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Deep groundwater mediates streamflow response to climate warming in the Oregon Cascades

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Abstract Recent studies predict that projected climate change will lead to significant reductions in summer streamflow in the mountainous regions of the Western US. Hydrologic modeling directed at quantifying these potential changes has focused on the magnitude and timing of spring snowmelt as the key control on the spatial–temporal pattern of summer streamflow. We illustrate how spatial differences in groundwater dynamics can also play a significant role in determining streamflow responses to warming. We examine two contrasting watersheds, one located in the Western Cascades and the other in the High Cascades mountains of Oregon. We use both empirical analysis of streamflow data and physically based, spatially distributed modeling to disentangle the relative importance of multiple and interacting controls. In particular, we explore the extent to which differences in snow accumulation and melt and drainage characteristics (deep ground water vs. shallow subsurface) mediate the effect of climate change. Results show that within the Cascade Range, local variations in bedrock geology and concomitant differences in volume and seasonal fluxes of subsurface water will likely result in significant spatial variability in responses to climate forcing. Specifically, watersheds dominated by High Cascade geology will show greater absolute reductions in summer streamflow with predicted temperature increases.

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1 Introduction

Predicting the consequences of climatic forcing on water dynamics and supply involves understanding complex and often highly variable feedbacks between atmospheric and land surface processes. This issue is particularly pressing in the Western US given limited summer water availability, a growing population of over 45 million people in California, Oregon and Washington alone, and burgeoning and competing demands for water for irrigation, municipal water supply, recreation, and aquatic ecosystems. Current analyses of projected climate change impacts show rising temperatures resulting in diminished snowpacks, leading to summer water shortages throughout the West (Service 2004). Knowles and Cayan (2002), for example, predict that the April–July fraction of total annual flow will be reduced by 30% in the Sierras by 2060 as a result of reduced snowpacks. More recent climate simulations taking different greenhouse gas emission pathways into account predict future snowpack reductions of 30–90% (Hayhoe et al. 2004).

While these studies suggest general trends across much of the mountainous Western US, not all regions will be affected to the same degree by these broad climatic shifts. Lower elevations of the North Sierras and Cascades, for example, are predicted to exhibit the greatest differences in the timing and magnitude of snowmelt recharge (Hayhoe et al. 2004; Payne et al. 2004). These broad regional-scale characterizations typically identify climatic gradients as first-order controls on the spatial variability of hydrologic response (e.g., Nijssen et al. 1997). The potential for other landscape controls, notably geology and vegetation, to mediate this response has received much less attention.

While it is generally acknowledged that geologically mediated factors controlling subsurface response (i.e., soil or bedrock storage capacity, hydraulic conductivity) as well as vegetation are important drivers of hydrologic response, such factors have generally not been considered as first-order controls on spatial variation in climate change sensitivity (Maurer and Duffy 2005). In model-based studies of climate change impacts on summer streamflow, estimation of streamflow response is typically based on hydrologic models driven by downscaled precipitation and temperature data from GCMs (i.e., Wood et al. 2004; Hayhoe et al. 2004). These modeling approaches implicitly capture integrated geo-topographic controls on watershed drainage efficiency through calibration of subsurface parameters with respect to streamflow hydrographs (e.g., Nijssen et al. 1997). Explicit characterization of how these controls vary across the landscape has not been done, and would require selection of a model scale to reflect the spatial scale at which geologic differences occur, as well as a comparison of calibrated parameters from different regions.

In this study we address the question of whether a broad-scale geologic framework can be used to reduce uncertainty in summer streamflow prediction, and identify hierarchies and the spatial structure of dominant controls on hydrologic response to climate variation. *We show that differences in groundwater dynamics are as important as topographic differences in snow regimes in determining the response of mountain landscapes to changing climate.* We develop this perspective by presenting a systematic analysis that explicitly incorporates geologic information in a modeling framework to explore the importance of significant geologic distinctions in predicting potential changes in streamflow in the western US. We examine two watersheds within the Oregon Cascade Range that differ both in terms of topographically driven snow accumulation and melt rates, and geologically controlled groundwater storage capacity and drainage rates.

Conceptually, geologic differences between the High and Western Cascades in Oregon represent a gradient of subsurface drainage efficiency, where drainage efficiency is defined as the time lag between recharge, as rain or snowmelt, and runoff. Our earlier empirical analysis suggests that a key control on streamflow differences between these two regions is the partitioning of water input between a fast draining shallow subsurface flow network versus a slow draining deeper groundwater system (Tague and Grant 2004). Differences in the magnitude and timing of water input also exist between the two systems, resulting from differences in the partitioning of precipitation into rain or snow and temperature controls on snowmelt. Seasonal and event-based hydrographs, as well as stream temperatures, demonstrate that High Cascade streams show a much more uniform flow regime, with higher, colder summer baseflows, slower recession rates, and significantly lower winter peak flows relative to streams draining the nearby Western Cascades (Grant 1997; Tague and Grant 2004) (Fig. 1). This geologically driven hydrologic variation is greater than that imposed by climatic differences between the High and Western Cascades, or as a result of forest land use (Jones and Grant 1996), and is comparable to the effect of large flood-control reservoirs (Grant 1997). Our earlier empirical analysis shows that geologic-based differences in High versus Western Cascade flow systems are evident in streamflow signatures even across elevation-based climate gradients (Tague and Grant 2004). Assessing the response of these systems to climate change must therefore take into account both *a priori* climatic gradients and these differences in groundwater dynamics.

In this paper, we use this case study approach to characterize the distinctive climate change responses for the young volcanic terrains comprising the High Cascades relative to the older volcanic landscapes of the Western Cascades. We utilize complementary techniques, including retrospective analyses of stream flow and climate, and a hydro-ecologic model, RHESSys (Regional hydro-ecologic simulation system), to disentangle the role played by both *a priori* climate setting and geology in determining summer streamflow sensitivity to climate variability and change. Empirical analysis of the two basins is used to illustrate differences in summer streamflow behavior and its relationship with climate

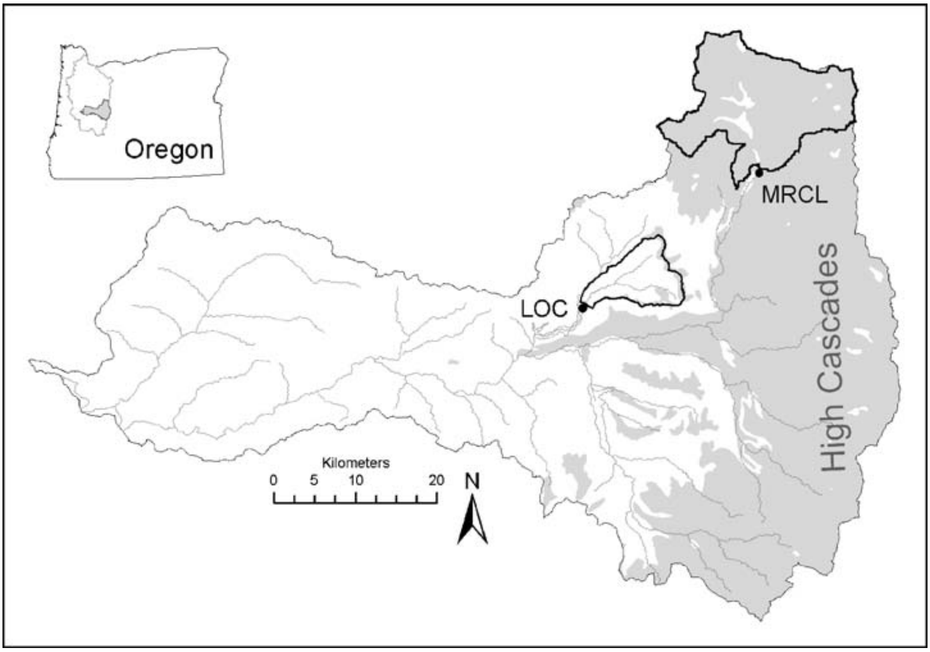


Fig. 1 Study site location

variability. Explanations for these relationships may depend upon a number of factors. Model-based analysis attempts to disentangle the relative importance of multiple and interacting controls. In particular, we use a physically based, spatially distributed modeling approach, RHESSys, to explore the extent to which differences in both snow accumulation and melt and drainage characteristics (deep ground water vs. shallow subsurface) mediate the effect of climate change on summer streamflow.

2 Study watersheds

We constructed this analysis around two watersheds with long-term USGS gaging records that represented relatively clear expressions of the High/Western Cascade geological framework (Fig. 2, Table 1). Lookout Creek (LOC) is a 64 km^2 watershed located within the Western Cascade geologic province and includes the H.J. Andrews Experimental Forest and Long-Term Ecological Research (LTER) site. Lookout Creek has been the site of numerous hydrologic and geomorphic studies (e.g., Swanson and James 1975; Grant et al. 1990; Jones and Grant 1996; Gooseff et al. 2003). Hillslopes are steep, and bedrock is a mixture of Tertiary volcaniclastic units and lava flows cut by scattered dikes (Swanson and Michael 1975; Sherrod and Smith 1989). The 239 km^2 McKenzie River at Clear Lake (MRCL) watershed represents the longest continuous gage record of a High Cascade watershed. It is mapped as 95% High Cascades geology, with extensive low-relief basaltic lava flows less than 5000 years old (Walker and MacLeod 1991; Sherrod et al. 2004).

At both sites, most precipitation falls between October and April as rain below 400 m and snow above 1,200; a transient snow zone between 400 and 1,200 m receives both rain and snow. Peak streamflows occur in winter in response to rain and rain-on-snow events; the summer is characterized by a prolonged (2–3 month) drought between June and September when virtually no precipitation occurs. The area is forested with a mix of old- and second-growth (following harvest) conifers, primarily Douglas fir (*Pseudotsuga menziesii*).

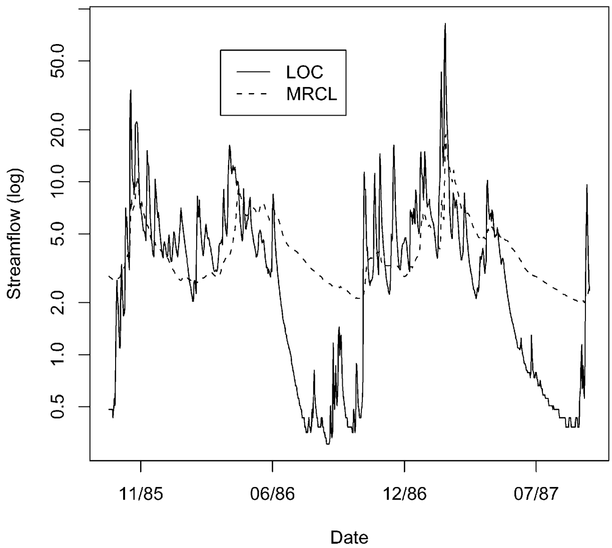


Fig. 2 Annual hydrographs for LOC and MRCL for 1985–1986 water years. Streamflow is normalized by basin area (mm/day)

Table 1 Study watersheds

Location	Geologic Classif.	Drainage area (km^2)	Basin elevation		USGS Gage	
			Range (m)	Mean (m)	elevation (m)	Period of record
Lookout Creek (LOC)	Western Cascade	64	410–1630	985	420	1937.10.01–2003.09.30
McKenzie River at Clear Lake (MRCL)	High Cascade	239	920–2035	1,254	920	1963.09.01–2003.09.30

Hydrologic differences between LOC and MRCL reflect the dominance of a well-developed and steep drainage network with shallow subsurface flowpaths in the older (Miocene to Pliocene) volcanic rocks of the Western Cascades, in contrast to a much deeper groundwater and spring-dominated drainage system that has developed in the younger (Pliocene to Recent) High Cascades. The High Cascades are characterized by extensive fields of thick, blocky lava flows with very low drainage densities and extremely high permeability due to substantial fractures and void spaces within the flows (Manga 1997; Gannett et al. 2003). While a shallow soil layer exists in some High Cascade areas, most vertical recharge drains rapidly to a slower draining and deeper groundwater aquifer. Large groundwater storage volumes and long (decadal-scale) residence times within the High Cascades aquifer result in pronounced lags over different timescales between seasonal input of precipitation and runoff (Manga 1996, 1999). In contrast, the steep, dissected topography of the Western Cascades, coupled with shallow soils and abundant clays forming aquitards result in rapid subsurface flows and little opportunity for groundwater storage. Streamflows are generally quite flashy in the winter, responding quickly to precipitation and snowmelt, with very low summer baseflows. These differences have implications not only for smaller headwater streams in each region, but also for the behavior of larger rivers, such as the McKenzie and Willamette, where both High and Western Cascade streams contribute to streamflow (Grant 1997; Tague and Grant 2004).

3 Empirical streamflow analysis and results

To explore how climate variation affects streamflow across geology, we analyzed historical streamflow, precipitation and temperature data for LOC and MRCL. Annual and August streamflow were calculated for water years 1958 through 2003. Annual precipitation for each watershed was estimated from the Watershed 2 station (CS2MET) located at 465 m elevation and maintained as part of the H.J. Andrews LTER. This meteorological station is within the LOC watershed and approximately 27 km from the MRCL watershed. Spatial interpolation of station precipitation and temperature data was based on PRISM (Parameter-elevation Regressions on Independent Slopes Model), a widely used tool that combines point meteorological data, surface topography and other information to generate spatial estimates of climate variables (Daly et al. 1994).

At an annual timescale, LOC and MRCL streamflow statistics are remarkably similar (Fig. 3, Table 2). Both receive approximately 2.2 m of precipitation annually and have mean runoff ratios (streamflow/precip) of 0.8. Inter-annual variation is also similar for the two watersheds, measured both as standard deviation and coefficient of variation (Table 2). Inter-annual variation in streamflow for both LOC and MRCL exhibit a strong relationship

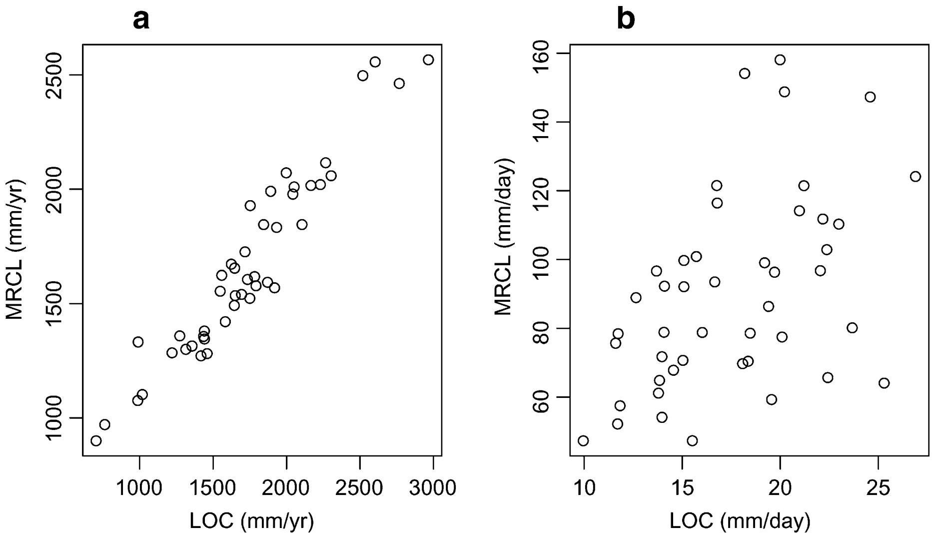


Fig. 3 Correlations between LOC and MRCL for **a** Annual and **b** August streamflow. August streamflow is given as mean daily streamflow (mm/day). Annual streamflow as mean annual streamflow (mm/year)

with annual precipitation; consequently annual streamflow for LOC and MRCL are highly correlated.

During the summer (represented here by August), however, the two watersheds exhibit clearly distinctive flow regimes. MRCL maintains significantly higher August streamflow (normalized by basin area) (Table 2). On average, August streamflow in MRCL is 6% of total annual streamflow while at LOC it is less than 1%. Late summer streamflow is also more variable in the MRCL watershed (measured as both standard deviation and coefficient of variation). Variation in MRCL summer streamflow is also more strongly correlated with annual precipitation (Table 2). Lower year-to-year variation in LOC August streamflow suggests that summer flows in this watershed are less sensitive to climate variability.

Most of the recharge in the Cascade region occurs during the winter months – 80% of precipitation falls between November and May producing a combination of rain and snow in winter and snowmelt of winter precipitation in spring. August streamflow reflects the magnitude of recharge during the previous winter and the rate at which this recharge exits storage. This rate is affected by both snow dynamics and hydrologic characteristics of shallow and deep groundwater storage. Warmer climates produce relatively more rain and earlier

Table 2 Summary of differences in streamflow regimes for study watersheds

Parameters	August (mm/day)		Annual (mm/year)	
	LOC	MRCL	LOC	MRCL
Mean streamflow	0.6	2.9	1,757	1,686
Interannual SD	0.14	1.1	549	463
Interannual coefficent of variation	0.24	0.35	0.3	0.3
Correlation coef (annual streamflow vs precip)	0.42	0.85	0.97	0.94

Streamflow is normalized by drainage area.

snowmelt resulting in more rapid recharge to groundwater systems. Other studies have shown that reducing snowpacks in the Western US will produce higher streamflows in early summer and lower summer streamflow during late summer and found that variation in April temperature was correlated with summer streamflow for the Columbia basin (Mote et al. 2003).

To test for evidence of a pattern of lower late summer streamflow with warmer temperature for our study watersheds in the historical record, we computed the Pearson correlation coefficient between August streamflow and average monthly temperature for each month in the recharge period prior to August (December through July). We also examined the relationship between air temperature during the recharge period and the percentage of total annual precipitation leaving the basin as August streamflow. These latter correlations remove the effect of the magnitude of winter recharge and illustrate more directly the role of snowmelt timing and both melt and watershed drainage rates. In general a negative correlation between August streamflow and monthly temperatures in the preceding months may occur for two reasons. Higher temperatures in spring may reduce snow accumulations that would otherwise persist into the summer, leading to lower summer streamflow. Higher temperatures in summer may also correspond to higher evapotranspiration rates and consequently lower late summer streamflow.

Significant (at the 95% level) correlations between average monthly air temperature and August streamflow were found for May through July for MRCL and only April and July for LOC (Table 3). Similarly, correlations between average monthly air temperature and proportional August streamflow (% of annual precipitation) were found for April through July for MRCL but only for April for LOC.

The negative correlation between August streamflow and July air temperature for both watersheds likely reflects a change in evapotranspiration – with higher air temperatures indicative of conditions leading to higher evapotranspiration rates. Correlations with other months include the impact of snow accumulation and melt rates. MRCL is higher in elevation and snowmelt occurs later in the season, with peak melt typically occurring in May as opposed to April for LOC. Further, MRCL has a more extensive seasonal snowpack with corresponding greater variability in the timing of peak melt within the basin, and therefore a longer time period over which air temperature is a relevant control (April through June vs. April for LOC). In addition to these differences in the relative timing of snowmelt, there are also geologic-based differences in groundwater drainage between the two watersheds, with MRCL showing slower drainage rates. The slower drainage efficiency of MRCL may also serve to maintain the impact of snowmelt, and its sensitivity to climate conditions, throughout the summer. Correlations between monthly air temperatures and late summer streamflow do not necessarily reflect causality and thus it is difficult to disentangle the relative importance of these different factors.

Table 3 Correlation coefficient between mean monthly temperature and August streamflow (in months before August)

Months	LOC	MRCL
Dec	*x*	*x*
Jan	*x*	−0.27
Feb	*x*	*x*
Mar	*x*	*x*
Apr	−0.33	*x*
May	*x*	−0.29
Jun	*x*	−0.38
Jul	−0.36	−0.43

Coefficients are only shown for correlation that obtained a *p*-value within the 90% confidence limit; *x* denotes months that do not have a significant correlation.

4 Process-based model comparisons

RHESSys has previously been applied to a number of catchments in the mountainous west of the US to model both hydrologic behavior (Baron et al. 2000, Tague and Band 2001a, b; Mackay et al. 2003) and carbon cycling dynamics (Landrum et al. 2002; Choate and Tague 2003). We utilized a revised version of the model that augments the shallow subsurface routing approach by including a second, deeper groundwater drainage system (Fig. 4). A complete description of process algorithms can be found in Tague and Band (2004); here we provide a brief overview of key processes represented in RHESSys in order to better interpret model results.

For this study, the ability of RHESSys to incorporate feedbacks between vegetation growth and hydrologic and climate forcing conditions was not used, and vegetation biomass was assumed to remain constant throughout scenarios. While changes to vegetation production, biomass accumulation and species change are likely to be important responses to climate change in the Western US (e.g., Lenihan et al. 2003; Bachelet et al. 2003; Neilson et al. 2005); they are not the focus of this study. The ability of RHESSys to model changes to vegetation production, however, sets the stage for subsequent research building on model applications developed in this study.

RHESSys partitions the landscape into a hierarchical set of landscape units (Band et al. 2000) that reflect the scale at which different hydro-climate-ecological processes are modeled. Most vertical hydrologic processes, as well as carbon and nitrogen cycling, are computed at the finest resolution patch layers – where patches are defined as either grid cells or spatially variable units designed to minimize within-unit variance in hydro-ecologic properties. Meteorological forcing (radiation, precipitation, temperature) is organized at the hillslope level, where hillslopes are defined as the area draining either side of a stream reach. Routing between patches within each hillslope level ultimately produces shallow subsurface and surface drainage to each stream as well as soil moisture patterns within each hillslope. In the modified RHESSys version used in this study, each hillslope is also assigned a single deeper

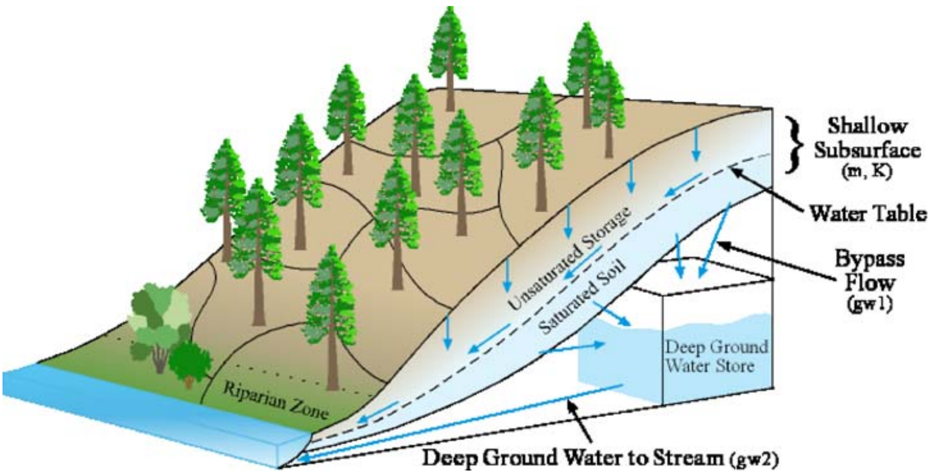


Fig. 4 RHESSys representation of shallow subsurface and deeper groundwater systems. *m* Defines rate of decay of hydraulic conductivity with depth, *K* defines saturated hydraulic conductivity at the surface, *gw1* defines percentage of effective (infiltrated) precipitation that becomes bypass flow to a deeper groundwater store, *gw2* defines drainage rate of linear deeper groundwater store

groundwater store, which is modeled as a simple linear reservoir. Two parameters control the dynamics of this reservoir: gw1, which determines percentage of infiltrated water that goes directly to the deeper groundwater store, and gw2, which controls the drainage rate of the groundwater linear reservoir. Infiltration that does not go to the groundwater store is assumed to enter the soil matrix and shallow subsurface flow system. The bypass flow to the deeper groundwater flow is assumed to follow macropores, cracks, or voids created by blocky lava flows prevalent in the High Cascades. Infiltrated water that is assigned as bypass flow to the deeper groundwater store is not available for plant or soil evapotranspiration.

Vertical hydrologic processes modeled in RHESSys include canopy interception, evaporation and transpiration, soil evaporation, snow accumulation and melt and infiltration. The snowmelt model is based on an empirical approach (Coughlan and Running 1997) adapted to consider both radiation and temperature (latent heat) controls on melt, the effect of canopy cover on radiation and wind attenuation and advective heat transfer associated with rain-on-snow events. The model currently sets minimum temperature at which incoming precipitation is assumed to be rain at −3°C and the maximum temperature at which snow occurs as 0°C. These values were selected to minimize error in partitioning precipitation into rain and snow based on both hydrograph and limited comparisons between RHESSys and SNOTEL measurements.[1]RHESSys models shallow subsurface hydrology using a simply two layer soil model with an unsaturated and saturated zone. Saturated, lateral transport between patches within a hillslope is based on topographic gradient and soil characteristics following Wigmosta et al. (1994).

Soil characteristics that control the rate of lateral shallow subsurface water transport in RHESSys are described by two parameters, *m* and *K*, which characterize saturated hydraulic conductivity at the soil surface and the decay of this conductivity with depth, respectively. Given the 900 to 100,000 m^2 patch size, *m* and *K* must be considered effective parameters which integrate the fine scale heterogeneity of measurable values. Further, values for these soil parameters are typically calibrated at the basin scale using a comparison between observed and modeled flow. Previous application of RHESSys in this region found calibrated values for *K* are typically higher than measured values for the soil matrix, suggesting that the calibrated value includes the effects of preferential flowpaths (Tague and Band 2001a). For this study, calibrated values for m and *K* are assumed to reflect the drainage efficiency of the shallow subsurface flow system at the hillslope scale.[2] Validation and climate change scenarios were run for the full 40-year record (water years 1960–1999). Measures were selected to test the model's ability to represent variation in streamflow at daily, monthly, seasonally and inter-annual timescales. Total streamflow is included to check for error in evapotranspiration estimation.

[1] Infiltration is based on the Green and Ampt model (1911); however, high infiltration capacity of soil in both Western Cascade and High Cascade watersheds means Hortonian overland flow is rarely observed (Harr 1977).

[2] Calibration of shallow subsurface parameters *m* and *K* applies a basin-wide scaling of initial values, and uses a Monte-Carlo based approach to maximize measures of fit between observed and modeled flow (Tables 4, 5, and 6). Calibration is performed at a daily time step for a 25-year period (water years 1960–1984). Four calibration metrics are used – the Nash-Sutcliffe Efficiency between observed and modeled daily streamflow, Nash-Sutcliffe Efficiency between log-transformed observed and modeled daily streamflow, error in total calibration period streamflow, error in total August streamflow in calibration period. Nash-Sutcliffe is computed following Nash and Sutcliffe (1970), where a value of 1.0 represents perfect correspondence.

Previous research suggests that drainage in the Western Cascades is dominated by shallow subsurface saturated flow, constrained by clay confining layers (Harr 1977). For LOC, we assume bypass flow to deeper groundwater is negligible and set gw1 and gw2 parameters to 0.0. This implementation for LOC is similar to other studies (e.g., Tague and Band 2001a, b) that have used RHESSys in this region. For MRCL, we assume that soil characteristics and the shallow subsurface flow system are similar to that of the Western Cascade site, and soil parameters (m and K) were set to values found through calibration in LOC. For the deeper groundwater system in MRCL, we consider two contrasting implementations: one in which the deeper groundwater system is included in addition to the shallow subsurface flow system and a second in which only the shallow subsurface flow system is modeled. For the first implementation, groundwater parameters (gw1 and gw2) are calibrated based on observed and modeled streamflow for MRCL using the same 4 measures of fit used to calibrate the shallow subsurface flow system in LOC.

The two model implementations in MRCL are designed to illustrate respectively: (1) the role of topographically mediated climate versus (2) the combined role of topography and geology as controls on hydrologic response. In the implementation of MRCL without the deeper groundwater system we are essentially modeling a hypothetical landscape with the topography and vegetation of MRCL but shallow subsurface dominated geology of the Western Cascades. Topography in this case primarily influences climate – with the higher elevations of MRCL (and much of the High Cascades) producing a more snow-dominated system relative to LOC (and much of the Western Cascades). When the deeper groundwater system is included, the impact of geology on drainage efficiency is added. Differences in response to historic and future climate forcing between these two implementations in MRCL provide insight into the relative roles of climate and geology in determining hydrologic behavior in this region.

The three model implementation, LOC (Look), MRCL with deeper groundwater (MRCL-GW), and MRCL without deeper groundwater but with shallow, subsurface flow (MRCL-SSF), is used to examine the potential impact of warmer temperatures. Current GCM model scenarios predict a 1 to 3°C warming in the Cascades with modest changes in precipitation. (Payne et al. 2004). We consider two scenarios, based on a 2020 and 2050 future climate prediction for the Pacific Northwest (Parson et al. 2001). These correspond to a 1.5°C warming and a 2.8°C warming, applied uniformly to 40 years of historic meteorologic data (1960–1999). We assume no change in precipitation. By applying warming to this 40 year historical record we are able to simulate variation in meteorological forcing, including ENSO (2–7 years) and, to a lesser extent, PDO (20–30) cycles that at least historically have been important drivers of variability in this region (Mote et al. 2003; Beebee and Manga 2004). Although simplistic, these climate change scenarios are broadly consistent with general predictions for this region (Parson et al. 2001) and serve to highlight differences in sensitivity across the three model implementations Fig. 5.

5 Results

5.1 Model performance

For LOC, reasonable model performance (Nash-Sutcliffe [N.S] >0.7 for both log- and non-log transformed streamflow) was obtained by constraining the value of m (decay of hydraulic conductivity with depth) (Tables 4, 5, and 6). Major seasonal trends in the observed hydrograph are well represented by the model (Fig. 7). Variation in model

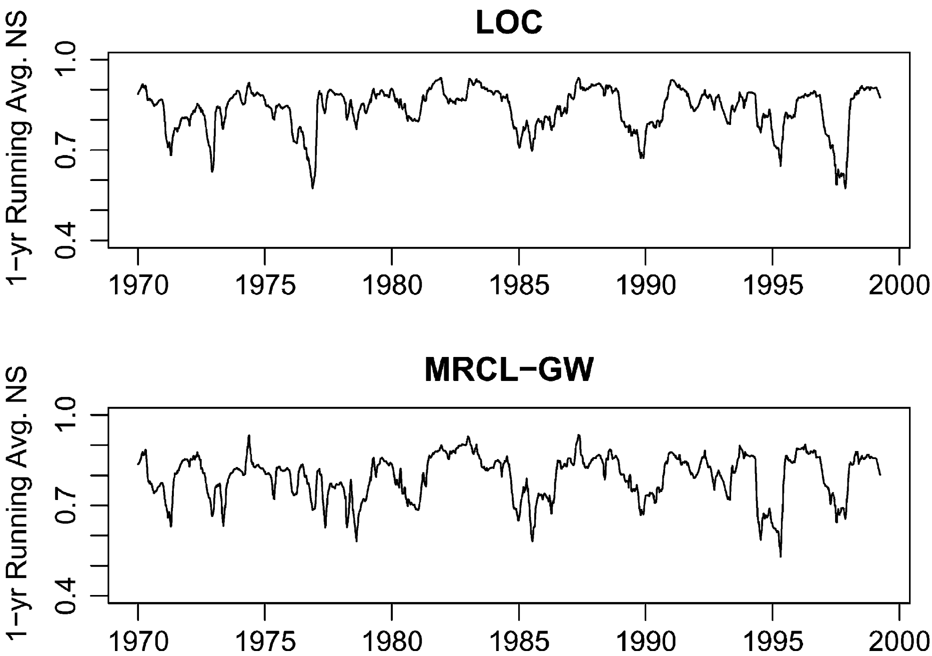


Fig. 5 Moving window (1 year) of Nash-Sutcliffe efficiency of log-transformed streamflow for water years 1970–1999

performance through time using the selected parameter set is assessed by computing an N. S. efficiency of a moving 1-year window over the simulation period (Fig. 6). Years when model performance is poor are typically those where the model misclassifies incoming precipitation as rain or snow. Analysis of modeled streamflow estimates found that these misclassification errors are not biased (e.g., they do not consistently over- or under-estimating rain versus snow) and thus should not impact estimation of multi-year flow statistics (e.g., mean August streamflow or annual minimum 7-day flow explored in this study).

For MRCL, calibration of gw1 suggests that approximately 50% of infiltrated water becomes part of the slow draining deeper groundwater systems (Table 3). The remaining infiltrated water in the shallow subsurface system generates storm flow as well as plant and soil evapotranspiration. Total modeled streamflow over the 30-year period of record is within 2% of the observed value, suggesting that the model does a reasonable job of estimating evapotranspiration and that the model maintains enough shallow subsurface water to support this level of ET. Again the model captures major seasonal trends in observed hydrograph (Fig. 7). Year-to-year variation in model performance is similar to that

Table 4 Calibrated parameter values

Parameter values	*m*	*K*	gw1	gw2
LOC	1.4	58	NA	NA
MRCL-GW	1.4	58	0.5	0.03
MRCL-SSF	1.4	58	NA	NA

Table 5 RHESSys predicted and observed streamflow statistics and measures of fit

Summary statistics		MRCL model	MRCL observed	LOC model	LOC observed
Mean	Annual (mm/year)	1,651	1,686	1,757	1,757
	August (mm/day)	2.9	2.9	0.5	0.6
SD	Annual (mm/year)	360	463	527	549
	August (mm/day)	0.7	1.1	0.2	0.14

of LOC with poor performance often due to misclassification of incoming precipitation (Fig. 6).

The importance of including the calibrated deeper groundwater store is evident in a comparison of calibrated model with and without the deeper groundwater system and observed hydrographs (Fig. 7). For MRCL, the model significantly underestimates summer baseflow for the approach without the inclusion of the deeper groundwater store. When deeper groundwater model is included, model performance improves and modeled streamflow maintains pattern of higher summer baseflow To test the models ability to capture snowmelt, RHESSys predictions of seasonal snowpack were compared with SNOTEL (Jump-off Joe, site 22E07S) measurements (Fig. 8a) and show that the model captures the general pattern of accumulation and melt.

5.2 Warming scenarios

The modeled impact of warmer temperatures shows a significant loss in snow accumulation and an increase in rates of snowmelt (Tables 4, 5, and 6, Fig. 8b). The impact on LOC is significantly greater, with 25 and 50% decreases in the number of days where there was snow cover for some fraction of the basin Table 7. Reductions were less for the higher elevation MRCL watershed. These results are consistent with other studies that have shown that intermediate elevations, which are often at the boundary between rain and snow precipitation events, are likely to experience the greatest changes in snow under a warmer climate (Mote et al. 1999; Leung et al. 2004).

There are interesting differences between MRCL-GW, MRCL-SSF and LOC in the modeled streamflow response to warmer temperature (Figs. 9, 10 and 11). All scenarios show an increase in winter flows and a reduction in spring and summer flows (Fig. 9). This is consistent with overall predictions made by other climate studies in the Western US. Percent reductions in mean August and annual minimum 7-day flows are similar for LOC

Table 6 RHESSys performance

Measures of fit (all computed for water years 1960–1999)		LOC	MRCL-GW
Nash-Sutcliffe efficiency			
Daily streamflow		0.7	0.7
Log transformed daily streamflow		0.8	0.9
Total error	All days	−2	0.1
(as percent)	August only	−2	−10

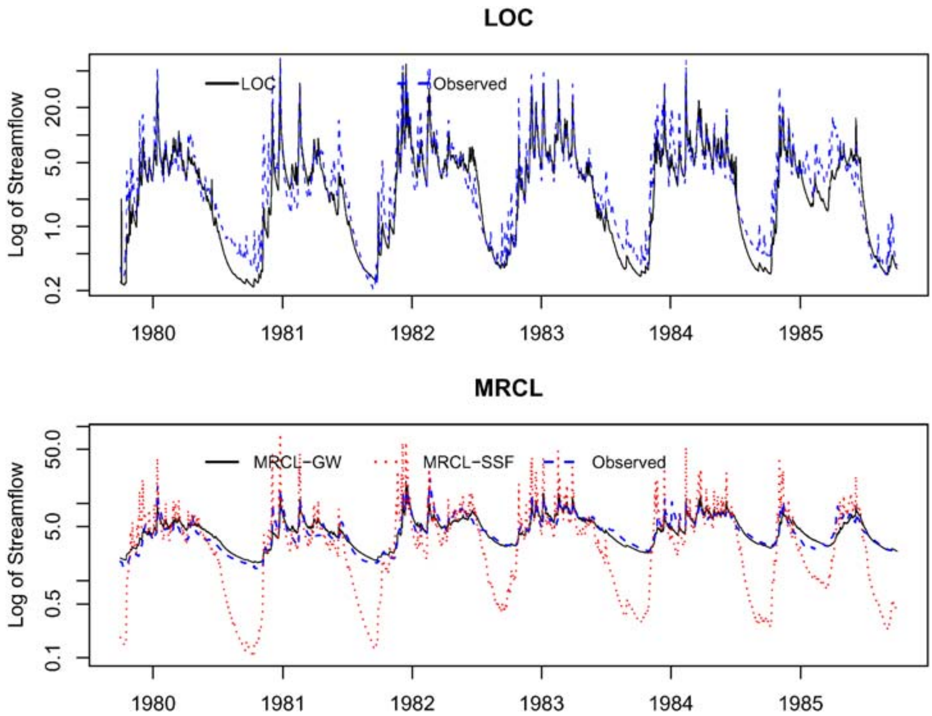


Fig. 6 Streamflow comparison for MRCL; with and without the inclusion of the deeper groundwater model – for water years 1980–1985

and MRCL-GW (Figs. 10a, and 11a). However, significantly greater percent reductions (>40 and 50% for mean August, and annual 7-day minimum, respectively) were obtained for the hypothetical catchment, MRCL-SSF, with MRCL topography and climate and only a shallow subsurface flow system. Higher percent reductions in summer streamflow associated with MRCL-SSF suggests that without the deeper groundwater component, the greater changes in snowpack associated with the higher elevation would result in a greater percent changes in summer streamflow. In reality, however, these changes are buffered by the deeper groundwater drainage system.

If reductions are expressed as an actual loss in streamflow (normalized by drainage area) results show a different pattern, with the greatest losses in both mean August and annual minimum flows occurring in the MRCL-GW catchment (Figs. 10b, and 11b). When actual loss in normalized summer flow is considered, LOC and MRCL-SSF behave similarly.

Table 7 Decrease in the number of days with snow cover for two warming scenarios −1.5°C and 2.8°C

Parameters	1.5°C Warming		2.8°C Warming	
	(days)	(%)	(days)	(%)
LOC	60	25	114	48
MRCL-GW	26	10	41	16

A day is considered to have snow cover if any patch within the watershed contains snow. Results are shown as number of days lost and as percent change (reduction) in the number of days.

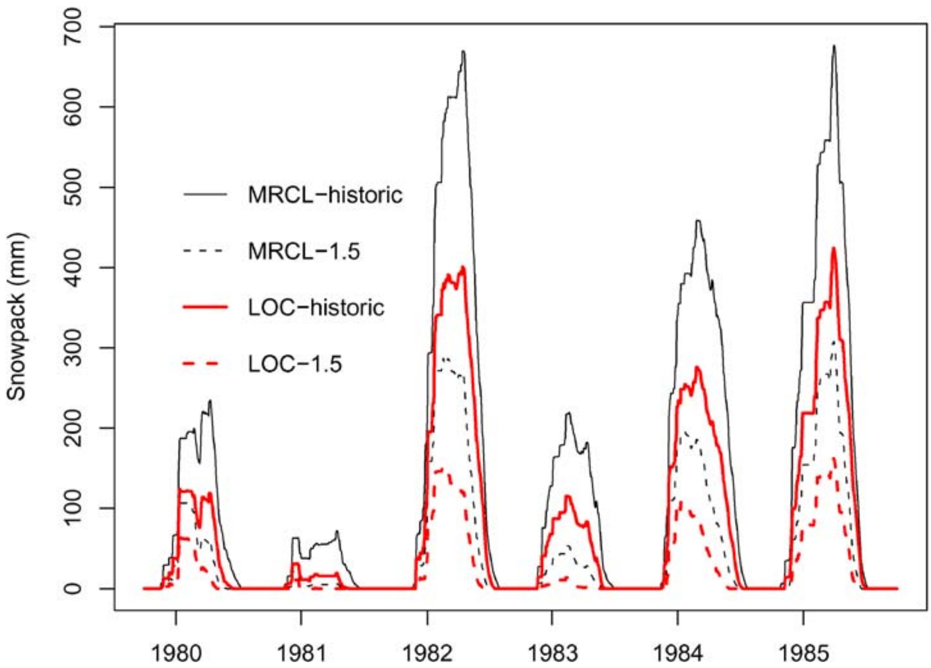
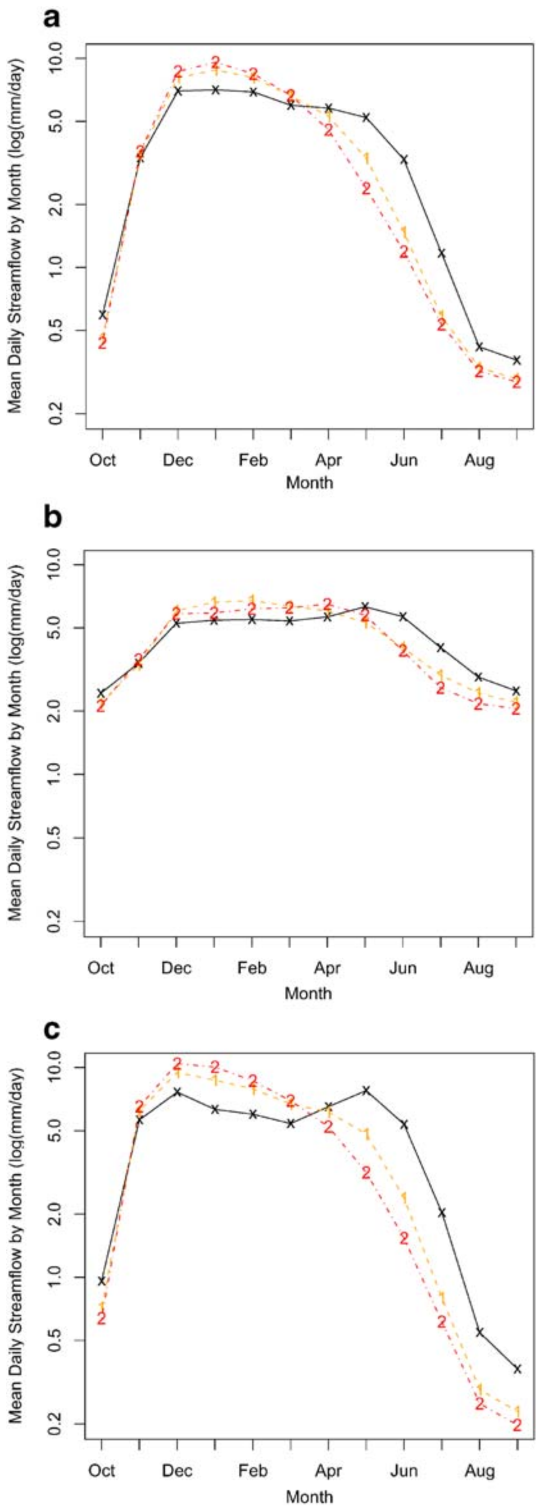


Fig. 7 Change in snowpack with warming scenario (1.5°C), shown for water years 1980 through 1985. *Solid lines* show baseline RHESSys predictions of mean basin snowpack (as water equivalent depth). *Dotted lines* show change with 1.5°C warming scenarios

These results illustrate the complex interaction between differences in sensitivity to snowmelt and differences in drainage efficiency. Understanding why the deep groundwater dominated system has a greater reduction in August flow volumes requires considering how the different watersheds translate the recharge signal into streamflow. In MRCL, there are two recharge peaks – the first in November as rain and the second in May during snowmelt. When the model includes the deeper groundwater system the streamflow response to recharge is delayed and extended to a much greater degree than a model using only the shallow subsurface system (Fig. 12a,b). Under climate warming, the recharge pattern collapses into a single peak in November as rain. Storage and subsequent slow metering out of this signal by the deeper groundwater system produces a three month delay in the timing of peak streamflow relative to peak recharge (Fig. 12c). In contrast, the shallow subsurface flow system has much less storage, and responds with only a one month delay in peak streamflow relative to recharge (Fig. 12d). In addition to delaying the timing of peak streamflow, the deep groundwater system also stretches both changes in recharge and consequent streamflow response over a longer time period and hence later into the summer. The greater absolute loss of August streamflow volumes under a warmer climate with the deep groundwater system is due to this extended storage effect. Greater relative changes in August streamflow occur with the faster shallow subsurface flow system since August volumes are very small and consequently highly sensitive to changes in recharge timing. These are greater for the shallow subsurface system in higher elevation MRCL relative to lower elevation LOC because the later snowmelt (even under warming) maintains the effect of changes in snowmelt to a later time in the season.

Fig. 8 Mean daily streamflow by month for historic data, 1.5°C (warm 1) and 2.8°C (warm 2) warming scenarios for **a** LOC, **b** MRCL-GW, and **c** hypothetical scenario of MRCL-SSF



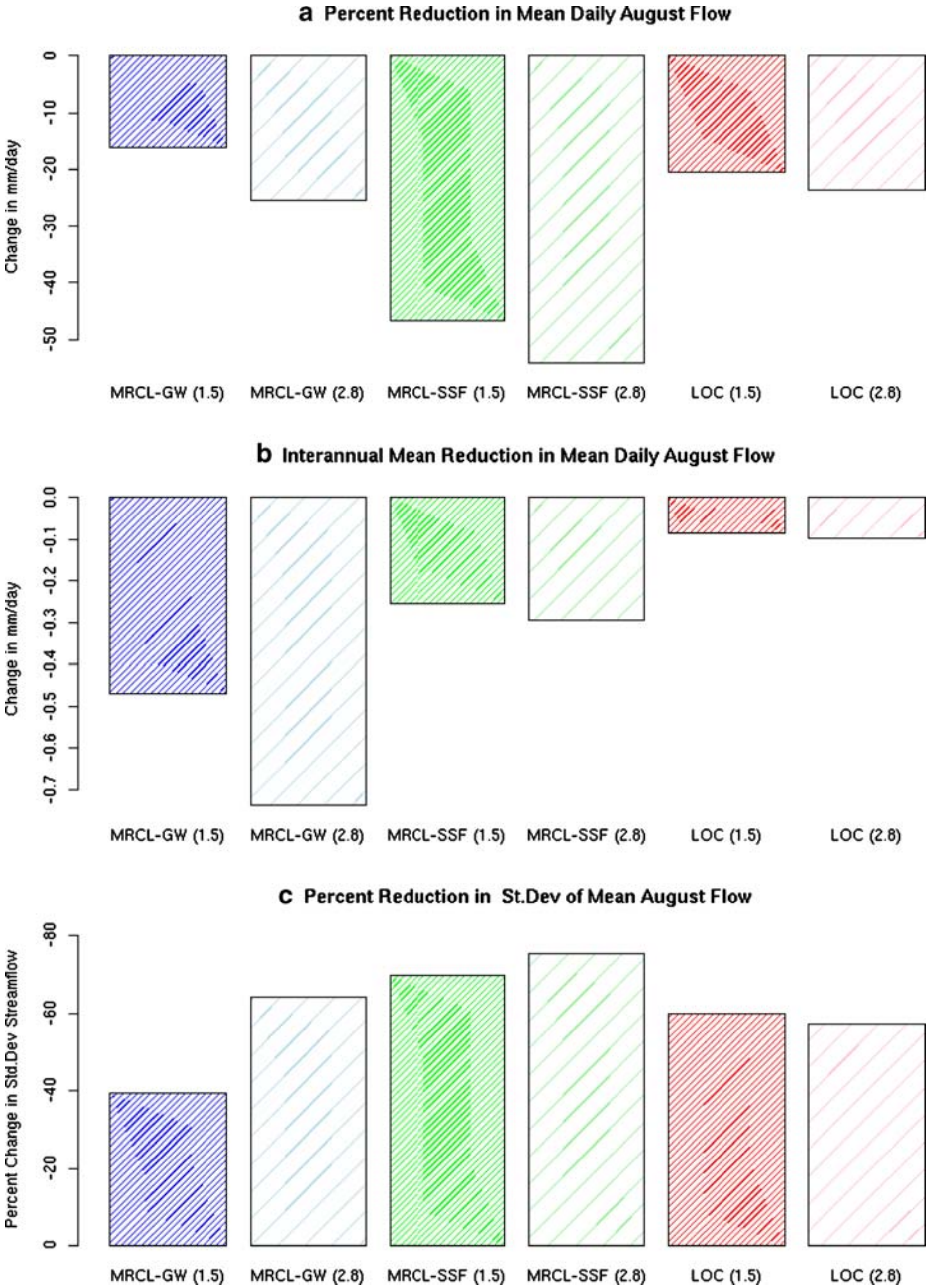


Fig. 9 RHESSys modeled change in mean August daily streamflow for 1.5°C (warm 1) and 2.8°C (warm 2) temperature increase scenarios. Comparisons are made between LOC, MRCL modeled with deeper groundwater included, and hypothetical scenario of MRCL modeled with only shallow subsurface flow system. Results shown as **a** percent reduction in flow, **b** actual reduction in flow (normalized by drainage area), and **c** as percent reduction in year-to-year standard deviation

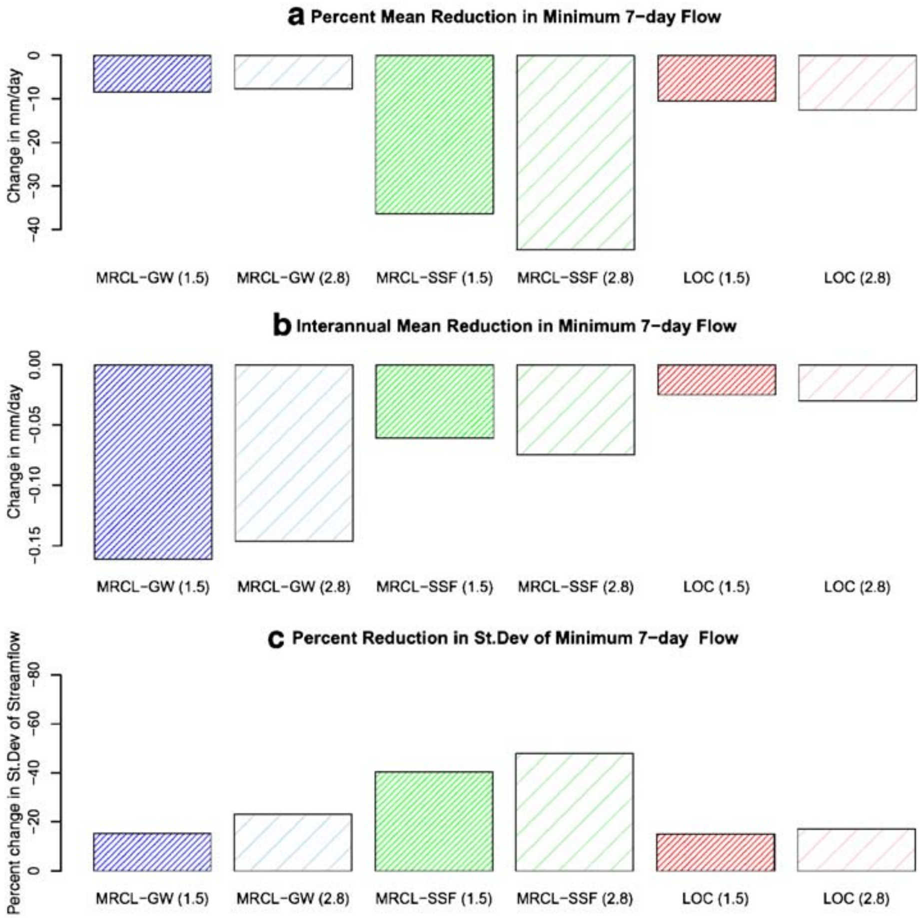


Fig. 10 RHESSys modeled change in mean minimum 7-day streamflow for 1.5°C (warm 1) and 2.8°C (warm 2) temperature increase scenarios. Comparisons are made between LOC, MRCL modeled with deeper groundwater included, and hypothetical scenario of MRCL modeled with only shallow subsurface flow system. Results shown as **a** percent reduction in flow, **b** actual reduction in flow (normalized by drainage area), and **c** as percent reduction in year-to-year standard deviation

The overall pattern of streamflow sensitivity is true for both moderate (1.5°C) and large (2.8°C) predicted warming scenarios, with only slight increases in effect with the more extreme warming scenarios. Further differences between watersheds are generally greater than differences between warming scenarios, suggesting that uncertainty in climate forcing, while important, can be overshadowed by within-region differences in watershed behavior.

Changes in snow accumulation and melt impact summer streamflow largely by changing the timing of runoff. Changes in temperature can also alter the total amount of streamflow by changing evapotranspiration. Warming scenarios for both watersheds do show a small reduction (<3%) in annual streamflow and hence a small increase in evapotranspiration but this is insignificant relative to changes due to snowmelt. Longer-term changes in evapotranspiration, however, may be greater if changes in biomass and fire frequency were included in the analysis.

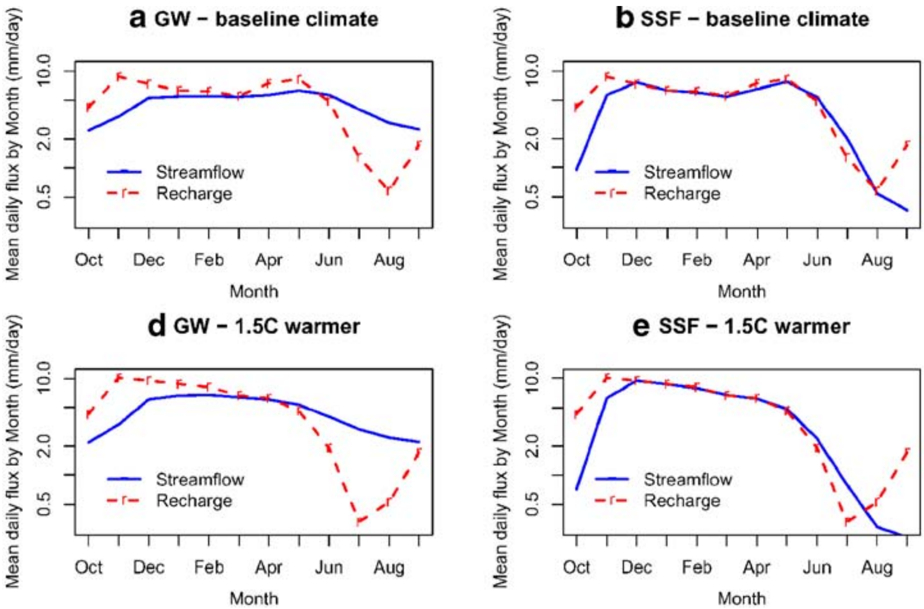


Fig. 11 RHESSys predictions of average (over historic climate variability) recharge and streamflow for 4 scenarios for the MRCL watershed: **a** model includes both shallow subsurface and deeper groundwater system. This scenario illustrates historical conditions; **b** model includes only a shallow subsurface flow system under historical climate variability; **d** model with deeper groundwater under a 1.5°C warmer climate; and **e** model with only shallow subsurface flow system under a 1.5°C warmer climate

6 Discussion and conclusions

This study was designed to gain insight into the importance of geologic-based distinctions in drainage efficiency in mediating the impact of a warming climate in the Cascade Range of the Pacific Northwest. Empirical analysis shows that watersheds within the High Cascades (groundwater-dominated) and those in Western Cascades (shallow subsurface flow dominated) have distinctive flow regimes and may have different sensitivities of summer streamflow to year-to-year variation in air temperature. Interpreting the cause of between-watershed differences in streamflow is complex given that High Cascade watersheds are also at a higher elevation and consequently have a different snow accumulation and melt regime as well as different geology. By using a physically based model that can separate the effect of snowmelt and geology as controls on streamflow, we show that while *a priori* differences in snow regimes are important, the presences of a geologically controlled deep groundwater system plays a key role in determining the impact of a warmer climate on summer streamflow.

The RHESSys model was successfully calibrated for two case study watersheds, LOC as a Western Cascade SSF system and MRCL as a High Cascade groundwater-dominated system. If actual reduction in flow (normalized by drainage area) is considered, the High Cascade catchment shows significantly greater reductions in both mean August streamflow and annual minimum 7-day flows under warming scenarios. Given that the MRCL study watershed is at a higher elevation than LOC, it is tempting to explain these differences as resulting from topographically controlled variation in snow accumulation and melt. The hypothetical modeling scenario (where MRCL is modeled as a faster shallow subsurface

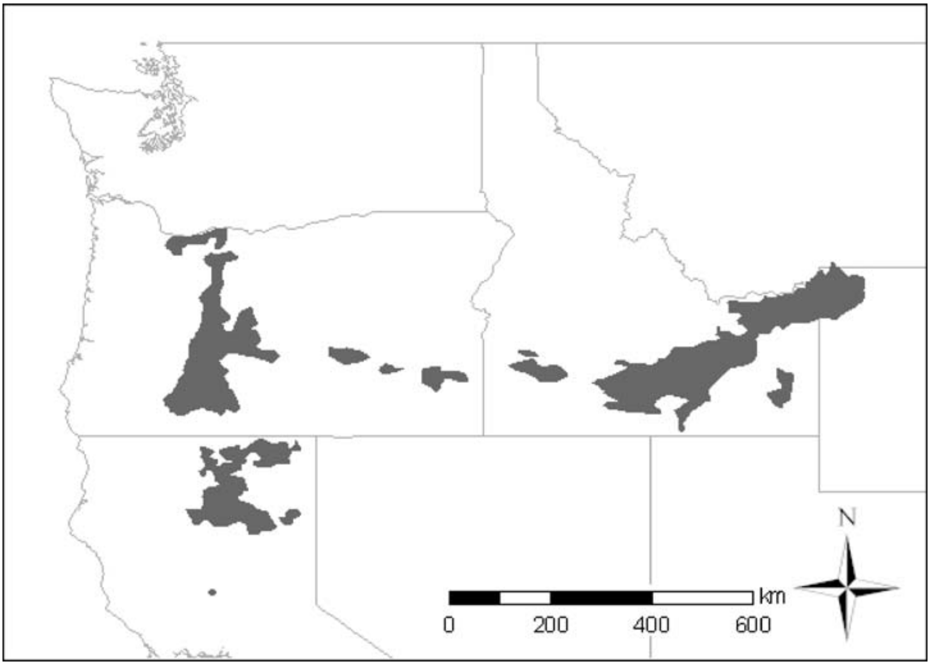


Fig. 12 Extent of Quaternary basalts in the Northwestern United States (Schruben et al. 1998)

flow system but with MRCL vegetation, topography and climate), shows, however, that these distinctions are instead due primarily to differences in rate of subsurface drainage.

The hypothetical MRCL-SSF generally exhibits behavior that is similar to LOC, producing lower August streamflow volumes relative to MRCL-GW, and lower absolute reductions in August flow with warming, although the percent impacts on August flow are higher. Large relative changes in summer flow volumes to some extent reflect the high sensitivity of very small summer flow volumes to even small absolute changes in magnitude. Greater relative changes in August streamflow for the hypothetical MRCL-SSF watershed relative to LOC can be explained as a difference in timing. Even under warming conditions, melt timing in the higher elevation MRCL-SSF still occurs later in the season than that of LOC, and thus changes are maintained for longer and are more evident in the late summer.

That a deep groundwater system can simultaneously buffer and amplify the effects of warming on summer flow is an unexpected and somewhat paradoxical result of this work. We emphasize that our results indicate that this is not due to the minor differences in ET between scenarios, but rather because the shape of the annual recharge and discharge hydrographs are fundamentally shifted and transformed by warming (Fig. 12). Shallow subsurface flow systems typical of the Western Cascades have higher winter flows, lose their distinct spring snowmelt peak and experience a more sustained recession throughout the spring. The deep groundwater dominated High Cascade system also sees higher and more sustained winter flows, loses its subtler snowmelt peak, and has lower sustained summer flows than under current conditions.

Based on these results we conclude that it is the difference in drainage efficiency rather than snow accumulation and melt regimes that allows the groundwater dominated High Cascade watershed to exacerbate in a volumetric sense, the effect of climate warming on summer flow. Whether relative or absolute changes in summer streamflow are important depends upon one's point of view. For individual headwater streams in the Western Cascades, the large percentage reduction in what are already low flow volumes may have significant ecological consequences through altering stream temperatures, pool volumes, and other habitat features (Sinokrot et al. 1995; Jager et al. 1999). Because snow is the primary storage term for Western Cascade systems, loss of winter snowpack will likely result in permanent streams becoming more intermittent and ephemeral, resulting in an overall contraction of the last summer drainage network. High Cascade streams will continue to flow year-round because of their groundwater component, but with reduced volumes.

From a water supply perspective, these flow volume reductions are a critical consideration. All results shown here have been normalized by drainage area. Conversion of unit area change to mean daily August non-normalized flow for the 1.5°C warming yields a loss of 1.30 m^3/s for MRCL but only 0.06 m^3/s for LOC . While MRCL is twice as large as LOC, the loss of flow associated with the moderate warming scenario is over 20 times greater than for LOC (or that of the hypothetical SSF MRCL system). For the 2.8°C warming, MRCL loses 1.96 m^3/s, which is 23% of current mean August flow and almost 30 times greater than LOC. Our earlier work has shown that in the summer, flow from High Cascade watersheds dominates the entire Willamette system (Tague and Grant 2004). Consequently, small changes in the High Cascade system will have large repercussions for water supply at the larger basin scale. To adequately address the impact on downstream water resources, differences in geologically mediated drainage efficiencies across the entire basin need to be considered.

Our analysis has only focused on climate warming — our assumption throughout has been that precipitation remains constant under all scenarios. The most recent climate simulations, however, generally predict wetter winters and drier summers for the Pacific Northwest (Mote et al. 2003). Although there is quite a range in model predictions, it is clear that predicting streamflow responses to future climates will require also joint consideration of temperature and precipitation changes within the broader geological context that we have emphasized here.

The young volcanic landscapes of the Oregon Cascades exert a strong control on the region's streamflow regimes and likely trajectories of change. Although the High Cascades have a number of unique features that predisposes them for the kind of geological contrasts and hydrologic behaviors examined here (i.e., extensive areas in young, undissected lava flows, thick basalt accumulations in structurally controlled basins), such features are not unique to the Cascades. Other areas with recent extrusive volcanism, such as the Snake River Plain of Central Idaho, the southern Cascades of northern California, and even Hawaii offer geologic analogues, though sometimes radically differing climates. Figure 12 illustrates the extensive areas of young volcanic landscapes through the Western US. These are all areas that are likely to maintain deep groundwater systems exhibit contrasting hydrologic responses to climate change when compared with older volcanic landscapes in the surrounding regions. Extension of findings from this study beyond the Oregon Cascades to places such as the Sierra Nevada of California and other western ranges relies on clear understanding and model representation of geologically controlled surface and sub-surface processes as well as topographically controlled climate forcing. In particular, groundwater storage must be viewed as potentially playing as important a role as snowpack dynamics.

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