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Impacts of Climate Change on Groundwater Availability and Spring Flows: Observations from the Highly Productive Medicine Lake Highlands/Fall River Springs Aquifer System

L.K. Mancewicz, L. Davisson, S.J. Wheelock, E.R. Burns, S.R. Poulson, and S.W. Tyler

Research Impact Statement: Increasing temperatures and decreasing precipitation will likely result in a decrease in spring discharge for highly productive volcanic aquifer-spring systems.

ABSTRACT: Medicine Lake Highlands/Fall River Springs Aquifer System, located in northeastern California, is home to some of the largest first-order springs in the United States. This work assesses the likely effects of projected climate change on spring flow. Four anticipated climate futures (GFDL A2, GFDL B1, CCSM4 rcp 8.5, CNRM rcp 8.5) for California, which predict a range of conditions (generally warming and transitioning from snow to rain with variable amounts of total precipitation), are postulated to affect groundwater recharge primarily by changing evapotranspiration. The linkages between climate variables and spring flow are evaluated using a water balance model that represents the physics of evapotranspiration and recharge, the Basin Characterization Model. Three of the four climate scenarios (GFDL A2, GFDL B1, CCSM4 rcp 8.5) project that by the year 2100, groundwater recharge (and consequently decreased spring flow) will decrease by 27%, 21%, and 9%, respectively. The fourth scenario (CNRM rcp 8.5) showed an increase in recharge of 32% due to a significant increase in precipitation (27%). Evapotranspiration increases due to a shift in the type of precipitation and a longer growing season. While the likelihood of each scenario is outside the scope of this work, unless total precipitation increases dramatically in the future, increased temperatures and decreasing precipitation will likely result in reduced spring flows, along with warmer water temperatures in downstream habitats.

(KEYWORDS: climate change; groundwater; recharge; volcanic aquifers; springs.)

INTRODUCTION

The impact of climate change (magnitude, timing and form of precipitation, as well as persistent temperature trends) on groundwater recharge and spring flow in volcanic aquifers is largely unknown, but potentially of critical environmental concern. Aquifers dominated by volcanic rocks can be highly productive, maintaining large springs, particularly when dominated by large open connected pathways at depositional contacts or in fracture zones (Singhal and Gupta 2010; Burns et al. 2016). About 75% (60 of 80) of the largest springs in the conterminous United States (U.S.) (~>3,000 L/s) are fed by volcanic aquifers (Meinzer 1927). Many of them, such as those in the Eastern Snake River Plain (Meinzer 1927; Lindholm 1996) of southern Idaho (U.S.), the Massif Central in France (Guglielmi et al. 2000), and on the Big Island of Hawaii (Paulsen et al. 1950) support large springs as well as significant

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Department of Civil, Environmental, and Geospatial Engineering (Mancewicz), Michigan Technological University Houghton, Michigan, USA; Department of Geological Sciences and Engineering (Mancewicz, Poulson, Tyler), University of Nevada, Reno Reno, Nevada, USA; ML Davisson & Associates, Inc. (Davisson), Livermore, California, USA; Lassen National Forest, U.S. Forest Service (Wheelock), Fall River Mills, California, USA; Department of Hydrology and Atmospheric Sciences (Wheelock), University of Arizona Tucson, Arizona, USA; and U.S. Geological Survey, Geology, Minerals, Energy, and Geophysics Science Center (Burns), Portland, Oregon, USA (Correspondence to Tyler: styler@unr.edu).

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human development. Young, permeable volcanic aquifers tend to support a large and steady spring flow to rivers that intersect them (Gannett et al. 2001; Tague et al. 2008; Gannett et al. 2010; Jefferson et al. 2010; Burns et al. 2017). Volcanic aquifers are also of critical importance to fisheries ecosystems where recharge on very permeable highlands feeds large, perennial springs that discharge relatively cold water for anadromous fishes (Tague et al. 2008; Jefferson et al. 2010). Volcanic terranes can also contain sources of geologic nitrogen and other nutrients, leading to highly productive river ecosystems (Lusardi et al. 2016). The cool discharge from the Fall River Springs (FRS) provides habitat for threatened and endangered salmonids and other species (Davisson and Rose 1997). Volcanic spring-fed river systems have much higher invertebrate densities when compared with runoff-dominated ones, though their species richness (the total number of species present) is often lower (Lusardi et al. 2016; Jeffres 2017).

In this work, we focus on the Medicine Lake Highlands (MLH)/FRS a volcanic aquifer-spring system in northeastern California. Volcanic aquifers often have very short residence times due to their low storage and high permeability, as well as large recharge rates associated with orographically driven precipitation (see, e.g., Manga 1999). Burns et al.'s (2017) modeling of the MLH-FRS system suggests that simple climate warming, independent of changes in recharge or precipitation, will impact spring discharge temperatures at time scales of 60–100 years. Work by Freeman (2001) suggests that the system is also sensitive to changes in precipitation. Freeman (2001) demonstrated a tight coupling between reconstructed spring discharge from the MLH-FRS aquifer system and annual variability in catchment precipitation. Typical time lags were 1–2 years, which is consistent with a very permeable but low storage aquifer system.

The upper Deschutes Basin (mid-Cascade Range in Oregon, east of the crest) is analogous to the MLH (southern Cascades, east of the crest). Climate-driven hydrologic modeling revealed that changes in the seasonal timing of recharge changed the timing of stream discharge in watersheds with relatively short groundwater flowpaths by 1–2 months (Waibel et al. 2013). However, Waibel et al. (2013) did not find statistically significant changes in the future simulated annual recharge over the chosen ensemble of futures.

While changes in precipitation may be an obvious driver to changes in recharge, recent work continues to shed light on projected variations in precipitation. For example, Massoud et al. (2019, 2020) used a novel framework for determining the ensemble mean from multiple GCMs in order to evaluate changes in precipitation magnitude and distribution. Direct warming is another somewhat obvious driver to changes in recharge, however, warming climate will impact several other processes that affect recharge (Ishida et al. 2019) to these aquifers including (1) a change from snow-dominated to more rain-dominated precipitation; (2) changes in the timing, duration, and magnitude of evapotranspiration; and (3) changes to the vegetation community (Hayhoe et al. 2004; Waibel et al. 2013). For example, Trinh et al. (2017) estimate that declines of snowfall (and transition to rainfall) of 15%–20% in the catchments of northern California, including the Fall River System, and a decrease in snow water equivalent (SWE) of up to 80% by 2099. Ishida et al. (2019) developed a physically based snow model of north-central California catchments that also predicted large declines in SWE and earlier snowmelt runoff. This was not only driven by warming under expected future climates, but also changes in wind and humidity under expected future climates. Results from Berghuijs et al. (2014) further suggest that a switchover to rain-dominated precipitation regime might decrease summer baseflows in a wide range of catchments. Most recently, Milly and Dunne (2020) documented the importance of changes in the land surface albedo when snow is no longer present. For the upper Colorado River Basin, they concluded that increases in evapotranspiration (and subsequent decreases in runoff) are due both to warming temperatures and to an increase in the amount of solar radiation absorbed on newly snow-free areas. In the MLH-FRS system, groundwater recharge is analogous to runoff in traditional catchment studies, as the MLH-FRS system has only one highly intermittent stream but continuous and large spring discharge. The interplay between the changing climatic processes, particularly in high recharge systems, will control the amount of groundwater recharge and ultimately affect the quantity of water reaching streams and rivers (Jeffres 2017).

The MLH/FRS system maintains both groundwater-dependent ecosystems and anthropogenic use. It is a reliable supply of cold water to Shasta Lake, the largest reservoir in California (Davisson and Rose 1997), which also produces substantial hydropower. It is representative of many volcanic aquifer systems in its proximity to maritime water sources, its current snow-dominated precipitation regime, and its high recharge rates (Davisson and Rose 2014). Using existing results from a climate scenario model, the impacts to spring flow from changes in both the timing and form of precipitation and ongoing warming were investigated.

Study Site and Hydrologic Characteristics

The FRS, located in northeastern California (Figure 1), are classified as first-magnitude springs (discharge >2.8 m^3/s, Meinzer 1927) with a historical range of discharge measurements of 33–42 m^3/s (Meinzer 1927). Subsequent five-year moving averages of spring discharge from 1924 to 2001 ranged from 24 to 38 m^3/s, with an increasing trend in discharge from 1932 onward (Freeman 2001).

The water from FRS provides approximately 17% of the average annual inflow to Shasta Lake and approximately 5% of the total hydropower for California (Davisson and Rose 1997). In combination with the hydrologically similar neighboring systems in the vicinity of Mt. Shasta and Mt. Lassen, these three systems of volcanic aquifer springs supply 10%–15% of the total power generation for the state (Davisson and Rose 1997; Nyberg 2019).

The volcanic aquifer that supplies water to the FRS is composed of lava flows that originate from Medicine Lake Volcano (MLV), the volcanically active portion of the MLH (Figure 1; Davisson and Rose 2014; Burns et al. 2017). MLV is a half million-year-old shield volcano which is the largest edifice by volume in the Cascade Range Volcanic Arc (Donnelly-Nolan 2010). The caldera attains a maximum elevation of 2,440 m and the lavas from these eruptions range from basalt through rhyolite covering an area of 2,000 km^2 (Donnelly-Nolan 2010).

One structure of hydrologic significance is the Giant Crater Lava Field. The corresponding Giant Crater Eruption occurred approximately 10,500 years ago, and is made up of six chemically different lavas with layer ranges in estimated thickness from 5 to 30 m (Donnelly-Nolan et al. 1991). The associated vent is ~50 km north of the FRS, on the southwestern flank of MLV (Donnelly-Nolan et al. 1991). This long, thin, and narrow flow complex accumulates and transmits groundwater recharge from the MLH to the FRS which lie at its terminus at ~1,010 m AMSL (Burns et al. 2017).

Discharge from the springs fluctuates slightly on a 14- to 16-year cycle that appears to mirror trends in precipitation (Freeman 2001). Following the work of Burns et al. (2017), topographic divides will be used to define the catchment (see Figure 1). It will be shown that the recharge zones of the aquifer likely extend beyond the surface divide.

MLV also contains a geothermal reservoir at 1,500–1,100 m below the surface (Cumming and Mackie 2007). It appears to be tightly sealed and isolated from the surface, based on the lack of surface expression and the occurrence of cold-water springs downstream (Burns et al. 2017). This is likely due to the presence of an overlying low-permeability layer which is evident in geophysical surveys. It was formed by hydrothermal alteration and likely serves as a confining layer for the geothermal reservoir (Cumming and Mackie 2007; Burns et al. 2017). At the top of the watershed, in the MLH and overlying the geothermal reservoir, is Medicine Lake, a shallow, 7.5 km^2 snow, spring, and rain-fed lake. Interpolation of geophysical evidence (Cumming and Mackie 2007) suggests that the lake is underlain by a low-permeability, bowl-shaped confining layer. This may

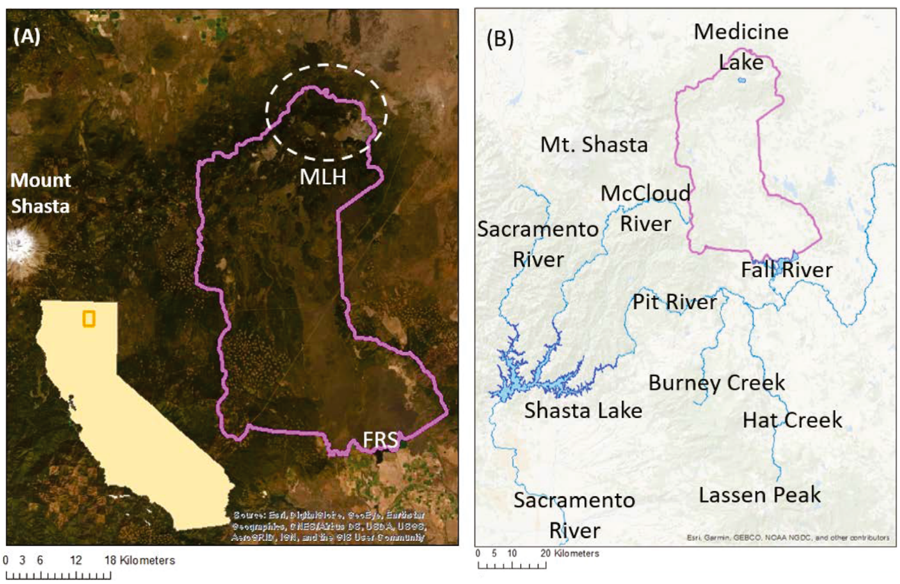

FIGURE 1. (a) Study area location with model domain outlined in purple and corresponding to the topographic watershed of the Fall River. The white dashed line indicates the Medicine Lake Highlands (MLH). Medicine Lake is located near the center of the highlands. The Fall River Springs (FRS) discharge approximately 45 km south and downgradient of MLH. (b) Study area with respect to surface water and landscape features.

be what allows the lake to exist despite the generally permeable terrane. The lake is perennial, but since it has no surface water outflows, it is presumably in hydraulic connection with the aquifer.

Recharge for the MLH-FRS Aquifer System originates mainly within the topographic basin (Figure 1). Most of the precipitation in the area falls on the MLH. Analyses of stable isotopes from FRS and surrounding areas indicate an isotopic composition similar to that of groundwater near the MLH (Davisson and Rose 2014; Mancewicz 2018). Stable isotope analyses for waters are reported using the normal δ notation vs. VSMOW, as $\delta^{18}O$ for oxygen and δD for hydrogen. This suggests that the majority of aquifer recharge also occurs in the MLH. Additional recharge is also likely to occur lower in the catchment, as soils are very thin and vegetation is much less dense on the lower portions of the Giant Crater Lava Field (Figure 1). The vegetation is likely sparse as a result of limited soil moisture storage capacity on these recently developed soils (typically coarse loams) and lower annual precipitation. Dominant species are overstory of lodgepole pine (*Pinus contorta*), sugar pine (*Pinus lambertiana*), and red and white fir (*Abies magnifica* and *Abies concolor*) and an understory of bitterbrush (*Purshia tridentata*), manzanita (*Arctostaphylos*), and snowbrush (*Ceanothus velutinus*). Due to the high permeability of the volcanic rock and thin soils, Davisson and Rose (2014) estimated that up to 75% of the average precipitation of 101 cm/year at MLV becomes recharge. Due to the substantial infiltration capacity of the soils and the highly transmissive aquifer below them, there are very few surface water features.

Knowledge of the dynamics and geometry of the MLV-FRS aquifer is surprisingly limited given the importance of the associated spring flows to California's water supply. With the exception of several geothermal exploration wells at the head of the catchment, no wells, and only a few small discharge springs exist upslope of the FRS. However, several factors support the concept of short groundwater residence times. Tritium analyses showed spring water ages of 20 years (Davisson and Rose 2014), although some samples were enriched in helium of mantle origin (Davisson and Rose 2014). Mancewicz (2018) resampled several of the springs reported in Davisson and Rose (2014) for deuterium, $\delta^{18}O$, and chlorofluorocarbons (Table S1). Little to no change in stable isotopic composition was observed from the large volume springs between 1997 and 2017 (FRS-1, FRS-4, FRS-12). Since CFC samples are unaffected by volcanic gas inputs and thus may be a better tracing tool in such environments, springs were sampled for CFCs following Busenberg et al. (2006) and Reston Groundwater Dating Laboratory (2017), and Table 1 documents the CFC-estimated ages of groundwater discharging at several of the FRS vents, showing ages of 33–46 years. Ages were very similar for each of the available CFCs (CFC-11, CFC-12, and CFC-113). While slightly older than the reported tritium ages, the consistency of the CFC ages across analytes and locations provide confidence that waters discharging at the major springs are modern and are consistent with fast travel times through the aquifer.

Somewhat discrepant CFC results were found at Paynes Springs, a small (1.5 L/s) high altitude spring 0.5 km from the top of the catchment near Medicine Lake (Schneider and Mcfarland 1995). A recent sampling from it showed significant differences in stable isotopic composition from 1997, including a 12‰ enrichment in deuterium. CFC ages range from ~30 to 32 years, inconsistent with expectations given its location 45 km upgradient from FRS, which show ages only a few years older (Mancewicz 2018). The time-variant stable isotope concentrations and unexpectedly old CFC ages suggest that Payne Spring may be influenced by surface water infiltration from the adjacent Medicine Lake.

While travel times are on the order of 30–40 years, the hydraulic response of FRS to precipitation (i.e., changes in flow rate due to drought or wet periods) is estimated to be in the range of 1–3 years (Mancewicz 2018). These times are generally consistent with the observed lag between spring discharge and multi-year precipitation patterns (Freeman 2001), as well as the conceptual model of a highly conductive aquifer system with but a small effective porosity (Mancewicz 2018).

In summary, like many volcanic aquifers, the MLV-FRS system is an "outlier" with respect to typical porous media aquifer systems, with recharge rates an order of magnitude higher, subsurface velocities of 2–5 km/year, and the very tight temporal coupling of spring discharge to precipitation variability. In these and other respects, the MLV-FRS System (and many other volcanic aquifer systems) is more analogous to a karst aquifer system. Thus, its response to climate change may be surprisingly dynamic.

Climate and Recharge: Current and Future

The average precipitation on MLH is 101 cm/year, much of which currently falls as snow. Because no well-developed surface water drainages exist, snowmelt directly recharges the aquifer system (Davisson and Rose 2014). Many studies predict that the decline in snow-dominated precipitation across the western U.S. (Knowles et al. 2006; Lute et al. 2015; Lynn 2015; Markovich et al. 2016; Safeeq et al. 2016) will continue to expand (Dettinger et al. 2015). Most studies predict rising temperatures and a 17%–80%

TABLE 1. Calculated CFC apparent ages from springs in the study area. CFC ages from all springs sampled at the major FRS discharge zone show ages between ~30 and 40 years (Mancewicz 2018).

Location	Date	Latitude	Longitude	Temperature (°C)	apparent age (years)		
					CFC-11	CFC-12	CFC-113
Paynes Springs	2017-09-09	41.5567	−121.5630	7.48	31.9	29.9	30.2
FRS-1	2017-10-22	41.1120	−121.5509	12.64	44.3	41.3	35.8
FRS-4	2017-10-22	41.1156	−121.5528	10.99	46.3	43.3	37.3
FRS-12	2017-10-21	41.1336	−121.4079	12.89	40.8	35.0	32.8

reduction in SWE (Lute et al. 2015; Trinh et al. 2017; Ishida et al. 2019). Klos et al. (2014) suggest that a major shift from snow to rain in the MLV-FRS watershed will occur by mid-century (Figure 2). However, Harpold et al. (2017) state that this may be an overestimate. A shift from snow to rain is generally thought to decrease the amount of recharge, because rainfall is considered to be more bioavailable, which allows an increase in actual evapotranspiration (AET) (Portge 2015; Meixner et al. 2016).

METHODS

To investigate the impact of climate change on flow through the MLH-FRS aquifer system, publicly available output for historic and predicted future climates from the Basin Characterization Model (BCM) was analyzed (Thorne et al. 2012). Water balance BCM output under current and future predicted climates (Flint and Flint 2014 and publicly available at https://ca.water.usgs.gov/projects/reg_hydro/basin-characterization-model.html) was statistically analyzed under four possible future climate predictions to evaluate the impact of climate change on the aquifer and spring system.

The BCM is a discretized surface and subsurface water and energy balance model that uses climate and landscape information to simulate hydrologic response at the basin scale. It was chosen for this study because it extends the traditional soil storage modeling approach by incorporating a more detailed treatment of bedrock permeability in order to more accurately estimate groundwater recharge (Flint et al. 2013). This model has been widely used in California and the western U.S. focusing long-term water balances and groundwater recharge. BCM simulates

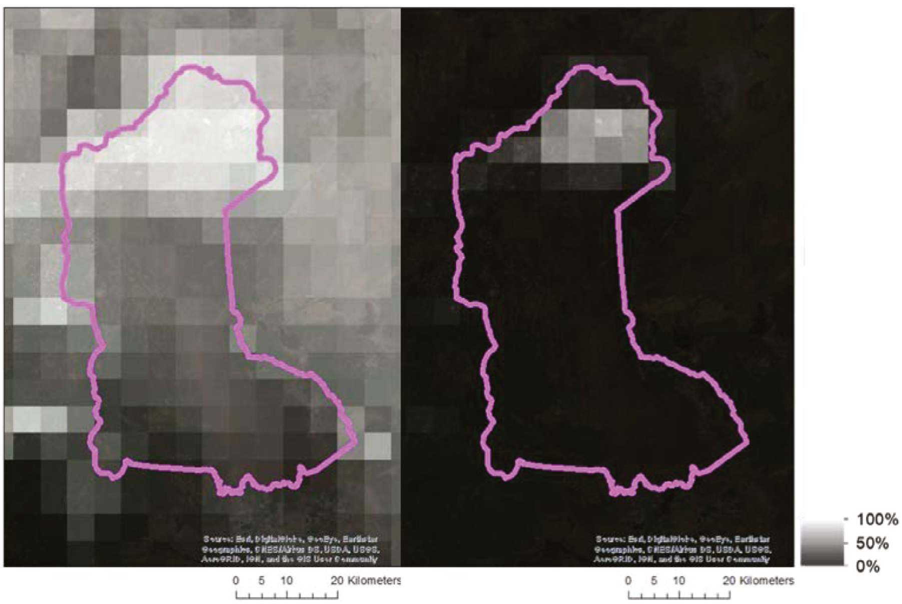

FIGURE 2. Probability of precipitation as snow from data by Klos et al. (2014), where 0% indicates no chance of precipitation falling as snow (precipitation falls as rain) and 100% indicates a 100% likelihood of precipitation falling as snow. The left image shows the average historical scenario (1979–2012) and the right shows an averaged modeled future scenario (2035–2065). The study area is outlined in purple.

the major components of the water budget, including runoff, AET, and recharge. Precipitation and snowmelt are each partitioned into runoff and infiltration. The latter is subsequently divided between AET and recharge using a physical model of soils that have been parameterized using the SSURGO database. Potential evapotranspiration (PET) is calculated via a modified Priestley–Taylor approach (Flint and Childs 1991) that accounts for the effects of snow cover on albedo, consistent with the approach used by Milly and Dunne (2020). The model operates on monthly time scales. The regional BCM for the southwestern U.S. (Flint and Flint 2007) was calibrated for California using stream gage data from 68 unimpaired watersheds throughout the state and validated using an additional 71 basins (Flint et al. 2013). While the BCM does not route streamflow, due to the lack of surface runoff observed in this basin as well and the unimpaired nature of the site, the MLH-FRS system was deemed a suitable locale for utilizing the BCM results.

The study area was defined as the topographic basin above FRS, including the entire MLV caldera, which preferentially drains to the south (Figure 1). BCM simulations used 270 m grid cells and monthly time steps. Data from Flint and Flint (2014) for each of the 21,626 grid cells, defined by the topographic basin draining toward the FRS, were analyzed both individually and summed as part of developing basin scale estimates of key hydrologic fluxes and reservoirs. Monthly air temperatures and precipitation data were obtained from PRISM (Parameter-elevation Regression on Independent Slopes Model: Daly et al. 2008). Future climate simulations were used as input parameters in order to evaluate the possible impacts of climate change at the regional and watershed scales (Flint and Flint 2007, 2012b; Micheli et al. 2012; Heller et al. 2015; Thorne et al. 2015).

Three time periods of BCM output (Flint and Flint 2014) were used to represent historical conditions (1920–2010) and three more time for future conditions (2010–2099). The period 1900–1920 is also used in order to establish reasonable initial conditions. The inputs for BCM simulation are based on PRISM data downscaled from the 4 km or 800 m (where available) to 270 m using a modified gradient-inverse distance squared method (Nalder and Wein 1998; Flint and Flint 2012a). Four downscaled General Circulation Models (GCM) were analyzed: GFDL A2, GFDL B1, CNRM rcp 8.5, and CCSM4 rcp 8.5. These were selected in keeping with recommendations from Lynn et al. (2015), who evaluated historical output for several GCMs with a set of metrics specifically related to California's climatic and hydrologic variability. The projected GCM scenarios span 2011–2099. The GFDL simulations suggest a warmer and drier climate for the region, whereas the CNRM rcp 8.5 simulation produces a warmer but wetter climate. CCSM4 rcp8.5 represents a warmer and wetter period, followed by continued warming, but with less precipitation.

BCM input, estimated precipitation, snowpack (as SWE), AET, and groundwater recharge are all compared to discern the influence of future climate on water resources. The model defines groundwater recharge as water that has infiltrated below the root zone. Since there are no well-developed and continuous surface water drainages in the study area, we include any modeled runoff was treated as groundwater recharge. Large volumes of water infiltrating the downgradient of the runoff area are also observed in the adjacent volcanic basins of the Upper Klamath and Modoc Plateau in northeastern California and southern Oregon, respectively (Flint et al. 2013).

The historical BCM simulations were evaluated against the records from two weather stations: one near Medicine Lake (SnowLite station ID MED) and the other in Fall River Mills, California near FRS (NWS Cooperative Network station number 042964-2). Temperature, SWE, and precipitation records were compared.

Nine model grid cells were selected for more detailed statistical analyses. Points 0–5 are along a transect that spans the transition from snow to rain (Figure 3; Table 2). This transition zone was defined by examining 30-year average historical April 1 snowpack for 1951–1980 and 1981–2010. This output was normalized by total precipitation and then historical and projected snowpacks were compared. The remaining points serve to capture any other variability across the study area. Point 6 is located in the high elevation area to the west of MLH. Point 7 is west of FRS at a lower elevation with relatively high precipitation. Point 8 is also near FRS and lowest in both elevation and annual precipitation (Figure 3; Table 2).

Simulation output, at the nine observation points, was smoothed using LOWESS (Locally Weighted Scatterplot Smoothing) in order to facilitate the evaluation of persistent changes over time (i.e., trends in high variability data). This is an iterative, robust, nonparametric smoothing technique (Helsel and Hirsch 2002). A specific LOWESS smoothing factor value (i.e., fraction of data used for local smoothing) was chosen to allow comparison of model output with the sub-decadal variability of FRS flow documented by (Freeman 2001). These methods were applied at each point for total yearly groundwater recharge, AET, yearly total precipitation, and snowfall (SWE), as well as normalized total yearly recharge and normalized total yearly AET. In this context, normalization refers to normalizing by the total yearly precipitation. "Yearly" time scale refers to calendar year, as opposed to water year.

The significance of trends in model output was examined using the Theil–Sen (also known as the

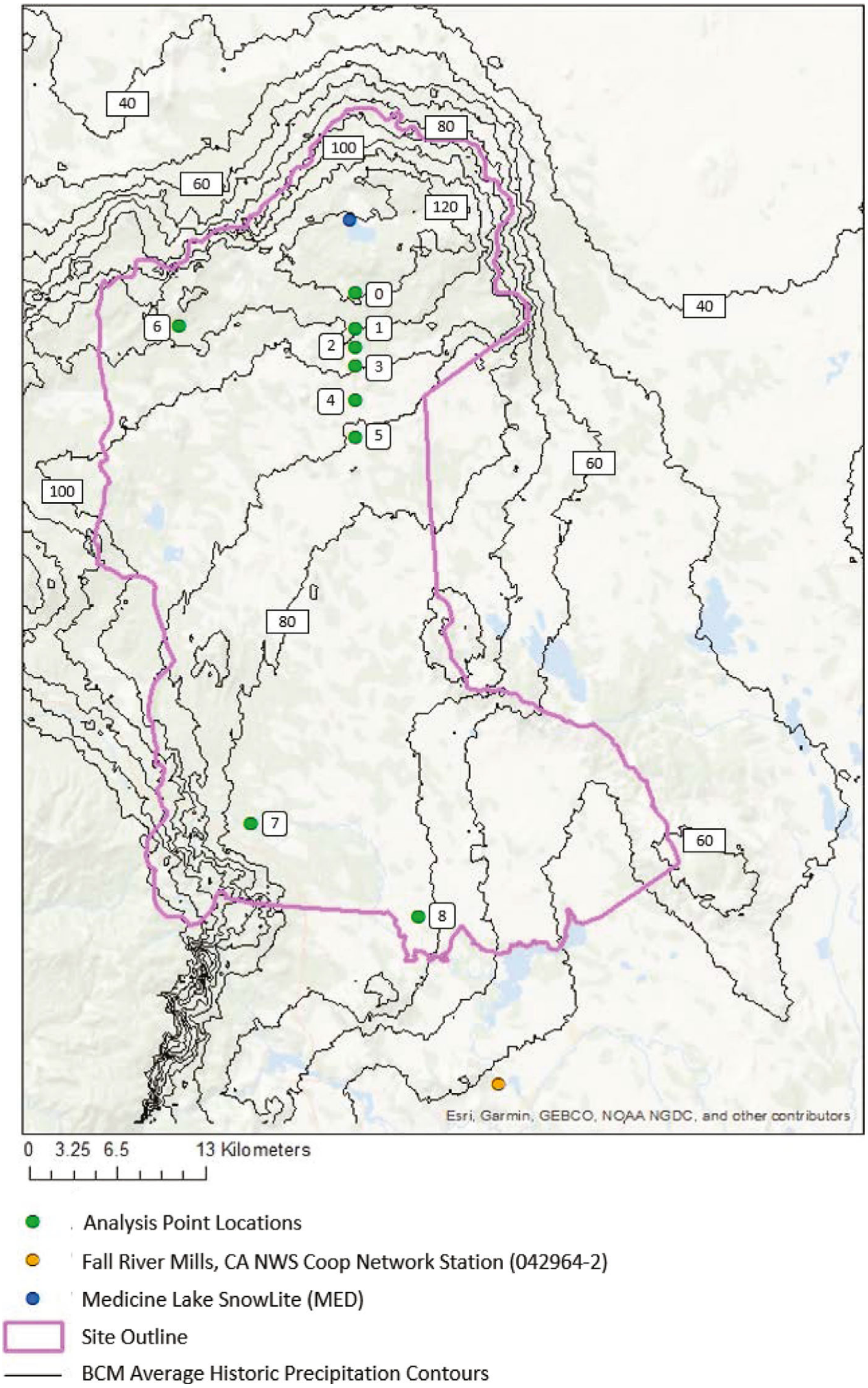

FIGURE 3. Study area map showing Points 0–5 which span the area where precipitation transitions from snow to rain and Points 6–8 which captures other variability across the site. The weather station near Fall River Mills, California is shown in orange just south of the site boundary and the weather station near Medicine Lake is shown in blue. Contours show the average historic precipitation (cm/year). BCM, Basin Characterization Model.

Kendall–Theil) Method. A positive slope of the Theil–Sen Line indicates an increasing trend and a negative slope indicates a decreasing one (Helsel and Hirsch 2002). Trends were determined based on whether the slope is statistically different from zero at the 95% confidence interval.

Correlation analysis was performed on snowfall and groundwater recharge. The correlation coefficient was

TABLE 2. Location and description of point locations for BCM output analysis. The *X* and *Y* coordinates are for the Teale Albers Projection.

Location name	General location	*X* (m)	*Y* (m)	Elevation (m)
Pt0	MLVH	−134,095.823	392,978.973	2,013
Pt1	MLVH	−134,015.893	390,314.671	1,800
Pt2	MLVH	−134,149.109	388,929.234	1,710
Pt3	MLVH	−134,042.536	387,570.440	1,665
Pt4	MLVH	−134,095.823	384,932.781	1,547
Pt5	MLVH	−134,015.893	382,188.549	1,468
Pt6	West of MLVH	−146,991.046	390,447.886	1,732
Pt7	West of FRS	−140,852.037	353,696.992	1,230
Pt8	Near FRS	−129,393.008	346,614.677	1,026

assessed to provide a measure of the strength of association between the two variables. Pearson's *r* was used, which provides a measure of the linear relationship between two variables (Helsel and Hirsch 2002).

AET, groundwater recharge, snowfall, and precipitation were plotted to assess changes in the timing of climate stresses and resulting hydrologic budgets. Monthly output was averaged over 30-year periods for both the historical and projected scenarios in order to examine the cumulative output on a seasonal basis.

Model output was also examined at the watershed scale. The publicly available output from the 30-year average BCM (Flint et al. 2013) was summed over the watershed area to determine the cumulative volumes of precipitation, groundwater recharge, and AET over each 30-year prediction interval. The April 1 snowpack, AET, and groundwater recharge were also evaluated.

RESULTS

BCM Output and Historical Observation Data

Total monthly precipitation observations and BCM historical output are well correlated (Pearson's $r = 0.78$, Figure 4) to the relatively short precipitation record from Medicine Lake. However, BCM appears to overpredict winter precipitation at Medicine Lake. For the precipitation observations at Fall River Mills, the historic observations are well correlated with BCM historical output.

BCM Historical Scenario

The BCM output shows no statistically significant trend in total annual precipitation at any of the observation points during the historical period. The average annual precipitation for the entire watershed is 87 cm/year. Approximately 46 cm/year. becomes groundwater recharge, 36 cm/year. becomes AET, and the remainder is lost to sublimation. Annual snowfall, AET, and groundwater recharge also exhibit no significant trends during the period 1900–2010.

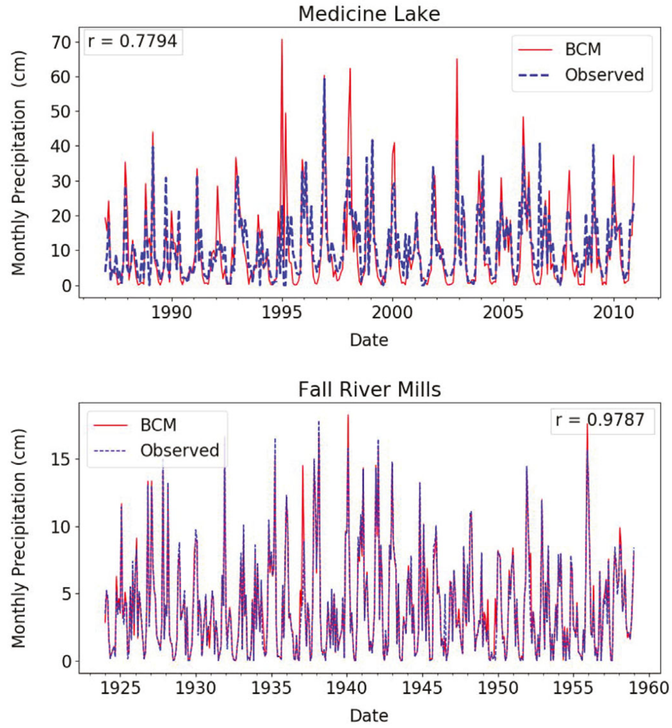

FIGURE 4. Comparison between historic BCM output and historic observational data from Medicine Lake and Fall River Mills.

Thus, precipitation has remained relatively constant throughout the historical period despite the large multi-year fluctuations characterized by Freeman (2001). As evident in Figures S1–S4 of the supplemental information, the 30-year average precipitation for 1920–1950, and consequently other hydrologic variables, appears to be significantly lower than the two subsequent 30-year historical periods. It is also consistent with the general drying which occurred in the 1920s and 1930s. When examining the yearly values over the entire historical period, however, there is no statistically significant trend.

Future Scenarios

GFDL A2. The changes reported in this and subsequent future scenario sections are determined by comparing the 30-year averaged future projected BCM output to the historic BCM output. The GFDL A2 model predicts a warmer and drier climate with

TABLE 3. Watershed-wide averages for selected water balance variables and their percent change from historical to projected scenarios. The historic baseline is taken as the average of the last two 30-year periods (1951–1980 and 1981–2010). For the projected scenarios, the latest projected period is used (2070–2099).

	Average temp		Precipitation			Groundwater recharge			AET			April 1 snowpack		
Scenario	°C	Change in °C	m^3/s	cm/year	% Change	m^3/s	cm/year	% Change	m^3/s	cm/year	% Change	m^3/s	cm/year	% Change
Historic	8.4	—	44.9	90.6	—	24.3	49.1	—	18.1	36.6	—	13.9	28.0	—
GFDL A2	12.2	3.8	38.2	77.0	−15.0	17.7	35.8	−27.1	18.7	37.7	3.1	2.0	4.1	−85.3
GFDL B1	10.6	2.2	40.5	81.7	−9.9	19.2	38.7	−21.2	19.3	38.9	6.5	5.4	10.9	−60.9
CNRM rcp 8.5	13.6	5.2	57.0	115.0	26.9	32.0	64.6	31.6	23.6	47.7	30.3	1.6	3.2	−88.7
CCSM4 rcp 8.5	12.6	4.2	45.1	91.1	0.6	22.2	44.7	−9.0	21.1	42.6	16.5	2.6	5.2	−81.3

Note: AET, actual evapotranspiration.

an increase in temperature of 3.8°C over the basin and a decrease in precipitation of 15%, or 136 mm/year. (Figure S1; Table 3). Both total precipitation and snowfall decrease. There is also a shift in the precipitation type, from snow to rain. Snowfall is correlated with groundwater recharge at each observation point. However, this decreases from the historical to the projected time period (Table 4).

Proportionally more of the total precipitation becomes AET and less becomes groundwater recharge. AET is maintained despite the overall decrease in precipitation, while groundwater recharge decreases by 27% or 13.3 cm/year. by the end of the century.

TABLE 4. Correlation between snowfall and recharge (R-value) at each observation point location for historic (1920–2010) and projected scenarios (2010–2099).

Point	Historic	GFDL A2	GFDL B1	CNRM rcp 8.5	CCSM4 rcp 8.5
0	0.72	0.70	0.62	0.27	0.57
1	0.73	0.55	0.49	0.12	0.52
2	0.72	0.53	0.48	0.11	0.52
3	0.70	0.57	0.57	0.12	0.52
4	0.70	0.53	0.60	0.10	0.50
5	0.66	0.50	0.41	0.05	0.43
6	0.71	0.60	0.57	0.15	0.52
7	0.64	0.47	0.54	0.12	0.47
8	0.79	0.50	0.77	0.30	0.61

GFDL B1. GFDL B1 is characterized by a temperature increase of 2.2°C by the end of the century and a decrease in precipitation of 10% or 8.9 cm/year. (Figure S2; Table 3). Similar to GFDL A2, the precipitation type shifts from snow to rain. Snowfall and groundwater recharge are correlated at each observation point, and the strength of this correlation decreases with time (Table 4). Again, proportionally more of the total precipitation becomes AET and less becomes groundwater recharge. The seasonal pattern of AET, groundwater recharge, and snowfall shifts 1–2 months earlier in the year. The average AET increases while groundwater recharge decreases by 21% or 10.4 cm/year. (Table 3) in the last 30 years of the century.

CNRM rcp 8.5. Unlike the GFDL simulations, the CNRM rcp8.5 simulations suggest an increase in total precipitation at the basin scale (by 27%) accompanied by a substantial increase in temperature (5.2°C: Figure S3; Table 3). The April 1 snowpack decreases as a result of the temperature increase and there is a shift from snow to rain. The portions of precipitation that become AET and groundwater recharge remain relatively consistent from the historical to projected time period. The total AET and groundwater recharge both increase by about 30% (Table 3) in the last 30 years of the century. The seasonal timing of AET, groundwater recharge, and snowfall shifts 1–2 months earlier in the year.

CCSM4 rcp 8.5. CCSM4 rcp 8.5 is characterized by an increase in temperature of 4.2°C and somewhat more variable precipitation relative to the other GCMs (Figure S4; Table 3). Precipitation first increases relative to historical values and then decreases during 2070–2099. The overall precipitation increase at the basin scale is small (0.6% or 0.5 cm/year; Table 3). The April 1 snowpack decreases across the basin and at each observation point. There is a shift in precipitation type from snow to rain, and proportionally more of the total precipitation becomes AET and less becomes groundwater recharge. At the basin scale, groundwater recharge decreases 9% or 4.4 cm/year. from the historical to projected period, while AET increases. The timing of snowfall and groundwater recharge, similar to the other three GCM results, shifts 1–2 months earlier in the year.

DISCUSSION

Historical Observation Data and Historical BCM Output

There is some discrepancy between the BCM historical groundwater recharge and the historical measurements of spring discharge from FRS. The simulated BCM groundwater recharge within the topographic boundary accounts for only 61%–77% of the historical range of observed spring discharge from FRS. The discrepancy between the simulated BCM recharge and observed FRS discharge may be due to the hydrologic parameters used within BCM, or it may be due to groundwater inflow from adjacent areas. The volcanic terranes of the Upper Klamath region and Modoc Plateau in northeastern California performed relatively poorly during calibration of the BCM basin discharge, indicating that the model may be somewhat less well-suited for such terranes (Flint et al. 2013). Additionally, the aquifer system is both large and remote, and there have been few studies on the hydrology, soils, vegetation, and climate of the MLV-FRS system. As a result, the hydrologic values used as input parameters to the BCM model are poorly constrained by observations and may also result in an underestimation of deep infiltration.

It is also probable that additional groundwater enters the aquifer system from outside of the topographic boundaries invoked in this study (see, e.g., Flint et al. 2011). Given the complexity of the faulted volcanic terrane, it is not surprising that recharge could be coming from outside of the topographic boundary. Northeast-southwest trending faults that transect the topographic boundary could route water from outside of the boundary toward FRS. However, since the surrounding area is similarly dominated by recent volcanism, the modeled changes may be representative of adjacent areas, as the mechanisms governing recharge are likely fundamentally similar. In other words, while total volumes differ, it is expected that temporal patterns will persist due to similar hydrology inside and outside of the estimated topographic basin.

Nonetheless, the topographic boundary was used to be consistent with past modeling work of the MLV-FRS system by Burns et al. (2017). Burns et al. (2017), however, used a simplifying assumption to estimate approximate groundwater recharge from each of three contributing areas: Recharge was scaled to be proportional to precipitation to match measured FRS discharge. This approach may be an overestimation of recharge within the topographic domain, while recharge from the BCM output used in our work is possibly underestimated. Additionally, the topographic boundary was used as there is limited data to further conclusively delineate alternative sources of spring discharge for FRS. As such, the following analysis strictly applies only to the portion of the groundwater that is from recharge within the topographic basin.

While basin recharge for the historical BCM output falls short of the measured historical range of observed spring discharge, it was found that, within the topographic basin, output from the BCM shows an average of 54% of annual precipitation falling on the topographic basin becomes groundwater recharge. In contrast, Scanlon et al. (2006), using data from large basins across the globe, would suggest annual recharge of 10 to perhaps 25% of annual precipitation for basins with 80–100 cm/year. precipitation. The results for the MLH-FRS system are consistent with those of Ingebritsen et al. (1994) and Tague and Grant (2009), who found the youngest volcanics of the Cascade Mountains in Oregon to be capable of supporting significant stream baseflow, while older, more weathered volcanics, with lower near-surface permeabilities have a much greater proportion of direct runoff.

Future Scenarios

Temperature increases under all of the four GCMs considered, resulting reduced groundwater recharge for three of the four cases (GFDL A2, GFDL B1, CCSM4 rcp8.5) and increased recharge for CNRM rcp 8.5, in which precipitation also increases substantially (Figure 5). In all cases, increasing temperature has a profound impact on both the precipitation type and AET, with 61%–89% less snow as of April 1 and 3%–30% increase in AET. The latter increases due to the longer growing season associated with higher temperatures, lower surface albedo, and because rainfall tends to be more available for ET than short-duration snowmelt pulses.

Shifts in precipitation type appear to influence the amount of AET in each of the projected scenarios. For example, for GFDL A2, an increase in continuous root water uptake may allow AET rates to be maintained despite a decrease in overall precipitation, because water that would be released rapidly during spring melt and potentially bypass root uptake is more readily available for plants throughout the winter. Periodic infiltration allows replenishment of the limited soil moisture storage reservoir. In contrast, a single snowmelt period will quickly overwhelm the available storage capacity. In the transition between these two extremes, it is likely that multiple melt periods will occur throughout the winter, trending toward periodic rain events. This will increase AET and decrease recharge. Changes in the timing of AET may also be related to precipitation type. In this Mediterranean climate, most precipitation occurs during the winter months. With precipitation falling as rain instead of snow, additional water is available for plants causing more AET to occur earlier in the year.

This earlier onset of AET contributes to the changes in groundwater recharge. For GFDL A2, AET seasonal response begins earlier in the year which allows AET rates to be maintained despite decreases in precipitation. Under the historical climate, winter precipitation would not have been available for AET early in the year, as it was primarily held as snow. Available radiation (for both PET and AET) was low due to snow cover (Milly and Dunne 2020). However, under a scenario with less snow, infiltration is available to fill soil water storage and there is additional radiation to drive evapotranspiration. Winter precipitation is apparently insufficient to exceed field capacity and increase recharge during these periods. Proportionally more of the precipitation is lost to AET and less becomes groundwater recharge, as is observed from the normalized output (Figure S1). The decline in groundwater recharge in GFDL A2 is a function of both greater losses to AET, as its AET seasonal response shifts earlier in the year, and a decrease in overall precipitation.

A somewhat similar phenomenon is observed for GFDL B1. In this scenario, the AET seasonal response also shifts earlier in the year. As in GFDL A2, this is due to increased temperatures and

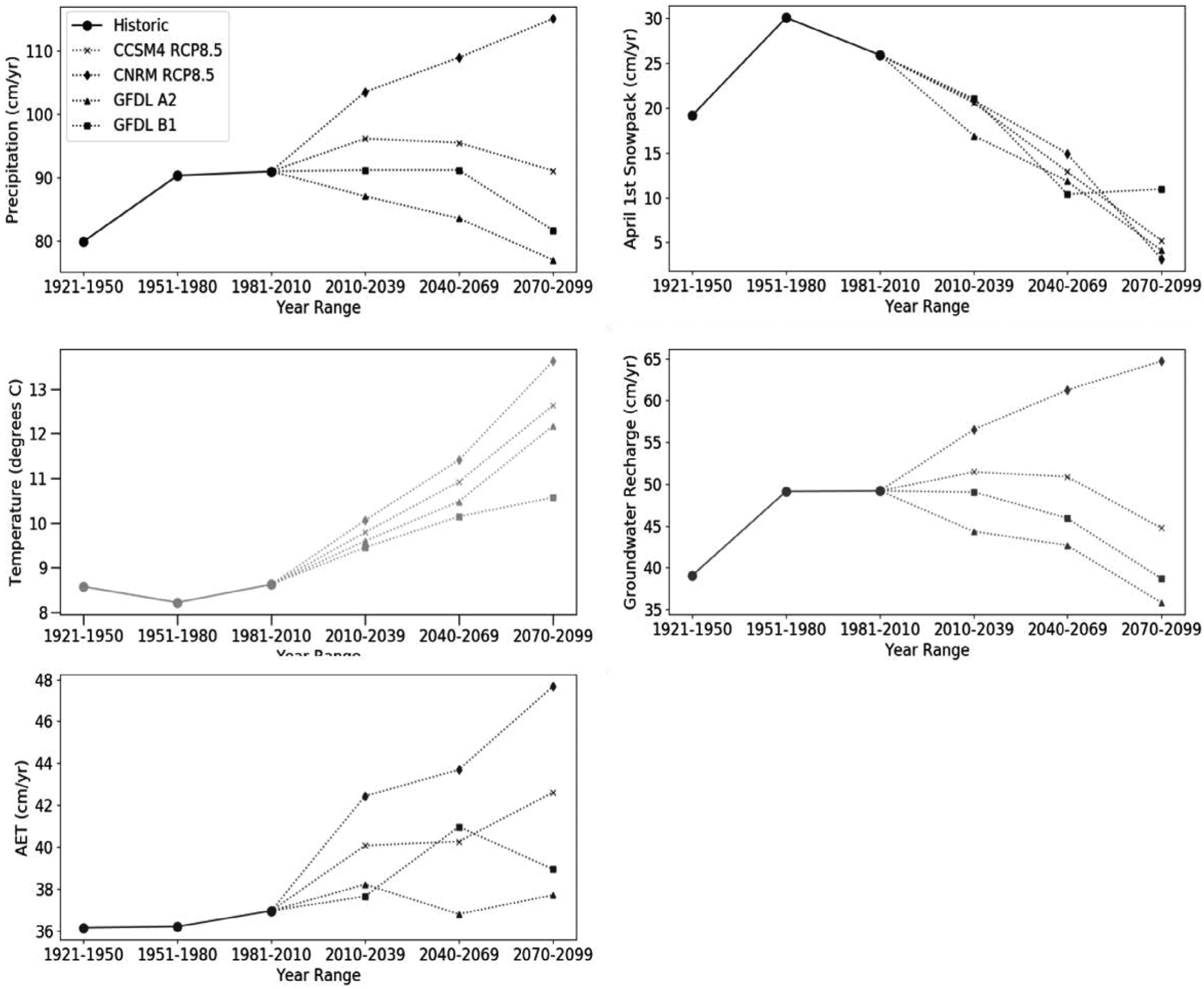

FIGURE 5. Comparison of 30-year average climate variables among each scenario and General Circulation Models (GCM). Precipitation, groundwater recharge, and AET are variable among GCMs. April 1 snowpack decreases and temperature increases for each GCM.

changing precipitation type resulting in a higher percentage of precipitation becoming AET rather than recharge. Precipitation amounts decrease after the latest historical period, and consequently groundwater recharge parallels these changes. The decrease in groundwater recharge for GFDL B1 results from a greater loss of incoming precipitation to AET combined with an overall decrease in precipitation.

For CNRM rcp 8.5, increased groundwater recharge occurs because of increased projected precipitation. Evapotranspiration demands are met despite the significant increase in temperature, and there is also additional water for groundwater recharge. In the CNRM rcp 8.5 scenario, increased precipitation during winter months, when PET is low, results in diversion of water from soil, where storage is now at full capacity, to runoff that — in this study — is counted as recharge.

The causation of changes to groundwater recharge in CCSM4 rcp 8.5 is more complex to resolve because of the variable precipitation projections over time. Groundwater recharge and precipitation are correlated. Thus, changes in groundwater recharge are partially due to similar changes in precipitation. During the historical period, AET is relatively consistent, with only a slight increase at the basin scale. At the start of the projected period, precipitation increases, as does groundwater recharge and AET, while snowpack declines. During this period, normalized AET and groundwater recharge are relatively consistent. This indicates that there is enough water available to accommodate both increasing AET and additional recharge. Later in the projected period, AET continues to increase, while precipitation and groundwater recharge decrease. April 1st snowpack significantly decreases and there is only ~5 cm of SWE. During this period, normalized AET sharply increases and normalized groundwater recharge decreases. This response is likely a function of infiltration mechanisms and changes in precipitation type, as was modeled in GFDL A2 and GFDL B1.

While the relationship between aquifer recharge and annual precipitation in all scenarios was expected, the result that the efficiency of recharge (as expressed by the normalized recharge/precipitation ratio, R/P, Figures S1–S4) decreases under a decreasing precipitation regime (Scenarios GFDL A2, and GFDL B1), yet remains constant under an increasing precipitation regime, (CNRM rcp8.5) is surprising and requires further discussion.

The differences in recharge behavior may be explained by the BCM model's approach to evapotranspiration, which gives priority to AET through its representation of soil water storage. Water available for evapotranspiration is fixed by the difference between field capacity and wilting point. This storage volume is not affected by the climate scenarios but remains fixed throughout all simulations. Because soil moisture storage for ET is always "filled first" in the BCM conceptual model, with any soil moisture above field capacity becoming deep infiltration, a decrease in annual precipitation will primarily affect recharge and runoff. This results in less basin outflow occurring. Total annual AET remains relatively insensitive to moderate decreases in precipitation. Figure 6a documents the similarity in annual AET for an example point located in the snow-rain transition zone in the GFDL A2 scenario of decreased precipitation throughout the warming period. In spite of slow decreases in precipitation and rising temperatures, the total AET grows very slowly. At the same time, recharge and basin outflow decrease significantly.

Only in a scenario where significant warming is coupled with a significant increase in precipitation (CNRM rcp 8.5) does the coupling between AET and precipitation appear to remain constant (Figure 6b). AET occurs earlier in the season (a consistent result for all scenarios) and total annual AET increases substantially over time. In this case, the increasing precipitation maintains soil moisture storage in the range between field capacity and wilting point longer, allowing for greater AET. The significantly increasing air temperatures in this scenario also drive up the estimated PET, increasing cumulative AET. For an increasing precipitation scenario, therefore, the ratio of AET to precipitation remains relatively constant, forcing the normalized recharge ratio to remain relatively constant. In this case, the basin maintains its "efficiency" for both AET and recharge.

Changes in recharge as a function of future climate can also be thought of in terms of the Budyko analysis (see, e.g., Sposito 2017). A nonlinear relationship exists between AET and PET, which are in turn related to the annual precipitation. In the Budyko concept, when precipitation significantly exceeds the PET, runoff will be approximately linearly related to the ratio of precipitation to PET. As the PET approaches and exceeds the precipitation, runoff will decrease more rapidly, with an ever-greater percentage of precipitation being used for AET and ever-decreasing amount available to runoff (or recharge). In the case of the MLS-FRS catchment, the historical ratio of PET (~110 cm) to precipitation (90.6 cm) is slightly greater than unity, indicating a "water-limited system." In Budyko space, the catchment would have a corresponding AET/PET ratio also near unity, however, the MLH-FRS BCM results suggest a value of ~0.40, well below the Budyko curve and consistent with the anomalously high observed rates of groundwater recharge. For an increasingly water-limited catchment, at unity, a decrease in P (and/or an increase in PET) requires that the AET increase

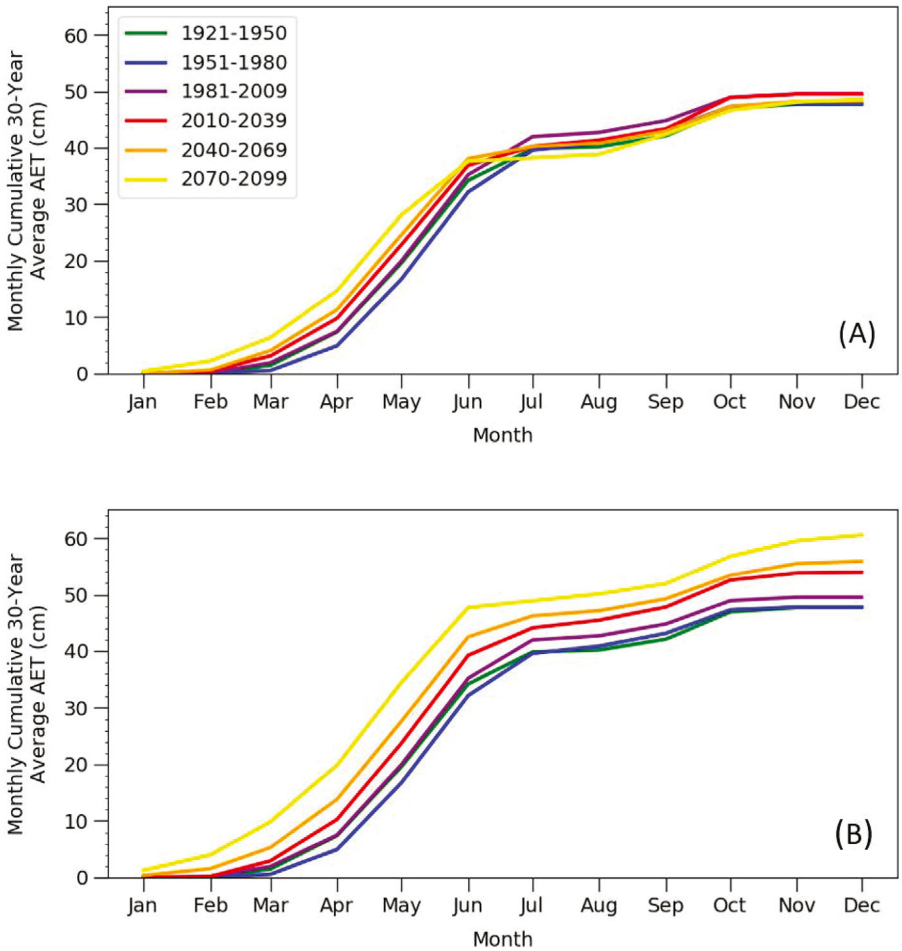

FIGURE 6. 30-year average AET at Point 2 for (a) GFDL A2 and (b) CNRM rcp 8.5.

to approach the Budyko curve. In all scenarios that show future decreases in recharge, the AET/P ratios increase from 0.40 during the historical record to ~0.47 for the drying scenarios. This increase, therefore, requires a decrease in groundwater recharge.

In contrast, under an increasing precipitation scenario (CNRM rcp 8.5), the simulated AET/P ratio at the end of the modeling period only increases from 0.40 to 0.41. In this case, while the actual evaporation does increase, it does not do so at the expense of recharge. Recharge also increases. In Budyko space, the catchment transitions to a more energy-limited state, where runoff (or recharge in this case) increases with increasing precipitation.

While these results suggest that under most warming scenarios groundwater recharge will decrease, several important processes have been ignored in these simulations. These may have far-reaching impacts on the water budget, including spring discharge. Primary among these is the assumption of a stable and stationary vegetation community. Under the modeled changes in temperature and precipitation, it is unrealistic to assume that there are no changes in vegetation. With a warming climate and less snow, understory vegetation may expand, and larger overstory trees may suffer mortality. These changes may actually increase the AET (and decrease recharge) as gains in water yield following thinning or clearing are often short-lived, and re-growth in understory quickly occur. (Keppeler and Ziemer 1990). Since wildfire can serve as a "resetting" factor on landscapes, it may also cause important shifts in the hydrology of the MLV-FRS system. If large portions of the catchment burn more frequently or at greater intensity as a result of warmer temperatures and the longer fire season. In this case, recharge may increase briefly, but then decline below pre-burn conditions as dense understory returns. Ongoing updates to BCM (BCMv8) and other models are addressing these limitations by providing more realistic and dynamic vegetation components for future hydrologic simulation (A. Flint, U.S. Geological Survey [USGS], written communication, 2020).

The simulations also ignore any other human-caused changes in the water cycle. Of particular importance in this study area is future groundwater pumping. Increases in pumping rate and changes in precipitation alone can have a significant impact on depleting groundwater storage without active groundwater management strategies (Massoud et al. 2018). The MLH-FRS aquifer system is currently not regulated under California's Sustainable Groundwater Management Act (Wyckoff and Floyd 2016) and the region contains significant geothermal resources that could lead to changes in pumping upstream of the springs. Water augmentation strategies such as enhanced cloud seeding or other geo-engineering options also could have significant impacts on recharge and spring flow. These could be included in more sophisticated, high-resolution basin scale models in the future.

CONCLUSIONS

All four selected climate projections indicate almost total elimination of annual snowpack in the upper portions of the MLH-FRS watershed, as it shifts to being rain-dominated. They also predict an earlier "green up" of vegetation, with a shift in the onset of evapotranspiration of up to two months earlier, beginning in what is now considered winter. Such changes will have profound impacts on the ecosystems in the watershed, including a significant lengthening of the forest fire season, changes in the food web, species migration and/or invasion and, at the discharging springs, warmer water temperatures as suggested by Burns et al. (2017). This includes downstream ecosystems that depend upon cold-water storage in Shasta Lake to support the endangered Chinook salmon of the Sacramento River.

Three of the four simulations predict that this shift will reduce the fraction of precipitation that becomes groundwater recharge and increase the amount of water lost to evapotranspiration. The average of these three models suggests recharge by the end of the 21st Century decreases by 26% or ~13 cm/year. This would not solely be a function of the magnitude of precipitation, but also its form and timing. In these three scenarios, the net recharge would decrease by 9%–27%, reducing spring flow from the historical mean by the same fraction.

Only in the scenario where significant warming is combined with a substantial increase in precipitation (CNRM rcp 8.5) is increased recharge predicted. This is consistent with a general conceptual model in which additional precipitation leads to more recharge. It is also consistent with the historical record of precipitation and FRS response. However, the scenario's predicted decoupling of the partitioning between AET and recharge in this scenario is surprising, because of the extreme increase in temperature which will likely lead to a corresponding rise in AET.

In summary, a warming climate is predicted to produce measurable changes in the water budget of this volcanic aquifer system. Such variations would be similar to those impacting many other volcanic systems of the Pacific Northwest, South America, and the North Pacific region. The effects of a warming climate should be easily recognizable by the end of the century. It will cause discharge to be warmer and is predicted to result in discharge decreases on the order of 10%.

SUPPORTING INFORMATION

Additional supporting information may be found online under the Supporting Information tab for this article: Spring discharge sampling results, BCM output 30-year average watershed scale summary for each GCM.

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AUTHOR CONTRIBUTIONS

L.K. Mancewicz: Formal analysis; Investigation; Writing-original draft. **L. Davisson:** Conceptualization; Formal analysis; Investigation. **S.J. Wheelock:** Funding acquisition; Project administration; Writing-review & editing. **E.R Burns:** Conceptualization; Formal analysis; Investigation; Writing-review & editing. **S.R. Poulson:** Formal analysis; Writing-review & editing. **S.W. Tyler:** Conceptualization; Investigation;

Supervision; Writing-original draft; Writing-review & editing.

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