**DYNAMIC THRESHOLD RELATIONS OF STORMFLOW RUNOFF IN HUMID HEADWATER CATCHMENTS OF THE COWEETA HYDROLOGIC LABORATORY**

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A thesis submitted to the faculty of the University of North Carolina at Chapel Hill in partial fulfillment of the requirements for the degree of Master of Arts in the Department of Geography

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## ABSTRACT

Charles Isaac Scaife: Dynamic Threshold Relations of Stormflow Runoff in Humid Headwater Catchments of the Coweeta Hydrologic Laboratory
(Under the direction of Lawrence E. Band)

This thesis examines the long-term threshold response of rainfall-runoff relationships at the Coweeta Hydrologic Laboratory located in the Southern Appalachian Mountains of North Carolina. Threshold relationships between total stormflow and antecedent wetness represent an important emergent behavior that has been observed in previous studies. These studies are limited to only several years of analysis, which raises questions about longer term non-stationarity in threshold response. To examine the influence of non-stationarity, this thesis uses 15 years of data collected by the Coweeta Long-Term Ecologic Research Study and USDA Forest Service at two long-term monitoring sites to supplement additional short-term observations. Results demonstrate that threshold behavior of stormflow generation exists in Coweeta as a function of total storm precipitation plus antecedent soil moisture. Long-term thresholds vary with respect to seasonality and interannual hydroclimate variability. Lastly, we found evidence of non-linear stormflow generation, which has implications for previously observed simple threshold behavior.

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# 1 INTRODUCTION

Runoff generation processes at the hillslope scale involve complex processes that are difficult to measure empirically and include an integration of flow times spanning several orders of magnitude (Sivapalan, 2003, Tetzlaff et al., 2008). The interactions of soil moisture (Western et al., 2002), preferential flow (Mosley, 1982), and the connectivity of saturated regions (Hewlett and Nutter, 1970; Dunne and Black, 1970) scale non-linearly 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. Alternatively, using a top-down approach, complex runoff generation processes can be simplified into emergent behaviors (Sivapalan, 2003).

Generalizing hydrology across spatiotemporal scales has become increasingly important for understanding the effects of land use (Walsh et al., 2005; Grimm et al., 2008) 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 hydrologic laws rather than isolated exceptions (Sivapalan, 2003, McDonnell et al., 2007). 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 (Tromp van Meerveld and McDonnell, 2006a, Detty and McGuire 2010, Graham et al., 2010); 2) stormflow traveltime distributions (McGuire et al 2005); and 3) catchment-scale hysteresis of soil moisture and streamflow (Rosenbaum et al., 2012).

Studies investigating emergent patterns are often limited by measurement period, and those with long-term datasets do not directly address issues of non-stationarity raising questions concerning long-term stability of emergent patterns. We use well-established soil moisture, rainfall, and discharge measurements collected from the Coweeta Hydrologic Laboratory 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. This study examines the formation of emergent patterns using threshold response behavior of stormflow and addresses the following questions:

(1) Do thresholds in stormflow response exist in Coweeta and do they persist over long-term monitoring sites?

(2) Is threshold stormflow response non-stationary and if so, can we characterize threshold variation through time?

# 2 BACKGROUND

Threshold behavior in stormflow response, as an emergent pattern, has the potential to discriminate between runoff generation processes linking sub-hillslope processes to catchment discharge. A major question is whether simple, threshold-based storage-runoff response models can adequately reproduce and summarize behavior of the number of complex, interacting processes active at the hillslope and catchment level. A number of recent research studies have investigated this question. Tromp van Meerveld and McDonnell (2006a) excavated a 20 m long trench in the Panola Mountain Research Watershed (PMRW) where they measured 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. Detty and McGuire (2010a, b) directly combined antecedent soil moisture integrated with depth and total storm precipitation to evaluate similar threshold response for the Watershed 3 catchment at the Hubbard Brook Experimental Forest. Their study 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 significantly high amounts of subsurface stormflow. Graham and McDonnell (2010) showed that stormflow response from 50 years of rainfall-runoff data at HJ Andrews Experimental Forest was related to interstorm period. Shorter interstorm periods were correlated with higher pre-event 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 non-linear 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 two years during June and July with portable soil moisture probes, in addition to four continuous soil moisture sites that measured from June to October.

Mechanisms of non-linear stormflow response in forested catchments have generally been attributed to the activation of subsurface macropore flow, but hillslope connectivity to near-stream riparian areas may be essential in forming non-linear stormflow responses (Freer et al., 2002). Tromp-van Meerveld et al. (2006b) proposed the fill and spill hypothesis that describes the importance of bedrock topography and flow at the soil-bedrock interface as a major contributor to non-linear stormflow response. 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. Transport at the soil-bedrock interface combined with activation of subsurface macropore flow contributes to strongly non-linear stormflow even at the hillslope scale. Uchida et al. (2005) showed that once significant pipe flow was generated, the relationship between pipe flow and hillslope discharge became strongly non-linear varying with rainfall intensity. Pipe geometry and connectivity with upslope regions prior and during storms was difficult to characterize, but likely important to non-linear runoff response. One measurable mechanism connecting upslope areas to streams is bedrock fracture flow. At PMRW, Tomp-van Meerveld et al. (2007) found that facture flow contributed up to 21% of all streamflow over a 2 year wet period and that flow through the bedrock at event time-scales continued to feed streams days after the storm event ended. They suggest that antecedent moisture and rainstorm size and intensity influence the relative contribution of pipe flow and fracture flow contributing to non-linear stormflow response (Tromp-van Meerveld et al., 2006a).

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 are shown to shift towards non-local controls (e.g. topography) with senescence and higher seasonal rainfall (Grayson et al., 1997). During the dormant season, less evapotranspiration and greater lateral redistribution increase runoff sensitivity to rainfall. Lateral redistribution during the growing season is inhibited by plant water uptake altering hillslope connectivity and subsurface flows. This observation highlights competition among runoff and transpiration outputs over limited root-zone soil moisture, particularly in deep forested soils where streamflow generation is predominately shallow subsurface flow (Hewlett and Hibbert, 1967). Significant runoff generation relies on increased levels of hillslope connectivity under wetter soil moisture in the shallow subsurface, but transpiration during the growing season actively removes soil water in this zone. The effects of shifting plant phenology and transpiration combined with interannual hydroclimate variability (Hwang et al., 2011; 2014) may alter well-established emergent patterns over time frames on the order of several years to a decade.

In forested ecosystems, transpiration is a major pathway for rainfall inputs comprising more than half of the annual rainfall in Coweeta (Swift et al., 1988). Recent research has shown that hydroclimate extremes, in particular drought, can have prolonged effects on tree transpiration rates, and their response to drought varies by severity and xylem anatomy resulting in delayed rates of recovery (Zweiniecki and Holbrook, 2009). Typical responses to dry periods are reduced leaf area (Grier and Running, 1977; Golz, 1982) and increased rooting depths (Hacke et al., 2001), but under severe droughts leaf (Maseda and Fernandez, 2006; Bucci et al., 2005) and fine root abscission (Maseda and Fernandez, 2006) can occur in addition to hydraulic impairment resulting from embolisms forming within the xylem (Tyree and Sperry, 1989). During recovery, transpiration capacity of trees is reduced and may persist from several months to years (Hacke et al., 2001).

Ring-porous xylem anatomies have transpiration rates limited by lower stem conductance in comparison to their stomatal conductance. As a result, species with this xylem anatomy sustain lower rates of transpiration overall and tend to leave stomata open even through dry periods making them more susceptible to cavitation and mortality during severe droughts (Ford et al., 2010). However, diffuse-porous xylem anatomies mediate water through stomatal closure and exhibit quicker response to changes in vapor pressure deficits. They are capable of higher rates of transpiration but can become stressed after numerous dry periods resulting in leaf senescence and reduced leaf area. Scaling differences in drought response of individual trees and functional types to the catchment can have profound effects on well-established rainfall-runoff relationships following droughts (Ford et al., 2010). Thus, characterizing long-term controls of vegetation on emergent behavior is important for expanding these principles to longer hydrologic studies and to other climates.

Methods for identifying thresholds are not well established especially in cases where emergent patterns are not clearly discernible (e.g. in long-term datasets). 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), but as hydroclimate variability increases with longer timescales this method may become subjective and inefficient. Also, threshold analyses typically assume zero or close to zero slopes below the threshold, which may overlook more gradual processes contributing to stormflow such as expanding variable source area.

To date, studies have utilized seasonal to several year datasets to characterize emergent patterns of stormflow response with respect to measured rainfall and storage in forested catchments. Penna et al., (2011) showed strongly non-linear response of streamflow and shallow groundwater to soil moisture limited to only two summer periods due to winter snowpacks. 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 2006a,b). Detty and McGuire (2010) demonstrated compelling results that isolate simple threshold behavior by combining meteorology and antecedent soil moisture. However, their study spanned a single wet-up period from August to January, which included 14 storms. By far the most comprehensive analysis was at HJ Andrews Experimental Forest (Graham and McDonnell, 2010), but their 50+ years of data only included rainfall and catchment runoff using interstorm period to approximate antecedent wetness, which makes identifying mechanisms of threshold behavior difficult. Also, most of the rainfall is received during a single winter season in a mature coniferous forest making phenologic drivers of thresholds moot.

Our study leverages continuous data at the Coweeta Hydrologic Laboratory to replicate threshold analyses conducted by Detty and McGuire (2010). Their results suggest strong but simple thresholds during wet-up periods, where stormflow response below the threshold is near zero, but increases linearly with antecedent soil moisture and storm event precipitation after the threshold. To test this in Coweeta, we utilized a multi-phase linear regression to compute thresholds by minimizing the least-squares residuals. This method does not assume zero slopes below the threshold, which may reveal more gradual processes of stormflow generation.

Previous studies have also shown the importance of subsurface flow in Coweeta supplying streamflow (Hewlett and Hibbert, 1967), especially within the rooting zone, which requires consideration of vegetation response in developing these emergent patterns. We pose a set of four hypotheses that serve our central objective of characterizing long-term threshold behavior:

- We hypothesize that watersheds in Coweeta exhibit well-defined thresholds at short timescales on the order of a single season to a year, but longer measurement periods (>10 years) introduce variability as a result of non-stationarity in these hydroecological systems.
- We hypothesize a correlation between threshold values and seasonal rainfall due to vegetation response to hydroclimate extremes. The influence of hydroclimate extremes will also affect long-term catchment memory resulting in a correlation of threshold values with long-term rainfall totals.
- We further hypothesize that differences in dominant vegetation types between catchments will result in distinct threshold sensitivities to seasonal rainfall. In particular, catchments dominated by ring-porous species will exhibit greater sensitivity to seasonal rainfall totals due to water-use mediated by stem conductance that favors partial cavitation, a long-term hydraulic impairment, as a result of severe drought. Catchments dominated by diffuse-porous species will have thresholds that are less sensitive to seasonal rainfall, as they regulate water through stomatal closure producing less severe and shorter drought responses, such as leaf senescence.

- Lastly, we hypothesize that thresholds computed over long measurement periods will shift with vegetation and hydroclimate suggesting non-linear stormflow response over long measurement periods.

# 3 DATA AND METHODS

## 3.1 Study Site

The Coweeta Hydrologic Laboratory, located in North Carolina, USA is a USDA Forest Service 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 in the 1930s to study the effects of land management practices and vegetation cover on streamflow generation. Rainfall-runoff records extend back 80 years for a set of the gaged catchments. Research at Coweeta has since expanded to include community ecology, forest ecophysiology and dynamics, stream biogeochemistry, and aquatic and terrestrial biota.

Coweeta is an east-facing, bowl-shaped basin with a drainage area of 1626 ha and elevation range of 1000 m (Swank and Crossley, 1988). It is further subdivided into paired watersheds 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 downslopes areas. The Nantahala Escarpment runs along the exterior boundaries of high elevation watersheds (Wooten et al., 2007; Band et al., 2012). Watersheds that border the Nantahala Escarpment transition from colluvial low slope bottomlands to steep rocky slopes with outcrops (Band et al., 2012).

Meteorology and discharge measurements date back to mid-1930s. 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 at Coweeta even at high elevations. The area is also prone to tropical storms that can trigger landslides especially around the Nantahala Escarpment (Wooten et al., 2007; Hales et al., 2009; Band et al., 2012). Furthermore, increasing long-term hydroclimate variability has increased extreme wet and dry periods that can impact timing of vegetation senescence (Hwang et al., 2014).

Forest community composition varies across elevation and moisture gradients from oak-pine dominance along dry, low elevation ridges 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 comprised of *Liriodendron tulipifera* (tulip poplar) and *Tsuga canadensis* (eastern hemlock), although recent infestation by woolly adelgid has killed off the hemlock canopy. 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). *Acer rubrum* (red maple) has become pervasive throughout the basin as a result of widespread *Castenea dentata* (American chestnut) mortality following chestnut blight in the 1930s (Elliot and Swank, 2007; Elliot and Vose, 2011). Much of the understory throughout the basin is comprised of two evergreen shrubs: *Kalmia latifolia* (mountain laurel) and *Rhodendron maximum* (Day et al., 1988). Changing stand composition along environmental gradients and increasing growing season lengths with elevation create complex phenologic patterns across the basin (Whittaker, 1956; Day and Monk, 1974; Hwang et al., 2011).

Our study focuses on three headwater catchments with variable basin morphology and environmental conditions (Figure 3.1). WS14 and WS18 are neighboring low elevation catchments with similar aspects. WS14 is a 62.4 ha northwest facing control watershed with an undisturbed mixed hardwood forest. Relief in WS14 is 285m implying little orographic effect on precipitation. 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 control catchment with an elevation range of 393m (Swank and Crossley, 1988) and relatively higher rainfall and runoff ratios (Swift et al., 1988). Due to milder temperatures, the growing season in WS27 is shortened by almost a month (Hwang et al., 2011). WS27 backs up to the Nantahala Escarpment giving it steeper slopes higher in the catchment that decrease towards the outlet (Wooten et al., 2008; Hales et al., 2009).

## 3.2 Long-term Climate and Hydrology

Daily precipitation is collected by four gaging stations and a climate station that span the elevation gradient of Coweeta. The base climate station (CS01; Figure 3.1), located on the valley floor, has been actively measuring daily precipitation, temperature, relative humidity, and radiation as early as the 1930s. Long-term data from CS01 is supplemented in this study by four additional rain gages to account for spatial heterogeneity and orographic effects in rainfall. 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 were averaged to account for precipitation gradients. Daily discharge from each catchment is measured using v-notched weirs with continuous discharge data dating back to the mid-1930s. This study subsets the long-term discharge measurements to coincide with *in situ* soil moisture measurements that began in 1999.

Climate and hydrology data for Watersheds 18 and 27 were trimmed to correspond to 15-years of soil moisture measurements recorded from 1999-2014. This time domain includes wet and dry precipitation extremes and several hurricanes. A prolonged drought occurred at the onset of the study lasting from 1999-2001. It peaked in 2000 and 2001 when rainfall was below the long-term annual average of 1796 mm by 31% and 22%, respectively. A wetter period between 2002 and 2005 followed where average precipitation was 1946mm, 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, where rainfall was 2374mm (+31.8%) and 2368mm (+32.1%), respectively. Several hurricanes also occurred, including Hurricane Frances and Ivan in 2004, which caused deadly landslides throughout the region. Overall, rainstorms at Coweeta are typically spaced 4 days apart but during droughts were as long as 39 days.

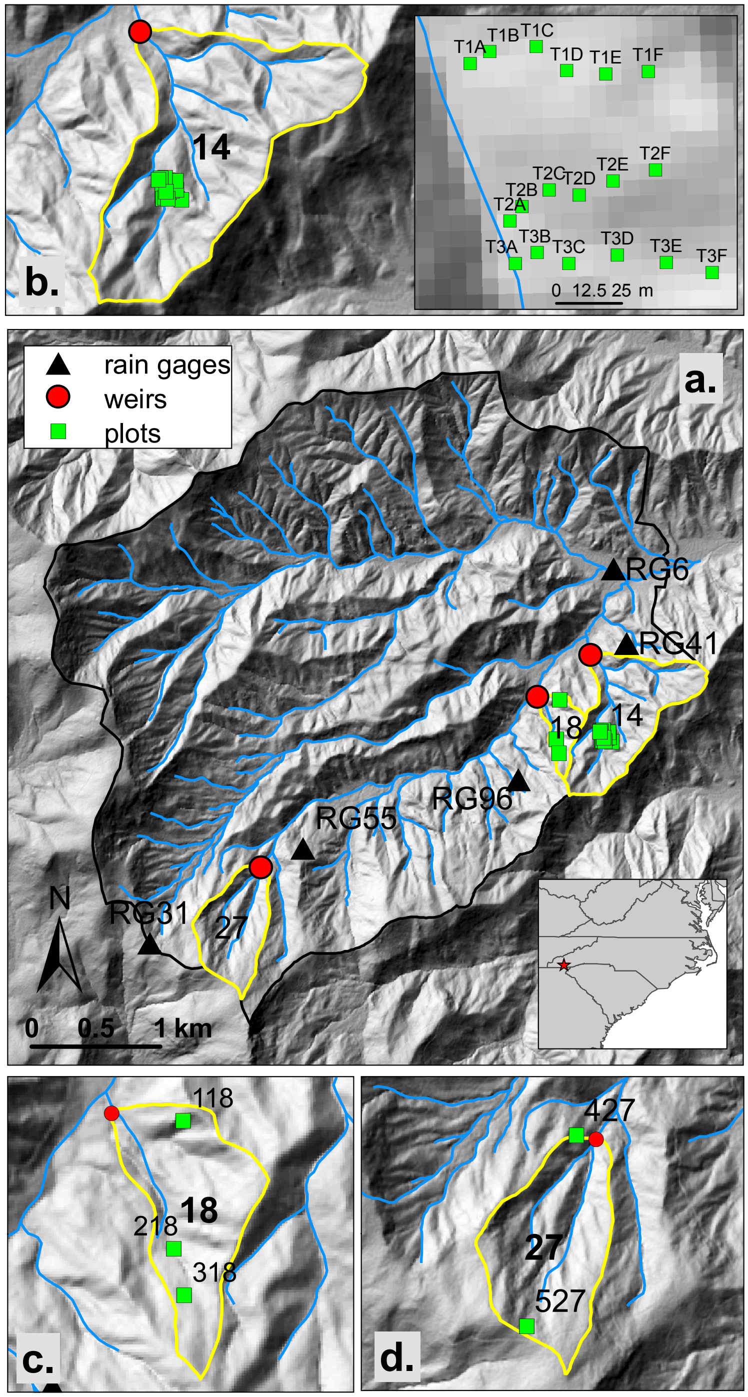


**Figure 3.1.** A hillshade map of the (a) the Coweeta Basin at the Coweeta Hydrologic Laboratory in SW North Carolina, USA, with precipitation stations, (b) WS14 with its 18 soil moisture plots in three transects, and both LTER Catchments (c) WS18 and (d) W27.

## 3.3 Field Measurements

In catchments WS18 and WS27, five soil moisture plots were established along a terrestrial ecohydrologic gradient by the LTER (Figure 3.1c-d). WS18 has three gradient plots located along a moisture gradient including a xeric pine-oak stand (118), an intermediate mixed oak stand (318), and a mesic cove hardwood stand (218). WS27 has one plot in a mesic mixed oak stand (427) and a second plot in a high elevation northern hardwood stand (527). Water content at each plot is measured every 15 minutes by four Campbell CS616 time-domain reflectometer (TDR; Campbell Scientific Inc., Logan, UT, USA) probes totaling 20 probes throughout both LTER catchments and has been collected continuously since 1999. The 15 minute interval measurements are averaged to daily times steps. Probes are inserted at two depths (0-30cm & 30-60cm) 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 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 were 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. Three near-stream plots (3A, 3B, 3C; Figure 3.1c) 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 0-30cm and 30-60cm in an upslope and downslope topographic position. Measurements from the Campbell CS616 TDR probes are calibrated using a second order polynomial factory calibration, where soil moisture is calculated as a volumetric wetness content (VWC) from period (T) in microseconds. Quadratic calibration functions are ideal for soils with high moisture variability because they reduce error at moisture extremes.

### 3.4 Storm and Threshold Definition

Following previous event-based studies, storms were defined as beginning with rainfall greater than 1 mm/day and ending when stormflow returns to zero (Hewlett and Hibbert, 1967; Detty and McGuire, 2010). This requires a hydrograph separation, which we estimated using the USGS HYSEP algorithm, adapted in MATLAB 2014b (The Mathworks, Inc.; Natwick, MA). This method utilizes a local minimum method (Pettyjohn and Henning, 1979) for calculating baseflow and quickflow from total discharge. We define storm events following Detty and McGuire (2010) and adapt it for daily timesteps to account for our extensive long-term discharge and soil moisture data that has a daily temporal resolution. We use this method for consistency and comparison with previous findings and do not claim it clearly separates specific sources of runoff.

Catchment wetness prior to storms was estimated from TDR measurements in each watershed. We used the 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 was combined with total storm precipitation to estimate an event total wetness from which significant threshold relations are shown to arise (Detty and McGuire, 2010). Time series data were partitioned into dormant and growing season with respect to topographically mediated phenology (Hwang et al., 2011) that produced growing seasons nearly 3 weeks shorter in WS27 than in WS18 and WS14.

Thresholds were calculated using an algorithm that minimizes the least-squares of residuals by computing the linear-regression of two discrete subsets of the data adapted in MATLAB 2014b (The Mathworks, Inc.; Natwick, MA). This method differs from previous rainfall-runoff studies by minimizing the sum of squares residuals on both sides of the threshold simultaneously rather than visually estimating. It also identifies strong breaks in slope between total event wetness and total stormflow indicating possible changes in subsurface flow processes. By identifying breaks in slope, this method can produce non-zero slopes below the threshold and signify accelerating subsurface stormflow generation processes (e.g. increasing near-stream saturation) whereas previous work assumed negligible stormflow response below the threshold (Detty and McGuire, 2010; Graham and McDonnell, 2010). Also, a least-squares approach can identify thresholds during periods of low stormflow variability. For example, during dry years, the maximum observed total stormflow may be significant (> 5mm) but a threshold may not be obviously identified using visual methods.

To calculate breaks in slope, the data was iteratively divided into upper ($s_u$) and lower ($s_l$) subsets that varied in size from $s_u=3$ to $s_u=n-3$ conserving the total number of storms ($n$), so that $s_u+s_l=n$. Within each iteration, a linear regression was computed for each subset using MultiRegressLines.m, an open-source statistical algorithm developed in MATLAB by Andrew Ganse at the University of Washington. The solution with the least-squares residuals became the working solution. Variance in the working solution was calculated using a Monte Carlo analysis that randomly subsampled model parameters from a Gaussian distribution of their standard deviations.

# 4 RESULTS

## 4.1 Threshold Stormflow Response

Total stormflow response from all three watersheds exhibit threshold behavior as a function of gross precipitation and ASI. Measurements made over 19 months in WS14 demonstrate a strong break in slope, as seen in Detty and McGuire (2010). Of the 67 storms observed occurring between July 2011 and January 2013, eight were not analyzed due to instrument outage of TDR and 11% of the remaining 59 storms showed significant stormflow response above 5 mm/day (Figure 4.1a-c). The maximum total stormflow peaked at 26.8 mm/day during the study period, which was notably smaller than long-term stormflow records from WS18 and WS27. Despite a smaller range in total storm discharge, gross precipitation was still significantly correlated ($r^2$=0.87; Figure 4.1a) with stormflow independent of ASI, when fit with a simple linear regression. ASI showed no linear relationship with total stormflow ($r^2$=0.01; Figure 4.1b). Thresholds became more evident when ASI and gross storm precipitation were combined, which was consistent across all three watersheds.

Applying the same method to 15 year stormflow response in WS18 and WS27 reveals less distinct threshold behavior than the 19 month record from WS14. In WS18 and WS27, there were 811 and 791 storms observed, respectively, between February 1999 and October 2014. WS27 had the largest observed stormflow of 222.7 mm followed by WS18 with 69.2 mm. In WS18, the amount of precipitation received was comparable to WS14 but the long-term data captured more extremes in stormflow generation. Also, WS18 has a slightly higher runoff ratio than WS14. Stormflow in WS18 was linearly related to precipitation (Figure 4.1d) but had no linear correlation with ASI (Figure 4.1e), which was consistent with WS14. Over 15 years of measurements, ASI plotted against total stormflow showed no linear trends in either WS18 and WS27, but variability in stormflow runoff did increase at higher values of ASI (Figure 4.1e,h). When ASI is low, relatively smaller stormflows are produced than at mid- to high-values of ASI where stormflow is highly variable. Combining ASI and total precipitation created a strongly non-linear relationship.

A comparison of thresholds across all three watersheds shows an overall weaker correlation in the long-term data (Figure 4.1c, f, i). Also, the multi-phase linear regression used to define threshold behavior showed stronger total correlations in data above the threshold when gross precipitation and ASI were combined, which is consistent with findings from Detty and McGuire (2010). Figure 4.1(c, f, i) shows the results of our threshold method applied to the sum of gross precipitation and ASI, where storms above and below the threshold are fit with respective linear regressions. Correlations of each regression are weaker compared to plots of total stormflow with only precipitation as the explanatory variable. However, computing the correlation coefficient from the threshold method over both linear fits (i.e. the fit above and below the threshold) generates a combined $r^2$-value greater than regressions computed from plots using only precipitation. WS14 had the strongest total correlation computed from the combined least-squares regressions of the lower and upper storm events. There were 14 storms that were larger than the threshold and 45 that were smaller. The linear regression through the upper and lower subsets had significant ($p<0.01$) coefficients of determination of 0.96 and 0.46 (Figure 4.1c), respectively. Differences between upper and lower coefficients of determination were not as great in WS18 and WS27 but still significant. In WS18, coefficients of determination on either side of the threshold were 0.46 (upper) and 0.12 (lower) with only 213 storms out of 811 (26.2%), occurring above the threshold. WS27 had a larger proportion of storms above the threshold (34.7%) that were also more strongly correlated than storms below the threshold, 0.68 and 0.37 respectively.

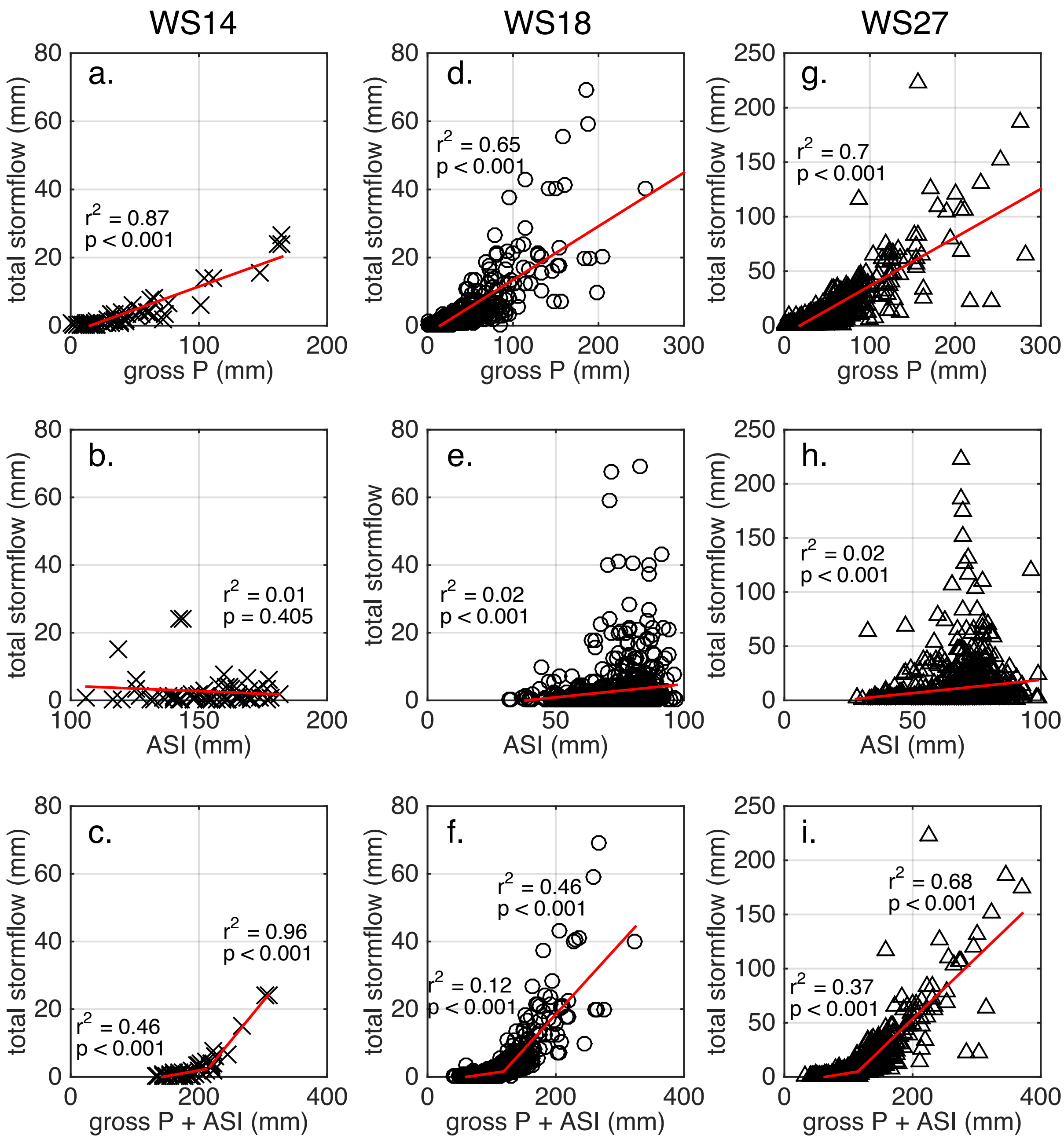


**Figure 4.1.** Summary of the storm event results where total stormflow is plotted as a function of (1) gross storm event precipitation (a, d, g), (2) antecedent soil moisture index (b, e, h), and (3) the sum of gross precipitation and ASI combined (c, f, i). Measurement period for WS14 is Jul. 2011-Jan. 2013 and for WS18/WS27 is Feb. 1999-Oct. 2014.

## 4.2 Temporal Dependence of Threshold Behavior

An analysis of soil moisture, precipitation, and discharge data over the 19 month period in WS14 show growing and dormant season shifts in thresholds that are not statistically significant. The computed growing season threshold was 192 mm, whereas the dormant season threshold was 206 mm with standard deviations of 11.1 mm and 6.7, respectively (Figure 4.2). Total stormflows were larger in the growing season but storms producing significant runoff (> 5mm) were more numerous in the dormant season. Seasonal differences in the coefficients of determination showed stronger correlations in the growing season. In Figure 4.2, linear regressions were strongest above the threshold in both seasons. Relative differences between lower and upper slopes within each season were greatest during the growing season. Differences in slope for rainstorms below the threshold were not significantly different between seasons, but this was not the case for slopes above the threshold (growing season $m_u$= 0.23; dormant season $m_u$= 0.08).

Applying the same analysis to our 15 year dataset in WS18 and WS27, long-term stormflow response patterns diverge from observations in WS14. Unlike WS14, the maximum total stormflow observed was not significantly higher in the growing season and the average long-term thresholds were greatest during the dormant season (Figure 4.3). The magnitudes of stormflow response were comparable between seasons, however their average thresholds for the study period were roughly 118 mm in the dormant season and 103 mm in the growing season in WS18 and WS27. Like WS14, seasonal threshold differences were not significantly different. Correlation coefficients were computed for linear regressions above and below the threshold and for the total fit of all storms. Table 1.1 summarizes the number of storms with respect to threshold and their combined correlation coefficients over the entire study period. On average, the dormant season had a greater proportion of storms above the threshold. WS18 had a greater number and proportion of storms below the threshold in both seasons, while WS27 had the highest proportion of storms above the threshold (41.3%). Correlation coefficients were similar in all watersheds and season ranging from 0.85 to 0.88.

Figure 4.3 also shows that over a 15 year measurement period, the thresholds that define significant stormflow production vary interannually. This variation is greatest during the growing season and the most in WS27 ($SD_{18,growing}$ = 21.7 mm; $SD_{27,growing}$ = 27.5 mm). Interannual thresholds during the dormant season are relatively more clustered ($SD_{18,dormant}$ = 18.1 mm; $SD_{27,dormant}$ = 14.2 mm). Analyzing yearly thresholds through time reveal that WS18 and WS27 follow similar paths (Figure 4.5). In the dormant season, interannual variations are gradual and lack obvious trends, but during the growing season thresholds slightly trend upward from 2000 to 2005 and drop drastically in 2006. They remain low until they increase again in 2010 and 2011 for WS27 and WS18, respectively. This continues for two years after which they decline again.

The relative patterns of thresholds between seasons also differ between the LTER catchments and WS14. This observation may be a result of differences in measurement periods between the LTER catchments and WS14. Figure 4.4 subsets the long-term data to coincide with data collected from WS14. Using identical time periods, patterns of threshold behavior are relatively consistent across all three catchments. Larger storms are observed during the growing season leading to steeper slopes above the threshold, which may provide evidence for non-linear behavior over strictly threshold behavior. Slopes below the threshold are comparable between both seasons.

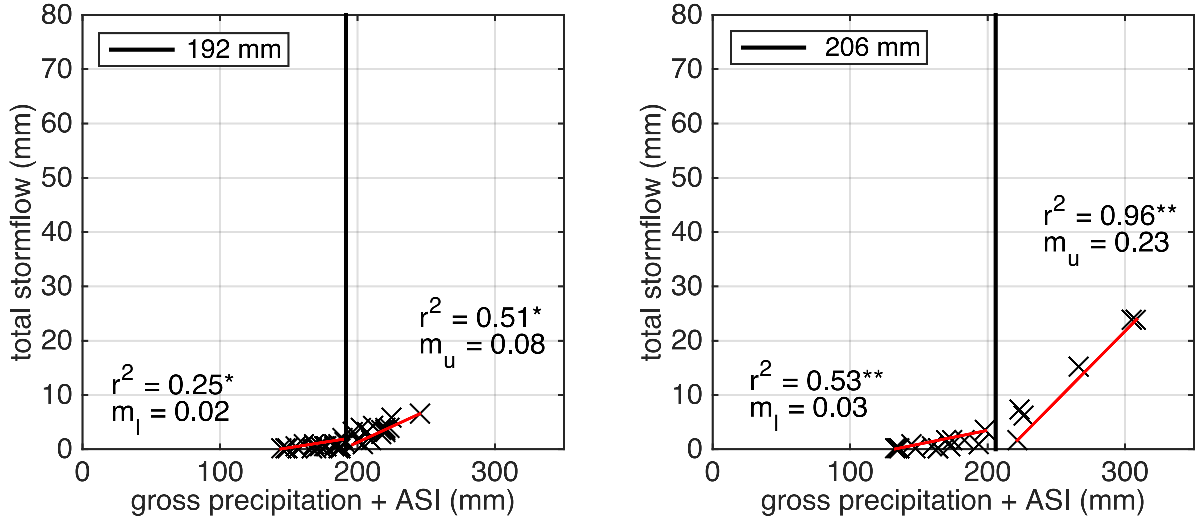


**Figure 4.2** Dormant (a) and growing (b) season response of total stormflow to gross precipitation and ASI in WS14 for the period 07/2011 to 1/2013. Threshold values are represented by black vertical lines and storms on either side of the threshold are fit separately. Their respective $r^2$ values and slopes ($m_{u,l}$) are shown. (* $p < 0.05$, ** $p < 0.01$)

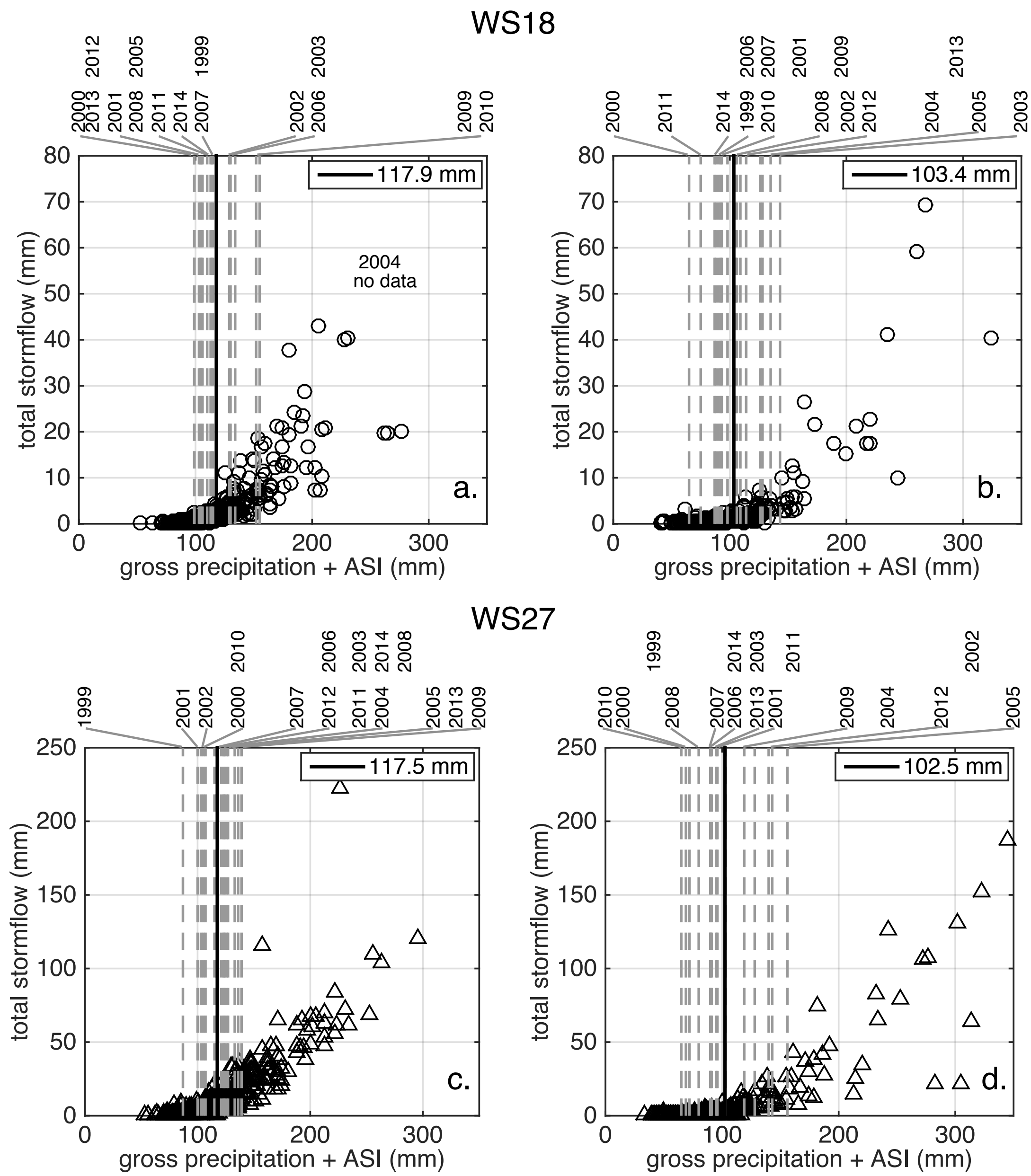


**Figure 4.3**. Long-term growing and dormant season response of total stormflow to gross precipitation and ASI in WS18 (a, b) and WS27 (c, d) between 1999-2014. Thresholds are averaged over all years for each season (black line) and over each individual year (dashed line). The distribution of yearly thresholds is expanded above each plot to show their corresponding year.

**Table 4.1** Summary statistics from interannual threshold analysis in WS18 and WS27 for the average number of storms below the threshold ($S_l$), above the threshold ($S_u$), and the combined correlation coefficient when each subset is fit with a linear regression.

| | **Dormant Season** | | | **Growing Season** | | |
|---|---|---|---|---|---|---|
| | $\overline{S_l}$ | $\overline{S_u}$ | $\overline{R^2}$ | $\overline{S_l}$ | $\overline{S_u}$ | $\overline{R^2}$ |
| WS18 | 17.9 (7.6) | 10.6 (5.7) | 0.85 (0.12) | 15.2 (4.7) | 7.1 (3.5) | 0.86 (0.18) |
| WS27 | 16.9 (6.8) | 11.9 (6.5) | 0.86 (0.12) | 11.6 (4.3) | 7.4 (3.5) | 0.88 (0.13) |

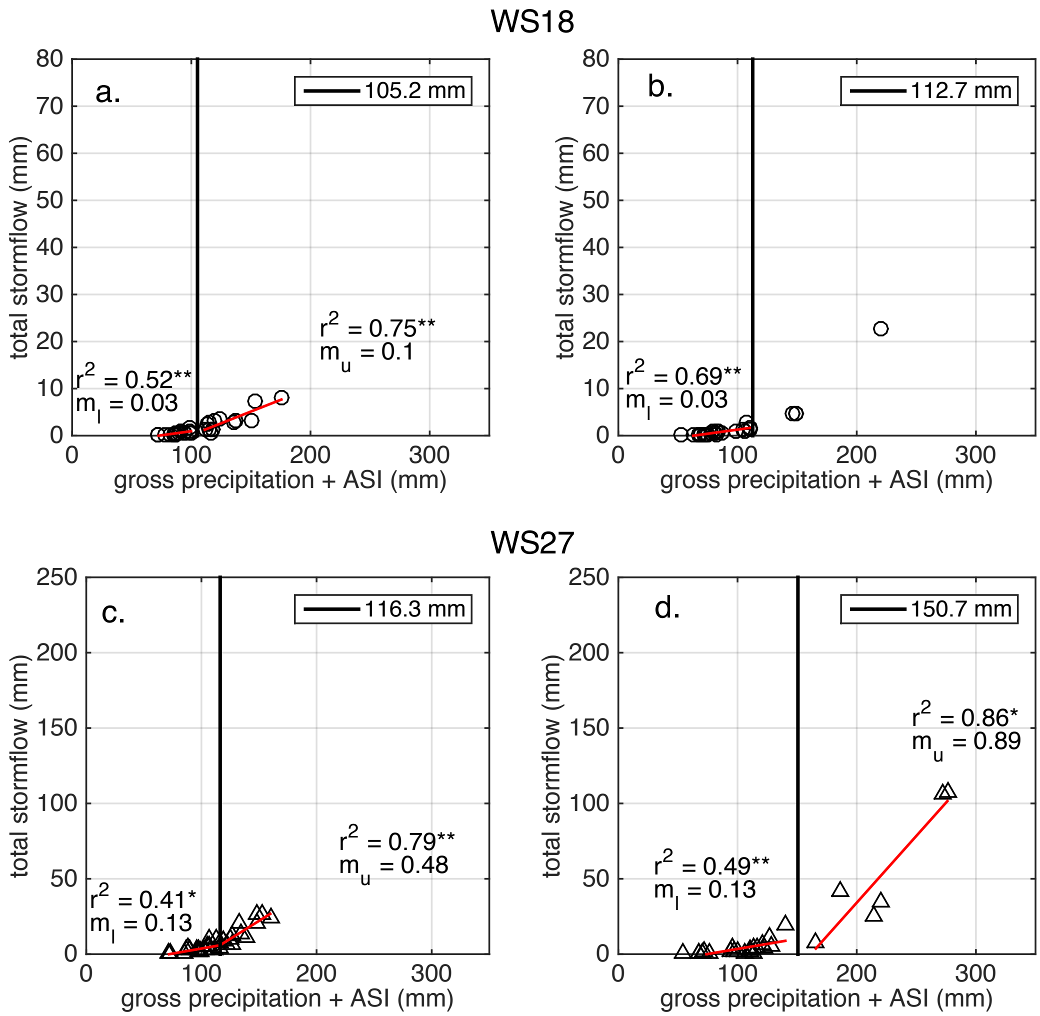


**Figure 4.4** WS18 and WS27 long-term datasets restricted to the WS14 measurement period from July 2011 to January 2013. Storm events are separated into dormant season (a, c) and growing season (b, c) with linear regressions (solid red line) computed for storms below and above the threshold. Respective $r^2$ (* p < 0.05, ** p < 0.01) and slopes ($m_{u,l}$) of each regression are also displayed.

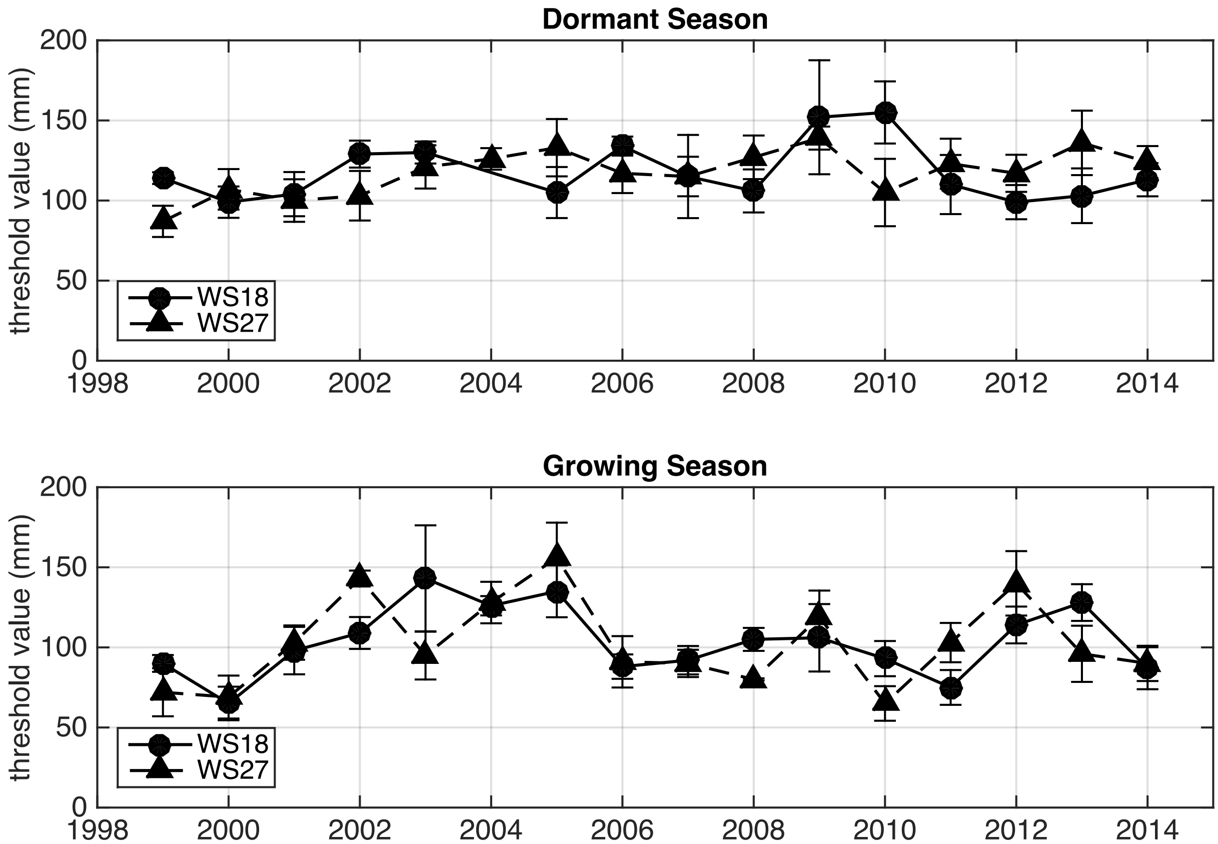


**Figure 4.5.** Dormant (a) and growing (b) season thresholds in WS18 (circles and solid line) and WS27 (triangles and dotted line) plotted as a timeseries.

### 4.3 Long-term hydroclimate and threshold pattern variation

Variations in thresholds are correlated with recent meteorology but have no relationship with longer term hydroclimate. Figure 4.6 shows interannual threshold behavior as a function of seasonal hydroclimate. Significant positive correlations between threshold values and total seasonal rainfall persisted during the growing season and were strongest in WS18. Meanwhile, during the dormant season significant correlations between threshold and seasonal rainfall totals do not exist. In WS27, weak but significant correlation during the dormant season appears signifying catchment differences that are unique from WS18. In Figure 4.7, threshold values are plotted against the prior seasons' rainfall total to show memory effects of moisture storage on rainfall-runoff relationships, however linear regressions did not show significant trends in WS18 or WS27 during the dormant or growing season.

Slopes were computed annually for linear regressions fit to storms above the threshold for both long-term catchments. Figure 4.8 shows how slopes vary with thresholds and seasonal rainfall within each season. In general, the slope of total stormflow over the sum of gross precipitation and antecedent soil moisture increase with larger thresholds. This positive trend is consistent throughout dormant and growing season and both watersheds. WS27 has a larger range of slopes than WS18 and it produces values greater than one in two out of 15 dormant seasons. Positive trends during the WS27 dormant period are relatively less significant than growing and dormant season in WS18 and the growing season of WS27. Slopes in WS27 range from 0.06-1.57, whereas WS18 ranges from 0.02-0.87. Total seasonal rainfall shows little pattern with slope during the dormant season, but in the growing season slopes tend to be steeper when total seasonal rainfall is greater.

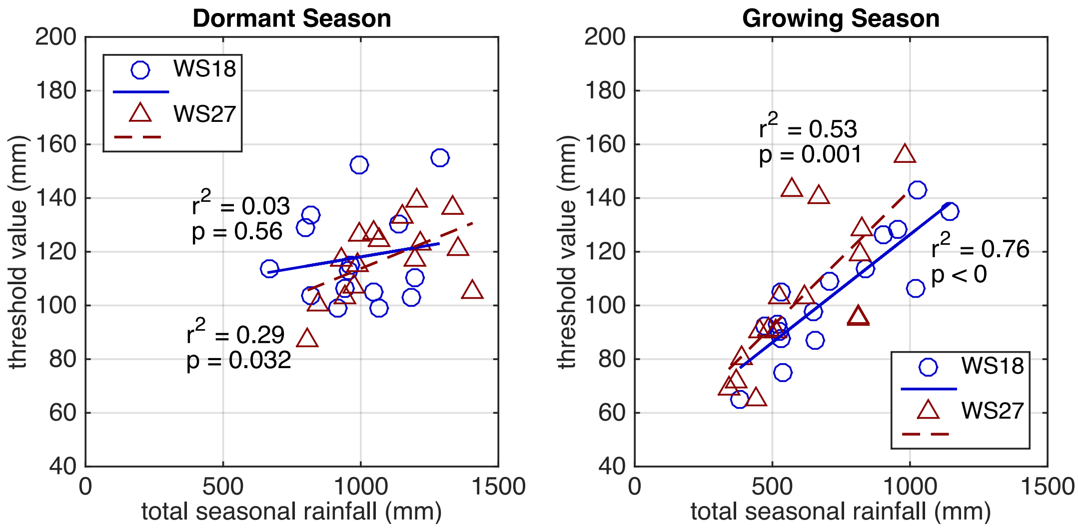


**Figure 4.6** Dormant and growing season response of rainfall-runoff thresholds to total seasonal rainfall in WS18 and WS27. Local phenology was determined for WS18 and WS27 separately following Hwang et al., 2011.

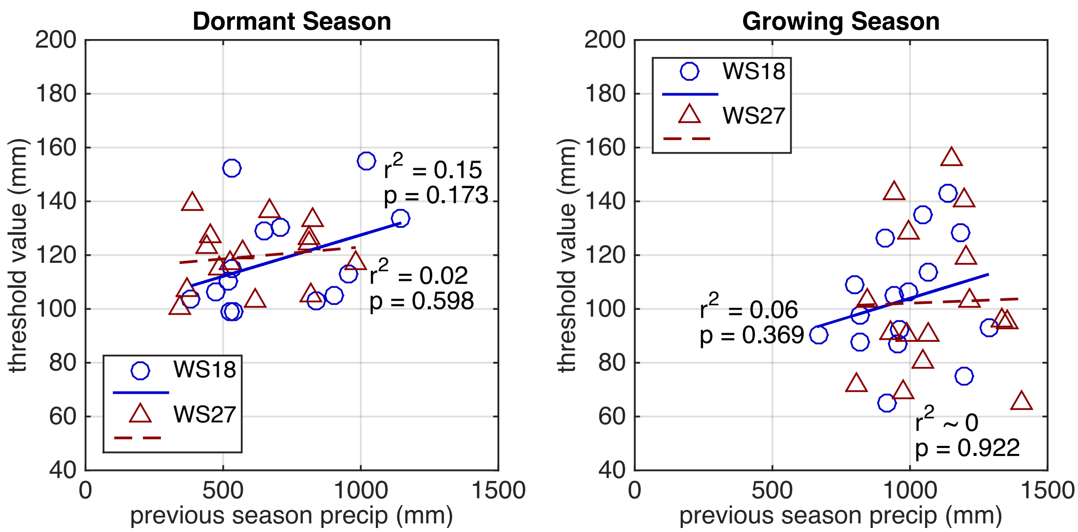


**Figure 4.7** Dormant and growing season threshold response to prior season precipitation in WS18 and WS27.

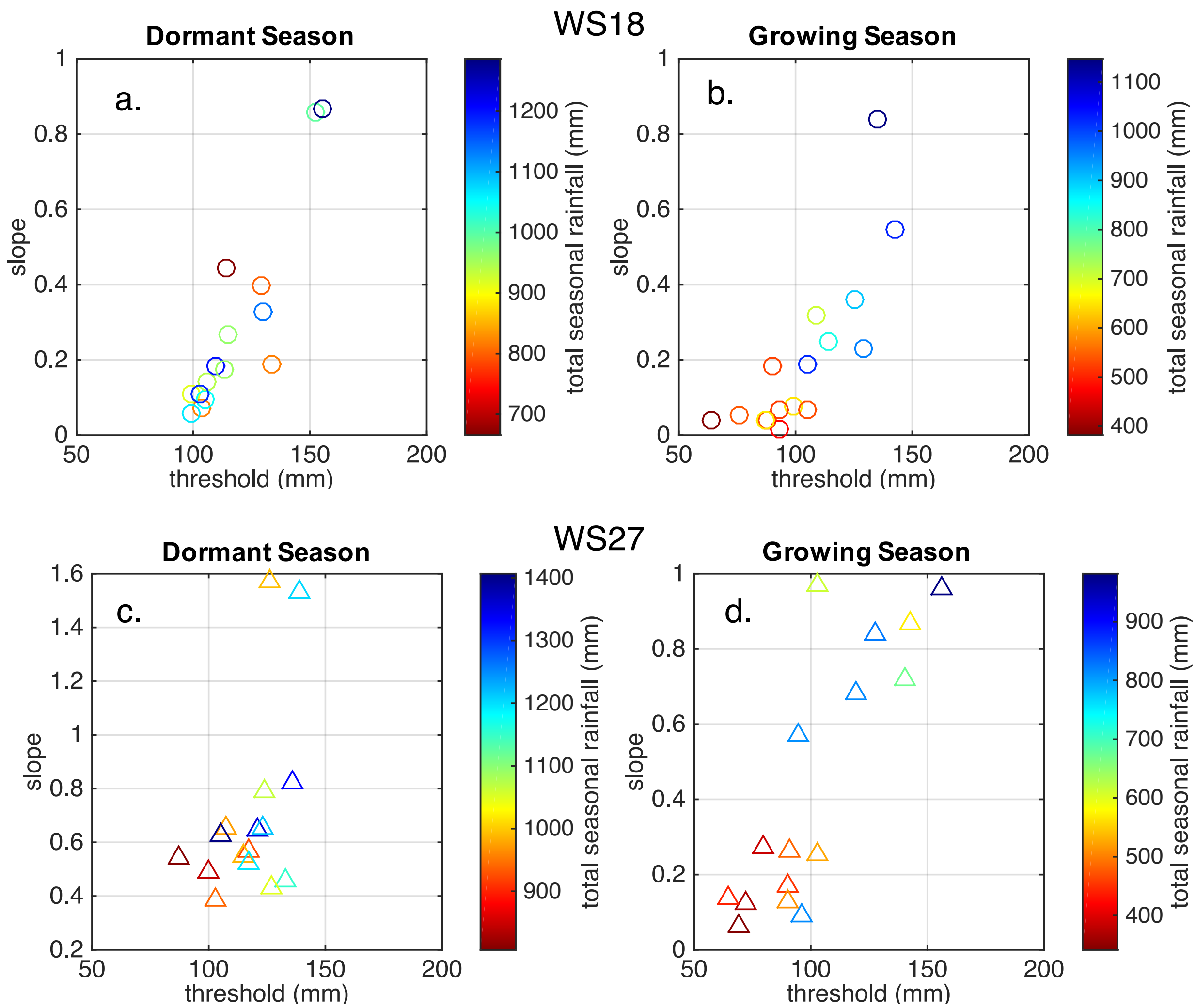


**Figure 4.8.** Interannual slope calculations of storms above the threshold identified for long-term dormant and growing season response in WS18 (a, b) and WS27 (c, d) between 1999-2014. Each point represents a single year and colors correspond to total seasonal rainfall.

# 5 DISCUSSION

## 5.1 Interannual variability in threshold dynamics

As in prior studies, our 15 year analysis of precipitation, soil moisture, and discharge data show that threshold relations are a function of antecedent soil moisture and total storm event precipitation at Coweeta, however we also found that thresholds vary with seasonal rainfall totals. When Detty and McGuire (2010) combined precipitation with antecedent soil moisture conditions to isolate threshold behavior during a single wet up period from August to January, they found stronger threshold definition than our long-term results, but were unable to look at multi-year relationships. Other long-term studies by Graham and McDonnell (2010) showed threshold response from 50 years of rainfall-runoff data was a result of pre-event wetness, however their study focused on identifying a single long-term threshold and did not directly account for interannual variability.

Our analysis suggests that threshold behavior persists both seasonally and interannually but that thresholds in rainfall-runoff relationships shift over longer timescales with climate. Interannual shifts were observed in WS18 and WS27 throughout the dormant and growing season, but were strongly correlated with seasonal rainfall during the growing season where low thresholds corresponded to dry growing seasons. WS27 also exhibited a weak but significant correlation during the dormant season, which may be due to non-linear response of total stormflow. Slight differences observed in threshold values between the LTER catchments (WS18/WS27) may be attributed to the spatial arrangement of soil moisture probes and differences in climate. Three soil moisture plots in WS18 are located along an ecological gradient from xeric to wet sites across the entire catchment, whereas WS27 has only a mesic and wet site.

Our interannual analysis also showed a greater response of stormflow with larger threshold values that covaried with climate during the growing season (Figure 4.6). During dry conditions, stormflow response above the threshold was relatively insensitive to changes in antecedent moisture and gross precipitation compared to wet periods. One hypothesis, is that dominant runoff generation mechanisms shift between seasons due to differences in the types of rainstorms. Our method does not allow us to explicitly distinguish between runoff mechanisms, but dormant period stormflow generation could reflect expanding variable source area, whereas during the growing season there is greater activation of the macropore network in accordance with seasonal storm intensity. Convective rain storms with high intensity precipitation are frequent in the summer months and transition to low intensity frontal storms in winter.

### 5.2 Drivers of threshold variation

The role of climate in structuring non-linear behavior supports previous short-term findings of stormflow response. However, our results suggest long-term climate does not produce a single, uniquely defined rainfall-runoff relationship in Coweeta, but instead generates year-to-year variation in non-linear stormflow response for the same catchment. Variation in thresholds can be explained during the growing season but appear more random in the dormant season. Growing season variation supports hypotheses set forth by Graham and McDonnell (2010). Their conceptual model highlights saturation deficit as the primary control over rainfall-runoff relationships at the catchment-scale. Saturation deficit is controlled at the catchment scale by two factors: (1) climate as it relates to interstorm length and PET, and (2) bedrock permeability. Longer drainage times between storms and greater PET increase thresholds, while low bedrock permeability decreases thresholds. Throughout Coweeta, bedrock permeability is low (Hewlett, 1961) and thus we assume it does not vary significantly interannually, leaving climate as the primary driver under this framework. The absence of fracture flow at Coweeta can be replaced by other mechanisms that contribute to both slow and fast runoff processes discussed in the next section.

Controlling for saturation deficit with ASI, our results show that shifting interannual thresholds may be in response to altered transpiration capacity of vegetation. Transpiration capacity is the total amount of water that can be transpired given the current state of the ecosystem. Recall that transpiration is responsible for over half of the annual water budget in Coweeta and removes a significant amount of water from the shallow subsurface – an important runoff generation zone in Coweeta (Hewlett and Hibbert, 1967). Small changes in transpiration capacity at the catchment scale may profoundly alter the amount of stormflow generation measured at the outlet. Figure 4.6 showed that thresholds have significant positive correlations with seasonal rainfall amounts only in the growing season. Because these correlations exist during the growing season when vegetation is most active, it suggests that the ecosystem capacity to transpire water may contribute to structuring rainfall-runoff relationships. In particular, response and recovery of ecosystems to dry periods may reduce transpiration capacity resulting in increased stormflow with similar antecedent conditions and gross precipitation (Figure 5.1). If patterns observed in Figure 4.6 were solely an effect of our threshold estimation method, we would also expect similar strength relationships during the dormant period.

Previous work has demonstrated that strategies for water regulation by various tree species can be impaired by droughts (Grier and Running, 1977; Maseda and Fernandez, 2006;

Bucci et al., 2005; Zweiniecki and Holbrook, 2009, Ford et al., 2010) modifying transpiration capacity across the catchment, which may account for positive trends in threshold values and seasonal rainfall observed in Figure 4.6. Recovery from survival mechanisms such as decreased leaf area (Grier and Running, 1977), early leaf senescence, abscission of fine roots (Maseda and Fernandez, 2006; Bucci et al., 2005), or in extreme cases emboli formation in xylem (Tyree and Sperry, 1989) can reduce transpiration rates even after drought subsides. In places like Coweeta, transpiration accounts for a large portion of rainfall (Swift et al., 1988), so reductions in transpiration capacity can significantly alter runoff ratios. A conceptual representation of this effect is presented in Figure 5.1. Under various drought response mechanisms given identical antecedent soil moisture and rainfall depth, the amount of stormflow generated from an event would be relatively greater in ecosystems still recovering from dry periods than in those unaffected by a drought. Figure 5.1 shows this hypothetical increase in stormflow production under these reduced rates of transpiration capacity.

Ford et al. (2010) also showed that differences in xylem anatomy were responsible for how tree species responded to drought in Coweeta. We hypothesized that differences between ring-porous and diffuse-porous species would result in significantly different threshold responses with seasonal rainfall between WS18 and WS27 during the growing season. WS18 is dominated by ring-porous and semi-ring porous species, like oaks and hickories, which were observed to be less sensitive to changes in vapor pressure deficit, a short-term response, and appeared to recover from droughts more slowly, a long-term response (Ford et al., 2010). Slow recovery following dry periods in oak-hickory forests represent a loss of stem conductance and long-term impairment of transpiration capacity. As a result, the transpiration capacity of these forests experience persistent reductions that suppress thresholds to a greater degree than in diffuse-porous dominated forests. WS27, a northern hardwood forest with diffuse-porous species, preserves stem conductance during drought through stomatal closure, but severe drought can lead to early leaf senescence and root abscission. WS27 could feasibly recover more quickly with the following growing season resulting in more stable transpiration capacities and less sensitivity of thresholds to seasonal rainfall. Patterns observed in Figure 4.6b do not show significant differences in their respective threshold sensitivity to seasonal rainfall. The lack of significance could result from several simultaneously occurring drought responses that are difficult to observe yet confound our results, including changes to leaf area, root distribution, and partial cavitation.

Vegetation-climate interactions span several timescales and thus we must consider both short- and long-term controls this relationship exerts on runoff production. Variability in thresholds may be a result of differences in response mechanisms to dry periods and the length of time required to recover. Roughly half of all rainfall at Coweeta has the potential to be evapotranspired, most of which occurs during the growing season. This underscores the importance of transpiration capacity, both seasonally and interannually, on stormflow production particularly during and after dry periods. Our study period spanned two severe, multi-year droughts, that may have caused long-term reductions in transpiration capacity due to severe water stress response. Time-lagged recovery can create greater runoff response in wetter years following a drought.

Response and recovery rates of ecosystems to drought is an actively researched area, so little is known about the degree to which drought affects discharge at the watershed scale. It is generally understood that during drought lower stomatal conductances as a result of water stress decreases transpiration slowing growth and recovery. The timing of rainfall even within a single growing season can cause a reduction in transpiration capacity lasting later into the season. Ford et al. (2010) observed that drought during the onset of leaf-out disproportionally affected oaks and hickories reducing transpiration even after the drought subsided. The effects of rainfall timing on transpiration capacity must be considered both seasonally and interannually, especially in mixed hardwood forests of the Southern Appalachians.

Other differences in catchment properties between WS18 and WS27, suggest an opposite correlation between thresholds and seasonal rainfall than what we observed. Referring to Graham and McDonnell's (2010) conceptual model for threshold behavior, variations in rainfall-runoff relationships are a function of climatic and geologic differences. WS27 is a high elevation catchment that receives greater rainfall, but has a growing season shortened by several weeks. Also, the Nantahala Escarpment, which traverses the Coweeta border, creates rocky outcrops and colluvial soils creating lower subsurface storage and higher runoff ratios compared to WS18 (Wooten et al., 2008; Hales et al., 2009). We would expect this to generate smaller thresholds in response to greater rainfall, which begs the question of which mediating factor is dominant, vegetation response to climate or catchment properties.

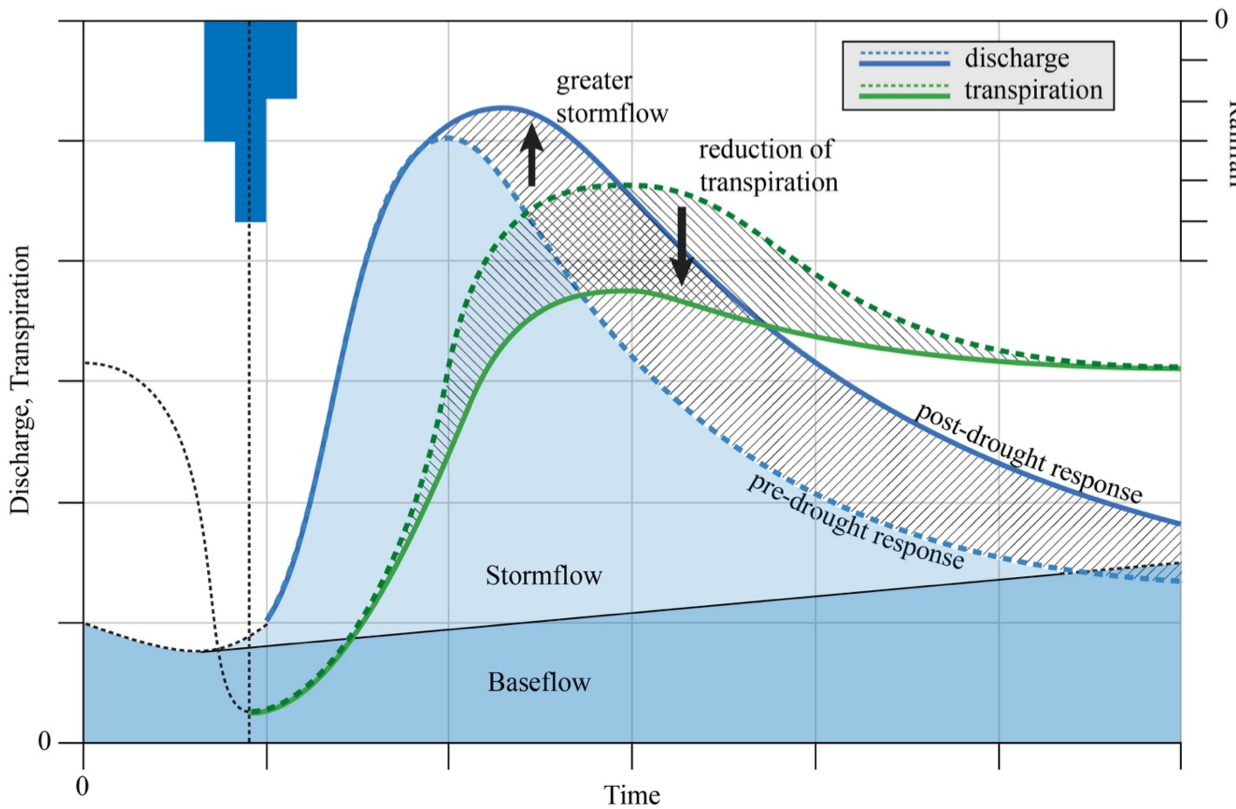


**Figure 5.1** Conceptual representation of stormflow response in an ecosystem that is healthy (dashed lines) and in the same ecosystem following a drought (solid lines) assuming identical antecedent soil moisture conditions and gross precipitation. Units are unspecified as stormflow response and time may vary with ecosystem, storm size, and drought magnitude.

## 5.3 Non-linearity and Runoff Generation Mechanisms

From our analysis, interannual variation of rainfall-runoff relationships manifests as simultaneous shifts in threshold and slope. Concomitant increases in threshold and steepening of slopes suggests that these relationships do not exhibit strict threshold behavior. Over several years or more, stormflow response to antecedent soil moisture and total precipitation forms a continuous, non-linear function. Subsetting the long-term data to specific years or seasons reveals well-defined thresholds, but even then these thresholds are not constant. Another indication of non-linear stormflow response is the steepening of slopes in years with large storms or during the growing season when convective storms are common. If simple threshold behavior was uniquely defined, we would expect the slopes above the threshold to be similar, independent of storm size and seasonal wetness. Also, if positive correlations of slope and threshold were solely an effect from our method being sensitive to years with high rainfall, then we would observe this pattern in the dormant season, as well. While dormant season demonstrates a correlation between threshold and slope like the growing season, the effect of total seasonal rainfall is more random in WS18 and to some degree in WS27 (Figure 4.8).

Mechanisms for stormflow response are difficult to isolate solely from rainfall, soil moisture, and discharge, but previous studies suggest several interacting subsurface flow controls on hillslope storm response that scale non-linearly to the catchment. Overland flow is uncommon in deep forested soils and diffuse porous media flow is typically thought of as being too slow to produce rapid stormflow response. Tritium experiments at Coweeta have suggested subsurface pressure waves in the unsaturated zone that displace hillslope soil water films during a rain event (Horton and Hawkins, 1965; Hewlett and Hibbert, 1967). Hewlett and Nutter (1970) also describe differences in seasonal storm hydrographs as a result of variable source area dynamics. Greater winter wetness causes a dual peak hydrograph caused by initial rain falling on saturated areas near the stream that contribute directly to flow, while the second peak is driven by slower subsurface processes. More recently, studies have explicitly focused on hillslope connectivity and macropore flow networks as drivers of rapid stormflow response (Band et al., 2014). Yeakley et al. (1998) showed increased connectivity and subsurface drainage response to storm events when soil moisture was high prior to rainfall. This pattern was disrupted by interstorm transpiration, which highlights the importance of throughflow contribution to streamflow, especially in the root zone. In Coweeta, subsurface flow is believed to be responsible for stormflow (Hewlett and Hibbert, 1967), making soil water, and subsequently stormflow, in the shallow root zone susceptible to evapotranspiration. At the PMRW, significant trench flow was measured showing greater macropore flow with higher antecedent wetness (Tromp van Meerveld and McDonnell, 2006a,b). Additionally, stormflow continues after rainfall has ceased. During this recessional period, transpiration may be most active yielding strong seasonal differences in the receding limbs of dormant versus growing season hydrographs. Data from our 15 year analysis lacks small enough time steps to resolve differences in seasonal storm hydrographs. In WS14, 15-min interval soil moisture and discharge data can demonstrate differences in seasonal storm hydrographs but this analysis is beyond the scope of this paper. These runoff generation processes combine and scale across the entire catchment to form repeatable emergent patterns of rainfall and runoff that are affected by catchment properties and ecosystem response to climate but do not clearly exhibit simple threshold response.

Non-linear behavior observed throughout our study watersheds suggests similar scaling of hillslope processes to catchment discharge as with other studies. Our method for determining thresholds allowed non-zero slopes to be calculated for storms occurring below the thresholds. Low slopes prior to the threshold, as opposed to zero slopes, represent gradual increases in stormflow response with increasing precipitation and antecedent moisture. These low slopes were demonstrated in our short-term analysis of WS14 (Figure 4.2) and persisted in our interannual analyses of WS18 and WS27. Representing low flow responses is especially important for storms that occur after relatively long interstorm periods with less total rainfall. These smaller storms occurring on drier hillslopes are likely activating slower processes with smaller variable source areas, generating modest stormflow. Slowly accelerating stormflow response with greater antecedent soil moisture and precipitation below the threshold may indicate that variable source areas are expanding. Subsurface stormflow via preferential flowpath activation, pore pressure waves or flow at the soil-bedrock interface produce faster and greater stormflow responses but require wetter antecedent conditions. Our study design cannot directly discriminate between these processes but our long-term data indicate a complex integration of runoff generation mechanisms that create accelerating stormflow response with increased antecedent soil moisture and precipitation.

The threshold behavior observed in WS14 and within each individual year of WS18 and WS27 indicate a switching between slow and fast dominant runoff generation mechanisms. Trench experiments have not been conducted in Coweeta, but highly transmissive and deep, forested soils may suggest faster macropore flow under wetter antecedent soil moisture and potentially slower displacement of hillslope water through film flow (Beven and Germann, 1981; 2013) when the hillslope is drier. These runoff mechanisms may be responsible for clear breaks in slope observed. Extensive trench work from PMRW shows that high antecedent soil moisture produces significant matrix and pore flow under large storm events creating strong thresholds in stormflow response (Tromp-van Meerveld and McDonnell, 2006a). However, the transition from low slope to high slope, as observed during the growing season in WS14 (Figure 4.2), was not observed as strongly in the long-term data due to interannual shifts in thresholds and slopes. This was not the case when the data was truncated to shorter time periods (Figure 4.4). Long measurement periods integrate processes over variable climate and vegetation conditions producing a non-linear stormflow response. While antecedent conditions have been shown to control stormflow generation, the degree to which this control is stable through time varies. It is possible that changes in preferential flowpaths, vegetation response to drought, and climate alters threshold behavior over long time periods. However, further research is required to explore their direct linkages to long-term stormflow characteristics and how these processes scale from hillslopes to the catchment.

# 6 CONCLUSIONS

In this study, we showed that threshold behavior in stormflow response could be identified in Coweeta by characterizing hillslope wetness as the sum of total storm precipitation with antecedent soil moisture. The Southern Appalachian Region has experienced greater hydroclimate variability, so non-stationarity in these emergent behaviors was also considered using a 15 year dataset of soil moisture, precipitation, and discharge from two watersheds. We also investigated the critical role of vegetation dynamics in threshold relationships by considering them with respect to dormant and growing season. Transpiration relies on root-zone soil moisture in the shallow subsurface, which is also an important zone for runoff generation in Coweeta.

We tested four hypotheses that stated that (1) watersheds in Coweeta will display well-defined threshold behavior at short timescales but that this varied with long-term data, (2) thresholds would vary interannually with seasonal rainfall, (3) threshold response to seasonal rainfall would vary by dominant forest type and (4) vegetation-climate interactions would suggest non-stationarity and non-linear stormflow response over long measurement periods.

From the four posed hypotheses we concluded the following:

(1) Threshold behavior was not uniquely defined using 15 years of soil moisture, rainfall, and discharge measurements and varied seasonally and interannually.

Seasonal variation between growing and dormant season produced lower thresholds in the growing season suggesting the importance of vegetation in structuring stormflow response.

(2) Interannual variation in thresholds is a function of hydroclimate potentially associated with shifting ecosystem transpiration capacity in the growing season. Transpiration capacity is the amount of water that can be transpired by an ecosystem given its current state. Drought-driven impairment and delayed recovery of transpiration capacity may explain smaller thresholds in drier years. Future studies will examine how changes in transpiration capacity alter storm hydrographs at shorter timescales using 15-min interval data.

(3) Threshold sensitivity to seasonal rainfall did not show significant differences between catchments dominated by ring-porous versus diffuse-porous species. This was likely an effect of several confounding drought-response mechanisms influencing transpiration capacity.

(4) Threshold values were also positively correlated with slope computed for storms above the threshold suggesting non-linear stormflow response rather than a simple well-defined threshold.

This study highlights the importance of considering emergent behavior over long measurement periods and argues the need to characterize variation in emergent patterns. Future work should also consider the the use of long-term data in determining the effects vegetation may have on thresholds. It is important, as we consider emergent properties in other catchments and with non-stationarity of hydroclimate, that we investigate how vegetation and climate structure both short- and long-term rainfall-storage-runoff patterns more thoroughly.

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