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Hydrogeologic controls on streamflow sensitivity to climate variation

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Abstract:

Climate models project warmer temperatures for the north-west USA, which will result in reduced snowpacks and decreased summer streamflow. This paper examines how groundwater, snowmelt, and regional climate patterns control discharge at multiple time scales, using historical records from two watersheds with contrasting geological properties and drainage efficiencies. In the groundwater-dominated watershed, aquifer storage and the associated slow summer recession are responsible for sustaining discharge even when the seasonal or annual water balance is negative, while in the runoff-dominated watershed subsurface storage is exhausted every summer. There is a significant 1 year cross-correlation between precipitation and discharge in the groundwater-dominated watershed ($r = 0{\cdot}52$), but climatic factors override geology in controlling the inter-annual variability of streamflow. Warmer winters and earlier snowmelt over the past 60 years have shifted the hydrograph, resulting in summer recessions lasting 17 days longer, August discharges declining 15%, and autumn minimum discharges declining 11%. The slow recession of groundwater-dominated streams makes them more sensitive than runoff-dominated streams to changes in snowmelt amount and timing. Copyright © 2008 John Wiley & Sons, Ltd.

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INTRODUCTION

The mountains of the Cascade Range provide critical water supplies for agriculture, municipalities, and ecosystems in the Pacific Northwest region of the USA. Recent analyses show that this region is sensitive to current and projected climate warming trends, specifically reduced snow accumulation and earlier spring melt, leading to a decline in summer streamflow. Temperature trends over the past half-century indicate warmer winters across the western USA (Folland *et al*., 2001) and particularly in the Pacific Northwest (Mote, 2003). These data are corroborated by ample proxy evidence of recent temperature increases in the Pacific Northwest (Cayan *et al*., 2001; Regonda *et al*., 2005; Stewart *et al*., 2005). Snowpacks in western North America have declined over the past 50 years, primarily due to an increase in winter temperatures (Mote *et al*., 2005). In accordance with earlier spring snowmelt, streamflow timing is earlier by 1 to 4 weeks, compared with the middle of the 20th century (Stewart *et al*., 2005). Climate models predict continued winter warming of 0·2 to 0·6 °C per decade in the Pacific Northwest (Christensen *et al*., 2007; Mote *et al*., 2003; Washington *et al*., 2000). By 2050, Cascade snowpacks are projected to be less than half of what they are today (Leung *et al*., 2004), potentially leading to major water shortages during the low-flow summer season. Regional-scale analyses identify climatic gradients as primary controls on changing streamflow regimes, but the potential for other hydrologic factors, notably groundwater, to influence this response has received much less attention.

Snowpack is the dominant water storage compartment in many watersheds, and snowmelt is quickly translated into streamflow. In the western USA, where there is little summer rainfall, the snowmelt peak is often the year's highest discharge event, and marks the beginning of a long decline in streamflow through the summer months. Groundwater systems can mediate the translation of water from snowpack to streamflow by providing an additional storage compartment and damping variations in precipitation and recharge. Extensive groundwater systems are often found where the bedrock is basalt, limestone, or fractured rock.

The object of this work is to develop an understanding of how discharge is controlled at the event, seasonal, and inter-annual scales by the interactions among groundwater storage, snowmelt dynamics, and regional climate signals. The over-arching goal is to better predict hydrologic responses to future climate change in groundwater-dominated watersheds of the western USA. The focus of this research is an analysis of the historical discharge, precipitation, and snowpack records from two adjacent watersheds on the western slope of the Oregon Cascades. One watershed has an extensive

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groundwater system (McKenzie River at Clear Lake), while the other watershed is runoff-dominated with little groundwater storage (Smith River). Results from the study watersheds should be broadly applicable to geologically similar watersheds in the Oregon Cascades and northern California.

Most watersheds on the western slope of the Oregon Cascades encompass elevations that receive winter precipitation as a mixture of rain and snow. These watersheds have complex winter hydrographs that are dependent on the distribution of rain and snow during individual events, which in turn is controlled by storm temperatures and catchment hypsometry. Snow cover typically accumulates at temperatures close to the melting point, and thus is at risk from climate warming because temperature affects both the rate of snowmelt and the phase of precipitation. With a projected 2 °C winter warming by mid-century, 9200 km^2 of currently snow-covered area in the Pacific Northwest would receive winter rainfall instead (Nolin and Daly, 2006).

Summer and autumn discharges are of interest because they occur during periods of low flows, high water demand, and critical temperatures and flows for salmonids and other fish species. Decreases in summer streamflows restrict junior water rights users, constrain hydropower generation, and make in-stream flow targets difficult to meet. Lowered summer discharge can also intensify stream temperatures warmed by hot summers (Webb *et al.*, 2003), and some species are very temperature-sensitive (Selong *et al.*, 2001), such as the threatened bull-trout that inhabits the McKenzie River watershed. Decreased minimum flows have also been linked to drought stress and mortality in riparian vegetation (Naiman *et al.*, 1998). Regional climate models predict that Pacific Northwest summers will become hotter and drier over the next century (Christensen *et al.*, 2007), exacerbating existing stresses.

Spatial patterns of summer streamflow in the Oregon Cascades exhibit significant differences between the geologically-distinct High and Western Cascade provinces (Tague and Grant, 2004). A key control on these differences is the partitioning of water to slow-draining basaltic aquifers in the High Cascades, versus shallow subsurface storm flow through the upper few metres of soil and bedrock in the Western Cascades. Saar and Manga (2004) estimated a groundwater recharge rate of 100 cm yr^{-1} for the High Cascades, while Ingebritsen *et al.* (1994) estimated a recharge rate of 3 cm yr^{-1} for the Western Cascades bedrock aquifers. High Cascades aquifer volumes are seven times greater than annual discharge (Jefferson *et al.*, 2006). Groundwater discharge to High Cascades streams also has a moderating effect on water temperatures, resulting in lower diurnal variability and colder summer temperatures (Tague *et al.*, 2007).

Tague *et al.* (2008) used the hydro-ecological model RHESSys to examine the influence of geology on streamflow response to 1·5 °C and 2·8 °C warming scenarios. Their study included the High Cascades watershed used in this project, McKenzie River at Clear Lake. Their model showed that warmer temperatures resulted in greater reductions in August discharge and annual minimum flows for the High Cascades than the Western Cascade watershed, both in terms of absolute volumes and normalized by drainage area. The Western Cascade stream, however, showed greater relative reductions in these summer streamflow metrics. Model results illustrate that differences between the responses of the two sites were primarily due to differences in groundwater flow, as manifested in drainage efficiency of the watersheds. Spatial differences in recharge characteristics and the timing of snow accumulation and melt were shown to be important, but secondary, in terms of explaining responses at the two sites. Further, they found that in the 2·8 °C warming scenario, August discharge of the McKenzie River at Clear Lake was reduced 23% from current mean values.

This study complements the previous model-based analysis through in-depth analysis of meteorological and streamflow records in order to more fully characterize climate–snowmelt–groundwater interactions for the Oregon Cascade region. To gain insight, analyses of discharge dynamics at the event, seasonal, inter-annual, and decadal time scales are presented utilizing time-series analyses and correlations between parameters. While dynamics are explored during all seasons, the focus is particularly on late summer streamflow because of its importance for water resources. Through these analyses, the primary and secondary controls on discharge variability are identified at each time scale. This will facilitate model development that appropriately factors in the influence of geologic setting and in order to better predict the effects of future climate variability and change.

STUDY WATERSHEDS

Study watersheds were selected on the basis of their location at rain–snow transitional elevations, availability of historical discharge and snow records, absence of stream regulation or diversion, and contrasting geology. The two study watersheds adjoin each other in the upper McKenzie River watershed on the west side of the Oregon Cascades (Figure 1). The US Geological Survey (USGS) gauge for the McKenzie River at Clear Lake is located at N 44° 22′, W 122° 0′. The watershed upstream from that point (the Clear Lake watershed) covers 239 km^2 and ranges in elevation from 918 to 2051 m, with a weighted average elevation of 1215 m. The USGS gauge at Smith River above Smith River Reservoir is located at N 44° 20′, W 122° 02′, with an area of 42 km^2 and elevations ranging from 803 to 1753 m; weighted average elevation is 1216 m. The Smith River is tributary to the McKenzie River 16 river km downstream of Clear Lake and 6 km downstream of the Smith River gauge. The McKenzie River provides water and electricity for Eugene, the second largest city in Oregon, supports a major recreational economy, and sustains superb salmonid habitat. The McKenzie River is a major tributary to the Willamette River, one of the largest rivers in the USA.

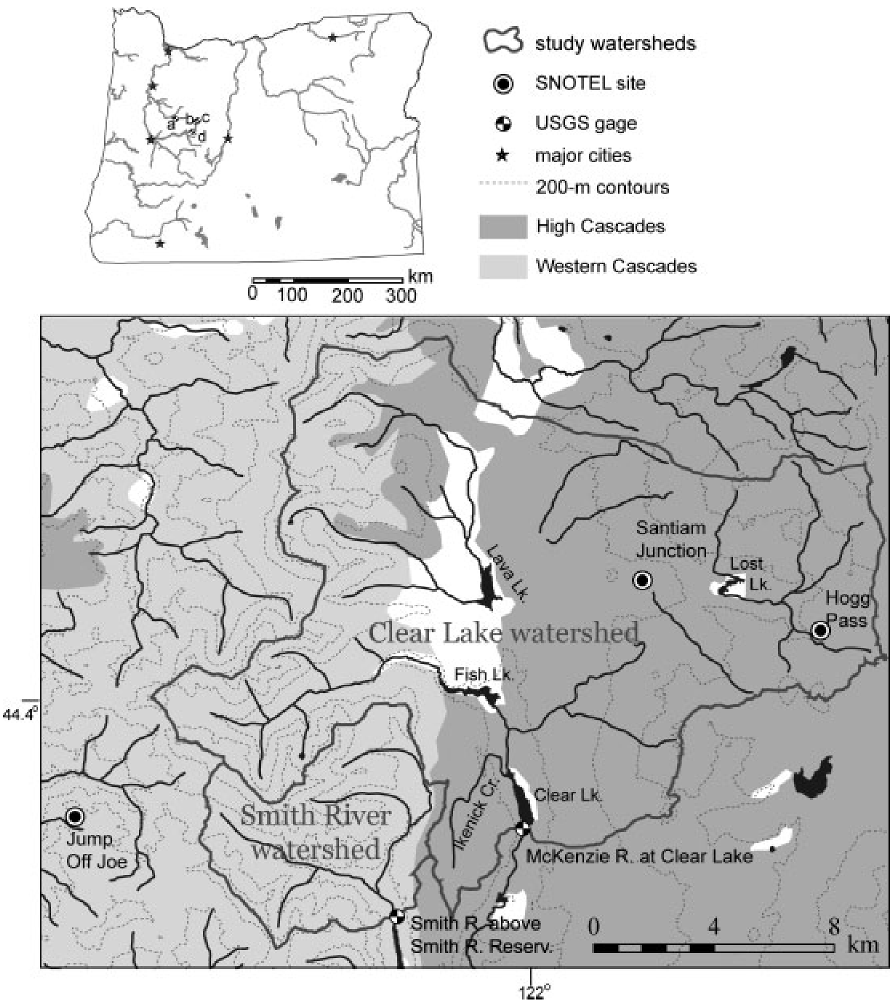

Figure 1. Map showing the study watersheds relative to their location in Oregon and to Cascades topography and geology. On the Oregon map, the letter 'a' next to a gauge symbol indicates the position of the stream gauge for South Santiam River below Cascadia, 'b' indicates Smith River above Smith River Reservoir, 'c' indicates McKenzie River at Clear Lake, and 'd' indicates McKenzie River at McKenzie Bridge

The study watersheds lie within the Cascades volcanic arc, which includes two distinct geologic provinces: the High Cascades and the Western Cascades. The High Cascades are known for their active composite volcanoes and extensive Quaternary basaltic lavas, while the Western Cascades are the products of Tertiary arc volcanism, and have been extensively folded, faulted, and weathered (Priest *et al.*, 1983). High Cascade lava flows have high permeability, resulting in extensive groundwater systems, low slopes, and low drainage densities (Jefferson *et al.*, 2006), whereas the Western Cascades are deeply dissected with well-developed drainage networks and shallow subsurface storm flow (Harr, 1977). We use the mapping of Sherrod and Smith (2000) to define the High Cascades as including all Quaternary igneous rocks and glacial till, while the Western Cascades includes all Tertiary igneous rocks. The Clear Lake watershed is 66% High Cascades, 25% Western Cascades, and 9% lakes and alluvium. The Smith River watershed is composed entirely of Western Cascades geology.

The study watersheds are largely forested, although cover is patchy on the youngest lava flows in the Clear Lake watershed. Forests include stands of Douglas fir (*Pseudotsuga menziesii*), noble fir (*Abies procera*), Pacific silver fir (*Abies amabilis*), and western hemlock (*Tsuga heterophylla*). Wet and dry meadows are also found throughout the watersheds. Land in the watersheds is primarily within the Willamette National Forest, although there are small in-holdings of private forestry land in the Clear Lake watershed. Forest harvest has occurred throughout the past 50 years in both watersheds, but less than 15% of either watershed was harvested or in an early regenerative stage in any year during the period. Both watersheds lie within the McKenzie River at McKenzie Bridge watershed, which has been deemed suitable for studying the effects of climate on water resources and is part of the Hydro-Climatic Data Network (Slack *et al.*, 1993).

Daily discharge data for the study watersheds are available from the USGS streamflow gauges. For the McKenzie River at Clear Lake, data are available for the 1913–1915 water years (water years begin October 1) and continuously since October 1947 (gauge 14 158 500) (Herrett *et al.*, 2005). Summer flow in the McKenzie River at Clear Lake consists largely of discharge from springs along the lakeshore. Additional wet season discharge includes direct precipitation onto the 0·6 km^2 lake, seasonal runoff from Fish Lake, and perennial runoff-dominated flow from Ikenick Creek (Figure 1). Based on occasional discharge measurements from August 2003 to September 2005, Ikenick Creek is a very minor contributor of summer flow to the McKenzie River at Clear, ranging from 3% on 6 June 2005 to 0·03% on 20 September 2005. Runoff from the creeks is probably much more significant during snowmelt periods. For Smith River (gauge 14 158 790), USGS daily discharge data are continuously available since October 1960 (Herrett *et al.*, 2005). Smith River is runoff-dominated with no lakes or mapped springs.

There are two Natural Resources Conservation Service snowpack telemetry (SNOTEL) sites within the Clear Lake watershed that collect real-time, automated measurements of snow water equivalent (SWE), precipitation, and temperature (http://www.wcc.nrcs.usda.gov/snow/). The Hogg Pass SNOTEL is located at 1451 m and Santiam Junction SNOTEL is at 1143 m. A third SNOTEL site, Jump Off Joe (1067 m), lies <5 km west of the study watersheds. This site was included in the analyses to provide additional information about low elevation areas. Approximately 20% of the Clear Lake watershed has an elevation higher than Hogg Pass, 63% is higher than Santiam Junction, and 83% is higher than Jump Off Joe. For Smith River, 12%, 64%, and 83% of the watershed area lies above the three SNOTEL stations, respectively.

Annual precipitation ranges from 1000 to 3200 mm, and 70% falls between November and March, based on the historical data from the SNOTEL stations. The watersheds receive a mixture of rain and snow at all elevations. At Hogg Pass, 56% of annual precipitation accumulates as snow, whereas at Santiam Junction it is 37%, and at Jump Off Joe it is 25%. From 400 to 1200 m elevation in the Cascades, snowpacks may accumulate and melt several times in the winter season and rain-on-snow events are common (Harr, 1981). Above this elevation, a seasonal snowpack of 2 to >7 m accumulates through the winter months and melts over the period March through June. In forested areas of the Cascades, up to 60% of falling snow can be intercepted by the forest canopy, but is later released by melt drip or falling snow masses. Sublimation accounts for only a minor component of ablation (Storck *et al.*, 2002). Peak snow accumulation occurs around 1 April at Hogg Pass, and around 1 March at Santiam Junction and Jump Off Joe. The highest monthly stream discharges at both Clear Lake and Smith River have historically occurred in May, due to snowmelt, but individual peak flows may be triggered by rain or rain-on-snow earlier in the water year.

DATA COMPILATION AND ANALYSIS

Daily time series datasets were compiled for discharge, precipitation, snow water equivalent, and temperature from the available USGS and SNOTEL records. Varying record lengths, as described below, were used for the analyses. Quality control on SNOTEL records was performed by the Oregon Climate Service (Daly, pers. commun., 2005). The Santiam Junction and Jump Off Joe SNOTEL sites precipitation and SWE records extend from water year 1979 through 2005, while for Hogg Pass precipitation and SWE records are from water years 1980 through 2005. Temperature records for Hogg Pass and Jump Off Joe span water years 1985–2003. For Santiam Junction, temperature records span 1983–2003. Reported minimum and maximum daily temperatures were averaged to generate a single daily value for temperature.

To understand the relationship between climatological and hydrological variables in space and time, Pearson's correlations, and auto- and cross-correlations were performed. Annual values of 49 parameters were derived from the discharge, precipitation, snow water equivalent, and temperature time series (Table I). Pearson's product moment correlations were calculated for 41 parameters, and significance was determined through F-tests and a 95% confidence limit. Daily values of discharge, precipitation, SWE, and recharge (rain + snowmelt) for 1979–2005 were normalized using daily means, and auto- and cross-correlations were computed following Box and Jenkins (1976). A similar analysis was conducted for annual time-series. Auto- and cross-correlation procedures generate correlation coefficients for data pairs separated by variable time lags. Auto-correlation is a measure of correlation between values in a single series, while cross-correlation is a measure of correlation between values in two different series. Confidence intervals for auto-correlation and cross-correlation coefficients were calculated using the effective number of observations following procedures outlined in Haan (2002) and Kite (1977). Auto- and cross-correlations were considered to be significant if they fell outside the 95% confidence interval around the null hypothesis of no correlation.

Monthly values of the Niño 3·4 Index of sea surface temperature (Trenberth and Stepaniak, 2001) and the Southern Oscillation Index of surface pressure (Ropelewski and Jones, 1987) over the period 1978–2004 were used to explore correlations between El Niño/Southern Oscillation (ENSO) with annual discharge and snow water equivalent. Discharge and SWE were also correlated with the Pacific Decadal Oscillation (PDO) data set from the University of Washington (Mantua *et al.*, 1997).

Simple water balances for the Clear Lake and Smith River watersheds were constructed for the 2001–2004 water years to examine the magnitudes and seasonal variations in water fluxes and stores. Discharge was calculated at each USGS gauge, and precipitation, SWE, and rain plus snowmelt at the areally weighted average elevation (1215 m) were derived by linear interpolation from values at Jump Off Joe and Hogg Pass. As a check on the appropriateness of linear interpolations of seasonal precipitation and rain plus melt, results were compared with observed values at Santiam Junction. Additionally, water

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Table I. Parameters derived from hydrological and climatic data. P indicates parameters for which Pearson's product moment correlations were calculated. R indicates parameters which were used in regression models of autumn minimum discharge. DOWY stands for day of water year, and DOY stands for day of calendar year

Parameter	Units	Clear Lake	Smith River	Hogg Pass	Santiam Jct.	Jump Off Joe
Water year	year	P	P	P	P	P
Mean discharge for water year	$m^3\ s^{-1}$	P	P			
Mean discharge for December–August	$m^3\ s^{-1}$	R	R			
Mean August discharge	$m^3\ s^{-1}$	P	P			
Autumn minimum discharge	$m^3\ s^{-1}$	PR	PR			
Date of autumn minimum discharge	DOY					
Temporal centroid	DOWY	P	P			
Mean discharge for previous December–August	$m^3\ s^{-1}$	R	R			
Autumn minimum discharge of previous year	DOY	R	R			
Total precipitation in water year	mm			PR	PR	PR
Accumulated precipitation on 1 April	mm			P	PR	P
Accumulated SWE	mm			PR	PR	PR
SWE present on April 1	mm			PR	PR	PR
Date of Peak SWE	DOWY			PR	PR	PR
Peak SWE	mm			P	P	P
Last date with SWE	DOWY			P	P	P
Temporal centroid of rain plus melt	DOWY			PR	PR	P
Average winter temperature (November–March)	°C			P	P	P
Average spring temperature (April–June)	°C			P	P	P
Average summer temperature (July–October)	°C			P	P	P

year 2004 precipitation was compared with spatially-distributed PRISM model outputs (Daly *et al.*, 1997, 2002). Basin-averaged evapotranspiration (ET) was calculated in RHESSys, a physically-based hydro-ecological model previously calibrated for the Clear Lake watershed (Tague *et al.*, 2008). ET estimates in RHESSys account for canopy transpiration, evaporation of canopy interception, litter/soil evaporation and snow sublimation losses. ET losses are calculated using Penman, for evaporative losses, and Penman–Monteith, for transpiration losses. Additional details on the RHESSys modeling approach are available in Tague and Band (2004). Watershed-averaged lake evaporation was estimated by multiplying the lake-covered area in the Clear Lake watershed (<2 km^2) by the evaporation rate of Crater Lake (1200 mm yr^{-1}; Redmond, 1990), located at 1883 m elevation in the Oregon Cascades. The resulting rate was <10 mm yr^{-1} for the Clear Lake watershed, and the Smith River watershed has no lakes, so lake evaporation was neglected in water budget calculations.

Predictive models of September–November minimum discharge using available climatic and hydrologic measurements were developed using stepwise regression in SAS 9·1 (Table I). Separate models were developed for the Clear Lake and Smith River watersheds. The models were based on data from the 1979–2005 water years, and stepwise selection of parameters proceeded until all variables in the model were significant at the 0·05 level and no other variables met the 0·05 significance level for entry into the model. Colinearity was minimized by manual elimination of strongly correlated variables.

Discharge records were investigated for secular trends in August discharge, minimum discharge, and temporal centroid. Less than 50 year record lengths for Clear Lake and Smith River watersheds may lead to ambiguous conclusions about the presence or absence of trends. To support these records, discharge data from two watersheds with longer historical records were also examined. The McKenzie River at McKenzie Bridge (USGS gauge 14 159 000) drains 901 km^2, including the Clear Lake and Smith River watersheds. It has a weighted average elevation of 1286 m, and a gauge elevation of 432 m. It is 56% High Cascade geology, and has a daily discharge record that extends from August 1910 to October 1995. A hydroelectric project has been in operation upstream of the gauge since 1963, but the reservoirs are not operated for flood control or low-flow supplementation. The South Santiam River below Cascadia (USGS gauge 14 185 000) drains 451 km^2, immediately to the west of the study watersheds. It has a weighted average elevation of 875 m, and a gauge elevation of 236 m. The watershed is 5% High Cascade geology, and it has daily discharge data from September 1935 to the present. Both watersheds are part of the Hydro-Climatic Data Network, a set of discharge records that have been deemed suitable for studying the effects of climate on water resources (Slack *et al.*, 1993).

EVENT AND SEASONAL SCALE VARIABILITY

The groundwater-dominated Clear Lake watershed exhibits a muted hydrograph response relative to the adjacent Western Cascade, runoff-dominated Smith River watershed (Figure 2). Base flows are higher, and peak flows are less frequent, smaller, and somewhat delayed relative to Smith River. Peak flows are generated by rain-on-snow, snowmelt, and rain events, with rain-on-snow events typically having sharp, isolated peaks. For example, on 7–9 January 2002 (event A in Figure 2), 53 mm

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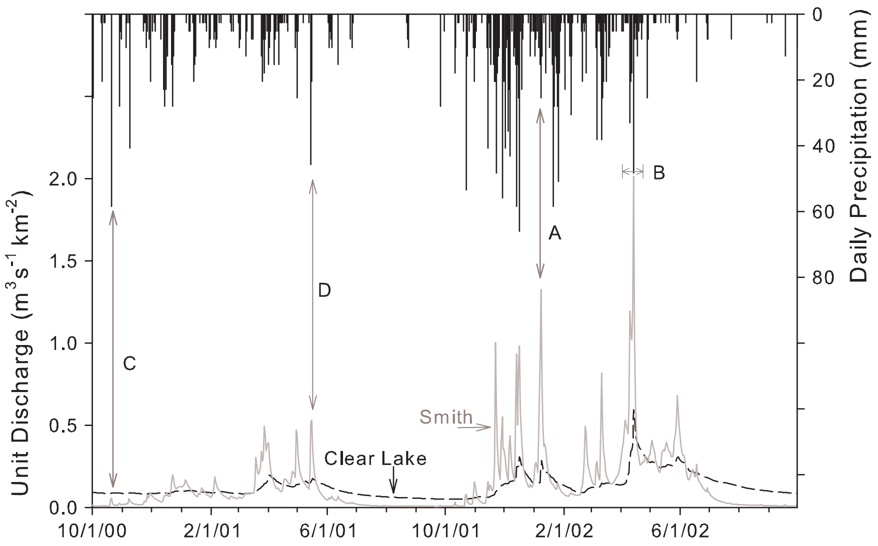

Figure 2. Water years 2001–2002 Santiam Junction precipitation and unit discharge hydrographs for the McKenzie River at the outlet of Clear Lake and Smith River above Smith River Reservoir. Event A corresponds to 7–9 January 2002; event B corresponds to 1–23 April 2002; event C corresponds to 21 October 2000; and event D corresponds to 15–16 May 2001. Events are discussed in the text

of precipitation fell at Santiam Junction yielding an additional 44 mm of melt. The McKenzie River at Clear Lake had a peak flow twice the pre-event flow, while Smith River peak flow was six times greater than before the event. Snowmelt events are generally sustained, building gradually to their peak flows, often with intermediate peaks. For example, between 1 and 23 April 2002 (event B), 188 mm of precipitation fell at Santiam Junction and 564 mm of snow melted. The peak flow of the McKenzie River at Clear Lake was four times greater than the pre-event discharge, while at Smith River it was eight times greater. Flows twice as large as the pre-event discharge were sustained for 14 days in the McKenzie River at Clear Lake and 6 days in the Smith River. The magnitude of response to rain events depends largely on the antecedent wetness in the watershed. For example, ~60 mm rain events in 21 October 2000 and 15–16 May 2001 (events C and D) produced very different hydrographs.

Annual unit discharge of the McKenzie River at Clear Lake and the Smith River are comparable, but the September unit discharge at Clear Lake is over six times higher than at Smith River, as illustrated in Figure 2. The ratio of May to September monthly discharge at Clear Lake is only 3·3, while for the Smith River it is 39·8. Tague and Grant (2004) provide an extended discussion of the low flow characteristics of High and Western Cascade streams and attribute the slow summer recession of High Cascade streams to their groundwater inputs.

Clear Lake and Smith River have very different auto-correlation structures (Figure 3a). The auto-correlation coefficient for Clear Lake is statistically significant at the 95% confidence level for 301 of the first 366 daily lags, whereas at Smith River statistically significant auto-correlations are observed for only 36 of the first 366 days, and recharge time series derived from rain and snowmelt at Hogg Pass has statistically significant correlations for only 22 days. These results suggest that both runoff-dominated and groundwater-fed stream systems have damped discharge variation relative to the input variation as estimated from a single SNOTEL site, reflecting the distribution of response times of shallow subsurface and groundwater flow and in the channel network. The prevalence of longer response times associated with groundwater systems of the Clear Lake watershed result in substantial smoothing of the hydrograph and sustained high auto-correlation values, relative to the faster shallow subsurface flow, runoff-dominated Smith River.

The maximum cross-correlation between discharge of Clear Lake and Smith River occurs at a one-day lag (Figure 3b), as do the peak cross-correlations of Clear Lake discharge and recharge at Santiam Junction and Jump Off Joe. These results suggest that the Clear Lake watershed experiences a minimal lag in discharge relative to the recharge time series or a runoff-dominated stream. This result is in stark contrast to the results of Manga (1999), who found a lag of 47–137 days for four snowmelt-dominated spring-fed streams on the east side of the Oregon Cascades compared to neighbouring runoff-dominated streams. Tague and Grant (2004) report lags averaging 30 days in their analysis of Western versus High Cascade streams. However, their analysis did not normalize the datasets to remove seasonal effects. Later work by Tague *et al*. (2008) show that discharge from Clear Lake includes both spring and shallow subsurface flow contributions. Thus, short lag times reflect the ‘fast’ shallow subsurface component of the Clear Lake system.

WATER BUDGET

Water budgets, constructed as described earlier, reveal divergent seasonal patterns of water storage and flux for the runoff- and groundwater-dominated watersheds. Over the 2001–2004 period, precipitation equalled 99%

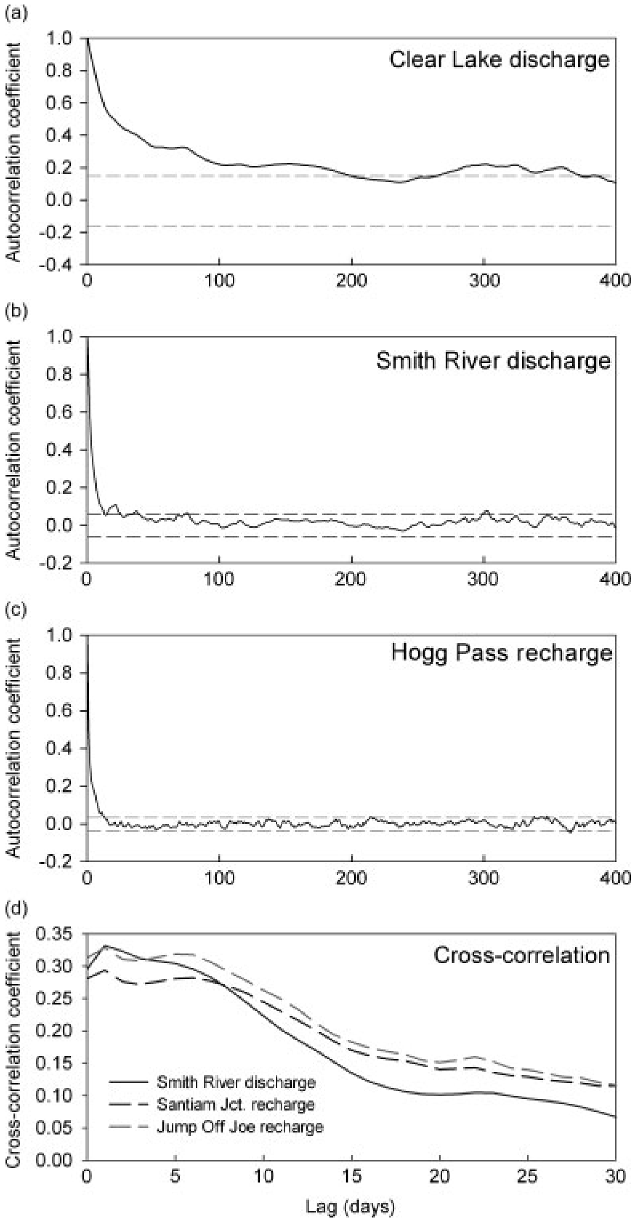

Figure 3. Auto- and cross-correlations of time-series data. Autocorrelation time series for (a) Clear Lake discharge, (b) Smith River discharge, and (c) Hogg Pass recharge (rain plus snowmelt). Correlations are considered to be significant if they fell outside the 95% confidence interval (dashed lines) around the null hypothesis of no correlation. (d) Cross-correlation time series for Clear Lake discharge versus Smith River discharge, Santiam Junction recharge, and Jump Off Joe recharge. All cross-correlations shown on the graph are outside the 95% confidence intervals of 0·0551 for Smith River, 0·0303 for Santiam Junction, and 0·0292 for Jump Off Joe

of discharge plus ET in the Clear Lake watershed and 80% of discharge plus ET in the Smith River watershed (Figure 4). In water year 2004, linear interpolation estimated precipitation as 2·14 m, and PRISM precipitation for the Clear Lake watershed is 2·11 m. In the Smith River watershed, PRISM precipitation was 2·49 m for 2004, 18% higher than in the Clear Lake watershed. This result suggests that the linear interpolation method correctly estimated precipitation in the Clear Lake watershed and underestimated precipitation by ~20% in the Smith River watershed. This difference accounts for the closure of the annual water budget in the Clear Lake watershed and the discrepancy in the Smith River watershed. The interpolations based on linear regression from Hogg Pass and Jump Off Joe over-predict the Santiam Junction record. This was expected because Santiam Junction lies in the rain shadow of the Western Cascades and often becomes snow-free earlier than Jump Off Joe.

The year was divided into three seasons based on the status of the snowpack. November through March constitutes the snow accumulation season; April through June is the snow ablation season; and July through October is the snow-free season. On average for 2001–2004, in the snow accumulation season, 35% of precipitation was stored as seasonal snowpack (450 mm), which was melted during the snow ablation season. Thus, despite the accumulation season having 350% of the precipitation of the ablation season, available water amounts were nearly equal in the two periods. Between 2001 and 2005, snowpack storage had varying importance for runoff, depending on the size of the snowpack, timing of the melt, and the amount of spring rain. In 2002, snowmelt was 137% of April through June Clear Lake runoff, but in 2005, it was only 29% owing to a warm dry winter combined with a wet spring.

For the Clear Lake watershed, during the 2001–2004 period, 71% of available water during the snow accumulation season was accounted for as runoff, and 5% was lost to ET (Figure 4). The remaining 24% (~200 mm), probably replenished the soil moisture supply and groundwater storage. During the snow ablation seasons in the Clear Lake watershed, snowmelt and rain supplied enough water to account for runoff and ET and provided ~125 mm to groundwater storage. In the snow-free season in the Clear Lake watershed, runoff and ET were approximately equal and accounted for 260% of precipitation. Groundwater storage diminished to sustain streamflow and supply water to vegetation.

For the Smith River watershed, available water during all seasons was increased by 20% over the Clear Lake watershed, as suggested by the PRISM outputs (Figure 4). During the 2001–2004 snow accumulation seasons, all available water was used for runoff and ET, with no groundwater recharge occurring. In the snow ablation seasons, ~160 mm of water was available to recharge groundwater and soil moisture, after runoff and ET were accounted for. In the snow-free seasons, runoff accounted for ~20% of precipitation and ET required 130%. The ET flux was six times as great as the runoff flux, suggesting that recharge during the snow ablation season was largely to the plant-available soil zone rather than to bedrock groundwater.

INTER-ANNUAL VARIABILITY

Generally, discharge, precipitation and SWE vary together throughout the period of record and among the three SNOTEL sites (Table II). Mean annual discharges at Clear Lake and Smith River are strongly correlated

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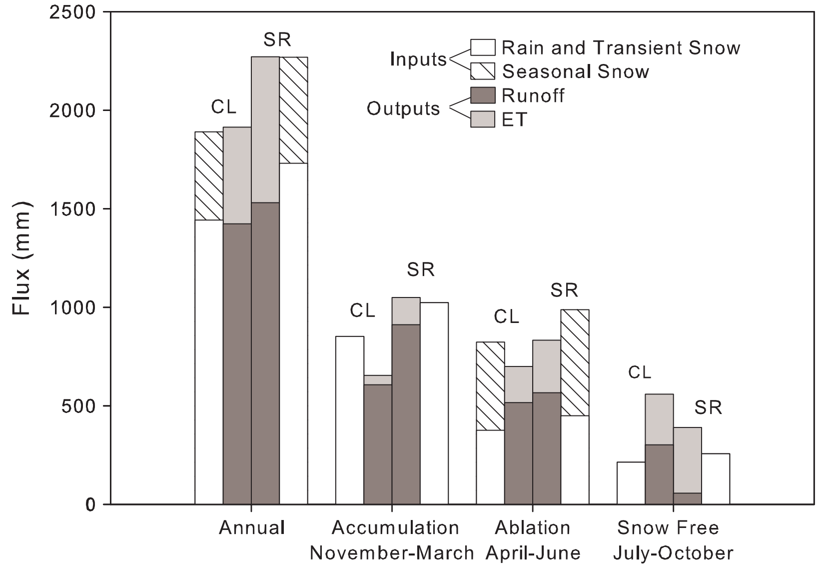

Figure 4. Water budget for Clear Lake (CL) and Smith River (SR) watersheds for the average of water years 2001–2004. The rain plus transient snow category includes all liquid precipitation and snow that falls and ablates within the same season. Seasonal snow falls and is stored during the accumulation season, and it melts and contributes to runoff during the ablation season. Inputs for SR are 120% of inputs for CL

with annual precipitation at all three sites ($r > 0{\cdot}88$), but only at Hogg Pass is there strong correlation between accumulated snow water equivalent (SWE) and mean discharge ($r = 0{\cdot}74$ for Clear Lake). The two watersheds show similar levels of variability in mean annual discharge, with the coefficient of variation (CV) equal to 0·25 for Clear Lake and 0·27 for Smith River. However, variability in August discharge is smaller for Clear Lake (CV = 0·32) than for Smith River (CV = 0·36). August discharge at Clear Lake is more strongly correlated with precipitation and SWE parameters than is August discharge at Smith River, which is more strongly tied to winter and spring temperatures. This suggests that summer streamflow in the groundwater-fed watershed is more strongly linked to hydroclimatic conditions than summer streamflow in the runoff-dominated watershed, where temperature-modulated ET demand may have more influence. Winter temperature was most strongly correlated with accumulated SWE at Jump Off Joe ($r = -0{\cdot}71$), which lies within the transient snow zone and is strongly affected by temperature-induced phase changes in precipitation. The amount of peak SWE and the date at which it is reached are only weakly correlated for any individual site ($r < 0{\cdot}55$).

There is an inter-annual memory in the Clear Lake watershed as a result of the High Cascades groundwater system. Discharge at Clear Lake is moderately cross-correlated with the previous year's precipitation ($r = 0{\cdot}52$ for Hogg Pass), and the cross-correlation is higher than the 1-year discharge auto-correlation ($r = 0{\cdot}45$). Both of the above correlations for Clear Lake discharge are higher than those for Smith River discharge with that of the previous year or the 1-year lagged auto-correlation of precipitation, which are not statistically significant. At 2 year lags and beyond all auto- and cross-correlations are not statistically significant.

The relationships between two major climate indices that have been correlated with streamflow in the Pacific Northwest were examined and quantified: El Niño/Southern Oscillation Index and the Pacific Decadal Oscillation (Beebee and Manga, 2004). Correlations between SWE and the monthly Niño 3·4 index for 1977–2000 are highest for December, and are stronger for the lower elevation SNOTEL sites. Hogg Pass SWE had a correlation of −0·54 with December Niño 3·4, while the correlation at Santiam Junction is −0·60, and that at Jump Off Joe is −0·63. These finding suggest that the temperature shifts associated with the ENSO cycle are more important than the precipitation signal in determining snow accumulation in the transient snow zone. There are no significant correlations with annual or August discharge in either watershed. These results indicate that ENSO, as explained by the Niño 3·4 index, is a moderate predictor of SWE, but not of discharge. Correlations were also explored using the SOI but it was found that the correlations are much lower than for the Niño 3·4 index. It was found that there are no significant correlations with annual discharge, August discharge or station SWE.

These results contrast with those of Beebee and Manga (2004), who found inverse correlations between annual discharge, peak runoff and ENSO indices for eight snowmelt-dominated watersheds in central and eastern Oregon. However, the current study area has a higher proportion of rain versus snow than the watersheds they studied, and it is also substantially more humid. They also used the SOI averaged over June–September and Niño 3·4 averaged over September–November. Individual monthly values of the indices were used here rather than seasonal aggregates to better understand the time lags between discharge and the climate indices. Lower correlations were found for seasonally averaged values of both SOI and Niño 3·4. Like Beebee and Manga [2004], no significant correlation with PDO was found.

AUTUMN MINIMUM DISCHARGE

The presence of groundwater also strongly influences the lowest flows of the year.

Annual minimum discharge occurred during September through November during 52 of 64 years of the Clear Lake record and 41 of 45 years of the Smith river record. In the remaining years, annual minimum discharge occurred later in the winter, probably as a result of snowy or dry autumns that didn't recharge the stream and prolonged the recession. For consistency of analyses, the minimum daily discharge between 1 September and 30 November was determined, henceforth called autumn minimum discharge (Q_{min}). On a unit area basis, Q_{min} is ~9 times greater in the Clear Lake watershed than the Smith River watershed. As discussed above, the greater Q_{min} in the groundwater-dominated watershed is the result of a slower summer recession. Clear Lake Q_{min} are also more strongly influenced by inter-annual fluctuations in precipitation (Figure 5). The CV for Clear Lake Q_{min} is 0·26, while for Smith River it is 0·22. It is proposed that subsurface storage is nearly exhausted in the Smith River watershed every year, whereas Q_{min} in the Clear Lake watershed are sensitive to aquifer levels, which in turn are controlled by the current and previous year's precipitation inputs. Using the previous year's December–August mean discharge at Clear Lake as a proxy for precipitation, the Pearson correlation with Clear Lake Q_{min} is 0·5.

Stepwise multiple linear regression was performed on the series of 1979–2005 Q_{min} for Clear Lake and Smith River, using 18 potential predictor variables for each watershed (Table I). Stepwise multiple linear regression for Smith River yielded a model using the accumulated snow water equivalent at Santiam Junction (S_{SJ}, mm) that explains 38·1% of the variation in the dataset. No other variables met the 0·05 significance level for entry into the model. The low association between Smith River Q_{min} and hydroclimatic variables supports the contention that subsurface storage in the watershed is nearly exhausted each year, and thus, insensitive to climatic variability.

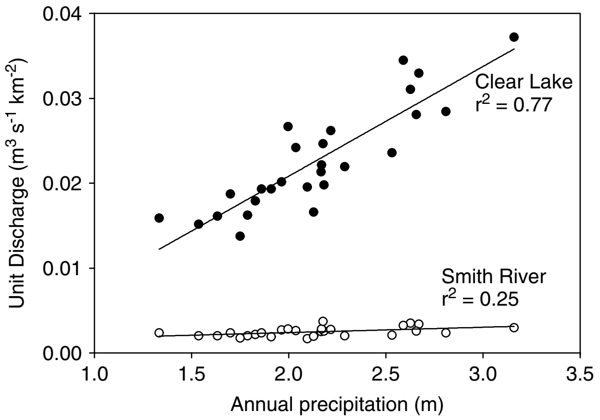

Figure 5. Relationship between annual precipitation at Santiam Junction and autumn minimum unit discharge for the Clear Lake and Smith River watersheds

In contrast to the results for Smith River, the best model for Clear Lake Q_{min} uses four variables to explain 93·4% of the variation in the dataset. The predictive variables are mean discharge from December–August (Q_{Mean}, m^3 s^{-1}), the previous year's autumn minimum discharge ($Q_{PrevMin}$, m^3 s^{-1}), S_{SJ}, and the day of the calendar year on which autumn minimum discharge occurs (A). The regression equation is:

$$Q_{min} = 0{\cdot}288 \times Q_{mean} + 0{\cdot}00175 \times S_{SJ} + 0{\cdot}292 \times Q_{PrevMin} - 0{\cdot}0197 \times A + 4{\cdot}73 \quad (1)$$

The high correlation between precipitation and mean discharge suggests that the discharge term represents the year's precipitation. The snow term is probably related to the timing of snowmelt and recharge to the aquifers. The previous minimum discharge term represents the inter-annual memory in the groundwater system and the state of the aquifer in the previous year. The autumn minimum flow date term represents the timing of the first major autumn rains that bring an end to the summer recession. Thus, Q_{min} is shown to be a function of subsurface storage, the year's precipitation, timing of snowmelt, and timing of the fall rains. Even when A is not included, the other three variables explain 90% of the variation in the historical record, enabling Q_{min} to be predicted by the end of August, using

$$Q_{min} = 0{\cdot}312 \times Q_{mean} + 0{\cdot}00174 \times S_{SJ} + 0{\cdot}260 \times Q_{PrevMin} - 1{\cdot}28. \quad (2)$$

SECULAR TRENDS

Investigation of secular trends in climatic and hydrograph parameters suggests that inter-annual variability masks potential trends in precipitation and SWE data derived from the <30 year SNOTEL record. The only SNOTEL-derived snow or precipitation parameter to exhibit a statistically significant relationship with time is the date of last snow cover at Santiam Junction. There is a weak ($r = -0{\cdot}39$) trend towards earlier loss of snow cover at this site, with a slope of 7·8 days per decade. There are no statistically significant trends at either the higher or lower elevation sites. Santiam Junction may be particularly susceptible to warming-induced earlier snow cover loss, because it lies near the lower limits of seasonal snow cover. Future warming may result in much more transient snow cover interspersed with winter rain.

Longer discharge records suggest that there are some secular trends, although the signal is still dominated by inter-annual variability. The hydrograph temporal centroid is the day of the water year when half of the annual discharge has occurred, and has been used as an indicator of climate change throughout the mountainous western USA (Regonda *et al.*, 2005). It was found that the

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Table II. Correlation coefficients for selected variables. Bold terms are statistically significant at the 95% confidence level. Mean, August, and autumn min. are daily discharges. CT is the temporal center of the hydrograph. SWE and Precip. are cumulative values for the water year. Peak day is the day of peak SWE accumulation, and Peak SWE is the amount. Winter temperature is average daily median temperature for November through March, and spring temperature is the average daily median temperature for April to June

		Clear Lake				Smith River			
		WY Mean	August	Autumn Min.	CT	WY Mean	August	Autumn Min.	CT
Clear Lake	Mean	**1·00**							
	August	**0·88**	**1·00**						
	Autumn Min.	**0·93**	**0·92**	**1·00**					
	CT	−0·10	0·28	0·04	**1·00**				
Smith River	Mean	**0·94**	**0·86**	**0·84**	−0·04	**1·00**			
	August	**0·50**	**0·76**	**0·59**	**0·43**	**0·86**	**1·00**		
	Autumn Min.	**0·56**	**0·74**	**0·74**	**0·41**	**0·49**	**0·62**	**1·00**	
	CT	−0·26	0·05	−0·14	**0·89**	−0·23	0·23	0·21	**1·00**
Hogg Pass	SWE	**0·74**	**0·77**	**0·69**	0·25	**0·80**	**0·44**	**0·47**	0·00
	Precip.	**0·91**	**0·75**	**0·81**	−0·19	**0·95**	**0·44**	**0·40**	**−0·34**
	Peak Day	0·14	0·25	0·17	0·15	0·21	**0·39**	0·11	0·21
	Peak SWE	**0·67**	**0·79**	**0·68**	**0·39**	**0·74**	**0·53**	**0·56**	0·11
	Winter T	−0·08	−0·33	−0·20	**−0·60**	−0·18	**−0·51**	**−0·43**	**−0·50**
	Spring T	−0·33	**−0·52**	**−0·47**	**−0·45**	−0·38	**−0·55**	**−0·52**	−0·34
Santiam Junction	SWE	**0·57**	**0·76**	**0·68**	**0·48**	**0·61**	**0·58**	**0·62**	0·27
	Precip.	**0·88**	**0·82**	**0·85**	−0·01	**0·95**	**0·55**	**0·51**	−0·22
	Peak Day	0·21	**0·34**	0·28	0·23	0·24	0·18	**0·43**	0·05
	Peak SWE	**0·51**	**0·73**	**0·60**	**0·51**	**0·57**	**0·60**	**0·60**	0·30
	Winter T	−0·09	−0·32	−0·23	**−0·58**	−0·20	**−0·40**	−0·36	**−0·52**
	Spring T	**−0·46**	**−0·64**	**−0·60**	−0·35	**−0·47**	**−0·56**	**−0·67**	−0·27
Jump Off Joe	SWE	**0·36**	**0·55**	**0·44**	**0·64**	**0·38**	**0·36**	**0·48**	**0·49**
	Precip.	**0·93**	**0·82**	**0·88**	−0·10	**0·93**	**0·51**	**0·50**	−0·26
	Peak Day	0·14	0·29	0·26	0·28	0·05	0·23	**0·36**	0·26
	Peak SWE	**0·34**	**0·54**	**0·40**	**0·69**	**0·39**	**0·43**	**0·50**	**0·53**
	Winter T	−0·11	−0·34	−0·26	**−0·55**	−0·14	−0·32	**−0·42**	**−0·53**
	Spring T	−0·15	−0·30	−0·26	−0·36	−0·23	−0·35	−0·31	−0·34

Table II. (*Continued*)

		Hogg Pass						Santiam Junction						Jump Off Joe					
		SWE	Precip.	Peak Day	Peak SWE	Winter T	Spring T	SWE	Precip.	Peak Day	Peak SWE	Winter T	Spring T	SWE	Precip.	Peak Day	Peak SWE	Winter T	Spring T
Hogg Pass	SWE	**1·00**																	
	Precip.	**0·74**	**1·00**																
	Peak Day	−0·06	0·13	**1·00**															
	Peak SWE	**0·92**	**0·66**	0·05	**1·00**														
	Winter T	−0·34	−0·14	**−0·43**	**−0·61**	**1·00**													
	Spring T	−0·40	−0·35	**−0·62**	**−0·44**	**0·46**	1·00												
Santiam Junction	SWE	**0·81**	**0·53**	0·02	**0·90**	**−0·59**	**−0·42**	**1·00**											
	Precip.	**0·77**	**0·95**	0·13	**0·75**	−0·27	**−0·39**	**0·68**	**1·00**										
	Peak Day	**0·33**	0·18	0·07	**0·55**	**−0·44**	0·05	**0·45**	0·24	**1·00**									
	Peak SWE	**0·75**	**0·48**	0·05	**0·91**	**−0·74**	−0·35	**0·94**	**0·64**	**0·55**	**1·00**								
	Winter T	−0·34	−0·17	−0·35	**−0·60**	**0·95**	**0·55**	**−0·60**	−0·33	**−0·41**	**−0·73**	**1·00**							
	Spring T	**−0·38**	**−0·41**	**−0·64**	**−0·42**	0·31	**0·94**	**−0·38**	**−0·45**	−0·06	−0·26	**0·38**	**1·00**						
Jump Off Joe	SWE	**0·67**	**0·33**	−0·05	**0·75**	**−0·59**	−0·37	**0·88**	**0·48**	0·29	**0·80**	**−0·59**	−0·28	**1·00**					
	Precip.	**0·69**	**0·95**	0·14	**0·64**	−0·16	−0·36	**0·57**	**0·97**	0·15	**0·51**	−0·21	**−0·45**	**0·39**	**1·00**				
	Peak Day	0·06	0·05	0·24	0·25	−0·30	−0·15	**0·35**	0·06	**0·43**	**0·32**	−0·21	−0·27	0·29	0·02	**1·00**			
	Peak SWE	**0·68**	**0·33**	−0·07	**0·77**	**−0·67**	**−0·39**	**0·86**	**0·47**	0·29	**0·84**	**−0·64**	−0·23	**0·96**	**0·36**	0·27	**1·00**		
	Winter T	−0·27	−0·08	**−0·42**	**−0·50**	**0·81**	**0·40**	**−0·65**	−0·23	**−0·48**	**−0·69**	**0·80**	0·31	**−0·71**	−0·13	**−0·44**	**−0·71**	**1·00**	
	Spring T	−0·23	−0·22	**−0·61**	−0·24	**0·41**	**0·93**	−0·25	−0·25	0·10	−0·19	**0·54**	**0·85**	−0·26	−0·21	−0·01	−0·29	**0·44**	**1·00**

temporal centroid is significantly correlated with temperature and SWE-derived variables, but not precipitation. Thus, early dates of the temporal centroid are indicative of warm years with winter rain or early snowmelt, regardless of the total precipitation or discharge. The regression between temporal centroid and water year for the Clear Lake discharge record is statistically significant ($P = 0{\cdot}06$), revealing that the temporal centroid moved earlier by 17 days between 1948 and 2007. This is consistent with the findings of Stewart *et al.* (2005), who reported temporal centroids moving earlier by 1–4 weeks between 1948 and 2002 in snowmelt-dominated basins of western North America. The regression between temporal centroid and water year is not statistically significant for Smith River ($P = 0{\cdot}85$), possibly reflecting the shorter record length in that watershed.

Warmer winters and earlier spring snowmelt also seem to be affecting low flows from groundwater-dominated watersheds disproportionately relative to runoff-dominated watersheds. In the Clear Lake watershed, August discharges, normalized by mean water year discharges, (Q_{Aug}/Q_{wy}) have experienced statistically significant declines, as have Q_{min}/Q_{mean} ratios (Figure 6). Q_{Aug}/Q_{wy} declined 15% from 1948 to 2005, and Q_{min}/Q_{mean} has declined 11% in the same period. In contrast, Q_{Aug}/Q_{wy} and Q_{min}/Q_{mean} ratios lack statistically significant trends in the Smith River watershed. Longer record watersheds mirror the results of the Clear Lake and Smith River watersheds: the groundwater-dominated McKenzie Bridge watershed shows significant declines, whereas the runoff-dominated South Santiam does not. These parallel results suggest that the trends are not a function of the period of record, but reflect the adjustment of flow regimes to changing climatic conditions.

In groundwater-dominated watersheds, summer and autumn low flows are decreasing as snowmelt occurs earlier, the slow, summer hydrograph recession begins sooner, and it occurs over a longer period of time. As the start date to the summer recession becomes earlier, discharge for a specified interval of the recession period (e.g. August discharge) decreases because the interval is later on the recession limb. In the same manner, with a longer recession, an ultimately lower endpoint (i.e. autumn minimum discharge) is reached. In runoff-dominated watersheds, the fast recession leaves summer and autumn low flows insensitive to climatic variability because subsurface storage is nearly exhausted early each summer.

DISCUSSION

The different behaviours of groundwater- and runoff-dominated watersheds on event, seasonal, and inter-annual scales are clearly revealed by the analysis of historical records in the Clear Lake and Smith River watersheds. These contrasting behaviours allow one to identify the primary and secondary controls on watershed response at each time scale (Table III). These different behaviours have implications for how the watersheds have responded to warming winters in the past 45 years and how they might respond to future climate warming.

At the event scale, discharge is controlled primarily by the magnitude and intensity of the rain or snowmelt and secondarily by the geology of the watershed. The recharge signal is spatially and temporally varying, because of topographic effects on both precipitation and temperature. Particularly important to determining discharge is the location of the transition between snow and rain during a precipitation event and the average temperature during a melt event. The bedrock hydraulic conductivity determines whether most recharge is quickly routed to streams, resulting in a flashy hydrograph, such as that of Smith River, or whether most recharge reaches a deeper aquifer where the signal becomes damped and lagged, as in the Clear Lake watershed.

Table III. Controls on discharge variability for Cascades watersheds

Scale	Primary Control	Secondary Control
Event	Precipitation	Geology
Seasonal	Geology	Winter–spring temperature
Annual	Precipitation	Geology (minor)

At the seasonal scale, variations in discharge are primarily controlled by the watershed geology, and secondarily by winter and spring temperature. For aquifers with seasonal recharge, damping of the hydrograph is a function of aquifer length, storativity, transmissivity and the periodicity of the recharge (Townley, 1995). The aquifer characteristics of the Clear Lake watershed produce significant damping of the discharge time series relative to the recharge time series. This damping results in the sustained release of groundwater during the summer drought,

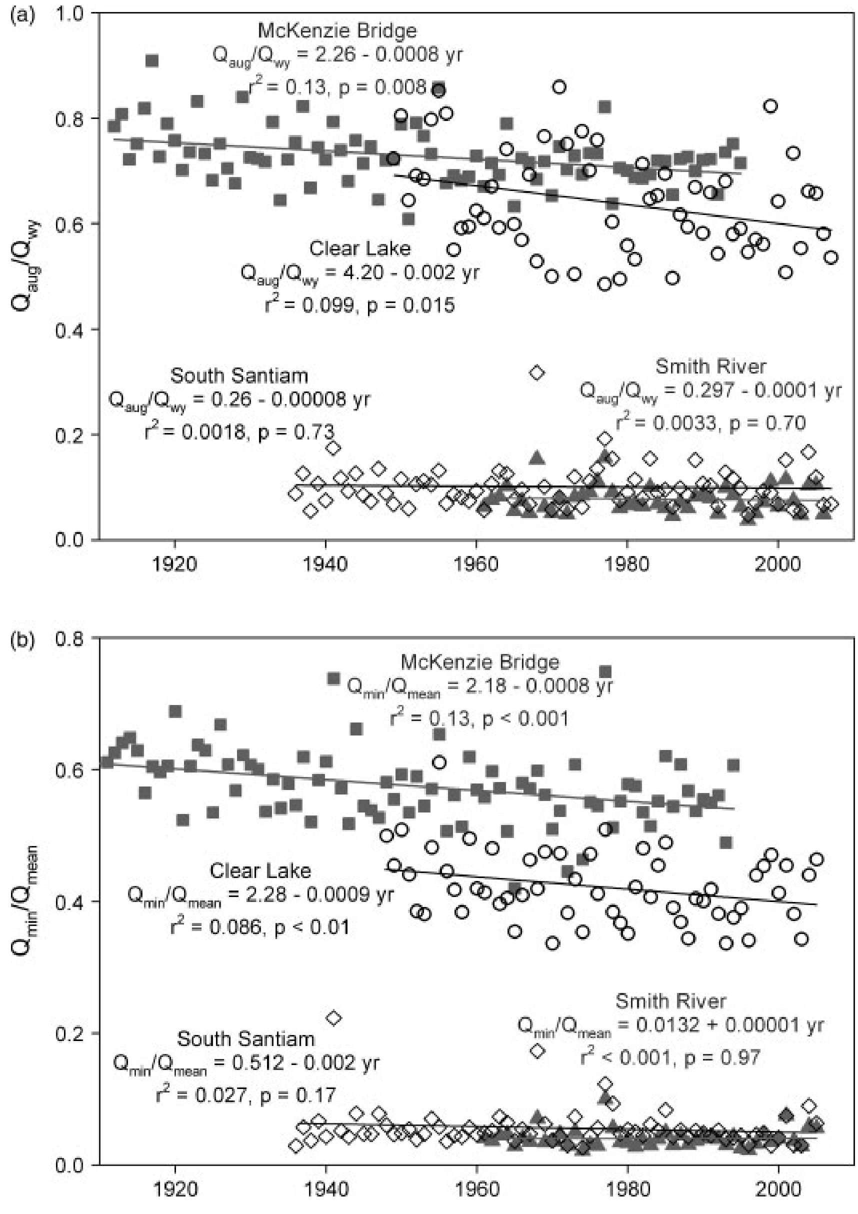

Figure 6. Relationship between low flow parameters and calendar year: (a) ratio of August mean discharge to water year mean discharge (*Qaug*/*Qwy*) and (b) ratio of autumn minimum discharge to the average discharge of the preceding nine months (*Qmin*/*Qmean*)

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even though the seasonal water balance is negative. In contrast, the shallow subsurface flow system of the Smith River watershed rapidly transmits pressure waves from recharge to produce changes in streamflow (Torres *et al.*, 1998). This results in a flashy winter hydrograph and rapid summer recession.

In the Clear Lake watershed, the streamflow seasonality, as measured by the ratio of discharge in the highest flow month to that in September, is related to the partitioning of precipitation between rain and snow and the timing of the snowmelt. Years with higher snow to rain ratios have lower winter flows and relatively more recharge in the spring, resulting in more water to sustain summer discharge. Years with colder spring temperatures have delayed snowmelt, resulting in a later onset of the summer recession and higher discharge during the autumn minimum flow period. Both of these scenarios result in less month-to-month variability in discharge.

At the annual scale, mean discharge is controlled by the current year's precipitation. Although it has been shown that the previous year's precipitation has some influence on mean discharge at Clear Lake, the inter-annual variability in discharge is similar in the two watersheds. This suggests that the groundwater system does not have a significant moderating effect on overall inter-annual flow variability. Drought and high water years show up with approximately the same frequency and severity in both groundwater- and runoff-dominated watersheds. However, during the summer recession, groundwater can have a more important moderating effect on inter-annual variability. Conversely, at the tail end of the summer recession groundwater increases the effects of inter-annual variability. This corresponds with the hypothesis that subsurface storage is nearly exhausted every year in the runoff-dominated watershed, resulting in less inter-annual variability.

Based on the patterns of the past several decades, one can begin to forecast what the next few decades will likely bring to streams in the Oregon Cascades. The correlations between accumulated SWE and winter temperature and ENSO indices, combined with the long-term decrease in the snow-covered season at Santiam Junction, suggest that there is very strong elevation sensitivity in SWE parameters. In particular, it seems that areas at the transition between transient and seasonal snowpacks are most sensitive to warmer winters. If winters continue to warm, as suggested by regional climate models (Christensen *et al.*, 2007), it is projected that the elevation of the transient snow zone will shift upward and substantial portions of the previously snow-dominated Cascades will experience a higher frequency of winter rain events. For these watersheds that in the future have more transient and less seasonal snow, winter discharge will increase and the snowmelt peak will become less important. Thus, watersheds in the Oregon Cascades most sensitive to warming will be those with much of their area between 1100 and 1300 m. For watersheds already dominated by transient snow (<1100 m), the proportion of winter rain will increase, but the hydrograph will likely be less affected than for higher elevation watersheds. For the highest elevation watersheds (>1500 m), seasonal snow cover will persist but the snowmelt peak will occur earlier in the spring. In the Willamette River watershed, 14% of the Cascade Range lies at an elevation range between 1100 and 1300 m, and 91% lies below 1500 m. Approximately 37% of the area of the Clear Lake watershed lies within the 1100–1300 m range, as does 43% of the Smith River watershed.

The historical record suggests that the annual hydrograph has already shifted and that Q_{min} in groundwater-dominated watersheds is already declining as snowmelt occurs earlier (Figure 7). Since the late 1940s the peak discharge month for Clear Lake has shifted from May to April, as a function of earlier snowmelt. This leads to longer summer recession periods, and ultimately lower Q_{min} in groundwater-dominated watersheds. If future warming proceeds as predicted by regional climate models, and simulated by Tague *et al.* (2008) for the Clear Lake watershed, the summer recession will further lengthen and minimum flows will decline even more. It is expected that in runoff-dominated watersheds, longer recession periods will exhaust subsurface storage sooner, but that the effect on Q_{min} will be smaller than in groundwater-dominated watersheds.

CONCLUSIONS

Analyses of the historical datasets for two watersheds in the Oregon Cascades clearly highlight the importance of groundwater systems in sustaining summer streamflow. Discharge and evapotranspiration in the groundwater-dominated watershed account for 260% of incoming precipitation between July and October, so groundwater storage and the associated slow recession are responsible for sustaining discharge even when the seasonal or annual water balance is negative. While groundwater significantly influences the shape of the annual hydrograph, climatic factors control the inter-annual variability of streamflow. Even though Cascades aquifers

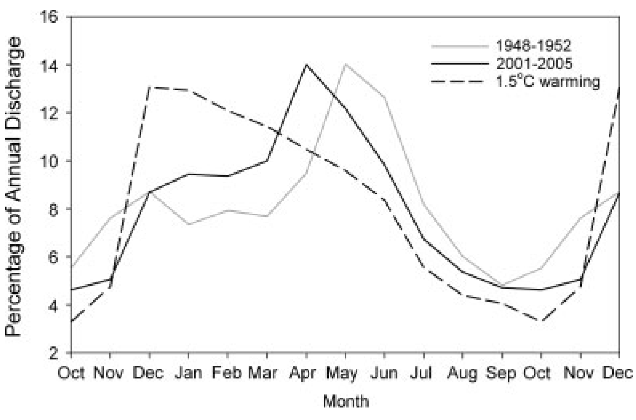

Figure 7. Monthly percentages of annual discharge for the Clear Lake watershed in two historical periods (1948–1952, 2001–2005) and a hydrograph based on 1·5° warming modelled by Tague *et al.* (2008). October to December are repeated to illustrate the low flow period

store multiple years worth of water (Jefferson *et al.*, 2006), and there is a one year memory associated with the groundwater system, these effects are not enough to dampen inter-annual variability in streamflow in groundwater-dominated streams, as evidenced by comparable coefficients of variation between the two study watersheds.

Warmer winters and earlier snowmelt over the past ~60 years have shifted the hydrograph for the McKenzie River at Clear Lake, resulting in an earlier temporal centroid (17 days), longer summer recessions and lowered August discharge (15%) and autumn minimum discharge (11%). Projections of future climate suggest that this trend will continue over the next 50–100 years. The slow summer recession of groundwater-dominated streams makes their low flows more sensitive to changes in snowmelt amount and timing than those of runoff-dominated streams. Groundwater-dominated watersheds that are transitioning between transient and seasonal snow regimes will be the most affected by projected climatic change, particularly in terms of late summer and fall streamflow. In the Oregon Cascades, the region of maximum sensitivity seems to be between 1100 and 1300 m, based on historical trends and projected changes over the next 50 years.

These results suggest that water resource managers in the mountainous western USA must identify groundwater-dominated watersheds and those that are perched at the transient/seasonal snow transition, and that they should be prepared for significant changes to the annual hydrograph in those areas. Downstream water users and hydroelectric dam operators will have to adjust their usage schedules or face shortfalls. Aquatic species, such as salmonids, will be subject to diminishing flows and possibly increasing water temperatures.

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