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Key Points:

- No consistent patterns in soil moisture responses among landscape positions laterally along hillslopes
- Dominant controls on soil moisture responses to storms varied with individual response metrics and during storms
- Storm depth, mean intensity, and antecedent conditions mediated the soil moisture-runoff relationships in space and time

Supporting Information:

Supporting Information may be found in the online version of this article.

Correspondence to:

N. K. Singh, nksingh2@ncsu.edu

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Soil Moisture Responses to Rainfall: Implications for Runoff Generation

Nitin K. Singh^{1,2}, Ryan E. Emanuel², Brian L. McGlynn³, and Chelcy F. Miniat⁴

¹Earth and Environmental Science, University of Waterloo, Waterloo, ON, Canada, ²Forestry and Environmental Resources, North Carolina State University, Raleigh, NC, USA, ³Division of Earth and Ocean Science, Nicholas School of Environment, Duke University, Durham, NC, USA, ⁴USDA Forest Service, Southern Research Station, Coweeta Hydrologic Laboratory, Otto, NC, USA

Abstract Soil moisture is a key control on runoff generation and biogeochemical processes on hillslopes. Precipitation events can evoke different soil moisture responses with depth through the soil profile, and responses can differ among landscape positions along a hillslope. We sought to elucidate the nature of these responses by estimating changes in water content, response time between peak precipitation and peak soil moisture, and wetting front velocities for 43 storms at 45 locations on three adjacent hillslopes within a headwater catchment of the southern Appalachian Mountains (NC, USA). We used a multivariate modeling approach to quantify the relative influences and the predictability of soil moisture responses by a combination of landscape and storm characteristics. We quantified the lag correlations between hillslope mean soil moisture and catchment runoff to demonstrate how storm properties and hillslope-scale characteristics may influence runoff at the catchment outlet. Soil moisture responses varied widely, and no consistent patterns were observed among response metrics laterally or vertically along hillslopes. In contrast to other studies, we found that the relative influence of hillslope properties and storm characteristics varied with soil moisture responses and during storms. Antecedent conditions and storm depths influenced the strength of lag correlations between soil moisture and runoff, whereas storm mean intensity was correlated with the lag times. These results highlight the utility of intensive observations for characterizing heterogeneity in soil moisture responses, suggesting, among other things, a need for better representation of the subsurface processes in rainfall-runoff models. Identifying the relative importance of drivers can be beneficial in building parsimonious hydrological models.

1. Introduction

Soil moisture influences an array of hydrological, ecological, and biogeochemical processes. These processes include runoff generation (Dunne & Black, 1970; R. B. Grayson et al., 1997; Harr, 1977; Helvey & Hewlett, 1962; Kaiser & McGlynn, 2018; Scaife & Band, 2017; Scaife et al., 2020; Western et al., 1999), hydrologic connectivity (Ali & Roy, 2010; Detty & McGuire, 2010; Western et al., 2001), and plant-water relations (Emanuel et al., 2010; Porporato et al., 2004; Rice & Emanuel, 2019; Tromp-van Meerveld & McDonnell, 2006). Soil moisture also influences carbon cycling in headwater catchments (Emanuel et al., 2011; Hawkes et al., 2017; Kaiser et al., 2018; Riveros-Iregui & McGlynn, 2009). In forested catchments with relatively deep soils, soil moisture sustains baseflow in headwater streams between storms and during dry times of the year (Hewlett, 1961; Nippgen et al., 2016; Singh et al., 2018). There is a need for continued research on soil moisture dynamics in the context of catchment hydrology, given the influence of soil moisture on various catchment processes (cf., Vereecken et al., 2008).

Links between climate and soil moisture are well studied, including the influence of storms on point-scale soil moisture distributions and vertical water fluxes (Rodriguez-Iturbe, 2000). However, headwater catchments and the hillslopes they comprise are complicated topological networks of hydrologic sources and sinks (Emanuel et al., 2010; Singh et al., 2016). The spatial heterogeneity of soil moisture is influenced not only by storms and vertical fluxes but also by interactions of storms with factors such as the lateral redistribution and preferential flow of soil water along hillslopes. Relatively few studies use spatially intensive observations (e.g., multiple depths and landscape positions) to characterize the soil moisture response to rainfall along hillslopes. Examples of existing studies include Yeakley et al. (1998), who used spatially intensive soil moisture measurements to observe the emergence of topographic gradients after events, analyzing changes in volumetric water content laterally at multiple depths along a forested hillslope. Kim (2009) used spatially intensive measurements across a dozen of landscape positions to identify distinct hillslope zones based on seasonal variations of soil moisture responses (i.e., volumetric water content and response time) to rainfall. Detty and McGuire (2010) monitored the timing of initial soil moisture responses for four landscape positions and categorized the responses into common landforms found in headwater catchments. Tymchak and Torres (2007) studied wetting front velocities at three landscape positions along a sandy forested hillslope.

Most of the existing studies focused on soil moisture patterns with some lateral or vertical resolution within a single hillslope. Further, our understanding of soil moisture dynamics at event scales relies on the studies that used one or the other aspects of soil moisture (i.e., changes in volumetric water content, response time, or wetting front velocity). Therefore, there is a need for more nuanced and comprehensive understanding of soil moisture dynamics based on multiple aspects of soil moisture response at high spatial resolution laterally and vertically along hillslopes. This gained knowledge will assist in elucidating the processes driving subsurface connectivity and runoff generation processes in general.

Soil moisture responses to rainfall are often attributed to storm properties, topography, soil properties, and vegetation (Rosenbaum et al., 2012). Storm depths and intensities are often reported as key drivers of soil moisture dynamics (Demand et al., 2019; Famiglietti et al., 1998; Graham & Lin, 2011; Wiekenkamp et al., 2016; D. J. Wilson et al., 2004; Tymchak & Torres, 2007), and some studies attribute variability in soil moisture dynamics to topography (Burt & Butcher, 1985; Detty & McGuire, 2010; Kim, 2009). In headwater catchments, topographic characteristics (e.g., slope, topographic wetness index, and plan curvature) may influence wetness state (Moore et al., 1991; Western et al., 1999) and the inception of lateral hillslope flow (cf., Anderson & Burt, 1978; Weiler et al., 2005). The influence of soil characteristics (e.g., soil depth, bulk density, and structure) on changes in volumetric water content and wetting front velocity is also well known (Mosley, 1982; Sidle et al., 2001). Generally, soil characteristics can affect flow paths, hydraulic properties, and preferential flows within pedons and at hillslope scales (Alaoui et al., 2011; Blume et al., 2009; K. Price et al., 2010; Scherrer, 1996; Weiler et al., 1998). Vegetation can also influence changes in volumetric water content via water uptake and throughfall. Root networks may further influence wetting fronts and soil moisture response times by triggering preferential flows (Sidle et al., 2001). Although past work has elucidated individual drivers of soil moisture responses to storms, the hierarchical influence of these drivers on soil moisture dynamics remains unknown.

Presently unaddressed issues include whether climatic and landscape drivers exert different controls on different aspects of soil moisture response (e.g., response time, wetting velocity), and the extent to which soil moisture responses to rainfall can be predicted by common drivers. Addressing these issues may improve our conceptual understanding of hillslope hydrology, subsurface processes, and to inform the development of parsimonious, processed-based hydrological models (Grayson & Blöschl, 2001; Merz & Plate, 1997; Scherrer & Naef, 2003; Uchida et al., 2005).

We investigated soil moisture responses at multiple landscape positions and depths for three adjacent hillslopes in a gauged headwater catchment at the Coweeta Hydrologic Laboratory (hereafter Coweeta) located in North Carolina, USA. We monitored soil moisture for more than 2 years using arrays of probes installed in soil pits. The experimental design allowed us to evaluate the influence of storm characteristics and landscape properties on soil moisture responses. Stream gauging data from the catchment outlet allowed us to evaluate relationships between soil moisture and runoff. We used Multivariate Adaptive Regression Splines (MARS) modeling to determine the hierarchy of storm and landscape characteristics that influence soil moisture responses. Lastly, lag-correlation analysis allowed us to study relationships between soil moisture and runoff responses and controls of these correlations. Our work addresses the following specific questions for Coweeta and similar forested headwater catchments. (a) How do soil moisture responses to storms vary spatially and temporally? (b) What are the relative controls on these soil moisture responses? (c) What factors influence relationships between runoff and soil moisture responses?





Figure 1. Delineated hillslopes (H1–H3) with landscape positions of soil moisture probes, rain gauge, and weir in WS02.

2. Study Site

We conducted fieldwork at Coweeta, a US Forest Service research facility located in the Nantahala National Forest of western North Carolina, USA (35°03'N, 83°25'W). The climate is classified as maritime and humid temperate with cool summers, mild winters, and frequent short-duration rainfall events distributed year-round. The mean annual rainfall varies between approximately 1,700 mm at low elevations and 2,500 mm at high elevations, and the mean annual temperature is about 12°C. A standard estimate of the growing season for this region is April 15 through October 14.

This study focuses on a small (12 ha) headwater catchment at Coweeta known hereafter as WS02. Vegetation within WS02 is almost entirely broadleaf deciduous forest characteristic of secondary succession in the southern Appalachian Mountains following logging and agricultural abandonment. The catchment aspect is predominantly south-facing with a mean slope angle of 28°. The catchment is underlain by Tallulah Falls formation originating from relatively low compositional maturity sedimentary protoliths (Hatcher, 1971; Velbel, 1985). The underlying bedrock geology is almost impermeable, and it is believed to be homogenous in composition throughout the catchment. Major rock forming minerals include Biotite, Garnet, Plagioclase, Allanite, Vermiculite, Kaolinite, and Gibbsite (J. R. Price et al., 2005; Velbel, 1985). There are no bedrock outcrops in the study hillslopes.

Our study focused on three hillslopes, H1–H3, which were chosen to represent the diversity of size, slope, and curvature among hillslopes at Coweeta (Figure 1 and Table S1). We define hillslopes as convergent but unchannelized terrain, synonymous with a zero-order basin. Two major soil types present in WS02 include Fannin (fine loamy, mica dominated), located near the ridges, and Cullasaja-Tuckasegee (fine loamy, oxidic), located closer to the stream channels (Thomas, 1996). The O, A, and BA horizons occupy depths up to a total of 30 cm. The B and BC horizons span depths between 30 and 100 cm, which is believed to be the maximum

rooting depth (Gaskin et al., 1989; Yeakley et al., 1998). Bulk densities measured across hillslope positions varied between 0.88 and 1.67 g/cm³ (Table S2), which agreed with prior research in the same catchment (Yeakley et al., 1998). Soil texture is 20%–35% clay and 45%–60% sand (Yeakley et al., 1998). The depth to bedrock ranges from approximately 1 to 2 m across monitoring locations. Singh et al. (2016) and Nippgen et al. (2016) provide additional details about soil, hydrological, and topographic characteristics of WS02.

3. Methods

3.1. Geospatial Analysis

Using bare Earth light detection and ranging (LIDAR) data coarsened from 1 to 5 m resolution, we estimated the drainage area for each pixel within the instrumented hillslopes following the multidirectional flow method of Seibert and McGlynn (2007) to estimate subsurface flow accumulation and to delineate the stream network using digital elevation data. We identified specific topographic variables of relevance to soil moisture based on past research that focuses on its variability at the catchment scale (Moore et al., 1991; Western et al., 1999). Variables computed from the 5×5 m DEM included slope (SLP), upslope accumulated area (UAA), topographic wetness index (TWI), plan curvature (PLC), and profile curvature (PRC). Supporting Information contain further information on these variables (Table S1).





Figure 2. Schematics of estimated response metrics during storms.

3.2. Field Measurements

We monitored soil moisture at 45 separate locations within WS02. The locations represent 15 different combinations of hillslope position and soil depth for three separate hillslopes (Figure 2). We instrumented four landscape positions for each hillslope: upslope (US), midslope (MS), lower slope (LS), and near stream (NS). It is worth acknowledging that the region near the ridge of the hillslope was not monitored in our study. In general, US and MS landscape positions were characterized by steeper, terrain than LS and NS landscape positions. At US and MS landscape positions, we excavated soil pits and installed volumetric water content probes at five depths within each pit: 10, 20, 30, 60, and 100 cm. In LS landscape positions, we excavated soil pits and installed probes at three depths within each pit: 20, 60, and 100 cm. In NS landscape positions, we excavated soil pits and installed probes (1–5) were assigned to each probe to indicate its relative depth

in the soil pit from 10 cm (1) to 100 cm (5). We recorded volumetric soil moisture via reflectometry probes (Model CS-650, Campbell Scientific, Logan, UT; HydraProbe, Stevens Water Monitoring Systems, Portland, OR) at 30-min intervals for 25 months, from October 2011 until December 2013 using a datalogger installed near the MS position of each hillslope. Probes and dataloggers were powered by rechargeable batteries and solar panels. Occasional probe malfunctions and battery charging issues resulted in minimal data loss, but we obtained high-quality soil moisture data for all storms of interest.

3.3. Data and Statistical Analyses

3.3.1. Response Metrics

We measured rainfall using a National Weather Service standard rain gauge (SRG17) in an open field located near WS02 (Miniat et al., 2017). We identified 43 separate storm events by inspecting rainfall data from SRG17 manually, using the same methods described in Singh et al. (2018). The following criteria were applied for event selection: (a) 30-min rainfall equaled or exceeded 0.5 mm, (b) total rainfall for the event exceeded 20 mm, (c) a minimum of 3 h separated events, and (d) hydrologic variables (i.e., soil moisture and runoff) were recorded immediately before, during, and for 3 days after each storm. Soils at Coweeta are highly conductive and have enormous storage capacity (Hewlett, 1961), so we focused on large storms that are likely to generate detectable response. We defined a minimum threshold for soil moisture response as a change of 0.01 m³ m⁻³ in 30 min. The response percentage was defined as the percentage of total storms that generated a detectable response for each soil moisture probe. We calculated three different storm response metrics for each soil moisture probe on a storm-by-storm basis (Figure 2).

The absolute change in volumetric water content (Δs , m³ m⁻³) was defined as the difference between the minimum and maximum soil moisture during each storm. The peak to peak time (T_{p2p} , h) was defined as the time difference between the time of maximum rainfall intensity at 30 min intervals and time of peak soil moisture response (Kim, 2009). Positive T_{p2p} values indicate that rainfall peaks earlier than soil moisture, and negative T_{p2p} values indicate that soil moisture peaks earlier than rainfall. The initial response time (T_i , h) was defined as the time between the beginning of a storm and a soil moisture response. We used T_i to compute the occurrence of preferential flow and wetting front velocity (V_{wf} , mm h⁻¹) for each probe and storm.

Prior studies propose using the sequential order of response time to detect preferential flow within the soil profile (Lin & Zhou, 2008; Wiekenkamp et al., 2016). As recommended, we estimated the occurrence of preferential flow by analyzing the sequential order of initial response timing (T_i, h) of probes with depth in each pit during storms. The occurrence of preferential flow was estimated in binary form 1 and 0 for each probe. If a probe at any greater depth (e.g., 20, 30, 60, and 100 cm) responded earlier than the shallower probe (i.e., 10 cm), we flagged the response as 1 otherwise 0 for that particular probe. The frequency of occurrence of preferential flow for each probe was defined as the ratio of storms that generated preferential flow to total storms that generated detectable response. At hillslope scale, the cumulative preferential flow



was estimated by aggregating all occurrences of preferential flows (i.e., summing "1") across landscape positions and depths during each storm.

The mean wetting front velocity $(V_{\rm wf})$ from the ground surface to the monitoring depth of each probe was defined as

$$V_{\rm wf} = \frac{L}{T_i},\tag{1}$$

where *L* is the depth of the probe, and T_i is the initial response timing of the probe (Tymchak & Torres, 2007). Because wetting front velocity was derived from initial response time, there was a high correlation (-0.68) between them (Table S3). Thus, we focused our analysis on the three soil moisture responses, Δs , V_{wt} , and T_{p2p} , that were least correlated (Table S3). We explored these response metrics vertically with depth and laterally along hillslopes. The Wilcoxon Rank Sum Test was used to test whether the medians of response metrics generated during all storms were significantly different among landscape positions (e.g., US, MS, LS, or NS) along hillslopes. For instance, one of the null hypotheses tested was US and MS landscape positions have equal median wetting front velocity along H1 during all storms. We defined antecedent moisture conditions (AMCs) for each probe as the volumetric soil moisture 1 h prior to the arrival of a storm. Antecedent precipitation index (API₇) was estimated by calculating the running sum of precipitation for the preceding 7-day period. Boxplots were used to explore the distribution and median of the response metrics for each probe during the study period. In addition, the coefficient of variation (CV) of response metrics was also estimated to understand the overall temporal variability of individual probes across storms.

3.3.2. Multivariate Adaptive Regression Splines (MARS) Modeling

Multivariate Adaptive Regression Splines is a nonparametric regression method suitable for simulating nonlinear and complex interactions between response and associated drivers (Friedman, 1991). Despite its robustness and potential, MARS is rarely used in catchment hydrology (Lall, 1995), but it has been widely used to rank the importance of explanatory variables in other disciplines (e.g., Leathwick et al., 2006). Here, we used MARS to rank the relative influence of landscape and storm characteristics on soil moisture response. In brief, MARS divides the data sets into segments (i.e., splines), and each segment is modeled individually, resulting in piecewise curves referred to as basis functions (BFs; Zhang & Goh, 2016). Basis functions offer flexibility to simulate thresholds and interactions among variables and are combined together, as shown in Equation 2 (Deo et al., 2017; Zhang & Goh, 2016):

$$f(x) = \alpha + \sum_{p=1}^{K} \alpha_k BF(x), \qquad (2)$$

where f(x) is a linear combination of BFs, α and α_k are coefficients, and k is the number of segments. The relative importance of variables and their predictive power was quantified in two steps. First, we developed a MARS model to simulate individual soil moisture response metrics (Δs , V_{wp} and T_{p2p}) to all explanatory variables. Second, using this MARS model, we conducted a variable importance analysis to identify the three most important variables for each soil moisture response. Variable importance was determined based on two parameters: $N_{subsets}$ and the residual sum of squares (RSS). $N_{subsets}$ value, the more important the variable. The RSS, a parameter that tracks model error derived from observed and predicted response; the greater the decreases in RSS, the more important the variable. The MARS models were developed in R (RStudio Team, 2016) using the package "Earth" (Milborrow, 2019). To test the influence of landscape positions on response metrics, we used landscape positions as one of the explanatory variables in the MARS models. Finally, to understand the direction (increase/direct or decrease/inverse) of the relationship between the most important variables and response metrics, we calculated the Spearman's correlation coefficient between each response metric and the respective key drivers.

3.3.3. Lag Correlations

We computed lag-correlation coefficients (Spearman's ρ) and the corresponding lags between streamflow at the catchment outlet and hillslope-scale mean soil moisture for ±5 h at the event scale. The length of time series used in the lag-correlation analysis started with the inception of a storm and ended 24 h after the storm. Discharge at the outlet of WS02 was measured by the US Forest Service using a 90° v-notch weir and stage recorder as part of their long-term streamflow monitoring efforts at Coweeta. These correlations

Table 1

Select Characteristics of	² 43 Storm Events
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Event	Arrival date	Depth (mm)	Period (h)	Average intensity (mm h ⁻¹)	Peak intensity (mm 30- min ⁻¹)
1	November 3, 2011	22.2	4.0	5.5	5.9
2	November 15, 2011	30.0	16.0	1.8	6.3
3	November 22, 2011	28.5	2.5	11.4	14.2
4	November 27, 2011	152.0	22.5	6.7	5.8
5	December 22, 2011	69.3	9.0	7.7	13.8
6	December 27, 2011	29.1	6.0	4.8	4.5
7	January 10, 2012	43.1	14.5	2.9	6.6
8	January 17, 2012	34.8	6.5	5.3	9.1
9	January 22, 2012	36.3	12.5	2.9	4.3
10	January 26, 2012	37.4	11.5	3.2	6.2
11	March 2, 2012	32.2	5.5	5.8	6.4
12	April 5, 2012	45.1	5.0	9.0	15.9
13	April 17, 2012	64.8	16.5	3.9	7.9
14	May 6, 2012	50.6	3.0	16.8	28.8
15	May 13, 2012	53.1	27.0	1.9	7.2
16	May 31, 2012	28.9	2.0	14.4	12.8
17	July 1, 2012	35.5	5.5	6.4	13.8
18	July 14, 2012	64.0	5.0	12.8	27.2
19	August 9, 2012	30.7	2.0	15.3	15.3
20	August 17, 2012	108.3	16.5	6.5	13.1
21	August 30, 2012	156.1	29.5	5.2	14.7
22	December 10, 2012	26.1	8.5	3.0	3.1
23	December 20, 2012	35.6	6.5	5.4	4.5
24	December 24, 2012	25.3	8.0	3.1	3.0
25	December 25, 2012	52.1	11.5	4.5	5.9
26	January 14, 2013	32.8	14.0	2.3	4.0
27	January 17, 2013	71.7	17.5	4.1	7.3
28	January 29, 2013	179.8	26.5	6.7	21.1
29	February 21, 2013	32.7	8.5	3.8	5.9
30	February 26, 2013	56.5	12.5	4.5	5.9
31	March 5, 2013	26.9	7.0	3.8	10.1
32	March 11, 2013	42.3	9.5	4.4	7.5
33	April 11, 2013	86.2	9.0	9.5	28.8
34	April 19, 2013	37.8	7.0	5.4	7.5
35	April 28, 2013	60.6	18.0	3.3	4.6
36	June 5, 2013	45.2	4.0	11.3	27.5
37	July 2, 2013	56.1	4.0	14.0	25.7
38	July 3, 2013	83.0	16.0	5.1	13.6
39	July 5, 2013	71.1	12.0	5.9	22.6
40	August 18, 2013	25.7	9.0	2.8	3.7

and corresponding lags revealed the linkages between hillslope-scale soil moisture and streamflow generation processes during storms. The lag-correlation coefficients determined the strength of the relationship between two time series, and the lag revealed the time difference between peak runoff and mean soil moisture peak during storms for each hillslope. Prior to estimating correlations, we detrended both time series to remove serial autocorrelations (Chatfield, 2004; Singh & Borrok, 2019). The time series was detrended via first-order differencing where the difference was taken between two consecutive observations. Negative lag times indicated that soil moisture peaked earlier than runoff at the catchment outlet, whereas zero lags indicated both soil moisture and runoff peaked during the same 30-min period. Later, we extracted the maximum correlation ($\rho_{\rm max}$) and the corresponding lags (here onward, $l_{\rm sr}$) between runoff and soil moisture for each event and hillslope. We explored $\rho_{\rm max}$ and l_{sr} and their relationships with the storm properties and hillslopescale soil moisture responses (e.g., AMC, wetting front velocity, and the cumulative preferential flows) via correlation analysis. Like many other studies (Brocca et al., 2010; Famiglietti et al., 2008; Penna et al., 2009; Tague et al., 2010), we used arithmetic means of soil moisture at hillslope scale to understand soil moisture-runoff relationships. However, means may not be true representation if the distribution is skewed and results should be interpreted with caution.

4. Results

4.1. Characteristics of Rainfall and Soil Moisture

The 43 storms that we identified are generally representative of the wide range of rainfall characteristics experienced in the southern Appalachian Mountains (Table 1). The study catchment received a total of 4,775 mm of rainfall from October 2011 to December 2013, corresponding to an average of 2,387 mm year⁻¹. The study period included one of the wettest years (2013) on record at Coweeta, in which 620 mm of rainfall fell between January and February 2013. The median 30 min rainfall for the entire study period was 1.6 mm, excluding periods with no rain. The 43 storms were distributed throughout two dormant seasons (26 events) and two growing seasons (17 events). Generally, the longer the storm event, the greater the storm total ($\rho = 0.55$, p < 0.001), and storms with high peak intensities were also more intense on average ($\rho = 0.76$, p < 0.05).

Temporal variability in soil moisture varied with soil depth and slope steepness for individual hillslopes (Table 2). Temporal variability in soil moisture generally declined with depth at all landscape positions in H1, but not in H2 or H3. Similarly, the temporal variability was higher for steeper landscape positions (i.e., US or MS) than corresponding flatter landscape positions (i.e., LS or NS). Finally, we found no gradient in mean soil moisture along hillslopes. In other words, soil moisture did not systematically increase or decrease moving from uphill to downhill locations.

4.2. Soil Moisture Responses

In the following subsections, we highlight key results associated with each of the three soil moisture response metrics (Δs , V_{wp} , T_{p2p}) that we

Table 1 Continu	ued				
Event	Arrival date	Depth (mm)	Period (h)	Average intensity (mm h ⁻¹)	Peak intensity (mm 30- min ⁻¹)
41	September 25, 2013	43.2	8.5	5.0	7.4
42	October 31, 2013	25.2	12.5	2.0	3.3
43	November 26, 2013	94.1	18.5	5.0	7.4

examined at Coweeta. These results are also summarized in Tables 3 and 4 and in Figures 3–6.

4.2.1. Response Percentage

In general, the response percentage (i.e., the fraction of storms that generated a detectable response) to storms varied with depth and landscape positions along hillslopes (Table 3). The response percentage was greatest at shallow depths (10–30 cm) and steep landscape positions (US, MS). For H1, the response percentage declined with depth for most landscape positions (Table 3). We observed no such systematic patterns in response percentage with depth for H2 and H3. The initial response timing (T_i) did

not increase linearly with depth. These findings may indicate the occurrence of preferential flow, in which deeper soil layers responded more rapidly to a storm than shallower layers (Figure 3).

The frequency of preferential flow varied among landscape positions and hillslopes (Table 3 and Figure 3). For H1, preferential flow occurred frequently at 30 cm depths for US and MS landscape positions. For H2 and H3, preferential flow occurred during most of the storms at all landscape positions and at 20, 30, and 100 cm depths. In particular, frequent occurrence (>5 times) of preferential flow was estimated at US (30, 60 cm), MS (30 cm), LS (100 cm), and NS (100 cm) area. For H1, MS (30 cm) and LS (100 cm) positions exhibited frequent occurrence (>5 times) of preferential flow (Figure 3). A significant but relatively weak relationship exists between storm depth and preferential flow for both H1 ($\rho = 0.33$, p = 0.02) and H2 ($\rho = 0.30$, p = 0.04).

4.2.2. Response Magnitude (Δs)

The change in volumetric water content (Δs) varied between 0.02 and 0.34 m³ m⁻³ for all storms and landscape positions (Figure 4 and Table 4). The median Δs declined with depth in the soil, and it was significantly different among all landscape positions in H1 (Wilcoxon p < 0.05). For H2, median Δs differed significantly among landscape positions (Wilcoxon p < 0.05), whereas no significant difference was detected in median Δs among landscape positions along H3. The median Δs did not vary unidirectionally (i.e., did not increase or decrease) moving along a hillslope. Further, some landscape positions within H2 and H3 exhibited an abrupt and unexpected increase in Δs with depth (e.g., US3, MS3, and LS5; Figure 4).

Table 2
Mean Soil Moisture (Coefficient of Variation) During the Study Period
Across Three Hillslones (H1–H3)

Positions	Depth (cm)	$H1 (m^3 m^{-3})$	$H2 (m^3 m^{-3})$	${ m H3}({ m m^{3}}{ m m^{-3}})$
Upslope	10	0.34 (0.13)	0.25 (0.15)	0.23(0.33)
	20	0.30 (0.15)	0.40 (0.06)	0.25 (0.27)
	30	0.30 (0.13)	0.29 (0.11)	0.20 (0.31)
	60	0.38 (0.07)	0.32 (0.10)	0.29 (0.17)
	100	0.44 (0.03)	0.32 (0.06)	0.29 (0.17)
Midslope	10	0.30 (0.17)	0.45 (0.05)	0.27 (0.25)
	20	0.44 (0.02)	0.40 (0.07)	0.34 (0.08)
	30	0.35 (0.04)	0.30 (0.09)	0.29 (0.06)
	60	0.38 (0.02)	0.38 (0.05)	0.23 (0.07)
	100	0.46 (0.01)	0.42 (0.02)	0.40 (0.13)
Lower slope	20	0.38 (0.04)	0.34 (0.10)	0.40 (0.11)
	60	0.36 (0.09)	0.39 (0.05)	0.44 (0.09)
	100	0.39 (0.01)	0.31 (0.12)	0.28 (0.28)
Near stream	60	0.39 (0.02)	0.21 (0.08)	0.29 (0.18)
	100	0.46 (0.02)	0.24 (0.11)	0.31 (0.22)

Based on mean soil moisture of all landscape positions and depths during all storms, the driest hillslope, H3, exhibited greatest variability in Δs (CV = 0.71), whereas the least variability in Δs (CV = 0.67) corresponded to the wettest hillslope, H1 (Figure S1, Tables 2 and 4).

4.2.3. Response Time (T_{p2p})

The peak to peak time $(T_{p2p}, the time difference between the peak rain$ fall intensity and peak soil moisture response) ranged from -8 to 32 h. Landscape positions along the wettest hillslope H1 (mean $T_{p2p} = 2.7$ h) peaked earlier than landscape positions along the other two hillslopes (H2 mean $T_{p2p} = 3.5$ h; H3 mean $T_{p2p} = 3.1$ h). Negative values, which indicate that soil moisture reached its maximum value before peak rainfall intensity, were most common on H3. For H1, median T_{p2p} increased with depth and was significantly different (Wilcoxon p < 0.05) among most of the landscape positions (Figure 5 and Table 4). For H2, several landscape positions (e.g., US and MS) exhibited a sudden decrease in $T_{\rm p2p}$ with depths, and median $T_{\rm p2p}$ was significantly different among all landscape positions along the hillslope (Wilcoxon p < 0.05; Figure 5c). For H3, median T_{p2p} was significantly different among all landscape positions (Wilcoxon p < 0.05), except for the two steeper landscape positions, US and MS (Wilcoxon p > 0.05; Figure 4). For the two drier hillslopes (H2 and H3; Table 2), $T_{p_{2p}}$ times for NS landscape positions were significantly longer than times for US landscape positions (Wilcoxon p > 0.05), but the same did not hold for H1, the wettest hillslope.

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Table 3

Response Percentage (Occurrence of Preferential Flow) for Three Hillslopes (H1–H3) and Depths

Positions	Depth (cm)	H1	H2	Н3
Upslope	10	100 (NA)	100 (NA)	100 (NA)
	20	100 (12)	91 (23)	100 (0)
	30	97 (18)	100 (49)	90 (50)
	60	75 (3)	86 (14)	92 (8)
	100	3 (3)	60 (5)	62 (15)
Midslope	10	100 (NA)	95 (NA)	100 (NA)
	20	45 (0)	98 (7)	96 (4)
	30	79 (42)	100 (42)	100 (30)
	60	15(3)	75 (3)	85 (10)
	100	0 (0)	0(0)	40 (25)
Lower slope	20	94 (NA)	98 (NA)	89 (NA)
	60	100 (58)	43 (0)	65 (12)
	100	No data	67 (15)	55 (15)
Near stream	60	48 (NA)	74 (NA)	65 (NA)
	100	16 (0)	50 (17)	31 (23)

Note. "NA" indicates preferential flow could not be estimated in the absence of shallower layer.

4.2.4. Wetting Front Velocity (V_{wf})

Wetting front velocities (V_{wf} the rate at which the storm signal propagated vertically through soil profile) varied widely, ranging from 10 to 2,000 mm h⁻¹ (Figure 6 and Table 4). For H1, median V_{wf} differed significantly among landscape positions along the hillslope (Wilcoxon p < 0.05). Only the MS location in H1 exhibited an increase in V_{wf} with depth. For H2 and H3, median V_{wf} increased with depth for most of the landscape positions and depths, and the patterns were not significantly different laterally along hillslope (Wilcoxon p > 0.05). We identified no predictable patterns in median V_{wf} among landscape positions in H2 or H3. Hillslope H3, the steepest hillslope (Table S1), had the greatest mean wetting front velocity of all three hillslopes (203 mm h⁻¹). Hillslope H2, where preferential flow was most prevalent (Figure 3 and Table 3), had the greatest variability in wetting front velocity (CV = 1.53).

4.3. The MARS Models and Relative Influences of Soil Moisture Response

The MARS models explained 66% of the variability in observed response magnitude (Δs), followed by 57% of variability in observed response time (T_{p2p}), and 34% of variability in observed wetting front velocity (V_{wf}). A follow-up analysis of variable importance did not identify consistent dominance by any single explanatory variable in the MARS models (Figures 7 and S2). Instead, the analysis suggests a codominance of multiple variables. For Δs , the most important explanatory variables were storm depth, AMCs, and soil depth. Slope, soil depth, and storm depth had the greatest influence

soil depth and topographic slope were critical to explaining variance in $V_{\rm wf}$

Soil moisture and runoff responses were well correlated in general, al-

on $T_{\rm p2p}$. The most important explanatory variables for $V_{\rm wf}$ were soil depth, mean storm intensity, and storm depth. Among all topographic variables, slope was consistently identified as an explanatory factor, followed by profile curvature and topographic wetness index (Figure S2). To summarize, landscape characteristics such as

Table 4And T_{p2p} , and storm depth was the key explanatory factor for Δs .Table 4And (Coefficient of Variation) of Response Metrics for All LandscapePositions and DepthsAnd Soil Moisture and Runoff

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Hillslopes	Pits	$\Delta s \left(\mathrm{m}^3 \mathrm{m}^{-3} \right)$	$T_{p2p}(h)$	$V_{\rm wf} ({\rm mm}~{\rm h}^{-1})$
H1	US	0.06 (0.53)	3.53 (1.28)	124.53 (1.22)
	MS	0.05 (0.79)	2.33 (1.56)	179.34 (1.05)
	LS	0.04 (0.53)	2.03 (1.54)	279.64 (1.05)
	NS	0.02 (0.28)	1.62 (1.78)	304.43 (1.40)
H2	US	0.05 (0.75)	2.86 (1.34)	151.25 (1.09)
	MS	0.06 (0.60)	1.45 (1.34)	160.65 (1.15)
	LS	0.05 (0.54)	4.94 (1.10)	220.61 (1.94)
	NS	0.05 (1.01)	8.97 (0.66)	307.30 (1.47)
H3	US	0.06 (0.56)	2.53 (1.82)	193.30 (1.05)
	MS	0.05 (0.64)	1.29 (1.69)	202.48 (1.09)
	LS	0.07 (0.90)	5.45 (1.29)	199.39 (1.08)
	NS	0.05 (0.84)	9.5 (0.83)	258.29 (1.07)

Note. Italics text show the greatest variability for each response and the study hillslope. $V_{\rm wP}$ wetting front velocity; $T_{\rm p2p}$, time to peak; Δs , change in volumetric water content; US, upslope; MS, midslope; LS, lower slope; NS, near stream.

$\begin{array}{ll} (\mathrm{mm}\,\mathrm{h}^{-1}) & \text{though the strength of the soil moisture-runoff relationship } (\rho_{\mathrm{max}}) \, \mathrm{varied} \\ \mathrm{somewhat among hillslopes} \, (\mathrm{Figure 8}). \, \mathrm{The relationship was the strong-} \\ \mathrm{est \ for \ H1} \, (\rho_{\mathrm{max}} = 0.85) \, \mathrm{and \ weakest \ for \ H3} \, (\rho_{\mathrm{max}} = 0.70; \, \mathrm{Figure \ 8}). \, \mathrm{The} \end{array}$

est for H1 ($\rho_{max} = 0.85$) and weakest for H3 ($\rho_{max} = 0.70$; Figure 8). The corresponding lag times (l_{sr}) of peak correlation between soil moisture and runoff were extremely short: 0 h for H1 and H3 and 0.5 h for H2. On average, correlations were similar among hillslopes, ranging from 0.41(H3) to 0.52 (H1), but average lags varied from 0.5 h (H3) to -0.5 h (H2 and H1). Hillslopes H3 and H2 exhibited relatively large numbers of positive (27%) and negative (49%) lags, whereas H1 exhibited a relatively large number of zero (35%) lags (Table S4). Similar to the soil moisture responses during storms (Table 4), variability in the strength of relationships and the corresponding lags between soil moisture and runoff were greatest for the driest hillslope H3 (ρ_{max} , CV = 0.45; l_{max} , CV = 15) and least for the wettest H1 (ρ_{max} , CV = 0.26; l_{max} , CV = -3.13).

Storm properties and mean AMCs appeared to influence soil moisture and runoff relationships, although the degree of influence varied among hillslopes (Figure 8). The more intense the storm, the more quickly soil moisture responded on the wettest hillslope H1, as evinced by correlation



Figure 3. Evidence of preferential flow (black filled circle) across hillslopes, H1 (a), H2 (b), and H3 (c) along with storm depth (blue bars) and peak intensity (gray filled circles) for the storms studied (d). Gray dots indicate no response and gray crosses represent missing data. Gray plus signs represent the depths that served as the shallowest layers to estimate preferential flow at greater depths. Horizontal green dash-dotted lines show responses corresponding to some of the large storms. SD, storm depth; PI, peak intensity; US, upslope; MS, midslope; LS, lower slope; NS, near stream, whereas indices (1–5) were assigned to each probe to indicate its relative depth in the soil pit from 10 cm (1), 20 cm (2), 30 cm (3), 60 cm (4), to 100 cm (5).

lag times being inversely related to mean storm intensity ($\rho = -0.50$, p = 0.005). For all hillslopes, the greater the storm depth, the stronger the relationship between soil moisture and runoff responses. For H1, the soil moisture and runoff relationship was sensitive to storm depth ($\rho = 0.40$, p = 0.03) and antecedent soil mois-



Figure 4. Changes in volumetric water content (Δs) across hillslopes, H1 (a, b), H2 (c, d), and H3 (e, f) along with storm depth (blue bars) and peak intensity (gray filled circles) for the storms studied (g). Boxplots show the temporal variability noted in Δs across landscape positions and depths. Symbol color indicates antecedent moisture conditions (AMCs, units m³/m³), and diameter indicates the magnitude of change in Δs . Gray dots indicate no response and gray crosses represent missing data in the top three panels. Gray rectangles highlight landscape positions and depths that exhibited relatively high antecedent conditions that resulted in poor infiltrability throughout the study period. Horizontal gray dash-dotted lines show responses corresponding to some of the large storms. SD, storm depth; PI, peak intensity; US, upslope; MS, midslope; LS, lower slope; NS, near stream, whereas indices (1–5) were assigned to each probe to indicate its relative depth in the soil pit from 10 cm (1), 20 cm (2), 30 cm (3), 60 cm (4), to 100 cm (5).





Figure 5. Soil moisture response timing (T_{p2p}) across hillslopes, H1 (a, b), H2 (c, d), and H3 (e, f) along with storm depth (blue bars) and peak intensity (gray filled circles) for the storms studied (g). Boxplots show the temporal variability noted in T_{p2p} across landscape positions and depths. Symbol color indicates antecedent moisture conditions (AMCs, units m³/m³), and diameter indicates the magnitude of T_{p2p} . Gray dots indicate no response and gray crosses represent missing data in the top three panels. SD, storm depth; PI, peak intensity; US, upslope; MS, midslope; LS, lower slope; NS, near stream, whereas indices (1–5) were assigned to each probe to indicate its relative depth in the soil pit from 10 cm (1), 20 cm (2), 30 cm (3), 60 cm (4), to 100 cm (5).



Figure 6. Soil moisture wetting front velocity (V_{wf}) across hillslopes, H1 (a, b), H2 (c, d), and H3 (e, f) along with storm depth (blue bars) and peak intensity (gray filled circles) for the storms studied (g). Boxplots show the temporal variability noted in V_{wf} across landscape positions and depths. Symbol color indicates antecedent moisture conditions (AMCs, units m³/m³), and diameter indicates the magnitude of V_{wf} . Gray dots indicate no response and gray crosses represent missing data in the top three panels. SD, storm depth; PI, peak intensity; US, upslope; MS, midslope; LS, lower slope; NS, near stream, whereas indices (1–5) were assigned to each probe to indicate its relative depth in the soil pit from 10 cm (1), 20 cm (2), 30 cm (3), 60 cm (4), to 100 cm (5).





Figure 7. Summary of relative importance for the most influential three variables for Δs (blue links), $T_{p_{2p}}$ (green links), and V_{wf} (black links) derived from Multivariate Adaptive Regression Spline models. For each response, wider links indicate greater relative importance. V_{wf} wetting front velocity; Δs , change in volumetric water content; $T_{p_{2p}}$, peak to peak time; SD, storm depth; SoD, soil depth; MI, mean intensity; AMCs, antecedent moisture conditions; SLP, slope.

ture conditions ($\rho = 0.69$, p < 0.001). For H2, storm depth alone was the dominant influence on the soil moisture—runoff correlation ($\rho = 0.62$, p < 0.001). For H3, a combination of storm depth ($\rho = 0.49$, p < 0.001), mean intensity ($\rho = 0.44$, p = 0.01), and mean AMCs ($\rho = 0.38$, p = 0.05) influenced the correlation. No significant influence of mean $V_{\rm wf}$ and occurrence of preferential flow was noted on the soil moisture and runoff relationships during the study period.

5. Discussion

5.1. Soil Moisture Variability

Soil moisture responses from 45 combinations of landscape position and soil depth revealed a few important behaviors. In general, both wetting front velocity ($V_{\rm wf}$) and the difference between the time of peak rainfall and time of peak soil moisture response ($T_{\rm p2p}$) gradually increased with depth in the soil. We only observed a trend in response magnitude (Δs) for H1, and that was a decline in Δs with depth. These results suggest that infiltration and percolation are dominant mechanisms through which wetting fronts propagate through soil profiles in H1, the wettest hillslope. However, hillslopes H2 and H3 did not exhibit clear depth gradients in soil moisture responses and indicated that preferential flow might be the dominant flow mechanism for these two hillslopes (Table 4 and Figures 3–6). These results are in agreement with studies that used similar methods to quantify preferential flow. These studies attributed the patterns to differences in antecedent conditions and topographic positions (Graham & Lin, 2011; Liu & Lin, 2015).

Hydrologists have long emphasized the complex task of studying preferential flow (Band et al., 2014; Beven & Germann, 1982, 2013; Weiler

& Flühler, 2005; Whipkey, 1965). Prior work has attributed preferential flow to mechanisms such as macropore activation, poor infiltration, restrictive layers, and heterogeneity in soil moisture distribution (cf., Lin, 2010). Preferential flow at Coweeta may occur due to any of these mechanisms, including combinations of mechanisms. For some landscape positions, preferential flows may have been infrequent (<15%), occurring only during large (>100 mm) storms (Table 3 and Figures 3 and 4). Large storms may initiate preferential flow by enabling connectivity between soil pores of different sizes (Sidle et al., 2000, 2001). This phenomenon is further supported by the observed correlation between storm depth and preferential flow ($\rho = 0.33$, p = 0.02), indicating that preferential flow may occur during larger storms at Coweeta due to this type of enhanced connectivity. Given the high temporal variability in soil moisture during events, the weak relationship is not surprising and still offers a valuable insight in explaining the patterns.

At other landscape positions, preferential flows occurred frequently (up to 50% of all storms), regardless of storm characteristics (Table 3 and Figure 3). These soil layers were superimposed by high AMCs (Figure 4), resulting in poor infiltrability and activation of preferential flow (Lin, 2010; Whipkey & Kirkby, 1978; G. V. Wilson et al., 1990). The influence of antecedent conditions on the occurrence of preferential flow has been established, but the role of rainfall properties in mediating preferential flow remains unclear. For instance, Hardie et al. (2013) reported no effect of rainfall properties on the occurrence of preferential flow in an agricultural landscape, whereas positive relationships have been observed between preferential flow and storm intensity and total depth (Demand et al., 2019; Wiekenkamp et al., 2016) in forested landscapes.

Soil moisture variability can inform the experimental design needed to fully understand soil water dynamics at hillslope or catchment scales (Tague et al., 2010). Prior studies have found consistent and systematic spatial and temporal variability in soil moisture responses associated with specific landscape positions (e.g., Detty & McGuire, 2010; Famiglietti et al., 1998; Kaiser & McGlynn, 2018; Kim, 2009; Penna et al., 2009). Our work further advances this topic by highlighting the wide variability noted among response metrics along hillslopes. We found (Figures 4–6 and Table 4) that at event scales variability in soil moisture is sensitive





Figure 8. Patterns of max correlation (ρ_{max}) and corresponding lag (l_{sr}) estimated between runoff and soil moisture along H1 (a), H2 (b2), and H3 (c). Significant Spearman's ρ correlations (p < 0.05) shown in color. The positive lag values suggested runoff peaked earlier than soil moisture, negative lag values suggested soil moisture peaked earlier, and zero lag values indicated both peaked at the same time during storms. Symbol color indicates antecedent moisture conditions (AMCs, units m³/m³).

to the response metrics analyzed (i.e., Δs , T_{p2p} , or V_{wf}). No specific landscape position showed consistently high temporal variability across all response metrics during storms (Table 4), indicating the insensitivity of the variability to any specific landscape position along hillslopes. On the other hand, temporal variability in all soil response metrics is inversely related to mean volumetric water content, irrespective of the response metric type (Tables 2 and 4 and Figures 4–6), suggesting greater water content led to less temporal variability in response metrics.

Overall, soil moisture response metrics are highly variable within a catchment (Figures 4-6 and Table 4). Studies that rely on single landscape positions or single soil depths may draw incorrect conclusions about soil water dynamics for the catchment as a whole. Shallow soil depths are often more responsive, and as a consequence, they have been prioritized in monitoring over greater depths. However, we note that deeper soil moisture responses can be quite heterogenous depending on the response metric (Figures 4-6 and Table 4). Overall, these findings highlight the heterogeneity in soil moisture patterns and point to the importance of spatially intensive soil water content monitoring. Differences among responses in each of the three hillslopes further highlight the potential pitfall of drawing inferences from an individual hillslope. Upscaling these responses to an entire catchment would require clear knowledge not only about the relationships between landscape positions, soil depths, and individual response metrics but also about the stability of these relationships with respect to storm characteristics.

5.2. Storm Characteristics

Individual storm properties interacted with landscape characteristics in various ways to influence soil moisture responses (Figures 7 and S2). Storm depth dominated the patterns of response magnitude (Δs), mean storm intensity influenced wetting front velocity (V_{wf}), and the storm depth and period had a greater influence on response time (T_{p2p}), as revealed with MARS modeling (Figure 7). Larger storms yielded greater

 Δs responses but shorter T_{p2p} , and they had almost no detectable influence on V_{wf} (Table S5). The positive correlation between mean intensity and wetting front velocity indicates that the more intense the storm, the faster the wetting front propagated through soil, and the shorter the initial response timing. These results suggest that at the inception of rainfall, storm intensity along with other landscape drivers might determine how rapidly the wetting front moves. By the time soil moisture would peak, storm depth begins to dominate the response (Figure 7). These results expand our understanding from the existing work that attributed the soil moisture responses to one or the other rainfall properties (Blume et al., 2009; Penna et al., 2011; Tani, 1997). For instance, Blume et al. (2009) and Lin and Zhou (2008) postulated the influence of storm intensity on initial response timing in forested catchments. Our findings refine these views by considering the relative influence of each storm's characteristics on various aspects of soil moisture response. Overall, these results demonstrate that the influence of storm properties on soil moisture varies during storms and highlight the need for careful representation of these processes in hydrological models.

5.3. Topography

We found that the dominant topographic influence varied depending on the specific soil moisture response (Figures 7 and S2). The topographic slope had a higher order of influence on soil moisture responses than the seven other topographic variables used in MARS models (Figures 7 and S2). Broadly, topographic slope can serve as a proxy to the hydraulic gradient and may thus influence wetting front and response time of subsurface flows (Moore et al., 1991; Western et al., 2002). In other words, the steeper the landscape, the larger the hydraulic gradient and the greater the rates of subsurface flows. This could possibly explain the

shortest response time (T_{p2p}) , including negative values, along H3, the steepest hillslope, where soil moisture reached its peak earlier than rainfall during most storms. Our findings expand our understanding of topographic influence on response time (T_{p2p}) by demonstrating the relative importance of slope over other topographic variables, as indicated by prior work (Kim, 2009).

Our results are consistent with longstanding research on the role of topographic heterogeneity in driving the lateral redistribution of soil water (Burt & Butcher, 1985; Jencso et al., 2009; Kaiser & McGlynn, 2018; Nanda et al., 2019; Western et al., 2002). However, our work also complicates the prevailing view that topography exerts a uniform influence on soil moisture responses (Figure 7). This work shows that for sites similar to Coweeta, where storms are frequent and deep soils have the potential to store substantial volumes of water, the topographic influence on soil moisture response is contingent on other factors. In other words, topographic variables may not be able to predict certain aspects of soil moisture responses to storms, because the influence of these variables may change from storm to storm, or within a single storm. Thus, the modelers that rely on topographic variables to represent soil hydrological processes should exercise caution when drawing conclusions about soil water dynamics at the storm scale.

5.4. Soil Properties and Antecedent Conditions

Soil properties mediate water redistribution and influence the spatial heterogeneity of AMCs due to varying structure, texture, depth, and hydraulic characteristics along the soil profile (cf., Famiglietti et al., 1998; Gwak & Kim, 2017; Hillel, 1998; Hopp & McDonnell, 2009). The MARS modeling confirmed that soil depth and antecedent conditions were among the key drivers of wetting front velocity (V_{wf}), response magnitude (Δs), and response time (T_{p2D}) at Coweeta (Figures 7 and S2).

Drier antecedent conditions can facilitate a greater response magnitude during storms (Table S5). Further, drier antecedent conditions also facilitate larger soil moisture gradients, resulting in greater cumulative infiltration within the soil profile (Gray & Norum, 1967). This mechanism may explain why we observed higher response percentages and larger values of Δs closer to the soil surface than deeper in the soil profile. Further, decrease in Δs with depth is supported by lower bulk density at shallow depths that encourages more storage than the greater depths. The role of bulk density in mediating the response magnitude (Δs) is supported by its relatively higher importance than the remaining landscape variables in the MARS models (Figure S2). For response timing, as the wetting front continued to percolate along the soil profile, the time taken to reach the peak increased with soil depth (Table S5). Similarly, due to the proximity to the surface and large evaporative gradients, pores in shallow soil layers were relatively dry, disconnected, and not conducting, resulting in lower velocities of wetting fronts than those at greater depths. These results agree with past work on the effects of soil depth and antecedent conditions on individual response metrics (e.g., Demand et al., 2019; Ritsema et al., 1998; Rosenbaum et al., 2012).

5.5. Soil Moisture and Runoff

Field-based, data intensive studies have shown a strong influence of soil moisture patterns on runoff generation in forested headwaters (Detty & McGuire, 2010; Dunne & Black, 1970; Haga et al., 2005; James & Roulet, 2007; Penna et al., 2011). Most of these studies identified subsurface runoff generation as the dominant runoff mechanism. For instance, at Coweeta, due to high infiltration rates and large storage capacities, soil moisture is believed to contribute substantially to subsurface storage that sustains streams throughout the year (Gaskin et al., 1989; Hewlett, 1961, Hewlett & Hibbert, 1967). Our findings show that for nearly 40% of storms, soil moisture peaked prior to runoff (Figure 8 and Table S4) indicating potential subsurface contributions from hillslopes to runoff at the catchment outlet. Similar results have been noted in other headwaters where subsurface flows from riparian zones influence streamflow (McGlynn & Mc-Donnell, 2003; Penna et al., 2011). Past work at Coweeta implicates shallow groundwater from NS areas as a major contributor to streamflow responses to storms (Hewlett & Hibbert, 1967; Singh et al., 2018). The results from this study suggest that unsaturated flow from NS areas of hillslopes may also contribute to streamflow generation. Our findings are further supported by recent work demonstrating the coupling of unsaturated flow and shallow groundwater at Coweeta, and the relationship of both fluxes to streamflow at the catchment outlet (Scaife et al., 2020). Storm properties appear to mediate relationships between soil moisture and runoff responses at Coweeta (Figure 8). Storm depth partly explained this relationship, and a possible mechanism could be enhanced hydrologic connectivity and greater subsurface flows from hillslopes to streams during larger storms (e.g., McGlynn & McDonnell, 2003; Penna et al., 2011; Scaife et al., 2020). Mean storm intensity also appears to mediate lags between soil moisture and runoff responses, likely because the rapid propagation of wetting fronts resulted in a shorter lag between soil moisture and runoff. Overall, these results emphasize the influence of storm properties on the soil moisture-runoff relationship at event scales.

Relationships between streamflow and soil moisture, including the spatial heterogeneity of soil moisture, remain a key area of interest in catchment hydrology (Blume et al., 2009; Freeze, 1980; R. B. Grayson et al., 1995; Merz & Plate, 1997). We found that hillslope H1 exhibited the shortest and most positive lags between soil moisture and runoff (Table S4), suggesting that soil moisture reached peak quickest along the wettest hillslopes for most storms. Further, the CV in the soil moisture and runoff relationship was highest (ρ_{max} , CV = 0.45; l_{sr} , CV = 15) for the driest hillslope (H3) and smallest (ρ_{max} , CV = 0.26; l_{sr} , CV = -3.13) for wettest hillslope (H1), indicating that antecedent conditions minimized the temporal variability in soil moisture–runoff relationship. Other field-based studies have reported the influence of hillslope-scale preferential flows on runoff (Angermann et al., 2017; Blume et al., 2009). In contrast, we did not detect any significant effect (p > 0.05) of preferential flow on soil moisture-runoff relationships. These findings contribute to the growing understanding of the role of preferential flow in streamflow generation processes in forested headwaters (Beven & Germann, 2013; Weiler, 2017). Overall, these results suggest an important role of hillslope characteristics such as antecedent conditions in determining correlation strength between soil moisture and runoff.

Above- and below-ground vegetation surveys could improve this work by elucidating the joint influence of topography and plant-water relations on soil moisture patterns at Coweeta. There are no major differences in vegetation characteristics among instrumented hillslopes in WS02, one of the reasons for choosing this site for our research (Table S1). Even so, the spatial heterogeneity of AMCs may be influenced by subtle variations in interception, evapotranspiration, shading, and other vegetation impacts on the soil water balance within the study watershed. Further research is needed to incorporate these impacts into studies of soil moisture responses to storms at Coweeta and in other landscapes characterized by frequent storms, deep soils, and forest cover.

6. Conclusions

We investigated soil moisture responses metrics during a wide range of storm conditions at Coweeta. Individual response metrics varied among landscape positions and soil depths, and no systematic variability was found in response metrics with depth and landscape positions along hillslopes. The MARS models confirmed no universal controls on response metrics, instead we found that drivers shifted with individual response metrics and during storms. These findings highlight the need for intensive observations to fully characterize heterogeneity in subsurface processes, including preferential flow. The soil moisture-runoff relationship varied with storm depth, mean intensity, and antecedent conditions, and the strength of the relationship was unique to each hillslope.

Storm properties and landscape characteristics interact to influence soil moisture responses. These soil moisture responses, together with storm properties, influence the soil moisture-runoff relationship during storms. Modelers should consider the relative importance of commonly used controls to refine process-based representations of soil moisture and its influence on streamflow. Further, this work highlights the potential limitations of topographic approximations of soil moisture status in applications that focus on storm events or relatively short time scales.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.



Data Availability Statement

Rainfall data are publicly available from USFS (Miniat et al., 2017) and soil moisture data can be retrieved from Figshare (doi:10.6084/m9.figshare.16441323).

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