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Soil moisture states, lateral flow, and streamflow generation in a semi-arid, snowmelt-driven catchment

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Abstract:

Hydraulic connectivity on hillslopes and the existence of preferred soil moisture states in a catchment have important controls on runoff generation. In this study we investigate the relationships between soil moisture patterns, lateral hillslope flow, and streamflow generation in a semi-arid, snowmelt-driven catchment. We identify five soil moisture conditions that occur during a year and present a conceptual model based on field studies and computer simulations of how streamflow is generated with respect to the soil moisture conditions. The five soil moisture conditions are (1) a summer dry period, (2) a transitional fall wetting period, (3) a winter wet, low-flux period, (4) a spring wet, high-flux period, and (5) a transitional late-spring drying period. Transitions between the periods are driven by changes in the water balance between rain, snow, snowmelt and evapotranspiration. Low rates of water input to the soil during the winter allow dry soil regions to persist at the soil–bedrock interface, which act as barriers to lateral flow. Once the dry-soil flow barriers are wetted, whole-slope hydraulic connectivity is established, lateral flow can occur, and upland soils are in direct connection with the near-stream soil moisture. This whole-slope connectivity can alter near-stream hydraulics and modify the delivery of water, pressure, and solutes to the stream. Copyright © 2005 John Wiley & Sons, Ltd.

KEY WORDS runoff generation; soil moisture; semi-arid; snowmelt; hydraulic connectivity

INTRODUCTION

Runoff generation in sloping terrain has been a topic of hydrologic research for decades, yet new research continues to unravel the complex mechanisms by which catchments in all environments translate vertical inputs of precipitation to lateral runoff. An important recent advancement is the recognition that hydraulic connectivity on hillslopes is a necessary condition for widespread lateral flow to occur. This has significant implications for index-based hydrologic modelling approaches that assume that regions in the catchment are hydraulically connected by topographically controlled lateral subsurface flow (Beven and Kirkby, 1979; Grayson *et al.*, 1997; Western *et al.*, 1999; Stieglitz *et al.*, 2003). In humid regions, a variety of topographically driven processes operate that encourage hydraulic connectivity during much of the year. However, in dry environments, where evapotranspiration exceeds or meets precipitation for much of the year, topography may have very little influence on soil moisture distribution, and lateral subsurface flow may be insignificant except in brief windows of time (Newman *et al.*, 1998; Puigdefabregas *et al.*, 1998; Flerchinger and Cooley, 1999; Gomez-Plaza *et al.*, 2001; Chamran *et al.*, 2002).

Grayson *et al.* (1997) showed that, for a temperate region in Australia, two distinct soil moisture states, bounded by brief transition periods, exist during a year: (1) a dry state, in which there are no topographic controls on moisture distribution, hillslope regions are hydraulically unconnected, the dominant flow direction

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is vertical, and spatial patterns of soil moisture are unorganized; (2) a wet state, in which topography is a primary control on moisture distribution, hillslope regions are connected via lateral flow and soil moisture patterns are organized spatially. Western *et al.* (1999) demonstrated that terrain-based indices work well to predict soil moisture distribution during wet conditions, but they do not work well during dry conditions, and that a variety of indices should be used that take into account the processes that occur in particular catchments under particular climates. A challenge in the multi-index modelling approach is to identify triggering mechanisms to initiate switches between moisture states that account for the lateral flow processes that operate within a catchment.

Grayson *et al.* (1997) suggested that lateral flow occurs when the relative saturation of the soil column is between 0·6 and 0·8. Shaman *et al.* (2002) incorporated a moisture-dependent switch based on a percentage of field capacity into a TOPMODEL formulation to allow lateral flow from perched water tables in the unsaturated zone. In a case study in the Panola Mountain Research Watershed, Freer *et al.* (2002) showed that, above a threshold moisture content, a small amount of bypass flow to depth drives saturated lateral preferential flow at the soil–bedrock interface. Stieglitz *et al.* (2003) implied that deep soil zones can act as barriers to whole-slope connectivity because relatively dry pockets of soil may exist at the soil–bedrock interface. When deep soil regions become sufficiently wet to allow for lateral downslope flow to occur, ridge to valley connectivity can be established, along with the associated delivery of enhanced solute concentrations to the valley-bottom stream.

An important consideration in this preferred moisture state concept is that the climatically controlled switch from the dry state to the wet state is synchronous with a switch from primarily vertical to primarily lateral flow, which is controlled by hydraulic properties of the soil and the rate of water delivery to the soil. In this paper we argue that these two switches may not be synchronous in regions where the occurrence of snow rather than rain delays the delivery of water to the soil. The rate of water input to the soil in such regions is primarily controlled by the energy balance between the soil, snowpack and atmosphere, rather than by precipitation depth and frequency. We propose that this asynchronous switching establishes the necessary conditions for deep soil connectivity barriers, and that whole-slope connectivity does not occur until these barriers are breached.

The focus of this paper is to use the concepts of preferred moisture states and hillslope hydraulic connectivity to explore relationships between hillslope soil moisture and streamflow in the Dry Creek Experimental Watershed (DCEW), a semi-arid, snowmelt-driven catchment. Specifically, we investigate the influence of hydrologic seasonality on the development and breakdown of deep soil connectivity barriers in the catchment. These ideas are presented and discussed following a site description in the next section, and a description of field methods in the section thereafter. Then follows a section presenting a one-dimensional soil water balance based on field observations and the Simultaneous Heat and Water (SHAW) model to provide insight into soil water flow and hillslope–streamflow connections. Subsequently, key results are taken from the water balance to discuss the 'hydrologic seasonality' of the site defined by five characteristic moisture periods with distinct connectivity conditions. There then follows a section that presents further modelling exercises using the SHAW model and HYDRUS2D (Simunek *et al.*, 1999) to clarify the influence of the energy balance over hydrologic seasonality and the development and subsequent breakdown of deep soil connectivity barriers within the hillslope. The penultimate section combines the observations and simulations into a conceptual model of how streamflow originates throughout the year with respect to hillslope–stream hydraulic connectivity. The final section presents a summary of our conclusions and recommendations for future work.

SITE DESCRIPTION

This study was conducted in a small catchment with an ephemeral stream draining 0·02 km^2 within the larger DCEW (27 km^2) in a range of hills near Boise, Idaho, USA, known as the Boise Front (Figure 1). The elevations in the DCEW range from about 900 to 2100 m. The lower elevation slopes are partly mantled

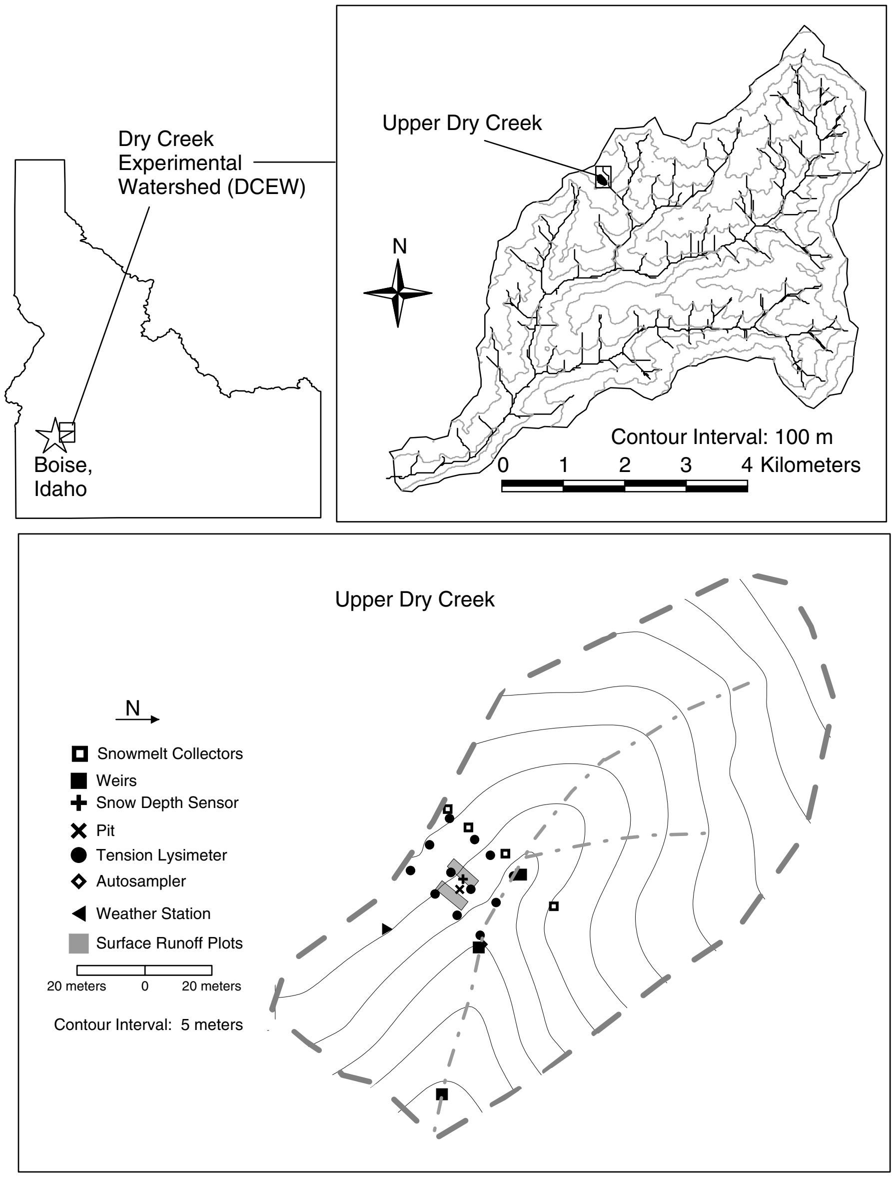

Figure 1. Location and instrumentation of the study site. The star indicates the location of Boise, ID

with sagebrush, which is succeeded by chaparral and then by fir, spruce, and pine at higher elevations. In the Koppen classification system, the lower portion of the DCEW is classified as a steppe summer dry climate (BSk), and the high elevations are classified as moist, continental climate with dry summers (Dsa) (Henderson-Sellers and Robinson, 1986). The 0·02 km^2 study catchment, called Upper Dry Creek (UDC), has a mean elevation of 1620 m, a relief of 35 m, and an east–west orientation. At this elevation the summers are hot and dry, but a persistent snowpack exists from around early November through to March or April. Approximately half of the annual precipitation falls as snow. Both the north- and south-facing aspects are steep slopes of approximately 20°, which converge in a narrow valley bottom with essentially no riparian zone for most of the catchment. Soils are formed from weathering of the underlying Idaho Batholith, a granitic intrusion ranging in age from 75 to 85 million years. Soils are classified as coarse-loamy, mixed mesic Ultic Haploxeroll (Harkness, 1997). A 10×10 m^2 grid survey at 55 locations found soil depths ranging from 7 to 105 cm, with the mean and the mode of the population being 29 cm and 22 cm respectively. Texture analysis by the hydrometer method found little variation in sand (74 to 78%), silt (15 to 17%) or clay (7 to 9%) among the A, B, and C horizons of a 70 cm deep soil pit excavated near midslope on the north-facing aspect. Various empirical approximations indicate that the field capacity for this soil texture is near 18% volumetric water content (Dingman, 1994).

The stream draining UDC is ephemeral and typically begins flowing soon after the onset of the seasonal snowpack each year. In early and mid winter, streamflow typically remains low, indicating basal snowpack melt, with small peaks associated with minor snowmelt events from the south-facing slopes. Winter snowmelt on the south-facing slopes is episodic, and occasionally intense enough to melt the snowpack completely; however, snowfall frequency is adequate to keep the slope covered most of the winter. Winter snowmelt on the north-facing slope is uncommon, which in combination with drift deposition of snow scoured from the adjacent catchment and the top of the slope results in much greater accumulation of snow on the north-facing slope. Yenko (2003) used end-member mixing analysis and chemical hydrograph separations to show that deep regional groundwater does not contribute to the ephemeral streamflow in this catchment.

FIELD METHODS

A monitoring programme has been operating in UDC since May 1999 to investigate cold-season runoff generation processes. A meteorological station on the north-facing slope records precipitation, snow depth, air temperature, relative humidity, wind speed, wind direction, incoming shortwave radiation, soil moisture, and overland flow on a Campbell Scientific CR10X datalogger. Rainfall and snowfall are measured in a weighing-bucket gauge using a load cell with an alter shield mounted on a post 1·5 m above the ground surface. Snow depth is monitored hourly at one point on the north-facing slope by a Judd sonic depth sensor (accuracy ± 1 cm, or 4% of distance to target). Occasional snow surveys are performed using a Mt Rose snow sampler to obtain basin-average snow water equivalent. Surface runoff is monitored in two 10 m $\times$ 3 m plots bounded by stainless-steel barriers. Overland flow from each plot is routed to 1892 l (500 gallon) collection tanks where the depth is recorded hourly using a Campbell Scientific shaft encoder. Electrical conductivity in the stream is logged hourly using a Campbell Scientific CS547A sensor (accuracy ± 5%). Streamflow and soil moisture measurements required the development of field calibrations.

Streamflow is gauged at three plywood weirs with a v-notch bevelled to a sharp edge. Each weir was grouted to the granite bedrock with bentonite. Stage in the ponds behind the weirs is monitored by Global Water WL14 pressure transducers (accuracy 0·5 cm). A stage–discharge relationship has been developed for each weir by timing the period required for the discharge from the weir to fill a bucket of known volume. The average of five to seven discharge measurements of this type was taken as discharge for that stage. Ten discharge measurements were taken at each weir during the study period. The average of standard deviations for all discharge measurements was 3%.

Soil moisture was monitored in two vertical profiles, 2 m apart and 15 m upslope from the stream channel on the north-facing slope. Moisture content is monitored at 15 min intervals with water content reflectometers (Campbell Scientific, Logan, UT) at depths of 5, 15, 30, 65, and 100 cm in one pit (henceforth called Pit 100), and at 5, 15, 30, 45 and 65 cm in another pit (henceforth called Pit 65). Bedrock was encountered a few centimetres below the deepest sensor in each pit. Sensor response was calibrated against occasional readings of collocated time domain reflectometry probes to an accuracy of ± 2% (Chandler *et al.*, 2004). Soil texture was determined when the pits were excavated. Subsequent soil depth surveys revealed that the two soil pits were serendipitously located in the deepest soils in the catchment.

Water samples were collected from the stream periodically through the winter and more frequently during the snowmelt period with an ISCO autosampler and analysed using ion chromatography for major inorganic constituents at Utah State University Analytical Labs (Logan, UT).

SOIL WATER FLOW AND HILLSLOPE STREAM CONNECTIONS

In this section we use field observations described in the previous section coupled with the SHAW model (Flerchinger *et al*., 1996) to compute the components of the one-dimensional soil water balance and its connections to streamflow.

The SHAW model

The SHAW model simulates heat, water and solute transfer in a one-dimensional profile extending from the top of a plant canopy or the snow, residue or soil surface to a specified depth within the soil. The system is represented by integrating detailed physics of vegetative cover, snow, residue and soil into one simultaneous solution (see Flerchinger *et al*. (1996) for details). Daily or hourly predictions include evaporation, transpiration, percolation, soil frost depth, snow depth, runoff and soil profiles of temperature, water, ice and solutes. With climate data including precipitation P as inputs, the SHAW model computes soil moisture time series at defined depths, as well as the daily water balance of a vertical profile, as follows:

$$P - \mathrm{INT} - \mathrm{ET} - \Delta S_{\mathrm{can}} - \Delta S_{\mathrm{snow}} - \Delta S_{\mathrm{res}} - \Delta S_{\mathrm{soil}} - \mathrm{POND} - \mathrm{RUN} - \mathrm{DP} + \mathrm{error} = 0 \quad (1)$$

where INT is intercepted precipitation on top of the canopy, ET is total evapotranspiration, ΔS_{can}, ΔS_{snow}, ΔS_{res}, and ΔS_{soil} are the changes in storage in the canopy, snow, residue, and soil respectively, POND is water ponded at the surface, RUN surface runoff generated by infiltration excess or saturation excess, and DP is deep percolation. The error term is the only residual and, therefore, represents the actual errors in the individual components.

In order to evaluate the potential timing for lateral flow along the soil–bedrock interface we redefine DP. DP is calculated from the Darcian flux of water moving between the deepest two soil nodes in the model profile. If vertically infiltrating water encounters a sloping impermeable boundary, then it will flow laterally along that boundary. Therefore, we redefine the DP component of the SHAW model as bedrock flow (BF) and recognize through Equation (3) that it is the sum of losses to deep groundwater in fractured bedrock ($\mathrm{GW}_{\mathrm{out}}$) and lateral subsurface flow along the soil–bedrock interface L_{out}:

$$\mathrm{DP} \equiv \mathrm{BF} \quad (2)$$

$$\mathrm{BF} = \mathrm{GW}_{\mathrm{out}} + L_{\mathrm{out}} \quad (3)$$

Although we cannot distinguish between $\mathrm{GW}_{\mathrm{out}}$ and $L_{\mathrm{out},}$ we assume that the timing of BF is the same as the timing of flow along the soil–bedrock interface.

Model performance

We ran a continuous SHAW simulation for one complete year using site climate data and for the soil data to represent each soil profile described in the 'Field methods' section. The model performed quite well using two criteria: comparison of the cumulative error component to total precipitation, and comparison of the predicted and observed soil water content time series. The cumulative daily error over the year was −25 mm, indicating that the SHAW model accounted for all but 4% of precipitation. The predicted soil water content matches the observed soil water contents in Pit 65 and Pit 100 very well for depth-averaged measurements

Table I. Nash-Sutcliffe goodness-of-fit coefficients comparing simulated and observed moisture contents in Pit 65 and Pit 100 (see Figure 2)

Depth (cm)	Nash–Sutcliffe coefficient	
	Pit 65	Pit 100
15	0·88	0·85
30	0·82	0·84
65	0·88	0·29
Average	0·92	0·91

(Nash–Sutcliffe coefficients of 0·92 and 0·91 respectively, and also for the individual soil depths with the exception of 65 cm in Pit 100 (Table I).

The SHAW model reproduces the near-surface moisture levels well (Figure 2a and b), although water inputs tend to be overpredicted for the peaks, particularly during the snowmelt period. With greater depth, the simulated soil moisture response may or may not lag behind the measured increase in soil moisture, depending on the soil pit (Figure 2c). The success of the model in predicting soil moisture at shallow depths, and at depth in Pit 65, supports the premise that soil moisture observations that diverge from model predictions occur by mechanisms that violate the one-dimensional or Darcian assumptions of the model. The following subsection explains why we believe that the gradual wet-up observed at 65 cm in Pit 65, and simulated by the SHAW model, represents the moisture profiles that occur when matrix flow, which the SHAW model is written to simulate, is the sole mechanism of vertical infiltration.

Vertical soil moisture movement

The SHAW model employs a Green–Ampt approach to translate soil moisture through the vertical profile, by matrix flow. It does not account for either vertical bypass (macropore) flow or lateral flow. We hypothesize that the overprediction of peak soil moisture and apparent advanced wetting at depth in Pit 100 may be caused by bypass flow within the hillslope, which is not accounted for by the one-dimensional SHAW model.

The measured soil moisture patterns for 30 cm are uniform between pits, but precede the simulated response, indicating a non-matrix flow component. The nearly simultaneous wetting at depths 15 cm (Figure 2a) and 30 cm (Figure 2b) in both pits indicates relatively uniform delivery of water throughout this layer. This pattern is not exhibited by the SHAW simulations, which agree with the measured timing of water delivery to 15 cm (Figure 2a) for both pits, but is similarly lagged at 30 cm, which would be an expected result from the advance of a regular wetting front.

The measured soil moisture patterns for 65 cm are, on the other hand, not uniform between pits, although the pattern at Pit 65 is matched quite closely by the SHAW simulation for that depth. The good agreement between field data and model simulation is expected, given the high degree of uniformity within the mineral soil horizons at this site (Chandler *et al*., 2004) and implies that the gradual wet-up is due to vertical flux through the soil matrix being the dominant hydrologic process. The rapid wetting at 65 cm in Pit 100 (Figure 2c) could occur as a result of preferential vertical flow to the base of the soil column or from lateral flow from upslope regions. An argument can be made for the existence of bedrock surface topographic concentration of upslope flow (Freer *et al*., 2002) based on the observation that Pit 100 was located in the deepest soil on the hillslope and probably in a bedrock hollow. This argument is further supported by the soil moisture record at 100 cm, which, following a very abrupt wetting, maintains the highest soil moisture content in the profile throughout most of the winter. In summary, observations and simulations suggest that soil water movement by matrix flow produces relatively dry pockets of soil that persist through the winter in the deep soil zones. However, preferential flow may cause wetting to the soil–bedrock interface.

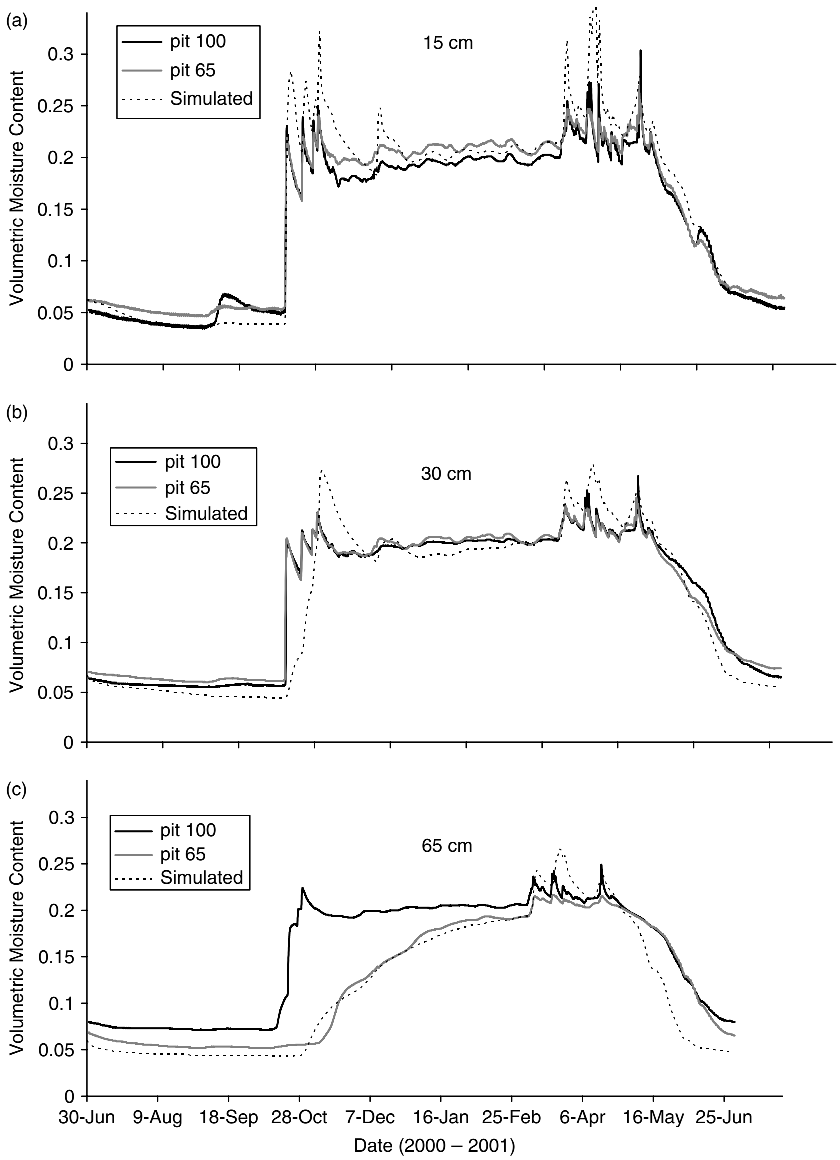

Figure 2. Observed and simulated volumetric moisture content at (a) 15, (b) 30, and (c) 65 cm depths in Pit 65 and Pit 100

Water budget

The terms POND, RUN, INT, ΔS_{can}, and ΔS_{res} from Equation (1) were negligible and are not included in Table II. The SHAW model never simulated non-zero values for POND and RUN, nor did we ever observe surface runoff at midslope runoff plots during the study. The term ΔS_{snow} is incorporated into the terms Snow and Snowmelt in Table II. The annual precipitation is nearly evenly divided between rain and snow (Table II). The difference (39 mm) between total precipitation (568 mm) and water input (529 mm) indicates that evaporative losses are a small fraction of the total ET (11%), confirming the observation that the snowpack provides efficient storage of accumulated precipitation. Total evapotranspiration is then dominated by plant demands and accounts for 62% of the annual precipitation. The SHAW model simulates bedrock flow of

Table II. Monthly water budget

	P[a] (mm)	Rain[b] (mm)	Snow[c] (mm)	Snowmelt[d] (mm)	Water input[e] (mm)	ET[f] (mm)	ΔS_{soil}[g] (mm)	Bedrock flow[h] (mm)	Error[i] (mm)
July 2000	3	3	0	0	3	17	−7	0	−5
August 2000	1	1	0	0	1	4	−2	0	0
September 2000	26	26	0	0	26	28	2	0	0
October 2000	130	130	0	0	130	29	76	0	0
November 2000	62	22	40	0	22	12	10	0	0
December 2000	78	5	74	2	8	8	11	0	0
January 2001	68	8	60	7	15	8	5	4	0
February 2001	29	7	22	9	16	8	2	8	0
March 2001	53	35	17	164	199	23	4	167	−1
April 2001	77	33	44	35	69	54	−15	61	−2
May 2001	20	20	0	0	20	124	−57	3	−9
1 June 2001	20	20	0	0	20	41	−24	0	−8
Total	568	311	257	217	529	354	5	244	−25

[a] Measured precipitation.
[b] The component of P that fell as rain calculated by the SHAW model.
[c] The component of P that fell as snow calculated by the SHAW model.
[d] Meltwater loss from the snowpack calculated by the SHAW model.
[e] Rain plus Snowmelt.
[f] Evapotranspiration.
[g] Change in stored soil moisture calculated by the SHAW model.
[h] Equivalent to deep percolation calculated by the SHAW model.
[i] Residual of water balance.

244 mm or 42% of total precipitation from Pit 100. Within the error reported by the SHAW model, the annual water balance seems logical.

Distinct seasonal patterns illustrated in Figure 3 are expressed in each of the monthly water balance components (Table II). Precipitation is inadequate to meet evaporative and plant demands from May through to August, and soil moisture storage is drawn down by plant demands. Fall precipitation is initially attended by an increase in actual ET during September and October, despite the decline in potential ET during that season. As the ET demands decline, the fall and early winter rains are stored in the soil (October–December) with no bedrock flow component predicted by the SHAW model. As the precipitation changes to snow with the onset of winter, from December through to February, water inputs decrease from the fall, but soil moisture storage is satisfied and bedrock flow begins in the shallow soil regions of the catchment. Spring snowmelt (March–April) drives a peak in bedrock flow, as ET begins to rise. Following snowmelt, soil moisture and bedrock flow decrease sharply, reflecting the negative balance between water input (primarily as rain) and ET.

CHARACTERISTIC SOIL MOISTURE PERIODS

Based on the observation of seasonal patterns within the monthly water balance values above, we use the difference between water input and ET as an indicator of hydrologic seasonality (Figure 3, Table III). This approach recognizes that the hydrologic status of the site is more dependent on the one-dimensional flux of water into (downwards) or out of (upwards) the soil than on precipitation. It also allows the division of the year into hydrologic seasons that are independent of the calendar year. We identify five characteristic soil moisture conditions in the water year: (1) a dry period; (2) a transitional wetting period; (3) a wet, low-flux period; (4) a wet, high-flux period; (5) a transitional late-spring drying period. Table III aggregates the daily

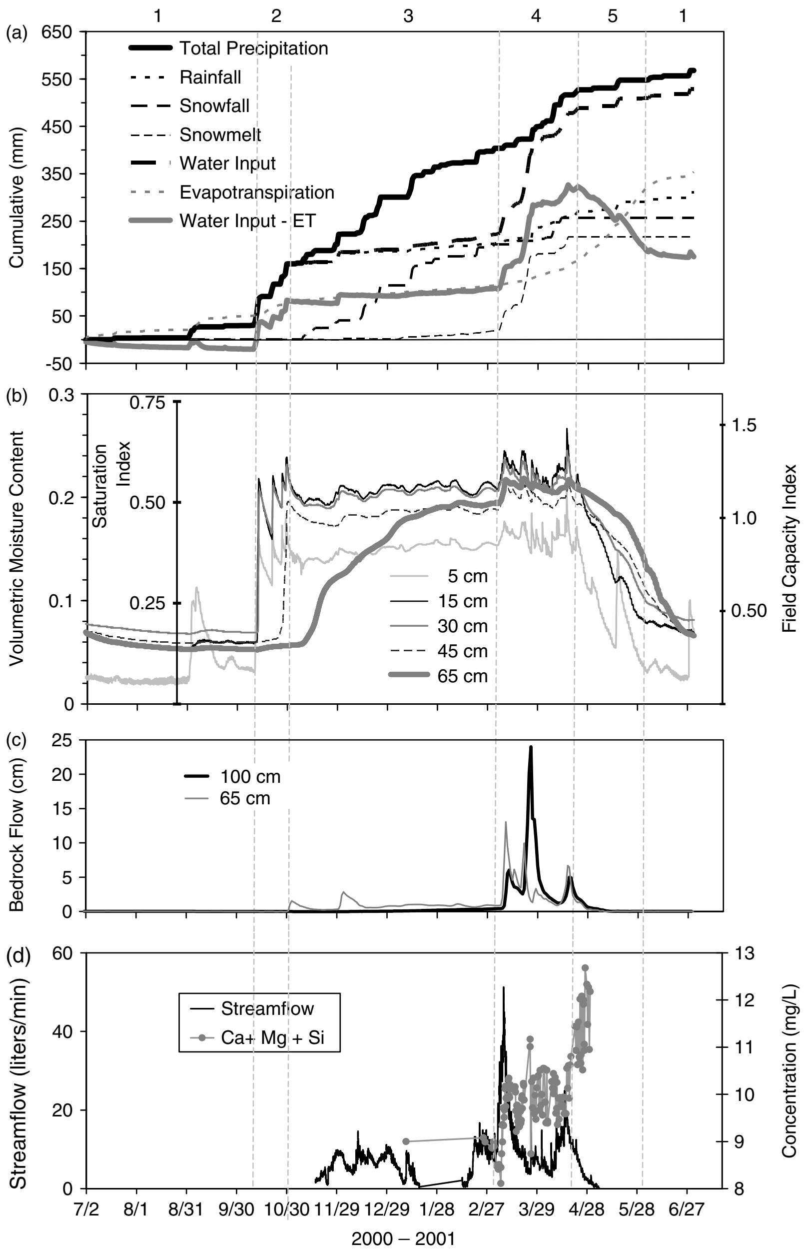

Figure 3. Timing of events during the 2001 water year (a) at the land—atmosphere interface, (b) in the soil column, (c) at soil–bedrock interface modelled by SHAW, and (d) in the stream. The numbers across the top and the grey vertical line refer to the characteristic moisture periods

values for the water balance components within those seasons. The wet state is split into low-flux and high-flux states due to a time lag between when the switch from a dry state to a wet state occurs and when the switch from primarily vertical to lateral flow occurs due to the presence of a persistent snowpack.

Figure 3 illustrates the soil moisture behaviour in Pit 65 (Figure 3b), along with the timing of other hydrologic processes (Figure 3a, c, and d). Most of the discussion will focus on Pit 65, rather than Pit

Table III. Water budget elements during the five characteristic moisture periods

Period	Duration (days)	ΔS_{soil}[a] (cm)	ET[b] (cm)	P[c] (cm)	Water input[d] (cm)	Bedrock flow[e] (cm)	ET[b] (mm day^{-1})
1	104	0	6	3	3	0	0·57
2	29	10	3	13	13	0	1·10
3	117	6	3	24	6	1	0·28
4	51	0	6	12	27	23	1·10
5	34	−16	17	4	4	0	2·71

[a] Change in stored soil moisture calculated by the SHAW model.
[b] Evapotranspiration.
[c] Measured precipitation.
[d] Rain plus Snowmelt.
[e] Equivalent to deep percolation calculated by the SHAW model.

100, because of the close agreement between the simulated and observed trends in Pit 65, which we attribute to a lack of lateral flow influence. Unless noted otherwise, all discussions below refer to Figure 3. The vertical, cross-cutting lines on Figure 3 correspond to boundaries between the five moisture periods, which are defined by sharp and sustained breaks in slope on the Water Input–ET line on Figure 3a. Note that there are three vertical axes on Figure 3b. The first is volumetric moisture content. The second is a saturation index (SI), which is calculated by dividing the volumetric moisture content by the porosity of the soil. The third is a field capacity index (FCI), which is calculated by dividing the volumetric moisture content by the field capacity of the soil. For graphical simplicity, one porosity (0·35) and one field capacity (0·18) were used for all depths.

Period 1: summer dry period

The summer dry period is marked by relatively stable, low moisture contents throughout the vertical profile. The surface is essentially air dry, and the lower soil depths remain between 5 and 7% volumetric water content. This period is equivalent to the preferred dry state described by Grayson *et al.* (1997). Evapotranspiration exceeds precipitation, and hence water input, throughout this period (Table III). Occasional summer thundershowers wet the surface, but most of the imbibed water is lost to evapotranspiration before the wetting front reaches 15 cm. Subsurface lateral flow is non-existent and the streambed is dry.

Period 2: transitional fall wetting period

The transitional fall wetting period is marked by a rapid increase in moisture content throughout most of the soil profile. As rainstorm frequency increases and day length and temperature decrease, water input exceeds evapotranspiration (Table III), allowing soil moisture to accumulate. Once the field capacity is reached in the upper 30 cm, the wetting front progresses downward into the mineral soil towards the bedrock interface. In the mineral soil, input fluxes are attenuated with depth, but the wetting response becomes similar to the near-surface soil, once the mineral soil has also exceeded field capacity, as shown by the response at 45 cm depth. During this transitional period, wetting front advance depends on the rate of water input in excess of evapotranspiration. As the cool season progresses, storms deliver snow rather than rain and the input of water to the soil is greatly reduced. The near-surface soil moisture levels stabilize marking the end of this transitional period. If the snow begins and shuts down the water supply soon enough, then deep soil zones will remain relatively dry at the end of this period.

Period 3: winter wet, low-flux period

This period is marked by relatively stable moisture levels near field capacity. The term wet is used to equate this period to the wet period described by Grayson *et al.* (1997) A key difference, however, is that precipitation falls as snow, so that moisture input into the soil is controlled by the energy balance of the snowpack, rather than by direct precipitation as during the wetting period. The precipitation/evapotranspiration ratio is higher than in the wetting period (Table III), but the snowpack reduces the actual water input to the soil so that the water input rate is less than during the wetting period. The shift into this period, therefore, depends on the form of the precipitation, rather than on the amount. Minor amounts of basal snowmelt occur, causing a gradual increase in near-surface moisture contents through the winter, but vertical moisture redistribution in the soil column is limited because of the low rate of water delivery to the surface and low hydraulic conductivity of the unsaturated soil. Given the limited vertical movement of soil moisture, lateral flow is even less likely, except at the soil–bedrock interface in shallow soil regions in the catchment. Bedrock flow occurs in the shallow upland and lowland soils, but the deeper soils in the midslope regions remain well below field capacity, except where vertical bypass flow has occurred. The upland bedrock flow may begin to establish local connectivity by flow along the bedrock interface, but the deep, relatively dry midslope soil limits its downslope impact. Upslope and downslope regions are hydraulically independent of each other.

Streamflow initiates at the beginning of this period and continues through the winter. Water exfiltrates from the stream bed high up in two concave regions draining south-facing exposures (Figure 1) with soil depths around 20 cm or less. Streamflow decreases between the upper and middle weirs, indicating a loss of water to the substrate and no further contributions from the hillslopes downstream of the source areas (Figure 4). The loss of water likely contributes to the growth of a near-stream saturated wedge on top of the underlying granite.

Period 4: spring wet, high-flux period

The onset of warm weather ripens then melts the snowpack and the rate of water input increases substantially, initiating a switch to a truly wet state with fully connected lateral hillslope flow. The rate of water input is much higher than the rate of evapotranspiration and is similar to that in the wetting period. In this period, however, water inputs produce an immediate response in streamflow, whereas similar water input events in the wetting period produce no streamflow. The moisture content in the deepest soils reached field capacity near the end of the previous period; now, bedrock flow occurs in response to water inputs. Evidence of flow along the soil–bedrock interface includes water contents above field capacity and rapid moisture responses to water input events at depth. Two critical observations are that the cation concentration in the stream increases substantially when bedrock flow is indicated, and that the reach between the upper and middle weirs switches from losing water to gaining water (Figure 4). In the penultimate section we argue that the occurrence of bedrock flow in the deepest soils establishes whole-slope hydraulic connectivity, which delivers solutes from a new source that has been previously disconnected from the stream.

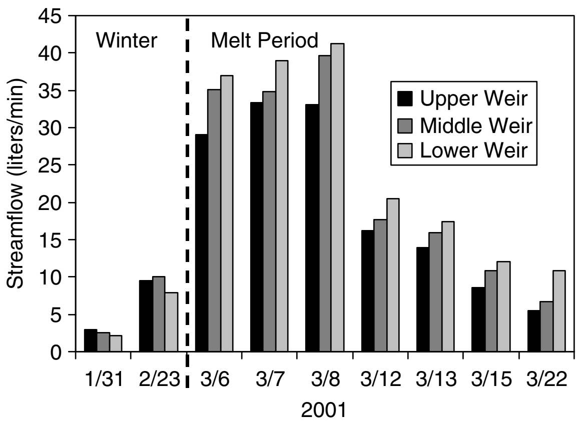

Figure 4. Measured values of streamflow at three weirs. Upper, middle, and lower weirs are in order from upstream to downstream. Note that the stream switches from losing water in the winter to gaining water in the snowmelt period

Period 5: late-spring drying period

Following the final snowmelt event, moisture contents begin a rapid decline as evapotranspiration greatly exceeds water inputs. Shallow moisture contents drop below field capacity early in this period, but bedrock flow and streamflow continue then cease simultaneously when basal midslope soil moisture drops below field capacity. Ion concentrations in the stream remain high through the receding flows.

VIRTUAL EXPERIMENTS ON CLIMATE, HILLSLOPE PROCESS AND STREAMFLOW

We used two model simulations to clarify key concepts from the 'Soil water flow and hillslope stream connections' section by testing two hypotheses: (1) that the vertical distribution of soil moisture for this site is most affected by the seasonality of the energy balance, which limits water input in the winter and augments water input in the spring; (2) whereas initiation and maintenance of streamflow are primarily dependent on the availability of water inputs to the soil, they are secondarily controlled by the hydraulic connectivity within the hillslope.

The first simulation was designed to test hypothesis 1 by exploring the hydrologic response and impact on the 'hydrologic seasonality' if the precipitation in the winter were rain rather than snow in a one-dimensional SHAW simulation for Pit 65. For this scenario we simply altered the air temperature data during the winter to be constant and above freezing, so that the SHAW model simulates all precipitation as rain. All other variables remained the same as actual conditions. Under this scenario the moisture content at the base of the soil column wets up sharply (Figure 5a), like all other depths, rather than gradually, as occurs under natural conditions when a snowpack develops (Figure 5b and c). The implication of this result is that, under the snowpack scenario, pockets of relatively dry soil may persist into the winter if the wetting front from fall rain does not satisfy the storage component before the onset of the winter snowpack. These dry pockets then buffer the translation of the low-flux winter water inputs into bedrock flow. We suggest that as long as these dry pockets exist along the soil bedrock interface they act as connectivity barriers and limit catchment-wide lateral subsurface flow.

The second simulation was designed to test the impact of a dry soil connectivity barrier on streamflow generation. This is a two-dimensional simulation of matrix flow using HYDRUS2D, a finite-element model for simulating movement of water using the Richards equation for saturated–unsaturated water flow (Simunek *et al.*, 1999). The model geometry consisted of a 44 m hillslope (slope distance) with an impermeable bottom sloping linearly at 20°. The soil thicknesses at midslope, ridgetop, and hillslope base were 65 cm, 30 cm, and 30 cm respectively. The soil's texture was uniform and identical to our field soils. The Brooks–Corey soil hydraulic functions with no hysteresis were used (Brooks and Corey, 1966), and vegetation was not included. Rainfall was added at a constant rate to produce a uniformly advancing wetting front until connectivity occurred along the bedrock interface. The constant addition of water was intended to eliminate the influence of discrete events on soil moisture behaviour. We emphasize that this simulation is not intended to represent actual conditions in the catchment, but serves to illustrate the isolated influence of deep-soil dry barriers on hydraulic connectivity and near-stream hydraulic head.

Flow along the bedrock interface began early in the simulation in the shallow regions, but was delayed in the deeper regions as the wetting front advanced uniformly (Figure 6). When the wetting front reached the base of the deep midslope soils the uplands became hydraulically connected to the lowlands and there was a dramatic jump in the hydraulic head at the base of the slope. Prior to this point the heads in the near-stream margins were influenced upslope only as far as the dry midslope soils. Because of the model conditions, the increase in hydraulic head in Figure 6 was a function of hydraulic connectivity. In a field situation, this jump in hydraulic head could cause a reversal in the hydraulic gradient in the near-stream saturated wedge, forcing substrate water into the stream.

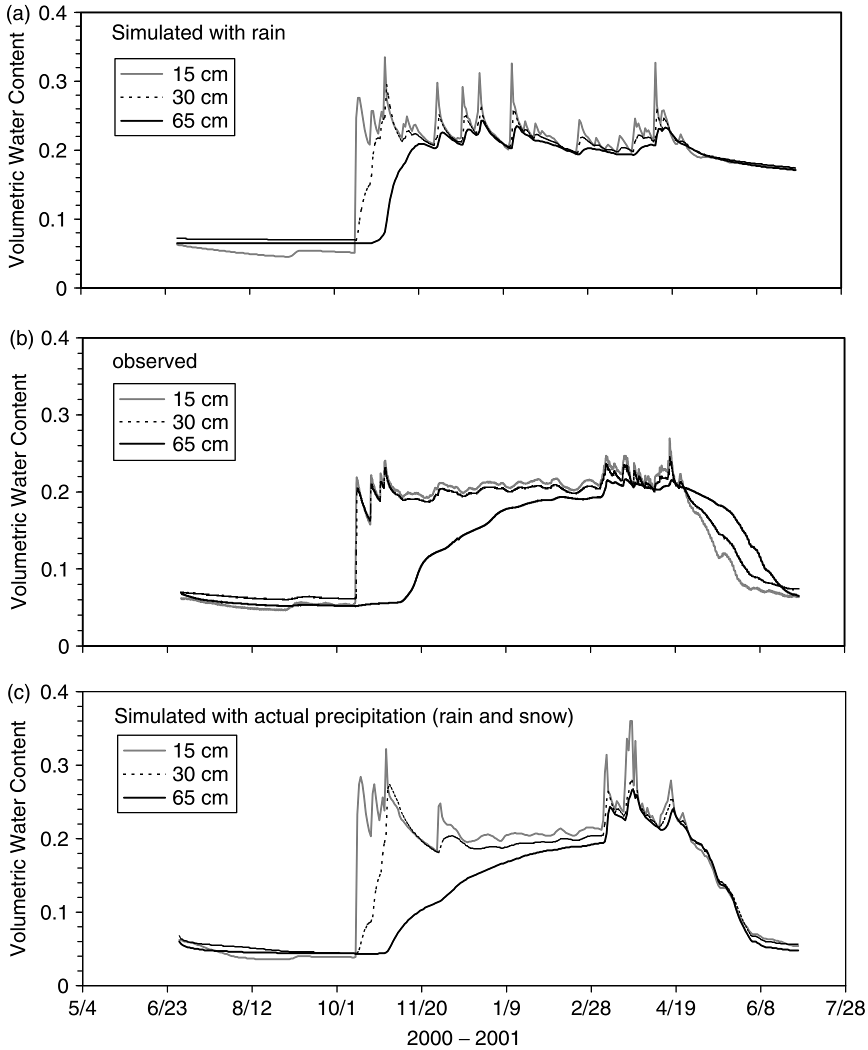

Figure 5. (a) Volumetric soil moisture content in pit4 simulated by the SHAW model when winter snow is changed to rain. Note the sharp wet-up at the base of the soil column. (b) Observed volumetric soil moisture content in pit4. (c) Volumetric soil moisture content in Pit 65 simulated by the SHAW model under actual meteorological conditions

CONCEPTUAL MODEL OF THE LINK BETWEEN HYDRAULIC CONNECTIVITY AND RUNOFF GENERATION

Here, we assemble the observations and simulations of the previous three sections into a conceptual model of how streamflow originates throughout the year. To summarize, the conceptual model must explain the following set of observations: (1) Streamflow is dilute and loses water to the substrate during the winter. (2) Bedrock flow occurs in the uplands and lowlands during the winter, but a relatively dry barrier exists at the base of the deep midslope soils. (3) When the deep soil pits become wet to depth, ion concentrations in the stream increase substantially in response to the next significant water input event. (4) The stream switches from losing water to gaining water coincident with the ion jump. The following discussion explains observations, but includes some speculation and should be regarded as a conceptual model only.

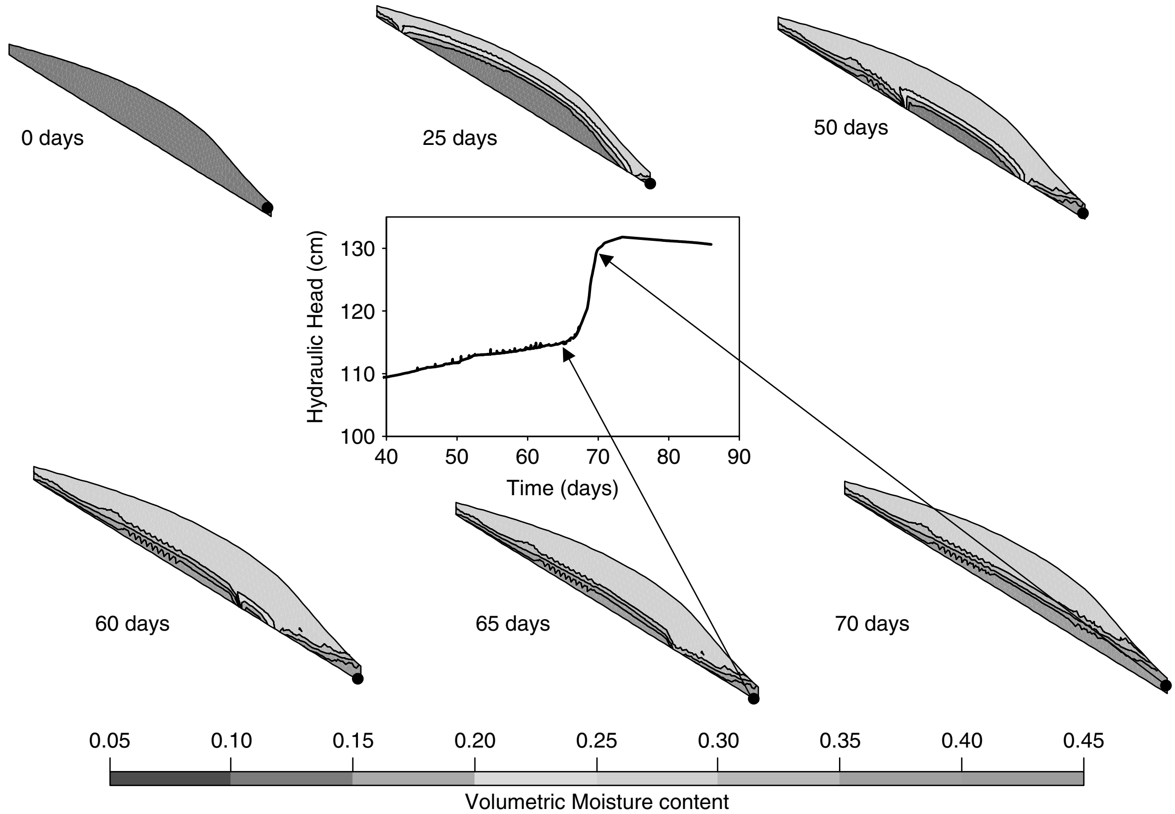

Figure 6. Cross-section of volumetric moisture content at various times during a HYDRUS2D simulation in an idealized hillslope under constant precipitation. The time series represents the hydraulic head at the base of the hillslope (black dot). Note that the pressure head at the base of the hillslope increases gradually as the wetting front advances, but increases dramatically after day 65 when the soil/bedrock interface becomes completely wetted

As summer changes to fall, a shift in the balance of evapotranspiration to precipitation causes the soil moisture in the catchment to switch from a dry state to a wet state. Thereafter, the onset of a snowpack reduces the downward advance of the wetting front. Unless the wetting front reaches the deepest soil regions in the catchment during the transitional period, large-scale connectivity via lateral flow does not occur because dry pockets inhibit downslope flow along the soil–bedrock interface. Streamflow during the winter originates high up in southeast-facing unchannelled gullies from direct input of new snowmelt. This dilute water picks up some solutes as it travels down the streambed, but it remains relatively dilute. The stream loses water to the underlying soil and a saturated wedge likely develops below the stream on the underlying granite. Meanwhile, localized connectivity grows via lateral flow along the bedrock interface in the shallow soils on the hillslopes near the ridgetops and streambed. Near-stream subsurface flow is independent of activities occurring higher in the slopes. When the deep midslope soils become wetted above field capacity, whole-slope hydraulic connectivity is established and upland soils are in direct connection with the near-stream soil moisture. Previously wetted upslope regions become hydraulically connected to downslope regions, raising the head in downslope regions due to the sudden connection to the higher heads in the upland soils. This can reverse the gradient in the substream saturated wedge and turn the losing stream into a gaining stream. This allows solutes from previously disconnected sources to enter the stream, causing a step increase in electrical conductivity of the streamwater. At this point, the source of solutes has not been identified. Solutes could be derived from hillslope regions or from the near-stream saturated wedge that has accumulated solutes through the winter. Regardless, as the catchment drains, the solute source continues to deliver concentrated water to the stream because basal connectivity is sustained.

SUMMARY AND CONCLUSIONS

We have identified five soil moisture conditions that occur during a year in a small snowmelt-driven, semi-arid catchment with shallow soils, and presented a conceptual model of how streamflow is generated with respect to the moisture conditions. The five moisture conditions are (1) a summer dry period, (2) a transitional fall wetting period, (3) a winter wet, low-flux period, (4) a spring wet, high-flux period, and (5) a transitional late-spring drying period. Transitions between the periods are driven by changes in the balance between rain, snow, snowmelt and evapotranspiration. If period 3 begins before the soil profile has wetted to the soil–bedrock interface by matrix flow, then dry soil pockets can persist through the winter. This causes a hydraulic disconnect at the bedrock interface. However, preferential flow will cause some locations to wet-up faster than matrix flow would allow. Only after this disconnect is breached can significant lateral translation of water, pressure, and solutes occur. We propose that, when dry-soil barriers are breached, the sudden connection to higher heads in the upland soils causes changes in the near-stream hydraulic conditions that allow delivery of water and solutes from previously disconnected sources such as a substream saturated wedge or hillslope regions. Future work will further investigate these ideas with respect to the hydraulic and chemical conditions in the near-stream saturated zone.

The results of this study have important implications for hydrologic modelling based on topographically controlled wetness indices. Grayson *et al*. (1997) and Western *et al*. (1999) pointed out that hydrologic models for regions in which annual soil moisture patterns settle into two preferred states must use a dual index approach to account for the different controls on moisture distribution during the different states. An additional caveat from this study is that, in addition to allowing for a climatically driven transition from a dry to a wet state in the near-surface soils, the rate of water delivery by snowmelt and the soil moisture condition at the soil–bedrock interface must also be considered.

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