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Storage dynamics and streamflow in a catchment with a variable contributing area

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Abstract:

Storage heterogeneity effects on runoff generation have been well documented at the hillslope or plot scale. However, diversity across catchments can increase the range of storage conditions. Upscaling the influence of small-scale storage on streamflow across the usually more heterogeneous environment of the catchment has been difficult. The objective of this study was to observe the distribution of storage in a heterogeneous catchment and evaluate its significance and potential influence on streamflow. The study was conducted in the subarctic Canadian Shield: a region with extensive bedrock outcrops, shallow predominantly organic soils, discontinuous permafrost and numerous water bodies. Even when summer runoff was generated from bedrock hillslopes with small storage capacities, intermediary locations with large storage capacities, particularly headwater lakes, prevented water from transmitting to higher order streams. The topographic bounds of the basin thus constituted the maximum potential contributing area to streamflow and rarely the actual area. Topographic basin storage had little relation to basin streamflow, but hydrologically connected storage exhibited a strong hysteretic relationship with streamflow. This relationship defines the form of catchment function such that the basin can be defined by a series of storing and contributing curves comparable with the wetting and drying curves used in relating tension and hydraulic conductivity to water content in unsaturated soils. These curves may prove useful for catchment classification and elucidating predominant hydrological processes. Copyright © 2009 John Wiley & Sons, Ltd and Her Majesty the Queen in right of Canada.

KEY WORDS streamflow; storage; Canadian shield; hysteresis; contributing areas

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INTRODUCTION

Storage has long been recognized as important for runoff generation potential. The role of storage in influencing infiltration capacities and dictating source areas for runoff was first demonstrated by Betson (1964). There have been numerous field experiments since Dunne and Black (1971) first confirmed the importance of storage thresholds to hillslope subsurface and surface runoff response. Recent theoretical developments that have incorporated newer findings of the importance of storage into hydrological theory of runoff generation mechanisms (Sidle *et al*., 2000; Spence and Woo, 2003; Tromp van Meerveld and McDonnell, 2006b) built upon eminent work by the likes of Hewlett and Hibbert (1967). These new theories are most applicable in dry landscapes with disorganized drainage networks (Stichling and Blackwell, 1957; Allan and Roulet, 1994; Quinton *et al*., 2003) or after extended dry periods in wetter environments (Branfireun and Roulet, 1998; Tromp van Meerveld and McDonnell, 2006a). Most of these studies have focused on how storage heterogeneity affects runoff response on hillslopes or small basins via controls on hydrological connectivity (e.g. Devito *et al*., 1996; McNamara *et al*., 2005; Mielko and Woo, 2006). Upscaling of hillslope runoff and storage patterns to larger catchments has been attempted by integrating hydrological, hydrochemical and spatial analysis techniques (Gibson *et al*., 2002; Tetzlaff *et al*., 2007). Soulsby *et al*. (2004, 2006) point out that it has sometimes proven difficult, however, to select *a priori* the first-order controls on these patterns and successfully upscale them to the catchment. In this paper, the approach is to focus instead on the hydrological processes, particularly that of storage. The approach is justified on the basis of evidence from many landscapes that the hydrological process of storage, and its thresholds and heterogeneity, is crucial to the lateral transfer of water through and from catchments (McGlynn and McDonnell, 2003; Spence and Woo, 2006; Laudon *et al*., 2007). It also negates the influence of any assumptions on the relationships between first-order controls and runoff. The objective of the study is to observe the distribution of storage in a heterogeneous catchment and evaluate its significance and potential influence on streamflow. The objective is to be achieved using a water budget approach, measuring hydrological fluxes over a range of conditions over a 6-month period across a mesoscale catchment. The catchment was selected because its landscape diversity was expected to result in the storage heterogeneity necessary to address the study objectives.

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BAKER CREEK

Baker Creek is a stream characterized by lakes connected by short channels that drains water from ~165 km^2 of subarctic Canadian Shield into Great Slave Lake in Canada's Northwest Territories (Figure 1). The portion of the watershed that was investigated is upstream of the Water Survey of Canada (WSC) hydrometric gauge Baker Creek at the outlet of Lower Martin Lake (07SB013), draining a ~155 km^2 basin area. Three terrestrial land cover types dominate the basin (Table I). Surface water accounts for 19% of the basin area. There are 349 perennial lakes in the basin, many small, and the median and mean lake areas are 5400 and 88 800 m^2, respectively. The basin is in the zone of discontinuous permafrost. Large changes in topography, vegetation, winter snow accumulation, local hydrology and surficial geology over short distances result in abrupt transitions from permafrost to non-permafrost conditions. Permafrost is typically found in peat bogs and plateaus where organic material contributes to forming and maintaining permafrost, particularly in glacio-lacustrine clays. Permafrost is generally absent from areas of exposed bedrock and well-drained coarse thin overburden (Wolfe, 1998).

The 1971–2000 climate is characterized with the record from the Meteorological Service of Canada (MSC) climate station Yellowknife A, 5 km from the southern end of the basin. Yellowknife is characterized by short, cool summers with a July daily average temperature of 17 °C, and long, cold winters with a January daily average temperature of −27 °C. Annual unadjusted precipitation for the same period averages 281 mm, with 42% of that falling as snow. Convective cells produce much of the summer precipitation, producing high inter-annual variability, especially in July and August. The weather becomes cool and damp in autumn when periodic synoptic conditions allow persistent travel of cyclones over the region (Spence and Rausch, 2005). Annual snow cover begins in October and lasts until the end of April and beginning of May.

In most years, the largest input of water to the basin is at this time of spring freshet (Wedel *et al.*,

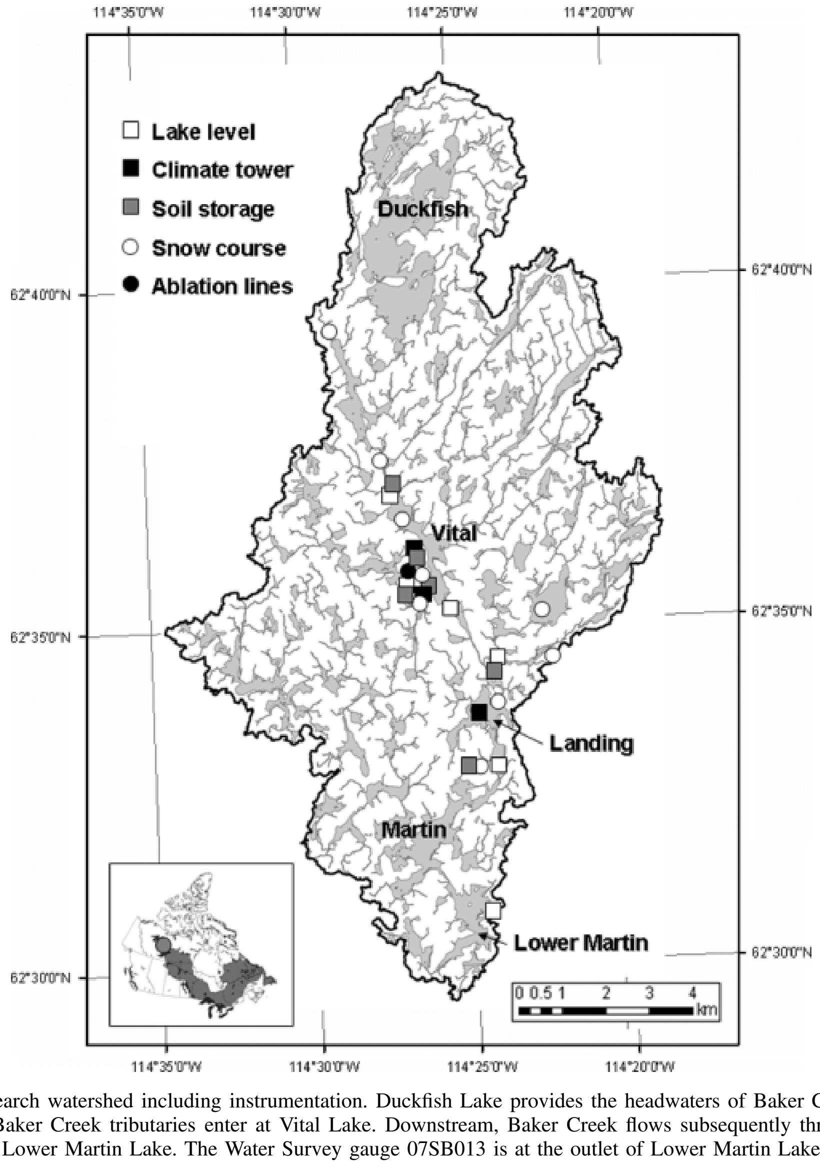

Figure 1. Baker Creek research watershed including instrumentation. Duckfish Lake provides the headwaters of Baker Creek, which flows towards Vital Lake. Many of the Baker Creek tributaries enter at Vital Lake. Downstream, Baker Creek flows subsequently through Landing, Martin and Lower Martin Lake. The Water Survey gauge 07SB013 is at the outlet of Lower Martin Lake

Table I. Predominant terrestrial land cover types and their characteristics in the Baker Creek catchment

Land cover type	% Cover	Primary vegetation species	Primary soil types
Exposed bedrock	30	*Cladonia* spp. *Sphagnum* spp. *Betula glandulosa* *Pinus banksiana*	n/a
Wetlands and peatlands	25	*Picea mariana* *Ledum groenlandicum* *Sphagnum* spp.	Organic possibly underlain by bedrock, sandy till or glaciolacustrine clays
Open forest	19	*Salix* spp. *Vaccinium augustifolium* *Cladonia* spp. *Picea mariana* *Eriophorum* spp. *Betula glandulosa*	Organic underlain by sandy till and glaciofluvial deposits. Some turbic and organic cryosols in poorly drained areas

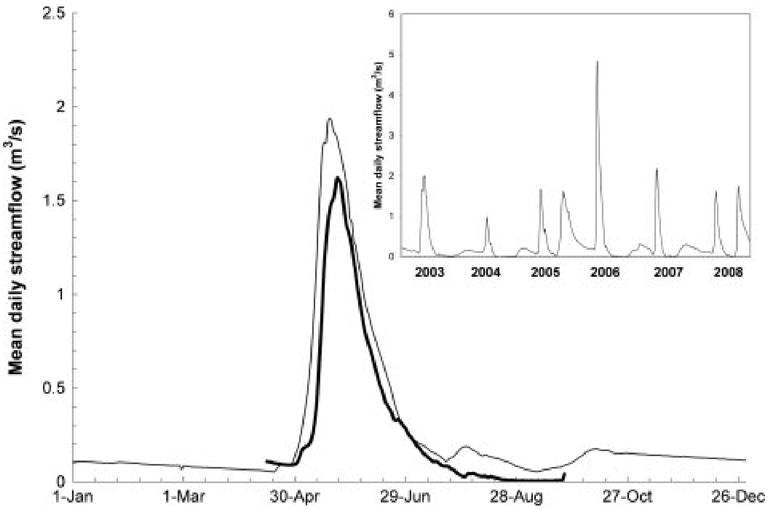

Figure 2. Baker Creek at the outlet of Lower Martin Lake average annual hydrograph (thin line) and 2008 streamflow during the study period (thick line). The inset shows the daily streamflow time series from 2003 to 2008 at the WSC hydrometric gauge 07SB013

1990), and the hydrological regime of the basin is described best as subarctic nival (Church, 1974) as this melt dominates the annual hydrograph of Baker Creek (Figure 2). The average annual streamflow at the outlet of Lower Martin Lake is 0·29 m^3/s or 59 mm/annum, providing an annual runoff ratio of 0·21. The runoff regime exhibits remarkable variation for a basin with almost 350 lakes, as the standard deviation of annual streamflow is 0·2 m^3/s or 40·5 mm/annum and annual runoff ratios range from 0·03 (1995) to 0·34 (2001). A maximum daily streamflow of 8·7 m^3/s has been observed, but common prolonged dry periods result in Baker Creek's intermittent discharge at the outlet of Lower Martin Lake (Figure 2).

METHODS

The Baker Creek catchment water budget, expressed as

$$\Delta S_c = P_c + M_c - Q_c - ET_c + \xi \qquad (1)$$

where ΔS_c is catchment change in storage, P_c is catchment rainfall, M_c is catchment snowmelt, Q_c is catchment streamflow at the outlet and ET_c is catchment evapotranspiration, was estimated for a study period of 15 April to 23 September 2008. ξ is any error in the calculation. All are expressed in terms of mm/day. The following sections outline how these terms were measured and/or estimated.

Meteorological terms

Meteorological terms necessary for calculations in Equation 1 were measured half hourly at four sites (Figure 1) for the entire study period. The first, representing large lake conditions, was located on a small rock outcrop exposure in Landing Lake. The second was on a ridge near the centre of the watershed surrounded by exposed bedrock and sparse jack pine. The third, representing more densely vegetated areas, was located in a black spruce, willow and alder wetland. The fourth was located just south of the watershed at a small headwater lake. Wind speed u (m/s), air temperature T_a (°C) and relative humidity RH (%) were measured at all climate towers. The wetland tower included a net radiometer to measure Q^* (W/m^2). Rainfall P_i (mm) was measured with tipping buckets at three of the climate towers. The manufacturer of the tipping bucket claims an accuracy of 5%. As daily differences measured at the stations were often <1 mm, values were averaged to attain P_c (mm/day), the catchment scale rainfall. The accuracy of P_c estimates is difficult to determine across the entire catchment, but measured study period differences among the tipping buckets did not exceed 10%.

Lake, soil and bedrock storage

Lake level data from the WSC gauge at the outlet of Lower Martin Lake were collected using WSC techniques and standards and were downloaded from the public WSC archive website (http://www.wsc.ec.gc.ca/hydat/H2O). Two lakes further up the main stem of Baker Creek: Landing Lake and Vital Lake, and three headwater lakes: Wetboot, Pocket and 690, were equipped with benchmarks and submersible pressure transducers that logged half-hourly water levels (Figure 1). When the transducers were not vented to the atmosphere, a single similar transducer mounted on shore near Vital Lake was used to correct for changes in barometric pressure. Transducer measurements were tied to local benchmarks during level surveys conducted at least once a month. Lake levels L at all the lakes are in reference to local datums (mald) except Lower Martin Lake which is in reference to sea level (masl). Lake storage S_l (m^3) can be divided into depression storage S_{dep} (m^3), water held in storage below the lake outlet elevation and detention storage S_{det} (m^3), water held in storage above the lake outlet elevation. A lake can be considered to have storage capacity S_{cl} (m^3) if the lake level is below the outlet elevation.

Detention storage and storage capacity were calculated for all observed lakes using the following formulae:

$$\begin{array}{lcccc} S_{cli} = A_{li} \cdot L & & L < L_T & & S_{cli} > 0 \\ & \text{if} & & \text{and} & \\ S_{deti} = A_{li} \cdot L & & L \geq L_T & & S_{cli} = 0 \end{array} \quad (2)$$

where A_{li} is individual lake area (m^2) derived from National Topographic System 1 : 50 000 scale mapsheets, and L_T is the lake outlet elevation. The measurements at Wetboot, Pocket and 690 were used to develop daily linear relationships between headwater lake area and S_{det} or S_{cl} and applied to estimate daily S_{det} or S_{cl} in unmonitored headwater lakes. Change in daily storage of every lake ΔS_{li} (mm/day) was calculated with

$$\begin{array}{lcc} \Delta S_{li} = (S_{cli_{(t)}} - S_{cli_{(t-1)}})/A_{li} & & L < L_T \\ & \text{if} & \\ \Delta S_{li} = (S_{deti_{(t)}} - S_{deti_{(t-1)}})/A_{li} & & L \geq L_T \end{array} \quad (3)$$

where (t) is the present and $(t-1)$ the previous time period. To attain change in storage for all the lakes ΔS_l (mm/day), each ΔS_{li} was multiplied by its lake area, the total summed and then divided by the total lake area. The accuracy of measured lake storage is a function of the accuracy of lake level and lake area measurements. Surveys with the local benchmarks imply that the transducers are accurate to within 5%. Tests of the storage extrapolation technique to gauged lakes suggest that the value of ΔS_l is within 15% of actual values.

A range of soil conditions occur within a range of typological, topographical and topological situations in the basin (Table II). Soil water storage was sampled at two locations of each type listed in Table II (Figure 1). Each of the six locations included site-specific calibrated ECH$_2$O-TE soil moisture and temperature probes installed horizontally at the soil surface and at 250 mm depth and recorded half hourly. Adjacent to the soil moisture strings were wells installed to the maximum fall thaw depth of 2006. Submersible pressure transducers inserted into the wells for the entire study period logged water table elevations half hourly. Manual measurements to confirm the transducer measurements of water table depth were taken at least every 6 weeks. Soil water storage at each location S_{si} (mm) was calculated as

$$S_{si} = \theta(z_T - z_w) + S_y(z_w - z_f) \quad (4)$$

where z_T is total soil depth (mm), z_w is height of the water table (mm) and z_f is the height of the frost table (mm). Measured specific yields S_y are summarized in Table II. The depth of the frost table was periodically measured using methods described in Spence and Woo (2003), which confirmed values applied in the relationship used to extrapolate ground thaw to unmeasured days (Woo and Steer, 1983):

$$z_f(\tau) = \beta\sqrt{\tau} \quad (5)$$

where τ is the period in days since the beginning of ground thaw and β is a constant defined as 124·81. Soil had storage capacity S_{csi} (mm) if S_{si} was less than the threshold S_{ti} (mm) at which the water table elevation was at the topographic surface and surface runoff from the land type occurs such that

$$S_{csi} = S_{ti} - S_{si} \quad (6)$$

Daily change in soil water storage ΔS_{si} (mm) was calculated as

$$\Delta S_{si} = S_{si_{(t)}} - S_{si(t-1)} \quad (7)$$

and change in storage for soil covered areas across the catchment ΔS_s (mm/day) was calculated as

$$\Delta S_s = \frac{\sum S_{si} \cdot A_{s_i}}{\sum A_{s_i}} \quad (8)$$

Table II. Properties of near-surface soils

Surface	Bulk density (kg/m^3)	Particle density (kg/m^3)	Porosity (%)	Specific yield
Wetlands	103·8	567	80	19
Peatlands	78	574	85	15
Arborous	113	644	83	16

where A_{si} is the area (m^2) of the soil covered land type. The expected accuracy of the soil storage measurements using these methods has been reported at 25% by Spence and Woo (2003)

In the instance of exposed bedrock, storage is the amount of water ponded on the bedrock surface and the storage threshold is controlled by depressions and the nature of soil patches on the hillslope (Spence and Woo, 2002). Daily bedrock storage S_b (mm) was calculated using a simple water budget model. The model was driven with data from the upland bedrock climate tower. The findings of Spence and Woo (2002) imply that a fixed infiltration rate of 1·3 mm/day and a storage threshold of 21 mm well represent the conditions across exposed bedrock ridges in the Baker Creek watershed. Change in bedrock storage ΔS_b (mm/day) was calculated using an equation similar to Equation 7. This model has reproduced bedrock storage values within 10% of those reported by Spence and Woo (2002). Catchment change in storage ΔS_c (mm/day) was calculated using

$$\Delta S_c = \Delta S_l + \Delta S_s + \Delta S_b \tag{9}$$

Evapotranspiration

Evapotranspiration over bedrock and lake surfaces was measured directly with an eddy covariance system consisting of a three-dimensional sonic anemometer and either an open path gas analyser (at the bedrock climate tower) or a krypton hygrometer (at the lake climate tower). Measurements of wind speed and water vapour content were taken at 10 Hz, and fluxes calculated over a half-hour period. Corrections to the eddy covariance measurements included coordinate rotation (Kaimal and Finnigan, 1994), the WPL adjustment (Webb *et al.*, 1980), sonic path length, high-frequency attenuation and sensor separation (Horst, 1997; Massman, 2000) and oxygen extinction.

Evapotransipiration from each of the three soil-covered land types ET_i (mm/day) was estimated using a Penman Monteith method as described in Shuttleworth (1993):

$$ET_i = \frac{1}{\lambda}\left(\frac{\Delta(Q^* - Q_g) + \dfrac{\rho_a c_p D}{r_a}}{\Delta + \gamma\left(1 + \dfrac{r_c}{r_a}\right)}\right) \tag{10}$$

where λ is latent heat of vapourization of water (J/kg), Δ is slope of saturated vapour pressure (kPa/°C), ρ_a is air density (kg/m^3), c_p is the specific heat of air (J/kg°C), D is the vapour pressure deficit (kPa) and γ is the psychrometric constant (kPa/°C). Q_g is ground heat flux (W/m^2) calculated with the Fourier heat flow equation with inputs from the ECH$_2$O-TE sensors noted earlier at each of the sites at two depths (z). Aerodynamic resistance r_a (d/m) was calculated following methods outlined in Brutsaert (1975), Shuttleworth (1993) and Monteith (1981). Canopy resistance r_c (d/m) was calculated with the revised version of the Jarvis (1976) and Verseghy *et al.* (1993) expressions (Lafleur and Schreader, 1994). Spence and Rouse (2002) reported an accuracy of 20% with methods similar to those selected for this study. Evapotranspiration from each land surface type was prorated by its fractional coverage to obtain the catchment-scale evapotranspiration ET_c (mm/day).

Snowmelt

An end of winter, stratified snow survey was conducted as close as possible to the maximum winter snow pack depth on 12 April 2008. The stratification included land cover, with measurements taken to include the diversity of vegetation cover, slope, aspect and wind conditions expected within each cover type. The snow water equivalent of the snow pack was calculated from snow density measured with an Eastern Snow Conference snow sampler, and snow depth was measured with an aluminium rod. At least one snow course was located in each land surface type present in the catchment, including coniferous forest, wetlands, exposed bedrock, deciduous forest and lakes (Figure 1). Each course included at least ten depth samples and six density samples. The accuracy of such snow surveys is expected to be within 15% (Pomeroy and Gray, 1995).

Daily snowmelt M_i (mm/day) was calculated by measuring the lowering of the snow surface and the snow density of the surface snow layer, along ablation lines (Heron and Woo, 1978) in different land surface types (i) near the bedrock and wetland climate towers (Figure 1). Catchment snowmelt M_c (mm/day) was calculated by prorating M_i by the fractional coverage of snow covered area in each land surface type. Heron and Woo (1978) report an accuracy of 25% with this method.

Streamflow

Streamflow data were derived from two sources. WSC gauge 07SB013 data were downloaded from the WSC public archive website noted above and is the streamflow loss from the catchment Q_c (mm/day). At other sites in the basin, streamflow was measured periodically using area–velocity methods with current meters, following WSC standards as close as possible. Stage discharge curves were constructed for the Landing, Vital and 690 outlets using observed streamflows and lake levels and used to estimate streamflow for days it was not measured. The regression coefficients of these curves were never less than 0·92. The WSC claims accuracy of within 10% of its published data.

Storage capacity and contributing area mapping

A July 2006 Landsat TM image underwent an unsupervised classification with channels 3, 4 and 5 in PCI Geomatica to determine the spatial distribution of land surfaces in the catchment. The initially identified nine land surface types were amalgamated into six classes: exposed bedrock, water, coniferous stands, deciduous stands, wetlands, and peatlands. The classified image was then imported into a geographic information system (GIS) within which daily storage capacity values were calculated for each lake with Equation 2, and daily storage capacity values were calculated for each terrestrial land surface type using Equation 6 for each day through the study period. Maps of saturated, or active, areas (Ambroise, 2004) were derived from the daily storage capacity maps by selecting those areas with storage capacity values of zero. These maps were further refined to produce maps of areas connected by surface runoff to the outlet of Lower Martin Lake: that is, contributing areas (Ambroise, 2004), by deleting any active polygons that were not contained within a contiguous extent that reached to the outlet of Lower Martin Lake. Five-hectare sub-watersheds delineated from a 10-m digital elevation model derived from a lidar survey were overlaid with the basin during this process to ensure that the contributing area maps respected drainage divides.

RESULTS

Catchment water budget

The 2008–2009 water year began when air temperatures approached freezing on 15 April 2008. A catchment-wide snow survey on 12 April 2008 implied that the spring snowpack averaged 93 mm of snow water equivalent. Average daily temperatures below freezing between 15 April and 27 April kept the average melt rate to 0·8 mm/day (Figure 3). The snow-covered fraction of the basin dropped from 1·0 to 0·88 by 27 April. Substantially warmer temperatures after 28 April increased the average daily melt rate to 10 mm/day over the next week (Figure 3) and removed the remaining snow cover by 3 May.

Most of the snowmelt was initially directed to storage on bedrock outcrops (Figure 3). A wet fall in 2007 saturated many of the soils in the catchment prior to freeze up. This prevented the meltwater from increasing soil storage, and the meltwater was subsequently directed to headwater lakes and then the higher order lakes along the main channel. The ~70 mm of snowmelt produced 27 mm of spring runoff and a runoff ratio of 0·39 at the outlet of Lower Martin Lake. It was not until the first rainfall in the second half of May that soil storage noticeably increased.

Rainfall was sparse through the first half of the summer of 2008. Only 7·3, 30·0 and 15·2 mm fell

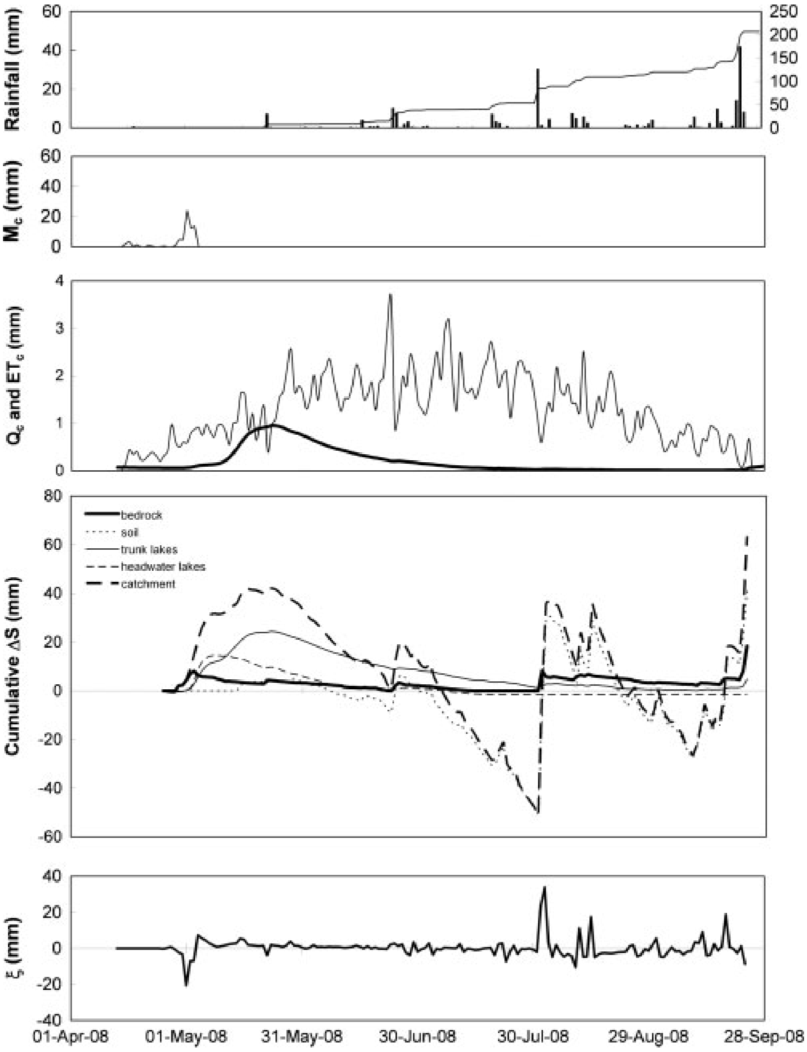

Figure 3. Water budget components from top to bottom for the study period: cumulative rainfall and rainfall time series, catchment snowmelt, catchment streamflow and evapotranspiration, cumulative change in storage for the catchment and landscape types, and error

in May, June and July, respectively. This resulted in a spring freshet almost exclusively dictated by the snowmelt pulse that peaked on 1 May. While Q_c reacted quickly to snowmelt inputs, there was a 20-day lag between snowmelt and streamflow peak. Ample water on the landscape in the interim supported terrestrial evapotranspiration rates larger than the streamflow loss from the basin. After a brief 2-day period on 21 and 22 May, when streamflow was larger than all losses to the atmosphere, lake evaporation became the greatest flux from the catchment through to the end of the study period. Storage steadily declined through the end of July.

Notable multi-day rainfall events of 24·7 and 36 mm at the end of June and beginning of August had little impact on Q_c. Both the June and August rainfall events generated ~0·4 mm of runoff and had runoff ratios of 0·014 and 0·013, respectively. The vast majority of the rainfall was directed to bedrock and then to soil storage (Figure 3). Wet conditions through August led to little change to streamflow response, as these inputs could not counteract storage demands and losses to the atmosphere. It was not until 9 September, the onset of senescence and decrease in evaporative demand, that storage began to significantly increase in response to rainfall inputs. A 56·7-mm event that began on 21 September caused an observed change in storage of 48 mm. The study period ended on 23 September due to logistical reasons, so water budget calculations could not be made after this date, but streamflow at the Water Survey gauge responded to this event and peaked on 10 October at 1·84 m^3/s, which was 0·13 m^3/s higher than the spring peak in May.

The smallest daily values of ξ occurred during the dry periods from May through July (Figure 3). The largest daily values of ξ coincided with large inputs to the watershed, notably early in the spring freshet and the 1 August rainstorm. The spatial variability of inputs from these two particular types of events, a snowmelt and convective rainfall storm, are difficult to quantify over an area as large as the Baker Creek catchment. The large 21 September event was measured remarkably well, perhaps because the storm was more frontal in nature and distributed rainfall more evenly through the watershed. Other large errors at the end of September may be due to observed but not measured falling snow and its subsequent melting. While there were compensating errors in the water budget measurements, error was low relative to estimated fluxes for much of the study period. This suggests that faith can be placed in the estimates of the components of the water budget.

Storage capacity

At the beginning of the study period, the largest values (>80 mm) of storage capacity were in the headwater lakes (Figure 4). The exposed bedrock was dry and at its maximum available storage capacity of 21 mm. Patches of peatlands and other soil-filled areas had no available capacity after they froze under saturated conditions during a wet fall in 2007. Some larger headwater lakes, notably Duckfish, were also at capacity and providing streamflow. This was also the case for the higher order lakes along Baker Creek, including Vital, Landing, Martin and Lower Martin (Figure 4a). Snowmelt inputs brought the entire watershed to capacity on 3 and 4 May. A drop in cumulative change in storage reflects an increase in storage capacity, and this occurred first on the exposed bedrock (Figure 3). By mid June, lake levels in all but the largest headwater lakes (e.g. Duckfish and 690) were below outlet elevations. Capacity had grown to as high as 35 mm in peatlands because of higher rates of evapotranspiration than rainfall (Figure 3), and is reflected in the storage capacity pattern illustrated on Figure 4c. This pattern enhanced in July (Figure 4d)

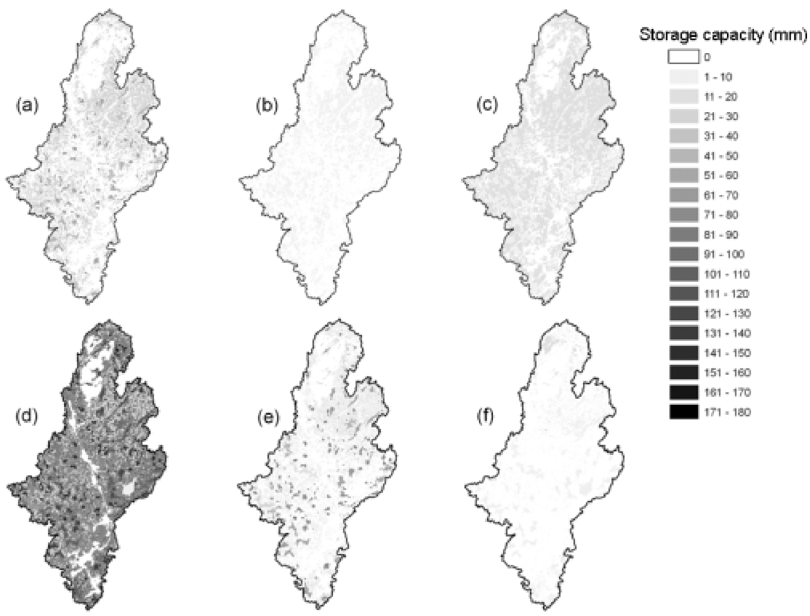

Figure 4. Storage capacity (mm) across the Baker Creek catchment on 26 April (a), 6 May (b), 10 June (c), 30 July (d), 19 August (e) and 23 September (f). The letters denote the same days as in Figures 5 and 6

as headwater lakes 150 mm below outlet elevations were common. Bedrock storage capacity was often at the maximum of 21 mm. Soil storage capacity often exceeded 50 mm in July. Increasing precipitation rates in early and mid August were able to exceed bedrock storage thresholds and generate runoff that was directed to soil downslope. Water was often efficiently shed from bedrock to permit some storage capacity to remain between rainfall events. With rainfall augmenting upslope runoff, soil storage approached storage thresholds several times in August (Figure 4e). Levels in lakes at the bottom of headwater basins rose, but most remained well below outlet elevations. Even wetter conditions in September allowed the wetting of the catchment to continue, whereby at the end of the study period all the hillslopes were above storage thresholds but most of the headwater lakes remained below as of 23 September (Figure 4f). Every headwater lake observed in the basin after 23 September during the final stages of field work was filled and providing water downstream.

The area at capacity across the watershed at the beginning of the study period was 94 km^2 (61%) (Figure 5). After a brief 2-day period in the beginning of May during which the entire watershed was at capacity, 108 km^2 (70%) remained at capacity from 5 to 28 May. As the basin dried because of sparse rainfall and ample evapotranspiration, the area at capacity fell to a minimum of 15 km^2 (10%) at the end of July. Changes in the area at capacity happened quickly in August and September. Rainfall at the end of July and in the first half of August brought over half the catchment above capacity for the second half of August. Vertical (evaporative) and lateral (runoff) losses reduced this area close to seasonal minimums again in September. Storage capacity values were much smaller than at the end of July; however, when large amounts of rainfall came in at the end of the month, 134 km^2 (89%) across the catchment was able to quickly reach capacity at the end of the study period.

Contributing areas

Throughout much of the study period, most headwater lakes were below their outlet elevations and soils contained storage capacity, even after significant rainfall was added to the watershed (Figure 4). This storage capacity prevented water generated from upslope active areas from proceeding to the basin outlet. These areas, while capable of generating runoff, were disconnected from and had no access to the basin outlet. This behaviour has been documented earlier in landscapes comparable to the Baker Creek catchment (Spence, 2000; Mielko and Woo, 2006). Contributing area to the outlet of Lower Martin Lake was always less than the active area except at the wet and dry extremes. Contributing areas to the outlet of Lower Martin Lake on 26 April were limited to the higher order lakes and immediately adjacent catchments (Figure 6a). The 70 mm of snowmelt input expanded the contributing area to the entire possible catchment on 3 and 4 May (Figure 5) after which it contracted into valley bottoms and lakes (Figure 6b). The trend through June and July was for contributing areas to contract along these locations (Figure 6c and d). Rainfall events would sometimes expand the contributing areas to a state well represented by conditions on 19 August (Figure 6e). As wetter conditions continued into September, active areas in headwater catchments across the basin grew. The expansion of contributing area to the Lower Martin Lake outlet grew first as active areas in catchments adjacent to the higher order lakes reached their maximum extent. This doubled contributing area from 15·9 to 27·9 km^2. Subsequent growth in the contributing area to Lower Martin came when levels in lakes at the bottom of larger sub-catchments reached outlet elevations and runoff was permitted into Baker Creek (Figure 6f). This growth lagged behind the increase in active area but still resulted in an increase in contributing area of an order of magnitude to 155·6 km^2 (Figure 5).

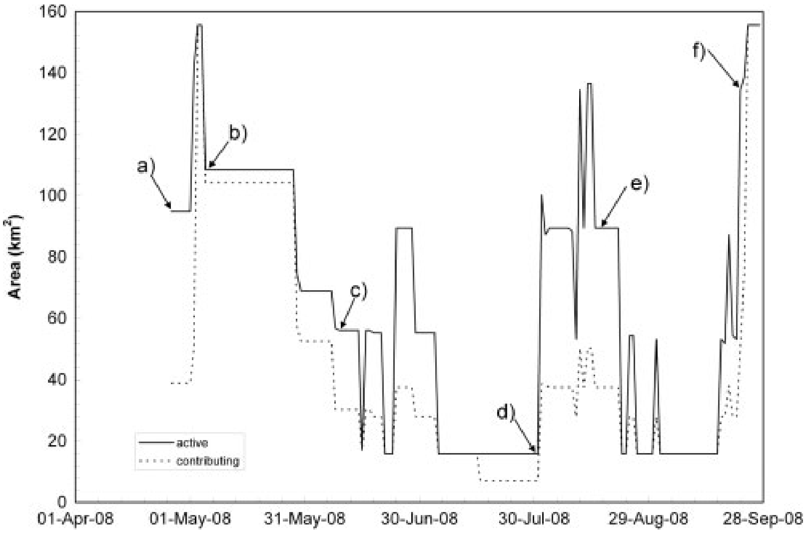

Figure 5. The change in areal extent of active and contributing locations through the study period. The letters denote the same days as in Figures 4 and 6

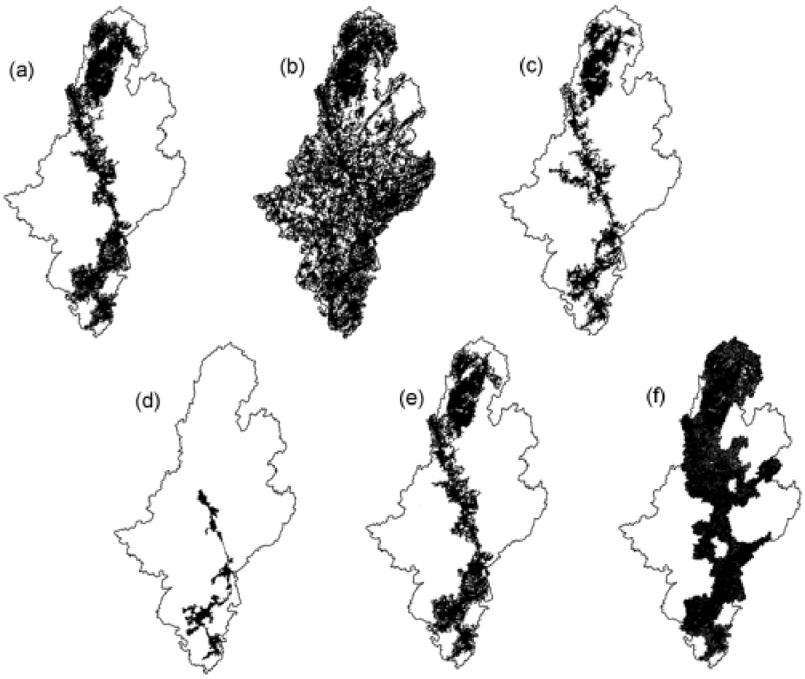

Figure 6. Contributing areas to streamflow at the outlet of Lower Martin Lake. The letters denote the same days as in Figures 4 and 5

Storage–discharge relationships

The rising and falling limb of the hydrograph during the spring freshet are apparent from the relationship between Q_c and S_c (inset to Figure 7). During the rising limb, storage remained relatively constant as it was transferred downstream to Lower Martin Lake. This is also reflected in Figure 3 from 7 May to 1 June. There were periods with large scatter between Q_c and S_c, suggesting that streamflow was not influenced solely by the amount of storage but also by its other characteristics. Even when summer runoff was generated on the hillslopes, storage capacities in intermediary locations, particularly headwater lakes, prevented water from transmitting to Baker Creek. Figure 6 illustrates how the topographic bounds of Baker Creek basin constitute the maximum potential contributing area to streamflow, not the actual contributing area. Usually, only a portion of catchment storage had direct access and control on streamflow at the basin outlet. Connected storage and streamflow exhibit a relationship that better reflects the physical processes acting within the catchment (Figure 7). The influence of an increase in storage at the beginning of the water year and the subsequent rise in streamflow until 16 May remain evident. Strong, sometimes linear, relationships between storage and streamflow during the recession occurred during distinct periods of stable contributing area to the outlet.

When a tributary stopped contributing streamflow, there was a notable decrease in connected storage. This reflects the bulk removal of a portion of catchment storage that was influencing the outlet response. On 24 June, the curve trended upwards in response to rainfall, but then dropped slowly tending along the same slope experienced earlier around 18 June. The decreases in contributing area on 5 and 16 July noted in Figure 5 are also evident. Increases to the contributing area with rainfall input at this time of year were limited to areas immediately adjacent to the higher order lakes (Figure 6).

CATCHMENT EFFICIENCY

Catchment streamflow can be interpreted as the outflow response from a series of reservoirs hydrologically connected to the outlet (Nash, 1957; Dooge, 1959; Wooding, 1965) where outflow can be expressed as a function of storage such that

$$Q_c = KS_{\leftrightarrow}{}^m \text{ and } Q_c = 0 \text{ if } S_{\leftrightarrow} = 0 \qquad (11)$$

where $S_{\leftrightarrow}$ is the hydrologically connected storage within active areas and m is a dimensionless coefficient. Spence (2007) interpreted K (units of $1/T$) as the efficiency with which a catchment can convert storage to streamflow. Figure 7 illustrates that each suite of reservoirs has a linear relationship between Q_c and $S_{\leftrightarrow}$ for short time periods during which m can have a value equal to 1.

The catchment exhibited increasing efficiency from the onset of the spring freshet to a peak on 9 June (Figure 8), which is comparable to the conditions illustrated in Figures 4c, 5c and 6c. The peak in K was subsequent to peak streamflow, suggesting the presence of a variable and hysteretic relationship between storage and streamflow similar to that often seen in open channel flow (Chow *et al*., 1988). This phenomenon would be due to a backwater effect as some water was held in storage during the rising streamflow, such that there was more discharge of water for the same level of storage during the falling limb. The number of lakes in this particular watershed likely enhanced this effect relative to what might be observed in a stream dominated watercourse.

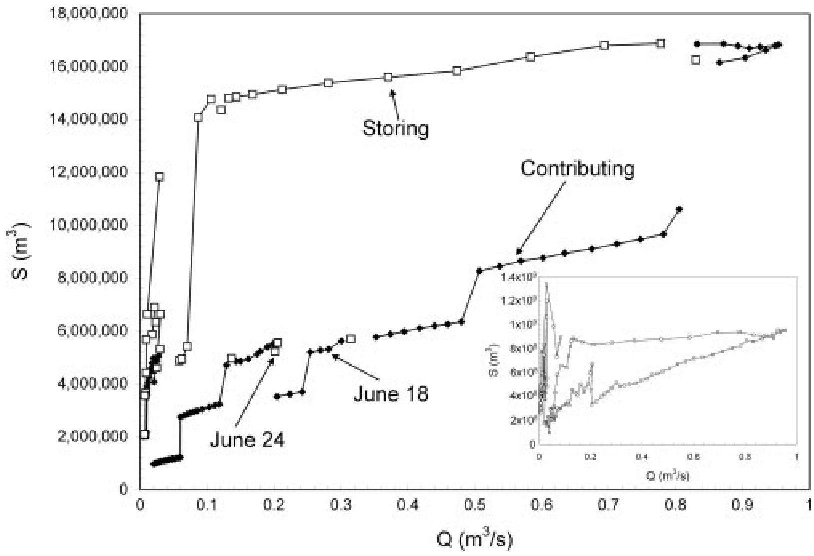

Figure 7. Streamflow at the outlet of Lower Martin Lake versus the volume of water stored in areas hydrologically connected to the outlet of Lower Martin Lake. The inset shows the relationship between streamflow and the entire volume of water stored across the Baker Creek catchment. The black diamonds denote days on which a contributing function was predominant, and the white squares a storing function

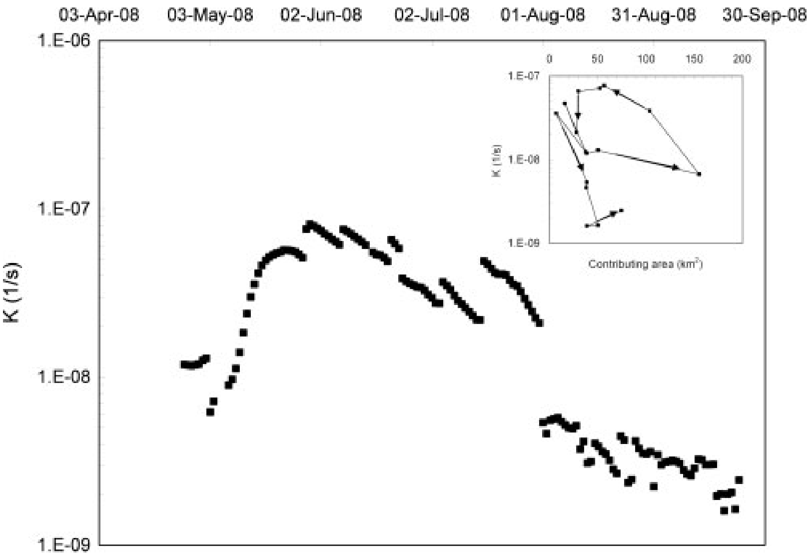

Figure 8. Basin efficiency changes through the study period with the inset showing the hysteretic relationship with contributing area

The relationship between contributing area and catchment efficiency was also hysteretic (inset to Figure 8). The most efficient period was not when the contributing area was at a maximum, but later, when water had been transferred downstream and Lower Martin was at its peak water level. Spence (2007) suggested that the largest store of water in a catchment will control basin scale K. These results instead imply that the basin efficiency is closely associated with the storage state of the element adjacent to the outlet and the specific controls on how water can drain from that location.

CONNECTIVITY

There were sometimes differences in what Ambroise (2004) defines as the active and contributing areas (Figures 4, 5 and 6) through the study period. Lack of physical stream connections between active areas for most of the study period meant that the contributing area remained relatively small. The contrast between spring and summer runoff ratios and a decrease in efficiency over the summer (Figure 8) illustrate the impact different degrees of connection can have on the ability of a watershed to generate streamflow. The results presented imply that there is a relationship between connectivity of active areas, the efficiency of the watershed to generate streamflow and the magnitude of streamflow from the Baker Creek catchment. Successful statistical approaches to measuring connectivity have concentrated on the plot, hillslope or headwater scale (Western *et al.*, 2001; James and Roulet, 2007). At these scales, there is not necessarily a need to account for catchment-scale topographic features such as drainage divides, which require a more physically and structurally based approach. Michaelides and Chappell (2009) state that process-based measures of connectivity provide a general description of the process but lack the stringent requirements for robust quantification. Conversely, pattern-based measures allow clear quantification of property connectivity, but lack an ability to incorporate a multitude of processes. An exercise to measure connectivity is beyond the scope of this paper. Given the presence of hysteresis in storage, streamflow and efficiency observed here at the catchment scale, future efforts to devise a measure for connectivity at the catchment scale should incorporate hydrological process, state and pattern within that metric.

THE FORM OF CATCHMENT FUNCTION

Black (1997) proposed that the three hydrological functions of a watershed are collection, storage and discharge. Spence and Woo (2006) proposed that the discharge, or what will be referred to here as the contributing function, begins with runoff, but Spence (2007) showed that if the runoff rate q is larger than the absolute value of the change in storage ds/dt, then the contributing function is predominant in the catchment. Vice versa, if change in storage is larger than the runoff rate, a storage function is predominant. As with basin efficiency, there was a hysteretic pattern to basin function (Figure 7). Once drained, either prior to spring freshet or near the end of July, the catchment stored any water introduced from snowmelt or rainfall, releasing relatively little of it. During these periods, even though q increased, ds/dt remained the larger of the two. During the spring freshet, it took almost a month before the streamflow rate exceeded the change in storage, and the basin switched to a contributing function. The patterns in Figure 7 could be interpreted as comparable to the wetting and drying curves commonly observed in unsaturated porous media (Freeze and Cherry, 1980). Secondary contributing and storing curves are visible near 24 June and also during low flows (Figure 7). These types of hysteretic patterns between storage and outflow have been documented at a range of scales, including soil columns (Huang *et al.*, 2005), hillslopes (Dunne, 1978) and stream channels (Chow *et al.*, 1988). The results provide evidence that they also manifest at the catchment scale and that they would imply a basin's functional state, as conceptually illustrated in Figure 9.

If we assume, as is done with unsaturated soils (van Genuchten, 1980), that the curve form is dictated by the statistical distribution of the terms under investigation, the log normal nature of storage and streamflow distributions imply that the form of the storing curve is

$$S = x \cdot \ln(Q) + y \tag{12}$$

where x and y for the primary curve in Figure 7 are $3{\cdot}28 \times 10^6$ and $1{\cdot}9 \times 10^7$, respectively. Conversely, the form of the contributing curve in Figure 7 is

$$S = a \exp^{bQ} \tag{13}$$

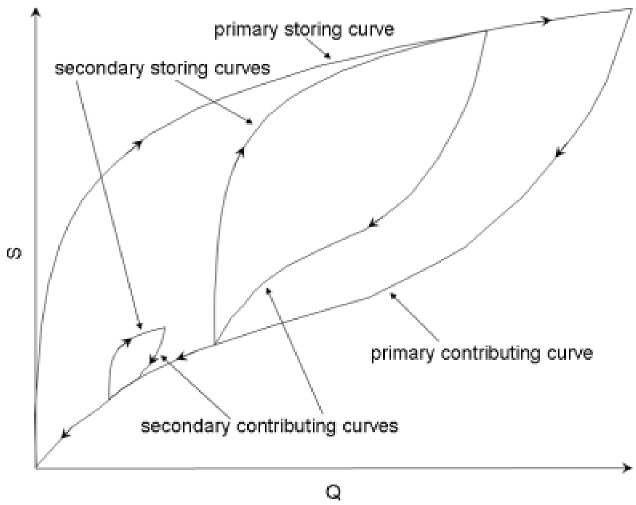

Figure 9. Conceptual catchment storing and contributing curves including primary and secondary curves

where the values for a and b are 2×10^6 and $2{\cdot}7285$, respectively. Primary wetting and drying curves are derived experimentally beginning with dry soils. The wet state of the Baker Creek watershed prior to the study period and the fact that 2008 was not a record high water year suggest that the values listed above for Equation 12 do not denote the primary storing curve for Baker Creek but a subordinate curve. The primary catchment storing curve can only be observed following a complete drying of the watershed and a wetting to the maximum storage and streamflow possible. In contrast, the values above for the contributing curve intercepted the primary contributing curve as streamflow approached zero at the outlet of Lower Martin Lake near the end of July. As noted above, the data show the existence of several subordinate curves during the study period. The values of x, y, a and b could not be derived for these curves, as the catchment often had a predominantly storing function for just 1 day.

Storing and contributing curves for other catchments could be interpreted from similar data. Recognizing that their existence is useful for understanding the runoff generation process and they permit a numerical representation of the physical relationship between storage in dynamic contributing areas and streamflow response at the catchment outlet. Just as with unsaturated soils, their derivation requires experimental data across a wide range of conditions within the catchment. Storage is notoriously difficult to quantify at any scale. Furthermore, as with many unsaturated soils, the coefficients of each subsequent curve are expected to be different. Finally, the curves for unsaturated soils are bounded by porosity and, while slope and topography may act as an index, catchments have no such easily quantified storage bound. The external influence of climate is also at play. In general, though, some insight could be gained by evaluating how first-order controls such as climate, topography and soils affect the storing and contributing curves of different catchments. Wetter climates may produce streamflow less influenced by thresholds, resulting in a flatter storing curve. The contributing curve also may be more linear as the relative loss to evapotranspiration is reduced. Steeper topography and shallow soils would permit less storage relative to streamflow and result in more linear curves. The topology of important landscape features for runoff production and their storage state (Soulsby *et al.*, 2006) would also have an influence on the shape of storage–discharge curves (Shaw *et al.*, in review). Compilation of curves from a diversity of catchments may aid in catchment classification (McDonnell and Woods, 2004). Application of the contributing and storing curves in a catchment classification system would permit basins to be classed not only by their streamflow response (e.g. Church, 1974) but also incorporate hydrological function and could infer the first-order controls responsible for catchment behaviour (Buttle, 2006).

CONCLUSIONS

The objective of this study was to observe the distribution of storage in a catchment and evaluate its significance and potential influence on runoff generation at the catchment scale. The spatial distribution of headwater storage controlled which active areas were able to transfer water to the basin outlet. This dictated the contributing area to the basin outlet and, in turn, the volume of the rainfall or snowmelt runoff response. The efficiency of the catchment to convert rainfall or snowmelt stored in the basin to runoff is controlled by (i) where the storage is, (ii) how accessible it is to the outlet and (iii) how well it can exit the outlet once it gets there. The connectivity of active areas is clearly important to basin response. Currently, hydrologists have only a few techniques to measure or quantify connectivity. Past approaches have examined connectivity of areas of similar moisture state with statistical approaches, but a more physically and structurally based approach is required at the catchment scale. Establishing a set of organizing principles to account for connectivity and the non-linear stream flow response to rainfall and snowmelt inputs at the basin scale, determined by sub-basin scale storage thresholds, has the potential to advance predictive capabilities, especially in small ungauged basins.

This study also documented a hysteretic relationship between storage and streamflow at the catchment scale. While hysteresis is well known to exist at smaller scales and in the stream channel, this is the first documentation of the pattern at the catchment scale. The recognition and parameterization of this hysteresis provides insight into basin behaviour, efficiency and function. It may also provide a means with which to compare catchments in different landscapes.

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