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Flow processes in a rangeland catchment in California

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Abstract

Emerging hydrology-related issues in California grasslands have directed attention towards the need to understand subsurface water flow within a complex, dynamic system. Tensiometers and neutron probes evaluated the subsurface hydrology of a rangeland catchment. Hydrological processes within the catchment varied both in space and time. Spatial variability was evident along the vertical profile and between the catchment slopes. Temporal variability in processes coincided with the seasons (i.e., wet winter, dry summer, and spring). From a water-balance equation developed for the catchment, we determined that there was significant variability both spatial and temporal in the amount of soil moisture lost to evapotranspiration and deep seepage. During the 16 month monitoring period there was a total of 50 cm of rainfall that fell in the catchment of which 9–55 cm was lost to evaporation and 37–79 cm to deep seepage. A simple deduction of the losses (evaporation and deep seepage) from the input (rainfall) shows that all monitored locations had a substantial decrease in the amount of water that was stored in the soil profile.

Key Words: California rangelands, subsurface flow, water budget

Resumen

Los problemas que están surgiendo en relacion a la hidrología de los pastizales de California, se han enfocado hacia la necesidad de entender el flujo subterráneo del agua dentro de un sistema complejo y dinámico. En una área de captación del pastizal se evaluó la hidrología subterránea mediante el uso de tensiómetros y dispersores de neutrones. Los procesos hidrológicos dentro del área de captación variaron en espacio y tiempo. La variabilidad espacial fue evidente a lo largo del perfil vertical y entre las pendientes del área de captación. La variabilidad temporal de los procesos coincidió con las estaciones (esto es, invierno húmedo, verano y primavera secos). A partir de una ecuación del balance hídrico desarrollada para esta área de captación, determinamos que hubo una variabilidad temporal y espacial significativa en la cantidad de humedad del suelo perdida por evaporación y filtración. Durante los 16 meses del período de monitoreo la precipitación recibida en el área de captación fue de 50 cm de los cuales de 9–55 cm se perdieron por evapotranspiración y de 37–39 cm se perdieron por filtración. Un simple substracción de las perdidas (evapotranspiración y filtración) de lo recibido (lluvia) muestra que todas la localidades monitoreadas tuvieron un decremento un substancial enla cantidad de agua que fue almacenada en el perfil del suelo.

A common omission in rangeland hydrology studies has been a rigorous treatment of the subsurface component of the hydrologic cycle. A need exists for the understanding of flow path dynamics (surface and subsurface) and the spatial and temporal variations in the hydrology of rangelands.

The movement of soil water in semiarid and arid climates has been investigated since at least 1949 when Maxey and Eakin attempted to measure ground water recharge in the desert basins of Nevada. By assuming basin recharge to equal basin discharge they concluded that in areas with less that 20 cm yr^{-1} of precipitation, recharge was essentially negligible. Nixon and Lawless (1960) monitored the movement of soil moisture from rainfall up to depths of 6.0 m in a region 250 km northwest of Los Angeles, California. Using a neutron probe and a nest of sparsely distributed tensiometers (up to a depth of 1.0 m), they concluded that 31% of the rainfall migrated to the deep profile over a period of 240 days following the last rainfall event. Holmes and Colville (1970), while investigating the water balance of a grassland in southern Australia with lysimeters and neutron probes, determined that less than 10% of the 63 cm of precipitation recorded in 5 years recharged a low lying water table.

In recent years both physical and chemical methods have been used to estimate recharge in arid and semiarid regions. Physical methods have included both direct and indirect measurements. While direct measurements of deep seepage have mainly involved the use of lysimeters (Allen et al. 1991, Gee et al. 1993), indirect physical methods have included the use of soil water balances (e.g. Rushton and Ward 1979), the zero flux plane method (Wellings 1984), and estimates of water fluxes from solutions to either Darcy's law or Richards' equation (Sophocleous and Perry 1985, Stephens and Knowlton 1986). Johnston (1987) and Sharma and Huges (1985) demonstrated the considerable variation in the rates of local recharge over a scale of a few meters in many soil types. Assessment of this spatial variability in recharge has been attempted with frequency-domain electromagnetic (EM) or transient electromagnetic methods (Allison et al. 1994). Table 1 summarizes some estimates of deep seepage for semiarid regions.

This paper presents results from the investigation of the subsurface hydrology of a rangeland watershed in California. The central thesis presented is that seasonal changes in the hydrologic cycle cause important variations in the flow dynamics in rangeland catchments in California. In this paper, observations of soil moisture content and associated energy levels are used to develop a water balance for a catchment dominated by annual grasses.

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Table 1. Estimates of deep seepage in semi-arid regions. Source Stephans et al. 1986.

Author	Method	Location	Estimate recharge rate
Enfield et al. 1973	Soil water potential	Washington state (south central)	1 cm yr^{-1}
Klute et al. 1972	Neutron moisture logging	High Plains, Colorado	0 cm yr^{-1}
Dincer et al. 1974	Tritium tracer	Saudi Arabia	25% annual precipitation
Allison et al. 1985	Chloride concentration	South Australia	1.4 cm yr^{-1}
Meyboom 1966	Water level hydrographs	Saskatchewan, Canada	7.5% annual precipitation
Boyle and Salem 1979	Temperature profiles	Illinois (north west)	7.8–31 cm yr^{-1}
Maxey and Eakin 1949	Steady State flow	Nevada	0 cm yr^{-1}
Watson et al. 1976	Steady State flow	Nevada	3.4% annual precipitation
Sammis et al. 1982	Hydraulic gradients	Phoenix, Arizona	18 cm yr^{-1}
Sammis et al. 1982	Temperature profiles	Phoenix, Arizona	9 cm yr^{-1}
Sammis et al. 1982	Tritium tracer	Phoenix, Arizona	40 cm yr^{-1}
Stephen and Knowlton 1986	Hydraulic gradients	Socorro, New Mexico	4 cm yr^{-1}

Materials and Methods

Study Site

Soil water potential and content was measured within a watershed in north-central California over a period of 16 months. The data was used to identify processes influencing the hydrology of the catchment, and to determine quantitatively, the moisture status along vertical soil profiles at specific locations.

The watershed is located within the Russell Tree Farm, a 115-hectare (ha) research station located, in Contra Costa County (Lat. 37° 54' N, Long 122° 03' W), California (Fig. 1). The amphitheater-shaped catchment has an area of 20 ha, with elevations ranging from 230 m above mean sea level (msl) in the valley bottom to nearly 360 m above msl on the ridge top. The slopes range from 2 to 75%, and the catchment contains at least 2 perennial springs. Annual grasses dominate the vegetation cover in the catchment. Sporadic clusters of predominantly coyote bush shrubs (*Baccharis pilularis*, D.C.) are found scattered throughout the catchment. The other major vegetation type is the California oak, which is restricted to the higher elevations close to the ridge and is dominated by coast live oak (*Quercus agrifolia*, Nee). The Mediterranean climate of the region results in warm to hot summers and winters that are relatively mild. Precipitation occurs as rainfall mainly in the winter months. While most climatic parameters remain relatively constant from year to year, precipitation in the region shows variability both within and between seasons (Fig. 2).

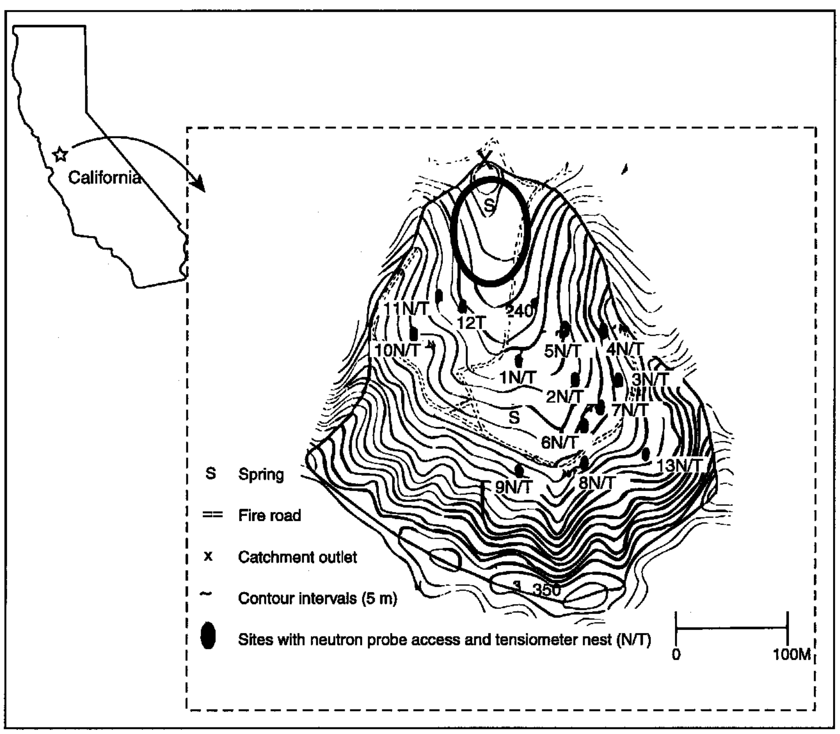

Fig 1. Location of the Russell Tree Farm with monitored sites.

Thirteen sites were located along the slopes for measurement of soil moisture content and potentials. These sites were selected to represent the 3 dominant vegetation types found in the catchment (annual grasses, baccharis shrubs, and oak trees) and 3 broad elevations viz. low, medium, and high.

Water content of soils was determined by the neutron-probe method (Gardner 1986). A Campbell Pacific Nuclear Model 503 with a 50 millicurie americium-beryllium source neutron probe was used. The probe was calibrated by a gravimetric analysis that utilized in situ samples from the access tube holes.

Thirteen neutron probe access tubes (Sites 1–13) were installed in February 1993. In most cases, the depth of augering was limited by bedrock. The deeper profiles monitored ranged from 3.75 to 5.70 m in depth and included 5 sites (6–9, and 13) located on the southeast corner of the catchment. Of the 7 shallow profiles, Sites 1 and 2 extended to 2.00 m, while the remaining were restricted to 1.20-m depths. During installation of the access tubes for the neutron probe, a 0.05-m diameter hole was augered, into which a 0.04-m inner diameter, schedule 40 PVC pipe was inserted. The PVC pipe in each hole extended 0.15 m above the ground surface. Bentonite clay was poured around the PVC pipe, 0.10 m below the surface to prevent surface run-off water from flowing down the sides of the tube.

Soil water potential along the catchment slopes was determined using tensiometers. Thirteen nests of tensiometers (Sites 1–13) were installed at a distance of about 1.0-m from the neutron probe access tubes (Fig.

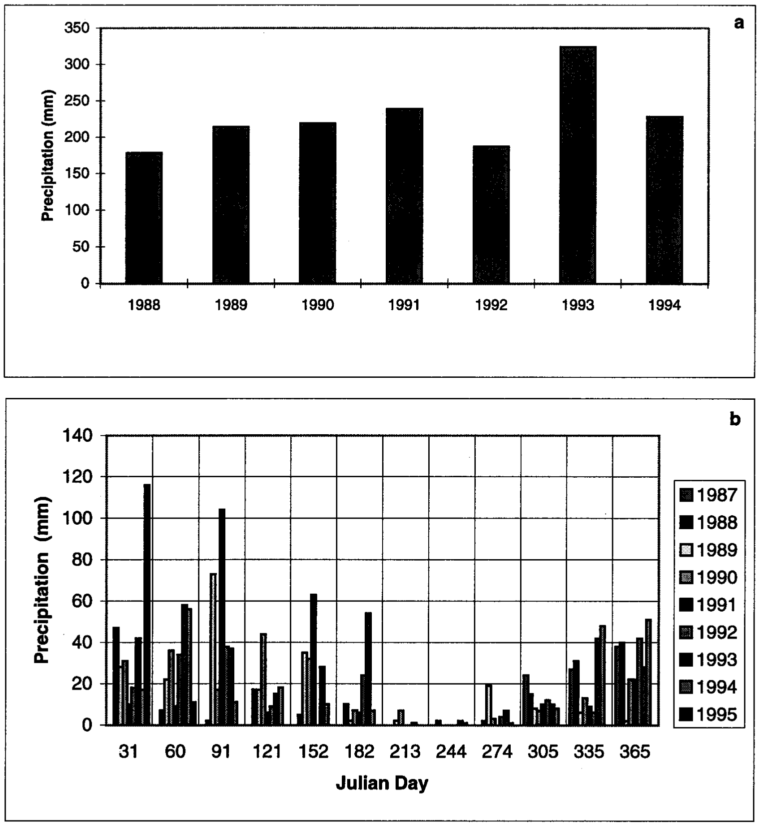

Fig. 2. (a) Annual rainfall at Walnut Creek located 10 Km east of Russell Tree Farm. (b) Variability in monthly rainfall 10 Km east of study site.

3a). A 0.10-m diameter soil auger was used to create the hole into which tensiometers were inserted. The soil removed during augering was collected in 0.15-m sections and stored in sealed plastic bags. The area around the ceramic tip of each tensiometer was back-filled with native soil removed from that particular depth. This soil was repacked to a density similar to that of the undisturbed native soil and filled to approximately 0.10 m above the center of the ceramic tip (The original packing density was achieved by reintroducing the augered soil from the 0.15 m sections back into the same zone). Above this backfill, bentonite (as a mixture of powder and pellets) was packed to a height of 0.10 m and wetted to provide a vertical hydraulic seal. Above this seal native soil was packed to the original density. Water was injected into the tensiometers through narrow diameter tubing which extended down to the ceramic cup.

At each site, tensiometers were installed in nests such that the ceramic tips were located at approximately 0.25-m intervals down to a depth of 1.0 m from the surface. Between 1.0 and 2.0 m, tensiometers were installed at 0.5 m intervals, and at greater depths the interval was increased to approximately 1.0 m (Fig. 3b). Of the 13 locations, 6 had tensiometers located at depths greater than 3.0 m (Sites 6–9, 12, and 13). The shallowest nest, Site 3, extended to a depth of 1.0 m, and the deepest, Site 6, extended to a depth of 6.0 m.

Hydraulic head measurements were determined as the sum of gauge pressure at the ceramic tip and the elevation of the ceramic tip relative to the ground surface at the nest. Thus, the hydraulic head measurements in each nest are referenced to the local elevation of the nest. Gauge pressure readings were taken using a portable pressure transducer, commercially referred to as the Tensimeter (Soil Measurement Systems, Tucson, Ariz., 2 Marthaler et al. 1983). Water potential readings were taken at weekly intervals during the winter and monthly intervals late in the summer.

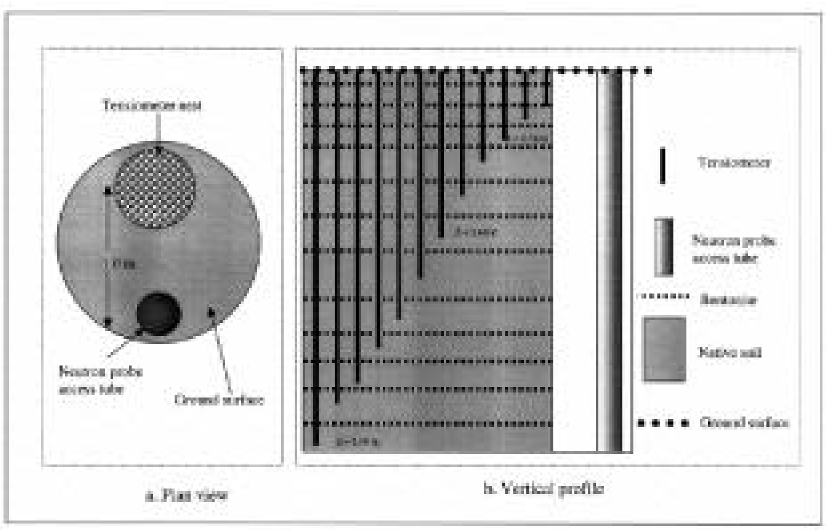

Fig. 3. Schematic of monitoring station at Site 6: (a) the location of nested tensiometers with respect to the neutron probe access tube and (b) vertical extent of tensiometer and neutron probe measurements.

Water balance estimates

Parameters used to evaluate the water balance of a location estimate how much usable energy and water are available (for evaporation and transpiration), how much evaporation demand is met by available water, and how much water is usable excess. In its simplest form, the equation describing annual conservation of water mass for a unit watershed can be written as:

$$P - Q - E = \Delta S \quad (1)$$

Where P is the average annual precipitation, Q is the average annual discharge, E is the annual evapotranspiration, and ΔS is the change in the amount of moisture stored. The signs indicate whether water is entering (positive) or leaving (negative) the system.

If an average is taken over many years of record, then the ΔS term can be assumed to equal zero, and Equation [1] becomes (Freeze and Cherry 1979):

$$P = Q + E \quad (2)$$

The terms on the right in Eq. [1] can be modified to various levels of detail, depending on the important characteristics of a region or on the available database.

For example, Eagleson (1978) presented the water-balance equation for a unit watershed as:

$$P_A = Q_{SA} + Q_{GA} + E_{TA} + \Delta S_S + \Delta S_g \quad \textbf{(3)}$$

Where P_A is the annual (seasonal) precipitation, Q_{SA} is the annual (season) surface runoff, Q_{GA} is the annual (seasonal) subsurface flow through the watershed, E_{TA} is the annual (seasonal) total evapotranspiration, ΔS_S is the annual (seasonal) change in surface storage, and ΔS_G is the annual (seasonal) change in soil moisture and groundwater storage.

For rangeland catchments in central California, large annual fluctuations in precipitation (Fig. 2a) results in large variability in the amount annual of recharge, which in turn results in large changes in the amount of moisture stored in the soil profile (ΔS_G). In the absence of surface runoff (Q_{SA}) and surface storage (S_S) on catchment slopes, Eq. [3] can be re-written for these rangelands as:

$$P_A = Q_{GA} + E_{TA} + \Delta S_G \quad \textbf{(4)}$$

In this study the moisture status for locations within the catchment was determined using a modified version of Eq.[1] where the dynamic water content for a vertical soil profile was evaluated using the equation:

$$P_P = \Delta S_E + \Delta S_Q \quad \textbf{(5)}$$

Here water infiltrating the surface (P_P) was partitioned as surface evapotranspiration or deep seepage based on the direction flow gradients as indicated by the position of a zero-flux boundary (Fig. 4). (The zero-flux boundary is determined along the vertical profile as the location where hydraulic gradients are directed, in opposite directions, away from the point). All changes in moisture content (ΔS_P) above the zero-flux boundary were assigned to ET (ΔS_E), while all losses below the boundary were assigned to deep seepage (ΔS_Q).

Since soil moisture content was monitored at monthly intervals and hydraulic gradients were measured at weekly intervals, the monthly changes in moisture content were proportioned according to the length of time the gradient direction was observed. For example, if the change in moisture content at a given depth, over a single month, was 10.0 cm and hydraulic gradients were positive for 3 of the 4 weeks, it was assumed that 7.5 cm of the moisture was lost to deep seepage, while 2.5 cm was released as ET to the surface. From data collected on moisture content and associated potentials in the soils the amount of moisture contributing to deep seepage was estimated by assuming that positive flow gradients lying below the zero flux plane directed flow into the deep profile. For example, in Site 9 the zero flux plain migrated to a depth of 5.0 m over a period of 7 months while in Site 13 it was located 3.0 m below the ground surface (Fig. 4).

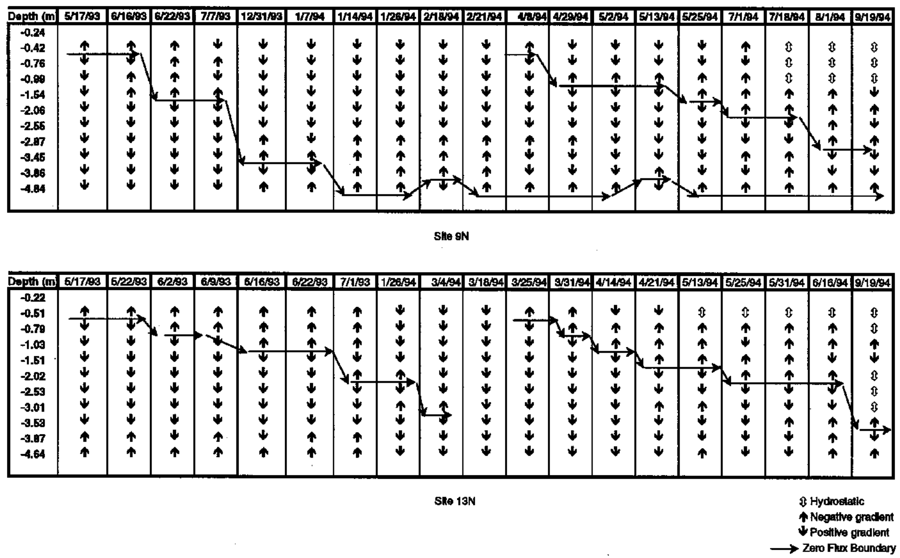

Fig. 4. Migration of the zero-flux boundary in the soil profile at Sites 9 and 13. Vertical arrows indicate the direction of the hydraulic gradient.

Observations

The data collected includes measurements of soil matric potential from 13 sites and water content from 12 sites within the slopes, and from 16 locations along the catchment valley. Site 12 was not included in the analysis because of installation errors in the neutron access tube. Monitoring began in April 1993, following a winter during which the recorded rainfall in the region was the highest in at least the past 7 years(324 mm), and continued through September 1994, a year which was much drier (Fig. 2).

Recharge events in the catchment

Effectively saturated conditions close to the surface were evident in the tensiometer measurements when monitoring began in April 1993, and recharge along the vertical profile of the monitored sites was indicated by positive hydraulic gradients (Table 2). Following the rains in early April there was a 2-week precipitation free period when tensiometers close to the surface began to record decreases in hydraulic potentials. At depths below 0.2 m from the surface, potentials also began to decrease slower such that by the second week of April hydraulic gradients in the top 0.5 m of the profile at all sites were directed towards the surface. A small rainfall event (12 mm) during the third week of April, which was spread over a period of 48 hours, did not change this trend. By the end of April, most of the sites were losing water to the surface from the top 0.5 m, while deeper in the soil, flow was directed into the profile.

During the winter of 1993–94, the first rainfall (12 mm) was recorded in mid-October. This event was not detected as soil moisture changes by any of the tensiometers located in the catchment. The second event (20 mm) occurred approximately 1 month later was insufficient to bring potentials at the near surface within functioning range of the tensiometers. At the end of November, 2 storms of 42 mm total precipitation briefly increased hydraulic potentials, but in the next 10 days these values again decreased below the tensiometer range. Surface water recharge deep enough to penetrate to depths below 0.20 m was observed following 5 rainfall clusters which occurred between December and April. The first recharge event was initiated after 74 mm of precipitation over a period of 6 days in mid-December. The second event occurred almost 45 days later following a 6-day period in late January when 66 mm of rain occurred. A week later, 40 mm of rain, which occurred over 3 days, provided sufficient moisture for the third recharge event. The largest rainfall of the season (90 mm), recorded over a 5 day period in late February resulted in the fourth recharge event. During this period, profiles 1.5 m below the surface were close to

Table 2. Hydraulic gradients measured along the vertical profile of 3 sites. Bold arrows indicate direction of flow. Clear arrow indicate hydrostatic conditions.

Depth (m)	4/1/93	7/21/93	2/25/94	7/18/94
Site 6				
-0.20				
	0.59 ↓	0.00 ⇳	0.91 ↓	0.00 ⇳
-0.51				
	0.09 ↓	0.00 ⇳	0.77 ↓	0.00 ⇳
-0.98				
	1.31 ↓	0.00 ⇳	1.04 ↓	0.00 ⇳
-1.32				
	1.72 ↓	-4.29 ↑	1.21 ↓	-3.05 ↑
-1.99				
	0.52 ↓	-1.35 ↑	10.78 ↓	-0.10 ↑
-2.48				
	1.58 ↓	-7.70 ↑	-1.72 ↑	-1.67 ↑
-2.99				
	1.34 ↓	0.94 ↓	-2.25 ↑	-3.65 ↑
-3.68				
	0.20 ↓	0.81 ↓	0.58 ↓	-0.71 ↑
-4.95				
	1.39 ↓	0.19 ↓	1.03 ↓	2.08 ↓
-5.90				
Site 9				
-0.24				
	4.42 ↓	6.46 ↓	1.59 ↓	0.00 ⇳
-0.42				
	1.50 ↓	-2.58 ↑	0.78 ↓	0.00 ⇳
-0.76				
	0.04 ↓	0.22 ↓	1.90 ↓	0.00 ⇳
-0.99				
	0.55 ↓	-6.23 ↑	0.94 ↓	-5.73 ↑
-1.54				
	0.84 ↓	1.57 ↓	12.69 ↓	-2.76 ↑
-2.06				
	1.21 ↓	1.33 ↓	1.09 ↓	2.68 ↓
-2.55				
	2.47 ↓	1.26 ↓	-0.28 ↑	2.70 ↓
-2.87				
	0.35 ↓	0.40 ↓	-2.21 ↑	-3.67 ↑
-3.45				
	1.72 ↓	2.32 ↓	-0.26 ↑	-2.70 ↑
-3.86				
	0.87 ↓	-0.06 ↑	-2.49 ↑	-0.06 ↑
-4.84				
Site 13				
-0.22				
	1.22 ↓	0.00 ⇳	0.94 ↓	0.00 ⇳
-0.51				
	1.57 ↓	-12.94 ↑	1.76 ↓	0.00 ⇳
-0.79				
	1.07 ↓	-6.42 ↑	1.07 ↓	-14.87 ↑
-1.03				
	-1.11 ↑	3.63 ↓	0.85 ↓	1.86 ↓
-1.51				
	3.24 ↓	-5.56 ↑	2.64 ↓	0.18 ↓
-2.02				
	0.92 ↓	0.70 ↓	2.93 ↓	0.56 ↓
-2.53				
	1.04 ↓	0.19 ↓	-1.12 ↑	0.14 ↓
-3.01				
	0.20 ↓	1.12 ↓	2.79 ↓	-1.12 ↑
-3.53				
	0.79 ↓	-0.08 ↑	0.52 ↓	-4.33 ↑
-3.87				
	0.41 ↓	-0.05 ↑	0.09 ↓	-0.73 ↑
-4.64				

saturation, the deepest observed during the entire wet season. The final recharge occurred 2 months later when at the end of April a cluster of rainfall events provided 45 mm of precipitation.

Thus during the wet season which extended over 206 days, there were only 5 significant periods (totaling 21 days, and 71% of the season's precipitation) when moisture deposited at the soil surface migrated below the top 0.50 m of the soil profile. Each of these periods had more than 40 mm of rainfall deposited. In 4 of these events, groundwater recharge (as detected in the magnitude and direction of hydraulic gradients) significantly altered the moisture profile of the top 1.0 m of the soil. During a single cluster of events in late February, 20% of the seasons' rainfall was recorded, near saturated profiles were detected up to 1.5-m depths. However at greater depths the changes in potential values were small during the entire winter season.

Hydrologic Processes on Catchments Slopes.

The vertical soil profiles of the 12 monitored sites showed a sinusoidal pattern of wetting and drying (Fig. 5). Early in May 1993 all the sites recorded high amounts of moisture at all depths (~0.15 to 0.35 cm^3 cm^{-3}). Over the next month all the shallow profiles recorded losses in moisture content, and in the ensuing months moisture losses from the profiles continued, but at decreasing rates. With the start of the winter rains, the profiles began to moisten with increases in wetness occurring during the first wet month (i.e. December 1993). In the next 2 months, when the bulk of the season's rain fell, small increases in soil moisture were detected in all profiles, but these were much smaller than those observed early in the winter (~0.20 to 0.25 cm^3 cm^{-3}). Shortly after the wet season ended in early March 1994, the shallow profiles began to record losses in moisture. The largest decreases were observed in April, and smaller reductions occurred in the following precipitation free months. This drying pattern was similar to that of the previous year with the exception that the drying process in 1994 began almost 60 Julian days earlier (Fig. 5).

At each of the monitored sites, the total moisture lost from the near surface profile during the summer of 1993 was replenished during the following winter. Similar amounts of moisture were then lost from the profiles by the end of August 1994. In essence, these profiles reached a fixed

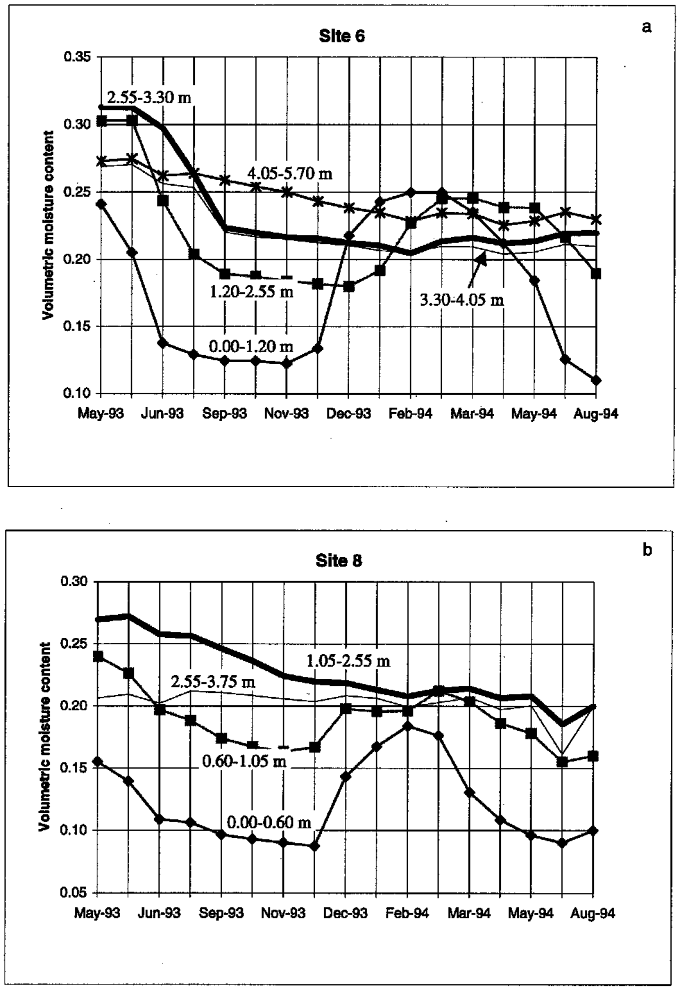

Fig. 5. Changes in volumetric moisture content at various depths along the vertical profiles of (a) Site 6 and (b) Site 8.

upper and lower limit in storing soil moisture towards the end of each season irrespective of the amount of rainfall received the previous wet season. At depths greater than 2.0 m, the amount and depth to which changes in volumetric moisture content occurred varied considerably with each profile showing a distinct response to seasonal changes in precipitation. Large losses in moisture, following a wet winter (1992–93), extended beyond the 2.0-m depth in 4 of the 5 deep sites. The single exception, Site 7, recorded losses in moisture, which were largely restricted to the top 1.9 m of the profile.

There were 2 distinct patterns of drying observed in the 5 deep profiles. At Sites 6, 7, and 8 drying began at the near surface zones early in the summer and migrated downwards as the summer progressed (Fig. 6a). In this downward migration, the near surface zones reached a critical moisture level early in the summer after which drying was observed lower in the profile. In the subsequent months this pattern was repeated such that the center of the drying zone was located at a depth of 3.3 m in Site 6 and 2.0 m in Site 8. This 'step-wise' drying pattern is significantly different from that observed in Site 9 and 13, where the drying was relatively uniform along the entire profile (Fig. 6b).

During the following winter (1993–1994), the depths to which the deep profiles wetted were less than that observed at the end of the previous winter rains (Fig. 7). Except for Site 13, in which the wetting front migrated to 2.4 m, wetting in the remaining sites was restricted to depths less than 2.0 m. Immediately following the last winter rains in March 1994, all profiles began to lose moisture along the entire wetted length. Unlike the previous summer, where there was 'step-wise' drying in some profiles, moisture losses were recorded at near equal rates at all the wetted lengths. In the deeper sections, which were not wetted by the winter rains, small decreases in moisture content continued to occur during the summer. For the first 6 dry months the loss in moisture in all the sites was much greater than the increases recorded during the wet period.

Comparing the moisture levels of different sampling dates with those observed in early May 1993, provides a measure of the 'relative wetness' within the catchment (Table 3). The relative wetness recorded in the deep profiles during May 1993 was never exceeded and only 1 profile (7N) registered levels similar to those at the start of the recording period. The peak in

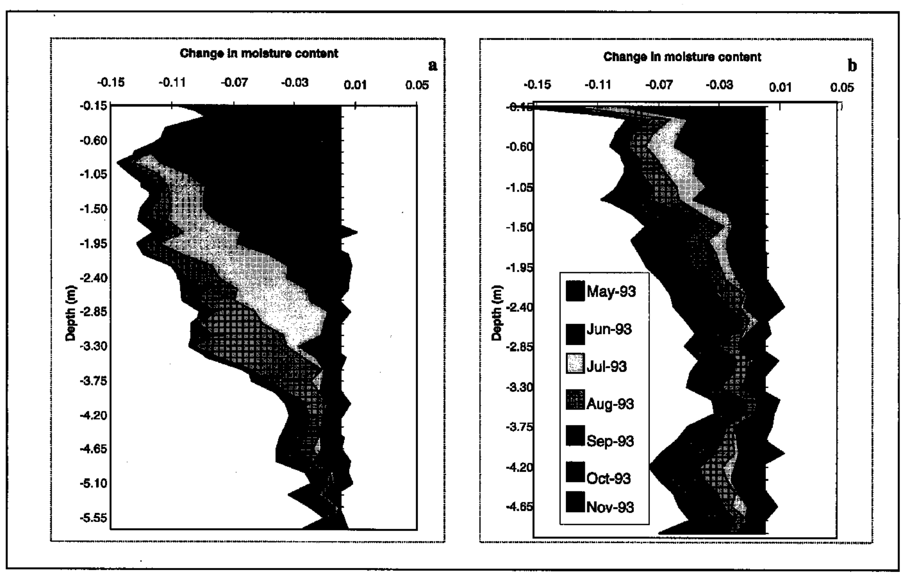

Fig. 6. Patterns of drying in 2 deep sites (6 and 9) during the summer of 1993. (a) At Site 6 drying was stepwise while (b) at Site 9 drying was relatively uniform along the entire vertical soil profile.

Table 3. Percentage of moisture present relative to early March 1993 along the vertical profile of monitored sites.

Site	1 May 1993	1 Jul. 1993	1 Sept.	1 Nov. 1993	1 Jan. 1993	1 Mar. 1994	1 May 1994	1 Aug. 1994
				(%)				
1*	100	82	79	77	95	102	91	82
2*	100	83	74	72	86	98	85	67
3*	100	70	67	64	101	108	80	57
4*	100	67	65	62	86	108	84	61
5*	100	78	74	71	90	100	85	70
10*	100	68	59	55	92	110	67	53
11*	100	75	73	69	86	109	92	65
6	100	85	73	71	77	84	80	69
7	100	88	88	88	97	100	93	83
8	100	92	89	83	88	89	82	77
9	100	91	85	79	81	88	83	75
13	100	88	86	84	86	92	84	75

*Shallow Profiles

moisture levels was in early March 1994, when the soil wetness ranged between 84–100% of the May 1993, values. The moisture content at all sites early in August 1994 had reached levels lower than those recorded prior to the start of the wet season in November 1993 indicating that the catchment was already drier than it had ever been in the previous year.

Spatial variability in total soil water content was detected both among and within all 12 profiles. Of the deep profiles monitored, Site 7 consistently contained the most water per unit volume of soil while Site 8 was always the driest (Fig. 8). Throughout the monitoring period the difference in wetness between these 2 sites remained fairly constant. In the other 3 deep sites the relative difference in wetness, however, continued to change at different times of year. Among the shallow profiles the pattern of changes in moisture content were similar to that observed for the deeper profiles.

Early in the summer of 1993, when steady vertical flow in the profile (which originated as surface infiltration) ceased, the contribution to deep seepage was through the draining of the soil profile. This draining process began close to the surface and as the summer progressed, migrated deeper into the profile. As the near-surface profiles began to dry as the summer progressed, a zero-flux boundary migrated into the deeper zones (Fig. 5). The zone above the boundary had a net negative gradient, resulting in moisture moving towards the surface while the profile below the boundary continued to release moisture to deep seepage.

Within and among the 3 zones dominated by transpiration from grasses and surface evaporation, transpiration from shrubs and trees, and deep seepage, respectively, continuous fluctuations occurred in the magnitude and directions of hydraulic gradients, resulting in the continuous redistribution of moisture in

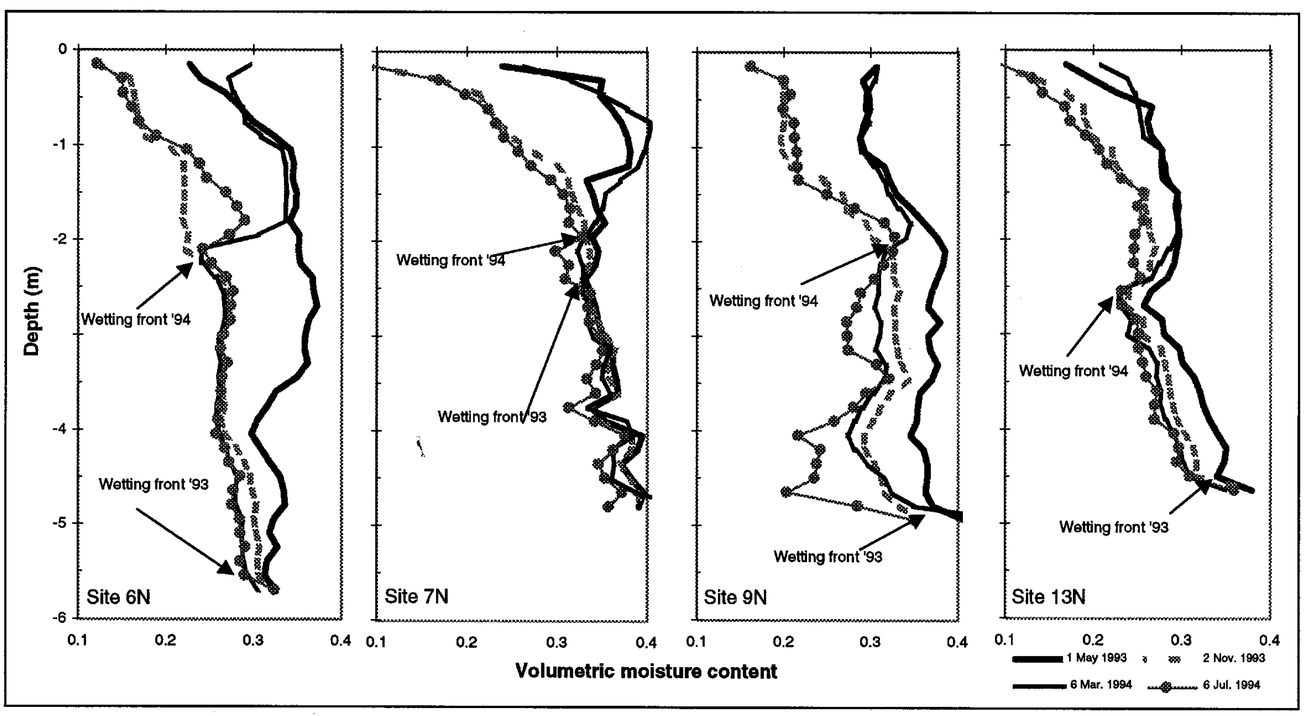

Fig. 7. Extent of migration of the wetting front in 4 sites (6, 7, 9, and 13) following a wet winter in 1993 and a significantly drier winter in 1994.

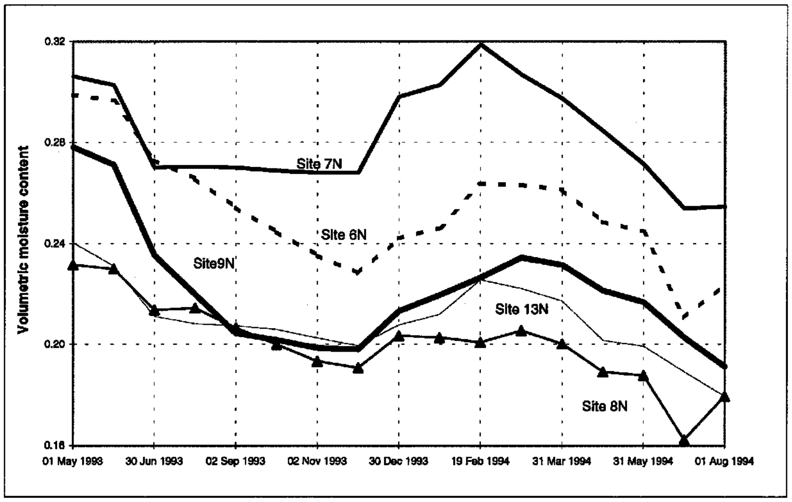

Fig. 8. Monthly changes in soil moisture content along the vertical profile of Sites 6, 7, 8, 9, and 13.

the soil (Table 2). Early in the summer, large negative gradients close to the surface resulted in the movement of water from below the grass root zone towards the surface. However, as the annual grasses dried and mulch covered the surface, little of the soil moisture was lost to the atmosphere. A significant portion of soil moisture was relocated within the grass root zones, where the soil was drier from transpiration losses that occurred before the grasses died. As this moisture was being redistributed, the continuous transpiration losses in the shrub/tree root zone allowed moisture from the near-surface profile and the deeper profiles to migrate toward this dry zone. This shifting of energy gradients in the profile resulted in a continuous movement of soil moisture between the surface and shrub/tree root zone. Within the shallow profile sites (where the bedrock lies between 1.0 and 2.0 m from the surface) there was a different hydrologic dynamic, with moisture migrating to the deeper profile in the winter. As the winter ended, the soil began to dry from gravitation drainage and from losses to ET. Once the grasses died, moisture losses were reduced, and the subsurface profile approaches a near-hydrostatic condition. Toward the end of the summer, most movement of soil water in the slopes was restricted to the deep profiles. In the shallower sections of the deep profile, the roots of transpiring trees and shrubs drew water from the soil. In the deeper profile moisture, continued to be lost to deep seepage.

Results

Parameters used in determining the site-specific water balance for Site 9 are summarized in Figure 9. Similar parameters were used in the other sites for which the dynamic moisture content was evaluated. In all 6 sites, for a brief period between January and February 1994, the amount of precipitation recorded was greater than the calculated ET and seepage losses during

Table 4. Changes in moisture content at 6 locations resulting from precipitation, ET and deep seepage.

	May–July ‘93	Aug–Oct ’93	Nov ‘93–Jan ‘94	Feb–Apr ‘94	May–July ‘94	Cummulative
Site 2N			(cm)			
Rainfall	0	1	21	25	3	50
Change water content	–13	–2	9	–3	–9	–17
ET loss	–13	–2	0	–7	–9	–31
Deep seepage loss	0	–1	–12	–21	–3	–37
Site 6N						
Rainfall	0	1	21	25	3	50
Change water content	–33	–12	12	1	–17	–50
ET loss	–22	–10	0	–4	–20	–55
Deep seepage loss	–12	–4	–9	–20	–1	–45
Site 7N						
Rainfall	0	1	21	25	3	50
Change water content	–17	–1	0	–11	–14	–43
ET loss	–10	–1	0	–6	–15	–32
Deep seepage loss	–7	–1	–21	–30	–3	–62
Site 8N						
Rainfall	0	1	21	25	3	50
Change water content	–6	–8	4	–5	–4	–20
ET loss	–5	–3	0	–6	–4	–18
Deep seepage loss	–2	–6	–16	–24	–3	–51
Site 9N						
Rainfall	0	1	21	25	3	50
Change water content	–16	–15	5	1	–12	–37
ET loss	–3	–1	0	–1	–3	–8
Deep seepage loss	–13	–15	–15	–23	–13	–79
Site 13N						
Rainfall	0	1	21	25	3	50
Change water content	–15	–3	4	–5	–10	–28
ET loss	–8	–2	–1	–7	–10	–28
Deep seepage loss	–8	–2	–15	–23	–3	–51

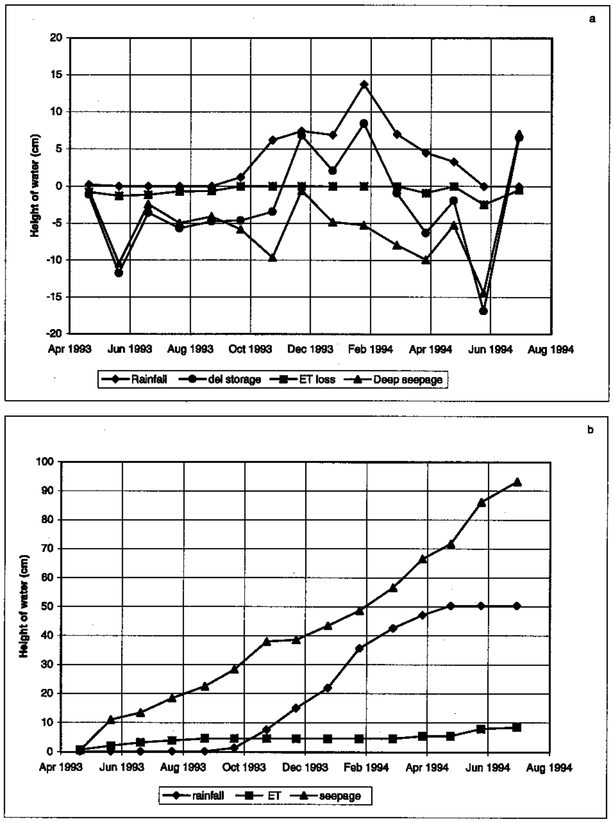

Fig. 9. Example of parameters used to calculate the (a) water balance at and (b) cumulative water losses and gains (for Site 9).

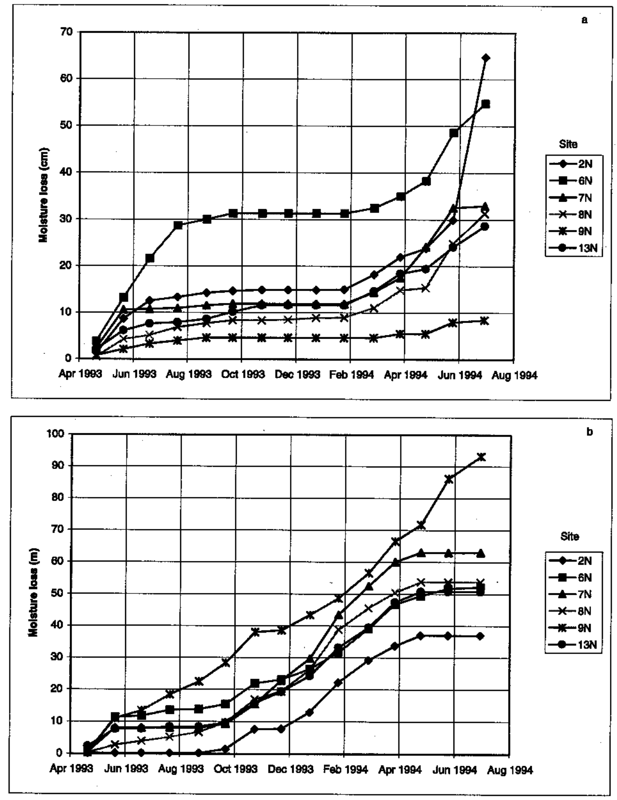

Fig. 10. a. Cumulative moisture losses to ET at Sites 2, 6, 7, 8, 9, and 13. b. Cumulative moisture losses to seepage at Sites 2, 6, 7, 8, 9, and 13.

the same period. For the remaining 14–month period, the monthly net moisture losses were much greater than net gains in all the sites. However, among sites where the monthly precipitation was assumed to be the same (since they were located in the same catchment), the monthly losses to ET and seepage varied significantly (Table 4).

At Site 2, ET losses were large (13 cm) early in the summer of 1993, and then gradually ceased over the next 6 months before increasing again in February 1994 until the end of the monitoring period (Fig. 10a). Deep seepage was detected with the start of the winter rains in 1994, and observed to continue until the end of April 1994 (Fig. 10b), during which time 33 cm of water was lost from this vertical profile. By late July1994, there was a cumulative loss of 68 cm of water to ET and deep seepage at this location.

In the remaining 5 sites, deep seepage losses were observed throughout the monitoring period. In 4 of these sites (6, 7, 8, and 13) the seepage losses were also largest (between 20–30 cm) immediately after the wet season in 1994. At Site 9, except for a brief period in November 1993, deep seepage continued at a steady rate throughout the monitoring period. By the end July 1994, the largest amount of moisture lost to deep seepage was at Site 9 (79 cm) and the least was at Site 2 (37 cm). The pattern of ET losses from these 5 sites was similar to that observed at Site 2. The largest ET losses were recorded in Site 2 (65 cm) and the least were at Site 9 (<10 cm), with the remaining sites showing losses ranging between 30 and 50 cm.

Conclusions

Solutions to a simple water-balance equation, comprising of precipitation, evapotranspiration, runoff, and storage as the key parameters, suggest that there was significant variability both spatial and temporal in the amount of soil moisture lost to evapotranspiration and deep seepage. During the 16-month monitoring period there was a total of 50 cm of rainfall that fell in the catchment. Measurements of water potential and soil moisture content at various locations within the catchment suggest that during this time losses to evaporation ranged from 9 to 55 cm while those to deep seepage ranged between 37 and 79 cm. A simple deduction of the losses (evaporation and deep seepage) from the input (rainfall) shows that all monitored locations had a substantial decrease in the amount of water that was stored in the soil profile.

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