Case No. 84739

IN THE SUPREME COURT OF THE STATE OF NEVADA

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ADAM SULLIVAN, P.E., NEVADA STATE ENGINEER, et al.

Appellants,

vs.

LINCOLN COUNTY WATER DISTRICT, et al.

JOINT APPENDIX

VOLUME 41 OF 49

Fundamental Concepts of Recharge in the Desert Southwest: A Regional Modeling Perspective

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Recharge in arid basins does not occur in all years or at all locations within a basin. In the desert Southwest potential evapotranspiration exceeds precipitation on an average annual basis and, in many basins, on an average monthly basis. Ground-water traveltime from the surface to the water table and recharge to the water table vary temporally and spatially owing to variations in precipitation, air temperature, root zone and soil properties and thickness, faults and fractures, and hydrologic properties of geologic strata in the unsaturated zone. To highlight the fundamental concepts controlling recharge in the Southwest, and address the temporal and spatial variability of recharge, a basin characterization model was developed using a straightforward water balance approach to estimate potential recharge and runoff and allow for determination of the location of recharge within a basin. It provides a means for interbasin comparison of the mechanisms and processes that result in recharge and calculates the potential for recharge under current, wetter, and drier climates. Model estimates of recharge compare favorably with other methods estimating recharge in the Great Basin. Results indicate that net infiltration occurs in less than 5 percent of the area of a typical southwestern basin. Decadal-scale climatic cycles have substantially different influences over the extent of the Great Basin, with the southern portion receiving 220 percent higher recharge than the mean recharge during El Niño years in a positive phase of the Pacific Decadal Oscillation, whereas the northern portion receives only 48 percent higher recharge. In addition, climatic influences result in ground-water traveltimes that are expected to vary on time scales of days to centuries, making decadal-scale climate cycles significant for understanding recharge in arid lands.

1. INTRODUCTION

The purpose of this study was to develop a simple model for basin characterization that allows interbasin comparison of recharge mechanisms and the potential for recharge under current, wetter, and drier climates, and to highlight the fundamental concepts and mechanisms that control recharge in the deserts of the Southwestern United States (Southwest). The method developed allows analysis of climate change, as changes in precipitation and air temperature, to evaluate the potential for changes in ground-water recharge in the Great Basin and eventually in other areas in the Southwest. Without further refinements, this modeling approach primarily is intended to provide a means for hydrologically characterizing basins on a basin-wide or regional scale on the basis of fundamental concepts of recharge as they apply to southwestern desert environments. Estimates of recharge in basins of the Great Basin are presented for the purpose of illustrating the approach, evaluating relative proportions of recharge and runoff to describe the dominant mechanisms controlling recharge, and providing a comparison with other methods that have estimated recharge in the Great Basin. They are not relied on as accurate enough at this time to be used for assessment of water availability.

A basin characterization model (BCM) was developed for this study to determine the spatial and temporal variability of net infiltration (all terms are defined below), which is assumed to be equal to recharge because the model assumes steady state con-

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ditions and no lateral subsurface flow. The BCM uses a mathematical deterministic water-balance approach that includes the distribution of precipitation and the estimation of potential evapotranspiration, along with soil water storage and bedrock permeability. The BCM was used with available GIS data (digital elevation model, geology, soils, vegetation, precipitation, and air temperature maps), and GIS data that was developed for this study.

The BCM can be used to identify locations and climatic conditions that allow for excess water, quantifying the amount of water available either as runoff or as in-place recharge on a monthly basis, and allows inter-basin comparison of recharge mechanisms. The model does not distinguish between mountain front and stream channel recharge, which are referred to in this paper as runoff, nor does it explicitly define the percentage of runoff that becomes recharge. Because the accurate estimates of recharge cannot be calculated without further refinement to the BCM to estimate the partitioning of runoff, it calculates *potential* in-place recharge and *potential* runoff, and provides the distribution of both in a basin. These values can be combined using assumptions of the amount of runoff that results in recharge to estimate total potential recharge.

A simple calculation of traveltime through the unsaturated zone can be estimated if steady-state conditions are assumed and if unsaturated zone thickness and permeability data are available [*Flint et al.*, 2000]. The BCM can also be used to evaluate the potential for recharge under current, wetter, and drier climates, and is used to evaluate the role of decadal-scale climate cycles (El Niño/La Niña and the Pacific Decadal Oscillation) on recharge at a pixel scale (generally 30–270 meters) across the Southwest.

1.1 Terms and Concepts

Because many terms related to infiltration and recharge often have different meanings to different researchers, the terms used in this paper are defined and are consistent with those in most current literature. Infiltration is the entry into the soil of water made available at the ground surface [*Freeze and Cherry,* 1979]. Net infiltration is the quantity of water that moves below the zone of surface evapotranspiration processes [*Flint et al.,* 2001]. Under steady-state conditions, net infiltration is equal to recharge unless diverted to an area of flow from a spring and thus lost to evapotranspiration; even under this condition, one could argue that some recharge occurs, even if only to a small local or perched aquifer. Percolation (or drainage) is the process by which water moves downward through the unsaturated zone [*Flint et al.,* 2001]. Recharge is the entry into the saturated zone of water made available at the water-table surface [*Freeze and Cherry,* 1979]. Discharge is the removal of water from the saturated zone across the water-table surface [*Freeze and Cherry,* 1979].

Traveltime in the unsaturated zone is the time it takes for water that has become net infiltration to recharge the water table (hours to millennia); it is controlled by net infiltration, the thickness of the unsaturated zone, and the effective porosity of subsurface flow paths [*Flint et al.,* 2000]. As climate changes, the traveltime of infiltrating water through the unsaturated zone may vary; the spatial distribution of recharge also may vary. Recharge that occurs today is spatially variable owing to the thicknesses of soil and alluvium, the thickness of the unsaturated zone, and to the layering and properties of geologic and sediment strata. Recharge is temporally variable owing to changes in processes controlling net infiltration (primarily climate) for time scales of years to centuries.

Recharge is often discussed as dominant within one of the following basin locations: mountain block, diffuse, mountain front, stream channel, and playa lake. Mountain block recharge occurs directly into the underlying bedrock without runoff and is widely distributed in areas of higher mountainous terrain particularly where there is permeable bedrock. Diffuse recharge is areally distributed in alluvial valleys but away from the stream channels (similar to mountain block recharge). Mountain block recharge and diffuse recharge occur in direct response to the infiltration of rainfall and snowmelt and will be referred to in this paper as in-place recharge. In-place recharge also can occur in response to the local-scale lateral redistribution of rainfall and snowmelt following runoff and subsequent overland flow that does not reach the larger stream channels. Water that does not recharge in place is referred to as runoff in this paper. Runoff may become mountain front recharge, which occurs at boundaries between mountain blocks and deeper alluvial valleys, or beneath ephemeral streams as the streams transition from upland areas with thin soils to alluvial valleys and basins with thick soils. Stream-channel recharge occurs in response to focused or coalescing surface-water flows in ephemeral streams away from mountain fronts, or in perennial streams. Playa lake recharge occurs from runoff that collects and eventually evaporates or recharges under the playa (*Stephens*, 1995).

1.2 Study Area

The climate regime of the Southwest is generally considered arid to semi-arid [*UNESCO,* 1979]. Recently, researchers in the United States Geological

Survey (USGS) have been evaluating climate cycles in the Southwest [*Schmidt and Webb,* 2001]. As part of that evaluation a study was initiated to define the boundaries of the "dry" Southwest and to classify each hydrologic basin by climate [*Flint et al.*, 2003]. The United States is divided and sub-divided by the USGS into successively smaller hydrologic units, which are classified into four levels. Surface water drainage divides primarily define the boundaries of the hydrologic units, with the larger drainage systems often subdivided into smaller sub-drainages or areas. Each hydrologic unit is identified by a unique hydrologic unit code (HUC) with the smallest unit having eight digits [*Seaber et al.,* 1987]. The approach to assessing the climate regime is to evaluate the relation between precipitation and potential evapotranspiration in each of these eight-digit HUCs using an international arid land classification index. The Man and the Biosphere Program under the direction of the United Nations Educational, Scientific, and Cultural Organization [*UNESCO*, 1979] developed a method based on the ratio of annual precipitation to potential evapotranspiration. The UNESCO method produces five classes based on this ratio: hyper-arid (< 0.05), arid (0.05-0.2), semi-arid (0.2-0.5), dry-subhumid (0.5-0.65), and humid (>0.65). In order to define a study area for application of the BCM in an arid or semi-arid environment, these classes were applied to the average conditions for eight-digit HUCs defined by the USGS for the Southwest (Plate 1). There are areas that are calculated to be hyper-arid on a grid cell basis, such as Death Valley, that do not appear when averaged for an entire basin.

The Great Basin represents an arid environment and is centrally located within the Southwest. It was selected for a preliminary analysis to determine the feasibility of applying a simple basin characterization model for estimating recharge because of the ability to compare it to previous analyses of recharge in the Great Basin. The Great Basin study area is 374,218 km^2 and contains a total of 258 hydrologic units (hydrographic areas and subareas), which will be referred to as basins in this paper (Figure 1). Net infiltration or recharge has been estimated by previous investigators for the basins within the Great Basin using methods such as chloride mass balance [*Dettinger,* 1989], transfer equations based on other variables, such as precipitation using the Maxey-Eakin method [*Maxey and Eakin,* 1950; *Harrill and Prudic,* 1998], basin discharge estimates using evapotranspiration [*Nichols*, 2000], and water-balance and soil physics techniques [*Hevesi et al.,* 2002; *Hevesi et al.,* 2003].

1.3 Conceptual Model

A conceptual model of recharge is essential for developing the GIS-based BCM (Figure 2) [*Flint et al.*, 2001]. The conceptual model for a basin can be simplified to identify areas within a basin where recharge processes are initiated. Recharge does not occur everywhere in a basin nor does it occur each year. It is likely that the majority of the area contributing to recharge is a relatively small portion of the basin and years with above average precipitation and snow accumulation provide the most recharge [*Flint et al.,* 2001]. The BCM is used to identify those areas and climate conditions that are conducive to direct recharge or to runoff (which, in turn, could lead to recharge downstream). In discussion of a conceptual model, the term net infiltration is used to describe the surface processes, whereas the BCM assumes steady state conditions and net infiltration is equal to recharge.

For most of the Southwest on a yearly basis, and in most basins on a monthly basis, potential evapotranspiration exceeds precipitation [*Flint et al.,* 2003]. However, in certain areas of a basin (in particular, for the higher elevations), precipitation can exceed potential evapotranspiration and storage and net infiltration and/or runoff may occur, depending on the rate of rainfall or snowmelt, soil properties (including permeability, thickness, field capacity, and porosity), and bedrock permeability. For many basins, snow accumulated for several months provides enough moisture to exceed the soil storage capacity and exceed potential evapotranspiration for the month or months during which snowmelt occurs.

The conceptual model assumes that all processes controlling net infiltration occur within the top 6 m of the surficial materials as shown by *Flint and Flint* [1995] for Yucca Mountain in the southern Great Basin. This is a conservative estimate for the Southwest, and is only likely to occur in riparian zones where deeper-rooted vegetation can retrieve water that has penetrated deeper than six meters. Although these zones are an extremely small percentage of the area in the Southwest, and particularly the Great Basin, if high resolution information on vegetation type for these areas is available, the model should be adjusted to use the appropriate rooting depth. The alternate process of exfiltration in arid environments whereby water is drawn upward from the soil profile under vapor density gradients and evaporates at the surface to provide a negative water balance, although important in characterizing deep alluvium, is considered negligible on a basin or regional scale for the purposes of this analysis.

The BCM uses spatially distributed estimates of monthly precipitation, monthly air temperature, monthly potential evapotranspiration, soil water storage, and bedrock permeability to determine the area in a basin where excess water is available. Potential evapotranspiration is modeled and partitioned on the basis of vegetation cover to represent bare soil evaporation and vegetation evapotranspiration. Depending on the soil and bedrock permeability, excess water is partitioned as either (1) in-place recharge, or (2) runoff that can potentially become mountain front recharge or stream channel recharge either at the mountain front or farther downstream in the alluvial basin.

Net infiltration occurs when enough water is made available to exceed the storage capacity of the soil (or rock); precipitation, snowmelt, or run-on provide the water; root zone, soil depth, porosity, and the soil drainage characteristics provide the storage; vegetation, bare soil surfaces, and the energy balance control potential evapotranspiration, which decreases soil water content thus increasing soil water storage between precipitation/snowmelt/run-on events. The topography and atmospheric conditions control much of the energy available for potential evapotranspiration.

For thin soils underlain by fractured bedrock the soil water content will approach saturation because the water entry potential of the fracture network must be exceeded before significant drainage into the underlying bedrock can occur (the fracture network is a capillary barrier to drainage from the soil). In locations with thick soil a greater volume of water is needed (compared to thin soil locations) to exceed the storage capacity of the root zone, which is deeper relative to locations with thin soil, (or the permeability must be high enough to quickly drain the root zone (e.g., young gravelly channels)). In general, bedrock permeability, soil storage capacity, and evapotranspiration are the factors that determine the vertical direction of water flow. In upland areas with thin soils, soil thickness is the most important factor affecting soil storage capacity. If the soil is thin and bedrock permeability is low then evapotranspiration has more time to remove stored water between precipitation, snowmelt, and run-on events. If the bedrock permeability is high then evapotranspiration has less time to remove stored water between events. In alluvial fans, basins, and valleys with thick soils and deeper root zones, if the soil field capacity is high and the permeability is low (for example, finer grained soils) then drainage through the root zone occurs slowly and evapotranspiration has more time to remove stored water between events. If the soil field capacity is low and the permeability is high (coarser grained soils) then drainage through the root zone occurs more rapidly and evapotranspiration has less time to remove stored water between events.

Where net infiltration occurs in the Southwest is very important, particularly if one intends to quantify or analyze it by means of field measurements. For example, measuring stream flow losses or calculating Darcy flux from data obtained under a stream channel would not provide an accurate estimate of recharge in a basin dominated by in-place recharge processes. To determine approximately where recharge is occurring and what mechanisms dominate, all available information was assembled (GIS coverages), combined with the conceptual model, to calculate locations within a basin where recharge is likely to occur. Because the spatial and temporal distribution of net infiltration is dependent on precipitation, soil water storage, bedrock permeability, and evapotranspiration, all of which can be estimated with available data on a regional scale, the most probable locations for potential in-place recharge and potential runoff can be identified. In the BCM, potential in-place recharge is calculated as the maximum volume of water for a given time frame that can recharge directly into bedrock or alluvium. Potential runoff is the maximum volume of water for a given time frame that will run off the mountain front or become streamflow. Total potential recharge is the combination of in-place recharge and runoff and assumes that all runoff becomes recharge. Analyses of basins using the water balance approach in the BCM can help determine when, where, and how the water-balance terms, the material properties, and the physical mechanisms can be combined to produce net infiltration or recharge.

1.4 Recharge and Groundwater Traveltime

An important issue to be addressed is the timing of recharge after net infiltration occurs. It is quite likely that if predictions of drier climate over the next 20 years [*Schmidt and Webb,* 2001] prove to be true, that this would reduce net infiltration values both spatially and temporally. It is also likely that some basins will not experience a change in recharge related to this climate change for hundreds or thousands of years. Therefore, an analysis of unsaturated zone traveltime is needed to determine when changes in surface processes will be reflected at the water table. Assuming negligible traveltime for net infiltration from 0-6 m and vertical flow through the unsaturated zone, traveltime is controlled by the net infiltration rate, unsaturated zone thickness, the effective porosity of the flow path, and the lowest permeability encountered along a given flow path (which would determine the maximum net infiltration rate at which the assumption of vertical flow would still apply). Unsaturated zone traveltime controls the timing of recharge; therefore ground-water responses to changes in climate (seasonal, yearly, or decadal) may be delayed, suggesting important implications for water availability under future climate scenarios.

Unsaturated zone traveltime can be calculated as $(\phi_{eff}Z_{uz})/ I_{net}$, where ϕ_{eff} is effective unsaturated zone porosity (m/m), Z_{uz} is the thickness of the unsaturated zone (m), and I_{net} is net infiltration (m/yr) [*Flint et al.*, 2000]. *Flint et al.* [2000] estimated the thickness of the unsaturated zone in the Death Valley region on the basis of the difference in elevation determined using a digital elevation model and the spatially interpolated water-table elevation. The effective unsaturated zone porosity is the most difficult parameter to assess. It can be evaluated using detailed geologic maps from the surface to the water table and an estimate of the porosity of the rock matrix and(or) fractures of the geologic material. An estimate of subsurface bedrock permeability of the matrix can also be useful in helping to estimate effective unsaturated zone porosity. If the estimated net infiltration is less than the matrix permeability then the flow is likely in the matrix and matrix saturation becomes a good estimate for ϕ_{eff}. If net infiltration is more than matrix permeability then the flow is likely in the fractures. In this case, an estimate of fracture porosity becomes a good estimate for ϕ_{eff}. Either case can help determine whether a high porosity (matrix flow dominated) or a much lower porosity (fracture flow dominated) should be used.

Flint et al. [2000] showed traveltime delays of 10's to 1,000's of years for the southern Great Basin due to variation in net infiltration rates and the thickness of the unsaturated zone, which is commonly 10-100 m thick, but can exceed 2,000 m in thickness. Although parts of the regional flow system may respond quickly to climate change, others may lag behind significantly. This variability may be significant in determining the rate and direction of groundwater flow and the resultant availability of groundwater as a resource.

2. METHODS

2.1 Water-balance Calculations

A series of water-balance equations were developed to calculate the area and the amount of potential recharge. For example, each model grid cell was analyzed for each month to determine water availability for recharge. This available water (AW) for potential recharge, potential runoff, or water to be carried over to the following month is defined as

$$AW = P + S_m - PET - S_a + S_s \qquad (1)$$

where P is precipitation, S_m is snowmelt, PET is potential evapotranspiration, S_a is snow accumulation and snow pack carried over from the previous month, and S_s is stored soil water carried over from the previous month. All units are in millimeters per month. Potential runoff was calculated as the available water minus the total storage capacity of the soil (soil porosity multiplied by soil depth). Potential in-place recharge was calculated as the available water remaining (after runoff) minus the field capacity of the soil (the water content at which drainage becomes negligible). Maximum in-place recharge on a unit grid cell basis is the permeability of the bedrock (cm^3 of water per cm^2 grid cell area per month). If the total soil water storage is reached, the potential in-place recharge is equal to the bedrock permeability. Any water remaining after the monthly time step would be carried over into the next month in the S_s term.

Soil water storage capacity and soil infiltration capacity were estimated using soil texture estimates from the State Soil Geographic Database (STATSGO; http:// www.ftw.nrcs.usda.gov/stat_data.html), a state-compiled geospatial database of soil properties that generally are consistent across state boundaries [*U. S. Dept. of Agriculture-National Resource Conservation Service*, 1994]. Soil thickness was estimated using available geologic maps to estimate soil depths of 6 m wherever quaternary alluvial deposits were mapped [*Hevesi et al.*, 2003]. Everywhere else, the STATSGO database was used, which provides soil depths to 2 m. Bedrock permeability was estimated using a bedrock geologic map and literature values for the estimation of permeability on the basis of geologic material [*Bedinger et al.*, 1989]. Macropore and fracture flow is considered within the bedrock permeability estimation, which assumes values on the basis of measured bulk permeabilities at the surface or borehole transmissivities. Uncertainties in soil and bedrock properties are discussed in *Hevesi et al.*, [2003].

The ratio of potential runoff versus potential in-place recharge determines whether mountain front and(or) stream channel recharge mechanisms dominate relative to in-place recharge in response to rainfall and snowmelt (in other words, the significance of surface-water flow to total recharge increases as the ratio of runoff to mountain block recharge increases). This ratio does not determine where the runoff infiltrates so it can not distinguish between mountain front or stream channel recharge that may occur farther into the basin. The BCM model allows snow pack and soil moisture to be carried over from month to month, which becomes important when temperatures are cold enough for precipitation to form snow. Since snow may persist for several months before melting, large volumes of water may be made available for potential recharge in a single monthly model time step.

2.2 Climate Distribution

Climate was simulated in this study for the Great Basin using two approaches to evaluate the difference in recharge estimates between (1) average climate conditions for 34 years, January 1, 1956, through December 31, 1999, where spatially distributed estimates of mean monthly precipitation and mean monthly maximum and minimum air temperature were used, and (2) time-varying climate conditions, where spatially distributed estimates of monthly precipitation and maximum and minimum monthly air temperatures for the 34 year period were used. These estimates were made using historical daily precipitation and air temperature data from a network of 448 monitoring stations in and adjacent to the Great Basin [*National Climatic Data Center*, 2000a,b] that existed between 1900-1999. Approximately 300 stations were active at any given time for the 34 year period. The measured values of precipitation and minimum and maximum air temperature were spatially distributed to all the grid cells for the Great Basin model domain (270 x 270 m) using a model from *Nalder and Wein* [1998] that combines a spatial gradient plus inverse distance squared weighting to monthly point data to interpolate to each grid cell with multiple regression. Parameter weighting is based on location and elevation following the equation:

$$Z = \left[\sum_{i=1}^{N} \frac{Z_i + (X - X_i) \times C_x + (Y - Y_i) \times C_y + (E - E_i) \times C_e}{d_i^2} \right] / \left[\sum_{i=1}^{N} \frac{1}{d_i^2} \right]$$

where Z = estimated climatic variable, Zi is the value of climate station I, Xi, Yi, Ei are easting, northing, elevation of climate station I, N is the number of climate stations, Di is the distance from the site to climate station I, and Cx, Cy, Ce are regression coefficients for easting, northing, elevation.

Snow depth was calculated for areas where precipitation occurs and air temperature is at or below freezing. Sublimation of snow was calculated as a percentage of evapotranspiration, and snowmelt was based on net radiation when air temperatures were above freezing.

2.3 Potential Evapotranspiration

Potential evapotranspiration was estimated using a computer program modified from *Flint and Childs* [1987] that calculates solar radiation for each grid cell in the model domain, and when combined with air temperature, is converted to net radiation and soil heat flux [*Shuttleworth*, 1993]. The result was used with the Priestley–Taylor equation [*Priestley and Taylor*, 1972] to estimate potential evapotranspiration, and was corrected for vegetated and bare soil area using estimates of vegetation cover from vegetation maps (National Gap Analysis Program; http://www.gap.uidaho.edu). Actual evapotranspiration is a function of soil moisture and is more rigorously addressed in *Hevesi et al.* [2002, 2003]. The regional scale approach used with the BCM assumes that potential evapotranspiration can be used to provide a potential estimate of recharge to bound the values for evaluating mechanisms and differences among basins. Following refinement and the incorporation of actual evapotranspiration in the BCM, recharge can more accurately be estimated for more intensive applications.

2.4 Basin Application

The BCM code is written in FORTRAN-90, and uses ASCII files of distributed upper boundary conditions and GIS grid files of surface properties as input for the calculations of potential recharge and potential runoff. The BCM was applied to the Great Basin using the two different simulation scenarios (mean monthly climate and 34-year monthly time series from 1956-1999) to evaluate the relative amount of recharge and the mechanisms that would dominate under wetter or drier climatic conditions. Consideration of snow accumulation can be critical because the accumulation can delay the application of water to the surface thus extending the possibility that in the following month the combination of precipitation and snowmelt will exceed the storage capacity of the soil causing net infiltration and(or) runoff. The BCM estimates for the Great Basin were compared with recharge estimates determined using the Maxey–Eakin approach from *Harrill and Prudic* [1998], chloride-mass balance estimates of *Dettinger* [1989] published in *Harrill and Prudic* [1998] for the Great Basin, basin discharge estimates determined using evapotranspiration [*Nichols*, 2000], and net infiltration estimates determined using a daily water-balance model *Hevesi et al.* [2002, 2003].

3. RESULTS AND DISCUSSION

Total mean annual potential recharge (mean annual potential recharge plus potential runoff) estimates were made on a grid cell basis for 258 basins in the Great Basin and are presented in Plate 2. Total mean annual potential recharge for each basin is presented in Plate 3 and Table 1. The results shown in Figure 4 indicate that most of the in-place recharge or runoff occurs at, or is generated from, the basin boundaries.

This was an expected result because the basin boundaries primarily occur along the drainage divides, and the divides tend to have higher elevations (thus higher precipitation and lower air temperature) and thinner soils relative to the soils in the central part of each basin.

3.1 Evaluation of Recharge Processes

Results of the mean monthly calculations indicate that there is 2.41 million acre-feet/year of potential in-place recharge in the Great Basin and 4.83 million acre-feet/year of potential runoff, or a total potential recharge of 7.24 million acre-feet/year. Results of the 34-year time series calculations indicate that there is slightly more recharge when water can be carried over between months: 2.43 million acre-feet/year of potential in-place recharge, and 5.24 million acre-feet/year of potential runoff, or a total potential recharge of 7.67 million acre-feet/year. Although the amount of recharge that occurs as a result of runoff is not known, based on analyses performed by *David Prudic* [*U.S. Geological Survey, personal communication,* 2001] and *Hevesi et al*. [2003], it was assumed that about 10 percent of runoff becomes recharge in the southern part of the Great Basin and as much as 90 percent in the northern part. This results in a total potential recharge for the 34-year time series of 2.95 million acre-feet/year. This is a conservative estimate, as it currently is not known what the spatial or temporal distribution of the recharge portion of runoff is. The percentage is probably a function of the timing of precipitation and snowmelt, topographic position, and the hydrologic properties of alluvium and bedrock, and deserves further investigation during future BCM refinement.

Grid-based estimates of the ratio of potential in-place recharge to potential runoff are presented in Plate 4, the ratio of the calculated means of potential in-place recharge to potential runoff for each basin is presented in Plate 5. The ratio of potential in-place recharge to potential runoff provides an indication of the mechanisms that likely are dominant in controlling recharge. The grid-based analysis provides the distribution within basins of the dominant mechanisms, whereas the mean basin values provide a larger scale representation for basin comparison and regional analysis. A ratio of 0.5 or less indicates that more than twice as much water has the potential to become runoff than to become in-place recharge. A ratio of 2.0 or greater indicates that water has at least twice as much potential to become in-place recharge than to become runoff. An example of the control that bedrock type, and thus bedrock permeability, has on the calculation of recharge is apparent in Plate 5 with the observation that the major assemblage of basins in which in-place recharge is dominant (> 2.0) (noted as extending from the southern portion of the Great Basin through the central region and to the northeast), coincides with the carbonate-rock province which is dominated by high permeability bedrock. The role that bedrock plays in the determination of recharge mechanisms is supported with the use of a detailed water-balance model for the Death Valley region by *Hevesi et al*. [2002], which showed much higher recharge in basins dominated by carbonate rock and lower recharge in basins dominated by thick soils and lower permeability volcanic rock types [*Hevesi et al.*, 2002; Figure 9].

3.2 Effect of Climate Variability

The role of climate variability is highlighted in an evaluation of potential recharge in two basins, Continental Lake Valley in the northern Great Basin, and Valjean Valley in the southern Great Basin (Plate 5). The total potential recharge is about 2,000 acre-feet/year for Continental Lake Valley and about 270 acre-feet/year for the Valjean Valley (Table 1). These values were calculated as all the in-place recharge plus 10 percent of runoff, using the 34-year time series approach. To illustrate the climatic conditions responsible for the resultant difference in recharge between the basins, the percentage deviation in annual potential recharge from the mean is shown in Figure 3, calculated as the difference between the total potential recharge for each year and the mean total potential recharge for the 34-year period from 1956-1999. The 34-year simulation period includes positive and negative phases of the Pacific Decadal Oscillation (PDO), a southern oscillation index of an approximately 40-year climatic cycle, and several El Niño cycles [*Dettinger et al.,* 2000]. Both basins appear to be influenced by the shift in the PDO in 1977 from a negative phase to a positive phase, and both basins are influenced by El Niño years, noted as open diamonds, during the positive PDO, although the influence is much stronger in the Valjean Valley. During El Niño years with a positive PDO, the mean annual total potential recharge in the Valjean Valley is about 220 percent higher than the 34-year mean; during El Niño years with a negative PDO, recharge is about 13 percent lower (recharge for all the years with a positive PDO is about 55 percent higher than the 34-year mean, and recharge for all the years with a negative PDO is about 48 percent lower) (Figure 3). In the Continental Lake Valley, annual total potential recharge for the El Niño years with a positive PDO is about 48 percent higher than the 34-year mean, and recharge for El Niño years with a negative PDO is about 43 percent lower (recharge during all years with a positive PDO is about 37 percent higher than the 34-year mean, and recharge during all the years with a negative PDO is about 37 percent lower). A comparison of annual total potential recharge estimated for non-El Nino years with the mean recharge for 1956 through 1999 showed that non-El Nino recharge was 97 percent lower than the 34-year mean for the Valjean Valley and 46 percent lower for the Continental Lake Valley. This suggests that actual climate data rather than a mean value for recharge should be used in the BCM.

The influence of climate on recharge can also be seen on a plot comparing the percentage deviation of mean annual total potential recharge from the 34-year mean and the percentage deviation of mean annual precipitation from the 34-year mean for the Valjean Valley and the Continental Lake Valley basins. The range of precipitation for the Valjean Valley is much wider than that for the Continental Lake Valley because of the stronger influence of El Niño years in the southern part of the Great Basin (Figure 4). This results in more scatter in the recharge estimates for years with high precipitation and occasionally much more recharge.

The expected climate conditions for the next 20 years, which have not yet been modeled, probably will provide less snow accumulation in the higher elevations [*Schmidt and Webb,* 2001] and therefore less net infiltration, which would greatly reduce the potential for mountain block and mountain front recharge. If the predicted warmer and drier climate occurs, recharge during the next 20 years will result in lower net infiltration to desert-basin aquifers, which eventually would result in lower recharge; the response to the predicted climate may be delayed 10s to 1000s of years. Only where traveltimes in the unsaturated zone are less than 20 years would there be a response to recharge for the drier and warmer climate scenario. When the details of the climate scenarios are better defined, the BCM can be used as a more direct indicator of recharge for each basin.

3.3 Comparison with Other Methods

Total potential recharge (shown on a log scale) calculated for 258 basins in Table 1 is presented in Plate 6, sorted from lowest to highest recharge calculated using the BCM and the 34-year time series approach. Estimates of total potential recharge, calculated as mean in-place potential recharge plus 10 percent of the potential runoff using the BCM was compared with estimates of recharge made using the Maxey–Eakin method [*Harrill and Prudic,* 1998], the chloride-mass balance method [*Dettinger,* 1989], and the daily water-balance model of *Hevesi et al.* [2002 (INFILv1); 2003 (INFILv3)]. The BCM improves estimates over that of the Maxey–Eakin method because it takes into account the spatially distributed features of the surface, such as bedrock permeability and soil storage capacity, as well as potential evapotranspiration, rather than only precipitation. The remaining methods compare to the BCM time-series results within an order of magnitude, with the exception of one chloride mass balance point and one water-balance model (INFILv1) point. The BCM results using the average monthly conditions have less total potential recharge for about half of the basins.

The range of estimates for each basin is indicated by a bar that is constructed by subtracting the 10 percent of runoff that is assumed to become recharge (which then assumes that no runoff results in recharge), and adding the other 90 percent (which then assumes that all runoff results in recharge). This results in a range of estimates for each basin that reflects the possible assumptions of no runoff resulting in recharge to all runoff resulting in recharge. The large range in total potential recharge given the possible assumptions regarding runoff, particularly at the higher recharge rates, indicates the need to further develop the BCM to differentiate between and quantify runoff that occurs in the mountain front areas to become recharge and the amount of runoff in the streams that becomes recharge. The red diamonds in Plate 6 show the results of the BCM determined using mean monthly climate estimates for a 12-month period rather than the monthly time series for a 34-year period. The basins with the largest recharge values have very similar estimates using either the time series or the monthly averages.

4. SUMMARY

Recharge is temporally and spatially variable and is controlled, to a large extent, by the near surface process of net infiltration. Net infiltration is a function of precipitation, air temperature, root zone and soil properties and depth, and bedrock permeability. Present-day net infiltration is assumed to be equivalent to potential future recharge on a regional basis but can be significantly delayed by traveltime through the unsaturated zone. The monthly water-balance method presented here provides a straightforward approach to compare the potential for net infiltration between basins for current or different climates. If using mean monthly precipitation, potential recharge in the Great Basin is estimated to be between 2.41 million acre-feet/year (including only in-place potential recharge) and 7.24 million acre-feet/year (including in-place potential recharge plus all potential runoff). Total estimated potential recharge including only 10 percent of potential runoff is 2.89 million acre-feet/year. A mean annual precipitation produces less recharge than the mean of the time series of years making climate variability an important consideration in analyzing recharge in desert environments. These calculations result in potential recharge estimated to be between 2.43 million acre-feet/year (including only in-place potential recharge) and 7.67 million acre-feet/year (including in-place potential recharge plus all potential runoff). Total estimated potential recharge including only 10 percent of potential runoff is 2.95 million acre-feet/year. Because net infiltration and recharge are temporally and spatially variable and often only occurs in 5 percent of a basin, an a priori estimate of the mechanisms and processes contributing to recharge and locations it occurs are an important precursor to locating field measurements used to quantify actual recharges rates.

Additional research is necessary to refine the BCM for use in providing more accurate estimates of in-place recharge and runoff, and particularly to quantify and differentiate between runoff that occurs in shallow alluvium at the mountain front or in ephemeral streams, and runoff that occurs in deeper alluvium under ephemeral or perennial stream channels. In addition, the apparent importance of using a time-series analysis in characterizing desert recharge suggests that a daily time scale would result in even more realistic estimates of recharge and runoff, better capturing the time scale at which precipitation and snowmelt occurs. Surface routing of water to adjacent grid cells would also improve the estimates of surface infiltration, especially if the BCM were used on a fine grid scale, such as 10 or 30 m. These refinements would likely require the use of coding for parallel processing for application of the BCM to basin-scale or regional-scale analyses. Finally, although it is beyond the scope of this paper to address this topic fully, the changes in vegetation type, density, and rooting depth, particularly in riparian zones, that likely would occur with decadal-scale changes in climate should be taken into consideration alongside the development of climate scenarios to include the associated changes in potential evapotranspiration.

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Figure 1. Map showing the aridity classification of groundwater basins in the southwestern United States. Classification based on arid land classification index of the United Nations Educational, Scientific, and Cultural Organization (UNESCO).

Figure 2. Hydrographic areas and subareas within the Great Basin and identifiers.

Figure 3. Schematic of mechanisms controlling net infiltration.

Figure 4. Total mean annual potential recharge, calculated from potential recharge plus potential runoff on a grid cell basis, for basins in the Great Basin.

Figure 5. Total mean annual potential recharge, calculated from potential recharge plus potential runoff as the mean of all grid cells for each basin in the Great Basin.

Figure 6. The ratio of potential in-place recharge to potential runoff, calculated on a grid-cell basis, for basins in the Great Basin, indicating locations where either in-place recharge or runoff are the dominant mechanisms.

Figure 7. The ratio of potential in-place recharge to potential runoff, calculated as the mean of all grid cells for each basin in the Great Basin indicating basins where either in-place recharge or runoff are the dominant mechanisms.

Figure 8. Annual potential recharge, as percentage deviation from the mean potential recharge for 1956–1999 for Continental Lake Valley in the northern part of the Great Basin and the Valjean Valley in the southern part of the Great Basin, indicating differences in recharge for El Niño years owing to negative and positive Pacific Decadal Oscillation (PDO).

Figure 9. Comparison of annual potential recharge, as percentage deviation from the mean potential recharge for 1956-1999, and annual precipitation, as percentage deviation from the mean for 1956-1999 for the Valjean Valley in the southern part of the Great Basin and the Continental Lake Valley in the northern part of the Great Basin.

Figure 10. Total potential recharge, calculated as potential in-place recharge plus 10 percent of potential runoff for 258 basins in the Great Basin determined using several methods of estimating recharge and the basin characterization model (BCM). Range bars indicate inclusion or exclusion of all runoff with in-place recharge.

TABLES

Table 1. Total mean potential recharge (acre-feet/year) calculated for 258 basins in the Great Basin calculated using several methods of estimating recharge and potential in-place recharge and potential runoff calculated several ways using the basin characterization model.

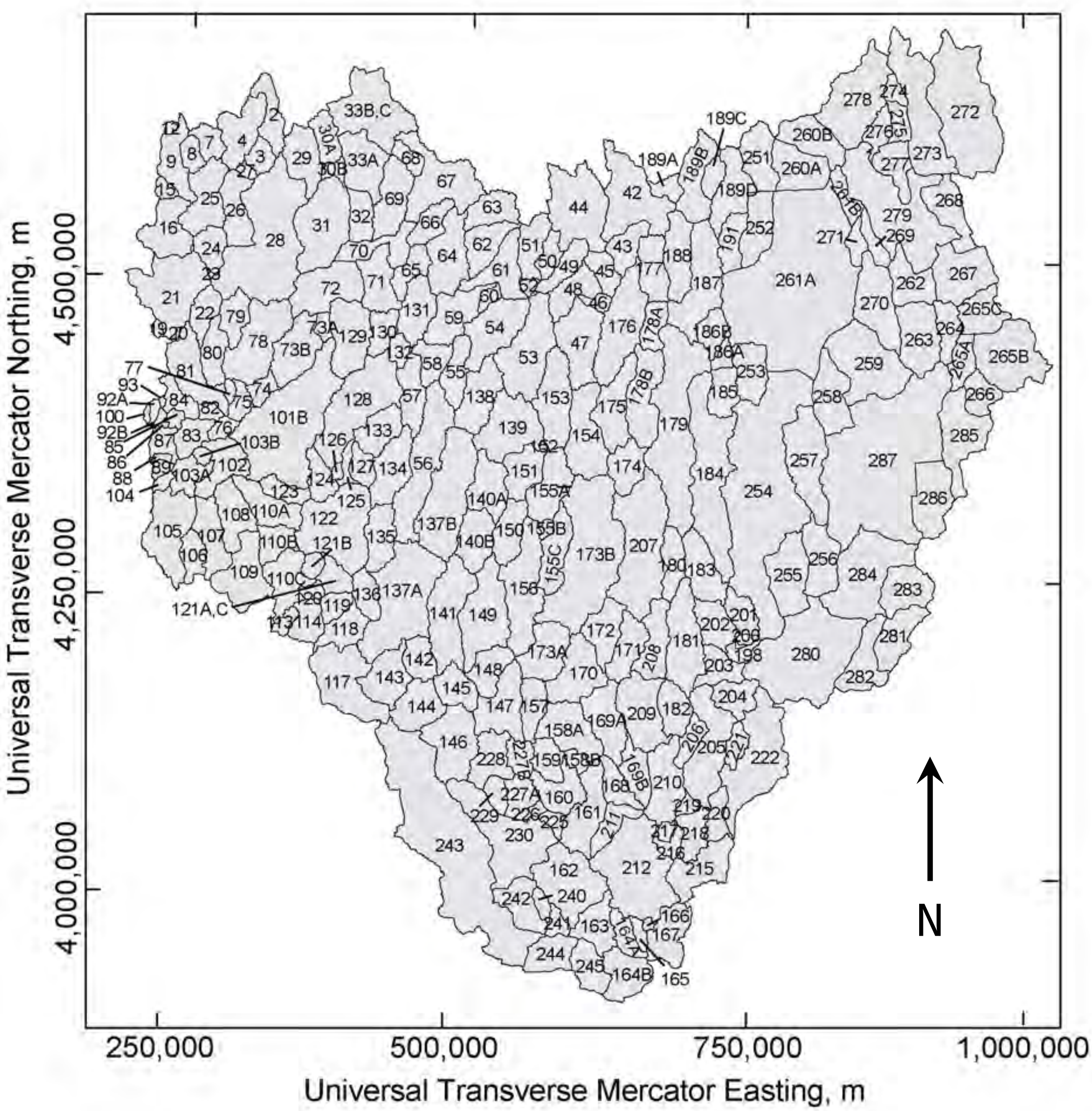

Figure 1. Hydrographic areas and subareas within the Great Basin and identifiers. (From *Harrill and Prudic* [1998])

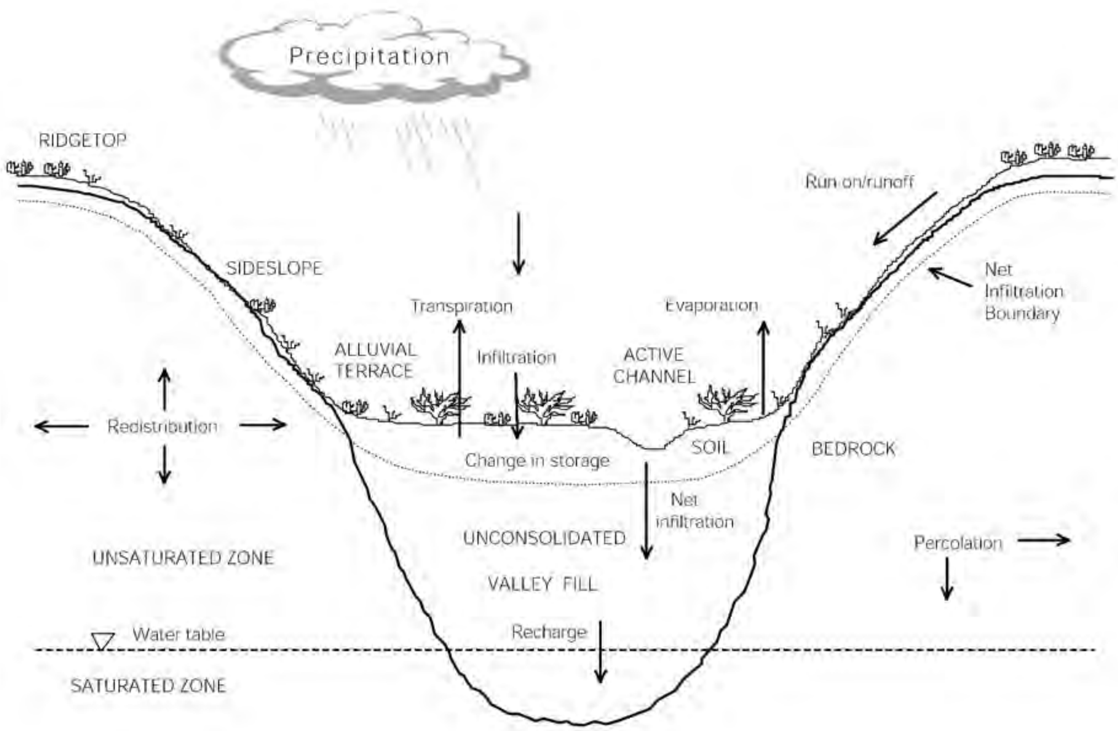

Figure 2. Schematic of mechanisms controlling net infiltration. (From *Hevesi et al.* [2003])

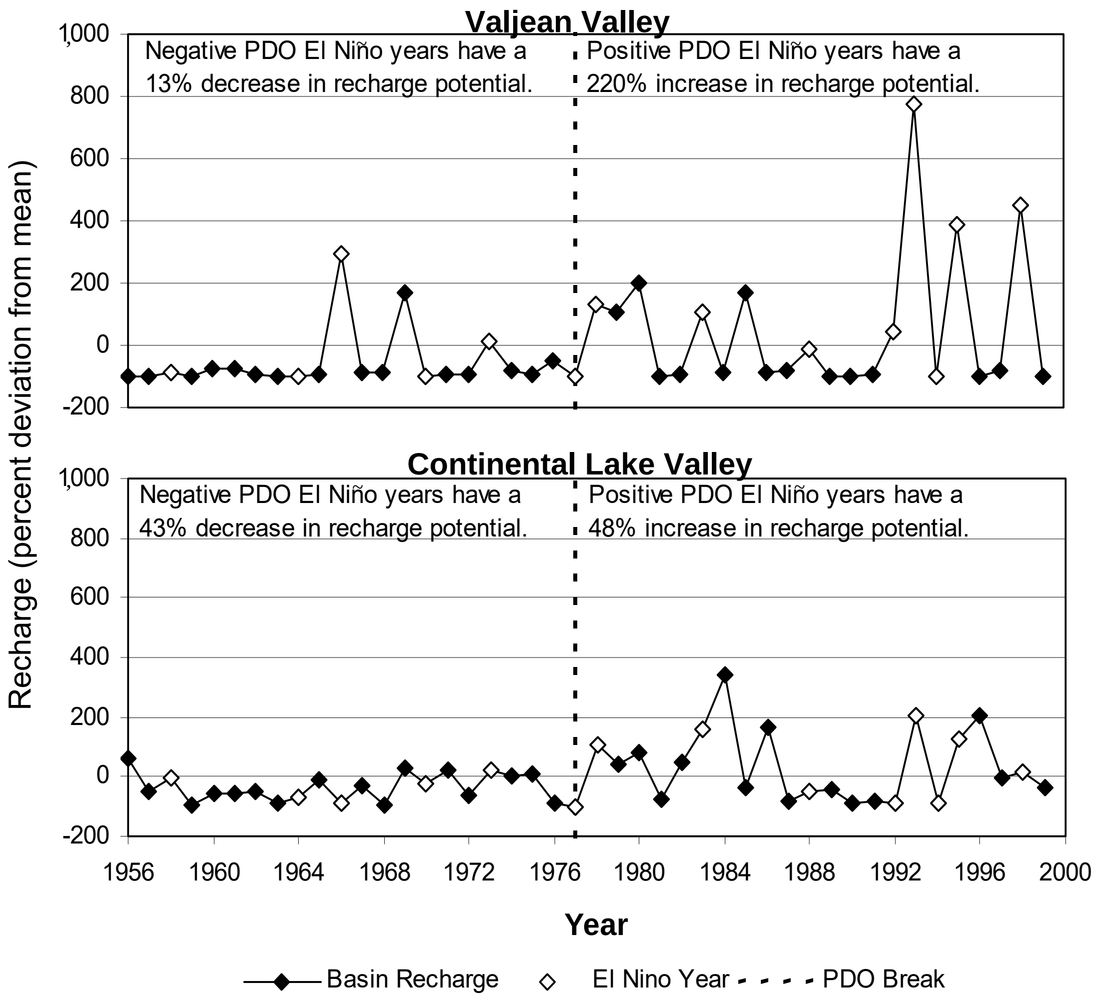

Figure 3. Annual potential recharge, as percentage deviation from the mean potential recharge for 1956–1999 for Continental Lake Valley in the northern part of the Great Basin and the Valjean Valley in the southern part of the Great Basin, indicating differences in recharge for El Niño years owing to negative and positive Pacific Decadal Oscillation (PDO).

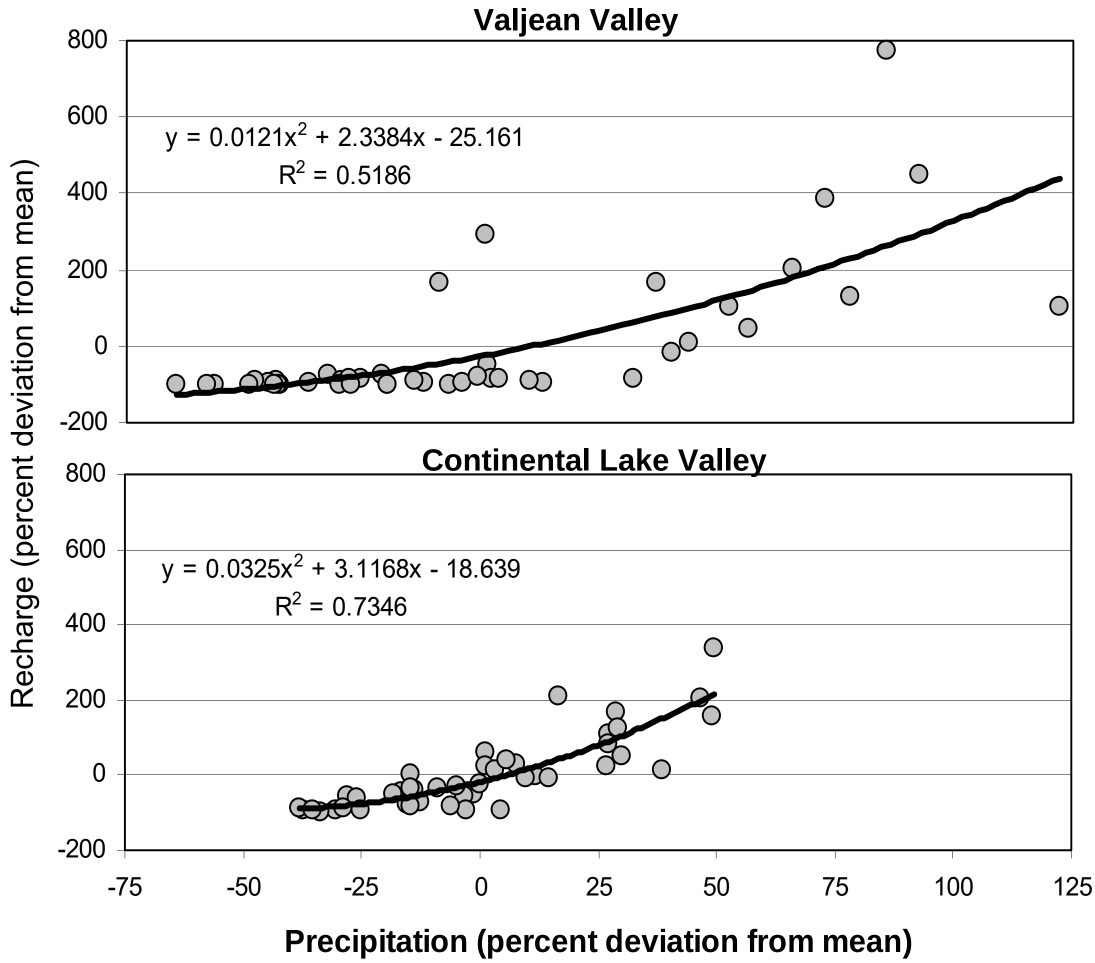

Figure 4. Comparison of annual potential recharge, as percentage deviation from the mean potential recharge for 1956-1999, and annual precipitation, as percentage deviation from the mean for 1956-1999 for the Valjean Valley in the southern part of the Great Basin and the Continental Lake Valley in the northern part of the Great Basin.

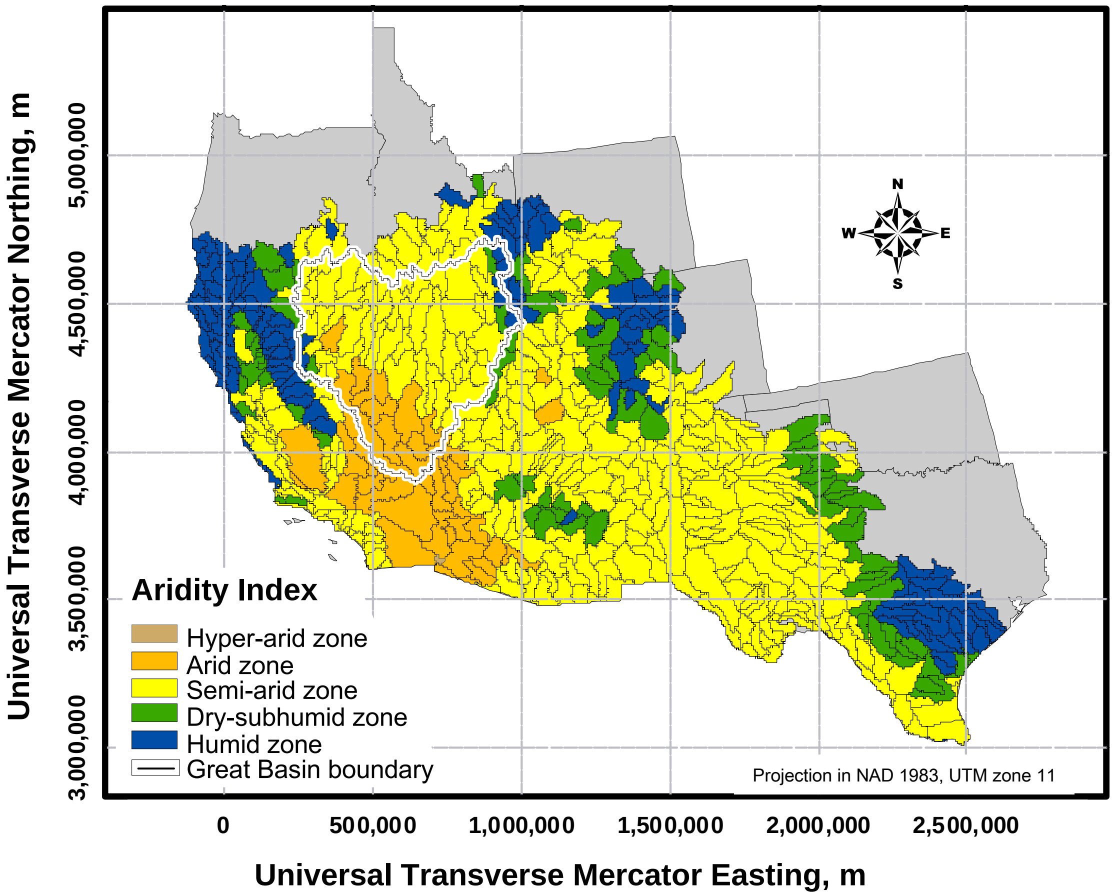

Plate 1. Map showing the aridity classification of ground-water basins in the southwestern United States. Classification based on arid land classification index of the United Nations Educational, Scientific, and Cultural Organization (UNESCO).

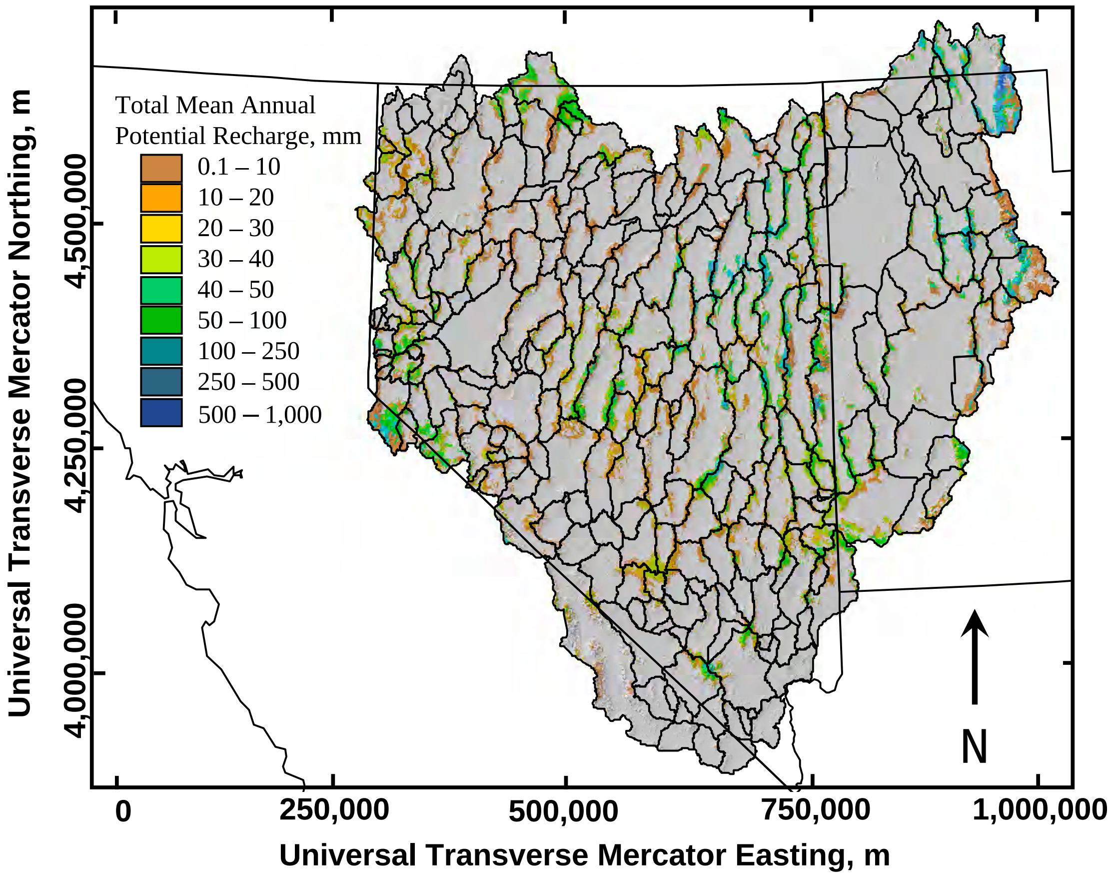

Plate 2. Total mean annual potential recharge, calculated from potential recharge plus potential runoff on a grid cell basis, for basins in the Great Basin.

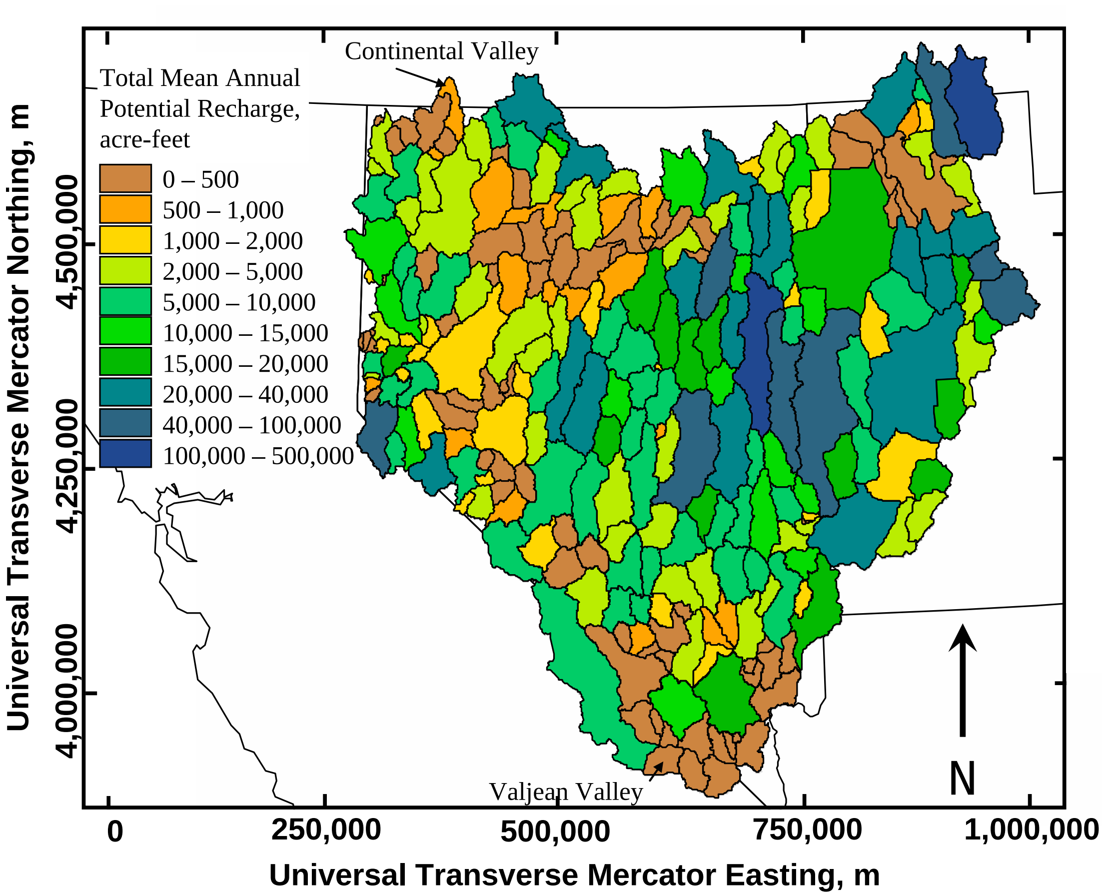

Plate 3. Total mean annual potential recharge, calculated from potential recharge plus potential runoff as the mean of all grid cells for each basin in the Great Basin.

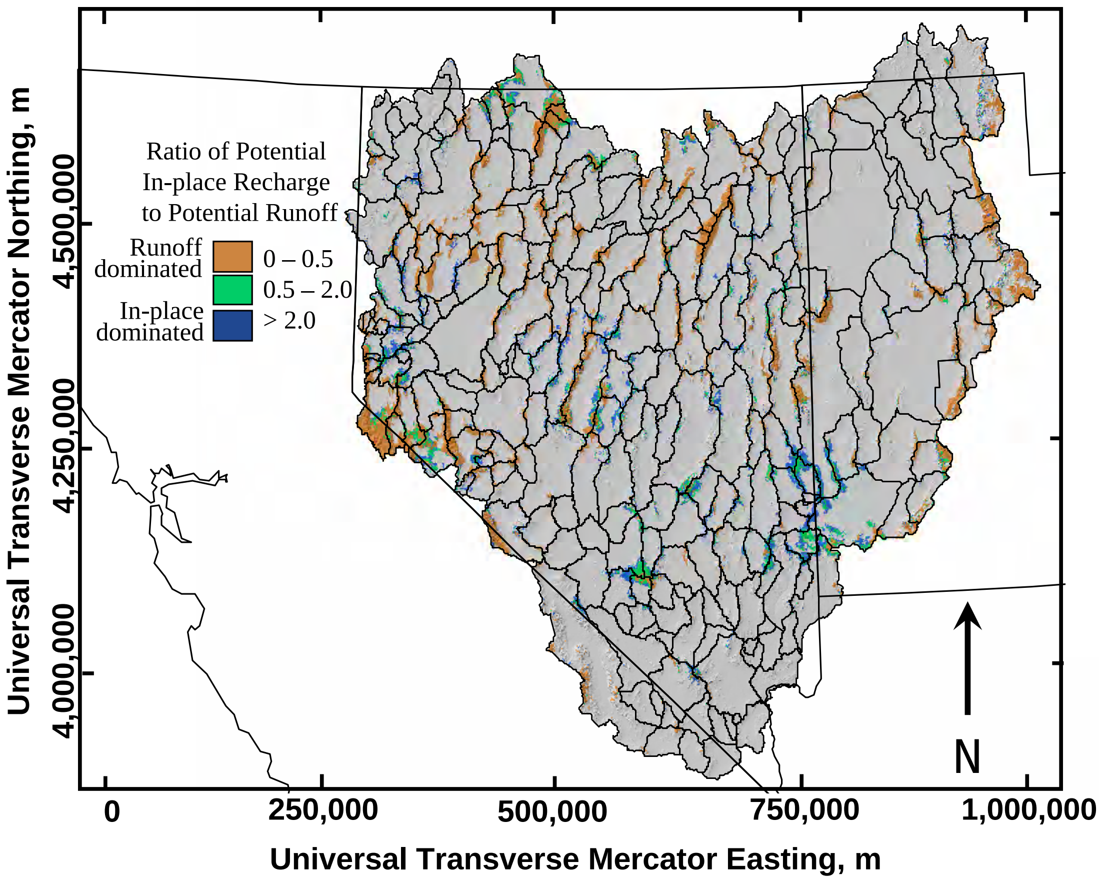

Plate 4. The ratio of potential in-place recharge to potential runoff, calculated on a grid-cell basis, for basins in the Great Basin, indicating locations where either in-place recharge or runoff are the dominant mechanisms.

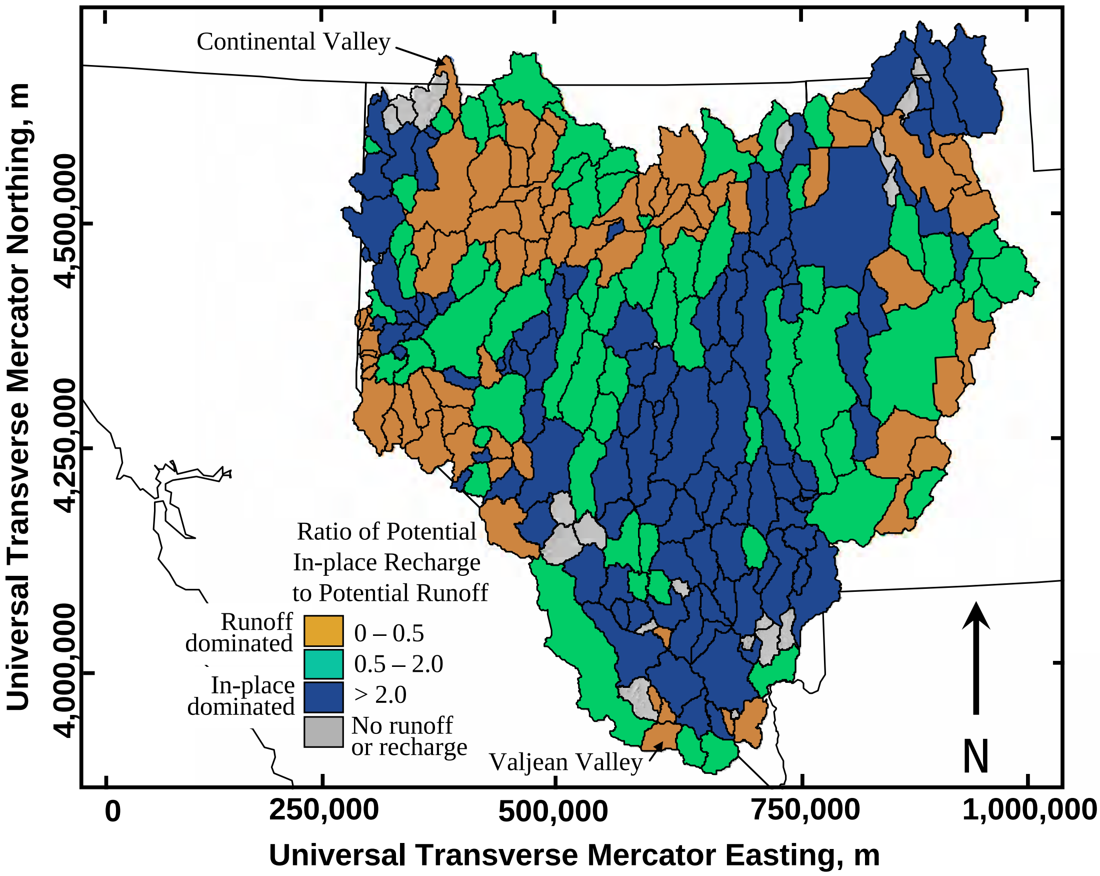

Plate 5. The ratio of potential in-place recharge to potential runoff, calculated as the mean for each basin, for the Great Basin, indicating locations where either runoff of in-place recharge are the dominant mechanisms.

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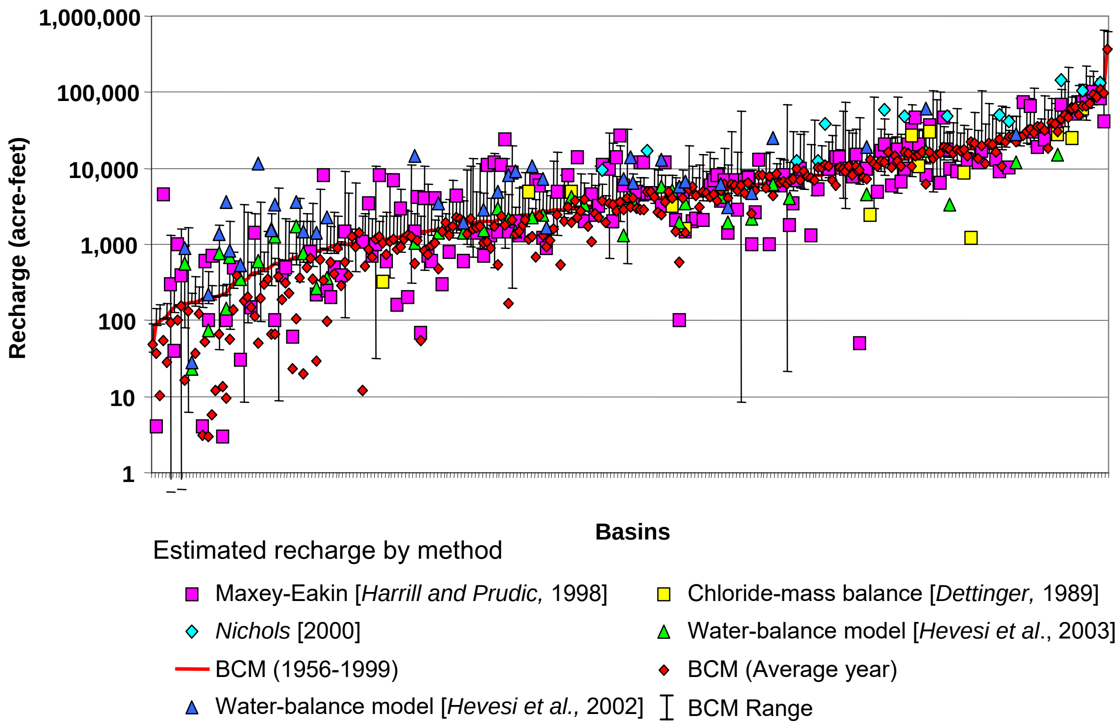

Plate 6. Total potential recharge, calculated as potential in-place recharge plus 10 percent of potential runoff for 258 basins in the Great Basin determined using several methods of estimating recharge and the basin characterization model (BCM). Range bars indicate inclusion or exclusion of all runoff with in-place recharge.

Table 1. Potential recharge (acre-feet/year) calculated for 258 basins in the Great Basin using the basin characterization model (BCM) for in-place recharge and runoff for a mean year and a time series of years, including estimates using the Maxey-Eakin method, chloride-mass balance. Total potential recharge for the Great Basin for BCM estimates shown at the bottom of table.

		Mean potential recharge, in acre-feet per year, by method										
							-------- Basin Characterization Model ---------					
							Mean year			Time series		
Hydro-graphic area or subarea identifier*	Hydrographic area or subarea*	Maxey--Eakin method*	Chloride mass balance method*	Estimates using discharge measure-ments**	Water-balance model (Hevesi et al., 2003)	Water-balance model (Hevesi et al., 2002)	Potential in-place recharge	Potential runoff	Total potential recharge for mean year	Potential in-place recharge	Potential runoff	Total potential recharge for time series
142	Alkali Spring Valley	100			141	3,544	9	0	9	221	82	229
230	Amargosa Desert	1,500			2,139	8,129	146	236	169	1,938	2,567	2,195
151	Antelope Valley (Eureka and Nye)						4,880	1,087	4,988	4,060	1,682	4,228
57	Antelope Valley (Humboldt System)	11,000					2,091	2,289	2,320	1,848	2,988	2,147
93	Antelope Valley (Lemmon Valley)	300					1	947	95	1	1,308	131
186A	Antelope Valley (south)						1,193	486	1,242	977	624	1,039
186B	Antelope Valley (north)						3,574	1,202	3,694	2,897	1,341	3,031
186	Antelope Valley (north and south)	4,700		16,824			4,767	1,688	4,936	3,874	1,965	4,071
106	Antelope Valley (Walker System)	18,000					5,045	75,829	12,627	4,678	82,497	12,928
283	Beaver Valley						15,201	64,886	21,689	15,551	55,149	21,066
280	Beryl-Enterprise Area						25,804	44,431	30,247	21,678	52,721	26,950
137A	Big Smoky Valley (north)	12,000					2,544	2,628	2,807	3,686	3,742	4,060
215	Black Mountains Area	70					51	25	54	1,376	939	1,470
28	Black Rock Desert	14,000					3,963	18,836	5,847	6,055	30,586	9,113
275	Blue Creek Valley	14,000					2,279	59	2,285	3,051	138	3,065
61	Boulder Flat						140	907	231	439	1,569	596
15	Boulder Valley	2,000					5,044	6,228	5,667	4,090	6,382	4,729
75	Bradys Hot Springs Area	160					812	542	866	1,088	1,290	1,216
129	Buena Vista Valley						588	9,755	1,563	670	12,681	1,938
131	Buffalo Valley						284	7,885	1,072	361	8,078	1,169
178A	Butte Valley (north)		2,400				12,653	3,923	13,045	10,465	3,570	10,822
178B	Butte Valley (south)		1,200				21,499	7,413	22,240	17,657	6,261	18,284
178	Butte Valley (north and south)	19,000					34,152	11,336	35,285	28,122	9,831	29,105
272	Cache Valley						339,819	226,765	362,495	372,607	245,166	397,124
148	Cactus Flat	600			1,410	1,969	1,818	1,603	1,978	1,612	2,142	1,826
241	California Valley				775	1,361	13	532	66	41	1,744	216
218	California Wash	60					23	1	23	639	130	652
55	Carico Lake Valley	4,300					1,826	4,080	2,234	1,435	3,582	1,793
101A,B	Carson Desert (Packard and Lahontan Valleys)	1,300					752	1,412	893	1,821	2,218	2,043
105	Carson Valley	41,000					39,856	589,167	98,772	41,627	617,008	103,328
180	Cave Valley	14,000					9,350	9,135	10,264	8,479	9,009	9,380
282	Cedar City Valley						3,275	29,899	6,265	2,696	27,149	5,411
264	Cedar Valley						16,024	12,075	17,231	16,370	12,688	17,639
240	Chicago Valley				569	903	11	57	17	80	873	167
102	Churchill Valley	1,300					6,470	10,420	7,512	6,718	14,298	8,148
143	Clayton Valley	1,500			1,051	14,347	524	306	555	1,300	1,190	1,419
204	Clover Valley (Colorado System)						14,512	17,614	16,274	12,367	20,215	14,389
177	Clover Valley (Independence Valley System)	21,000		58,802			8,065	38,353	11,900	8,223	36,675	11,890
64	Clovers Area						2,250	5,458	2,796	2,493	6,088	3,102
171	Coal Valley	2,000			3,325		3,575	2,643	3,839	2,740	3,701	3,110
100	Cold Springs Valley						7	1,764	184	8	3,355	344
118	Columbus Salt Marsh Valley	700					633	420	675	983	1,207	1,104
2	Continental Lake Valley	11,000					643	4,364	1,079	1,233	7,889	2,022
126	Cowkick Valley						290	91	300	442	352	477
210	Coyote Spring Valley	2,600					5,037	1,467	5,184	5,659	2,924	5,951
229	Crater Flat	220			268	1,424	29	9	30	782	382	820
54	Crescent Valley						1,043	10,935	2,136	910	9,933	1,903
278	Curlew Valley	75,600					26,646	2,177	26,863	26,276	2,728	26,548
103A	Dayton Valley (Carson Plains)						5,522	14,372	6,959	7,090	19,847	9,074
103B	Dayton Valley (Stagecoach Valley)		320				932	990	1,031	1,018	1,357	1,154
103	Dayton Valley (Stagecoach Valley and Carson Plains)	7,900					6,454	15,362	7,991	8,108	21,204	10,228
243	Death Valley	8,000			16,891	60,997	4,960	11,712	6,131	11,755	28,056	14,560
253	Deep Creek Valley	17,000					9,743	25,765	12,319	9,004	23,970	11,401
182	Delamar Valley	1,000					6,627	11,366	7,764	5,308	10,958	6,404
31	Desert Valley	5,000					1,218	12,203	2,438	1,292	15,250	2,817

Table 1. (cont.)

		Mean potential recharge, in acre-feet per year, by method										
							-------- Basin Characterization Model ---------					
							Mean year			Time series		

Hydrographic area or subarea identifier*	Hydrographic area or subarea*	Maxey--Eakin method*	Chloride mass balance method*	Estimates using discharge measurements**	Water-balance model (Hevesi et al., 2003)	Water-balance model (Hevesi et al., 2002)	Potential in-place recharge	Potential runoff	Total potential recharge for mean year	Potential in-place recharge	Potential runoff	Total potential recharge for time series
153	Diamond Valley	21,000	10,500				13,081	20,431	15,124	12,199	19,417	14,141
128	Dixie Valley	6,000					1,909	4,347	2,343	2,199	5,154	2,714
82	Dodge Flat	1,400					1,527	1,460	1,673	1,627	3,337	1,961
181	Dry Lake Valley	5,000					10,307	3,207	10,627	10,666	6,316	11,298
19	Dry Valley (Black Rock Desert System)	200					552	314	584	839	857	925
198	Dry Valley (Colorado System)						2,065	1,278	2,192	1,555	2,603	1,815
16	Duck Lake Valley	9,000	8,900				16,185	11,988	17,384	16,060	20,458	18,106
259	Dugway-Government Creek Valley	7,000					4,489	17,112	6,200	3,714	14,735	5,187
104	Eagle Valley (Carson System)	8,700					219	18,933	2,112	266	19,625	2,228
200	Eagle Valley (Colorado System)						810	796	890	848	1,508	999
268	East Shore Area						3,530	98,590	13,389	4,993	101,225	15,116
109	East Walker Area	31,000					21,032	84,308	29,463	19,215	92,571	28,472
127	Eastgate Valley Area						1,032	1,319	1,164	1,194	1,707	1,364
133	Edwards Creek Valley	8,000					2,722	3,453	3,067	2,503	4,239	2,927
167	Eldorado Valley	1,100					1	112	12	933	1,384	1,072
49	Elko Segment						244	3,823	626	340	4,909	831
158A	Emigrant Valley (Groom Lake Valley)	3,200			5,739	12,910	2,279	1,409	2,420	3,655	4,574	4,112
158B	Emigrant Valley (Papoose Lake Valley)	4					2	7	3	151	359	187
124	Fairview Valley	500					124	163	140	265	521	317
76	Fernley Area	600					888	647	953	1,307	2,001	1,507
77	Fireball Valley	200					1,239	968	1,336	1,213	1,563	1,369
117	Fish Lake Valley	33,000	26,800				5,855	48,812	10,737	7,743	60,393	13,783
258	Fish Springs Flat	4,000					1,016	384	1,054	1,460	664	1,526
227A	Fortymile Canyon (Jackass Flat)	900			1,583	1,665	857	535	910	2,524	2,535	2,778
227B	Fortymile Canyon (Buckboard Mesa)	1,400			1,959	3,113	3,727	3,287	4,056	4,684	6,436	5,327
160	Frenchman Flat	100			1,903	5,683	537	396	576	4,299	2,207	4,520
122	Gabbs Valley	5,000	4,900				1,023	1,238	1,147	2,195	2,367	2,431
172	Garden Valley	10,000			3,323		16,542	14,325	17,974	13,866	16,939	15,559
120	Garfield Flat	300					1,371	1,257	1,497	1,382	2,265	1,609
216	Garnet Valley	400					288	60	294	989	109	1,000
147	Gold Flat	3,800			4,205	6,287	4,637	3,701	5,007	4,595	5,847	5,180
187	Goshute Valley	10,400		40,911			25,210	9,048	26,115	22,410	9,498	23,360
23	Granite Basin	400					1	1,535	154	1	1,599	160
78	Granite Springs Valley	3,500					5,044	22,631	7,307	5,046	25,213	7,567
138	Grass Valley	13,000					6,891	11,266	8,018	5,030	10,926	6,123
71	Grass Valley (Humboldt System)	12,000					410	13,387	1,749	502	15,453	2,048
279	Great Salt Lake						3	1,320	135	6	1,647	171
261B	Great Salt Lake Desert (east)	4,500					54	0	54	106	0	106
261A	Great Salt Lake Desert (west)	47,000					14,026	4,685	14,494	13,365	5,116	13,876
3	Gridley Lake Valley	4,500					933	1,666	1,099	2,588	5,981	3,186
251	Grouse Creek Valley	14,000					2,369	3,490	2,718	3,265	4,606	3,726
276	Hansel and North Rozel Flat	8,000					331	4	332	864	28	867
68	Hardscrabble Area	9,000					12,833	46,734	17,506	12,248	48,868	17,134
217	Hidden Valley (north)	400					188	6	188	566	57	571
166	Hidden Valley (south)				23	28	0	0	-	169	63	175
25	High Rock Lake Valley	13,000					13,762	8,367	14,599	16,559	16,145	18,173
156	Hot Creek Valley	7,000		5,756			4,512	1,805	4,692	5,380	4,034	5,783
24	Hualapai Flat	7,000					3,700	7,727	4,473	4,088	9,248	5,013
47	Huntington Valley						34,668	59,713	40,639	29,248	52,667	34,514
113	Huntoon Valley	800					1,226	1,012	1,327	1,440	2,439	1,683
72	Imlay Area	4,000					226	6,056	831	462	10,260	1,488
188	Independence Valley	9,300		50,065			22,907	8,347	23,742	20,525	8,863	21,411
161	Indian Springs Valley	10,000			4,591	18,978	6,912	3,904	7,302	9,966	7,901	10,756
135	Ione Valley	8,000					1,176	689	1,245	1,026	984	1,125
164A	Ivanpah Valley (north)				1,399	3,482	438	418	480	1,487	896	1,576
164B	Ivanpah Valley (south)				1,569	1,519	53	126	66	293	2,261	519
164	Ivanpah Valley (North and South)	1,500			2,968	5,001	491	545	546	1,779	3,158	2,095
174	Jakes Valley			38,203			10,761	2,131	10,974	8,082	2,280	8,310
165	Jean Lake Valley	100			73	217	0	28	3	167	276	195

Table 1. (cont.)

		Mean potential recharge, in acre-feet per year, by method										
							-------- Basin Characterization Model ---------					
							Mean year			Time series		

Hydrographic area or subarea identifier*	Hydrographic area or subarea*	Maxey--Eakin method*	Chloride mass balance method*	Estimates using discharge measure-ments**	Water-balance model (Hevesi et al., 2003)	Water-balance model (Hevesi et al., 2002)	Potential in-place recharge	Potential runoff	Total potential recharge for mean year	Potential in-place recharge	Potential runoff	Total potential recharge for time series
132	Jersey Valley	800					557	955	652	677	1,366	813
206	Kane Springs Valley						4,579	8,416	5,421	5,262	10,659	6,328
157	Kawich Valley	3,500			3,688	6,563	3,788	3,008	4,089	3,454	5,143	3,968
66	Kelly Creek Area						3,730	5,497	4,279	3,408	6,654	4,073
30A	Kings River Valley (Rio King Subarea)						8,386	21,333	10,520	7,698	24,428	10,141
30B	Kings River Valley (Sodhouse Subarea)						26	23	28	109	62	116
30	Kings River Valley (Rio King and Sodhouse subareas)	15,000					8,412	21,357	10,547	7,808	24,490	10,257
139	Kobeh Valley						7,793	5,852	8,378	5,942	5,413	6,483
79	Kumiva Valley	1,000					36	11,208	1,157	31	10,742	1,105
183	Lake Valley	13,000					13,213	15,049	14,718	10,858	14,946	12,353
45	Lamoille Valley						20	62,875	6,308	21	69,928	7,014
212	Las Vegas Valley		28,000		15,147		28,072	21,349	30,207	33,697	28,483	36,545
285	Leamington Canyon						3,786	31,981	6,984	4,388	38,152	8,203
92A	Lemmon Valley (west						8	3,787	386	9	5,521	561
92B	Lemmon Valley (east						7	1,906	197	99	3,519	451
92	Lemmon Valley (east and west	1,500					14	5,693	584	108	9,040	1,012
144	Lida Valley				610	11,335	50	6	50	406	118	418
150	Little Fish Lake Valley	11,000		9,628			3,501	2,996	3,801	3,010	3,131	3,324
67	Little Humboldt Valley	24,000					26,022	58,057	31,828	25,338	64,651	31,803
155A	Little Smoky Valley (north)						7,881	1,466	8,028	6,122	1,561	6,278
155B	Little Smoky Valley (central)						391	93	400	317	167	334
155C	Little Smoky Valley (south)						1,889	567	1,946	1,542	963	1,638
155	Little Smoky Valley (north, central and south)	5,400		12,681			10,161	2,126	10,374	7,981	2,692	8,250
9	Long Valley	6,000					5,908	5,164	6,424	5,913	7,486	6,662
175	Long Valley (Colorado System)	10,000		47,740			15,875	4,139	16,289	13,186	3,495	13,536
73A	Lovelock Valley (Oreana Subarea)						39	1,672	206	95	2,542	349
73B	Lovelock Valley (Upper and Lower Valley subareas)						1,732	2,826	2,015	2,290	5,810	2,871
73	Lovelock Valley (Orena, and Upper and Lower Valley subareas)	3,200					1,771	4,498	2,220	2,385	8,352	3,220
242	Lower Amargosa Valley				767	1,475	17	26	20	590	1,420	732
205	Lower Meadow Valley Wash						10,883	8,004	11,683	18,126	19,659	20,092
220	Lower Moapa Valley	40					0	0	-	128	193	147
59	Lower Reese River Valley						354	5,804	935	445	5,995	1,044
51	Maggie Creek Area						695	8,759	1,571	1,748	10,529	2,801
273	Malad-Lower Bear River Area						81,639	43,703	86,010	84,159	44,066	88,566
52	Marys Creek Area						35	17	37	154	228	176
42	Marys River Area						19,014	36,806	22,694	18,977	43,651	23,342
108	Mason Valley	2,000					1,438	19,162	3,354	1,635	19,694	3,604
8	Massacre Lake Valley						1,086	247	1,110	2,613	1,829	2,796
225	Mercury Valley	250			359	2,256	75	243	99	751	1,165	867
163	Mesquite Valley	1,500	1,600		3,470	6,696	1,370	582	1,428	4,328	2,492	4,577
58	Middle Reese River Valley	7,000					1,065	1,119	1,177	1,045	1,274	1,173
284	Milford Area						1,509	6,091	2,118	1,734	6,919	2,426
140A	Monitor Valley (north)						8,536	15,375	10,074	6,981	12,882	8,269
140B	Monitor Valley (south)						13,827	22,150	16,042	10,260	17,665	12,026
136	Monte Cristo Valley	500					190	1,179	308	399	1,756	575
12	Mosquito Valley	700					6	1	6	185	106	196
26	Mud Meadows	8,000					3,439	3,346	3,774	4,590	4,711	5,061
219	Muddy River Springs Area						12	0	12	207	1	207
154	Newark Valley	17,500		49,092			16,721	17,077	18,428	13,852	15,380	15,390
44	North Fork Area						7,189	34,246	10,614	17,330	49,380	22,268
137B	Northern Big Smoky Valley	65,000					25,680	70,153	32,695	20,720	62,976	27,018
266	Northern Juab Valley						12,996	24,774	15,474	12,878	27,698	15,648
228	Oasis Valley	1,000			2,209	4,698	2,445	744	2,519	5,512	3,919	5,903
209	Pahranagat Valley	1,800			4,046		6,620	4,234	7,043	6,665	5,211	7,186
208	Pahroc Valley	2,200					4,275	1,564	4,432	4,531	3,015	4,832
162	Pahrump Valley				11,759	28,437	20,976	17,319	22,708	23,716	25,591	26,275
203	Panaca Valley						4,535	2,059	4,741	4,506	4,779	4,984
69	Paradise Valley	10,000					2,902	63,905	9,293	2,971	70,503	10,022
260B	Park Valley (east)						256	10,171	1,273	317	10,772	1,394

Table 1. (cont.)

Mean potential recharge, in acre-feet per year, by method

Hydrographic area or subarea identifier*	Hydrographic area or subarea*	Maxey--Eakin method*	Chloride mass balance method*	Estimates using discharge measure-ments**	Water-balance model (Hevesi et al., 2003)	Water-balance model (Hevesi et al., 2002)	Basin Characterization Model: Mean year: Potential in-place recharge	Mean year: Potential runoff	Mean year: Total potential recharge for mean year	Time series: Potential in-place recharge	Time series: Potential runoff	Time series: Total potential recharge for time series
260A	Park Valley (west)						319	1,736	493	585	1,923	777
260	Park Valley (east and west)	24,000					575	11,907	1,765	902	12,696	2,171
281	Parowan Valley						6,718	24,701	9,188	5,368	24,572	7,825
202	Patterson Valley	8,000					6,201	4,427	6,643	6,046	7,132	6,759
286	Pavant Valley						20,068	56,338	25,701	19,957	64,934	26,450
170	Penoyer Valley	4,300	3,200		5,160		3,797	2,551	4,052	3,828	4,460	4,275
191	Pilot Creek Valley	2,400					2,239	2,778	2,517	2,871	3,187	3,189
252	Pilot Valley	3,400					613	2,543	867	837	2,551	1,092
29	Pine Forest Valley	10,000					5,493	15,310	7,024	5,452	23,193	7,771
255	Pine Valley (Great Salt Lake Desert System)	21,000					14,027	18,308	15,858	11,982	16,365	13,619
53	Pine Valley (Humboldt System)	46,000					16,331	27,297	19,060	13,026	23,031	15,330
130	Pleasant Valley (Dixie Valley System)	3,000					601	3,188	920	801	4,544	1,256
88	Pleasant Valley (Truckee System)	10,000					746	27,585	3,505	663	28,877	3,550
274	Pocatello Valley						7,766	102	7,777	8,008	121	8,020
277	Promontory Mountains Area						1,888	490	1,937	3,373	954	3,468
65	Pumpernickel Valley						101	2,591	360	321	3,754	697
81	Pyramid Lake Valley	6,600					9,830	9,656	10,796	11,443	16,877	13,130
33A	Quinn River Valley (Orovada Subarea)						8,128	68,406	14,969	7,865	73,280	15,193
33B,C	Quinn River Valley (McDermitt and Oregon Canyon)						40,294	103,185	50,612	35,080	103,920	45,472
33	Quinn River Valley (Orovada, McDermitt, and Oregon Canyon subareas)	73,000					48,422	171,590	65,581	42,945	177,200	60,665
173A	Railroad Valley (south)		4,900		4,135		1,853	892	1,942	2,682	2,539	2,936
173B	Railroad Valley (north)		24,800	61,083			57,421	39,280	61,349	46,876	38,659	50,742
173	Railroad Valley (north and south)	52,000					59,274	40,172	63,291	49,558	41,199	53,678
141	Ralston Valley	5,000					3,708	3,683	4,076	4,028	5,410	4,568
123	Rawhide Flats	150					144	42	149	394	179	412
119	Rhodes Salt Marsh Valley	500					318	882	406	756	1,880	944
62	Rock Creek Valley						442	849	527	921	1,581	1,079
226	Rock Valley	30			352	532	0	0	-	324	110	335
199	Rose Valley						48	4	48	38	52	43
176	Ruby Valley	68,000		145,636			35,382	88,306	44,212	29,133	82,288	37,362
263	Rush Valley	34,000					33,806	42,371	38,043	31,493	40,184	35,511
267	Salt Lake Valley						28,193	182,454	46,439	29,827	184,549	48,282
22	San Emidio Desert	2,100					3,862	9,961	4,858	3,747	11,559	4,903
20	Sano Valley	4					37	2	37	87	54	93
146	Sarcobatus Flat	1,200			2,466	7,315	1,230	707	1,301	2,532	2,398	2,772
287	Sevier Desert						17,238	30,771	20,316	17,924	33,064	21,230
245	Shadow Valley				1,731	3,634	89	145	104	528	1,506	679
32	Silver State Valley	1,400					52	634	115	183	2,344	418
271	Sink Valley	1,000					99	0	99	154	5	154
270	Skull Valley						16,969	44,502	21,419	14,624	39,740	18,598
134	Smith Creek Valley	12,000					3,279	2,935	3,572	3,738	4,550	4,193
107	Smith Valley	17,000					11,313	87,974	20,111	10,359	94,692	19,828
21	Smoke Creek Desert	13,000					14,993	14,351	16,428	18,729	25,829	21,311
254	Snake Valley	100,000					80,079	126,490	92,728	69,738	122,176	81,955
121A,C	Soda Spring Valley (east and central)						242	1,483	390	598	4,188	1,017
121B	Soda Spring Valley (west)						257	871	344	367	1,108	478
121	Soda Spring Valley (east, central, and west)	700					499	2,354	735	965	5,297	1,494
46	South Fork Area						8	59,056	5,914	8	55,920	5,600
85	Spanish Springs Valley	600					695	474	743	991	1,685	1,159
201	Spring Valley (Colorado System)	10,000					9,549	13,249	10,874	7,486	14,436	8,930
184	Spring Valley (Great Salt Lake Desert System)	75,000	61,600	103,569			57,629	93,577	66,987	48,116	80,635	56,179
43	Starr Valley Area						2,905	84,762	11,381	2,986	82,405	11,226
179	Steptoe Valley	85,000		131,469			104,285	71,344	111,419	88,282	61,094	94,391
152	Stevens Basin						1,390	10	1,391	1,055	113	1,067
125	Stingaree Valley						9	13	10	90	73	97
149	Stone Cabin Valley	5,000					2,843	1,628	3,006	3,673	3,139	3,987
145	Stonewall Flat	100			1,241	3,393	65	6	65	540	110	551
27	Summit Lake Valley	4,200					1,000	1,072	1,107	1,248	2,204	1,469
86	Sun Valley	50					5,657	36,757	9,333	6,260	40,549	10,315
50	Susie Creek Area						178	1,684	346	525	2,907	816

Table 1. (cont.)

Hydrographic area or subarea identifier*	Hydrographic area or subarea*	Mean potential recharge, in acre-feet per year, by method										
							-------- Basin Characterization Model ---------					
							Mean year			Time series		
		Maxey--Eakin method*	Chloride mass balance method*	Estimates using discharge measure-ments**	Water-balance model (Hevesi et al., 2003)	Water-balance model (Hevesi et al., 2002)	Potential in-place recharge	Potential runoff	Total potential recharge for mean year	Potential in-place recharge	Potential runoff	Total potential recharge for time series
7	Swan Lake Valley						514	248	539	2,697	1,688	2,866
114	Teels Marsh Valley	1,300					1,284	1,887	1,473	2,035	3,527	2,387
48	Tenmile Creek Area						3,608	17,122	5,320	2,954	16,702	4,624
189A	Thousand Springs Valley (Herrell Siding-Brush Creek subarea)						1,192	5,092	1,701	1,197	5,707	1,768
189B	Thousand Springs Valley (Toano-Rock Spring subarea)						2,206	4,322	2,638	3,505	5,960	4,101
189C	Thousand Springs Valley (Rocky Butte subarea)						1,728	0	1,728	3,160	74	3,167
189D	Thousand Springs Valley (Montello-Crittenden Creek subarea)						7,573	358	7,609	10,436	1,462	10,582
189	Thousand Springs Valley (Herrell Siding-Brush Creeak, Toano-Rock Spring, Rocky Butte and Montello-Crittenden Creek subareas)	12,000					12,699	9,772	13,676	18,299	13,202	19,619
168	Three Lakes Valley (north)	2,000			1,490	9,031	1,317	472	1,364	2,182	903	2,272
211	Three Lakes Valley (south)	6,000			1,298	7,335	2,725	1,773	2,903	3,631	1,981	3,830
169A	Tikapoo Valley (north)				3,971	13,767	3,028	947	3,123	3,756	2,050	3,961
169B	Tikapoo Valley (south)				2,295	10,819	1,230	263	1,256	2,419	581	2,477
169	Tikapoo Valley (north and south)	6,000			6,266	24,586			-			-
185	Tippett Valley	6,900		12,389			9,364	3,534	9,717	7,367	2,918	7,659
137A	Tonopah Flat	12,000					2,544	2,628	2,807	3,686	3,742	4,060
262	Tooele Valley						23,941	24,445	26,386	23,885	23,766	26,262
83	Tracy Segment	6,000					9,768	6,750	10,443	10,613	14,424	12,056
87	Truckee Meadows	27,000					1,983	15,837	3,566	2,013	17,699	3,783
221	Tule Desert	2,100					1,319	1,512	1,470	4,126	3,456	4,472
257	Tule Valley	7,600					6,206	2,992	6,505	5,559	2,736	5,833
56	Upper Reese River Valley	37,000	30,000				13,529	30,683	16,598	12,137	29,699	15,107
265A	Utah Valley Area (Goshen Valley						1,561	2,526	1,814	2,056	3,630	2,419
265C	Utah Valley Area (north)						42,897	76,850	50,582	45,816	78,973	53,714
265B	Utah Valley Area (south)						62,634	85,648	71,199	63,401	94,892	72,890
244	Valjean Valley				671	820	2	533	56	77	1,921	269
222	Virgin River Valley						16,014	23,837	18,398	29,392	30,078	32,400
4	Virgin Valley	7,000					615	615	676	2,377	1,561	2,533
256	Wah Wah Valley	7,000					5,869	1,886	6,057	5,186	2,319	5,418
110A	Walker Lake Valley (Schurz Subarea)						351	13,684	1,720	897	10,780	1,975
110B	Walker Lake Valley (Lake Subarea)						487	35,034	3,991	560	32,806	3,841
110C	Walker Lake Valley (Whiskey Flat-						4,599	54,355	10,035	4,096	53,332	9,429
110	Walker Lake Valley	6,500					5,438	103,074	15,745	5,553	96,918	15,245
84	Warm Springs Area	6,000					3,446	7,044	4,150	3,738	12,722	5,010
269	West Shore Area	600					53	1	53	188	6	189
60	Whirlwind Valley						119	55	125	169	104	179
74	White Plains	3					13	0	13	212	80	220
207	White River Valley						33,443	14,818	34,925	29,192	15,673	30,759
63	Willow Creek Valley						2,629	5,052	3,134	4,189	6,954	4,885
80	Winnemucca Lake Valley	2,900					4,099	9,894	5,088	4,292	11,791	5,471
70	Winnemucca Segment						622	7,321	1,354	990	8,478	1,838
159	Yucca Flat	700			1,557	2,815	874	1,732	1,047	1,677	3,002	1,977
		Total potential Great Basin recharge					2,406,022	4,828,227	2,888,844	2,428,874	5,239,825	2,952,856

* Harrill and Prudic (1998)
** Nichols (2000)

SE ROA 51184

Prediction of the Effects of Changing the Spatial Distribution of Pumping in the Lower White River Flow System

Project #: 117-0524303
July 3, 2019

PRESENTED TO

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July 3, 2019

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1.0 INTRODUCTION

A regional carbonate-rock aquifer has the potential for being a productive source of additional water in southern Nevada. However, this same regional aquifer is also the source of several large-volume warm springs that discharge on Federal and private lands, and in some cases provide baseflow to streams. The effects of pumping the carbonate-rock aquifer could include the eventual capture of the water that discharges from these springs and thus depletion of their flow. Reduction or cessation of spring discharge on Federal and private lands not only would have an adverse effect on sensitive habitat and species, but also senior water rights associated with stream baseflow provided by some of these springs.

Springs and water-related resource attributes are important features in the Overton Arm area of Lake Mead National Recreation Area (Lake Mead NRA). The springs provide water for vegetation and wildlife habitat and create an environment that many visitors use and enjoy. Many of these springs are fed by regionally- and locally-derived groundwater (Pohlmann et al., 1997), and could be affected by up-gradient groundwater diversions. Springs include Rogers Spring, Blue Point Spring, Corral Spring, and other smaller, unnamed springs. Visitation to Rogers and Blue Point Springs in recent years has been conservatively estimated at 30,000 visitors per year.

The National Park Service (NPS) is entitled to Federal reserved water rights for reserved lands within Lake Mead NRA. The priority dates for these reserved rights are the dates when the lands were reserved. These rights have not been judicially quantified. The NPS also has a State appropriative water right to water from Rogers Spring. The priority date for this water right is February 16, 1937.

Desert bighorn sheep are also dependent upon the springs in the northern part of Lake Mead NRA. The relict Las Vegas Valley leopard frog, *Rana onca*, has been found at Rogers and Blue Point Springs. The relict leopard frog, previously believed extinct, had been petitioned for listing in 2002 as protected under the Endangered Species Act, but the United States Fish and Wildlife Service (USFWS) subsequently determined in 2016 that such a listing is not warranted at this time. There are three endemic springsnail species found in the short springbrook above the flow measurement weir at Blue Point Spring. In 2009, several entities petitioned for the listing of 42 springsnail species in the Great Basin, including the Blue Point pyrg. In 2017, the USFWS decided that listing the Blue Point pyrg is not warranted at this time.

1.1 SETTING

The basins located north of Lake Mead NRA are underlain by a regional groundwater flow system, originally referred to as the White River groundwater flow system (Eakin, 1966), and later as the Colorado River groundwater flow system (Prudic et al., 1995, and Harrill and Prudic, 1998) (Figure 1-1). Groundwater generally flows from north to south in this flow system, which discharges most of its flow at springs near the headwater area for the Muddy River, thus supplying the base flow to the Muddy River (Eakin, 1966; Harrill and Prudic, 1998; and Prudic et al., 1995). Several of the previously mentioned Lake Mead NRA springs are also discharge points for this same regional groundwater flow systems (Harrill et al., 1988 and Prudic et al., 1995).

The White River groundwater flow system contains basin-fill and carbonate-rock aquifers that appear to be hydraulically connected (Prudic et al., 1995). As a result, large-scale development of groundwater in the basin-fill aquifers could induce groundwater to flow from the regional carbonate-rock aquifer to the basin-fill aquifer, lowering the hydraulic head in the regional carbonate-rock aquifer, and eventually causing depletion of the discharge of large regionally sourced springs, such as the Muddy River Springs, and smaller regionally sourced springs such as the previously mentioned Lake Mead NRA springs. Similar concerns would also apply to large-scale development of groundwater directly from the carbonate-rock aquifer, if withdrawals are large enough and occur over a sufficiently long period of time.

Geologic mapping by Page et al. (2005) and later geologic cross-sections developed by Page et al. (2011) indicate that much of the Paleozoic carbonate-rock section is present beneath the Black Mountains Area basin (see cross-sections D-D’, E-E’, F-F’, G-G’ and I-I’), extending to the Lake Mead and Rogers Spring fault zones which transect the northern portion of Lake Mead NRA. These Paleozoic rock formations also are present in the upper and lower plates of the Muddy Mountain thrust fault that also transects portions of the Black Mountains Area basin as shown in several of the aforementioned cross-sections (Figure 1-2). Page et al. (2011) state that “the Muddy Mountain thrust in the Muddy Mountains juxtaposes Paleozoic carbonate rocks in the upper plate against Mesozoic and Paleozoic rocks in the lower plate (G–G'); such a relationship suggests that the less permeable Mesozoic rocks below the thrust may act as a groundwater flow barrier, and the thrust has been characterized as a barrier in local groundwater models. Although the lower plate rocks may act as a barrier in localized zones along strike, we think that overprinting of the thrust by Cenozoic faults (Langenheim and others, 2002) provides linkage between rocks in the upper and lower plates, allowing for some groundwater flow across the thrust. This example may apply to other Mesozoic thrust faults in the map area, especially where the thrusts are highly modified by younger Cenozoic extensional faults.”

Geochemical modeling (Appendix A) has indicated that the areas around the Muddy River Springs and in Garnet Valley both could be along groundwater flow paths to Rogers Spring and Blue Point Spring (Geochemical Technologies Corporation and Tetra Tech, Inc., 2012). The study determined that while a central flow path through California Wash (and the Moapa River Indian Reservation) is plausible, the more plausible flow path is a southern flow path from Garnet Valley, beneath California Wash and the Black Mountain Area basins to discharge at Rogers Spring and Blue Point Spring. In both cases, the conceptual direction of groundwater flow crosses the Muddy Mountain Thrust Fault and requires that groundwater from the Paleozoic carbonate aquifer pass through Mesozoic clastic sediments and the Tertiary basin-fill evaporite lithologies (Muddy Creek and/or Horse Spring) in the lower thrust plate. Alternatively, groundwater could flow in the lower plate carbonate rocks, and upwards along the Roger Springs Fault, mixing with water discharging from the Mesozoic rocks as it ascends to the surface along the Rogers Spring fault zone (see cross-sections F-F’ and G-G’, Page et al., 2011). The study indicated that capture zone modeling of Rogers Spring and Blue Point Spring using a recently developed groundwater flow model of this region indicated that the more likely southern flow path is consistent with the geochemical data.

Both of these flow paths are further supported by recent potentiometric surface mapping of the upper carbonate-rock aquifer in southern Nevada (Wilson, 2019), which indicates a southeast flow direction through these basins toward the NPS’ springs. This mapping was part of a larger study involving the drilling, installation and sampling of six monitoring wells completed in the carbonate-rock aquifer system underlying portions of Clark County. Two of these new wells [Buffington Pockets Well (BUFPKTS-01), which may be completed in the Mesozoic rock section below the Muddy Mountain thrust fault, and Rogers Bay Well (RB-01), which is completed near or within the Rogers Spring fault zone], provide important supporting hydrogeologic information on the carbonate rock flow system in California Wash near the Muddy Mountain thrust fault and in the Black Mountains Area near Rogers Springs, respectively. In particular, the water level for BUFPKTS-01, is about 105 feet and 187 feet higher than the water levels in RB-01 and Blue Point Spring, providing sufficient head potential to accommodate groundwater flow through permeable structures or formations within the Mesozoic rock section toward the NPS’ springs. Oxygen/deuterium sampling results for BUFPKTS-01 are very similar to the results reported by Geochemical Technologies Corporation and Tetra Tech, Inc. (2012) for the Valley of Fire Well, which is completed in the Mesozoic section nearer to Blue Point Springs. The similarity in stable isotope results for both of these wells supports Geochemical Technologies’ earlier suggestion of representing the composition of recharge from more local sources, probably the Muddy Mountains, but may themselves be a mixture of incoming water from the carbonate aquifer plus local recharge from the Muddy Mountains. The close similarity in the analytical results between RB-01, Rogers Spring and Blue Point Spring support the contention that much of their source water is probably originating from depth and ascends to the surface along the Rogers Spring fault zone at various spring discharge sites in the northern portion of Lake Mead NRA. The significant increase in dissolved solids and major ion content at Rogers Spring and Blue Point Spring likely is attributable to evaporite dissolution in the Mesozoic section, in Tertiary volcanic rocks, or in the Tertiary basin-fill sediments (Hershey and Mizell, 1995; Laney and Bales, 1996, from Geochemical Technologies Corporation and Tetra Tech, Inc., 2012).

Based on the information presented above, the NPS is concerned that increases in groundwater withdrawals along this Garnet Valley-California Wash-Black Mountain Area flow path, and possibly the central flow path through California Wash, may cause a decline in discharge from Rogers Spring and Blue Point Springs.

1.2 MANAGEMENT OF GROUNDWATER DEVELOPMENT IN THE LWRFS

The Nevada State Engineer, who regulates water rights within the State of Nevada, traditionally has managed groundwater development in Nevada within individual basins, using the concept of perennial basin yield to determine the amount of groundwater available for appropriation. This approach inherently assumes that groundwater flow is contained within an individual basin, where precipitation recharges the flow system and groundwater discharges from the flow system, primarily by evapotranspiration. As a result, the perennial yield has been defined as the amount of natural discharge that can be reasonably salvaged, without causing long-term depletion of the aquifer, and in no case should it exceed the estimated annual recharge of the basin. Unfortunately, this approach presents difficulties when the groundwater flow system within a basin is actually part of a more extensive regional groundwater flow system that hydraulically interconnects several individual basins.

In 2001, the NPS, the U.S. Bureau of Land Management (BLM), and the USFWS, collectively the Department of Interior (DOI) bureaus, participated in an administrative hearing held by the Nevada State Engineer concerning proposed groundwater development in Coyote Spring Valley, about 40 miles north of Las Vegas Valley. This hearing was one of the earliest examples where the individual basin management approach was challenged by the DOI bureaus' contention that the Coyote Spring Valley basin was part of a hydrologically-connected regional flow system in which the component basins should be managed collectively instead of individually. As part of their preparations for hearing, the DOI bureaus cooperated in the development of a preliminary numerical groundwater flow model, prepared by Tetra Tech, Inc., to simulate regional groundwater flow in the area and to evaluate and demonstrate the potential effects of large-scale groundwater pumping on water levels in the regional aquifer and on nearby spring flows.

Following this hearing, the Nevada State Engineer issued Order 1169 in 2002 holding all pending groundwater applications in Coyote Spring Valley and several surrounding hydrographic areas in abeyance, until further evaluation of the effects of pumping groundwater associated with existing permits was completed. During this abeyance period, the DOI bureaus participated in several scientific investigations with the goal of producing reliable information that would enable Tetra Tech, Inc. to develop refinements to the preliminary numerical model and improve the model's accuracy in predicting the effects of regional groundwater development on nearby Federal water resources. Some of the more notable investigations included gain-loss studies on the Muddy River (Beck and Wilson, 2006) and the Virgin River (Beck and Wilson, 2005), development of a regional geologic map (Page et al., 2005) and associated geologic cross-sections (Page et al., 2011) for the model area, geophysical studies of selected basins in the model area (Scheirer, Page and Miller, 2006), and estimation of evapotranspiration within the model area (DeMeo et al., 2008). Tetra Tech, Inc. completed refinements to the numerical model in 2012 and conducted several simulations to assess the pumping effects from existing water rights and most of the largest water rights applications held in abeyance. This work was documented in a model development report (Tetra Tech, 2012a) and a predictive modeling simulations report (Tetra Tech, 2012b).

During the abeyance period, the Nevada State Engineer required certain water-right holders to conduct a two-year pumping test, in which they were required to pump at least half of their existing water rights in Coyote Spring Valley annually. The purpose of the pumping test was to assess the potential effects on nearby water resources before ruling on the applicants' pending water-right applications in this valley and other surrounding valleys. This pumping test was started in late 2010 and completed in late 2012, at which time the Nevada State Engineer issued a modification to Order 1169 (Order 1169A) that invited any study participant to submit interpretive reports on the results of the pumping test in June 2013. The DOI bureaus submitted an interpretive report that also included additional simulation results generated using the updated predictive numerical model completed by Tetra Tech, Inc. All interpretive reports submitted were considered by the Nevada State Engineer in making their 2014 decisions to deny all pending applications in Coyote Spring Valley and several surrounding valleys. The NPS continues to support the analyses and conclusions presented in the 2013 data interpretation report submitted by the DOI bureaus.

In the years since the Order 1169 pumping test was completed, water level, spring discharge and pumping data continued to be collected in these same basins to monitor the effects of water-level recovery and ongoing groundwater pumping. In 2018, the Nevada State Engineer expressed concern to water-right holders in these basins that water levels and spring discharges affected by the earlier Order 1169 pumping test had not recovered to their pre-pumping conditions, and that higher levels of pumping were likely to adversely affect senior water right holders and an endangered fish (Moapa dace) in the Muddy River Springs Area. This subsequently led the Nevada State Engineer to issue Interim Order 1303 in January 2019, which designated several of the affected basins as a jointly managed administrative unit to be known as the Lower White River Flow System (LWRFS). Interim Order 1303 also seeks to maintain the status quo on groundwater withdrawals in this area, while allowing for the submittal of additional data to inform the Nevada State Engineer on groundwater sustainability, and for progress to continue on the development of a voluntary conjunctive management plan for the LWRFS.

In issuing Interim Order 1303, the Nevada State Engineer ordered that any stakeholder with interests that may be affected by water right development within the LWRFS may file a report by the established deadline, which should address the following matters:

a. The geographic boundary of the hydrologically connected groundwater and surface water systems comprising the LWRFS;

b. The information obtained from the Order 1169 aquifer test and subsequent to the aquifer test, and Muddy River headwater spring flow as it relates to aquifer recovery since the completion of the aquifer test;

c. The long-term annual quantity of groundwater that may be pumped from the LWRFS, including the relationships between the location of pumping on discharge to the Muddy River Springs, and the capture of Muddy River Flow;

d. The effects of movement of water rights between alluvial wells and carbonate wells on deliveries of senior decreed rights to the Muddy River; and

e. Any other matter believed to be relevant to the State Engineer's analysis.

On behalf of the National Park Service, Tetra Tech, Inc. utilized the updated numerical groundwater flow model developed for the DOI bureaus to qualitatively evaluate some of the matters of interest to the Nevada State Engineer noted in Interim Order 1303. This includes evaluating the spatial relationships between the location of pumping on discharge to the Muddy River Springs and capture of Muddy River flow, the effects of moving water rights between alluvial aquifer wells and carbonate aquifer wells on the deliveries of senior decreed rights on the Muddy River, and the possible expansion of the current geographic boundary of the LWRFS. These are also matters of interest to the NPS, as such movement of water rights within the LWRFS could increase the threat potential to NPS-managed groundwater resources at Lake Mead NRA.

The updated numerical groundwater flow model was used to perform three simulations to qualitatively evaluate the effects of redistributing pumping within and between the alluvial and carbonate aquifers in the LWRFS, and where possible, make recommendations on geographic boundary adjustments to the hydrologically connected groundwater and surface water systems comprising the LWRFS. The three simulations, described in more detail in Section 3, include:

1. Simulation evaluating pumping by priority date,

2. Simulation evaluating the redistribution of increased pumping to the carbonate aquifer, and
3. Simulation evaluating the redistribution of pumping from the alluvial aquifer to the carbonate aquifer.

Each of these simulations limited pumping in the LWRFS to approximately 14,535 acre-feet/year. The results of this study are presented for consideration by the Nevada State Engineer and stakeholders with direct or indirect interests that may be affected by water right development in the LWRFS.

2.0 SUMMARY OF PREVIOUS GROUNDWATER MODELING

As stated in Section 1, Tetra Tech completed an update of their groundwater flow model of the LWRFS in 2012 (Tetra Tech, 2012a). A predictive scenario modeling report followed later that year describing seven scenarios of future pumping within the LWRFS (Tetra Tech, 2012b). After the Order 1169 aquifer testing was completed, a post audit was conducted on the 2012 Tetra Tech Groundwater Flow Model to see how well the 2012 model simulated this aquifer testing (Tetra Tech, 2013). The major conclusions from these three reports and observations of the interpretation of their results are provided below focusing on the five basins of interest.

2.1 2012 GROUNDWATER MODEL CALIBRATION

The 2012 groundwater flow model was completed for all or parts of 13 hydrographic areas within the lower portion of the Colorado River Regional Flow System in southeastern Nevada and parts of Arizona and Utah. Several of these hydrographic areas overlap with those in the Lower White River Flow System. This model simulates the movement of groundwater in an area ranging from the Clover and Delamar Mountains on the north to the Las Vegas Valley Shear Zone and Lake Mead on the south, and from the Sheep Range on the west to the Virgin and Beaver Dam Mountains on the east.

The 2012 model was calibrated based on many different types of information, including measurements of water levels and drawdown, discharge rates for springs, streamflow measurements, reported pumping rates that varied through time, seasonal estimates of ET based on field measurements and satellite mapping of plant communities, and estimates of model boundary fluxes based on regional information. A “pre-production” model was developed to match water levels and water-budget information. Simulated water levels agree well with observed water levels. The correlation between measured and simulated water levels was 0.96. The largest model residuals are in high gradient areas, where model errors can result in large differences, in the Clover Mountains where the volcanic stratigraphy is greatly simplified, and in the Tule Desert where some of the structural complexity may not be incorporated in the geologic model and the model grid is relatively coarse.

The model was also calibrated to the effects of time-varying pumping and seasonal ET during the period October 2008 through December 2011, primarily in the area of the Muddy River Springs, Coyote Spring Valley, and California Wash. This included calibrating the groundwater flow model to data collected during Year 1 of the two-year pumping test required by the Nevada State Engineer under Order 1169. The simulated drawdowns agree reasonably well with the observed drawdowns. In California Wash, the seasonal variation observed in the measurements is not present in the simulated water levels, but the longer-term trends are present.

The simulated discharge rates in the Muddy River Springs area and at Rogers and Blue Point springs agree very well with measured values. The simulated streamflow in the Muddy River near Moapa is less than measured, indicating that more water discharges directly into the stream above the gage than is being simulated, rather than downstream of the gage. In the lower part of Meadow Valley Wash, simulated water levels are higher than observed, causing simulated discharge into the stream over a larger area than it occurs. Similarly, simulated water levels are higher than measured in the lower parts of Beaver Dam Wash, causing simulated flow in the stream over a larger reach than observed. In these areas, the model is likely to underestimate the drawdown that occurs because of the larger area in which buffering of drawdown is simulated to occur. The effects of the drawdown on the Muddy

River Springs and discharge into the Muddy River may occur sooner than would be predicted by the model because of the simulated capture in lower reaches of the Muddy River and in the lowermost reaches of Meadow Valley Wash.

Pertinent observations and comments about the model are provided so that the user of the model is aware of limitations that may affect decisions made related to modeling predictions presented in this report and future reports until the model is re-calibrated:

1. The responses of the groundwater system to pumping are determined primarily by the local geology and the hydrologic properties of the aquifers being pumped. Pumping in the carbonate aquifer in the western part of the model produces widespread drawdown because of the high transmissivity and low storativity of the carbonate aquifer. The model predicts that pumping in the Virgin River basin causes more local (less widespread) drawdown of greater magnitude. Elsewhere, current groundwater development is more limited. In the volcanic rocks in the Clover and Delamar Mountains, the complex stratigraphy of the volcanic rocks will likely limit the extent of drawdown, and the productivity of the rocks will likely be highly variable. The complex stratigraphy is not incorporated in the model. The drawdown is reduced by proximity to large-volume springs, and to perennial reaches of streams. This local effect is caused by the buffering of drawdown caused by capture of water at these locations by pumping.
2. The use of the Well Package to simulate ET (so that seasonal changes in ET rates could be used as a driving function during model calibration) may cause head changes to be exaggerated during the long-term predictions of pumping in areas where ET rates are high and where drawdown from pumping occurs. In nature, as the water table declines, the ET decreases which in turn decreases the drawdown. However, in the simulation the rate of ET will remain constant and produce a greater drawdown. This effect has only been observed in a small reach of Meadow Valley Wash, where it appears that drawdown could be oversimulated by tens of feet over a small area. Effects in other areas do not appear to be significant but are unknown.

Prediction of the effects of groundwater pumping will be more reliable in areas where data are available on the responses to pumping and time-varying ET. The best dataset is from the vicinity of the Muddy River Springs and nearby areas (Coyote Spring Valley and California Wash). Thus, predictions for these areas will be most reliable. An evaluation of the uncertainty in model predictions would be a significant effort, and certainly was outside the scope of this current evaluation. An estimate (based on experience with this model and the sensitivity testing that was performed) of the prediction uncertainty for drawdown in these areas would be in the range of 20 to 30 percent over a period of 20 to 30 years. With increasing distance from the area of the Muddy River Springs, the uncertainty increases. In other areas where pumping is occurring (Garnet Valley and the Virgin River Valley), the simulated drawdowns are reasonable, but cannot be compared with measured drawdowns. Thus, there is more uncertainty of the model results to pumping in these areas.

In summary, this model is a great improvement over previous models of the area, because of the advances in information on the geology and hydrology of the study area, and improvements in modeling codes available. This is also the first model to include the Virgin River Valley and Tule Desert, the lower White River Flow System, and the area of Lower Meadow Valley Wash. It can be used to evaluate cumulative effects of pumping in different areas within the model, and to estimate the magnitude and timing of changes that will occur as a result of use of the groundwater. Predictions made using the model will be approximate but can be used to guide decisions about management of the groundwater resource and to determine whether there will be impacts on sensitive environments and on other users of the water. The uncertainty in the predictions will primarily affect the timing of when impacts become significant, not whether there will be impacts.

2.2 2012 PREDICTIVE MODEL SCENARIOS

Seven different predictive scenarios were evaluated, ranging from a continuation into the future of current pumping rates only, through pumping of all existing rights plus all pending applications filed through 2009. Major findings from these scenarios are presented below. For specific findings on the actual seven scenarios, please refer to Tetra Tech (2012b).

1. The impacts of pumping on spring discharge and stream flow will increase as time passes, and as the rates of pumping increase.
2. The effects of drawdown will cause impacts outside the modeled area, and capture flow from adjoining basins, including those in Utah and Arizona. The magnitude of this impact is not known but could be estimated by linking this model with models of other areas.
3. In some areas, the aquifers may not be able to sustain the projected pumping, regardless of effects elsewhere. In Scenarios 4 through 7, the maximum predicted drawdown exceeded 3,000 feet. The model also lowered the rate of production as water levels were lowered to below the assigned screen intervals of the wells.
4. There is uncertainty in these projections that needs to be evaluated further. A detailed uncertainty analysis is recommended. However, it is unlikely that the general conclusions will be altered substantially, but changes in new equilibrium discharge rates (for lower pumping rates) or rates of depletion would be expected to become better defined through the uncertainty analysis.

2.3 2013 POST AUDIT SUMMARY AND CONCLUSIONS

The Tetra Tech (2012a) model was calibrated using information available through December 2011. The pumping of MX-5, and the related collection of water-level and discharge information during the Order 1169 pumping test, has provided additional information that was used in evaluating the predictions made with the model pertaining to the effects of pumping in Coyote Spring Valley. The pumping dataset for the model was updated with monthly pumping information for 2012, and the model was run with this revised dataset. Results indicate that the model under-simulates the amount (i.e., calculates less effect) of drawdown and reduction of spring discharge than occurred as a result of MX-5 pumping during the Order 1169 pumping test period. The observed drawdown is more widespread, and is of greater magnitude, than simulated by the model during this period. The model simulates that the discharge from springs is not affected to a measurable amount, but the real effects are measurable. Thus, predictions that have been made with the model that evaluate the effects of pumping in Coyote Spring Valley should be considered conservative. More specifically, the actual impacts from pumping would be larger and more widespread than simulated by the model.

In addition, a 15-year period after the end of the Order 1169 pumping test on December 31, 2012 was simulated to determine how quickly water level (and spring discharge) recovery is likely to occur. This evaluation indicates that recovery from the 28-month pumping test will occur over years. In the Muddy River Springs area, it was estimated that recovery will be approximately 70% complete after 15 years. In areas that are “distant” from MX-5, results suggest that drawdown can still be increasing 15 years after pumping of MX-5 stopped. If pumping were to occur for longer than 28 months (the total time of the pumping at MX-5 as part of the Order 1169 test), the rate of recovery can be expected to be slower. In addition, the model predicts that drawdown will continue to increase throughout the LWRFS with pumping of approximately 11,500 acre-feet/year (afy).

The data collected during 2012 and the six years since the completion of the Order 1169 pumping test could be used to improve the calibration of the model to the observed effects of pumping in Coyote Spring Valley and neighboring LWRFS basins. A revised model would be expected to simulate greater and more widespread drawdown than the current model, more impact on spring flow, and shorter recovery times. This additional work was beyond the scope and timeframe for the modeling simulation effort conducted as part of this report.

3.0 DEVELOPMENT OF CURRENT PREDICTIVE SCENARIOS

3.1 GROUNDWATER WITHDRAWAL TRENDS IN LWRFS BASINS (2007 – 2017)

Table 3-1 presents the estimated annual withdrawals for each LWRFS basin from 2007 through 2017, as reported by the Nevada State Engineer in Appendix B of Interim Order 1303. Figure 3-1 presents a graphical representation of this annual withdrawal data for each basin. Additional summary statistics are included in Table 3-1, as estimated by the NPS. These statistics include summaries of estimated annual amounts of alluvial aquifer versus carbonate aquifer withdrawals, and the annual withdrawals grouped by the northern basins [Coyote Spring Valley and the Muddy River Springs Area (MRSA)] versus the southern basins (Garnet Valley, California Wash and Black Mountains Area). Figure 3-2 and Figure 3-3 present graphical representations of the additional summary statistics for each basin. This information was important in understanding the spatial and temporal trends of groundwater withdrawals over the last decade and helped to inform our development of the predictive scenarios described in the next section.

Annual pumping in the LWRFS basins has averaged about 11,900 acre-feet/year (afy) from 2007-2017 (Table 3-1). During this same period, approximately 34% (4,000 afy) of the average annual pumping occurred in the alluvial aquifer, most of it in the Muddy River Springs Area, and 66% (7,900 afy) occurred in the carbonate aquifer. Additionally, about 73% (8,700 afy) of the pumping during this period occurred in the northern basins and about 27% (3,200 afy) occurred in the southern basins. By individual basin, the estimated average annual withdrawals in descending order includes: Muddy River Springs Area – 5,800 afy (49%), Coyote Spring Valley – 2,900 afy (25%), Garnet Valley – 1,600 afy (13%), Black Mountains Area – 1,500 afy (12%), and California Wash – 100 afy (1%).

Examination of this same withdrawal data over shorter periods of time within this period of record indicates that greater amounts and percentages of withdrawals have been shifting from the alluvial aquifer to the carbonate aquifer and from the northern basins to the southern basins. In the years prior to the Order 1169 pumping test (2007-2010), annual withdrawals in the LWRFS basins averaged about 12,000 afy. During this same period, approximately 41% (4,900 afy) of the average annual pumping occurred in the alluvial aquifer and 59% (7,100 afy) occurred in the carbonate aquifer. During the Order 1169 pumping test period (2011-2012), average annual withdrawals in the LWRFS basins increased to about 14,535 afy, with approximately 29% (4,300 afy) of the average annual pumping occurring in the alluvial aquifer and 71% (10,300 afy) occurring in the carbonate aquifer. In the two years following the pumping test (2013-2014), average annual withdrawals generally returned to pre-test levels, averaging about 12,600 afy, along with similar pre-test percentages and amounts of annual pumping occurring in the alluvial aquifer (41% or 5,100 afy) and carbonate aquifer (59% or 7,500 afy).

In the last three years of record (2015-2017), the annual withdrawals in the LWRFS basins decreased substantially to about 9,400 afy, with approximately 18% (1,700 afy) of the average annual pumping occurring in the alluvial aquifer and 82% (7,600 afy) occurring in the carbonate aquifer. This drop in groundwater withdrawals is primarily due to reduced alluvial aquifer pumping by NV Energy in the Muddy River Springs Area, as the Reid Gardner power plant was decommissioned and the need for this water diminished. This decrease is reflected in the estimated alluvial aquifer pumping information for the Muddy River Springs Area contained in Appendix B of Interim Order 1303, in which NV Energy's average annual pumping from the alluvial aquifer was about 4,150 afy from 2007-2014, and dropped to about 900 afy from 2015-2017.

In the years prior to the Order 1169 pumping test (2007-2010), approximately 76% (9,100 afy) of the average annual pumping occurred in the northern basins and 24% (2,900 afy) occurred in the southern basins. During the Order 1169 pumping test period (2011-2012), average annual pumping in the northern basins increased slightly to about 80% (11,600 afy), while average annual pumping in the southern basins decreased slightly to 20% (2,900 afy) of the total pumping in the LWRFS basins. In the two years following the pumping test (2013-2014), average annual pumping generally returned to pre-test levels, with similar pre-test percentages and amounts of annual pumping occurring in the northern basins (75% or 9,400 afy) and southern basins (25% or 3,200 afy).

In the last three years of record (2015-2017), the average annual pumping in the northern basins noticeably decreased to about 61% (5,800 afy), while average annual pumping in the southern basins increased to 39% (3,600 afy) of the total pumping in the LWRFS basins. Again, this drop in groundwater withdrawals in the northern basins is primarily due to reduced alluvial pumping by NV Energy in the Muddy River Springs Area. Prior to this period, average annual pumping in the northern basins was about 9,800 afy, and decreased to about 5,700 afy, which represents about a 41% reduction by comparison. The noticeable percentage increase in groundwater withdrawals in the southern basins is mainly the result of this pumping making up a larger percentage of the whole during this 3-year period, as pumping in the northern basins decreased. It should be noted that average annual pumping in the southern basins has been slowly rising in recent years from an average of about 2,900 afy (2007-2012) to about 3,400 afy (2013-2017), which represents about an 18% increase by comparison.

	2007	2008	2009	2010	2011	2012	2013	2014	2015	2016	2017	Average
Coyote Spring V. (210)	3,147	2,000	1,792	2,923	5,606	5,516	3,407	2,258	2,064	1,722	1,961	2,945
Black Mtns Area (215)	1,585	1,591	1,137	1,561	1,398	1,556	1,585	1,429	1,448	1,434	1,507	1,476
Garnet V. (216)	1,412	1,552	1,427	1,373	1,427	1,351	1,484	1,568	1,520	2,181	1,981	1,571
California Wash (218)	27	27	21	26	33	28	66	241	460	252	88	115
MRSA (219)	7,076	6,811	6,379	6,167	6,302	5,852	6,712	6,520	3,898	4,048	3,553	5,756
TOTAL	**13,247**	**11,981**	**10,756**	**12,050**	**14,766**	**14,303**	**13,254**	**12,016**	**9,390**	**9,637**	**9,090**	**11,863**
Alluvial Aquifer	5,361	4,877	4,722	4,764	4,892	3,627	4,629	5,691	2,070	1,856	1,288	3,980
	40%	41%	44%	40%	33%	25%	35%	47%	22%	19%	14%	34%
Carbonate Aquifer	7,886	7,104	6,034	7,286	9,874	10,676	8,625	6,325	7,320	7,781	7,802	7,883
	60%	59%	56%	60%	67%	75%	65%	53%	78%	81%	86%	66%
Basins 210 + 219	10,223	8,811	8,171	9,090	11,908	11,368	10,119	8,778	5,962	5,770	5,514	8,701
	77%	74%	76%	75%	81%	79%	76%	73%	63%	60%	61%	73%
Basins 215 + 216 + 218	3,024	3,170	2,585	2,960	2,858	2,935	3,135	3,238	3,428	3,867	3,576	3,161
	23%	26%	24%	25%	19%	21%	24%	27%	37%	40%	39%	27%

Table 3-1. Summaries of estimated annual withdrawals (acre-feet/year) from LWRFS basins and aquifers, 2007-2017.
[Sources: NDWR Interim Order 1303 (App. B) & NDWR pumping inventories for LWRFS basins and aquifers

3.2 CURRENT PREDICTIVE SCENARIOS EVALUATED

The updated numerical groundwater flow model was originally calibrated to data collected during Year 1 of the two-year pumping test required by the Nevada State Engineer under Order 1169. Subsequent simulations indicated the model conservatively under-predicted the effects of pumping observed during the Order 1169 pumping test (See Section 2 for further discussion). The NPS has requested additional model simulations to qualitatively assess the spatial relationships between the location and redistribution of withdrawals within and between the alluvial and carbonate aquifers in the LWRFS, and the accompanying effects on the groundwater resources in the Muddy River Springs headwater area and Lake Mead NRA.

The updated groundwater flow model was used to conduct a set of three (3) simulations, which allows for the qualitative evaluation of the potential effects of redistributing pumping within and between the alluvial and carbonate aquifers in the LWRFS. In all 3 simulations, the total groundwater withdrawals in the LWRFS basins was maintained at or near the average annual withdrawals achieved during the Order 1169 pumping test (approximately 14,535 afy), so that the effects of progressively moving greater volumes of withdrawals further from the Muddy River Springs Area can be assessed under the same level of pumping stress within the LWRFS. This pumping level was chosen for this set of simulations because pumping impacts to the Muddy River Springs headwater area were observed not only during the Order 1169 pumping test, but also in pumping Simulation #1, which was reported in the earlier modeling simulation work conducted in 2012 (Tetra Tech, 2012b). Under Scenario #1, the pumping level simulated in the LWRFS basins was approximately 14,470 afy, which represented the average pumping conditions occurring in the 2009-2011 time frame, the latter part of which the Order 1169 pumping test had started. By extension, Scenario #1 modeled the estimated long-term pumping impacts in the LWRFS associated with withdrawals estimated to have occurred during the Order 1169 pump test period. Given that similar pumping impacts were observed on the ground and in the groundwater flow model results under similar aquifer stress conditions, this provides confidence in using the groundwater flow model as a tool to qualitatively evaluate whether or not similar pumping impacts may occur under similar aquifer stress conditions, but different spatial pumping arrangements for the 3 simulations modeled as part of this report.

Simulation #1

In the first simulation, a scenario of pumping by "priority date" in the LWRFS was evaluated. Table 3-2 and Figure 3-4 presents the annual withdrawals simulated in each LWRFS basin for the three withdrawal scenarios evaluated in this report. In the first scenario, nearly all of the senior water rights are concentrated in the northern basins. Under this scenario, approximately 96% (13,926 afy) of the simulated annual withdrawals occurs in the northern basins and the remaining 4% (609 afy) of withdrawals occurs in the southern basins. However, withdrawals are distributed between the alluvial and carbonate aquifers in relatively equivalent amounts of about 6,904 afy (47%) and 7,631 afy (53%), respectively (see Table 3-2 and Figures 3-5 and 3-6).

This first simulation serves as a baseline from which to qualitatively evaluate the effects on the Muddy River Springs headwater area resulting from redistributing withdrawals to other basins within the LWRFS, which were evaluated in the two subsequent modeling simulations. This first simulation also may help to inform the State Engineer and stakeholders on the potential pumping effects that might result if stakeholders are unable to voluntarily develop a conjunctive management plan and the State Engineer is forced to "manage by priority date."

	Simulation #1	Simulation #2	Simulation #3
Coyote Spring Valley (210)	6,440	1,934	1,730
Black Mountains Area (215)	0	1,374	1,638
Garnet Valley (216)	519	2,103	4,234
California Wash (218)	90	2,362	3,316
Muddy River Springs Area (219)	7,486	6,761	3,616
TOTAL	14,535	14,534	14,534
Alluvial Aquifer	6,904	2,234	1,376
	47%	15%	9%
Carbonate Aquifer	7,631	12,300	13,158
	53%	85%	91%
Basins 210 + 219	13,926	8,695	5,346
	96%	60%	37%
Basins 215 + 216 + 218	609	5,840	9,188
	4%	40%	63%

Table 3-2. Summaries of simulated annual withdrawals (acre-feet/year) from LWRFS basins and aquifers.

Simulation #2

In the second simulation, the model simulated a scenario in which the existing water rights associated with all active 2017 pumping sites in the LWRFS were apportioned equally (approximately 82.5% of each annual duty) to reach the same level of total annual pumping achieved during the Order 1169 pumping test. Under this scenario, approximately 60% (8,695 afy) of the withdrawals occurs in the northern basins and the remaining 40% (5,840 afy) of the withdrawals occurs in the southern basins. Approximately 15% (2,234 afy) of the pumping occurs in the alluvial aquifer and 85% (12,300 afy) occurs in the carbonate aquifer under the second simulation (see Table 3-2 and Figures 3-5 and 3-6). These percentages are nearly identical to those estimated for the actual 2017 pumping total reported in the LWRFS by the Nevada State Engineer.

Pumping locations in 2017 are more widely distributed throughout the LWRFS compared to the concentration of pumping that occurs under the first simulation. As a result, this second simulation should provide a convenient way to qualitatively evaluate the effects of redistributing larger portions of the alluvial and/or carbonate withdrawals to other areas within the LWRFS, when compared to the effects predicted for the first simulation. This simulation also should help to provide similar qualitative information on the potential effects of redistributing increased amounts of pumping closer to the groundwater resources of concern managed by the NPS.

Simulation #3

In the third simulation, the apportioned pumping occurring at the 2017 pumping sites in the LWRFS under the second simulation was modified to accommodate the redistribution of pumping from alluvial aquifer sites in the northern basins to carbonate aquifer sites in the southern basins. Under this modified scenario, water rights associated with 2 sets of pending change applications submitted by NV Energy (Application Nos. 87735, 87736, 87738 and 88181 for a total combined duty of 1,800 afy) and the Moapa Band of Paiute Indians (Application Nos. 86738 and 86739 for a total combined duty of 500 afy) are redistributed from alluvial aquifer pumping sites in the Muddy River Springs headwater area to carbonate aquifer pumping sites in Garnet Valley and California Wash, respectively. While the actual NV Energy change applications seek to move a portion of an unused junior carbonate aquifer water right in Coyote Spring Valley to existing carbonate aquifer pumping sites in Garnet Valley, this simulation instead chose to examine a possible alternative scenario where a similar amount of senior alluvial water rights owned and used by NV Energy in the Muddy River Springs headwater area is moved to the same existing carbonate aquifer pumping sites in Garnet Valley. Under this third simulation, it is assumed that all water rights held by NV Energy at alluvial aquifer pumping sites in the Muddy River Springs headwater area are not pumped, as NV Energy would have no need for this water in this area. In the case of the Moapa Band of Paiute Indians, their change application actually involves moving senior alluvial aquifer water rights recently acquired from NV Energy from the Muddy River Springs headwater area to existing carbonate aquifer pumping sites on the Moapa River Indian Reservation. Under the third simulation, these senior alluvial rights were pumped in full, while the remainder of the Moapa Band of Paiutes carbonate aquifer water rights were prorated accordingly with other water rights in the southern basins.

Under the third simulation, the existing water rights associated with all 2017 pumping sites in Coyote Spring Valley and the Muddy River Springs Area were used in equal proportions (approximately 74% of each annual duty) to help make up the difference in achieving the same level of total pumping (14,535 afy) tested in the first two simulations. The one exception was pumping of Moapa Valley Water District (MVWD) water rights, which were held at their reported 2017 total pumping level (2,823 afy). This one exception allowed for an intermediate pumping level (compared to the first two simulations) to be evaluated for the MVWD as part of the three simulations performed. Similarly, the existing water rights associated with the remaining 2017 pumping sites in Garnet Valley, California Wash and the northwest corner of the Black Mountains Area were used in equal proportions (approximately 98% of each annual duty), but at a higher level than those in the northern basins. The proration of pumping in these southern basins was higher in order to further accommodate increased levels of pumping in the southern basins, while the overall level of total pumping within the LWRFS remained consistent with the first two simulations. The one exception in the southern basins involved the NV Energy carbonate aquifer pumping site in Garnet Valley, where only the changed amount of 1,800 afy was simulated.

Under this modified scenario, approximately 37% (5,346 afy) of the pumping occurs in the northern basins and the remaining 63% (9,188 afy) of pumping occurs in the southern basins. Approximately 9% (1,376 afy) of the pumping occurs in the alluvial aquifer and 91% (13,158 afy) occurs in the carbonate aquifer under the third simulation (see Table 3-2 and Figures 3-5 and 3-6).

The third simulation should serve to qualitatively evaluate the potential effects of redistributing greater amounts of alluvial pumping away from the Muddy River Springs headwater area and to the carbonate aquifer, when compared to the effects predicted for the first two simulations. This simulation also should help to provide similar qualitative information on the potential effects of redistributing greater amounts of pumping closer to the groundwater resources of concern managed by the NPS.

In addition to the groundwater pumping simulated in the LWRFS basins, groundwater pumping was also simulated in Kane Springs Valley in all three simulation runs to qualitatively evaluate potential pumping interactions with groundwater withdrawals occurring in Coyote Spring Valley and the Muddy River Springs Area. Such interactions could indicate that Kane Springs Valley may be hydrologically connected to Coyote Springs Valley, and therefore may need to be considered for inclusion in the LWRFS. To date, groundwater rights have been permitted, but not used, in the amount of 1,000 afy in Kane Springs Valley. Pumping in Kane Springs Valley was simulated at four proposed points of diversion associated with these existing water rights.

Groundwater pumping was also simulated in the Tule Desert basin and the Virgin River Valley basin to qualitatively evaluate if there could be drawdown interference effects between these two pumping centers and the pumping simulated in the LWRFS. The amount of permitted groundwater simulated in all three scenarios for the Tule Desert basin and the Virgin River Valley basin was 9,340 afy and 12,272 afy, respectively. This approach of simulating the permitted water rights in neighboring basins beyond the current boundary of the LWRFS is consistent with pumping Scenario #2 reported in the earlier modeling simulation work conducted in 2012 (Tetra Tech, 2012b).

4.0 PREDICTION SIMULATIONS

4.1 MODEL SETUP

For the predictions of the effects of the three pumping scenarios, some model datasets described in Tetra Tech (2012b) were modified to represent changes in hydrogeologic conditions since the release of the 2012 groundwater model or improve model performance, while others remained unchanged.

Initial conditions of hydraulic head for the predictive scenarios were calculated by the model at the end of a long-term historic simulation representing December 31, 2017. The existing pumping records used to generate the 2012 model predictive scenarios' initial conditions represented actual groundwater withdrawal through 2011. Recent pumping records published by the Nevada State Engineer's Office (NSEO) were used to update most of the pumping rates through calendar year 2017.

When available, monthly pumping rates were used from 2012 through 2017. Otherwise, annual rates were distributed evenly across the year for which they were reported to NSEO. An exception was made for pumping wells in the Virgin River Valley, for which we applied the wells' average annual pumping rate between 2009 and 2011 through the end of the long-term historic simulation in 2017.

The long-term historic simulation used a combination of monthly and yearly stress periods with single time-steps within each. Model simulations for predictive scenarios used a single stress period of 500 years, split into one hundred 5-year time steps.

No changes were made to the material properties of the 2012 model.

Most boundary conditions were unchanged from the previous predictive scenario model simulations. For the long-term historic simulation, the constant-head boundaries representing Lake Mead were updated to represent changes in lake stage between 2011 and 2017. For the predictive scenarios, the lake stage was set to an elevation of approximately 1,133 feet above mean sea level, the lake stage used for the previous predictive modeling runs, to be consistently conservative. Additionally, some pumping wells were added to the long-term historic simulation, to account for pumping at new locations since 2011 reported to the NSEO. Predictive scenario pumping rates are summarized in Table 4-1, and shown in full in Appendix B.

4.2 PREDICTION RESULTS

The results of the model predictions are presented through a series of the maps showing simulated drawdown at specific times, and graphs of simulated spring discharge and streamflow versus time at select locations. The scales for both map and graphical figures are kept constant across time and scenarios to allow the reader to more easily compare the differences in simulation results.

Information is provided on both drawdown and discharge because of the relationship between the two. For example, when drawdown occurs beneath a stream that is hydraulically-connected with the groundwater system, the drawdown will reduce groundwater discharge to a gaining stream or increase losses from a losing stream. In this case, pumping causing drawdown is capturing water from the stream. Similarly, when a portion of the water being pumped from a well is being captured from a stream, drawdown is reduced. The amount of drawdown is buffered by stream-water capture. Similar effects occur between groundwater and groundwater-fed streams and springs. If the stream or spring is dry or poorly connected to the groundwater system, however, this capture and buffering will not occur. In this way, drawdown maps and stream and spring discharge plots can provide information on whether a stream or spring is flowing and able to buffer drawdown. In the following drawdown maps, streams or springs that appear to be affecting drawdown are likely to be flowing. Streams or springs not affecting drawdown are likely either dry or poorly connected to the groundwater system.

Each of the three scenarios simulate 500 years of pumping, beginning on January 1, 2018. Each of the three simulations also simulate the pumping of approximately 14,535 afy from the LWRFS basin. Pumping was also simulated in Kane Springs Valley, Virgin River Valley, and Tule Desert at rates equal to their full permitted annual duty and kept constant across all three simulations. While alluvial aquifer pumping generally is moved away from the Muddy River Springs Area in Simulations 2 and 3, an exception is noted for a well owned by NV Energy located in California Wash near the Muddy River Springs Area. This well is pumping from the alluvial aquifer in close proximity to the Muddy River and its pumping rate was increased in Simulations 2 and 3 due to the methodology applied to increase pumping in the southern basins, which includes California Wash. As a result, the alluvial pumping associated with this well is likely contributing to the simulated reduction of streamflow in the river. Relative magnitude and distribution of pumping across the three scenarios is shown in Figures 4-1 through 4-3. Table 4-1 presents the withdrawal amounts for several of the larger water right holders in the LWRFS basins that were modeled in the three simulations. Appendix B presents a similar summary of the withdrawal amounts for all water-right holders that were modeled in the three scenarios.

Water Right Holder Pumping Basin	2017 Withdrawals (ac-ft)	Simulation #1 Withdrawals (afy)	Simulation #2 Withdrawals (afy)	Simulation #3 Withdrawals (afy)
Coyote Springs Investment				
Coyote Spring Valley	1,399	4,140	1,650	1,477
SNWA				
Coyote Spring Valley		1,957		
Garnet Valley	1,048		1,433	1,709
Moapa Valley Water District				
California Wash		90		
Muddy River Springs Area	2,823	1,000	5,079	2,823
Moapa Band of Paiutes				
California Wash	43		2,063	2,960
Muddy River Springs Area		500		
NV Energy				
California Wash	29		299	356
Garnet Valley	75	75	62	1,800
Muddy River Springs Area	296	3,160	795	
LDS Church				
Muddy River Springs Area	240	2,329	655	586
Nevada Cogeneration Associates				
Black Mountains Area	1,507		1,374	1,638

Table 4-1. Summary of the largest simulated annual withdrawals for selected water-right holders in the LWRFS basins and aquifers.

4.3 DRAWDOWN MAPS

Simulated drawdown for model layer 1 is shown on Figures 4-4 through 4-12 for simulated times of 10, 100, and 200 years for Simulations 1 through 3. An inset has been added to these figures to show a close-up of drawdown around the Muddy River Springs Area. Model layer 1 represents the water table, and drawdown in deeper layers will differ from that simulated for layer 1, depending on the depth of pumping and geology. In addition, simulated streams are located in layer 1, and their effect on the drawdown is greatest in layer 1. The model-predicted drawdown is calculated based on the simulated water levels at the end of the long-term historic simulation (December 2017). It should be noted that annual pumping within the LWRFS basin at the end of the long-term simulation was approximately 9,500 afy, substantially less than the pumping rate assigned for the three scenarios (14,535 afy).

4.3.1 Results at 10 years

To reiterate, withdrawals in Simulation 1 were modeled on a "priority date" basis, resulting in the vast majority of the pumping (96%) occurring in the northern basins and with pumping being distributed between the alluvial aquifer and the carbonate aquifer in roughly equivalent proportions (47% and 53%, respectively). Under this scenario, the largest withdrawals involved pumping of alluvial-aquifer rights held by NV Energy and the LDS Church in the Muddy River Springs Area, and pumping of carbonate-aquifer rights held by Coyote Springs Investment (CSI) and Southern Nevada Water Authority (SNWA) in Coyote Spring Valley and MVWD in the Muddy River Springs Area (Table 4-1 and Appendix B).

The simulated drawdown for Simulation 1 after 10 years is between 2 and 20 feet in the immediate vicinity of Muddy River Springs Area (Figure 4-4). In the carbonate aquifer beneath Coyote Spring Valley, Hidden Valley (North), Garnet Valley, California Wash, and the rest of Muddy River Springs Area, drawdown is widespread and ranges from 1 to 5 feet, with larger drawdown occurring near the CSI and MX-5 wells. In Kane Springs Valley, drawdown from the three southernmost wells, pumping from the carbonate aquifer, have coalesced with the drawdown in the LWRFS basin. The northernmost well in Kane Springs Valley is pumping from the less hydraulically conductive volcanic aquifer and has induced a larger drawdown of over 20 feet that extends over a more restricted area. Drawdown induced by pumping in Kane Springs Valley and Muddy River Springs Area is also observed in Lower Meadow Valley Wash.

In Simulation 2, withdrawals for existing water rights associated with active 2017 pumping sites in the LWRFS were apportioned equally (approximately 82.5% of each annual duty) resulting in about 60% and 40% of the pumping occurring in the northern basins and southern basins, respectively, and with pumping being distributed between the alluvial aquifer and the carbonate aquifer in amounts of 15% and 85%, respectively. Under this scenario, the largest withdrawals involved pumping of carbonate aquifer rights held by MVWD in the Muddy River Springs Area, the Moapa Band of Paiute Indians in California Wash and CSI in Coyote Spring Valley, and pumping of alluvial aquifer rights held by NV Energy and the LDS Church in the Muddy River Springs Area (Table 4-1 and Appendix B).

The simulated drawdown for Simulation 2 at 10 years shows smaller magnitude and less extensive drawdown in the Muddy River Springs area compared to Simulation 1 (Figure 4-5). Maximum drawdown is reduced to below 10 feet, and the area of drawdown exceeding 2 feet has been substantially reduced. Drawdown in the carbonate aquifer beneath Coyote Spring Valley and surrounding basins has expanded to the south and has reached the southern model boundary (Las Vegas Valley Shear Zone), and the area of increased drawdown of 2 to 5 feet has shifted to the Moapa Band of Paiute Indians' wells in California Wash. Drawdown in Kane Springs Valley carbonate aquifer has been isolated from the area of drawdown in Coyote Springs Valley, but the relative extent and magnitude has remained the same.

In Simulation 3, existing water rights in the northern basins were used in lower proportions (about 74% of each annual duty) while existing water rights in the southern basins were used in higher proportions (about 98%), compared to Simulation 2. Additionally, this scenario also simulated the transference of alluvial-aquifer rights held by NV Energy and the Moapa Band of Paiute Indians in the northern basins to carbonate aquifer pumping sites in the southern basins, thus resulting in about 37% and 63% of the pumping occurring in the northern basins and southern basins, respectively, and with pumping being distributed between the alluvial aquifer and the carbonate aquifer in amounts of 9% and 91%, respectively. Under this scenario, the largest withdrawals involved pumping in the carbonate aquifer by the Moapa Band of Paiute Indians in California Wash, MVWD in the Muddy River Springs Area, NV Energy and SNWA in Garnet Valley and CSI in Coyote Spring Valley. Pumping of alluvial aquifer rights by several entities on the order of a few hundred acre-feet/year occurred mostly in the Muddy River Springs Area (Table 4-1 and Appendix B).

The simulated drawdown for Simulation 3 shows an even less extensive area of drawdown in the Muddy River Springs Area, but still shows an area of drawdown exceeding 5 feet, centered on the LDS East well (Figure 4-6). Drawdown in the carbonate aquifer beneath Coyote Spring Valley and surrounding basins is now mostly concentrated in Hidden Valley, Garnet Valley, and California Wash, and shows drawdown exceeding 2 feet over an area much larger than either Scenario 1 or 2. The pattern of drawdown in Kane Springs is very similar to what is seen in Simulation 2.

Patterns of drawdown in the Virgin River Valley and Tule Desert remain the same across all three scenarios, which is expected because pumping rates remain the same. Less permeable aquifer units in both basins result in larger drawdown cones over relatively restricted areas compared to drawdown seen in the LWRFS carbonate aquifer to the west and southwest. Changes in Lake Mead stage between the historic and predictive simulation causes drawdown to be calculated at some model cells adjacent to the lake that are not associated with pumping.

4.3.2 Results at 100 years

The simulated drawdown for Simulation 1 after 100 years shows an increase in drawdown in the Muddy River Springs Area, with the magnitude of drawdown increasing to over 5 feet for most of the inset area and the area of drawdown exceeding 10 feet expanding to the west to include the Lewis and Arrow Canyon wells (Figure 4-7). Drawdown in the carbonate aquifer beneath Coyote Spring Valley and surrounding basins has expanded and now extends over most or all of Coyote Spring Valley, Hidden Valley (North), California Wash, Garnet Valley, and northeastern Las Vegas Valley, with drawdown between 5 and 20 feet. Drawdown in the Kane Springs Valley carbonate and volcanic aquifers has coalesced with both drawdown in the LWRFS and Tule Desert, in the Lower Meadow Valley Wash area.

Simulated drawdown for Simulation 2 is very similar to Simulation 1, with some exceptions (Figure 4-8). Drawdown in the Muddy River Springs Area is limited to less than 10 feet. However, the area of drawdown exceeding 5 feet has expanded south to the model boundary in Las Vegas Valley, Garnet Valley, California Wash, and the eastern edge of the Black Mountains Area.

Simulated drawdown for Simulation 3 after 100 years is again similar to that of Simulation 1 (Figure 4-9). The area of drawdown exceeding 5 feet has been further reduced compared to Simulation 2, and drawdown is visibly buffered by the Muddy River. Drawdown has exceeded 10 feet in California Wash, Garnet Valley, Hidden Valley (North), northeastern Las Vegas Valley, and a small portion of the Black Mountains Area.

Drawdown in the Virgin River Valley and Tule Desert have exceeded 100 feet and 200 feet, respectively, and coalesced. Flow in the Virgin River continues to buffer drawdown, preventing drawdown from expanding into Utah, and almost entirely separating drawdown cones in Nevada.

4.3.3 Results at 200 years

Simulated drawdown in Simulation 1 follows a similar pattern as simulated drawdown after 100 years (Figure 4-10). Drawdown exceeding 10 feet in the Muddy River Springs Area has expanded and the buffering of drawdown by the Muddy River is less impactful. Drawdown in the northern portion of the LWRFS carbonate aquifer, southern Kane Springs Valley, and southwestern Lower Meadow Valley Wash has exceeded 10 feet.

Simulated drawdown in Simulation 2 after 200 years closely resembles Simulation 1 (Figure 4-11). Drawdown exceeding 10 feet is less extensive than in Simulation 1 after the same amount of time, and the Muddy River is more effective at buffering drawdown. Drawdown increases to above 10 feet in the carbonate aquifer below the southern portion of the LWRFS and extends to the Las Vegas Valley shear zone.

Drawdown for Simulation 3 after 200 years varies only a small amount from Simulation 2 (Figure 4-12). There is a small reduction in the area of drawdown exceeding 10 feet in the Muddy River Springs Area and southwestern Lower Meadow Valley Wash, while the area and magnitude of drawdown increases slightly in northeastern Las Vegas Valley and the Black Mountains Area.

4.4 FLOW HYDROGRAPHS

Hydrographs of simulated spring discharge and stream flow for the three simulations are presented below. Locations of the flow measurements within the model are shown in Figure 4-13.

4.4.1 Muddy River Springs Area Flow Comparison

Changes in spring discharge at the Muddy River Springs Area are shown in Figure 4-14. Spring discharge is relatively consistent through all three simulations. After ten years, Simulation 3 shows the highest discharge rate for all springs, followed by Simulation 2, then Simulation 1, each with marginally lower rates. Between 200 and 300 years, discharge for Simulation 2 decreases to below the discharge rate for both Simulations 1and 3. The relative amounts of discharge then remain the same for the reminder of the simulation, with Simulation 3 having the highest discharge and Simulation 2 the lowest. Across all springs, Muddy Spring shows the largest difference in discharge between the three scenarios, while the Cardy-Lamb and Pederson springs show the smallest differences. Pederson Spring is predicted to dry up at approximately 250 years, but because the model underpredicted drawdown by several fold in this area, the spring would likely become dry many years sooner.

The primary effect on moving the locations of pumping to the southern basins is a slight increase in flow rate, causing a delay of 10 to 25 years for flow from Muddy Spring to match the Simulation 1 rates for the other simulations. Delays are shorter for the other springs.

4.4.2 Streamflow Comparison

Simulated streamflow along Muddy River are shown in Figures 4-15 through 4-18. Figure 4-15 shows all simulated streamflow for all locations at the same scale for Simulation 1. Figures 4-16 through 4-18 show streamflow at select areas for all three scenarios.

Simulated streamflow at the upper confluence of the Muddy River is approximately 5 cubic feet per second (cfs) at the beginning of all simulations and decreases to about 3 cfs after 500 years. Simulation 3 shows the highest discharge throughout the simulation with Simulation 2 showing a just slightly lower value, and Simulation 1 the lowest.

Simulated streamflow at Muddy River near Moapa is approximately 25 cfs at the beginning of the simulation and decreases to approximately 17 cfs after 500 years. Scenario 3 has the highest flow rate, followed by Scenario 2. Scenario 2 and 3 flow about 1.1 cfs and 1.5 cfs higher than Scenario 1, respectively, at the beginning of the simulation, with the differences reducing over time.

Streamflow for the Muddy River above Overton, near Glendale, and near Bowman Reservoir all show similar flow rates throughout the simulation. Flow rates range from 61 to 63 cfs at the beginning of the simulation and decrease to 45 to 46 cfs after 500 years. Simulation 3 shows the highest discharge among the three scenarios throughout the simulation, with the difference in flow rates across scenarios decreasing with time. The large increase in streamflow between Moapa and Glendale is caused by displacement along the Muddy Mountain Thrust Fault, placing lower permeability Mesozoic rocks against the carbonate aquifer (Figure 1-2), damming up the water and raising water levels on the upgradient side of the fault.

Simulated streamflow for the Muddy River at Lake Mead decreases from about 53 cfs down to 38 cfs over 500 years of simulation. Again, Simulation 3 shows a higher discharge than Simulation 1 or 2, with a decrease in the differences between them over time. The changes in rate of streamflow decline during the first forty years of simulation is largely caused by the difference in lake stage between the historical and predictive simulations.

Moving the locations of pumping to the southern basins is predicted to result in higher surface-water flow rates but not enough to prevent the decline of streamflow. The time required for the flow rate to reach a certain value (e.g., 60 cfs) is increased at the different downstream locations (for example, near Glendale and Bowman Reservoir) is increased by 25 to 40 years for Simulation 2, and 35 to 60 years for Simulation 3. Moving pumping will not prevent impacts to surface-water rights, but does delay the impact.

4.4.3 Rogers and Blue Point Springs Comparison

Simulated discharge from the Rogers and Blue Point Springs are shown in Figure 4-18. Discharge from the springs is highest at the beginning of the simulations, about 2.25 cfs. Discharge from the springs is reduced by the smallest amount in Simulation 1, down to 2.1 cfs. Simulation 2 and 3, in which greater amounts of pumping were moved from the northern to southern basins, simulated a slightly lower discharge of about 2.0 cfs after 500 years.

5.0 DISCUSSION

5.1 SIGNIFICANT OBSERVATIONS FROM THE PREDICTIVE RUNS FOR THE THREE SCENARIOS

<u>Predicted decline in spring and stream discharge</u> - The predictive runs assume a constant pumping rate of about 14,535 afy. This rate is consistent with the average annual pumping that occurred during the Order 1169 test, and is greater than the rate in 2017 (approximately 9,300 afy) that was used to develop the starting conditions for the predictive simulations. As a result, the increase in pumping in the predictive runs causes an increase in the rate of decline in discharge at the Muddy River Springs Area that decreases over several decades. Pumping at the lower historical rates caused smaller declines in simulated discharge. Because the model under-simulated the effects of the Order 1169 pumping, it should not be used to predict the magnitude of the change in discharge for future pumping without additional calibration. However, it can be used to predict the direction of change, and to compare the effects of pumping from different areas. The model suggests that re-establishing hydrologic equilibrium with a LWRFS pumping rate of 14,535 afy will require centuries (longer than 500 years in the simulations), and that discharge rates will continue to decrease for centuries.

<u>Temporal and spatial changes in drawdown</u> – The spatial distribution of drawdown is affected by the locations of pumping at early time, but less so for longer time. During the first several decades, there are very apparent differences between the maps of predicted drawdown for the three scenarios. Shifting of pumping to the southern basins (Simulations 2 and 3) causes drawdown in the northern basins to decrease initially. However, with continued pumping, the differences between the maps decrease significantly as the drawdown extends over long distances from the pumping wells. A strategy of moving pumping away from the MRSA to protect the springs and streamflow will be beneficial for a relatively short period of time (decades) but is unlikely to be a long-term solution, because of the continuity of the carbonate aquifer and its hydraulic properties.

<u>Limited effect of changing pumping from northern to southern basins on discharge rates</u> – With the total pumping rate from the LWRFS held constant in all 3 simulations, changing the locations of pumping cause small changes in the predicted impacts on discharge rates at early time. Because the drawdown effects extend over large areas (as demonstrated by the Order 1169 test), as pumping continues the differences between the scenarios diminish. In

other words, moving pumping away from the MRSA and Coyote Spring Valley will provide small benefits to protecting spring flows in the MRSA and stream flows in the Muddy River for several decades. As pumping continues over time, these benefits will diminish as drawdown expands and captures spring and river discharge. The similarity in the results from all three simulations reaffirms the DOI Agencies' 2013 conclusion that due to the high degree of hydrologic connectivity throughout the LWRFS basins, carbonate pumping anywhere within the connected basins (even under different spatial pumping configurations) will affect groundwater levels throughout these basins and eventually capture the major forms of natural discharge in the area – spring/stream discharge and evapotranspiration present within the connected basins.

Because the model underpredicted the drawdown and reduction in discharge caused by the Order 1169 pumping, it should not be used in its current state to determine what the safe yield of the LWRFS aquifer is. In addition, it has not been calibrated to data on the effects of pumping in the southern basins because of the limited data available at the time it was calibrated. Recalibration of the model using data collected since 2011 is recommended to improve the accuracy of predictions.

We emphasize again that simulation of the recovery from ceasing MX-5 pumping at the end of the Order 1169 test indicated that recovery in locations distant from the pumping well occurs slowly, consistent with the observed water-level data. Thus, we support the approach of phased development of the aquifer, in which limited pumping is performed, accompanied by monitoring of water levels and discharge rates both near and distant from the productions well(s). If wells are drilled in the southern basins, pumping should be limited initially and data collected on pumping rates and water level changes. Use of recording pressure transducers is highly recommended so that barometric pressure and earth-tide effects can be measured and removed.

While the model predicts that discharge from Rogers and Blue Point Springs will be decreased a small amount by moving larger volumes of pumping from the northern basins to the southern basins, we emphasize that there is a paucity of information on the effects of pumping in the southern part of the flow system on these springs. Continued monitoring of spring discharge and aquifer water levels is essential.

5.2 NSEO REQUESTED INPUT

Three different predictive scenarios were simulated to qualitatively evaluate the potential effects of redistributing pumping within and between the alluvial and carbonate aquifers in the LWRFS. These three simulations, described in more detail in Section 3, include:

1. Simulation evaluating pumping by priority date,
2. Simulation evaluating the redistribution of increased pumping to the carbonate aquifer, and
3. Simulation evaluating the redistribution of pumping from the alluvial aquifer to the carbonate aquifer.

Major findings from the groundwater modeling simulations as they relate to matters of interest to the State Engineer defined in the Interim Order 1303, are discussed below.

5.2.1 Geographic Boundary of LWRFS

Interim Order 1303 requested analysis to assist in addressing the geographic boundary of the groundwater / surface water system comprising the LWRFS. Results from Simulations 1, 2, and 3 clearly show that under the same aquifer stress (pumping) levels but different pumping configurations, the potential impacts from future pumping are migrating into or out of the existing boundaries of the five basins in Interim Order 1303. The adjacent basins of interest include Black Mountains, Kane Springs Valley, and Las Vegas Valley. Each of these three basins are discussed below.

5.2.1.1 Black Mountains Area (HA 215)

Drawdown has the potential to extend into the Black Mountain Area basin further than just the northwestern corner. The drawdown maps presented previously show drawdown only at the water table (layer 1 in the model). The carbonate aquifer extends continuously in the subsurface further to the east, beneath the Muddy Mountain thrust sheet exposed at the surface, until it is truncated by the Rogers Spring fault (Page and others, 2011). Rogers and Blue Point Springs, which discharge carbonate-rock sourced water, are located along the Rogers Spring fault. Drawdown occurs in the model within this deeper carbonate layer in all three simulations.

These springs were simulated as discharging from all layers present at their locations, including the Mesozoic section and the underlying deep carbonate aquifer. The simulated discharge declines a small amount in all three scenarios, with the greatest decline in Simulations 2 and 3 when greater amounts of pumping in the carbonate aquifer are moved closer to these springs. Given that other lines of hydrogeologic evidence also strongly support a pathway for groundwater flow to Rogers and Blue Point Springs in this area (see Section 1.1), we would recommend including all of the Black Mountains Area basin within the final boundary of the LWRFS. The effects of pumping on these springs should be considered when permitting and management decisions are made. A phased approach for development and monitoring is recommended for wells in the southern part of the LWRFS.

5.2.1.2 Kane Springs Valley (HA 206)

The three pumping simulations assumed that 1,000 afy was pumped from 4 wells situated along Kane Springs Valley. The three southwestern wells were assumed to be completed in the carbonate aquifer, and the northeastern-most well was assumed to be completed in volcanic rocks based on available geologic mapping of the valley. In Simulation 1, the drawdown cones of the three carbonate wells are predicted to coalesce with the drawdown caused by pumping from the wells in Coyote Spring Valley. In Simulations 2 and 3, in which the pumping from the Coyote Spring Valley wells was reduced, the drawdown in both valleys had not coalesced by 10 years, but had coalesced by 100 years. Thus, the model predicts that the carbonate aquifers in Kane Springs Valley and Coyote Spring Valley are connected. Observations of water levels in wells CSVM-4 and KMW-1 show drawdown caused by the pumping in MX-5 during the Order 1169 test, showing that pumping effects are transmitted into this area in a few months. Based on this evidence, we would recommend including all of Kane Springs Valley within the final boundary of the LWRFS.

5.2.1.3 Las Vegas Valley (HA 212)

The Las Vegas Valley Shear Zone is considered to be the down-gradient end of the LWRFS. While there is a gradient across the shear zone indicating that there may be groundwater flowing from the LWRFS into the rest of Las Vegas Valley, the amount of flow is believed to be very low. The model simulates that boundary flow to be 0 afy, using a no-flow boundary condition, based on estimates developed by USGS hydrologists Jim Harrill and Doug Bedinger. Thus, the model does not simulate a change in the flow rate as a result of drawdown along the boundary. In reality, there is likely to be a small, but probably insignificant, decrease in the flux into Las Vegas Valley across the shear zone. Thus, it is appropriate to manage Las Vegas Valley groundwater separately from the LWRFS.

5.2.2 Aquifer Recovery Since Order 1169 Test

The predictive modeling discussed here did not provide additional insight into aquifer recovery. Refer to section 2.3 for conclusions of the Order 1169 post-audit modeling.

5.2.3 Sustainable Quantity of Groundwater Pumping and Relationship of Pumping Location on Spring and River Flow

As indicated previously, the model under-simulated the amount of drawdown that occurred during the Order 1169 test. Flow measurements indicated significant changes in the discharge of Pederson Spring (to about one-third of the pre-test flow rate). The model indicates that pumping at approximately 14,535 afy under several different pumping configurations continues to cause declines in discharge in the MRSA area for more than 500 years, so that it is very likely that continuation of the Order 1169 test would have caused Pederson Spring and maybe other Muddy River springs to dry up over time. Thus, the annual, sustainable quantity of groundwater available is less than 14,500 afy. Recall that the post-audit of the model involved a simulation of the system to evaluate recovery times, in which the LWRFS was pumped at approximately 11,500 afy. Besides showing that drawdown can continue to increase for years in areas distant from the pumping well, there was a general decline in water levels that was not associated with the Order 1169 test caused by the simulated pumping at this rate.

The simulations indicate that there would be short-term benefit on the flows of the Muddy River Springs and the Muddy River from moving greater amounts of alluvial and/or carbonate withdrawals from the northern basins into the southern basins, but that after several decades, the drawdown caused by the southern pumping will affect the springs in the MRSA and the surface flow in the Muddy River to the same degree as if the pumping locations were not changed. Moving pumping to the south and from the alluvial aquifer delays impacts by a few decades, but does not avoid impacts. The similarity in the results from all three simulations reaffirms the DOI Agencies' 2013 conclusion that due to the high degree of hydrologic connectivity throughout the LWRFS basins, carbonate pumping anywhere within the connected basins (even under different spatial pumping configurations) will affect groundwater levels throughout these basins and eventually capture the major forms of natural discharge in the area – spring/stream discharge and evapotranspiration present within the connected basins.

5.2.4 Effects of Replacing Alluvial Well Pumping with Carbonate Well Pumping on Delivery of Decreed Rights on the Muddy River

It would seem that decreasing the pumping from the alluvium along the Muddy River would reduce the capture of the surface flow by this alluvial groundwater pumping, and as a result, would provide more surface water to meet the delivery of decreed rights on the Muddy River. The simulations indicate that this is true in early years, but not enough to fully offset the reduction in surface flow that is predicted in later years. Changing the location of the wells delays, but does not prevent, the impact. Recall that the alluvial groundwater is primarily derived from the underlying carbonate aquifer in the area north and west of Glendale. Increased pumping of the carbonate aquifer, as a result of moving alluvial aquifer pumping to the carbonate aquifer, reduces the discharge of groundwater from the carbonate aquifer into the overlying alluvial aquifer over time as drawdown effects expand throughout the LWRFS.

6.0 REFERENCES

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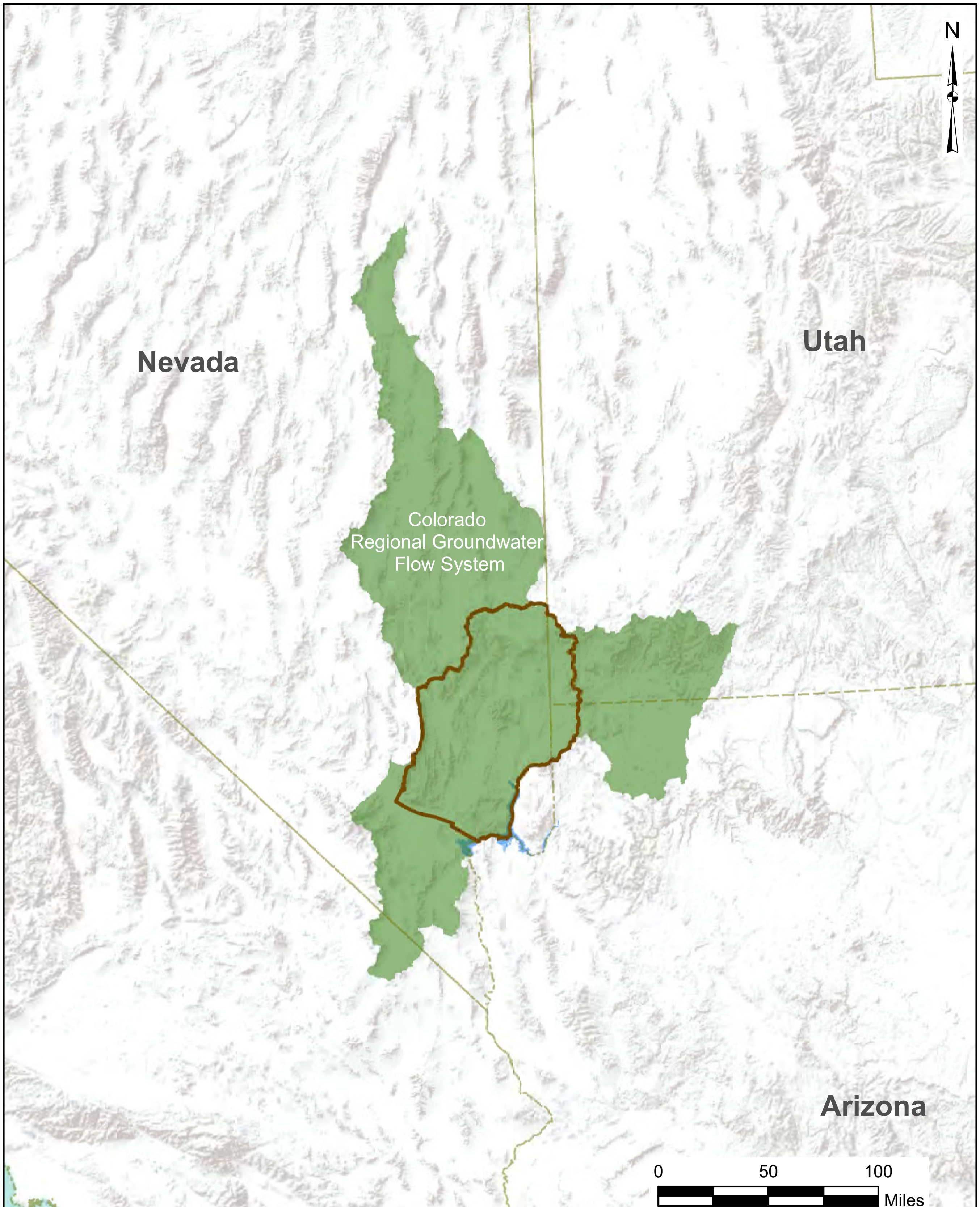
Figures

SE ROA 51468



Legend

Active Model Domain

State Boundary

Colorado River Regional Groundwater Flow System

TITLE:

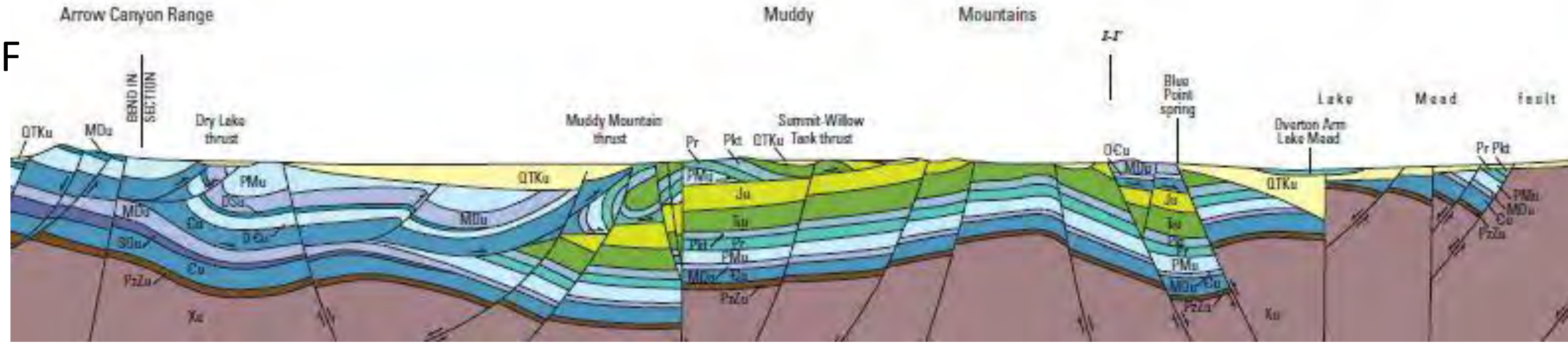
Location of Study Area

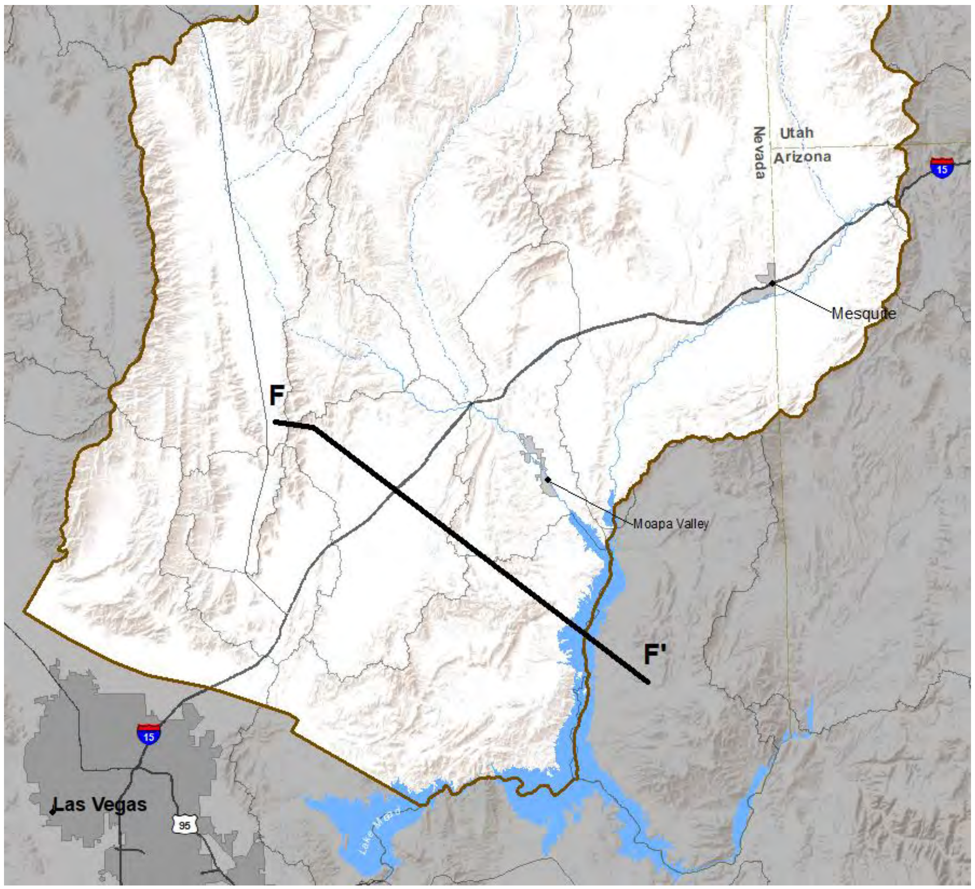
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada, Utah, and Arizona**

TETRA TECH

FIGURE

1-1



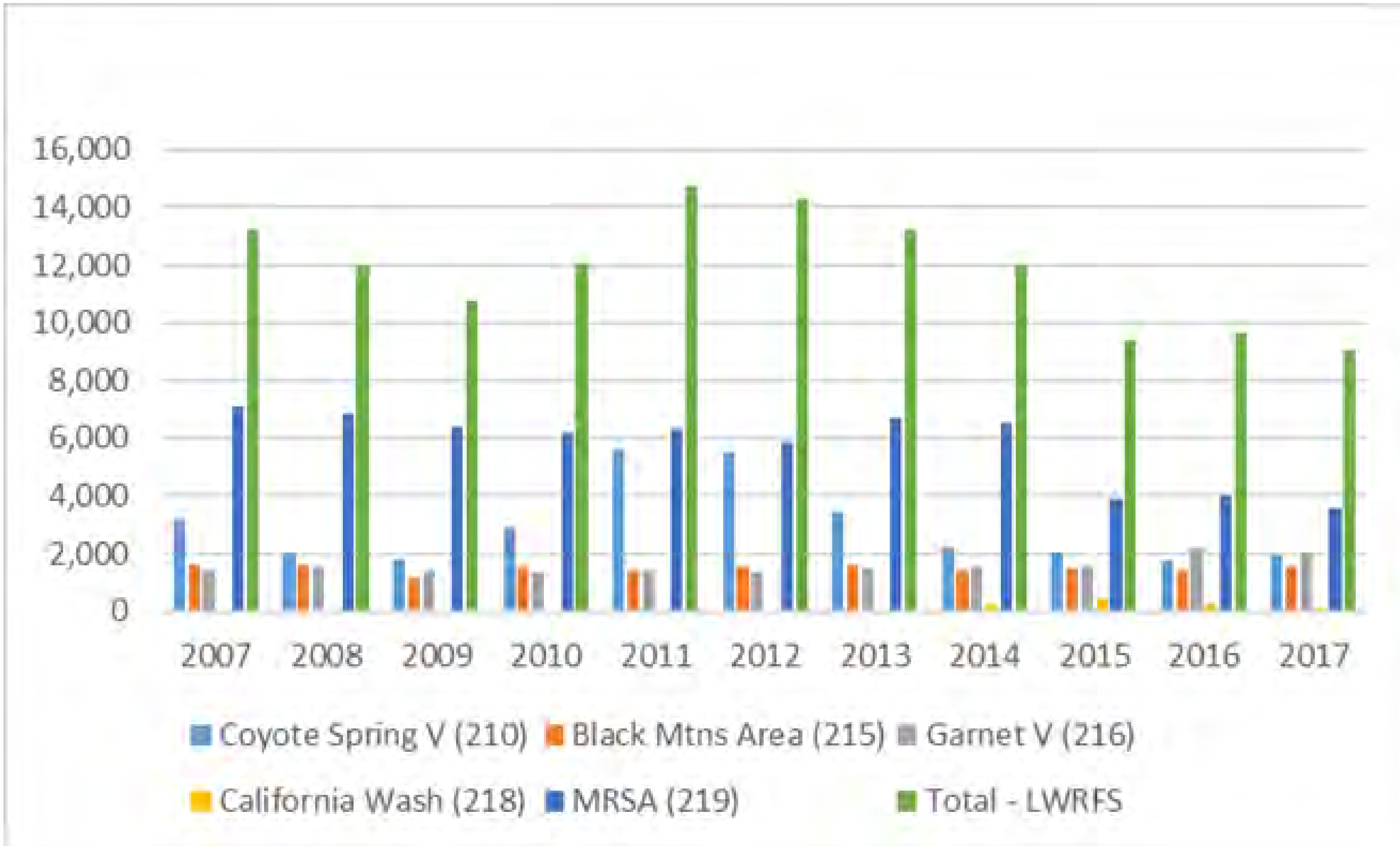


TITLE:	PORTION OF SECTION F-F' (PAGE ET AL., 2011)
LOCATION:	Lower Colorado River Flow System

TETRA TECH

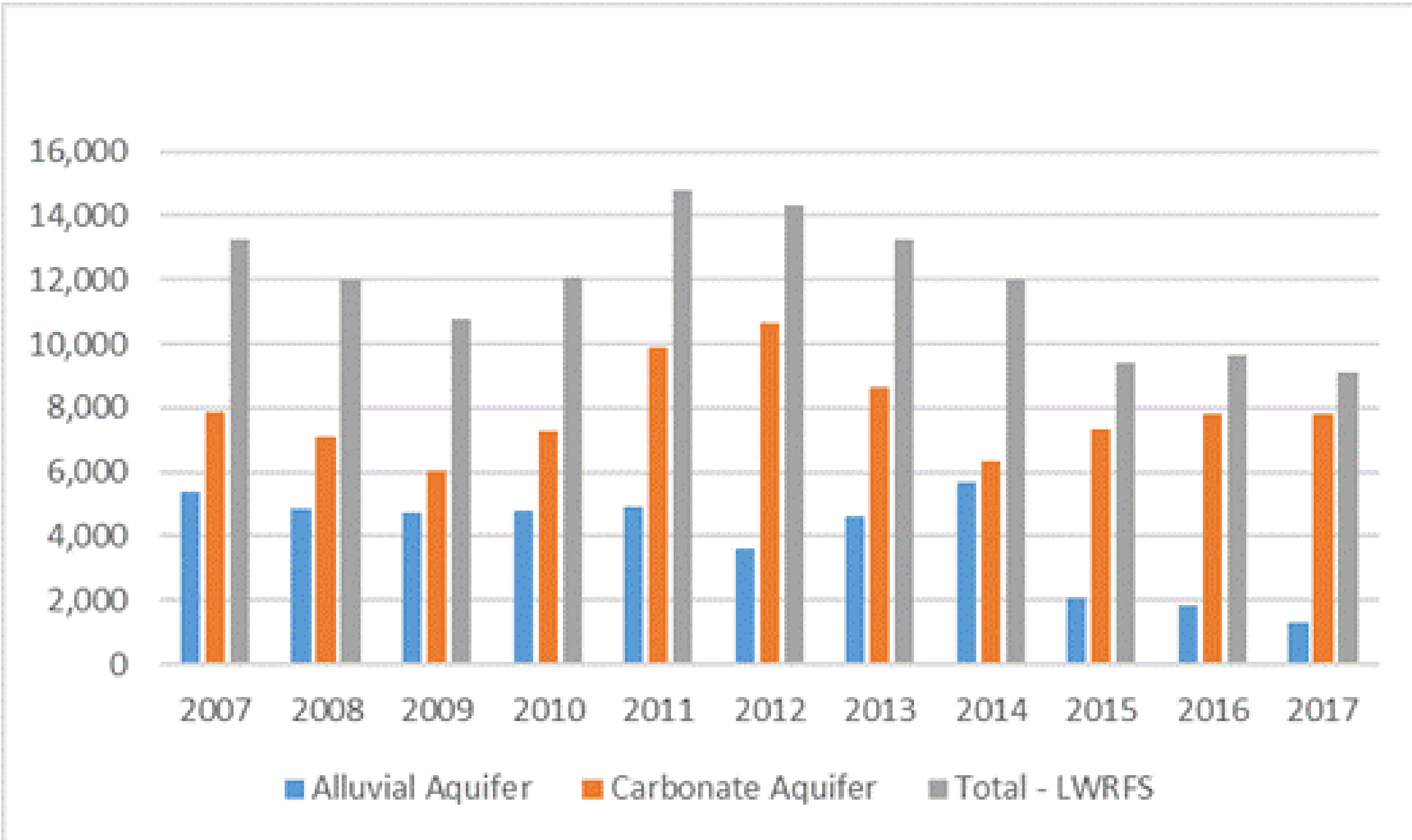
APPROVED		FIGURE
DRAFTED	SRH	1-2
PROJECT #	117-0524303	
DATE	6/24/2019	

SE ROA 51470



[SOURCES: NDWR INTERIM ORDER 1303 (APP. B) & NDWR PUMPING INVENTORIES FOR LWRFS BASINS]

TITLE:			
ESTIMATED ANNUAL WITHDRAWALS (AFY) FROM LWRFS BASINS, 2007-2017			
LOCATION: Lower Colorado River Flow System			
National Park Service	APPROVED		FIGURE
	DRAFTED	GK	3-1
	PROJECT #	117-0524303	
	DATE	6/24/2019	



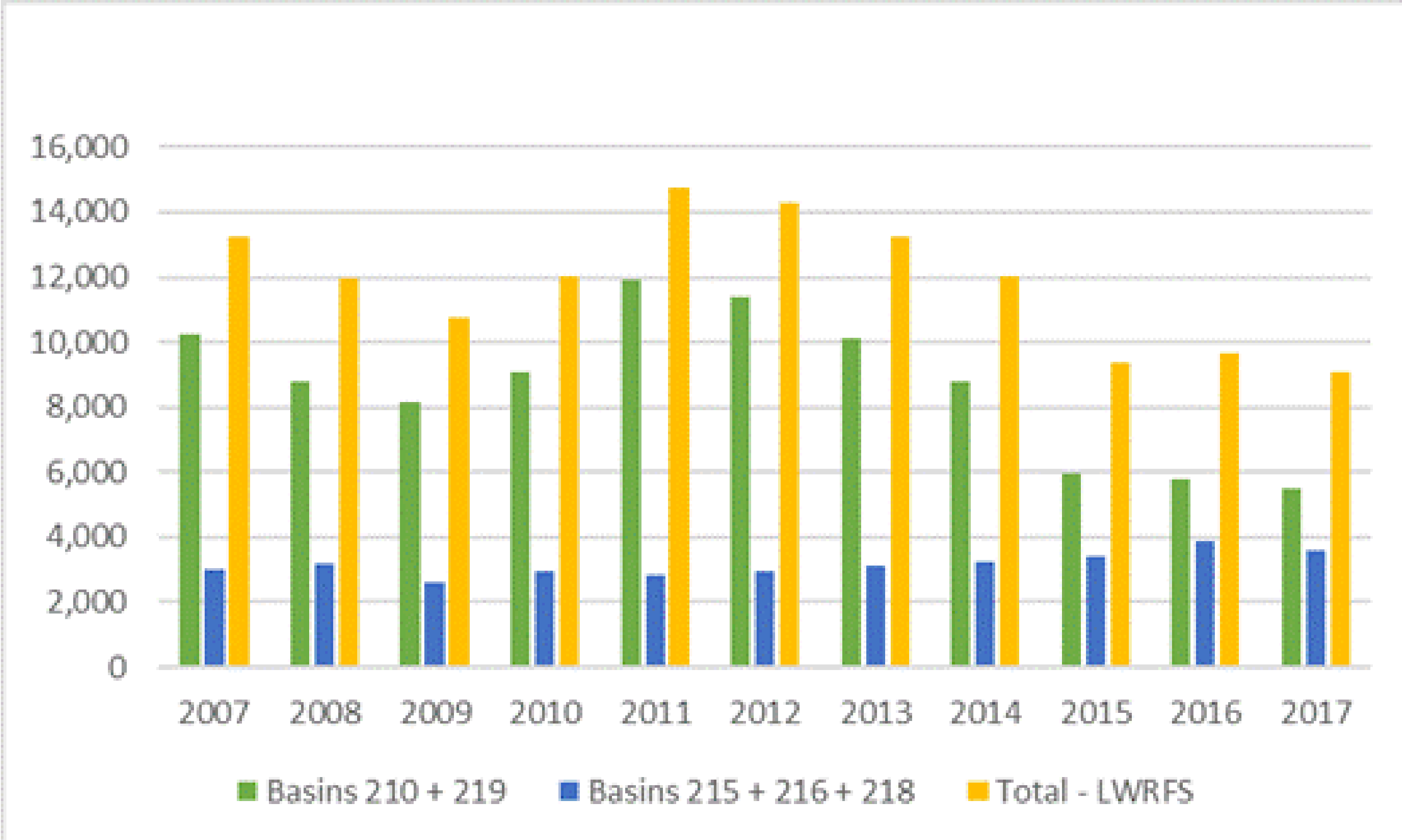
[SOURCES: NDWR INTERIM ORDER 1303 (APP. B) & NDWR PUMPING INVENTORIES FOR LWRFS BASINS]

TITLE:
ESTIMATED ANNUAL WITHDRAWALS (AFY) FROM THE ALLUVIAL AQUIFER AND CARBONATE AQUIFER IN THE LWRFS, 2007-2017

LOCATION:
Lower Colorado River Flow System

National Park Service

APPROVED		FIGURE
DRAFTED	GK	**3-2**
PROJECT #	117-0524303	
DATE	6/24/2019	



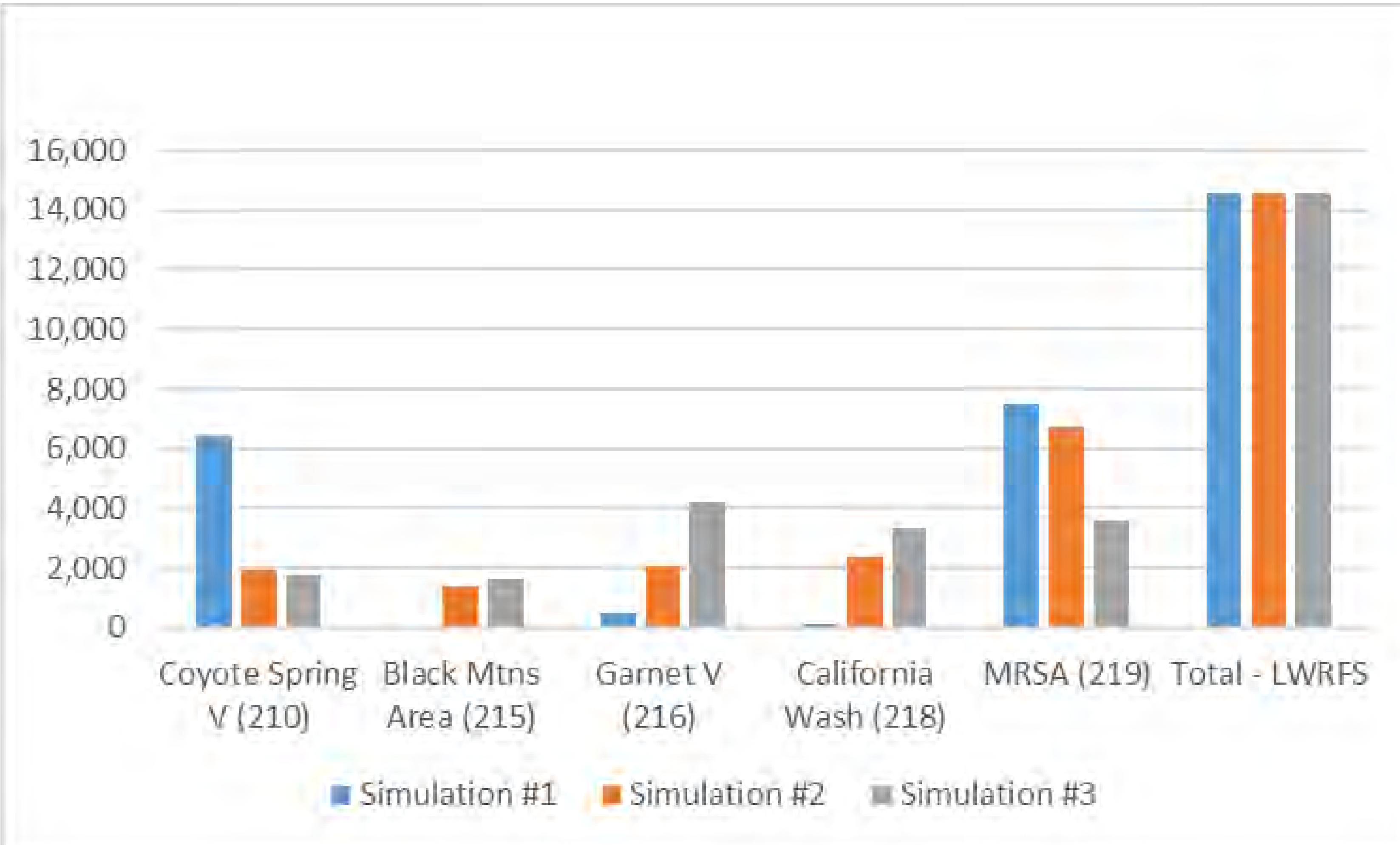
[SOURCES: NDWR INTERIM ORDER 1303 (APP. B) & NDWR PUMPING INVENTORIES FOR LWRFS BASINS]

TITLE: **ESTIMATED ANNUAL WITHDRAWALS (AFY) FROM THE NORTHERN BASINS (BASINS 210 & 219) AND THE SOUTHERN BASINS (BASINS 215, 216 & 218) IN THE LWRFS, 2007-2017**

LOCATION: **Lower Colorado River Flow System**

National Park Service

APPROVED		FIGURE
DRAFTED	GK	**3-3**
PROJECT #	117-0524303	
DATE	6/24/2019	



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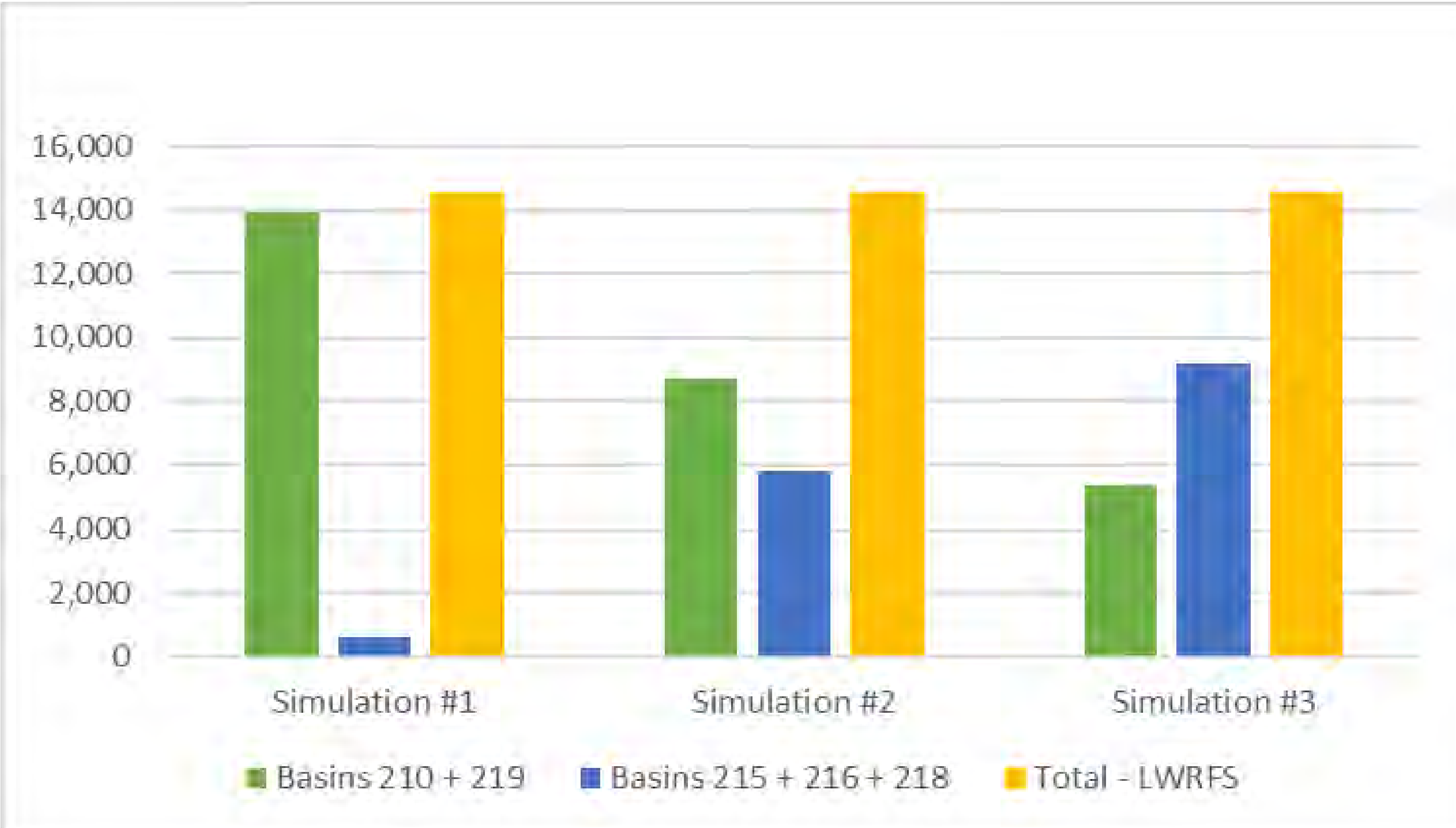
LOCATION: Lower Colorado River Flow System			
National Park Service	APPROVED		FIGURE
	DRAFTED	GK	3-4
	PROJECT #	117-0524303	
	DATE	6/24/2019	

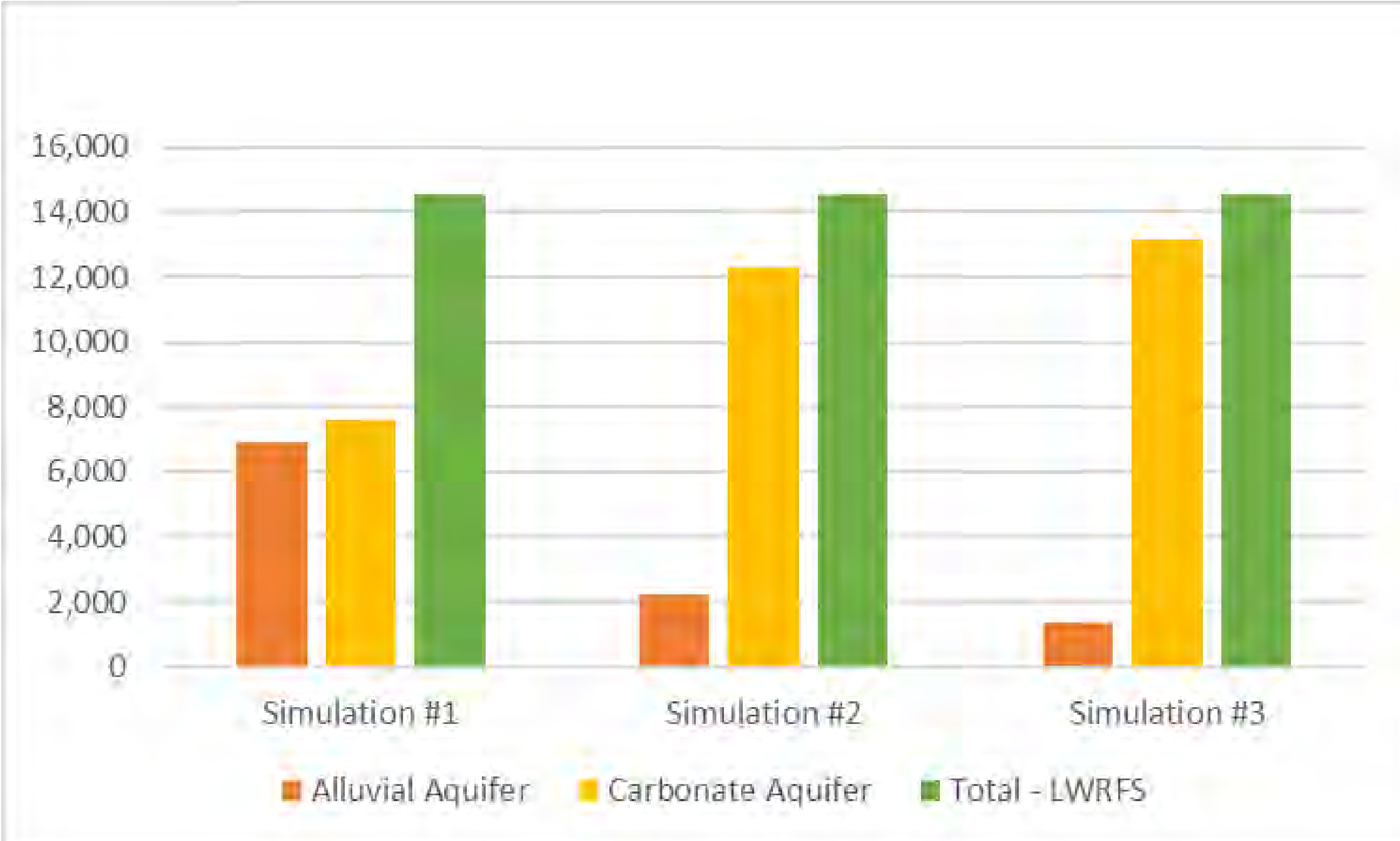


TITLE: **SIMULATED ANNUAL WITHDRAWALS (AFY) FROM THE NORTHERN BASINS (BASINS 210 & 219) AND THE SOUTHERN BASINS (BASINS 215, 216 & 218) IN THE LWRFS.**

LOCATION: **Lower Colorado River Flow System**

National Park Service

APPROVED		FIGURE
DRAFTED	GK	**3-5**
PROJECT #	117-0524303	
DATE	6/24/2019	



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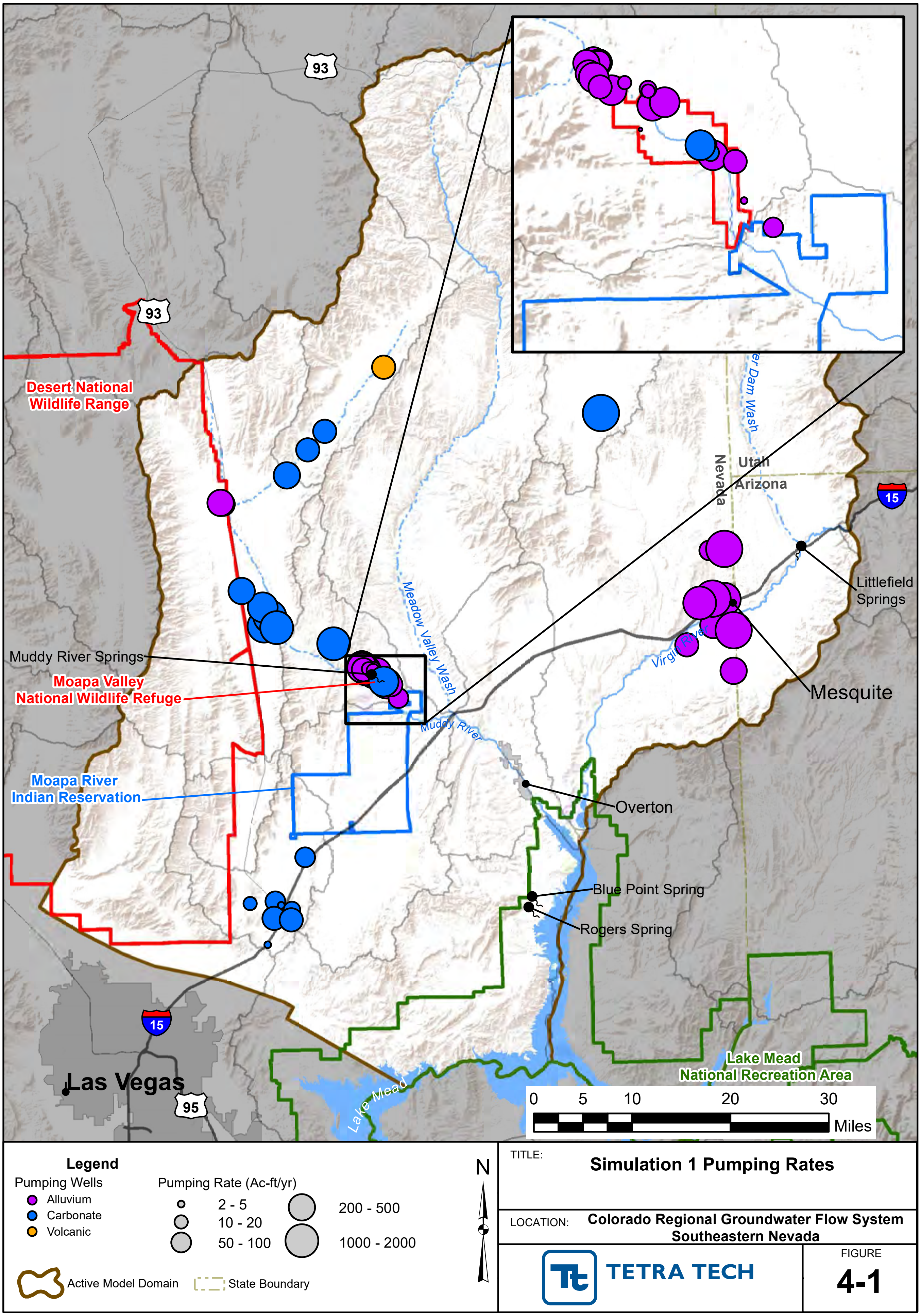
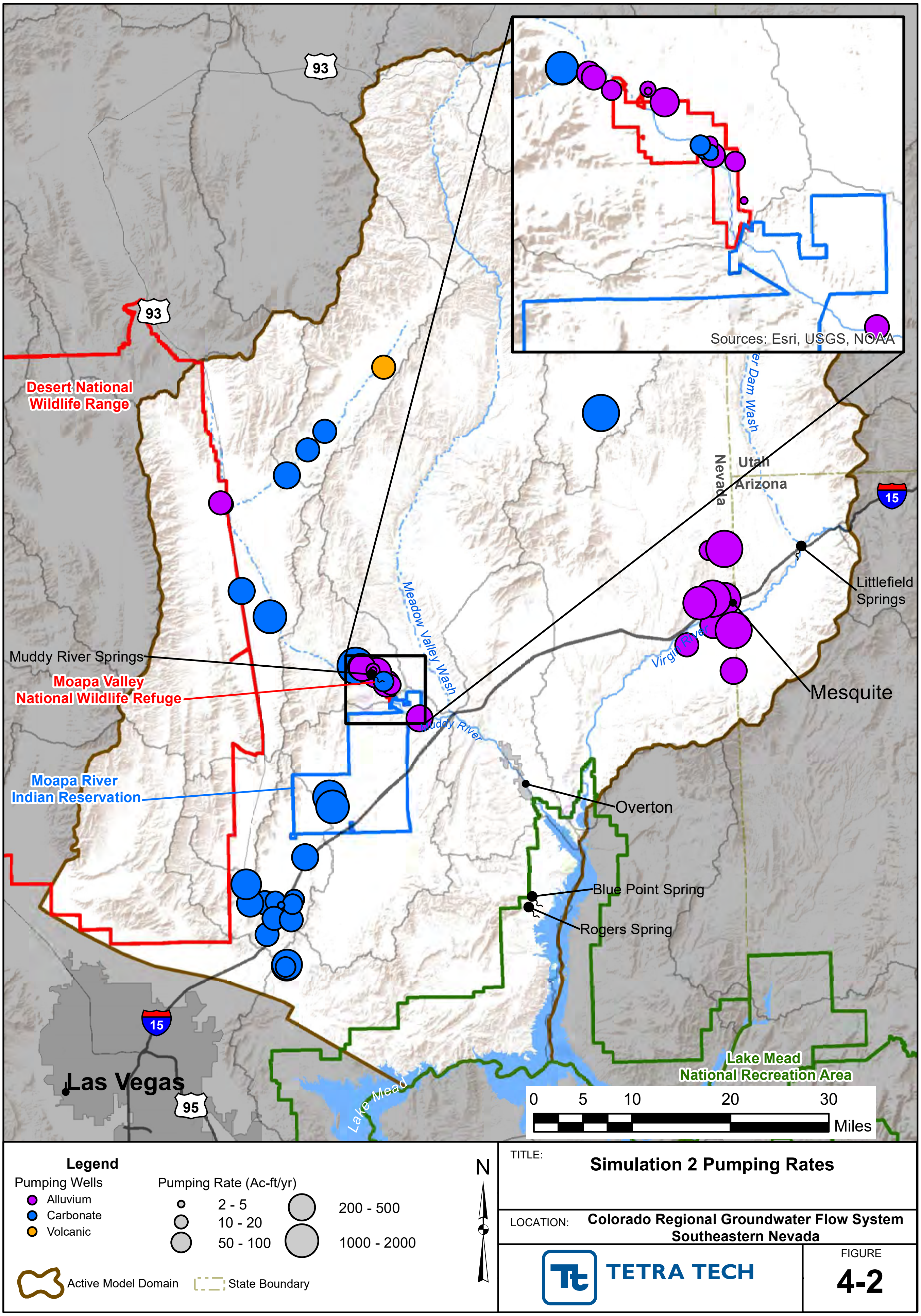
LOCATION: Lower Colorado River Flow System			
National Park Service	APPROVED		FIGURE
	DRAFTED	GK	3-6
	PROJECT #	117-0524303	
	DATE	6/24/2019	



93
93
Desert National
Wildlife Range
Muddy River Springs
Moapa Valley
National Wildlife Refuge
Moapa River
Indian Reservation
Meadow Valley Wash
Muddy River
Virgin River
Beaver Dam Wash
Nevada
Utah
Arizona
15
Littlefield
Springs
Mesquite
Overton
Blue Point Spring
Rogers Spring
Lake Mead
National Recreation Area
Lake Mead
15
Las Vegas
95
0 5 10 20 30
Miles
Legend
Pumping Wells
Alluvium
Carbonate
Volcanic
Pumping Rate (Ac-ft/yr)
2 - 5
10 - 20
50 - 100
200 - 500
1000 - 2000
Active Model Domain
State Boundary
N
TITLE:
Simulation 1 Pumping Rates
LOCATION:
Colorado Regional Groundwater Flow System
Southeastern Nevada
TETRA TECH
FIGURE
4-1

SE ROA 51477



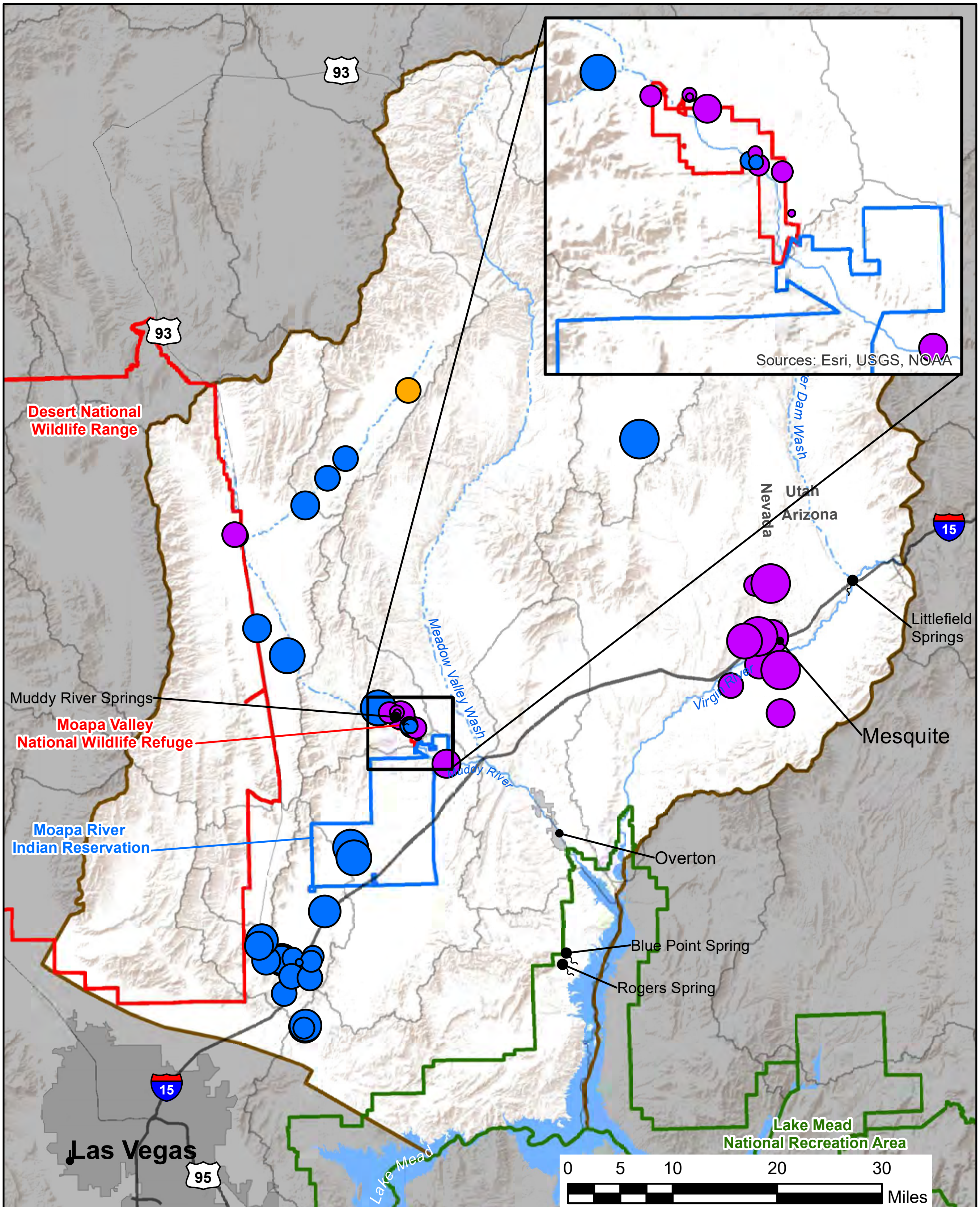
TITLE: **Simulation 2 Pumping Rates**

LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-2**

SE ROA 51478



Legend

Pumping Wells
- Alluvium
- Carbonate
- Volcanic

Pumping Rate (Ac-ft/yr)
- 2 - 5
- 10 - 20
- 50 - 100
- 200 - 500
- 1000 - 2000

Active Model Domain
State Boundary

N

TITLE: **Simulation 3 Pumping Rates**

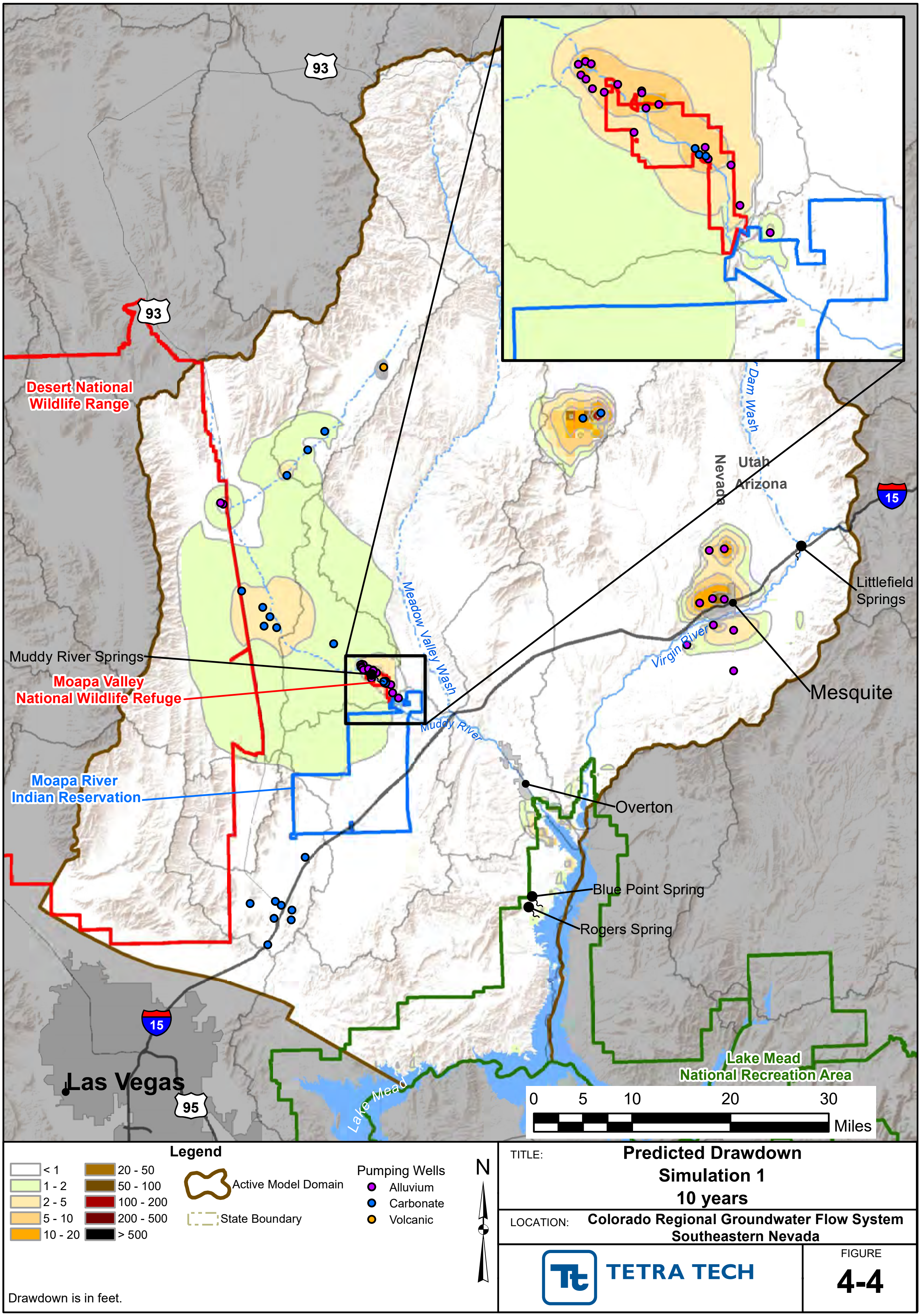
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-3**

SE ROA 51479

JA_16610



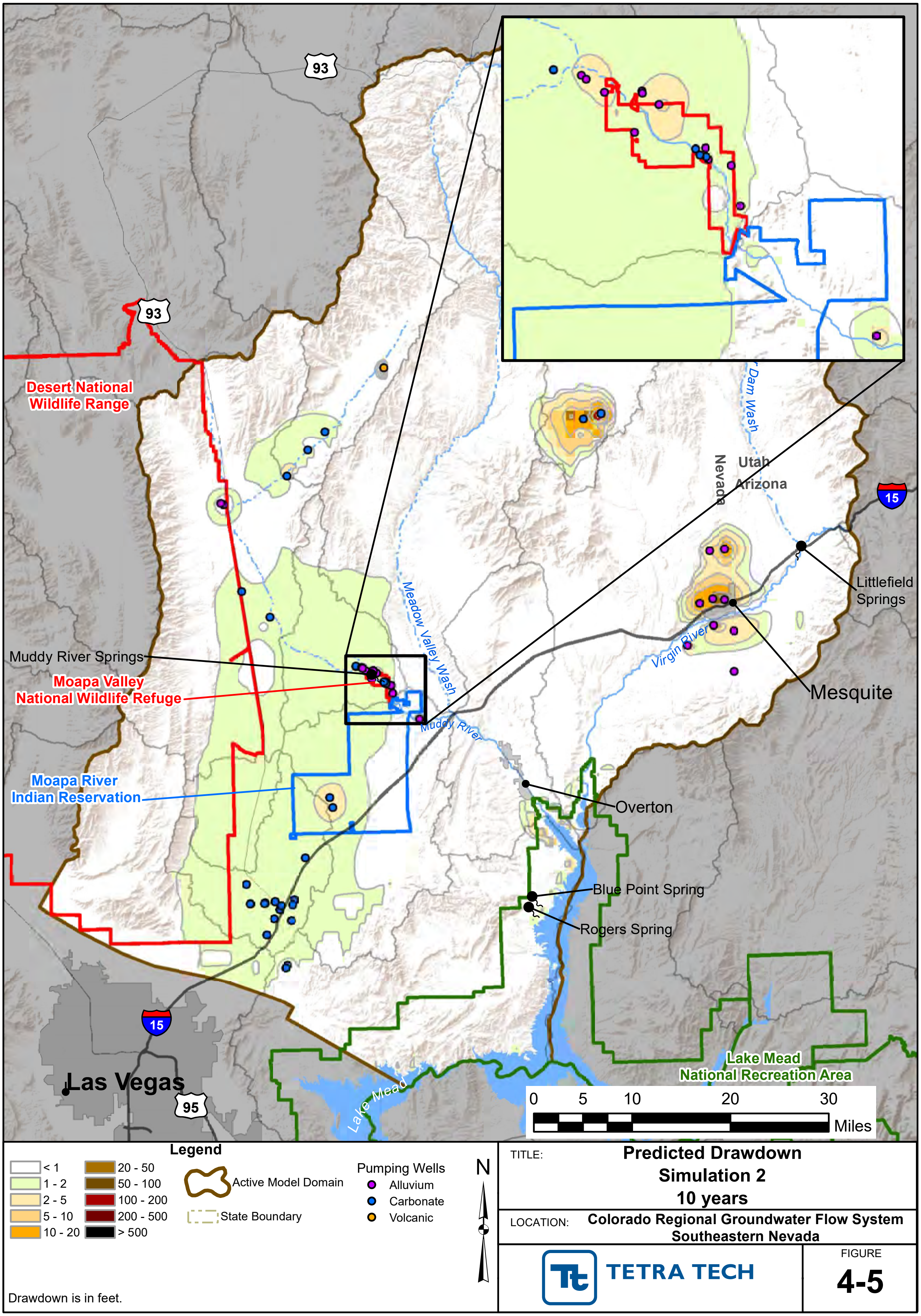
TITLE: **Predicted Drawdown**
Simulation 1
10 years

LOCATION: **Colorado Regional Groundwater Flow System**
Southeastern Nevada

TETRA TECH

FIGURE
4-4

SE ROA 51480



TITLE: **Predicted Drawdown Simulation 2 10 years**

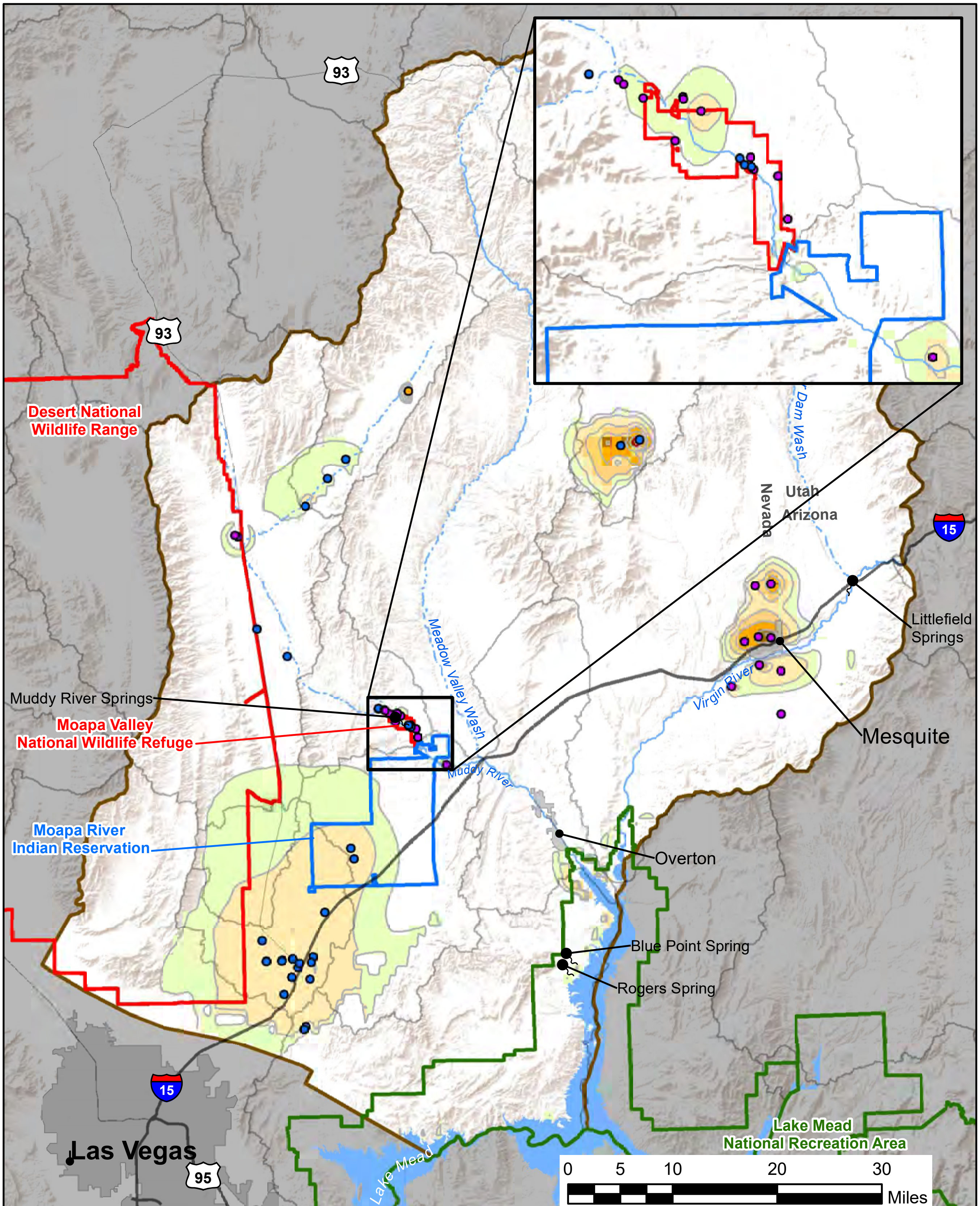
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-5**

SE ROA 51481

JA_16612



Legend

< 1	20 - 50
1 - 2	50 - 100
2 - 5	100 - 200
5 - 10	200 - 500
10 - 20	> 500

Active Model Domain

State Boundary

Pumping Wells
- Alluvium
- Carbonate
- Volcanic

N

Drawdown is in feet.

TITLE: **Predicted Drawdown Simulation 3 10 years**

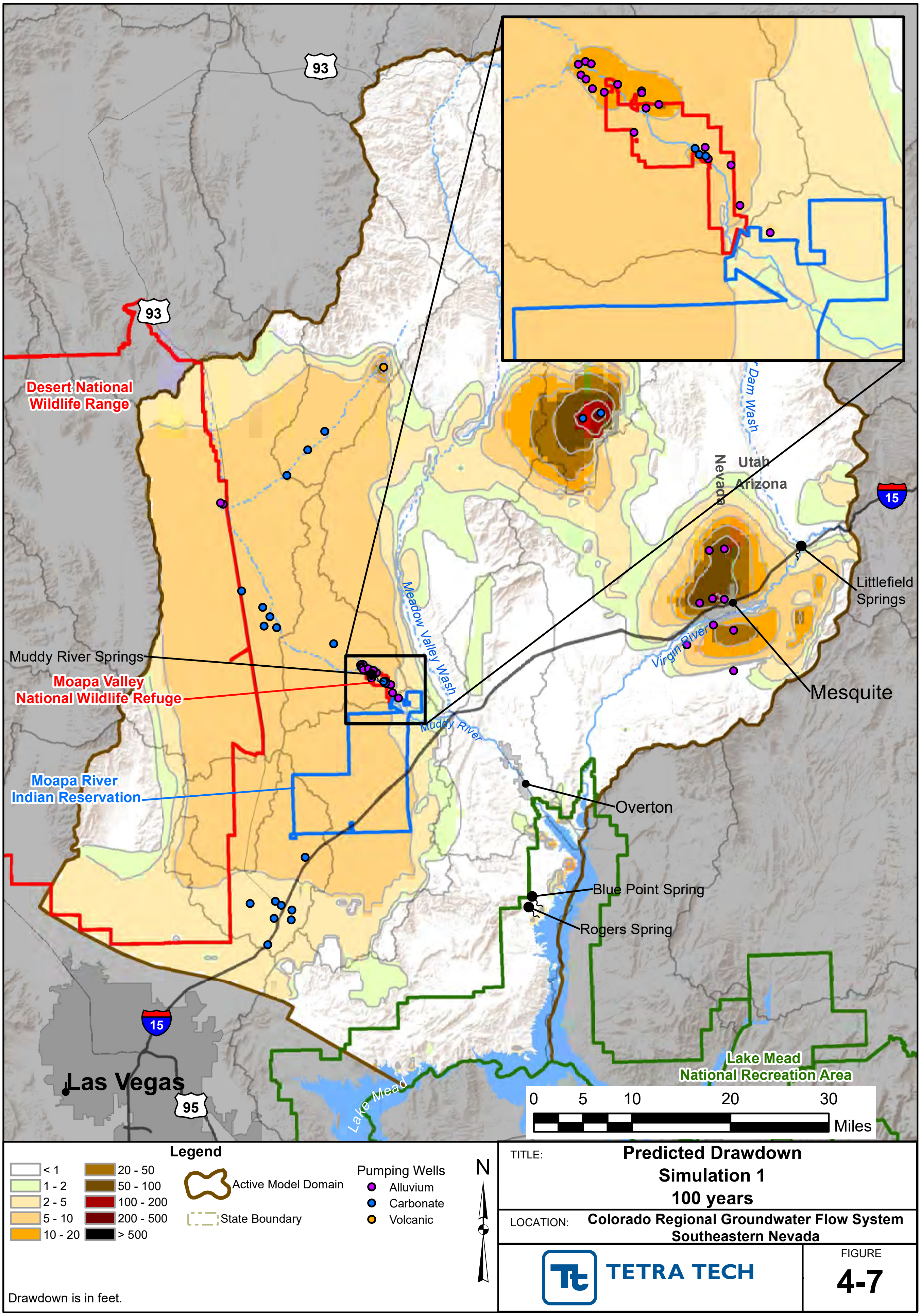
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-6**

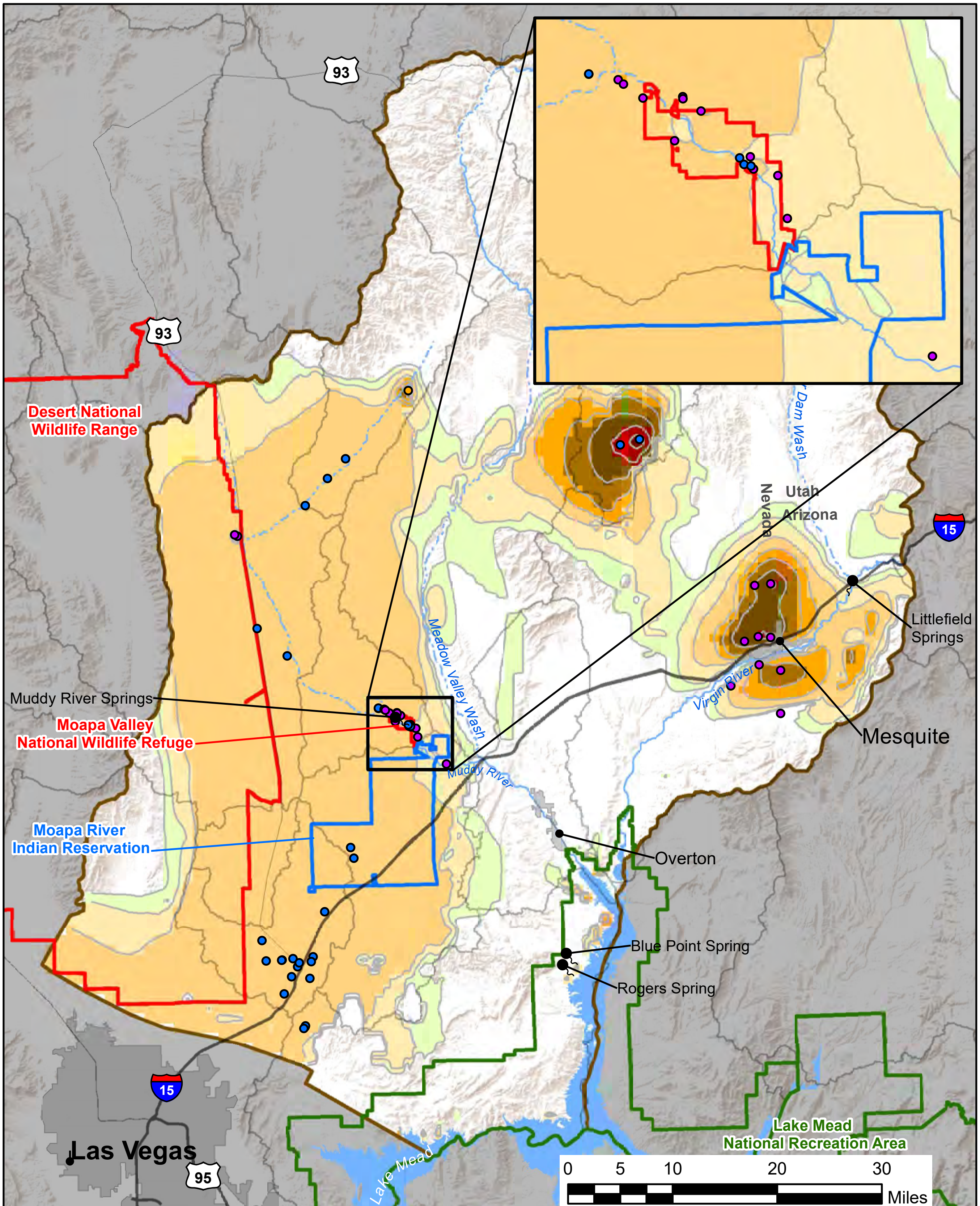
SE ROA 51482

JA_16613



SE ROA 51483

JA_16614



Legend

- < 1
- 1 - 2
- 2 - 5
- 5 - 10
- 10 - 20
- 20 - 50
- 50 - 100
- 100 - 200
- 200 - 500
- > 500
- Active Model Domain
- State Boundary

Pumping Wells
- Alluvium
- Carbonate
- Volcanic

N

Drawdown is in feet.

TITLE: **Predicted Drawdown**
Simulation 2
100 years

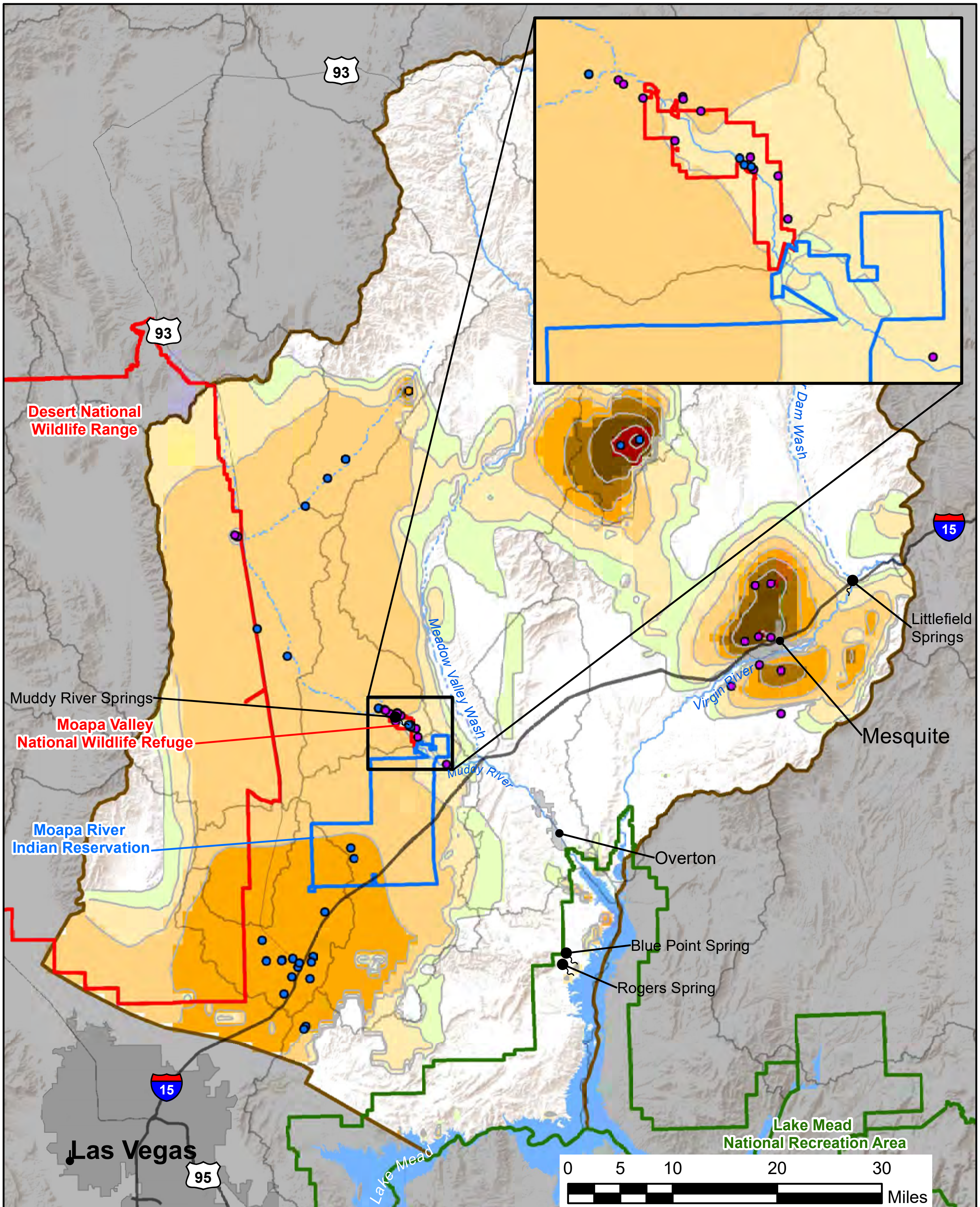
LOCATION: **Colorado Regional Groundwater Flow System**
Southeastern Nevada

TETRA TECH

FIGURE
4-8

SE ROA 51484

JA_16615



Legend

< 1	20 - 50
1 - 2	50 - 100
2 - 5	100 - 200
5 - 10	200 - 500
10 - 20	> 500

Active Model Domain

State Boundary

Pumping Wells
- Alluvium
- Carbonate
- Volcanic

N

Drawdown is in feet.

TITLE: **Predicted Drawdown**
Simulation 3
100 years

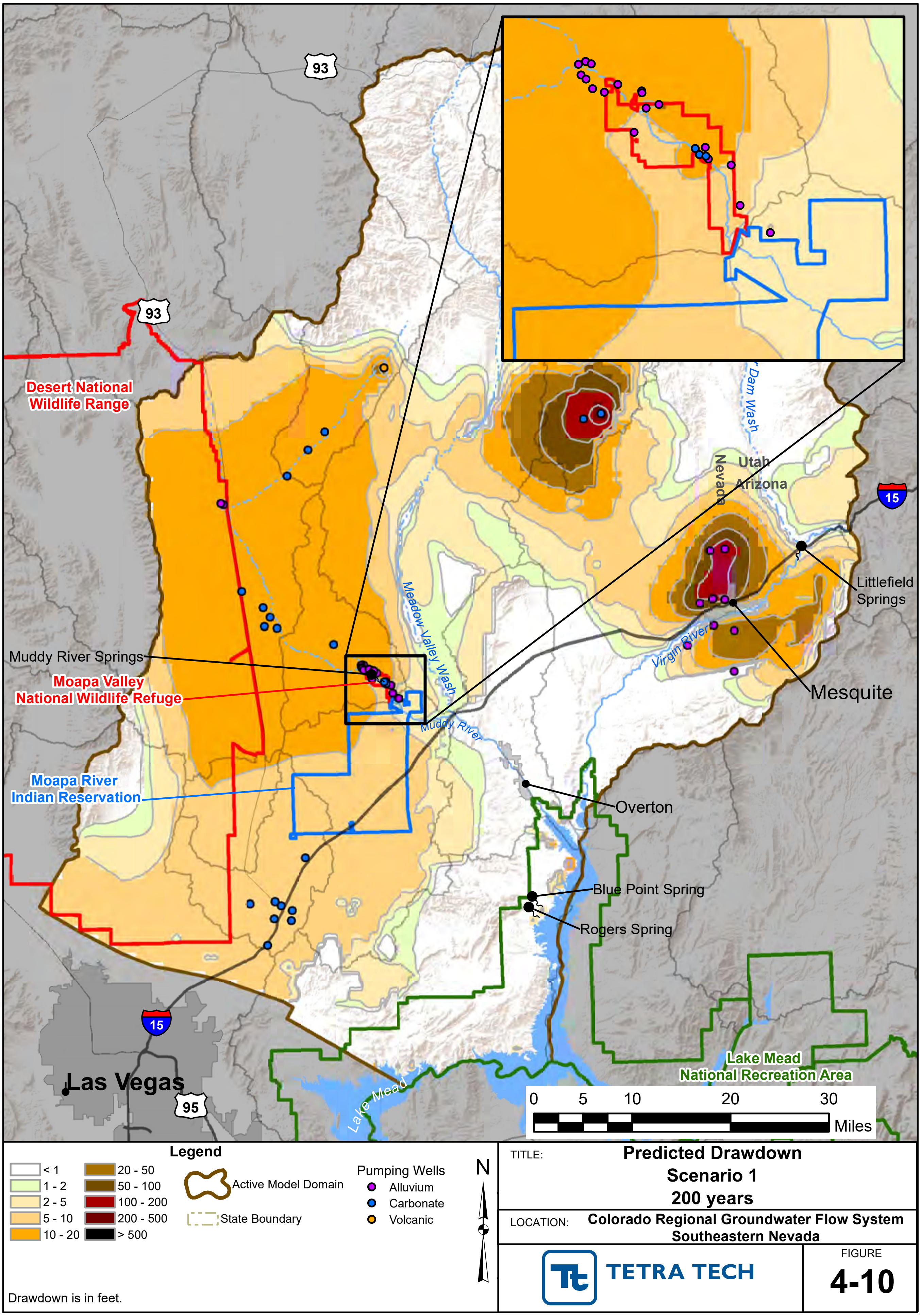
LOCATION: **Colorado Regional Groundwater Flow System**
Southeastern Nevada

TETRA TECH

FIGURE
4-9

SE ROA 51485

JA_16616



TITLE: **Predicted Drawdown Scenario 1 200 years**

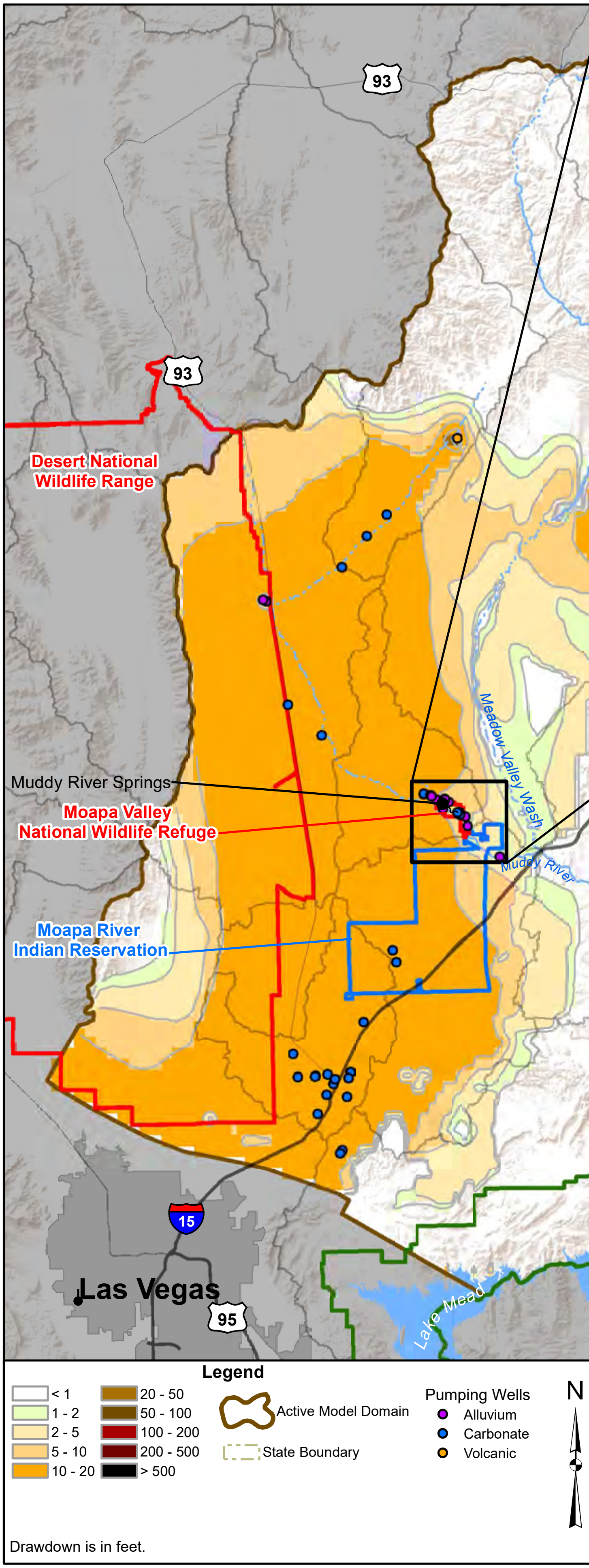
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-10**

SE ROA 51486

JA_16617



TITLE: **Predicted Drawdown**
Simulation 2
200 years

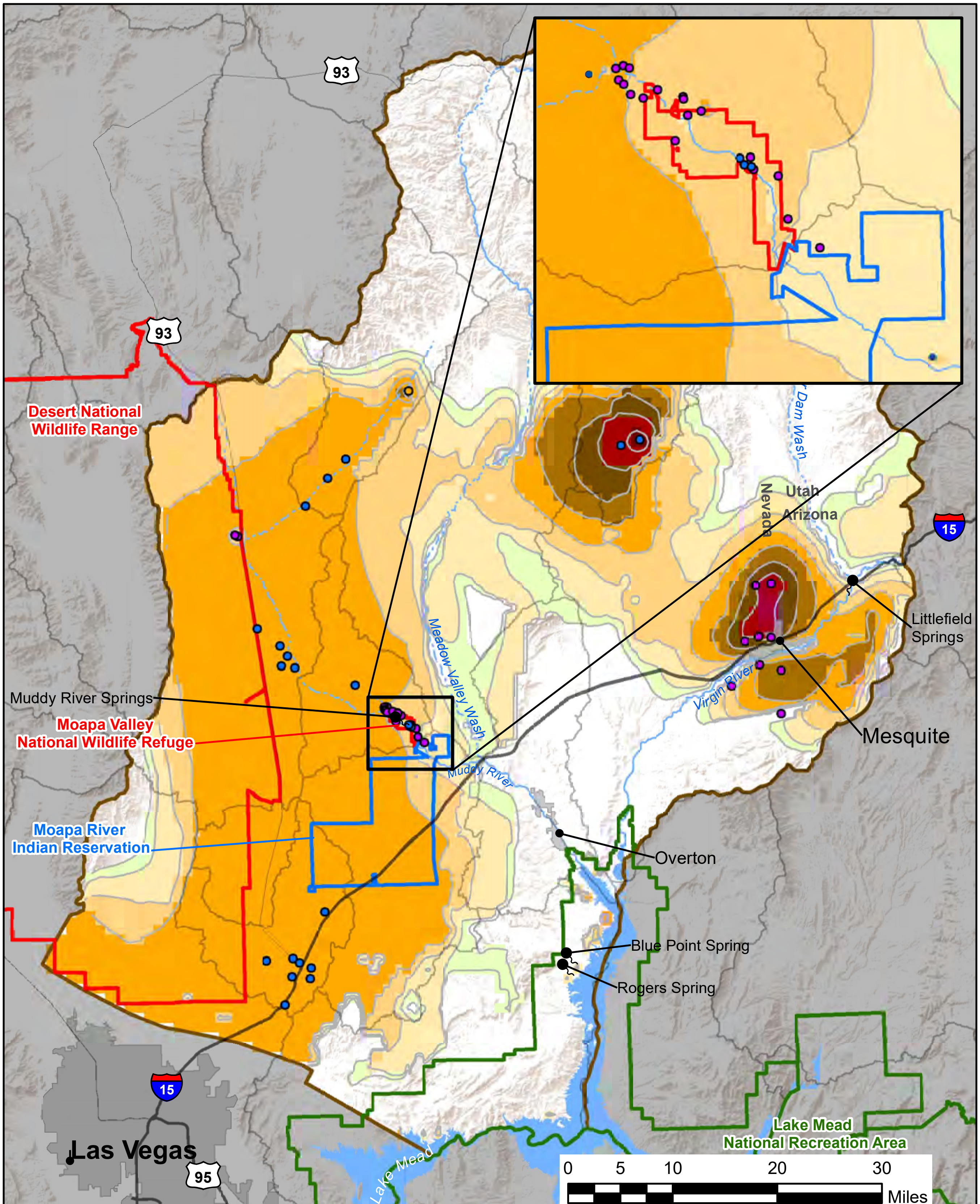
LOCATION: **Colorado Regional Groundwater Flow System**
Southeastern Nevada

TETRA TECH

FIGURE
4-11

SE ROA 51487

JA_16618



Legend

- < 1
- 1 - 2
- 2 - 5
- 5 - 10
- 10 - 20
- 20 - 50
- 50 - 100
- 100 - 200
- 200 - 500
- > 500

Active Model Domain

State Boundary

Pumping Wells
- Alluvium
- Carbonate
- Volcanic

N

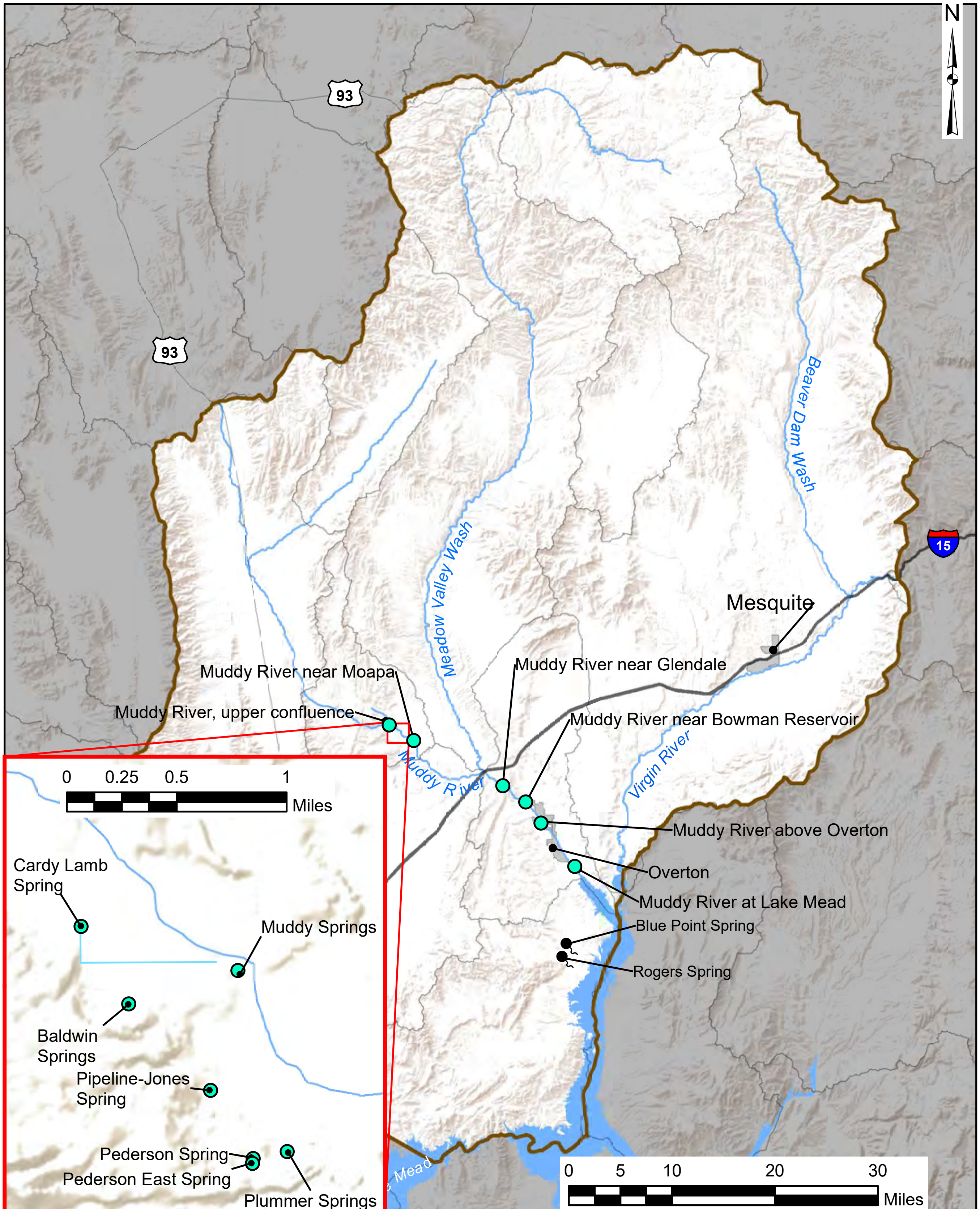
Drawdown is in feet.

TITLE: **Predicted Drawdown Scenario 3 200 years**

LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-12**



Legend

- Stream
- Active Model Domain
- State Boundary
- Hydrographic Region
- Stream or Spring Observation Point

TITLE: **Locations of Simulated Spring Discharge and Streamflow Measurements**

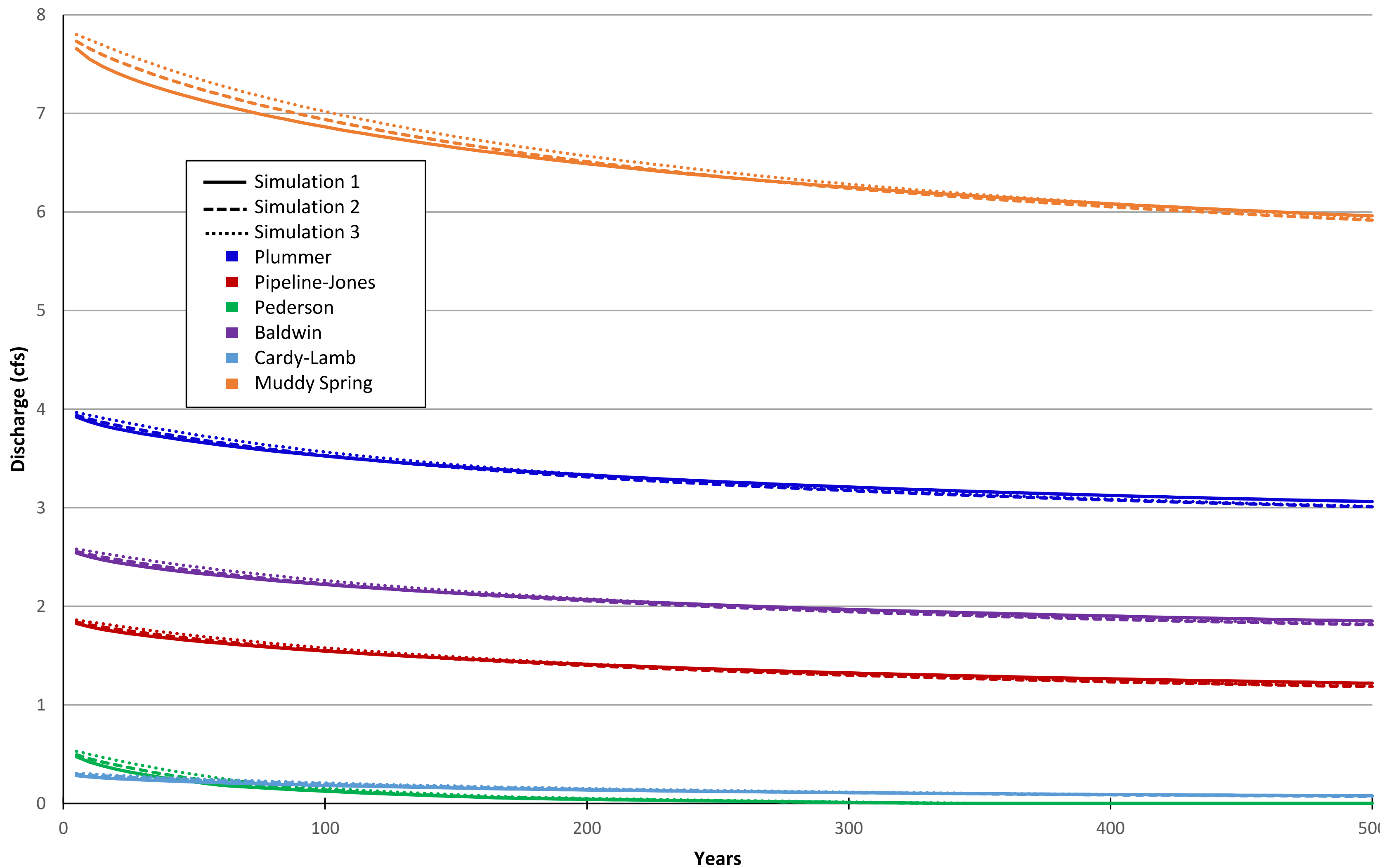
LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **4-13**

SE ROA 51489

JA_16620



TITLE: SIMULATED SPRING DISCHARGES IN THE MUDDY RIVER SPRINGS AREA			
LOCATION: Lower Colorado River Flow System			
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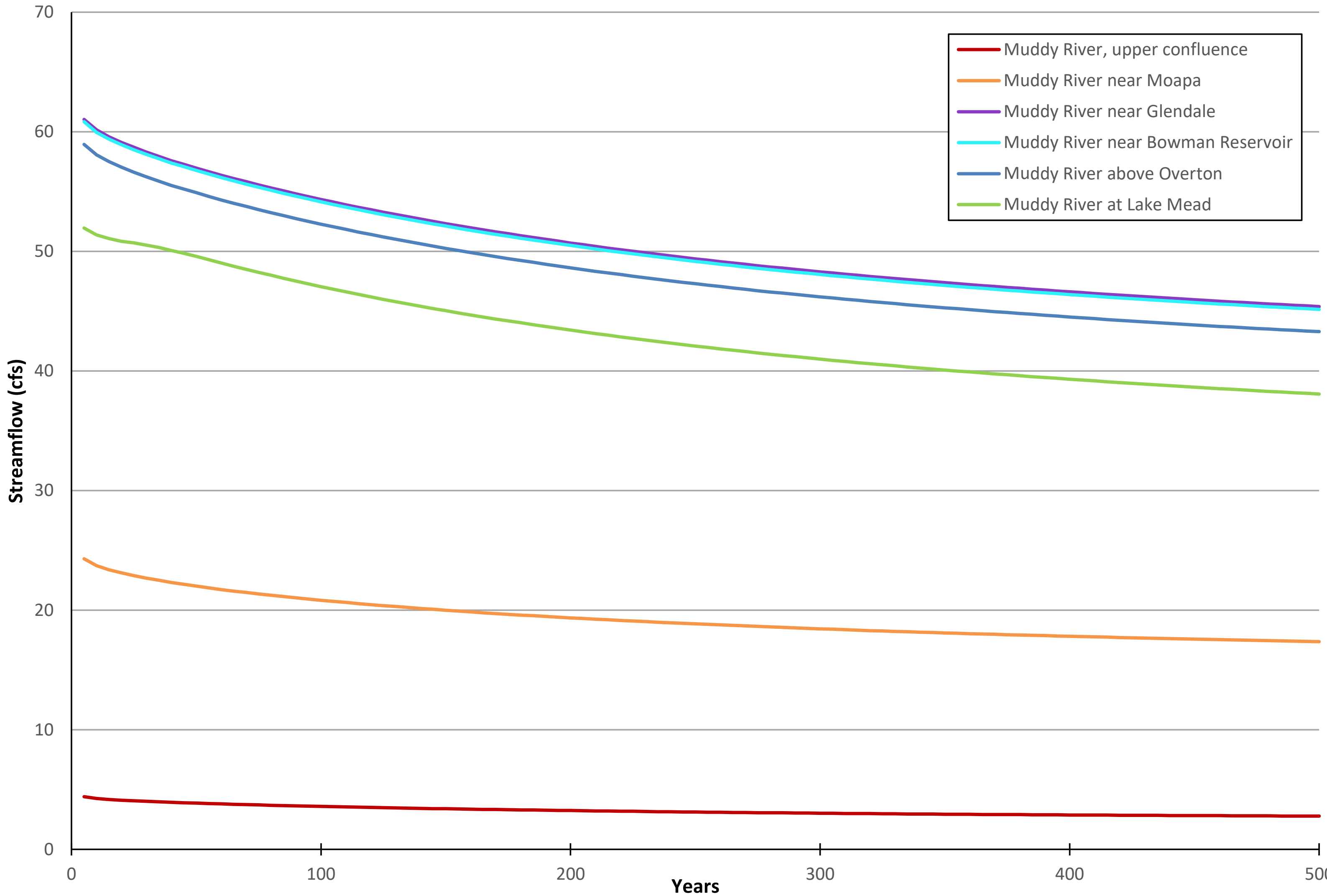
	DRAFTED	SRH	4-14
	PROJECT #	117-0524303	
	DATE	6/24/2019	

SE ROA 51490

JA_16621



TITLE: SIMULATED STREAMFLOW ALONG THE MUDDY RIVER, SIMULATION 1			
LOCATION: Lower Colorado River Flow System			
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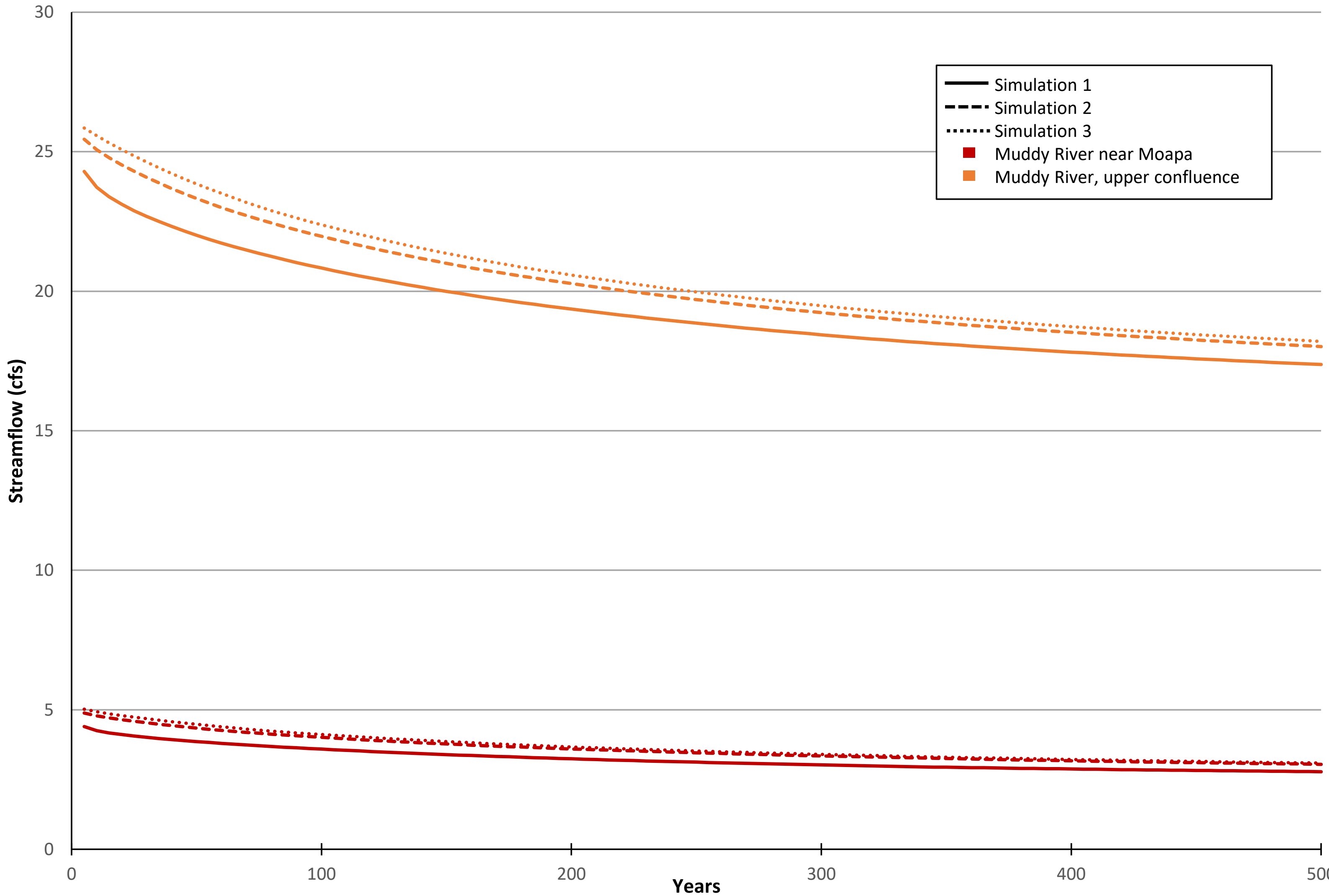
	PROJECT #	117-0524303	
	DATE	6/24/2019	

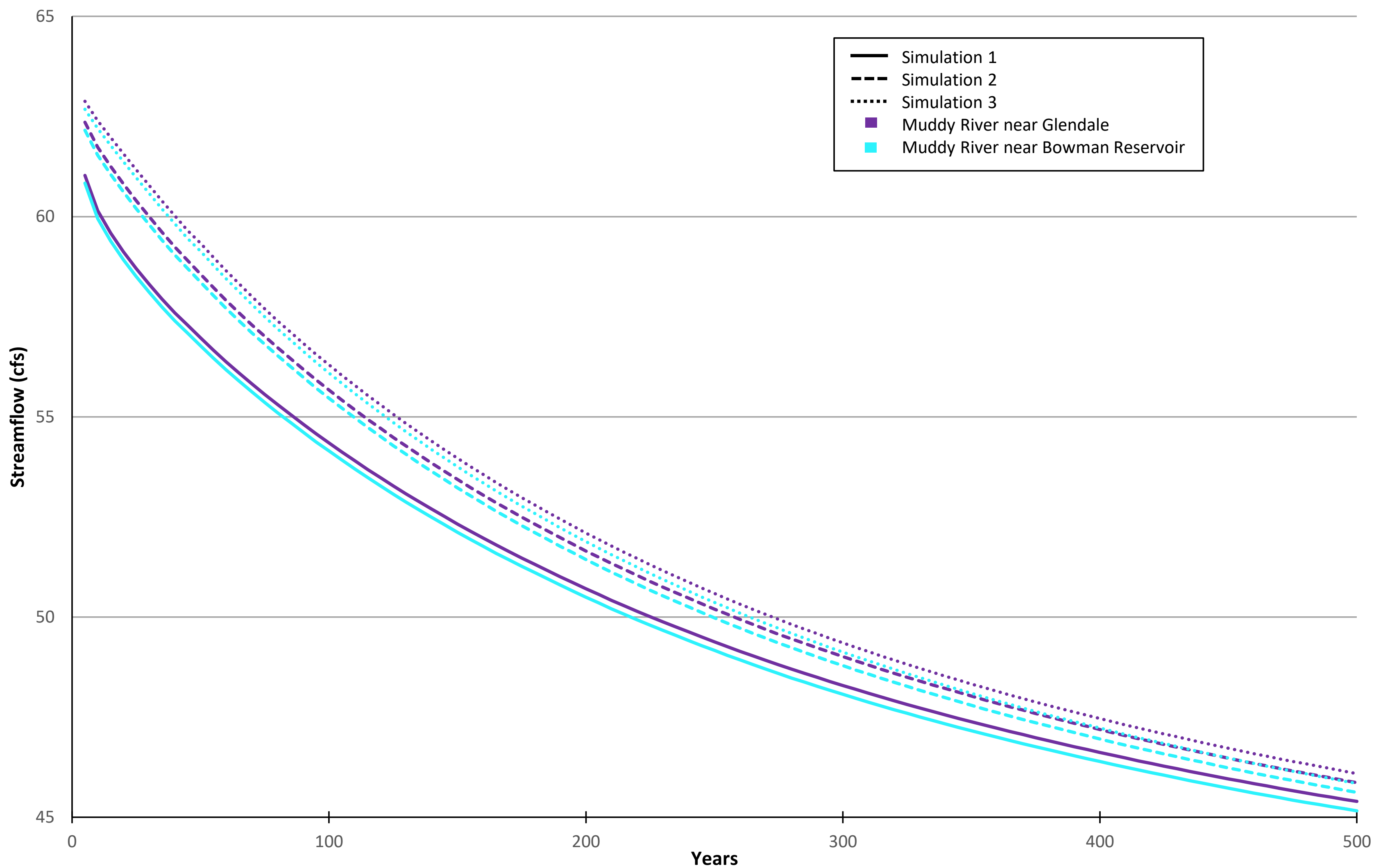


TITLE: SIMULATED STREAMFLOW OF THE MUDDY RIVER, UPPER CONFLUENCE AND NEAR MOAPA

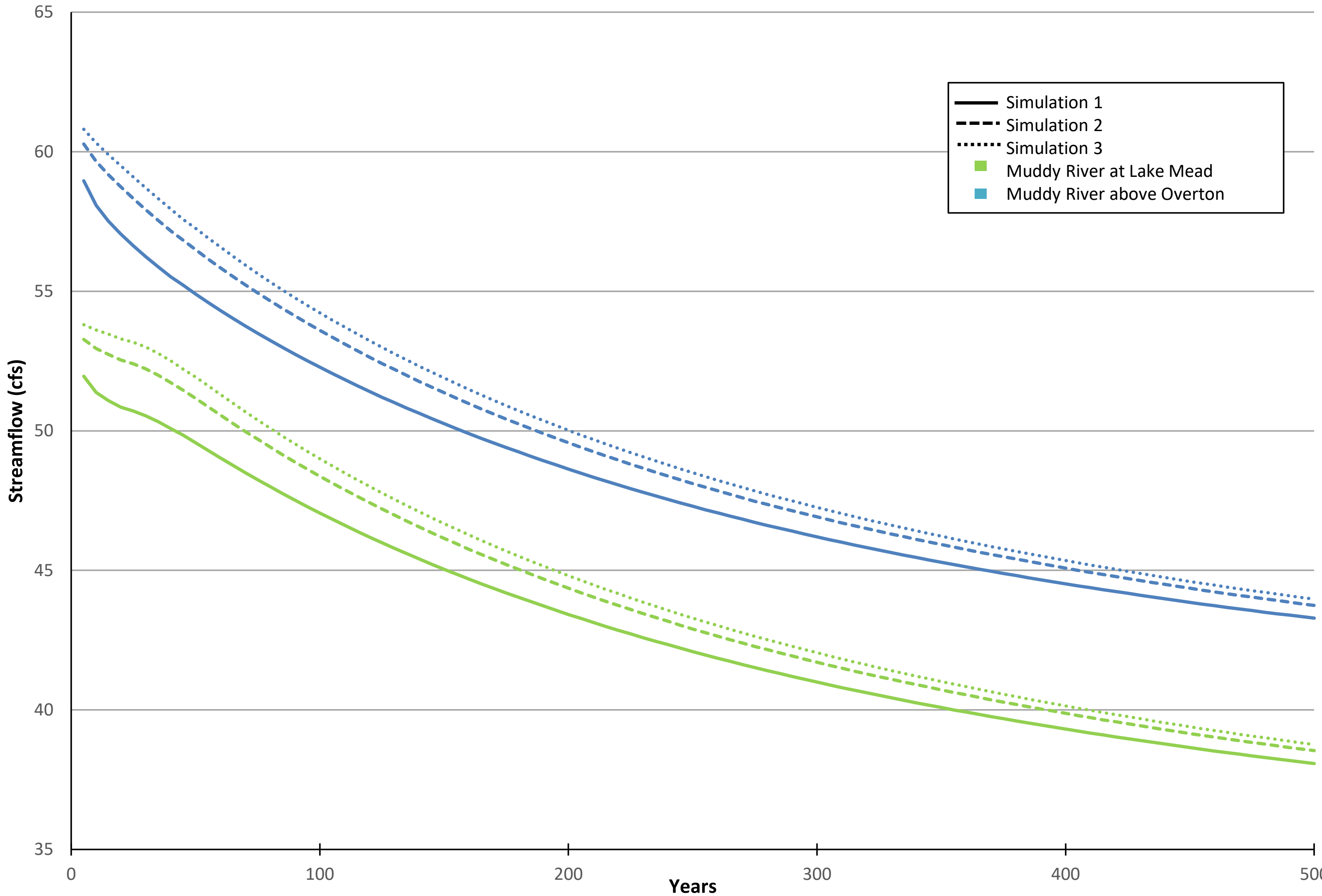
LOCATION: Lower Colorado River Flow System

TETRA TECH

APPROVED		FIGURE
DRAFTED	SRH	4-16
PROJECT #	117-0524303	
DATE	6/24/2019	



TITLE:	SIMULATED STREAMFLOW OF THE MUDDY RIVER, NEAR GLENDALE AND BOWMAN RESERVOIR		
LOCATION:	Lower Colorado River Flow System		
TETRA TECH	APPROVED		FIGURE
	DRAFTED	SRH	4-17
	PROJECT #	117-0524303	
	DATE	6/24/2019	



TITLE:	SIMULATED STREAMFLOW OF THE MUDDY RIVER, ABOVE OVERTON AND AT LAKE MEAD
LOCATION:	Lower Colorado River Flow System

TETRA TECH

APPROVED		FIGURE
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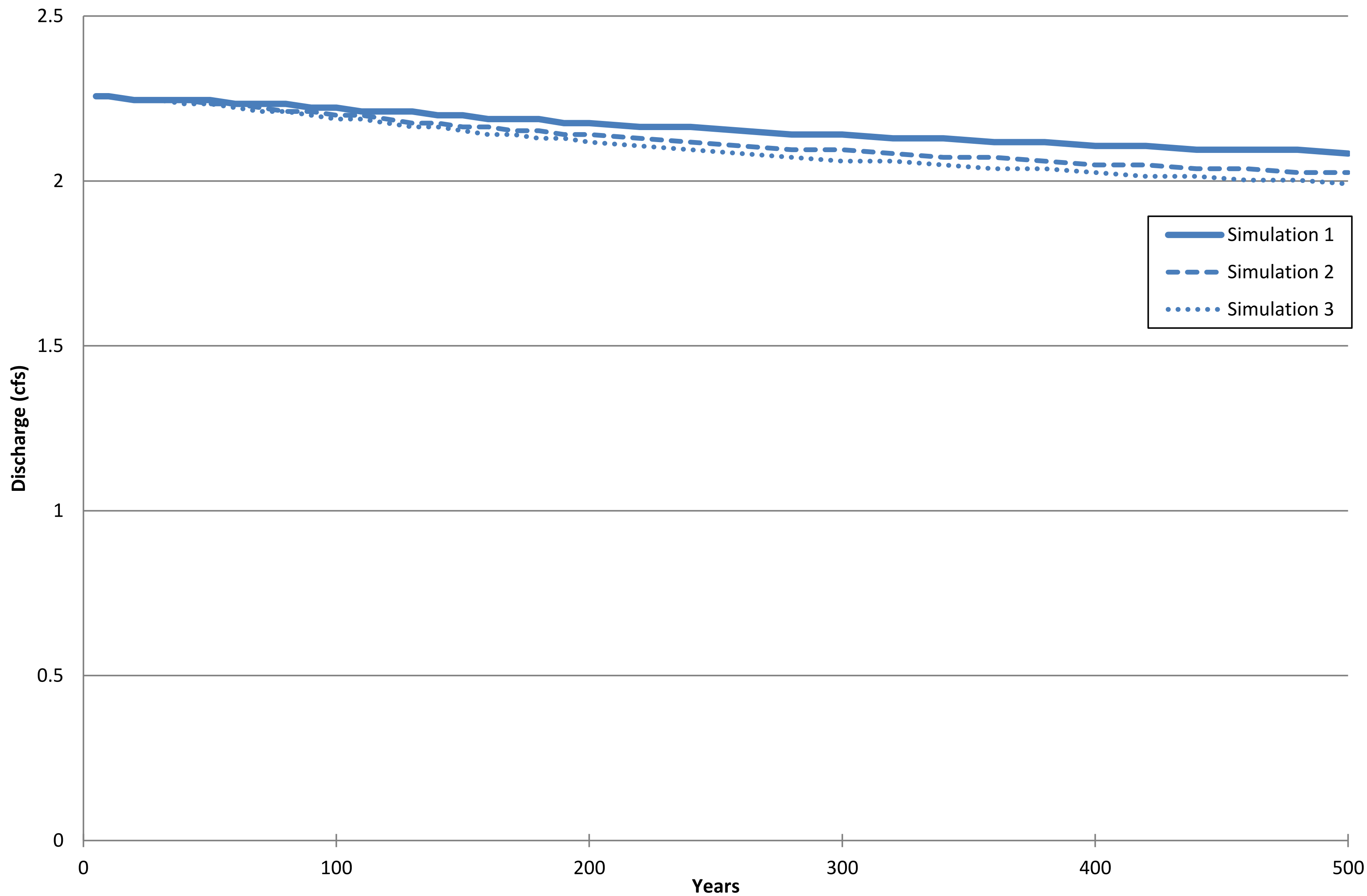
PROJECT #	117-0524303	
DATE	6/24/2019	



TITLE: SIMULATED COMBINED DISCHARGES FROM ROGERS AND BLUE POINT SPRINGS			
LOCATION: Lower Colorado River Flow System			
TETRA TECH	APPROVED		FIGURE
	DRAFTED	SRH	4-19
	PROJECT #	117-0524303	
	DATE	6/24/2019	

Appendix A

SE ROA 51496

Preliminary Geochemical Evaluation of Sources of Water Discharging at Rogers and Blue Point Springs, Southeastern Nevada

Prepared For

National Park Service

U.S. Fish & Wildlife Service

Bureau of Land Management

Prepared By

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SEPTEMBER 2012

SE ROA 51497

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Preliminary Geochemical Evaluation of Sources of Water Discharging at Rogers and Blue Point Springs, Southeastern Nevada

Abstract

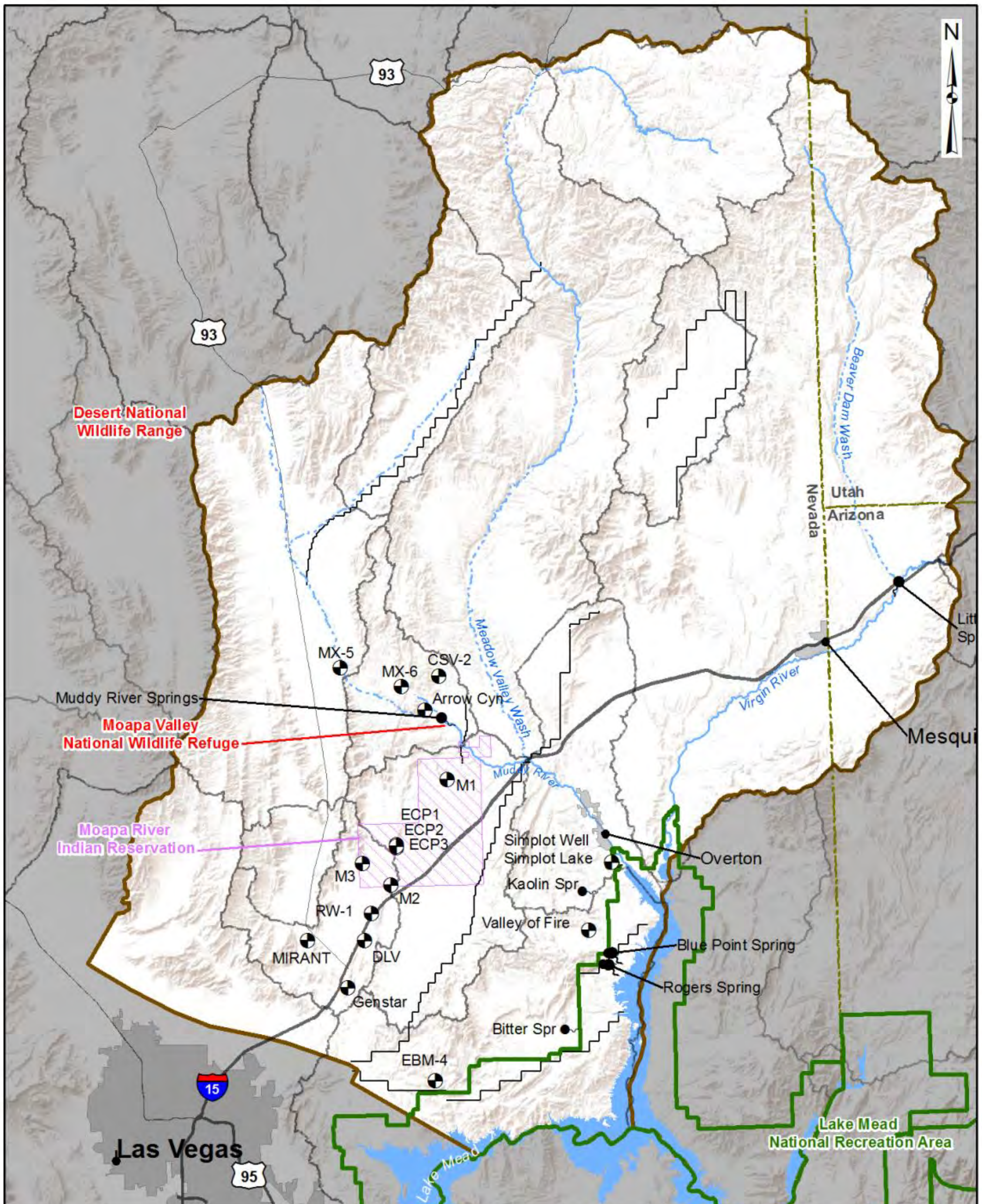
Groundwater resources in southeastern Nevada are in high demand, and competing water needs require advanced understanding of the hydrogeology. In this study the plausible interconnections between the regional Paleozoic carbonate aquifer and fault-related spring discharge at Rogers and Blue Point Springs near Lake Mead are investigated. Chemical and isotopic (δD, $\delta^{18}O$, $\delta^{11}B$, $\delta^{13}C$, $\delta^{34}S$, ^{3}H, ^{14}C, ^{36}Cl) analyses and geochemical reaction path modeling through the complex structural setting adjacent to Lake Mead were combined to determine most likely pathways of flow. Geochemical modeling indicated that the areas around the Muddy River Springs and in Garnet Valley could be along the flow path to Rogers and Blue Point Springs, but the path through Garnet Valley provided results more consistent with the geochemical data. The evolution of groundwater composition between California Wash and the springs near Lake Mead requires the input of water from the carbonate aquifer which reacts with Mesozoic and/or Tertiary sediments along the flow path, but is also dependent on recharge from the Muddy Mountains. The capture zone of Rogers and Blue Point Springs was then evaluated using a recently developed flow model of the area. Geochemical modeling indicates that this flow path is consistent with the geochemical data. The results indicate that groundwater is likely flowing from Garnet Valley, beneath California Wash and the Black Mountain Area Hydrographic Basins to discharge at Rogers and Blue Point Springs.

1.0 INTRODUCTION

In the arid southwest, the decision on uses of groundwater involve political, legal, engineering, and scientific issues, and integration of these disparate disciplines is needed to resolve conflicts of competing use for this important water resource. Although most regions of the southwest could serve as examples for the problem of competing water use, southeastern Nevada has hydrogeologic circumstances that make it particularly instructive.

Several assessments of the Great Basin Paleozoic carbonate aquifers in Nevada, Utah and Arizona had been performed, even before the USGS conducted the Regional Aquifer Systems Analysis (RASA) in 1995-96; these studies as well as compilations of baseline data for hydrogeology and geochemistry are well documented and are not reviewed here (Dettinger et al, 1995; Prudic at al., 1995; Thomas et at., 1996; Winograd and Pearson, 1976; Winograd and Friedman, 1972). Although the Great Basin extends over more than 140,000 mi^2, the study area for this project is a small subset of this region. The project area for this geochemical evaluation is within the downgradient part of the Colorado Regional Groundwater Flow System (CRFS), and encompasses the Muddy River Springs on the north edge, the Arrow Canyon Range on the west, the Muddy Mountains in the south central area, and the northwest shores of Lake Mead in the southeastern corner (Fig. 1). The hydrogeology and water resources of the project area have been evaluated in several reports because of the importance of springs and seeps in and near the Lake Mead National Recreation Area (Laney and Bales, 1996; Pohlman et al., 1988).

This investigation focuses on the portion of the hydrologic system that terminates in the area of two key springs, Rogers and Blue Point, which discharge near Lake Mead (Figure 1.)



0 5 10 20 30 Miles

TITLE: **Well and Spring Sampling Locations Used in Geochemical Analysis**

LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **1**

Culturally and environmentally, Rogers and Blue Point Springs have historical significance and are part of the Lake Mead National Recreation Area. Several springs in the immediate area are part of a linear trend of springs and seeps along the Rogers Springs Fault system. As groundwater development proceeds in this region, the discharges from these springs may be depleted if the natural spring discharge is derived from the regional aquifer system, rather than local recharge.

The specific objectives of the study are to:

1. Present the results of geochemical sampling of several wells and springs within a portion of the Colorado River Flow System;

2. Briefly discuss the results of preliminary geochemical reaction path modeling performed by Geochemical Technologies Corporation (GTC); and

3. Present an evaluation of whether a flow path analysis of Roger and Blue Point Spring source areas determined by the three-dimensional flow model presented in Tetra Tech (2012) is geochemically feasible.

This report relies on chemical and isotopic analyses from previously published reports as well as data from sixteen water samples obtained during this study from the locations indicated in Figure 1. New samples were collected to provide a wider range (spatially and elementally) of isotopic analyses, also giving the previously available forensic data a more constrained interpretation of the flow path, travel time, and mixing because of the complementary chemically conservative characteristics. The multiple isotopic analysis approach includes results for δD, $\delta^{18}O$, $\delta^{34}S$, $\delta^{11}B$, ^{3}H, $\delta^{13}C$, ^{14}C, and ^{36}Cl. The justification for the use of a broad range of analytical entities is that each provides different but potentially correlative information lending support to the interpretation process.

The forensic approach taken here relies on previous geologic and hydrologic investigations, a recent compilation of geologic mapping by Page and others (2005), interpretations of structural geology and cross sections through the region by the USGS (Page et al., 2011), and chemical and isotopic data for aquifers and spring samples (Johnson et al., 2001; Pohlmann et al., 1988; Thomas et al., 1996). Tetra Tech (2012) used the geologic information to construct a three-dimensional geologic model, as part of the development of a groundwater flow model of selected basins within the CRFS. The flow model is of a larger area than the study area for this geochemical evaluation. This flow model was used to estimate the “capture zone” for Rogers and Blue Point Springs.

This study proceeded through several steps:

1. Compilation of existing data on the geochemistry of groundwater sampled at springs and wells;

2. Collection of additional samples, to provide both chemical and isotopic information;

3. Evaluation of the subsurface geology to provide information on likely water/solid reactions;

4. Preliminary geochemical modeling to determine feasible sources of water discharging at Rogers and Blue Point Springs [the results from these models are summarized in Section 5.2.];

5. Analysis of the capture zone of Rogers and Blue Point Springs using the recently developed flow model by Tetra Tech (2012) to determine flow paths to the springs; and

6. Additional geochemical modeling along the flow path developed from the flow model to determine if the geochemistry data are consistent with this simulated flow path.

2.0 METHODS

2.1 SAMPLE COLLECTION

Personnel from Geochemical Technologies Corporation (GTC) collected nine new samples, and obtained three archived samples collected for earlier projects by Mifflin & Associates, Inc. (Johnson, et al., 2001). Four additional samples (CSV-2 well, Mirant well, MX-5 well and Blue Point Spring) were collected and analyzed cooperatively among the three organizations: GTC, the United States Geological Survey (USGS) and the Southern Nevada Water Authority (SNWA). The co-operatively obtained samples allowed these organizations to extend the coverage of data for locations that overlapped in three different independent projects by cost sharing some component of the field or analytical costs.

The selection of sampling sites was optimized based on location, access, installed pumps, and reported composition from previous investigations (Berger et al., 1988; Hershey and Mizell, 1995; Johnson et al., 2001; Laney and Bales, 1996; Pohlmann et al., 1998; Thomas et al., 1991, 2001). Prior studies provide data for most wells and all the springs, but earlier results did not include isotopic analyses needed for this study. Historical data establish a baseline, indicate change, provide confirmation of original analyses, and fill data gaps. Collection of samples was completed in three field excursions over a fourteen-month period from June 2003 to January 2004. Samples from four locations were collected by cooperating agencies for isotopic analyses; the Mirant well and the MX-5 well by SNWA personnel, and Blue Point Spring and the CSV-2 well by USGS personnel.

Wells in production did not require purging prior to sampling; the remainder were purged for at least three well bore volumes. Field activities included acquiring hydraulic and hydrogeologic data about the wells from the owner agency or company, measuring field parameters (temperature and pH) and collecting, filtering, or preserving samples. All procedures were performed according to standard accepted professional practices. Filtered samples were passed through 0.45 micrometer effective pore diameter material. Samples for metals were acidified to pH less than 2. All samples were collected in plastic bottles except for the isotopic samples for hydrogen, oxygen, and radiocarbon. Spring samples were collected at the orifice where possible to minimize atmospheric exposure.

2.2 ANALYTICAL METHODS - CHEMICAL DATA

Chemical analyses for samples collected by GTC were performed by Evergreen Analytical, Inc. in Wheat Ridge, CO; samples collected by SNWA and the USGS were analyzed in the USGS laboratories. Relevant EPA or USGS methods were used and noted on the analytical result forms.

2.3 ANALYTICAL METHODS - ISOTOPIC DATA

All isotopic measurements were performed according to published professionally accepted procedures. Hydrogen, oxygen, and sulfur were measured on a Finnigan Delta mass spectrometer; expected precision of each to one standard deviation are δD (0.9 ‰), $\delta^{18}O$ (0.08 ‰), and $\delta^{34}S$ (0.2 ‰). Boron ($\delta^{11}B$) with a precision of 0.5 ‰ was measured on a VG Thermal Ionization Mass Spectrometer (TIMS) under the direction of GTC in the Isotope Laboratory of the Department of Hydrology and Water Resources, University of Arizona.

The analyses for enriched tritium (^{3}H) were done by beta counting with a detection limit of approximately 0.5 tritium units (TU) in the Isotope Geochemistry Laboratory of the Department of Geosciences, University of Arizona. Radiocarbon analyses were measured in the Accelerator Mass Spectrometry (AMS) laboratory in the Physics Department, University of Arizona. The ^{36}Cl measurements were determined by AMS at PRIME laboratory, Purdue University.

3.0 ANALYTICAL RESULTS

The chemical and analytical results for samples collected in this study are given by location in Tables 1-3. Sodium chloride is probably derived from dissolution of playa lake halite in the lacustrine evaporites of the Tertiary sediments, the presence of which is well documented, and was confirmed by coring operations conducted by Stauffer Chemical Company (Laney and Bales, 1996). The most likely source of the elevated boron in spring waters along the Rogers Spring Fault is from these halite-rich sediments; in the absence of any samples of halite from the study area, core samples were obtained from the Arizona Geological Survey. The core samples are from the Detrital Valley within the region of Cenozoic evaporite deposits that includes the Tertiary sediments of the Muddy Creek Formation adjacent to Lake Mead. Detrital Valley is south of the study area about 20 miles and Grand Wash is 25 miles east.

Table 1. Summary of Well Data and Field Measurements.

Site Name	Surface Elevation (ft)	Well Depth (ft)	Screen Interval (ft)	Water Level Depth (ft)	Date Measured	pH	T °C	Date Measured
Wells								
Arrow Canyon	1859	565	205-565	45	2/8/91[1]		32.1	1/20/04
CSV-2[2]	2186	478	Open hole	392	10/27/85[3]	7.20[8]	28.6[8]	7/8/03
DLV		575				7.30	29	7/1/85
EBM-4[7]	2391	1129	608-1129	593	2/5/92		30.0	1/20/04
ECP-1	2230	1170	600-1051	416	7/30/00[4]	7.5	30.5	7/27/00
ECP-2	2229	1228	Open hole	416	12/4/00[4]	8.0	29.7	12/7/00
ECP-3	2242	1500	Open Hole	429	10/31/00[4]			
G. P. Apex		1205				7.00	31	9/30/86
Genstar		500				7.40	24.0	3/31/86
M1	1896	403	358-398	80	10/10/00[4]	8.10[4]	29.2[4]	10/11/00
M2	2109	683	640-680	297	10/18/00[4]	8.10[4]	29.0[4]	10/17/00
M3	2235	673	633-673	422	10/24/00[4]	8.10[4]	27.8[4]	10/21/00
Mirant 1	2566	1979	1197-1979	755	3/1/02[1]	7.19[1]	27.3[1]	6/4/03
MX-5	2169	628	Open hole	352	5/6/81[3]	7.30[8]	35.5[8]	5/28/03
MX-6	2275	937	Open hole	458	6/3/81[3]	7.20	33.5	9/28/86
RW-1	2069	833	553-833	260	7/3/01[1]	7.75[9]	30.0	7/2/01
Simplot	1640	820	420-820	122	12/18/02[1]	7.35	25.6	6/30/03
Valley of Fire	2240	1140	780-1140	570	1/15/85[1]	7.56	28.1	6/30/03
Surface Water Sites								
Bitter Spring	1660		-		10/3/95[5]	7.85	14.9	1/21/04
Blue Point Spring	1542	-	-		10/4/95[5]	6.90[8]	30.8[8]	6/5/03
Kaolin Spring	1440	-	-		10/4/95[5]	7.95	12.1	1/21/04
Rogers Spring	1601	-	-		10/3/95[5]	7.00	30.3	3/30/03
Simplot Lake	1810	-	-		6/30/03[6]	8.47	26.9	6/30/03

Notes:
(1) Nevada Div. of Water Resources.
(2) Written Communications, Moapa Valley Water District (2004)
(3) Berger et al. (1988).
(4) Johnson et al. (2001).
(5) Pohlmann et al. (1988).
(6) Simplot Co. (2003).
(7) Nevada Cogeneration (2004)
(8) USGS
(9) SRK Consulting (2001)

Table 2. Summary of General Chemistry.

Site Name	TDS (mg/L)	Na (mg/L)	K (mg/L)	Ca (mg/L)	Mg (mg/L)	Cl (mg/L)	Alk(1 (mg/L)	SO_4 (mg/L)	SiO_2 (mg/L)	B (μg/L)
Wells										
Arrow Canyon	568	98	12	60	26	54.9	213	158	32.1	310
CSV-2(	591	101	11	61	26	62	215	158	31.6	293
DLV		120	13	110	48	170	210	360	21	
EBM-4	1040	130	15	110	56	184	158	404	32.1	380
ECP-1	754	110	14	110	46	110	160	270	10	
ECP-2	750	93	14	110	44	120	180	260	20	
ECP-3	742	99	13	100	45	120	180	260	11.0	320
G. P. Apex		130	13	120	47	200	230	380		
Genstar		140	1.3	120	47	180	230	370	23	
M1	636	110	13	94	41	74	240	220	23	
M2	817	110	15	110	50	140	170	290	17.0	
M3	779	79	14	130	49	100	190	300	14.0	
Mirant 1	932	79	10	79	30	148	184	337	17.4	300
MX-5	476	84	13	49	21	36	241	93	35.7	318
MX-6		87	10	58	25	53	271	160	30	318
RW-1	895	100	14	99	54	157	159	316	14.3	320
Simplot	754	89	15	95	30	88	144	338	10.9	454
Valley of Fire	464	24	5	69	33	22	125	229	11.6	180
Surface Water										
Bitter Spring	3980	250	19	570	180	171	99	2390	25.7	1400
Blue Point Sp.	3680	353	26	490	162	374	130	1910	17.7	1390
Kaolin Spring	293	39	21	34	16	15	172	53	15.0	830
Rogers Spring	3250	280	21	430	130	358	132	1910	17.3	1020
Simplot Lake	1100	180	20	79	49	144	122	537	14.1	901

Notes:
(1) Total Alkalinity as mg $CaCO_3$/L

Table 3. Summary of Isotopic Composition.

Site Name	Sample Date	δD (‰)	$\delta^{18}O$ (‰)	$\delta^{13}C$ (‰)	$\delta^{11}B$ (‰)	$\delta^{34}S$ (‰)	^{14}C (pmc)	^{3}H (TU)	$^{36}Cl/Cl x 10^{15}$
Wells									
Arrow Canyon	1/20/04	-97	-12.9		6.6(1)	13.7			
CSV-2	7/8/03	-97	-12.7		5.4	14.1			107
DLV									
EBM	1/20/04	-98	-13.2		-2.7	15.1			23
ECP-1	7/27/00	-97	-12.9						
ECP-2	12/7/00	-98	-13.4						
ECP-3	10/31/00	-97	-12.7		-2.7	15.3			6
G. P. Apex									
Genstar									
M1	10/1/00	-95	-12.5						
M2	10/17/00	-98	-13.3						
M3	10/21/00	-95	-13.0						
Mirant 1	6/4/03	-98	-13.2	-3.97	25	18.5	2.71	<0.5	25
MX-5	4/8/03	-99	-12.9	-5.4	4.7	13.9	9.77	<0.5	390
MX-6	9/28/86	-97	-13.0	-8.0	4.4		8.4	0.63	
RW-1	1/20/04	-99	-13.4			24.4			43
Simplot	6/30/03	-83	-10.2	-8.8	-19.6	12.4		0.7	203
Valley of Fire	6/30/03	-79	-11.0	-7.8	-4.8	11.0	23.88	<0.9	276
Surface Water									
Bitter Spring	1/21/04	-76	-10.2	-4.3(2)	-13.1	13.7			47
Blue Point Sp.	6/30/03	-93	-12.4	-2.0	-14.2	13.1	3.30	<0.5	25
Kaolin Spring	1/21/04	-90	-11.7	-6.5(3)	-42.8	11.0			250
Rogers Spring	6/30/03	-92	-12.4	-3.9(4)	-11.6	12.8		<0.6	28
Simplot Lake	6/30/03	-75	-7.8		0.1	13.4		0.6	181

Notes:
(1) Sample collected 7/2/03.
(2) Sample collected 2/6/96
(3) Sample collected 2/9/96
(4) Sample collected 3/19/92.

4.0 HYDROGEOLOGIC CONCEPTUAL MODEL

Rogers and Blue Point Springs are located along the upthrown side of the Rogers Spring Fault on the north side of Lake Mead. The springs discharge from Paleozoic carbonate rocks that are in the upper plate of the Muddy Mountain thrust fault. This upper thrust sheet structurally overlies Mesozoic rocks, which themselves depositionally overlie a deeper buried section of carbonate rocks. These deeper carbonate rocks extend to the west beneath California Wash and to the north beneath the Muddy River (Page et al., 2011).

Potential sources of water at these springs include:

1. Lateral flow in the deeper carbonate aquifer to the Rogers Spring Fault and upward flow along the fault to the springs. As water levels in the deeper carbonate rocks are unknown, flow could be (a) from the west beneath California Wash and further upgradient areas, (b) from the vicinity of the Muddy River Springs and Coyote Spring Valley, and (c) from the area of the Morman Mountains, which would involve flow beneath Lower Moapa Valley, the Muddy River, and the Muddy Mountains;

2. Lateral movement from saturated basin-fill sediments in California Wash through Mesozoic rocks and through the upper plate carbonate rocks in the Muddy Mountains; and

3. Local recharge in the Muddy Mountains that could flow through upper plate carbonate rocks and the underlying Mesozoic rocks.

In addition, mixing of water from these potential sources may occur.

The available hydrologic data are insufficient to determine the source of the water discharging from the springs. The elevations (NGVD 29 datum) of the spring orifices are approximately 1576 feet (Rogers Spring) and 1562 feet (Blue Point Spring). In order for groundwater tom move from an area to the springs, the hydraulic head in that area must be greater than the spring elevations. This is a necessary, but not sufficient, condition. Hydraulic heads in California Wash, Coyote Spring Valley, and the Muddy River Springs area are higher than 1800 feet, so water could potentially flow from these areas to Rogers and Blue Point Springs within the lower plate carbonate rocks (Sources 1a and 1b), if the head in the lower plate carbonate rocks beneath the springs is also greater than the spring elevation. Water from basin-fill material in California Wash (Source 2) could also flow to the springs. In order for water to flow from the northeast side of the Muddy River (for example, from the area of the Morman Mountains, Source 1c), the hydraulic head in the carbonate rocks beneath the Muddy River along this potential flow path would need to be greater than 1576 feet. The hydraulic head in shallow rocks near the Muddy River in this area ranges from 1400 to 1500 feet, but hydraulic heads at greater depths are unknown. They could be greater than 1576 feet. Thus, a source near the Morman Mountains cannot be ruled out on the basis of the available water-level data.

The combined discharge of Rogers and Blue Point Springs is approximately 1,600 acre feet per year (af/y). In contrast, the historic discharge in the Muddy River at Moapa was approximately 30,000 af/y prior to significant pumping of groundwater near the Muddy River Springs. There is no evidence of significant groundwater discharge into the area occupied by Lake Mead from this groundwater system, either before or after the filling of Lake Mead. Thus, nearly all of the flow occurring the carbonate rocks discharges at the Muddy River Springs area, and the discharge at Rogers and Blue Point Springs is a relatively small component of the water budget. It is therefore unlikely that there is significant convergence of flow from different areas (for example, from the west, north, and northeast) to support the discharge at Rogers and Blue Point Springs. A later section of this report will provide a depiction of the source areas of the springs as estimated from a groundwater flow model (Tetra Tech, 2012) which supports this statement.

5.0 DISCUSSION

5.1 CHEMICAL AND ISOTOPIC COMPOSITIONS

Table 4 provides the chemical compositions of the waters, grouped by area. Water issuing from Rogers and Blue Point Springs has a $Ca\text{-}SO_4$ dominated, high TDS (3,250 and 3,680 mg/L), water composition, significantly different from that of the $Ca\text{-}HCO_3$, low TDS (~800 mg/L), carbonate aquifer to the north and to the west. This significant increase in dissolved solids content could be attributed to evaporite dissolution in the Mesozoic section, in Tertiary volcanic rocks, or in the Tertiary basin-fill sediments (Hershey and Mizell, 1995; Laney and Bales, 1996).

Table 4. Major ion composition by area.

Site Name	TDS	Na	K	Ca	Mg	Cl	ALK(1)	SO_4	SiO_2	B
	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	mg/L	µg/L
Coyote Spring Valley and Muddy River Springs area										
Arrow Canyon										336
Arrow Canyon	568	98	12	60	26	54.9	213	158	32.1	310
CSV 2										310
CSV 2(2)	591	101	11	61	26	62	215	158	31.6	293
MX 5										370
MX 5(3)	448	83	13	45	20	29	245	91	26.4	
MX 5(2)	476	84	13	49	21	36	241	93	35.7	318
MX 6		87	10	58	25	53	271	160	30	318
Mean	**521**	**91**	**12**	**55**	**24**	**47**	**237**	**132**	**31**	**320**
California Wash										
ECP 1	754	110	14	110	46	110	160	270	10	
ECP 2	750	93	14	110	44	120	180	260	20	
ECP 3										
ECP 3(4)	742	99	13	100	45	120	180	260	11.0	320
M1										
M1(4)	636	110	13	94	41	74	240	220	23.0	
M2										
M2(4)	817	110	15	110	50	140	170	290	17.0	
M3										
M3(4)	779	79	14	130	49	100	190	300	14.0	
Mean	**746**	**100**	**14**	**109**	**46**	**111**	**187**	**267**	**16**	**320**

Garnet Valley										
Mirant No. 1										300
Mirant No. 1[3]	932	79	10	79	30	148	184	337	17.4	
Mirant No. 1[2]	984	106	13	111	50	154	179	329	18.8	286
RW 1	895	100	14	99	54	157	159	316	14.3	320
RW 1[5]	836	120	14	100	49	170	170	350		
Mean	**912**	**101**	**13**	**97**	**46**	**157**	**173**	**333**	**17**	**302**
Valley of Fire	464	24	5	69	33	22	125	229	11.6	180
Blue Point[2]	3680	353	26	490	162	374	130	1910	17.7	1390

Notes:

(1) Alkalinity as $CaCO_3$

(2) Analytical results provided by the USGS

(3) Analytical results provided by SNWA

(4) Johnson et al. (2001)

(5) SRK Consulting, Inc. (2001)

(6) Pohlmann et al. (1998)

Tritium (^{3}H). The tritium content in groundwater has historically been most useful in identifying groundwater that contains a fraction of water, which was recently recharged. Nuclear testing in the late 1950s and early 1960s contributed large concentrations of tritium into the atmosphere; water recharged over the subsequent decades contained elevated tritium and thus was a label for "modern" or “post-bomb” recently recharged water. The atmosphere has almost returned to pre-nuclear testing or "pre-bomb" levels of tritium or <10 tritium units (TU), and since the half-life of tritium is 12.3 years, and detection limits are slightly below about 10% of that value or <1TU, it is progressively less and less useful. All of the samples of well water and even the spring water are so close to the detection limit that it does not allow for much interpretation other than to say the recharge was probably more than 50 to 60 years ago.

Hydrogen and Oxygen Environmental isotopes (δD, $\delta^{18}O$). In contrast to the ^{3}H values, the δD, $\delta^{18}O$, and ^{36}Cl data provide useful information. The δD and $\delta^{18}O$ are commonly used to indicate whether waters in a given region have been recharged from different elevations, climates, or locations. These isotopes are usually plotted together because they are both components of the same water molecule and it was discovered that the meteoric water tested around the globe when cross-plotted had a linear relationship. This is because the processes of evaporation and condensation cause a definable and consistent change in the isotopic values. Meteoric water values generally exhibit a linear relationship known as the Global Meteoric Water Line (GMWL) with local variation due to climate (See Clark and Fritz, 1997, for detailed discussion of the processes involved). Although the GMWL is defined for meteoric waters, e.g. rainfall, snow, etc., it has been well established that the environmental isotopes for wells and springs in carbonate aquifers of southeastern

Nevada consistently plot along the GMWL (Hershey and Mizell, 1995; Thomas et al., 2001; Thomas et al., 1996). This alignment with the GMWL indicates that recharge of meteoric water occurs from many different elevations, and from storms in different seasons, but evaporation prior to recharge is in general not reflected in the data. In the study area for this report, the same observation can be made; the samples collected from the carbonate aquifers plot along the GMWL (Figure 2). Note also that the data for Kaolin Springs and Bitter Springs plot off of the GMWL, as do the samples collected from the Simplot lake and adjacent well from the Simplot Industries property. Rogers and Blue Point Springs are not on the GMWL but are clearly as close to the line as the mean value of the carbonate aquifer samples but further up the line toward more enriched values. The Valley of Fire well plots on the GMWL but represents an even more enriched sample.

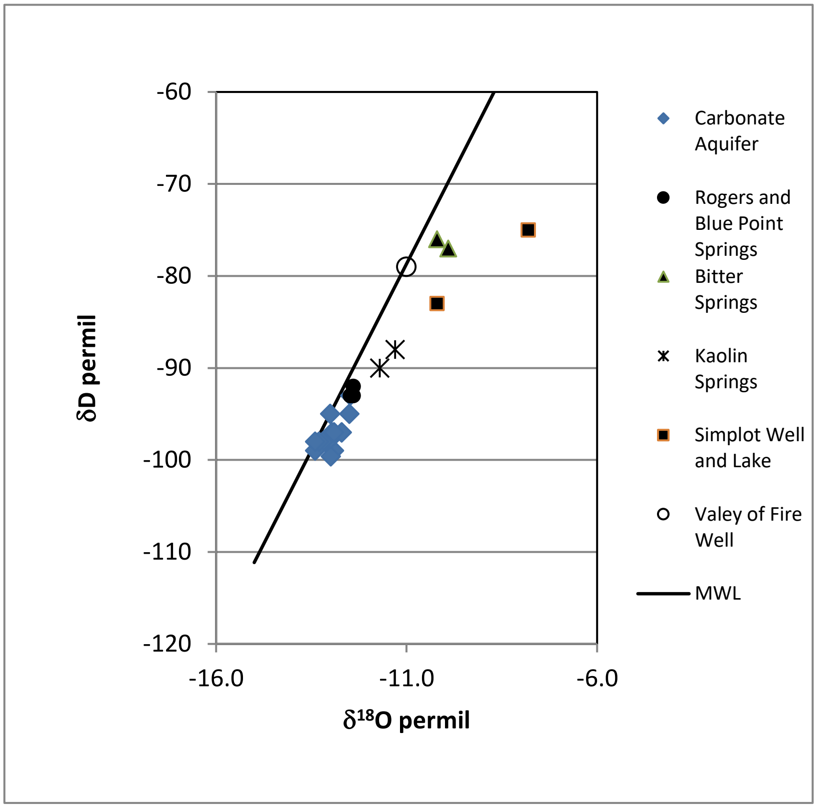


Figure 2. Environmental isotopic data.

The process of evaporation yields a kinetically driven isotopic fractionation that results in plotted values moving to new positions to the right of the line and toward more enriched values of both isotopes. This indication of evaporation is not surprising for the Bitter Springs and Kaolin Springs samples. Bitter Springs was little more than a seep and evaporation was inevitable even though the sample was collected as close to the point of emergence as possible (Figs. 3, 4). Similarly, the sample at Kaolin Springs was actually from the seep that was flowing into a ponded water area; nevertheless

Figure 3. Orifice and sampling point for Bitter Spring.

Figure 4. Bitter Creek drainage.

evaporation would be probable (Figure 5).

The water samples from Rogers and Blue Point Springs are isotopically enriched compared with water sampled from the carbonate aquifer. Blue Point Spring δD of -93 ‰ and $\delta^{18}O$ of -12.4 ‰ contrast with the carbonate wells that have an average δD of -97 ‰ and $\delta^{18}O$ of -13 ‰. This difference is significant and results in a separation of the plotted values (Figure 2). If the water emerging from Rogers and Blue Point Springs is principally water that originates in the carbonate aquifer, and is modified only by reaction with reactions with rocks or other water along the flow path, then there are only two explanations for the difference in the isotopic values. The shift to more enriched values would occur if the spring water has been evaporated or has mixed with a water with a more enriched signature.

Figure 5. Pond at the Kaolin Spring orifice.

It is clear that relying on evaporation as a cause is not plausible since both springs have values that plot on the MWL. It then appears that some mixing is occurring. The Valley of Fire well is just north of the Muddy Mountains and the well is screened entirely within the Mesozoic section. It provides the only well sample that could be inferred as representing the composition of recharge from more local sources, probably the Muddy Mountains, but it may itself be a mixture of incoming water from the carbonate aquifer plus recharge from the Muddy Mountains. This is a key sample because it has δD and $\delta^{18}O$ that are clearly enriched, but not evaporated, and thus could be representative of a water that is mixing with incoming carbonate aquifer type water, and emerging at Rogers and Blue Point Springs. Its isotopic composition indicates that recharge of this water occurred at lower elevations and/or warmer conditions than for the other carbonate water, which was largely recharged in central and northern Nevada, and during cooler conditions.

Radiocarbon (^{14}C). The use of radiocarbon in groundwater investigations has a long history and is based on the idea that the fraction of dissolved inorganic carbon in the water that was derived from the soil zone during recharge carries with it a ^{14}C signature . This ^{14}C value originates from the solution of respired carbon dioxide from roots of plants, which were in equilibrium with the reservoir of cosmogenic ^{14}C in the atmosphere. There are correction factors that must be applied to adjust for the different sources of carbon, some of which may be derived from the dissolution of carbonate minerals, such as the calcite in limestone, that are devoid of ^{14}C. The half life of ^{14}C is 5730 years and the most accurate age dating can yield a date up to around 50,000 years before present.

Chlorine-36 (^{36}Cl). The cosmogenic ^{36}Cl is created naturally in the upper atmosphere and is essentially always entering the hydrologic cycle; additionally a pulse of ^{36}Cl many times background level entered the atmosphere during the post WWII testing of nuclear weapons in the atmosphere. As in the case of tritium, the post-bomb levels date the time frame of the water relative to the nuclear tests; but because of the 3.01×10^5 year half-life of ^{36}Cl, it also has been useful in dating very old groundwater (see Clark and Fitz, 1997 for an overview of the application). The concentration of ^{36}Cl can be measured in a Tandem accelerator at very low concentrations and thus it is often reported as the number of atoms of ^{36}Cl in 10^{15} atoms of stable chlorine (termed the ^{36}Cl ratio herein). ^{36}Cl continues to decay along a flow path but the half-life is so long the decay only becomes useful in deep basin studies or circumstances of long flow path. Here it is more useful as a label of the source water. ^{36}Cl data are available for 13 samples, which range in ^{36}Cl ratios from 6 (ECP-3) to 390 (MX-5). Rogers and Blue Point Springs had ratios of 28 and 25 respectively. Samples from Coyote Spring Valley had ratios of 107 and 390, while those from the southern carbonate wells had values of 23, 6, 25, and 43. Samples for locations near Rogers and Blue Point Springs have ratios of 203, 276, 47, 250, and 181. Thus, the ^{36}Cl ratios of Rogers and Blue Point Springs are most similar to samples from carbonate wells in Garnet Valley and the Black Mountains area.

Without knowing how the aquifers obtained these ratios there is uncertainty in whether the measured values are representative of the local region of the aquifer. That analysis requires evaluation of each recharge source and the likely age or the recharge so that initial atmospheric ^{36}Cl can be estimated. Nevertheless, whatever the origin, the key point is that the measured values are instructive in this study. Chloride-36 is treated as a chemically conservative constituent and mixing can be computed without correction for decay, or contribution from *in situ* generation, which is small. There are too few measurements available at this point to make a mixing calculation that would be credible; however, it is clear that the ^{36}Cl mixed with a small percent of the Valley of Fire water composition is plausible within the range of error of the data currently available.

Boron-11 (^{11}B). The variability in $\delta^{11}B$ suggests that these data may be useful in evaluating the sources of the discharge at Rogers and Blue Point Springs. Most samples from wells yielded values ranging from -3 to +7 ‰. Exceptions include Mirant 1, which had a value of 25 ‰. The values near 0 ‰ are consistent with a marine biogenic carbonate source, while the Mirant 1 value may suggest an enriched biogenic carbonate source (Coplen et al., 2001, p. 19). In contrast, waters near Rogers and Blue Point Springs yielded $\delta^{11}B$ values ranging from -42.8 (Kaolin Spring) to -4.8 (Valley of Fire well) ‰, suggesting non-marine evaporite sources. Rogers and Blue Point springs had values of -11.6 and -13.1 ‰, respectively, suggesting a mixed source of water.

Sulfur-34 (^{34}S). Most of the samples, including those from Rogers and Blue Point Springs, had $\delta^{34}S$ contents in the range of 11 to 15 ‰. Samples collected from carbonate wells in the southern part of the area (California Wash, Garnet Valley, and Black Mountains Area) had higher values, ranging from 15 to 24 ‰. These data would suggest that the discharge from Rogers and Blue Point Springs has a low component of water in the southern part of the carbonate aquifer. However, the sulfate concentration in the Rogers and Blue Point Springs samples was considerably higher than in the carbonate well samples, indicating the dissolution of sulfate minerals, which could significantly affect the $\delta^{34}S$ content of the spring discharge.

Carbon-13 (^{13}C). The $\delta^{13}C$ data do not provide information on source areas, but will be used to estimate the correction of ^{14}C ages based on the dissolution of "dead" carbon, which reduces the ^{14}C concentration and therefore yields uncorrected ages that are too old unless corrected.

5.2 PRELIMINARY GEOCHEMICAL MODELING

Geochemical modeling was performed to develop a general understanding of the information provided by the chemical and isotopic information, and to evaluate whether reasonable chemical reactions (given the geology of the area) could provide information with which to determine the source of the water discharging at Rogers and Blue Point Springs. The water composition in the Paleozoic carbonate aquifer changes from a Na-Ca-HCO_3-SO_4 to Ca-Na-SO_4-HCO_3 water type from north to south across the study area (from Coyote Spring Valley, to California Wash and to Garnet Valley); this increase in calcium sulfate is accompanied by a gradual increase in dissolved solids (Table 4). The compositional change is significant enough that for purposes of modeling, three regions were designated as initial endmember water compositions. The endmembers are defined as northern, central and southern, and will be used to evaluate whether these differences will provide information regarding what the water sources to the Lake Mead are. Three wells were chosen to represent the endmember compositions, MX-5, ECP-3, and Mirant, rather than use the averaged concentrations in samples from each of these areas. This derives from the fact that a charge-balanced analysis is needed for detailed calculations of additions and losses of each dissolved constituent tested by the model. Averages are neither true nor sufficiently accurate representations of the actual water composition; however, the converse applies in that the well composition selected for modeling is assumed to be representative of the endmember group. Wells within the endmember groups were selected based on the completeness of both the chemical analysis as well as the range of isotopic values measured for this study.

The geochemical modeling domain extends from the Muddy River Springs Area, Garnet Valley, and California Wash to the springs near Lake Mead. The objective is to determine if any or all of the suspected water sources could plausibly evolve into a water composition like that observed in Rogers & Blue Point Springs. The required processes to match the chemistry at the springs may provide information on whether the spring discharge is local recharge, regional groundwater, or a mixture of local recharge and regional flow system water.

Geochemical modeling is by definition a numerical process with many scenarios and constructions. Inverse geochemical models are a subset to this process in which all the plausible reactions between minerals and gases, or mixing with other water sources, are considered in defining the evolution of an initial endmember into a final endmember. Geochemical modeling approaches are described in detail in Bassett (1997), and Bassett and Melchior (1989). Assumptions include the following:

- Identified groundwater gradient is correct;
- Endmember compositions are representative;
- Structural, geologic, and lithologic interpretation is correct; and
- Needed phases with estimated isotopic content are correct.

The following pathways were considered:

1. Northern: Muddy River Springs (represented by MX5) to Blue Point Spring;

2. Central: California Wash (represented by ECP-3) to Blue Point Spring; and

3. Southern: Garnet Valley (represented by Mirant No. 1) to Blue Point Spring.

Preliminary modeling of the three flow paths demonstrated:

1. The major ion chemistry and stable isotopic compositions of S, B, and C at Rogers and Blue Point Springs could be derived from the waters at each of the three different source areas through a process involving gypsum or anhydrite dissolution, calcite precipitation, B-containing halite dissolution, and ion exchange. The need for dissolution of B-containing halite implies contact with Tertiary basin-fill (Muddy Creek and/or Horse Springs) sediments;

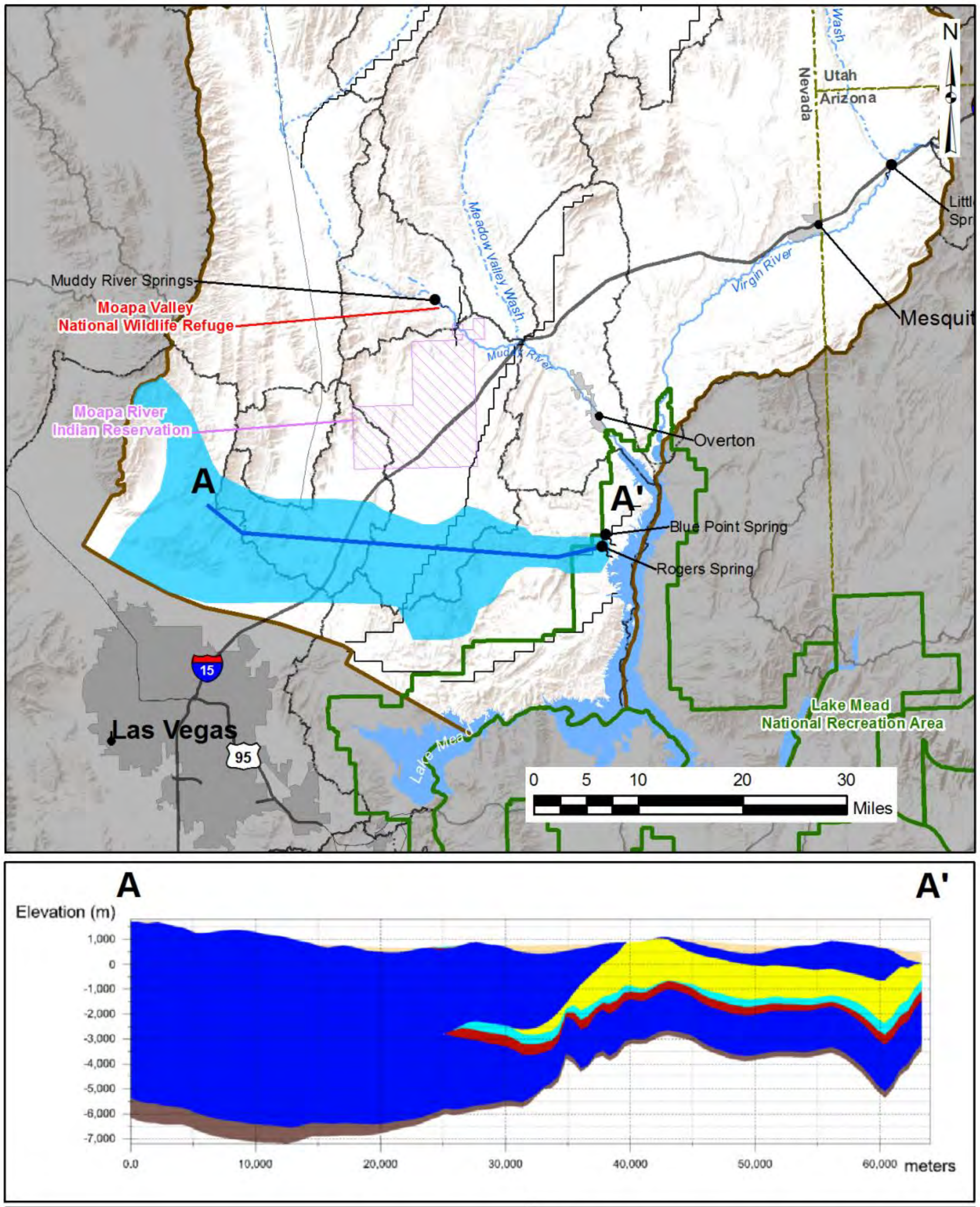
2. The δD and $\delta^{18}O$ compositions of Blue Point Spring can only be matched by addition of recharge water with a heavier isotopic composition than the water from the three postulated source areas. This implies mixing of local recharge. The Muddy Mountains are the most feasible source of recharge. ^{14}C data from the Simplot and Valley of Fire wells support this conclusion. The ^{14}C value from Blue Point Spring is best matched by using a source of water in the southern part of the study area; and

3. The water at Rogers and Blue Point Springs cannot be derived from the Valley of Fire water alone. The δD, ^{14}C, and ^{36}Cl data require mixing with another source of water.

In summary, the preliminary modeling indicates that the water discharging from Rogers and Blue Point Springs is likely a mixture of water from the southern part of the study area (represented by a well in Garnet Valley) and local recharge, accompanied by contact with basin-fill deposits and perhaps Mesozoic rocks. The southern flow path is the most plausible, followed by the middle flow path. However, no data were available from Paleozoic carbonate wells near the Morman Mountains, and the geochemical feasibility of that pathway has not been evaluated.

5.3 EVALUATION OF SIMULATED FLOWPATH TO ROGERS AND BLUE POINT SPRINGS

In 2012, a new model of groundwater flow in the southern part of the CRFS was completed (Tetra Tech, 2012). This model was calibrated using geologic and hydrologic information, and the geochemical interpretation that the water discharging from Rogers and Blue Point Springs was a mixture of carbonate-aquifer water and local recharge discussed above. The interpretation that the southern pathway was the most likely pathway was not used to guide the model's development. After calibration, this model was used to estimate the "capture zone" for Rogers and Blue Point Springs. The groundwater model head file for the long-term model run was used to generate a flow-vector field for each model cell. The vector field was exported to ArcGIS, where a backward analysis of vectors terminating at Rogers and Blue Point Springs was conducted. Figure 6 portrays the capture zone (map, in blue) for these two springs simulated by the model. It includes parts of Las Vegas Valley, Garnet Valley, California Wash, and the Black Mountains Area Hydrographic Areas. Also shown is a section showing the hydrogeologic units (HGUs) along the indicated section line. Rogers and Blue Point Springs are located near the east end of the section, at the contact between the carbonate rocks (in dark blue) in the upper thrust plate and the basin-fill sediments (in tan). The upper plate carbonate rocks are underlain by undivided Mesozoic rocks (in yellow), the Kaibab and Toroweap (in teal), Permian redbeds (in red), and the lower plate of carbonate rocks. The Proterzoic clastic rocks and crystalline basement are shown in brown and white, respectively. In order for shallow groundwater to flow from Garnet Valley to Rogers and Blue Point Springs, it flows through carbonate rocks, Mesozoic clastic rocks, and Tertiary basin-fill sediments (Muddy Creek/Horse Spring) (along the section line), and the Kaibab-Toroweap and Permian redbeds (to the north of the section line). Alternatively, groundwater could flow in the lower plate carbonate rocks, and upwards along the Roger Springs Fault (not shown), mixing with water from the Mesozoic rocks.



TITLE: **Model Simulated Flow-Path Outline for Rogers and Blue Point Spring Flow**

LOCATION: **Colorado Regional Groundwater Flow System Southeastern Nevada**

TETRA TECH

FIGURE **6**

The objective of this part of the study was to use geochemical data to test whether the flow path as simulated by the flow model is consistent with the geochemical data. Simulated flow from the southern end of Garnet Valley passes beneath California Wash and the Muddy Mountains to Rogers and Blue Point Springs. The model simulates recharge occurring in parts of the Muddy Mountains.

The well density in the study area is low; however, three wells in southern Garnet Valley were selected for measurement of chemical and isotopic composition because they are in locations that represent the bounding area defining the flowpath to the springs (the Mirant-1, Genstar, and Dry Lake Valley (DLV) wells). The Valley of Fire well is the only well that provides information on the composition of water that has been in contact with the Mesozoic rocks and Tertiary basin-fill sediments. It is located a short distance north of the simulated flow path.

The geochemical models PHREEQC and NETPATH were used to rigorously compute the set of plausible mass transfer reactions that could be responsible for the change in composition observed from the wells in Garnet Valley to the springs. A mathematically consistent set of plausible reactions, based on the lithology of the rocks and sediments along the flow path, would provide support for the numerical flow model; an inability to define a reaction pathway would raise questions of uncertainty in the conceptual model or the underlying dataset.

The groundwater composition changes significantly along the flow lines between the southern end of Garnet Valley to the discharge point at Blue Point spring. It should be noted that the intervening mineral composition between wells in Garnet Valley and the springs could plausibly provide the source for the needed solute increase without the requirement of mixing with recharge from the Muddy Mountains. However, the isotopic composition of the Blue Point springs, most notably the hydrogen (δD) and oxygen ($\delta^{18}O$) isotopic composition, cannot be explained without a second water source that has a more enriched isotopic content, one which is consistent with recharge at lower elevations in this region. This chemical and isotopic requirement is satisfied by mixing approximately 74% groundwater with the composition of the wells in Garnet Valley with 26% water with the composition of the Valley of Fire well. The mineral reactions are minimal for the bulk of the observed mass transfer: dissolution of a small amount of calcite (<0.46 mmoles/L), larger mass of NaCl and gypsum/anhydrite (6.8 and 16.5 mmole/L respectively), a small loss of carbon dioxide and silica, and some ion-exchange reactions. All reactions are thermodynamically valid. The geochemical reaction path modeling completed in this study supports this pathway.

The conceptual direction of groundwater flow transects the Muddy Mountain Thrust Fault and requires that groundwater from the Paleozoic carbonate aquifer pass through Mesozoic clastic sediments and the Tertiary basin-fill evaporite lithologies (Muddy Creek and/or Horse Spring). Constraining the mixing percentages of California Wash and Valley of Fire groundwater by the required δD and $\delta^{18}O$ composition of Blue Point Spring, it was determined that the other two stable isotopic systems ($\delta^{34}S$ and $\delta^{11}B$) also fit the model. The sulfur ($\delta^{34}S$) isotopic composition of the sulfate dissolved from gypsum/anhydrite yields the observed isotopic value of Blue Point Spring, when dissolved in the amount needed to increase the sulfate from the two groundwater components (300 mg/L and 229 mg/L) to the

1916 mg/L SO_4 observed in Blue Point Spring. Similarly the increase of boron concentration by more than a factor of 4 is also due to dissolution. The Mirant well ($\delta^{11}B$ of 25.0 ‰) mixes with the Valley of Fire groundwater ($\delta^{11}B$ of -4.8 ‰) and acquires dissolved boron from the halite rich evaporite section ($\delta^{11}B$ of -15.96 ‰) to yield the isotopic composition of Blue Point Spring ($\delta^{11}B$ of -14.2).

The age of the groundwater is estimated by measuring the tritium content of the water and the radiocarbon content of the dissolved inorganic carbon (DIC). The tritium content of groundwater and of the springs is below detection limit indicating the water is not comprised of recent recharge (< 60 years before present). The measured ^{14}C content is adjusted to the more reasonable ^{14}C or corrected value, by using the computer model NETPATH which incorporates the dilution or enrichment of ^{14}C as the result of reaction with carbonate minerals or carbon dioxide, using the stable isotopic composition ($\delta^{13}C$) of these phases. The estimates for the corrected ^{14}C values for travel time along the three pathways including the mixing with the Valley of Fire well are 24,324 years before present (ybp) for the Mirant path, 25,555 ybp for the Genstar well pathway, and 24,559 ybp for the Dry Lake Valley well. These ages would imply a carbon-14 transport rate of approximately 6 ft/yr.

6.0 CONCLUSIONS

The chemical and isotopic data provide an independent assessment of the conceptual model and test the concepts in ways that flow models cannot. The approach used here can be used to determine if the flow pathways are plausible, if they are supported by the water composition, and if the time frames of flow are corroborated by time-related isotopic ages.

The results of the preliminary geochemical modeling study suggest that Rogers and Blue Point are most likely fed by groundwater from the Paleozoic carbonate aquifer in the southern part of the study area (represented by water sampled from Garnet Valley) and with some minimal <30%) recharge contribution from the Muddy Mountains. This model fits the geochemical data better than a flow path from the vicinity of MX-5. It also matches the ^{36}Cl data better.

It should be noted that this evaluation did not consider a possible flow path from the area of the Morman Mountains. Additional data from deep wells would aid such an evaluation.

Geochemical modeling performed along the capture zone of the Rogers and Blue Point Springs as simulated by the recently completed flow model shows that, assuming that the Valley of Fire well water chemistry is representative of the recharge water beneath the Muddy Mountains, approximately 30% of the spring discharge is derived from the local recharge. Further, the net rate of ^{14}C transport from Garnet Valley to the springs is approximately 6 ft/yr.

Several factors need to be considered in using geochemical data and models. These affect this evaluation.

1. There is spatial and temporal variability in measured chemical and isotopic compositions that will result in different quantitative results;

2. Wells typically penetrate multiple lithologies at different depths. As a result, water samples will be mixtures of waters that are likely different in their compositions and ages. Also, wells that have long completion intervals may sample waters that have taken significantly different flow paths to reach the well. The interpretation of these data assumes that the mixture is representative of the aquifer at the location of the well;

3. Geochemical modeling requires information on the isotopic compositions of reactive mineral phases. This information is rarely available along the flow paths being modeled, and reasonable assumptions are frequently made based on data from others areas and geology. In addition, spatial variability should be expected, but is typically not considered; and

4. The geochemical model assumes that equilibrium between aqueous and solid phases is occurring. Slowly occurring reactions may affect the results.

As a result of the assumptions and limitations, the results of geochemical models should be interpreted to determine which sets of reactions are consistent with the data and which are not, but the numerical results should be considered to be approximate.

It is also important to realize that the geochemical evaluation does not provide information on the possible effects of pumping on spring discharge. It is incorrect to assume that only the pumping within flow paths that provide water to a spring will impact the spring discharge. The distribution of hydraulic-head changes (which will affect spring discharge rates) is independent of the flow paths, and will generally cover a much larger area than the area which contributes water to the spring.

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Appendix B

APPENDIX TABLE B-1 - Summary of Pumping Rates Used in Modeling Simulations #1 - #3

Basin	Permit	Priority Date	Annual Duty (afy)	Owner of Record	2017 Withdrawals (ac-ft)	Simulation 1 Pumping Rates (afy)	Simulation 2 Pumping Rates (afy)	Simulation 3 Pumping Rates (afy)
Coyote Spring Valley	*85249	10/22/1919	109.8	BEDROC LIMITED LLC	109.80	109.80	90.63	81.08
Coyote Spring Valley	*85250	10/22/1919	233.2	BEDROC LIMITED LLC	449.77	233.20	192.48	172.19
Muddy River Springs Area	*50733	8/13/1947	70	LDS		70.00		
Muddy River Springs Area	*50723	8/13/1947	88	LDS	88.00	88.00	72.64	64.98
Muddy River Springs Area	*50729	8/13/1947	120	LDS	55.15	120.00	99.05	88.61
Muddy River Springs Area	*50728	8/13/1947	158	LDS		158.00		
Muddy River Springs Area	*50731	8/13/1947	586	LDS	96.44	586.00	483.68	432.70
Muddy River Springs Area	*50732	8/13/1947	930	LDS		930.00		
Muddy River Springs Area	*29296	2/4/1948	300	NEVADA POWER COMPANY	88.26	300.00	247.62	
Muddy River Springs Area	38871	2/4/1948	75	EGTEDAR, ASCAR	11.10	75.00	61.91	55.38
Muddy River Springs Area	*86209	4/20/1948	14.01	3335HILLSIDE LLC	14.01	14.01	11.56	10.34
Muddy River Springs Area	*71026	4/20/1948	3.993	PARSON, BILLY & LINDA	13.27	3.99	3.30	2.95
Muddy River Springs Area	*71344	4/20/1948	6.067	PARSON, BILLY & LINDA		6.07		
Muddy River Springs Area	*82096	4/20/1948	1.903	CLOUD, MARY K		1.90		
Muddy River Springs Area	*82097	4/20/1948	2.891	CLOUD, MARY K	2.29	2.89	2.39	2.13
Muddy River Springs Area	*77381	4/20/1948	6.069	WILLIAM O`DONNELL		6.07		
Muddy River Springs Area	*77382	4/20/1948	9.221	WILLIAM O`DONNELL		9.22		
Muddy River Springs Area	59257	4/20/1948	15	BRUNDY, LARRY	9.54	15.00	12.38	11.08
Muddy River Springs Area	63504	4/20/1948	15	KOLHOSS, KELLY	9.54	15.00	12.38	11.08
Muddy River Springs Area	59256	4/20/1948	28.875	WHITMORE, DAN	18.37	28.88	23.83	21.32
Muddy River Springs Area	59253	4/20/1948	43.875	LEAVITT, UTE	27.96	43.88	36.21	32.40
Muddy River Springs Area	*24186	8/14/1948	310	NEVADA POWER COMPANY		310.00		
Muddy River Springs Area	*64840	10/7/1948	19.8	CLARK COUNTY	0.04	19.80	16.34	14.62
Muddy River Springs Area	*50851	10/7/1948	30	CLARK COUNTY		30.00		
Muddy River Springs Area	*22633	12/20/1948	297.5	NEVADA POWER COMPANY		297.50		
Muddy River Springs Area	*50724	10/4/1949	162.55	LDS		162.55		
Muddy River Springs Area	*50275	10/7/1949	32.88	NEVADA POWER COMPANY	55.00	32.88	27.14	
Muddy River Springs Area	*22636	6/19/1952	260	NEVADA POWER COMPANY	2.09	260.00	214.60	
Muddy River Springs Area	*22632	6/19/1952	315	NEVADA POWER COMPANY	134.08	315.00	260.00	
Muddy River Springs Area	*22635	12/18/1958	25	NEVADA POWER COMPANY		25.00		
Garnet Valley	83553	7/24/1959	3	TECHNICHROME		3.00		
Muddy River Springs Area	*50934	11/20/1959	55.4	NEVADA POWER COMPANY	16.11	55.40	45.73	
Muddy River Springs Area	18437	11/20/1959	20.15	COYOTE SPRINGS INVESTMENT LLC	1.00	20.15	16.63	14.88
Muddy River Springs Area	21466	8/15/1963	183.2	CASA DE WARM SPRINGS LLC		183.20		
Muddy River Springs Area	*50730	4/28/1965	25	LDS		25.00		
Muddy River Springs Area	*50725	4/28/1965	65	LDS		65.00		
Muddy River Springs Area	27216	8/25/1965	1.381005	UNITED STATES OF AMERICA	0.18	1.38	1.14	1.02
Muddy River Springs Area	22738	8/25/1965	18.81	DAVIS, DON J. & MARSHA L.	18.81	18.81	15.53	13.89
Muddy River Springs Area	*22949	2/2/1966	433	NEVADA POWER COMPANY/MOAPA BAND of PAIUTE INDIANS		433.00		
Muddy River Springs Area	**22950	2/2/1966	0	NEVADA POWER COMPANY		0.00		
Muddy River Springs Area	**22951	2/2/1966	0	NEVADA POWER COMPANY		0.00		
Muddy River Springs Area	**22952	2/2/1966	0	NEVADA POWER COMPANY		0.00		
Muddy River Springs Area	**24185	2/2/1966	0	NEVADA POWER COMPANY		0.00		
Garnet Valley	*64880	7/24/1967	133.81	CHEMICAL LIME COMPANY	117.17	133.81	110.45	131.67

APPENDIX TABLE B-1 - Summary of Pumping Rates Used in Modeling Simulations #1 - #3

Basin	Permit	Priority Date	Annual Duty (afy)	Owner of Record	2017 Withdrawals (ac-ft)	Simulation 1 Pumping Rates (afy)	Simulation 2 Pumping Rates (afy)	Simulation 3 Pumping Rates (afy)
Muddy River Springs Area	*25310	10/9/1969	160	MOAPA BAND of PAIUTE INDIANS		160.00		
California Wash	26371	11/18/1969	90	MOAPA VALLEY WATER COMPANY		90.00		
Muddy River Springs Area	*50272	7/7/1970	99.51	NEVADA POWER COMPANY		99.51		
Muddy River Springs Area	*50273	7/7/1970	289.91	NEVADA POWER COMPANY		289.91		
Muddy River Springs Area	*85156	7/7/1970	322.17	NEVADA POWER COMPANY		322.17		
Muddy River Springs Area	*29298	7/7/1970	327.5	NEVADA POWER COMPANY		327.50		
Muddy River Springs Area	*79068	7/7/1970	432.7	NEVADA POWER COMPANY		432.70		
Muddy River Springs Area	*50727	7/29/1970	60	LDS		60.00		
Muddy River Springs Area	*50726	7/29/1970	65	LDS		65.00		
Garnet Valley	*74399	7/20/1981	74.57	NEVADA POWER COMPANY	74.57	74.57	61.55	1,800.00
Garnet Valley	*63261	10/20/1981	100	CHEMICAL LIME COMPANY OF ARIZONA	55.61	100.00	82.54	98.40
Garnet Valley	*83715	10/20/1981	37	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC		37.00		
Garnet Valley	*83714	10/20/1981	157	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC	209.30	157.00	129.59	154.49
Garnet Valley	63348	10/20/1981	4	WESTERN MINING & MINERALS, INC.	2.71	4.00	3.30	3.94
Garnet Valley	77745	10/20/1981	10.02	NORTH LAS VEGAS-CITY	10.02	10.02	8.27	9.86
Coyote Spring Valley	*74095	3/31/1983	500	COYOTE SPRINGS INVESTMENT, LLC	172.78	500.00	412.70	369.20
Coyote Spring Valley	*74094	3/31/1983	1000	CLARK COUNTY COYOTE SPRINGS WATER RESOURCES GID		1,000.00		
Coyote Spring Valley	*70430	3/31/1983	1140	COYOTE SPRINGS INVESTMENT, LLC		1,140.00		
Coyote Spring Valley	*70429	3/31/1983	1500	CLARK COUNTY COYOTE SPRINGS WATER RESOURCES GID	1,226.64	1,500.00	1,238.10	1,107.60
Coyote Spring Valley	*70430R01	3/31/1983	460	COYOTE SPRINGS INVESTMENT LLC				
Coyote Spring Valley	*77292	3/31/1983	400	SNWA		400.00		
Muddy River Springs Area	*46932	5/19/1983	1000.15451	MOAPA VALLEY WATER DISTRICT		1,000.15		
Coyote Spring Valley	**77293	9/27/1985	4000	SNWA		1,557.08		
Garnet Valley	**86961T	9/27/1985	0	SNWA	217.38		222.86	265.68
Garnet Valley	**86962T	9/27/1985	0	SNWA				
Garnet Valley	**86959T	9/27/1985	0	SNWA				
Garnet Valley	**86960T	9/27/1985	0	SNWA				
Coyote Spring Valley	77164	12/30/1985	2500	NEVADA POWER COMPANY				
Coyote Spring Valley	*77294	1/27/1986	100	SNWA				
Coyote Spring Valley	**77295	1/27/1986	0	SNWA				
Coyote Spring Valley	**77296	1/27/1986	0	SNWA				
Muddy River Springs Area	**52520	4/14/1986	0	MOAPA VALLEY WATER DISTRICT	1,447.93		1,195.12	664.17
Coyote Spring Valley	*77297	7/15/1986	4500	SNWA				
Coyote Spring Valley	**77298	7/15/1986	0	SNWA				
Coyote Spring Valley	**77299	7/15/1986	0	SNWA				
Coyote Spring Valley	**77300	7/15/1986	0	SNWA				
Coyote Spring Valley	**77301	7/15/1986	0	SNWA				
Coyote Spring Valley	**77302	7/15/1986	0	SNWA				
Coyote Spring Valley	**77303	7/15/1986	0	SNWA				
Coyote Spring Valley	**77304	7/15/1986	0	SNWA				
Coyote Spring Valley	**77305	7/15/1986	0	SNWA				
Coyote Spring Valley	**77306	7/15/1986	0	SNWA				
Garnet Valley	56855	10/28/1986	144.146233	GEORGIA PACIFIC CORPORATION	94.27		118.98	141.84
California Wash	**50559	2/2/1987	0	NEVADA POWER COMPANY	28.97		298.77	356.18

APPENDIX TABLE B-1 - Summary of Pumping Rates Used in Modeling Simulations #1 - #3

Basin	Permit	Priority Date	Annual Duty (afy)	Owner of Record	2017 Withdrawals (ac-ft)	Simulation 1 Pumping Rates (afy)	Simulation 2 Pumping Rates (afy)	Simulation 3 Pumping Rates (afy)
California Wash	50558	2/2/1987	28.970416	NEVADA POWER COMPANY				
California Wash	50560	2/2/1987	28.970416	NEVADA POWER COMPANY				
Garnet Valley	*66784	3/6/1987	156.84	DRY LAKE WATER, LLC				
Black Mountains Area	*68351	6/21/1988	542.98	DRY LAKE WATER, LLC				
Garnet Valley	*83707	10/3/1988	0.11	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83709	10/3/1988	0.11	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83710	10/3/1988	0.11	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83712	10/3/1988	3.7	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83713	10/3/1988	23.8	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83711	10/3/1988	40.78	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Garnet Valley	*83717	10/3/1988	68.39	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC	271.07		56.45	67.30
Garnet Valley	*83708	10/3/1988	68.5	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC	67.99		56.54	67.40
Garnet Valley	*83716	10/3/1988	68.5	REPUBLIC ENVIRONMENTAL TECHNOLOGIES INC				
Black Mountains Area	*68350	10/18/1988	119.44	DRY LAKE WATER, LLC				
Black Mountains Area	*68352	10/18/1988	137.55	DRY LAKE WATER, LLC				
California Wash	75198	4/4/1989	25	COYOTE SPRINGS INVESTMENT LLC				
California Wash	*70257	10/17/1989	2500	MOAPA BAND of PAIUTE INDIANS	12.82		1,031.75	1,480.00
California Wash	**70258	10/17/1989	0	MOAPA BAND of PAIUTE INDIANS				
California Wash	**70259	10/17/1989	0	MOAPA BAND of PAIUTE INDIANS				
Garnet Valley	**79002	10/17/1989	0	SNWA				
Garnet Valley	**79003	10/17/1989	0	SNWA				
Garnet Valley	**79004	10/17/1989	0	SNWA	233.33		192.59	229.60
Garnet Valley	**79005	10/17/1989	0	SNWA	230.34		192.59	229.60
Garnet Valley	**54073	10/17/1989	0	SNWA				
Garnet Valley	**86967T	10/17/1989	0	SNWA				
Garnet Valley	**86968T	10/17/1989	0	SNWA				
Garnet Valley	**86969T	10/17/1989	0	SNWA				
Garnet Valley	**83490	10/17/1989	0	SNWA	17.50		247.62	295.20
Garnet Valley	**86970T	10/17/1989	0	SNWA				
Garnet Valley	**79001	10/17/1989	0	SNWA				
Garnet Valley	**68822	10/17/1989	0	SNWA	350.00		577.78	688.80
Hidden Valley	*54074	10/17/1989	2200	SNWA				
Garnet Valley	*87169T	10/17/1989	5	SNWA				
Black Mountains Area	*55269	10/30/1989	96	NEVADA COGENERATION ASSOCIATES #1	33.05		79.24	94.46
Black Mountains Area	*58031	10/30/1989	824	NEVADA COGENERATION ASSOCIATES #1	834.72		680.13	810.82
Black Mountains Area	*58032	9/13/1990	745	NEVADA COGENERATION ASSOCIATES	639.54		614.92	733.08
Muddy River Springs Area	*55450	11/9/1990	2171.906564	MOAPA VALLEY WATER DISTRICT	1,013.76		1,792.69	996.25
California Wash	57441E	4/16/1992	32.591718	NDOT				
Muddy River Springs Area	*58269	10/27/1992	1085.94	MOAPA VALLEY WATER DISTRICT				
Muddy River Springs Area	*66043	10/27/1992	2533.9	MOAPA VALLEY WATER DISTRICT	361.79		2,091.48	1,162.30
Muddy River Springs Area	61427	7/26/1995	1.350316	S & R, INC.				
Black Mountains Area	*68353	4/17/1998	592.06	DRY LAKE WATER, LLC				
Garnet Valley	*81344	8/25/2000	8	DRY LAKE WATER. LLC	8.00		6.60	7.87
Garnet Valley	*72098	8/25/2000	13.16	DRY LAKE WATER, LLC	13.00		10.86	12.95

APPENDIX TABLE B-1 - Summary of Pumping Rates Used in Modeling Simulations #1 - #3

Basin	Permit	Priority Date	Annual Duty (afy)	Owner of Record	2017 Withdrawals (ac-ft)	Simulation 1 Pumping Rates (afy)	Simulation 2 Pumping Rates (afy)	Simulation 3 Pumping Rates (afy)
Garnet Valley	**79948	8/25/2000	0	DRY LAKE WATER LLC	8.97		24.76	29.52
Garnet Valley	**66785	8/25/2000	0	DRY LAKE WATER, LLC				
Garnet Valley	**77389	8/25/2000	0	DRY LAKE WATER, LLC				
Muddy River Springs Area	*75161E	12/6/2006	905.81	NEVADA POWER COMPANY				
California Wash	**76643	1/18/2008	0	MOAPA BAND of PAIUTE INDIANS	30.06		1,031.75	1,480.00
Coyote Spring Valley	**77291	8/13/2008	0	SNWA				
Garnet Valley	**79009	11/2/2009	0	SNWA				
Garnet Valley	**79008	11/2/2009	0	SNWA				
Garnet Valley	**79010	11/2/2009	0	SNWA				
Garnet Valley	**79007	11/2/2009	0	SNWA				
Garnet Valley	**79006	11/2/2009	0	SNWA				
Garnet Valley	**86965T	11/2/2009	0	SNWA				
Garnet Valley	**86964T	11/2/2009	0	SNWA				
Garnet Valley	**86963T	11/2/2009	0	SNWA				
Garnet Valley	**86966T	11/2/2009	0	SNWA				
Muddy River Springs Area	**80843	5/9/2011	0	NEVADA POWER COMPANY				
Muddy River Springs Area	**80844	5/9/2011	0	NEVADA POWER COMPANY				
Muddy River Springs Area	**80845	5/9/2011	0	NEVADA POWER COMPANY				
Muddy River Springs Area	**80846	5/9/2011	0	NEVADA POWER COMPANY				
Muddy River Springs Area	*71766	7/21/2011	21.29	3335HILLSIDE, LLC	24.20		17.57	15.72
Garnet Valley	**84041	7/1/2014	0	DRY LAKE WATER LLC				
				Cumulative Pumping Totals:	**9,028.30**	**14,535.00**	**14,534.73**	**14,534.53**

National Park Service's Response to July 2019 Interim Order 1303 Reports

Project #: 117-0524303
August 16, 2019

PRESENTED TO

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National Park Service's Response to

July 2019 Interim Order 1303 Reports

As part of Order 1303 by the Nevada State Engineer, interested parties were allowed to provide reports to the State Engineer related to the establishment of a joint administrative unit for an area called the Lower White River Flow System (LWRFS). Interested parties are also allowed to provide rebuttal reports as part of this process.

This report provides general comments for consideration in the deliberations by the State Engineer. Tetra Tech provides these as a contractor to the National Park Service. These comments are not intended to refute the statements by other parties on a point by point basis, but to provide an overall context with which to consider them.

1. **Joint Administrative Unit** - We fully support the management of use of the groundwater system involving the regional carbonate aquifer as a joint administrative unit, and support using data to inform management decisions. The Order 1169 test is an excellent example of how pumping from an aquifer combined with careful collection of data over a large area can provide valuable information about an aquifer and the area affected by the pumping. Similar testing should be performed in other areas, with establishment of appropriate monitoring during testing and for long-term monitoring.
2. **LWRFS is not unique** - The Order 1169 test demonstrated that water levels were affected over a large part of the LWFRS, but that the effects are not uniform. For example, pumping of MX-5 does not appear to have affected water levels in CSVM-5 located a relatively short distance to the southwest. In contrast, water levels in wells up to 20 miles away to the south were affected (see Figure 5 of the July 3, 2019 report submitted by the U.S. Fish and Wildlife Service). These results were very similar to those obtained as part of a multi-month converging tracer test in the carbonate aquifer performed in Yucca Flat by the U.S. Department of Energy located in the Death Valley Regional Flow

System (DVRFS). In this test, water level changes were observed in the carbonate aquifer at Tracer Well 2 upgradient of Devils Hole. However, no changes were observed in some wells close to the pumping well (ER-6-1 #2) that were in separate fault blocks.

3. **Analytical Tools** – The Department of Interior agencies have used two analysis tools to evaluate water-level changes. The first is a data-analysis tool called SeriesSEE, an Excel add-in developed by the USGS (Halford and others, 2012) for determining causes for changes in groundwater levels. It can incorporate information on barometric pressure changes, earth tides, recharge, and groundwater pumping from multiple wells to decipher the measured signal and attribute the changes to these different processes. Pumping effects are calculated using the Theis equation. [Thus, comments implying that simple Theis solutions are better than SeriesSEE are inaccurate.] It uses a non-linear regression routine to apportion the effects of the different stresses as well as estimate the modeling parameters. If the input stresses have similar temporal signatures (such as seasonal pumping over the entire period of analysis), there is insufficient information with which to separate the effects of the individual stresses. However, if the input stresses are different (such as wells having similar seasonal patterns but clearly different periods of pumping) there is a better chance of separating their effects. The capability is increased if the patterns are distinctly different. Thus, an aquifer test to determine the extent of drawdown effects should not have a seasonal pattern and should last for more than a year, as was done with the Order 1169 test. The near-continuous pumping, rather than seasonal pumping, of MX-5 during the Order 1169 test allowed its effects to be determined.

   The second tool that can be used to predict effects of pumping is a calibrated three-dimensional flow model. Models, by their nature, are not 100% accurate, but they incorporate available information on geology, aquifer parameters, water budgets, responses to stress, etc. to allow predictions to be made based on mass-balance constraints. It is accepted that the predictions will not be exact, but as knowledge is gained, the models can be improved. We recommended to the State Engineer years ago that major water users be required to support the State Engineer's office in the development and maintenance of predictive groundwater models. Although we recognize that this may not be possible in the current regulatory framework, we again make this recommendation.

4. **Seasonal Pumping Effects** - Information on the widespread effects of pumping near the Muddy River Springs Area (MRSA) is provided by temporal changes in water levels in response to seasonal pumping that preceded the Order 1169 test and other more recent pumping in Garnet Valley. These seasonal signals are observed in most of the wells in Coyote Spring Valley, including as far south as CSVM-2 and CSV-3. To the north, they are present in alluvial well CSV3011M, and probably in CSVM-4. Seasonal effects are observed in Garnet Valley, but a careful analysis would be needed of the temporal pumping in Garnet Valley before determining whether this is the result of MRSA or Garnet Valley pumping.
5. **Capture zones are not indicative of drawdown zones** - The Moapa Band of Paiute Indians (MBPI) use an incorrect concept in arguing that pumping in most of California Wash would not affect the MRSA. They try to relate the area that would be affected by groundwater pumping to the capture areas of discharge locations. They present a simple model of flow toward the MRSA and toward Las Vegas Valley incorporating the area included in the vicinity of the LWRFS and the Death Valley Regional Flow System (DVRFS). The model is a very simple one-layer, steady-state model that simulates both groundwater flow and heat transport, and unfortunately is too simple to do a good job of either. But the present comment has to do with their argument that the boundary between the simulated capture zones of the MRSA and Las Vegas Valley is also a boundary that delineates where pumping effects would occur. Their simulated capture zone boundary is near the western boundary of Coyote Spring Valley and south of the MRSA, and includes approximately the northern 20% of California Wash. The last paragraph of the Appendix III of the MBPI report says:

   > The southern and westernmost areas of Coyote Spring Valley may be tributary to Las Vegas Valley, as is most of California Wash; neither of

> these areas, Garnet Valley, or Hidden Valley is tributary to the MRSA, which appears hydrodynamically isolated from the Las Vegas capture zone at the present time.

The reasoning associated with this statement was stated on page 61 of the MBPI report:

> The information presented here adds support to the idea that Las Vegas Valley is the terminus of a regional groundwater flow system originating north of White River Valley, but not ending at the Muddy River Springs Area as Eakin (1966) proposed, which instead is fed by a separate capture zone that includes Panaca Valley and terminates at the MRSA. [Note that this statement is firmly contradicted by water level data, which indicate that most of the simulated flow paths that end in Las Vegas Valley should actually discharge at Ash Meadows in the DVRFS.] <u>The implication for water management is that developments in areas tributary to Las Vegas Valley may not cause harm or even be sensed by monitoring, whereas developments in areas tributary to the MRSA, which might exclude alluvial-aquifer systems demonstrably isolated from the carbonate-rock aquifer, would impact endangered species and senior water-rights holders.</u> [Underlining added.]

This last statement is totally false. Drawdown from pumping is not sensitive to the boundary between capture zones. Whether a well is up-gradient, down-gradient, or cross-gradient from a discharge area is immaterial in whether drawdown from the well will affect the discharge area. The well does not need to be within the capture zone to affect water levels in the capture zone. Location affects changes in the capture zone boundary, but does not affect the distribution of drawdown. Only if the groundwater system behaves strongly non-linearly (i.e., when the drawdown is large compared with the saturated thickness of the aquifer) will the location have a significant impact, because the distribution of transmissivity would change in this instance. This is not an issue in the LWRFS, where the carbonate aquifer is commonly thousands of feet thick and under confined conditions in most areas.

Consider, for example, two wells located near each other, pumping at equal rates. The capture zone boundary will be located between the two wells. However, the drawdown in one well will be greater than if only that single well was being pumped, because of drawdown from the other well. This is a typical well-interference problem. The capture zone boundary is not a drawdown boundary.

In summary, it is not the source of the water that is important in evaluating pumping effects, it is the degree of hydraulic connectedness, which is a function of geology and hydrologic properties.

6. **Kane Springs Wash should be part of the joint administrative unit** - One of the questions asked of the interested parties is whether other basins should be included within the joint administration unit. It is worthy to note that in the data reports submitted, Kane Springs Valley was recommended most often for inclusion into the LWRFS, with recommendations provided by the NPS, U.S Fish & Wildlife Service (USFWS), Moapa Valley Water District (MVWD) and the Center for Biologic Diversity (CBD). Additionally, while SNWA retreated from an earlier position for inclusion of Kane Springs Valley into the LWRFS, they still provided data that would seem to support inclusion. Similarly, Coyote Springs Investment (CSI) also recognizes that Kane Springs Valley contributes flow to the LWRFS, but does not commit to recommending inclusion of this basin.

   Tetra Tech concurs with this recommendation. Water levels measured in CSVM-4 and KMW-1 show a response to the Order 1169 pumping, although less pronounced that observed in wells to the south. Water-level response data to the northeast of these wells are not available to allow prediction of pumping effects in most of Kane Springs Valley. However, Kane Springs Valley is the result of left-lateral strike-slip faulting which probably produces enhanced permeability parallel to these faults. Whether or not Kane Springs Valley is initially included in the joint administration unit, an aquifer test similar to that of the Order 1169 test should be performed before there is appreciable development in Kane Springs Valley.

In their data report, Lincoln-Vidler claim that there is no correlation in water level and pumping trends in Kane Springs Valley and northern Coyote Spring Valley, thus providing evidence that the two basins are hydrologically disconnected. As further proof, they presented hydrographs for wells KMW-1 and CSMV-4, in which they claimed the water levels during the Order 1169 pumping test period were still being affected by the wet period response from 2004-05 and therefore was masking or overriding any pumping signal from the test. The USFWS and SNWA also provided similar hydrographs for these two wells in their data reports and concluded that there appeared to be an Order 1169 pumping signal reflected in the hydrographs for both wells. The NPS and MVWD also concluded in their data reports that the available water level data in the area suggested that a pumping signal was detectable in Kane Springs Valley, without providing similar hydrographs.

The NPS has subsequently plotted the available water level data for both of these monitoring wells (Figure 1) that include best-fit trend lines projected through the data points representing the pre-pumping test period (2007-late 2010), the pumping test period (late 2010 – early 2013) and the post-pumping test period (early 2013 – present). The pumping test period was extended into early 2013 to account for the continuation of pumping at MX-5 into April 2013, which continued to affect water levels. As shown in Figure 1, both hydrographs show similar trends during all three of these periods. Of particular note is the nearly identical slopes and correlation coefficients for the trend lines through the portion of the hydrographs showing Order 1169 pumping effects (yellow data points). We agree with Lincoln-Vidler that the hydrograph data for CSVM-4 clearly shows the 2004-05 wet year response, as do many other monitoring wells in the LWRFS, which was clearly demonstrated by the USFWS and SNWA in their data reports. In fact, many of those wells exhibit similar trends during all three of the periods noted above. The NPS disagrees with Lincoln-Vidler that the effects of the 2004-05 wet period are masking/overriding Order 1169 pumping effects in these two wells, as both hydrographs clearly show a distinct declining trend during the Order 1169 pumping test (compared to the pre-test period), which suggests the pumping was of such magnitude as to override any effects that might still have been occurring from the 2004-05 wet year. The NPS will be providing additional rebuttal data in a later section suggesting that much of southern Nevada has been experiencing rising groundwater levels for the last several decades and that pumping in the LWRFS basins has been of such magnitude as to override this increasing trend.

In Coyote Spring Valley, transmissivity appears to be lower north of CSVM-6 than it is to the south. This results in higher hydraulic gradients in the northern part of the valley than in the south. Figure 2 shows recent water-level measurements and calculated gradients for selected wells in Coyote Spring Valley. [The water level for CE-VF-2 is estimated from data collected prior to loss of integrity of the well casing, and reflects carbonate aquifer levels.] The gradient between CSVM-2 and CSVM-6 (in the southern part of Coyote Spring Valley) is approximately 0.0008 ft/ft. The gradient between CSVM-6 and EH-4 is approximately 0.00003 ft/ft. Gradients between well pairs to the north are 1 to 2 orders of magnitude higher, with the highest between CSVM-5 and CSVM-6 (0.006). While the groundwater flux (Darcy flux) is not uniform, the higher gradients indicate that either the permeability of the rock mass is lower than in low-gradient areas, or that there are faults that impede flow across them. However, in spite of the reduced transmissivity, the propagation of drawdown to the mouth of Kane Spring Valley did occur during the Order 1169 test.

The Controlled Source Audio Magneto Telluric (CSAMT) geophysical surveys conducted by CSI suggest the Kane Springs Wash strike-slip fault zone may be continuous down-valley on the eastern side, which would direct groundwater toward many of the down-gradient production wells (e.g., MX-5 and CSI wells) in southern Coyote Spring Valley. Many of the low resistivity areas located within the Kane Springs Wash Fault Zone demarcated in Lines 1, 2, 10, 11 and 12 of the Lincoln-Vidler geophysical surveys (see Figures 4-5 through 4-8) confirm the presence of groundwater at depths known or suspected to be within the carbonate aquifer [e.g., see Well KMW-1 on Line 1 (Figure 4-6) and Well CSVM-4 on Line 12 (Figure 4-8)]. A similar low resistivity pattern is reflected on the east side of Coyote Spring Valley in Lines A, B and C (Figures 11, 12 and 13) of the CSI data report, indicating that similar flow conditions extend down-valley in the carbonate aquifer.

Our interpretation that the Kane Springs Wash fault zone may be acting to channel groundwater into northern Coyote Spring Valley along the various faults within this zone is further supported by the conditions observed during Lincoln-Vidler's 2006 pumping test at KPW-1. A 7-day constant-rate aquifer test was successfully completed at well KPW-1 pumping at a sustained rate of 1,800 gpm. The effects of the pumped well were monitored in a nearby monitor well KMW-1. The following summary of selected conclusions (*italicized*) presented in the URS (2006) well completion report (pages 7-1 and 7-2) indicated:

- *The well was sited in close proximity to the Willow Springs Fault and Kane Springs Wash fault zone in an area of extensive tectonic activity, leading to significant fracturing of the carbonate-rock aquifer. By locating the well in a highly fractured geologic terrain, the well is drilled in rocks with a secondary permeability that has been enhanced by faulting and fracturing. This is observed in the drill cuttings by the existence of fracture planes and small-scale solution features (i.e., vugs) indicative of karst development.* This conclusion would seem to be consistent with the DOI Agencies' and other stakeholders' position that much of the carbonate aquifer underlying the LWRFS is hydraulically connected as a result of the highly fractured, faulted and karstic carbonate-rock terrain that produces the aforementioned secondary permeability. Due to the resulting high diffusivity (ratio of transmissivity to storativity) and confined conditions of the carbonate aquifer beneath the LWRFS, drawdown responses can be expected to extend over large areas.

- *Analysis of the pumping test data indicated a moderately high transmissivity (90,000 – 300,000 gpd/ft) in the southern Kane Springs Valley area.* It's worth noting that the transmissivity and storativity (~ 2 x 10^{-4}) values calculated from this test were utilized as input parameters for the current groundwater flow model developed by Tetra Tech (Tetra Tech, 2012, Table 3-1), which was used to conduct the modeling simulations presented in the NPS' Order 1303 data report.
- *Residual drawdown data demonstrated that <u>hydraulic barriers to groundwater flow were not encountered</u> during the 7-day aquifer test.* This observation is counter to what Lincoln-Vidler are currently claiming, which is this fault zone is acting as a barrier or impediment to groundwater flow into Coyote Spring Valley. Instead, the pumping drawdown response indicates these faults impart recharge boundary effects, as groundwater underflow recharge along these conduit faults is intercepted by the pumping, thus capturing flow that is would otherwise contribute to Coyote Spring Valley and other basins within the LWRFS.
- *It was evident from testing that the residual recovery data represents fault-induced high transmissivity*. This observation is consistent with the NPS' position that faults and fractures within this fault zone naturally act to channel groundwater underflow recharge from Kane Springs Valley southward into the LWRFS. Such conduits are likely to have a relatively high transmissivity associated with them, especially if dissolution of the carbonate rock has occurred along the faults and fractures.
- *The carbonate-rock aquifer in this area behaves as a porous media similar to an alluvial aquifer system and thereby can be analyzed as such*.

Based on differences in the CSMAT lines 10 and 11, Lincoln-Vidler interpreted that a down-to-the-southwest normal fault (presumptively termed the Northern LWRFS Boundary Fault) is present in the area between these two lines. Lincoln-Vidler's introduction of a new normal fault is not supported by the geophysical data presented in Line 12 (Figure 4-8) in our opinion. Based on the information presented in Figure 4-9 of their report where the trace of this new fault intersects

Line 12 (near Stations 34000 and 35000), we see no indication of the presence of a normal fault near these stations.

Gravity data (Phelps and others, 2000) indicate that Coyote Spring Valley is a gravity low that is bounded by faults on the east side of the valley, based on the steepness of the gravity gradient (Figure 3). There appears to be a fault that strikes approximately N45°W near the mouth of Kane Springs Wash. This fault is southwest of the Northern LWRFS Boundary fault interpreted by Lincoln County/Vidler (Figure 4-9 of their report) and appears to cross CSMAT section 11 (Figure 4-7) at approximately Station 26500. It may not be visible on Line 12 because it may be below the effective depth of investigation of CSAMT. We point this out to encourage use of gravity data, and to recommend that additional gravity data on the west side of the valley would assist in understanding the structure below the depth of investigation of CSAMT.

Lincoln County/Vidler present temperature data which they used to indicate that the groundwater in eastern and western Coyote Spring Valley exits the valley in different directions. We agree, in general, with their analysis. The highest temperatures measured were for wells KMP-1 and CSVM-4, at the mouth of Kane Spring Valley. The most likely source of this heat is associated with the volcanic rocks and the relict associated magma chambers. They demonstrate that this water likely flows along the eastern side of Coyote Spring Valley and then toward the MRSA. These data indicate that Kane Spring Valley contributes a significant portion of the discharge at the MRSA, confirming hydraulic continuity of Kane Spring Valley with Coyote Spring Valley.

Groundwater modeling simulations (4 different simulations to date) by the DOI agencies qualitatively demonstrate that pumping drawdown effects in Kane Springs Valley and Coyote Spring Valley will likely coalesce with each other causing cumulative pumping impacts on the Muddy River Springs and Muddy River within 100 years, in contrast with Lincoln-Vidler's assertion that pumping effects in Kane Springs Valley will not be felt outside of the valley for at least 100 years.

Following the issuance of Order 1169, the NPS and others petitioned the Nevada State Engineer to include Kane Springs Valley as part of the Order 1169 set of basins due to the suspected hydrologic connection of this valley with Coyote Spring Valley. While this early petition was denied, additional data has been collected since that time which the NPS believes strongly supports this hydrologic connection and, therefore, inclusion of Kane Springs Valley into the LWRFS. If the Nevada State Engineer is unconvinced by the prevailing evidence for inclusion of Kane Springs Valley into the LWRFS administrative unit, the NPS would recommend that Lincoln-Vidler be ordered to conduct a longer-term pumping test similar to that recommended by USFWS for KPW-1 and monitor the response in existing and new monitoring wells in this area. Lincoln-Vidler should be required to pump the total amount of their existing water rights during the test to demonstrate that pumping in this area will not affect water levels and spring discharge in the current LWRFS basins, before granting any additional water rights in Kane Springs Valley.

7. **Effects of pumping in Coyote Spring Valley -** Coyote Springs Investment (CSI) contends that pumping from the carbonate aquifer in Coyote Spring Valley has less impact on groundwater levels in EH-4 than nearby carbonate pumping in the Muddy River Springs Area (Arrow Canyon wells). Further, they interpret the presence of a structurally high block of the carbonate aquifer to provide a barrier to flow in an east-west direction. The first of these is unimportant, while the second is unproven and is inconsistent with prevailing opinions and data about the carbonate aquifer.

   Whether pumping in Coyote Spring Valley has a larger or smaller impact than pumping in the MRSA is not the correct question. The correct question is "Does pumping in Coyote Spring Valley have an impact in the MRSA?" If so, then senior water rights and spring and surface-water discharge are impacted.

   Still, we are compelled to discuss CSI's approach to compare Coyote Spring Valley and MRSA pumping effects. CSI employs a simple Theis analytical solution to estimate impacts to groundwater levels in Muddy River Springs Area, due to pumping in Coyote Spring Valley, and claim that while this approach may not provide an absolute solution of groundwater level drawdown, it does provide a relative comparison. They further contend that the impact to groundwater levels is constrained by the value of transmissivity used to estimate drawdown; smaller values of transmissivity have a greater impact on drawdown when compared to higher values. While they recognize there is a hydraulic connection between Muddy River Springs Area and Coyote Spring Valley, their claim is that high values of transmissivity minimize the impact of the distal pumping on carbonate groundwater levels.

It is interesting to note that CSI's use of a simple Theis solution follows a discussion in their data report (Section 2.3) about concerns they had with the DOI agencies' use of the USGS' SeriesSEE water level modeling tool in their Order 1169 data report. One of the primary concerns CSI noted was this tool does not account for boundary conditions which may affect the drawdown responses that were observed and modeled with the tool. This discussion is followed by their discussion (Section 2.4) on the use of the Theis solution to demonstrate a relative comparison of drawdown impacts between the two areas of pumping. Within this discussion, they point out the analytical constraints under which the Theis solution is applicable, one of which is that this approach assumes an infinite aquifer extent for which boundary conditions do not have to be accounted. By this admission, their analysis commits the same error they accuse the SeriesSEE modeling tool of committing.

With respect to CSI's use of a simplistic Theis analytical solution to estimate impacts to groundwater levels in the Muddy River Springs Area due to pumping in Coyote Spring Valley, their analysis solely focuses on the effects that transmissivity has on predicting the degree and extent of drawdown related to pumping from the carbonate aquifer. As a result, they fail to evaluate the effect that differing values of aquifer storativity have on predicting drawdown effects. In the NPS' opinion, the storativity coefficient of 0.03 is too high for a confined carbonate aquifer setting, and that values on the order of 10^{-4} or 10^{-5} are more representative of the prevailing storage conditions in this confined aquifer, based on published storativity values for confined aquifers of this type of lithology.

While CSI correctly points out that a high aquifer transmissivity allows for drawdown effects to expand over a long distance, they fail to evaluate the effect that low storativity has on drawdown expansion. As noted by the DOI agencies, SNWA and others, the high hydraulic diffusivity (i.e., the ratio of T to S) of the carbonate aquifer is largely responsible for why we are observing pumping effects over wide areas of the LWRFS. Storativity has a significant effect on the hydraulic diffusivity and therefore the magnitude and extent of drawdown in the carbonate aquifer. Therefore, the effects of distal pumping in the carbonate aquifer of the LWRFS is sufficient to cause considerable impacts on the Muddy River Springs, especially when cumulative pumping effects are considered.

Another factor in CSI's use of the Theis solution is that the effects of capture of spring and surface discharge are ignored. Thus, the analysis will tend to overestimate the effects of Arrow Canyon pumping at EH-4, as discharge capture reduces drawdown.

Even if pumping of the carbonate and alluvial aquifer in the MRSA is reduced and that same amount of pumping is transferred to the carbonate aquifer in LWRFS basins further away from the Muddy River Springs, the pumping impacts will eventually expand from these basins to the Muddy River Springs and capture spring and ET discharge in this area. This was qualitatively demonstrated in the groundwater pumping simulations presented in the NPS' data report, at the pumping level that was modeled. Based on these simulation results, we can conclude that the amount of sustainable annual groundwater pumping in the LWRFS suggested by CSI in their data report(30,630 afy) is almost certainly unachievable without impacting the Muddy River Springs and senior water rights on the Muddy River.

Based on discussions presented in Sections 3.5, 3.6 and 5.0 of their data report, CSI contends that existing and new geological information indicates that the carbonate flow system in CSV is split into a western flow path that discharges into Hidden and Garnet Valleys and an eastern flow path that is connected to the MRSA. This view is largely based on their new CSMAT data that indicates the presence of a high-resistivity structural block within the carbonate aquifer that they interpret to separate the flow system in the vicinity of wells CSI-1, CSI-2 and CSI-3. Furthermore, they state that groundwater level responses in Coyote Spring Valley show that CSI pumping in the western flow path has no measurable impact on flow to the MRSA due to the presumed barrier effect provided by this structural block. Additionally, CSI contends that results from their Theis solution and observations of flow in the Warm Spring West gage since 2013 also show minimal or no impact in the Muddy River Springs Area due to pumping on the eastern side of Coyote Spring Valley.

CSI's CSMAT geophysical surveys have identified what appears to be a high-resistivity carbonate block or ridge located in the subsurface in the vicinity of wells CSI-1, CSI-2 and CSI-3, which correlates with an outcropping of Silurian-, Devonian- and Mississippian-aged carbonate rocks at land surface. We agree with the CSI's interpretation of the low-resistivity signature as representing carbonate rock. However, it is unlikely that the carbonate rock acts as a barrier. MX-5 is completed in this block and was quite productive. Wells to the north of MX-5, also in this structural block, had pronounced drawdown signals from MX-5 pumping as well as seasonal effects from MRSA pumping. The fault on the east side of this gravity and structural high did not prevent transmission of pumping effects on water levels into the MRSA. The effect of the fault on the west side is uncertain. MX-5 pumping effects were not observed in CSVM-5, but there are likely other faults that could limit the propagation of drawdown to the west of MX-5. The report states (p. 48) "Coyote Spring Valley monitoring wells CSVM-2, -3, -4, -5 (Figure 20), and CE-VF-2 (Appendix E) do not show a response to pumping that occurred in either Muddy River Springs Area or the eastern portion of Coyote Spring Valley." Except for CSVM-5, water levels in these wells show a strong seasonal pattern (caused by pumping in alluvium and the carbonate aquifer in the MRSA), and response to Order 1169 pumping.

CSI's assertion that groundwater level responses in Coyote Spring Valley show that CSI pumping in the western flow path has no measurable impact on flow to the MRSA, due to the presumed barrier effect provided by this structural block, cannot be confirmed by the data presented in their report. Claims that several monitoring wells in the vicinity of this carbonate block didn't show pumping impacts appear erroneous based on the USFWS' evaluation of the hydrographs for several of these wells. Some of these monitoring wells are located on the western and eastern side of this carbonate block. Wells CSI-3 and CSI-4, which are located on the western side of this structural block, and well MX-5, which appears to be located within this structural block, were all pumping as part of the Order 1169 test. The productivity of MX-5 refutes the contention that this block of carbonate rock can serve as a barrier. To claim that these CSI wells had little or no effect on flow to the MRSA cannot be substantiated by their analysis of the water level data. Until more persuasive evidence can be presented by CSI, stakeholders should conclude that the cumulative pumping of CSI-3, CSI-4 and MX-5 contributed to the Order 1169 pumping effects that are detectable in various hydrographs throughout the LWRFS.

With respect to CSI's claim that pumping on the eastern side of this structural block shows minimal or no impact in MRSA based on their lines of analysis, we refer the Nevada State Engineer and stakeholders to our earlier rebuttal comment relating to the Theis analysis conducted by CSI which presents some of the failings of their analysis.

If wells were to be drilled west of the "MX-5 structural block", long-term pumping tests should be performed to determine the distribution of drawdown. It would be inappropriate to assume that drawdown would not extend to the MRSA without testing data. In summary, we disagree with the interpretation that the structural high occurrences of the carbonate rock represent barriers to groundwater flow.

8. **Regional water levels are rising** - Within the carbonate aquifer in the LWRFS, there is a trend of declining water levels. Many hydrologists have attributed this to drought conditions. This interpretation does not take into account that water

levels are rising in many other areas in southern Nevada. During preparation of version 3 of the model of the Death Valley Regional Flow System (DVFRS), and in studies in general support of the Department of Energy, USGS hydrologists observed that water levels have been rising, not declining. This is documented in a presentation for the Nevada Water Resources Association by Jackson and others (2017). A copy of this presentation is provided in Appendix A of the NPS rebuttal report for reference. A more thorough discussion of the analysis behind the information in this presentation will be included in a USGS report publication related to groundwater modeling of the Death Valley Flow System that will published be in the near future (personal communication with Tracie R. Jackson, July 26, 2019).

The USGS presents nearly 70 hydrographs from the area covered by the DVRFS model that have rising water levels. Two of these hydrographs are located at Devils Hole, where the recent record shows rising levels since 2005. The observed rising groundwater levels occur in wells completed in several different types of aquifers including carbonate-rock, volcanic and basin fill aquifers that are present within this regional flow system. The USGS also noted that water level rises tended to be more pronounced in wells located nearer suspected recharge areas and in areas with lower aquifer transmissivity (personal communication with Tracie R. Jackson, July 26, 2019).

They evaluated temporal changes in recharge through the use of a standardized precipitation index for southern Nevada from the early 1900s through 2015 (Appendix A, Slides 43 & 44) which shows that the area was experiencing an extended period of dry conditions from about 1900 – 1970, followed by an extended period of wet conditions from about 1970 – 2015. This interpretation is different from those that claim that the area is currently undergoing drought conditions because of the longer time frame that the USGS evaluated. With the additional data, it is apparent that the period from 1970 to 2015 is wetter than normal, but contains dry periods within it.

During wet years, precipitation recharges the groundwater system while discharge occurs at the same time. During dry years, recharge is greatly reduced, but discharge occurs at about the same rate. Discharge is relatively constant over time as reflected by spring flows, because hydraulic gradients that are driving this flow remain relatively unchanged in the absence of pumping. The USGS also indicated during their presentation that groundwater level rises may be plateauing in southern Nevada, which could have management implications in basins where pumping is occurring.

It is worth noting that the USGS presented some hydrographs for wells located in basins where pumping was occurring (Appendix A, Slides 30-32) to demonstrate that even in pumped basins, wet year responses can still be observed in water level records even though water levels are generally declining in these basins. These hydrographs have many similarities to well hydrographs within the LWRFS, where wet year responses (e.g., 2004-05) are visible in many of the hydrographs. Similarly, water levels in the LWRFS have been declining for years in response to pumping occurring within the LWRFS basins. Increasing groundwater level trends in the LWRFS are commonly followed by a return to declining water levels during drier years, similar to the trends shown in Slides 30-32.

Based on the evidence of rising groundwater levels presented by the USGS in their presentation, we performed a search of available groundwater level data in neighboring basins that are distant from the pumping occurring in the LWRFS basins to see if rising water level trends were occurring in any of these basins. If numerous examples exist, then it can be argued that the climatic conditions causing the rise in groundwater levels in the non-pumping basins are also applicable to the LWRFS basins where pumping is occurring. Our cursory search of water level information included data accessible on the Nevada State Engineer's online Nevada Hydrology Data mapping application, as well as in the USGS' NWIS data base. The search was not exhaustive given the time constraints for submitting this rebuttal report, but we did find a total of 33 examples (Appendices A and B) in the following basins (Figure 4):

- Three Lakes Valley (Basin 211), Las Vegas Valley (Basin 212) and Tikapoo Valley (Basin 169B) on the west side of the Sheep Range [Appendix A, Slide 27];
- Coyote Spring Valley (Basin 210) [Appendix B];
- Black Mountains Area (Basin 215) [Appendix B];
- Lower Moapa Valley (Basin 220) [Appendix B];
- Lower Meadow Valley Wash (Basin 205) [Appendix B];
- Tule Desert (Basin 221) / Virgin River Valley (Basin 222) [Appendix B];
- Dry Lake Valley (Basin 181) [Appendix B];
- Coal Valley (Basin 171) [Appendix B]; and
- Garden Valley (Basin 172) [Appendix B].

The fact that there are numerous examples of rising groundwater levels in neighboring basins to the west, south, east and north of the LWRFS basins suggests that the LWRFS basins also should have been affected by rising groundwater levels during the same period of time reflected in the water level records recorded in these valleys. Evidence for this actually exists in the hydrograph for well CSVM-5 in Coyote Spring Valley and well BM-ONCO-2 in the Black Mountains Area (see Appendix B). Both of these wells are completed to depths of 1,600 feet or more. In the case of CSVM-5, which is completed in carbonate bedrock, water levels were unaffected by pumping lower on the valley floor, which may be the result of this well being isolated in a higher-elevation structural block on the eastern flanks of the Sheep Range, as suggested by some of the stakeholders in their data reports. The water level record for this well has steadily increased (nearly 8 feet) for the most part from 2003 – 2019, and clearly shows the influence from the 2004-05 wet year in its record, similar to many other monitoring wells within Coyote Spring Valley and elsewhere in the

LWRFS basins, which largely exhibit a declining water level trend during this same period of time. In the case of BM-ONCO-2, which is technically located outside of the existing LWRFS boundary, the water level has steadily increased (> 14 feet) over a similar period of time. This well is completed in clastic bedrock.

Wells in three different valleys (Three Lakes Valley, Las Vegas Valley and Tikapoo Valley) on the west-southwest side of the Sheep Range (see Slide 27, Appendix A) corroborate the rising groundwater levels observed in CSVM-5 on the east side of the Sheep Range. These wells are all completed in the carbonate aquifer associated with the Death Valley Flow System, and show continuous rising water levels of about 2 to 4 feet over the last 10 to 20 years.

East of the LWRFS area, hydrographs for several wells located in the far southern end of Lower Meadow Valley Wash also have been exhibiting rising groundwater levels for many years. This includes wells EH-6, EH-8a, NPC-2, NPC-4a, NPC-5 Old, TH-8, TH-12, TH-31 and TH-35. Available well logs indicate that most, if not all, of these wells are completed in basin fill sediments. Water levels in these wells largely have been increasing since the late 1980s – early 1990s through the present, many on the order of 15-20 feet. Additionally, most of these well hydrographs show clear seasonal pumping signals throughout this period of rising water levels, which is not surprising as many of these wells were used to provide water to the former Reid Gardner power plant. While many of these wells were pumped, their water levels continued to rise throughout much of this period, indicating one or more sources were contributing recharge flow to this area, and that pumping over this period was not of sufficient magnitude to overwhelm the rising water level response during this period. One possible recharge source could include the Mormon Mountains to the northeast. It is noteworthy that groundwater levels consistently rose in this area of pumping during this period of time. Because of the observed responses to pumping, it is possible, although unlikely, that the rising water levels may be totally in response to decreases in pumping prior to the period of record.

Further to the northeast in the Tule Desert basin, groundwater levels in nearly all of the wells have been steadily rising for much of the last 10 years, as much as 15 feet in some cases. Wells exhibiting this behavior include FF-1, FF-2B, PW-1, PW-2, MW- 1S, MW-1D, MW-2S, MW-2D, MW-4, MW-5, MW-7 (when not used for periodic stock pumping), MW-8 and MW-10. These wells are predominantly completed in carbonate bedrock, but are also completed in clastic bedrock, volcanic bedrock and basin fill sediments.

North of the LWRFS area, there is some evidence of rising water levels occurring in Dry Lake Valley, Coal Valley and Garden Valley. Wells exhibiting rising water levels are completed in the carbonate aquifer and basin fill aquifer in these valleys. Depending on the well, the water level rise over the last 30 years in has ranged from about 7 to 14 feet.

Based on the hydrograph data collected and discussed above, we believe that the rising trend in groundwater levels observed by the USGS in many of the basins distant from pumping in the Death Valley Flow System is also observed in several basins in the southern portion of the White River Flow System that tend to be distant from pumping as well. In the LWRFS where pumping has been occurring throughout much of this period of rising groundwater levels in surrounding non-pumped basins, groundwater levels have been on a decline during much of this same period. This indicates to us that the groundwater pumping that has been occurring in the LWRFS has been of sufficient magnitude to overwhelm the rising water level response that likely would have been observable in the LWRFS basins in the absence of any pumping. Even though a significant reduction in alluvial pumping in the MRSA since 2015 has resulted in noticeable recovery of groundwater levels and spring discharge in the MRSA, continued pumping at current levels still appears to be limiting (or extending the period to) full recovery from the pumping effects observed from the Order 1169 pumping test.

If the water level rises observed in the aforementioned White River Flow System basins truly represent the aquifer response to frequent wet year recharge that has been occurring from 1970 to the present in southern Nevada, as advanced by the USGS for the Death Valley Flow System, then these responses also may represent decadal-scale (short-term) water level fluctuations that are part of a century-scale (long-term) period of steady state. Therefore, climate variability and effects viewed on these scales may be a significant factor in the degree of water level declines (and rises) we might expect to observe in basins where pumping is occurring. The fact that water levels have been rising in these distant basins during ongoing drought conditions clearly suggests that significant aquifer lag times can be expected for decadal-scale recharge pulses (and decadal-scale drought impacts) to dissipate and turnover.

There are a couple of water management implications that arise from this view of a long-term, century-scale steady-state response that is imbedded with short-term, decadal-scale periods of wet and dry period responses. First, if southern Nevada has been undergoing several decades of rising water level conditions as indicated by measurements in many wells in the area of interest, and groundwater levels in the LWRFS basins have been declining, then this decline can only be explained by pumping in the LWRFS basins and not by the current drought conditions we appear to be experiencing. While the drought effects are real, they may not manifest their effects in the aquifer for years or decades according the working model proposed. The USGS indicated in their presentation that there is evidence that the rising water level trends may be plateauing (dissipating) which may be signaling a turnover to a decadal-scale dry period in which groundwater levels can be expected to decline in basins distant from pumping.

Second, if the current pumping levels are sufficient to overwhelm the decadal-scale recharge pulse that has been traveling through the aquifers of the LWRFS and create the observed water level declines, then there are larger management concerns awaiting when this recharge pulse dissipates and turns over to a decadal-scale drier period where aquifer recharge will be significantly reduced for extended periods. If pumping levels are maintained at the same levels that have been causing the recent declining water level trend, then this declining trend is going to be magnified not only on groundwater levels, but also on spring and river discharges in the Muddy River Springs Area and elsewhere in the LWRFS. Conjunctive management in the LWRFS should factor in long-term monitoring of groundwater levels in several basins distant from pumping in the LWRFS basins to gauge the real-time climatic response being transmitted through the aquifers in southern Nevada so that adjustments to pumping can be made accordingly to manage against excessive declines in groundwater levels, and spring and river discharges.

In summary, water-level monitoring shows that water levels are rising in areas with limited pumping, and declining in areas where water is being pumped. The SeriesSEE evaluation performed by the U.S. Fish and Wildlife Service included pumping from 39 major pumping wells, which produced declines in water levels over the period of analysis. It did not include the effects of rising levels throughout southern Nevada. As a result, the evaluation underestimated the effects of the long-term pumping, but did capture the effects of pumping MX-5, because it's pumping began and ended during the analysis period and because of the rapid response in water levels that occurred.

9. **Las Vegas Valley Shear Zone** - Some reports suggest that there is significant flow across the Las Vegas Valley shear zone into Las Vegas Valley. The shear zone has long been considered as a permeability barrier to flow perpendicular to its strike, because of the presence of springs at Corn Creek and Indian Springs, and the increase in hydraulic gradients near the shear zone. In the area of interest, hydraulic-head maps typically show flow to the southeast parallel to the shear zone on both sides of the shear zone, rather than across it. Water production in Las Vegas Valley has been from wells completed in basin-fill materials so that drawdown has not spread like it does in the carbonate aquifer. There are few wells near the shear zone for defining hydraulic gradients across it, and for detecting water-level changes caused by groundwater pumping. Thus, rates of flow across it are poorly known.

Estimates of flux across the shear zone that have been recently presented are based on water balance estimates, which are very uncertain. Several of these estimates are being used to argue that water is available for pumping within the LWRFS. We would like to point out that if there is significant flow across the shear zone that is "available" for production, water users in Las Vegas Valley are likely to be impacted.

10. **Available Water** – Several of the parties revisited the water budget of the LWRFS. Without doubt, knowing fluxes of water into, within, and out of the LWRFS is useful. But it is more important to have information on the hydraulic connections within the LWRFS and across its boundaries in order to know which areas (and thus uses) will be affected. The Order 1169 test demonstrated that within the carbonate aquifer, the system is well connected. In basin-fill aquifers, pumping typically removes water from aquifer storage in the aquifer pores or through compaction in the vicinity of the wells being pumped, and the area of drawdown is limited. In the carbonate aquifer, the storativity of the aquifer is small, and as a result, the area of impact of pumping is large. Thus, pumping in one area will impact other users long distances away, if the areas are hydraulically connected.

    If more water is available, on the basis of the water budget, than currently believed, the problem is how to manage that water without impacting other users. In the carbonate aquifer, this is a question that has a simple answer. Multi-year pumping tests are needed to test the degree of connectedness. The pumping stress must have a different temporal pattern than the seasonal pumping pattern that is prevalent in the aquifer in order to differentiate the effects. The Order 1169 test was successful because the pumping was not seasonal and lasted more than a year. The rate of pumping must be high enough that effects can be measured. One of the parties suggested that pumping rates could be lower, but this may have been intended to hide the effects because of a low signal-to-noise ratio.

The apparent “surplus” of water is hypothesized as discharging to Las Vegas Valley across the shear zone. Additional monitoring wells are needed to measure effects of pumping in Garnet Valley within and south of the shear zone to determine whether capture of outflow across this boundary of the LWRFS administrative unit will occur. Similarly, monitoring is needed to provide early detection of any effects that would diminish flow at Rogers and Blue Point Springs.

REFERENCES

Halford, K., Garcia, C.A., Fenelon, J., and Mirus, B., 2012, Advanced methods for modeling water-levels and estimating drawdowns with SeriesSEE, an Excel add-In, (ver. 1.1, July, 2016): U.S. Geological Survey Techniques and Methods 4–F4, 28 p., https://dx.doi.org/10.3133/tm4F4.

Jackson, T.R., Fenelon, J.M., and Halford, K.J., 2017, “What Drought? Water Levels on the Rise in Southern Nevada”, February 16, 2017.

Phelps, G.A., Jewel, E.B., Langenheim, V.E., and Jachens, R.C., 2000, Principal Facts for Gravity Stations in the Vicinity of Coyote Spring Valley, Nevada, with Initial Gravity Modeling Results: U.S. Geological Survey Open-File Report 00-420.

Tetra Tech GEO, 2012, Development of a Numerical Groundwater Flow Model of Selected Basins within the Colorado Regional Groundwater Flow System, Southeastern Nevada (Version 1.0), 70 p. plus figures and 2 appendices.

URS, 2006, Final Well Completion Report, Kane Springs Valley, Lincoln County, Nevada.

Figures

TETRA TECH

SE ROA 51557

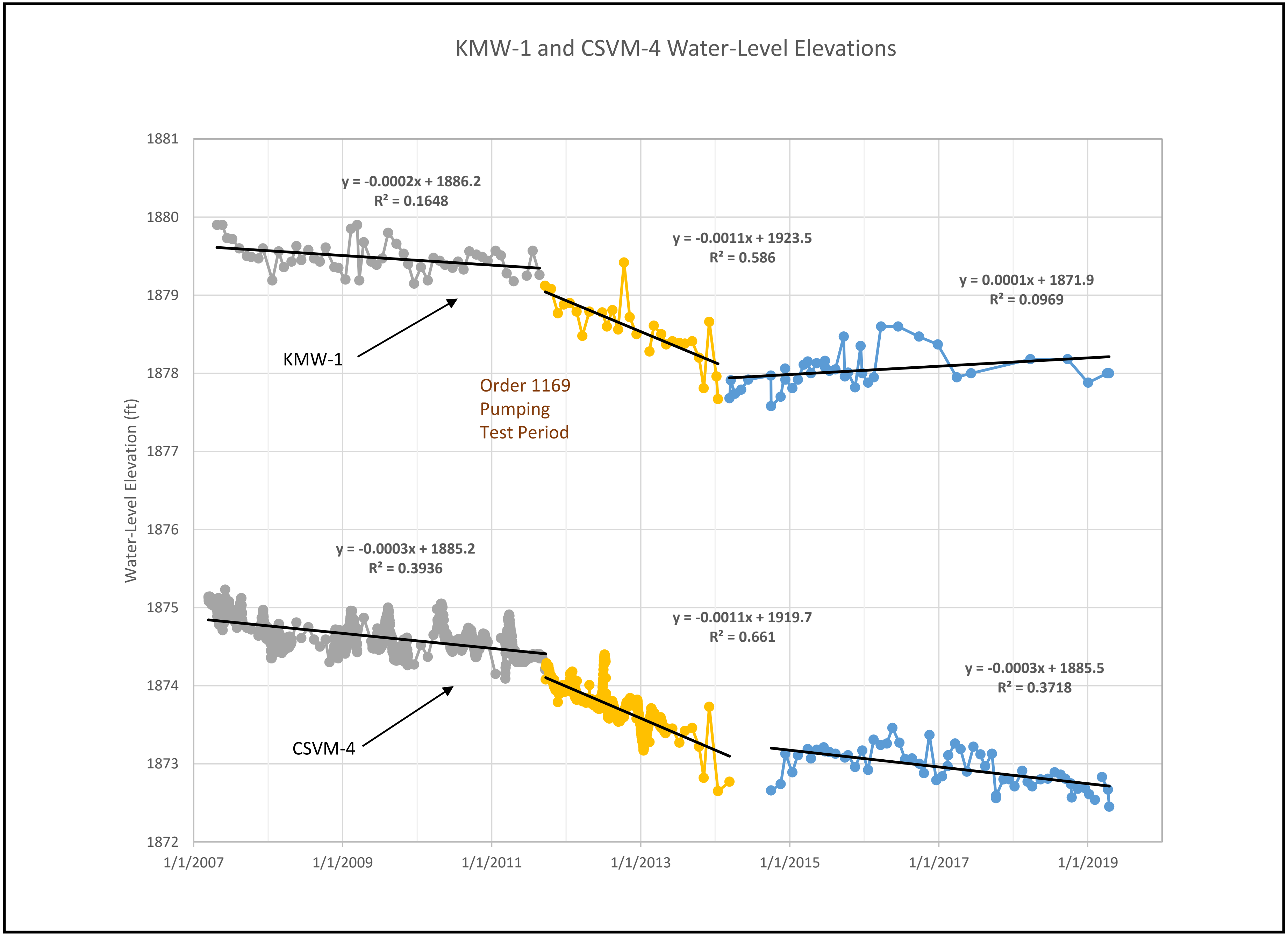


Figure 1

Hydrographs for Select Wells in
Coyote Spring Valley & Vicinity
8/16/2019

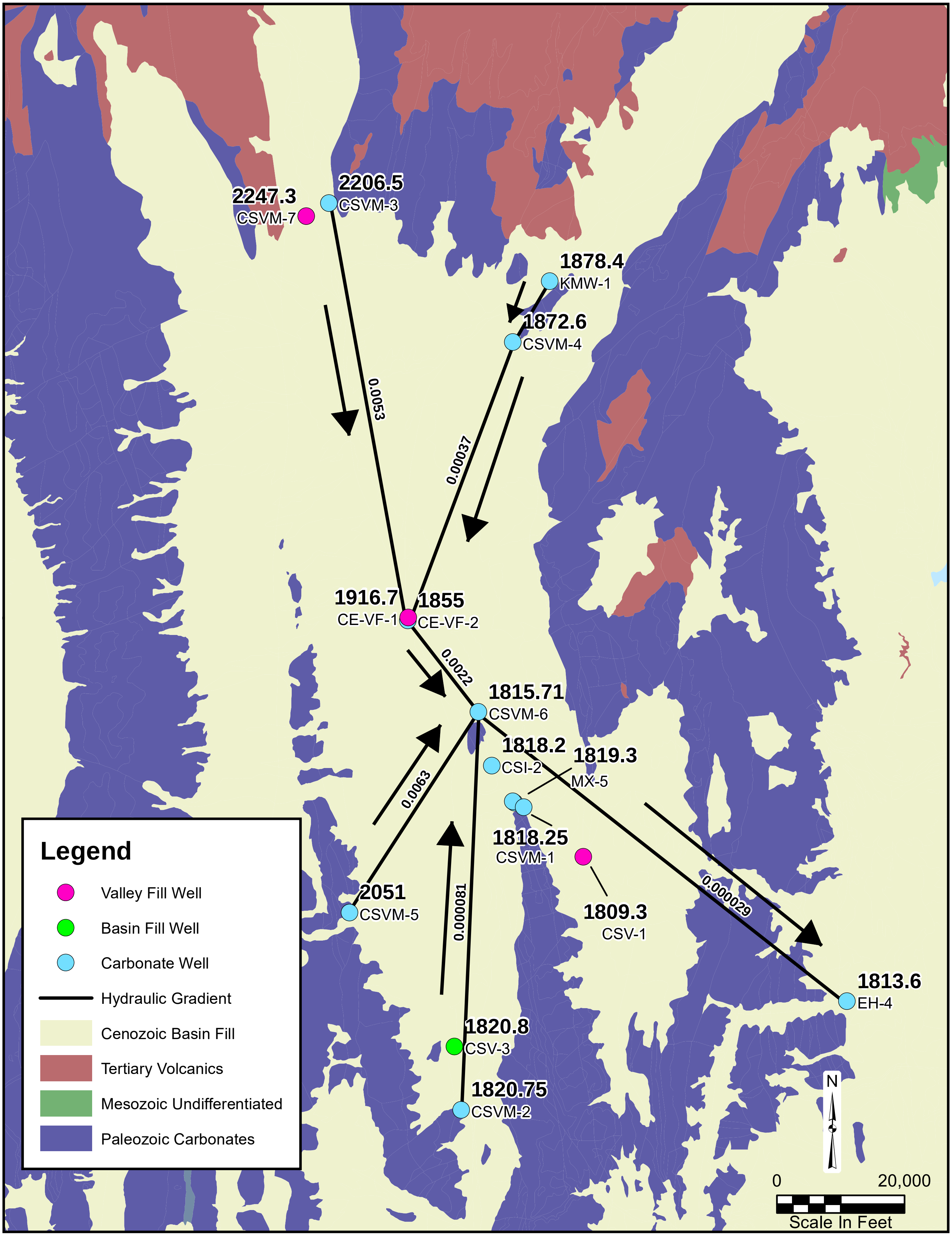


Figure 2
Horizontal Hydraulic Gradients Between Selected Well Points,
Coyote Springs Valley and Vicinity

SE ROA 51559

8/16/2019

JA_16690

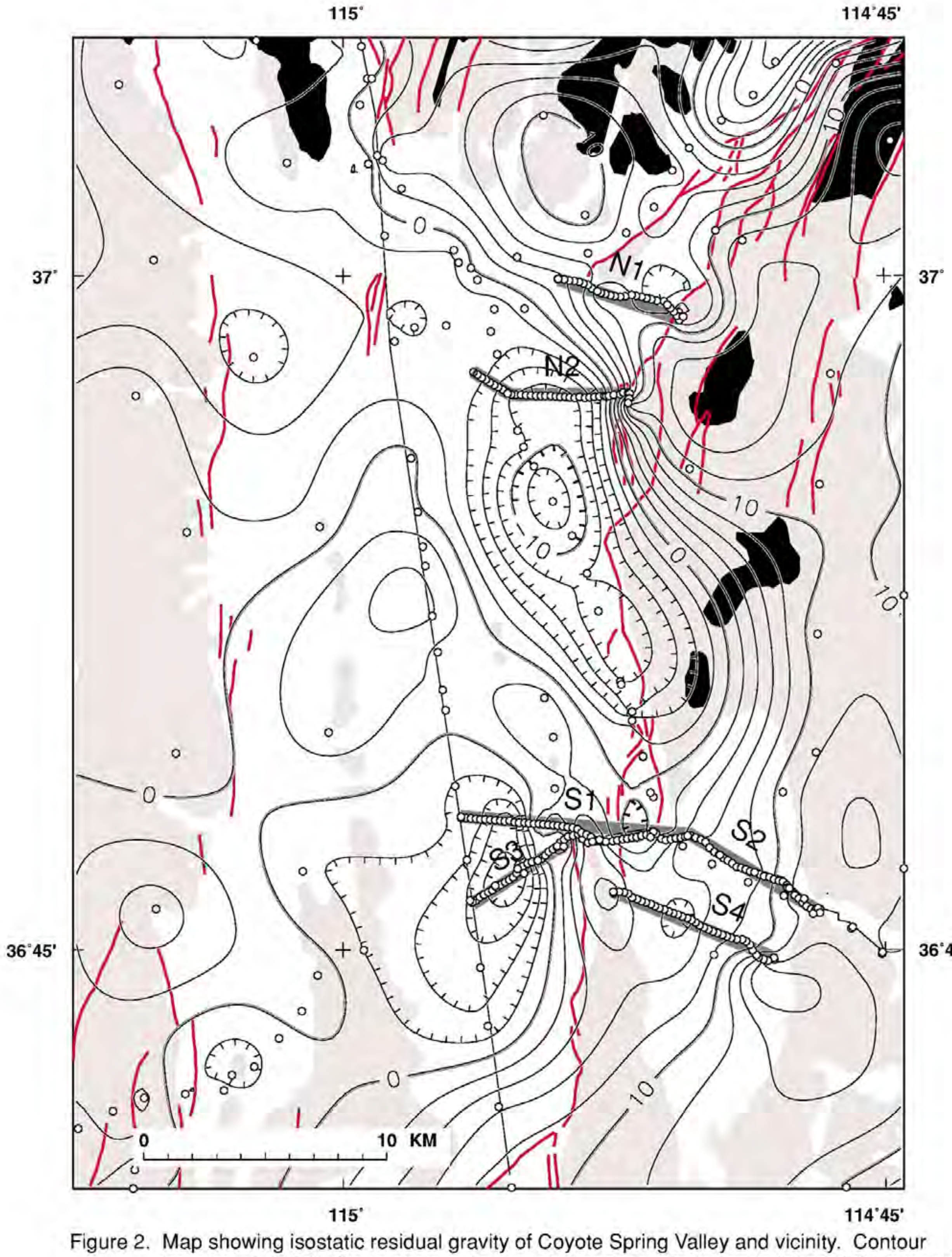


Figure 2. Map showing isostatic residual gravity of Coyote Spring Valley and vicinity. Contour interval = 2 mGal. Open circles show gravity stations. Gray bands labelled N1-N2 and S1-S4 are detailed gravity profiles that were modeled to define basin shape. Red lines indicate faults mapped by Dohrenwend and others (1996). See figure 1 for geology and culture. Refer to Plate 1 for larger scale presentation of these data.

Source: Phelps, G.A., Jewel, E.B., Langenheim, V.E., and Jachens, R.C., 2000, Principal Facts for Gravity Stations in the Vicinity of Coyote Spring Valley, Nevada, with Initial Gravity Modeling Results: U.S. Geological Survey Open-File Report 00-420.

Figure 3

Isostatic Residual Gravity in Coyote Spring Valley & Vicinity

8/16/2019

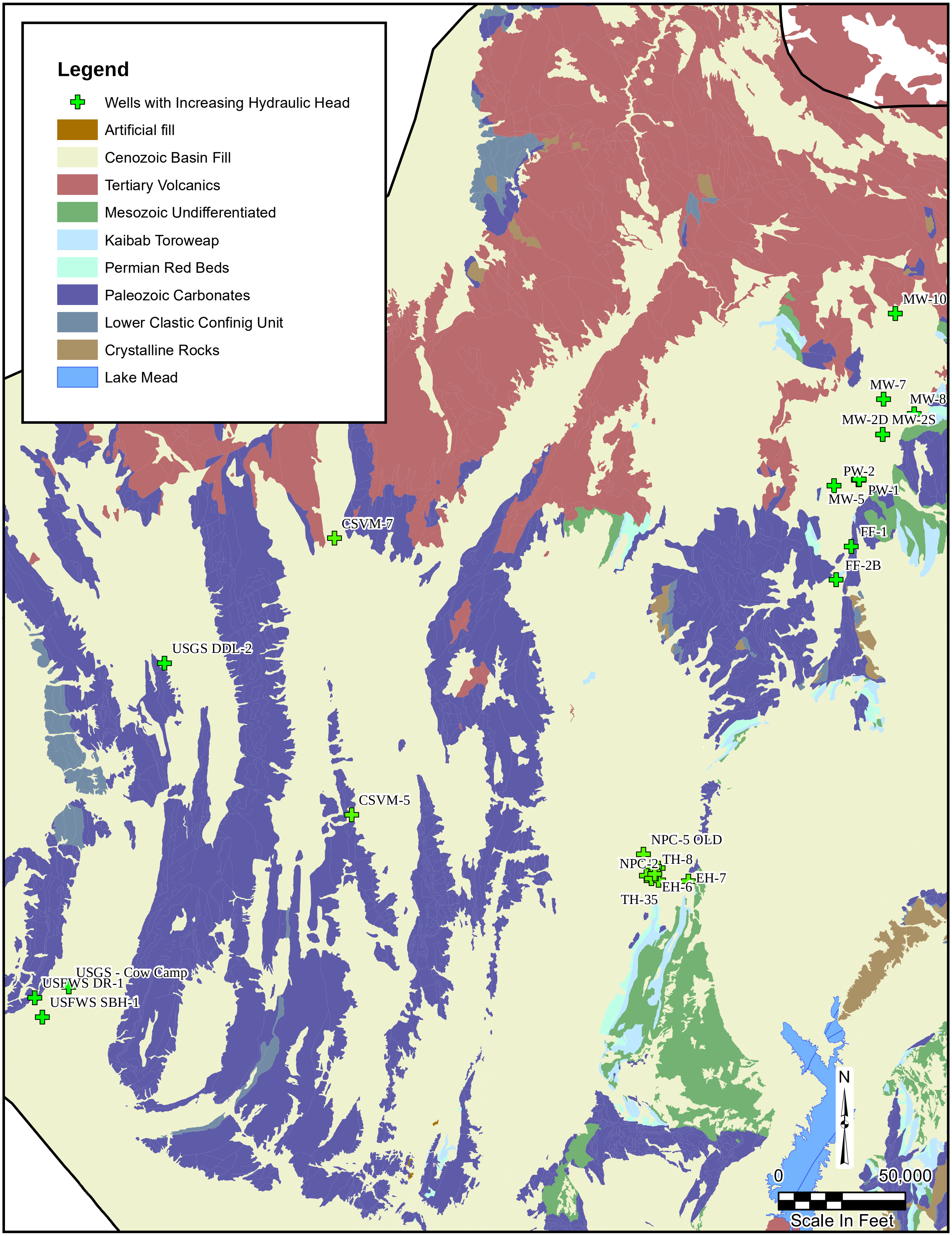


Figure 4
Wells with Rising Water Levels,
Coyote Spring Valley & Vicinity
8/16/2019

SE ROA 51561

JA_16692

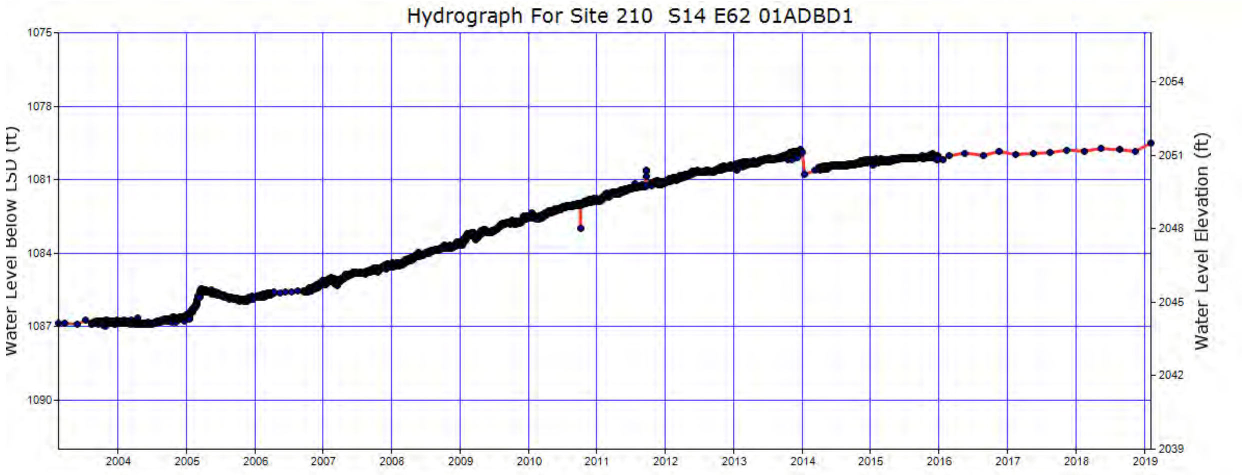
Appendix A

TETRA TECH

SE ROA 51562

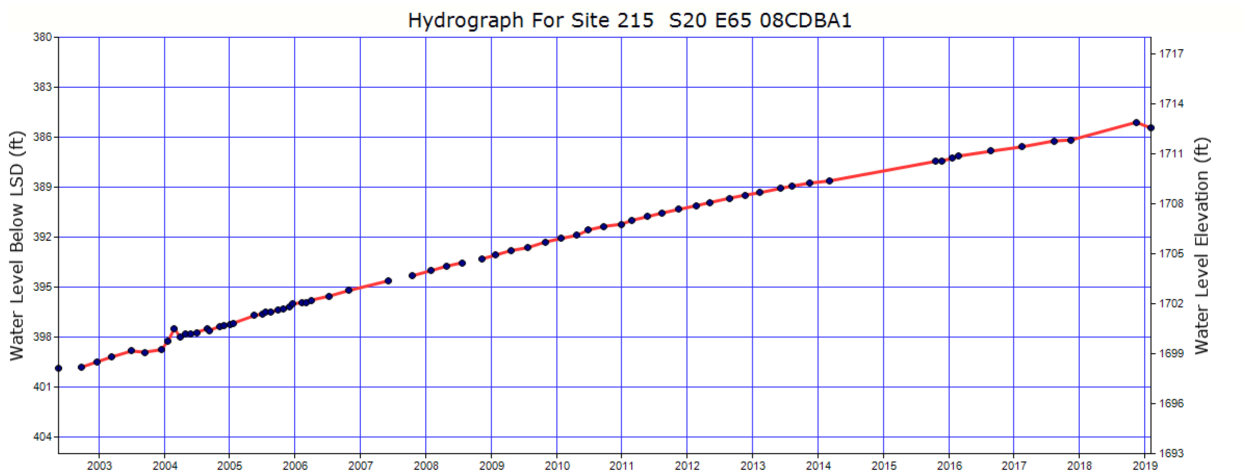
Coyote Spring Valley (Basin 210)

Well CSVM-5 (Carbonate Aquifer)



Black Mountains Area (Basin 215)

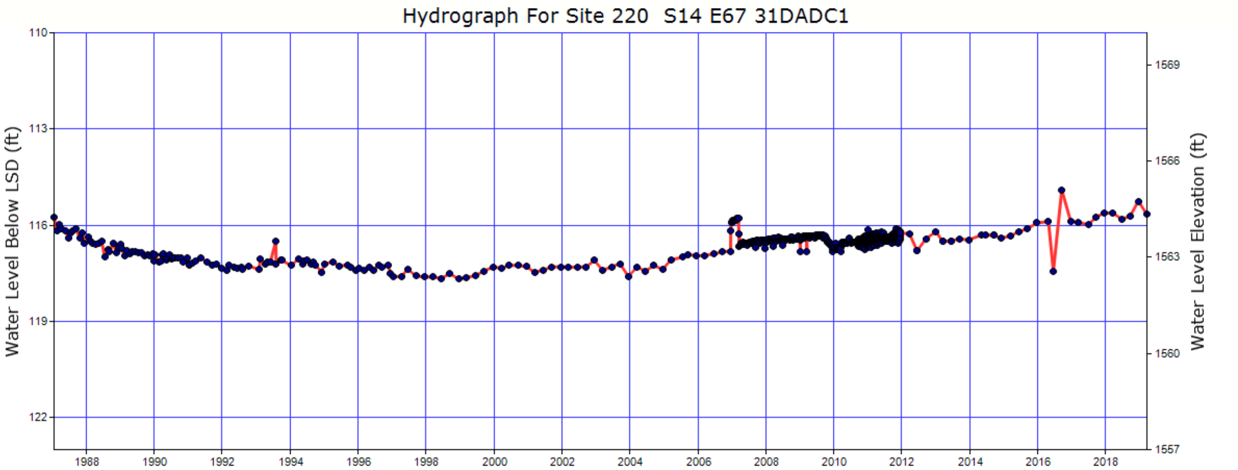
Well BM-ONCO-2 (Clastic Aquifer)



SE ROA 51563

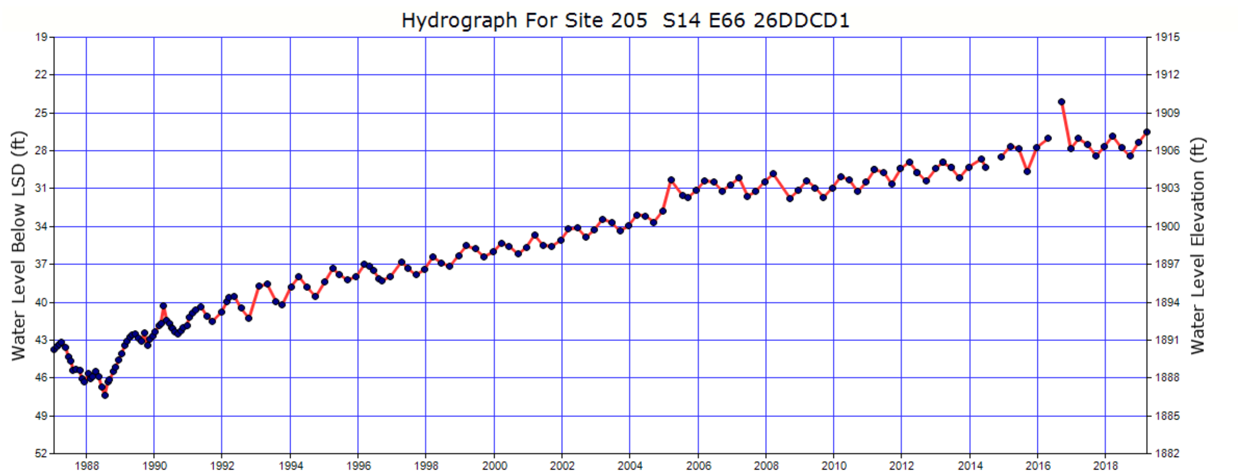
Lower Moapa Valley (Basin 220)

Well EH-7 (Carbonate Aquifer)

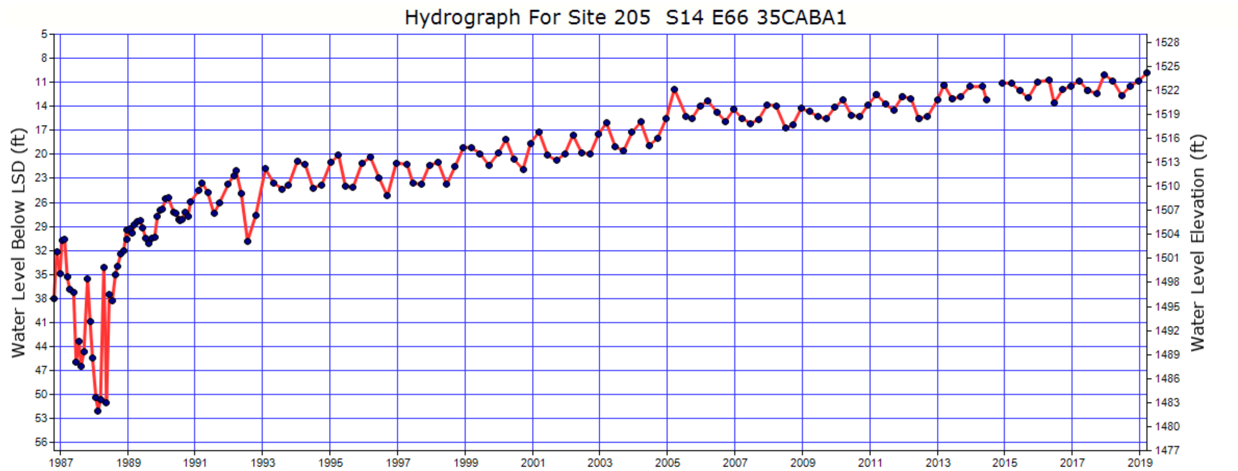


Lower Meadow Valley Wash (Basin 205)

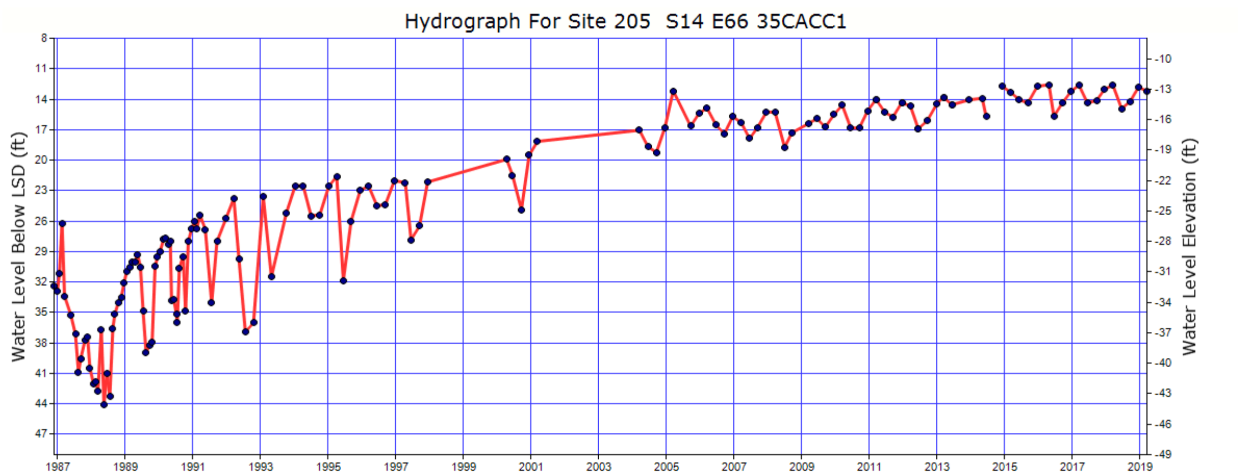
Well EH-6 (Basin Fill Aquifer)



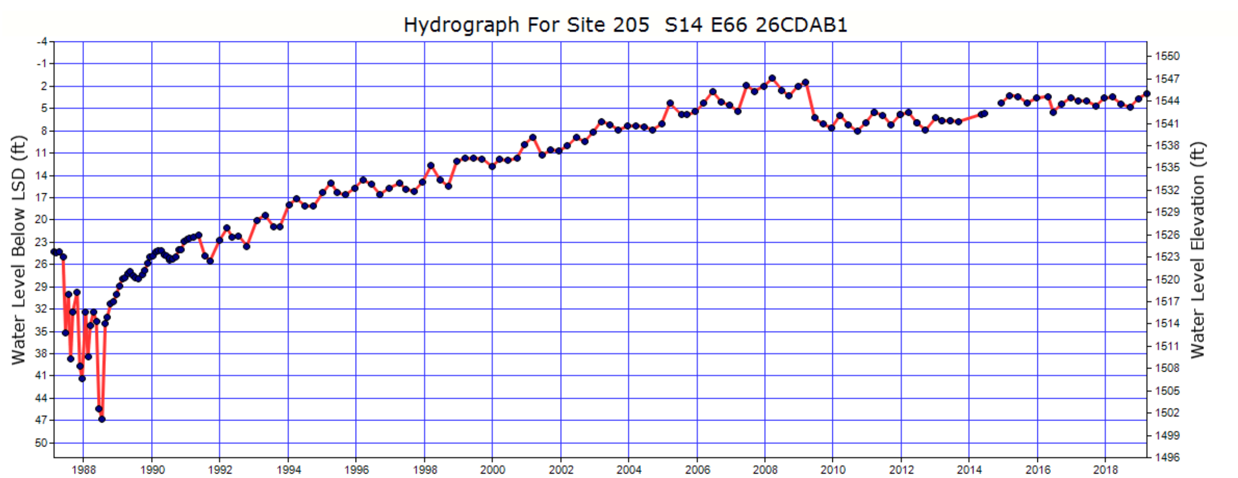
Well EH-8a (Basins Fill Aquifer)



Well NPC-2 (Basin Fill Aquifer)

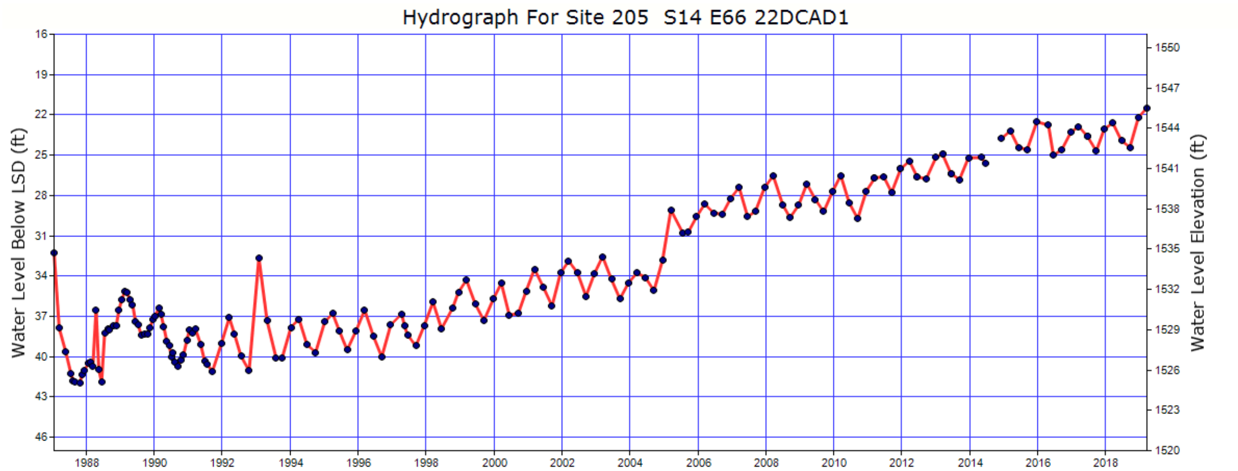


Well NPC-4a (Basin Fill Aquifer)

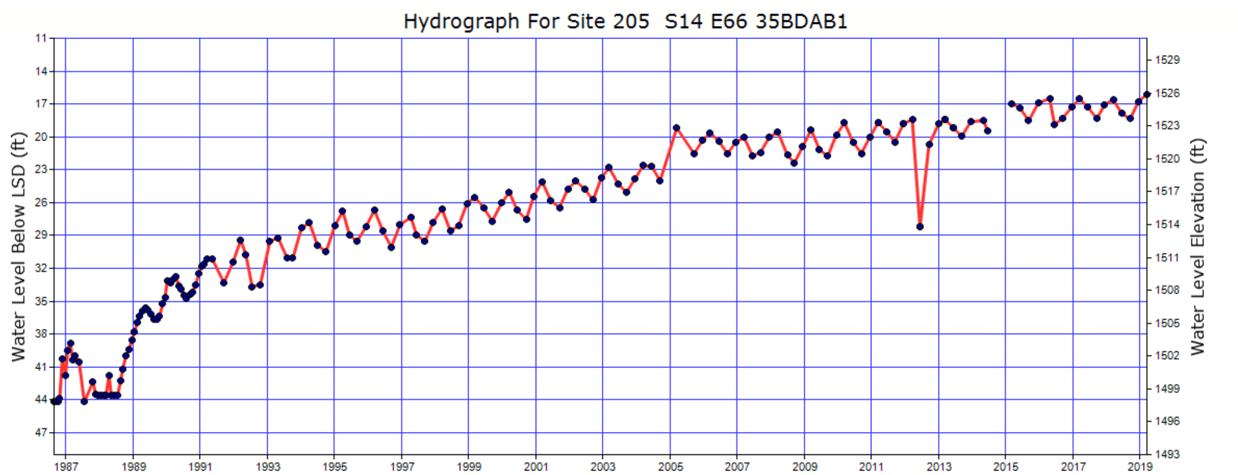


SE ROA 51565

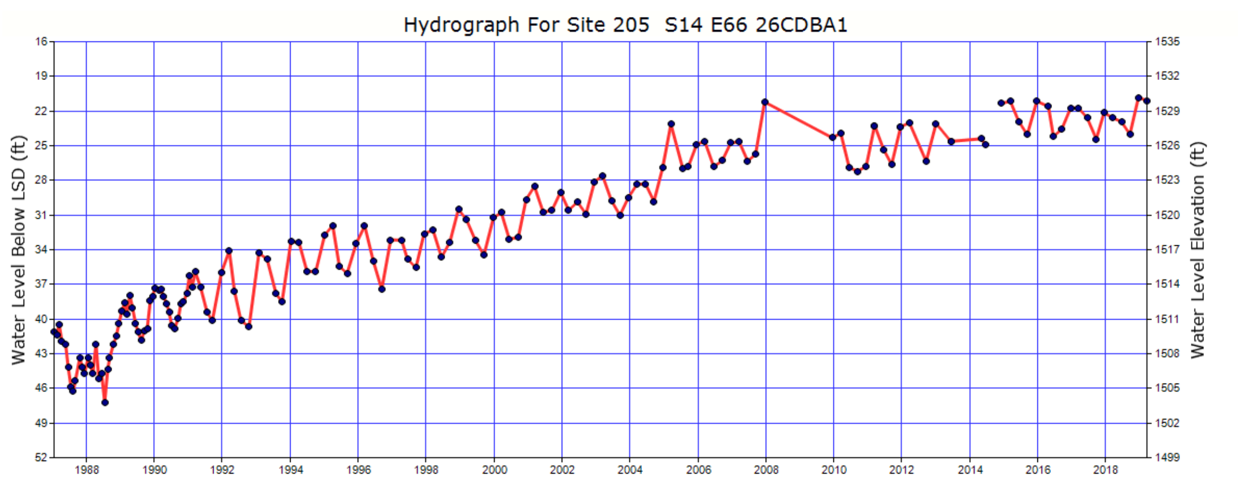
Well NPC-5 Old (Basin Fill Aquifer)



Well TH-8 (Basin Fill Aquifer)

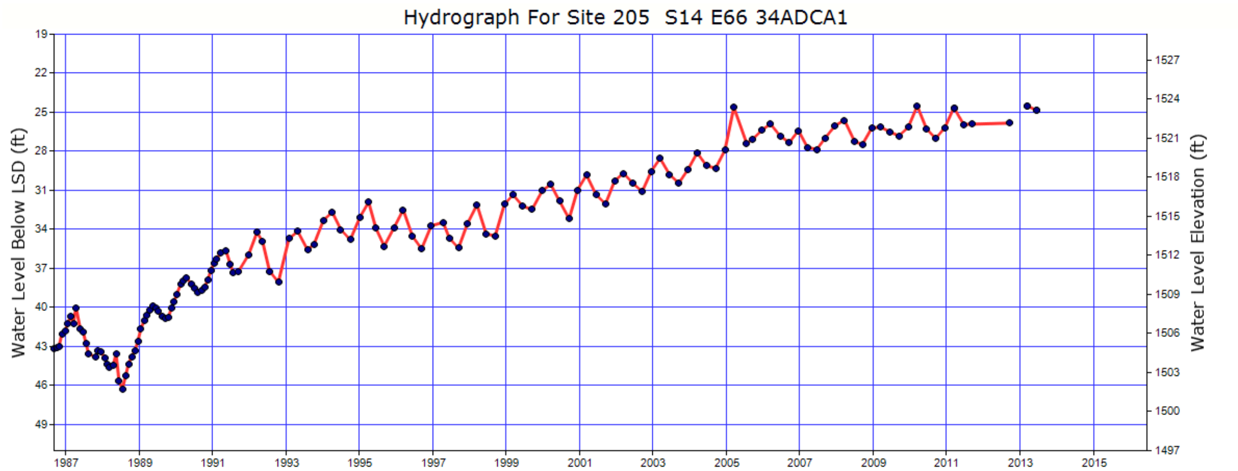


Well TH-12 (Basin Fill Aquifer)

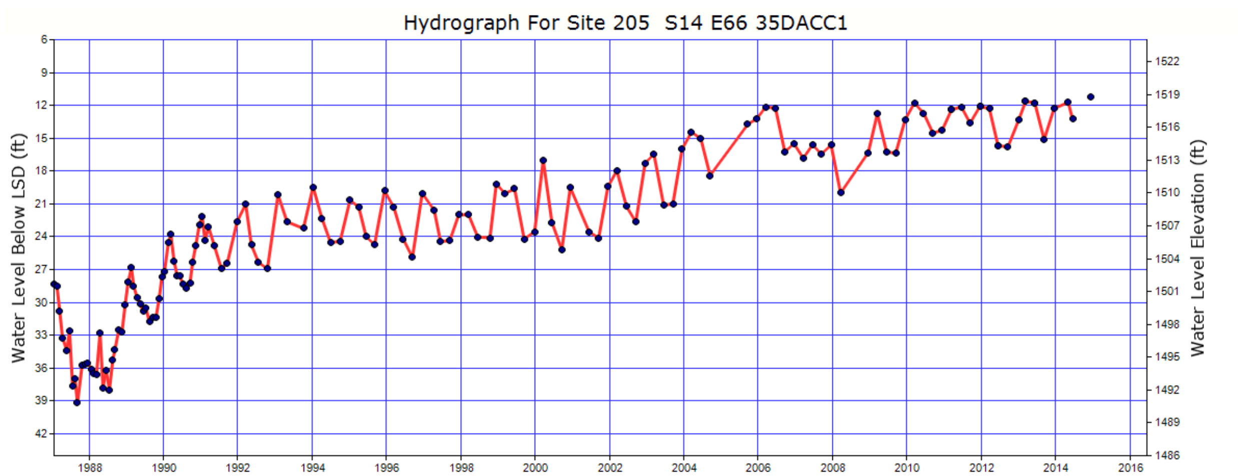


SE ROA 51566

Well TH-31 (Basin Fill Aquifer)



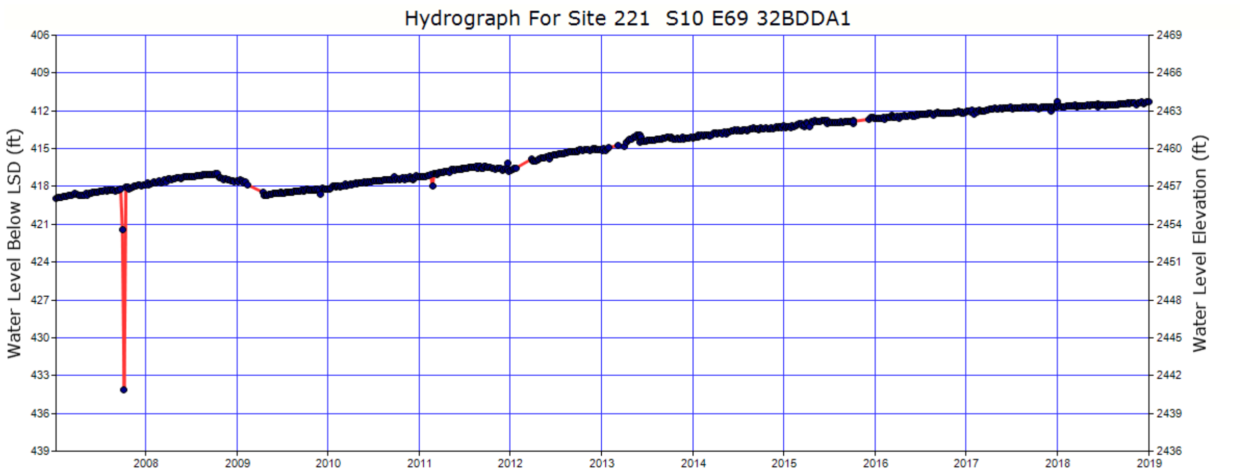
Well TH-35 (Basin Fill Aquifer)



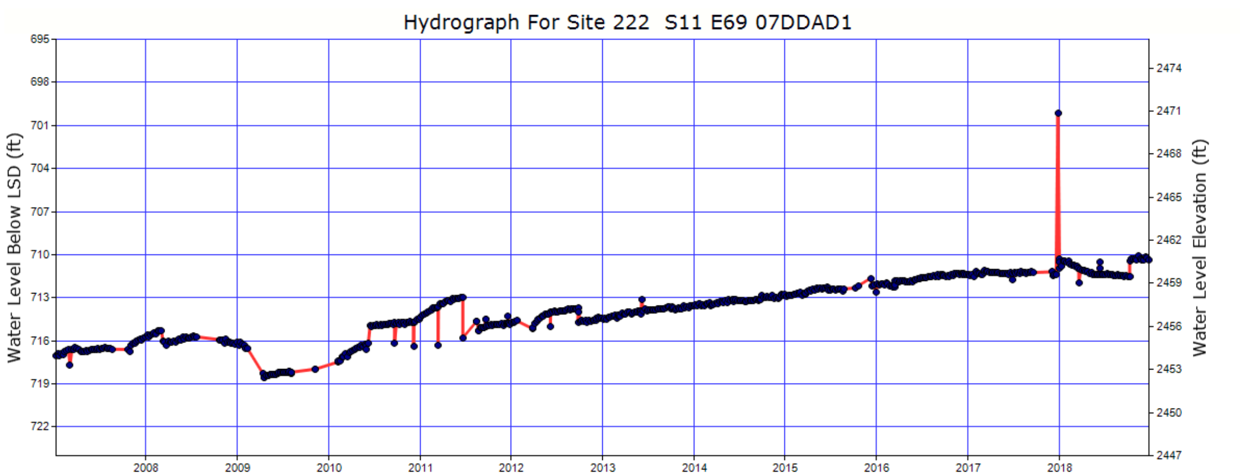
SE ROA 51567

Tule Desert (Basin 221) and Virgin River Valley (Basin 222)

Well FF-1 (Carbonate Aquifer)

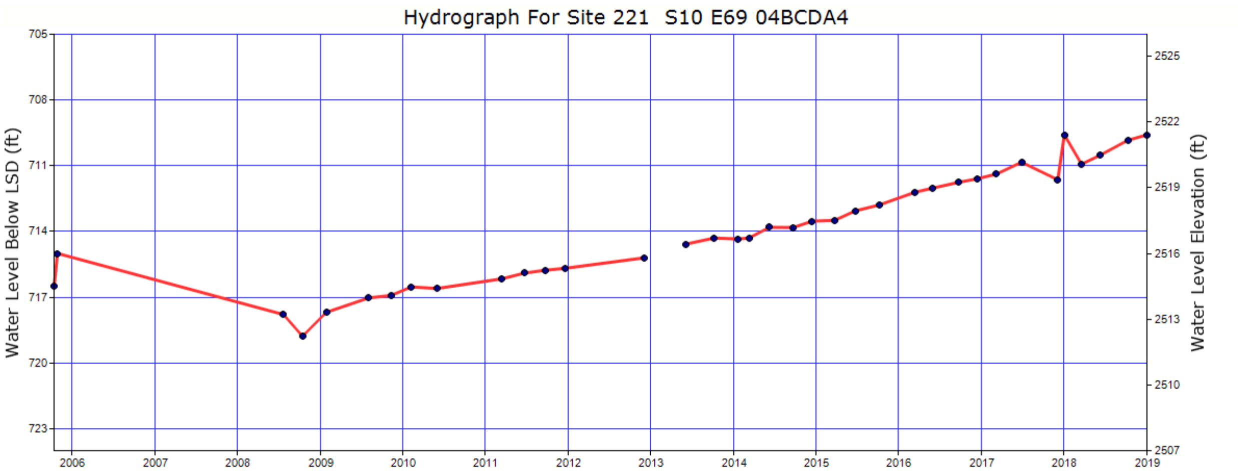


Well FF-2B (Carbonate Aquifer)

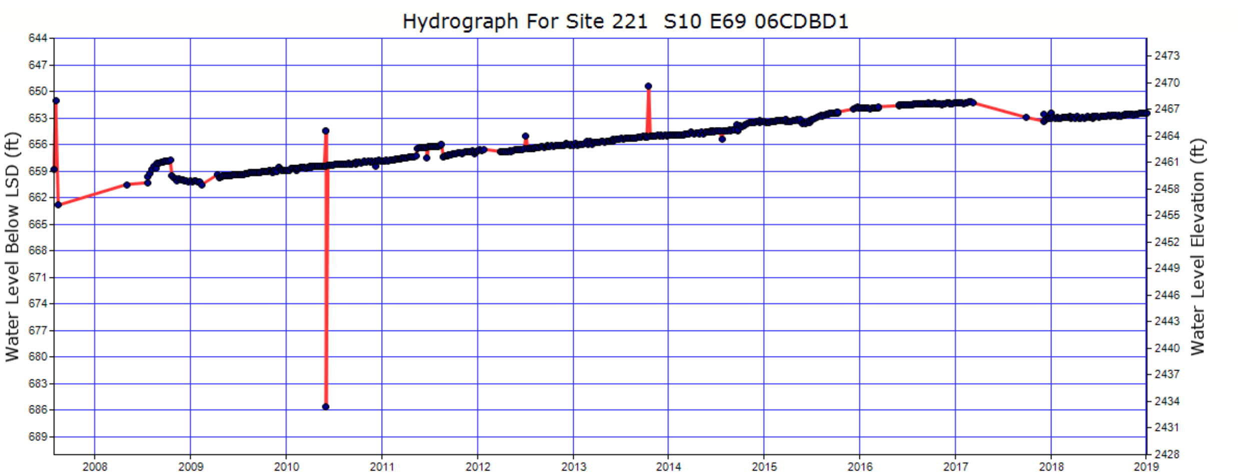


SE ROA 51568

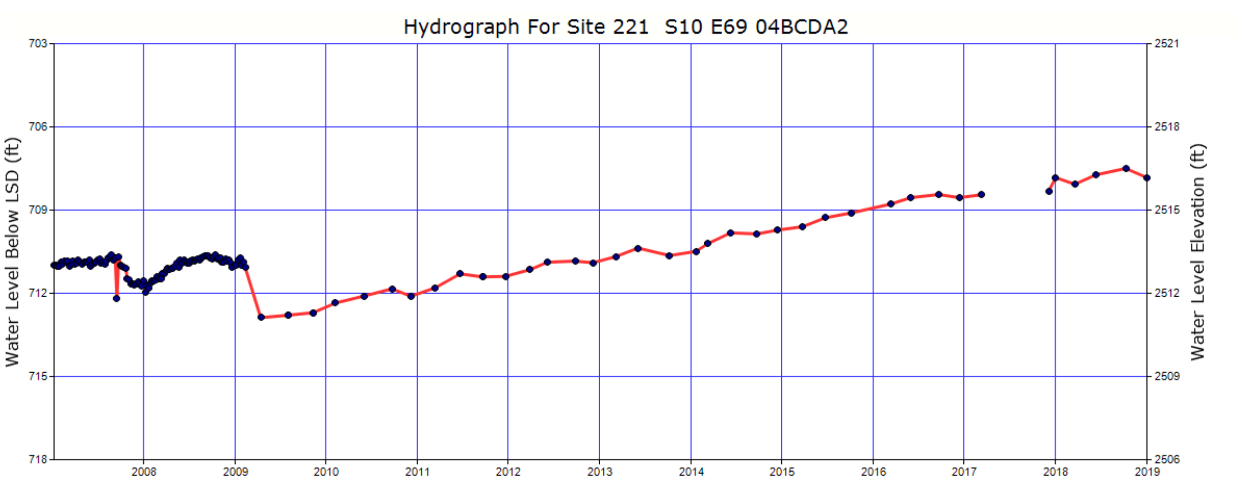
Well PW-1 (Carbonate Aquifer)



Well PW-2 (Carbonate Aquifer)

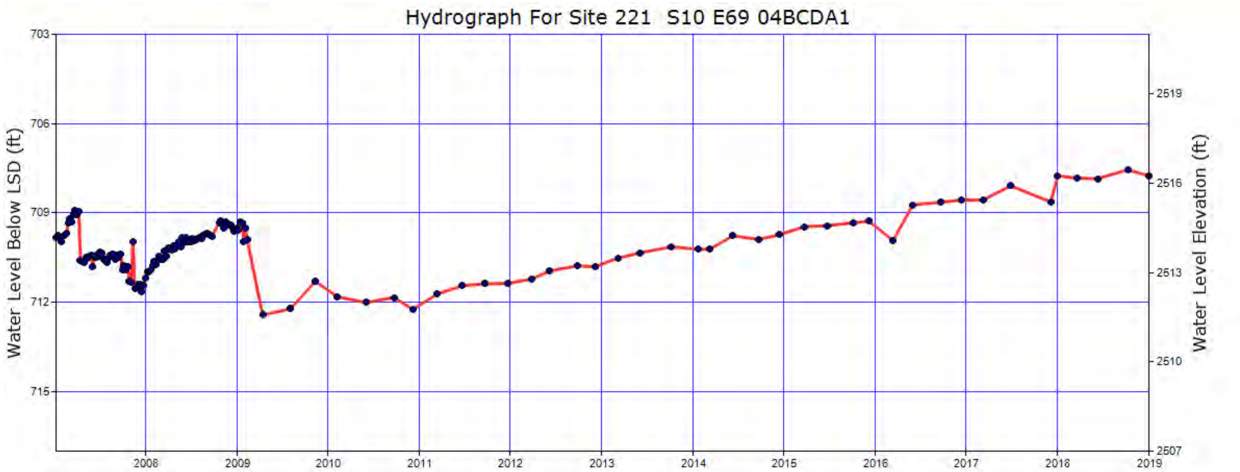


Well MW-1S (Basin Fill Aquifer)



SE ROA 51569

Well MW-1D (Carbonate Aquifer)



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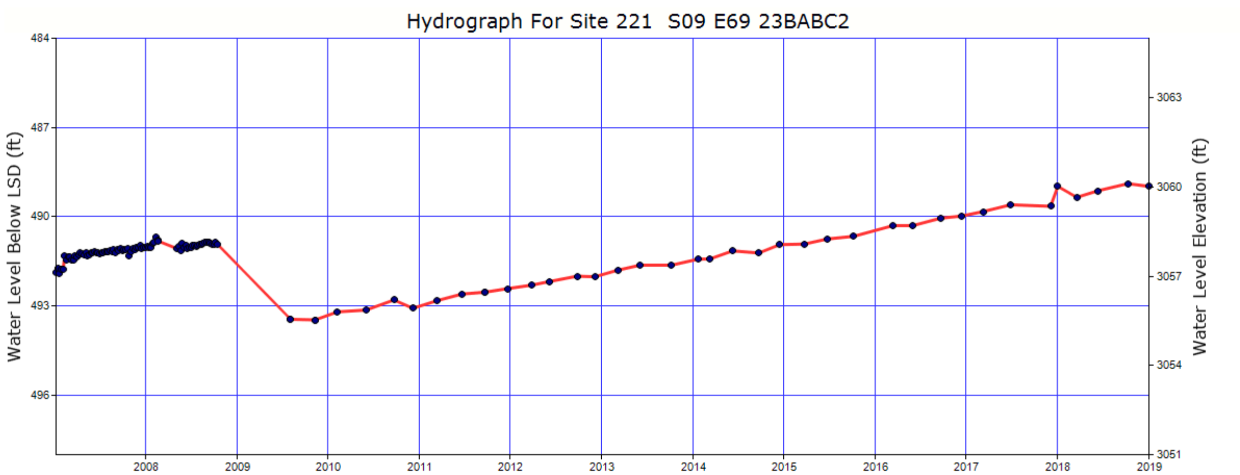

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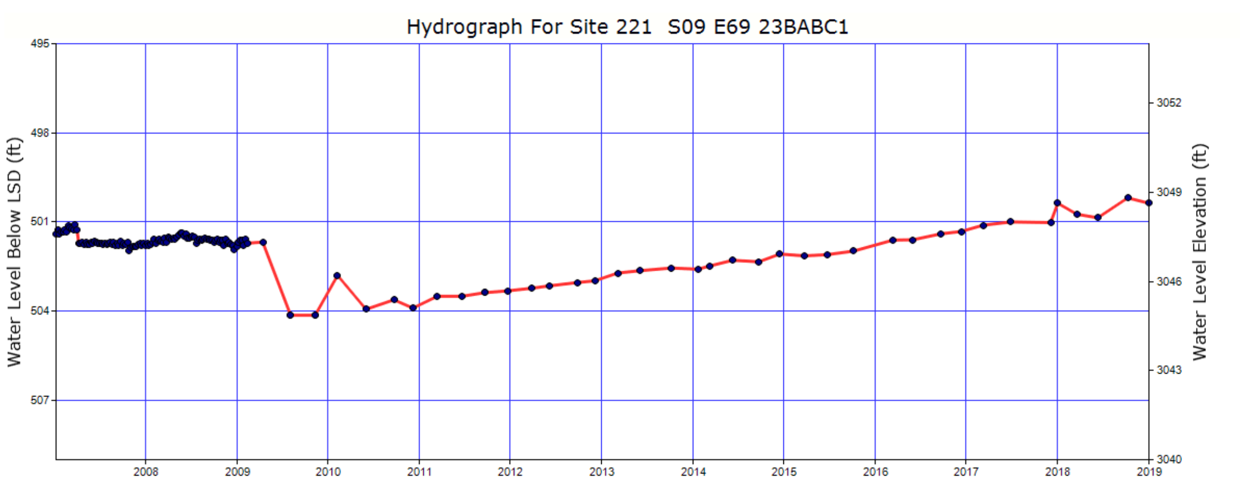
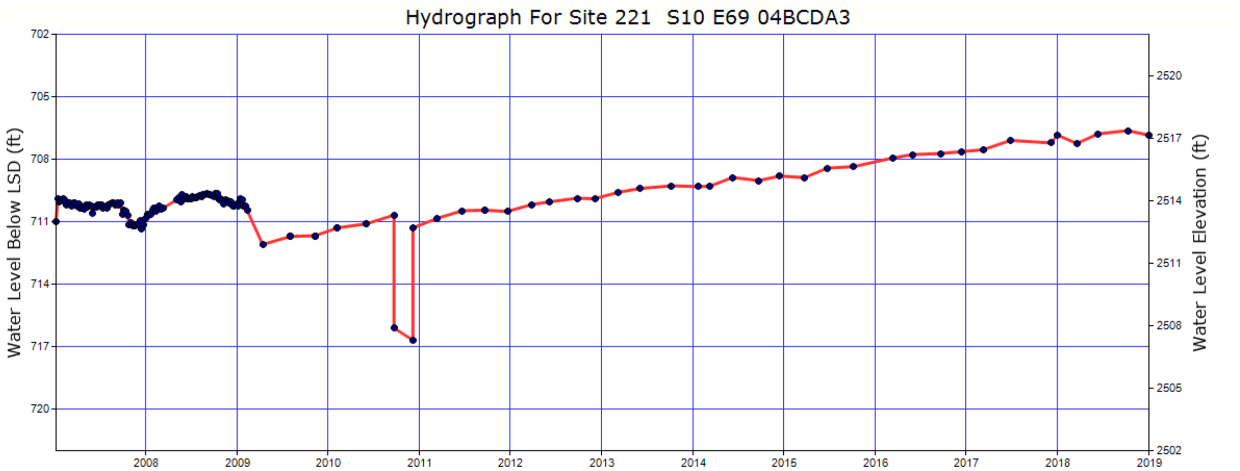
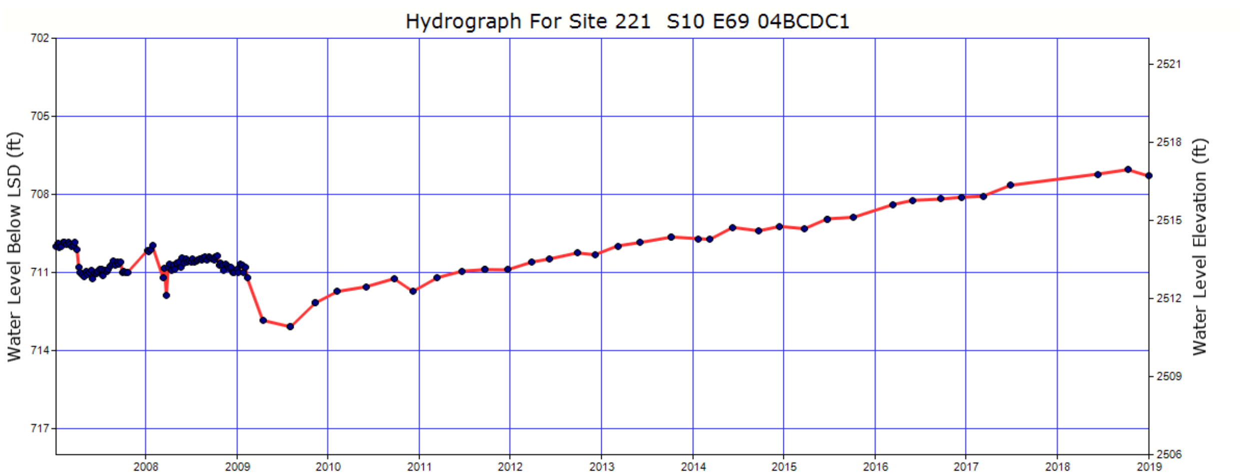


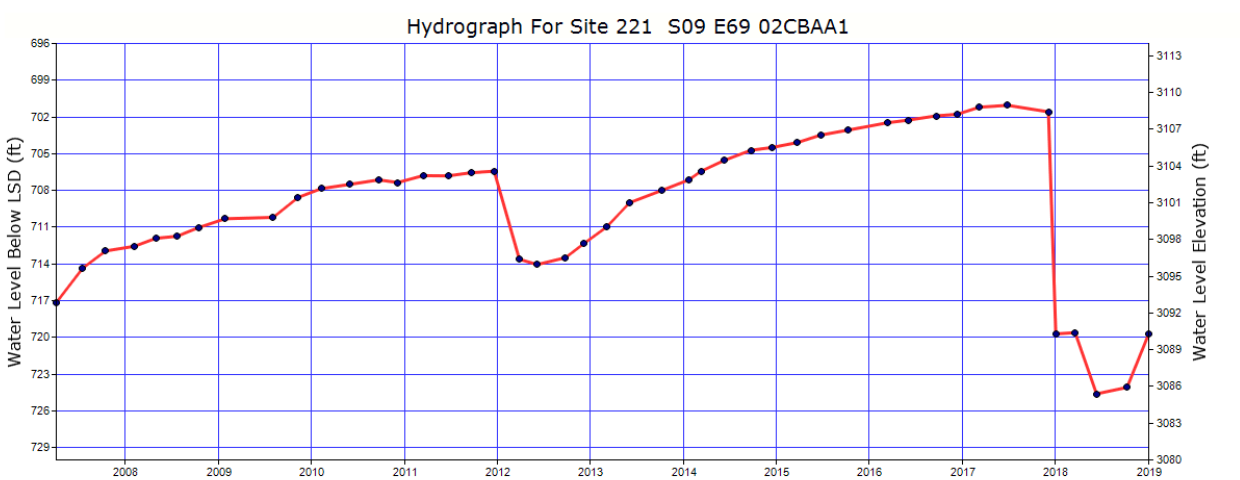
Well MW-4 (Carbonate Aquifer)



Well MW-5 (Basin Fill Aquifer)

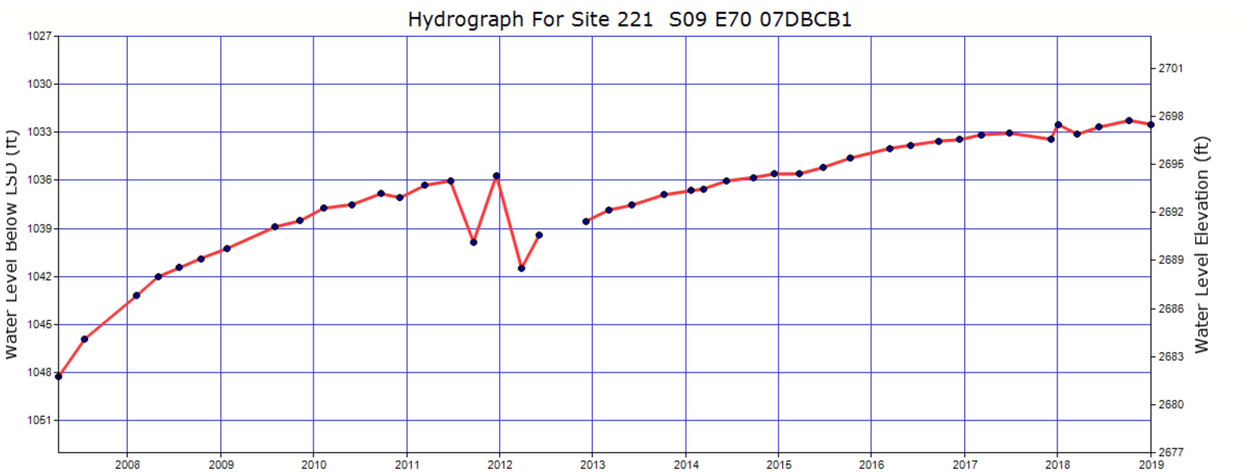


Well MW-7 (Carbonate/Clastic Aquifer)

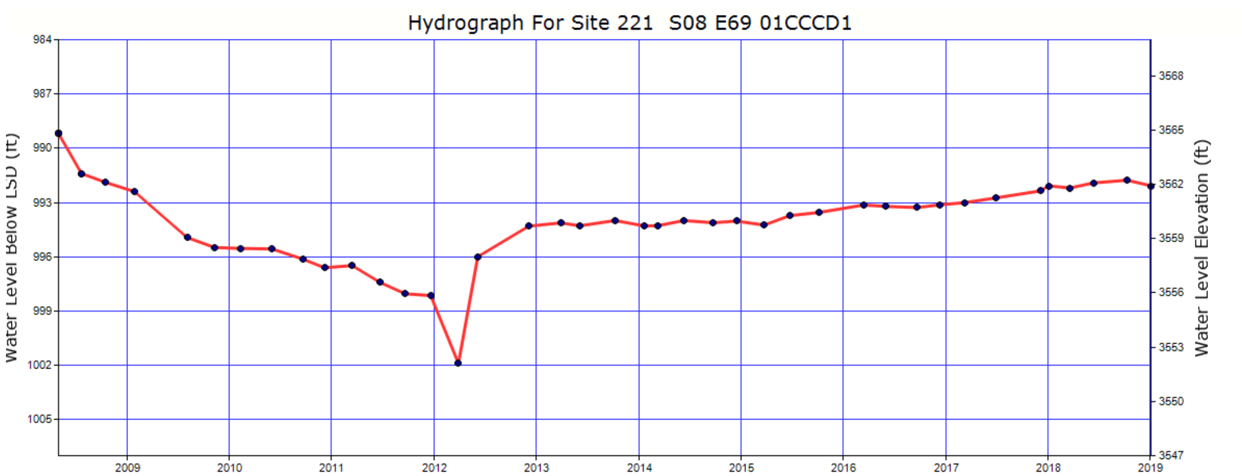


SE ROA 51571

Well MW-8 (Carbonate Aquifer)



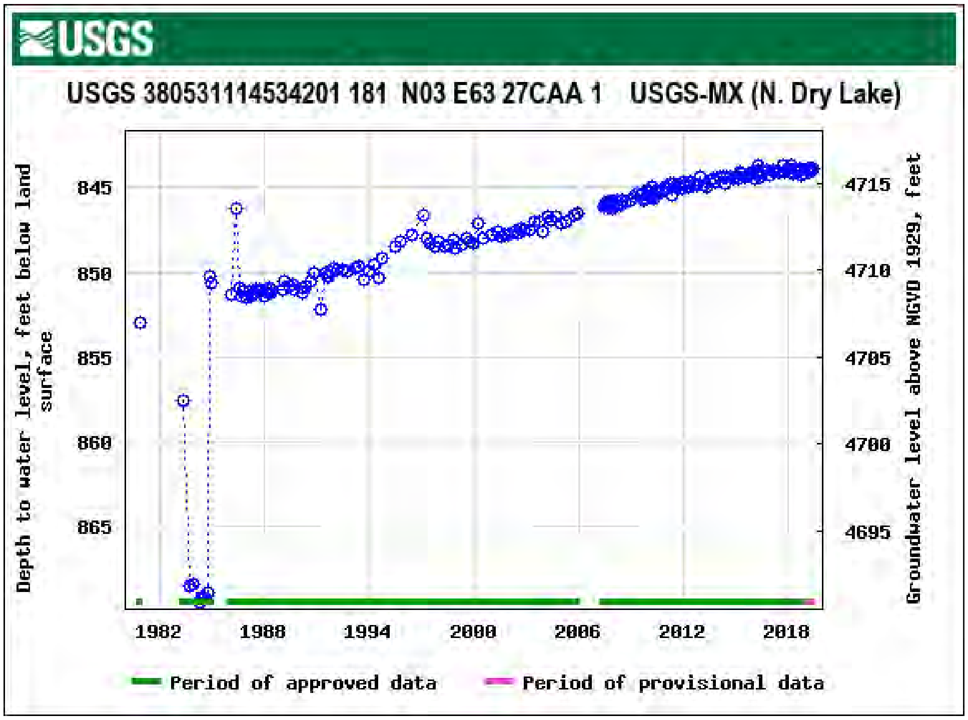
Well MW-10 (Volcanic Aquifer)



SE ROA 51572

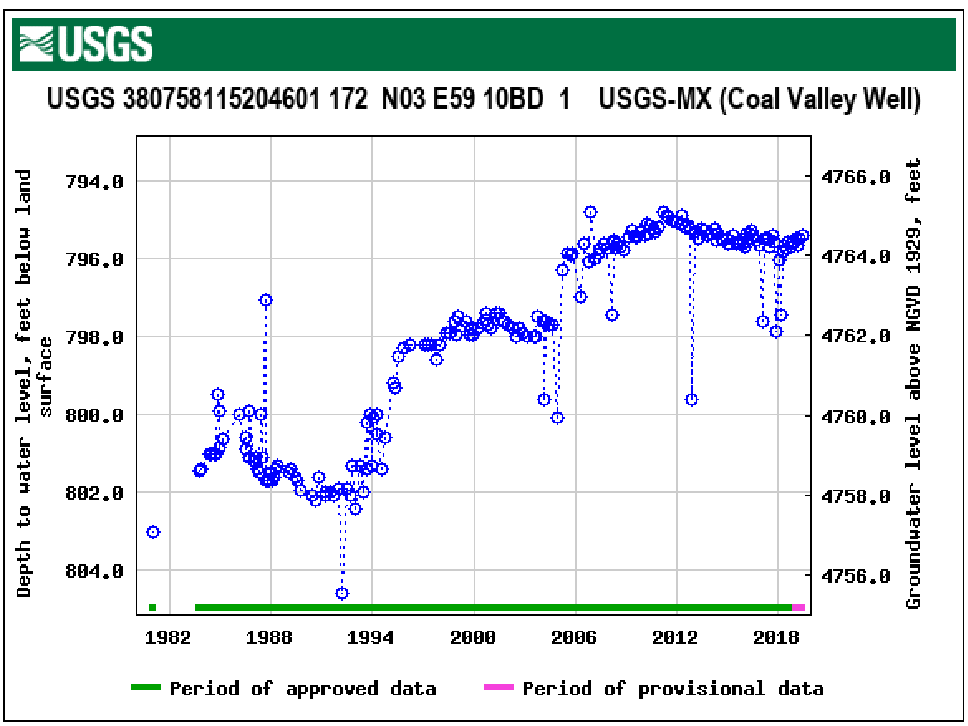
Dry Lake Valley (Basin 181)

Well USGS-MX N. Dry Lake (Carbonate Aquifer)



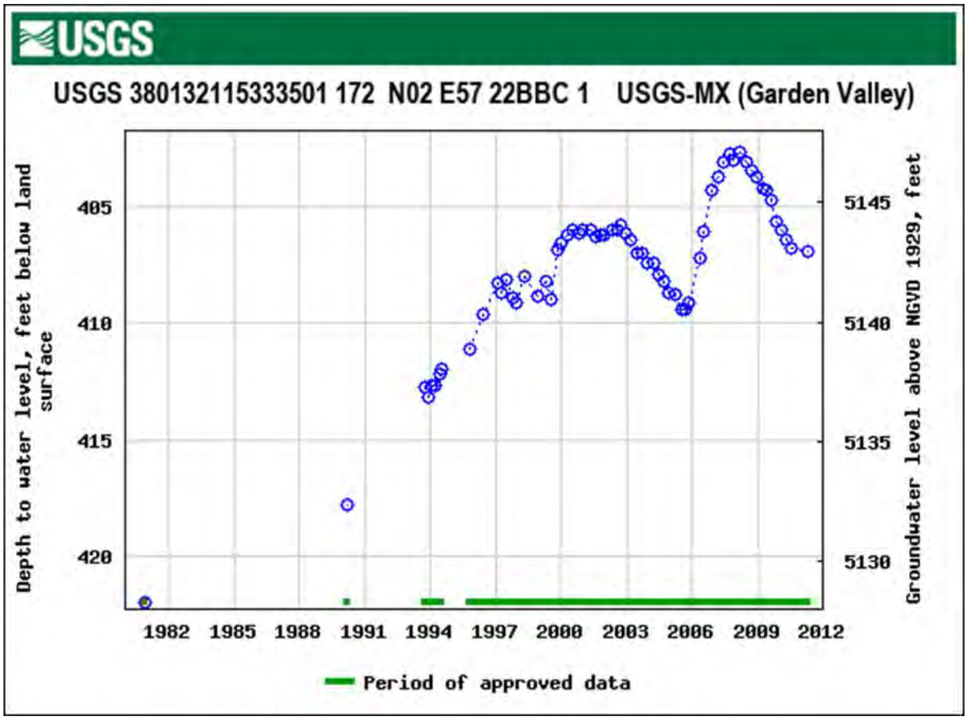
Coal Valley (Basin 171)

Well USGS-MX Coal Valley Well (Carbonate Aquifer)



Garden Valley (Basin 172)

Well USGS-MX Garden Valley Well (Basin Fill Aquifer)

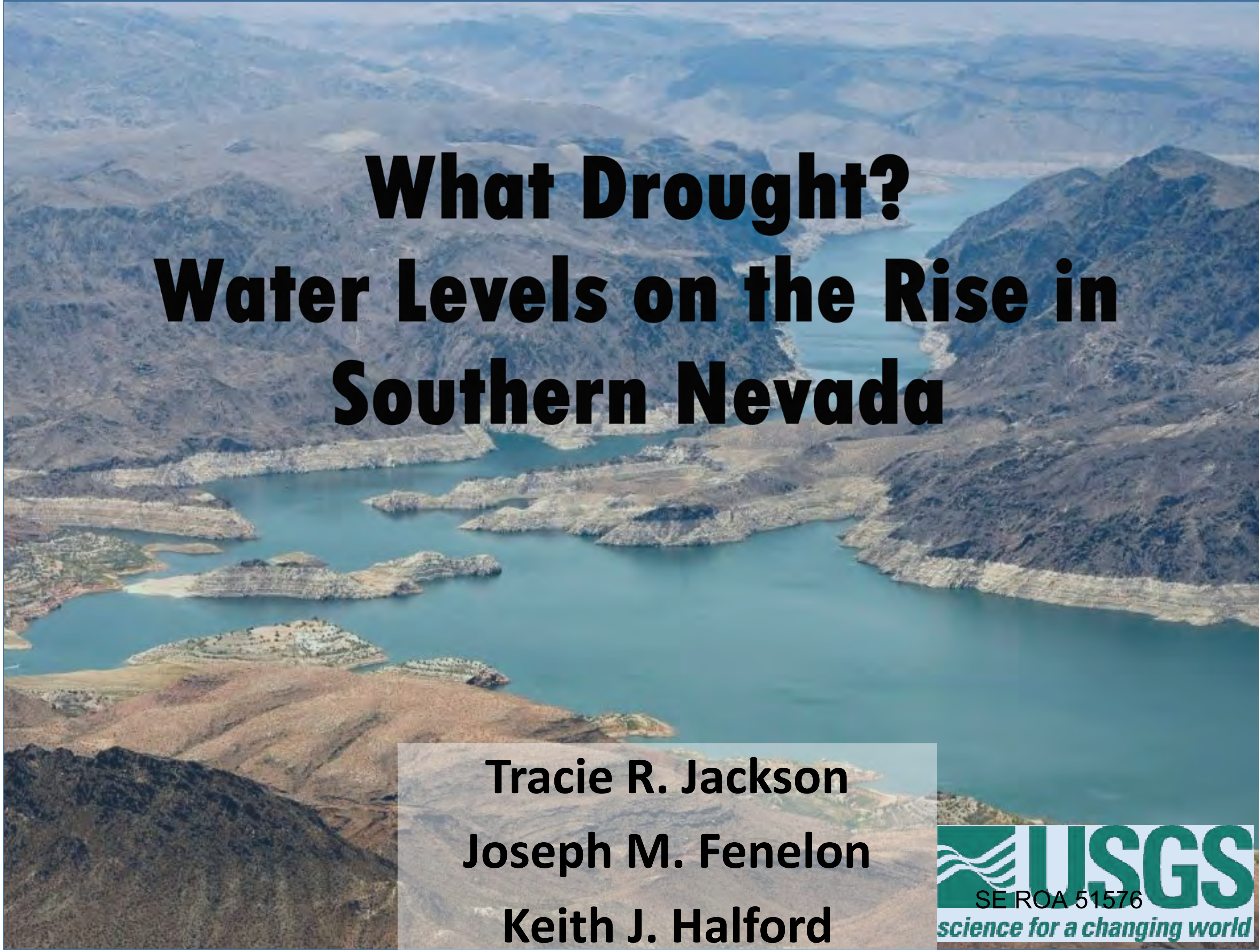


Appendix B

TETRA TECH

SE ROA 51575

What Drought? Water Levels on the Rise in Southern Nevada



Tracie R. Jackson
Joseph M. Fenelon
Keith J. Halford

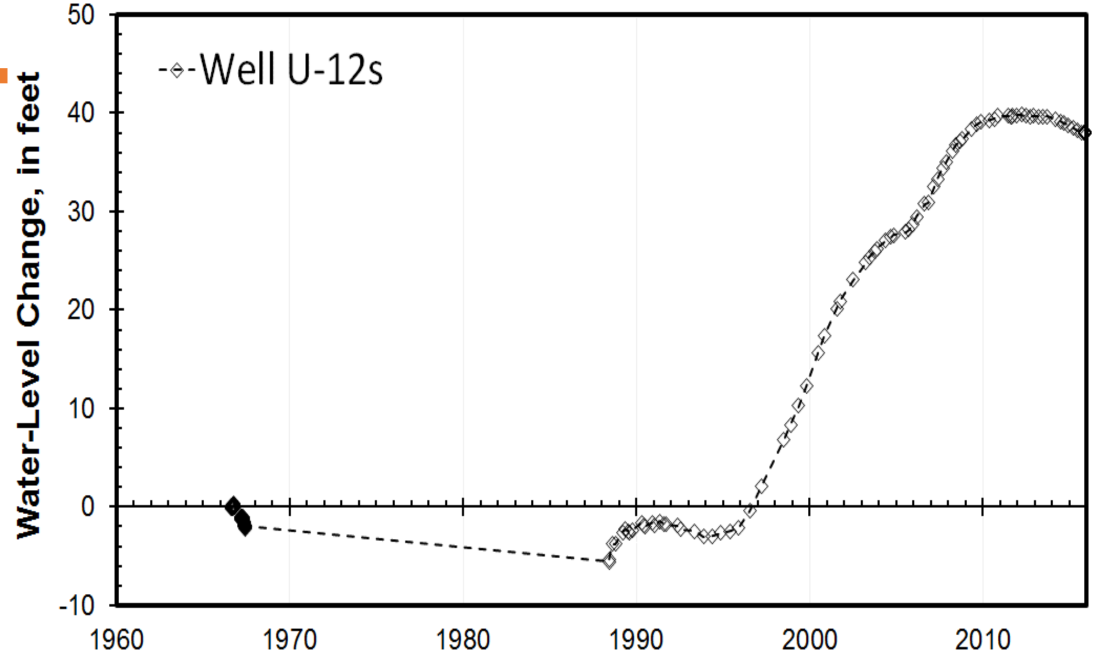
science for a changing world

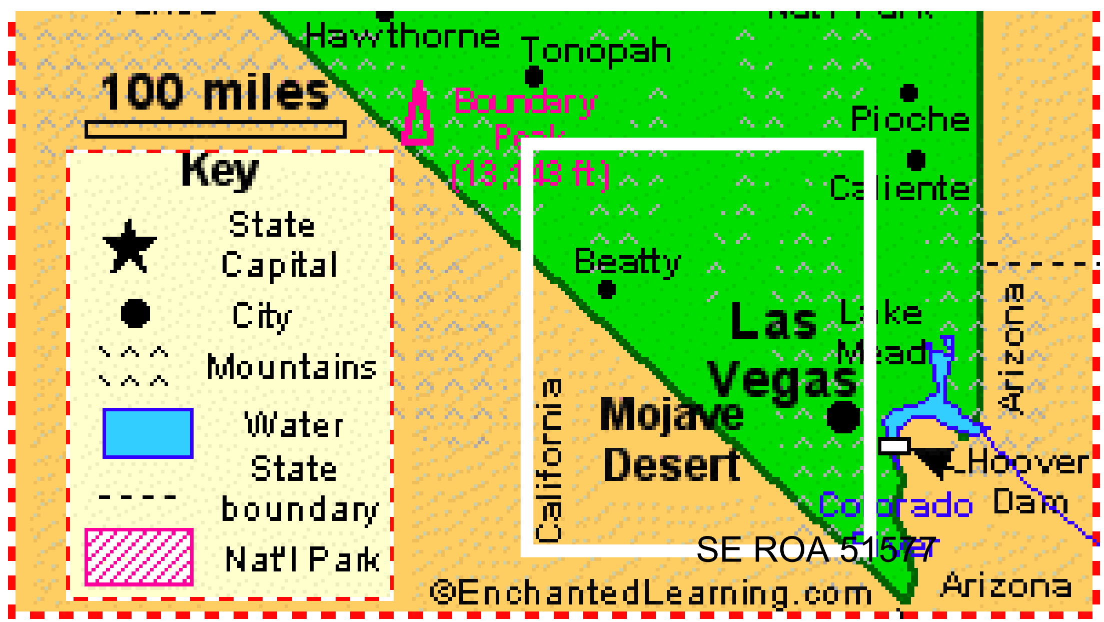
SE ROA 51576

JA_16707

DROUGHT?

- Water levels distant from pumping are rising in southern Nevada

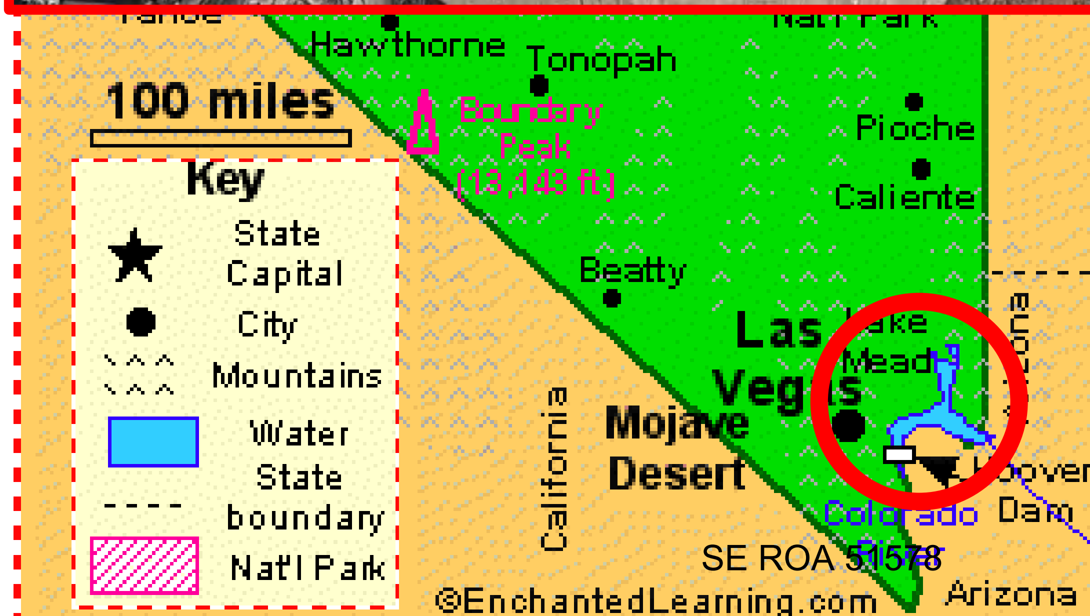




SE ROA 51577

Drought?

- Water levels distant from pumping are rising in southern Nevada
- Las Vegas – Drought



JA_16709

LAKE MEAD

Lake Mead declines to lowest level in history

-The Desert Sun

When will Las Vegas run out of water?

-Las Vegas Sun

LAKE MEAD WATER LEVEL AT ALL-TIME LOW

THE RACE TO STOP LAS VEGAS FROM RUNNING DRY

-THE TELEGRAPH

What Lake Mead's Record Low Means for California

-News Deeply

As Lake Mead levels drop, the West braces for bigger drought impact

-Nevada Public Radio

SE ROA 51579

JA_16710

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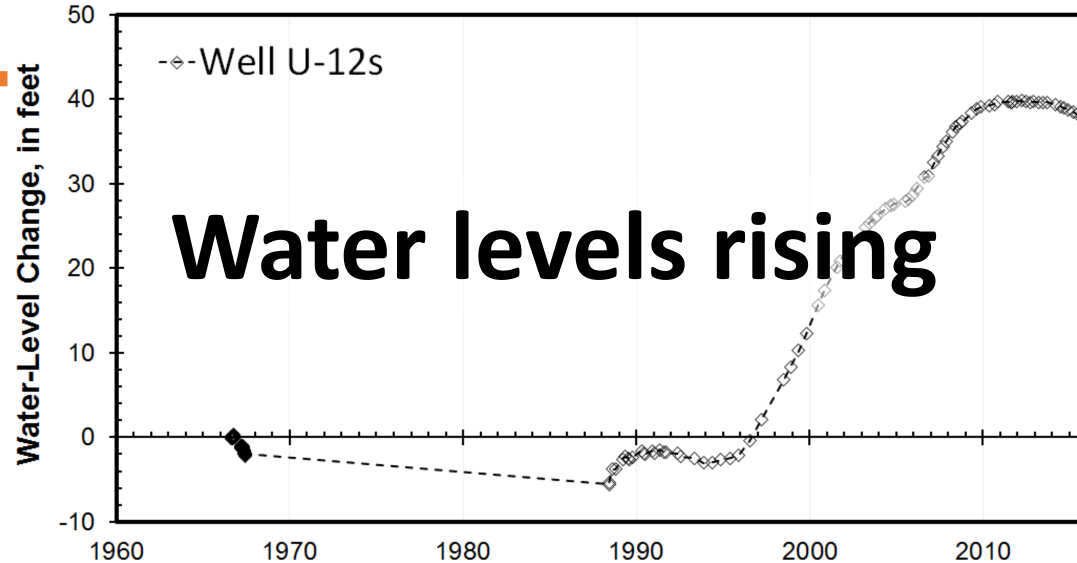
PROBLEM

How do we reconcile drought with rising water levels?

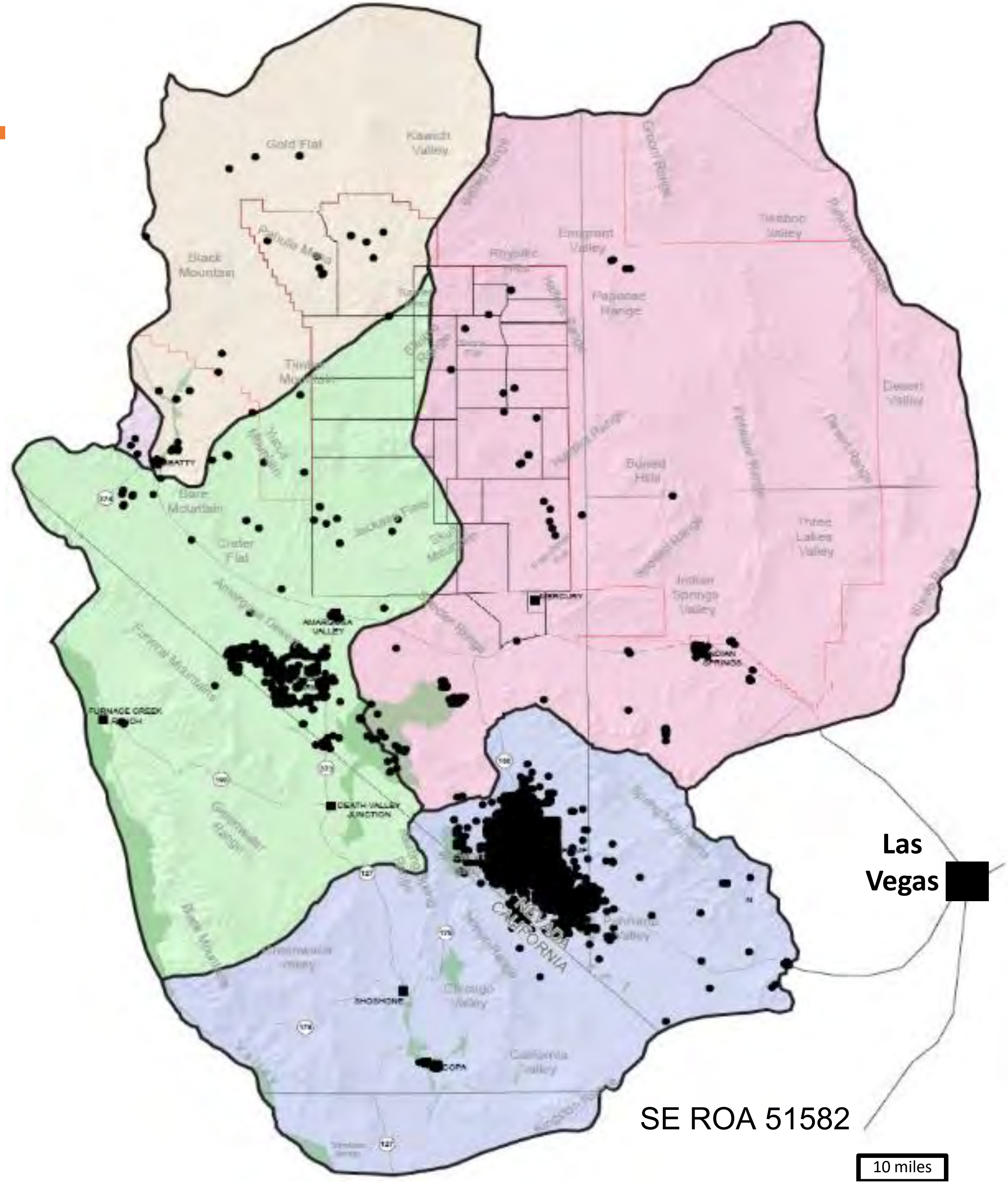


Hydrographs

SE ROA 51581

Death Valley Regional Flow System

Pumping wells



SE ROA 51582

JA_16713

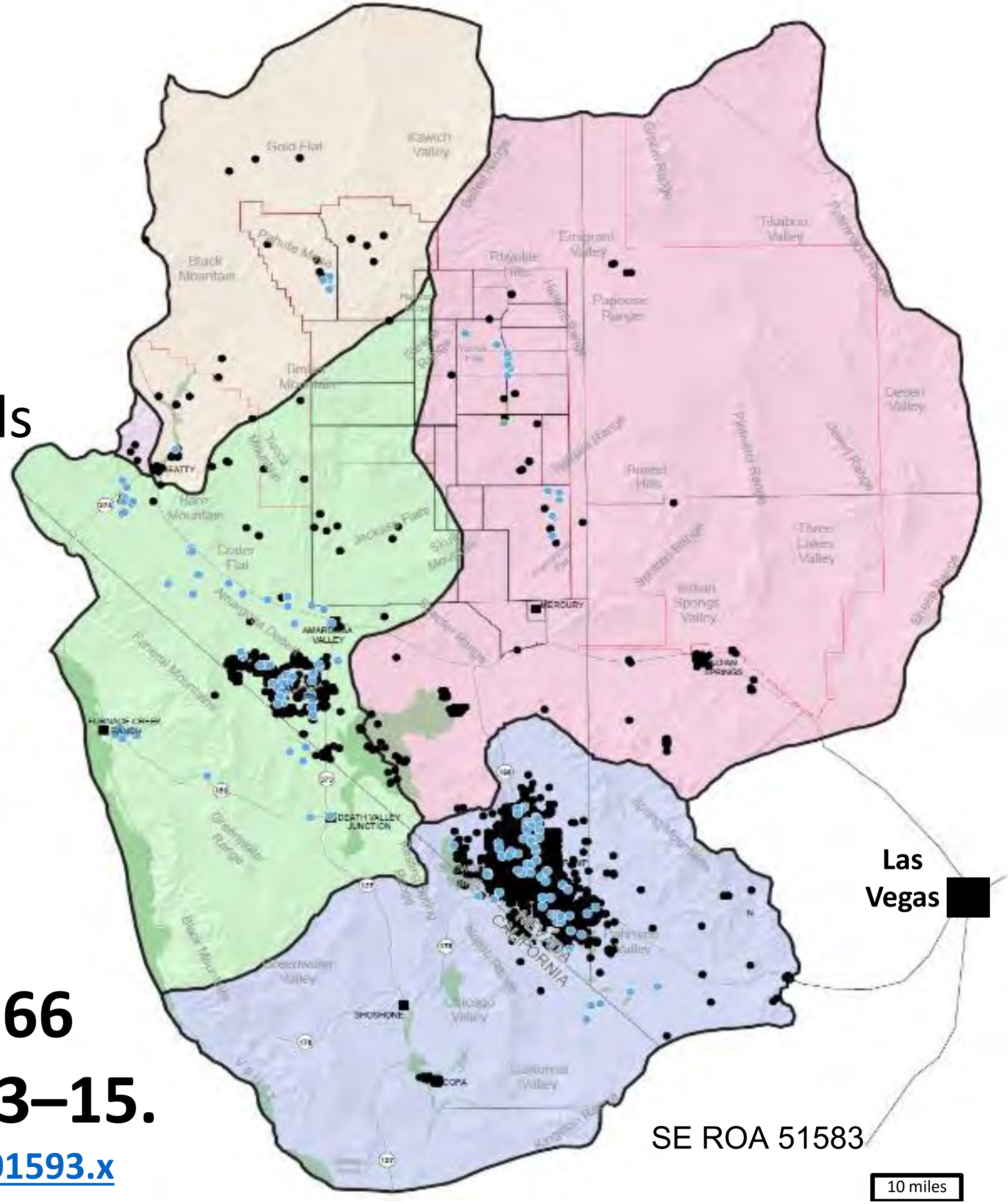
Death Valley Regional Flow System

Wells with water levels affected by:

- Pumping
- Nuclear Testing

Atomic Tools in Developing Water
Arthur M. Piper, 1966
Ground Water, 4: 13–15.

doi:10.1111/j.1745-6584.1966.tb01593.x

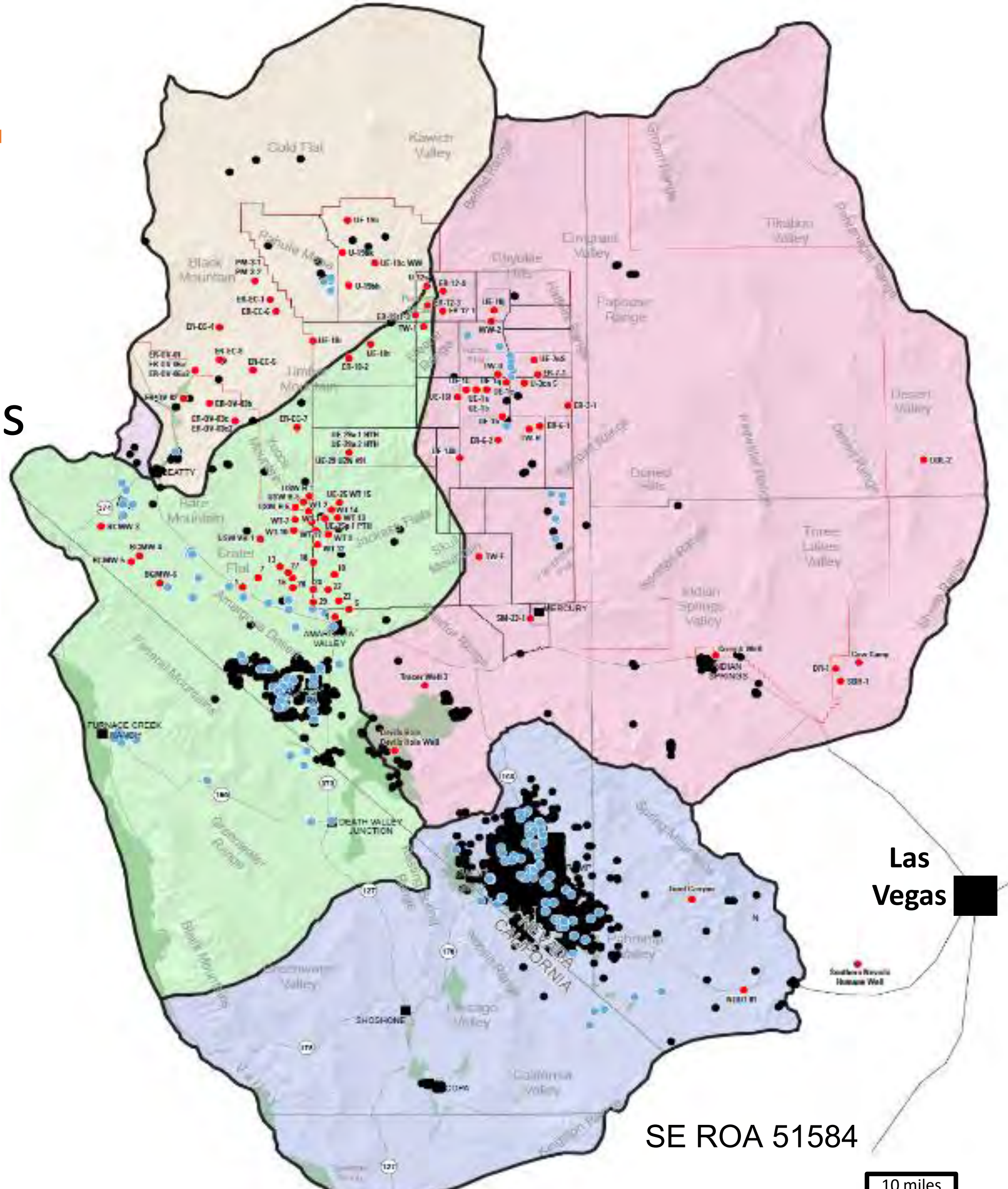


SE ROA 51583

JA_16714

Death Valley Regional Flow System

Wells with water levels distant from pumping

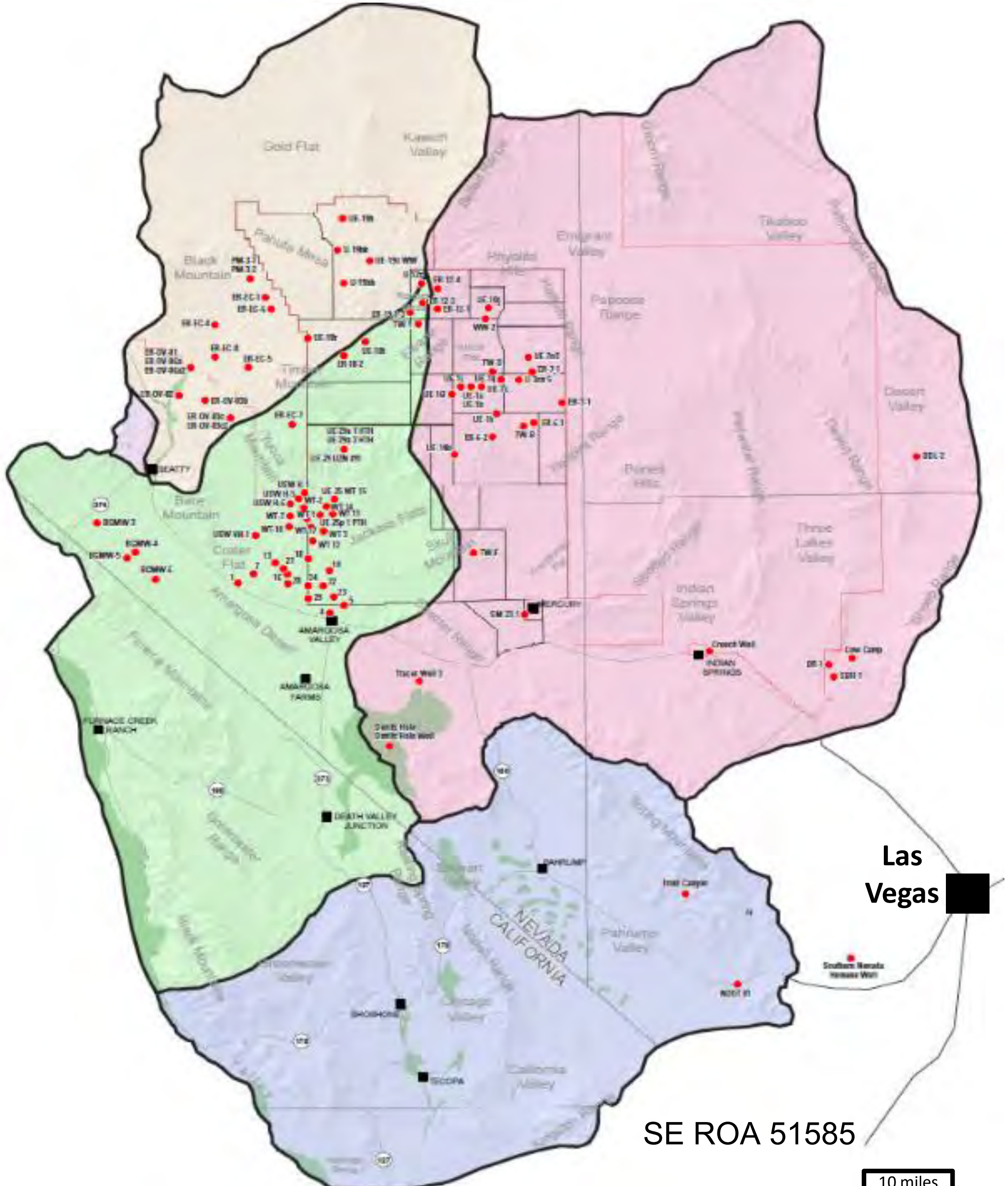


SE ROA 51584

10 miles

Death Valley Regional Flow System

Focus on wells distant from pumping

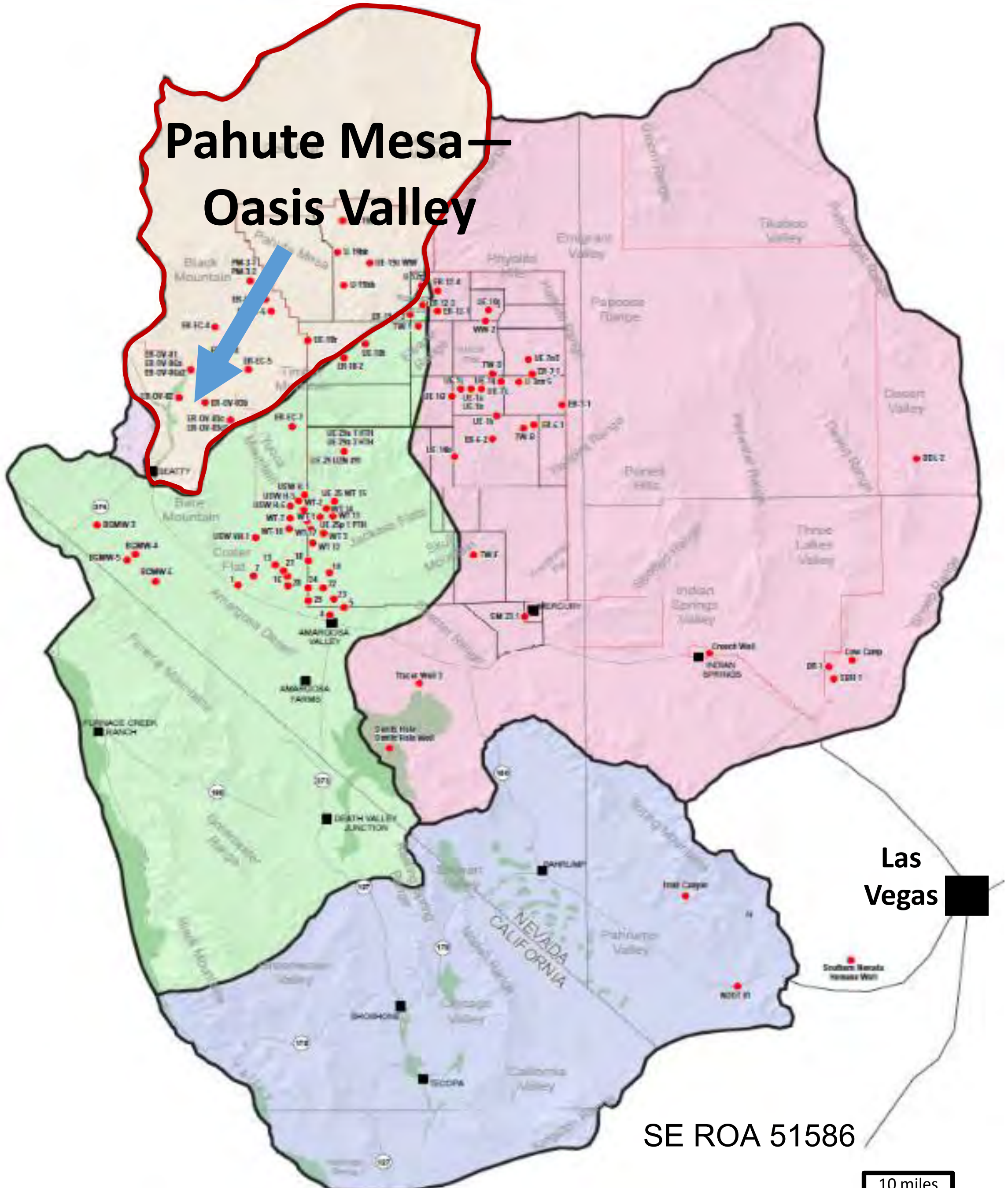


SE ROA 51585

JA_16716

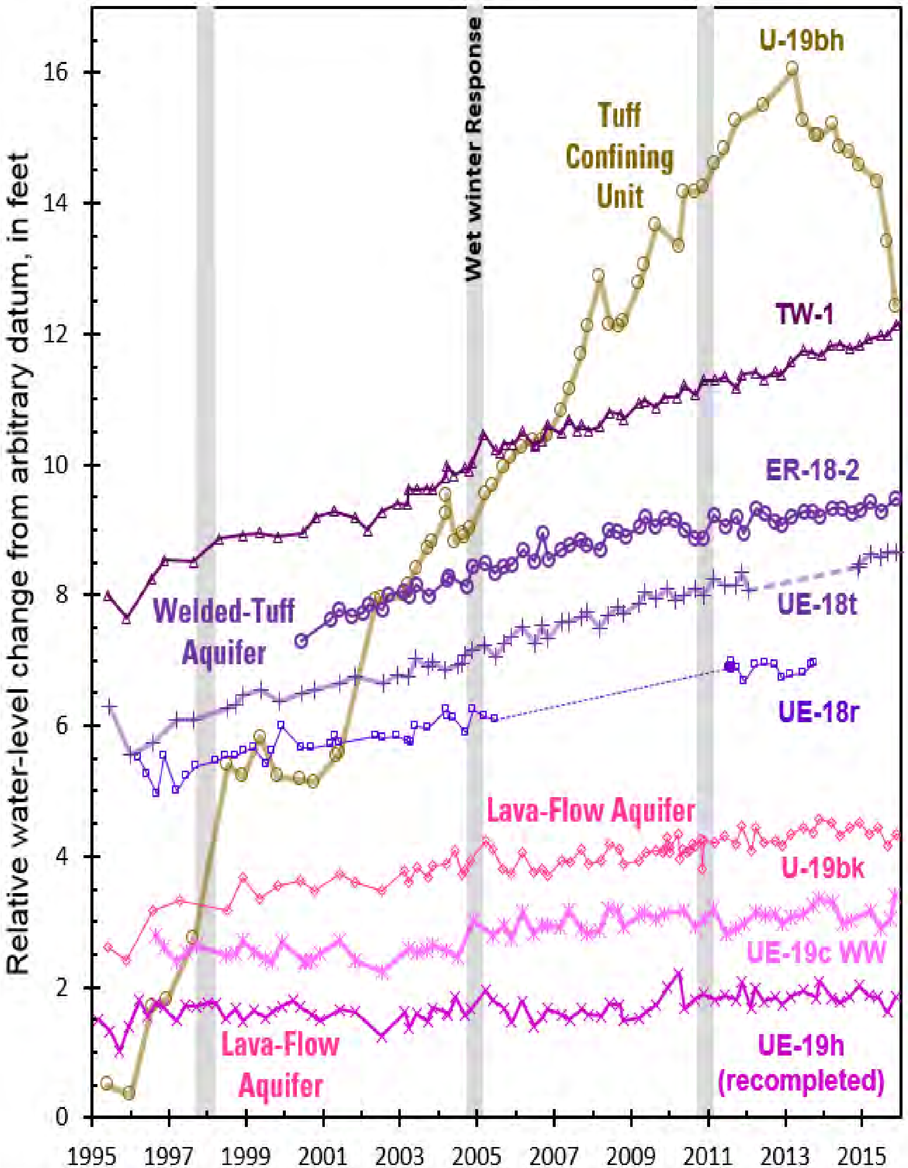
Death Valley Regional Flow System

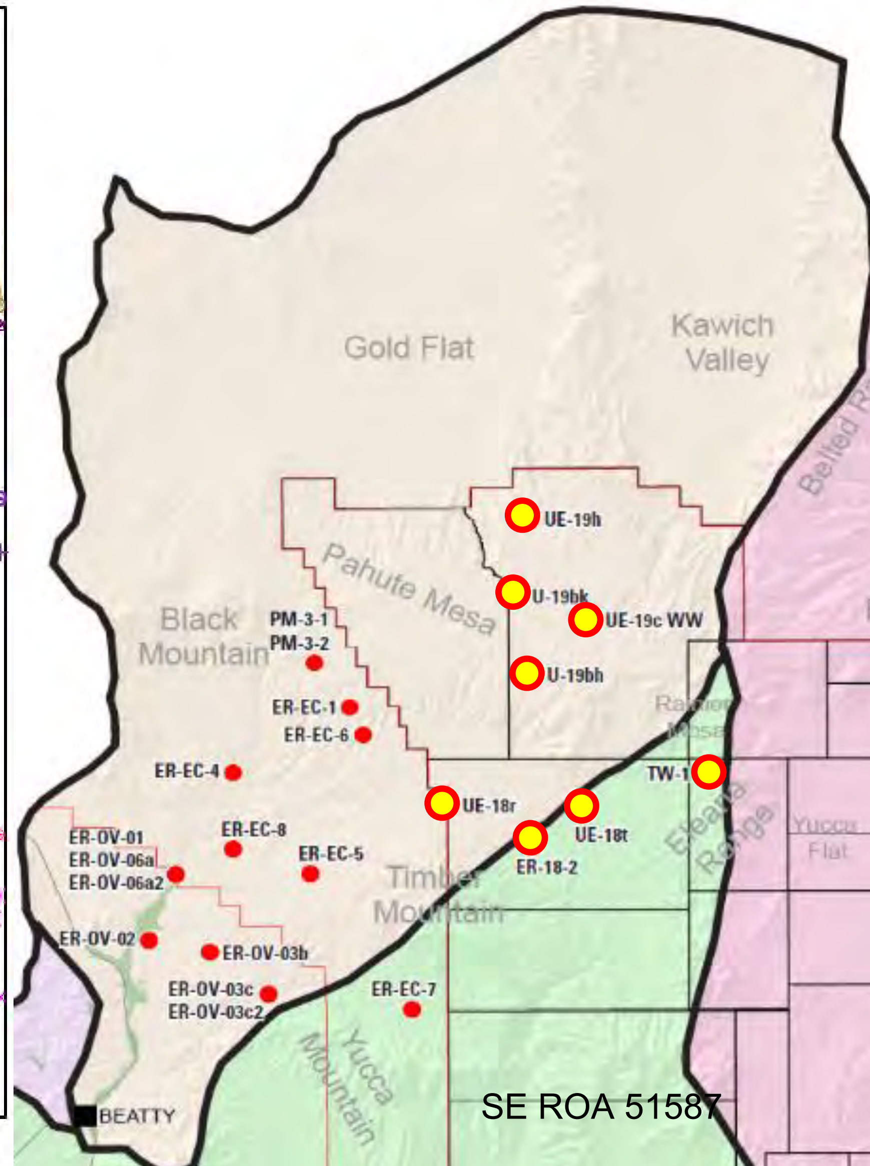
Focus on wells distant from pumping



SE ROA 51586

PAHUTE AND BUCKBOARD MESAS





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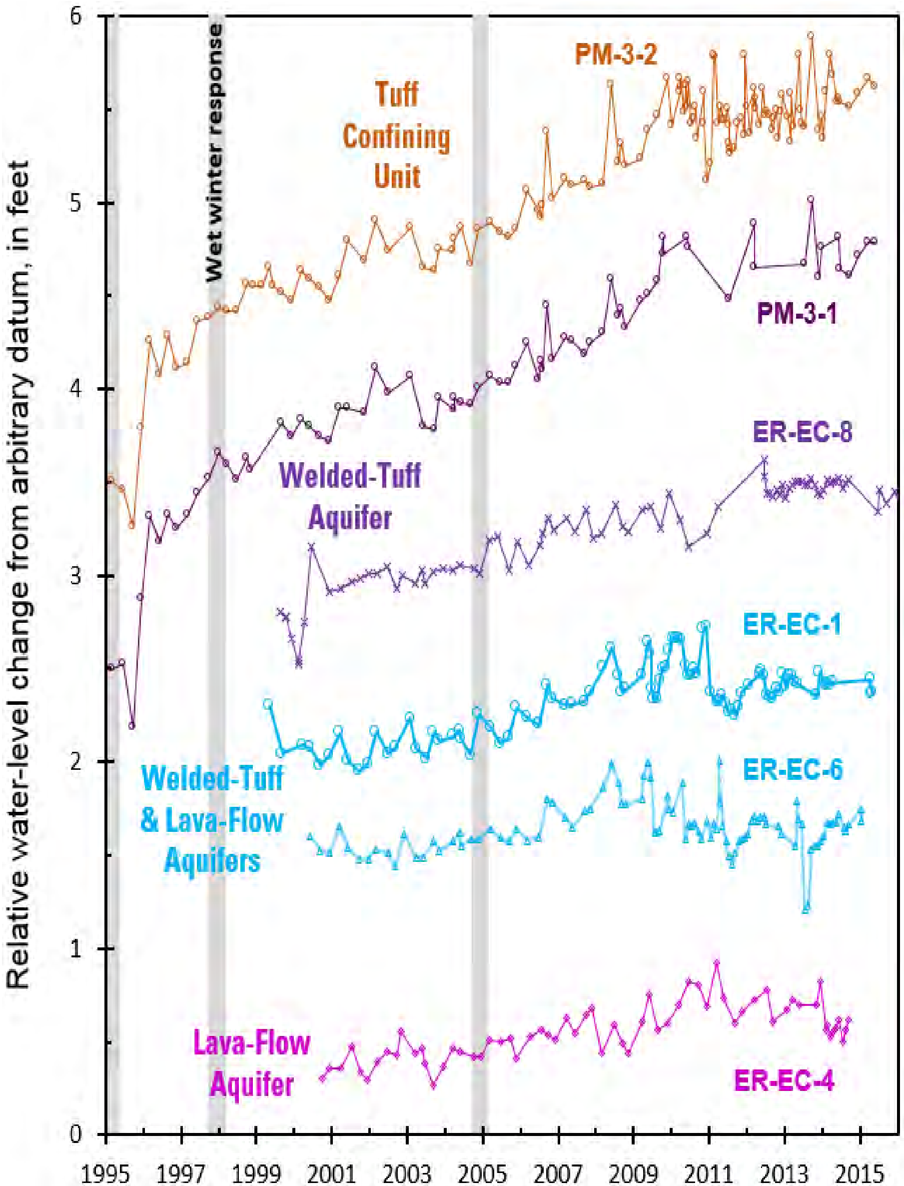
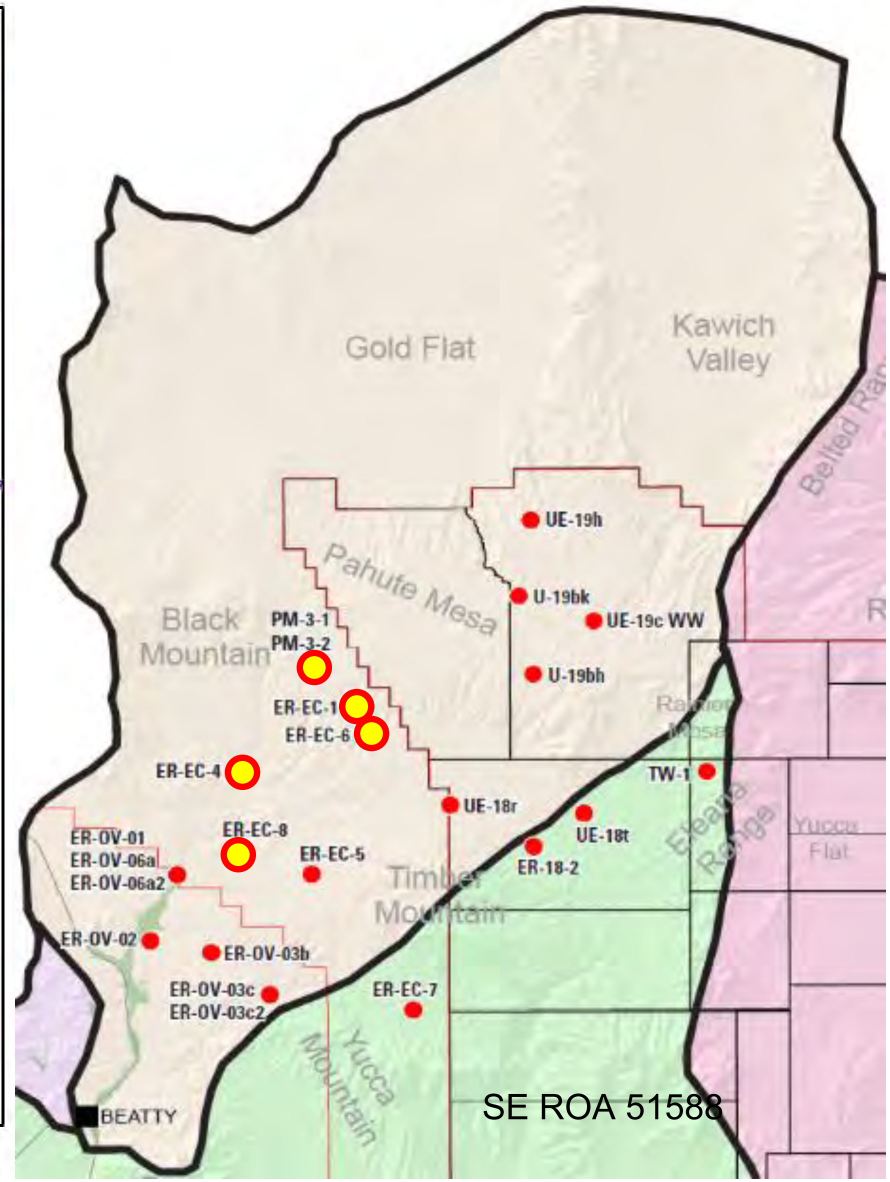
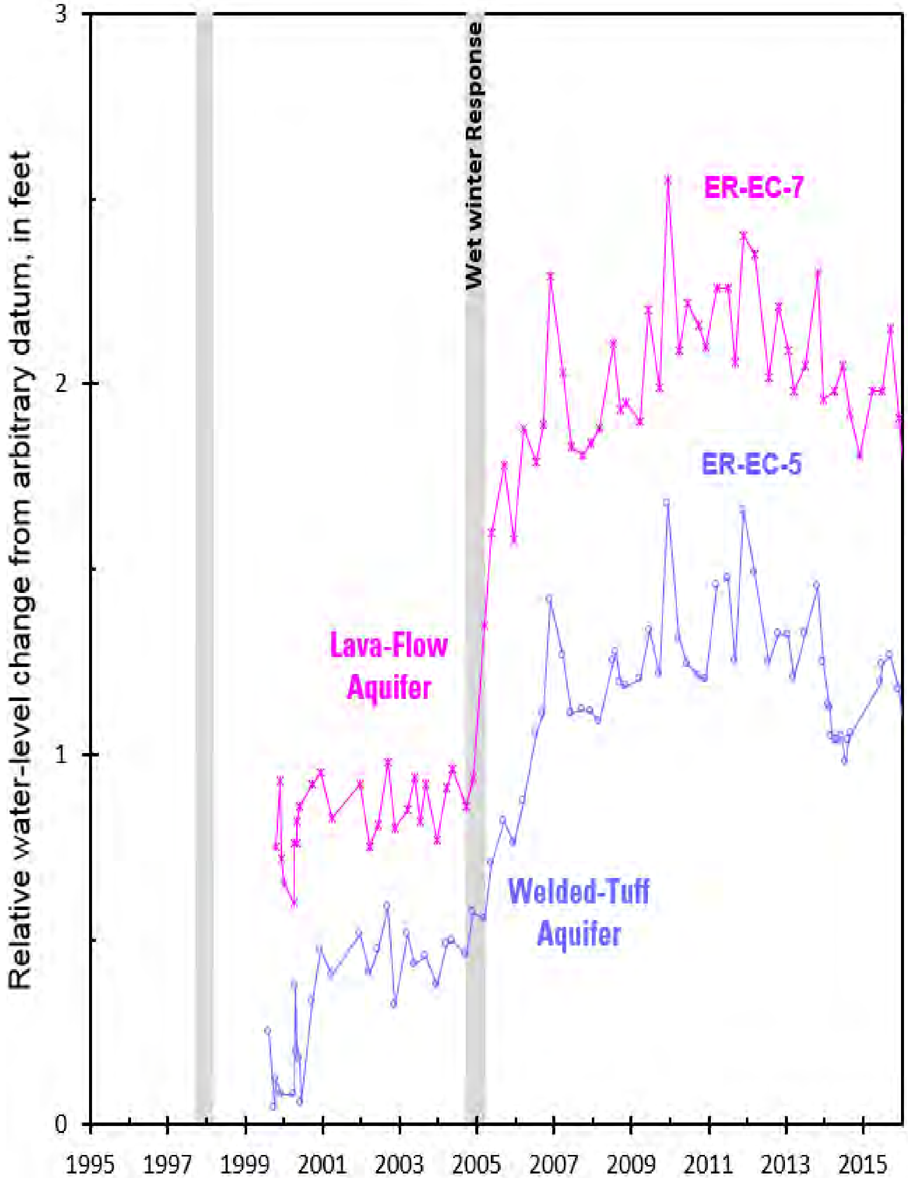
THIRSTY CANYON AND ROCKET WASH

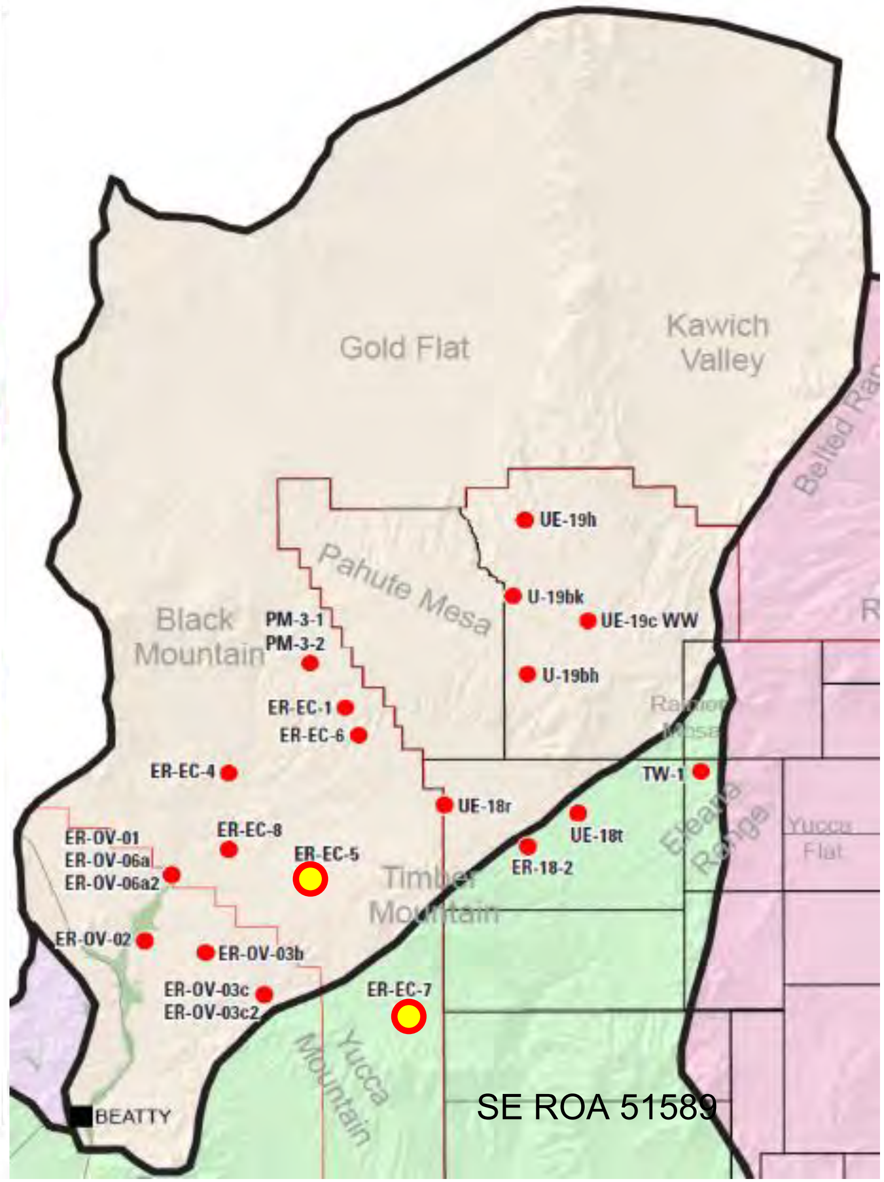




JA_16719

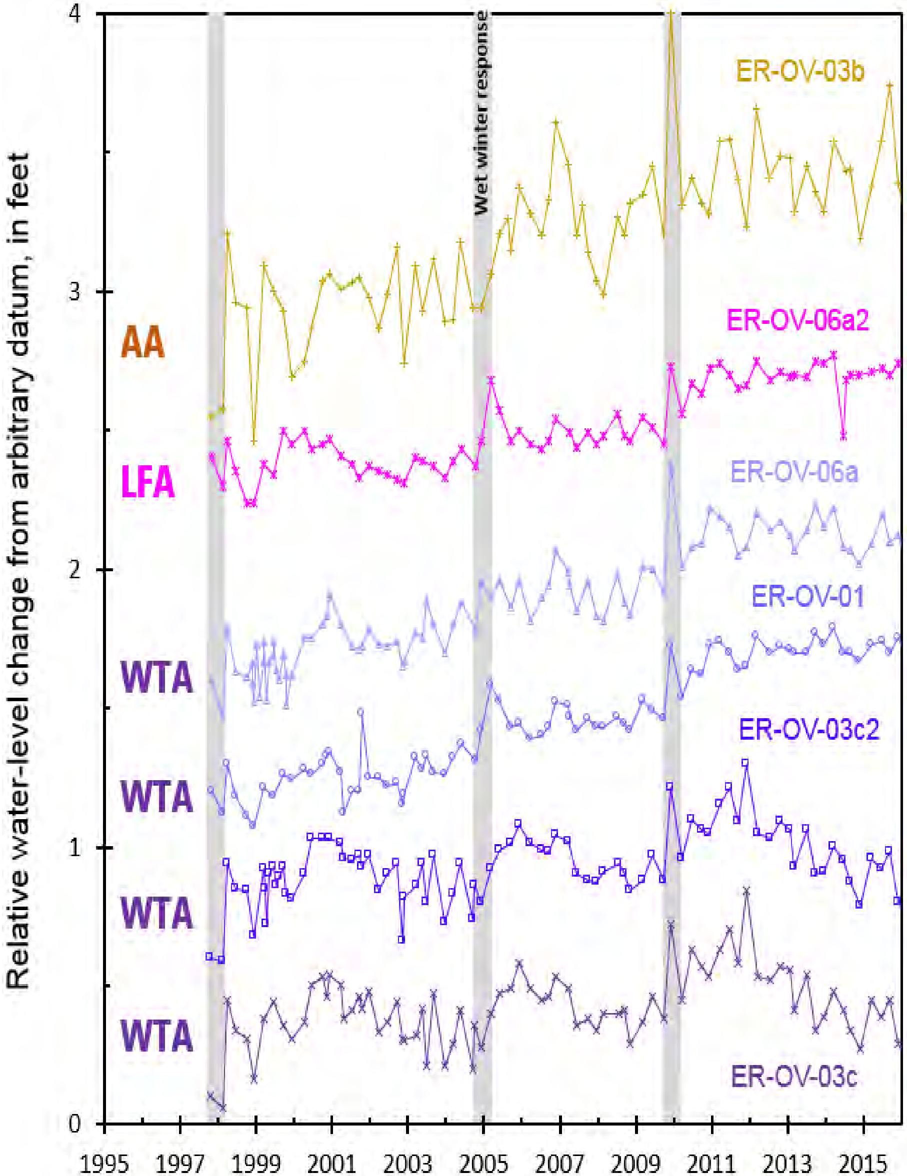
TIMBER MOUNTAIN

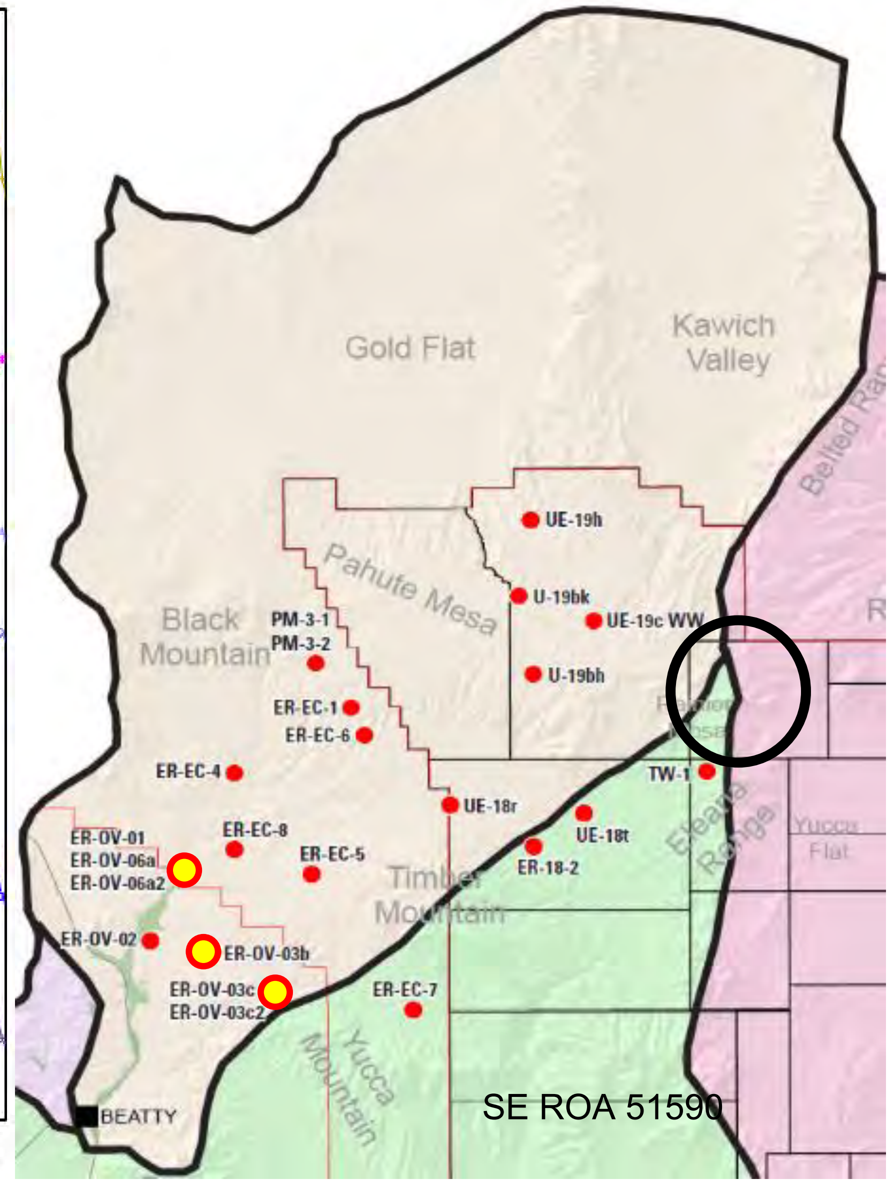




JA_16720

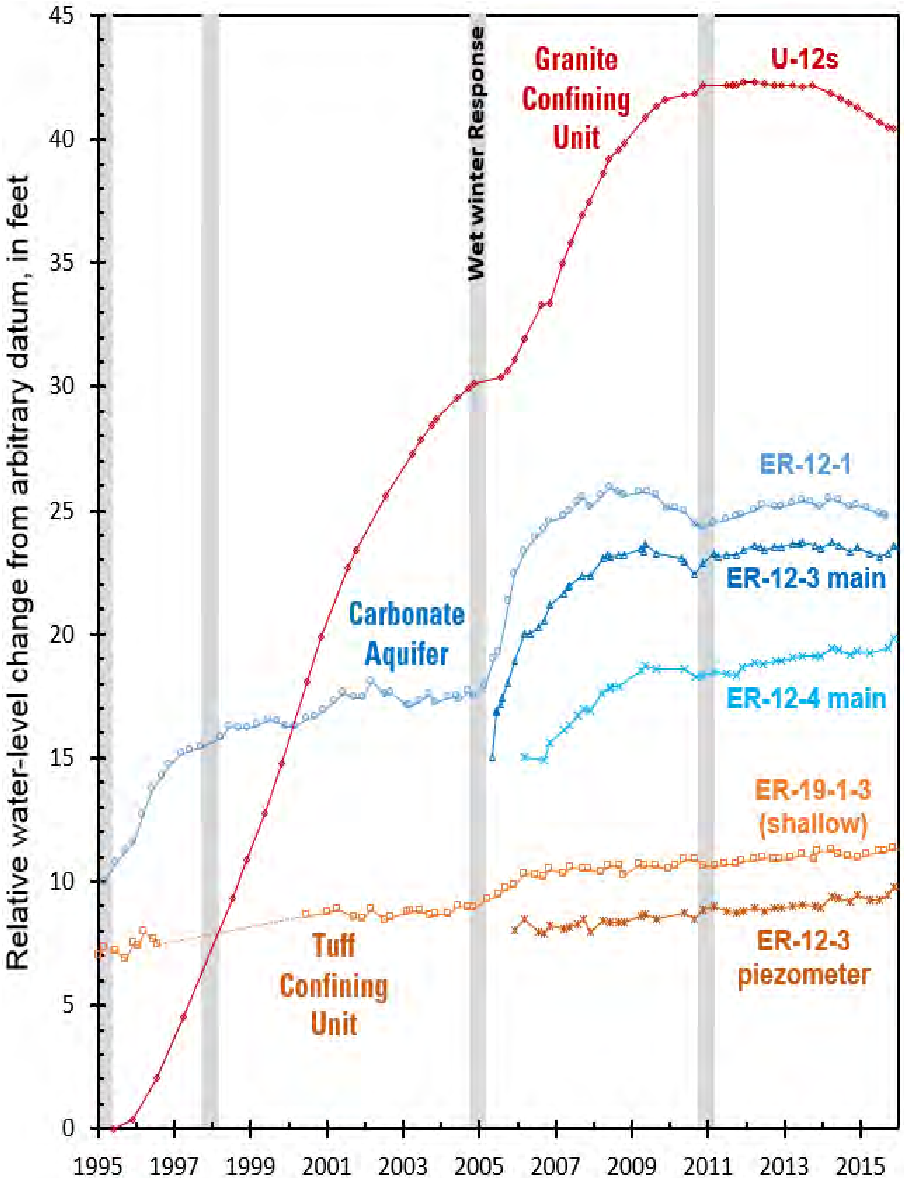
OASIS VALLEY

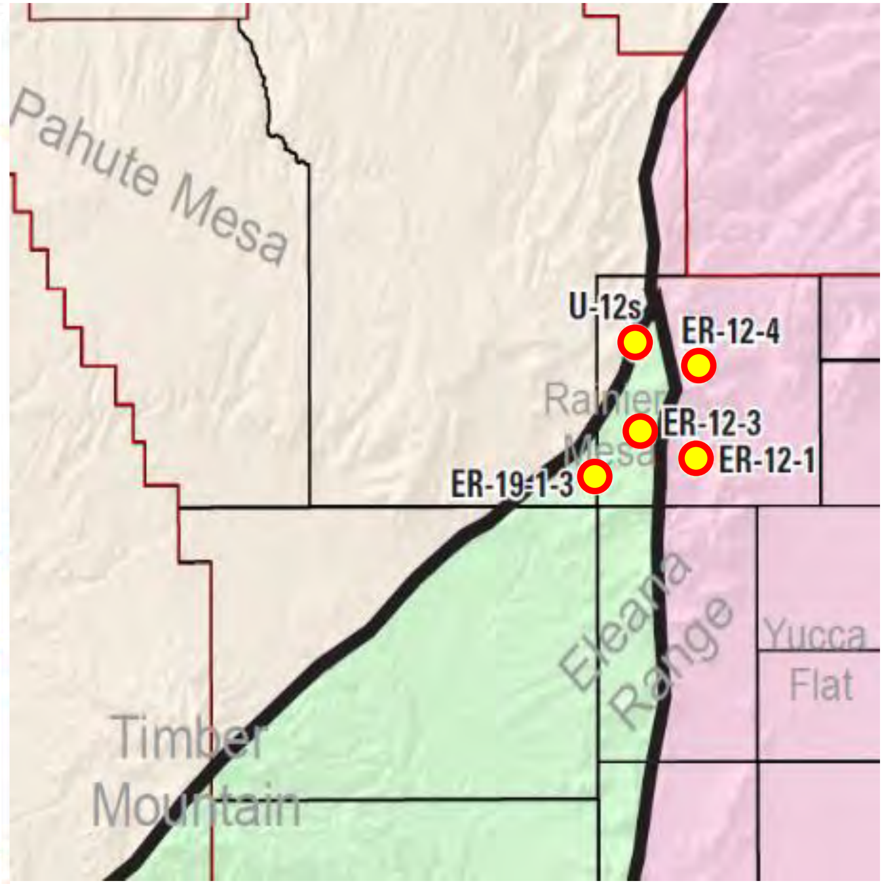




JA_16721

Rainier Mesa

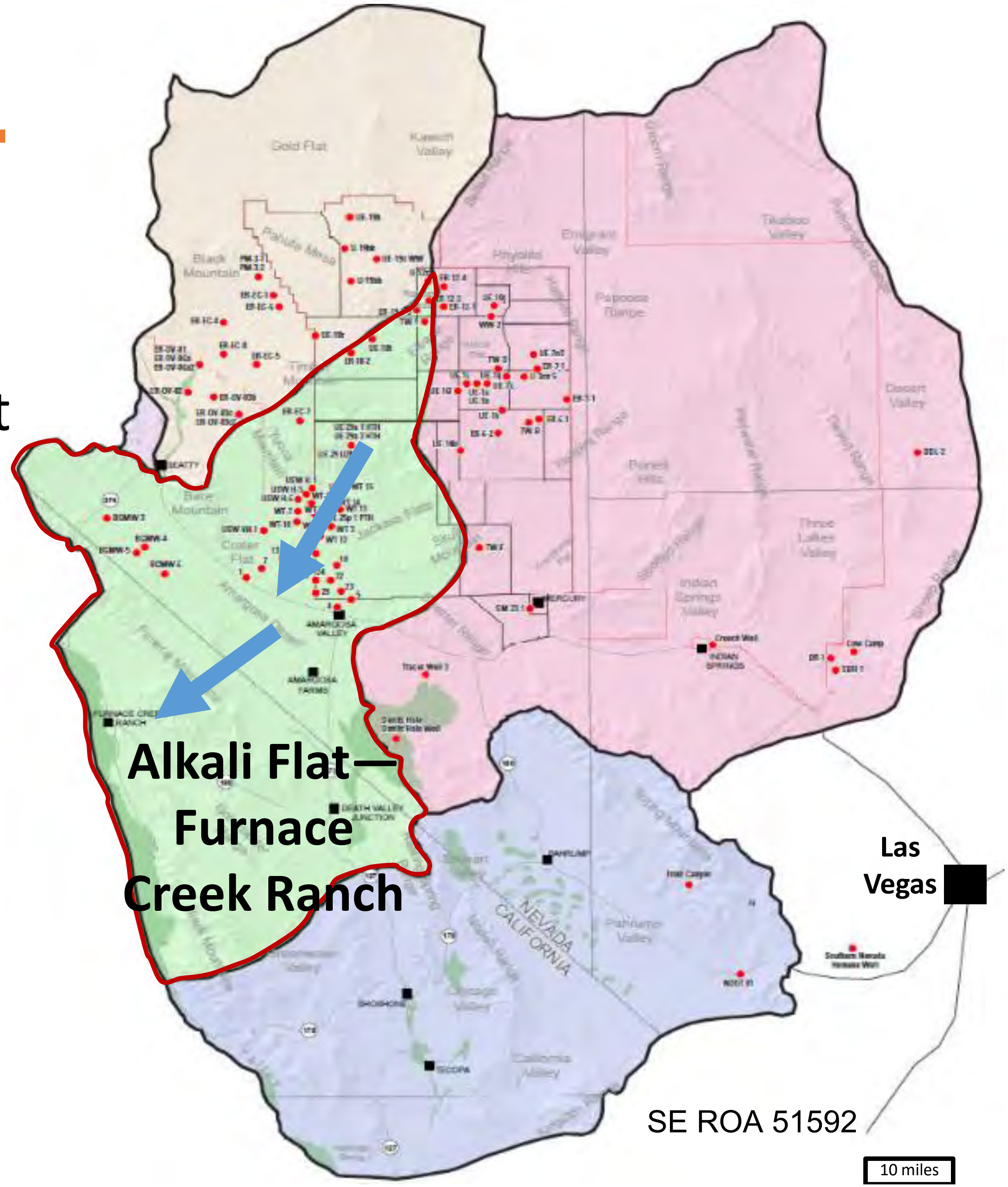




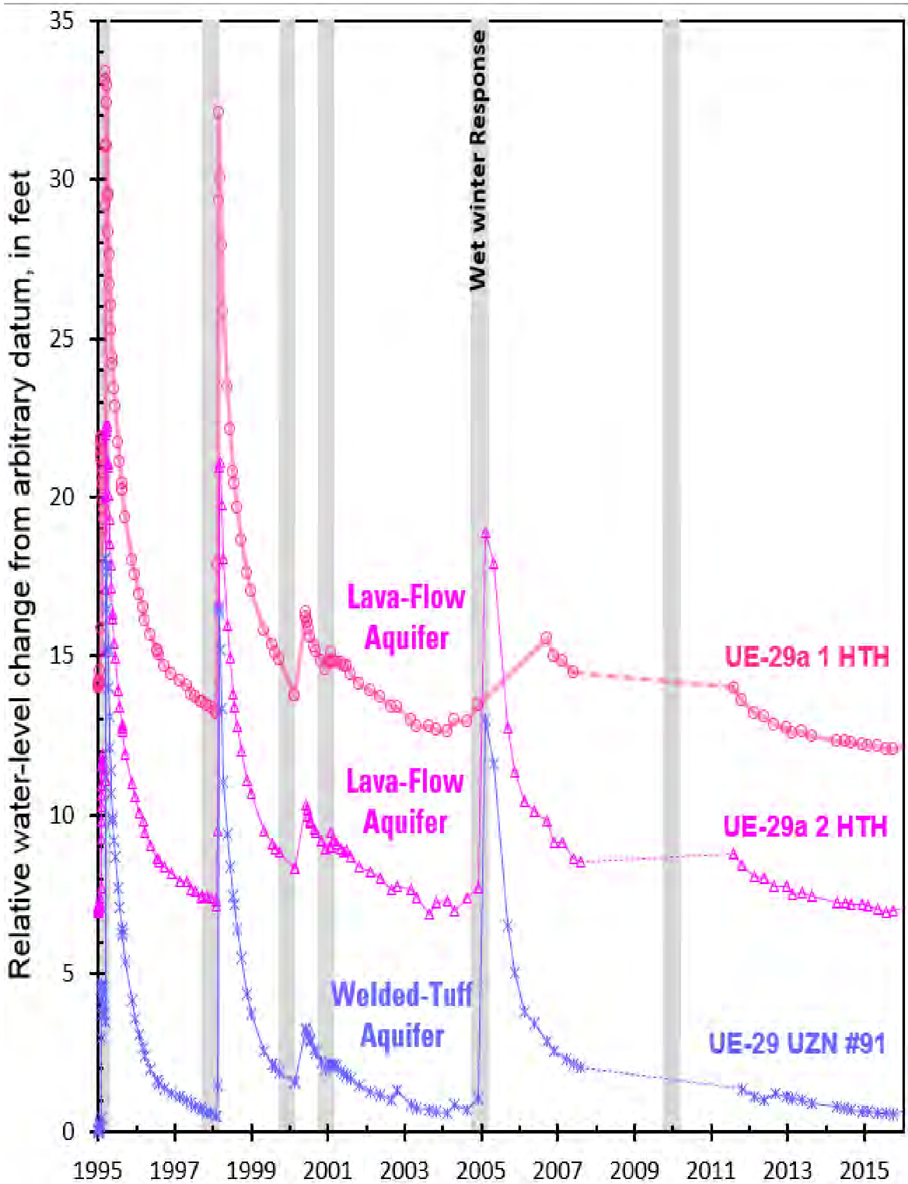
SE ROA 51591

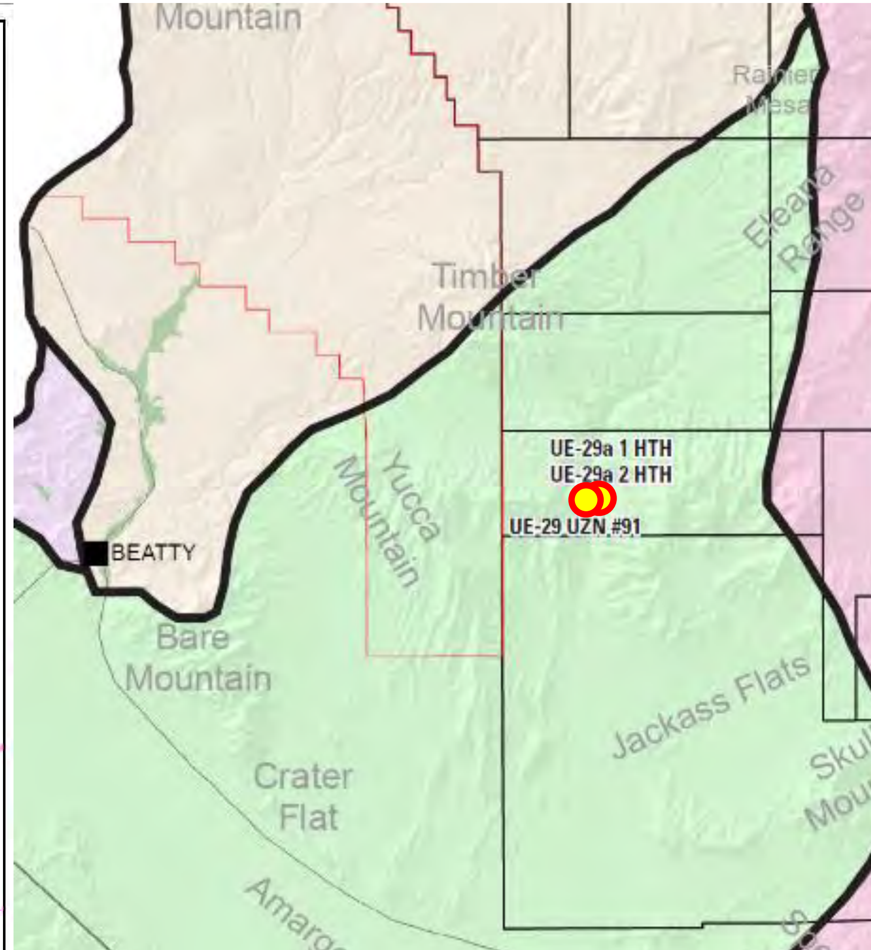
Death Valley Regional Flow System

Focus on wells distant from pumping



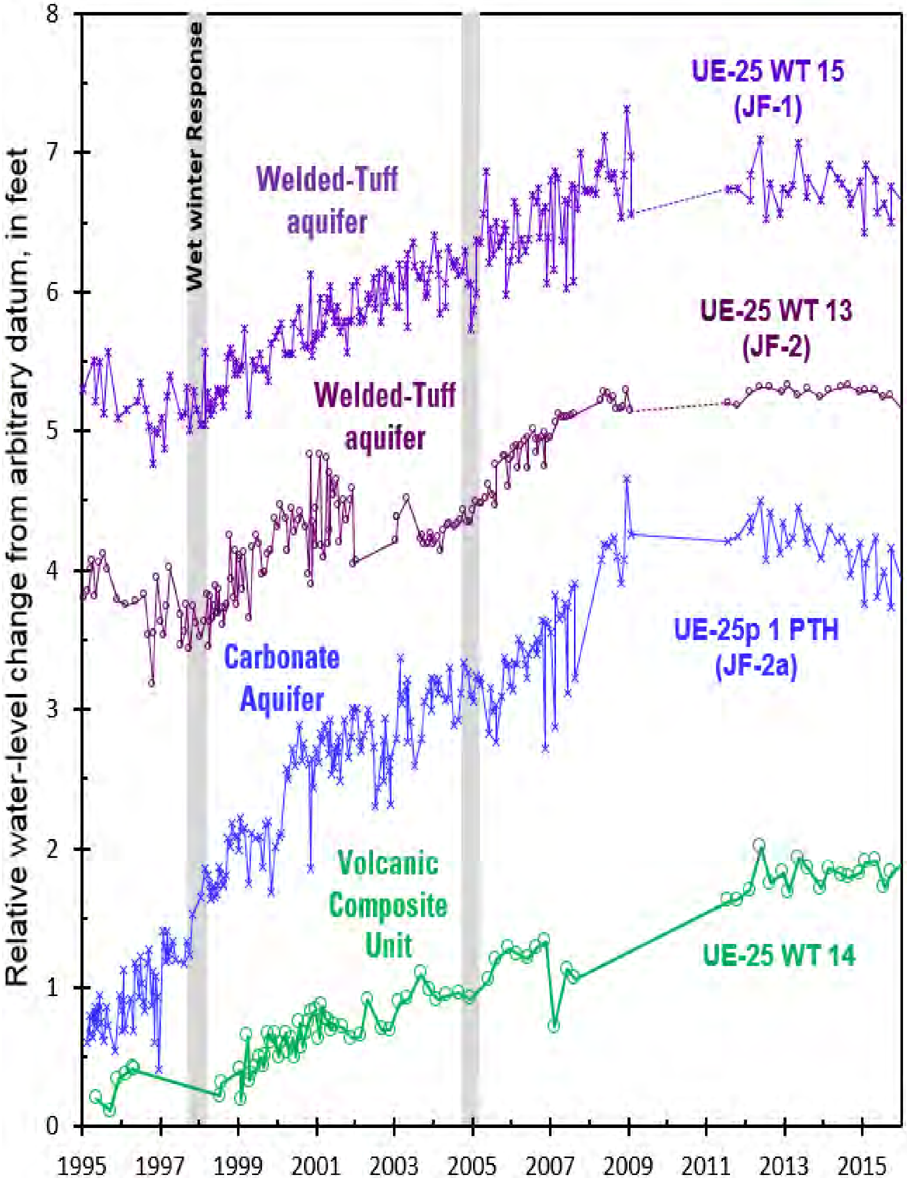
FORTYMILE WASH

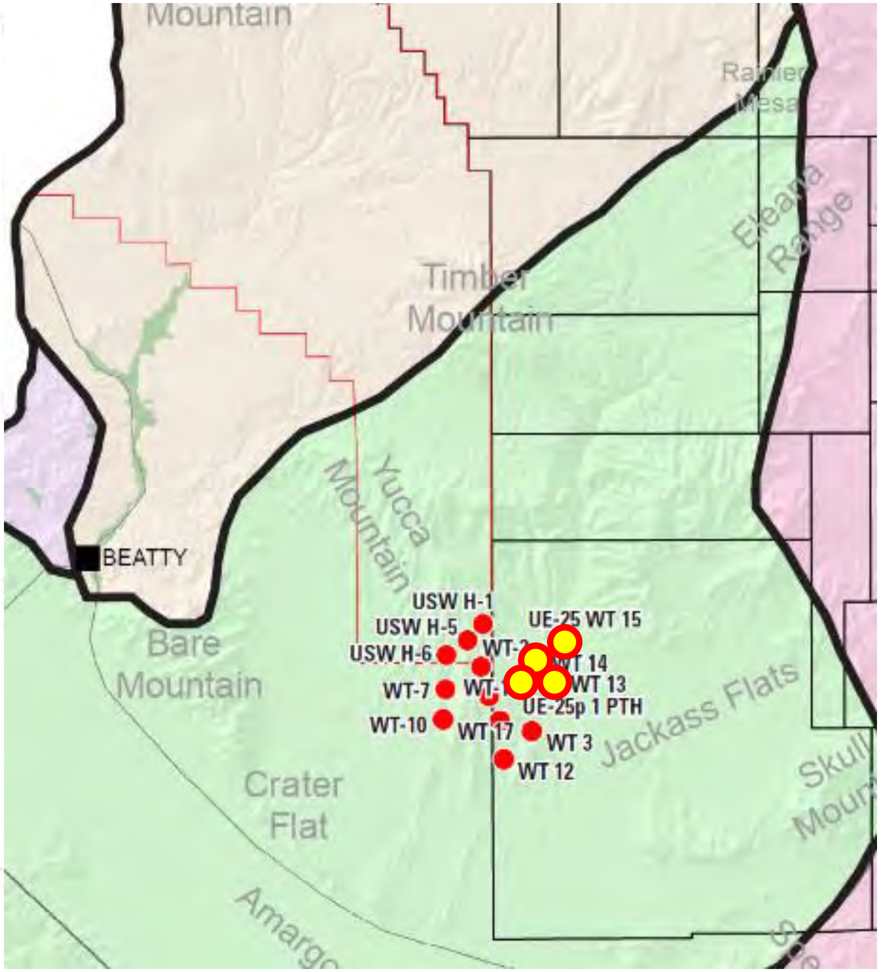




SE ROA 51593

JACKASS FLATS





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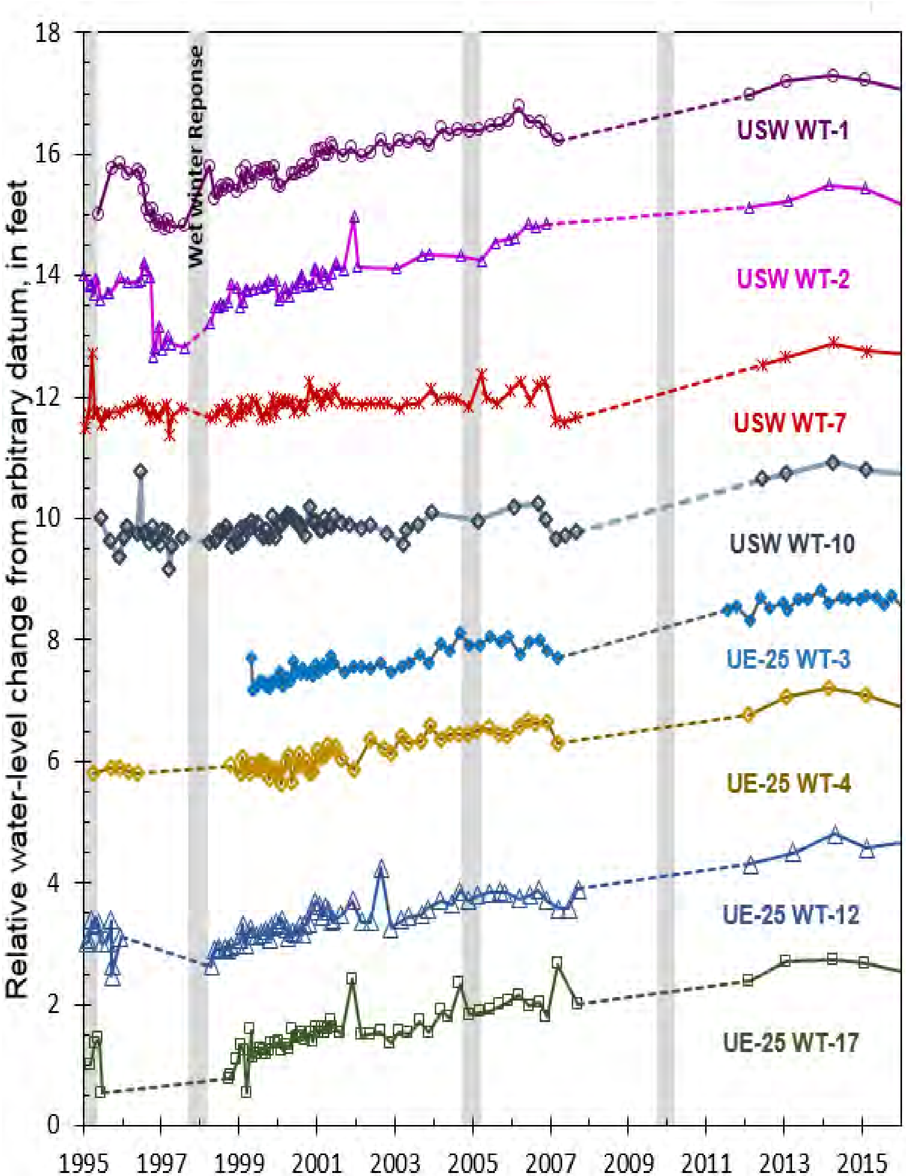
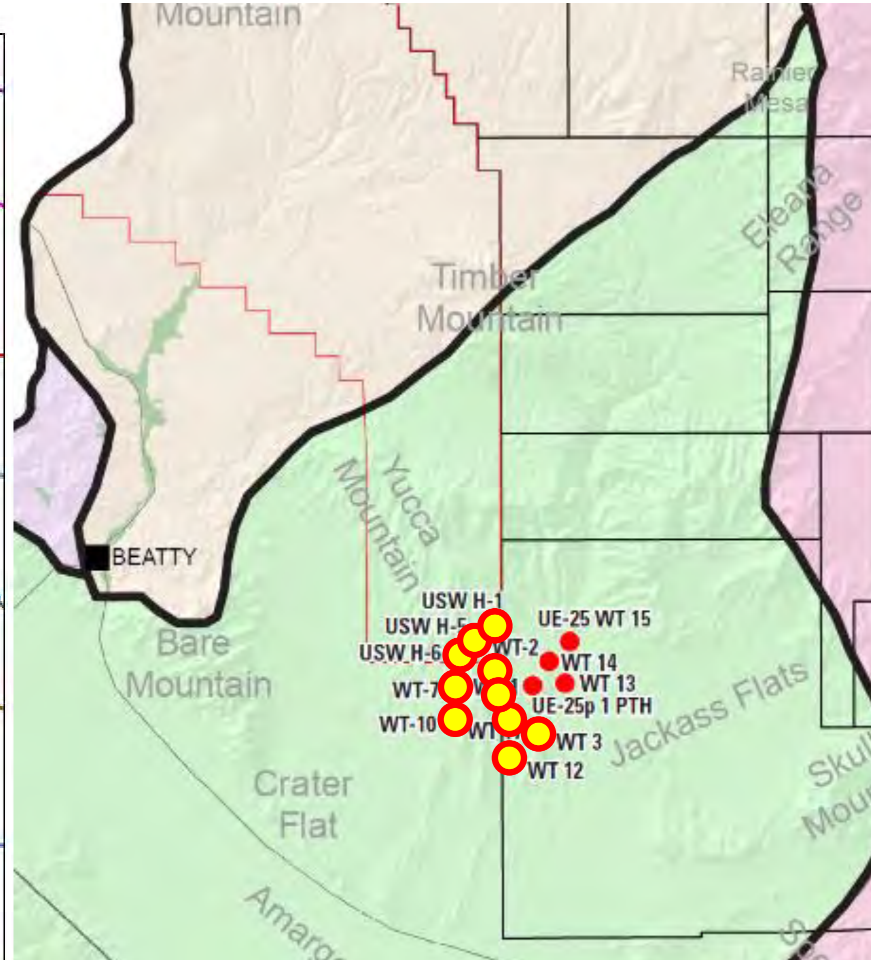
JA_16725

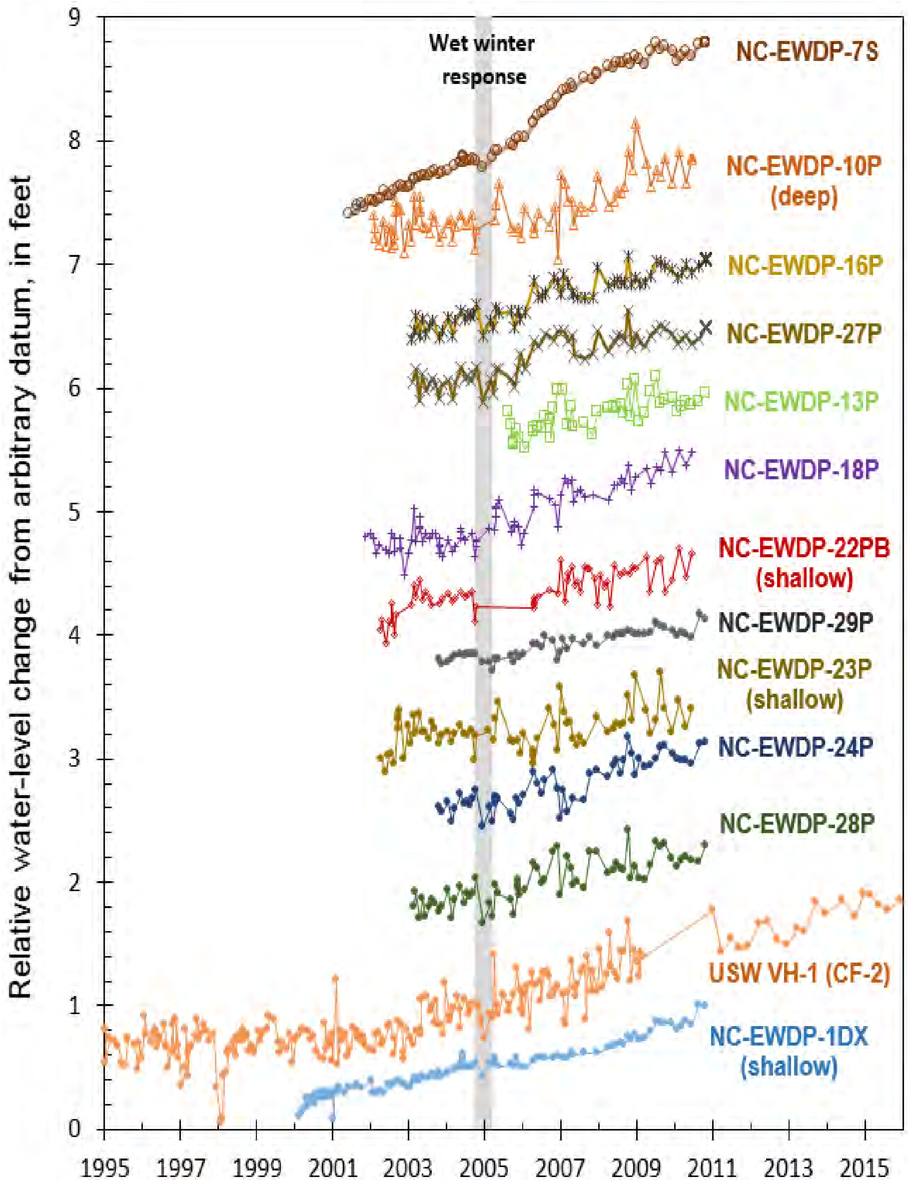
YUCCA MOUNTAIN

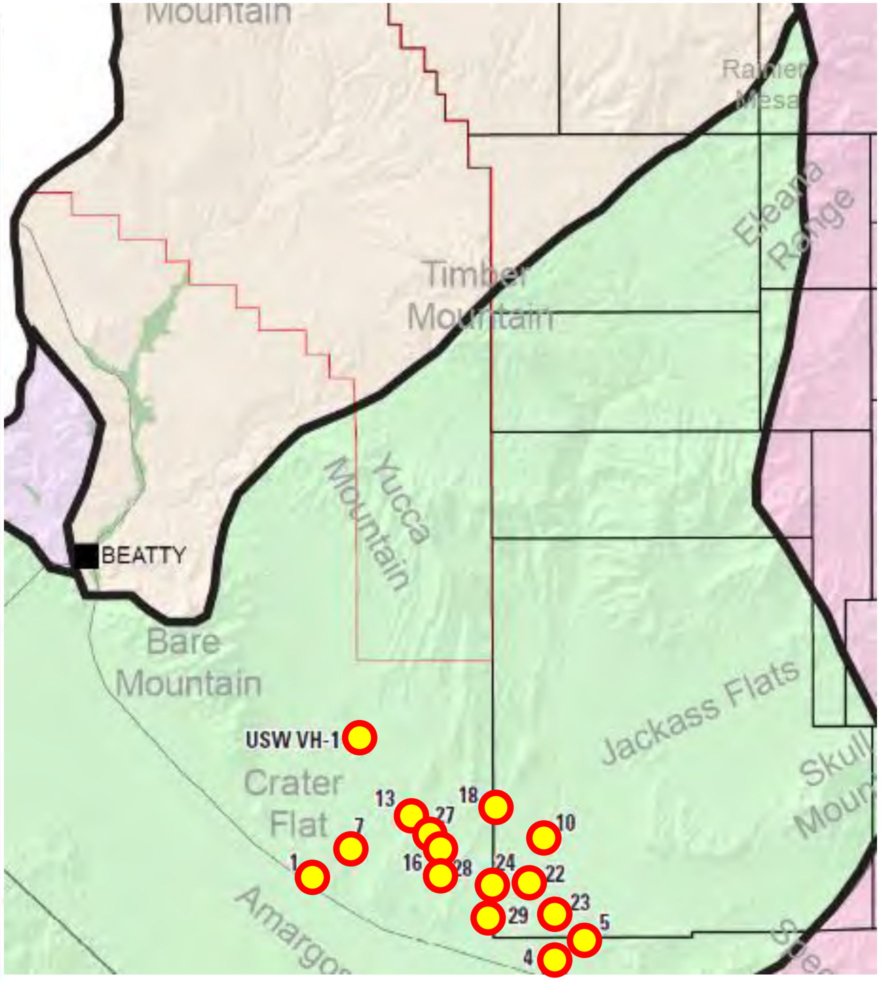




SE ROA 51595

NYE COUNTY EWDP WELLS





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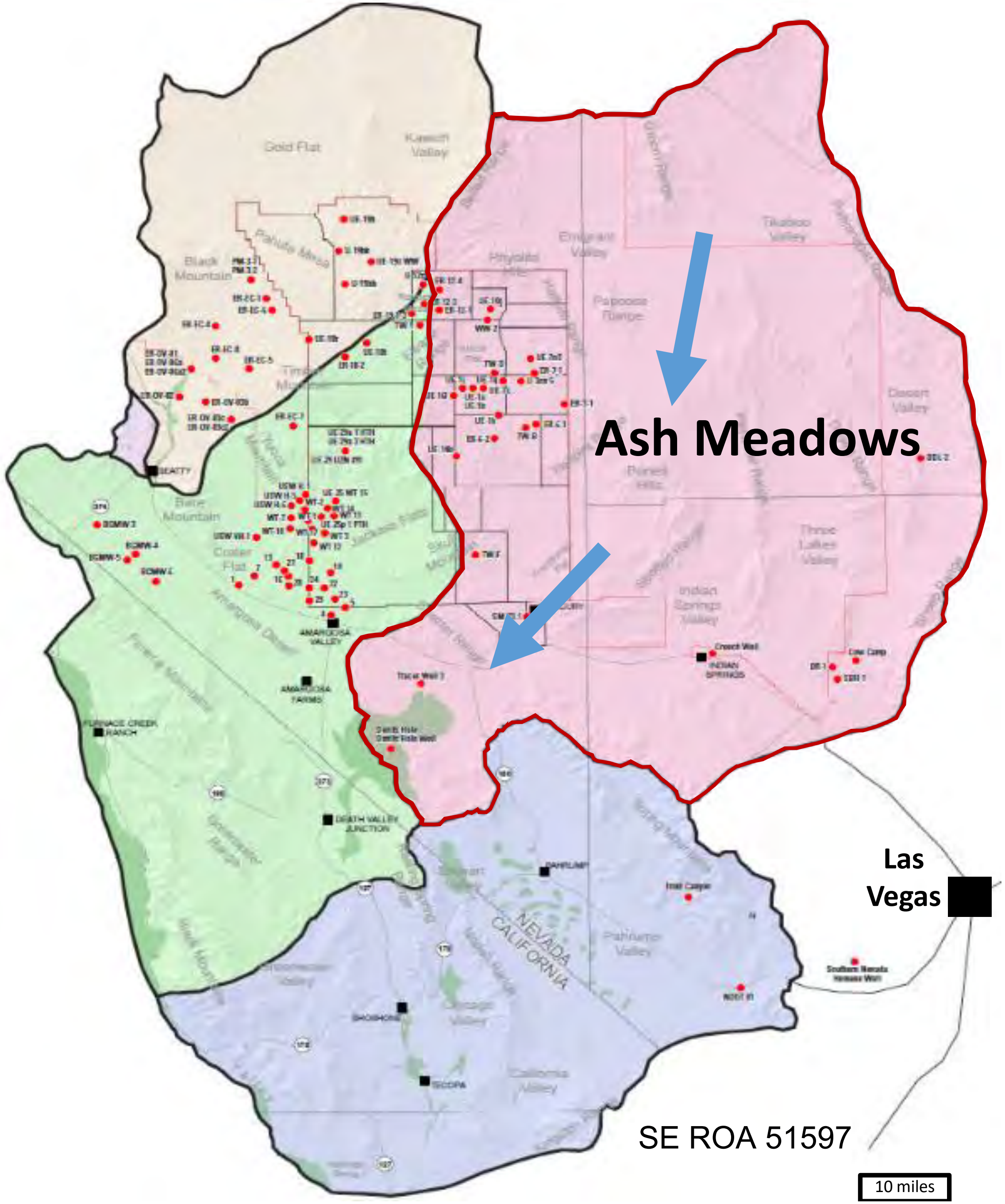
JA_16727

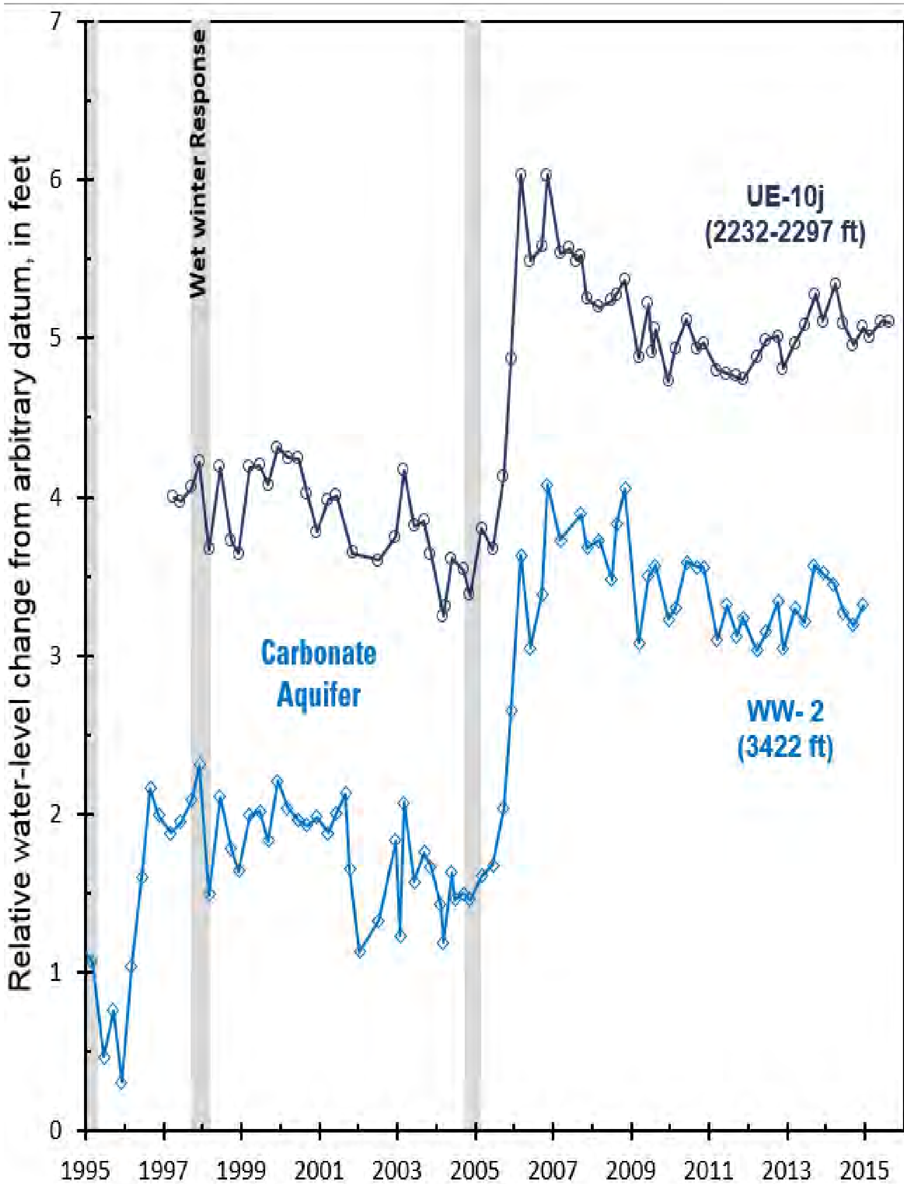
DEATH VALLEY REGIONAL FLOW SYSTEM

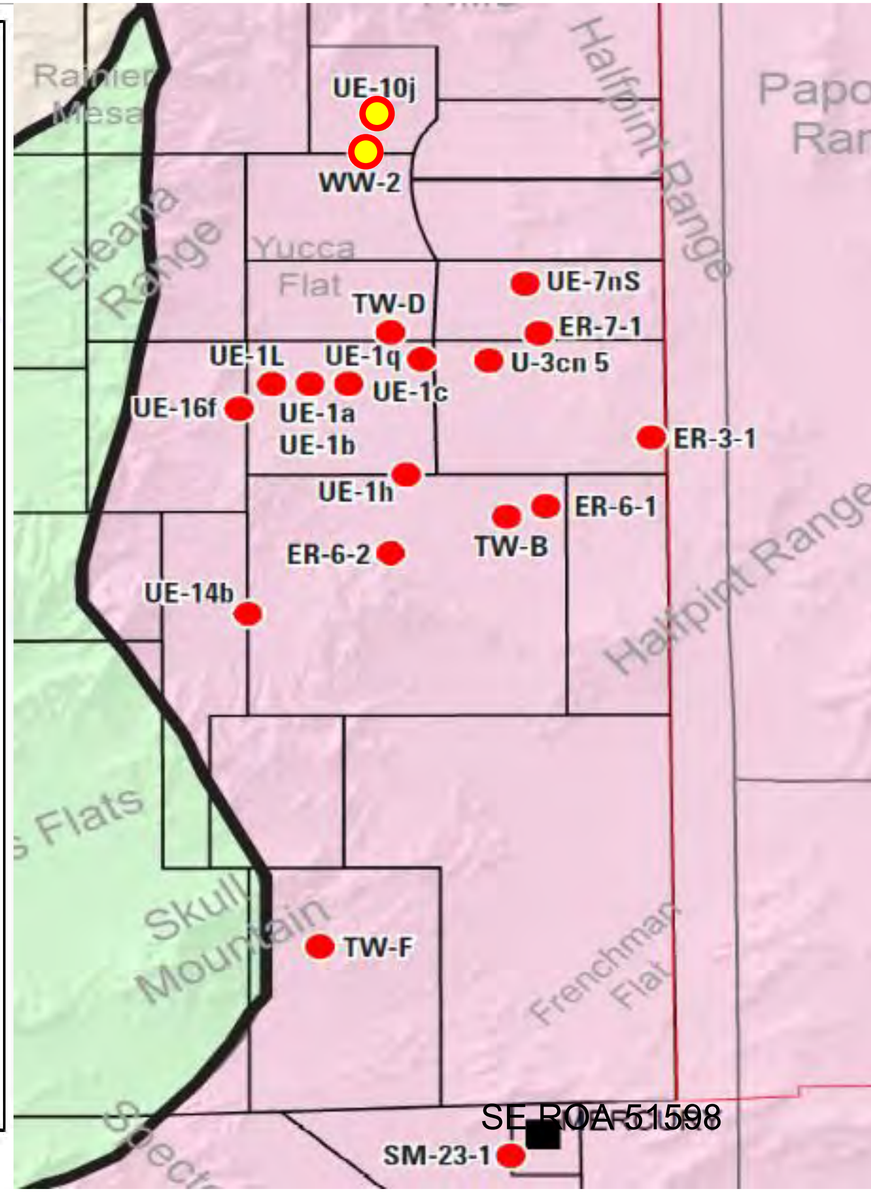
Focus on wells distant from pumping



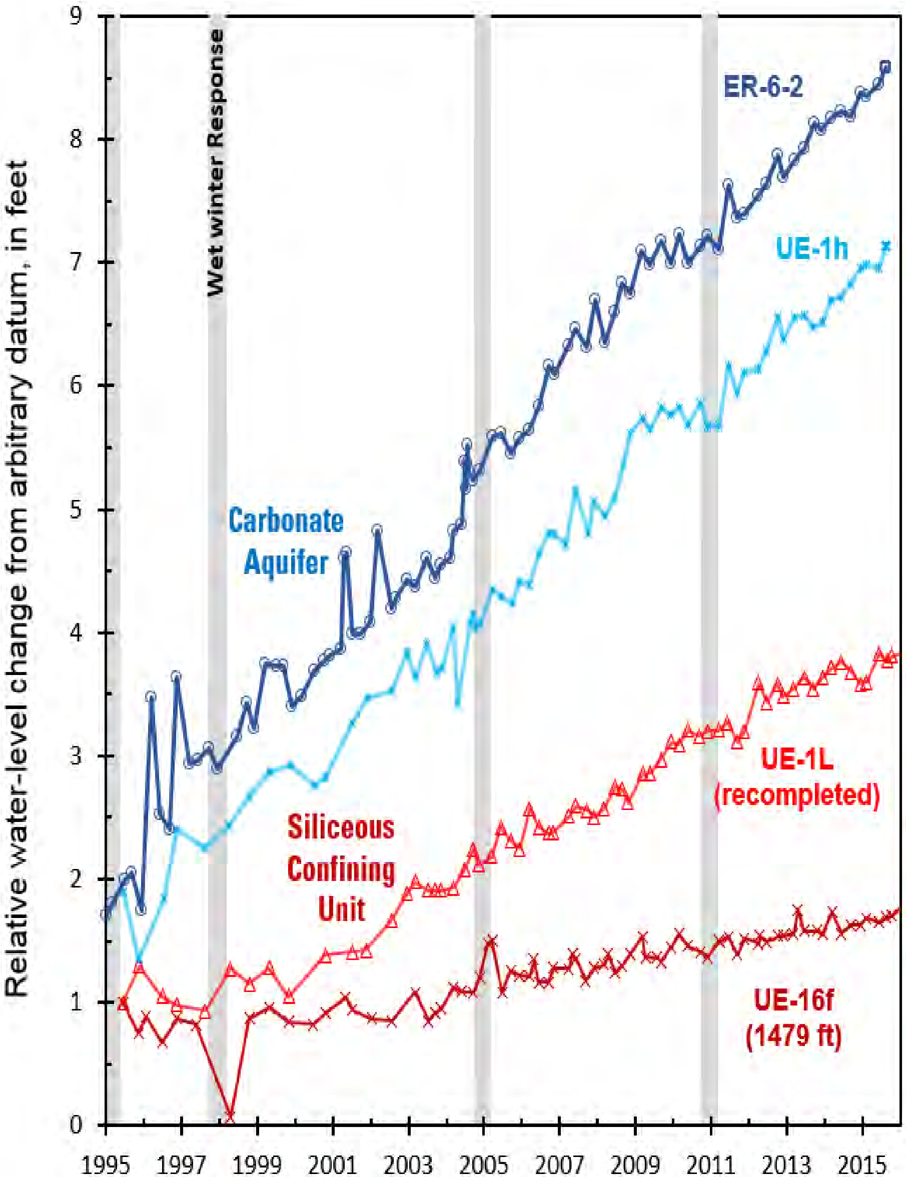
JA_16728

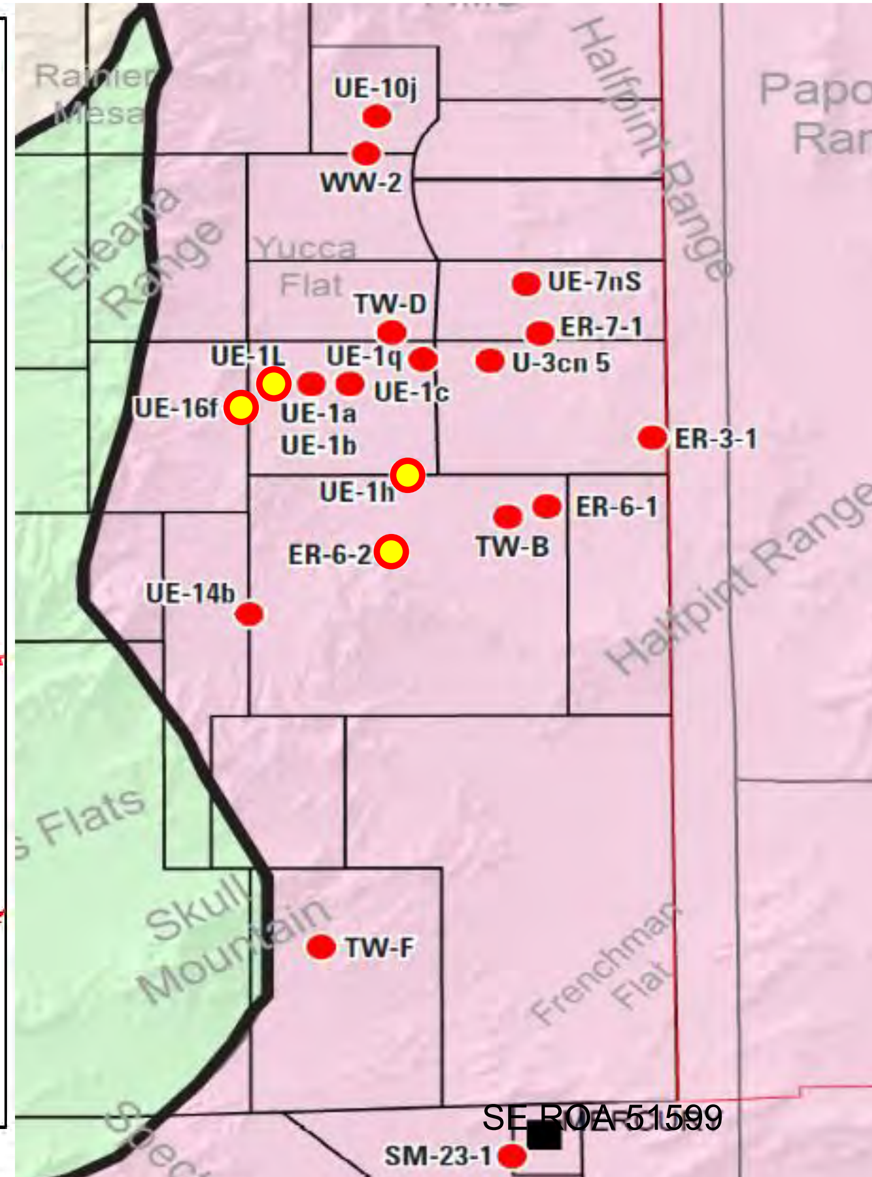
Northern Yucca Flat



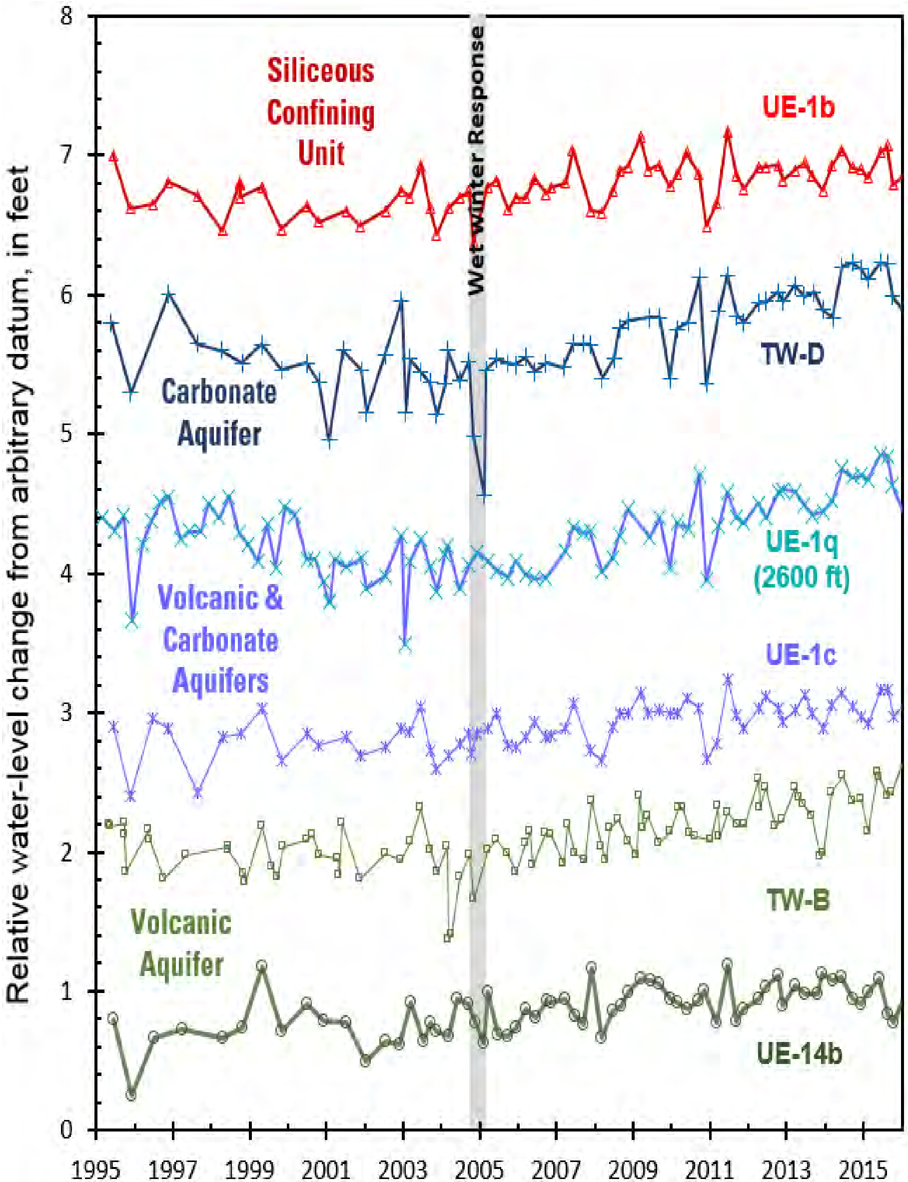


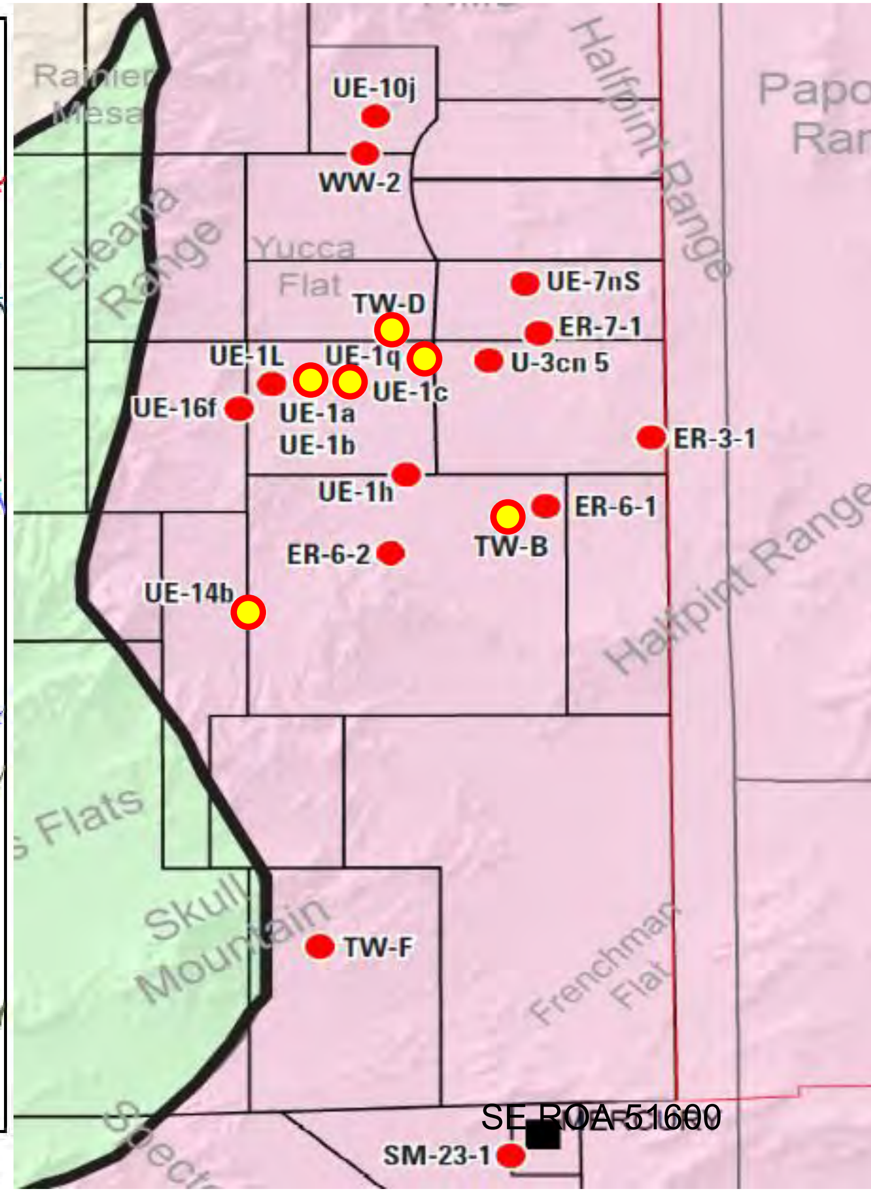
WESTERN YUCCA FLAT





CENTRAL YUCCA FLAT





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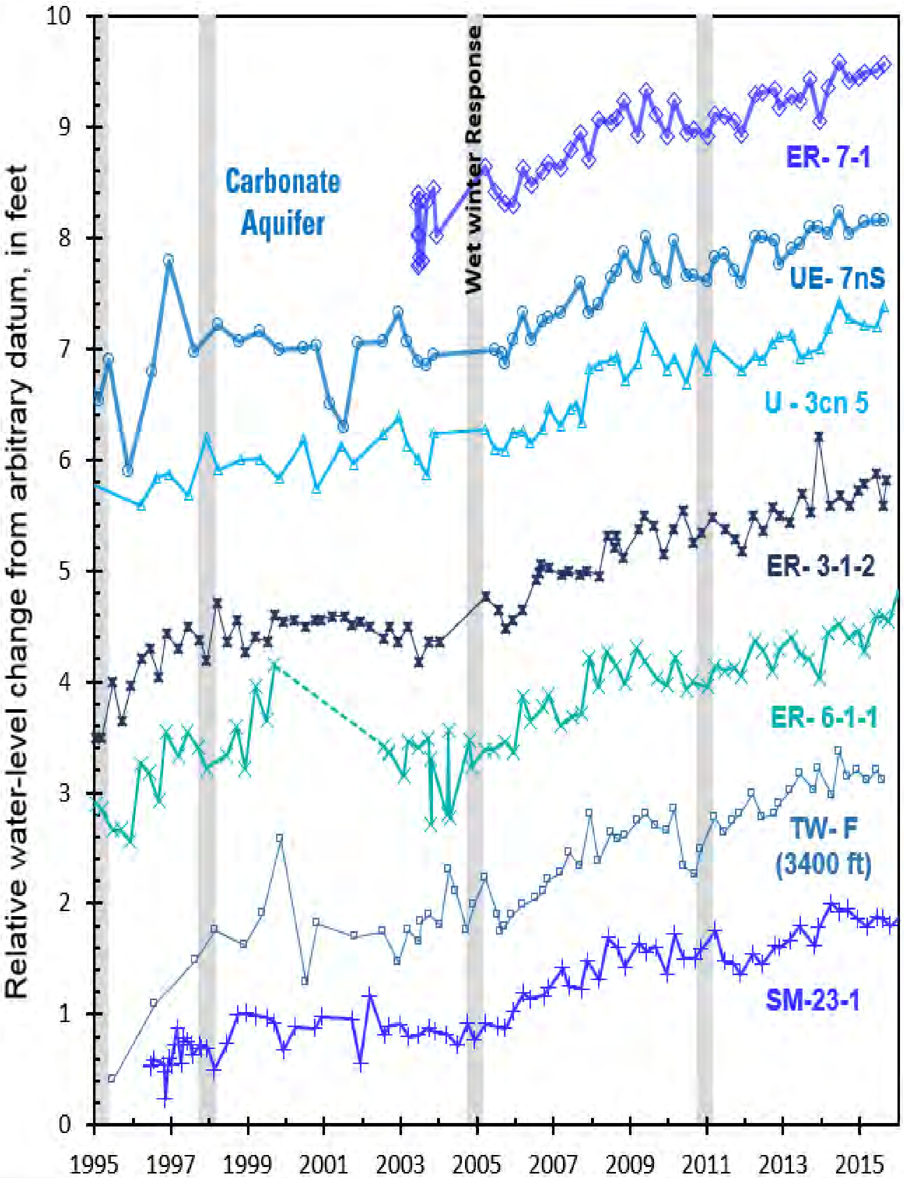
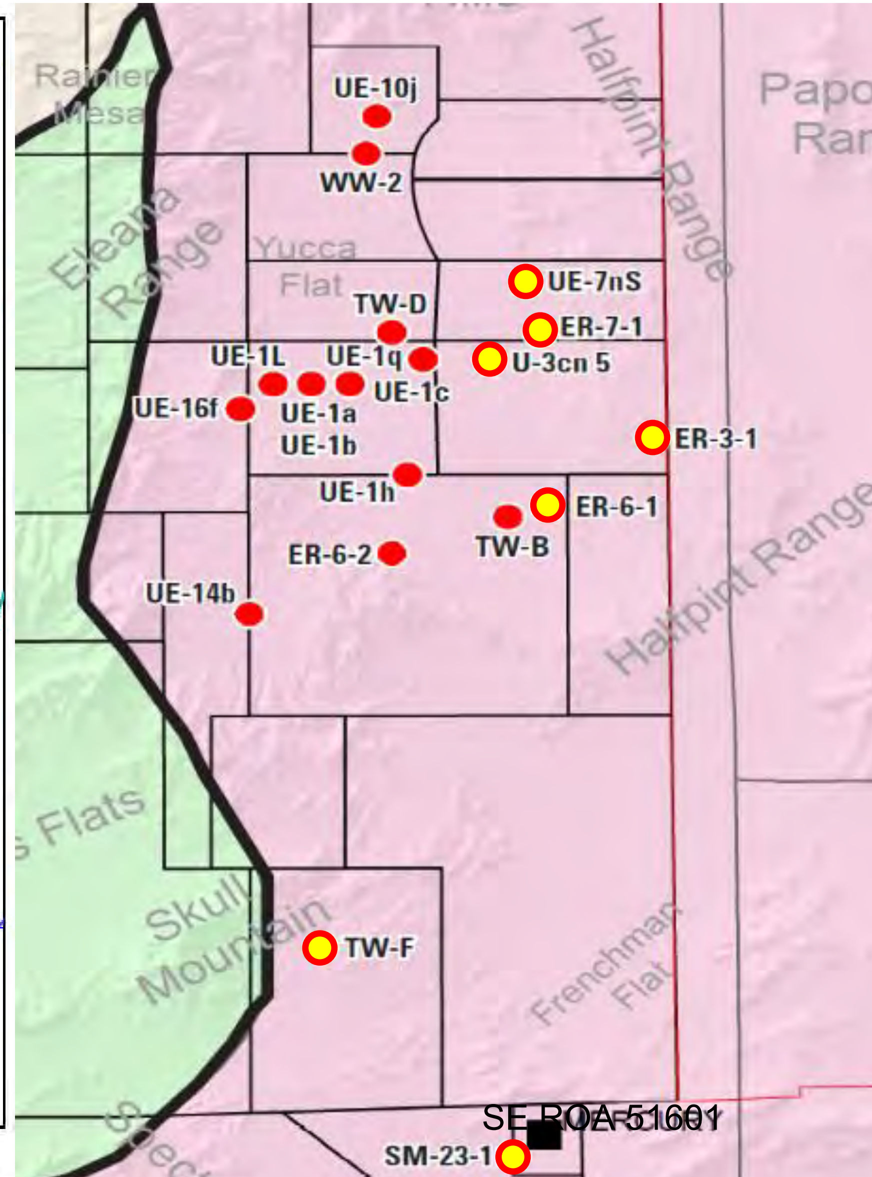
EASTERN AND SOUTHERN YUCCA FLAT





SE ROA 51601

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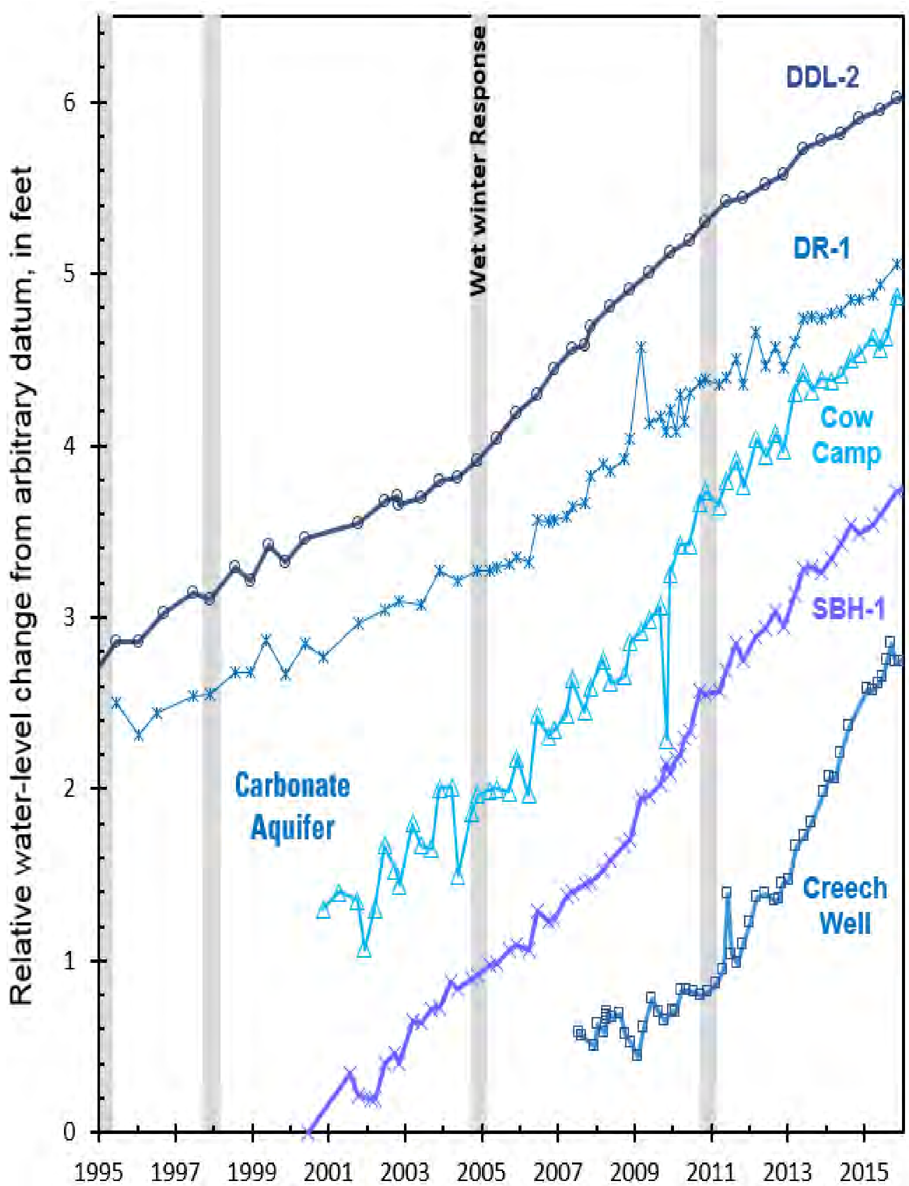
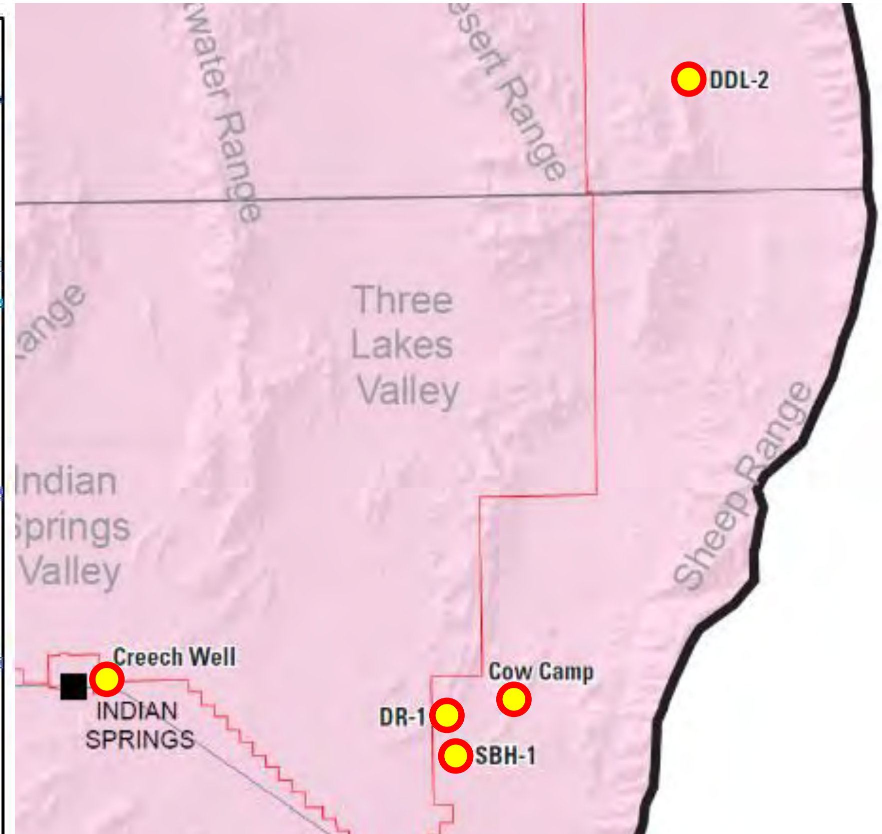
SHEEP RANGE – INDIAN SPRINGS





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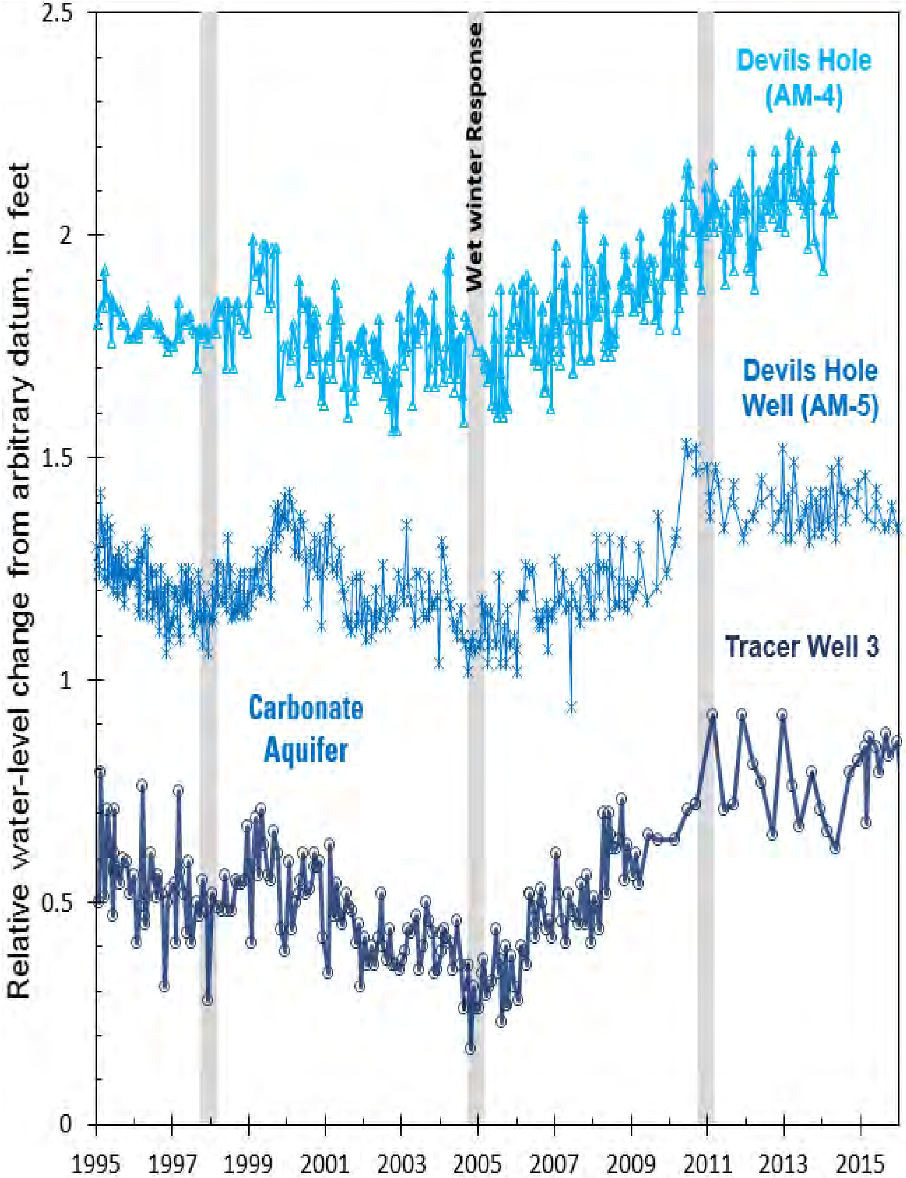
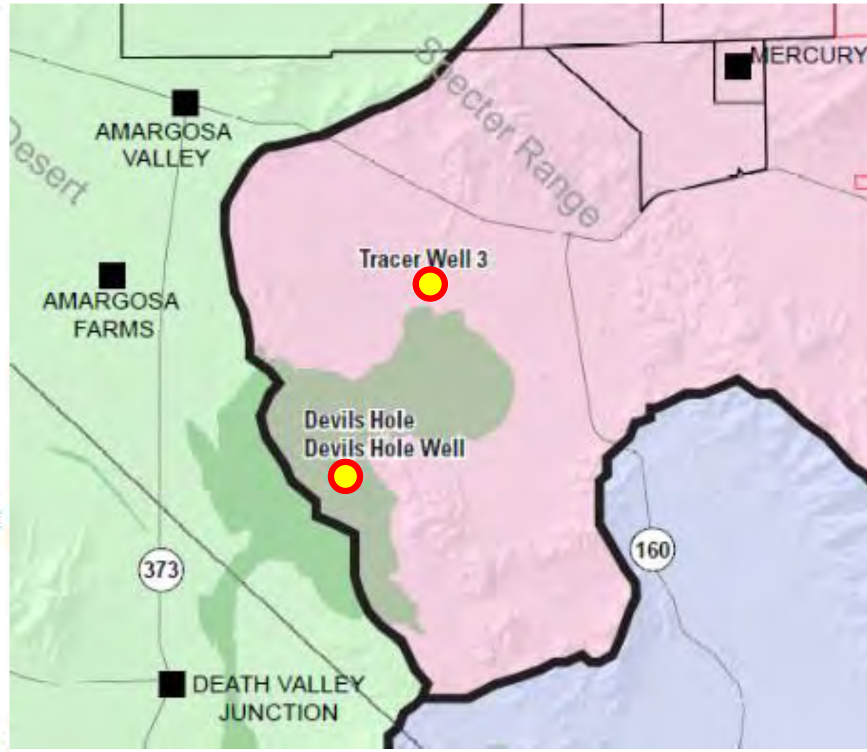
JA_16733

DEVILS HOLE



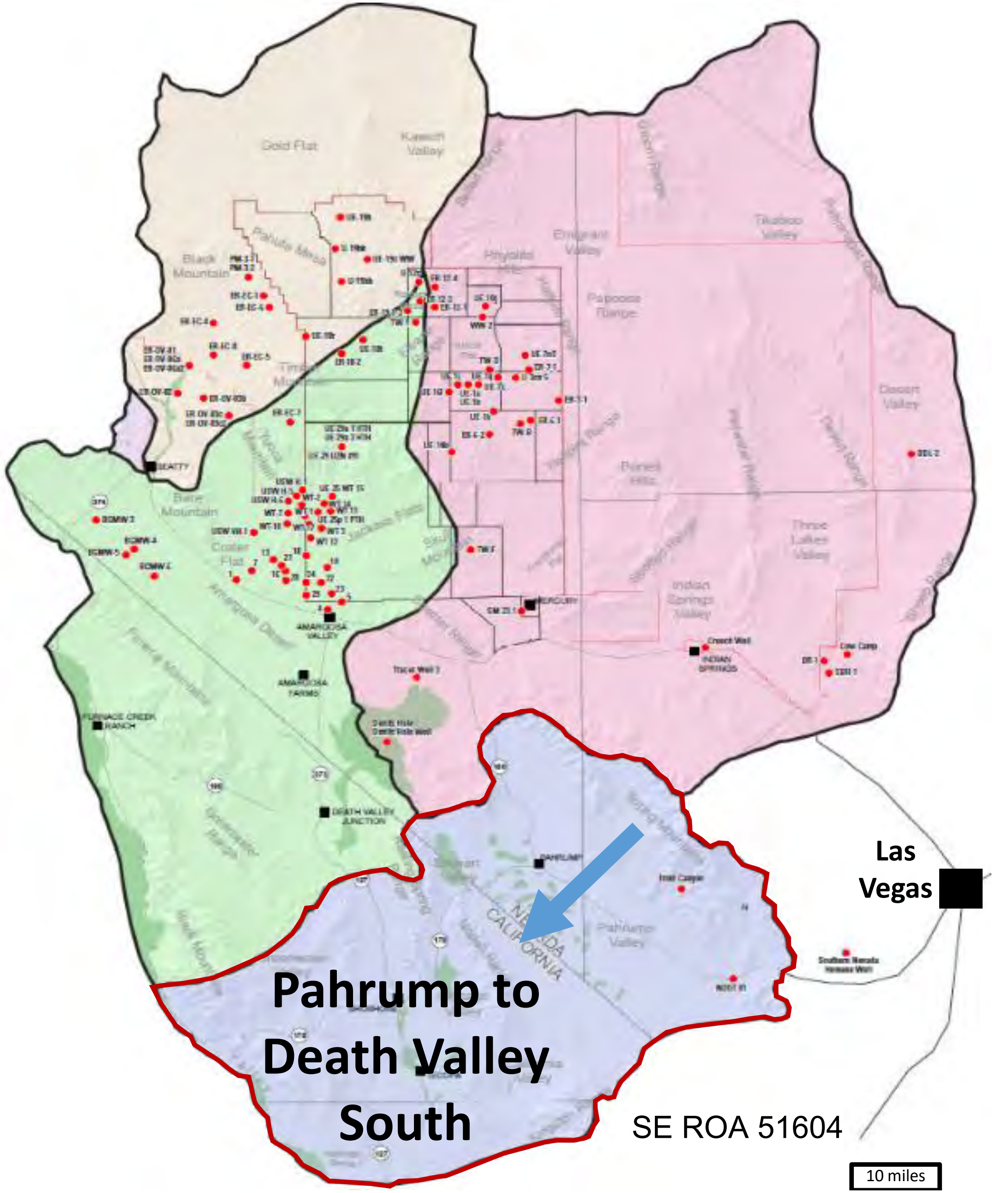


SE ROA 51603

JA_16734

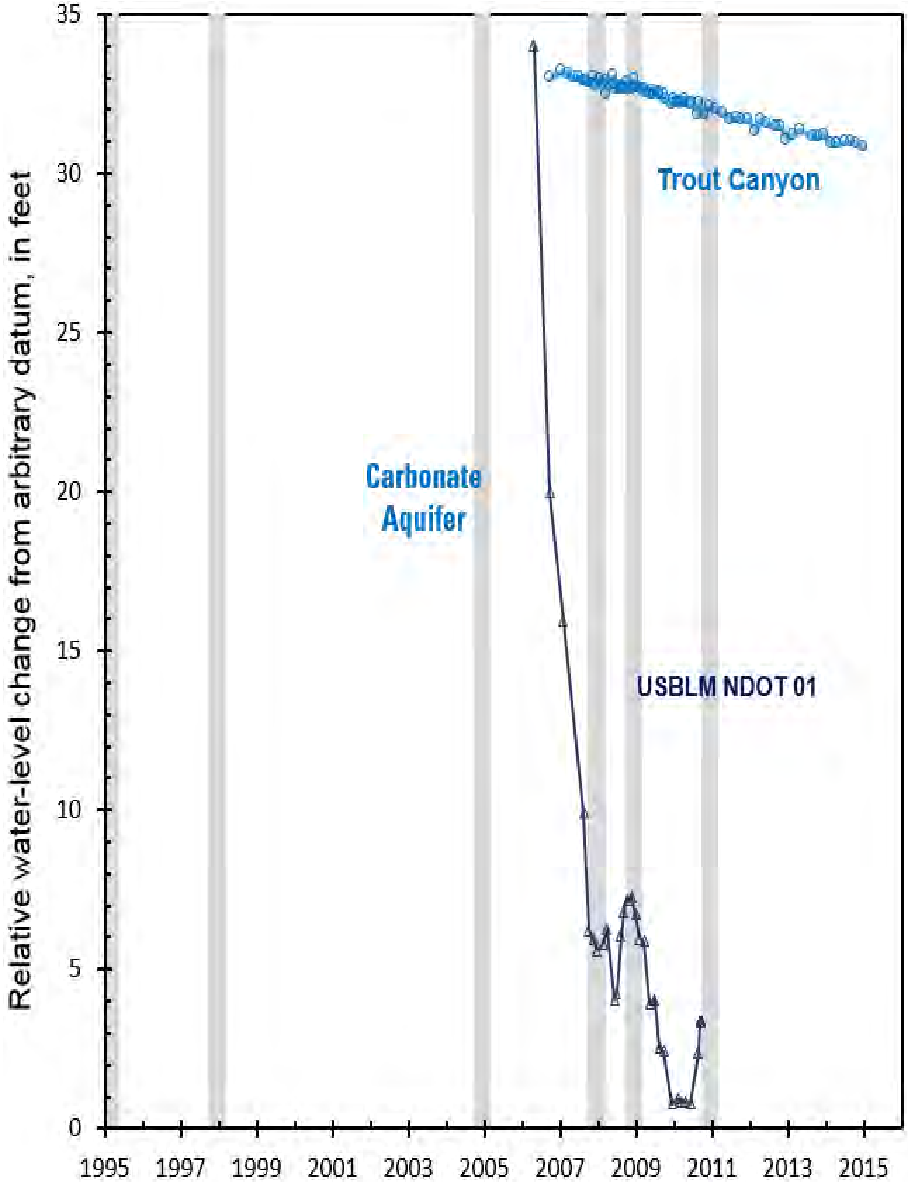
DEATH VALLEY REGIONAL FLOW SYSTEM

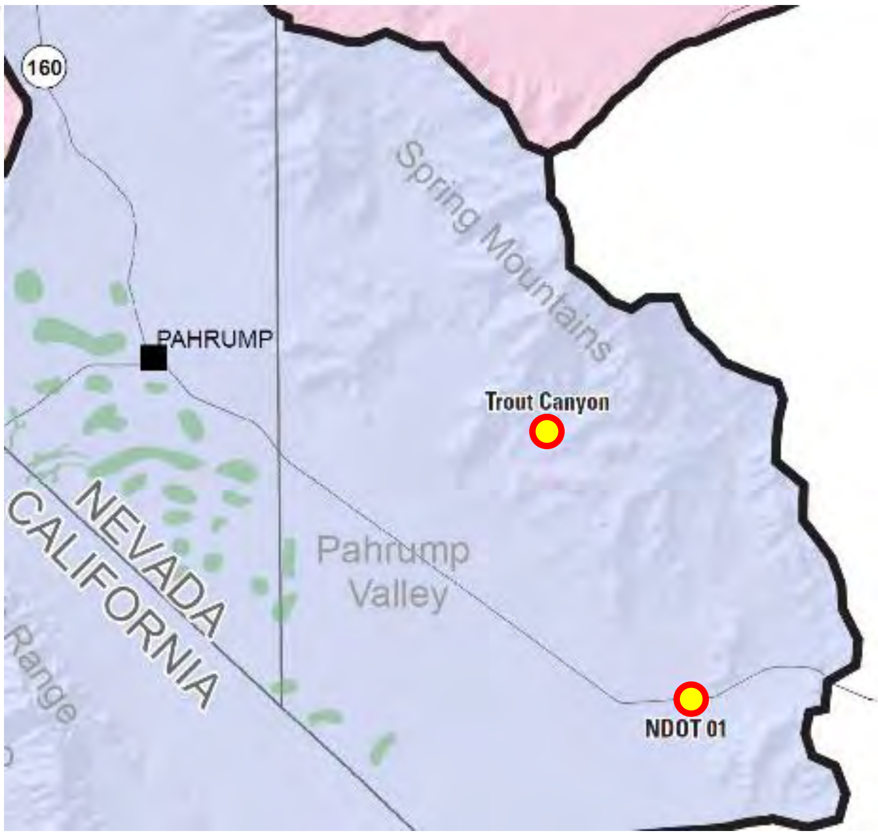
Focus on wells distant from pumping



JA_16735

Spring Mountains





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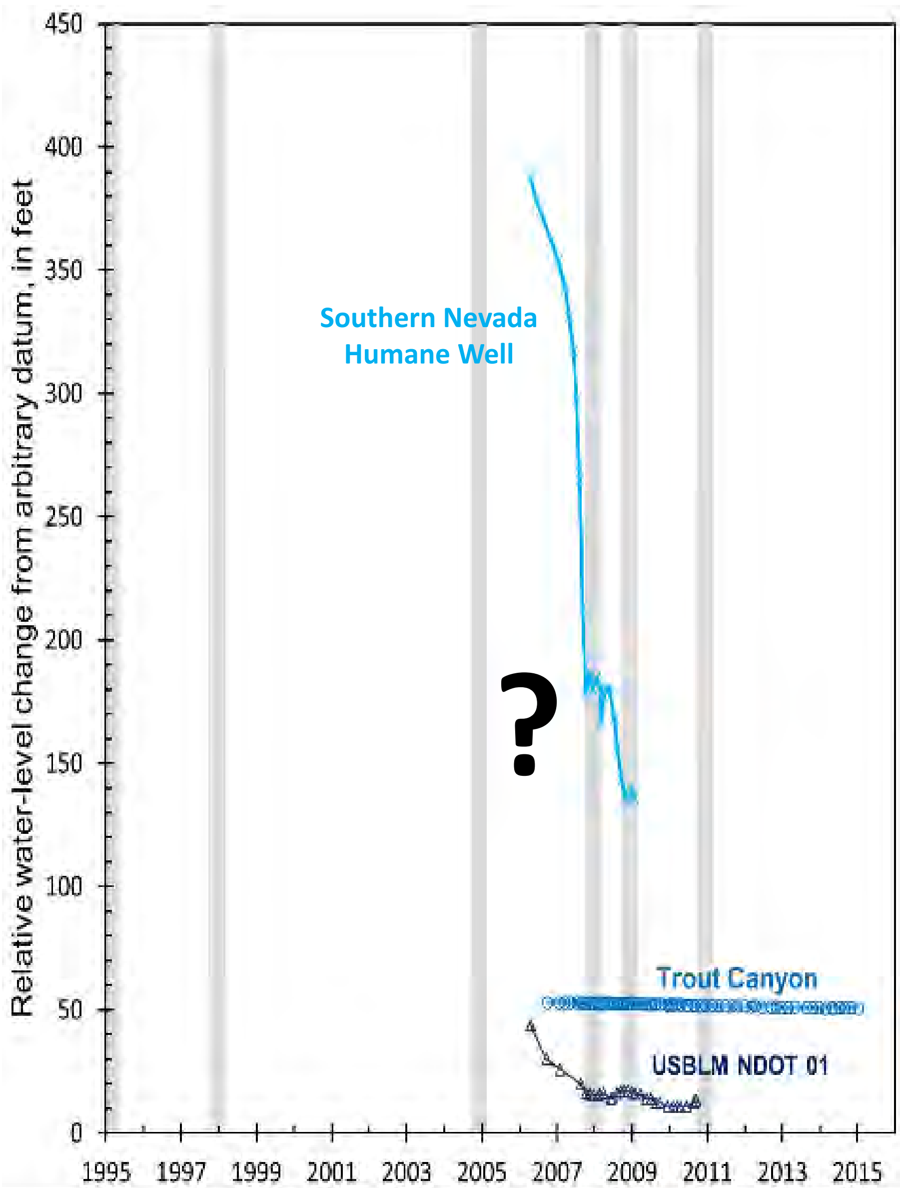
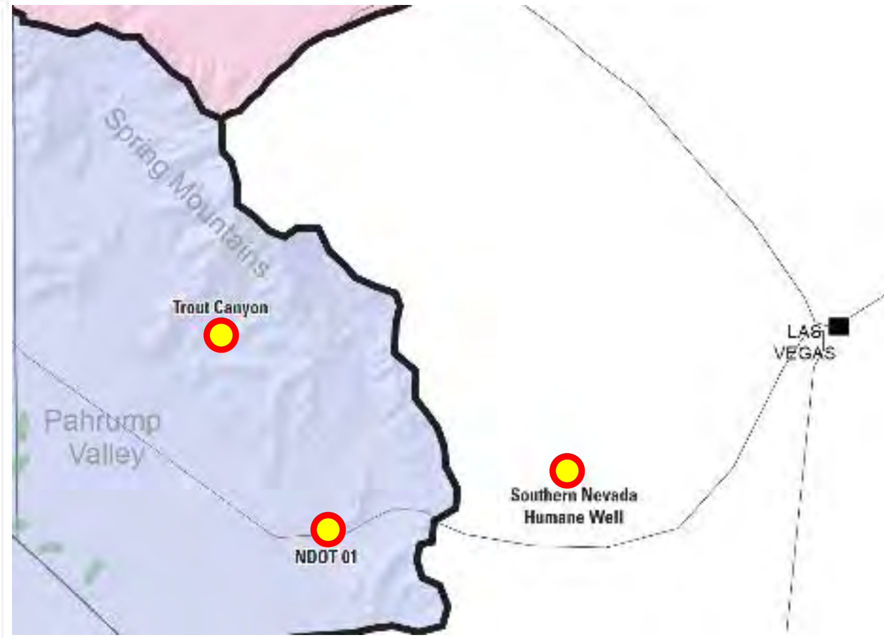
SPRING MOUNTAINS





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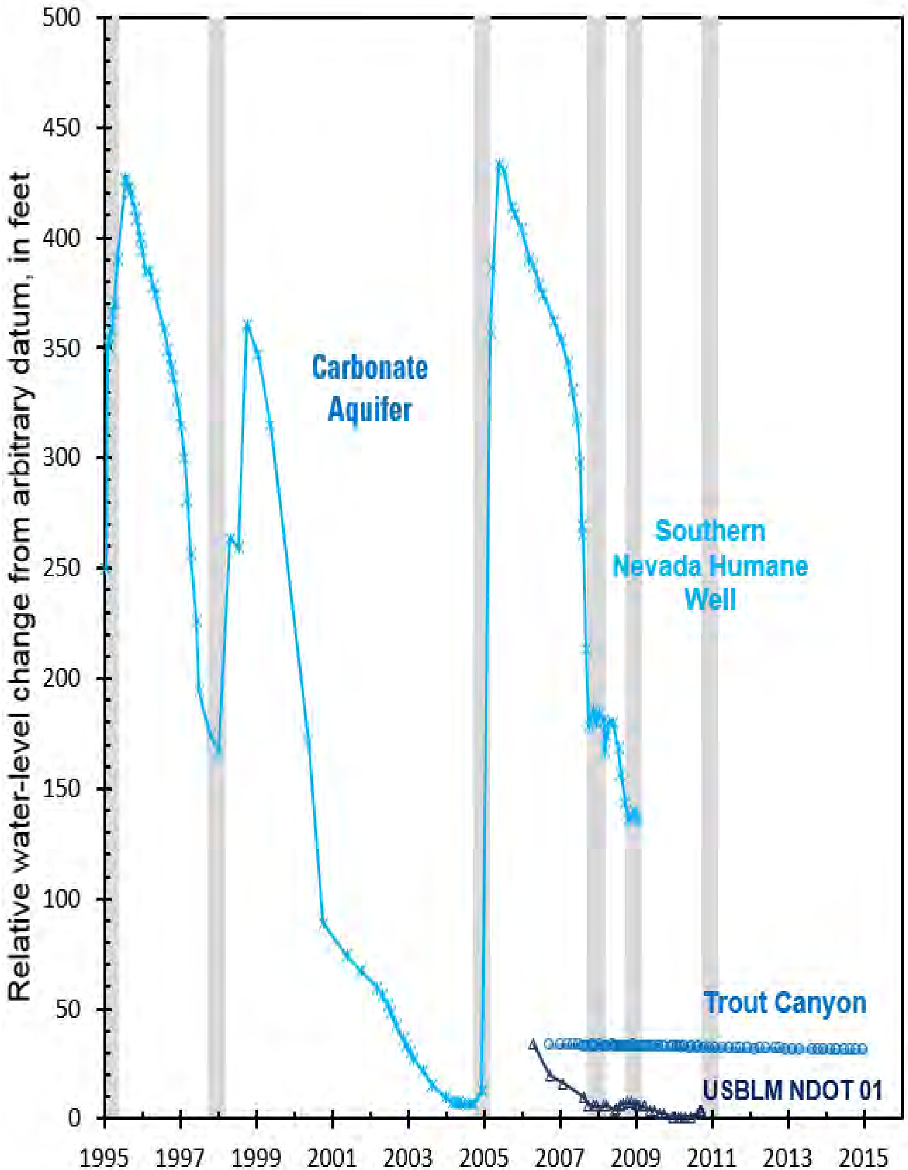
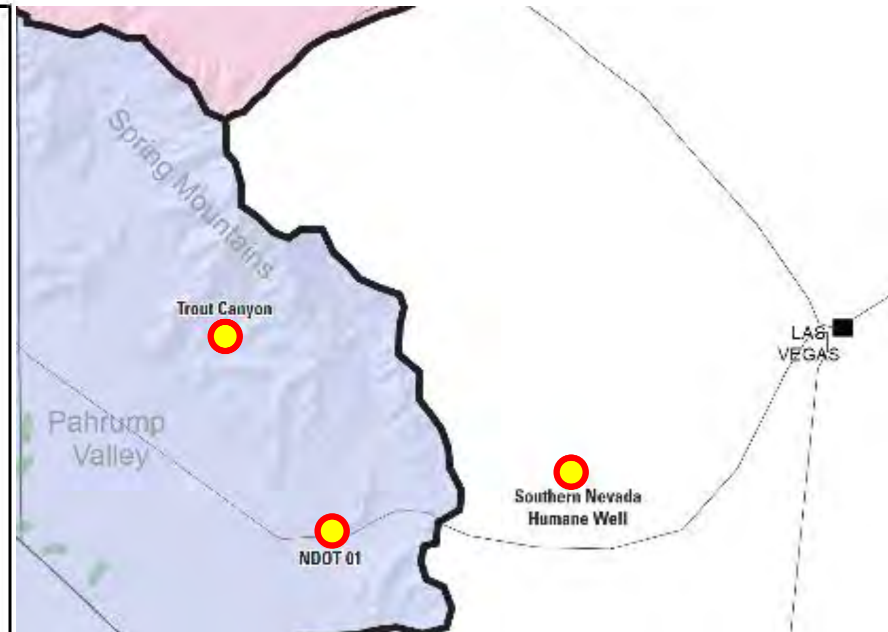
SPRING MOUNTAINS





SE ROA 51607

Main Point #1

Water levels in wells distant from pumping have been rising in southern Nevada

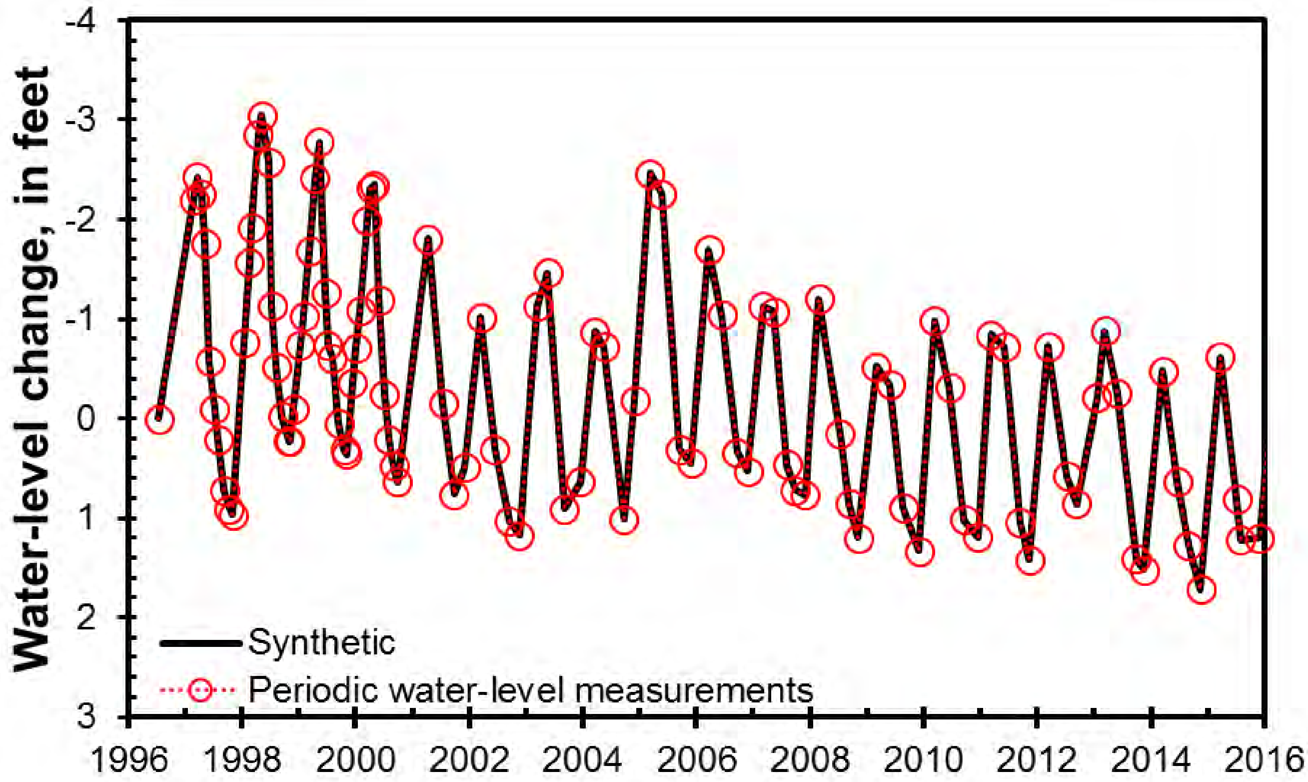
SE ROA 51608

Can you attribute rising water-level trends to groundwater recharge?

SE ROA 51609

WATER-LEVEL MODELS (WLM)

- WLMs used to analytically model water-level trends by fitting periodic water levels to a synthetic curve



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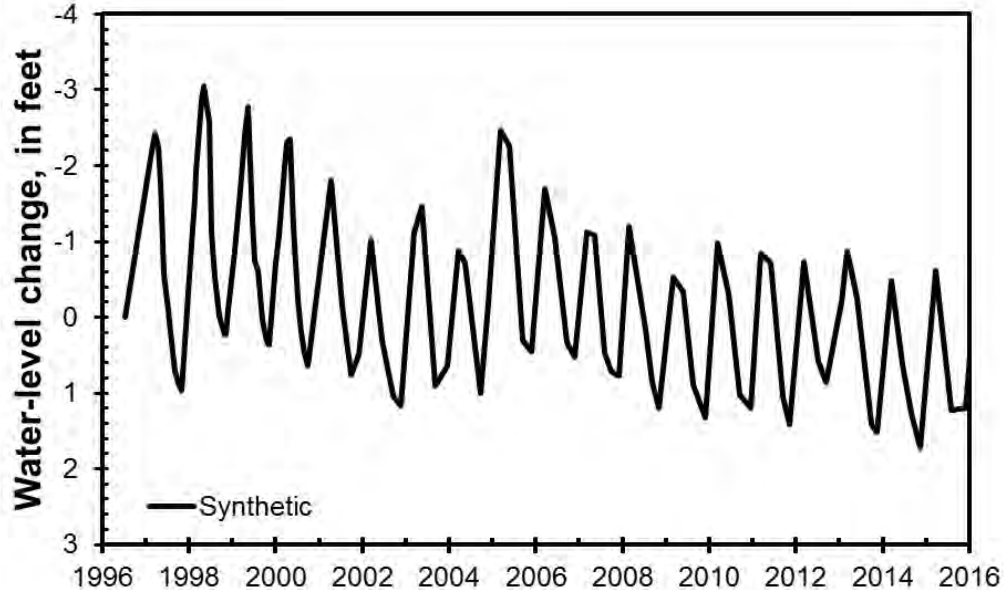
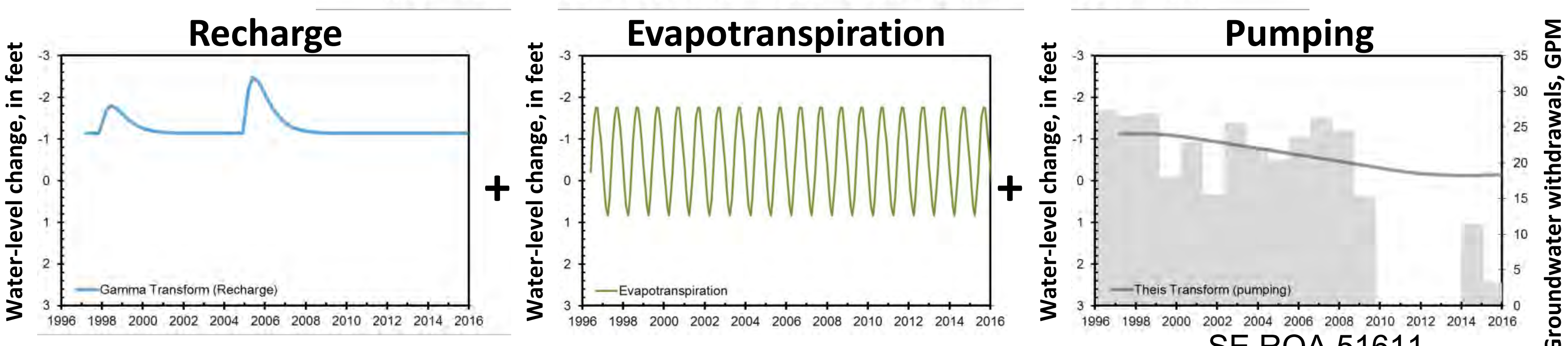
WATER-LEVEL MODELS (WLM)

- Synthetic curve is sum of one or more time series components that likely explain water-level fluctuations





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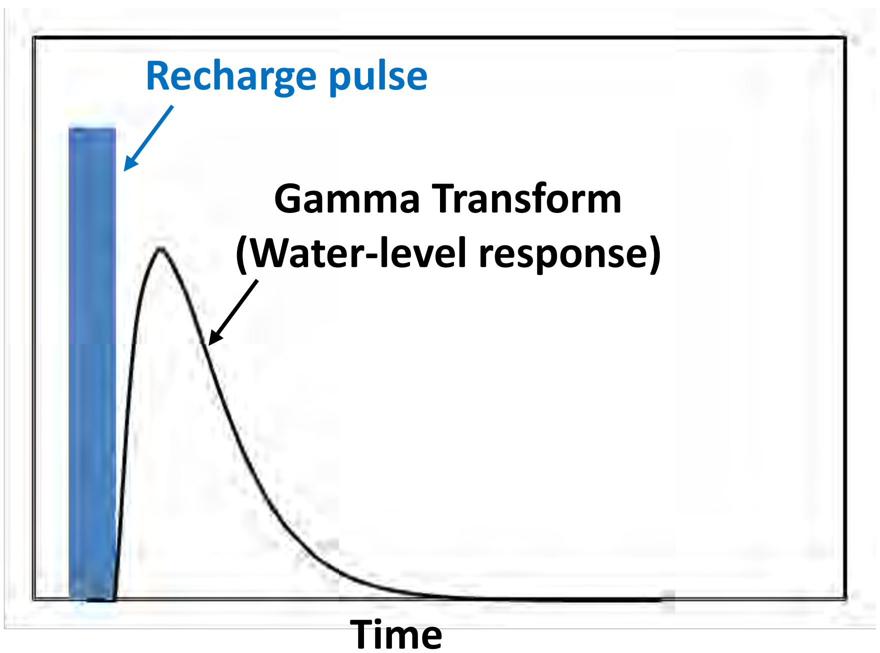
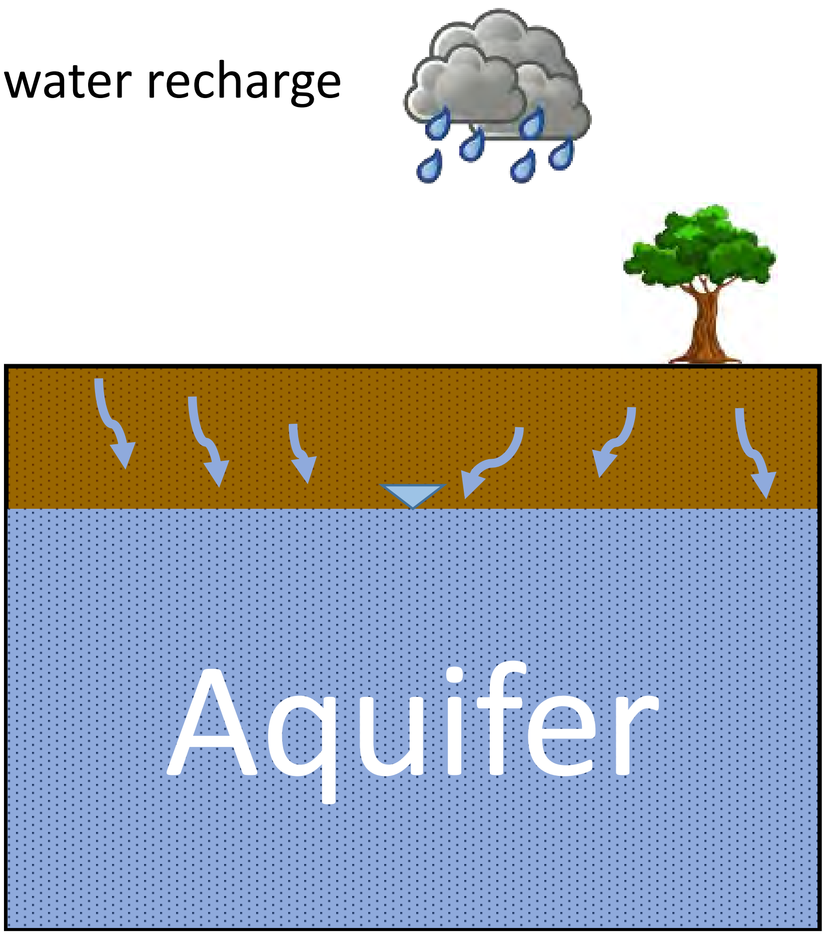
WATER-LEVEL MODELS (WLM)

Time series components

Gamma transform

- simulates precipitation-derived groundwater recharge





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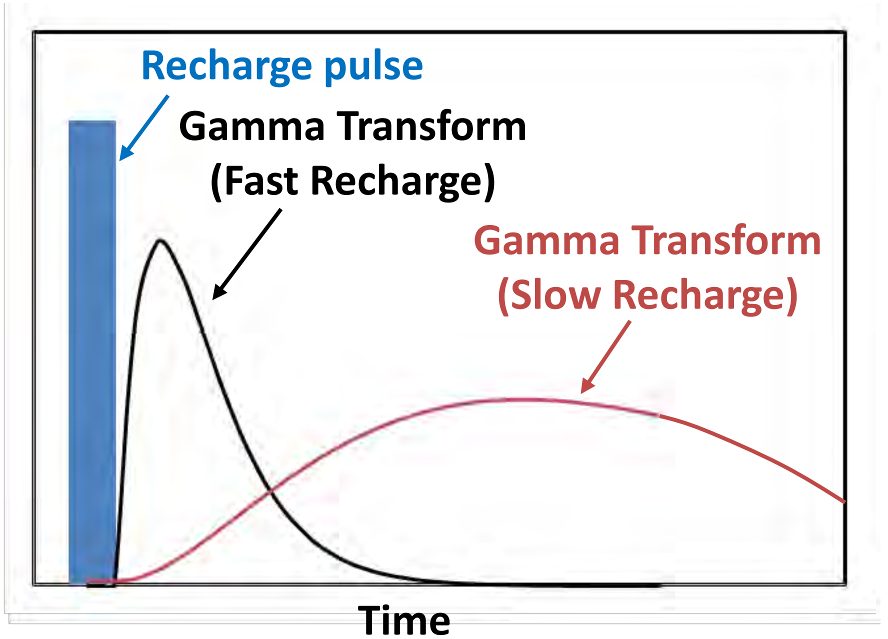
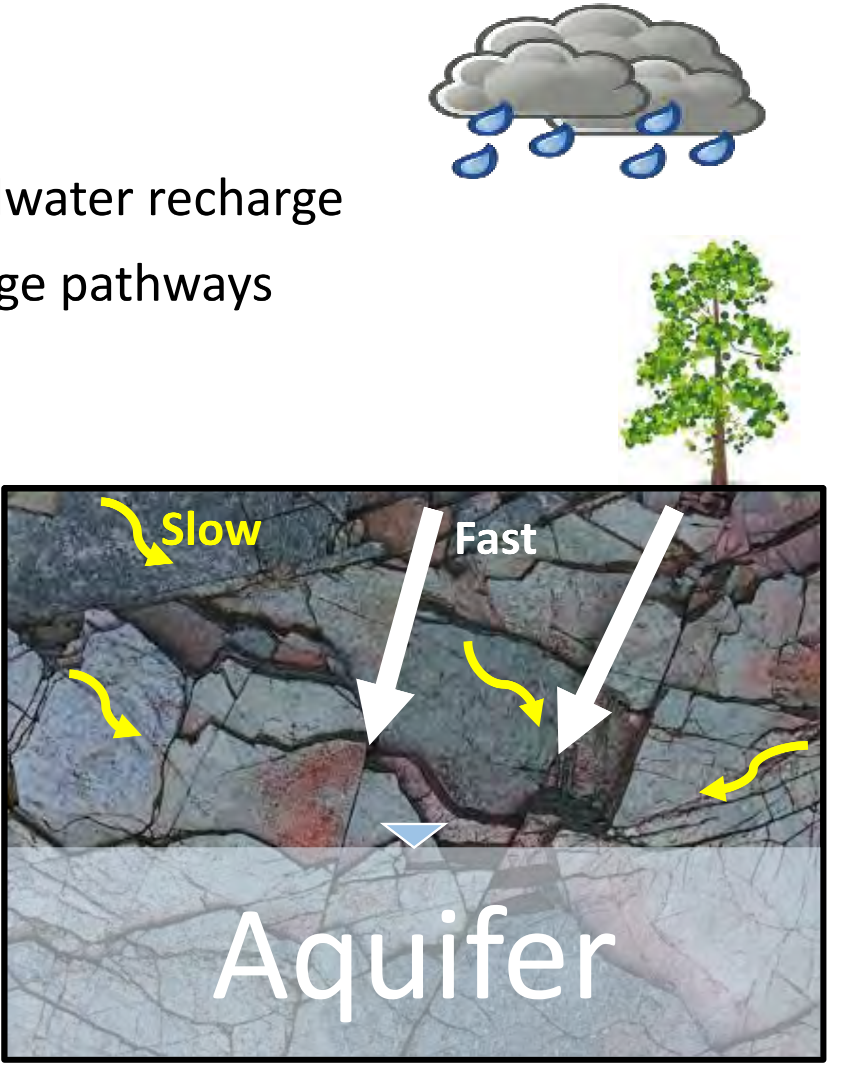
WATER-LEVEL MODELS (WLM)

Time series components

Gamma transform

- simulates precipitation-derived groundwater recharge
- used to represent fast and slow recharge pathways in a thick unsaturated zone





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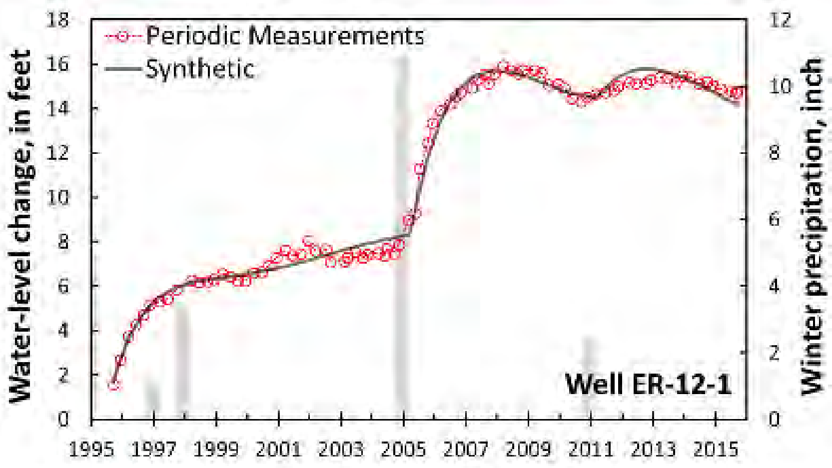
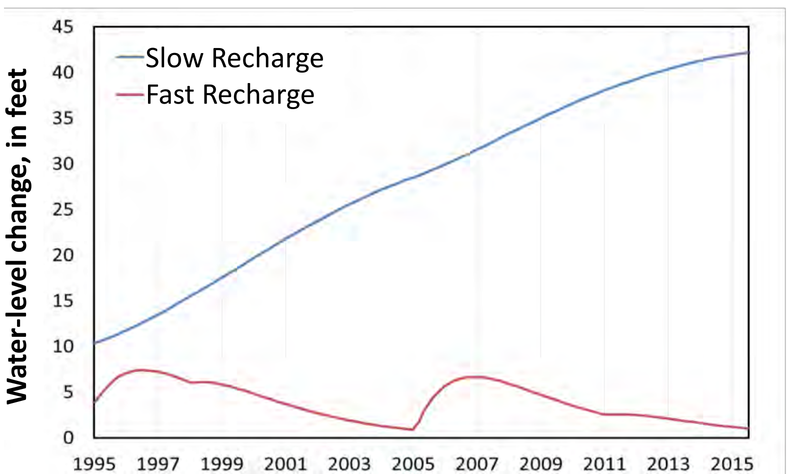
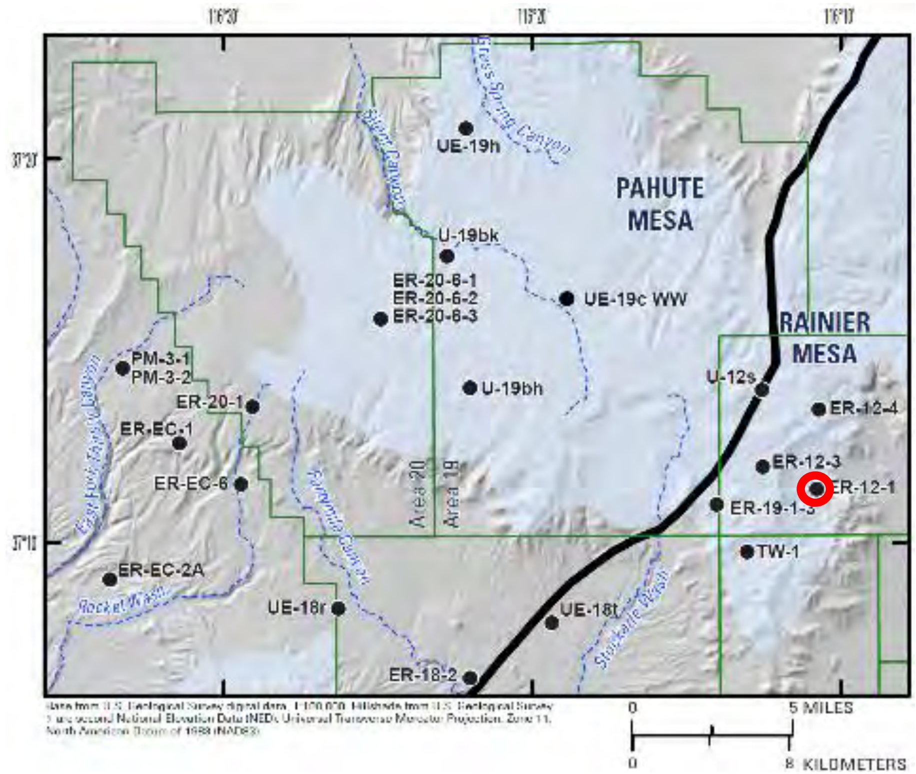
PRECIPITATION WLMs







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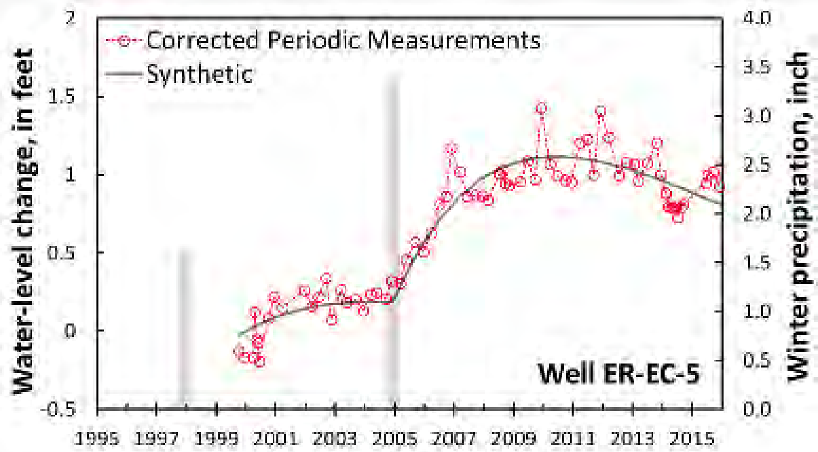
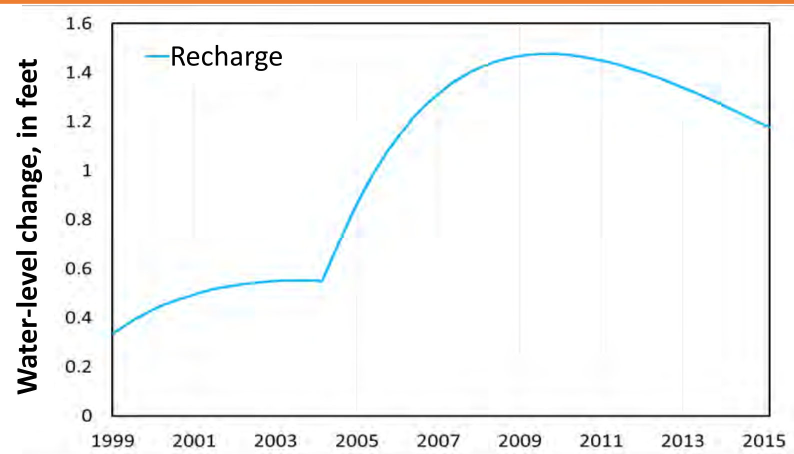
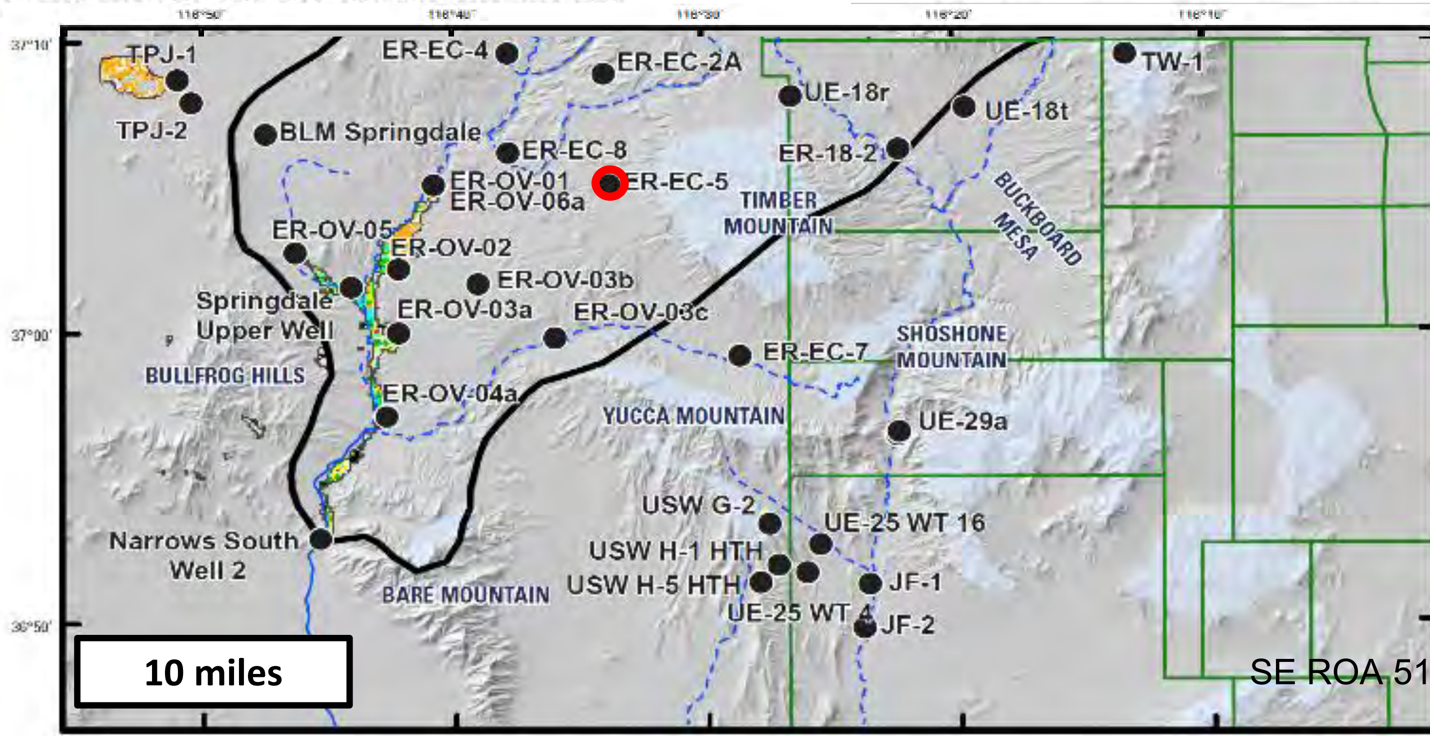
PRECIPITATION WLMs







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JA_16746

Main Point #2

Rising water-level trends attributed to groundwater recharge from wet winters

How can water levels be rising due to recharge in a drought?

Conceptual model

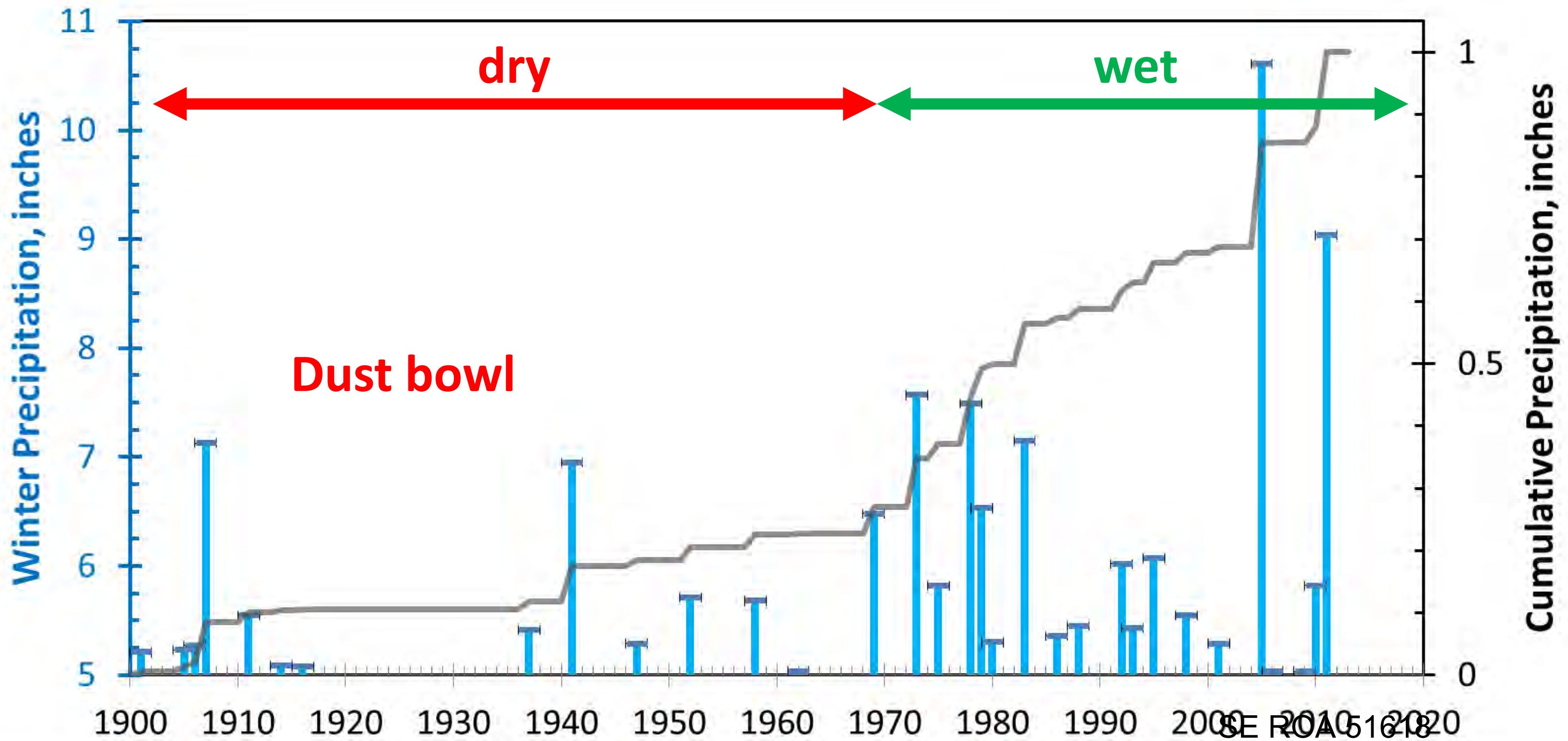
SE ROA 51617

How do we define steady state?

Steady state: water levels do not change over a period of time

What is that period of time?

- DVRFS: assume steady state is on a century scale



*Precipitation data from Western Regional Climate Center, 2015, Standardized precipitation index: accessed February 2015 at URL http://www.wrcc.dri.edu/spi

SE ROA 51618

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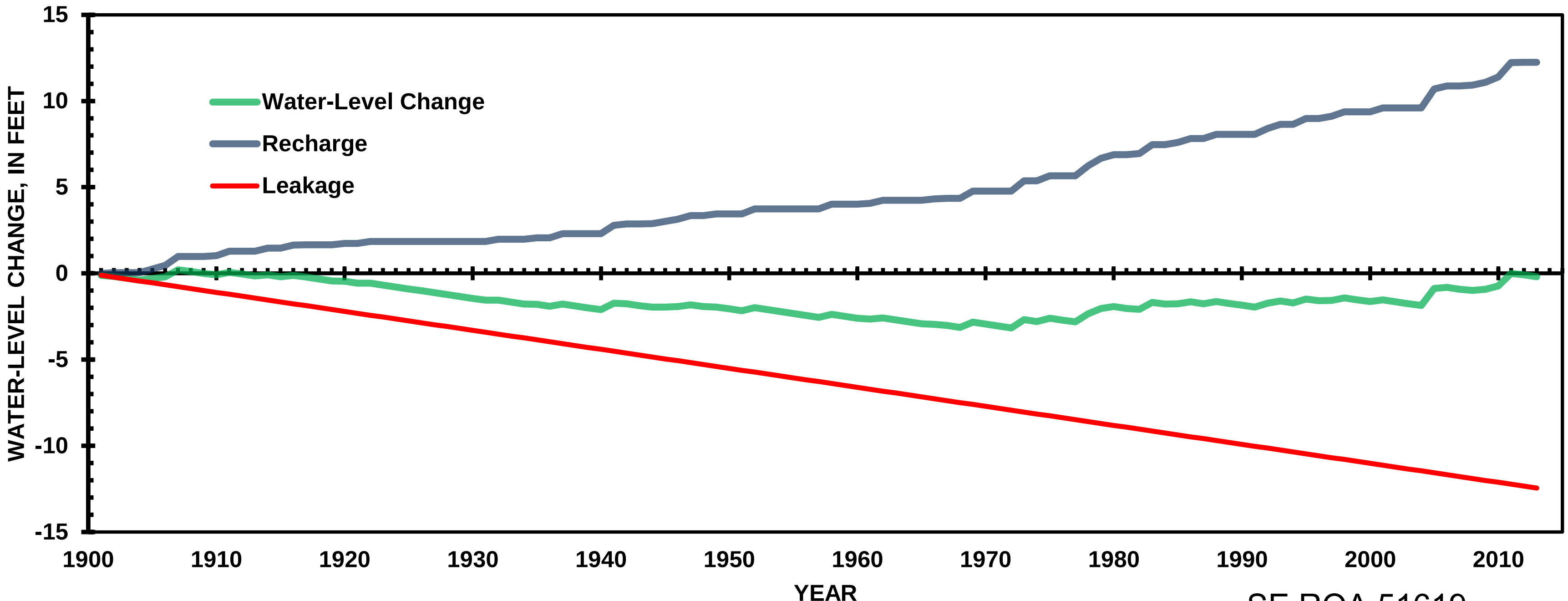
HOW DO WE DEFINE STEADY STATE?

In steady state: Recharge = Discharge

Hypothetically, expect water-level decline from 1900-1970

water-level rise from 1970-2015



SE ROA 51619

Main Point #3

Short-term (decadal) rising trend part of longer-term (century-scale) steady-state condition

SE ROA 51620

Main Points

Water levels are rising in southern Nevada

Water-level rise is attributed to groundwater recharge from wet winters

Water-level rise is a ***short-term*** water-level fluctuation in the ***long-term*** period of steady state

Questions?

SE ROA 51622

Groundwater

Depletion and Capture: Revisiting "The Source of Water Derived from Wells"

by L.F. Konikow[1] and S.A. Leake[2]

Abstract

A natural consequence of groundwater withdrawals is the removal of water from subsurface storage, but the overall rates and magnitude of groundwater depletion and capture relative to groundwater withdrawals (extraction or pumpage) have not previously been well characterized. This study assesses the partitioning of long-term cumulative withdrawal volumes into fractions derived from storage depletion and capture, where capture includes both increases in recharge and decreases in discharge. Numerical simulation of a hypothetical groundwater basin is used to further illustrate some of Theis' (1940) principles, particularly when capture is constrained by insufficient available water. Most prior studies of depletion and capture have assumed that capture is unconstrained through boundary conditions that yield linear responses. Examination of real systems indicates that capture and depletion fractions are highly variable in time and space. For a large sample of long-developed groundwater systems, the depletion fraction averages about 0.15 and the capture fraction averages about 0.85 based on cumulative volumes. Higher depletion fractions tend to occur in more arid regions, but the variation is high and the correlation coefficient between average annual precipitation and depletion fraction for individual systems is only 0.40. Because 85% of long-term pumpage is derived from capture in these real systems, capture must be recognized as a critical factor in assessing water budgets, groundwater storage depletion, and sustainability of groundwater development. Most capture translates into streamflow depletion, so it can detrimentally impact ecosystems.

Introduction

In a classic and often-cited paper, Theis (1940) explains the sources of water derived from a pumping well. Among other things, Theis (1940) concludes that "All water discharged by wells is balanced by a loss of water somewhere. This loss is always to some extent and in many cases largely from storage in the aquifer. Some groundwater is always mined." He then notes that "After sufficient time has elapsed ... further discharge by wells will be made up at least in part by an increase in the recharge if previously there has been rejected recharge. ... further discharge by wells will be made up in part by a diminution in the natural discharge." The combination of increased recharge and decreased discharge is termed "capture" (Lohman et al. 1972; Bredehoeft and Durbin 2009; Leake 2011).

These generic relations show that at early times the principal source of water to a well is from depletion of storage in the aquifer (Figure 1). With increasing time, the fraction of pumpage derived from storage depletion (a nondimensional "depletion fraction") tends to decrease, and the fraction derived from capture increases. Eventually, provided that sufficient potential increases in recharge and decreases in discharge are available, a new equilibrium will be achieved when no more water is derived from storage and heads or water levels in the aquifer stabilize. The actual response time for an aquifer system to reach a new equilibrium is a function of the dimensions, hydraulic properties, and boundary conditions for the specific case. The response time will change as these conditions are varied. For example, the response time will decrease as the hydraulic diffusivity of the aquifer increases (see Theis 1940; Barlow and Leake 2012). The response time can range from days to millennia (Bredehoeft and Durbin 2009; Walton 2011). An important corollary to Theis' (1940) principles is that the average predevelopment rate of natural recharge itself is largely irrelevant to storage depletion and capture responses (Bredehoeft et al. 1982; Bredehoeft 1997; Barlow and Leake 2012). However, the natural recharge does serve as a constraint on capture—in the sense that it controls the natural predevelopment groundwater discharge, which is subject to capture by pumping wells.

Capture includes several factors and processes, but is often considered synonymous with (or dominated by) streamflow depletion (e.g., Alley et al. 1999; Barlow and Leake 2012). This includes increased recharge through

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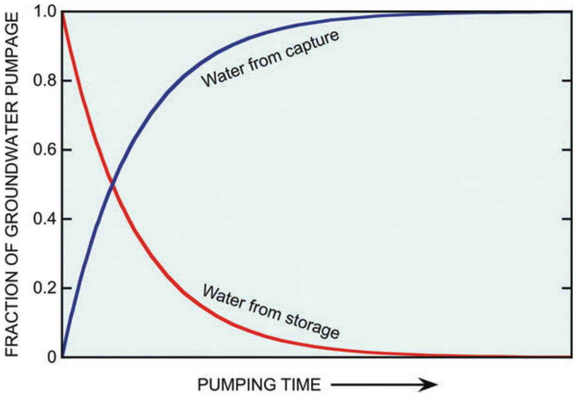

Figure 1. Sources of water to a well can change with time. Time scale for curves depends on the hydraulic characteristics of the aquifer and the distance of the well from recharge and discharge locations. Modified from Alley et al. (1999) and Barlow and Leake (2012).

induced infiltration from streams (and other surface water bodies), as well as decreases in groundwater discharge to springs, streams, and other surface water bodies (i.e., decreases in base flow). However, capture can also include (1) increased recharge facilitated by water-table declines in areas where potential recharge from precipitation under natural conditions is rejected and runs off the land surface because high water tables preclude infiltration, and (2) decreased evapotranspiration in areas where the water table is close to the land surface but declines due to pumpage-induced drawdown (Theis 1940, 1941; Bredehoeft et al. 1982; Walton and McLane 2013). If recharge were to increase or discharge were to decrease, either coincidentally or through intentional water management policies (e.g., artificial recharge, especially using imported water, or phreatophyte control), the effects of well pumpage would be additionally offset or balanced accordingly.

Theis (1940) notes that aquifers are bounded; Walton and McLane (2013) expand on this point and note that because of bounds, full capture of supply components may not be feasible. What are the consequences if sufficient capture is not available to meet the demands imposed by substantially increased pumpage? Then the response will be constrained and a new equilibrium may never be achieved (Bredehoeft and Durbin 2009; Barlow and Leake 2012). As explained by Theis (1940), if the amount of pumping in an area exceeds the amount available for capture, water levels will continue to decline and pumping therefore will continue to be derived from storage depletion. Pumping under these constraints is clearly unsustainable. However, most studies in the literature of the effects of pumping on storage depletion and streamflow capture have assumed the presence of at least some boundary conditions that would allow a new equilibrium to be achieved. For example, there is always some flow in a stream or river that bounds an aquifer, as assumed by Theis (1941).

Surface-water bounding conditions that place limits on capture potential are more likely to occur in arid climates, but how common or large this constraint might be is uncertain. Lack of recognition of constraining boundary conditions might lead to erroneous estimates of storage depletion and sustainability. On the other hand, too much weight can be given to bounds on capture. For example, Wada et al. (2010) assume that groundwater storage depletion equals the excess of groundwater pumpage over natural recharge, except in humid climates. Pokhrel et al. (2012) estimate "unsustainable groundwater use" on the basis of estimated total water demand, and further assume that such groundwater use is equivalent to groundwater depletion. By essentially ignoring capture, they may substantially overestimate groundwater depletion (Konikow 2013a). Such analyses effectively assume that the storage depletion fraction of pumpage is 1.0 and the capture fraction is 0.0, and ignore the time dependency of the relations as shown in Figure 1.

The purpose of this study is to further characterize the partitioning of sources of well pumpage between capture and storage depletion. The study examines quantitatively the responses to pumpage under capture-constraining conditions, which have not been well elucidated in the literature, and compares responses under such conditions with those under unconstrained conditions. The study also examines storage depletion in real-world aquifer systems on the basis of long-term records and assessments in specific aquifer systems, in part to document the long-term depletion fractions in large-scale systems developed for long periods of time, and in part to assess the reasonableness of assumptions that estimate storage depletion on the basis of pumpage while ignoring capture.

Capture-Constrained Case

Aquifers are often bounded by surface-water features, such as streams or lakes. If such features have a limited availability of surface water, it is hypothesized that the balancing of well withdrawals by increases in recharge and/or decreases in discharge to or from that bounding feature would be constrained, and the general balance between storage depletion and capture (Figure 1) would be disrupted. This might occur, for example, if the stream or lake goes dry. If growth of capture is limited, then the abatement of storage depletion with time is thereby also diminished and storage continues to provide water to the well. In this case, the relative fraction of pumpage balanced by storage depletion is greater than would otherwise occur. If pumpage is so large that capture can never balance the withdrawals, then a new equilibrium cannot be attained, water levels would continue to decline, and the system will continue to be depleted until well yields are necessarily reduced or eliminated (Bredehoeft and Durbin 2009).

To test and evaluate this hypothesis in a quantitative framework, a hypothetical desert-basin aquifer was simulated. The model was based closely on the hypothetical desert-basin aquifer, as developed and documented by Barlow and Leake (2012), which includes a through-flowing river along the eastern edge of the basin. For

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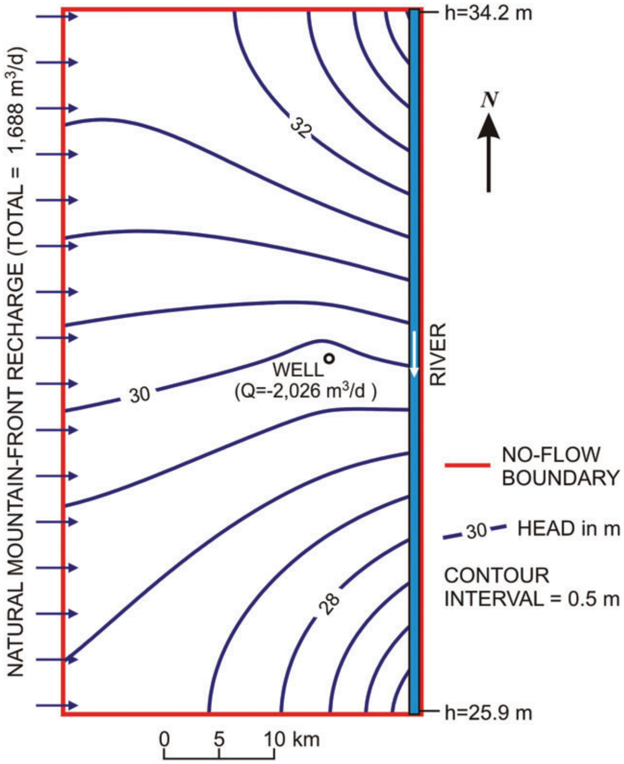

Figure 2. Map view of hypothetical desert-basin aquifer with a through-flowing river along the eastern edge of the basin, showing boundary conditions and calculated heads after 200 years of pumping from a single well. Modified from Barlow and Leake (2012).

simplification, it is assumed that the river lies on the easternmost edge of the alluvial aquifer, and that impermeable bedrock exists beyond the location of the river. Relative to the model developed by Barlow and Leake (2012), all length units were converted to the metric system, and a finer grid spacing was imposed in the *y*-direction. This is a two-dimensional model with no areally diffuse recharge from precipitation—noting that the natural predevelopment recharge rate would not affect the total streamflow depletion (e.g., see Bredehoeft et al. 1982; Barlow and Leake 2012). The only sources of recharge to this hypothetical aquifer are from natural mountain-front recharge at a fixed rate along the western boundary of the model and from head-dependent leakage from a river along the east side of the basin (Figure 2). Mountain-front recharge is a common and important phenomena in arid alluvial basins and typically represented as a boundary condition in groundwater models of a basin (Wilson and Guan 2004). As such, the specified recharge cannot be affected or directly captured by wells pumping from a basin aquifer. However, this type of recharge creates a downgradient discharge from the system that indeed can be captured. Also for simplification, it is assumed that there are no evapotranspirative losses from the water table that could potentially be captured.

Properties and characteristics of the aquifer and the model are listed in Table 1. The aquifer is approximately 32.2 km wide and 64.4 km long. It is discretized into a

Table 1
Characteristics of Hypothetical Aquifer System and Model

Property	Value
Basin dimensions	32.2 × 64.4 km
K_h	15.24 m/d
Aquifer thickness	157.4 m
Specific yield	0.2
Natural mountain-front recharge	1688 m^3/d
River width	10.0 m
River depth	0.001 m
Streambed K_z	3.05 m/d
Streambed thickness	0.305 m
Grid spacing	804.7 m
Number of rows	80
Number of columns	40

single-layer grid of 80 rows and 40 columns yielding square cells with a length of about 805 m on each side. The base case for development includes one pumping well located approximately 8.05 km west of the river and halfway between the northern and southern impermeable boundaries of the rectangular basin. The assumed pumping rate for the well (2026 m^3/d) represents the actual pumpage (and consumptive use), further assuming that none of the pumped water subsequently recharges the aquifer (e.g., see Bredehoeft 2011a). (If some of the pumped water subsequently recharged the aquifer, the net effect on the hydraulic responses in the aquifer would be the same as if the well discharge were reduced by the amount of return flow.) The river flows southward, and the streambed elevation varies linearly downstream from an elevation of 34.2 m to 25.9 m at the two ends of the stream reach. For a base case simulation, it is assumed that the flow rate entering the upstream end of the river is 20,000 m^3/d.

The aquifer system is simulated numerically using MODFLOW-2005 (Harbaugh 2005). The river is represented using the Streamflow Routing (SFR) Package, with a specified depth of water in the stream (Niswonger and Prudic 2005). The model computes a fluid flux between a stream and underlying aquifer at each relevant node of the grid based on head gradients between the stream and aquifer, and routes streamflow downstream after adjusting for the computed aquifer flux at a given location. It further assumes that as long as there is flow in the river, it remains connected to the aquifer (as opposed to disconnected, as described by Brunner et al. 2011). If the stream goes dry at a particular location, then no flow can be routed downstream and the boundary condition is automatically adjusted to preclude a stream-aquifer flux where a stream cell is dry. If the groundwater head at a downstream location is higher than the streambed elevation, then groundwater discharge will restart flow in the stream. Computational methods and assumptions are described in detail by Prudic et al. (2004) and Niswonger and Prudic (2005). A base-case simulation was run starting with an

Table 2
Hydrologic Budget for Base-Case Simulation (Flux Values Are in m^3/d)

	Steady-State Predevelopment	50 Years	200 Years
Mountain-front recharge	1688	1688	1688
Recharge from river infiltration	5785	6585	6859
Groundwater discharge to river	7473	6855	6649
Well pumpage	0	2026	2026
Change in storage	0	608	128

initial steady-state simulation to represent predevelopment conditions. Then a 200-year transient simulation was run using annual time steps and a constant rate of pumping from a single well (location and pumping rate are shown in Figure 2). The resulting hydrologic budgets at three reference times are shown in Table 2. The flow field at the end of the simulation, as depicted by the head distribution (also shown in Figure 2), shows the effect of the pumping well, recharge and flow from the upper reaches of the river into the aquifer, lesser recharge from the western boundary of the model, and groundwater discharge to the lower reaches of the river.

After 47 years, most of the total cumulative pumpage is derived from capture and the amount derived from storage depletion has nearly stabilized by the end of the 200-year simulation period (Figure 3A). While the pumping rate remains constant in time, the rate of capture increases exponentially and the rate of storage depletion decreases exponentially (Figure 3B). The components (sources) of capture include increases in recharge (arising from increased stream leakage into the aquifer induced by declining groundwater levels) and decreases in groundwater discharge (base flow to the river) relative to predevelopment conditions. At early times, the latter is somewhat larger, but the two sources of capture stabilize in a few decades to where increases in recharge contribute about 56% of capture and decreases in discharge account for 44% of capture. For the conditions of this simulation, both factors combined (i.e., total capture) result in (and are equivalent to) streamflow depletion.

In terms of sources of water derived from the well, capture increases exponentially while storage depletion decreases exponentially. Although the generic relations shown in Figure 1 offer no specific time scale, in this test case the relative contributions still had not stabilized after 200 years (Figure 4). Storage depletion and capture can be compared to pumping on the basis of either cumulative volumes or instantaneous rates. Because the storage depletion rates decrease with time, the depletion curve based on cumulative volumes, which integrate system responses over time, will reflect higher fractional values at any particular time during the transient evolution of the response than that based on instantaneous rates. Conversely, capture (or streamflow depletion) fractions based on cumulative volumes will be smaller at any particular time than those based on instantaneous rates (also see Barlow and Leake 2012, 16 to 17). On the basis of cumulative volumes, in the base case simulation the results were depletion dominated for the first 47 years and then were capture dominated after that (Figure 4). When the fractions are computed on the basis of flow rates (for annual time steps in this case), the cross-over occurs earlier—after only about 17 years. A large difference is also present for the two calculations at any given time. For example, after 100 years, the depletion fraction based on cumulative fluxes was about 36% whereas the depletion fraction based on flow rates was about 18%.

The timing and relative magnitude of the response of the stream-aquifer system depends on the hydraulic properties of the aquifer, its boundary conditions, and the distance of the well from the recharge and discharge boundaries (Theis 1940). The sensitivity of the response in the hypothetical desert-basin aquifer to well location was evaluated by varying the well position in an east-west direction between the two lateral boundaries. The results (Figure 5) show that the time it takes for the system to reach a new equilibrium condition increases with distance of the well to the river (it was assumed that steady-state conditions are attained when 99.9% of the ultimate storage depletion has occurred). The storage depletion fraction was even more sensitive to well location and varied from 0.01 to 0.18 over the range of tested distances. The total storage depletion volume at steady state also increased by a factor of almost 20 (from 4.5×10^6 to 8.8×10^7 m^3) as the distance to the river increased from 0.805 km to 30.6 km.

In the base case, the river never goes dry during the 200-year simulation, so the potential for increasing recharge in response to drawdown in the aquifer is never limited. To evaluate the affects of constraints on increases in recharge, the specified inflow to the upstream end of the river was reduced to 10,000 and 6400 m^3/d. The downstream flow profiles for the three different specified stream inflows show that the river goes dry only for the lowest inflow case (Figure 6). The changes in flow over space and time are identical when $Q_{in} = 20{,}000$ or 10,000 m^3/d. However, when Q_{in} is reduced to 6400 m^3/d, the river starts to go dry during the 22nd year of the simulation when the increasing stream leakage into the aquifer equals the total flow in the river. As time progresses, the dry reach advances further upstream, as indicated by the difference between the 50-year and 200-year curves for the case of $Q_{in} = 6400$ m^3/d. The differences also show that the change in flow during the first 50 years was much greater than the change during the next 150 years. Streamflow increases in the downstream part of the river because of groundwater discharge into the river. The difference between the predevelopment profiles and the curves at a given time after pumping began represents capture (and streamflow depletion).

The calculated flow rates for the streamflow-limited case (Figure 7) can be compared to the same elements in the base case (Figure 3B). The results show that

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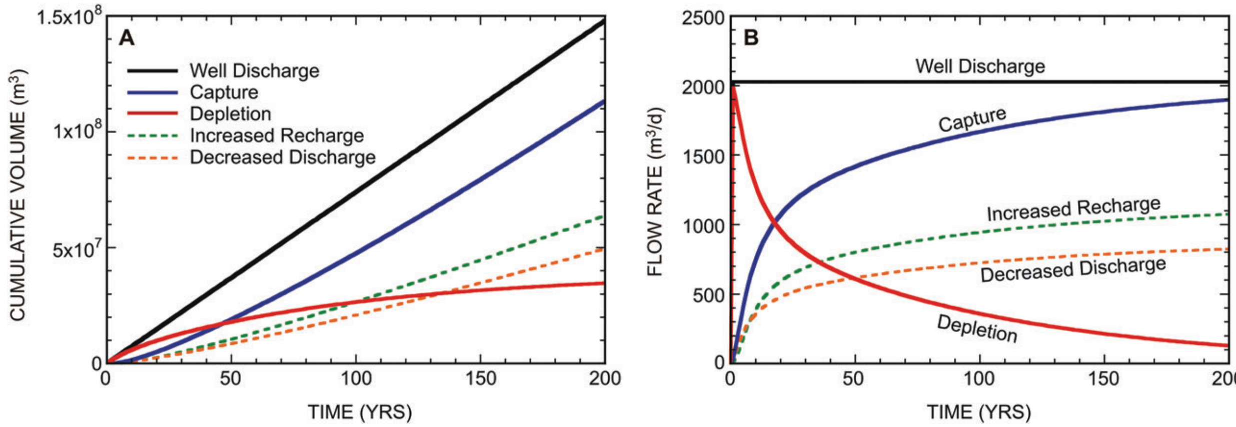

Figure 3. Simulated hydrologic budget elements for base case simulations of hypothetical aquifer expressed as cumulative volumes (A) and instantaneous flow rates (B).

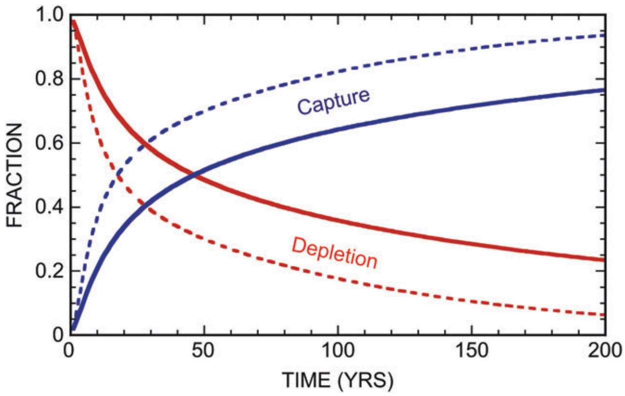

Figure 4. Simulated storage depletion (red curves) and capture (blue) fractions relative to pumping on both a cumulative (solid lines) and annual rate (dashed lines) basis.

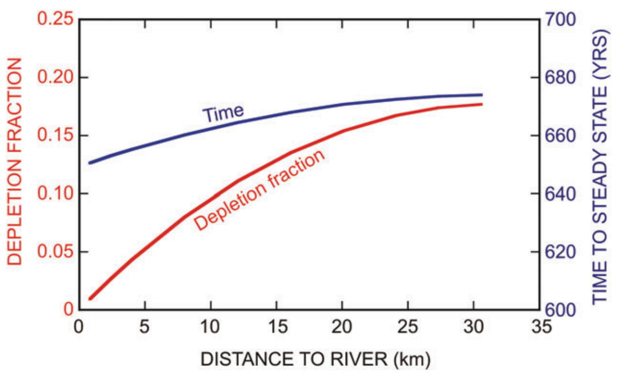

Figure 5. Sensitivity of time to achieve equilibrium, and of storage depletion fraction at equilibrium, to distance of pumping well to river. Depletion fraction is based on cumulative volumes at the time that a new steady-state condition is achieved.

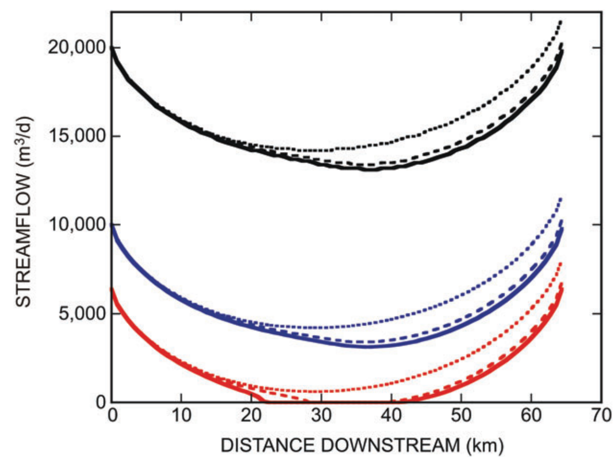

Figure 6. Change in flow along the length of the river when the specified upstream inflow is 20,000 m^3/d (black curves), 10,000 m^3/d (blue curves), and 6400 m^3/d (red curves) for $t = 50$ years (dashed lines) and 200 years (solid lines). Streamflow profiles for steady-state predevelopment conditions are shown for comparison (dotted lines).

when there is insufficient water in the river to meet the drawdown-induced demand, the amount of pumpage derived from capture decreases and the amount derived from storage depletion increases. The effect is most noticeable on the increase in rate of recharge derived from stream leakage (induced infiltration), which reaches its maximum in year 22 and becomes constant after that because 100% of the upstream inflow to the river has been captured. Simultaneously, there is an increase (relative to the base case) in the amount of groundwater discharge to the downstream reaches of the river that is captured, though not enough to offset the constrained increase in recharge. Thus, after 22 years, the total capture is reduced and storage depletion is increased relative to the base case. The decreased capture relative to the base case is also reflected in a plot of storage depletion and capture fractions (Figure 8). After 200 years, the annual capture fraction is 0.84 for the streamflow-limited case, whereas it is 0.94 in the base case.

In the previous analysis, the total well pumpage is less than the total available capture and the rate of capture is still increasing after 200 years (Figure 7). But this should change if the total well pumpage were greater than the available capture. This was tested by adding nine

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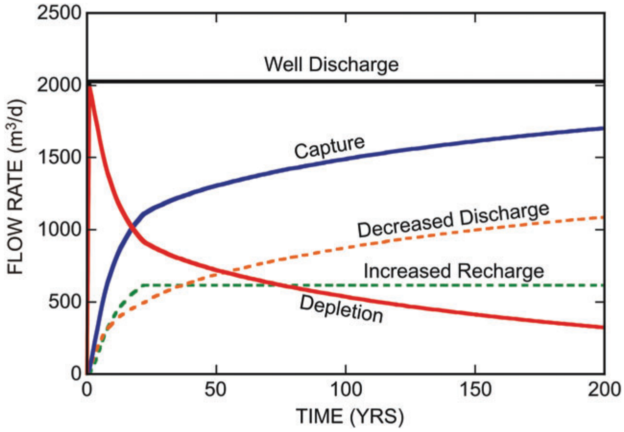

Figure 7. Simulated hydrologic budgets for streamflow-limited case ($Q_{in} = 6400\,m^3/d$) showing calculated flow rates for selected boundary conditions.

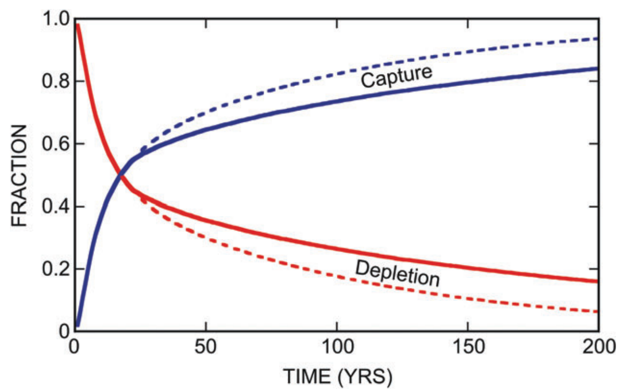

Figure 8. Simulated annual storage depletion (red curves) and capture (blue) fractions relative to pumping for the streamflow-limited case (solid lines) and the high streamflow base case (dashed lines).

more wells pumping at the same rate to the downstream (southern) half of the aquifer, so that the total well withdrawals are ten time greater than in the previous case ($Q_{tot} = -20{,}260\,m^3/d$). With the increased pumpage, the stream first goes dry in the 6th year of the simulation, and captures all of the groundwater discharge in the 104th year (when the river outflow from the basin becomes zero). This is reflected in the changing downstream flow profiles at various times (Figure 9), which also illustrates the progressively longer length of the dry reach with time.

In terms of the hydrologic budget for the system under the higher pumping scenario, the rates of storage depletion and capture (streamflow depletion) become steady after 104 years (Figure 10A). Similarly, the fractions of annual pumpage derived from storage depletion (0.60) and capture (0.40) do not change after this time either (Figure 10B). This means that the cones of depression around the pumping wells will not stabilize and will continue to expand as long as the pumping continues and the boundary conditions remain the same. This is a classic groundwater mining situation, though slow recovery is possible if well pumpage is eliminated.

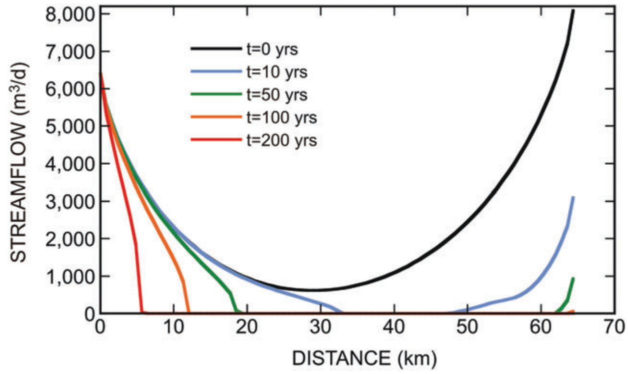

Figure 9. Change in flow along the length of the river for the streamflow-limited, high-pumpage case for selected times. Specified upstream inflow is $6400\,m^3/d$. Total well pumpage is $20{,}260\,m^3/d$.

Analyses of real systems as well as of hypothetical desert-basin aquifers clearly demonstrate that streamflow depletion (capture) can continue long after pumping has ceased (Bredehoeft 2011b; Barlow and Leake 2012). They note that the rate of recovery depends on a number of factors, including hydraulic properties and boundary conditions. In the analysis by Barlow and Leake (2012) of the hypothetical desert-basin aquifer, following 50 years of pumping, it required an additional 100 years after pumping ceased to recover most (but not all) of the storage depletion. Under more severe streamflow-limited scenarios, such as reflected in the budgets of Figures 7 and 10, the recovery would take much longer.

Responses in Real Systems

Theis (1940) states that the source of water that balances well discharge is "always to some extent and in many cases largely from storage in the aquifer. Some groundwater is always mined." But he also points out that "in most artesian (i.e., confined) aquifers—excluding very extensive ones, such as the Dakota sandstone—little of the water is taken from storage." Were Theis' assessments basically correct? In real aquifer systems that have been developed (pumped) for decades, how much of the pumpage has been derived from storage? The previous analyses show that it can take many decades, if not centuries, for a stream-aquifer system—especially an areally extensive system in an arid climate—to reach a new equilibrium in response to long-term pumping stresses (also see Bredehoeft and Durbin 2009; Barlow and Leake 2012). During the transient response phase, the fraction of pumpage derived from storage depletion would tend to decrease with time, and the complementary capture fraction would correspondingly increase. The range of experiences in real aquifer systems that have been developed (pumped) for decades is examined, with analyses limited to aquifers, time periods, and areas for which adequate data are available for both estimates of storage depletion volume and estimates of total well withdrawals. Of course, at the scale of an aquifer system,

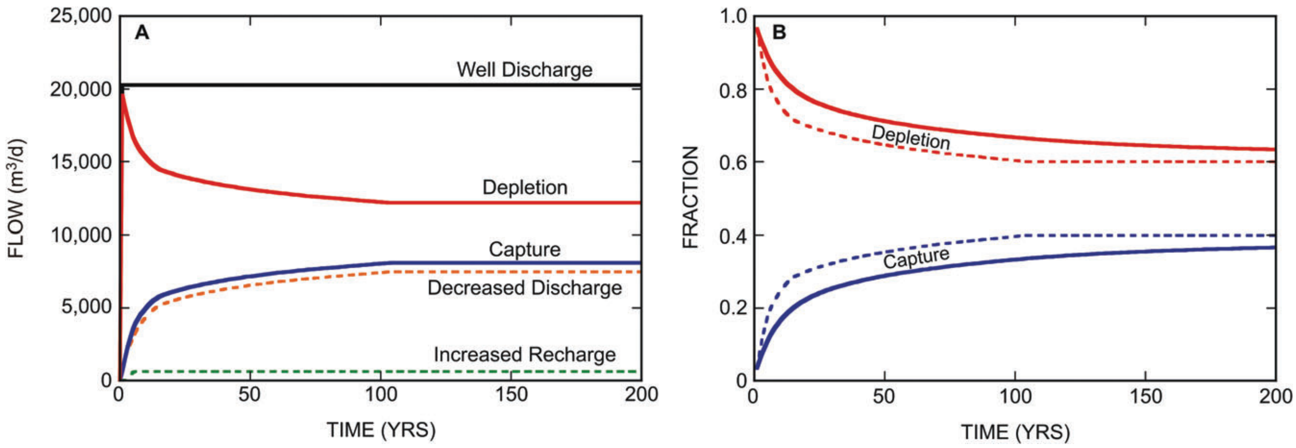

Figure 10. Simulated hydrologic budgets for streamflow-limited, high-pumpage case (A) and storage depletion (red curves) and capture (blue) fractions relative to pumping (solid lines show cumulative data and dashed lines show annual rates) (B).

the observed cumulative depletion is a complex response function of the interactions of multiple transient stresses, both natural and engineered, consistent with the principles of superposition.

Leakage from low-permeability confining units into pumped aquifers is a well-known and important process affecting the propagation of responses through an aquifer system. Typically, head declines will propagate slowly through confining layers, and the leakage will be derived largely from storage depletion in the confining unit until a new steady-state head distribution is eventually achieved (see Konikow and Neuzil 2007). Leakage also acts to slow the lateral propagation of head declines through an aquifer, thereby delaying the interaction with aquifer boundaries. Thus, streamflow depletion caused by pumping wells will take longer to occur and longer to reverse than in a nonleaky system.

Various estimates of long-term storage depletion in specific aquifers are available (e.g., see Konikow 2011, 2013b). There are 31 aquifers or areas in the United States and two outside the United States for which adequate data are available to estimate depletion and capture fractions (see Table S1, which shows that estimates for almost all areas represent cumulative volumes over periods of several decades). In many cases, the estimates of volumetric depletion include depletion in overlying and/or underlying confining units (methods and specific analyses are described by Konikow 2013b). Also, an estimate can be made for the United States as a whole based on cumulative withdrawals and depletion volumes over more than five decades. These aquifers and areas include a broad range of hydrogeologic settings and climates, so should be representative to some extent of global conditions. The areas for which data are available are mostly areas that have experienced relatively large-scale and long-term development of groundwater supplies. Because the estimates are generally based on long-term cumulative volumes, the depletion fractions for the most recent time would likely be smaller than the value computed on the basis of cumulative volumes and capture fractions during the most recent time increments would likely be larger (see Figure 4). Note that these fractional values are not static. Rather, they would be changing slowly with time, although after several decades, the cumulative fractions are relatively stable and tend to change only very slowly.

In the United States, the distribution of depletion fractions shows a wide variance (Figure 11). The highest depletion fraction (0.97) is in the Death Valley regional flow system, which has an arid climate and few surface water resources. Outside the United States, the Nubian aquifer in North Africa has essentially zero recharge, no potential for increasing recharge, and an increasing magnitude of development. Even without the effects of development the system is undergoing a slow transient evolution of heads from a wetter period with recharge thousands to millions of years ago (Voss and Soliman 2013). Residual discharge is balanced by storage decreases. Yet a model study (CEDARE 2001) calibrated to 38 years of record (1960 to 1998) indicates that in 1998, the end of the study period, the storage depletion fraction was only 0.84 and the capture fraction was therefore 0.16, with the capture representing reductions in natural discharge (e.g., by a reduction in the discharge of springs at oases).

Theis' insight about confined aquifers was generally correct. For example, for 1901 through 1980 only about 30% of the pumpage in the areally extensive Cambrian-Ordovician aquifer in the Midwestern United States was derived from storage depletion. Theis' exception for the Dakota aquifer was also reliable, as about 78% of the withdrawals from the Dakota in South Dakota during 1881 through 1980 was balanced by a reduction in storage. However, as concluded by Konikow and Neuzil (2007), most of the storage depletion originated in the adjacent thick confining units—an aspect not noted by Theis. At the other end of the spectrum, intense groundwater development has occurred in the Floridan and adjacent aquifers in Florida and parts of Georgia and South Carolina. These areas have relatively high precipitation.

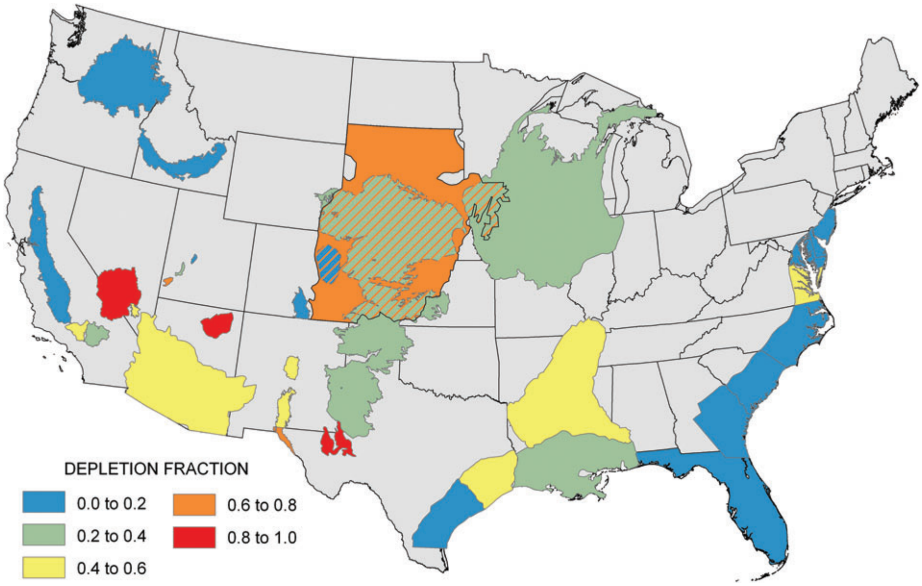

Figure 11. Estimated long-term cumulative storage depletion fraction in 31 areas and aquifers within the United States. Hatched patterns reflect areas where one aquifer overlies another. Also see data in Table S1.

The depletion fraction for this combined area is only about 0.01 for 1950 through 2005, so that about 99% of the pumping is derived from capture. For the United States as a whole for 1950 through 2005, the total net groundwater storage depletion volume is about 812 km^3 (Konikow 2013b) and the cumulative withdrawals are approximately 5340 km^3 (Kenny et al. 2009). Thus, the long-term depletion fraction is about 0.15 and the capture fraction is about 0.85.

Considering all 31 areas in the United States, the United States as a whole, and two aquifer systems outside the United States (Nubian aquifer [CEDARE 2001] and North China Plain [Cao et al. 2013]), an analysis of the frequency distributions (Figure 12) indicate that most systems have evolved to low cumulative depletion fractions (mean = 0.39) and high cumulative capture fractions (mean = 0.61). However, there can also be a wide variation within any particular areally extensive aquifer system. For example, the largest volume of storage depletion in the United States occurs in the High Plains Aquifer system. This large system underlies parts of eight states, and state by state data are also available (e.g., see McGuire et al., 2003; McGuire 2007). For cumulative volumes during 1950 through 2000, the depletion fraction for the entire High Plains Aquifer was about 0.27, but it ranged from 0.00 in the Nebraska portion (where there were slight water-table rises during this time period) to 0.42 in the Texas portion.

The storage depletion fractions also show some correlation with climate (Figure 13). The 33 data points in Figure 13 include separate values for the Texas and Nebraska parts of the High Plains Aquifer, but exclude averaged values for the United States as a whole. In general, where precipitation is higher and water tables are higher, one would expect a greater potential for pumping-induced drawdown to cause increases in recharge and/or

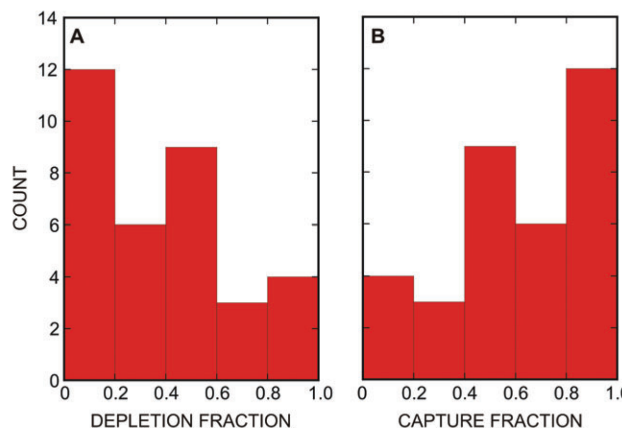

Figure 12. Histograms for (A) depletion fractions and (B) capture fractions in 34 areas and aquifers (from data in Table S1).

decreases in discharge. Also, in more humid climates, drainage densities tend to be higher, so that the effective distances from wells to surface water boundaries are generally shorter, especially in shallow aquifers; consequently, response times for inducing increased recharge or decreased discharge are shorter, which would tend to reduce relative storage depletion. The correlation coefficient (R) for this relation is 0.40, indicating a mild relation rather than a strong one, which can also be seen by the large spread of values about the regression line. It would be erroneous to assume that the cumulative depletion fraction can be accurately predicted on the basis of climate alone. Other factors that influence the cumulative depletion fractions include variability in the distances to influential boundary conditions, in hydraulic properties, and in time histories of well development and total aquifer withdrawals.

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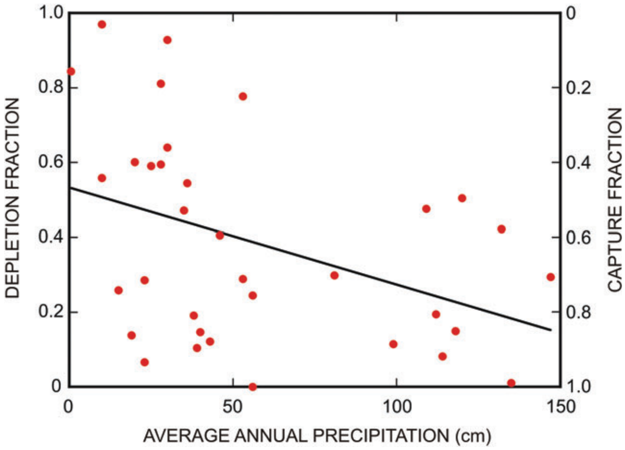

Figure 13. Relation between average annual precipitation and long-term cumulative depletion and capture fractions for 33 aquifer areas and subareas, showing a best-fit linear regression line (from data in Table S1).

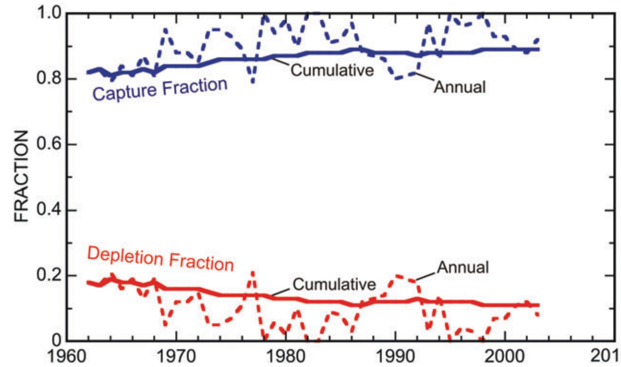

Figure 14. Results of water budget calculations of the Central Valley, California, calibrated groundwater-flow model (Faunt et al. 2009b), showing storage depletion (red) and capture (blue) fractions (solid lines for cumulative fractions; dashed lines for annual rates).

Well-calibrated and well-constructed simulation models of long-term responses in aquifer systems offer a means to analyze the sources of water derived from wells and how they vary with time. For this type of analysis, a well-constructed model would be free of artificial boundaries that would affect calculations of groundwater storage depletion and capture for a groundwater system. This will be illustrated briefly using two representative examples of such well documented model analyses.

The Central Valley of California is a major agricultural area in a large valley with an area of about 52,000 km^2 (Williamson et al. 1989; Bertoldi et al. 1991). The Central Valley has an arid to semiarid Mediterranean climate, where the average annual precipitation ranges from 13 to 66 cm (Bertoldi et al. 1991). Streamflow is an important factor in the water supply of the valley. Groundwater development began around 1880. By 1913, total well pumpage was about 0.44 km^3 annually (Bertoldi et al. 1991). During the 1940s and 1950s, the pumpage increased sharply, and by the 1960s and 1970s averaged about 14.2 km^3/yr. By the 1980s there were approximately 100,000 high-capacity wells in the Central Valley for either irrigation or municipal supply. During 1962 through 2003, withdrawals from irrigation wells averaged about 10.6 km^3/yr (Faunt et al. 2009a).

A transient groundwater-flow model of the Central Valley was developed for 1961 through 2003 (Faunt et al. 2009b). The model indicates that the decrease in groundwater storage from 1961 through 2003 was about 71.2 km^3. However, the total decrease in groundwater storage from predevelopment conditions until 1961 was about 58 km^3 (Williamson et al. 1989, 95), and this is not accounted for in the 1961 through 2003 model. As expected, the cumulative fractions are smoother than the annual fractions (Figure 14), and the year-to-year variability in annual fractions is largely controlled by variations in annual pumpage and precipitation. The depletion and capture fractions (both cumulative and rate based) for the first year of the simulation period are 0.18 and 0.82, respectively. But over the 42-year simulation period, the fractional rates did not change greatly, as reflected by the relatively small change in the cumulative storage depletion and capture fractions to 0.11 and 0.89, respectively, indicating that such long-term cumulative fractions (such as presented in Figure 4) are relatively stable and representative of conditions in the aquifer. Compared with the generic fractional curves (Figure 1), it is evident that this model of the Central Valley of California, which begins about 80 years after the start of pumpage, cannot and does not represent the expected early-time system responses of high depletion fractions and low capture fractions, so that the cumulative depletion fraction would be too small (and cumulative capture fractions too high) in the early years of these simulation results.

Antelope Valley, California, is a small (2400 km^2) topographically closed basin with an arid climate (average annual precipitation is less than 25 cm). The basin contains a thick (more than 1500 m in places) sequence of unconsolidated alluvial and lacustrine sediments. Surface water is limited, and the area includes several springs, playas, and intermittent streams that drain into the playas (Leighton and Phillips 2003). Delivery of some imported water began in 1986. Leighton and Phillips (2003) note that recharge to the groundwater system is primarily from the infiltration of precipitation runoff near the valley margins, and discharge from the aquifer system was primarily from evapotranspiration. Development of the groundwater system began around 1915 and increased rapidly into the 1950s. Pumpage peaked at more than 0.37 km^3/yr in the 1950s and 1960s, but by the mid-1980s had declined to about 0.12 km^3/yr (Galloway et al. 2003). Groundwater pumping has caused large water-level declines in the basin, resulting in a major decrease in evapotranspirative discharge (Leighton and Phillips 2003).

A 3D transient MODFLOW model was developed and calibrated to simulate groundwater-flow and aquifer-system compaction in the area (Leighton and Phillips 2003). The model was first calibrated to represent predevelopment conditions prior to 1915. Then the transient model was developed using 81 1-year stress periods to simulate the period of 1915 through 1995 inclusive. More detailed descriptions of model parameters, boundary conditions, and the calibration process are presented by Leighton and Phillips (2003). The results of the transient simulation indicate that more than 10.5 km^3 of groundwater was removed from storage during 1915 through 1995, with most of the storage change occurring between about 1945 and 1975 (Leighton and Phillips 2003). The model-computed water budgets indicate that most of the pumpage was derived from storage depletion during the first few decades of development, but that the capture fraction generally increased with time—becoming dominant since the early 1970s (Figure 15). As would be expected in this type of basin, there appears to be a strong direct correlation between pumpage and storage depletion (Figure 15B). Similar to the Central Valley (Figure 14), in the Antelope Valley the fractions based on annual flow rates show greater variability than the cumulative fractions (Figure 15A), and the variability in annual fractions is largely controlled by variations in annual pumpage. During 1988 through 1990, the annual pumpage was the smallest since 1925, and during these 3 years there were small increases in net storage. Capture was comprised largely of increased recharge from irrigation return flows, but during the first 5 decades decreased evapotranspiration also contributed to capture. After 1985, increased recharge from imported water also provided a substantial offset of the effects of pumping.

The evolution of storage depletion and capture fractions in Antelope Valley indicate the transient nature of these factors, and the data (Figure 15A) show that the system is still continuing to evolve. It has not achieved a new permanent equilibrium state and storage depletion—though temporarily halted during 1988 through 1990—continues to increase even though annual pumpage has decreased substantially since its previous peak rates. The change in the fractions during the historical period of record covers the full range of values. The difference between the cumulative and rate-based fractions is much greater than seen in the results for the Central Valley. For example, at the end of this study period (1995) the annual storage depletion fraction was 0.18 while the cumulative depletion fraction was 0.59. Achieving a sustainable groundwater development practice would require that the annual depletion fraction approach and be maintained at values at or close to zero and that the environmental consequences of capture be acceptable.

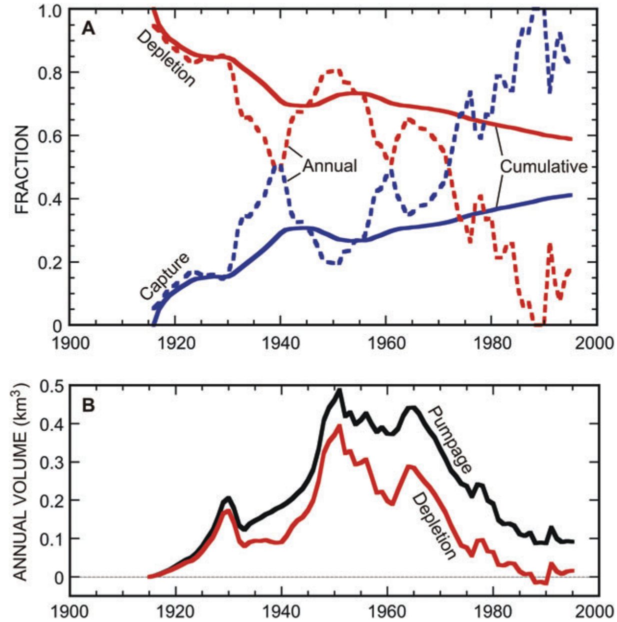

Figure 15. Results of water budget calculations of the Antelope Valley, California, calibrated groundwater-flow model (Leighton and Phillips 2003), showing (A) computed storage depletion fractions (red) and capture fractions (blue), with solid lines representing fractions based on cumulative data and dashed lines representing annual values, and (B) estimated annual pumpage (black) and calculated annual storage depletion volume (red).

Conclusions

Nearly 75 years have passed since Theis (1940, 1941) published his classic papers that clearly elucidated the sources of water derived from wells and the effect of pumping a well on flow in a nearby stream. His principles and guidance have stood the test of time, and are not only still relevant today, but should be required reading for every groundwater analyst. His overriding principle is the simple message that all water discharged by a well must be balanced by a loss of water somewhere—either from storage or by capture. This study expands a little on Theis' work by examining two aspects that he did not focus on. First, we analyze how the balance is affected if capture is constrained by a limited availability of water. Theis (1941) had assumed "that the stream maintains a flow past the pumped area." Second, we analyze a number of real systems in which sufficient data are available to assess the partitioning of the balancing components into storage depletion and capture fractions after a long history of pumpage.

Groundwater storage depletion and capture can be measured in terms of nondimensional fractions relative to pumpage. These measures can be computed on the basis of either cumulative volumes or flow rates. The former will yield more moderated values that reflect long-term averaged responses (i.e., rates integrated over time), but may not accurately indicate system status at any particular time years after development started. These measures will tend to change exponentially with time, and the complementary fractional values of storage depletion and capture, based on flow rates, will effectively reach 0.0 and 1.0, respectively, if sufficient water for capture is available at aquifer boundaries. When this occurs, the aquifer system has attained a new equilibrium condition and continued development should be sustainable. However, if prior to equilibrium aquifer bounds are reached that preclude any further increases in recharge and decreases in discharge, then a new equilibrium cannot be attained and storage depletion will continue to occur.

The potential for well withdrawals to be balanced by capture would be constrained if there is insufficient water available at aquifer boundaries to meet the increased demands imposed by drawdown-induced steepening of hydraulic gradients. Evidence of constraining conditions includes streams or springs going dry following an extended period of pumpage within the aquifer. When capture is constrained, the relative amount of pumpage balanced by (or derived from) capture decreases and the amount derived from storage depletion increases. In severely constrained cases, all sources of capture can reach their limits. Then, discounting natural fluctuations in recharge from precipitation, with continued steady pumpage the fractions of the pumping rate derived from capture and storage depletion will stabilize with time. This means that groundwater levels will continue to decline—a classic groundwater mining situation. This can then continue until drawdowns themselves start to limit the pumpage because of increased lifts and higher costs of pumping or because reduced saturated thicknesses decrease well yields. In this sense, groundwater storage depletion itself should eventually be self-limiting and unsustainable.

In an illustrative test problem representing pumping in a hypothetical desert-basin aquifer, the only source of capture was from the stream. In this case, rates of capture (streamflow depletion) exceeded storage depletion after 17 years. As long as the stream did not go dry at any point, the largest contributor to capture was increased recharge from induced infiltration. But if the stream did go dry and capture was thereby constrained, then the amount of pumpage derived from storage depletion increased relative to the nonconstrained condition, and the amount derived from capture correspondingly decreased. Also, under capture-constraining conditions, decreases in groundwater discharge to the stream became the larger contributor to total capture after 36 years because the central reach of the river went dry and induced infiltration could no longer increase.

There are 31 specific areas or aquifers within the United States and two outside the United States for which adequate data are available for both total withdrawals and cumulative storage depletion to allow estimates to be made of long-term storage depletion and capture fractions. The mean depletion fraction is 0.39 and the mean capture fraction is 0.61. For the United States as a whole during 1950 through 2005, about 15% of total pumpage was derived from a reduction of storage of groundwater—a depletion fraction of 0.15. But depletion fractions vary widely within the United States and even within any given large aquifer system. For example, the fraction of long-term (1950 to 2000) pumpage derived from storage depletion in the High Plains aquifer is about 0.27, but ranges from 0.0 in Nebraska (where there was a slight water-table rise) to 0.42 in Texas. In general, storage depletion fractions tend to be higher in arid regions, but the relation is not strong and depletion fractions cannot be accurately predicted on the basis of climate alone. These fractions are time dependent, but analyses from the Central Valley and Antelope Valley, both in California, support the notion that cumulative fractions tend to be relatively stable at late times (typically a few decades after major development begins).

Well-calibrated simulation models offer a means to analyze the sources of water derived from wells and how the fractions vary with time—a modern tool not available to Theis. To reliably simulate the history of storage depletion and capture in a groundwater system, groundwater-flow models must start with initial conditions representative of predevelopment times and conditions. Such models also provide water managers with a tool to predict future changes in storage and streamflow depletion in response to possible changes (or no changes) in water management policies.

Groundwater storage depletion and capture problems must be confronted on local and regional scales, where water managers faced with unsustainable withdrawals will necessarily have to take actions to reduce demand and/or increase supply through managed aquifer recharge, desalination, and/or developing alternative sources. Otherwise, storage depletion of the aquifer system will itself ultimately limit withdrawals—in ways that are economically and environmentally less than optimal.

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Supporting Information

Additional Supporting Information may be found in the online version of this article:

Table S1. Supporting Data and References for Estimates of Groundwater Depletion

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