegetation types were greatest during the dormant season from November through May. Additionally, interception evaporation from leaf surfaces is also higher for coniferous trees due to a much higher leaf area index in both dormant and growing season [e.g., *Ford et al*., 2010; *Helvey*, 1967; *Neary and Gizyn*, 1994; *Swank*, 1968].

There was no strong aspect effect evident in runoff ratios. While WS18 exhibited a slightly higher runoff ratio than WS02 throughout the year, the runoff ratios for WS01 and WS17 were almost identical in the winter months, and in the summer WS01 even exceeded the runoff ratios of WS17. However, one-sample *t* tests on the differences between the runoff ratios of the deciduous (WS18–WS02) and coniferous watersheds (WS17–WS01) determined the differences in *RR*s were significantly different from 0 for both deciduous as well as coniferous watersheds, although the actual differences in annual *RR*s were small (on average 0.05 for the deciduous watersheds, 0.02 for the coniferous watersheds).

3.2.1. Annual Analysis

Annual runoff ratios ranged from a minimum of 0.15 in WS01 to a maximum of 0.91 in WS36 (Figure 5). Average annual *RR*s for the coniferous watersheds were 0.33 (WS01) and 0.31 (WS17), while the deciduous watersheds averaged 0.45 (WS02) and 0.50 (WS18) (Figure 5). The high elevation watershed (WS36) had consistently higher *RR*s with an average annual *RR* of 0.77 (Table 4). The deciduous and coniferous watersheds plot as closely matched pairs with no apparent aspect effect.

A lag-correlation analysis between annual *P* and annual *RR* (Figure 6, top) showed significant correlations for all watersheds without lag (one year's *RR* with the same year's *P*) as well as a lag of one year (one year's *RR* with the previous year's *P*), indicating a strong 1 year memory effect. The lag-correlations became insignificant after the first year. There were no strong differences between the low-elevation coniferous and deciduous watersheds. WS36 exhibited similarly strong correlations for both no lag and a one-year lag ($r_p = 0.68$). Interestingly, it appears that the strength of the correlation depends on the start date for the water balance calculation. Choosing the wettest time of the year (April) as the start date (year running from 1 April to 31 March), the runoff ratio of a year is mostly affected by the same year's *P*, while when beginning the year in the drier part (e.g., the regular water year, 1 October to 30 September) the previous year's *P* has a greater or equally great effect on a year's *RR* (Figure 7). This shift could highlight a pronounced influence

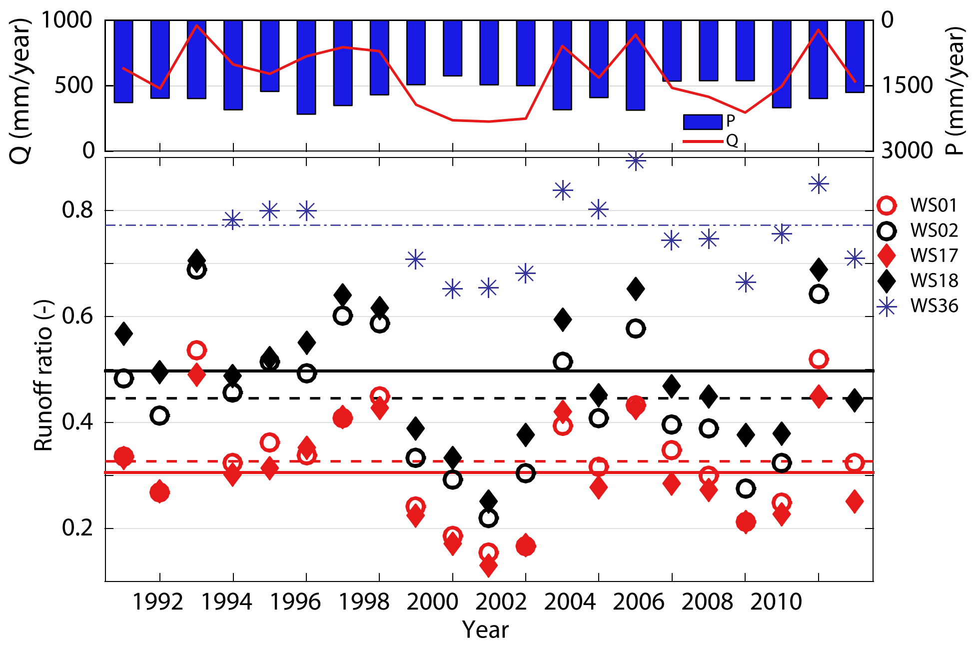

Figure 5. Variability in annual runoff ratios for the five watersheds. Diamonds denote the south-facing watersheds (WS01 and WS02) and open circles denote the north-facing watersheds (WS17 and WS18). The gray crosses denote the high elevation watershed (WS36). The red color-coding denotes coniferous watersheds (WS01 and WS17) and black denotes deciduous watersheds (WS02 and WS18). The dotted lines represent the mean runoff ratios over the 21 year study period. For context, the top plot shows *P* and *Q* for WS01 representative for all watersheds.

of the growing season *P* closest to the start date of the water balance calculation and may hint to a strong seasonal dependence on the *RR*. For the remainder of the paper, we will use the 1 October to 30 September definition for a water year, as outlined by the United States Geologic Survey (USGS).

There was no significant autocorrelation in the *P* time series (1 year lag averaged over all five watersheds $r_p = 0.28$, $p_{val} = 0.27$), suggesting the correlations between *P* and *RR* were not affected by autocorrelation in the *P* data.

3.2.2. Seasonal Analysis

Growing season (15 April to 14 October) *P* was on average 14–20% less (depending on watershed) than dormant season (15 October to 14 April) *P* (Table 5); however, these differences were not statistically significant when testing the medians (Kruskal-Wallis test) of growing and dormant season *P* or differences (Wilcoxon rank sum test) between dormant and growing season pairs (i.e., one dormant season paired with the following growing season). Comparison of growing season and dormant season *RR* series for each watershed showed different behavior for the low-elevation watersheds and the high-elevation watershed (Figure 8). While the *RR*s of growing and dormant season within the low-elevation watersheds were not significantly different from each other (two-sided Wilcoxon rank sum test at the 5% significance level), the high-elevation watershed showed distinct and significant growing and dormant season *RR*s (two-sided Wilcoxon rank sum test at the 5% significance level, Table 5, Figure 8).

For the low-elevation watersheds, correlations between seasonal *P* and *RR* were strongest at a lag of one season (Figures 9 and 6, middle), while WS36 exhibited stronger correlations ($r_p = 0.46$, $p_{val} < 0.001$, rightmost column Figure 9 and Figure 6, middle) without a lag. In contrast to the lower elevation watersheds, the correlation at a lag of one season was low and insignificant ($r_p = 0.22$, $p_{val} = 0.2$). However, when considering separately growing season *P* to dormant season *RR*, and dormant season *P* to growing season *RR* (Figure 10), the correlation at a lag of one season also became significant for the

Table 4. Mean Annual Values for Runoff (*Q*), Precipitation (*P*), Runoff Ratios (*RR*), and Mean *ET* ($P - Q$)[a]

	WS01	WS02	WS17	WS18	WS36
P (mm)	1722	1776	1973	1929	2116
Q (mm)	576	812	623	982	1654
RR	0.33	0.45	0.31	0.5	0.77
ET (mm)	1146	964	1351	946	462

[a]Please note that *RR* were calculated for each year and then averaged, so due to rounding errors *Q*/*P* may not equal *RR*.

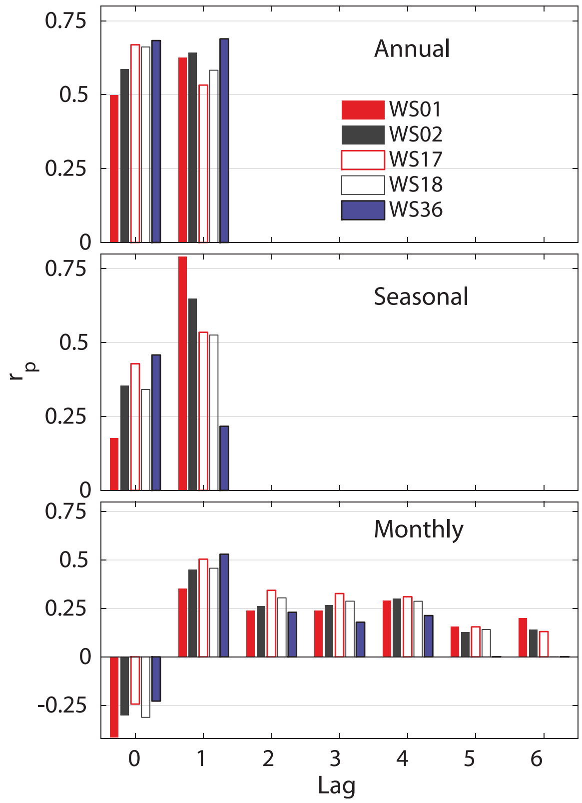

Figure 6. Lag-correlations for between precipitation and runoff ratios for (top) annual, (middle) seasonal, and (bottom) monthly time scales. The lag on the *x* axis is (top) months, (middle) seasons, or (bottom) years. All correlations shown are significant at the 0.05 level.

high elevation watershed. The correlations improved for all watersheds and were similar for the low-elevation watersheds (average $r_p = 0.74$ and $p_{val} < 0.01$ for dormant season *P* to growing season *RR*, and $r_p = 0.83$ and $p_{val} < 0.01$ for growing season *P* to dormant season *RR*), however the increase in correlation strength was greatest for the high-elevation watershed ($r_p = 0.56$ and $p_{val} < 0.05$ for dormant season *P* to growing season *RR*, and $r_p = 0.83$ and $p_{val} < 0.01$ for growing season *P* to dormant season *RR*). This is due to the stronger separation between dormant season and growing season *RR*s at the high elevation watershed. Since the dormant seasons are slightly wetter than the growing seasons, the global regression line (regression for all seasons) has a nonsignificant slope.

Correlations for all watersheds became nonsignificant for more than one season, e.g., one year's dormant season *P* did not affect next year's dormant season *RR* (Figure 6, middle).

3.2.3. Monthly Analysis

Lag-correlation analysis between monthly *P* and *RR*s for all watersheds showed significant lags for up to 6 months, with the highest correlations at a lag of 1 month (averaged $r_p = 0.45$, averaged $p_{val} \ll 0.01$, Figure 6, bottom). Due to the large sample size (n = 252 months over 21 years), correlations were significant even for small values of r_p. The highest correlation was at a lag of 1 month, meaning that the previous month's *P* had a stronger influence on the *RR* than the same month's *P*. The lag-correlations gradually leveled off after 1 month and became insignificant after 6 months (after 4 months for WS36). The negative but significant correlation at 0-lag (averaged $r_p = -0.29$, averaged $p_{val} < 0.01$) is likely caused by the general *P* regime with alternating low and high *P* months (see Figure 4, top). The lag correlations followed a similar pattern for all five watersheds irrespective of vegetation type or elevation (Figure 6, bottom).

3.3. Storage

Solving the water balance for storage required the calculation of average-watershed *ET*. Average annual *ET* calculated from the water balance over the 21 year study period was lowest for WS36 (462 mm), highest for coniferous WS01 and WS17 (1146 and 1351 mm, respectively), and intermediate for deciduous watersheds WS02 and WS18 (964 and 947 mm, respectively) (Table 6). ΔS and *S* followed similar patterns in the four low-elevation watersheds (Figure 11). The similarities were even more pronounced within the same vegetation type. While the two coniferous watersheds apparently maintained higher *S* than the deciduous watersheds, it is important to note that the 0 mm storage lines in Figure 11 represent undetermined and arbitrary storage start values at the beginning of the 21 year time period. Likewise, negative storage values simply imply that storage was below the undetermined storage state at the beginning of the time series. *S* fluctuated with *P*, but the absolute change in storage also depended on *Q* due to the nature of the water balance calculation. The highest *S* gain (greatest ΔS) was typically reached in high *P* years with below average annual *Q*, e.g., in 2009 where *P* was high but *Q* was still low because of the prior dry period. The majority of the annual *P* then replenished storage. On the other hand, the greatest *S* decreases occurred during years with normal to low *P*, but high *Q*, e.g., 1993.

In general, a net increase in *S* lead to higher *Q* in the following year. However, the current year's *P* also affected *Q*, so a particularly dry or wet year could reverse or dampen the effect of storage state on *Q*

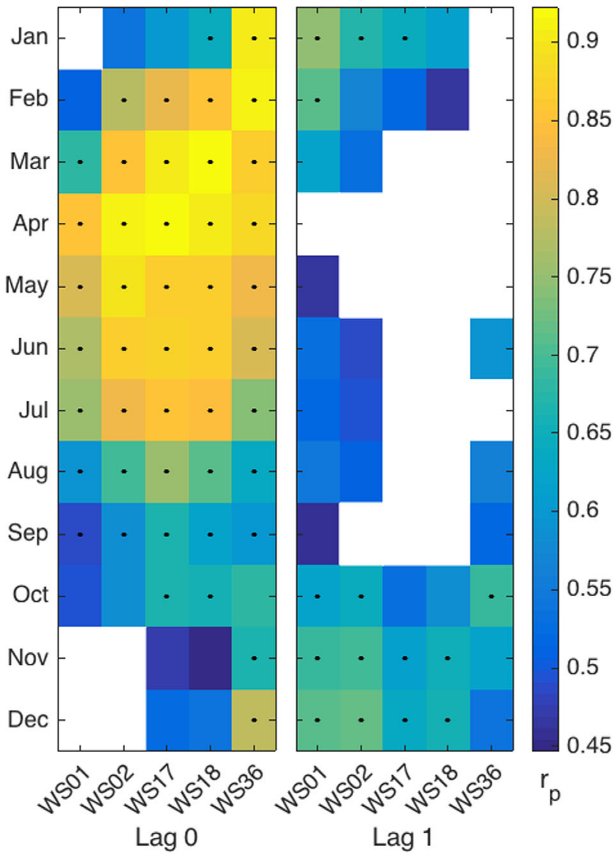

Figure 7. Annual lag-correlations with shifting start times for the year between *P* and *RR* at (left) lag 0 and (right) lag 1. Black marker denotes greater correlation strength as compared to the other lag. All correlations shown are statistically significant at the $\alpha = 0.05$ level.

(Figure 11). Changes in *S* could be great even if *P* were almost identical in two adjacent years (e.g., 1992 and 1993 or 2006–2008). For example, *S* can explain why from 1992 to 1993 *Q* increased despite similar *P*: the *S* built up in 1992 was carried over into 1993 and led to an increase in streamflow. Low *P* in 1993 lead to a decrease in *S* and a subsequent drop in *Q* in 1994.

Linear regressions between annual *P* and *Q* indicated strong, significant correlations, especially for WS36 (Figure 12, top row). The stronger correlation for WS36 suggests less storage capacity for the steeper, high-elevation watershed. We evaluated the effect of storage on the linear regressions between annual *P* and *Q* by analyzing the relationship between the residuals of the linear regressions and *S* with zero-lag and 1 year lag. In this case, zero-lag refers to the storage at the end of the year, while a 1 year lag denotes the storage at the end of the previous year and hence represents antecedent conditions for the year in question. At zero lag (1 year's residual correlated with the same year's storage), the correlations were weak and insignificant for WSs 01, 02, and 17, but significant for WSs 18 and 36 (Figure 12, middle). However, at a lag of one year (a year's residual correlated with the previous year's storage state), significant positive correlations emerged for all low-elevation watersheds (Figure 12, bottom row). The correlation for WS36 became positive as well but was weaker than without lag, indicating no significant influence of storage on the residuals of the regression between annual *P* and *Q*. This corroborates the interpretation that storage has no strong influence on the regression between *P* and *Q* for the high-elevation watershed. In contrast, the strong positive correlations for the low-elevation watersheds suggest a strong influence of antecedent storage state on the relationship between *P* and *Q*. Interestingly, while all low-elevation watersheds exhibit significant correlations at lag one, there appears to be a difference in correlation strength for the different vegetation types, with the coniferous watersheds having higher correlation coefficients than the deciduous ones. While the insignificant correlation for WS36 is likely caused by limited storage capacity and steeper slopes, these factors are likely not the cause for the weaker correlation in the deciduous watersheds since the physical characteristics of all four low-elevation watersheds are similar. Here it may be caused by the more pronounced wet and dry periods experienced by the deciduous watersheds as a result of limited dormant season transpiration.

4. Discussion

4.1. The Effect of Watershed Memory on Hydrologic Response Across Different Time Scales

Lag-correlations between monthly, seasonal, and annual precipitation totals and runoff ratios indicate strong influence of past precipitation on present runoff (Figure 6). Our storage correlations corroborate the lag correlation results that system memory significantly influences runoff behavior across time scales. The shorter the observed time scale (e.g., monthly versus annually), the more important was the previous time step's precipitation for the observed runoff ratio of any given time period, e.g., the difference in correlation between zero-lag and a lag of one time step. At the monthly time scale, correlations remained

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Table 5. Climate Statistics for Precipitation (*P*), Runoff (*Q*), and Runoff Ratios (*RR*) for Growing (grow) and Dormant (dorm) Season for All Five Catchments

	WS01	WS02	WS17	WS18	WS36
P_{dorm}	924	957	1093	1064	1124
P_{grow}	795	814	875	859	972
Q_{dorm}	321	488	382	591	1021
Q_{grow}	255	322	238	387	619
RR_{dorm}	0.34	0.5	0.34	0.54	0.91
RR_{grow}	0.32	0.39	0.26	0.45	0.61

significant for lags of up to 7 months (Figure 6, bottom). The strong correlation at shorter monthly lags highlights the importance of system memory on shorter time scales, as has been documented by others [e.g., *Kim et al*., 2005]. Significant but weaker correlations at longer monthly lags are in agreement with memory effects of up to 3 months for the entire North Carolina Coastal Plain region [*Anderson and Emanuel*, 2008]; *Rose* [1998] found memory effects of more than 6 months for the coastal plain of Georgia, USA. The dissipation of water could be expected to be much faster in the mountains relative to coastal areas because of higher gradients and different geology; however, deep soils and high soil water storage capacity may offset differences in gradients.

The influence of past precipitation was also evident at seasonal time scales. Similar to the monthly time scale, the lag-correlations were strongest at a lag of one season (Figure 6, middle). In contrast to the monthly lag-correlations, correlations at 0 lag were positive at the seasonal scale. This indicates that when integrated over a longer period of time, precipitation of one time step gains importance for the runoff ratio of that time step. Even WS36, where *Hewlett and Hibbert* [1966] found higher quick flow ratios because of decreased storage capacity, shallower soils, and steeper slopes, exhibited a significant correlation at a lag of one season (Figure 10).

Increasing the time scale from seasonal to annual, both the same as well as the previous year's precipitation have a nearly equal effect on the runoff ratio of a year (Figure 6, top), highlighting the assumption that when integrated over longer time steps, the current time step gains more importance.

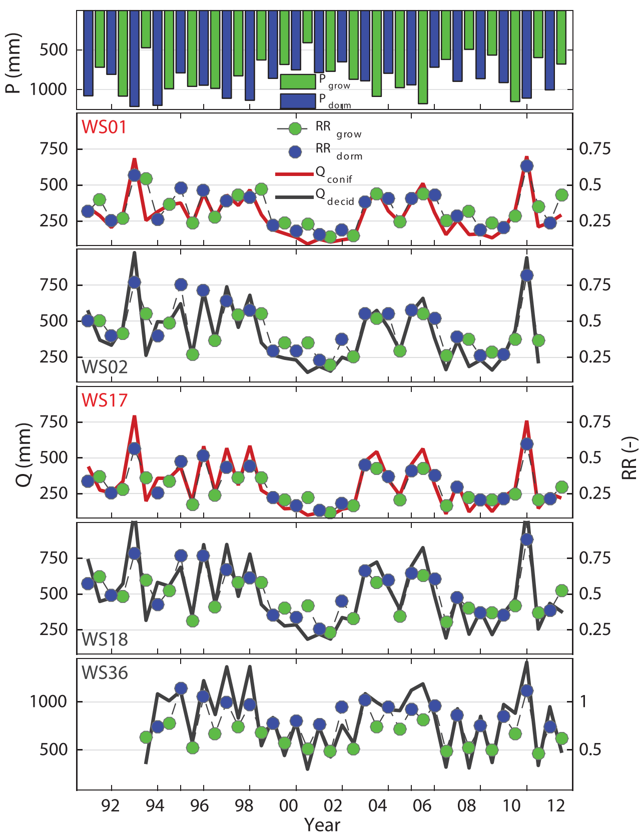

Figure 8. Seasonal precipitation (*P*), runoff (*Q*), and runoff ratios (*RR*) for all five watersheds.

While the lag-correlations showed a significant memory effect across all watersheds and temporal scales, there were no detectable effects of vegetation type or aspect. This suggests that past precipitation influences runoff ratios similarly across watersheds, despite differences in the absolute *RR* values caused by different vegetation types. The importance of antecedent conditions on runoff has long been documented [e.g., *Kohler and Linsley*, 1951] in numerous studies, often centered on storm runoff assessment [e.g., *Fedora and Beschta*, 1989; *Kim et al*., 2005; *Linsley et al*., 1949; *Sidle et al*., 2000; *Sittner et al*., 1969]. However, the lag-correlation calculations demonstrated that it may be necessary to consider antecedent precipitation over different lengths of time. For example, to make predictions about a month's water balance it may be sufficient to know the precipitation of several months back, but to assess a full year's water balance it may be necessary to know the

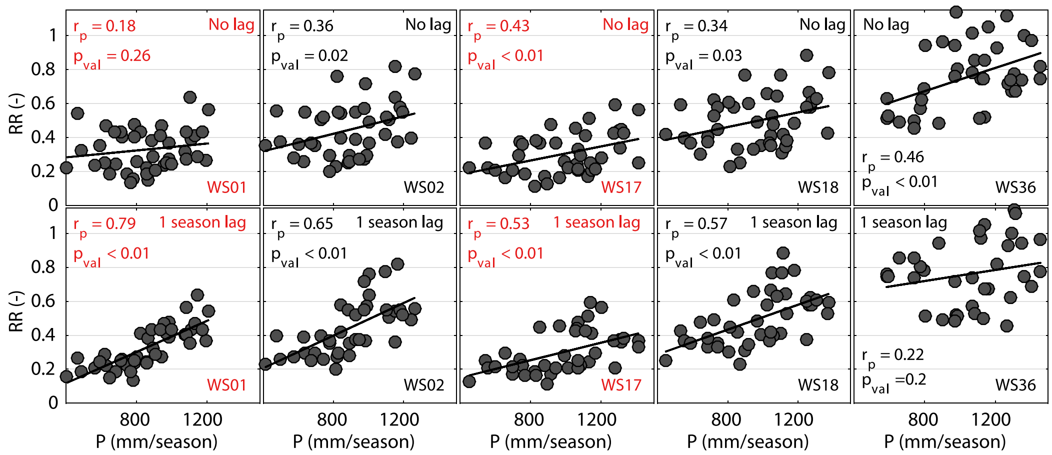

Figure 9. Scatterplot of seasonal *P* versus seasonal *RR* without (top) lag and (bottom) lagged by one season for all watersheds.

precipitation history of a much longer time period. *Merz and Bloschl* [2009] showed that for Austrian watersheds even mean annual precipitation can inform prediction of event runoff ratios, suggesting that besides exerting control on the annual water balance, annual precipitation totals also influence event runoff ratios by setting longer-term storage conditions.

4.2. The Effect of Storage on the Annual Water Balance in Coniferous and Deciduous Watersheds

While the lag-correlations suggest that watershed memory is likely responsible for the observed patterns, no actual observations of storage were used in calculating the correlations. Our main assumption was that annual *ET* at Coweeta was relatively consistent across years (as described in the Methods section).

Here, the storage calculated from the annual water balance using the 21 year averages of *ET* ($\Delta S = P_i - ; Q_i = ET_{mean}$) revealed a similar temporal pattern of storage over the study period for all five watersheds (Figure 11). This is not surprising since the general rainfall and runoff patterns were similar among the five watersheds. It is important to note, however, that storage (Figure 11, dashed black lines) should not be treated as an absolute value nor compared across watersheds, as the absolute storage states, i.e., the amount of water stored in the watershed at any given time, were undetermined. While the average annual

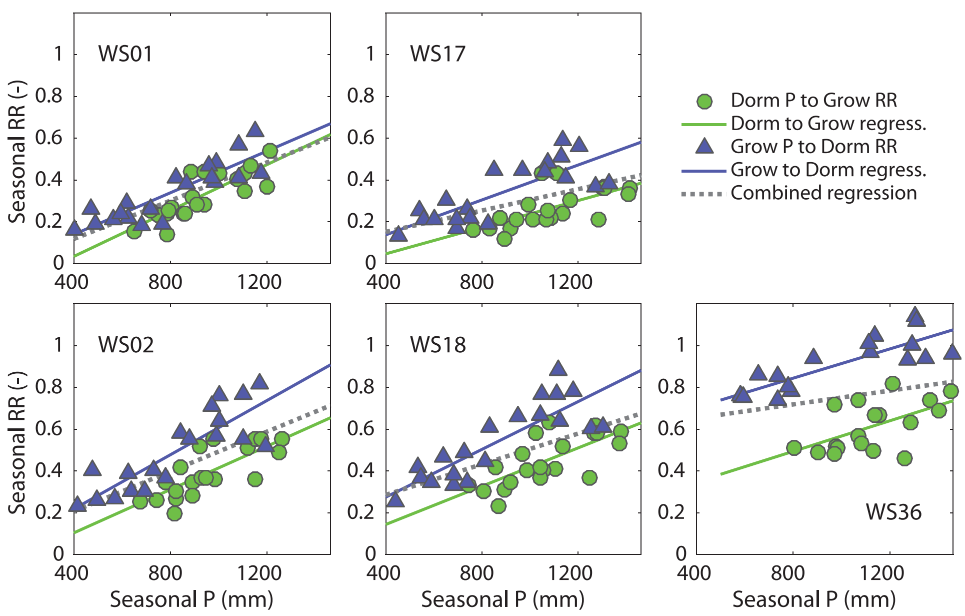

Figure 10. Seasonal *P* versus seasonal *RR* for dormant season *P* to growing season *RR* and for growing season *P* to dormant season *RR*.

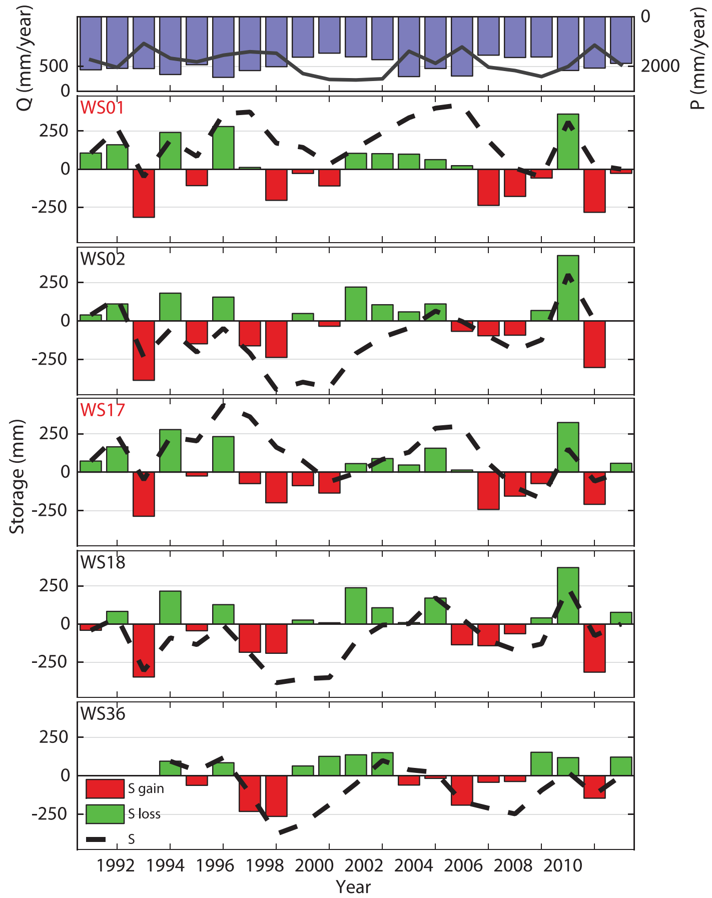

Figure 11. Time series of storage changes. ΔS was calculated as $P - Q - ET_{mean}$. Bars are changes in storage from year to year (ΔS), black line is cumulative *S*. (top) *P* and *Q* from WS01, representative of the general trend in all watersheds. Please note that the cumulative storage values (dashed black line) are not indicative of absolute storage values because of the arbitrary start value.

change in storage among all watersheds from was just 141 mm, the storage fluctuations within 1 year can be much greater as shown by *Sayama et al.* [2011] who calculated storage changes for watersheds in California within one wet season as up to 500 mm.

Watershed memory greatly affected annual storage dynamics across both vegetation types. One year's storage gain or loss was primarily determined by the precipitation of the prior year: in years following high *P* years all watersheds tended to "lose" or discharge extra water (relative to average annual *Q*), often leading to a negative ΔS, while after low *P* years all watersheds tended to "gain" water, mostly resulting in a positive ΔS. However, the deciduous low-elevation watersheds underwent greater changes in storage than the coniferous watersheds, resulting in a greater (relative) storage range (Figure 11 and Table 7). While the smallest storage range for the steeper high-elevation WS36 is in agreement with results from *Troch et al.* [2003] who found faster response and less storage dynamics on the steeper of two idealized hillslopes using a hillslope-storage Boussinesq model, the storage dynamics in the low-elevation watersheds cannot be explained as easily. Furthermore, regardless of aspect, both low-elevation deciduous watersheds and both coniferous watersheds exhibited very similar storage changes (Figure 13). In most years even the ratios of ΔS_{conif} to ΔS_{decid} were similar for both aspects, suggesting that at Coweeta vegetation type has a stronger effect on storage of a watershed than aspect. However, the solar insolation values for each aspect (Table 1) indicate that especially during the growing season the differences in energy input are likely negligible, resulting in nearly identical 21 year average annual *ET* values in the deciduous watersheds (18 mm difference, 2% of the mean annual *ET* for WS02), while the 21 year average *ET* difference for the coniferous watersheds was larger (205 mm difference, 15% of the mean annual *ET* of WS17). The difference in annual *ET* in the coniferous watersheds is likely the result of greater leaf area in WS17 as compared to WS01 (LAI of 7.2 [*Ford et al.*, 2010] and 5.3, respectively [*Vose and Swank*, 1990]). This suggests that at least in these aggrading deciduous watersheds the N-S aspect contrast may not have a strong influence on water balances. This is corroborated by comparable runoff ratios (Figures 4 and 5) and watershed memory (Figure 6). However, in contrast, *Hibbert* [1966] found that north and south-facing watersheds showed a very different streamflow recovery after logging, with south facing watersheds generating much less runoff than the north facing ones. However, this might simply imply different water use behavior between young and old stands [*Donovan and*

Table 6. Mean Annual Watershed Evapotranspiration, Calculated From the Water Balance

	Mean Annual *ET* (mm)
WS01	1146
WS02	964
WS17	1351
WS18	946
WS36	462

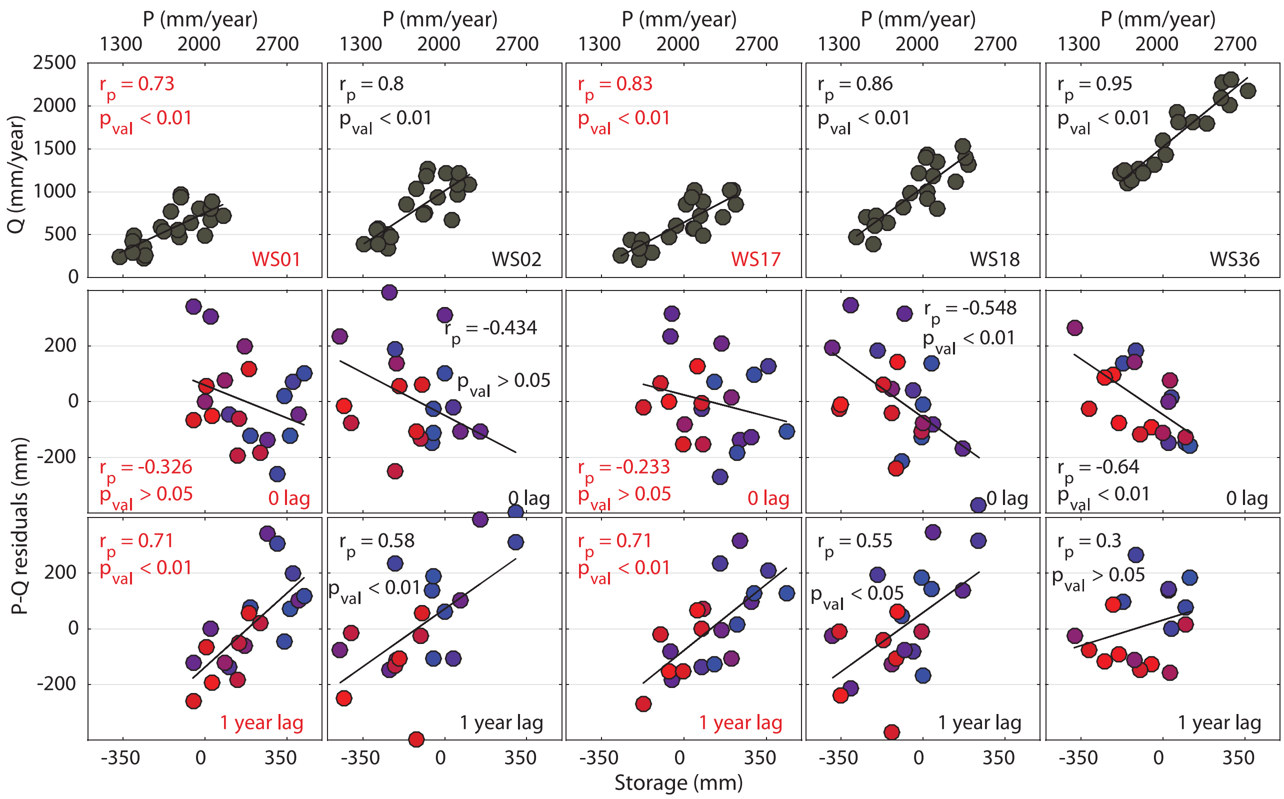

Figure 12. Linear regression between annual *P* and *Q* (top) for each of the five watersheds (each column represents one watershed). Linear regression for the residuals of the regressions between annual *P* and *Q* versus cumulative watershed storage without lag (middle) and a lag of 1 year (previous year's storage versus current year's residuals, bottom). The two bottom rows share the same *x* axis. Color-coding blue to red denotes annual *P* (red = more dry, blue = more wet).

Ehleringer, 1991], where the transpiration rates of young stands can vary much more with insolation but become more similar to each other as they mature. In addition, *Swank and Vose* [1988] reported increased leaf litter evaporation after clear cutting, which could be more susceptible to differences in insolation.

While storage values of the watersheds (Figure 11, dashed black lines) cannot be compared directly, it is possible to make inferences about the general storage state in the coniferous and deciduous watersheds based on different runoff behavior. We hypothesize that because of greater rates of evapotranspiration the

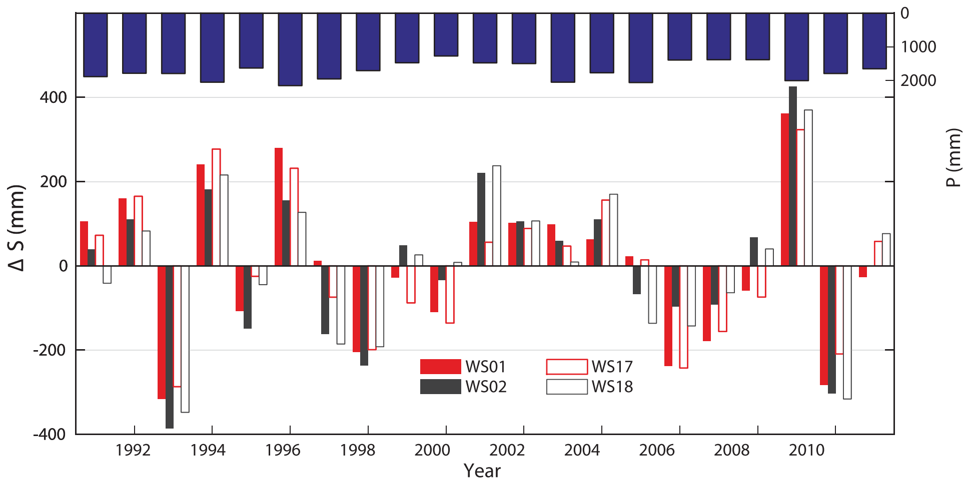

Figure 13. Comparison of annual storage changes in the low-elevation watersheds. Red denotes coniferous watersheds, black denotes deciduous watersheds. The filled bars denote the south facing watersheds. The annual precipitation of WS01 is shown and is representative of precipitation dynamics. The patterns of storage changes between the north and south-facing watersheds are almost identical.

Table 7. Relative Storage Range Over 21 Year Study Period

	Storage Range (mm)
WS01	677
WS02	811
WS17	610
WS18	717
WS36	417

actual storage state in the coniferous watersheds is likely lower than the storage state in the deciduous watersheds. This would lead to different magnitudes of runoff between the two vegetation types as storage state increases or decreases over time. The foundation of this concept is a simple storage-release function, as presented for example in *Grayson et al*. [1997]. This assumes that the relationship between storage state and water flux (i.e., runoff) can be described by a power function (Figure 14). According to the nature of the relationship between storage and flux, a small change in storage at a high storage state would lead to a greater change in runoff than the same change in storage at a low storage stage (Figure 14). This is corroborated by *Ford et al*. [2011], who found increased differences in annual runoff with increasing annual precipitation for the two paired watersheds used in this study as well as other paired watersheds at Coweeta. If the actual storage state in the deciduous watersheds is generally higher than the storage in the coniferous watersheds, mainly because of higher *ET* in the coniferous watersheds, this could have ramifications for runoff magnitudes. During the growing season, both vegetation types shift toward a lower storage state. Due to the low runoff at low storage states, slightly higher storage in the deciduous watersheds would not have a significant effect on runoff and hence both watershed vegetation types would exhibit roughly equal amounts of runoff. The data from Coweeta are consistent with this as demonstrated by small differences in monthly runoff totals during the growing season (Figures 3 and 14). In contrast, during the dormant season as both watershed types shift toward higher storage states, the runoff from the deciduous watersheds would be greater than the runoff from the coniferous watersheds. This would be due to the position of the watersheds on the storage-flux curve and its high degree of nonlinearity. Similar differences in storage between the watersheds would result in greater runoff differences at the higher, wetter end than at the lower, drier end of the storage-flux curve. Those differences could even increase as the coniferous watersheds continue to transpire during the dormant season. The data are consistent with this hypothesis as demonstrated by the monthly runoff differences (Figures 3 and 14). Continued *ET* in the coniferous watersheds during the dormant season further decreases storage and amplifies this effect. Basically, the longer duration transpiration in the coniferous watersheds leads to less storage than in the deciduous watersheds and therefore lower *Q*. This behavior could also explain the slightly weaker correlation for the deciduous watersheds between the previous year's storage state and the residuals of a regression between *Q* and *P*, as it appears that the storage effect (i.e., the lag-correlations) may be affected by overall storage state at the beginning of the water balance calculation (see Figure 7). In addition to the annual fluctuations in storage and resulting fluxes, this conceptual model is also transferable to dry and wet time periods. During the two droughts in the study period, overall storage in both vegetation types declined (i.e., the watersheds shifted toward a drier state on the storage-flux curve in Figure 14), resulting not only in less runoff overall, but also smaller differences in runoff between the two vegetation types during dormant seasons (e.g., 1999–2001 and 2006–2008). In the

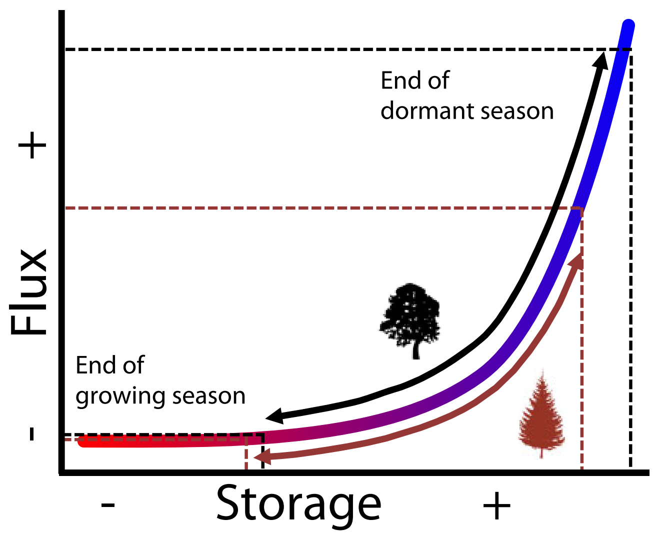

Figure 14. Conceptual model of storage states and associated water fluxes in the deciduous and coniferous watersheds. The thick line denotes the general storage-flux relationship, indicating low (red) and high (blue) storage and flux states. The arrows denote the range of storage for the deciduous watersheds (black arrow) and the coniferous watersheds (red arrow). Fluxes are similar at the end of the growing season, due to more similar storage states and minimal differences in resulting fluxes. Storage and flux differences are greatest at the end of the dormant season, when evapotranspiration continues to decrease storage in the coniferous watersheds.

wetter periods (e.g., 1994–1996), storage increased in all watersheds (i.e., the watersheds shifted toward the right on the storage-flux curve in Figure 14), thus leading to greater differences in monthly runoff.

In summary, the history of precipitation would determine a watershed's general location on a storage-flux curve, e.g., low or high storage (Figure 14). Vegetation could modify this location by water losses through transpiration. The degree of influence would then be a function of vegetation type, with deciduous watersheds being wetter than coniferous watersheds due to lesser *ET*.

4.3. Implications

The latest IPCC report suggests drought frequencies and intensities may increase in the later part of the 21st century. At the same time, we will likely see increases in the frequency and intensity of heavy precipitation events [*IPCC*, 2013; *Hartmann et al*., 2013]. These effects have already been documented at Coweeta [*Laseter et al*., 2012]. How watersheds respond to the anticipated changes in climate is largely a function of storage that acts as a buffer between water inputs and streamflow response. Variability in buffering capacity can even be observed over small spatial scales such as within the Coweeta Hydrologic Laboratory. Depending on the location of a watershed within the greater Coweeta basin, storage becomes a more important descriptor of hydrologic response, as demonstrated by the contrast between the low and high elevation watersheds (Figure 12). While we identified watershed memory in a system like Coweeta for a period of up to 1 year (Figure 6, top), watershed recovery from severe droughts may actually take much longer. In fact, our results suggest that it can take several years after a drought period to refill storages (Figure 9, dashed lines, from relative minimum to relative maximum). While we did not observe vegetation influences on watershed memory, both precipitation history and vegetation type can play a role in determining a watershed's storage state. This storage state difference in turn can partially explain the effects of different vegetation (coniferous versus deciduous) on hydrologic response. This can lead to greater differences not during dry periods, but during wet periods. The location of a watershed on the storage-flux curve (Figure 14) thus has implications for a watershed's response to individual precipitation events. As a consequence, drought periods could decrease the effect that different vegetation types have on the water balance. Decreasing precipitation could result in similar water balances for coniferous and deciduous watersheds, especially if the drought periods extended through the dormant seasons, when runoff differences are typically greatest due to differences in storage. Conversely, increases in annual precipitation could increase the differences in runoff between deciduous and coniferous watersheds.

Our observations can also inform interpretations of the effect of land cover change, e.g., clearcutting, on peak flows and storm runoff that is well documented at Coweeta and elsewhere [e.g., *Harr et al*., 1975; *Hewlett and Helvey*, 1970; *Pierce et al*., 1970; *Verry et al*., 1983]. Several studies highlight that the differences between cut and uncut watersheds were greatest in late summer and fall, when the moisture differences between the watersheds was greatest [e.g., *Harr et al*., 1975; *Pierce et al*., 1970]. Here we propose that the differential position on the storage-flux curve of cut and uncut watersheds or watersheds with different vegetation types serves as a mechanism to explain this observed phenomenon.

5. Conclusions

Utilizing 21 years of precipitation and runoff data from five watersheds at the Coweeta Hydrologic Laboratory, we found strong memory effects (i.e., influence of past precipitation on present runoff) in all watersheds across a range of temporal resolutions. Differences in runoff between the coniferous and deciduous watersheds can be explained largely by different storage dynamics that developed under deciduous and coniferous vegetation. We summarize our findings as follows:

1. The Coweeta watersheds with their generally steep slopes and deep soils exhibited considerable memory effects. Past precipitation influenced the runoff behavior of watersheds on monthly, seasonal, and annual time scales. For monthly, seasonal, and annual resolutions, the precipitation of the previous time step had an equal or greater influence on the runoff ratio than the precipitation of the same time step. Neither aspect nor vegetation type appears to have had an influence on lag-correlations between precipitation and runoff ratios at any time scale.

2. All watersheds exhibited similar temporal patterns of storage, although the range of relative storage differed in the low-elevation coniferous and deciduous watersheds, and between the deciduous low-elevation and deciduous high-elevation watershed.
3. The previous year's storage state explained much of the variability in the relationship between annual *P* and *Q*, and explanatory power was greater for the low-elevation watersheds with deep soils than for the steeper and wetter high-elevation watershed with shallower soils. Additionally, the storage effect on the residuals was greater in the coniferous than the low-elevation deciduous watersheds. We demonstrated that watershed storage state set by the previous years' precipitation can be an important component in determining annual runoff.
4. During growing seasons and during multiyear dry periods, runoff differences between deciduous and coniferous watersheds decreased. In dormant seasons and during wetter periods, runoff differences between the vegetation types increased (greater in deciduous watersheds). This suggests that the deciduous low-elevation watersheds at Coweeta had higher storage states on average than the coniferous watersheds. These differences, combined with nonlinearity of the storage-discharge relationship, can partially explain this behavior and further demonstrates the role of vegetation in the watershed water balance.
5. Depending on the geographic region, projected increases or decreases in precipitation as a result of climate change could alter the relative runoff behavior of systems with differing degrees of *ET*, e.g., coniferous and deciduous watersheds. Decreases in precipitation, e.g., increases in drought frequency and intensity, could lead to smaller differences in the observed runoff between different vegetation types; increases in precipitation could increase differences in runoff magnitudes between coniferous and deciduous watersheds.

This study provides insight into how watershed memory can affect runoff in steep headwater catchments. We showed how past precipitation affects storage dynamics under different vegetation and we postulated a conceptual model to explain differences in runoff between coniferous and deciduous watersheds. Given the results from this study, we hypothesize that climate change may affect watershed memory and the runoff response from watersheds with different vegetation or *ET* magnitudes. Future research could include distributed modeling to disentangle the spatial patterns of storage and the degree of hydrologic memory across physiographic and climatic gradients.

Appendix A: *ET* Sensitivity Analysis

A1. Generation of Variable *ET* Time Series

An important assumption of our approach to calculate a time series of storage changes is that interannual variability in *ET* at Coweeta is relatively stable. However, we acknowledge that even in environments with stable climatic and atmospheric conditions annual *ET* can vary within some limited bounds across years. Therefore, we performed a sensitivity analysis by generating annual *ET* times series to demonstrate that even with variable *ET* the storage calculations still yield meaningful correlations with the runoff ratio time series.

We used annual pan evaporation and radiation data from the weather station located near the main outlet at Coweeta, as well as the annual precipitation time series for the individual watersheds to generate time series of watershed *ET* that were allowed to vary across years. The adjustments were made using the following principle: if the annual precipitation was greater (smaller) than the mean precipitation for a watershed, the adjusted annual *ET* was greater (smaller) than ET_{mean} of said watershed. The mean of the adjusted *ET* time series, ET_{adj_mean}, had to be equal to ET_{mean} in order to conserve the long-term water balance. We describe here the procedure for adjustment with the precipitation time series for one watershed; the same method was applied to all watersheds with the pan evaporation and radiation time series as well as the randomized time series.

A2. *ET* Adjusted Based on Climatic Variables

We calculated the mean and the range of the precipitation time series. If in a given year the precipitation equaled the mean precipitation, then that year would receive the previously calculated mean *ET* for the watershed. If precipitation of the year were greater (smaller) than the mean precipitation, the annual *ET* would be greater (smaller) than mean *ET*. The maximum adjustment was fixed to 10% of mean watershed

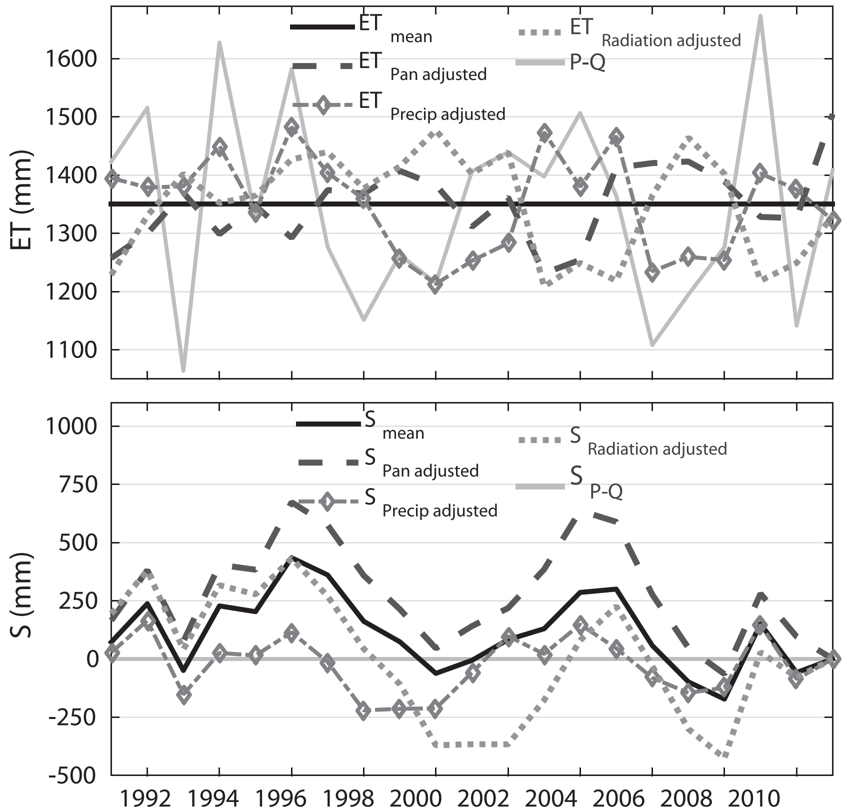

Figure A1. Time series for climate adjusted (top) *ET* and (bottom) storage for WS17.

ET, ET_{mean}, to match or exceed the ranges in PET computed by *Rao et al.* [2011] ($\sim$200 mm). Therefore, the wettest year received 110% of mean *ET* and the year with the lowest precipitation was assigned an annual *ET* of 90% of mean *ET*. Between minimum and maximum precipitation values, the *ET* adjustment factor was calculated with a simple linear regression. The adjusted *ET* for a given year and watershed, ET_{adj_i}, was calculated as follows:

$$ET_{adj_i}=\begin{cases} ET_{mean}, \quad P_i=P_{mean} \\ ET_{mean}\times(1+0.1((P_i-P_{mean})/(P_{max}-P_{mean}))), \quad P_i>P_{mean} \\ ET_{mean}\times(1-0.1((P_{mean}-P_i)/(P_{mean}-P_{min}))), \quad P_i<P_{mean} \end{cases} \tag{A1}$$

where ET_{mean} the mean watershed *ET*, P_i the precipitation for a given year and watershed, P_{mean} is the mean precipitation for a given watershed, P_{max} and P_{min} are the maximum and minimum precipitation for a given watershed.

Since the mean of the adjusted *ET* time series, ET_{adj_mean}, could be slightly different than ET_{mean}, we multiplied each adjusted annual *ET* value with the ratio of ET_{mean} to ET_{adj_mean}. In addition to *ET* time series constraint by climatic variables, we generated 1000 completely randomized annual *ET* time series within the 90%–110% boundaries around ET_{mean}.

Table A1. Lag-Correlations (r_p) Between Annual Cumulative Storage States, Calculated With Different *ET* Approximation Methods, and Runoff Ratios[a]

	ET_{mean}	ET_{pan}	$ET_{radiation}$	$ET_{precipitation}$	$ET_{randomized}$
Lag 0	0.02	0.1	0.3	0.19	0
Lag 1	0.69*	0.68*	0.72*	0.64*	0.59**

[a]Values were averaged over all five watersheds. $ET_{randomized}$ was averaged over 1000 realizations (95% confidence intervals $0.40 \le r_{p\_ET_{randomized}} \le 0.81$). *denotes statistical significance at the 0.01 level, **denotes statistical significance at the 0.05 level.

A3. Storage Calculation and Lag-Correlations

Subsequently we generated time series of cumulative storage changes from the newly derived annual *ET* time series following equations (2) and (3):

$$\Delta S_{adj_i} = P_i - Q_i - ET_{adj_i} \tag{A2}$$

$$S_{adj\_i} = \sum_{i=1}^{n} \Delta S_{adj\_i}. \tag{A3}$$

We then compared the lag-correlations between the adjusted cumulative storage time series and runoff ratios to the lag-correlations for the original cumulative storage time series calculated with ET_{mean}.

The range of *ET* across all methods was 230 mm for Ws01, 193 mm for WS02, 270 mm for WS17, 190 mm for WS18, and 93 mm for WS36. The range for WS17 is thus much larger than the range in potential *ET* ($\sim$200 mm) computed by *Rao et al*. [2011]. While the adjusted *ET* time series were different from each other (Figure A1, top, example shown for WS17), the resulting cumulative storage times series behaved similarly (Figure A1, bottom).

The lag-correlations derived from all *ET* time series exhibited nonsignificant and weak correlations at lag 0, but strong, significant correlations at lag 1 (storage state from previous year explains this year's runoff ratio, see Table A1). This is even the case for the randomly generated *ET* time series.

While the actual storage values would differ slightly in case of variable *ET*, the sensitivity analysis suggests that even if *ET* were highly variable at Coweeta, the general evolution/pattern of a storage time series would likely not be affected, as the differences in annual $P - Q$ are usually much greater than a potential variability in *ET*.

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