QUANTIFYING RECHARGE DURING THE LAST GLACIAL MAXIMUM IN THE
DEATH VALLEY REGIONAL FLOW SYSTEM

by Joel Hecker

Paleo-hydrologic surface deposits signify wetter than present conditions during the Last Glacial Maximum (LGM) in the American Southwest. Detailed knowledge exists about the location of these surface deposits, as well as the climate and hydrologic conditions during the LGM. Using groundwater modeling programs in conjunction with paleo-hydrologic deposits allowed for an opportunity to quantitatively characterize past climate for a 100,000 km$^2$ area near Death Valley, approximating recharge and the level of the potentiometric surface. Simulations using the Death Valley Regional Flow System model indicated that the average recharge rate in the model domain during the LGM was approximately 1.63 cm/yr, or 7.71 times the calibrated modern recharge rate. As a result, the simulated potentiometric surface was generally 0-50m higher throughout the model domain. Assuming LGM precipitation was twice the modern value (Forester et al., 1999), ~6.80% of total precipitation became recharge during the LGM, compared to only ~1.75% during modern day.
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Introduction

The Earth is warming at an unprecedented pace and global temperatures have increased by ~0.75°C since 1860 (IPCC Working Group 1, 2007). This warming, which has increased the water holding capacity of the atmosphere by ~7%, is also influencing precipitation. Precipitation events globally have increased in intensity and decreased in duration and frequency (Trenberth et al., 2003). Arid regions are especially susceptible to climate change with predicted and already observed increased frequency of sustained droughts and heavy rainfall events (IPCC Working Group 1, 2007; Huntington, 2006).

The American Southwest is one such semi-arid area that is particularly sensitive to climate change. Current climate models predict that the southwestern USA will receive 20% less precipitation, and have temperatures rising by 1°C by as soon as the year 2020. By 2090, temperatures will be as much as 3.5 to 4°C higher than current temperatures (IPCC 4th Assessment, 2007). Currently, the American Southwest has no natural standing bodies of fresh water. The only fresh-water lakes in this arid environment are dammed reservoirs which speak to the power man has on the environment. These reservoirs are already showing the effects of climate variability. Lake Mead, which is the main source of water for the city of Las Vegas, shrank to 57% of its capacity in 2008 due to long-term drought, and some predictions have it running dry by 2021 if these drought conditions persist (Slater, 2010).

The American Southwest has not always been a desert environment. During the Last Glacial Maximum (LGM), ~20,000 years ago, numerous lakes and wetlands existed, but have since dried up due to the warming climate (Quade et al., 1995; Quade 1996; Forester et al., 1999). Quantifying the changes to climate since the LGM can be beneficial in giving more accurate estimates of future climate change. To understand impacts of future climate change on water resources, it is necessary to fully understand the past. Future climate predictions can be refined with quantified estimates of past climate, as future climate models are largely based on conditions that have already occurred. As economic growth and climate change continue, better estimates will allow for more accurate predictions of future available water. Communities need to know what level of water use will be sustainable for manufacturing, irrigation, and municipal consumption.
Our detailed knowledge of the climate and hydrologic conditions of the American Southwest during the LGM, coupled with the existence of modern-day, complex groundwater models, provides a perfect opportunity to better quantify climate during the LGM. Evidence of a wetter climate during LGM exists in the form of paleo-hydrologic deposits. Paleolake and paleowetland deposits are common in the American Southwest and provide evidence higher effective moisture during the LGM (Benson, 1990; Brook et al., 2006; Quade and Broecker, 2009; Tchakerian and Lancaster, 2002). Paleowetland deposits signify the surficial expression of the paleo-water table; the modern water table lies more than 100m below the land surface in many regions (Quade et al., 1995). Shoreline indicators of paleolakes can be dated and reconstructed to show lake elevations at different times. These paleo-hydrologic deposits can be used along with other indicators of past climate, including packrat middens, speleothems, and pollen records to qualitatively reconstruct climate.

Over the past 20 years, a plethora of groundwater models have been developed for the Death Valley region of the American Southwest, with each model building on the previous one (Figure 1). The majority of the models were created for the Nevada Test Site and the Yucca Mountain Waste Repository. The early models from the 1980s were simplified two-dimensional, finite-element models (Rice, 1984; Czarnecki and Waddell, 1984; Sinton, 1987), but the most recent models are three-dimensional representations of the hydrogeologic framework and groundwater flow systems. The modern models are highly detailed, and have a large range of applicability, from following contaminant flow paths to analyzing the effects of water depletion caused by pumping. These models can be used with the paleoclimate proxies to quantify specifics of past climate, such as the elevation of the potentiometric surface over a large region, and the amount of recharge that would be required to engender the elevated water levels of the LGM. D’Agnese et al. (1999) performed such an analysis using an older three-dimensional groundwater flow model to estimate how much recharge was required to raise the water table up to LGM levels inferred from paleowetland deposits.

This study builds upon the work of D’Agnese et al. (1999) by applying a similar analysis to an updated groundwater-flow model of the Death Valley Regional Flow System (DVRFS) (Belcher and Sweetkind, 2010). I have worked with this model to
simulate conditions during the LGM, estimating the hydrologic conditions necessary for the creation of over 100 lakes and wetlands represented by paleodeposits that were used by D’Agnese et al. (1999) and other paleodeposits discovered since that study.

**Study Area and the Death Valley Regional Flow System Model**

The DVRFS model domain covers approximately 100,000 km$^2$ in Nevada and southeastern California, covering parts of the Mojave Desert, Amargosa Desert, the Great Basin Desert, and Death Valley (Figure 1). The N-S boundaries of the model are located at 35°00’N and 38°15’N, and the E-W are located at 115°00’W and 118°00’W (Belcher and Sweetkind, 2010). Land-surface elevations in the model area range from 86 meters below sea level in Death Valley (on the western side of the model), to more than 3,600 meters above sea level in the Spring Mountains (southeastern edge of model). Mountain ranges occupy ~25% of the study area (Peterson, 1981), and the remainder comprises intermontane basins filled with alluvial and playa deposits. The central part of the DVRFS contains Yucca Mountain and the Rainier Mesa, both situated within a large volcanic plateau. The majority of the model area is within the Mojave Desert and Death Valley, with the transition zone and the Great Basin Desert occurring at the northernmost portions.

The modern climatic conditions of the DVRFS are largely controlled by altitude. Death Valley and the eastern parts of the Mojave Desert are characterized by hot dry summers and warm dry winters (Benson and Darrow, 1981). The Great Basin Desert to the north experiences warm dry summers and cold dry winters. The strip between these two areas is a transition zone of their mixing climates. Precipitation affecting the DVRFS originates from two different storm patterns. Winter precipitation from the west generally covers large tracts of land, and is of a long duration and of minimum severity. These storms originate as large cloud masses in the Gulf of California and rise over the mountains, heading towards the low pressure desert air (Hales, 1972). The summer precipitation arrives in strong thunderstorms in very short, high intensity episodes, originating from localized convective clouds (Hales, 1974). High amounts of precipitation fall in the mountains, with the Spring Mountains receiving over 55 cm/yr
Figure 1. DVRFS study area and previous Regional Hydrologic Framework Models of the DVRFS (Modified from Belcher and Sweetkind, 2010).
(Scanlon et al., 2006). However, in the valley bottoms, precipitation can average less than 5 cm/yr (Winograd and Thordarson, 1975). Based on figures from Daly et al. (1994), on average, the DVRFS receives about 12 cm of precipitation per year.

With such small amounts of precipitation, very little perennial surface water exists in the study area, with the exception of man-made reservoirs. Las Vegas receives 88% of its water resources from Lake Mead. Located southeast of Las Vegas, Lake Mead is 112 miles long, with a variety of widths from 500 feet to 8.5 miles wide, and if full can hold 28 million acre-feet of water (~32 km³). This makes it the largest reservoir in the United States, but it is currently only half full due to extensive droughts and overconsumption. Lake Powell, upstream of Lake Mead, has a slightly lower capacity of 27 million acre-feet. Lake Powell is also in a state of rapid decline. It has very low odds of recovering and most likely will empty out (Gertner, 2007).

Very little of the precipitation that falls upon the DVRFS becomes groundwater recharge. The infiltration of water into the ground is determined by the vegetation, slope, aspect, and surface soil type, which are all highly variable and complex. Soils range from highly weathered bedrock, with low moisture holding capacity in the mountains, to fine-grained distal alluvial fan and playa deposits on basin floors. Alluvial fans and terraces contain thick coarse soils with high infiltration rates (Belcher and Sweetkind, 2010). Many alluvial soils, on the other hand, are cemented and calcified, and therefore have low infiltration capacities. Vegetation types differ based on altitude, latitude, water availability, soil

![Figure 2: Paleoclimate Records of the American Southwest](image-url)
type, and temperature (Beatley, 1976). The plant community residing in the DVRFS is very heterogeneous.

**Climate During the LGM**

Our knowledge of the LGM climate of the American Southwest, ~26 ka to ~18 ka, is highly detailed due to an abundance of paleoclimate evidence (Figure 2). Since the LGM, climate has substantially changed, warming by at least 4 to 7°C. However, that warming occurred at an average rate 10 times slower than the warming that occurred during the 20th century (IPCC Working Group 1, 2007).

Paleolake records with increased lake-surface area from Bonneville, Manley, Searles, Owens, Estancia, Lahontan, and Russell basins provide evidence for a wetter hydrologic budget during the LGM (Benson, 1990; Brook et al., 2006; Quade and Broecker, 2009; Tchakerian and Lancaster, 2002). Nearly 100 other basins contained lakes within the Great Basin during the Late Quaternary(Figure 2). To perennially fill these terminal lake basins, large storms would have needed to occur frequently, much more often than in modern times (Enzel and Wells, 1997). Only the most intense floods reach the basins today. Lake stands began to decrease around 14ka. A salt core from Death Valley complements the evidence for standing lakes during the LGM and shows the drying out of the Early Holocene. Lacustrine deposits within the core indicate that during the Late Pleistocene perennial lakes existed, transitioning to Early Holocene salt pans and mud flats (Yang et al., 1999).

Noble gas concentrations of groundwater that date to the late Pleistocene add strength to the argument that more intense precipitation events occurred during the LGM. 14C-dated groundwater, from California, has high concentrations of excess air, indicating more intense precipitation events and more vigorous recharge than today (Kulongoski, 2009). Conditions were El Nino-like. Noble gas concentrations within groundwater are also used to quantify paleotemperatures, as those collected from deep aquifers closely reflect the mean annual ground temperature. These paleotemperature estimates suggest that during the LGM, temperatures in the American Southwest were 5.5 ± 0.7°C lower than the average Holocene temperature (Stute et al., 1995).
Evidence from pollen records and packrat middens indicates that numerous vegetation types, such as *Pinus monophylla* and *Juniperus osteosperma*, were more prevalent, grew at lower altitudes than present day, and in places that previously received more moisture (Hall, 1985; Koehler et al., 2005). Modern day conditions support Junipers and Sagebrush at elevations above 1800m (Quade, 2001). However, during the LGM, packrat middens give evidence of Junipers existing as low as 425m (Wells and Woodcock, 1985) and Sagebrush as low as 365m (King and Van Devender, 1977). In locations such as Corn Creek Flats, sagebrush grew in the valley bottoms, in what are now xeric conditions unable to support such life (Hughie and Passie, 1963). Plant macrofossils have indicated that the mean annual precipitation was highly variable, but typically as much as twice that of the modern mean annual precipitation (Forester et al., 1999). These vegetation records suggest that northern parts of the Mojave Desert were part of a transition community, whereas the southern portions of intermediate elevation were covered by mesic woodlands (Koehler et al., 2005). Furthermore, there is a high species richness that indicates cooler summers, but winters that were similar to those of today.

The vegetation records also show that most of the precipitation during the LGM occurred from winter until spring. Winter annuals mainly grew in the southern deserts, suggesting conditions were most favorable for growth during winter (Connin et al., 1998). Speleothems also give evidence for a much wetter climate than today. The Georgia Giant speleothem from New Mexico indicates that during OIS 2, which is roughly the time of the LGM, it was much wetter than today. The speleothems also show that temperatures were about 5.5°C lower than today (Brook et al., 2006). Some evidence suggests that the total volume of precipitation during the LGM may actually be similar to modern times, differing only in timing and amount of ‘effective precipitation’ (Galloway, 1965; Brugger, 2010). The same volume of precipitation still has the power to result in higher water levels, due to lower temperatures and more effective precipitation.

There are several causes for the differing climate and precipitation of the LGM. Firstly, a section of the polar front jet stream was displaced to the south due to the high pressure cells that ensued from the continental ice sheets (Anderson et al., 1988). The displaced jet stream allowed Pacific-based storms to discharge their precipitation on what
are now the southern deserts. During the LGM, the waters in the subtropics of the Pacific Ocean were much warmer, particularly in winter. This allowed for higher moisture content in the air overlying the ocean, and subsequently, more precipitation (Imbrie et al., 1984). Other factors that led to changes in climate during the Late Pleistocene included modifications to the ocean currents, lowering of sea level by ~120 m, enhanced latitudinal gradients in atmospheric temperatures, and increased wind speeds from the southern displacement of the jet stream. The increase in wind speed resulted in increased atmospheric dust and therefore an increase in the cloud cover (Tchakerian and Lancaster, 2002).

Precipitation levels were not only changing in long term scales, but also in decadal long pulses, varying by as much as 13 cm per decade, as evidenced by stream discharge deposits in the Estancia Basin of central New Mexico (Allen and Anderson, 1993).

Groundwater discharge deposits from the LGM exist in the American Southwest at elevations 10-30 meters higher than the modern potentiometric surface, proving that the potentiometric surface must have been higher. Some localized areas contain deposits indicating a potentiometric surface that was as much as 120 m higher than present day (Quade et al., 1995). Paleoclimate proxies indicate that conditions varied within the DVRFS during the LGM. Corn Creek Flats is a valley bottom located in the southeast section of the DVRFS,
approximately 50 km northwest of Las Vegas. LGM conditions were right for spring fed channels and marshy areas. Areas only 100km to the North (Groom Lake and Kawich Lake) were able to support pluvial lakes, as they were located at higher elevations and received substantially more recharge (Quade, 1986). Also within the central Great Basin, Emigrant Valley and Gold Flat each contained shallow lakes less than 1.3 meters deep (Figure 4). Active springs in the Sheep Range and Spring Mountains resulted in groundwater discharge at the base of alluvial fans (Belcher and Sweetkind, 2010). After 14 ka, water levels gradually began to decrease at the same time as the many paleolakes began receding (Quade, 1986). Correlative sites are present all around the California Valley and Great Basin, including the Pahrump Valley, Tecopa Basin, and Amargosa Desert (Mahan et al., 2007).

In summary, during the LGM, several lines of evidence indicate a wetter hydrologic budget. Lake levels were higher, evapotranspiration rates were lower, humidity levels were higher, and temperatures were lower by 4 to 7°C. Woodland expansion occurred, as well as increased frequency and intensity of storms, increased effective moisture, and more cloud cover.

**Previous Groundwater Flow Models and Simulations**

The first numerical groundwater-flow models of the Death Valley region were created in the mid-1980s and were relatively simplified. These models contained many estimates and groupings of system parameters. The most recent groundwater models of the Death Valley region were more complex, defining the geometry and structure of hydrogeologic units, as well as more accurately depicting the groundwater flow systems (Figure 1).

The first groundwater model of the Death Valley region was a two-dimensional finite-element model of the Nevada Test Site (NTS) near Yucca Mountain (Waddell, 1982). However, simulated hydraulic heads did not correspond well to observations from wells (Belcher and Sweetkind, 2010). Czarnecki and Waddell (1984) also used a two-dimensional finite-element model to evaluate sub-regional groundwater flow in the Amargosa Desert. Czarnecki (1985) improved on this model by adding low permeability zones. Rice (1984) created a model similar to that of Czarnecki and Waddell (1984), but
Rice’s model differed in that it had detailed estimates of discharge and recharge. Sinton (1987) created a quasi-3D model to characterize the regional groundwater flow system of the NTS. This model incorporated two transmissive layers, making the NTS flow system more accurate than previous models. True three-dimensional models of the region were first developed by Prudic et al. (1995), who constructed a regional-scale numerical model of the Great Basin. This model was the basis for the models that followed, and it described the flow patterns of the Death Valley Regional Flow System (DVRFS). The early simplified models built upon one another, but were inferior to the complex models of today. They did not adequately depict important aspects of the groundwater flow system such as sub-basin groundwater fluxes and the steep hydraulic gradients found in some regions.

The modern three-dimensional flow models incorporate much more of the spatial and process complexities of the hydrogeologic system. These newer versions are based on hydrogeologic framework models used to incorporate more details about the geometry and structure of the hydrogeologic units. Models of the Yucca Mountain Project and the underground test site at the NTS were constructed in the late 1990s and early 2000s (Belcher and Sweetkind, 2010). The IT Corporation designed a steady state, three-dimensional regional geologic model (DOE/NV-UGTA Model, 1996) describing the hydrogeologic framework for the flow system surrounding the NTS. The 20-layer model encompassed a 17,700-km\(^2\) area divided into cells that were 2,000 meters on a side. Information from geologic reports, maps, cross sections, stratigraphic sections, well logs, and geophysical interpretations were used to develop the model. The main goal of the model was to predict the movement of contaminants from beneath the underground nuclear testing areas.

**D’Agnese et al. (1997): Modern Day**

D’Agnese et al. (Yucca Mountain Project/ Hydrologic Resource Management Program Model, 1997) described the hydrogeologic framework of the flow system around Yucca Mountain. The three-layer, steady-state MODFLOWP model covered 70,000 km\(^2\) divided into 163 rows and 153 columns with cells that were 1500 m on a side. The model was based on DEMs, geologic maps, cross sections, and nearly 700 well
logs. This model used the geometry, composition, and hydraulic properties that control regional groundwater flow in the present day system. Recharge values used by D’Agnese et al. (1997) relied on an altered version of the Maxey and Eakin method (1949) to estimate recharge. Maxey and Eakin developed an empirical relationship of recharge versus altitude specifically for basins in southern and eastern Nevada. D’Agnese et al. (1997) altered this to include other critical factors affecting recharge: altitude, slope aspect, vegetation, and rock/soil permeability. Quantitative recharge potential maps for each factor were created and overlaid, producing a recharge potential map that gave a final percentage of precipitation that became recharge.

The D’Agnese and IT Corporation models were combined to produce a single, steady state, pre-pumping groundwater flow model (D’Agnese et al., 2002). Discrepancies between the two models were largely ignored, due to the project’s limited scope, which was to estimate aquifer-system parameters (Belcher and Sweetkind, 2010). The combined groundwater-flow model was an improvement over the two individual models, but there were several uncertainties and/or errors that remained.

D’Agnese et al. (1999): Last Glacial Maximum

D’Agnese et al. (1999) estimated past and future recharge conditions, using the Yucca Mountain Project/ Hydrologic Resource Management Program model from 1997. Water levels were simulated beneath Yucca Mountain relative to present day. Their model assessed regional groundwater flow under LGM climatic conditions as well as predicted changes to the flow system based on assumptions of much warmer climatic conditions. The overarching purpose of this modeling effort was to evaluate the effect that changing climate would have on water levels under the Yucca Mountain Waste Repository.

D’Agnese et al. (1999) modified modern recharge values based on climatic conditions simulated for the LGM using a combination of nested and regional climate models developed by the National Center for Atmospheric Research (Thompson, 1998). Also, a global circulation model represented the large scale climatic forcing due to Earth’s orbit and changes in CO₂ levels. Predicted increases in precipitation led to a rise in recharge. Several areas in D’Agnese et al.’s (1999) model simulated a drop in the
potentiometric surface compared to modern day, as a result of lower recharge rates. These areas contained no paleo-hydrologic deposits for comparison, and the accuracy of the potentiometric surface simulations in these areas is unknown. D’Agnese et al. (1999) used four simplifying assumptions when creating their LGM groundwater simulation. First, because the complexities of fracture flow are unfeasible to model, all flow in the region was assumed to occur through porous media. Zones of differing hydraulic conductivities helped to account for the differences in flow rates. Second, hydraulic conductivities were assumed to be homogenous (within a zone) and horizontally isotropic. Third, the system was assumed to be at steady state, representing long-term conditions with, of course, no pumping. Last, the saturated thickness in the model layers was constant. Simulated groundwater discharge occurred at most known paleodischarge sites, indicating the recharge distributions were generally valid.

**Belcher and Sweetkind (2010): Death Valley Regional Flow System**

The 2010 Death Valley Regional Groundwater Flow Model is the most up to date model of the region. The model was developed by the United States Geological Survey, in conjunction with the U.S. Department of Energy, the Office of Environmental Management, the Office of Civilian Radioactive Waste Management, and the National Park Service. This transient model employed the numerical code MODFLOW-2000 (Harbaugh et al., 2000; Hill et al., 2000), the USGS three-dimensional, finite-difference modular groundwater-flow modeling code. MODFLOW-2000 incorporates a nonlinear least-squares regression technique to estimate aquifer parameters based on collected data such as measured heads and groundwater discharge rates. Pre- and post-processers (Microsoft Excel, ArcGIS) were used with the numerical code of MODFLOW-2000 to define parameter arrays, display results, and assist in analyzing results.

The model covered over 100,000 km² gridded into 194 rows, 160 columns, and 16 layers, resulting in 496,640 individual cells, 314,780 of which were active. The grid was oriented N-S and E-W; cell spacing for both rows and columns was 1,500m. The model extended 4,000 m below modern sea level to encompass nearly all of the aquifer units in the region.
Recharge in the DVRFS was determined based on a net infiltration model, INFIL v. 3 (Hevesi et al., 2003). Net infiltration approximates recharge, because most net infiltration eventually moves downward through the unsaturated zone to recharge the groundwater flow system (Belcher and Sweetkind, 2010). In a steady state model, net infiltration equals recharge, due to the absence of a time factor.

The model boundaries generally extended to surrounding mountain ranges with low permeability rock (i.e. Sheep Range in the East, Cottonwood Mountains in the West, Kingston Range and Avawatz Mountains in the south). The external boundaries were assumed to be no-flow boundaries, except in small localized areas in the far north and far south of the model area, where constant-head boundaries simulated potential groundwater flow into the model area.

Hydraulic conductivities were divided into different zones based on 27 different Hydrogeologic Flow Units and ranged from $3 \times 10^{-8}$ m/day to over 800 m/day. Discharge occurred as evapotranspiration, flowing springs, and pumpage. Evapotranspiration rates in the model were based upon temperature, wind speed, altitude, and humidity. Mean annual groundwater discharge in the model was calculated to equal the difference between mean annual evapotranspiration and the sum of mean annual precipitation and any surface-water inflow. Pumping rates were

![Figure 4. Location of Lakes and Wetland Areas during LGM](image)
assigned, associated with areas where groundwater withdrawal is occurring.

Methods

Table 1. Overview of Methods

- Acquired the transient model of the DVRFS from the USGS (Belcher and Sweetkind, 2010)
- Converted the model to steady-state
- LGM recharge from D’Agnese et al. (1999) was expanded to fit the DVRFS boundaries
- Drains were added to paleo-hydrologic deposits to signify water leaving the model
- Lowered recharge in areas where the water level rose above land surface, but no surface deposits existed (Some areas within model domain were left at D’Agnese et al. (1999) recharge values)
- Completed uncertainty and sensitivity analyses for the entire model domain to quantify the overall uncertainty, and the sensitivity to five parameters

Modifications to the DVRFS model

For this study, the 2010 DVRFS model was modified to simulate assumed steady-state conditions during the LGM, based upon the well-documented and well-constrained climate records of that period.

The model domain, most parameter values, and the exterior boundaries used in this LGM simulation were identical to those of the modern-day DVRFS model (Belcher and Sweetkind, 2010). The model was modified in terms of recharge rates, pumping rates, and the number and location of drains used to represent discharge to paleowetland/paleolake areas and steep mountain areas.

Pumping Rates

The model simulation represents the Last Glacial Maximum (~20 ka), and therefore all pumps were removed from the simulation.
Incorporation of Paleowetland Data and Addition of Drains

Drains were added to the model at all locations where wetland, lake, and perennial river deposits dating to the LGM have been identified; these drains simulate water leaving the model (Figure 5). Quade et al. (1995) and Forester et al. (1999) have dated such locations to the LGM, providing evidence of a wetter climate. The drains were head-dependent, allowing water to leave the model at a rate proportional to the height of the water above the drain. Water is discharged only when water levels rise above the land surface. The drain conductances were defined using the hydraulic conductivity and thickness of materials through which water flowed along with the length and width of the discharging area.

Application of D’Agnese et al. (1999) Recharge

To simulate the potentiometric surface during the LGM, the amount and distribution of recharge must be estimated. Climate during the LGM was characterized by average annual temperatures 4 to 7°C lower than present along with lower evapotranspiration rates, increased frequency and intensity of storms, increased effective moisture, and more cloud cover. These and other factors

Figure 5. Location of Drains used in LGM model simulations. UP_DVN_DRN and SPG_MT_DRN groups represent mountain drains for calibration purposes. The remaining five drain groups occurred in areas with known lakes and wetlands.
caused the LGM to have greater recharge rates when compared to the present. Recharge is based upon the amount of precipitation a region receives along with evapotranspiration rates (controlled in part by vegetation type and amount), altitude, slope aspect, and relative rock and soil permeability (D’Agnese et al., 1997).

Recharge rates were used as the principal calibration parameter to bring the water levels up to elevations of the paleowetland and paleolake deposits. As a starting point, the LGM recharge array from D’Agnese et al. (1999) was adapted to Belcher and Sweetkind’s 2010 DVRFS model. The LGM recharge of D’Agnese et al. (1999) did not cover the full extent of the Belcher and Sweetkind (2010) model domain (Figure 1). Therefore, the first step in LGM recharge estimation using the newer model was to estimate recharge for the areas not covered in the D’Agnese et al. (1999) model. Regions not covered in the newer model were matched with similar regions covered in the D’Agnese et al. (1999) model to approximate the value of the LGM recharge. Comparisons in vegetation, slope aspect, topography, and rock/soil permeability were used to expand the recharge to the full model domain. For example, unvegetated high elevation, west-oriented slopes in the northeastern section of the model were given LGM recharge rates similar to a matching region of D’Agnese et al. (1999).

Calibration: Using Recharge to Obtain LGM Water Levels

After expanding the LGM recharge array based on D’Agnese et al.’s (1999) values, and running MODFLOW-2000, recharge was adjusted so that the potentiometric surface was below the land surface, surface water did not exist in large tracts of land not containing LGM hydrologic deposits, and so that the potentiometric surface met the land surface where paleowetland and paleolake deposits indicated.

Uncertainty Analyses

The two main changes to this version of the DVRFS groundwater flow model, from the 2010 version, related to drains and recharge. The 2010 DVRFS has been previously calibrated and undergone a wide variety of sensitivity and uncertainty analyses of many parameters (Belcher and Sweetkind, 2010) that were not altered in the LGM version of the model. However, there is still uncertainty associated with the
parameters, and those uncertainties contribute to the uncertainty associated with the estimation of recharge during the LGM. Small changes within the acceptable range of hydraulic conductivity and drain conductance values may significantly alter the amount of recharge necessary to replicate LGM conditions. The uncertainty of the LGM recharge has been quantified based on the collective uncertainties of the five of the model parameters. Three hydraulic conductivity values and two drain conductance values were used to quantify the uncertainty of the recharge (Table 2). Parameters 1 and 2 were drains, chosen because they represented a large amount of water leaving the model. Parameters 3-5 were hydraulic conductivity values that covered a large expanse, were near the surface, or were high compared to other conductivities. The uncertainty analyses were limited to five parameters due to time constraints. Two additional conductivity values were used in the initial analysis, but the model was not sensitive to changes in those parameters (K2412_LCA and K232_LCA).

### Table 2. Altered Hydraulic Conductivity and Drain Conductance Values for Uncertainty Analyses

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Name (and parameter type)</th>
<th>0.042Q (1)</th>
<th>0.5 (2)</th>
<th>0.958Q (3)</th>
<th>Reasoning for Parameter</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>UP_PLY_DRN (Drain Conductance)</td>
<td>1.13 x10^3</td>
<td>8.39 x10^1</td>
<td>1.56 x10^2</td>
<td>Springs in playa deposits</td>
</tr>
<tr>
<td>2</td>
<td>UP_PAH_DRN (Drain Conductance)</td>
<td>2.64 x10^3</td>
<td>1.95 x10^2</td>
<td>3.64 x10^2</td>
<td>Lakes in SE quadrant of DVRFS</td>
</tr>
<tr>
<td>3</td>
<td>K3LFU_am (Hydraulic Conductivity)</td>
<td>1.56 x10^3</td>
<td>5.09 x10^2</td>
<td>1.67 x10^1</td>
<td>Large spatially and very near the surface</td>
</tr>
<tr>
<td>4</td>
<td>K321521_PP (Hydraulic Conductivity)</td>
<td>1.99 x10^5</td>
<td>1.66 x10^2</td>
<td>1.38 x10^1</td>
<td>Central part of DVRFS with high K values</td>
</tr>
<tr>
<td>5</td>
<td>K422DV_VSU (Hydraulic Conductivity)</td>
<td>7.47 x10^3</td>
<td>5.80 x10^2</td>
<td>4.01 x10^1</td>
<td>Large expanse in Death Valley</td>
</tr>
</tbody>
</table>

Uncertainty was quantified using the approximation of the three-point Gauss-Hermite quadrature approach (Raftery and Zeh, 1993). The approximation of the three-point Gauss-Hermite quadrature approach is an alternative to the full Gauss-Hermite quadrature approach and the Monte Carlo approach. For analysis using five parameters, the approximation requires only 51 model runs compared to 243 for the full Gauss-Hermite and typically hundreds to thousands of model runs for the Monte Carlo approach. The Gauss-Hermite method requires running the model using a low, medium and high value for each parameter considered in the analysis; specifically, the 0.042
quantile, the mean, and the 0.958 quantile value must be estimated. These quantiles were determined for the 5 parameters using the coefficients of variation, from Belcher and Sweetkind (2010). The full Gauss-Hermite method utilizes every possible combination of the three quantile values for each parameter, but the approximation includes only model runs with no more than two parameters not at the 0.5 quantile. This allows quantification of parameter main effects (i.e., sensitivity of the model estimation to changes in a single parameter) and two-way interactions; all higher-order interactions are assumed to be small and are ignored. With each model run, the parameter values were changed, and recharge was recalibrated to get the best agreement between simulated surface water and observed paleo-hydrologic deposits. After calibration of each of the 51 model runs, the average recharge value was calculated.

**Results**

D’Agnese et al. (1999) Recharge Array Expanded to DVRFS Model Area

The recharge array used by D’Agnese et al. (1999) to simulate LGM conditions was expanded and used directly in the 2010 DVRFS model (Belcher and Sweetkind, 2010). That expanded recharge array represented about 9 times more recharge than modern day and resulted in an overestimation of the potentiometric surface. The potentiometric surface was above the land surface in many areas (Figure 6). The LGM recharge as estimated by D’Agnese et al. (1999) and expanded and applied to the DVRFS model, was far too much recharge according to the newer model. This was likely a result of lower hydraulic conductivity values in Belcher and Sweetkind’s (2010) model.

Description of Modifications during Calibration

**Recharge Array**

To better approximate the LGM potentiometric surface using the DVRFS model, the amount of recharge had to be lowered in those areas where the potentiometric surface was initially simulated above the land surface. Recharge was lowered in many areas until the potentiometric surface was below the land surface, but still resulted in surface water or spring flow (at drains) at known paleo-hydrologic deposits. The model domain was broken into small areas and various multipliers were applied to either raise or lower the
potentiometric surface and resulting locations of surface water (Figure 7). This was done location by location, until reaching the best possible agreement between simulated surface water and observed deposits. The average LGM recharge for the calibrated result was 30% less than the expanded array based on D’Agnese et al. (1999). Some areas of D’Agnese et al.’s (1999) array were not altered, as no evidence existed to suggest that water levels should have been raised or lowered(Figure 8). These unaltered areas of the model domain are relics of the simulations completed by D’Agnese et al. (1999).

**Drains**

After lowering recharge rates and reaching agreement between simulated water levels and known paleo-hydrologic deposits, the model still simulated water levels above land surface in areas of the Spring Mountains and Funeral Mountains. This was most likely due to the inability of the regional-scale model to completely resolve steep hydraulic gradients present in these areas of great topographic relief. To provide a more accurate simulation of the steep hydraulic gradients in these areas, additional drains were inserted in these locations to lower the water level to just below the land surface (Figure 6).
5). The model by D’Agnese et al. (1999) similarly relied on drains in steep mountainous areas.

**Calibrated LGM Simulations using the DVRFS model**

**Comparison of Observed and Simulated Paleowetlands and Paleolakes**

The calibrated LGM recharge values were determined by altering the recharge component in various areas to reach an agreement with the known surface deposits (Figure 5). The final agreement between the location of hydrologic deposits and LGM surface water is shown in Figure 9. Overall, the LGM locations of surface water matched up well with the known location of deposits. The largest discrepancies occurred in the northwest part of the model domain, as a result of inaccuracies within the present-day DVRFS. In this quadrant, the present-day simulations had the poorest agreement between simulated and observed heads (Belcher and Sweetkind, 2010), and this carried over to LGM simulations. The south, east, and west portions aligned well with groundwater discharge and paleolake deposits.

![Figure 7. Calibrated LGM Surface Water Levels](image-url)
Figure 8. Areas within DVRFS with recharge differing from D’Agnese et al. (1999) values
Figure 9. Known LGM lakes and wetlands superimposed onto LGM surface water levels
Potentiometric Surface and Flow Budget Changes

The simulated LGM potentiometric surface had the same general character as the modern day surface (Figures 10, 11) with some distinct differences. Simulated LGM water levels were higher throughout the domain with few exceptions. Heads were highest in the Spring Mountains at 2800 m and lowest in Death Valley/Lake Manley at 5 m. Both the LGM and modern day simulations contained higher heads in the north and mountain regions, with lower heads in southeast and Death Valley/Lake Manley. The general direction of groundwater flow was from the northwest towards the southeast and southwest. The LGM potentiometric surface varied from steep to very shallow gradients. The highest hydraulic gradient occurred along the foothills of the Amargosa, Spring, Sheep, Halfpint, and Eleana Ranges (See Figure 1 for locations). Shallow gradients occurred in the central and far northern realms of the DVRFS.

Recharge accounted for 80% of the water entering the LGM model area. Water flowed into the model from the northern and western portions of the model. Outflow occurred across the southeastern edge of the DVRFS. 59% of water entering the LGM model was removed via the drains. Lake Manley (Figure 5) and the other lake and wetland drains were responsible for removing the largest amount of water (Table 3). The water simulated as discharging via mountain-top drains (SPG_MT_DRN and UP_DVN_DRN) was, in reality, probably discharged close to the recharge area, as surface water runoff, springs, or evapotranspiration (D’Agnese et al., 1999).

The modern DVRFS model simulation contains no stable bodies of water and only a few ephemeral playas and springs. For the LGM simulation, the DVRFS model predicted numerous lakes and wetlands, most in similar locations as known paleo-hydrologic deposits (Figure 9). In some high elevation areas in the south and east, the simulated LGM potentiometric surface was

<table>
<thead>
<tr>
<th>Cumulative Volumes (L)</th>
<th>In</th>
<th>Out</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storage</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>Constant Head</td>
<td>441298.3047</td>
<td>916234.9144</td>
</tr>
<tr>
<td>Drains</td>
<td>0</td>
<td>1321433.935</td>
</tr>
<tr>
<td>Recharge</td>
<td>1796202.656</td>
<td>0</td>
</tr>
<tr>
<td>Total In</td>
<td>2237500.961</td>
<td>2237668.849</td>
</tr>
</tbody>
</table>

In - Out  -167.8884000
Percent Discrepancy -0.01

*Percent Discrepancy reflects primarily numerical errors associated with the convergence of the model solution.
as much as 500 m higher than present day (Figure 12). Alternatively, the simulated LGM potentiometric surface was 100 m lower than today’s in parts of the Amargosa Valley and the Grapevine Mountains. This result was an artifact of using D’Agnese et al.’s (1999) LGM recharge in these areas. No paleo-hydrologic deposits exist in this region; therefore no better approximation of the LGM potentiometric surface could be simulated.

Throughout most of the study area, the simulated LGM potentiometric surface was no more than 50 m higher than the modern day, despite the higher recharge, largely due to the high conductivity values of the sediments in low lying areas.
Figure 10: Modern Day Potentiometric Surface
Figure 11: LGM Potentiometric Surface, based on adjustments to the array of D’Agnese et al. (1999)
Figure 12. Difference between LGM and modern day potentiometric surface, based on adjustments to the array of D’Agnese et al. (1999)
Relative Recharge Rates for Different Models and Simulations

The calibrated results from the LGM simulation showed that on average, the DVRFS received 1.63 cm/yr of recharge throughout the model domain (Table 4, Figure 13). This calibrated LGM recharge value is 7.7 times modern average recharge (Figure 14). The modern day recharge had a slightly higher maximum recharge value (39 cm/yr versus 35 cm/yr), but the extent of higher recharge covered much more of the model area during LGM simulations. Four times more area received a non-zero amount of recharge during the LGM than modern day. Areas receiving at least 2 cm/yr of recharge covered ~20% of the model domain during the LGM simulation, but only 2.2% of the model domain during modern day. Those regions receiving greater than 0.02 cm/yr simulated recharge covered only 31% of the present-day domain, but 85% of LGM domain.

The highest simulated recharge for modern day occurred in the Spring Mountains along the southeastern portion of the DVRFS; however, in the LGM simulation, the highest recharge occurred in the Panamint Range along the western side of Death Valley. The majority of simulated LGM recharge fell in high elevation areas in both the northeastern and southern portions of the model domain. Modern day recharge was simulated as spatially limited and large portions of the model domain received no recharge throughout the year.

Uncertainty Analysis

To quantify the uncertainty associated with the recharge estimation for the LGM, an uncertainty analysis was conducted focused on a single model prediction: the average recharge rate of the entire model area. This uncertainty analysis therefore covered both areas in which recharge was altered from the D'Agnese et al. (1999) values and those areas that were unaltered due to a lack of paleowetland deposits. In hindsight, a better approach would have been to focus on those areas with paleowetland deposits and use the parameters that were most important in those areas. The approach used here

<table>
<thead>
<tr>
<th>Recharge Version</th>
<th>Recharge (m/d)</th>
<th>Recharge (m/year)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Modern Day Recharge</td>
<td>5.77x10^-6</td>
<td>2.11x10^-3</td>
</tr>
<tr>
<td>Uncalibrated LGM Recharge</td>
<td>4.53x10^-5</td>
<td>1.65x10^-2</td>
</tr>
<tr>
<td>~D'Agnese LGM Recharge</td>
<td>6.32x10^-5</td>
<td>2.31x10^-2</td>
</tr>
<tr>
<td>Calibrated LGM Recharge</td>
<td>4.46x10^-5</td>
<td>1.63x10^-2</td>
</tr>
</tbody>
</table>

Table 4. Recharge values for different versions of DVRFS

28
encompassed the five parameters thought to have the most influence on hydraulic heads over the entire model domain and may have therefore underestimated the uncertainty of recharge estimates.

The uncertainty analysis was performed using the approximation of the three-point Gauss-Hermite quadrature approach based on the collective uncertainties of the five model parameters as described earlier. The analysis resulted in an expected recharge value and the standard error and 95% confidence interval associated with that estimate (Table 5). The expected value was 1.65 cm/yr, slightly different than the calibrated best estimate of 1.63 cm/yr, as a result of the parameters being log-normally distributed. The 95% confidence interval ranged from 1.53-1.73 cm/yr, with a standard error of 5.04x10^{-4} m/yr.

**Sensitivity Analyses**

The single parameter effects (Table 6) showed which parameters played the biggest role in the uncertainty of the recharge. The single parameter effects were evaluated by analyzing the relationship between the sum-of-squares main effects (SSME) for each parameter, defined as the sum of the squares of the 0.042 and 0.958 quantile components, standardized by taking the square root and dividing by the expected value of the prediction (Levy et al., 1998). The largest SSME values have the largest role on the uncertainty of the model prediction. The two-way interactions (Table 7) indicate whether any two parameters are affecting the outcome in a way that is not accounted for by the single-parameter effects. Sum-of-squares two-way interactions (SSTWI) were used to

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**Table 5. Recharge Uncertainty Statistics**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>SSME</th>
</tr>
</thead>
<tbody>
<tr>
<td>Variance (m²/d²)</td>
<td>6.97x10^{-10}</td>
</tr>
<tr>
<td>Standard Error (m/yr)</td>
<td>5.04x10^{-4}</td>
</tr>
<tr>
<td>95% Confidence Interval</td>
<td></td>
</tr>
<tr>
<td>Low (m/yr)</td>
<td>1.53x10^{-2}</td>
</tr>
<tr>
<td>Expected Value (m/yr)</td>
<td>1.63x10^{-2}</td>
</tr>
<tr>
<td>High (m/yr)</td>
<td>1.73x10^{-2}</td>
</tr>
</tbody>
</table>

**Table 6. Single Parameter Effects**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>SSME</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 (UP_PLY_DRN)</td>
<td>5.43x10^{-3}</td>
</tr>
<tr>
<td>2 (UP_PAH_DRN)</td>
<td>5.69x10^{-4}</td>
</tr>
<tr>
<td>3 (K3LFU_am)</td>
<td>5.01x10^{-4}</td>
</tr>
<tr>
<td>4 (K321521_PP)</td>
<td>3.42x10^{-4}</td>
</tr>
<tr>
<td>5 (K422DV_VSU)</td>
<td>3.20x10^{-13}</td>
</tr>
</tbody>
</table>

**Table 7. Two Way Interactions**

<table>
<thead>
<tr>
<th>Parameters</th>
<th>SSTWI</th>
</tr>
</thead>
<tbody>
<tr>
<td>1,2</td>
<td>3.72x10^{-3}</td>
</tr>
<tr>
<td>1,4</td>
<td>2.26x10^{-3}</td>
</tr>
<tr>
<td>1,3</td>
<td>2.16x10^{-3}</td>
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<tr>
<td>4,5</td>
<td>1.93x10^{-3}</td>
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<td>2,4</td>
<td>1.54x10^{-3}</td>
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<td>2,5</td>
<td>1.47x10^{-3}</td>
</tr>
<tr>
<td>1,5</td>
<td>1.36x10^{-3}</td>
</tr>
<tr>
<td>3,5</td>
<td>1.36x10^{-3}</td>
</tr>
<tr>
<td>3,4</td>
<td>1.27x10^{-3}</td>
</tr>
<tr>
<td>2,3</td>
<td>1.11x10^{-3}</td>
</tr>
</tbody>
</table>
Figure 13: LGM Recharge (m/day)
Figure 14: Modern Day Recharge (m/day)
assess the role that two way interactions had on the uncertainty of the prediction. Similar to SSME, SSTWI were evaluated by taking the sum of the squares of the individual components, and were standardized in the same manner as SSME. The drain conductances had the largest effect on the uncertainty of the recharge estimation, but the hydraulic conductivity values also had some effect. Parameter 5 (SSME of 3.20 x10^{-13}), in comparison, did not contribute to the prediction uncertainty. The Two Way Interaction effects were all similar; there were no interactions that stood out as particularly strong.

**Discussion**

**Simulated Heads vs. Observed Paleo-Hydrologic Deposits**

There were several problem areas in the comparison between simulated surface water and paleo-hydrologic deposits. In these problem areas, simulated water levels were consistently higher than the observed deposits. This was a result of an issue with the present-day 2010 DVRFS model. According to Belcher and Sweetkind (2010), “the fit of simulated heads to observed hydraulic heads [in the present day model] is good in most areas of nearly flat hydraulic gradients, and moderate in the remainder of the areas of nearly flat hydraulic gradients. The poorest fit of simulated heads to observed hydraulic heads is in steep hydraulic-gradient areas and the north-northwestern part of the model domain.”

The problem locations described by Belcher and Sweetkind (2010) were identical to the problem areas in the LGM simulation. The north-northwest quadrant and areas with steep gradients had the poorest agreement between deposits and simulated surface water. Belcher and Sweetkind (2010) attributed these inaccuracies to insufficient representations of the hydrogeology, misinterpreted water levels, and model error caused by grid cell sizes.

**Uncertainty and Limitations of Climate Change Simulations**

Predictive groundwater modeling has significant limitations and can only be as accurate as the model on which it is based. For this simulation of hydrogeologic conditions during the LGM, the model was based upon the 2010 DVRFS groundwater flow model (Belcher and Sweetkind, 2010). The uncertainty analysis completed
exclusively for this simulation was purely based on five model parameters thought have the largest impact on the uncertainty of recharge. Parameters 1 and 2 were drains, chosen because they represented a large amount of water leaving the model. Parameters 3-5 were hydraulic conductivity values that covered a large expanse, were near the surface, or were high compared to other conductivities.

Flux out of the discharge areas was simulated by inserting drains into the model at elevations corresponding to known present-day or paleo-hydrologic deposits. The rate of discharge from these deposits is largely unknown, especially as many of them are no longer active. The appropriate MODFLOW conductances were estimated based on the present day conductance rates of the DVRFS model created by Belcher and Sweetkind (2010) which were based on the conductivity of sediment in the drain cells. Recharge was sensitive to the conductance parameters, and therefore the validity of the recharge estimate relied on the quality of the conductance rates.

Interpolation issues with the model were another source of error. Even simply converting the present day transient model of Belcher and Sweetkind (2010) to a steady state model resulted in water at the surface in several locations where it didn’t belong, mainly in areas where there was a steep gradient of the land surface. This was probably a result of a raster interpolation issue when the land-surface elevations for the MODFLOW grid were selected from a finer grid of land-surface elevations. During this process, it is possible to get MODFLOW cells with land surface elevations lower than the average elevation of the potentiometric surface. Therefore, the average water level in some cells was calculated higher than the interpolated land surface elevations.

Uncertainty in quantifying the recharge was a result of the previous concerns, as well as limitations associated with climate change simulations. D’Agnese et al. (1999) identified the following limitations:

1. Paleo-hydrologic evidence is important in order to run accurate simulations, and the evidence is incomplete. Many paleodischarge sites have been eroded away, while others may only be of local significance and not play a role in the regional flow.
2. Climate models, like those used to determine recharge in D’Agnese et al. (1999), are used to estimate past conditions and predict precipitation patterns,
and may not be entirely accurate. If the climatic patterns don’t mimic the true conditions during the LGM, the model results will be inaccurate.

3. D’Agnese et al.’s (1999) recharge estimates (used for the initial model run to assess LGM recharge rates with the DVRFS model) were based upon a version of the Maxey-Eakin method (Maxey and Eakin, 1949). The Maxey-Eakin method to determine recharge may not be valid under different climate conditions, as it was created specifically for the modern day climate of the American Southwest.

4. This model involved converting a transient model to a steady state model, implying that the LGM climate reached a state of equilibrium, which may not be true (D’Agnese et al., 1999).

**Implications of the Recharge Rate**

Simulations using the DVRFS model indicate that the average recharge rate in the model domain during the LGM was approximately 1.63 cm/yr. The total amount of recharge for the LGM simulation was 7.7 times the calibrated modern recharge rate. Previous estimates for the LGM predict slightly higher recharge than this study. In a study of the \(^{14}\)C concentration in ground water in the American Southwest, Schwartz et al. (2010) noted that LGM recharge to the flow system near Yucca Mountain was about an order of magnitude higher than present. The expanded recharge array for D’Agnese et al.’s (1999) simulation resulted in about 9 times the modern recharge rate. The modeled rate from this study using the DVRFS model indicated that recharge was not as high as previously thought. Even with the higher rates, estimated future recharge rates (from D’Agnese et al.’s (1999) future climate simulation) predicted that they would result in water levels substantially below a level of concern for the Yucca Mountain Waste Repository. Results using the DVRFS model may signify even lower estimates for the future as well, which implies that there is no concern of the potentiometric surface reaching the Yucca Mountain site.
Recharge vs. Precipitation

Only a portion of the water that enters the DVRFS becomes recharge. The percentage of recharge is largely based on the climate and seasonality of precipitation, as well as the topography and soil and vegetative cover of a region. For the modern day DVRFS, most recharge occurs from the infiltration of precipitation (usually from winter storms) and runoff from the mountain ranges. Water also may infiltrate from melting snowpack, and through streams flowing over alluvium (Belcher and Sweetkind, 2010). Based on figures by Daly et al. (1994) an average of ~12 cm/yr of precipitation falls on the modern day DVRFS. In semi-arid to arid regions such as the American Southwest, only 0.1–5% of annual precipitation becomes recharge (Scanlon et al., 2006). The percentage of precipitation becoming recharge is highly variable however. Near Yucca Mountain, 3-6% of precipitation becomes recharge (Flint et al., 2001), but in the spring mountains, ~500 mm of recharge occurs for 550 mm of precipitation (Scanlon et al., 2006). Not all mountain ranges in the DVRFS receive high amounts of recharge. The Paramint Range receives <2mm/yr of recharge out of ~400mm/yr of precipitation, because the rocks are highly impermeable (Scanlon et al, 2006). In Death Valley, between 1950 and 1999, only 2.8 mm/yr of recharge occurred out of 170 mm/yr precipitation, which is only 1.6% (Hevesi et al., 2003).

With cooler temperatures and rainfall more evenly distributed, a higher percentage of precipitation becomes recharge. Changes in paleoclimate have led to less effective moisture during the modern semiarid period in the American Southwest, compared to the LGM (Scanlon et al., 2006). Supporting evidence comes from numerous proxies such as vegetation records, speleothems, and packrat middens. Numerous plant types existed in the DVRFS during the LGM, relying on wetter conditions. Based on plant macrofossils, Forester et al. (1999) suggested that as much as twice modern day precipitation occurred during the LGM. Assuming LGM recharge was ~7.7 times the modern day recharge, and precipitation was about twice modern precipitation, approximately 3.9 times more precipitation became recharge during the LGM than modern day. Out of about 12 cm of modern yearly precipitation, about 0.21 cm/yr become recharge, or ~1.75%. Assuming LGM precipitation was twice the modern value
and LGM recharge was 1.63 cm/yr, then ~6.8% of precipitation became recharge during the LGM.

**Summary and Conclusions**

Paleo-hydrologic surface deposits signify wetter conditions during the Last Glacial Maximum (LGM) in the American Southwest. Detailed knowledge about the location of these surface deposits, as well as the climate and hydrologic conditions of the Mojave Desert during the LGM provide an opportunity to quantitatively characterize climate for the Death Valley region. This is necessary if these present-day groundwater flow models are to be used to manage water resources under the different climatic conditions that are predicted in the future. Future climate models predict large increases in temperature and major decreases in precipitation in arid regions worldwide. However, during LGM times, there were higher lake levels, lower evapotranspiration rates, higher humidity, temperatures lowered by 4 to 7°C, woodland expansion, increased frequency and intensity of storms, increased effective moisture, and more cloud cover.

Using the most up to date version of the Death Valley Regional Flow System (DVRFS) groundwater flow model, the effects of a full glacial climate were simulated. This simulation expanded on the work of D’Agnese et al. (1999), simulating LGM conditions, but with the more up to date 2010 DVRFS model. The United States Geological Survey and the U.S. Department of the Interior, with cooperation from the U.S. Department of Energy created the 2010 DVRFS model, which was used as a platform for this past climate simulation. The DVRFS model domain encompasses approximately 100,000 km² in Nevada and southeastern California, covering parts of the Mojave Desert, Amargosa Desert, the Great Basin Desert, and Death Valley. Modern day recharge in the DVRFS model was replaced with a calibrated LGM recharge value. This recharge value was calculated by altering the recharge component from D’Agnese et al. (1999) in various areas near paleo-hydrologic surface deposits to reach an agreement between these deposits and the simulated potentiometric surface. If no paleo-hydrologic deposits existed in a region, and the potentiometric surface was beneath the land surface, the recharge component from D’Agnese et al. (1999) remained unaltered; as no evidence existed suggesting it should be changed. D’Agnese et al.’s
(1999) recharge values were based on climate models developed for them by the National Center for Atmospheric Research. D’Agnese et al.’s (1999) LGM model recharge was expanded to the larger DVRFS boundaries based on similar regions covered in their model, with the purpose of matching previously uncharacterized regions to already documented regions. According to the model results, the expanded D’Agnese et al. (1999) recharge rates, approximately 9 times more recharge than present day, were too high for the system. Recharge values were lowered in areas where the model showed water ponding at the surface, and an agreement between surface deposits and simulated surface water was reached.

The resulting LGM average recharge rate was estimated to be 1.63 cm/yr, 7.7 times the present day average rate for the entire region. In the LGM simulations, groundwater recharge was applied over a much greater area (~4 times) than occurs today. High elevation areas and regions in the northeast and south had the greatest increases in simulated recharge. These regions also saw the largest rise in the simulated potentiometric surface. Due to the high conductivity values of sediments in low lying areas, the simulated potentiometric surface rose no more than 50 m in the majority of the study area. However, the simulated potentiometric surface of the Spring Mountains and several other mountainous regions in the east and south portions of the model area rose by as much as 500 m. The overall trend of the simulated potentiometric surface was similar, with highest heads in the north, and water flowing towards the southeast and southwest. Uncertainty analyses were completed using 5 parameters from the model, and gave a 95% confidence interval of 1.53-1.73 cm/yr for the average annual recharge. These analyses showed that drains had the largest effect on the uncertainty of the recharge component.

Schwartz et al. (2010) determined that recharge near Yucca Mountain was about an order of magnitude higher during the LGM, by studying \(^{14}\)C concentrations in groundwater. He also estimated that the water table was more than 100 m higher than present day. Schwartz et al.’s (2010) study, along with D’Agnese et al.’s (1999) simulation support the validity of this LGM simulation, as their results were similar to the results of this study. However, this study’s simulations estimated slightly less recharge than previous estimates. Only a portion of precipitation becomes recharge. Forester et al.
(1999) estimated that LGM precipitation was approximately twice modern day precipitation. An LGM recharge rate that is 7.7 times the modern-day recharge rate suggests that the percentage of precipitation becoming recharge was almost 4 times higher than today.

Based on the agreement between simulated surface water, and paleo-hydrologic deposits, this simulation provided a good approximation of the potentiometric surface and recharge rates during the LGM. However, it must be remembered that this is only an approximation, and that predictive modeling has many substantial limitations. These approximations are only as accurate as the models upon which they are based.
References


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Appendix A: Supplemental Information

Confidence Interval Information (From Belcher and Sweetkind, 2010):

Confidence intervals for the estimated parameter values are calculated using sensitivities calculated for the optimal parameter values. The sensitivity analysis focused on identifying parameter values that could be estimated by regression and identifying key observations that supported each parameter. The calibrated present day DVRFS model was evaluated to assess the likely accuracy of simulated results. As part of the model evaluation, the regional water budget, the model fit, values of parameter estimates and their associated sensitivities, and boundary flows were evaluated.

Additional Drain Conductance Information (From Belcher and Sweetkind, 2010):

The Drain package simulates groundwater discharge through a head-dependent boundary. Groundwater is simulated as discharging from a finite-difference cell in which a drain is defined when the simulated head in the cell rises above a specified drain altitude. The simulated discharge is calculated as the drain conductance multiplied by the difference in altitude between the simulated head and the drain. The drain conductances are defined using the hydraulic properties of materials through which water flows to the surface. The drain conductances were estimated as part of model calibration. The drain altitudes were set equal to 10 m below the lowest land-surface altitudes for each group of cells. This value is assumed to represent a reasonable altitude below which ET would not occur and to account for springs being located in land-surface depressions that are lower than would be evident in the top surface of the HFM. This altitude would approximate the extinction depth for ET as well.

This is a past climate, steady state simulation, where evapotranspiration was not used as a means for removing groundwater. Drains have this same effect, but only if the water is at the land surface. Evapotranspiration removes water from the model even if it is not at the land surface.
Appendix B: Future Climate Predictions

As the climate has changed in the past, it will change in the future. The IPCC 4th Assessment combined 18 independent climate models and produced estimates of future climate change worldwide, broken down into regions. Under the A2 climate scenario, which is a business as usual climate scenario, the southwestern USA will have temperatures rising by 1 °C by as soon as the year 2020, and by 2090, as much as 3.5 to 4 °C higher. Warming of the planet will be greatest at higher latitudes, over land, and in the Northern Hemisphere (IPCC Working Group 1, 2007).

Large tracts of the planet are predicted to have increased precipitation (especially in latitudes greater than 50°), and overall global mean precipitation will increase by 4% over the oceans and 5% over land masses. However, the subtropics and areas between 20° and 40° latitude are likely to have decreasing precipitation, by as much as 20% (by the year 2099). The drying out will be to the greatest degree at the high latitude margins of the subtropics, caused by increased water vapor transport out of the subtropics and the extension of subtropical high pressure systems (IPCC Working Group 1, 2007). The climate projections all point to the southwestern United States experiencing increasing temperatures, less precipitation, more intense but less frequent rainfall events, and more severe droughts in the near future.

With changes in climate, there are likely to be negative effects on both the biota and the landscape. Globally the number of people living in stressed river basins is expected to rise from 1.4 billion to 4.3-6.9 billion by 2050 due to climate change. 20 to 30 percent of all species will be at a high risk of extinction with increases in temperature exceeding 2-3°C. Already, every available gallon of the Colorado River has been appropriated by farmers, industries and municipalities, and the region’s population is expected to continue growing (Gertner, 2007). Also, with decreased precipitation, groundwater recharge is likely to decrease considerably in these already water stressed areas. Increases in population and water demand will only exacerbate the rate of water depletion. Higher water pollution also will occur due to higher water temperatures, precipitation intensity, and longer periods of low flow. Without adequate adaptations to the coming climate changes, there will be very adverse human health effects through heat and pollution related problems (IPCC Working Group 1, 2007).