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Spatial variation and temporal stability of soil water in a snow-dominated, mountain catchment

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Abstract:

Soil is a critical intermediary of water flux between precipitation and stream flow. Characterization of soil water content (θ, m^3 m^{-3}) may be especially difficult in mountainous, snow-dominated catchments due to highly variable water inputs, topography, soils and vegetation. However, individual sites exhibit similar seasonal dynamics, suggesting that it may be possible to describe spatial variability in terms of temporally stable relationships. Working in a 0·36 km^2 headwater catchment, we: (i) described and the spatial variability of θ over a 2 year period, (ii) characterized that variability in terms of temporal stability analysis, and (iii) related changes in temporally stable soil water patterns to stream flow generation. Soil water data were collected for 2 years at representative sites and quantified in terms of θ and water storage to a depth of 75 cm (S_{75}, cm). Both S_{75} and θ were normally distributed in space on all measurement dates. Spatial variability was high relative to other studies, reflecting catchment heterogeneity. However, the ranking of S_{75} values displayed temporal stability for all site locations, seasonally and annually. This stability was attributed to soil texture. Further temporal analysis indicated that estimates of catchment mean and standard deviation of S_{75} may be characterized with relatively few measurements. Finally, we used temporal linear regression to define catchment soil water conditions related to stream-flow generation. Static, high S_{75} conditions in late winter and early spring indicate that stream-flow response is highly sensitive to inputs, whereas static, low S_{75} conditions in late summer and early fall indicate minimum stream-flow sensitivity to water inputs. The fall transition was marked by uniform S_d across the catchment. The late spring transition was marked by nonuniform S_{75} decreases, with the highest S_{75} sites decreasing most. Threshold S_{75} values identifying catchment sensitivity to water input were identified. Copyright © 2004 John Wiley & Sons, Ltd.

KEY WORDS soil water; spatial organization; temporal stability; snowmelt; runoff

INTRODUCTION

Soil stores intermittent water inputs from rain and/or snowmelt and slowly releases that water to streams, vegetation, and groundwater recharge. This release is partly controlled by soil characteristics that modify and modulate stream flow timing and quantity. Snow in headwater catchments in the intermountain west region of the USA provides most of the water supply to the region for irrigation, hydroelectric power, recreation, fish habitat, municipal drinking water, etc. The interactions and feedback between soil water, plants, and climate affect vegetation yield, carbon cycling, and the energy balance, all of which have implications for local land management in terms of grazing, logging, recreation, and wildlife.

Point-scale measurements of volumetric soil water (θ, m^3 m^{-3}) impose a fundamental constraint to quantifying the role of soil water in watershed hydrology. Variability of point measurements at a given time (spatial variability) results in uncertainty in estimation of larger scale values. Aside from variability due to the measurement technique used, spatial variability may be attributed to variability of landscape characteristics (e.g. soil type, topography), or simply may be random (Burrough, 1993). Reynolds (1970) characterized θ

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variability in small plots using gravimetric surveys and attributed variability to two sources: static properties (e.g. soil type) and dynamic influences (e.g. precipitation). Seyfried (1998) identified vegetation, soil type and elevation (a surrogate for climate) as effective sources of variability at different scales in the Reynolds Creek Experimental Watershed (RCEW) in Idaho. In general, the spatial variability of θ in mountainous regions is expected to be high relative to other landscapes due to heterogeneous conditions of surface and bedrock topography, soil characteristics, wind patterns affecting snow deposition, and vegetation.

In addition to the amount of variability, the spatial distribution or organization of θ may have a critical impact on hydrological processes such as stream-flow generation. Dunne *et al*. (1975) documented regions of seasonal variation of soil saturation as a runoff-generating mechanism in humid areas. Grayson *et al*. (1997) found that θ patterns in a rain-dominated, temperate region of Australia switch seasonally from a random pattern, which generated no runoff, to a spatially organized pattern dictated by topography during periods of runoff generation. Working in southern Idaho, McNamara *et al*. (2004) found that source areas for stream flow depended on the pattern of θ.

Statistical approaches for the description of spatial variability include geostatistics (Cressie, 1991), connectivity statistics (Allard, 1994) and temporal stability analysis (Vachaud, 1985; Kachanoski and de Jong, 1998). Geostatistics and connectivity statistics require relatively large data sets. With geostatistics, the semi-variograms used to determine the spatial correlation structure within stationary spatial fields are highly dependent upon the quantity and spacing of data pairs. Russo and Jury (1987a,b) approached the issues of semi-variogram estimation by using simulated data and found important constraints on the technique, such as a minimum lag distance of half the correlation range or less and that data sets with less than 72 data pairs per lag distance resulted in 25% or greater error in range estimation. Yates and Warrick (2002) compiled many field studies using semi-variogram techniques, citing that a minimum of 30 pairs per bin is necessary for accurate characterization of the semi-variance.

Temporal stability analysis, which was introduced by Vachaud *et al*. (1985), describes the persistence of spatial patterns of θ over time in three European agricultural fields. Vachaud *et al*. (1985) found that soil texture was responsible for the high degree of temporal stability they observed. They went on to note that some locations represented statistical parameters of the normal probability distribution, such as mean and standard deviation (SD) of θ, potentially reducing the number of measurements necessary to characterize θ distribution. Kachanoski and de Jong (1988) expanded the definition of temporal stability, considering it to be defined as the degree of linear correlation between θ measured at consecutive times. They also showed that the temporal stability of soil water recharge is scale dependent at their site, such that drying periods showed scale independence and wetting periods are scale dependent. Others have applied these analyses with varying degrees of success (e.g. Comenga and Basille, 1994; Grayson and Western, 1998; Gomez-Plaza *et al*., 2000; Martinez-Fernandez and Ceballos, 2003). Notably, Grayson *et al*. (1997) found that some specific patterns of temporal stability are indicative of stream-flow generation and referred to these patterns as preferred states. To date, relatively few temporal stability analyses have been performed in areas of complex terrain or in regions affected by snow processes.

This study is part of a larger project designed to improve our understanding and prediction of stream-flow timing and amount in mountainous, snow-dominated watersheds. Our objectives were to: (i) describe the spatial variability of soil water in a snow-dominated watershed; (ii) characterize that variability in terms of temporal stability; (iii) examine the relationship between spatially and temporally varying soil water and stream flow dynamics.

MATERIALS AND METHODS

Site description

The location of this study is the Reynolds Mountain East (RME) catchment at the southern edge of RCEW in the Owyhee Mountains. The RCEW was established in 1960 by the Agricultural Research Service (ARS),

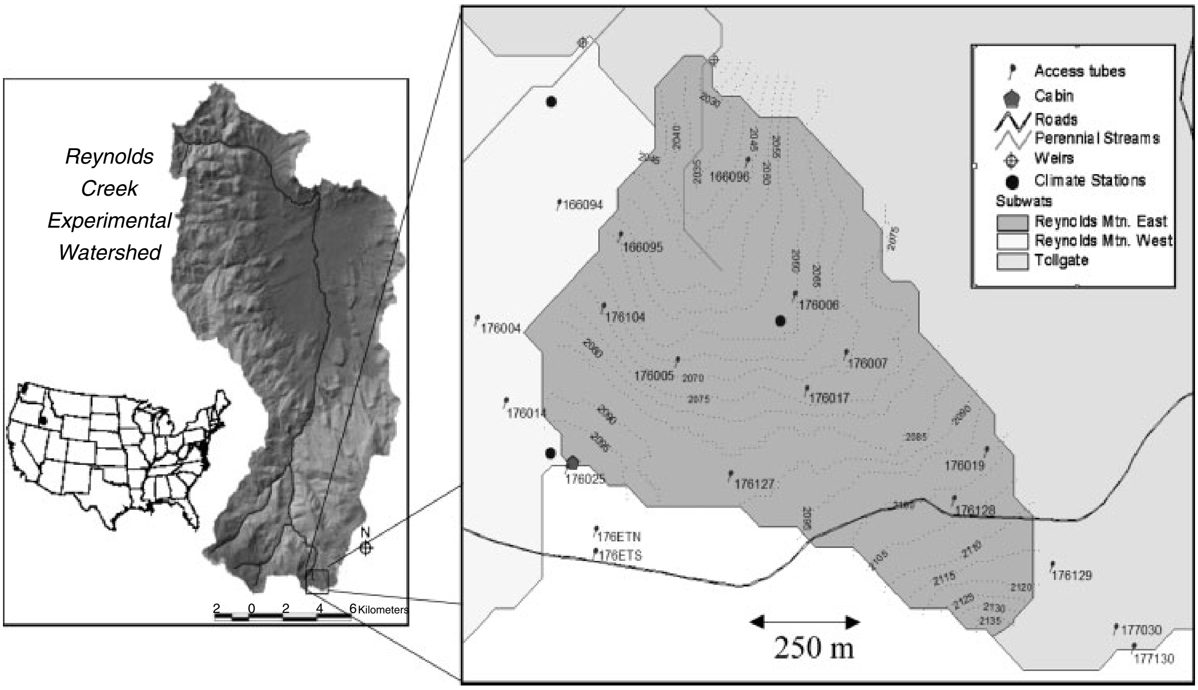

Figure 1. Site location map of the RCEW and the RME catchment

US Department of Agriculture, and is located 80 km southwest of Boise, Idaho (Figure 1). Reynolds Creek is a perennial stream with a watershed area of 239 km^2, flowing north to meet the Snake River (Seyfried *et al.*, 2001a).

The RME catchment encompasses a 0·36 km^2 portion of the Reynolds Creek headwater region with elevations ranging from 2020 to 2140 m (Figure 1). Slopes vary from nearly level to 40%. Soil textures range from fine loam to clay with intermixed cobbles. Surficial loess deposits of varying thickness are present throughout the watershed. Both clay and coarse fragments tend to increase with soil depth towards fractured bedrock. Parent material consists of basalt and latite. Rocky outcroppings generally coincide with ridges, and soil depth may exceed 3 m in forested regions. Vegetation at the RME catchment is dominated by Vaseyana big sagebrush, with scattered juniper, patches of quaking aspen and Douglas fir. Willows line the riparian corridors. Plant density is highly variable, ranging from dense thickets near the stream, to relatively open fir forest to sparse sagebrush cover dominated by bare ground.

The main drainage is oriented south to north. Average annual precipitation at RME catchment is 866 mm (averaged over 30 years) and concentrated in the winter months, with July and August typically very dry (Hanson *et al.*, 2001). Rainfall is generally spread evenly over the watershed. Snowfall, accounting for over 70% of the mean annual precipitation (Hanson and Johnson, 1993), is strongly affected by wind patterns and amasses in drifts (Marks and Winstral, 2001). The snowpack is persistent through the winter and spatially continuous, except on a few exposed ridges. Areas of drifted snow contribute 15–20% of the snow water input, although only covering 9% of the land area; sheltered areas comprise another 25% of snow water input (Marks *et al.*, 2002). The main stream in the RME catchment is perennial with mean annual flow of 523·1 mm, averaged from 1963–96 data, a period nearly congruent with those data used for precipitation. The greatest flows are nearly concurrent with the maximum snowmelt events, typically occurring in late April to early May (Pierson *et al.*, 2001). Flow diminishes during the dry summer months to a very low discharge ($<0{\cdot}000\,14$ m^3 s^{-1}) that remains consistent into fall and early winter.

Data acquisition and quality

Point-scale soil water data were obtained using a neutron probe, manufactured by Troxler Electronic Laboratories, Inc., series model 4300. The method has a long history (Gardner and Kirkham, 1952) and the strengths and weaknesses are well documented (Hignett and Evett 2002). Since soil water dynamics occur in context of specific soil structure and extant vegetative roots, disturbance by the measurement method may skew data. The neutron probe accesses the soil via an imbedded tube, commonly made of aluminium, and measures θ in a spherical volume roughly 15 cm in diameter. The access tube sites in RME catchment were installed almost 40 years ago, thus any disturbance is now imperceptible. Neutron probe measurements are unaffected by frozen soil conditions; thus, aluminium tubing was extended above the ground surface for this recent survey, allowing for winter measurement. Locations are distributed throughout the catchment or immediately adjacent to it and are intended to represent different topographic, soil and vegetative characteristics within the catchment (Table I), as discussed in Stephenson and Zuzel (1967). Two extreme regions lack coverage for logistical reasons: rocky ridges have little soil to secure a tube, and areas of drifting snow often bury and bend the aluminium access tubing.

Soil water data were collected for two waters years, WY2001 and WY2003. The access sites vary in depth from 55 to 330 cm, according to soil thickness. Measurements consisted of 30 s neutron counts at depths of 15 cm, 30·5 cm, and then at intervals of 30·5 cm, providing continuous θ information through the root zone to bedrock, allowing for water storage calculations of the column. We calibrated the probe using a polyethylene standard (see Seyfried *et al*. (2001b)). On-site tests of water storage using a lysimeter confirmed the calibration accuracy in terms of differences in soil water storage (Seyfried *et al*., 2001b). Precision testing, using 40 30 s counts, yielded an SD of less that 0·01 $m^3\ m^{-3}$ for wet (0·45 $m^3\ m^{-3}$) and dry (0·1 $m^3\ m^{-3}$) soils. Snow depth was also measured at each access tube site when accumulation was present.

Table I. Access site characteristics. If no maximum snow depth is given, the access site was inactive for the year. N/I indicates that no information is available

	Neutron tube	Elevation (m)	Tube depth (cm)	Soil texture, average over depth (%)				Vegetation type common name	Maximum snow depth (cm)	
				Sand	Silt	Clay	Coarse		WY2001	WY2003
1	166094	2058	190	20	47	25	8	Douglas fir		78
2	166095	2060	285	N/I	N/I	28	59	Mountain sagebrush	87	45
3	166096	2045	185	25	13	25	37	Sagebrush/snowberry	80	53
4	176004	2076	111	13	10	20	57	Mountain sagebrush		35
5	176005	2069	358	13	10	20	57	Aspen/Douglas fir	100	44
6	176006	2060	127	30	20	45	5	Aspen	97	59
7	176007	2073	72	20	34	33	13	Mountain sagebrush	71	46
8	176014	2091	114	38	5	17	40	Douglas fir/Mnt. sagebrush		56
9	176017	2075	195	22	14	33	31	Mountain sagebrush		138
10	176019	2103	135	25	13	31	31	Sparse mountain sagebrush		43
11	176025	2099	100	38	5	17	40	Forbs	65	25
12	176104	2069	88	10	10	20	60	Mountain sagebrush	85	48
13	176127	2084	175	25	13	31	31	Mountain sagebrush	100	69
14	176128	2118	81	25	30	30	15	Sagebrush/snowberry	90	61
15	176129	2122	179	25	30	30	15	Dense mountain sagebrush		56
16	176ETN	2092	112	30	6	24	40	Mountain sagebrush/forb	88	23
17	176ETS	2087	112	30	6	24	40	Mountain sagebrush/forb	58	31
18	177030	2133	181	25	13	31	31	Low sagebrush		41
19	177130	2135	273	25	5	30	40	Sparse mountain sagebrush		40
20	176627	2085	157	25	10	35	30	Mountain sagebrush	85	

Data are available for 12 locations in the RME sub-basin during WY2001 (Tribbeck, 2002). Data acquisition was limited to those sites that were outfitted with extensions. Also, a faulty cable prevented measurement at depths greater than 1·2 m during the first two dates of measure (29 November 2000 and 8 January 2001). We expanded the measurement site network to 19 access tube sites in WY2003, 11 of which coincided with sites from WY2001. During the spring of 2003, water flooded the access tubes at sites 176 128 and 177 130, rendering the deepest position immeasurable from 28 April 2003 to the end of WY2003. The manual measurement technique limited the number of access sites measurable in a single day. This constraint and the remote location caused variation of temporal resolution of soil water data from 2 weeks (during wet periods) to 2 months (during the dry season) between measurements. Thus, we did not observe short time-period fluctuations, such as diurnal variation and immediate water input response.

Two climate stations, each with two types of precipitation gauge (shielded and unshielded), are located in the watershed. These provide standard climate data, including air temperature, relative humidity, dew-point temperature, vapour pressure, and solar radiation. Stream flow is measured with a 90° V-notch weir instrumented with a stage recorder and covered by a heated shelter to prevent freezing.

Data analysis

Spatial variability of θ measured at different depths was quantified using the mean and SD statistics, which allows for comparison of data collected in this study with that in others. In addition, the total water stored in the soil column S to depth d was calculated for the greatest common depth of 75 cm for each site during both years as

$$S_d = \sum_{i=1}^{n} \theta_i z_i \tag{1}$$

where z is the thickness of the soil layer measured, θ is the measured volumetric soil water content, n is the number of soil layers measured and the subscript i is the layer measured (following Seyfried *et al.* (2001b)). Given the precision of θ measurements, S_{75} measurement SD ranged from 0·2 cm for dry soils to 0·4 cm for wet soils.

We did not employ commonly used geostatistical analysis in this study because of the insufficient sample size (e.g. Yates and Warrick, 2002) and assumed that the data are spatially independent, which we consider to be reasonable for the following reasons. First, in a previous study on the catchment, Pandit (1999) found that the correlation length of θ was on the order of 30 m, which is much less than the spacing between almost all of our measurement sites. Second, we found that a comparison of the difference between S_{75} values measured at different sites and the distance between those sites yielded no trend, consistent with a relatively short correlation length. Third, the sampling design was intended to represent each landscape feature within the catchment and not the spatial variability within those characteristics, the locations of which are not necessarily spatially correlated. For example, sites characterized by deep soils with mountain sagebrush on north aspects are found at opposite ends of the catchment and are separated in space by the full range of conditions found in the catchment.

The mean relative difference $\overline{\delta}(j)$ was computed for each location for each water year by computing the time average of

$$\delta_t(j) = \frac{S_t(j) - \overline{S_t}(j)}{\overline{S_t}(j)} \tag{2}$$

Thus, $\delta_t(j)$ is the difference between the measured and expected value (i.e. $\overline{S_t}(j)$), divided by the expected value. By plotting these values according to rank, this analysis determines whether certain sites are consistently indicative of basin characteristics such as mean soil water content, theoretically reducing the necessary measurements to determine the distribution of soil water.

Using the S_{75} values, temporal stability analysis was performed on every site pair (both intra-annual and interannual). The first test was the Spearman rank test, a non-parametric statistic proposed by Vachaud *et al.*

(1985). For a given measurement date, S_{75} values were ranked, $R_t(j)$, according to total water stored at time t for each location j. Successive times were compared using the correlation coefficient r_s:

$$r_s = 1 - \frac{6\sum_{i=1}^{n}(R_{t_2}(j) - R_{t_1}(j))^2}{n(n^2-1)} \quad (3)$$

where n is the number of spatial data points and r_s gauges the ranking observed at t_2 described by t_1. Kachanoski and de Jong (1988) suggested that linear regression and correlation are indicators of temporal stability. Storage values at two dates (t_1 and t_2) are regressed according to

$$S_{t_2}(j) = a_{t_2-t_1}S_{t_1}(j) + b_{t_2-t_1} + \varepsilon_{t_2-t_1} \quad (4)$$

where $a_{t_2-t_1}$, $b_{t_2-t_1}$, $\varepsilon_{t_2-t_1}$ are the regression slope, intercept and error respectively, and $S_t(j)$ is the value of soil water at time t at location j. Pearson's correlation coefficient r_P was used to evaluate the degree of correlation. Note that values of r_s (rank) and r_P (correlation) are not necessarily correlated. The cumulative probability function of S_{75} was also computed using ranks and the assumption of an underlying normal probability distribution.

We examined temporal stability further using the linear regression parameters of a and b in Equation (4). Since a is a linear function of independent measurements of S_{75}, which are normally distributed, confidence intervals may be calculated, testing for a significantly different from one (Devore, 1995). By analogy, b may be tested for b significantly different than zero. We used a two-tailed t-test, with $\alpha = 0{\cdot}05$ for all analyses.

Given that both r_s and r_P show significance, we describe four temporal stability conditions, defined by combinations of a and b (Table II). Condition 1 occurs if a and b are not significantly different than one and zero respectively. This condition implies there is effectively no net change in S_{75} between the two dates compared, and may occur due to relatively little water input or output, or if output from each point equalled the

Table II. The four conditions possible according to the significance of the parameters from linear regression[a]

	Slope	Intercept	Explanation
Condition 1	Significantly near 1	Significantly near 0	Water contents at different times are nearly identical; may be due to relatively small input/output or to a extended period that buffers catchment response, i.e. time erases the memory of the antecedent condition
Condition 2	Significantly near 1	Significantly different than 0	Nearly uniform change in soil water pattern over catchment, caused by uniform input, such as rain, and/or output such as shallow evaporation and drainage from the soil column
Condition 3	Significantly different than 1	Significantly near 0	Change in watershed pattern is nonuniform over time, in the absense of uniform input or output, e.g. the intercept slays the same and while some locations are drying or wetting more than others or redistribution is occurring
Condition 4	Significantly different than 1	Significantly different than 0	Nonuniform change in catchment soil water pattern caused by nonuniform input (snowmelt) or output (transpiration), or combination of uniform input/output and redistribution.

[a] Significance is calculated using a two-tailed confidence interval with $\alpha = 0{\cdot}05$. No significance (NS) occurs when either r_s or the r_P are not significant. The explanations given are based upon changes that would cause the associated condition.

input. Condition 2 is characterized by b significantly different than zero while a is not significantly different than one. This condition occurs when changes in S_{75} are approximately equal at each measurement site, as might result from a uniform rain or uniform plant water uptake. Condition 3 applies when b is zero and a is different from one. This implies that the mean S_{75} changes between measurement dates and that the changes were not evenly distributed, but rather that the wettest sites changed more than the drier sites. For Condition 4, a is significantly different from one and b is significantly different from zero, indicating nonuniform changes between measurement dates (e.g. snowmelt, transpiration). Note that, for each of these conditions, the pattern of ranks and values are temporally persistent, but the mean and/or spread among values may change.

RESULTS AND DISCUSSION

Spatial trends and variability

Seasonality of climate produces general trends in θ that have long been observed on the RCEW (e.g. Seyfried, 1998; Seyfried *et al*., 2001b). The entire catchment dries to minimum values in the late summer, with nearly all soil depths reaching permanent wilting point (0·03–0·1 $m^3 m^{-3}$). Fall rains and short-lived snow moisten the upper layers of the soil. The snowpack further moistens the soil column with small, mid-winter melt events, mostly from diurnal temperature variations. Field capacity (0·15–0·25 $m^3 m^{-3}$) is reached in mid winter. Spring snowmelt brings θ to maximum values, near saturation (0·35–0·45 $m^3 m^{-3}$). Soil water progressively decreases through the summer shortly after the snowmelt.

These patterns are repeated with some variation each year. The 2 years of soil water data in this study were distinct due to different weather patterns (Figure 2). WY2001 did not show a significant increase in soil water until late February, because most of the precipitation arrived as snow. WY2003 includes an example of a rare winter rain-on-snow event that caused an early rise in θ.

The spatial variability of θ, as indicated by the SD, was roughly constant during the year, and much greater at 120 or 150 cm than at 15 cm (Figure 2, Table III). This was consistent with the general trend of increased spatial variability with depth (Figure 3). This variability was roughly independent of the average θ for all depths except the 15 cm. Other studies showed some dependence of the SD on θ, but these data were rarely collected at depths below 30 cm (e.g. Famiglietti *et al*., 1998). The increase in SD with depth was partly

Table III. Summary of soil water statistics for WY2001 and WY2003 at two depths (cm) each. Units are ($m^3 m^{-3}$) × 100(%). Dates were chosen to represent seasonal variation of soil water

Parameters	WY2001					WY2003				
	23 Jan	26 Mar	26 Apr	5 Jul	25 Oct	28 Jan	18 Mar	28 Apr	10 Jul	25 Sep
15 cm						15 cm				
Mean (%)	20·8	25·4	25·2	5·4	8·8	27·8	29·8	28·5	11·0	9·5
Minimum (%)	16·6	19·5	17·4	1·4	2·4	20·1	22·5	18·5	4·9	4·8
Maximum (%)	25·4	33·6	35·5	9·5	15·5	36·6	36·9	37·5	16·6	17·9
Range (%)	8·7	14·1	18·1	8·1	13·1	16·5	14·5	19·0	11·7	13·2
SD (%)	3·3	4·2	5·0	2·2	4·2	4·0	3·5	4·6	3·1	3·2
Observations	12	12	11	12	12	17	19	19	19	19
120 cm						150 cm				
Mean (%)	20·6	25·7	32·7	26·5	20·3	25·6	28·0	31·4	27·1	21·7
Minimum (%)	7·5	13·7	18·6	14·5	5·9	12·6	12·8	18·6	13·4	8·1
Maximum (%)	41·7	50·8	49·2	46·7	41·8	42·9	46·1	50·9	45·6	41·0
Range (%)	34·1	37·1	30·6	32·2	35·9	30·3	33·3	32·3	32·2	32·9
SD	10·4	12·2	9·7	10·1	10·9	9·8	11·1	9·9	9·4	9·7
Observations	7	8	8	8	8	7	9	9	9	9

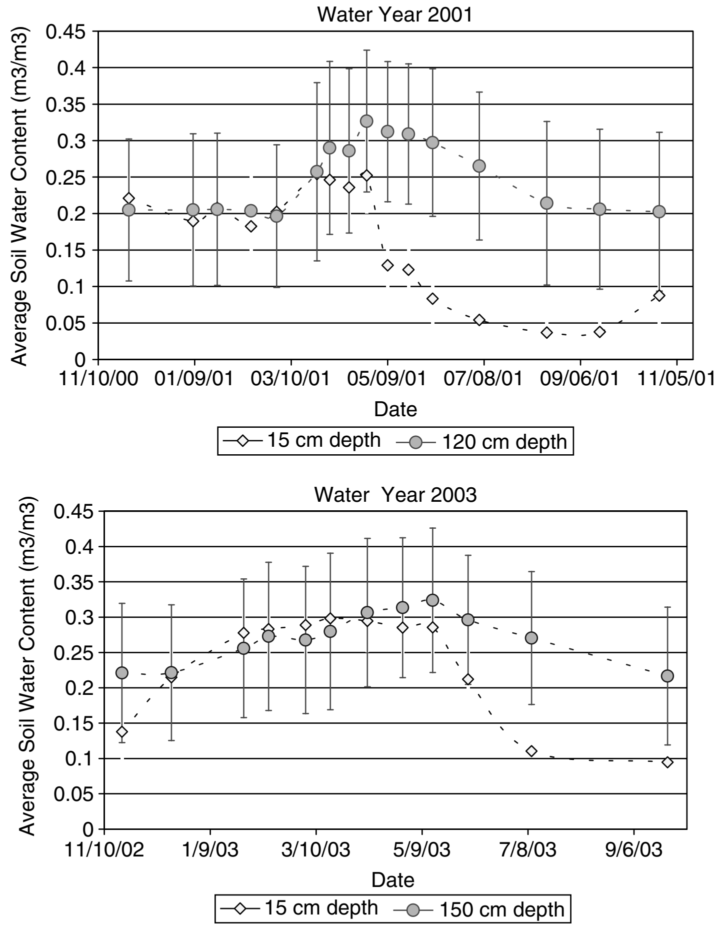

Figure 2. Average water content for WY2001 and WY2003. Error bars represent one standard deviation. Data are connected by a spline function to show temporal connection of the successive measurement for visualization, which does not represent the full time series of soil water. Number of samples varies by depth and measurement date, as given in Table III

attributed to the relatively uniform textures characteristic of loess deposits near the surface, relative to the variable mixtures of high clay content and high coarse fragments, which may range from 10% to nearly 100% as depth increases. Another factor causing an increase in spatial variability with depth was that one of the sites was influenced by groundwater, resulting in unusually high θ values at greater depths. In general, the SDs measured were high relative to most studies (Western *et al.*, 2003), which may be attributed to the wide range of conditions within the watershed.

These levels of spatial variability have important implications for characterizing the soil storage component of the overall water balance. Given an SD of 0·1, and if 10 measurements are made, the 0·95% confidence interval is $\pm 0{\cdot}0715$ m^3 m^{-3}, which is roughly the entire range of values measured. Even with 30 measurements, the interval is 0·037 m^3 m^{-3}, which is still about half the total range. Thus, regardless of the technique used, accurate characterization of soil water within the watershed is a very costly proposition unless more is understood about the nature of the spatial variability.

Despite relatively high spatial variability, the temporal patterns at each site were very similar. A general progression of wetting from top to bottom was observed at all sites, but this was modified at one site by groundwater. We illustrate the general trends and intersite variations in soil water dynamics with three

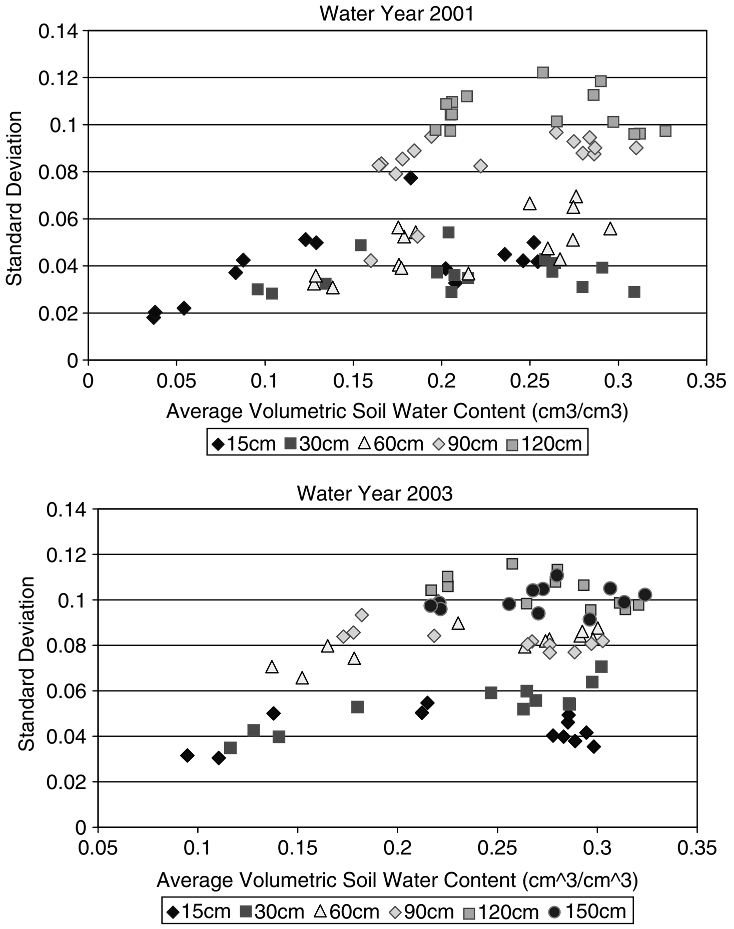

Figure 3. SD as a function of average soil water content, WY2001 and WY2003. Both years of measurement show similar trends of SD as a function of average soil water over all dates, for various measurement depths within the soil column

representative sites during WY2001 (Figure 4). At site 176 127, soil water progressed through the soil column during the fall and winter, with the peak of all layers in mid spring, then dried through the summer. There was also a damping effect of the water input signal with depth, as seen from the smoothing of the two major peaks at 90 and 150 cm. The trend of delayed and damped response was also established in wells elsewhere in RCEW (Seyfried, 1998). The dynamics for site 176 006 were distinct from other sites, in that θ increased progressively with depth and the site maintained exceptionally high θ values at depth even in summer. Nearly linear drainage indicated the influence of a water table. This was supported by field observations of gleyed soil in the vicinity and nearby ephemeral stream flow. The final example, site 176 005, was the deepest site in the study area at 330 cm depth. This site consistently displayed relatively low θ at 210 and 300 cm, suggesting fractured bedrock with low porosity. There was also a delayed θ response at those depths, reflecting the slow progression of the wetting front, followed by a slow decline indicative of drainage due to gravity as opposed to plant uptake.

Temporal stability

Mean relative differences of S_{75} were calculated (Equation (2)) and values were plotted in increasing order to show how individual site values deviate from the mean during the year (Figure 5). Error bars indicate the

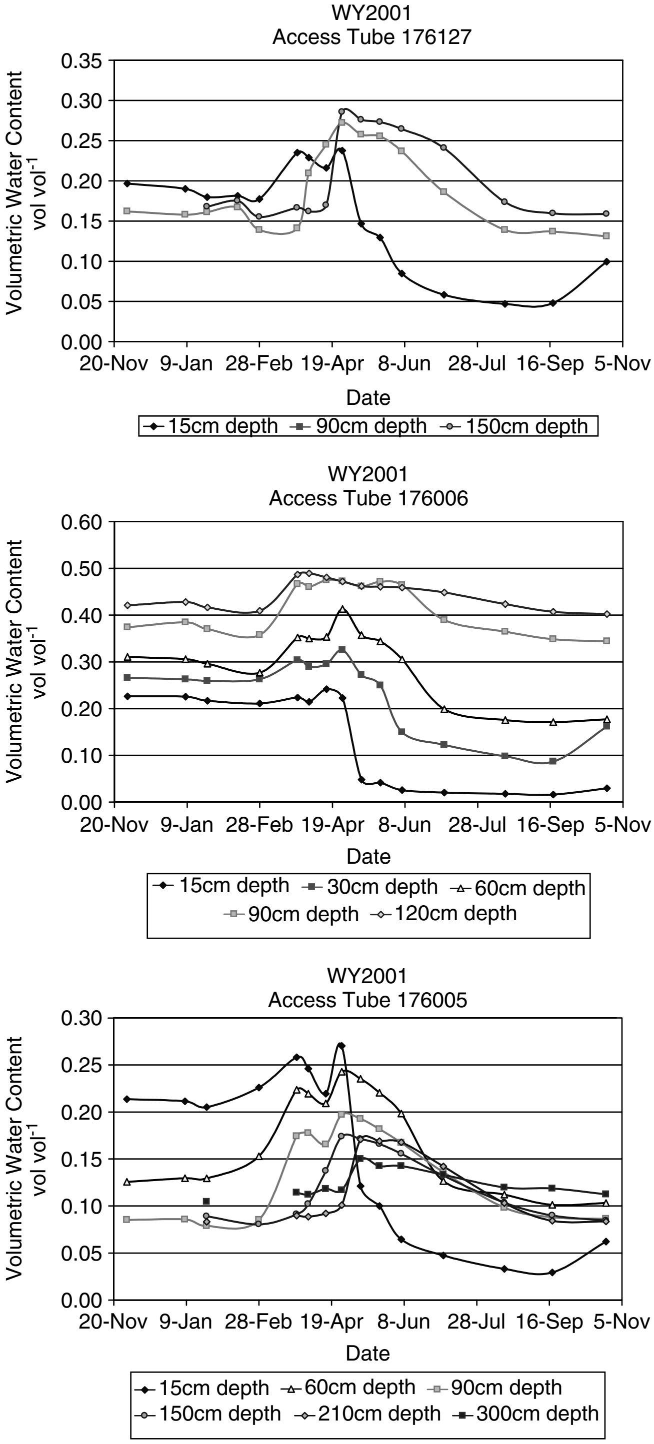

Figure 4. Three examples of the WY2001 displayed with values from the measured soil layers showing distinct soil water dynamics with overall similar annual trends of soil water

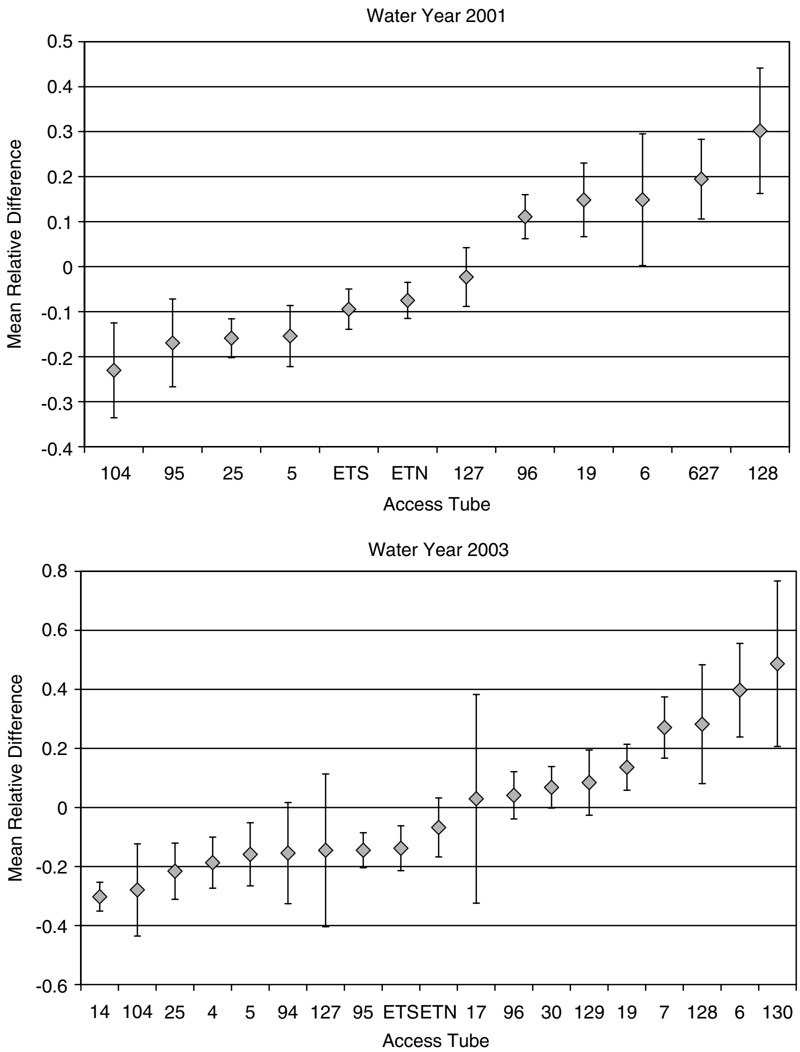

Figure 5. Plots of mean relative differences for S_{75}, for WY2001 and WY2003. Vertical bars are associated temporal SD. Site location labels are truncated to the last three characters

associated temporal SD. The relative differences for the set of access sites are fairly symmetric about the mean, with a slight tendency for fewer, larger values above the mean. The temporal variability, as demonstrated by the temporal SD, does not appear to be preferential towards wetter or drier locations. Interannual comparison shows consistency between sites, with sites common to both years occupying similar positions. Site-specific temporal SD is also consistent between the 2 years. These data indicate that individual sites not only exhibited similar temporal dynamics, but that the relative ranking of the sites in terms of S_{75} remained fairly constant throughout the year and for different years. Thus, the high spatial variability displayed in Figure 2 is largely the result of individual site values progressing parallel to one another during the season rather than due to random scatter.

The cumulative probability plots are given for dates of extreme conditions, i.e. dry and wet, for both years of record, as well as for an approximately average date (Figure 6). The measured S_{75} at site 176 127 (note that site labels are truncated to three digits in Figure 6) for both years is consistently near the mean, regardless of

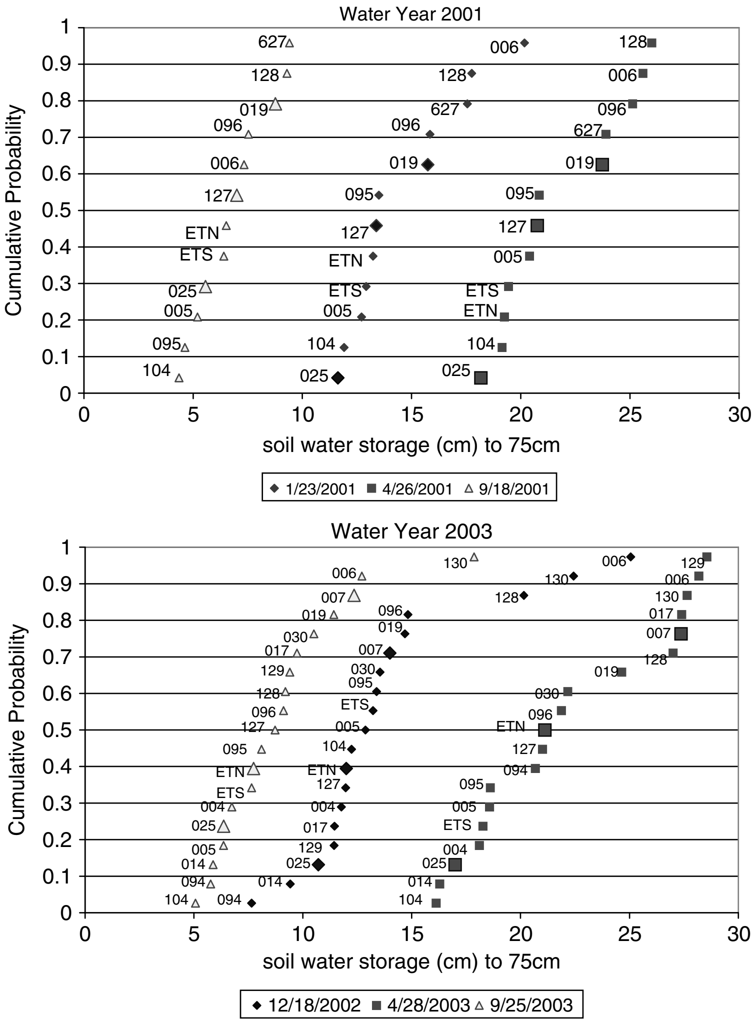

Figure 6. Cumulative probability plots for WY2001 and WY2003; different realizations of water storage are given, showing extreme and median soil water conditions. Labelled data points indicate sites with values that represent mean, +1SD, and −1SD of the respective data

conditions. This indicates that the mean S_{75} on any given date may be estimated from a single measurement at that site. The catchment variability can be similarly estimated using sites consistently approximately one SD above (site 176 019) and below (site 176 025) the mean. Thus, it is possible to estimate soil water statistics for the catchment from a limited sample size for the years of data.

For purposes of extrapolating values within the watershed and understanding the processes that affect soil water variation, it is of interest to know what landscape features are correlated with the relative rankings. Correlation coefficients were calculated between nine site characteristics and the site ranking by mean relative difference for both water years (Table IV). Surprisingly, only the clay fraction of soil texture and coarse fragment content, which both affect soil water storage, were significantly correlated with the relative rank. Maximum snow depth, which largely controlled the amount of input water, apparently was not critical because input water in excess of the storage capacity rapidly passed through the 0 to 75 cm measurement zone. Vegetation density, which controls the rate of dry down, was not important because uptake dynamics were probably similar at these relatively shallow depths. Parameters related to basin geometry, elevation and

Table IV. The correlation coefficient results between specific site characteristics and the rank of their respective mean relative difference are displayed. Accordingly, the static properties of textures of percentage clay content and percentage coarse material have a significant relationship with the wetness of sites. The dynamic property of maximum snow depth appears uncorrelated for these data

Property	Correlation coefficient	
	WY2001	WY2003
%Sand	0·229	0·090
%Silt	0·641	0·299
%**Clay**	**0·789**	**0·813**
%**Coarse**	**−0·840**	**−0·597**
Elevation (m)	0·142	0·367
Topographic index	−0·042	0·025
Vegetation LAI	0·470	−0·029
Depth of soil	−0·161	0·015
Max snow depth	0·232	0·082

Table V. Summarized results for r_s and r_P from the entire data collection period using two critical values. Coefficients were calculated from the integrated 75 cm soil water storage values

	Number of active access tubes	Number of data pairs	$\alpha = 0{\cdot}01$		$\alpha = 0{\cdot}05$	
			$r_s(\%)$	$r_p(\%)$	$r_s(\%)$	$r_p(\%)$
WY2001	12	105	64	70	96	97
WY2003	19	78	83	82	95	91
Interannual	11	195	62	53	91	84

topographic index would be expected to be important if there were substantial lateral redistribution within the upper 75 cm of soil; but, consistent with other work at Reynolds Creek (e.g. Flerchinger *et al.*, 1992), this was apparently not the case.

Overall results from the significance testing using two-tailed confidence intervals ($\alpha = 0{\cdot}01$ and $0{\cdot}05$) show very high significance and, thus, temporal stability, for both the temporally successive data and the majority of the data collection time period (Table V). Despite the high degree of spatial variability, it is reasonable to look at average values of S_{75} ($\overline{S_{75}}$) across the watershed due to the high temporal stability of the sites. We compared daily precipitation with $\overline{S_{75}}$ (normalized with respect to the maximum value, $S_{N,75}$) and temporal stability values for consecutive dates during both years of measure (Figure 7). The few dates that showed low or no significance were related to periods of rapid change (i.e. slope) in S_{75}. In particular, note that the period when the correlation coefficients displayed the lowest significance was coincident with the winter rain-on-snow events in WY2003.

Although both r_s and r_P indicated a high degree of temporal stability, quantifiable changes of these measures were analysed using the slope a and Y-intercept b from Equation (4), revealing further information concerning the dynamics of S_{75} during the year, which were roughly consistent for both years (Figure 8). We categorized these changes in terms of 'conditions' described previously and in Table II. The matrices in Figure 8 were coded to compare dates and describe the status of spatial soil water changes with time, while rank remained stable. For example, changes between 20 November WY2003 and subsequent, higher S_{75} dates (Figure 7) were almost all Condition 2 increases, whereas changes between 20 November and the

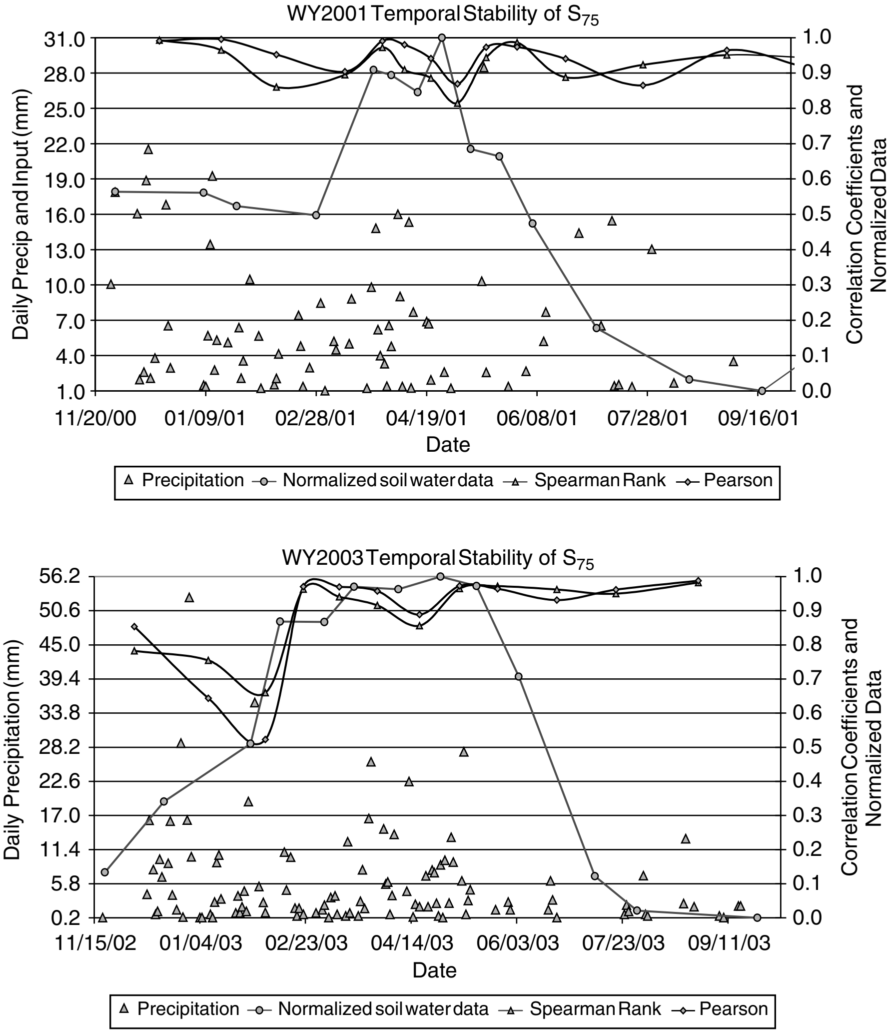

Figure 7. Comparison of Spearman rank test and Pearson correlation coefficient with normalized annual soil water patterns for the 2 years of record. Precipitation given in millimetres for comparison

even drier late summer dates were Condition 3. Comparison between the years displayed similar patterns over the seasons.

Condition 1 indicates no effective change in the S_{75} values and was observed in three situations. The first was during the late summer, the driest part of the year. This situation is probably underrepresented in our data owing to the sampling frequency, and can be expected to persist during the late summer and early fall. At this time, rainfall was very low and plant uptake had effectively ceased in the upper 75 cm of soil due to lack of plant-available water. The second situation occurred when the soil was effectively buffered from inputs due to a persistent snow cover and drainage was negligible. And the third situation occurred in the late winter and early spring, when the soil reached field capacity. At that time, melt water inputs were rapidly transmitted through the soil column and plant uptake was negligible, resulting in constant S_{75} values.

Condition 2 occurs when there is a roughly uniform change in S_{75} across all sites, thus changing b without affecting a. This was observed most commonly during the fall and a few winter dates prior to soils reaching maximum measured S_{75} values. It indicates dry (less than field capacity) soil conditions and spatially uniform

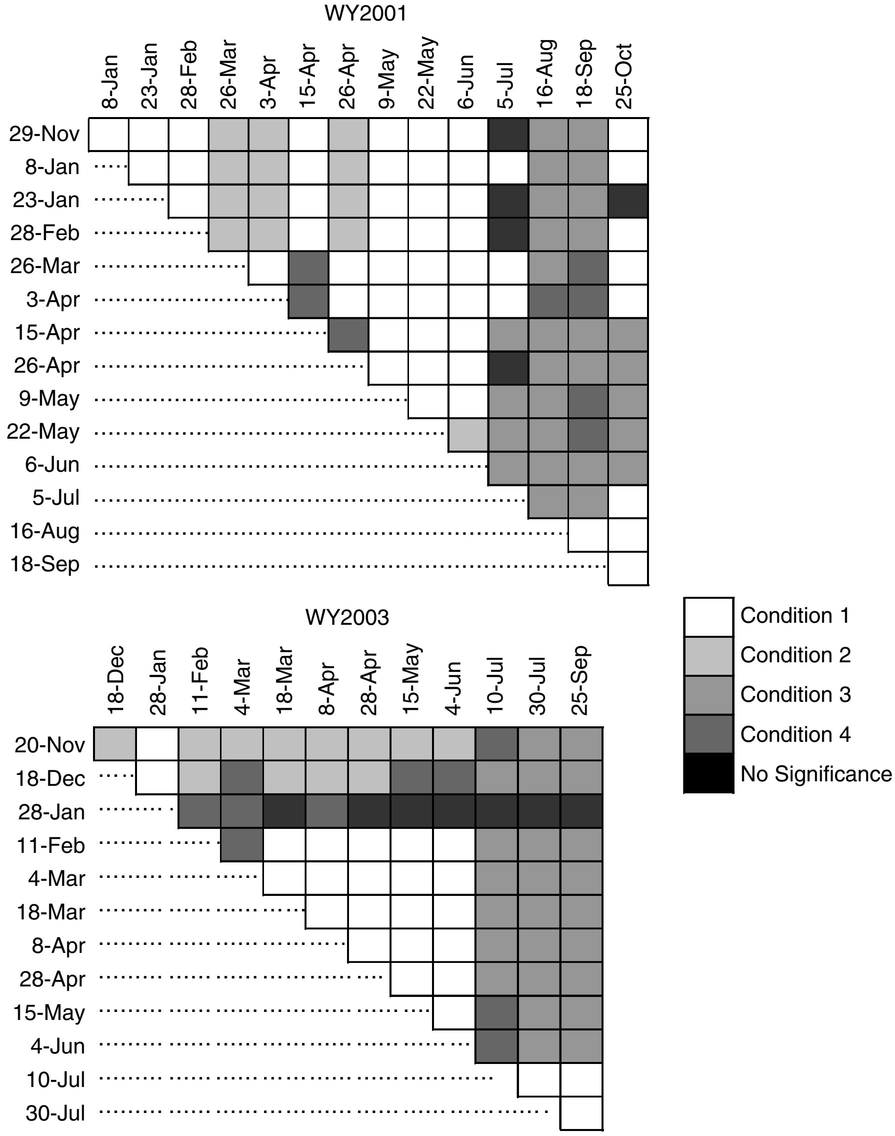

Figure 8. Matrices showing the four conditions attributed to linear regression analysis during temporal stability. The annual cycle is clearly partitioned into regimes. No significance occurs when either r_S or r_P is not significant

inputs of precipitation or snowmelt. This condition is also present with uniform loss of soil water, i.e. during spring, when evapotranspiration and throughflow are fairly evenly distributed.

Condition 3, in which *a* changed (reduced) but *b* remained constant, prevailed during late spring and early summer, the period of dry down. This indicates that soil drying, primarily by plant uptake, occurred relatively rapidly at the wetter sites.

Condition 4 is attributed to changes in both *a* and *b*. This condition appeared short lived, only occurring during events with unevenly distributed fluxes, such as during the rain-on-snow events, when drier sites increased more than the wetter sites. We expect that a greater frequency of sampling would have revealed more such dates.

Two of the situations described by Condition 1, when soils are either near field capacity (Condition 1_{fc}) and when they are near the limit of plant water extraction (Condition 1_{pwp}), are somewhat analogous to the two preferred states conceptualized by Grayson *et al.* (1997), and the other three conditions represent transitional environments between those preferred states. Grayson *et al.* (1997) were able to relate preferred states to the level of spatial organization and potential for stream-flow generation in a snow-free environment. In what follows, we use the slope–intercept analysis and the related temporal stability conditions to relate the soil water conditions to stream-flow generation.

Soil water dynamics

Stream flow, represented by normalized values for visualization, is related to changes in $S_{N,75}$ and daily precipitation in Figure 9. Both water years were drier than average, with total precipitation of 649·9 cm and 709·3 cm for WY2001 and WY2003 respectively. However, the general trends of stream flow and soil water dynamics were similar to other years (Pierson *et al.*, 2001; Seyfried *et al.*, 2001b). Differences in melt dynamics resulted in higher peak flows in WY2001 (0·105 m^3 s^{-1}) than WY2003 (0·048 m^3 s^{-1}). The rain-on-snow events in 2003 caused an unusual spike in stream flow in February.

Conspicuously, stream-flow generation was rarely related directly to precipitation. This is partly because water inputs to dry soil were effectively absorbed and stored in the soil generating no runoff, and partly because inputs of snow were stored above surface until the spring melt. Thus, precipitation and/or snow water input may be poor indicators of stream-flow generation. The relationship between water inputs and stream-flow generation may be clarified if considered in terms of the soil water status as described by the conditions described above.

The end of summer was characterized by low $S_{N,75}$, as any precipitation was rapidly lost to evapotranspiration and Condition 1_{pwp} temporal stability dominated. Essentially no flow was generated through the soil and stream flow was near the minimum due to lack of inputs. Rainfall events generated only small spikes in the hydrograph, as the contributing area was immediately adjacent to the stream channel. Similarly, increases in $S_{N,75}$ with uniformly distributed fall rain and lower evaporative demand (Condition 2) resulted in virtually no stream flow increase, as all incoming water was absorbed by the dry soil. Once additions were sufficient to achieve field capacity ($S_{N,75} = 0{\cdot}7–0{\cdot}8$), soil water patterns were surprisingly constant (Condition 1_{fc}). However, once this threshold was achieved, stream flow was highly responsive to melt events, as evidenced by the observed dramatic spikes in stream flow. Values of $S_{N,75}$ remained constant because incoming water was rapidly transmitted through the soil. Thus, Condition 1_{fc} represents soil conditions 'ripe' for stream-flow generation. We note that in the case of the rain-on-snow event of 2003, Condition 4 prevailed. This is probably due to the dry conditions prior to the event and possibly due to a unique distribution of input water, which included both snowmelt and rainfall.

Shortly after snowmelt, plant uptake of water was apparently most rapid from the wettest portions of the catchment (Condition 4) and $S_{N,75}$ decreased below field capacity. Once this threshold was crossed, upland contributions to stream flow were diminished and stream flow declined rapidly. There was a short lag seen in the hydrograph, as soil water drained through to the stream and then stream flow quickly returned to values near the summer minimum.

McNamara *et al.* (2004), working in a similar environment, identified similar relationships between upland soil water conditions and stream-flow generation. This work extends their findings using a quantitative measure of spatially variable soil water conditions throughout the catchment, as opposed to a single profile and hillslope.

CONCLUSIONS

At all measurement dates, both θ and S_{75} were normally distributed, as observed elsewhere (Famiglietti *et al.*, 1998; Western *et al.*, 2002). In general, the spatial variability in θ, as measured by the SD, was greater than that observed in other studies, reflecting the heterogeneity of conditions in the catchment. This variability

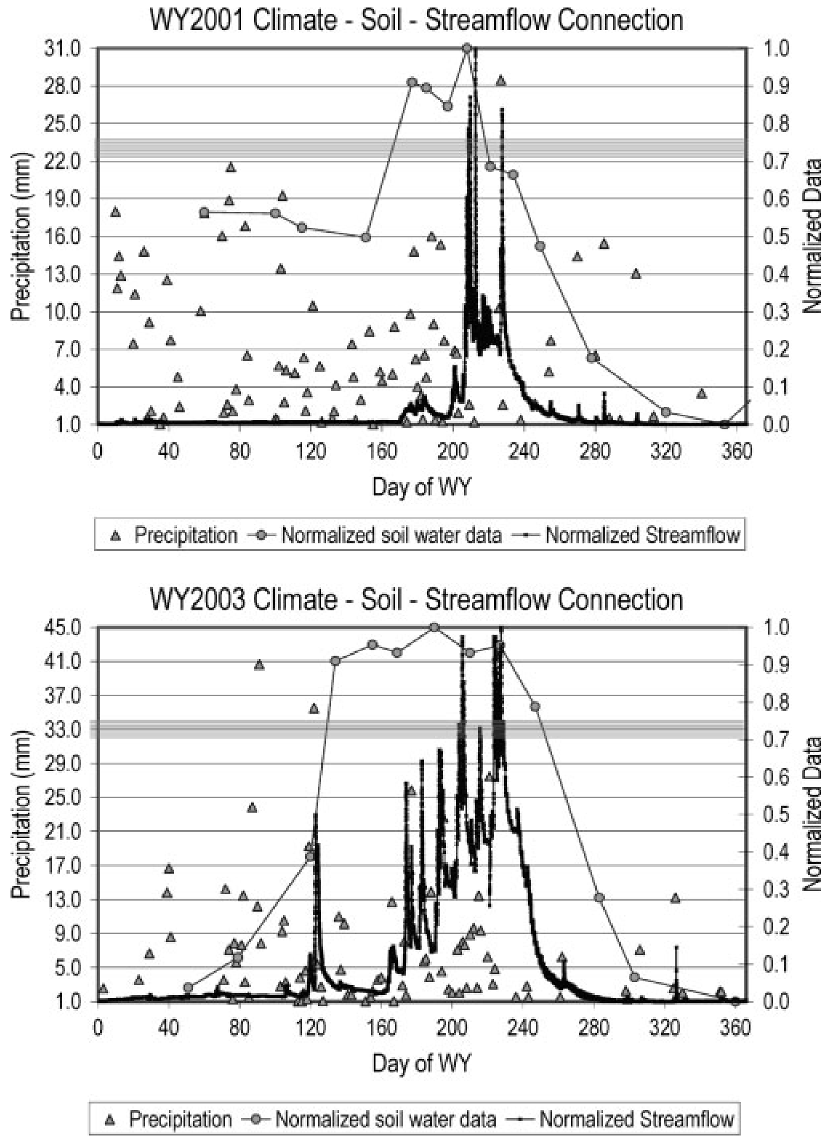

Figure 9. Climate–soil–stream-flow connection: WY2001 and WY2003. Annual trends of climate, soil water and stream flow. The neutron probe (NP) data are average S_{75} values (cm), normalized: for WY2001, minimum and maximum values were 6·8 cm and 21·9 cm; for WY2003, they were 5·8 cm and 21·8 cm. Grey horizontal bar gives the approximate stream-flow generation threshold. Date is in day of water year, starting on October 1

increased with measurement depth, attributed to increased heterogeneity of soil properties with depth and convergent catchment geometry. With the exception of near-surface measurements, we did not observe the correlation between θ and SD described in other work (Famiglietti *et al.*, 1998; Grayson and Western, 1998; Martinez-Fernandez and Ceballos, 2003). These findings illustrate the difficulty of characterizing soil water in heterogeneous mountain settings.

Statistically significant temporal stability, using both the Spearman rank test and temporal linear regression analysis, was generally observed across all sites for both years, indicating that S_{75} changes occurred in a roughly parallel fashion with time. The relative ranking of each site is largely controlled by soil properties affecting water storage (percentage clay and coarse fragments), which were highly correlated with the ranking, whereas parameters related to topographic position, vegetation density and snow water input were not. These consistent relative rankings for each site were related to the probability distribution of S_{75} values.

Temporal linear regression analysis of temporally stable changes of S_{75} was used to describe changes in the mean and/or if changes were weighted toward the wet or dry sites. Periods of no change were most common during the dry summer, when both inputs and outputs of water were minimal, and during snowmelt, when they were maximal. Other conditions described the distribution of change during transitions between those conditions. There was a threshold average S_{75} above which stream flow was highly sensitive to upland melt or rainfall inputs, because those inputs passed rapidly through the soil. When the average S_{75} was below that threshold, stream flow was insensitive to upland inputs because water was absorbed by the soil.

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REFERENCES

Allard D. 1994. On the connectivity of two random set models: the truncated Gaussian and the Boolean. In *Geostatistics Troia '92. Quantitative Geology and Geostatistics*, Soares A (ed.). Kluwer: Dordrecht, The Netherlands; 197–211.

Burrough PA. 1993. Soil variability: a late 20th century view. *Soils and Fertilizers* **56**(5): 529–562.

Comenga V, Basille A. 1994. Temporal stability of spatial patterns of soil water storage in a cultivated Vesuvian soil. *Geoderma* **62**: 299–310.

Cressie N. 1991. *Statistics for Spatial Data*. Wiley: New York.

Devore J. 1995. *Probability and Statistics for Engineering and the Sciences*, 4th edition Brooks/Cole Publishing Co.: Pacific Grove, CA.

Dunne T, Moore TR, Taylor CH. 1975. Recognition and prediction of runoff-producing zones in humid regions. *Hydrologic Sciences Bulletin* **20**: 305–327.

Famiglietti JS, Rudnicki JW, Rodell M. 1998. Variability in surface moisture content along a transect: Rattlesnake Hill, Texas. *Journal of Hydrology* **210**: 259–281.

Flerchinger GN, Cooley KR, Ralston DR. 1992. Groundwater response to snowmelt in a mountainous watershed. *Journal of Hydrology* **133**: 293–311.

Gardner WR, Kirkham D. 1952. Determination of soil moisture by neutron scattering. *Soil Science* **73**: 392–401.

Gomez-Plaza A, Alvarez-Rogel J, Albaladejo J, Castillo VM. 2000. Spatial patterns and temporal stability of soil moisture across a range of scales in a semi-arid environment. *Hydrological Processes* **14**: 1261–1277.

Grayson RB, Western AW. 1998. Towards aerial estimation of soil water content from point measurements: time and space stability of mean response. *Journal of Hydrology* **207**(1–2): 68–82.

Grayson RB, Western AW, Chiew FHS, Blöschl G. 1997. Preferred states in spatial soil moisture patterns: local and nonlocal controls. *Water Resources Research* **33**(12): 2897–2908.

Hanson CL, Johnson GL. 1993. Spatial and temporal precipitation characteristics in southwestern Idaho. Management of irrigation and drainage systems: integrated perspectives. In *Proceedings of the 1993 National Conference on Irrigation and Drainage Engineering*, Allen R (ed.). Park City, Utah, American Society of Civil Engineers. New York, NY; 394–400.

Hanson CL, Marks DG, VanVactor SS. 2001. Long term climate database, Reynolds Creek Experimental Watershed, Idaho, United States. *Water Resources Research* **37**(11): 2839–2841.

Hignett C, Evett SR. 2002. Neutron thermalization. In *Methods of Soil Analysis, Part 4, Physical Methods*, Dane JH, Topp GC (eds). *Soil Science Society of America Book Series* 5. SSSA Press: Madison, WI; 501–521.

Kachanoski RG, de Jong E. 1988. Scale dependence and the temporal persistence of spatial patterns of soil water storage. *Water Resources Research* **24**(1): 85–91.

Marks D, Winstral A. 2001. Comparison of snow deposition, the snow cover energy balance, and snowmelt at two sites in a semiarid mountain basin. *Journal of Hydrometeorology* **2**(3): 213–227.

Marks D, Winstral A, Seyfried M. 2002. Simulation of terrain and forest shelter effect on patters of snow deposition, the snowmelt, and runoff over a semi-arid mountain catchment. *Hydrological Processes* **16**: 3605–3626.
Martinez-Fernandez J, Ceballos A. 2003. Temporal stability of soil moisture in a large-field experiment in Spain. *Soil Science Society of America Journal* **67**: 1647–1656.
McNamara JP, Chandler D, Seyfried MS. 2004. The role of a snowpack on relationships between preferred soil moisture state, lateral flow, and stream flow generations in a small catchment. *Hydrologic Processes* **18**: in press.
Pandit KN. 1999. *A study of soil moisture variability and vegetation greenness dynamics in a mountain rangeland watershed using direct measurements, remote sensing, and modeling*. PhD dissertation, Department of Biological and Agricultural Engineering, University of Idaho, Moscow.
Pierson Jr FB, Slaughter CW, Cram ZK. 2001. Long-term stream discharge and suspended sediment database Reynolds Creek Experimental Watershed, Idaho, USA. *Water Resources Research* **37**(11): 2857–2861.
Reynolds SG. 1970. The gravimetric method of soil moisture determination part III: an examination of factors influencing soil moisture variability. *Journal of Hydrology* **11**: 288–300.
Russo D, Jury WA. 1987a. A theoretical study of the estimation of the correlation length scale in spatially variable fields. 1. Stationary fields. *Water Resources Research* **23**: 1257–1268.
Russo D, Jury WA. 1987b. A theoretical study of the estimation of the correlation length scale in spatially variable fields. 2. Nonstationary fields. *Water Resources Research* **23**: 1269–1279.
Seyfried M. 1998. Spatial variability constraints to modeling soil water at different scales. *Geoderma* **85**: 231–254.
Seyfried MS, Harris R, Marks D, Jacob B. 2001a. Geographic database, Reynolds Creek Experimental Watershed, Idaho, United States. *Water Resources Research* **37**(11): 2825–2830.
Seyfried MS, Murdock MD, Hanson CL, Flerchinger GN, Van Vactor, SS. 2001b. Long-term soil water content database, Reynolds Creek Experimental Watershed, Idaho, United States. *Water Resources Research* **37**(11): 2847–2852.
Stephenson G, Zuzel J. 1967. Seismic refration studies in watershed hydrology. *5th Annual Symposium on Engineering Geology and Soil Engineering Bulletin*. Pocatello, ID, USA.
Tribbeck M. 2002. Modelling the effect of vegetation on the seasonal snow cover. PhD dissertation, University of Reading, UK.
Vachaud G, Passerat de Silans A, Balabanis P, Vauclin M. 1985. Temporal stability of spatially measured soil water probability density function. *Soil Science Society of America Journal* **49**: 822–828.
Western AW, Grayson RB, Blöschl G. 2002. Scaling of soil moisture: a hydrologic perspective. *Annual Review of Earth and Planetary Sciences* **30**: 149–180.
Yates SR, Warrick AW. 2002. Soil sampling and statistical procedures. In *Methods of Soil Analysis: Part 4—Physical Methods*, Dane JH, Topp GC (eds). *Soil Science Society of America Book Series* 5. SSSA Press: Madison, WI, Chapter 1, Section 5: 81–118.

