Nonstationarity in threshold response of stormflow in southern Appalachian headwater catchments

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Abstract Threshold behavior of stormflow response is an emergent pattern observed in several studies demonstrating subsurface storage controls on catchment rainfall-runoff dynamics. These studies demonstrate a distinct transition from negligible stormflow discharge response to rapid, linearly increasing stormflow identified by a single, uniquely defined threshold as a basic catchment attribute that relates to geophysical properties. Utilizing precipitation, streamflow, and soil moisture data spanning 15 years from three catchments at the Coweeta Hydrologic Laboratory (CHL), we analyze how threshold behavior forms and varies at several timescales. We pose three hypotheses: (1) stormflow thresholds form at CHL as a function of antecedent soil moisture and gross precipitation, (2) thresholds vary seasonally and interannually, and (3) threshold variation through time implies greater long-term complexity of runoff controls beyond catchment geophysical properties, including forest canopy ecohydrologic feedbacks. We isolate threshold behavior of stormflow using piecewise regression analysis in short to long-term data sets with respect to antecedent soil moisture index and gross precipitation. We use this to investigate threshold variation over seasonal, interannual, and decadal timescales that encompass hydroclimatic extremes. Seasonal analysis reveals that thresholds are more variable between growing seasons than between dormant seasons. In growing seasons with greater water stress, stormflow thresholds are lower after controlling for soil moisture storage suggesting more complex, long-term rainfall-runoff relationships as a result of forest canopy response to water stress. We present a conceptual model of how vegetation-climate interactions influence long-term rainfall-runoff relationships creating interannual variability of stormflow thresholds and linear stormflow response.

1. Introduction

Runoff generation at the hillslope scale involves complex processes that are difficult to measure empirically and include an integration of flow travel times spanning several orders of magnitude [Sivapalan, 2003b; Tetzlaff et al., 2008]. The interactions of soil moisture [Western et al., 2002], preferential flow [Mosley, 1982; Band et al., 2014], and the connectivity of saturated regions [Dunne and Black, 1970; Hewlett and Nutter, 1970] scale nonlinearly to catchment runoff generation [Dunne et al., 1975]. A number of complex, process-oriented simulation models have been developed over the past few decades to attempt to integrate surface, subsurface, and ecosystem processes to predict catchment runoff behavior under different conditions, constituting a “bottom-up” approach [Band et al., 1993; Vertessy et al., 1993; Wigmosta et al., 1994; Tague and Band, 2004; Ivanov et al., 2008a, 2008b; Fatichi et al., 2012a, 2012b].

Alternatively, using a top-down approach, complex runoff generation processes can be simplified into emergent behaviors [Sivapalan, 2003a]. Generalizing hydrology across spatiotemporal scales has become increasingly important for understanding the effects of land use [Walsh et al., 2005; Grimm et al., 2008], ecosystem dynamics [Vertessy et al., 1996; Watson et al., 1999], and global climate change [Vörösmarty et al., 2000] on local and regional water resources. However, efforts to broadly characterize hydrologic patterns in space and time have been impeded by a legacy focused on empirical research at individual sites without sufficient general theory [McDonnell et al., 2007]. Much discussion in the hydrologic sciences has recognized this knowledge gap and stresses new paradigms that challenge the field to pursue more generalizable hydrologic behavior rather than isolated assessments [Sivapalan, 2003a, 2003b; McDonnell et al., 2007; Ali et al., 2012]. This requires a top-down approach where macroscale properties are defined and interpreted to smaller hydrologic units [Dooge, 1986]. The expectation is that the properties that emerge with increasing
scale integrate hydrologic complexities and heterogeneities into well-defined spatiotemporal patterns. Common forms of these emergent patterns include (1) threshold flow generation based on storage-flux relationships [Tromp-van Meerveld and McDonnell, 2006a; Detty and McGuire, 2010; Graham and McDonnell, 2010; Ali et al., 2012]; (2) stormflow travel time distributions [McGuire et al., 2005]; and (3) catchment-scale hysteresis of soil moisture and streamflow [Rosenbaum et al., 2012].

Threshold behavior in stormflow response, as an emergent pattern, has the potential to discriminate between slow and fast runoff generation processes linking subhillslope processes and connectivity to catchment discharge [Zehe et al., 2005; Lehmann et al., 2007; Mirus and Loague, 2013]. The intent of threshold behavior models is to simplify spatially and temporally complex hydrologic behavior by relating hydrologic flux patterns to observable state variables. An example of this relationship that has been well-studied is thresholds of antecedent soil moisture that create binary stormflow response patterns [Detty and McGuire, 2010; Graham and McDonnell, 2010; Oswald et al., 2011]. Drier antecedent soil moisture below some threshold will produce little or no stormflow, whereas wetter antecedent soil moisture will produce stormflow that is linearly correlated to rainfall depth. In these models, correlation between rainfall depth and stormflow is negligible below the threshold and strongly linear above it. A major question is whether linear threshold-based storage-flux response models can adequately reproduce and summarize behavior of the number of complex, interacting processes active at the hillslope and catchment level in the context of increasing hydroclimate variability. To answer this question requires testing stormflow thresholds over long measurement periods.

1.1. Stormflow Threshold Studies

A number of recent studies utilize seasonal to several year data sets to characterize emergent patterns of stormflow response with respect to measured rainfall and subsurface storage in forested catchments. Tromp-van Meerveld and McDonnell [2006a] excavated a 20 m long trench in the Panola Mountain Research Watershed (PMRW) to measure subsurface pipe flow and matrix flow. They found total pipe flows greater than 1 mm at the trench face when total storm event rainfall exceeded 55 mm and when soil moisture prior to storm events was high. Uchida et al. [2005] analyzed four watersheds, including PMRW comprising up to 2.5 years of continuous data ranging from 16 to 147 storm events [Tromp-van Meerveld and McDonnell, 2006b, 2006a] and found similar thresholds. Detty and McGuire [2010] used soil moisture records integrated with depth and directly combined antecedent soil moisture with total storm precipitation to evaluate threshold response for the Watershed 3 catchment at the Hubbard Brook Experimental Forest during a single wet-up period from August to January. Their study included 14 storms and showed total quickflow generation followed a strongly linear correlation when the sum of storm event precipitation and antecedent soil moisture exceeded some threshold. Events above this threshold had characteristically greater runoff ratios and more responsive groundwater with no observed saturation overland flow, indicative of high amounts of subsurface stormflow. The longest analysis was at HJ Andrews Experimental Forest [Graham et al., 2010] and showed that stormflow response over 50 years was related to interstorm period. Interstorm period was an indicator of antecedent storage prior to rainfall because measured soil moisture was unavailable. Shorter interstorm periods were correlated with higher preevent wetness, which resulted in greater stormflow response. In events with interstorm periods that exceeded 10 days, stormflow response to rainfall was delayed requiring almost 80 mm of precipitation before significant stormflow was observed. Penna et al. [2011] showed similar controls of antecedent soil moisture on stormflow generation and runoff ratios. Their study in Rio Vauz Basin of the Italian Alps consisted of 26 plots sampled synoptically over 2 years during June and July with portable soil moisture probes, in addition to four continuous soil moisture sites that measured from June to October.

Threshold response has also been studied in a number of northern catchments, where threshold was visually assessed for nine sites across Europe and North America [Ali et al., 2015]. In this study, antecedent saturation deficit was estimated by computing the sum of all water inputs between the start of an event to the initial stormflow response. They showed stronger threshold response in catchments that are rainfall-dominated as opposed to snowmelt-dominated. Oswald et al. [2011] conducted a similar study in Ontario, Canada, where they examined stormflow thresholds as a function of storage capacity using piecewise regression analysis from 164 events between 2001 and 2009. Their study was constrained to purely rain driven events, excluding snowmelt and rain-on-snow, limiting their analysis to May through October. Findings from this study stress the importance of catchment properties for
controlling both threshold formation and linear stormflow response and also the role of antecedent storage on runoff ratio.

1.2. Processes Contributing to Thresholds

Mechanisms of threshold stormflow response in forested catchments have generally been attributed to the activation of subsurface macropores and pipe flow, but hillslope connectivity to near-stream riparian areas may be essential in forming more gradual stormflow responses [Freer et al., 2002; Lehmann et al., 2007]. Spence and Woo [2003] proposed the fill-and-spill hypothesis that describes the importance of bedrock topography and lateral flow at the soil-bedrock interface as a major contributor to nonlinear response of runoff. As storm events progress, depressions in the bedrock fill from subsurface stormflow and as they breach their downslope ridge, water spills into downslope depressions or into the stream. Tromp-van Meerveld et al. [2006b] attributed threshold response of stormflow at hillslope scales to fill-and-spill mechanisms combined with activation of subsurface macropore flow. Uchida et al. [2005] showed significant pipe flow occurred when thresholds of antecedent wetness and gross precipitation were surpassed. The relationship between pipe flow and hillslope discharge following the threshold became strongly nonlinear varying with rainfall intensity. Pipe geometry and connectivity with upslope regions prior to and during storms was difficult to characterize, but likely important to nonlinear runoff response. Connectivity of upslope areas to streams can create thresholds due to activated bedrock fracture flow. At PMRW, Tromp-van Meerveld et al. [2007] found that fracture flow contributed up to 21% of all streamflow over a 2 years wet period and that flow through the bedrock at event timescales continued to feed streams days after the storm event ended. They suggest that antecedent moisture and rainfall size and intensity influence the relative contribution of pipe flow and fracture flow contributing to threshold stormflow response [Tromp-van Meerveld and McDonnell, 2006a].

In simulation studies, vertical heterogeneity of soil conductivity profiles dominates vertical and lateral subsurface stormflow [Mirus, 2015]. Decreasing hydraulic conductivities with depth can create transmissivity feedbacks [Kendall et al., 1999; Bishop et al., 2004], particularly when catchments have high antecedent wetness, resulting in threshold formation [Detty and McGuire, 2010].

Vegetation also plays a critical role in hillslope hydrology that manifests as emergent patterns. Phenology, fine root growth, succession, rates of evapotranspiration, rooting depth, and landscape pattern influence runoff generation processes with increasing spatiotemporal complexity [Western et al., 1999, 2002]. Local hydrologic controls exerted by vegetation and edaphic properties [Lin et al., 2006] describe spatial patterns of soil moisture, but are shown to shift toward nonlocal controls (e.g., topography) with senescence and higher seasonal rainfall [Grayson et al., 1997]. During the dormant season, less evapotranspiration relative to rainfall leads to higher soil moisture and enhanced lateral drainage resulting in increased runoff sensitivity. Lateral drainage during the growing season is inhibited by plant water uptake altering hillslope connectivity and subsurface flows. These observations highlight competition among runoff and transpiration over limited root-zone soil moisture, particularly in deep forested soils where streamflow generation is predominately shallow subsurface flow [Hewlett and Hibbert, 1967].

Significant stormflow generation relies on higher levels of hillslope connectivity under wetter soil moisture in the shallow subsurface [Hewlett and Hibbert, 1962], but transpiration actively removes soil water in this zone. This raises questions regarding ecosystem water use controls on stormflow at both event and seasonal scales. Peak growing season evapotranspiration rates estimated from sap flow data [Ford et al., 2011a] can approach daily stormflow totals in humid, temperate ecosystems. The effects of transpiration on stormflow recession limbs have been observed using recession analysis [Shaw and Riha, 2012], but not explored in terms of storage-flux threshold relationships.

Recent research has shown that increasing hydroclimate extremes and drought severity exacerbate prolonged reductions in tree transpiration rates, and their response to drought varies by tree functional types resulting in delayed rates of transpiration and photosynthetic recovery [Zwieniecki and Holbrook, 2009]. During recovery, transpiration rates of trees are reduced and may persist from several months to years [Hacke et al., 2001]. The impacts of this at the storm event scale are not well studied, but raise questions pertaining to the onset of stormflow thresholds and the amount of total stormflow generated.
1.3. Purpose
Studies investigating emergent patterns in runoff threshold response are often limited by measurement period. Those with long-term data sets do not directly address issues of seasonality and nonstationarity or lack methods for statistically computing thresholds of stormflow generation raising questions concerning long-term stability of observed emergent patterns. Methods for identifying thresholds are not well-established especially in cases where emergent patterns are not clearly discernible (e.g., in long-term data sets). This study investigates the stationarity of rainfall-storage-runoff relationships in small forest catchments, and the impacts of hydroclimate and ecohydrologic feedbacks of the forest canopy on short to longer-term system dynamics. We use well-established soil moisture, rainfall, and discharge measurements collected from the Coweeta Hydrologic Laboratory (CHL) in the southwest Appalachian Mountains of North Carolina, which span 15 years of continuous measurement allowing us to consider short and long-term rainfall-storage-runoff behavior. We pose a set of three hypotheses that serve our central objective of characterizing long-term threshold behavior:

1. At short timescales, on the order of a single season to a year, watersheds at CHL exhibit well-defined runoff thresholds and linear stormflow response as a function of the sum of integrated antecedent soil moisture and total storm precipitation.
2. Longer measurement periods (>10 years) introduce variability as a result of nonstationarity in these hydroecological systems, which produce seasonal and interannual variability in thresholds.
3. Due to the hydroclimatic nonstationarity and ecohydrological feedbacks, thresholds computed interannually over long measurement periods confound linear threshold methods that assume a single threshold and linear stormflow response, determined by catchment geophysical properties.

2. Data and Methods
2.1. Study Site
The Coweeta Hydrologic Laboratory (CHL) located in North Carolina, USA is a US Forest Service (USFS) site, jointly funded by the National Science Foundation Long-Term Ecological Research (LTER) program. It is located in the southern Appalachians and was originally established by the USFS in the 1930s. CHL is an east facing, bowl-shaped basin with a drainage area of 1626 ha and elevation range of 1000 m (Figure 1a)

Figure 1. A hillshade map of (a) the Coweeta Basin at the Coweeta Hydrologic Laboratory in SW North Carolina, USA, with rain gages, (b) WS14 with 3 near stream soil moisture plots, and both LTER catchments, (c) WS18 and (d) W27.
[Swank and Crossley, 1988]. It is further subdivided into subwatersheds that represent a variety of treatments and environmental gradients. The soils are sandy loam Inceptisols and Ultisols [Velbel, 1988] underlain by folded schist and gneiss formations [Hatcher, 1988; Hales et al., 2009]. Soil catenary processes lead to coarser loams on ridges and side slopes with finer colluvium of sandy loams in hollows and downslope areas. The Nantahala Escarpment runs along the divide of high-elevation watersheds [Wooten et al., 2008] and watersheds that border the Nantahala Escarpment transition from colluvial low slope bottomlands to steep rocky slopes with outcrops [Band et al., 2012].

The climate is classified as marine, humid temperate with an average annual rainfall at the base weather station (CS01) around 1700 mm and a strong orographic effect that leads to annual rainfall amounts exceeding 2500 mm at the highest elevations [Swift et al., 1988]. Rainfall is evenly distributed throughout the year with higher-intensity convective storms during the summer and frontal precipitation in the winter [Laseter et al., 2012]. Snow is uncommon even at high elevations. The area is prone to tropical storms that can trigger landslides especially around the Nantahala Escarpment [Wooten et al., 2008; Hales et al., 2009; Band et al., 2012]. Despite high annual rainfall totals, summer potential evapotranspiration rates can lead to water stress in low-elevation catchments, particularly along ridges and midslopes [Hwang et al., 2012], and as a result of longer growing seasons [Hwang et al., 2014]. Increasing long-term hydroclimate variability has increased extreme wet and dry periods that impact timing of vegetation senescence [Ford et al., 2011b; Hwang et al., 2014].

Forest community composition varies across elevation and moisture gradients from drier oak-pine-dominated stands located along ridges at lower elevation to northern hardwood dominance in wetter, high-elevation regions [Day and Monk, 1974]. At low elevations, forest types are predominantly Quercus spp. (oaks) codominated by Carya spp. (hickories) with wetter, downslope coves and riparian corridors primarily comprised of Liriodendron tulipifera (tulip poplar) and Tsuga Canadensis (eastern hemlock [Elliott et al., 1999; Elliott and Vose, 2011]); however, eastern hemlocks have declined due to hemlock woolly adelgid. Ample moisture in high-elevation catchments supports a variety of hardwood trees including Betula lutea (yellow birch), Tilia heterophylla (basswood), and tulip poplar with some Quercus rubra (northern red oak). Much of the understory throughout the basin is comprised of two evergreen shrubs: Kalmia latifolia (mountain laurel) and Rhododendron maximum [Day et al., 1988]. Changing stand composition along environmental gradients [Elliott et al., 1999] and varying growing season lengths with elevation create complex phenologic patterns across the basin [Whittaker, 1956; Day and Monk, 1974; Hwang et al., 2011b].

Our study focuses on three headwater catchments with variable basin morphology and environmental conditions (Figure 1). Watersheds 18 (WS18) and 27 (WS27) are low and high elevation catchments, respectively, associated with the LTER program comprising over 15 years of soil moisture data and watershed 14 (WS14) is associated with a recent hillslope-scale study comprising a shorter data set. WS14 is a 62.4 ha northwest facing watershed with an undisturbed mixed hardwood forest. Relief in WS14 is 285 m implying an orographic effect on precipitation of 10% or less from the outlet to the ridge. The watershed is divided into two main drainage stems of first and second stream orders. A thrust fault intersects WS14 midway downslope [Hatcher, 1988] creating rocky, steep channels downslope that transition to gentler gradients and banks upstream. WS18 is 12.3 ha with a similar mixed hardwood-oak stand that has been undisturbed since 1927. The basin is elongated with steeper slopes, but a relief comparable to WS14. WS27 is a 39.8 ha, high-elevation catchment with an elevation range of 393 m [Swank and Crossley, 1988] and relatively higher rainfall and runoff ratios [Swift et al., 1988]. Due to cooler temperatures, the growing season in WS27 is shortened by up to 3 weeks [Hwang et al., 2011a]. WS27 backs up to the Nantahala Escarpment giving it steeper slopes higher in the catchment that decrease toward the outlet [Wooten et al., 2008; Hales et al., 2009].

2.2. Long-Term Climate and Hydrology

Daily precipitation is collected by four gaging stations and a climate station that span the elevation gradient of CHL. The base climate station (CS01; Figure 1a), located on the valley floor, has been actively measuring daily precipitation and temperature since 1937, relative humidity since 1983, and radiation since 1960 [Miniat et al., 2015]. Long-term data from CS01 are supplemented in this study by four additional rain gages to account for spatial heterogeneity and orographic effects in rainfall [Miniat et al., 2017]. The two lower watersheds (WS14/WS18) are situated between rain gage 41 (RG41) and RG96 whose measurements are
averaged across the two catchments. WS27 has a larger elevation range, so rain gages above (RG31) and below (RG55) the catchment are averaged to account for precipitation gradients. Rainfall totals are aggregated to daily timesteps to match the temporal resolution of our long-term data. Daily discharge from each catchment is measured using v-notched weirs with continuous discharge data dating back to the mid-1930s [Miniat et al., 2016]. This study subsets long-term, daily discharge and climate measurements to coincide with in situ soil moisture measurements that began in 1999.

Our time domain for analysis runs from 1999 to 2014 and includes wet and dry precipitation extremes and several hurricanes. A prolonged drought occurred at the onset of the study lasting from 1999 to 2001. It peaked in 2000 and 2001 when rainfall was below the long-term annual average measured at CS01 of 1796 mm by 31 and 22%, respectively. A wetter period between 2002 and 2005 followed where average precipitation was 1946 mm, the wettest of which was 2003 (+17%). Another severe drought occurred starting in 2006, which included the driest year on record in 2007 where rainfall was 32% below average. The drought ended in 2009 with subsequent years switching between wet and dry. This included the two wettest years on record, 2009 and 2013, when rainfall was 2374 mm (+31.8%) and 2368 mm (+32.1%), respectively. Several hurricanes also occurred, including Hurricane Frances and Ivan in 2004, which caused landslides throughout the region [Wooten et al., 2008; Band et al., 2012; Hwang et al., 2015]. Overall, rainstorms at CHL are spaced 4 days apart on average but spacing between events during droughts was as long as 39 days.

2.3. Field Measurements

In catchments WS18 and WS27, five soil moisture plots are established along a terrestrial ecological gradient as part of the LTER program (Figures 1c and 1d). WS18 has three plots located along a moisture gradient including a xeric pine-oak stand (plot #118), an intermediate mixed oak stand (plot #318), and a mesic cove hardwood stand (plot #218). WS27 has one plot in a mesic mixed oak stand (plot #427) and a second plot in a high-elevation wet northern hardwood stand (plot #527). Water content at each plot is measured every 15 min by four Campbell CS616 time domain reflectometer (TDR; Campbell Scientific Inc., Logan, UT, U.S.) probes totaling 20 probes throughout both LTER catchments and has been collected continuously since 1999. In this study, the 15 min interval measurements are averaged to daily times steps. Probes are inserted vertically at two depths (0–30 cm and 30–60 cm) and are located at two topographic positions (upper and lower) within each plot. A more detailed description of the terrestrial gradient plots can be found at the Coweeta LTER website (https://coweeta.uga.edu/dbpublic/dataset_details.asp?accession=1013).

WS14 is part of a separate hillslope-scale study concerned with ecohydrologic patterns, with soil moisture plots confined to a single hillslope, so sites in WS14 are not meant to be statistically representative of the entire catchment. Convergence zones along topographic flowlines spanning a ridge to stream gradient are preferentially selected as sites for soil moisture plots because this region is likely where expanding variable source areas will first occur. Water content measurements are a combination of synoptic samples taken at biweekly to monthly intervals and continuous TDR measured between July 2011 and January 2013. Three near-stream plots (plot #3A, #3B, #3C; Figure 1c) measure hourly soil moisture with buried Campbell CS616 TDR probes. TDR probes are arranged similar to WS18 and WS27 with four probes at each plot buried from 0–30 and 30–60 cm in an upslope and downslope topographic position.

2.4. Storm and Threshold Definition

Studies typically assess thresholds visually, which may be sufficient for short-term studies with well-defined thresholds [Detty and McGuire, 2010; Penna et al., 2011; Ali et al., 2015], but as hydroclimate variability increases with longer timescales this method may become subjective and inefficient especially for interannual analyses. To address this, we quantitatively test for thresholds at CHL utilizing piecewise regression analysis (PRA) to compute threshold values and slope parameters derived from each linear segment of the PRA similar to Oswald et al. [2011]. We acknowledge that stormflow generation often forms nonlinear, complex relationships, but searching for thresholds and linear slope parameters using PRA conforms to top-down approaches that seek emergent hydrologic behavior. We use PRA for consistency and comparison with previous findings. Components of the PRA are also useful for understanding broad controls of shifts between slow and fast stormflow generation. For example, previous studies that assume zero slope below the threshold may miss potential slower linear stormflow generation denoted by low, positive slopes.
Following other event-based studies, storms are defined as beginning with rainfall greater than 1 mm/d and ending when stormflow returns to zero [Hewlett and Hibbert, 1967; Detty and McGuire, 2010]. This requires a hydrograph separation, which we estimate using the USGS HYSEP algorithm, adapted in MATLAB 2014b (The Mathworks, Inc., Natwick, MA). HYSEP utilizes a local minimum method [Pettyjohn and Henning, 1979] for calculating baseflow and stormflow from total discharge. We define storm events following Detty and McGuire [2010] and adapt it for daily time steps to account for our extensive long-term discharge and soil moisture data that has daily temporal resolution. Table 1 summarizes catchment properties and storm event statistics for each catchment.

Catchment wetness prior to storms is estimated from TDR measurements in each watershed. We use average daily soil moisture measured the day prior to a storm event and integrate it over the depth of the TDR sensors [Haga et al., 2005] to estimate an antecedent soil moisture index (ASI). This is combined with gross precipitation, defined as the total event rainfall falling on the canopy [Helvey and Patric, 1965], to estimate an event total wetness from which significant threshold relations are shown to arise [Detty and McGuire, 2010]. Time series data are partitioned into dormant and growing season with respect to topographically mediated phenology [Hwang et al., 2011a] that produce growing seasons nearly 3 weeks shorter in WS27 than in WS18 and WS14.

### Table 1. Overview and Comparison of Measurement Period, Catchment Properties and Storm Analysis for Each Watershed

<table>
<thead>
<tr>
<th>Measurement Period</th>
<th>Drainage Area (ha)</th>
<th>Elevation (m)</th>
<th>Runoff Ratio</th>
<th># of Events</th>
<th>Avg. Stormflow Duration (Days)</th>
</tr>
</thead>
<tbody>
<tr>
<td>WS14 Jun 2011 to Jan 2013</td>
<td>62.4</td>
<td>878</td>
<td>0.40</td>
<td>67</td>
<td>3.7</td>
</tr>
<tr>
<td>WS18 Feb 1999 to Oct 2014</td>
<td>12.3</td>
<td>823</td>
<td>0.49</td>
<td>811</td>
<td>3.4</td>
</tr>
<tr>
<td>WS27 Feb 1999 to Oct 2014</td>
<td>39.8</td>
<td>1256</td>
<td>0.66</td>
<td>791</td>
<td>4.1</td>
</tr>
</tbody>
</table>

3. Results

3.1. Threshold Stormflow Response

Using daily precipitation and stormflow, we identify events lasting from 1 to 14 days (Figure 2). The range of total stormflow varies across the three catchments. Maximum total stormflow in WS14 is 26.8 mm, which is notably smaller than maximum total stormflows in long-term records for WS18 and WS27 that are 69.2 and 222.7 mm, respectively. WS14 has fewer large events compared to WS18 and WS27 due to shorter measurement period (Figures 3a, 3d, and 3g). In WS18, the amount of annual precipitation is comparable to WS14 but the long-term data captures more extremes in stormflow generation. Also, WS18 has a slightly higher runoff ratio than WS14 (Table 1). Throughout all three catchments, total stormflow is significantly correlated with gross precipitation (Figures 3a, 3d, and 3g). ASI has no initial linear relationship with stormflow for both the short-term analysis in WS14 and long-term analysis in WS18 and WS27 (Figures 3b, 3e, and 3h).

![Figure 2. Distribution of stormflow duration in days for (a) WS18 and (b) WS27 spanning February 1999 to October 2014.](image-url)
Thresholds become more visually evident when ASI and gross precipitation are combined, which is consistent across all three watersheds (Figures 3c, 3f, and 3i). However, 15 years of stormflow response in WS18 and WS27 reveals less distinct threshold behavior compared to our 19 months record from WS14 denoted by lower r² values summarized in Table 2. PRA also identified the largest threshold value in WS14 (214.1 mm), followed by WS18 (127.0 mm) and WS27 (117.0 mm). Computing PRA for the full data set, irrespective of season, produced similar slope parameters both below (m₁) and above (m₂) the threshold. Due to differences in ASI measurements, we cannot make direct comparisons of absolute threshold values between catchments.

3.2. Temporal Dependence of Threshold Behavior

Seasonally fitting total stormflow as a function of gross precipitation and ASI using PRA strengthens threshold behavior as shown by relative increases in r² in all but two instances (Table 2). Despite relative decreases in correlation in dormant seasons of WS14 and WS27, values of r² are still high. In WS14, the range of total stormflows are generally larger in the growing season but storms producing significant runoff (>5 mm) are more numerous in the dormant season (Figure 4). This contrasts patterns of stormflow totals in the long-term data where ranges are comparable between seasons. Differences between lower and upper slope parameters (m₁ and m₂, respectively) computed from the PRA are significantly different during the growing season for all watersheds (Table 2). Absolute m₂ values are also consistent across all catchments during this season.
period. Seasonal thresholds in WS14 shift from 192.5 mm in the dormant season to 205.8 mm in the growing season but this offset is not statistically significant. Seasonal threshold values reported in Table 2 for WS18 and WS27 differ from those computed for individual years (Figure 5). Thresholds computed for individual years are more variable and smaller, but their mean values are roughly 118 and 103 mm for dormant and growing season, respectively, which is consistent across both long-term catchments.

The relative seasonal patterns of thresholds also differ between the long-term (WS18/WS27) and short-term data (WS14). This observation may be a result of differences in measurement periods or the spatial distribution of soil moisture between the LTER catchments and WS14. To test if differences are a result of measurement period, we subset the long-term data to coincide with data collected from WS14 (supporting information Figure S1). Using identically short-time periods, patterns of threshold behavior are relatively consistent across all three catchments. Larger storms are observed during the growing season leading to larger \( m_2 \) values, which may provide evidence for nonlinear behavior over piecewise linear behavior.

### 3.3. Long-Term Hydroclimate and Threshold Pattern Variation

Analyzing yearly thresholds through time for WS18 and WS27 reveal divergent seasonal patterns (Figure 6). In the dormant season, interannual variation is small, without clear trends so that WS18 and WS27 thresholds are not synchronized. Synchronization between WS18 and WS27 is greater during the growing season.

#### Table 2. Summary From Threshold Analyses in WS14, WS18, and WS27 With Respect to Season

<table>
<thead>
<tr>
<th></th>
<th>( r^2 )</th>
<th>Threshold (mm)</th>
<th>SD</th>
<th>( m_1 )</th>
<th>( m_2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>WS14</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dormant</td>
<td>0.84</td>
<td>192.5</td>
<td>11.1</td>
<td>0.02</td>
<td>0.08</td>
</tr>
<tr>
<td>Growing</td>
<td>0.98</td>
<td>205.8</td>
<td>6.7</td>
<td>0.03</td>
<td>0.23</td>
</tr>
<tr>
<td>All</td>
<td>0.96</td>
<td>214.1</td>
<td>3.5</td>
<td>0.03</td>
<td>0.23</td>
</tr>
<tr>
<td>WS18</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dormant</td>
<td>0.73</td>
<td>137.6</td>
<td>4.5</td>
<td>0.03</td>
<td>0.1</td>
</tr>
<tr>
<td>Growing</td>
<td>0.83</td>
<td>134.1</td>
<td>10.4</td>
<td>0.03</td>
<td>0.25</td>
</tr>
<tr>
<td>All</td>
<td>0.58</td>
<td>117.0</td>
<td>9.1</td>
<td>0.03</td>
<td>0.23</td>
</tr>
<tr>
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*SD, threshold standard deviation; \( m_1 \), slope parameter below the threshold; \( m_2 \), slope parameter above the threshold.

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**Figure 4.** (a) Dormant and (b) growing season response of total stormflow to gross P and ASI in WS14 for the period July 2011 to January 2013. Threshold values are represented by black vertical lines and storms on either side of the threshold are fit separately. Their respective \( r^2 \) values and slopes below (\( m_1 \)) and above (\( m_2 \)) the threshold are shown. (*\( p < 0.05 \), **\( p < 0.01 \)).
Figure 5. Long-term growing and dormant season response of total stormflow as a function of the sum of gross P and ASI in (a, b) WS18 and (c, d) WS27 between February 1999 and October 2014. Thresholds are averaged for each season over all years (black line) and for individual years (dashed lines). Thresholds computed for individual years are colored according to their normalized rainfall totals (dimensionless).

Figure 6. (a) Dormant and (b) growing season thresholds in WS18 (circles) and WS27 (triangles) plotted as a timeseries. Blue lines are the seasonal rainfall totals shown as percent change from seasonal means.
particularly from 2000 to 2005 when thresholds trend upward and drop drastically in 2006. Beyond 2006, they remain low until they increase again in 2010 and 2011 for WS27 and WS18, respectively. In Figure 6, seasonal precipitation anomalies show similar desynchronization with thresholds during the dormant season that becomes synchronized during the growing season. Interannual threshold behavior plotted as a function of seasonal rainfall totals show significant positive correlations that persists during the growing season (Figure 7) and are strongest in WS18. Meanwhile, during the dormant season significant correlations between threshold and seasonal rainfall totals exist in WS27, but not in WS18. Threshold values are also plotted against the prior seasons’ rainfall total to show interannual memory effects of moisture storage on rainfall-runoff relationships (not shown), but linear regressions did not show any significant trends in WS18 or WS27 during either season.

Figure 8 shows variation of m2 parameters computed from the annual PRAs with thresholds and seasonal rainfall. In general, m2 increases with larger thresholds and values vary from 0.06 to 1.57 in WS27 and 0.02 to 0.87 in WS18. Positive trends between m2 and thresholds are consistent throughout dormant and growing season and across both watersheds. Total seasonal rainfall shows little pattern with m2 during the dormant season, but in the growing season m2 tends to be steeper when total seasonal rainfall is greater.

4. Discussion

4.1. Stormflow Runoff Thresholds

Our long-term analysis of precipitation, soil moisture, and discharge data demonstrate threshold behavior and linear stormflow response as a function of ASI and gross precipitation at CHL, which can vary interannually. This supports other short-term studies by Detty and McGuire [2010] that show linearly increasing stormflow response with greater ASI and gross precipitation after a well-defined threshold. Previous studies observing threshold effects tend to focus on limited time ranges [Detty and McGuire, 2010; Oswald et al., 2011; Penna et al., 2011] or do not directly include measured soil moisture [Graham and McDonnell, 2010; Ali et al., 2015] complicating seasonal and interannual characterization of storage-flux relations.

Our findings show threshold variation with seasonal and interannual rainfall totals suggesting that thresholds shift with long-term hydroclimate variation. Interannual shifts in WS18 and WS27 thresholds occur in both the dormant and growing seasons, but are strongly correlated with seasonal rainfall totals only during the growing season (Figure 7). Low growing season thresholds correspond with dry summers and high thresholds with wet summers. This observation initially appears contradictory considering changes in subsurface storage, but we note that ASI in our analysis directly controls for differences in saturation deficit between events. Our analysis suggests ecohydrological feedbacks involving forest canopy evapotranspiration rates in response to water stress contribute to the seasonal to interannual system “memory.”
High rates of evapotranspiration during the growing season account for over half of the annual rainfall at CHL leading to a partitioning between localized water use and nonlocal redistribution of water down hydrologic flow paths to the stream [Brooks et al., 2010; Hwang et al., 2012]. Direct soil evaporation is negligible due to the lack of bare soil but evaporative losses due to canopy and litter interception may reinforce threshold behavior to a lesser degree. At CHL, roughly 18% of annual gross precipitation is lost to canopy interception, 2–5% of which is intercepted by litter [Blow, 1955; Helvey, 1964; Helvey and Patric, 1965]. Canopy interception during growing season rainstorms suppress transpiration rates that recover only after leaves dry. However, this effect is dampened because our daily long-term analysis produces multiday stormflow events that include nonrain days when transpiration has recovered.

Removal rates of water from the shallow subsurface via transpiration can alter baseflow [Bond et al., 2002]. At CHL, drainage from the shallow subsurface is critical for stormflow and baseflow generation [Hewlett and Hibbert, 1967], but the top 60 cm of this zone also coincides with 95% of all root biomass [McGinty, 1976]. In terms of stormflow generation, other studies show that higher rates of transpiration lead to faster declines in recession hydrographs resulting in relatively less stormflow production [Shaw and Riha, 2012] and larger thresholds. Small but persistent changes in transpiration as a result of short-term atmospheric and long-term climatic and forest canopy conditions scaled to the catchment can have profound effects on both baseflow and stormflow measured at the outlet [Ford et al., 2011a] driving interannual threshold variation.

**4.2. Nonunique Thresholds and Hydroclimate Nonstationarity**

Nonuniqueness of long-term interannual thresholds and linear stormflow response slopes ($m_2$) may be linked to nonstationarity [Milly et al., 2008] in vegetation-climate interactions. Changing runoff response as
a result of hydroclimatic variability is observed in studies of precipitation elasticity of streamflow [Chiew et al., 2006] during the Millennium Drought in southeastern Australia between 1997 and 2009 [Saft et al., 2015]. Findings from Saft et al. [2015] describe shifts in runoff response throughout catchments in Australia as driven primarily by changes in biophysical properties of the catchment. Persistent drought stress over several years can lead to changes in vegetation composition, biomass, and resilience. Other factors may include changes in rainfall intensity or changes to vertical heterogeneity of soil properties and preferential flow paths.

Vegetation-climate interactions that contribute to structuring rainfall-runoff relationships span several timescales, so both short and long-term controls of vegetation must be considered. The long-term effects of drought and drought recovery times of vegetation can introduce variability in thresholds. Previous work demonstrates that strategies for water regulation by various tree species can be impaired by droughts [Grier and Running, 1977; Buccè et al., 2005; Maseda and Fernández, 2006; Zwieniecki and Holbrook, 2009; Ford et al., 2011a] reducing maximum transpiration rates across the catchment. Recovery from physiological responses to water stress, such as decreased leaf area [Grier and Running, 1977], early leaf senescence, abscission of fine roots [Buccè et al., 2005; Maseda and Fernández, 2006], or in extreme cases embolism formation in xylem [Tyree and Sperry, 1989] can reduce transpiration rates even after drought subsides.

Drought-induced mortality was not observed at Coweeta during our study period despite drought mortality in the past [Clinton et al., 1993], however a study conducted by Hwang et al. [2014] shows reductions in ET and growing season length following severe drought in 2001. Sap flux data from WS18 between 2004 and 2006 also demonstrates the effects of early season dry spells on reductions in late season transpiration [Ford et al., 2011a] potentially lowering thresholds in 2006 after controlling for ASI. Lower leaf area following dry periods can increase throughfall also lowering thresholds. While interception likely reinforces threshold behavior, Ford et al. [2010] found that patterns of ET in response to climate are driven more by changes in transpiration than in interception. The relative importance of direct soil evaporation is considered negligible at CHL [Ford et al., 2007], but may increase during periods of water stress contributing to smaller stormflow volumes. Time lags in response and recovery of transpiration rates can create greater runoff response in wetter years following drought when controlling for antecedent storm conditions, which may account for positive trends in threshold values and seasonal rainfall observed in Figure 7.

4.3. Is Threshold Representation of Stormflow Adequate?

Previous studies observing stormflow threshold behavior as a function of gross precipitation and antecedent soil moisture assume effectively no stormflow response below the threshold [Detty and McGuire, 2010; Graham and McDonnell, 2010]. PRA relaxes this assumption by optimizing fits on both sides of the threshold enabling nonzero m1 slope parameters. Threshold behavior in WS14 and within individual years of WS18 and WS27 indicates slow and fast runoff processes denoted by increases from m1 to m2 slope parameters. Relatively low m1 values may depict more gradual stages of runoff generation. Oswald et al. [2011] showed similar transitions from m1 to m2 that were an order of magnitude different.

Like other threshold behavior studies, our design cannot directly discriminate between specific runoff processes but other studies at CHL discuss increasing connectivity and subsurface drainage leading to gradually more stormflow [Yeakley et al., 1998]. Increasing connectivity of saturated areas is shown to be weakly associated with threshold relationships in humid, temperate catchments with deep, forested soils [Ali and Roy, 2010]; however, these periods of slowly increasing stormflow response, denoted by small m1 values, may still be indicative of small but expanding variable source areas (VSAs) [Hewlett and Hibbert, 1967; Dunne and Black, 1970; Hewlett and Nutter, 1970]. As VSAs extend and increase into soils with greater hydraulic conductivity, rapid lateral drainage can occur [Mirus, 2015] along with greater macropore connectivity and flow [Band et al., 2014] demonstrating quick transitions from low to high stormflow generation. The question becomes whether these transitions in the long-term resemble linear threshold behavior or are more nonlinear in nature.

From our analysis, interannual variation of rainfall-runoff relationships manifest as concomitant increases in thresholds and steepening of slopes suggesting that stormflow behavior diverges as a result of measurement period (Figure 8). Stormflow thresholds are well-defined at short timescales, but coalesce into apparent nonlinear relationships at long timescales. At CHL, analysis of long-term data where thresholds are not discernable become more apparent when data were subset to specific years or seasons to reflect other
studies (Supporting Information Figure S1). In general, we find that thresholds are constant at short timescales and variable at long timescales. These long-term, nonunique interannual thresholds suggest that stormflow response is not solely a function of catchment geophysical characteristics [Oswald et al., 2011] or climate [Graham and McDonnell, 2010], but also ecosystem transpiration rates driven by changing vegetation dynamics [Saft et al., 2015].

At event timescales, higher rates of ET are shown to accelerate hydrograph recession attenuating runoff peaks and stormflow volume [Shaw and Riha, 2012]. Controlling for initial conditions prior to storm initiation, we would expect greater rates of ET to produce less stormflow overall. We observe this as greater thresholds during wetter growing seasons when ET rates are high and unaffected by drought (Figure 7). Greater values of $m_2$ during the growing season (Table 2) can be driven by high-intensity convective storms which reduce transit times. This effect occurs during two rare high-intensity rainfall events during the dormant seasons of 2008 and 2009 (Figure 7). High-intensity summer storms can also lead to smaller thresholds for similar gross precipitation and ASI totals, although we did not observe significant seasonal threshold differences in our study. If linear threshold response is unique, we would expect thresholds and $m_2$ slope parameters to be similar, independent of storm size, intensity, season, and year.

To summarize how long-term vegetation-climate interactions might alter interannual thresholds and linear stormflow response, we consider the effects of predrought and postdrought conditions on ecosystem transpiration rate. In Figure 9, predrought and postdrought scenarios are posed to conceptually demonstrate how growing season stormflow hydrographs may vary. Vegetation affected by drought may suffer from tissue damage or reduced biomass preventing efficient water uptake even when conditions are ideal. This is in contrast to vegetation prior to drought, when transpiration occurs uninhibited. For simplicity, we assume identical ASI, antecedent baseflow, and storm characteristics. Baseflow conditions prior to storm onset likely vary between predrought and postdrought, but are not represented in Figure 9.

Under a predrought scenario, transpiration peaks following a rainstorm and draws water from the shallow subsurface. Large rainstorms produce enough subsurface drainage to last several days to weeks in Coweeta (Figure 2). Because most tree rooting depths are well within this drainage zone, transpiration can quickly utilize actively draining water that would otherwise flow to the stream. This produces relatively smaller stormflow volumes, in addition to attenuating peaks and creating faster streamflow recession characteristics [Wittenberg and Sivapalan, 1999; Shaw and Riha, 2012]. In a postdrought scenario with identical antecedent conditions, impaired vegetation will have lower maximum transpiration rates. This means vegetation utilizes less actively draining water resulting in greater stormflow volumes reaching the stream.

**Figure 9.** Conceptual representation of stormflow response in an ecosystem that is healthy (dashed lines) and in the same ecosystem following a drought (solid lines) assuming identical antecedent soil moisture conditions and gross precipitation. This model assumes a continuous multiday storm hydrograph (blue). Transpiration (green bars) is represented by daily averages. Units are unspecified for rainfall, discharge, and transpiration are unspecified.
Competition for water resources in the shallow subsurface between transpiration and streamflow generation has implications for the two water worlds hypothesis [McDonnell, 2014] that describes the ecohydrologic separation between water for ecological use versus for hydrological use [Evaristo et al., 2015]. Evaristo et al. [2015] show groundwater and stream water are isotopically distinct from plant xylem and soil water, globally. Our findings suggest that this hydrologic disconnect may not always persist in watersheds at CHL. Seasonal differences in storm characteristics produce high-intensity summer storms [Laseter et al., 2012] that activate preferential flows facilitating greater ecohydrologic interactions in the shallow subsurface, but also faster stormflow generation. Isotopic studies at Coweeta suggest similar transience between periods when shallow groundwater and precipitation are well-mixed versus when they are distinct [Singh et al., 2016]. Lastly, sources of runoff at CHL are thought to primarily originate from shallow subsurface flow [Hewlett and Hibbert, 1967], as opposed to other humid, temperate sites where deep groundwater or fracture flow are important [Tromp-van Meerveld and McDonnell, 2006b] and support greater separation of ecological and hydrological uses of water. Our findings emphasize the importance of long-term vegetation-climate interactions on event-scale stormflow response, but more research is required to fully understand subsurface interactions between transpiration and streamflow generation.

5. Conclusions

Our study identifies variable threshold behavior of total stormflow response in watersheds throughout CHL by characterizing hillslope wetness as the sum of gross precipitation and antecedent soil moisture. The Southern Appalachian Region where CHL is located is experiencing increased hydroclimatic variability [Laseter et al., 2012] affecting vegetation communities throughout the basin [Elliott et al., 2015]. These effects can lead to vegetation-climate interactions that alter well-defined thresholds and linear stormflow response over long periods of time. Our study leverages short and long-term data encompassing 15 years of soil moisture, precipitation, and discharge data to investigate three hypotheses. We hypothesize that (1) watersheds at CHL display well-defined threshold behavior at short timescales, (2) long measurement periods thresholds vary seasonally and interannually, and (3) that variation is driven by nonstationarity in vegetation-climate interactions leading to variable thresholds over long measurement periods.

From the three posed hypotheses, we conclude the following:

1. Stormflow threshold behavior is uniquely defined as the sum of gross precipitation and ASI, but only at subyearly measurement intervals.
2. Thresholds vary interannually, largely during the growing season, and in response to seasonal precipitation.
3. Threshold behavior and linear stormflow response vary with hydroclimatic nonstationarity indicating that they are not solely determined by catchment geophysical properties and climate, but can incorporate seasonal and interannual feedbacks of forest canopy evapotranspiration rates.

This study emphasizes the importance of considering emergent behavior over long measurement periods and argues the need to characterize variation in emergent patterns, particularly with nonstationarity of climate and vegetation-climate interactions.

References


McIntyre, D. T. (1976), Comparative root and soil dynamics on a white pine watershed and in the hardwood forest in the Coweeta Basin, PhD dissertation, Athens, Ga.


