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Impact of Water Use by Utility-Scale Solar on Groundwater Resources of the Chuckwalla Basin, CA: Final Modeling Report

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Abstract

The DOE and BLM identified 285,000 acres of desert land in the Chuckwalla valley in the western U.S., for solar energy development. In addition to several approved solar projects, a pumped storage project was recently proposed to pump nearly 8000 acre-ft-yr of groundwater to store and stabilize solar energy output. This study aims at providing estimates of the amount of naturally-occurring recharge, and to estimate the impact of the pumping on the water table. To better provide the locations and intensity of natural recharge, this study employs an integrated, physically-based hydrologic model, PAWS+CLM, to calculate recharge. Then, the simulated recharge is used in a parameter estimation package to calibrate spatially-distributed K field. This design is to incorporate all available observational data, including soil moisture monitoring stations, groundwater head, and estimates of groundwater conductivity, to constrain the modeling. To address the uncertainty of the soil parameters, an ensemble of simulations are conducted, and the resulting recharges are either rejected or accepted based on calibrated groundwater head and local variation of the K field. The results indicate that the natural total inflow to the study domain is between 7107 and 12,772 afy. During the initial-fill phase of pumped storage project, the total outflow exceeds the upper bound estimate of the inflow. If the initial-fill is annualized to 20 years, the average pumping is more than the lower bound of inflows. The results indicate after adding the pumped storage project, the system will nearing, if not exceeding, its maximum renewable pumping capacity. The accepted recharges lead to a drawdown range of 24 to 45 ft for an assumed specific yield of 0.05. However, the drawdown is sensitive to this parameter, whereas there is insufficient data to adequately constrain this parameter.

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1. Introduction

The DOE and BLM identified 285,000 acres of desert land in the western U.S., for solar energy development. Encompassing 17 solar energy zones (SEZs) distributed over six states, this landscape was primarily identified based on solar potential, existing environmental factors and the proximity to electrical grids. A key goal of the federal and state governments was to develop utility-scale solar energy installations (USSIs) on all the SEZs. Since 2008, a number of solar energy plants have been located in the semi-arid desert in Southern California, especially in the Chuckwalla valley, our study area (Figure 1).

However, these projects are likely to pose a range of environmental issues as they are constructed, operated, and ultimately dismantled. Some issues are specific to a particular technology, such as the potential for toxic fluid spills at concentrated solar plant (CSP) facilities, while others stem from the relatively large footprint of USSIs, and from the potential impact to already existing land-uses or basic ecosystem processes. Most notably, the water demands by USSIs in desert landscapes is a significant challenge to meet. Water is essential to constructing and operating USSIs. Besides personal consumption on site, water is also used for dust control, cooling instruments, energy storage, building materials, and for washing mirrors and panels. The operations of these plants requires cooling because the efficiency of the photovoltaic (PV) panels significantly reduces as a function of rising temperature [Akbarzadeh and Wadowski, 1996; Moharram *et al.*, 2013]. Without access to the Colorado River water, USSIs in the valley have to utilize groundwater, which leads to impacts on the regional groundwater balance. Water constraints in the Southwest have halted some proposed solar-energy projects, while others have undergone modifications in scale, technology, and proposed water source. With project delays from modifications, costs to project proponents are significantly greater. In addition, pumped-storage projects [Rehman *et al.*, 2015], specifically, the Eagle Mountain Pumped Storage Project (EMPS) [GEI, 2010a], has been proposed to store solar power as water potential to smooth its supply to the grid. Pumped-storage projects need large amounts of water for initial fill and re-supply. The approved solar plants collectively extract a total of $2.3 \text{ Mm}^3 \text{ yr}^{-1}$ (1850 acre-ft-yr, or afy) from the aquifer, and almost $10 \text{ Mm}^3 \text{ yr}^{-1}$ (8100 afy) has been proposed for the initial-fill phase from the EMPS.

While the demand for groundwater is increasing, the aquifers receive limited recharge, which means only limited water can be renewably extracted from the aquifer. Average annual rainfall in the region is less than 100 mm (3.9 in) and recharge is a small fraction of the precipitation due to the large evaporative demand. Estimating recharge in this desert basin is especially challenging because it only occurs through spatio-temporally sporadic infiltration of runoff along many ephemeral washes descending from the mountains surrounding the valley [CADWR, 1979]. Most rainfall that falls on the valley floor is returned as evapotranspiration (ET). Accurate and long-term data collection of infiltration data in the many ephemeral washes is logistically prohibitive. Data is often scarce and the type of data is varied, a situation representative of many other desert regions.

In the past, the groundwater system was often evaluated in isolated groundwater models such as MODFLOW [Harbaugh, 2005]. In this paradigm, recharge needs to be estimated through independent means, e.g., empirical methods such as Maxey-Eakin, which calculates recharge as

a percentage of precipitation [Maxey and Eakin, 1949], or precipitation-runoff regression [Wilson and Guan, 2004], which is impractical here due to few runoff records. Earlier, environmental impact assessment (EIA) for solar plants in the region was done using a MODFLOW model that assumed Maxey-Eakin-estimated spatially uniform recharge, from 2 to 10 percent of precipitation [WorleyParsons, 2009; GEI, 2010b]. However, this method is known to have strong limitations as it does not consider location and mechanism of recharge [Maurer and Berger, 2006]. The uncertainty bounds were also too large. Physically-based integrated hydrologic models, e.g., GEOTop [Rigon *et al.*, 2006], GSFlow [Markstrom *et al.*, 2008], HydroGeoSphere [Therrien *et al.*, 2006], MIKE-SHE [McMichael *et al.*, 2006], ParFlow [Maxwell and Miller, 2005], PAWS [Shen and Phanikumar, 2010], calculate recharge as an internal flux. Given climate forcing data, they can serve as practical tools for recharge estimation.

Here, an observationally-constrained dual-model approach is used to study the groundwater system in the Chuckwalla basin. This approach integrates a surface-subsurface processes model that simulates both surface and subsurface processes and a groundwater flow and parameter estimation package. The integrated modeling system is constrained by meteorological and soil moisture data collected during the study, and the groundwater calibration is constrained by recharge provided from the integrated model and groundwater head observations. The purpose of this modeling effort is to:

1. Determine the distribution and amount of recharge to aquifers in the part of basin to the west of the Mesa Verde valley.
2. Evaluate the impact of USSI pumping on the water table.
3. Develop a model that can be used by the BLM to manage water resources in the basin.

2. Data and Methods

2.1. Environmental setting of the Chuckwalla Basin

2.1.1 Physiographic properties

The Chuckwalla Basin (6712 km²) is located within the central portion of the Riverside East Solar Energy Zone (RESEZ), to the west of the city of Blythe (Figure 1), between the Mojave Desert and the Colorado-Sonoran Desert. It is a ~40 mile (64 km) long, northwest-trending intermontane basin surrounded by mountain ranges. The basin is relatively flat, with elevations ranging between 300 to 900 ft (~100 to 300 m), while the surrounding mountains are rugged, reaching elevations of more than 2700 ft (900 m) [Rotstein *et al.*, 1976]. The northern portion of the Chuckwalla Valley, lying between the Eagle and Coxcomb mountains, slopes to the southeast. Along this section of the valley ground surface elevations decrease from 820 ft (~250 m) at the base of the mountains to less than 650 ft (200 m) at the center of the valley. The central section of the Chuckwalla Valley is nearly flat, with a slight slope in the south-southeast direction towards Ford Dry Lake located at an elevation of 293 ft (~120 m). The Palen valley slopes to the south-southeast with dry stream channels suggesting a preferred surface flow direction towards Palen Lake located at 426 ft (~130 m). Further east, the Palo Verde Mesa also slopes to the southeast, dropping to 320 ft (~100 m) at the border of the Palo Mesa Verde (Colorado River) Valley.

The mountains are overlain by thin, sandy soils. The valley surficial materials contain (i) coarse, permeable and steep alluvial fans at the mountains feet; (ii) loamy sand alluvium deposits, and (iii) flat, fine-grained lake playa in the center of basin [USGS, 1995]. Desert pavement is seen in various parts of the basin. While the main soil classes are known, their spatial locations are poorly mapped. Desert shrubs and grass are found on the valley floor, and near mountain feet. Flora are found in the Pinto-to-Chuckwalla neck, in the Alluvium fans and along washes (Figure 2b).

2.1.2 Climate

Climate in the region is characterized by high aridity, and low precipitation, between ~3- 5” annually, which is highly variable and occurs mainly in the winter, and to a lesser extent, in the summer. The hot dry summer extends from April through September, when daytime temperatures remain close to 40°C, and often exceed 45°C.

2.1.3 Geology

The basin contains unconsolidated sedimentary material deposited largely during the Quaternary period [CADWR, 1963]. Borehole logs indicate that the sandy alluvium layer, which is interbedded with silt, clay and gravel, varies between 210 m (700 ft) to 366 m (1,200 ft) [CADWR, 1979] in thickness. Most of this material is water lain as alluvial-fan, stream-channel, lake or play deposits, with some sand that was deposited by wind. Underneath the alluvium we find the productive Bouse formation [Metzger *et al.*, 1973], which is composed of basal limestone, clay, silt and sand [Owen-Joyce *et al.*, 2000], in the lower Chuckwalla valley. The Bouse is a sedimentary marine and estuarine sequence of Pliocene age. Many wells logs show that its surface is very flat [Stone, 2006; WorleyParsons, 2009]. However, the Bouse formation was not noted in the upper Chuckwalla valley, to the west of Desert Center in a previous report [GEI, 2010b]. The western end of the Bouse is likely around the Palen Solar project site. A Miocene Fanglomerate aquifer unconformably underlies the Bouse in parts of the basin, but their interface is fuzzy. The alluvium is the main source of water in the upper basin while the Bouse is that in the lower basin, but the water table typically resides in the alluvium.

From their gravity investigations, Rotstein *et al.* [1976], determined that at the northernmost portion of the Chuckwalla Valley, east of the Eagle Mountains, the sedimentary basin (consisting of the alluvium and the Bouse) is more than 0.75 mile (1.2 km) deep. Further south, at locations north of the Chuckwalla Mountains, they estimated the basin to be more than 1.5 km deep. Within the Palen Valley the gravity anomalies they observed suggested that the valley is divided in to two parts, with the southern portion more than 1.5 km deep, and the northern portion shallower and less uniform.

Underlying the sedimentary basin is a pre-tertiary crystalline basement complex composed mainly of plutonic igneous rocks with pendants of metasedimentary and metavolcanic rocks [Rotstein *et al.*, 1976]. In some parts of the basins the basement complex is overlain by volcanic rock of Tertiary age. These rocks are regarded as impervious. The mountains are also thought of as impervious.

2.1.4 Estimates of water balance for the study region in the literature

There is no detailed estimate of recharge in this area. To provide some context, a review by U.S. Geological Survey (USGS) suggested 3-7 percent of precipitation becomes recharge in similar arid basins. A consultant report for a nearby basin (Joshua-Tree area) suggested recharge is 2.8% to 5.2% of precipitation (Whitt and Jonker in [CGB, 2004; Nishikawa *et al.*, 2005a]). Another distributed modeling report estimated a 1.1% recharge-to-precipitation ratio for Joshua-Tree (Table 11 in [Nishikawa *et al.*, 2005b]). An earlier report of the Chuckwalla basin analyzed a large range of recharge, from 1 to 10% of precipitation [WorleyParsons, 2009].

Surface drainage from the eastern portion of the valley is towards Ford Dry Lake while in the western portion flows are towards Palen Lake, both ephemeral playas. Groundwater flow into the basin is thought to be from multiple sources, which include precipitation, and subsurface flows from the adjacent Orocopia and Pinto Valley basins [Godfrey *et al.*, 2012]. Discharge from the basin is through evaporation from Ford and Palen Lakes and deep seepage eastwards (Steinemann, 1989) into the Colorado River via the Palo Verde Mesa Basin. As is with most desert basins, there is insufficient data to assess storage, recharge, discharge and water-use amounts in the Chuckwalla Valley (e.g., [CADWR, 2003]). Godfrey *et al.*, [2012], found a wide range of estimates for groundwater recharge into the valley and suggested that a recharge rate between 3000 and 6000 acre-feet per year (afy) was likely most accurate. Their estimates for discharge to the Colorado River range between 400-1200 afy.

2.1.5. Land Use

About 80% of the land in the RESEZ is administrated by the BLM. The presence of private land including some residences, the state highway, extensive MWD facilities, a small airport, the inactive Kaiser Mine, and more recently, the presence of two USSIs under construction, give the area a more developed setting.

Numerous parcels of private land totaling about 11,000 acres (45 km²) also are scattered throughout the SEZ, with additional private lands in near proximity to its external boundaries. There is also one section of state land surrounded by the SEZ. The city of Blythe, California, on the eastern side of the SEZ, is surrounded by an extensive block of agricultural lands irrigated with water from the Colorado River as well as groundwater.

Desert Center, the single town with the Chuckwalla Basin, is located at the junction of Interstate 10 and Route 177 at an elevation of 656 feet (200 m). It was founded in 1921, but continued to remain sparsely populated until 1942 when over a period of two years, the US army established the Desert Center Air Field to support operations in the California-Arizona Maneuver Area (Bischoff). In the 1980's, Jojoba (*Simmondsia chinensis*) was cultivated in farms around Desert Center. It is a hardy desert shrub, whose oil is used chiefly in cosmetic products. The recent census (2010) has the population of Desert Center at 204, which lives in couple of mobile home parks, and the Lake Tamarisk Community. This retirement community features single family homes, duplexes and mobile homes, situated around the lake and includes a 9-hole golf course.

In the center part of the basin, near Desert Center, there was historically a groundwater decline of more than 40 m (120 ft) caused by groundwater-fed agricultural ventures in the 1980s, which was discontinued in 1986, followed with recovery [WorleyParsons, 2009]. In the East, an

agricultural region near Blythe, CA withdraws a significant amount of Colorado River water and groundwater for irrigation and municipality water use. This groundwater use is connected to the Colorado River, and there is a lack of detailed groundwater use data by different major wells.

2.2. Available basin-specific datasets

2.2.1 Groundwater observations

Groundwater head observations from the California Department of Water Resources (CADWR) have been compiled for Chuckwalla Valley [CADWR, 1963]. Observations are also available from various environmental impacts statements (e.g., [GEI, 2010b; WorleyParsons, 2010; BLM, 2011]). A large part of the CADWR data is for the Blythe region, which is outside of the model domain for groundwater calibration as will be explained in Section 2.4.1.

Groundwater aquifer hydraulic conductivities (K) have been estimated using well specific capacity and drawdown data for a number of existing wells from previous EIAs and the current work. Such estimates are available for both the alluvium aquifer in the upper basin and the Bouse in the lower basin (Figure 1). The BLM has compiled a table of groundwater head observations and conductivity estimates. Multiple wells have records of lithology along the depths, which were also included in the compilations.

We used a gravity-data-derived bedrock topography model to determine the bottom depth of the lower (Bouse/Fanglomerate/Clay) layer (i.e., top of basement bedrock; Figure 4). In the lower Chuckwalla valley, Bouguer gravity data [Mariano *et al.*, 1986] was modeled and calibrated to known bedrock depths from lithology records of wells reaching the bedrock. Parts of the domain was also included in Figure 6 in Appendix C of [GEI, 2010b]. A buried ridge, shown in Figure 4, is set as the western boundary of the Bouse Formation. With respect to the boundary between alluvial materials and Bouse, in the lower basin, we assumed a constant elevation for the top of the second layer, since, as a marine/estuarine formation, the Bouse is observed to be flat. In the upper valley, as there is no clear divide between formations nor detailed data coverage, a constant thickness of ~90 m from geophysical surveys along a transect describes the sandy layer above the lake deposit layer.

2.2.2 Soil moisture and meteorological observations

At the request of the BLM, the Natural Resource Conservation Service (NRCS) established two monitoring stations with the Chuckwalla Basin, in 2011, as part of the Soil Climate Analysis Network (SCAN). The stations collect real-time microclimate and shallow soil moisture data and can be accessed through the NRCS website. In the summer of 2012, BLM and LBNL established a third microclimate station within the valley. This installation also includes real time measurements of soil moisture and temperature up to a depth of 20 inches from the ground surface. Rainfall, temperature and humidity data are available from 5 weather stations in the region along with the two newly installed weather stations. Solar radiation is modeled using sun-earth relationships, with atmospheric transmittance estimated using the algorithm of [Spokas and Forcella, 2006]. (SCAN site URL <https://wcc.sc.egov.usda.gov/nwcc/site?sitenum=2183&state=CA>)

2.3. The modeling approaches

2.3.1 PAWS+CLM

We employ an integrated, hydrologic-land surface-ecosystem modeling system that simulates both surface and subsurface processes. The model, with a full physically-based representation of hydrologic components, is driven by climatic forcing data and simulates recharge as a spatio-temporally dynamic variable. As a result the model does not require estimated recharge as an input. In this section we briefly describe the model and its features. A more detailed description of the model physics are described in Appendix A1.

The Process-based Adaptive Watershed Simulator (PAWS) coupled to the Community Land Model (PAWS+CLM) is a comprehensive model representing the whole land-phase of the hydrologic cycle [Shen and Phanikumar, 2010; Shen et al., 2013, 2014, 2016; Niu et al., 2014; Riley and Shen, 2014; Ji et al., 2015] and transport [Niu and Phanikumar, 2015]. A total of 10 peer-reviewed journal papers, as cited above, have been published, all of them in top-tier hydrology journals. The model efficiently solves the governing partial differential equations for major hydrologic processes, including ponding, overland flow, channel flow, subsurface (soil and groundwater) flow, depression storage/wetlands and the dynamic, multi-way exchanges among these components.

Due to its physically-based flow descriptions, the model is able to capture surface-groundwater interactions, run-ons, and recharge resulting from percolation from washes at a reduced interface area using the leakance concept [Gunduz and Aral, 2005]. The ponding domain contributes runoff to the flow domain while the latter may inundate the former during heavy flows. The flow domain is concentrated in a fraction of the cell, which captures the effects of flow concentration into smaller channels or washes compared to the entire land area.

The computational efficiency of the PAWS model allows for long-term, large-scale simulations and makes parameter estimation and uncertainty analysis feasible. By reducing the dimensionality of the fully three-dimensional (3D) subsurface problem using simplifying assumptions for the Richards' equation, while maintaining the salient nature of the equation, the model significantly reduces the computational demand with little loss of physics. Being a physically-based model, PAWS has been demonstrated to reproduce experimental results, benchmark analytical solutions, idealized test cases, and numerical solutions from models that solve the full 3D Richards' equation [Shen and Phanikumar, 2010]. PAWS compared well with other physically-based models in the model inter-comparison project hosted by U.S. National Science Foundation (NSF) [Maxwell et al., 2014]. The model performed well in simulating the hydrology of a range of watersheds of varying sizes in the U.S. Midwest.

PAWS has been coupled to the NCAR (National Center for Atmospheric Research) Community Land Model (CLM) [Shen et al., 2013, 2016], a process-based land surface model [Niu and Yang, 2006; Lawrence and Chase, 2007; Sakaguchi and Zeng, 2009; Zeng and Decker, 2009; Oleson et al., 2010]. CLM encompasses comprehensive land surface processes including surface heat/momentum/vapor transfer, surface radiation balance, snow/soil heat transfer and freeze-thaw phase changes, photosynthesis and plant growth, carbon and nitrogen fluxes and other ecosystem processes, mostly using process-based descriptions. However, the flow modules in CLM, e.g. surface runoff and channel flow, subsurface flow domain and characterization of geologic and soil properties, all tend to be overly simplified as compared to other components. The coupling of PAWS and CLM brings together surface flow, three-dimensional subsurface flow,

carbon/nitrogen dynamics, energy cycle and land surface processes to offer a higher degree of model realism.

2.3.2 Input to numerical models

A 800 x 800 m² horizontal grid and 40 vertical layers, which are exponentially finer near the surface, were used to discretize the terrain for PAWS+CLM. As described in *Shen et al.* [2014], we incorporated 30 m digital elevation model (DEM), 30 m National Land Cover Database (NLCD) [*Homer et al.*, 2015], SSURGO soils survey [NRCS, 2010] (only for initial setup), National Hydrography Dataset (NHD, only for the Colorado River) [NHD, 2014], and weather data from the National Climate Data Center (NCDC) [NCDC, 2014] in conjunction with our *in-situ* meteorological stations. Most of the valley is barren land but there are also vegetation simulated as tropical brush.

Two layers of aquifers are used in both MODFLOW and PAWS+CLM models. We used results from a bedrock topography model to determine the bottom of the second layer (Bouse/Fanglomerate), which is also the top of impervious basement bedrock. This model was constructed using Bouguer gravity data mentioned above (Appendix 1 in [WorleyParsons, 2009]). A constant elevation for the bottom of the first layer (alluvium) was assumed, reflecting the fact that the Bouse, as a marine/estuarine deposited formation, is observed to be very flat. In the absence of more detailed information, we fitted a straight line to the porosity data and used that to change θ_s with depth.

The model is warmed up with nearly 60 years of simulations by cycling through the climate forcing data from 2002-2016 for 4 times. The extraction of simulated variables starts in 2005 in the 5th cycle.

2.3.3 Soil parameter adjustment

Soil parameters, including vertical conductivity, K_s , and van Genuchten parameters α and n , were adjusted on a trial and error basis for the alluvium and playa deposits based on *in-situ* moisture measurements. We tried to match not only the moisture peaks but also inter-peak minima. After suitable adjustment factors (multipliers and additions) had been found, we applied the parameters to their respective soil regimes. A different set of parameters are derived for the alluvium and the playa using data from the Desert Center site and the Palen Dry Lake site, respectively.

2.4. Calibration of groundwater conductivity using MODFLOW+PEST

The USGS MODFLOW [Harbaugh, 2005] is the USGS's three-dimensional (3D) finite-difference groundwater model. MODFLOW is considered an international standard for simulating and predicting groundwater conditions and groundwater/surface-water interactions. The Model Independent Parameter Estimation package (PEST) [Doherty, 2010; Doherty et al., 2010] allows the estimation of spatially-distributed K fields by adjusting K values so that simulated groundwater head (H) is close to the observed data.

In theory, the MODFLOW is formulated similarly to the groundwater flow model internal to PAWS. However, calibrating the spatial K field requires the PEST package, which is third party software with an existing coupling to MODFLOW. A MODFLOW model including the

Chuckwalla groundwater valleys has been set up for the calibration. MODFLOW+PEST is used to calibrate the K fields to observation wells, incorporating known K values from specific capacity data in the first layer and constraining the value between [1- 30] m/day. For the second layer, as pumping tests are rarer and most estimates are close to 1.5 - 4 m/day, the conductivity is constrained between [0.1 - 6] m/day.

Spatially-distributed PAWS+CLM simulated recharge is stored as input to MODFLOW+PEST. PAWS simulates direct soil-column recharge, run-on infiltration (which include run-off from upstream inundating downstream soil, and percolation beneath ephemeral washes), and mountain-front subsurface flow (the flow entering the valley aquifer from thin soils in the mountains). Recharge water takes many years to reach the water table and the long-time integration required is a major practical obstacle. Therefore, the flux that goes downward through the cell interface 5 meters below ground surface is recorded. The fluxes that moves below this interface is regarded as the soil-column recharge that eventually reaching the water table. Percolation below washes is added to the recharge. Recharge is averaged in time and provided to the MODFLOW model, which has identical horizontal grid spacing.

In addition, PAWS+CLM-simulated inflows from the Pinto sub-valley are used as an inflow condition. The Pinto sub-valley connects to the Chuckwalla basin through a thin neck, and there are no groundwater observations there, so an average K value is used there in PAWS+CLM and it is excluded from MODFLOW+PEST calibration to reduce overfitting.

2.4.1 Source and sink terms and boundary conditions

The Eastern boundary of the MODFLOW model ends at the perimeter of the agricultural zone of Blythe, where USGS well data is available to help us build a fixed head boundary condition. This boundary condition is set because our interest is not on the agricultural zone in the lower basin, but on the upper and middle Chuckwalla valley. It is also out of necessity because modeling the agricultural zone involves modeling irrigation and water intake from the Colorado River, both of which are challenging given the availability of data. The other side of the MODFLOW model is no-flow boundary condition.

Presently, a California State Prison and a resort pump about 7100 m³/day (~2100 afy) and 3684 m³/day (~1090 afy) from the Bouse formation and the alluvium, respectively [WorleyParsons, 2010] (Figure 1). These sink terms have existed over two decades and they have been included for calibrating the steady-state model. Some other minor pumping terms are not simulated. For future projections, we have added approved solar plants and the proposed Eagle Mountain pumped-storage project (EMPS), as in Table 2. Their water use values are obtained from their respective project Environmental Impact Assessment (EIA) report.

2.5. Model rejection and consideration of uncertainty

Since there are only two sites that allow the characterization of soil parameters, there is a great deal of uncertainty with respect to these parameters. Earlier research has found that soil parameters can vary significantly depending on the age of soils [Young *et al.*, 2004; Mirus *et al.*, 2009]. Here a dual-model approach is introduced to reduce the uncertainties (Figure 3). An ensemble of simulations with perturbed soil parameters that generate a range of different recharges are examined. The parameters are perturbed simultaneously using global multipliers to

generate recharges from high to low, as high recharge is expected to lower the impact of pumping. Preliminary sensitivity tests allowed determination of sensitive parameters and in which direction they influence recharge. Then, a plausible range is determined for each parameter. The parameters are set to change from generating the smallest recharge to the largest, according to the schedule provided in Table 3.

In PAWS+CLM, there are three possible recharge sources: run-on infiltration in the washes, mountain-front subsurface flow, and direct soil column recharge. The long period (many years) required for recharge to reach the water table is a major practical obstacle. Therefore, we recorded the flux that travels downward through each cell interface five meters bgs. The flux that passes below this interface was regarded as the recharge that eventually reaches the water table. While at local scales there may be (discontinuous) clay layers that impede vertical flow, we are concerned with large-scale, long-term-average fluxes. We also added mountain-front subsurface inflow to the recharge. Time-averaged recharge was provided to the MODFLOW model, which has identical horizontal grid spacing as the PAWS+CLM model. MODFLOW+PEST was used to calibrate the K fields to water-table levels in observation wells. Constraining the possible range of K is important for reducing overfitting, which means K is adjusted unrealistically to fit the noise rather than true signal. For the top aquifer layer, we added pumping-test-estimated K as known values and constrained K between [0.1-30] m/day. For the second layer, as pumping tests are rarer and most K estimates are close to 1.5-4 m/day, we constrained the conductivity to [0.1-6] m/day. We used a warm-up period of 4 years before extracting recharge.

The MODFLOW-PEST calibration penalizes (or, in term terminology, regularizes) large local variations of the K field, in order to reduce overfitting [Doherty, 2003]. In addition to maximizing the regularization parameter, here a detrended local variation metric, σ_l , is utilized to quantify the magnitude of local variation. First, a bi-quadratic surface is fitted to the calibrated K field. σ_l is then the standard deviation of the residual (calibrated K minus the fitted surface). Models with large σ_l are rejected.

We rejected a recharge if the calibrated head differed significantly from observed head despite the calibration, assessed using a z-test of the mean. To be lenient, we use 4 times of the residual variance from the best-calibrated case, $4 \cdot var_{min}$, for the z-test. We conducted chi-squared test on the residual variance and regression test with elevation as a predictor. If residuals are correlated to elevation, there is a regional pattern to the error which suggest the model/recharge is flawed. Furthermore, when the calibration overfits to data, it tends to force local K adjustment leading to large small-scale variations. To detect overfitting, we fitted a bi-quadratic surface to the K field, and calculate the standard deviation for the K residual from the surface. Five calibrations were conducted for each recharge case shown in Table 3, using different initial guesses of K.

Three more sources of uncertainties must be considered: the specific yield of the alluvium (S_y), the specific storage of the Bouse (S_s), and the uncertainty associated with the PEST calibration itself. As the calibration is for steady state simulations, neither S_y nor S_s influences the calibrated K. To address the third source of uncertainty, 5 calibrations were conducted using different initial guesses for each recharge. Altogether, 60 models were calibrated and passed through the rejection tests. For transient (forward) simulations, the plausible ranges of S_y and S_s

were included in future projections. Three values were tested for S_y : [0.05, 0.10, 0.15]. The small value of 0.05 was estimated for Desert Center [WorleyParsons, 2009]. However, other estimates place the value around 0.15. For S_s , earlier studies for aquifers in this area have bounded the range in $5e-6$ to $1e-4$, so three values were tested in this work: [1×10^{-6} , 5×10^{-6} , 5×10^{-5}].

3. Results

In the following, the comparison with the observations are first presented. Then, the effects of perturbed recharge are examined using the observed groundwater head. The retained simulations are then used to estimate the bounds for the impacts of the solar plant pumping.

3.1. Soil moisture comparisons

After soil parameters are adjusted, the Richards'-Equation-based PAWS+CLM model was able to match the soil moisture time series at both stations (Figure 5). The calibrated vertical saturated conductivity (K_s) values are similar on both sites (Table 1), taking a value of around 0.1 m/day, which is lower than the expected range from sandy soils.

To provide some context, the calibrated K_s value is near the lower range of the values reported for Mojave desert soils, which ranges between 0.07 and 350 m/day from old to new soils [Young *et al.*, 2004]. However, the observed near-surface moisture always stayed in a low range, <0.15 . In this range, the adjusted van Genuchten parameters are more influential than K_s for estimating infiltration and recharge, and might compensate for an inaccurate estimate of K_s . In addition, desert soils that are under the influence of extreme heat and aridity are hydraulically different from soils from elsewhere with the same sand/clay/silt composition [Young *et al.*, 2004; Mirus *et al.*, 2009].

3.2. Assessing and rejecting perturbed simulations

Five of the recharge fields, which are near either the high end or the low end of recharge rates from the experiments, were completely rejected due to their inability to fit the groundwater head (Tables 4 & 5). Figure 7 presents the observed vs calibration groundwater head for some examples of accepted and rejected simulations. Experiments #1 through #5, rejected by all tests, over-estimate the groundwater head (Tables 5), suggesting their recharge rates are too large. On the contrary, experiment #12 under-estimates groundwater head regardless of calibration, suggesting its recharge rate is too low. For simulation #5, we can see the calibration algorithm either over-estimates the head, or generates a large local variation in K in the middle of the domain (Figure 7 and Figure 8), which is indicative of incorrect recharge leading to overfitted K fields. The *z-test* alone was able to rule out most of the cases from recharges 1-5 & 11-12. The elevation-regression test and detrended K variation by themselves rejected some cases for recharges #6-#10. The variance test by itself did not reject any cases. One calibration using recharge 11 was considered a borderline case.

Using recharge generated by the default parameter set above, the spatially-distributed hydraulic head compares well with the observations (Figure 7), and the resulting K field is smooth. Overall the magnitude and variation of K conform to our knowledge of the area. In

addition, the simulated groundwater contour (Figure 9) is in agreement with trends shown in earlier studies (WorleyParsons 2010).

3.3. Water balance of the basin under uncertainty

The lower bound estimate of total inflow is 3.07 mm/yr, between #10 and #11 (7,107 afy, see Table 4 caption). The upper bound of our inflow estimate is 4.99 mm/yr (11,564 afy), the average between #5 and #6. Our estimates range from 3.4% to 5.6% of precipitation. In the literature, recharge estimates in arid and semi-arid basins in the southern Mojave range from 3%-7% of precipitation (Stonestrom *et al* 2007). Reports in nearby basins range from 2.8%-5.2% (Whitt and Jonker in CGB 2004), down to 1.1% (Nishikawa *et al* 2005). Simulated recharge is focused on ephemeral washes and alluvial fan on mountain fringes (Figure 6a). As runoff reaches the alluvial fans, the thick sediment provides more volume for storage and infiltration.

The proposed withdrawal during the initial-fill phase of the EMPS Project (13,140 afy, from Table 2), is larger than the upper bound of the recharge estimate. Even if we assume there is no outflow to the Mesa Verde Valley, for the purpose of estimating maximum renewable extraction, groundwater storage will likely decline significantly during the initial-fill stage. If the initial fill is evenly distributed into 20 years, the annualized pumping is still more than the lower bound estimate. Therefore, the system may be nearing, if not exceeding, its full sustainable groundwater production capacity after the EMPS initiates.

3.4. Projections of the impacts of pumping on groundwater sustainability

Recharges from the retained simulations and their respective calibrated K fields were used to estimate drawdown in response to new solar plant groundwater pumping. At EMPS, the largest drawdown occurs at the end of the initial fill period and has a range of 24 to 35 ft when $S_y = 0.05$ (Figures 10a and 11). Without rejection of overfitted simulated recharge rates, this range would have been 31 to 46 ft. The reduction of uncertainty depends on the site, as Desert Sunlight sees a large reduction (Figures 10b) while Genesis almost sees no effect (Figures 10c).

For EMPS at $S_y = 0.05$, the drawdown reduces by 10~12 ft within one year after the initial-fill phase, then linearly declines over the next 16-year re-fill period. Heavy pumping induces a large hydraulic gradient and a deep cone of depression. Once the pumping ceases, the large aquifer transmissivity lead groundwater flow to rapidly fill the cone. The water table then gradually declines during the project's re-fill phase. After the simulated cessation at the EMPS, the drawdown can reduce by 4 m in one year, and at the end of simulation the water table recovers to 6-7 m from initial values. This pattern suggests that the system may be able to recover fast from the assumed pumping, but the recovery speed does not imply it can go to pristine conditions. If there is a boom of projects pumping groundwater, groundwater levels will not be sustainable, as can be seen from the mass balance analysis. We also note the specific yield has larger impact than recharge (Figures 11).

If, as in conventional methods, we had assumed a uniformly distributed recharge before calibrating K, the results would have been much different, even with the same total recharge. The uniform recharge tends to over-estimate head in the lower basin (Figure 12a). While the RMSE is not very high, the resulting K fields have higher local variation. Also, the retained range of pumping drawdown for the EMPS is larger (Figure 12c-f). However, such an effect is not

spatially homogeneous, as at the Genesis site uniform recharge leads to under-estimation of pumping drawdown. This difference is because the EMPS Project is closer to mountain-front and wash recharge. Since Genesis is located in the valley center and far from recharge locations, a uniform recharge will over-estimate the recharge near the site. Therefore, the impacts of the uniform-recharge assumption cannot be generically described.

4. Discussion

Water managers may find fast water table recovery to be re-assuring and use it as a guideline to manage water. However, as heavy pumping induces large hydraulic gradient. It is likely always followed by rapid recovery after cessation, even if pumping rates far exceed recharge and result in large storage loss. Therefore, the speed of recovery itself cannot indicate sustainability as the water may not recover to before-pumping levels.

Drawdown is a function of recharge, pumping rates and aquifer transmissivity. The Calibrated K fields significantly influences possible drawdown and recovery, which is also why an ensemble of simulations are required and the assimilation of groundwater head observations is critically important. Well drawdown data helped constraining the K calibration process. Without these constraints, the estimated K and drawdowns will have a larger uncertainty bound. We also see that our estimates are bounded by the recharges that have been rejected, which indicates that there is a good likelihood the true recharge will be bounded in our accepted simulations (red lines).

The cones of depression after 20 years of pumping (before recovery) in the two layers of the MODFLOW model for the largest and smallest accepted recharges are presented in Figure 11. Assuming a $S_y = 0.05$, the cone of depression reaches 18 ft in the second layer even toward the mouth to Pinto. The drawdown is more sensitive to assumption of S_y than to the recharge employed. If $S_y = 0.2$, the cone of depression is much shallower. On the other hand, the drawdown is insensitive to specific storage in the range tested. The curves look mostly the same so they are omitted here.

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Figure 1. (a) Satellite image of the Chuckwalla basin and the modeling domain. The MODFLOW+PEST (Section 2.4) model domain is smaller than PAWS+CLM (Section 2.3.1) model domain. A fixed head boundary condition (green line), which was constructed by groundwater head observations, is set to encompass the agricultural region, so that dynamics east to this line does not impact the calibration. The “water balance budget mask” refers to the area over which mass balance is reported in Table 4. The reason the fluxes are reported for this region is that the agricultural region in the East and Pinto valley in the Northwest have no effect for groundwater model calibration; (b) terrain map showing locations of observations, soil moisture stations, existing K estimates, existing pumping sources (the state prison, Lake Tamarisk resort, and solar plants), and the Eagle Mountain Pumped Storage project. There are no perennial streams, lakes or wetlands (WorleyParsons 2009).

Figure 2. (a) A well in the basin surrounded by soils with visible desert pavement; (b) A picture taken within an alluvial fan looking upslope to higher elevations. Note that vegetation is visibly denser on the alluvial fan. Washes are also visible; (c) A zoomed-in satellite image of the Chuckwalla Valley, with annotated patterns of ephemeral washes and vegetation.

Figure 3. The proposed dual-model approach. We collected 4 years of field soil moisture measurements to estimate base soil properties. We then generated a range of recharge estimates by making perturbations to the calibrated soil parameters. Groundwater observations are used to constrain K in MODFLOW+PEST and, more importantly, retain or reject some of the recharge estimates. The retained recharges were used to produce the range of possible drawdowns induced by solar plant pumping, given the available information.

Figure 4. Depth to the basement bedrock map. The black thick line indicates a buried ridge that is visible in Figure 6 of Appendix C in (GEI 2010) and multiple well-based transect profiles. In the lower valley, the bottom of the Bouse/Fanglomerate layer in the lower basin is available through gravity modeling. This model was constructed using Bouguer gravity data (Mariano *et al* 1986) and calibrated to bedrock depth measured from wells reaching the bedrock (Appendix 1 in (WorleyParsons 2009)). North to the buried ridge, Bouguer gravity data is also available from GeoPentech, which was reproduced in Figure 6 of Appendix C in (GEI 2010).

Figure 5. Soil moisture comparisons at the Desert Center site (upper three panels) and the Ford Dry Lake site (lower three panels). At the Desert Center site, the 8-inch probe appears to malfunction as it records moisture rises that are much larger than those detected at the surface. At 20-in depth, while the timing of the moisture wave is not completely correct, the amplitude of seasonal fluctuation is similar between observed and simulated.

Figure 6. Simulated recharge maps from the base parameter set (experiment #6) (a) The total recharge, consisting of run-on percolation, soil matrix recharge, and mountain front subsurface recharge (Q_{MSub}). Note the percolation through flow paths along washes in annotated regions A and B, which agree with the vegetation pattern seen from Satellite images in Figure 2. Recharge also occurs at the alluvium that is at the feet of mountains; (b) mountain-front subsurface recharge, which is lateral subsurface flow from thin mountain soils. Note that Q_{MSub} only occurs at the interface between mountain and valley. The cross hatched areas are the bedrock / mountain exposures. The Palo Verde Mesa Basin / Colorado River Floodplain (white area in the east) are not considered in the calibration. The Pinto Valley (white area to the northwest) is outside of the groundwater modeling domain, however, groundwater inflow to the Chuckwalla Basin is considered.

Figure 7. Observed vs. calibrated groundwater head for several recharges. "rch6-c3" means the calibration realization 3 (with a particular initial guess for K) using recharge from simulation #6. Other data series are defined similarly. We can see that with recharge #6 the calibrated head matches very well with the observed after calibration, with only a few meters of differences at the maximum for each data point. However, for recharge #4, the groundwater head is always over-estimated, regardless of the calibration effort and the initial guesses for K. Recharge #5 also tend to slightly over-estimate head in the lower basin (around observed head = 80 m). While the over-estimation is reduced in some calibration runs (rch5-c5), the K field tends to be overfitted. Recharge #11, on the other hand, is apparently under-estimated.

Figure 8. Maps of calibrated first-layer conductivity for (a) the calibrated K for the base case; (b) the calibrated K for the case using recharge #5. Although for this case the calibrated groundwater head matches decently with the observed (rch5-c5 in Figure 7), the K field has large local variations, suggesting overfitting. Note that areas close to the Eastern boundary is not considered in model rejection due to limited observations. There are no used observations in the agricultural region due to the fixed head boundary condition.

Figure 9. Simulated groundwater head map. This map does not include the effects of the assumed solar plant pumping. The Palo Verde Mesa Valley groundwater basin near Blythe (to the East of the mountain mouth) is controlled by the fixed head boundary condition.

Figure 10 (a) Influence of assumed pumping on the water table at the Eagle Mountain Pumped Storage (EMPS) Project pumping site. The first 4 years is the initial fill phase. 5-20 years is the re-fill period. The pumping is terminated after 20 years to examine the rate of recovery. The red lines indicate accepted recharges. The magenta lines are “less-likely” recharges that have higher error statistically but could not be completely rejected. The drawdown is sensitive to the specific yield of the alluvium layer. It is not sensitive to the specific storage (S_s) in the range tested. The gray lines are the rejected recharges indicating the extent of uncertainty facing the prediction if no model rejection was applied. Note that the model rejection procedure reduces the uncertainty for EMPS. After 20 years of pumping, the maximum decline is likely around 35 ft for the case $S_y = 0.05$. (b) Same figure as in (a) but for the Desert Sunlight solar plant. The model rejection greatly reduces the uncertainty at this site; (c) the same Figure as (a) but for the Genesis solar plant. This site has more uncertainty than other pumping sites and the model rejection did not effectively reduce the uncertainty.

Upper 4 panels: Model layer 1

Lower 4 panels: Model layer 2

Figure 11. The cones of depression formed by the drawdowns (groundwater head from simulations without pumping minus that with pumping). To the west of desert center, the first model layer, with a thickness of ~50 ft, becomes dry after pumping. Rch6 is the highest accepted recharge while Rch10 is the lowest accepted recharge, which results in a deeper cone of depression. The drawdown is more sensitive to the assumption of S_y than the recharge employed.

Figure 12. Comparing model-estimated recharge vs. uniform recharge for $S_y = 0.15$. No model rejection is applied to uniform-recharge simulations as they would have been rejected. Dashed lines indicated distributed-recharge simulations that have been rejected. (a-b) calibrated vs observed groundwater head: in the lower basin, uniform recharge tends to over-estimate groundwater head, which is due to under-estimating impacts of pumping; (c-f) projected impacts of pumping at EMPS and Genesis: uniform recharges produce a wider range of projected drawdown at EMPS but smaller range and less drawdown at Genesis. Not that the bottom 2 lines in the distributed-recharge case at Genesis have been rejected.

Table 1. Calibrated soil parameters on two field sites. K_s , N , α are kept constant throughout different depths. θ_r is adjusted at different depths to better fit the data.

Ford Dry Lake						
Depth	K_s (m/day)	N (-)	α (m^{-1})	θ_r (-)	θ_s (-)	λ (-)
2 in (layer 7)	0.1	1.6	4	1.00E-10	0.3805	-1.2155
4 in (layer 9)	0.1	1.6	4	1.00E-10	0.4221	-0.1059
8 in (layer 10)	0.1	1.6	4	0.02	0.4221	-0.1059
20 in (layer 12)	0.1	1.6	4	0.05	0.4221	-0.1059
Desert Center						
Depth	K_s (m/day)	N (-)	α (m^{-1})	θ_r (-)	θ_s (-)	λ (-)
2 in (layer 8)	0.12	1.8	3.2	1.00E-10	0.3877	-1.3
4 in (layer 11)	0.12	1.8	3.2	1.00E-10	0.3824	-1.3
8 in (layer 13)	0.12	1.8	3.2	0.025	0.3969	-1
20 in (layer 15)	0.12	1.8	3.2	0.06	0.3969	-0.8

Note: the van Genuchten water retention formulation is written as:

$$S = \frac{\theta(\psi) - \theta_r}{\theta_s - \theta_r} = (1 + |\alpha\psi|)^{-(N-1)/N}$$

where S is relative saturation, ψ is the pressure head, θ is the moisture content, θ_r is the residual moisture content, θ_s is the saturated moisture content (porosity), and α and N are parameters. The unsaturated conductivity is calculated by:

$$K_z(S) = K_s S^\lambda \left[1 - (1 - S^{N/(N-1)})^{(N-1)/N} \right]^2$$

where K_s is the saturated conductivity and $K_z(S)$ is the soil unsaturated vertical hydraulic conductivity at the relative saturation S .

Table 2. Pumping sources from the solar plants

	acre-ft-yr	m3/day
Genesis	1525	5154
Desert Sunlight	52	176
Desert Harvest	53	179
Palen	220	743
Eagle Mountain (1-4 yrs)	8100	27373
Eagle Mountain (5-20 yrs)	1800	6083
Eagle Mountain (21-24 yrs)	0	0
Total (period 1, 1-4 yrs)	9950	33625
Total (period 2, 5-20 yrs)	3650	12335
Total (period 2, 21-24 yrs)	1850	6251
20-year annualized pumping	4910	16590

Table 3. Parameter perturbations for the numerical experiments. These changes are applied as multipliers to additions to default values. N/C means no change is applied. Generally, going from Sim #1 to Sim #11, the resulting recharge decreases.

Parameter	Ks	α	K	K_S for mountain areas	Deep layer porosity for non mountain areas	N
'sim#1'	$\times 10$	$\times 1.5$	$\times 3$	N/C	$\times 1.2$	N/C
sim #2'	$\times 8$	$\times 1.4$	$\times 2.5$	N/C	$\times 1$	N/C
'sim#3'	$\times 6$	$\times 1.3$	$\times 2$	N/C	$\times 1$	N/C
'sim#4'	$\times 4$	$\times 1.2$	$\times 1.5$	N/C	$\times 1$	N/C
'sim#5'	$\times 2$	$\times 1.1$	$\times 1.25$	N/C	$\times 1$	N/C
'sim#6'	$\times 1$	$\times 1$	$\times 1$	N/C	$\times 1$	N/C
'sim#7'	$\times 1$	$\times 1$	$\times 1$	=1.6 m/day	$\times 0.8$	N/C
'sim#8'	$\times 0.75$	$\times 0.85$	$\times 0.5$	N/C	$\times 0.7$	N/C
'sim#9'	$\times 0.5$	$\times 0.7$	$\times 0.3$	N/C	$\times 0.55$	N/C
'sim#10'	$\times 0.5$	$\times 0.7$	$\times 0.3$	=1.6	$\times 0.45$	N/C
'sim#11'	$\times 0.5$	$\times 0.7$	$\times 0.3$	N/C	$\times 0.55$	-0.2

Table 4. Mass balance (in afy) and model acceptance status from the perturbed simulations. These fluxes are summed up for the “water balance budget mask” area in Figure 1a. ‘mfront’ means mountain-front subsurface recharge. A recharge is rejected if none of the 5 realizations was retained. Taking from #6, the upper bound of recharge is estimated as 11,564 afy. Because only one case from recharge #11 is narrowly retained, we take the average of the #10 and #11 to calculate the lower bound of recharge is estimated as 7,107 afy. Recharge #12 is the same simulation as #11, but the recharges are 80% of #11.

Recharge #	Soil & wash recharge	Pinto underflow	mfront	Total inflow	Prp	Annualized pumping	Results
'sim#1'	18509	2236	298	21043	205,376	8101	Reject – always overestimate head
'sim#2'	18564	2335	316	21215	205,376	8101	Reject – always overestimate head
'sim#3'	16908	1777	241	18926	205,376	8101	Reject – always overestimate head
'sim#4'	15051	1212	223	16486	205,376	8101	Reject – always overestimate head
'sim#5'	12744	1012	225	13980	205,376	8101	Reject—either GW is over-estimated or K variation is too large
'sim#6'	10478	877	210	11564	205,376	8101	Accept 2 runs
'sim#7'	10594	825	182	11602	205,376	8101	Accept 1 run
'sim#8'	9487	522	173	10183	205,376	8101	Accept 1 run
'sim#9'	8539	372	136	9047	205,376	8101	Accept 3 runs
'sim#10'	7899	388	107	8394	205,376	8101	Accept 2 runs
'sim#11'	5309	320	191	5820	205,376	8101	Mostly rejected. One narrow retention retained as “unlikely”
'sim#12'	4247	320	191	4758	205,376	8101	Reject – always underestimate GW head

Table 5. Detailed metrics for model rejection. Green-filled cases pass all statistical tests. As shown in the legend, for every calibrated field (12 recharges, each with 5 calibration realizations), the numbers shown for each field are mean bias (upper left) of residuals (calibrated-observed head), root-mean-squared error (rmse, upper right), p -value for the elevation regression test (P_E , lower left), and standard deviation of the detrended K residuals (σ_K , lower right), respectively. To be lenient in retaining simulations, we implement relaxed rejection criteria for 3 statistical tests, using a confidence level of 2% and an assumed variance that is 4 times that of the best calibrated field (var_{min} , from recharge #10, realization 2). A field is rejected if one of the following is true for the calibrated head residuals: (a) the residuals fails the z-test for zero mean (upper left cell is then flagged red); (b) data rejects the null hypothesis that the residual variance is smaller than $4 \times var_{min}$ using a one-sided chi-squared test (upper right cell is then shallow blue); (c) the p -value for regressing residual to elevation (lower left cell is yellow); (d) $\sigma_K > 4.5$ (lower right is flagged dark blue). The hatched case, Rch#11 case 5, is a “border-line” case. It is the only retained case from Recharge #11 and it would have been rejected if, instead of 4, we had used 2.25 times var_{min} . Therefore we label it as “unlikely”. We tried increasing soil conductivity on the mountains in simulation #7 but it was more often rejected.

Recharge	Realizations									
	<i>i</i>		<i>ii</i>		<i>iii</i>		<i>iv</i>		<i>v</i>	
1	13.4	14.0	5.6	13.5	11.4	12.8	35.7	36.4	6.9	13.2
	0.0%	6.1	0.0%	6.5	0.0%	7.4	0.0%	2.1	0.0%	8.1
2	86.9	87.1	49.9	51.4	5.9	13.9	45.5	51.2	12.2	13.6
	6.0%	1.3	0.0%	1.3	0.0%	6.6	0.0%	4.1	0.0%	7.8
3	34.6	36.3	4.4	11.0	34.6	36.3	8.9	10.6	8.9	10.6
	0.0%	6.8	0.0%	6.6	0.0%	2.3	0.0%	7.4	0.0%	7.4
4	1.1	11.0	5.2	7.3	1.7	10.4	14.0	15.4	7.1	7.9
	0.0%	6.1	0.0%	6.8	0.0%	6.0	0.6%	6.2	0.3%	6.8
5	11.3	15.3	6.1	8.2	4.5	5.9	6.2	18.9	3.1	4.5
	0.0%	1.3	0.0%	3.6	0.1%	3.7	0.0%	2.2	0.1%	6.8
6	2.0	3.8	1.1	2.7	1.1	2.6	1.1	2.6	1.0	2.6
	0.6%	3.9	20.4%	5.5	2.5%	4.6	8.6%	3.9	4.8%	4.3
7	4.5	5.1	6.7	7.6	1.1	2.6	-0.2	3.5	3.7	4.5
	92.4%	2.0	5.6%	2.2	5.2%	4.3	0.1%	4.5	17.7%	3.6
8	2.8	3.6	0.6	2.3	0.1	4.0	2.6	3.9	-0.3	3.1
	15.3%	2.6	21.7%	3.8	0.0%	5.3	0.1%	5.2	0.2%	5.1
9	0.2	2.2	1.9	3.1	0.4	2.3	1.8	3.2	0.2	2.3
	29.8%	3.9	1.2%	2.2	24.4%	3.4	2.7%	7.9	28.3%	4.1
10	0.1	2.2	0.1	2.2	3.8	5.4	-0.5	2.5	1.7	5.4
	13.4%	3.9	18.9%	2.1	13.3%	3.8	1.0%	4.0	2.4%	6.3
11	-11.5	14.6	-9.0	10.8	-11.7	15.1	-8.2	9.8	-1.6	4.5
	0.0%	4.1	0.0%	4.6	0.0%	5.7	0.1%	5.5	14.7%	3.8
12	-6.5	7.4	-11.1	13.3	-15.5	19.8	-6.4	7.0	-14.7	18.9
	15.2%	5.0	0.1%	3.5	0.0%	7.8	93.0%	4.4	0.0	6.9

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Appendix A1. Descriptions for the flow modules in the PAWS+CLM model

Figure S1. Sketch of PAWS+CLM hydrologic and ecosystem processes (reprinted from [Shen *et al.*, 2016]). Vegetation photosynthesis, energy cycle, soil freeze-thaw, ET and carbon/nutrient cycling are provided by CLM, while hydrology is provided by PAWS; (b) multi-way exchange between the flow domain, ponding domain, soil water and groundwater. Surface water is divided into the flow domain, which can circulate laterally, and the ponding domain, which is connected to the soil matrix. The ponding domain contributes runoff to the flow domain while the latter may inundate the former during heavy flows. The flow domain is concentrated in a fraction of the cell termed f_{sat} . Flow domain water can evaporate at potential rate as calculated by f_{sat} multiplied by the Penman-Monteith equation, and can also percolate through the wash bed which will eventually reach the groundwater using the leakance concept [Gunduz and Aral, 2005].

Figure S1 illustrates the model structures of PAWS+CLM and Figure S2 shows the program structure. The flow module starts with throughflow + stem flow + snowmelt as precipitation input, and converts the evapotranspiration term computed by CLM to a sink. The flow modules in PAWS include physically-based two-way exchanges between hydrologic compartments. In PAWS+CLM, the landscape is discretized using a structured grid, with channels as spatially explicit (represented by specialized model elements with parameterized attributes and geographic information), vector objects that exchange water with the unconfined aquifer as governed by the Darcy-type leakance formulation [Gunduz and Aral, 2005]. Overland flow can contribute to channel flow while channel flow may flood nearby cells, allowing the floodplain size to be freely determined by the input elevation data and flow conditions. Overland flow is connected to the groundwater via the lowland storage compartment, which is designed to simulate wetlands, potholes, washes and other ponds. More details regarding the surface flow modules and river-land exchanges are provided in Text S4 in the Supporting Information. The lowland storage compartment reduces scale dependence, allowing groundwater to exfiltrate through local

topographic minima prior to saturating the entire soil column. Overland flow gathers runoff from the ponding domain, which is integrated to soil water flow (Richards' equation), which is in turn dynamically coupled to the saturated groundwater flow. The flow modules are solved sequentially to reduce computational demand. The flow modules have been found to reproduce analytical solutions and experimental results, and compare favorably with models solving the full 3D Richards' equation without simplifications [Maxwell *et al.*, 2014].

The fraction of overland-flow inundated area, f_{sat} , can describe the fraction of the land area occupied by water as a function of water depth. It impacts the area of water in the ephemeral washes under evaporative demand and can percolate. It is estimated according to the formulation proposed in CLM4.5 [Oleson *et al.*, 2013]:

$$f_{sat} = \frac{1}{2} \left(1 + \operatorname{erf} \left(\frac{d}{\sigma_{micro} \sqrt{2}} \right) \right)$$

$$\sigma_{micro} = (\beta + \beta_0)^\eta$$

where d is the water height above elevation, β is the topographic slope, $\beta_0 = (\sigma_{max})^{1/\eta}$ determines the maximum value of σ , and η is a parameter. By default, $\sigma_{max} = 0.4$. In this study, η is set as -1.5.

Figure S2. Program Flow Diagram for the essential processes of the PAWS+CLM model. Enclosed in the dashed rounded rectangle are essential processes from CLM. V/M/H Transfer means Vapor/Momentum/Heat transfer. PET estimated by PAWS (Penmon-Monteith) is used for statistical purposes and is not used for canopy flux calculations.

The surface water layer is divided into the flow domain, which can flow laterally, and the ponding domain, which exchanges with the main soil column and does not circulate laterally. The flow domain water is routed downstream as overland flow, described by 2D diffusive wave equation (DWE). Infiltrated water is governed by the Richards' equation. Water reaching the phreatic water table may move laterally, as described by the Dupuit-Forchheimer flow in the unconfined aquifer. 1D columns of vertical soil flow are coupled to the saturated lateral flow at the bottom. The confined aquifers below are described by a 3D saturated groundwater flow equation. The channel flow is governed by DWE in a 1D cascade network.

The 2D overland flow component in PAWS employs a 9-point scheme involving a cell and its eight lateral neighbors. The fluxes are weighted to calculate the divergence at a cell, with face fluxes weighted by $2/3$ and corner fluxes weighted by $1/6$, which produces 2nd order accuracy for smooth solutions [Lynch, 1992]. This approach allows water to be routed to the outlet on a sink-filled landscape. In many other model applications, a fill-sink operation is applied to the landscape to remove sinks that can retain water (e.g., [Camporese *et al.*, 2010]).