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Soil moisture gradients and controls on a southern Appalachian hillslope from drought through recharge

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Abstract

Soil moisture gradients along hillslopes in humid watersheds, although indicated by vegetation gradients and by studies using models, have been difficult to confirm empirically. While soil properties and topographic features are the two general physiographic factors controlling soil moisture on hillslopes, studies have shown conflicting results regarding which factor is more important. The relative importance of topographic and soil property controls was examined in an upland forested watershed at the Coweeta Hydrologic Laboratory in the southern Appalachian mountains. Soil moisture was measured along a hillslope transect with a mesic-to-xeric forest vegetation gradient over a period spanning precipitation extremes. The hillslope transect was instrumented with a time domain reflectometry (TDR) network at two depths. Soil moisture was measured during a severe autumn drought and subsequent winter precipitation recharge. In the upper soil depth (0–30 cm), moisture gradients persisted throughout the measurement period, and topography exerted dominant control. For the entire root zone (0–90 cm), soil moisture gradients were found only during drought. Control on soil moisture was due to both topography and storage before drought. During and after recharge, variations in soil texture and horizon distribution exerted dominant control on soil moisture content in the root zone (0–90 cm). These results indicate that topographic factors assert more control over hillslope soil moisture during drier periods as drainage progresses, while variations in soil water storage properties are more important during wetter periods. Hillslope soil moisture gradients in southern Appalachian watersheds appear to be restricted to upper soil layers, with deeper hillslope soil moisture gradients occurring only with sufficient drought.

Introduction

Soil moisture distribution and controls on hillslopes have long been subjects of inquiry (e.g. Dreibelbis and Post, 1940; Hack and Goodlett, 1960; Helvey *et al.*, 1972; Dunne *et al.*, 1975; Burt and Butcher, 1985; Boyer *et al.*, 1990, Afyuni *et al.*, 1993). Two general physiographic factors control soil moisture distribution on hillslopes: soil properties and topographic features. The relative importance of these controls depends on a complex set of factors, including rainfall magnitude and frequency, geologic structure, geomorphic history, and vegetation type. Few studies, however, have quantified both topographic features and soil property distribution simultaneously; as a result it remains unclear whether topographic features (Burt and Butcher, 1985; Petch, 1988) or soil properties (Helvey *et al.*, 1972; Afyuni *et al.*, 1993) provide more control. Further, it remains unclear how these controls operate dynamically under various rainfall regimes.

Ecologists in the Appalachian mountains of North America have inferred the existence of hillslope soil moisture gradients from hillslope distributions of forest vegetation (Whittaker, 1956; Hack and Goodlett, 1960; Day and Monk, 1974). Physical and simulation models have indicated the existence of soil moisture gradients along hillslopes in humid temperate watersheds (Hewlett and Hibbert, 1963; Sloan and Moore, 1984), but field measurement has not verified the existence of such moisture gradients (Dreibelbis and Post, 1940; Helvey and Patric, 1988). Decades of measurement in the southern Blue Ridge have been summarised: 'We conventionally think of cove sites as wet and upper slopes as drier, but this generalization did not hold in the study area because there was no consistent relationship between soil moisture content and slope position' (Helvey and Patric, 1988).

The objectives of this study were: (1) to determine whether significant hillslope soil moisture gradients exist along steep hillslopes in humid upland forested watersheds in the southern Appalachian mountains; and

(2) to determine the relative importance of topographic and of soil property controls on hillslope soil moisture during both dry and wet seasonal conditions.

Methods

SITE SELECTION

The Coweeta Hydrologic Laboratory is in the Coweeta syncline in the eastern part of the southern Appalachian Blue Ridge. The soils of Coweeta are predominantly Ultisols and Inceptisols underlain by a deep saprolite layer. Overall average weathering profile thickness (depth to bedrock) is about 6 metres (Swank and Douglass, 1975). The major physical distinction between Ultisols and Inceptisols is morphological, as their chemical and mineral properties at Coweeta are very similar (Velbel, 1988).

Selection criteria for the experimental site were approached using the inference made by ecologists by selecting a hillslope with a distinct mesic-to-xeric vegetation gradient. The hillslope was further selected so that soil moisture was isolated as the only probable environmental gradient affecting vegetation distribution. Other criteria included a hillslope with relatively continuous slope, a lack of rock outcrops and control on environmental variables other than soil moisture that could distribute vegetation. A hillslope fitting the criteria was found on the lower western side of Watershed 2 (WS 2), approximately 200 metres north of the weir. The selected hillslope was fairly short at 84.7 metres in length (plan view). Visual inspection, later confirmed by vegetation stem-mapping, indicated that forest vegetation changed from a mesic *Rhododendron maximum-Tsuga canadensis-Quercus alba* association near the stream to a xeric *Kalmia latifola-Pinus rigida-Quercus prinus* association on the ridge. The hillslope had an eastern aspect and an elevation change of roughly 60 metres. The slope of the study transect was relatively smooth and possessed a steepness typical for watershed slopes at Coweeta (Table 1). Solar radiation received during the day was uniform from cove to ridge due to a relatively low opposing hillslope. Near sunset, the ridgetop received more solar input than the cove. With an elevation change of just 60 metres and a nearly constant solar input, variation in temperature along the hillslope was negligible. Two soil series for lower WS 2 were previously identified: Fannin (fine-loamy, micaceous, mesic Typic Hapludult) on upper slopes and Cullasaja-Tuckasegee (fine-loamy, oxidic, mesic Typic Haplumbrept) near the stream (Thomas, 1996). Both series are mostly sandy loam to sandy clay loam and are derived from mica gneiss parent material, so mineralogical differences that can cause vegetation changes (Strahler, 1972) were minimal. In summary, other environmental gradients (temperature, incident radiation, soil type) that distribute vegetation were relatively constant. The existence of a strong vegetation gradient, coupled with the absence of other environmental controls on vegetation, indicated a high probability of a soil moisture gradient on this hillslope.

EXPERIMENTAL DESIGN AND SAMPLING

A transect approximately perpendicular to the stream was established along the center of the hillslope using survey level and rod. A time domain reflectometry (TDR) network was installed vertically through two depths (0–30 cm, 0–90 cm) along the transect. The 0–30 cm depth was chosen to represent upper horizons (O, A, BA) (Gaskin *et al.*, 1989); the 0–90 cm depth was chosen to represent the approximate root zone (McGinty, 1976). Sample plots were placed at 5 m intervals through the first 40 m and then every 10 m to ridge. Sample plots were also placed on the streambank and at the divide, giving a total of 14 plots. A plot consisted of 2 depths, each with 3 replications, for a total of 6 sample points. Three replicates have been shown to be sufficient to estimate mean soil moisture content for 5 m plots with no more than a 3% error (Kamgar *et al.*, 1993). A sample point consisted of two 3-mm diameter stainless steel welding rods (i.e. TDR rods) set 5 cm apart and inserted vertically. Litter was removed during rod emplacement and then replaced. Approximately 2 cm of each rod was left above the surface for connection to the TDR meter (Trase 6050XI, Soil Moisture Equipment Corporation). The TDR method uses an empirically-determined polynomial relationship between dielectric constant (K_a) and water content (θ) of a soil, which is essentially independent of soil type, density, salt content, and temperature for a wide range of soils (Topp *et al.*, 1985). Knowing time (t) to reflection, the dielectric constant of soil material is given by $K_a = (ct/L)^2$, where c is speed of light and L is length of TDR rods (Trase manual). Recent work has called for calibration of TDR to individual measurement sites and for visual interpretation of TDR traces to avoid automated meter interpretation errors (Gray and Spies, 1995). Individual sites were not calibrated in this work, because of studies that have calibrated the TDR method to within 1.3% for a wide range of soils, including sandy loams and sandy clays loams prevalent in the Coweeta Basin (Topp *et al.*, 1980, 1985). Recalibration for anomalous soils has sometimes resulted in differing y-intercepts, yet it has not resulted in significant slope differences (Gray and Spies, 1995). While it is possible that absolute moisture contents in the present study had errors due to not calibrating for each of the 42 sites, relative moisture changes would not have been affected. In all cases in this study, TDR traces were interpreted manually by the same individual. A backup TDR meter (Tektronix 1502B) was also used; this measured an equivalent distance to reflection. Soil moisture using this meter was determined by $K_a = (S/L)^2$, where S is measured distance to reflection (F.N. Dalton, pers. comm., 1990). Close agreement (within 1%) was found for several comparisons between Trase and Tektronix metres.

Table 1. Hillslope physiographic characteristics. Each value below is the mean of three independent replicates, with standard deviations shown in parentheses.

Dist[a]	Slope	D_b	Org[%]		Horizon depth[b][cm]				
[m]	[°]	[g cm^{-3}]	O	A	O	A	BA	B	BC
0	33(3.8)	.81(.03)	20(.48)	11(1.3)	7(3.6)	11(6.7)	15(8.2)	17(7.1)	27(25.)
5	27(7.8)	.86(.07)	16(5.2)	8(0.7)	6(0.0)	8(2.0)	9(2.5)	23(15.)	30(16.)
10	42(1.2)	.89(.14)	16(1.1)	8(0.3)	7(1.0)	7(2.1)	6(1.5)	7(2.1)	36(26.)
15	28(5.7)	.85(.08)	16(1.6)	9(1.6)	7(2.5)	8(2.9)	10(4.0)	32(8.7)	25(8.1)
20	28(4.4)	.91(.07)	15(1.3)	9(0.8)	6(0.6)	9(1.0)	8(1.5)	22(16.)	39(22.)
25	28(2.3)	.94(.10)	13(1.5)	8(0.5)	3(0.6)	7(0.6)	7(3.8)	68(3.1)	4(4.6)
30	32(4.0)	.94(.13)	14(2.3)	6(2.1)	6(0.6)	9(1.2)	8(4.2)	36(15.)	29(13.)
35	34(1.2)	.87(.06)	13(4.2)	8(0.8)	8(1.0)	8(1.5)	9(3.6)	28(16.)	35(15.)
40	33(2.1)	.91(.06)	12(.78)	9(0.4)	10(1.5)	6(2.3)	6(1.0)	17(8.5)	45(8.5)
50	37(3.5)	.96(.05)	16(2.6)	8(0.3)	7(2.0)	7(0.0)	7(1.0)	21(7.8)	46(5.9)
60	37(1.5)	.96(.09)	17(4.9)	8(0.3)	8(1.2)	5(1.0)	10(3.1)	14(5.3)	51(3.8)
70	33(1.5)	.86(.19)	27(12.)	12(0.6)	8(3.8)	5(1.2)	5(0.6)	9(7.4)	62(2.6)
80	25(1.5)	.92(.15)	47(8.2)	11(0.7)	6(1.5)	3(0.6)	4(2.0)	18(3.1)	56(6.7)
85	22(6.2)	.96(.10)	37(2.7)	15(2.4)	6(1.0)	3(1.2)	3(1.2)	11(7.5)	46(14.)

Dist[a]	Clay[%]				Sand[%]			
[m]	A	BA	B	BC	A	BA	B	BC
0	22(3.2)	31(3.1)	25(2.6)	26(2.1)	55(3.5)	48(2.3)	51(3.9)	53(1.7)
5	29(3.9)	31(4.1)	29(6.8)	27(3.3)	50(4.7)	46(2.0)	47(1.9)	56(1.4)
10	27(4.4)	28(3.3)	30(4.2)	21(5.5)	51(5.1)	48(4.7)	48(4.1)	56(3.0)
15	26(2.2)	27(5.7)	29(1.6)	26(6.3)	53(3.0)	47(7.0)	45(1.7)	52(3.1)
20	24(3.5)	31(0.7)	35(1.2)	25(4.4)	54(3.1)	47(0.6)	45(3.1)	52(3.9)
25	29(2.2)	31(0.9)	34(1.3)	26(0.3)	49(3.3)	47(3.1)	44(3.5)	50(2.1)
30	25(1.7)	31(5.5)	29(0.5)	30(2.7)	52(3.5)	46(6.3)	48(0.5)	49(2.2)
35	25(4.4)	27(3.5)	29(3.9)	26(0.9)	52(3.4)	52(4.5)	53(3.0)	54(1.2)
40	25(1.8)	29(1.3)	34(0.5)	26(5.3)	52(2.4)	49(1.5)	45(1.5)	56(2.3)
50	24(3.4)	29(2.4)	29(7.4)	26(5.0)	52(4.4)	48(3.9)	47(9.1)	53(1.7)
60	21(3.2)	24(1.3)	27(4.6)	27(2.8)	56(4.5)	53(0.6)	53(5.1)	52(1.9)
70	21(6.2)	29(3.3)	37(1.4)	24(9.4)	60(7.6)	49(3.9)	43(4.0)	54(12.)
80	22(3.2)	29(5.3)	34(2.0)	28(5.1)	56(0.5)	47(3.8)	46(3.2)	51(2.2)
85	19(0.7)	26(2.5)	23(5.1)	22(4.0)	59(0.2)	49(3.8)	58(7.0)	57(3.2)

[a] measured horizontally from stream
[b] measured vertically

All TDR rods were in place by 10 November, 1991; the first sample was taken on 11 November. Samples were collected intermittently depending on rainfall (16 samples in 30 days) through the period of precipitation recharge. A collection period lasted 2–2.5 hours. With a few exceptions due to rain or meter failure, collections were conducted within 3 hours of noon. After recharge, collection was conducted less frequently, with a final sample on 3 February, 1992.

PHYSIOGRAPHIC VARIABLES

Indices of topographic and soil variation were used to contrast the relative importance of controls on soil moisture distribution. For humid watersheds in relatively steep terrain, topographic factors are a primary control on streamflow (Hewlett and Hibbert, 1967; Wood *et al.*, 1990). Soil moisture at a point is positively related to the cumulative upslope watershed area draining to that point, or upslope source area (a). Soil moisture at a point is also inversely related to local slope angle ($tan\beta$). Based on these observations, Beven and Kirkby (1979) developed an index of topographic similarity that could be applied at any given point on a watershed surface:

$$Topographic\ Index = ln(a/tan\beta) \quad (1)$$

Such indices that explain cumulative topographic effects have been used to classify areas within a watershed by

topographic similarity to predict streamflow (Hornberger *et al.*, 1985; Wood *et al.*, 1990) as well as soil moisture distribution (O'Loughlin, 1986; Jackson, 1991).

Variation in soil properties affecting soil moisture may occur vertically as soil horizons with varying water holding capacities change and laterally along hillslopes with changing depths of soil horizons due to geomorphic processes (Conacher and Dalyrmple, 1977; Buol *et al.*, 1989). Soil properties affect the shape of the soil moisture characteristic equation. Several studies have predicted soil moisture content at fixed matric potentials based on texture, bulk density and/or organic matter content alone with correlation coefficients ranging from 0.80 to 0.97 (Williams *et al.*, 1992). In general, the greater the clay content, the greater the water content at any particular suction, and the more gradual the slope of the curve. In contrast, an inverse relationship is generally found between soil moisture content and sand (Hillel, 1980; Cosby *et al.*, 1984). Based on these relationships, in this study a storage index was developed to represent soil moisture holding capacity integrated over a given soil depth. The storage index accounts for the relative percentages of clay and sand in a soil particle size distribution weighted by horizon depth, and is calculated by:

$$Storage\ Index = \frac{\Sigma(D_i.Cl_i)}{\Sigma(D_i.Sa_i)} \qquad (2)$$

where for horizon i, D_i is depth, Cl_i is clay particle size fraction and Sa_i is sand particle size fraction. Organic matter content has also been correlated with soil moisture distribution (Rawls *et al.*, 1982). To account for control by organic matter distribution in upper horizons, a depth weighted organic matter index was used:

$$Organic\ Matter\ Index = \Sigma(D_i \cdot Or_i) \qquad (3)$$

where for horizon i, D_i is depth and Or_i is the organic matter fraction.

FIELD MEASUREMENTS

Soil horizons in each sample plot were determined using a soil auger rather than soil pits to obtain a closer estimation of soil variation near all 42 TDR sample sites and to minimize damage to the soil profile on the hillslope. Successive 15 cm segments were extracted to determine the depths of each horizon down to 85 cm (auger length). Horizon determination was based on textural and colour differences, and classified as either O, A, BA, B, or BC (Buol *et al.*, 1989). Three soil cores were extracted for each plot (each within a metre of a TDR sample point) and horizon depths were averaged to represent the plot. To obtain sufficient sample for textural analyses in the case of thin horizons, additional soil was collected within a metre of each core measurement. Slope angle (β) at each TDR site was measured using a clinometer. Length for upslope contributing area (a) was determined by surveyed distance to the divide. Width for upslope contributing area was set equal to the widest spacing in TDR sites on any plot, 20 m, as terrain analysis has shown this hillslope to be divergent in planform (Yeakley *et al.*, 1995). All soil samples collected for soil property analyses were first oven dried at 105 °C for 48 hours. Particle size analysis was conducted using the hydrometer method (Kalra and Maynard, 1991) for the mineral horizons (n = 168: 4 horizons × 3 cores × 14 plots); at least 50 g of sample was used for each hydrometer test. Bulk density, D_b, was measured using a Soiltest field density sampler (10.1 cm diameter × 11.8 cm depth). Bulk density samples (n = 42) were extracted from the upper 30 cm of soil, after first removing the O horizon. Organic matter was determined using loss-on-ignition for samples from the O and A horizons (n = 84). At least 5.00 g of sample was burned in a muffle furnace at 375 °C for at least 16 hours (Kalra and Maynard, 1991).

ANALYSIS

Means and variances were computed for each measurement period for soil moisture values by depth and by plot. Although only directly measured soil depths (i.e., 0–30 cm and 0–90 cm) were admitted for use in analyses of gradients and physiographic controls, an estimate was made of 30–90 cm layer soil moisture to contrast whole-slope response of the 0–30 cm layer with that of the 30–90 cm layer. Each point value $\theta_{30\text{–}90}$ was estimated using the relation $\theta_{0\text{–}90}{}^{*}d_{0\text{–}90} = \theta_{0\text{–}30}{}^{*}d_{0\text{–}30} + \theta_{30\text{–}90}{}^{*}d_{30\text{–}90}$, where d is depth. Mean soil moisture values for both the 0–30 cm and estimated 30–90 cm soil layers were areally-weighted by plot size along the transect to determine a value for each soil layer over the entire hillslope.

Moisture gradients were approximated by the regression coefficient (or slope), using linear regression for both measured depths (0–30 cm, 0–90 cm). For moisture gradient determination, the independent variable was normalised plane distance along the hillslope, varying from 0 to 1. The dependent variable was fractional moisture content, varying from about 0.05 to 0.45. A moisture gradient could then vary from 0% (i.e. no change along the hillslope) to ±40% (i.e., maximum change in soil moisture content from ridge to stream). In using the regression coefficient to represent a soil moisture gradient, no assumption was made of a linear process distributing soil moisture or that moisture gradients along hillslopes are best described as linear relations. Rather, a linear approximation was just as used as the most straightforward representation of whether a change in soil moisture from ridge to stream was found.

Multiple regression analyses were used to determine relative importance of controls on soil moisture. Independent variables included all primary physiographic measurements (Table 1), plus the topographic index, the organic matter index and the storage index over the appropriate depth (i.e. either 0–30 or 0–90 cm). Stepwise regression was used as a screening analysis to determine which inde-

pendent variables were significantly correlated in partial F tests ($p<.05$) with soil moisture distribution in each measured soil layer. For those independent variables found significantly correlated, partial regression analyses of the appropriate order (Zar, 1984) were conducted to isolate their relative importance in controlling soil moisture on the study hillslope from drought through recharge. Two-tailed t-tests were used to determine the significance level of resulting partial correlation coefficients.

Results

SOIL MOISTURE RESPONSE

Only 4.8 cm rain fell during September and October, 1991, 80% below the mean for those months (Fig. 1). Shortly after the TDR network was installed, precipitation recharge began with a 12.7 cm rain over 21–23 November, 1991. A second large storm front deposited 11.6 cm from 30 November to 3 December, 1991. Several subsequent lighter rains occurred during the following two months. Streamflow approximately doubled following the two storms (Fig. 1).

Contrasting the areally-weighted mean soil moisture response between the 0–30 cm and 30–90 cm layers showed that before the recharge, the lower layer was about 3% higher in moisture content (18% vs. 15%). Following the 12.7 cm event, average moisture content in the top layer peaked one day later at 29%, while the lower layer peaked two days later at 32%. During the second event of 11.6 cm, moisture content in both layers was near 32% (Fig. 1). Drainage after these events followed a negative exponential function. Sporadic rain events kept the mean value of both layers between 25 and 30% for the remainder of the study period. Along the transect, peak values corresponded to one of the two major rain events (21–23 November, 1991, and 30 November–3 December, 1991).

SOIL MOISTURE GRADIENTS

The period of September through October, 1991, was the second driest two month period on the 63 year rainfall record at Coweeta. Considering the severity of the drought as well as the amount of rain (24.3 cm) falling between 11 November and 2 December, 1991, these

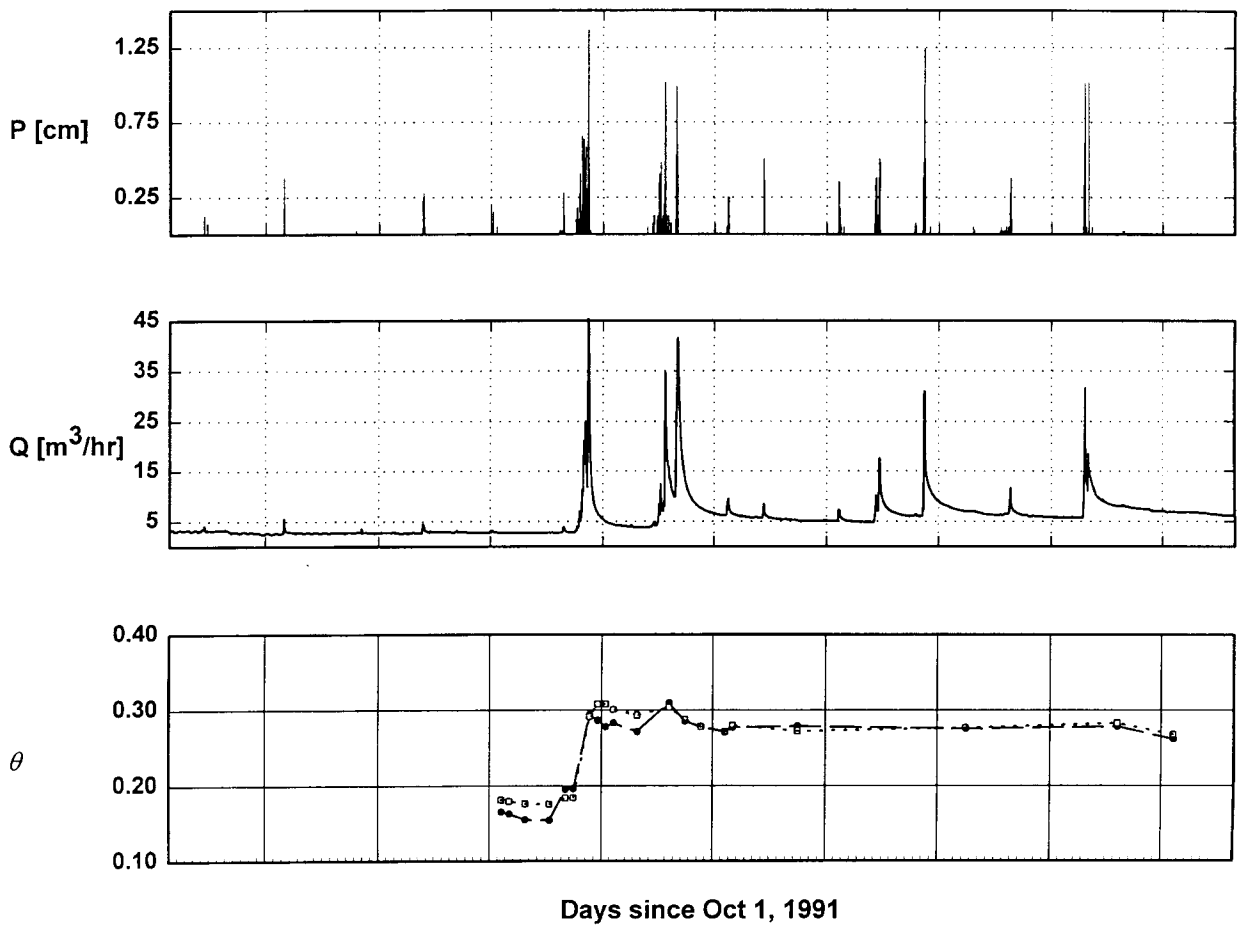

Fig. 1. *Precipitation (top) and streamflow (middle) for WS 2 from 1 October, 1991, to 10 February, 1992. The bottom graph shows corresponding average soil moisture on a WS 2 hillslope at two depths: 0–30 cm (-●-) and 30–90 cm (-□-).*

graphs approximate the range of soil moisture response for this hillslope profile. Soil moisture varied with position along the transect in both the 0–30 cm and 0–90 cm layers during the drought. Least squares regression of hillslope soil moisture content (Fig. 2) showed that a gradient was apparent in both layers before recharge (–14%, r^2 = .77, p<.01 in the 0–30 cm; –9%, r^2 = .47, p<.01 in the 0–90 cm). After recharge, the gradient was reduced in the upper layer (–10%, r^2 = .55, p<.01) and became insignificant for the 0–90 cm depth (–3%, r^2 = .08, p = .33). During sample collection, position along the hillslope never again explained more than 15% of the variation in soil moisture for the 0–90 cm soil layer.

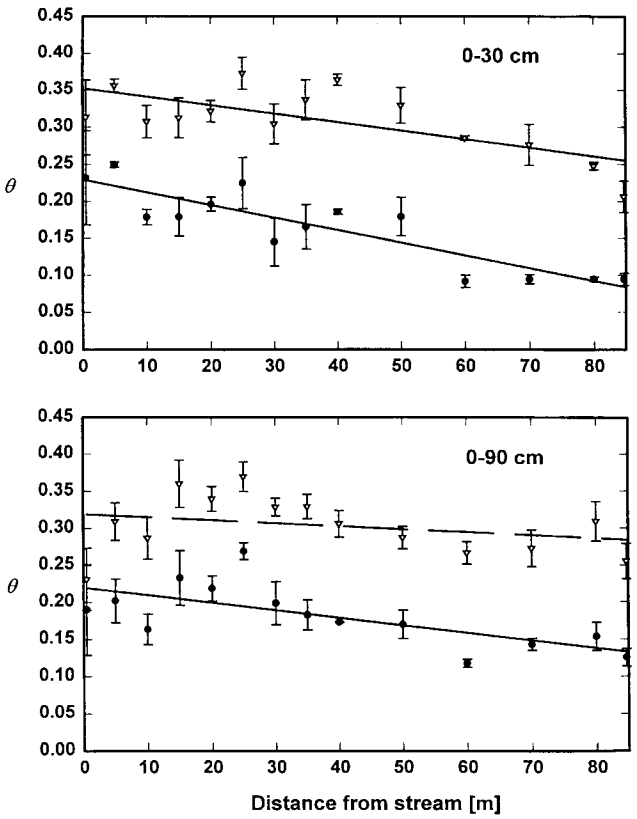

Fig. 2. *Soil moisture response ranges: from drought (11 November, 1991; -●-) through 24.3 cm precipitation recharge (2 December, 1991; -▽-). Each point is a mean of 3 measurements. Least squares regressions of soil moisture content on distance along hillslope in the shallow layer showed significant (p<.05) gradients before and after recharge. Regressions in the deeper layer showed a significant gradient during drought, but not after recharge.*

PHYSIOGRAPHIC CONTROLS

Measurement of soil properties showed bulk density for the upper 30 cm ranging from .81 to .96 g cm $^{-3}$ and soil organic matter in the A horizon ranging from 6 to 15% along the transect (Table 1). These values were more similar to those reported for Cullasaja-Tuckasegee than for Fannin soils (Thomas, 1996). Organic matter in both the O and A horizons was highest near the divide. The A horizon was thicker nearer the stream. The B horizon, however, was thickest in the lower midslope, 25–30 m from the stream. Soils were predominantly sandy loam and sandy clay loam, as expected. Clay content in the A and BA horizons was higher nearer the stream; however, clay content in the B horizon was highest both near the divide (70–80 m from stream) and in the lower midslope (20–25 m from stream). Overall, clay content was significantly higher (p<.01) in the BA and B horizons than in the A and BC horizons (Table 1).

Stepwise regression of the 0–30 cm layer moisture measurements on physiographic variables found three significant independent variables: topographic index, organic matter index, and storage index for 0–30 cm. Stepwise regression of the 0–90 cm layer moisture measurements on physiographic variables found only the storage index for 0–90 cm and the topographic index significant. For the shallow depth (0–30 cm), the topographic index was positively correlated with soil moisture in 2nd order partial correlation analysis for the entire measurement period (Fig. 3, top). Topography alone explained from 40 to 72%

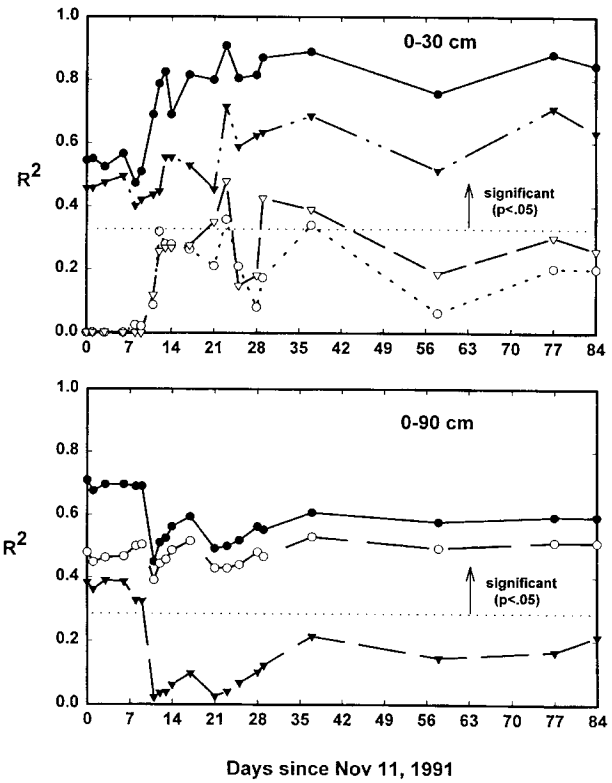

Fig. 3. *Squared partial correlations for physiographic variables vs. soil moisture content during the measurement period. Physiographic variables shown are: topographic index (-▼-), soil storage index (-○-), and organic matter index (-▽-). Also shown are adjusted r^2 values for multiple regressions of physiographic variables on soil moisture (-●-); only those variables that were significant in partial F tests (p<.05) were included.*

of the variation in soil moisture. Soil property variables were significantly correlated during the recharge event, but explained less variation than topography throughout the measurement period for the 0–30 cm soil layer.

For the 0–90 cm depth (Fig. 3, bottom), topographic and storage indices each explained at least 33% of the variation during drought. During recharge (22 November–2 December, 1991), control shifted markedly to the storage index (Fig. 3, bottom). After recharge (2 December–3 February, 1991), the topographic index began to show higher correlations once again, although none were statistically significant. The storage index remained the only significantly correlated control over soil moisture in the period after recharge, explaining from 39 to 53% of the variation in soil moisture distribution for the 0–90 cm layer.

Discussion

SOIL MOISTURE RESPONSE

These results suggest an extreme case for hillslope soil moisture gradients in the southern Appalachians. This relatively short hillslope exhibited a mesic-to-xeric vegetation species gradient without significant variation in other primary controls on vegetation, such as aspect (i.e. solar radiation), elevation (i.e. temperature), or soil mineralogy (i.e. nutrient availability) (Whittaker, 1956; Strahler, 1972). The measurement period coincided with the period of the year that has the lowest average precipitation at Coweeta (Swift *et al.*, 1988). During autumn 1991, this seasonal drying effect was enhanced due to the unusually severe drought, only 20% of the longterm average rainfall. Such an extreme drought followed by 24 cm rain within two weeks allowed capture of the soil moisture response range on the instrumented hillslope over a short sampling period.

In addition to the drought, late summer and early autumn evapotranspiration lowered average soil moisture content on the hillslope. Due to closed canopy forest over the full length of the hillslope, it is unlikely that late summer transpirational differences along the slope significantly affected the gradients found prior to recharge, particularly in comparison with topographically-driven drainage. Diurnal evapotranspirational effects on WS 2 streamflow largely ended at least a month before recharge, during mid October, 1991 (Fig. 1). The timing of the measurement period minimized the effect of transpiration due to leaf fall having already occurred from the dominant deciduous canopy trees on the hillslope. Transpirational losses during the period of recharge from the evergreen understory were relatively insignificant, as incident solar radiation for this eastern aspect and temperate latitude is low near winter solstice (Swift *et al.*, 1975).

These results showing that upper soils had the largest response range (Fig. 2) correspond with the finding that ridge soils had the greatest amount of annual variation in soil moisture (Helvey *et al.*, 1972). Soil moisture content in the root zone (0–90 cm) was higher near the ridge than on the midslope after recharge, which also concurs with Helvey *et al.* (1972). The primary difference between studies was sampling resolution. Helvey *et al.* (1972) took monthly samples over a multi-year period at 3 hillslope positions broadly designated as cove, midslope and ridge. In contrast, this study sampled with replication at each of 14 points along a hillslope with a roughly 2 day timestep across the range of precipitation conditions.

This study represented the transient nature of soil moisture dynamics in response to large events on a hillslope previously drained by severe drought. Drainage due to topographic factors from higher, steeper portions of the hillslope has been found to have both vertical and lateral components in Coweeta soils (Hewlett and Hibbert, 1963; Gaskin *et al.*, 1989) and may produce a moisture gradient in upper soil layers, as shown in physical models at Coweeta (Hewlett and Hibbert, 1963). It was found in the present study that a soil moisture gradient persisted in the upper 30 cm throughout the measurement period. For the entire root zone (0–90 cm), however, a significant gradient was present only during the drought. These results indicate that root zone soil moisture gradients on steep humid forested hillslopes are ephemeral, occurring only with sufficient drought.

PHYSIOGRAPHIC CONTROLS

That an index of topography would be correlated with soil moisture distribution is supported by several studies (e.g. Anderson and Kneale, 1982; Burt and Butcher, 1988; Boyer *et al.*, 1990). For example, significant correlations between the topographic index, as well as with an index of landform convexity (i.e. plan curvature), and soil saturation depth above bedrock were found along a grass-covered hillslope (Burt and Butcher, 1985). Neither topographic index was entirely satisfactory, however, and they suggested that knowledge of varying soil depths would have been helpful. In another study of first order basins having both forest and pasture vegetation, topographic control on soil moisture was particularly evident on shallow slopes and in steeper areas where hillslope form was strongly concave (Petch, 1988).

Correlation between soil moisture content and an index of varying soil properties is also supported by others (e.g. Parker, 1978; Pierson, 1980; Helvey and Patric, 1988). Observations at the North Appalachian Experimental Watershed indicated soil type had more influence on soil moisture than relative elevation (Dreibelbis and Post, 1940). Work in the southern Appalachians showed that mean annual soil moisture in the top 2.1 m on ridges was comparable to that found in cove sites. Further, midslope soil moisture content was consistently lowest (Helvey *et al.*, 1972). More recent studies have shown variation in soil

properties to be at least as important as topography in determining soil moisture distribution (Boyer *et al.*, 1990; Afyuni *et al.*, 1993).

Our measurements of soil property distribution (Table 1) correspond with various studies over both plot and landscape scales. At the plot scale, soil clay content tends to increase with depth through the solum, then decrease or remain constant going from B to C horizons (Buol *et al.*, 1989). On the landscape scale (Gerrard, 1993), upper horizon thickness is generally smallest on the backslope (i.e. upper midslope) due both to high rates of erosion and low rates of soil profile development (Walker *et al.*, 1968). Summit or upland soils are less prone to erosion and are usually more clayey, as found by Afyuni *et al* (1993) for hillslopes in the North Carolina Piedmont and in studies of road effects on erosion in the Coweeta Basin (Swift, 1984). Footslope soils show the least amount of coarse fraction due to mechanical sorting of weathered drift into finer components (Conacher and Dalrymple, 1977).

Topography was the dominant control in the upper layer throughout the measurement period. The storage index was generally not significantly correlated with moisture content in the upper layer (Fig. 3, top). For the deeper layer, however, both indices were significantly correlated with soil moisture content prior to recharge. During drought and then again as drainage occurred after recharge, the influence of topography became more important as moisture moved deeper and downslope. The storage index, however, became the only significant control during and after recharge (Fig. 3, bottom). The higher clay B horizon along the transect was thickest in the lower footslope; also higher clay content was found in the B horizon near the divide than on the rest of the transect (Table 1). In other areas the more sandy BC horizon, with a lower moisture holding capacity, occupied a larger portion of the measured depth.

IMPLICATIONS

The results of this study indicate that both topography and soil properties play important roles in distributing soil moisture along hillslopes in humid watersheds. Variation in correlation between these controls is a function of antecedent soil moisture conditions driven by the frequency, intensity and duration of precipitation. Topographic factors assert more control during drier periods as drainage progresses, while variation in soil water storage properties is more important during and following rain events (Fig. 3). These results suggest that spatially distributed watershed hydrology models predicting soil moisture redistribution, as well as other hydrologic processes such as evapotranspiration, at fine resolutions within watersheds should be parameterized with plot-scale information on soil property variation in addition to the more standard information on topographic variation.

For the overall root zone on upland forested hillslopes, physical control on soil moisture gradients is suggested to vary between drainage and storage controls, with the variation driven by temporal patterns of precipitation. The magnitude of hillslope soil moisture gradients in this study was a function of local slope, upslope contributing area, thickness of horizons with higher clay content, and depth in the soil. Drainage due to topographic factors from higher, steeper portions of the hillslope may produce a moisture gradient in the upper soil layers. Deeper soils have a lagged response due to percolation time in the upper layers, as well as having more clay content and lower hydraulic conductivities due to diminished macroporosity. As a result, deeper soil layers may not show hillslope moisture gradients during short duration droughts. Further, ridge soils with more extensive clay layers than midslope soils may retain soil moisture longer, delaying topographically driven drainage. These factors cause soil moisture gradients along hillslopes in humid watersheds to be ephemeral and primarily restricted to upper soil layers.

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