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SUMMARY

Long-term water balance investigations are needed to better understand hydrologic systems, especially semi-arid mountainous catchments. These systems exhibit considerable interannual variability in precipitation as well as spatial variation in snow accumulation, soils, and vegetation. This study extended a previous 10-year water balance based on measurements and model simulations to 24 years for the Upper Sheep Creek (USC) catchment, a 26 ha, snow-fed, semi-arid rangeland headwater drainage within the Reynolds Creek Experimental Watershed in southwestern Idaho, USA. Additional analyses afforded by the additional years of data demonstrated that the variability between streamflow and annual precipitation ($r^2 = 0.54$) could be explained by the timing of precipitation and antecedent moisture conditions. Winter–spring precipitation and soil moisture deficit at the beginning of the water year accounted for 83% of the variability in streamflow, which was almost as accurate as applying the more complex physically-based Simultaneous Heat and Water (SHAW) numerical model ($r^2 = 0.85$) over the three dominant land cover classes. A conceptual model was formulated based on field observations, numerical simulations and previous studies. Winter precipitation and spring snowmelt must first replenish the deficit within the soil water profile and ground water system before water is delivered to the stream. During this period, surface water and ground water are tightly coupled and their interaction is critical to streamflow generation. Shortly after snow ablation, however, water flux in the root zone becomes decoupled from the ground water system and subsequent precipitation does little to contribute to streamflow for the current year, but serves to offset ET and the soil moisture deficit at the beginning of the following year. This study demonstrates the merits of long-term catchment-scale research to improve our understanding of how climate and land cover interact to control hydrologic dynamics in complex mountainous terrain.

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1. Introduction

The majority of the land that constitutes the intermountain western United States is semi-arid rangelands that are predominately composed of shrubs, forbs, and grasses (Branson et al., 1981; Krueger, 1988) with forested patches in the higher elevations and along riparian corridors. In these regions, a large portion (40–80%) of annual precipitation is delivered between October and March as snow in upland areas. The interaction of the complex topography and heterogeneous vegetation patterns with snowcover processes in these mountainous environments results in meltwater being highly variable in space and time. A more detailed characterization of semi-arid rangeland hydrology is needed to understand how these hydrologic systems will respond to both climatic variability and management practices. This requires a thorough knowledge of the interactions of patterns and processes controlling hydrologic fluxes. Understanding how the interannual variability in climate and precipitation dynamics affects hydrologic fluxes is limited. To improve our understanding of semi-arid rangeland hydroclimatology, long-term water balance investigations are needed.

A catchment water balance quantifies the components of the hydrologic cycle at the watershed scale. This provides the most fundamental information about the hydrology of a watershed and is necessary to assess the importance of climate and land cover in determining water availability (Dingman, 2002). In addition to providing a baseline understanding of the hydrologic processes

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occurring within a catchment, the water balance components can be compared over long periods of time to track the hydrologic response of a catchment to climatic and land cover variability. The intrinsic spatiotemporal land surface and hydroclimatic variability of these areas necessitates long-term and spatially distributed data to produce an accurate water balance (Branson et al., 1981). These computations can be challenging given the considerable spatio-temporal variation of climate and land cover, but are essential to better characterize the hydroclimatology of the semi-arid rangelands of the intermountain west.

Because of the difficulty in collecting long-term spatially distributed records, relatively few long-term water balance studies have been reported, and most have focused on forested systems. For example, Lewis et al. (2000) reported a 17-year study of a 103 ha oak woodland and demonstrated a threshold in oak harvesting to significantly influence water yield. Peters et al. (2006) reported water and solute balances for five watersheds (40–3300 ha) over six water years to ascertain regional controls over solute yields. Shimuzu et al. (2003) conducted an 8-year study of a 3.7-ha forested watershed to quantify evapotranspiration losses. Ishii et al. (2004) analyzed 11 years of water balance data from a 1.15 km^2 snow-dominated watershed and correlated premelt snow water equivalent with November and March air temperature deviations. Zhuravin (2004) attributed discrepancies in the water balance of three watersheds over a 15-year period in a mountainous taiga zone to changes in ice volumes in the permafrost zone. In snow-dominated semi-arid systems where a large proportion of annual precipitation may be required to replenish the vadose zone reservoir before streamflow occurs, and where the intra-annual variations in rain to snow ratios control the spatial distribution of moisture inputs, long-term investigations are needed to capture the full range of variability.

Flerchinger and Cooley (2000) reported a 10-year water balance of the 26-ha Upper Sheep Creek (USC) catchment in southwestern Idaho covering water years 1985 through 1994, however several factors limited the conclusions that could be drawn from the study. Five of the years examined were very dry and produced no streamflow and 1 year produced an extremely small amount (3 mm) of flow out of the catchment. Further analyses of water balance components over a longer period of time is therefore needed to understand the relationships between climate and streamflow in the watershed and to propose a conceptual model of the catchment. This extension also allows for a more complete analysis of the impacts of antecedent conditions on these components and the response of the system both to relatively short-term interannual and intra-annual hydroclimatic variability.

The general objective of this study was to advance our understanding of mountainous semi-arid rangeland hydroclimatology by constructing a 24-year water balance of the USC catchment. The specific objectives were to: (1) describe the hydroclimatic conditions for the 24-year period, validate the model simulations, and compare the results to the previous 10-year study period; (2) assess critical precipitation thresholds needed to generate streamflow and analyze the correlation between streamflow and hydrological variables; (3) investigate the influence of hydroclimatic variability and antecedent conditions on the water balance; and (4) present a conceptual model of streamflow generation processes within the catchment.

2. Previous work at Upper Sheep Creek

The 10-year water balance of the USC catchment by Flerchinger and Cooley (2000) contributed to the evolving hydrologic understanding of the catchment that has been studied since the early 1970s, thus demonstrating the advantages of long-term catchment studies. Pioneering work on ground water modeling in the catchment was performed by Schreiber et al. (1972) and Stephenson and Freeze (1974). Schreiber et al. (1972) applied a steady-state two-dimensional ground water model to a hillslope transect within USC and concluded that equilibrium heads were not achieved and that a steady-state model did not adequately describe the system. Stephenson and Freeze (1974) subsequently applied a variably-saturated two-dimensional model; they identified that subsurface flow of snowmelt was the primary source of streamflow, but with limited geologic information could not simulate the rapid response of ground water through the hill slope and could not establish a conceptual model of sub-surface flow. Springer et al. (1984) applied the Simulation of Production and Utilization of Rangelands (SPUR) model to simulate streamflow from USC, but found it was necessary to use much larger values for the sub-surface return flow time (34–40 days) in USC compared to other catchments (5–15 days) in the Reynolds Creek Experimental Watershed.

An instrument network and snow surveys were established in USC in 1984 to further advance hydrologic understanding of the catchment. Cooley (1988) demonstrated a consistent snow drift pattern within USC and the influence of the variability in snow accumulation and melt on streamflow. With the information provided by geophysical surveys (Winkelmaier, 1987; Stevens, 1991), Flerchinger et al. (1992, 1993) proposed and tested a conceptual model describing the rapid piezometer and streamflow response to snowmelt events observed by Stephenson and Freeze (1974). During years with adequate snowmelt, the shallow (<25 m) ground water flow system becomes confined beneath a semi-dense basalt layer between the drift area and the stream, and the rapid response is primarily due to a pressure pulse through the ground water. Using isotope tracers, Unnikrishna et al. (1995) confirmed that snowmelt flowed primarily through the altered basalt zone beneath the confining semi-dense basalt layer and highlighted the limitations of topographically-driven models within the USC catchment. They suggested a decoupling of the saturated flow above and below the semi-dense basalt layer. This is somewhat in contrast with streamflow generation studies in forested semi-arid systems that identified lateral sub-surface flow in soil macropores as a dominant component of streamflow (Wilcox et al., 1997; Newman et al., 1998; Liu et al., 2008) with minor components of infiltration excess overland flow (Wilcox et al., 1997) and thermal meteoric waters (Liu et al., 2008) in some cases.

Using simulated snowmelt provided by Flerchinger et al. (1994), Deng et al. (1994) simulated the response of the shallow ground water system to spatially variable snowmelt and demonstrated the limitation of using a uniform snowpack across the basin. Flerchinger et al. (1998) proposed dividing the catchment into three zones based on vegetation and snow accumulation across the watershed. Luce et al. (1998) demonstrated that detailed snow drifting information within the catchment is equally or perhaps more important than modeling the effects of local topography on radiation. Luce et al. (1999) subsequently developed an approach to capture sub-grid snowpack variability that when combined with a lumped snowmelt model performed as well as a distributed model operating on a 30-m grid. However, Artan et al. (2000) showed that a 15-m grid was necessary to capture the variability in radiometric surface temperature, and sensible and latent heat during the spring and summer.

Flerchinger and Cooley (2000) computed a 10-year water balance of USC by computing a partial water balance for each of three hydrologically homogeneous landscape units that were subsequently aggregated for the entire watershed. This study showed that the spatial variability of precipitation between landscape areas increased during years with higher precipitation and indicated that most of the contribution to streamflow and ground water came from a large snowdrift area that forms each winter and sustains the aspen stands in the catchment. Precipitation adjusted for redistributed and drifted snow was approximately 15% higher than measured precipitation. Evapotranspiration was found to account for nearly 90% of the drift-adjusted precipitation in the catchment. It was also found that catchment streamflow was generally correlated with precipitation above a critical threshold of roughly 450 mm necessary to generate streamflow (r^2 = 0.52) and that a very good correlation (r^2 = 0.90) existed between streamflow and model simulated percolation of water beyond the root zone above a critical threshold of approximately 50 mm. The average water balance error in this study was approximately 10% of the drift-adjusted precipitation for the 10-year period that was largely attributed to deep percolation losses through bedrock fractures, confirming suspicions of Flerchinger et al. (1998) that the dense basalt underlying the basin is more transmissive than initially assumed.

3. Methods

The approach for establishing a long-term water balance was to collect field data from the study site, prepare and gap fill hourly hydrometeorological data to form a continuous 24-year data set, and quantify components of the annual water balance using a physically-based energy and water transfer model.

3.1. Study site

The USC study area is a semi-arid rangeland catchment located within the Reynolds Creek Experimental Watershed (RCEW) in the Owyhee Mountains of southwestern Idaho, USA (Marks, 2001; Slaughter et al., 2001). It is a 0.26 km^2 headwater catchment with an elevation range of 1840–2036 m asl (Fig. 1). Average annual precipitation during the 24-year study period was 426 mm, with approximately 60% occurring as snow. Streamflow is intermittent with an average annual yield of approximately 44 mm, equating to an annual runoff ratio of 10.3%.

The site is characterized by high spatial variability in vegetation, snow distribution, and soils. Prevailing winter storms are from the southwest and cause deep snow drifts to form below the crest of the leeward northeast-facing slopes. The southwest-facing slopes are sparsely vegetated with low sagebrush (*Artemisia arbuscula*) and some grasses. These exposed areas have little or no snow cover in the winter. Soils here are generally high in rock content (>50%), shallow (~30 cm), and contain relatively high clay content (~25%) argillic horizons and thin (<10 cm) silt loam surface horizons. Mountain big sagebrush (*Artemisia tridentata vaseyana*), snowberry (*Symphoricarpos spp.*), and grasses cover the lower portions of the northeast-facing slopes. These areas typically accumulate about a meter of snow over the winter and soils here are deep (>100 cm) loess silt loam having low rock content. The upper portions of the northeast-facing slopes are predominantly vegetated by aspen (*Populus tremuloides*) and willow (*Salix spp.*) thickets. Large snow drifts (varying in depth from 1 m to greater than 8 m) form annually in these areas. Soils here are virtually rock free and are very deep (>200 cm) loess silt loam. Three landscape units in the USC catchment have been identified based on similarity in vegetation, snow accumulation, and soils (Flerchinger and Cooley, 2000; Flerchinger et al., 1998). These are referred to as low sagebrush, mountain big sagebrush, and aspen (Fig. 2) and comprise 58.9%, 26.6%, and 14.5% of the catchment, respectively.

Vegetation data indicate that leaf area index of the low sagebrush, mountain big sagebrush, and understory of the aspen at the peak of the growing season is approximately 0.4, 1.2, and 1.0, respectively, based on point frame measurements. Transect measurements and destructive sampling of a single representative aspen yielded a tree leaf index of 2.04 for the aspen. Because of natural "pruning" of the aspen thickets due to shifting of the massive snow drifts during the winter, the aspen leaf area effectively remained the same over the 24-year study period. Leaf area index of the aspen computed by difference between measurements taken using a light interception instrument (LAI-2000, Li-Cor, Inc., Lincoln, Nebraska) at peak growth during 2005 and after leaf-fall indicated a leaf area index (LAI) of 2.0.

The geology of the USC catchment consists of variably fractured and altered basalt underlain by thick dense basalt at a depth of 20–30 m mid-slope and shallower near the stream and ridge. Geophysical studies of the area indicate that the surface of the dense basalt closely follows surface topography (Mock, 1988; Stevens, 1991; Winkelmaier, 1987), hence the catchment boundary for the ground water flow is approximately the same as for the surface. Flerchinger et al. (1992, 1993) demonstrated that the shallow ground water system flows laterally between the drift area and the

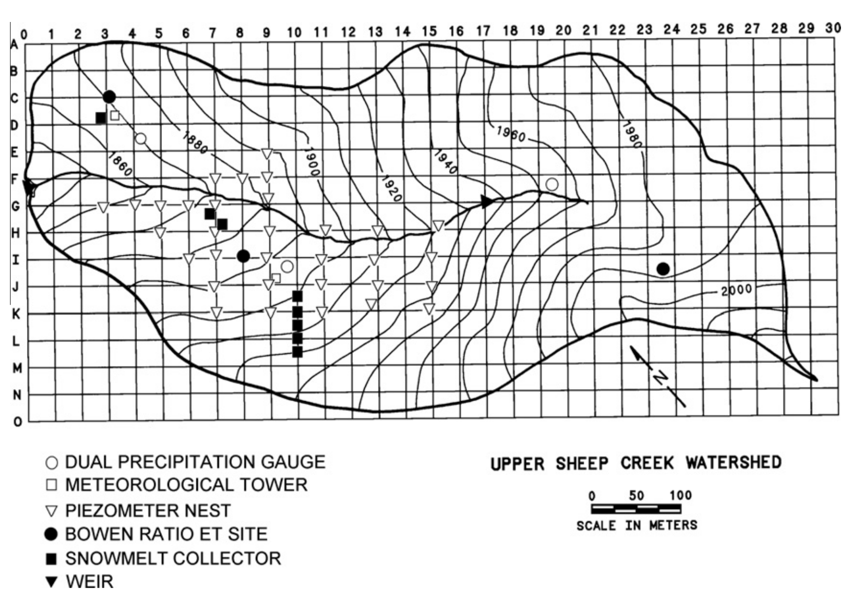

Fig. 1. Upper Sheep Creek catchment orientation, elevation range, and instrumentation.

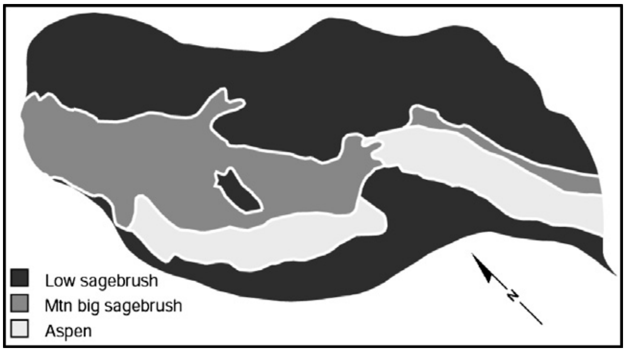

Fig. 2. The three landscape units of the Upper Sheep Creek catchment identified based on similarity in vegetation, snow accumulation, and soils. The low sagebrush, mountain big sagebrush, and aspen units comprise 58.9%, 26.6%, and 14.5% of the catchment, respectively.

stream within an altered basalt layer between a semi-dense basalt layer 7 m deep and the dense basalt layer 25 m deep. Spring snowmelt percolates to ground water, flows over the dense basalt and is delivered to the stream via springs exiting along the stream channel.

3.2. Field data

The instrumentation network present at the USC catchment was installed in 1984 and consists of meteorological stations, piezometers, precipitation gauges, snowmelt collectors, and weirs (Fig. 1). This network was described in detail by previous studies (Flerchinger et al., 1998; Flerchinger and Cooley, 2000). Much of the instrumentation was removed in 1996, but reinstalled after 2002.

Hourly meteorological and precipitation data collected at two sites within the USC catchment were used for this study. Air temperature, precipitation, and wind speed measurements were recorded throughout the study period at the southwest-facing site near grid coordinate D4 in Fig. 1; relative humidity and solar radiation were added in 2002. Air temperature, precipitation, relative humidity, solar radiation, and wind speed measurements were collected on the northeast-facing site near J9 from October 1983 through September 1996 and from July 2003 onwards. Dual-gauge systems especially designed for the windy and snow-dominated conditions prevalent in the area were used to measure precipitation (Hanson et al., 2004). Streamflow was monitored continuously using a permanent v-notch weir installed at the outlet of the catchment.

A Bowen ratio system was used to estimate evapotranspiration at each of the landscape units within the study area during focused field campaigns. This instrument was rotated at weekly intervals between the three vegetation types when vegetation was actively transpiring in 1990. Three separate Bowen ratio units were operated continuously at C4, I8, and I24 during the 1993 growing season. Soil water content measurements were recorded bi-monthly from 1990 to 1993 at 11 profiles ranging in depth from 100 to 260 cm using a neutron probe after the snow had melted. In 2005, measurements at six neutron probe profiles were resumed in each the aspen and mountain big sagebrush areas, and time domain reflectometry measurements (0–30 cm) were initiated at ten locations in the low sagebrush area and read bi-monthly during the growing season. A total of 53 piezometers were used to monitor ground water elevation from October 1985 through September 1994. These were placed at 32 locations within the catchment at depths ranging from 4 to 22 m. Six of these piezometers (located at G3, G5, H7, I7 and K7) were re-instrumented in the fall of 2004. Water levels in the piezometers were recorded hourly using pressure transducers. Snow depth and water equivalent data were collected at each grid point on the catchment using standard Natural Resources Conservation Service (NRCS) snow sampling techniques and a Rosen-type snow tube (Goodison et al., 1981). Snow measurements were taken approximately every two weeks after peak accumulation until the end of the snow melt period (March through May). Snow measurements were collected for water years 1984 through 1994 and 2003 through 2007.

3.3. Hydrometeorological data preparation

The approach for computing a 24-year water balance for the USC catchment required the compilation of a complete hourly data set of hydrometeorological variables at two sites (near grid coordinates D4 and J9) within the catchment for the entire time period. Gaps or erroneous data that existed for 3 h or less were filled using a spline fit method. Data were not available for the site near J9 from 1996 to 2003 and were filled in using simple linear regressions for each month with the site near D4; correlation coefficients exceeded 0.83 for precipitation, 0.92 for temperature, 0.81 for dew point, 0.92 for solar radiation, and 0.55 for wind speed. During a few brief periods when D4 was not operational, data were filled in using three nearby (within 1–7 km distance and 90–250 m elevation) hydrometeorological stations located within the RCEW.

3.4. Model implementation and validation data

After validation of the model, simulations using the Simultaneous Heat and Water (SHAW) Model (Flerchinger et al., 1996; Flerchinger and Pierson, 1991; Flerchinger and Saxton, 1989) were used to supplement field measurements and fill gaps in the water balance components. Model simulations provided continuous evapotranspiration estimates as well as changes in soil water storage when measurements were not available. The SHAW model has been tested and applied extensively over a range of vegetation types in semi-arid and arid environments, particularly in the USC catchment and the surrounding RCEW. The model simulates the surface energy balance, evaporation, transpiration, and soil water status and fluxes within a system composed of a multi-species plant canopy, making it ideal for use in this study.

The SHAW model simulates a vertical, one-dimensional system composed of a vegetation canopy, snow cover (if present), plant residue, and soil profile. Weather conditions above the upper boundary and soil conditions at the lower boundary define heat and water fluxes into the system. A layered system is established though the model domain, with each layer represented by a node. After computing a full solution of the surface energy balance at the upper boundary, the interrelated heat, liquid water, and vapor fluxes between layers are simulated down through the profile using implicit finite-difference equations.

Soil water and ET simulations from the SHAW model were compared with field measurements for validation. Measured changes in soil water content profiles were compared to SHAW model simulations for water years 1991–1993 and 2006–2007. Neutron tubes located at D4 and I8 were representative of the low sagebrush and mountain big sagebrush units, respectively, and tubes at I23 and I24 were representative of the aspen unit for 1991–1993. Measurements for the 2006 and 2007 water years were taken at six access tubes distributed in each of the aspen and mountain big sagebrush units, and ten locations with TDR probes in the low sagebrush unit. Additionally, ET values from the 24-year simulations were compared with measurements from Bowen ratio units near grid points C4, I8, and I24 within the USC catchment during 1990 and 1993.

Continuous model simulations for each landscape unit were conducted from October 1982 through September 2007 for each landscape unit of the USC catchment. The 1983 water year was simulated to allow a "spin-up" year to distance the simulations from assumed initial conditions. Hourly air temperature, wind speed, relative humidity, solar radiation and drift-adjusted precipitation were used to drive the model. Average soil, slope and vegetation conditions were assumed for each landscape unit. Simulations were carried out to a depth of four meters where soil temperature was assumed constant and gravity flux was assumed for water flow at the lower boundary.

3.5. Water balance computations

The USC catchment water balance was simulated for water years (October 1 through September 30) 1983 through 2007. A water balance may be expressed as:

$$P - \mathrm{ET} - R - \Delta S - \mathrm{DP} = 0, \quad (1)$$

where P is precipitation, ET is evapotranspiration, R is streamflow runoff, ΔS is the change in water storage, and DP is ground water that exits the basin by deep percolation. Flerchinger and Cooley (2000) initially assumed that the dense basalt layer underlying the catchment restricts deep percolation, i.e. DP = 0, but attributed the residual error in the water balance to DP. Unfortunately data are not available to assess deep percolation loss.

Annual drift factors specific to each landscape unit were applied to measured precipitation data to account for wind-driven snow redistribution. A drift factor was applied during the snow-covered period to the precipitation measured near D4 for the low sagebrush landscape unit and near J9 for the mountain big sagebrush and aspen landscape units. Drift factors were calculated by comparing measured precipitation (minus sublimation simulated by the SHAW model) with snowmelt collector and snow depth measurements (Flerchinger and Cooley, 2000; Flerchinger et al., 1998). Drift factors used for water years 1984 through 1994 are given by Flerchinger and Cooley (2000) and range from 0.68 to 1.22 for the sagebrush areas and 1.80–3.20 for the aspen area. Drift factors for the aspen during 2004 through 2007 were computed identically to the 10-year study using snow surveys, yielding values of 2.56, 1.99, 1.64 and 2.08, respectively. Calculation of year-specific drift factors was not feasible for water years 1995 through 2003 for the aspen and 1995 through 2007 for the sagebrush sites because of data constraints. Therefore, the average of the drift factors from water years 1985 through 1994 was applied for water years 1995 through 2004 for each landscape unit. The average drift factors used were 0.98, 0.93, and 2.29 for the low sagebrush, mountain big sagebrush, and aspen landscape units, respectively. While alternative approaches exist to distribute snow across a watershed (e.g. Pomeroy et al., 1993; Winstral et al., 2002), the approach of Winstral et al. (2002) is hampered when there is only one precipitation gauge within the basin, as was the case for 1995 through 2007, and application of the Prairie Blowing Snow Model (Pomeroy et al., 1993) was beyond the scope of the current study. A single station algorithm for precipitation redistribution is under development (Winstral et al., 2009), but was not available for the current study.

Evapotranspiration (ET) was simulated by the SHAW model. Change in catchment storage (ΔS) was partitioned into change in soil water storage and change in ground water storage. Estimates of change in soil water storage were based on SHAW model simulations for all years except for those years when measured data were available (water years 1991–1993 and 2006–2007). Change in ground water storage was estimated using piezometer data for water years 1985 through 1994, 2005 through 2007 when data were available. Change in ground water storage along each numbered transect in Fig. 1 was estimated using the average measured change in ground water depth along the transect, the estimated extent of the saturated zone based on topography of the dense basalt layer (Stevens, 1991), and an estimated porosity of 0.10 for the altered basalt layer where much of the ground water flow occurs (Flerchinger et al., 1993).

An aggregated approach was used to calculate the 24-year water balance of the USC catchment, as was done in the 10-year water balance investigation (Flerchinger and Cooley, 2000). This approach involved computing annual partial water balances for the low sagebrush, mountain big sagebrush, and aspen landscape units (described above) and then computing annual aggregated water balances for the entire catchment using area-weighted averages of components from each landscape unit. Compared to a uniform approach that treats the entire catchment as homogeneous, the aggregated approach more adequately addresses the precipitation and ET differences exhibited over the watershed and therefore better represents the heterogeneity of the USC catchment (Flerchinger et al., 1998).

4. Results and discussion

4.1. Hydroclimatic conditions of 24-year study period

The climatic variability characteristic of USC is exemplified by the hydrometeorological data of the 24-year study period. Average air temperature for the study period was 6.7 °C with annual averages fluctuating between 5.0 and 8.0 °C. Average annual daily maximum and minimum air temperatures were between 8.9–12.1 and 1.5–4.2 °C, respectively. Average annual precipitation varied from 425 mm measured in the low sagebrush zone to 576 mm measured in the mountain big sagebrush. The range of measured annual precipitation was 270–780 mm and estimates of total annual snow precipitation based on dew-point temperature ranged from 102 to 523 mm. The wettest month in the study period was December of water year 1984, which had 226 mm of precipitation. November and December were typically the wettest months, each receiving an average of 63 mm of precipitation, and August was the driest month receiving on average just 11 mm of precipitation. Fig. 3 presents the total annual precipitation for the catchment adjusted for snow drifting, and emphasizes the influence of the first 6 months of the water year on the hydrology of the catchment. Approximately 70% of the drift-adjusted precipitation for the catchment was received between October and March.

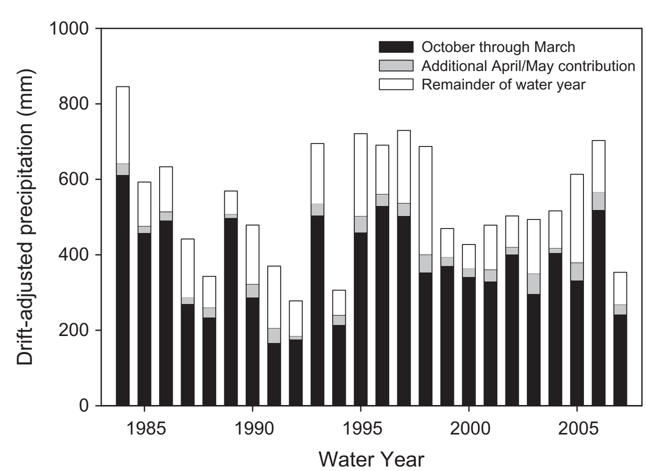

Fig. 3. Total annual precipitation adjusted for drifting of snow separated into: the first 6 months of the water year (October through March); and additional April precipitation for the mountain big sagebrush zone and April/May precipitation for the aspen zone to show the total winter–spring precipitation contributing to streamflow.

4.2. Model validation

ET values from the 24-year SHAW simulations were within 1–30% of measured data (Table 1). Overall, these values were consistent with modeled values from the previous 10-year simulations, indicating that minor corrections made to the model and forcing data during the interim had a minimal effect. The 1990 simulated ET data for the low sagebrush and aspen landscape units more closely matched the measured data than those for the mountain big sagebrush landscape unit. When ET values for the different landscape units were aggregated into a representative value for the entire catchment, the values from the 24-year simulations were within 2% and 3% of the measured values from 1990 and 1993, respectively. Values of change in annual soil storage from the 24-year simulations were within 2–17 mm of measured data (Table 2). Simulated changes in storage were similar to, or better than those from the 10-year study. Based on these results, the model can be used to estimate evapotranspiration and change in soil water storage with reasonable confidence.

4.3. Ten-year and 24-year water balance comparison

A comparison of the water balance components between the 10- and 24-year studies is provided to assess the effect of more than doubling the amount of years used in the computations. For each of the 24 years examined, the annual partial water balance is plotted by landscape unit in Fig. 4 and the annual aggregated water balance for the entire catchment is given in Table 3. The average annual precipitation adjusted for snow drifting in the USC catchment over the 24-year study period was 539 mm, which is 14% higher than the average annual drift-adjusted precipitation of the 10-year study (471 mm). The average ET over the catchment increased by 6%, however the average annual measured streamflow from the catchment increased by 47%, or 14 mm. This demonstrates that for semi-arid systems, a relatively small amount of water can have a large relative impact on the average streamflow.

Water balance errors for water years 2005 through 2007 when piezometer data were available are 6%, 10% and 7% of drift-adjusted precipitation, which are comparable to the 10% average error for the 10-year study. We attribute these errors to ground water exiting the basin by deep percolation, as did Flerchinger and Cooley (2000).

Additional years of data in this 24-year water balance study also yielded some revealing insights on the ET dynamics. While cumulative ET in the aspen area is quite consistent from year to year, ET in the sagebrush areas is dependent on the variable precipitation and plant available water (Fig. 4). This is because the deep snow drifts and meltwater in the aspen unit replenishes the deep soil profile so that the growing season begins with the soil at field capacity, which is sufficient to meet ET demand in most years. Ishii et al. (2004) also reported small variations in ET for a forested, snowy, mountainous watershed, and Lewis et al. (2000) noted only marginal increase in ET with increased rainfall in an oak woodland. These observations are consistent with other work suggesting that forest transpiration is a relatively conservative hydrological process that exhibits low interannual variability (Roberts, 1983). Snow accumulation and subsequent spring snowmelt in the mountain big sagebrush zone, however, is not always sufficient to return the soil water profile to field capacity and is mostly absorbed by the deep loess soil. The short-statured low sagebrush landscape unit is not capable of trapping very much snow and the shallow soil profile allows for only a small storage capacity; hence the low sagebrush is more dependent on the frequency of precipitation events rather than winter snows. These results are consistent with observations by Lewis et al. (2000) who noted that ET is not correlated with precipitation above 600 mm, but is increasingly variable and correlated with precipitation below this threshold.

Table 1
Comparison of measured ET with continuous 24-year simulations.

Year	Days measured	Measured ET (mm)	Simulated ET (mm)
Low sagebrush			
1990	24	41	48
1993	73	145	141
Mountain big sagebrush			
1990	27	74	52
1993	86	279	286
Aspen			
1990	23	85	92
1993	48	196	217

More importantly, the extension of the 10-year water balance of the USC catchment to 24 years permitted greater assessment of the temporal hydroclimatic variability of this system including a series of wet and dry periods that followed the previous study period. The current study captured 11 additional years with streamflow, more than tripling the number of years with streamflow and allowing additional analysis that would not have been possible with only 5 years of observed streamflow. Thus, in environments such as this with high interannual variability and inconsistent streamflow, longer records are needed to capture the full range of hydroclimatic variability.

4.4. Threshholds, hydroclimatic variability, and antecedent conditions

The relationship between drift-adjusted precipitation and measured streamflow for the catchment is plotted in Fig. 5. In the 10-year water balance study, this relationship indicated that a threshold in precipitation of approximately 450 mm must be reached to generate streamflow and that above this threshold the relationship between precipitation and streamflow was fairly linear (Flerchinger and Cooley, 2000). Observations from the current study tend to support this observation. Although three water years (1999, 2000, and 2002) produced considerable streamflow with less than or slightly exceeding 450 mm, this may be attributed to uncertainty in using an average drift factor. With the additional years of data, however, Fig. 5 suggests an upper bound of streamflow for a given amount of precipitation and displays considerable amount of variability ($r^2 = 0.54$) below this upper bound. Several years in particular (1993, 1995, 1998, and 2005) received relatively high amounts of precipitation, yet yielded relatively small amounts of streamflow, even with the uncertainty in the drift factor.

The aggregated, simulated percolation beyond the root zone for the catchment is plotted against measured streamflow in Fig. 6. Percolation beyond the root-zone simulated by the model represents water available for lateral ground water flow to the stream and possible exchange with deeper groundwater, even though the 1-dimensional model does not simulate streamflow generation in 3-dimensions. Threshold behavior in the current study is consistent with observations from the 10-year study; only water year 2000 had less than 50 mm of simulated percolation beyond the root zone and still produced streamflow. Beyond the apparent 50 mm threshold, the relationship remains somewhat linear. Although there is some scatter (namely 1995, 1996, 1999, and 2002), none of these years can be considered outliers given the uncertainty in the drift factor. The correlation is not as high as suggested by the 10-year study (r^2 of 0.85 versus 0.90) due in part to the uncertainty in drift factors.

Variations in the timing and phase of precipitation and antecedent moisture conditions can explain a considerable amount of the variability displayed in Fig. 5. Recognizing that winter precipitation

Table 2
Comparison of measured change in annual soil water storage for the USC catchment with estimates from continuous 24-year simulations. (Values are aggregated from the three landscape units).

Water year	Measured change in soil water storage (mm)	Simulated change in soil water storage (mm)
1991	11	−7
1992	−18	−20
1993	100	110
2006	−5	−26
2007	−13	−29

and snowmelt are the primary sources of streamflow, the r^2 of 0.54 for the relationship in Fig. 5 is increased to 0.70 by including drift-adjusted precipitation from only October though March (Fig. 3). This relation can be further refined by recognizing that snowmelt and the contribution of the different vegetation zones to streamflow extends to varying lengths within the snowmelt season. When considering only precipitation through March for the low sagebrush zone, through April for the mountain big sagebrush zone, and through May for the aspen, as depicted in Fig. 3, the r^2 is increased to 0.74 (although not significantly better than 0.70). The implication is that precipitation after snowmelt and during the growing season contributes very little to streamflow for the current water year. Indeed, simulated percolation beyond the root zone typically ceased in mid-June. Thus, with much of the saturated sub-surface flow to the stream occurring over 6 m deep (Flerchinger et al., 1992), water flow in the root zone becomes largely decoupled from the ground water system shortly after snow ablation when downward flux of water to the ground water system becomes insignificant.

The influence of antecedent conditions is important in reevaluating the relationship between drift-adjusted precipitation and streamflow. As noted previously, water years 1993 and 1995 both had relatively low streamflow for the relatively large drift-adjusted precipitation. Both of these years were near the end of a drought period and the relationship between precipitation and streamflow for these years was likely influenced by the dry antecedent moisture conditions. Although considerable precipitation occurred in these years, moisture deficits created during the preceding drought periods had to be replenished, hence streamflow was reduced. Indeed, the largest increases in soil water storage in Table 3

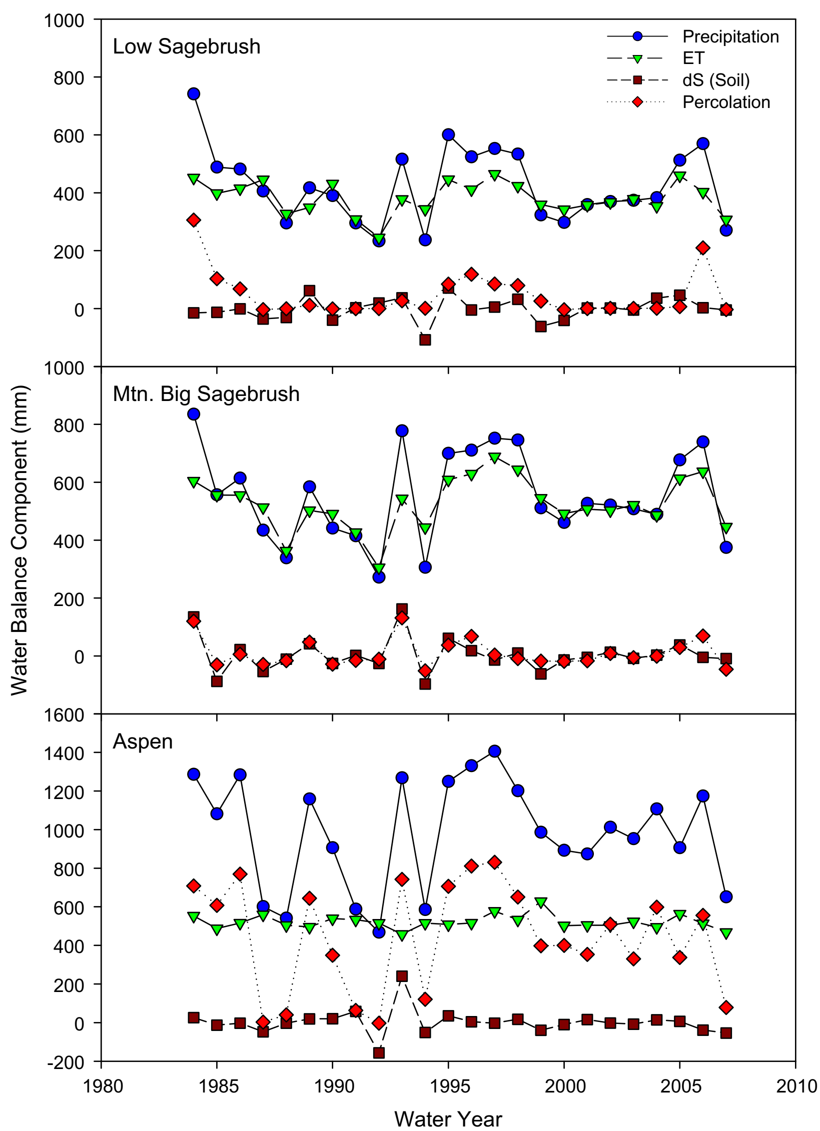

Fig. 4. Annual drift-adjusted precipitation, simulated evapotranspiration, change in soil water storage, and simulated percolation beyond the root zone for each landscape unit. (Change in soil water storage was simulated when measurements were not available.).

Table 3
Annual water balance summary (in mm) from aggregating the three landscape units.

Water year	Precip.[a]	Winter–spring precip.[b]	ET[c]	Streamflow	ΔS (soil)	ΔS (ground water)	Error[d]
1984	846	643	507	239	31[c]	No data	na
1985	593	476	453	82	−33[c]	−32	123
1986	633	514	467	95	5[c]	14	54
1987	442	289	480	0	−43[c]	−30	35
1988	343	261	362	0	−22[c]	−22	25
1989	569	509	411	54	50[c]	8	43
1990	479	323	463	3	−28[c]	−2	43
1991	370	206	372	0	11	−28	33
1992	278	185	300	0	−18	−29	28
1993	695	537	433	61	100	75	12
1994	306	240	395	0	−97[c]	−19	28
1995	721[e]	503	498	29	63[c]	24	103
1996	691[e]	561	484	82	2[c]	9	102
1997	729[e]	537	540	98	−1[c]	No data	na
1998	687[e]	401	498	53	24[c]	No data	na
1999	470[e]	395	448	42	−59[c]	No data	na
2000	427[e]	364	405	12	−29[c]	No data	na
2001	478[e]	361	419	0	2[c]	No data	na
2002	503[e]	421	423	49	4[c]	No data	na
2003	494[e]	351	437	0	−6[c]	No data	na
2004	516	418	409	11	24[c]	No data	na
2005	613	379	515	15	38[c]	8	37
2006	703	567	481	135	−5	22	70
2007	354	269	367	0	−13	−26	26
Average	539	405	440	44	0	na	na

[a] Precipitation adjusted for drifting of snow.
[b] October through March/April/May (depending on landscape unit) precipitation adjusted for drifting of snow.
[c] Indicates model-simulated values.
[d] Water balance error may be attributed to ground water exiting the basin by deep percolation.
[e] Values based on average drift factors.

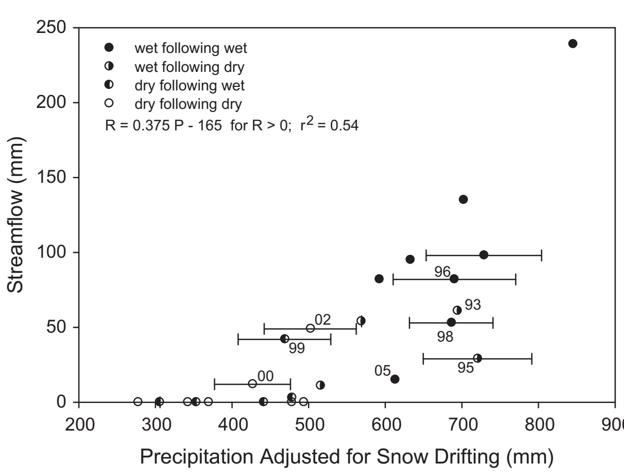

Fig. 5. Measured streamflow versus average areal precipitation adjusted for snow drifting on the Upper Sheep Creek Watershed. "Wet" or "dry" years are defined as years with above or below the mean annual precipitation of the 24-year period. Error bars denote uncertainty associated with using average drift factors (one standard deviation). Labeled points refer to years that are discussed in the text.

occurred during these 2 years; the wettest year on record, 1984, had only a modest increase in soil water storage because it followed a year with above average precipitation. Because streamflow within the watershed is generated by springs discharging to the stream, winter precipitation and spring snowmelt must first replenish the deficit within the soil profile before water percolates to the ground water system and is transported to the stream. Therefore water available for percolation to ground water and streamflow is the drift-adjusted winter–spring precipitation (Fig. 3) minus the soil moisture deficit. This can be expressed as:

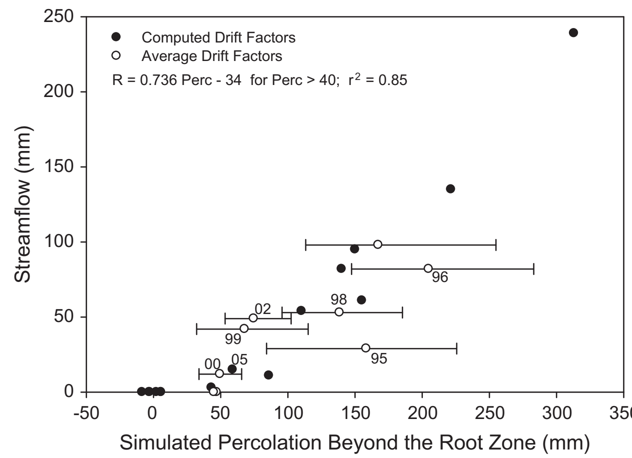

Fig. 6. Measured streamflow versus simulated percolation beyond the root zone using the SHAW model. Points are plotted to distinguish years where drift factors were computed from those where average drift factors were used; error bars denote uncertainty associated with one standard deviation in the average drift factor. Labeled points refer to years that are discussed in the text.

$$WAS = \sum_{i=1}^{3} A_i (P_{w/s,i} - S_{def,i}), \tag{2}$$

where WAS is the water available for sub-surface flow, A_i is the fractional area for each respective landscape unit, $P_{w/s,i}$ is its drift-adjusted winter–spring precipitation (i.e. through March for the low sagebrush, April for the mountain big sagebrush and through May for the aspen; Fig. 3), and $S_{def,i}$ is the simulated soil moisture deficit at the beginning of the water year required to bring the soil profile up to field capacity, taken as −0.1 bar water potential. Catchment average S_{def} ranged from 239 to 344 mm, with 1993 and 1995 representing two of the three highest soil moisture deficit years that produced streamflow. WAS could not include the water deficit below the root zone because ground water measurements were not available for the entire period. Fig. 7 presents the relation between streamflow and WAS. The relation has an r^2 of 0.83 for all positive values and therefore effectively accounts for the variability in streamflow as influenced by timing of precipitation and antecedent soil moisture. The slope and intercept suggest a threshold of 41 mm is necessary for streamflow that is likely re-

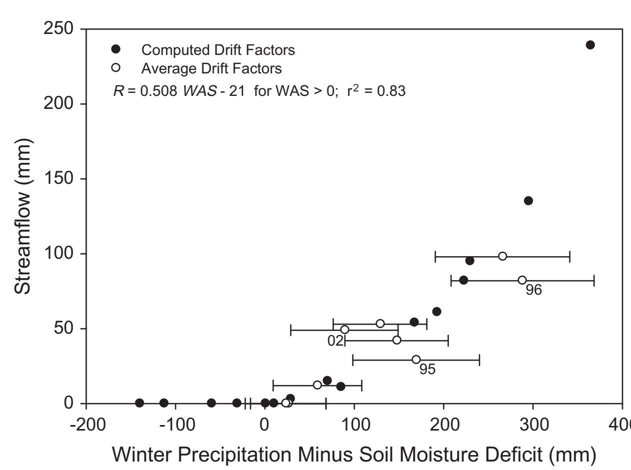

Fig. 7. Measured streamflow versus water available for sub-surface flow, defined as drift-adjusted winter–spring precipitation minus the simulated soil moisture deficit at the beginning of the water year. Points are plotted to distinguish years where drift factors were computed from those where average drift factors were used; error bars denote uncertainty associated with one standard deviation in the average drift factor. Labeled points refer to years discussed in the text.

lated to the deficit within the ground water system below the root zone. Comparison of Figs. 5 and 7 indicates that annual streamflow can be estimated based on precipitation and two parameters consisting of a runoff threshold value and slope term, but that use of the winter–spring precipitation and a soil water deficit term considerably improves the prediction.

Years 1998 and 2005 were both wet years following wet years, but both produced less streamflow than expected based on drift-adjusted precipitation (Fig. 5). Water year 1995 also had very low streamflow relative to precipitation. These were the only years with above average annual drift-adjusted precipitation with 70% or less occurring in the winter–spring period. This may be attributed to extremely high rainfall in May after a relatively dry winter for these years, especially 2005. May rainfall for 1995, 1998 and 2005 was 132, 153 mm and 131 mm, respectively, compared to the average May precipitation of 53 mm. Without the wet May, streamflow would not likely have occurred in 2005. Therefore, the streamflow for these years is primarily a response to rainfall distributed over the entire watershed as opposed to snowmelt concentrated in a small portion of the watershed. Because the low and mountain big sagebrush zones were already transpiring, much of the rainfall satisfied the already established soil water deficit and these areas contributed relatively little to streamflow. More of the water was therefore retained in the soil profile over the entirety of the watershed, resulting in less streamflow than expected based solely on annual precipitation. This observation indicates that the hydrologic regime in this semi-arid system is controlled by both the timing and phase of precipitation, relative to the timing of transpiration and degree of soil moisture abstraction.

The variability in streamflow due to antecedent moisture conditions, snow accumulation, and weather conditions is explained reasonably well by the SHAW simulations, even for the years where streamflow was dominated by rainfall, as illustrated in the relation between simulated percolation and streamflow in Fig. 6. The r^2 between streamflow and simulated percolation beyond the root zone is 0.85. Points that deviate from the general trend in Fig. 7 are mostly the same years as in Fig. 6, and all 4 years in Fig. 6 (1995, 1996, 1999 and 2002) are years where average drift factors were used. Neglecting only these 4 years or all of the years where the average drift factor was used both yield an r^2 of 0.96. Based on the regression parameters in Fig. 6, a threshold of 46 mm is required to produce streamflow, after which approximately 74% of the percolation results in streamflow. The threshold is very similar to the one in Fig. 7 (41 mm). Thus, the threshold in percolation beyond the root zone required to satisfy the average deficit in ground water storage is approximately 45 mm. Strongly threshold-driven event flow behavior has been observed in vastly different catchments (e.g.: Spence and Woo, 2003; Tromp-van Meerveld and McDonnell, 2006), and this work demonstrates that this can also be the case for annual flow in snow-dominated, semi-arid catchments where infiltrating water must replenish and connect a series of sub-surface reservoirs, in this case the soil profile and bedrock ground water system, before streamflow can occur.

4.5. Conceptual model formulation

Flerchinger et al. (1992, 1993) presented a conceptual model explaining rapid response of the lateral ground water flow system within the USC catchment. This long-term approach and additional analyses, based in part on model simulations, are critical to the development of a more complete conceptual model of the catchment. The following discussion should be considered somewhat speculative given that the model does not explicitly simulate 3-dimensional sub-surface flow, however it does provide a reasonable explanation of results from both the current and prior investigations. Inspection of Fig. 7 reveals that winter precipitation

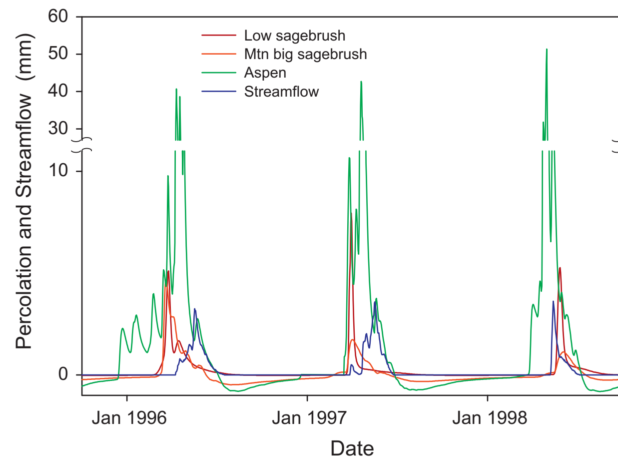

Fig. 8. Simulated percolation beyond the root zone for each landscape unit and measured streamflow for three selected water years.

and spring snowmelt must first satisfy the soil and ground water deficit before water is delivered to the stream. During this period, percolation of surface water to the ground water is critical to streamflow processes. The simulated contribution of each zone to the areal percolation beyond the root zone is plotted in Fig. 8 along with streamflow for a few selected years. Except for years such as 1998 that had a dry winter and percolation for the sagebrush was due to heavy rains in May, peak percolation from the sagebrush always precedes that for the aspen. Shortly after snow ablation, however, water flux in the root zone becomes decoupled from the saturated ground water system (7–25 m deep), as indicated by trivial water flux percolating beyond the root zone to the water table below the simulation domain (4 m). During those years when there was significant percolation beyond the root zone, this decoupling occurred on average around day 120 (late April) for the low sagebrush and day 140 (mid May) for the mountain big sagebrush unless there were heavy May rains. Decoupling for the sagebrush zones occurred around day 155 (early June) during years with heavy May rains (1995, 1998 and 2005). Conversely, decoupling for the aspen area typically occurred around day 170 (mid-June). This decoupling of the root zone from the saturated ground water system is similar to the transition between spring high-flux, late-spring drying, and summer dry periods in the conceptual model for the nearby Upper Dry Creek Watershed proposed by McNamara et al. (2005), however in this case the transitions occur in each landscape unit during different times.

Variability in Fig. 5 with respect to years 1995, 1998 and 2005 can be explained in part by precipitation variability associated with Oceanic Niño Index (ONI) anomalies. These were the only years that were warm phase ONI years (National Weather Service, 2010), typically referred to as El Niño years, that also had above average precipitation. This suggests that relatively wet El Niño years may produce less flow than for similarly wet non-El Niño years due to: (1) less spatially variable precipitation due to more of the precipitation occurring as rain; and (2) late-season replenishment of soil moisture reservoirs due to precipitation occurring later in the growing season. However, the variability associated with these anomalies (Fig. 5) is captured by the conceptual model, as illustrated in Figs. 6 and 7.

5. Conclusions

A long-term annual water balance of the semi-arid mountainous USC catchment was established by extending field data and model simulations from a 10-year study to 24 years. This 24-year study reaffirmed that the hydrology of the USC catchment is characterized by: (1) a distinct spring snowmelt peak dominated by snow drifts in the aspen; (2) a large ET component; (3) intermittent streamflow; and (4) strongly threshold-driven streamflow. However, extension of the USC catchment water balance to 24 years captured greater hydroclimatic variability that impacts the hydrology of the system and therefore enabled an improved fundamental understanding of the system.

Influence of the system variability on observed streamflow was captured reasonably well by hydrologic simulation. After validation, the SHAW model was used to provide estimates of ET and soil moisture storage for the water balance. Correlation between streamflow and percolation beyond the root-zone simulated by the SHAW model yielded an r^2 of 0.85.

Expanded analysis afforded by the additional years of data demonstrated that much of the variability between streamflow and annual precipitation ($r^2 = 0.54$) can be explained both by timing of precipitation and antecedent moisture conditions. Drift-adjusted winter–spring precipitation and antecedent moisture conditions accounted for 83% of the variability in streamflow (Fig. 7). The threshold in drift-adjusted winter–spring precipitation necessary to produce streamflow was attributed to the deficits in soil moisture and ground water at the beginning of the water year. This simple relation was nearly as accurate as applying the much more complex SHAW model and can be used to greatly simplify streamflow analyses for this and similar watersheds dominated by snowmelt and sub-surface flow.

This long-term approach is critical to the development of a more complete conceptual model of the system, beyond that for the lateral ground water flow system developed by Flerchinger et al. (1992, 1993). Because streamflow within the watershed is generated by water percolating through the soil profile and being delivered to the stream via shallow sub-surface and ground water flow (7–25 m deep; Flerchinger et al., 1992), winter precipitation and spring snowmelt must first replenish the deficit within the soil water profile and ground water system before water is delivered to the stream (Fig. 7). During this period, interaction between surface water and ground water is critical to streamflow processes. However, water flux in the root zone typically becomes decoupled from the ground water system shortly after snow ablation. The timing of this decoupling varies across the watershed, occurring first in the low sagebrush area around April and extending into June in the aspen area. Precipitation occurring after this decoupling rarely contributes to streamflow for the current water year, but serves to offset ET and reduce the soil moisture deficit at the beginning of the following water year. Finally, this study demonstrates the need for long-term catchment-scale research to improve our understanding of how climate and land cover interact to control hydrologic fluxes in complex mountainous terrain.

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