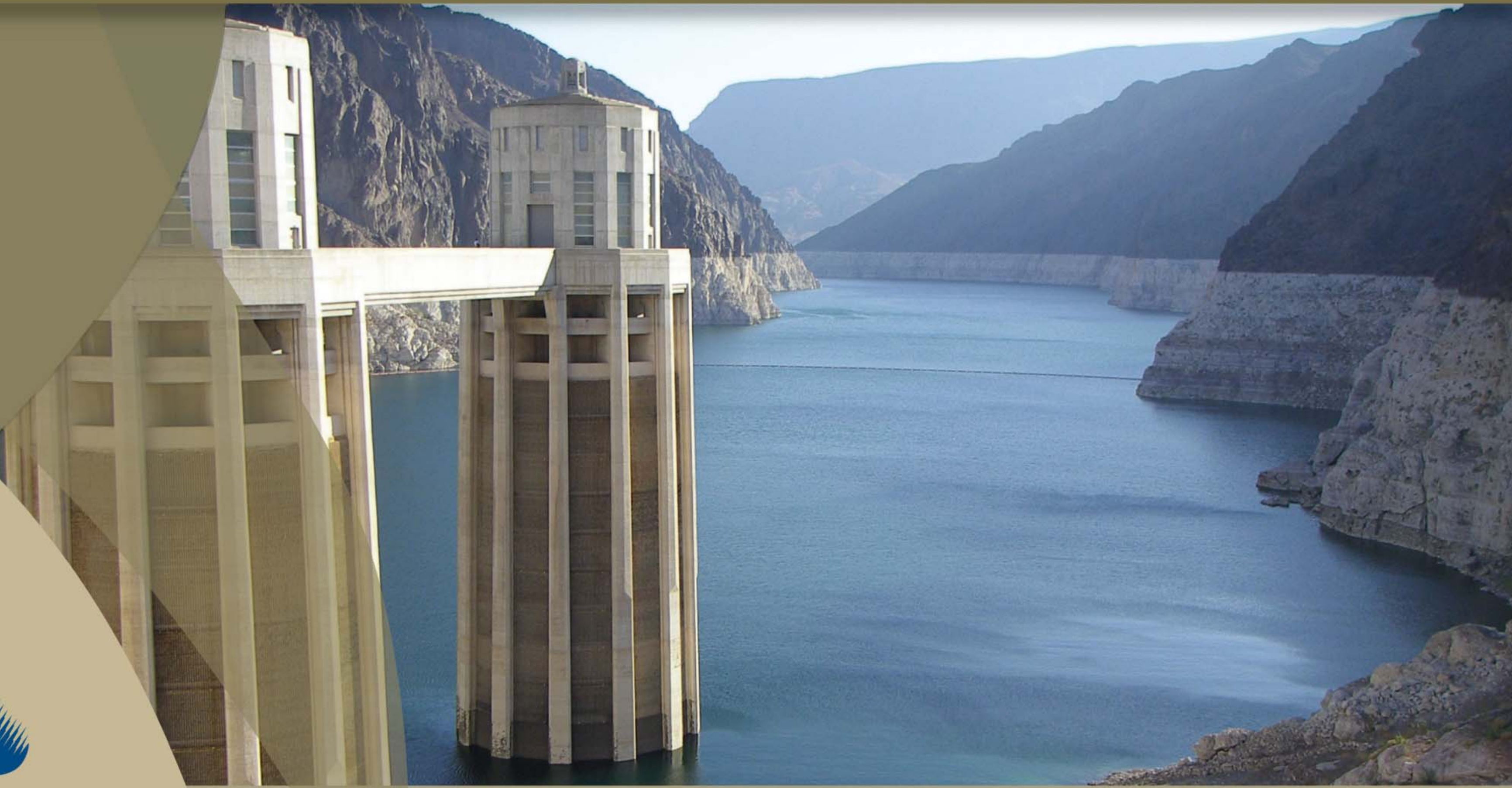
Spring, Cave, Dry Lake and Delamar Valleys



SOUTHERN NEVADA
WATER AUTHORITY

Presentation for
D'Agnese and Luptowitz Testimony

Lisa M. Luptowitz

Southern Nevada Water Authority
100 City Parkway, Suite 700
Las Vegas, Nevada 89106

e-mail: lisa.luptowitz@snwa.com
phone: (702) 862-3789

SUMMARY

Over 20 years experience in environmental compliance and permitting. Project manager and technical lead for preparation of environmental compliance documents and permitting for federal, state, local and private projects. Conduct environmental analyses and obtain necessary permits and authorizations to ensure compliance with federal and state environmental regulations, including National Environmental Policy Act, Endangered Species Act, National Historic Preservation Act, and Clean Water Act. Supervise a diverse team of resource specialists. Establish cooperative relationships with federal and state regulatory personnel and other project stakeholders, and represent the agency or client in public meetings and presentations.

PROFESSIONAL EXPERIENCE

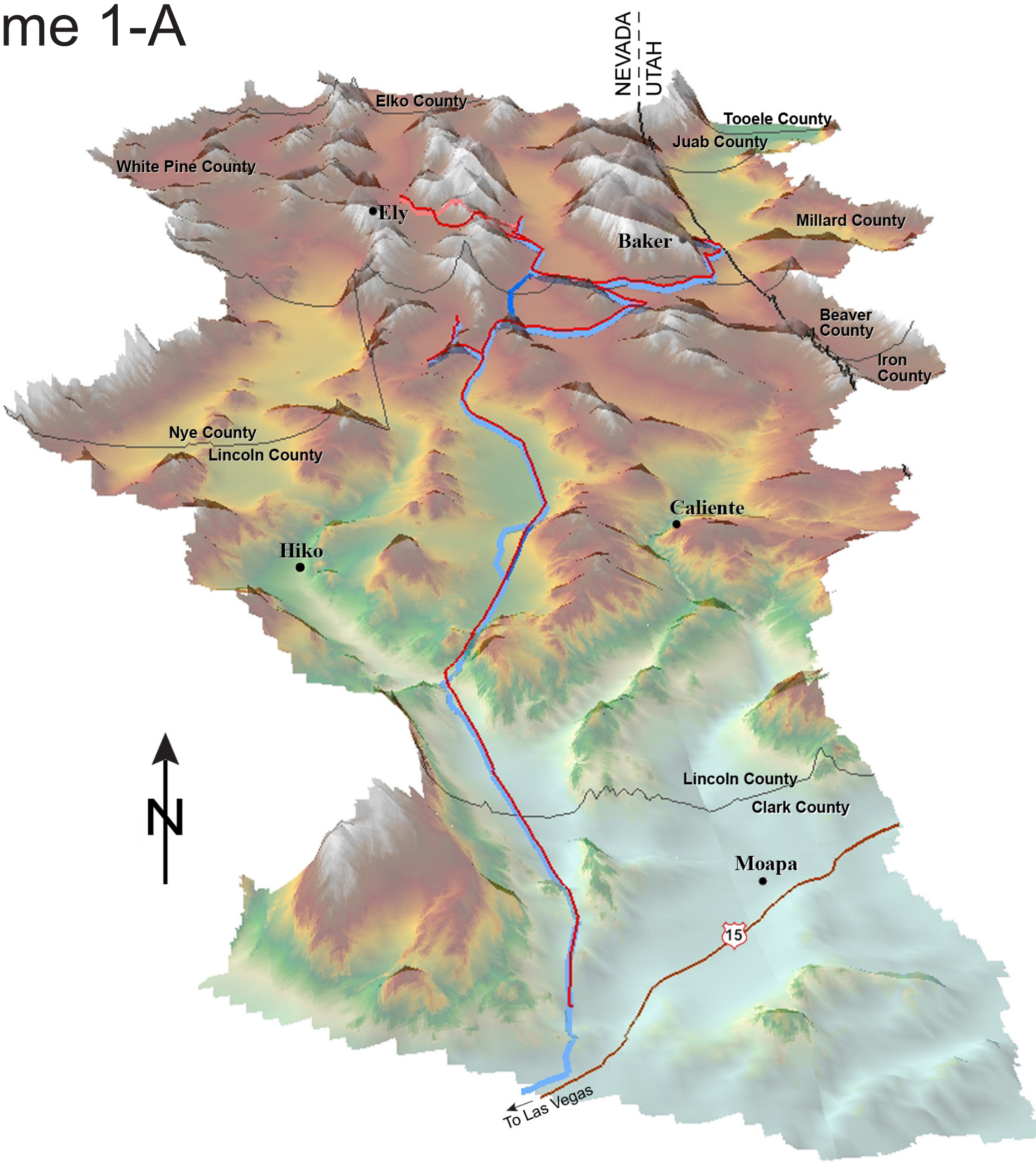
1999 – Present **Southern Nevada Water Authority**
100 City Parkway, Suite 700
Las Vegas, Nevada 89106

<u>Division Manager, Environmental Resources</u>
Supervise 15 professional and administrative staff in two working group teams for environmental planning and biology. Prepare short and long term staffing plans, prioritize workload and job assignments, complete performance evaluations, perform staff counseling, and conduct other administrative functions. Prepare annual budgets and track expenditures for division budget.

Manage and provide technical input on the environmental compliance and permitting for Clark, Lincoln, and White Pine Counties Groundwater Development project. Serve as SNWA lead in coordinating with the Bureau of Land Management and 16 cooperating agencies on development of an Environmental Impact Statement for National Environmental Policy Act compliance. Prepare project applicant materials, including development of project and alternative descriptions and applicant environmental measures. Assist in the development of and participate in hydrological and biological technical review committees for development of reports and technical information. Review and comment on interim drafts to ensure accuracy and regulatory compliance. Conduct public outreach and provide public presentations. Negotiate schedules and contractor budgets. Supervise contractors performing biological and cultural field studies and investigations. Assist in the development of a Programmatic Agreement for National Historic Preservation Act compliance. Participate in informal consultation with Fish and Wildlife Service for Endangered Species Act Section 7 compliance, including preparation of draft Biological Assessment and other consultation materials.

Clark, Lincoln, and White Pine Counties Groundwater Development Project Draft Environmental Impact Statement

Volume 1-A



Bureau of Land Management

June 2011

DES 11-18

Cooperating Agencies:

Army Corps of Engineers
Bureau of Indian Affairs
Bureau of Reclamation
Central Nevada Regional
Water Authority
Clark County, NV
Juab County, UT
Lincoln County, NV
Millard County, UT
National Park Service
Nellis Air Force Base
Nevada Department of Wildlife
State of Utah
Tooele County, UT
U.S. Fish and Wildlife Service
U.S. Forest Service
White Pine County

BLM

Nevada Groundwater Projects Office

SNWA Exhibit 408

Frank A. D'Agnese

500 N. Tucson Blvd., Suite 150 **Business: (520) 829-7127**
Tucson, AZ 85716 **Mobile: (520) 440-2657**

EDUCATION

Colorado School of Mines, Golden, CO

1991-1994 **Doctor of Philosophy**, Geological Engineering
Emphasis: Hydrogeology, Ground-water Modeling, and GIS
Minor: Environmental Sciences / Ecology
Thesis: *Using Geoscientific Information Systems for Three-Dimensional Modeling of Regional Ground-water Flow Systems, Death Valley Region, Nevada and California*

1989-1991 **Master of Engineering**, Geological Engineering
Emphasis: Engineering Geology and Applied Geomorphology
Thesis: *A Regional Aggregate Evaluation of Surficial Materials using a GIS*

1985-1989 **Bachelor of Science**, Geological Engineering
Emphasis: Geology and Engineering Science
Minor: Environmental Science

CONTINUING EDUCATION

U.S. Geological Survey, Water Resources Division, Lakewood, CO

Fall 1995 **Advanced Modeling of Ground-water Flow**
Coordinator: Stan Leake (U.S. Geological Survey)
- Survey of expanded capabilities of MODFLOW; Particle tracking using MODPATH; Stream-flow routing; Rewetting of model cells; Transient leakage from confining units; Low-permeability barriers to horizontal flow; Issues of parameter estimation techniques.

Winter 1995 **Parameter Estimation for the Modular Ground-water Flow Model**
Instructor: Mary Hill (U.S. Geological Survey)
- Capabilities of MODFLOWP; Parameter estimation using non-linear regression; Least-squares estimation; Error analysis for regression solution; Analysis of residuals; Predictive uncertainty; Field examples.

A Summary of the Development of the Central Carbonate-Rock Province Groundwater Flow Model

PRESENTATION TO THE OFFICE OF THE NEVADA STATE ENGINEER

Prepared for

SOUTHERN NEVADA WATER AUTHORITY

Prepared by

EarthKnowledge
Information for Sound Decisions

June 2011

SNWA Exhibit 87

Transient Numerical Model of Groundwater Flow for the Central Carbonate-Rock Province: Clark, Lincoln, and White Pine Counties Groundwater Development Project

November 2009

A Summary of the Development of the Central Carbonate-Rock Province Groundwater Flow Model

PRESENTATION TO THE OFFICE OF THE NEVADA STATE ENGINEER

Prepared for

SOUTHERN NEVADA
WATER AUTHORITY

Prepared by

EarthKnowledge
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June 2011

Table 4-10
Setup of Regional and Intermediate Springs in the Numerical Model of the Central Carbonate-Rock Province
(Page 1 of 2)

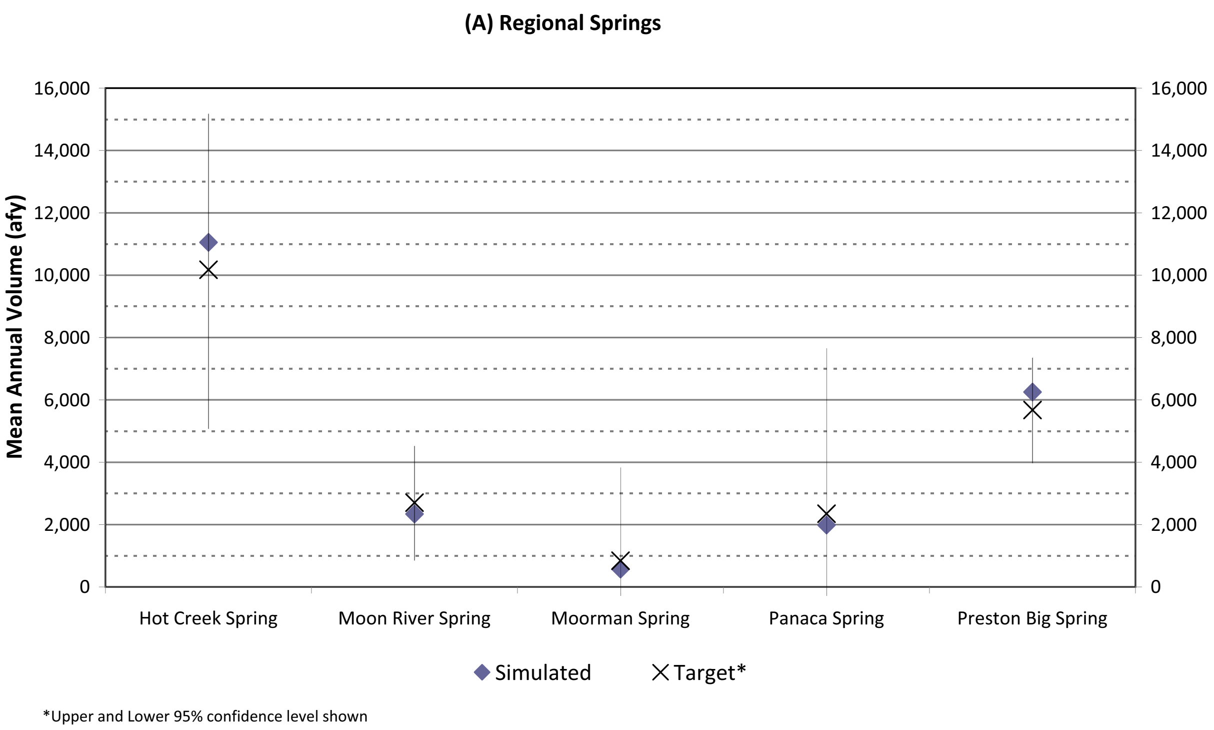
Model Type[a]	Spring Name	DRN and SFR2 Spring Name[b]	Observation Type	Comment
CHD	Deadman Spring	---	---	---
CHD	North Springs	---	---	In an active cell next to CHD
CHD	Walter Spring	---	---	---
CHD	Wilson Hot Spring 1	---	---	---
CHD	Wilson Hot Spring 2	---	---	---
CHD	Wilson Hot Spring 3	---	---	In an active cell next to CHD
CHD	Wilson Hot Spring 5	---	---	---
DRN	Arnoldson Spring	SPiw07_2_01	Flow	---
DRN	Blue Point Spring	SDiw15_2_##	Flow	---
DRN	Brownie Spring	SPis09_4_01	Flow	---
DRN	Butterfield Spring	SPib07_10_01	Flow	---
DRN	Caine Spring	SPis95_3_01	Flow	---
DRN	Campbell Ranch Springs	SPib79_5_01	Flow	---
DRN	Cherry Creek Hot Springs	SPr79_2_01	Flow	---
DRN	Cold Spring	SPiw07_3_01	Flow	---
DRN	Cold Spring	SPis79_4_01	Flow	---
DRN	Currie Spring	SPib79_6_01	Flow	---
DRN	Emigrant Springs	SPib07_15_01	Flow	---
DRN	Flag Springs 1	SPiw207_7	Flow	Flag Springs 1, 2, and 3 in same cell; flow aggregated in one observation
DRN	Flag Springs 2			
DRN	Flag Springs 3			
DRN	Foote Res. Spring	SPib95_12_01	Flow	---
DRN	Four Wheel Drive Spring	SPis84_11_01	Flow	---
DRN	Hardy Spring NW	SPis07_11_01	Flow	Hardy and Hardy NW in same cell; flow aggregated in one observation
DRN	Hardy Springs			
DRN	Hot Creek Spring	SPr07_1_01	Flow	---
DRN	Keegan Spring	SPis84_12_01	Flow	---
DRN	Kell Spring	SPis95_13_01	Flow	---
DRN	Knoll Spring	SPis95_4_01	Flow	---
DRN	Layton Spring	SPis84_7_01	Flow	---
DRN	Lund Spring	SPib07_5_01	Flow	---
DRN	McGill Spring	SPiw79_1_01	Flow	---
DRN	Minerva Spring	SPis84_13_01	Flow	---
DRN	Monte Neva Hot Springs	SPr79_3_01	Flow	---
DRN	Moon River Spring	SPr07_14_01	Flow	---
DRN	Moorman Spring	SPr07_6_01	Flow	---
DRN	Nicholas Spring	SPiw07_13_01	Flow	---

Table 4-10
Setup of Regional and Intermediate Springs in the Numerical Model of the Central Carbonate-Rock Province
(Page 2 of 2)

Model Type[a]	Spring Name	DRN and SFR2 Spring Name[b]	Observation Type	Comment
DRN	North Millick Spring	SPis84_3_01	Flow	---
DRN	North Spring	SPiw84_8_01	Flow	---
DRN	Osborne Springs	SPis84_10_01	Flow	---
DRN	Panaca Spring	SPr03_1_01	Flow	---
DRN	Preston Big Spring	SDr07_4_##	Flow Change	---
DRN	Preston Big Spring	SDr07_4_58	Flow Change	---
DRN	Preston Big Spring	SPr07_4_01	Flow	---
DRN	Rogers Spring	SDiw15_1_##	Flow Change	---
DRN	Rogers Spring	SPiw15_1_01	Flow	---
DRN	South Bastian Spring	SPis84_5_01	Flow	---
DRN	South Bastian Spring 2	SPis84_6_01	Flow	---
DRN	South Millick Spring	SPib84_4_01	Flow	---
DRN	Stonehouse Spring	SPis84_14_01	Flow	---
DRN	The Seep	SPiw84_15_01	Flow	---
DRN	Twin Spring	SPib95_15_01	Flow	---
DRN	Unnamed 5 Spring	SPis84_16_01	Flow	---
DRN	Unnamed Spring	SPis95_14_01	Flow	---
DRN	Warm Creek near Gandy, UT	SPiw95_2_01	Flow	---
DRN	Willard Springs	SPis84_2_01	Flow	---
DRN	Willow Spring	SPiw84_1_01	Flow	---
SFR2	Ash Springs	GdASH_61	Flow	---
SFR2	Big Springs	GdBIG_SPR_61	Flow	---
SFR2	Crystal Springs	GdXTL_61	Flow	---
SFR2	End of Lake Creek	GdLKCK_END_##	Flow	---
SFR2	End of Pahranagat Wash	GdPW_7_##	Flow	---
SFR2	Hiko Spring	GdHIKO_01	Flow	---
SFR2	Muddy River at Lake Mead	GdLK_MEAD_01	Flow	---
SFR2	Muddy River at Overton	GdOVERTON_61	Flow	---
SFR2	Muddy River near Glendale	GdmrGLEND_08	Flow	---
SFR2	Muddy River near Moapa	GdmrMOAPA_##	Flow	Baldwin Spring, Jones Spring, M-10, M-11, M-12, M-13, M-15, M-16, M-19, M-20, Muddy Spring, Pederson East Spring, Pederson Spring, and Warm Springs East aggregated in Muddy River near Moapa SFR2 gage observation

[a]DRN: MODFLOW-2000 Drain package; SFR2: MODFLOW-2000 Streamflow-Routing package;
CHD: MODFLOW-2000 Constant-Head package (Springs within CHD cells not represented in the model).
[b]Used as MODFLOW-2000 and UCODE_2005 observation names in DRN and SFR2 packages. ## indicates two-digit number corresponding to stress period.

SNWA Exhibit 89



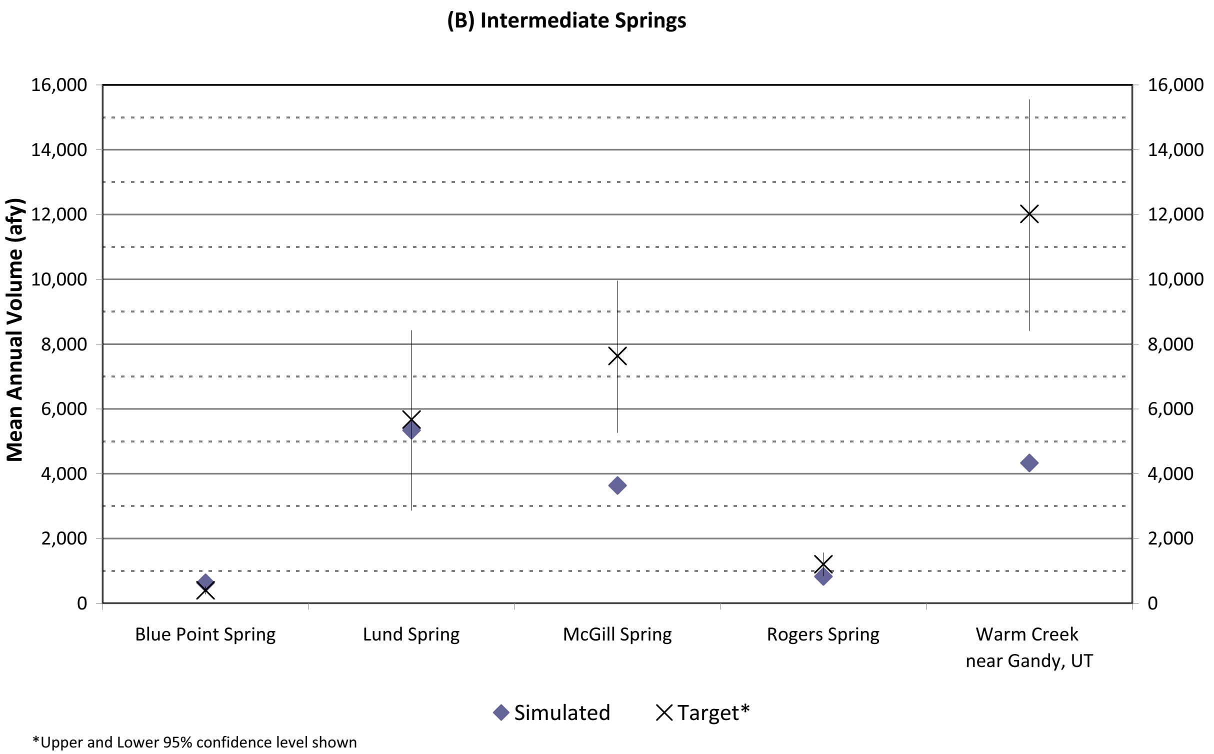


Figure 6-33
Groundwater Discharge from Regional (A) and Intermediate (B) Springs Simulated and Target with ±2 Standard Deviations

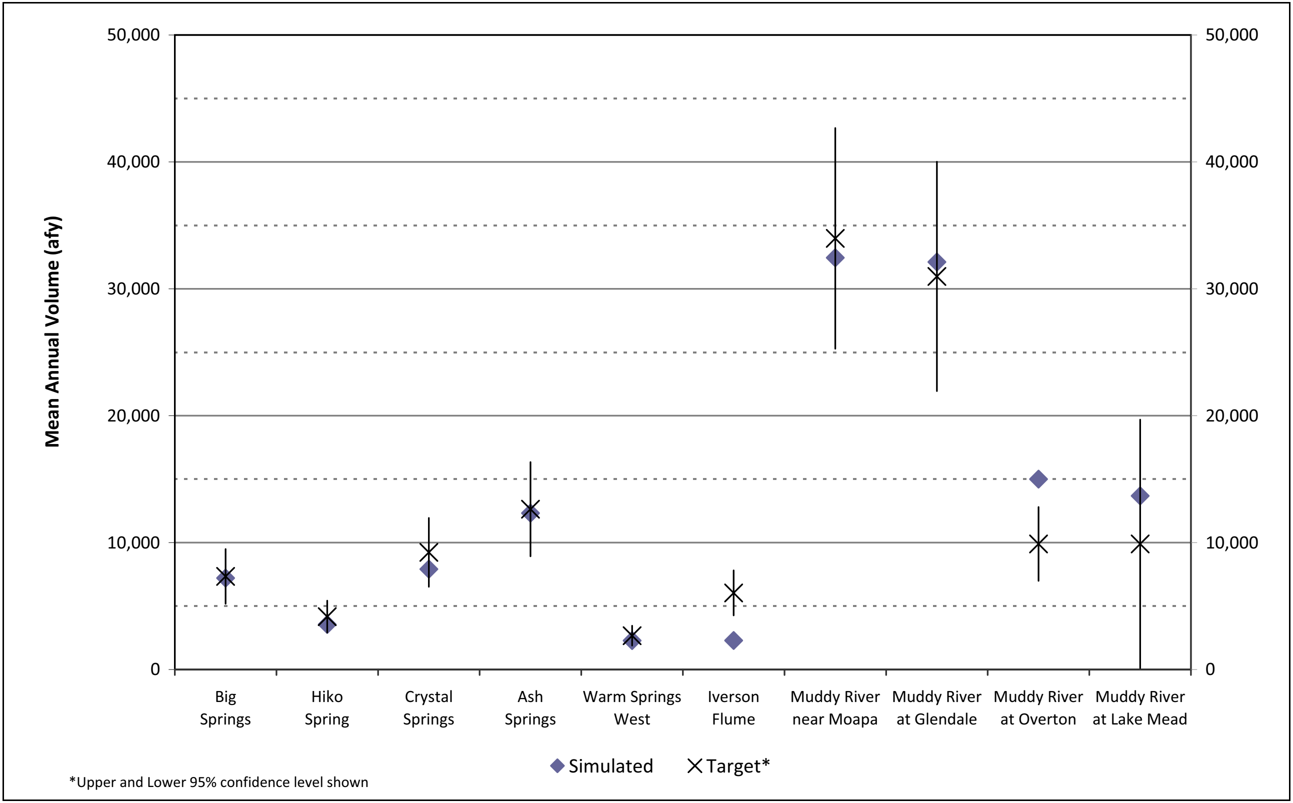
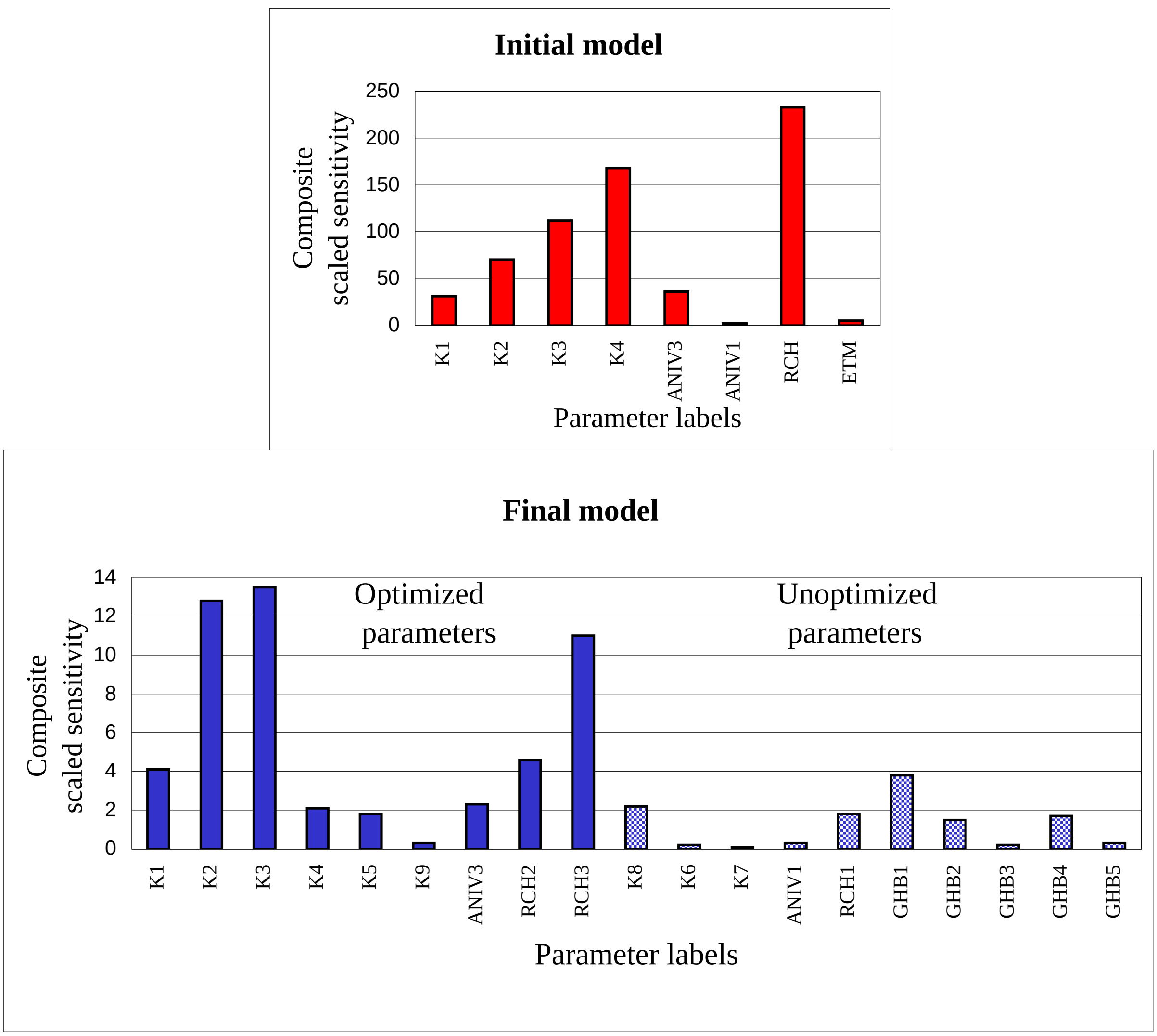


Figure 6-34
Groundwater Discharge at Stream Flow Routing Gages
Simulated and Target with ±2 Standard Deviations

METHODS AND GUIDELINES FOR EFFECTIVE MODEL CALIBRATION

U.S. GEOLOGICAL SURVEY
WATER-RESOURCES INVESTIGATIONS REPORT 98-4005

With application to
UCODE, a computer code for universal inverse modeling, and
MODFLOWP, a computer code for inverse modeling with MODFLOW



SNWA Exhibit 92

Review and Evaluation of the Spring Valley Groundwater Model Developed by Myers (2011b)

PRESENTATION TO THE OFFICE OF THE NEVADA STATE ENGINEER

Prepared for

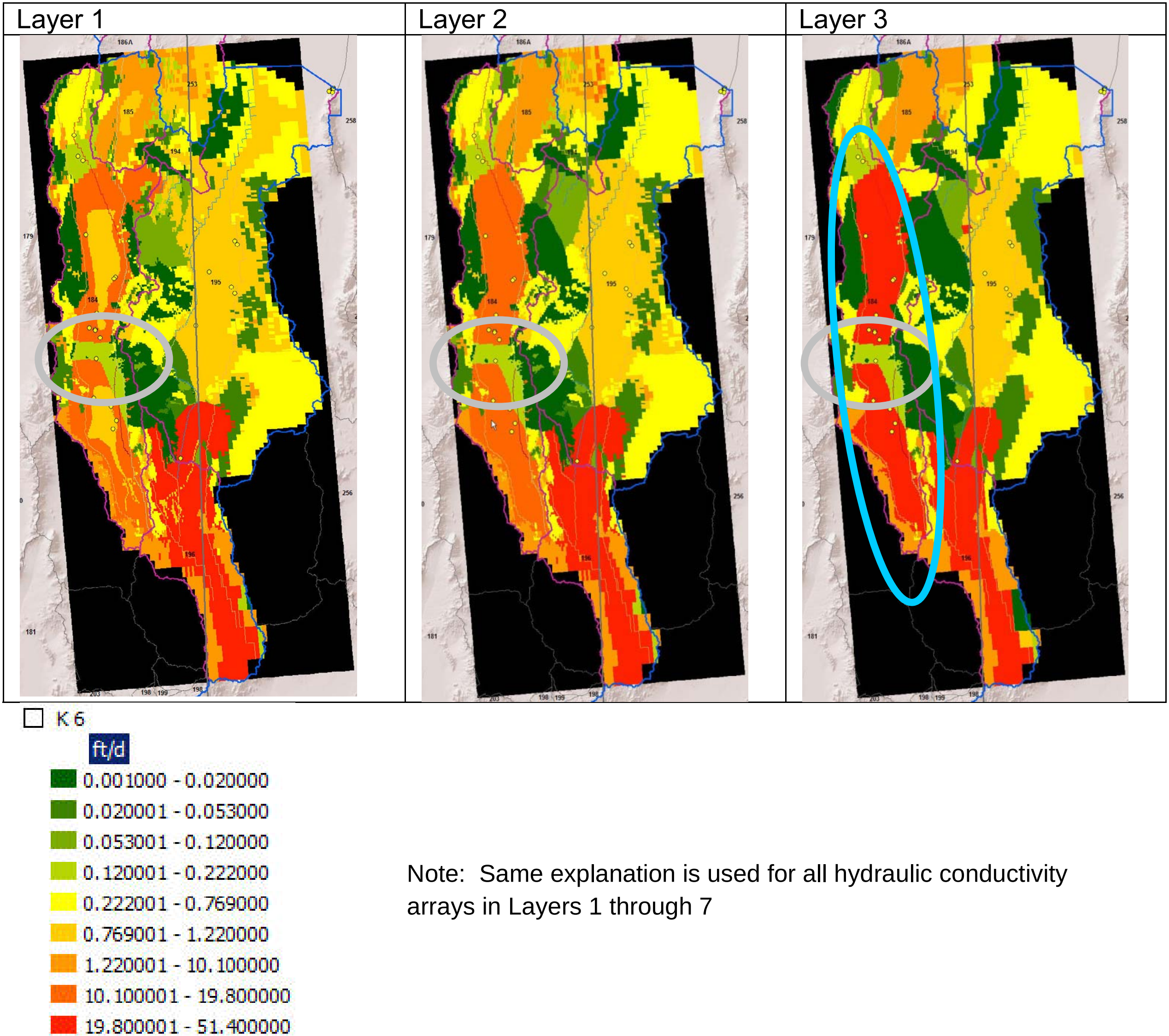
SOUTHERN NEVADA WATER AUTHORITY

Prepared by

EarthKnowledge
Information for Sound Decisions

August 2011

SNWA Exhibit 404



Note: Same explanation is used for all hydraulic conductivity arrays in Layers 1 through 7

Figure 1
Hydraulic Conductivity Distribution in Layers 1, 2, and 3

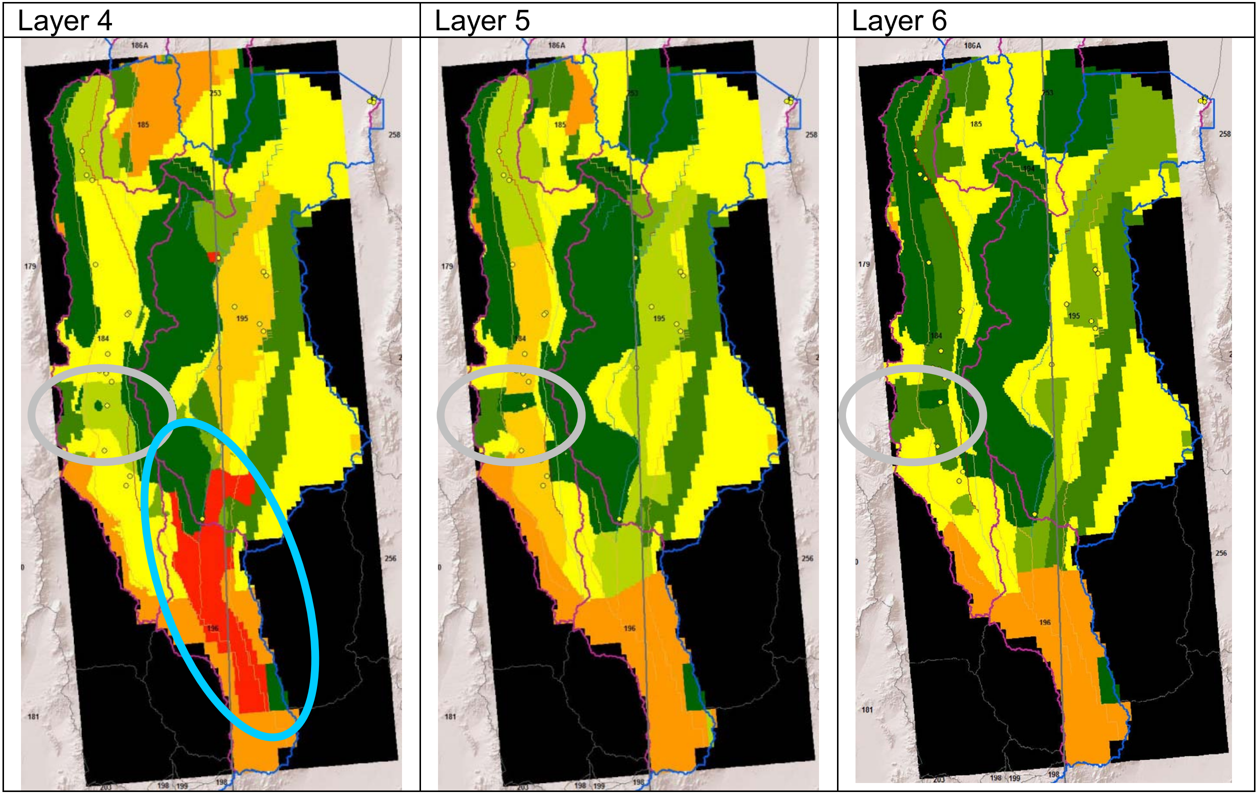


Figure 2
Hydraulic Conductivity Distribution in Layers 4, 5, and 6

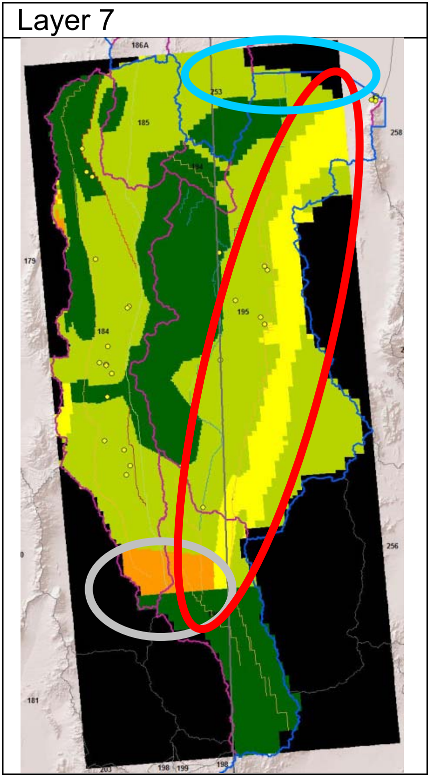


Figure 3
Hydraulic Conductivity Distribution in Layer 7

SNWA Exhibit 404

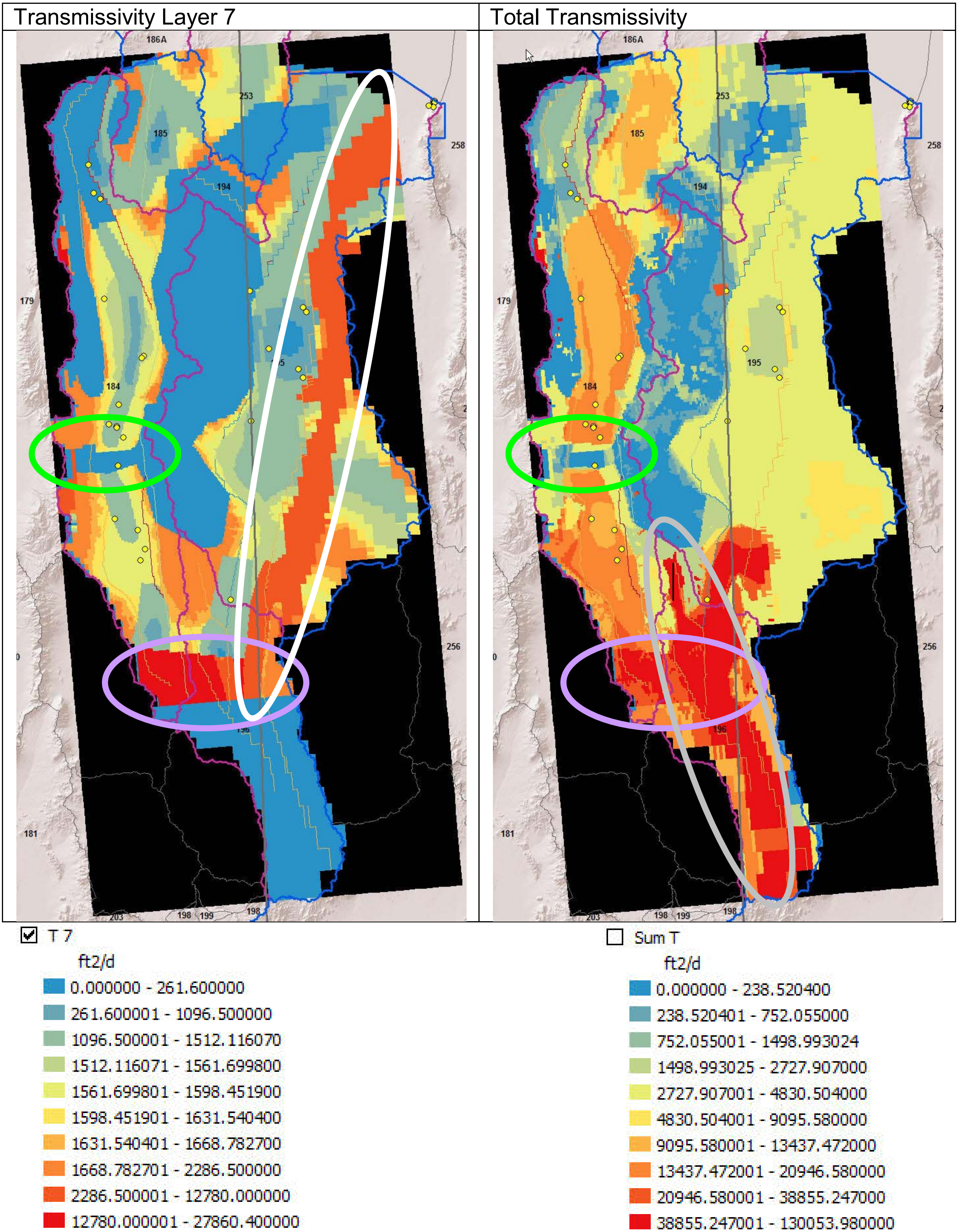


Figure 4
Transmissivity Distribution for Layer 7 and for the Total Model Thickness

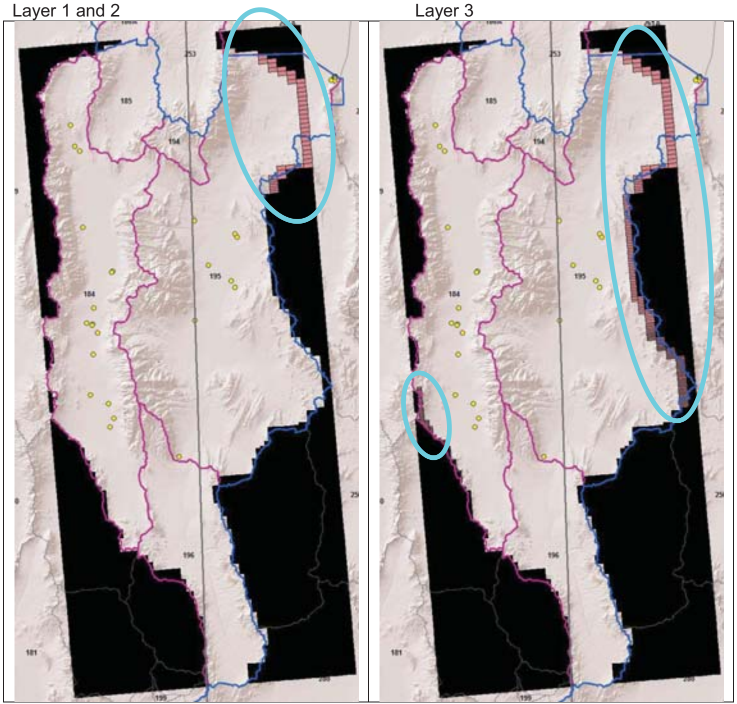


Figure 5
General Head Boundaries for Layers 1, 2, and 3

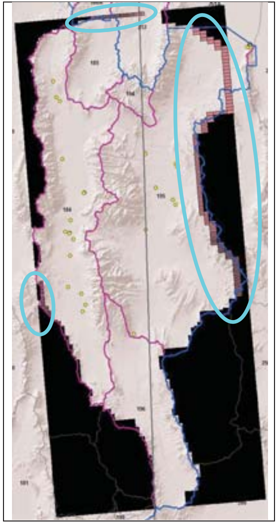


Figure 6
General Head Boundaries for Layers 4 through 7

SNWA Exhibit 404

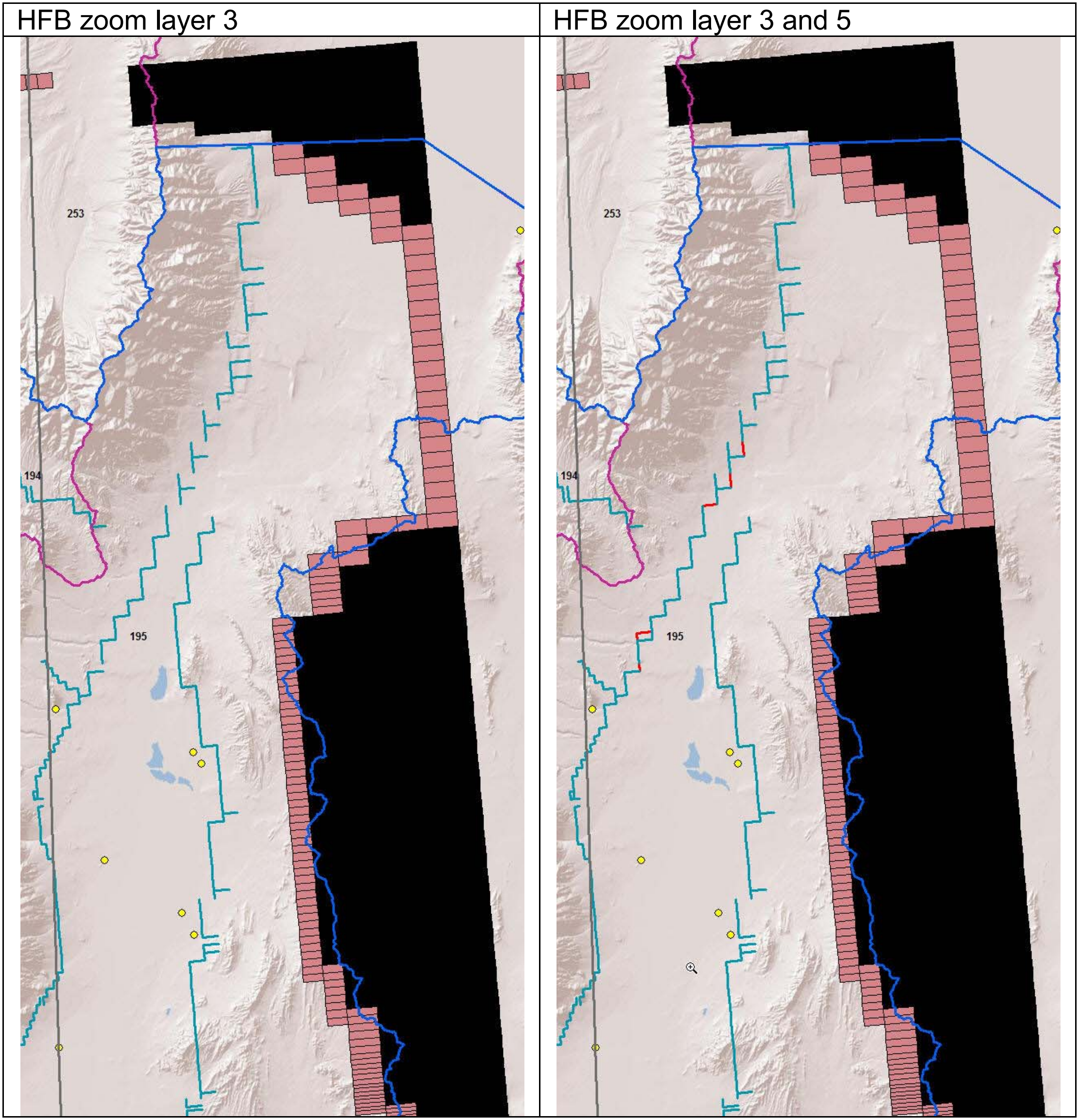
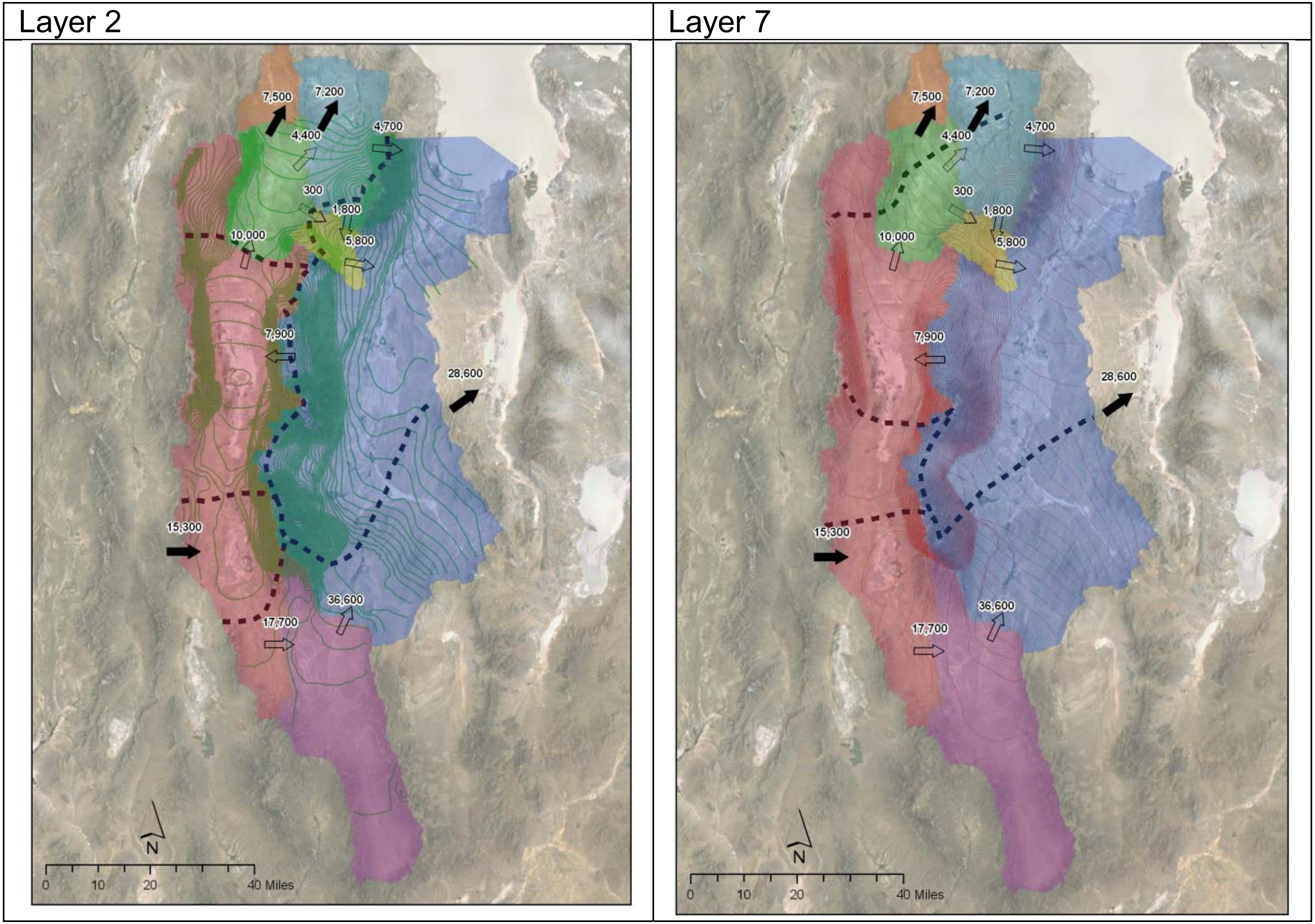


Figure 7
Horizontal Flow Barriers Layers 3 and 5



Note: Interbasin flow represents flow across a hydrographic area boundary for the entire model thickness.

Figure 8
Flow Regions Based on Simulated Water Levels for Layers 2 and 7

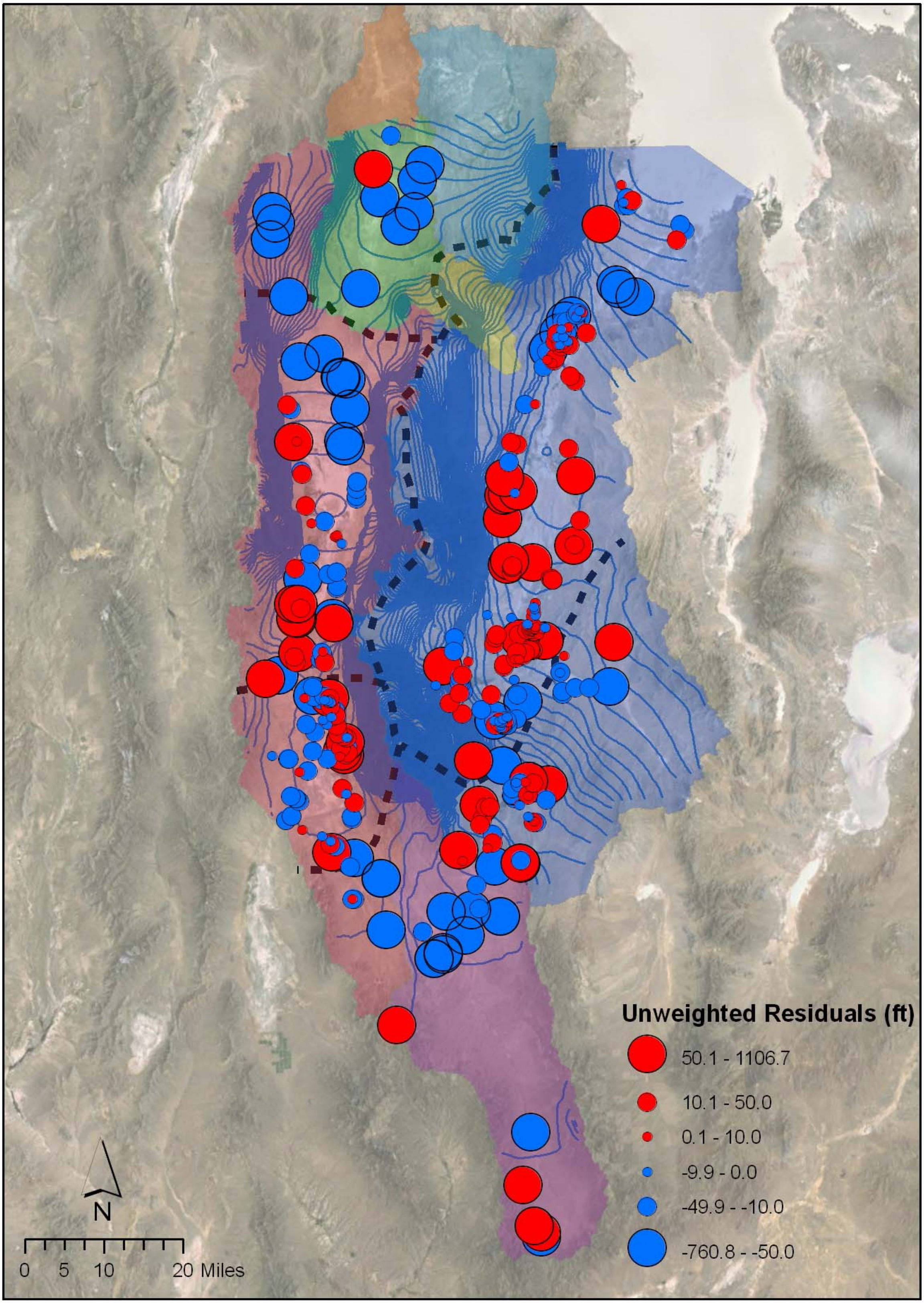


Figure 9
Map of Unweighted Residuals Based on CCRP Observations

Table 1
Maximum and Myers Model Simulated ET Rates and ET Extinction Depths

Myers Model ET Zone	Type	Valley	Calibrated Maximum ET Rate		BARCASS ET Rate		Myers Simulated ET Rate[a]		Extinction Depth (ft)
			(ft/d)	(ft/yr)	(ft/d)	(ft/yr)	(ft/d)	(ft/yr)	
1	Playas	All	0.00073	0.27	0.00197	0.72	0.000608	0.22	30
2	Sparse shrub	Snake	0.00236	0.86	0.00236	0.86	0.000830	0.30	50
3	Sparse shrub	Spring	0.0004	0.15	0.00258	0.94	0.000352	0.13	50
4	Moderate shrub	Snake	0.00288	1.05	0.00288	1.05	NA	NA	50
5	Moderate shrub	Spring, Tippett	0.00301	1.10	0.00201	0.73	0.001808	0.66	50
6	Moist bare soil	Spring	0.00548	2.00	0.00548	2.00	0.004214	1.54	20
7	Avg of marsh and meadowland	Snake[b]	0.00908	3.31	0.00908	3.31	0.005649	2.06	20
8	Avg of marsh and meadowland	Spring[b]	0.00738	2.69	0.00933	3.41	0.004497	1.64	20
9	Sparse shrub	Tippett	0.00271	0.99	0.00271	0.99	0.002144	0.78	50
11	Riparian marshland	Spring	0.0114	4.16	0.01123	4.10	0.010275	3.75	20
NA	Close to BARCAS agriculture areas	Snake	0.003501	1.28	---	---	0.002375	0.87	50

[a]ET rate at water table.
[b]Valley not detectable in Table 3 of Myers (2011b)

Review and Evaluation of the Cave, Dry Lake, and Delamar Valleys Groundwater Model Developed by Myers (2011d)

PRESENTATION TO THE OFFICE OF THE NEVADA STATE ENGINEER

Prepared for

SOUTHERN NEVADA WATER AUTHORITY

Prepared by

EarthKnowledge
Information for Sound Decisions

August 2011

SNWA Exhibit 405

CONCEPTUAL EVALUATION OF REGIONAL GROUND-WATER FLOW IN THE CARBONATE-ROCK PROVINCE OF THE GREAT BASIN, NEVADA, UTAH, AND ADJACENT STATES

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1409-D

Simulated Effects of Proposed Ground-Water Pumping in 17 Basins of East-Central and Southern Nevada

By DONALD H. SCHAEFER and JAMES R. HARRILL

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 95-4173

Prepared in cooperation with the
NATIONAL PARK SERVICE,
U.S. FISH AND WILDLIFE SERVICE,
BUREAU OF LAND MANAGEMENT, and
BUREAU OF INDIAN AFFAIRS

Carson City, Nevada
1995

770,000 in 1990. As the number of people in the province increases and surface-water supplies become less available, additional sources of water will be needed. One such source that has been proposed (Hess and Mifflin, 1978) is the water stored in the carbonate rocks beneath much of western Utah and eastern Nevada.

In most other RASA studies, enough information exists for comprehensive model simulations and evaluations of ground-water flow in regional aquifer systems. Although numerous wells have been drilled within the carbonate-rock province, most have been drilled into unconsolidated deposits in the valleys and usually to shallow depths, except at the Nevada Test Site. Thus, little is known about the deeper and more regional ground-water flow in the carbonate rocks. However, because of the greatly increased demand for water and because of the potential for contamination of ground water from underground testing of nuclear weapons at the Nevada Test Site (fig. 2) and from the possible storage and disposal of nuclear and hazardous wastes, an improved understanding of ground-water flow in the province is needed.

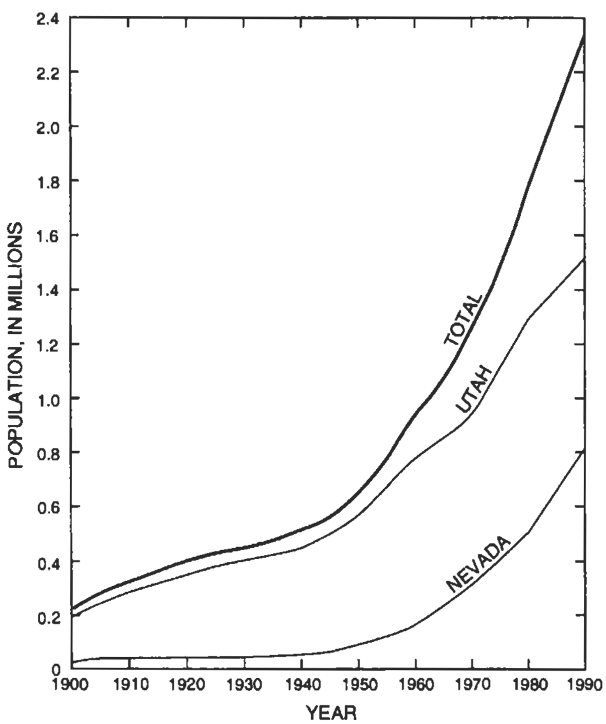


FIGURE 3.—Population growth in study area between 1900 and 1990. Data from U.S. Bureau of Census (1913, 1921, 1952, 1983, 1991a, b).

PURPOSE AND SCOPE

The purpose of this report is to present a conceptual evaluation of ground-water flow in the carbonate-rock province, mainly in Nevada and Utah. The evaluation is based on simulation results using the three-dimensional ground-water flow model of McDonald and Harbaugh (1988). The basic conceptual model for the province includes relatively shallow flow from recharge areas in the mountains to discharge areas in the adjacent valley lowlands, superimposed over deeper, more regional flow through carbonate rocks. The concept is based on theoretical analyses of regional flow by Freeze and Witherspoon (1967, p. 623–634) where, in regions of hummocky terrain, numerous relatively shallow flow systems are superimposed over fewer deeper flow systems. Results of the model analysis include: transmissivity distributions, identification of shallow and deep flow systems, and comparisons of simulated flow and discharge to estimates presented in previous reports.

The original version of this report was published in January 1991 as a U.S. Geological Survey interim Open-File Report and in September 1991 as a U.S. Geological Survey Professional Paper. In November 1991, an error that resulted from an inadvertent coding transposition of the cell-dimension variables DELR and DELC (McDonald and Harbaugh, 1988, chap. 5, p. 8) was discovered. This error produced an unintended regional anisotropy in the model transmissivities (Stillwater and others, 1992). As a result, the model grid cell dimensions have been corrected and the model recalibrated. David E. Prudic did the recalibration and, along with James R. Harrill, has revised the report to reflect changes resulting therefrom. In addition, Donald H. Schaefer and James R. Harrill assisted in checking information used in the model.

PREVIOUS INVESTIGATIONS

Surveys of geologic features in the Great Basin began in the late 1860's under the leadership of Clarence King, J.W. Powell, G.K. Gilbert, A.R. Morvine, and E.E. Howell. Nolan (1943) summarized available geologic information pertaining to the entire Great Basin. Between 1938 and the late 1970's, numerous geologic investigations were completed in the Great Basin region. The results of all these studies and studies before 1938

Total simulated discharge (as evapotranspiration and leakage to head-dependent flow boundaries) along the Humboldt River is 52,000 acre-ft/yr (fig. 35). This quantity represents only a fraction of the total estimated evapotranspiration and streamflow in the Humboldt River valley above Palisade (Eakin and Lamke, 1966, p. 59, 60). Simulation of regional ground-water flow with the model did not account for the local circulation of water adjacent to the Humboldt River; rather, the model is designed to assess the potential for regional flow from distant sources to regional discharge areas. In the upper Humboldt River region, the quantity of simulated deep flow (flow through the lower model layer) to the Humboldt River is small (a few thousand acre-ft/yr) compared to local flow between the river and its alluvium.

POTENTIAL USES OF MODEL

The ground-water flow model of the carbonate-rock province is unlike most models in that the extent of aquifers and their hydraulic properties are generally unknown in the province; thus, the model greatly simplifies flow through a complex geologic region. Simulation results are based on assuming recharge to the province is known with the distribution of transmissivities simulated to match the general distribution of water levels and estimates of discharge. However, water levels in consolidated rocks are generally unknown, and estimates of recharge and discharge are known only approximately. Consequently, other, equally valid distributions of transmissivities may be found that permit the model to be calibrated to the existing water-level data and estimates of recharge and discharge. The model may be best suited for:

- Simulating alternative transmissivity distributions to evaluate potential source areas of regional springs,
- Simulating the effects of differing recharge rates on regional ground-water flow, and
- Simulating the effects of changing location of discharge on regional ground-water flow.

Therefore, the potential uses of the model are limited. The model is not suited to predict accurate water-level declines that would result from pumping ground water in the province. Also, the model is not suited to predict the accurate rate of change in natural discharge caused by pumping, because the model has not been calibrated to any transient simulations.

SUMMARY AND CONCLUSIONS

This report describes the results of a computer-model-based analysis of regional ground-water flow in the eastern Great Basin, a 100,000-mi^2 area that lies mostly in eastern Nevada and western Utah, with small parts in northwestern Arizona, eastern California, and southern Idaho. The original version of this report, published in 1991, presented results that subsequently proved to be adversely affected by a transpositional error in the computer data files that define the model-cell dimensions. This error produced an unintended regional anisotropy in hydraulic conductivity. The results reported herein constitute a reanalysis of regional flow after the transposition was corrected and the computer model recalibrated.

Ground-water flow in the eastern Great Basin has been evaluated as part of the U.S. Geological Survey's Great Basin Regional Aquifer-System Analysis. The area is referred to as the carbonate-rock province because, during the Paleozoic era, thick sequences of limestone and dolomite were deposited in a shallow sea that inundated the area. Since then, many episodes of sediment deposition, volcanic activity, erosion, and tectonic deformation by both compressional and extensional forces have altered the extent and thickness of the carbonate rocks. The present-day physiography, which is characterized by north- to northeast-trending mountain ranges separated by intervening valleys that are partly filled with sedimentary deposits eroded from the adjacent mountains, is the result of normal faulting caused by extension that began about 20 million years ago. Relief between the block-faulted mountains and the adjacent valley floors ranges from 1,000 ft to more than 7,000 ft.

Shallow ground-water reservoirs in the basin fill supply most of the current (1992) pumpage from wells in this geologically complex terrain. Aquifers in the underlying carbonate rocks are largely undeveloped; regionally, however, these aquifers are important because they provide an avenue for interbasin ground-water flow. The source of ground water in the province is precipitation, most of which falls in the higher mountain ranges. Ground-water discharge is mostly by evapotranspiration in the low parts of the many valleys. Some ground water also discharges from small, local springs. Such springs are fed by recharge that originates nearby. In contrast, ground water discharging at larger, regional springs have a value of 0.15. Conductances range from 0.01 ft^2/s for the upstream reaches of the Sevier River to 0.15 for Sevier Lake; conductances for the Sevier River average 0.07 ft^2/s. Conductances for the Virgin River are 0.01 ft^2/s, except for the northernmost cell, which is 0.02 ft^2/s. Conductances for Lake Mead are also 0.01 ft^2/s, except for the two cells nearest the dam, which are 0.5 ft^2/s. The conductance for the one cell representing the Colorado River below the dam is 0.005 ft^2/s. Conductances for Death Valley are 0.1 ft^2/s.

Because flow to and from the head-dependent boundaries are generally controlled by the estimated transmissivities of the model cells, changing conductances does not greatly affect the simulation results. For example, decreasing the conductances for cells that have a value of 0.5 ft^2/s to 0.1 ft^2/s resulted in a slight decrease (0.1 ft^3/s) in discharge and recharge along the Humboldt River and no change to discharge at Lake Mead. Increasing the conductances for nine cells along the Sevier River which had values less than 0.1 ft^2/s by a factor of 10 resulted in a 10-percent increase in discharge (5 ft^3/s increase) to the Sevier River, a corresponding decrease in simulated evapotranspiration, and consequently, no change in the simulated discharge from the area.

Total simulated spring discharge from the lower model layer is only 0.5 percent greater than the total estimated discharge (table 1). However, the percentage difference between simulated and estimated discharge for individual springs is generally more. For example, simulated discharge at Warm Springs (table 1) is 152 percent of the estimated discharge.

During final model calibration, conductance values used to simulate spring discharge were changed to test their sensitivity. Initially, a uniform value of 3 ft^2/s was assigned to each spring. This value is more than two orders of magnitude greater than the initial conductance value assigned between layers (vertical leakance multiplied by cell area). Increasing the conductance value for springs to 10 ft^2/s did not affect discharge from the lower layer, indicating that the discharge was dependent on flow from adjacent model cells. The higher conductance values resulted in slightly reduced water levels in cells where spring discharge was simulated, because not as much head difference was needed to simulate flow through the springs. A value of 10 ft^2/s was used during final model calibration. Spring discharge was extremely sensitive to changes in both transmissivity and vertical leakance.

Land-surface altitude assigned to each model cell in the upper layer controlled the distribution of evapotranspiration and water levels in cells where evapotranspiration was simulated. Initially, land-surface altitudes assigned to each cell were averaged values. This did not produce a reasonable distribution of evapotranspiration and water levels in some areas of the model. Adjusting transmissivities and vertical leakances did not always improve results. Areas of evapotranspiration are generally confined to the lowest parts of a valley. Consequently, minimum land-surface altitudes from the one-minute data were used in areas of known evapotranspiration.

Because evapotranspiration did not reach a maximum rate when water levels exceeded land surface (fig. 10*B*), simulated water levels in cells with evapotranspiration were compared with the assigned land-surface altitude. Whenever water levels exceeded land surface, transmissivity and leakance values in that cell, and sometimes in surrounding cells, were changed to lower heads below land surface. Evapotranspiration of ground water was assumed to occur only from basin fill in the valley lowlands. Thus, the transmissivity and leakance values were increased in a model cell corresponding to consolidated rocks whenever evapotranspiration was simulated in such a cell. Final distribution of simulated evapotranspiration is shown in figure 17. The simulated distribution generally corresponds to areas mapped by Harrill and others (1988, pl. 2). Areas mapped by Harrill and others are shown in figure 18.

The model was deemed calibrated when simulated discharge approximated the mapped distribution and estimated discharge in each hydrographic area. In addition, computed water levels were matched as closely as practical with estimated values. For the best-fit simulation, 86 percent of the simulated water levels (666 out of 773 model cells) were within 250 ft of the estimated water levels for the upper layer and 76 percent (109 out of 144 cells) were within 250 ft for the lower layer.

The 250-ft criterion used for calibration purposes is only 3 percent of the total water-level difference in the model. The maximum simulated water level is more than 7,000 ft above sea level, along the eastern side of the model; in contrast, the minimum is below sea level, in Death Valley. Water-level differences between adjacent model cells commonly exceed 250 ft; in a few locations, they exceed 500 ft. The distribution of water levels in both model layers for the best-fit calibration is shown in figure 19.

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Mifflin (1968) classified springs in Nevada as local, intermediate, and regional on the basis of water chemistry, water temperature, and fluctuation of flow from the springs. Regional springs presumably represent the discharge of deep flow through carbonate rocks. Locations of the regional springs, as delineated by Thomas and others (1986) using similar criteria, are shown in figure 7. The largest concentration of regional springs is in a small area at Muddy River Springs. The flow of these springs totals about 36,000 acre-ft/yr (Eakin and Moore, 1964).

Most ground-water withdrawals in the province are from wells drilled into the basin fill beneath the valley floors because (1) people settled in the valleys where the climate is less severe than the mountains and where the land is more suitable for agriculture, (2) ground water in many of the valleys is generally within a few feet to several tens of feet below land surface in contrast to generally deeper water levels in mountain areas, and (3) the basin fill generally yields large quantities of water to wells. Eakin and others (1976, p. 15) reported yields as much as 8,600 gallons per minute from large-capacity wells in north-central Utah.

Prior to World War II, most of the ground-water withdrawals were from flowing wells drilled into basin fill. Areas of flowing wells were concentrated largely along the eastern side of the province in valleys adjacent to the Wasatch Range, although several other valleys, including Las Vegas Valley, also had flowing wells. Ground-water withdrawals were generally small and constant until after World War II, when more efficient pumps and inexpensive energy greatly increased the quantity of ground water withdrawn to irrigate crops and to supply a rapidly increasing population. The total quantity of ground water withdrawn in the province during 1975 was approximately 1 million acre-ft. Major areas of ground-water withdrawals during 1975 are shown in figure 8.

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FIGURE 7.—Principal source areas for ground-water recharge, areas where ground water is consumed by evapotranspiration, and regional springs (discharge exceeds 100 gallons per minute; water chemistry indicates long flow time, mostly within carbonate rocks). Recharge and evapotranspiration areas from Mifflin (1988, pl. 3); spring locations and criteria from Thomas and others (1986, pl. 2).

CONCEPTUAL EVALUATION OF GROUND-WATER FLOW

Computer models are tools that can be used effectively to help understand complex ground-water flow systems. However, rarely are computer models used to simulate ground-water flow over such a large and geologically complex area as the carbonate-rock province. Endless arguments could be invoked as to the validity of the assumptions and hydrologic values used in simulating ground-water flow within the carbonate-rock province. For this reason, it must be stressed that the computer simulation discussed in this report is conceptual in nature. Only broad concepts and large-scale features can be inferred from the results of this study. Although a fairly detailed analysis of ground-water flow will be discussed, it does not intend to indicate that the study results presented here are adequate; in fact, the objective in presenting a detailed analysis of ground-water flow is to examine the possibility of the relatively shallow flow regions being interconnected by deep flow through carbonate rocks, and how regional geologic features might affect the direction of flow and water levels.

GENERAL ASSUMPTIONS

In the carbonate-rock province, ground-water flow takes place through the pores of basin-fill sedimentary deposits and through the fractures and solution openings of consolidated rocks. On a regional scale, flow through fractures and solution openings in the consolidated rocks is assumed to be the same as flow through a porous medium; that is, it was assumed that Darcy's Law is applicable. This assumption may be reasonable because the model grid used to simulate regional flow results in the averaging of hydraulic properties over 37.5-mi^2 areas. However, not enough information is available for the study area to substantiate the assumption.

Model simulations assume steady-state conditions prior to development, in which estimates of current recharge (1950–80) equal estimates of natural discharge prior to ground-water development. That is, the model does not include ground-water withdrawals. Whether current recharge equals natural discharge is unknown. During the late Wisconsin glaciation (from about 20,000 to 10,000 years ago), ground-water flow in the prov-

Conductance values used to simulate the interaction of ground water with surface water (general-head boundaries in fig. 9) were changed during model calibration until the simulated water-level gradients near the boundaries approximated the estimated gradients.

LIMITS OF CALIBRATION

Results from the model simulation are only approximate because uncertainties exist in the distribution and quantity of recharge and because water levels in the consolidated rocks are unknown over much of the area. Although discussed in detail, the model results are conceptual because actual values are not known for any of the variables in the ground-water flow equation. In particular, other, equally valid, distributions of transmissivity may be found that permit the model to be calibrated to the existing information. Model results are also dependent on the general assumptions discussed previously.

Transmissivities estimated for both model layers are in part dependent on the quantity and distribution of recharge used in the model, particularly for model cells that correspond to mountains. Recharge is simulated in the mountains except where head-dependent flow boundaries are used to simulate the interaction of ground water with surface water. Simulating all recharge in mountains that consist of carbonate rocks is probably reasonable because little surface water flows to the nearby valleys. But in mountains that consist of low-permeability rocks, much of the water flows into nearby valleys where recharge occurs mostly on the adjacent alluvial fans. Thus, the transmissivities estimated for model cells that represent these mountains are probably higher than the actual transmissivities.

Transmissivities in the upper model layer are highly sensitive to changes in both the quantity and location of recharge. Transmissivities for the lower model layer are not as sensitive to changes in recharge, because recharge is not added directly to cells in this layer. Recharge to the lower layer is dependent on the leakage between the upper and lower layers, which is controlled by the vertical leakance.

Errors in the estimates of recharge are unknown but locally could be well in excess of 100 percent. If recharge is increased in the model by 100 percent, a similar distribution of water levels could be simulated by proportionately increasing transmissivities and vertical leakances. Because the model assumes steady-state conditions, discharge would also increase by 100 percent. However, a different distribution of transmissivity and vertical leakance near regional springs would be needed if the additional recharge was forced to discharge as evapotranspiration instead of allowing spring discharge to increase as well.

Estimates of water levels used to calibrate transmissivities in the lower model layer are based on limited data. Locally, transmissivities could be changed an order of magnitude, and model results might still be reasonable with respect to areas of estimated water levels and quantities of simulated discharge. Large cell sizes and the generalization of transmissivities result in a more gradual change in simulated water levels than might be expected from abrupt lateral and vertical changes in geologic units observed in the study area. Where geologic structures are barriers to flow in south-central Nevada, water-level differences between adjacent valleys are as much as 2,000 ft (Winograd and Thordarson, 1975, p. 63). With cell sizes of 5 mi by 7.5 mi, the model tends to smooth such large differences.

The model is designed to simulate ground-water flow at a regional scale. Orientation of the columns in the model grid corresponds to the general trend of range-front faults. These faults are thus parallel and perpendicular to the two directions of horizontal transmissivity. However, range-front faults are not the only faults present in the province. The mountains are extensively faulted, as presumably are the rocks beneath the basin fill. Orientation of the model grid to coincide with the range-front faults therefore may be unnecessary. Also, transmissivity in one of the two principal directions could be changed with respect to the other direction over the entire modeled area, although no compelling reason was discovered to simulate such a condition. Anisotropy probably exists on a more localized scale, but available computer programs do not allow anisotropy to be specified by individual model cells. Localized anisotropic conditions could be simulated by reducing the dimensions of the model cells. The simulation of ground-water flow with smaller cell dimensions is not beyond the scope of this study. However, insufficient data over large areas preclude such a detailed simulation.

SIMULATION RESULTS

Discussion of the simulation results has been divided into three sections: (1) estimated transmissivities, (2) correlation of ground-water flow to regional geologic features, and (3) distribution of flow into regions.

ESTIMATED TRANSMISSIVITIES

Transmissivities in both model layers were estimated by adjusting the initial values until simulated water levels generally agreed with estimated water levels and the quantity and distribution of simulated discharge approximated those of the estimated discharge. The transmissivities are also dependent on the quantity and distribution of recharge assigned to cells corresponding to mountain ranges. Estimated transmissivities for the upper and lower model layers are shown in figure 20.

Errors in transmissivities are unknown, but the estimates could be off by a factor of 5 or more. Other uncertainties used in the model also result in unknown errors, especially the assumption of isotropy in each 37.5-mi^2 model cell in an area of complex geology. Consequently, transmissivities are discussed using the qualitative terms listed as follows:

Qualitative term	*Transmissivity range (feet squared per second)*
Lowest	<0.0006
Low	0.0006–0.006
High	0.006–0.18
Highest	0.18–0.66

In the upper model layer, no distinct pattern of transmissivities is simulated (fig. 20*A*), perhaps because of areal variability in the quantity and distribution of recharge. Highest transmissivities are scattered in small groups of cells throughout much of the province. Lowest transmissivities are concentrated in the Great Salt Lake Desert, in the vicinity of Death Valley, and in the extreme southern part of the province. Low values are assigned in the Great Salt Lake Desert to match estimated ground-water discharge. Circulation of fresh ground water in this area is assumed minimal because the area is underlain by an extensive body of saline ground water. Low values are assigned in the vicinity of Death Valley and in the southern part of the province to simulate large hydraulic gradients between Death Valley and adjacent basins. Outcrops of Cambrian and Precambrian clastic rocks, assumed to be poorly permeable, are common in the mountains surrounding Death Valley.

In the lower model layer, high transmissivities are generally grouped in areas associated with regional springs or in the vicinity of basins where ground-water discharge is considerably more than the estimated recharge from tributary drainage areas (fig. 20*B*). Highest values are simulated in narrow bands near regional springs in the White River Valley in eastern Nevada, near the Muddy River Springs area in southern Nevada, and near Fish Springs in west-central Utah. Elsewhere in the province, low transmissivities are simulated. Lowest transmissivities are simulated in the Great Salt Lake Desert, Death Valley, and the extreme southern end of the province, with an areal distribution similar to that of the upper layer.

Transmissivities in the upper and lower model layers are summarized in table 2.

The geometric mean transmissivity of the upper layer is greater than that of the lower layer even though the minimum, median, and maximum values in the upper layer are less than those in the lower layer. However, the 25th-and 75th-percentile values are nearly an order of magnitude greater in the upper layer. The reason for this seeming disparity is that the estimated transmissivities in the model cells are assigned values that generally differ by an order of magnitude. In the upper layer, about 40 percent of the active cells (979 of 2,456 cells) are assigned an estimated transmissivity of 0.022 ft^2/s, whereas in the lower layer approximately half of the active cells (1,187 of 2,456 cells) are assigned an estimated transmissivity of 0.0033 ft^2/s.

As a result of model calibration, estimated transmissivities in both model layers are generally less than the initially assigned values. Initially, one of three transmissivity values was assigned to groups of model cells in the upper layer on the basis of surficial geology (that is, carbonate rocks, basin fill, or consolidated rocks of low permeability; fig. 15), and one value representing carbonate rocks was assigned to all cells in the lower layer. The frequency distribution of estimated transmissivities for the three groups of rocks in the upper layer is shown in figure 21. Also shown is the frequency distribution for

Conductance values used to simulate the interaction of ground water with surface water (general-head boundaries in fig. 9) were changed during model calibration until the simulated water-level gradients near the boundaries approximated the estimated gradients.

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Table 1
Estimate Discharge of Regional Springs Compared with Simulated Discharge Following Model Calibration

Regional Spring	Map No. (fig. 11)	Discharge (acre-feet per year) Estimated	Simulated	Source of Discharge Estimate	Absolute Residual (afy)	cfs	% of Est.
Manse Springs	1	4,300	3,900	Maxey and Jameson, 1948, p. 9-10	400	0.55	9%
Ash Meadows area (several springs)	2	17,000	17,000	Winograd and Thordarson, 1975, p. C78-C80	0	0.00	0%
Rogers and Blue Point Springs	***3***	***1,500***	***1,200***	***Rush, 1968b, p. 39***	***300***	***0.41***	***20%***
Muddy River Spring Area	4	36,000	37,000	Eakin, 1966, p. 264	1,000	1.38	3%
Grapevine and Stainigers Springs	***5***	***1,000***	***720***	***Miller, 1977, table 4***	***280***	***0.39***	***28%***
Pahranagat Valley (several springs)	6	25,000	24,000	Eakin, 1963, p. 20	1,000	1.38	4%
Panaca Warm spring	***7***	***7,900***	***9,900***	***Rush, 1964, table 9***	***2,000***	***2.76***	***25%***
Hot Creek Ranch Springs	***8***	***1,800***	***2,000***	***Rush and Everett, 1966a, table 9***	***200***	***0.28***	***11%***
Lockes (several springs)	***9***	***2,400***	***2,800***	***Van Denburgh and Rush, 1974, p. 23, 50-52***	***400***	***0.55***	***17%***
Blue Eagle and Tom Springs	***10***	***3,700***	***3,200***	***Van Denburgh and Rush, 1974, p. 25, 50-51, Mifflin 1968, table 4***	***500***	***0.69***	***14%***
Moon River and Hot Creek Springs	11	13,000	13,000	Maxey and Eakin, 1949, p. 37	0	0.00	0%
Mormon Hot Spring	***12***	***3,100***	***2,200***	***Maxey and Eakin, 1949, p. 37***	***900***	***1.24***	***29%***
Northern White River Valley (several springs)	***13***	***12,000***	***10,000***	***Maxey and Eakin, 1949, p. 39***	***2,000***	***2.76***	***17%***
Duckwater (Big and Little Warm Springs)	***14***	***11,000***	***13,000***	***Van Denburgh and Rush, 1974, p. 23, 50-52***	***2,000***	***2.76***	***18%***
Fish Creek Spring	***15***	***3,900***	***2,800***	***Rush and Everett, 1966a table 9***	***1,100***	***1.52***	***28%***
Twin Spring	***16***	***2,900***	***4,000***	***Hood and Rush, 1965, table 9***	***1,100***	***1.52***	***38%***
Campbell Ranch Spring	17	7,700	7,400	Eakin et al., 1967, table 4	300	0.41	4%
Shipley Hot Springs and Bailey Spring	***18***	***5,700***	***4,400***	***Harrill, 1968, p. 31***	***1,300***	***1.79***	***23%***
Fish Springs	19	27,000	26,000	Bolke and Sumsion, 1978, p. 10	1,000	1.38	4%
Nelson Springs (Currie Springs)	***20***	***2,200***	***1,800***	***Eakin et al., 1967, table 4***	***400***	***0.55***	***18%***
Blue Lake and Little Springs	***21***	***18,000***	***20,000***	***Gates and Kruer, 1981, table 8***	***2,000***	***2.76***	***11%***
Warm Springs	***22***	***3,300***	***5,000***	***Eakin et al. 1951, p. 108***	***1,700***	***2.35***	***52%***
Total discharge, all regional springs (rounded)		210,000	211,000				

Source: From Prudic et al. (1995)

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FIGURE 7.—Principal source areas for ground-water recharge, areas where ground water is consumed by evapotranspiration, and regional springs (discharge exceeds 100 gallons per minute; water chemistry indicates long flow time, mostly within carbonate rocks). Recharge and evapotranspiration areas from Mifflin (1988, pl. 3); spring locations and criteria from Thomas and others (1986, pl. 2).

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Model simulations assume steady-state conditions prior to development, in which estimates of current recharge (1950–80) equal estimates of natural discharge prior to ground-water development. That is, the model does not include ground-water withdrawals. Whether current recharge equals natural discharge is unknown. During the late Wisconsin glaciation (from about 20,000 to 10,000 years ago), ground-water flow in the province may have been more than that of the present day because the climate was significantly wetter, with numerous lakes in the closed basins (Hubbs and Miller, 1948). Ground-water levels and spring discharge may not be in equilibrium with the present-day recharge because of the long distances between areas of recharge and discharge. That is, the water levels and spring flows may still be declining in response to the drier climate of today relative to that of 10,000–20,000 years ago.

Evidence of a long-term water-table decline at Ash Meadows, in the southern part of the province near Death Valley (fig. 1), is presented by Winograd and Szabo (1986). They estimated a slow rate of decline—0.07 to 0.26 ft per 1,000 years. This range of rates is based on (1) uranium-disequilibrium dating of calcitic veins as much as 160 ft (reported as 50 meters) higher than the highest present-day water level at Ash Meadows and as much as 8.7 mi (reported as 14 kilometers) up the hydraulic gradient, and (2) the assumption that the rate of decline has been constant for the past 510,000 to 750,000 years. The calcitic veins are associated with other features indicative of paleo-ground-water discharge. Further evidence for a slow rate of water-table decline near Ash Meadows is presented by Jones (1982) in which he reports the water table beneath an alluvial fan at the Nevada Test Site has been within 160 ft (reported as 50 meters) of the present level through most of Quaternary time. In contrast, the water table in some of the northern valleys and, in particular, the Great Salt Lake Desert must have declined at least several hundred feet over the past 10,000–20,000 years as ancestral Lake Bonneville shrank to the present level of the Great Salt Lake.

The assumption of steady-state conditions cannot be validated. However, the lack of long-term trends in measured water levels in basin fill (in areas not influenced by pumping) suggests that a dynamic equilibrium or steady state exists (at least prior to pumping) in many of the basins. Because estimates of hydraulic properties and the length of flow through the consolidated rocks are generally unknown, deeper flow through carbonate aquifers may not be in equilibrium throughout the province. If deeper flow is not in equilibrium, then present-day discharge may be responding to residual water levels related to recharge from previous wet periods, such as the last glacial epoch, and the analysis of flow presented herein may not represent actual flow everywhere.

Transmissivity in the province is assumed heterogeneous because horizontal hydraulic conductivities can change abruptly as a result of changes in lithology. Heterogeneity is simulated by varying the transmissivity among the model cells. Transmissivity within a model cell, however, is assumed homogeneous and isotropic, and is assumed to represent an average for the cell. Abrupt changes in transmissivities within a model cell are not simulated in the model. Consequently, the model is designed to simulate flow across regional changes in transmissivity.

The assumption of isotropy within a model cell is reasonable for cells corresponding to basin fill, but may be unreasonable for cells corresponding to consolidated rocks. Where flow is through fractures, the fractures may have a preferred orientation that could produce a greater transmissivity in one direction. However, anisotropic conditions may not be the same throughout the province because the orientation of fractures in consolidated rocks is not the same everywhere. Even though some types of consolidated rock may be anisotropic, there is no compelling reason to assume a regional anistropy for the entire modeled area, and the model is not capable of simulating anistropy in individual cells. Furthermore, data is lacking to calibrate a model whereby every cell corresponding to consolidated rocks could have a greater value of transmissivity in one direction.

MODEL DEVELOPMENT

A three-dimensional finite-difference ground-water flow model developed by McDonald and Harbaugh (1988) was used for the computer simulations. The model uses the basic partial differential equation for ground-water flow in an anisotropic, heterogeneous porous medium with a constant water density:

$$\frac{\partial}{\partial x}\left(Kxx\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(Kyy\frac{\partial h}{\partial y}\right)+\frac{\partial}{\partial z}\left(Kzz\frac{\partial h}{\partial z}\right)-W=S_s\frac{\partial h}{\partial t} \quad (1)$$

⇐

FIGURE 8.—Distribution of estimated ground-water withdrawals by hydrographic areas for 1975. Hydrographic areas from Harrill and others (1988); estimates of ground-water withdrawals for Utah from Sumison and others (1976); estimates for Nevada from Bedinger and others (1984).

Total simulated discharge (as evapotranspiration and leakage to head-dependent flow boundaries) along the Humboldt River is 52,000 acre-ft/yr (fig. 35). This quantity represents only a fraction of the total estimated evapotranspiration and streamflow in the Humboldt River valley above Palisade (Eakin and Lamke, 1966, p. 59, 60). Simulation of regional ground-water flow with the model did not account for the local circulation of water adjacent to the Humboldt River; rather, the model is designed to assess the potential for regional flow from distant sources to regional discharge areas. In the upper Humboldt River region, the quantity of simulated deep flow (flow through the lower model layer) to the Humboldt River is small (a few thousand acre-ft/yr) compared to local flow between the river and its alluvium.

POTENTIAL USES OF MODEL

The ground-water flow model of the carbonate-rock province is unlike most models in that the extent of aquifers and their hydraulic properties are generally unknown in the province; thus, the model greatly simplifies flow through a complex geologic region. Simulation results are based on assuming recharge to the province is known with the distribution of transmissivities simulated to match the general distribution of water levels and estimates of discharge. However, water levels in consolidated rocks are generally unknown, and estimates of recharge and discharge are known only approximately. Consequently, other, equally valid distributions of transmissivities may be found that permit the model to be calibrated to the existing water-level data and estimates of recharge and discharge. The model may be best suited for:

- Simulating alternative transmissivity distributions to evaluate potential source areas of regional springs,
- Simulating the effects of differing recharge rates on regional ground-water flow, and
- Simulating the effects of changing location of discharge on regional ground-water flow.

Therefore, the potential uses of the model are limited. The model is not suited to predict accurate water-level declines that would result from pumping ground water in the province. Also, the model is not suited to predict the accurate rate of change in natural discharge caused by pumping, because the model has not been calibrated to any transient simulations.

SUMMARY AND CONCLUSIONS

This report describes the results of a computer-model-based analysis of regional ground-water flow in the eastern Great Basin, a 100,000-mi^2 area that lies mostly in eastern Nevada and western Utah, with small parts in northwestern Arizona, eastern California, and southern Idaho. The original version of this report, published in 1991, presented results that subsequently proved to be adversely affected by a transpositional error in the computer data files that define the model-cell dimensions. This error produced an unintended regional anisotropy in hydraulic conductivity. The results reported herein constitute a reanalysis of regional flow after the transposition was corrected and the computer model recalibrated.

Ground-water flow in the eastern Great Basin has been evaluated as part of the U.S. Geological Survey's Great Basin Regional Aquifer-System Analysis. The area is referred to as the carbonate-rock province because, during the Paleozoic era, thick sequences of limestone and dolomite were deposited in a shallow sea that inundated the area. Since then, many episodes of sediment deposition, volcanic activity, erosion, and tectonic deformation by both compressional and extensional forces have altered the extent and thickness of the carbonate rocks. The present-day physiography, which is characterized by north- to northeast-trending mountain ranges separated by intervening valleys that are partly filled with sedimentary deposits eroded from the adjacent mountains, is the result of normal faulting caused by extension that began about 20 million years ago. Relief between the block-faulted mountains and the adjacent valley floors ranges from 1,000 ft to more than 7,000 ft.

Shallow ground-water reservoirs in the basin fill supply most of the current (1992) pumpage from wells in this geologically complex terrain. Aquifers in the underlying carbonate rocks are largely undeveloped; regionally, however, these aquifers are important because they provide an avenue for interbasin ground-water flow. The source of ground water in the province is precipitation, most of which falls in the higher mountain ranges. Ground-water discharge is mostly by evapotranspiration in the low parts of the many valleys. Some ground water also discharges from small, local springs. Such springs are fed by recharge that originates nearby. In contrast, ground water discharging at larger, regional springs

In keeping with the conceptual nature of the model, the simulation provides information about the probable areas that may be affected, the general magnitude of possible water-level declines or other effects, and the general period of time over which changes may be expected to occur. Prediction of specific, detailed water-level changes throughout the area would require that effects of the proposed pumping be superimposed on the effects of existing and other anticipated future pumping. That was beyond the scope of this analysis.

The second assumption was that storage values used for transient simulations for the upper layer were based on the predominant aquifer material in each cell, determined from surficial maps. This distribution may not be totally correct because the material may be different at depth in the zone of saturation. Storage coefficients in the upper layer also assume dewatering of the sediments.

Rock and deposit types were divided into three categories—basin-fill materials, carbonate rocks, and other consolidated rocks. Distribution of these units is shown by Prudic and others (1993, fig. 15). Average values for storage coefficients in layer one were assigned to each of these materials. For basin-fill material, a value of 0.1 was assigned on the basis of average values of specific yield used in U.S. Geological Survey reconnaissance evaluations of ground-water resources in most basins of the study area. For carbonate rocks, a value of 0.05 was assigned on the basis of an average porosity value of 0.047 determined from geophysical logs of five wells in the Coyote Spring Valley area (Berger, 1992, p. 18). For other rocks, a value of 0.01 was assigned on the basis of a range of values for fractured rocks given by Snow (1979, table 1).

The storage coefficient for the lower layer was estimated on the basis of the probable average porosity of the rocks present (0.01 to 0.05), the effective thickness of aquifer material (probably between 5,000 and 10,000 ft), the bulk modulus of elasticity of water (3×10^5 lb/in^2), and the bulk modulus of elasticity of the solid skeleton of the aquifer (for limestone, about 4.8×10^6 to 5.4×10^6 lb/in^2; Krynine and Judd, 1957, table 2.5). The following equation from Lohman (1972, p. 9) was used to estimate the coefficients:

$$S = \theta \gamma b\left(\frac{1}{E_w} + \frac{C}{\theta E_s}\right), \qquad (1)$$

where S is storage coefficient (dimensionless);
θ is porosity, as a decimal fraction;
γ is specific weight per unit, 62.4 lb/ft^3 ÷ 144 in^2/ft^2 = 0.434 (lb/in^2)/ft;
b is thickness, in feet;
E_w is bulk modulus of elasticity of water;
C is a dimensionless ratio, which may be considered unity in an uncemented granular material; in a solid aquifer, such as limestone with tubular solution channels, C is apparently equal to porosity; and
E_s is bulk modulus of elasticity of the solid skeleton of an aquifer.

Estimates of storage values based on the above numbers ranged from 7.6×10^{-5} to 1.2×10^{-3}. For purposes of this report, the storage coefficient for the lower layer was set at the midrange of these values, 6×10^{-4}, for the entire layer. The data set for storage values used in the model is listed in appendix 1.

The third major assumption used in the model is from the previous steady-state model and concerns the lower layer. The individual basin-fill aquifers underlying the various ground-water basins can be adequately described in the upper layer as a series of high-transmissivity zones (the basin-fill valleys) separated from each other by low-transmissivity zones (the intervening mountain ranges). The lower layer represents the distribution of carbonate-rock aquifers in the system in a limited way that may affect the calculated drawdowns in that layer.

The fourth and final assumption was that all input values used in the conceptual steady-state model remain constant during the transient simulations. No changes were made to transmissivity, leakance, recharge, or the other input data sets described by Prudic and others (1993) and Schaefer (1993).

RESULTS OF SIMULATIONS

Simulation of Conditions Prior to Proposed Pumping

The steady-state conditions simulated by Prudic and others (1993) represent a conceptualization of ground-water flow in the carbonate-rock province of the Great Basin before ground-water pumping within the province commenced. Figure 2 shows the general distribution of simulated steady-state heads (water

Sensitivity of Model Results to Storage Values

To test the sensitivity of the model to input values, several additional simulations were made by varying the values of aquifer storage. Transmissivity values from the original model (Prudic and others, 1993) were not tested during this study. Previous sensitivity analyses were deemed sufficient, and although transmissivity values may be more variable than storage values in a given geologic unit, storage values may be more responsible for long-term effects in the simulation.

The storage values for both the basin-fill and carbonate aquifers are not well known, and may cause the results of the model to vary significantly. Changing the storage values of the upper layer by a range of ± 50 percent, and changing the storage values of the lower layer to the two endpoints of 7.6×10^{-5} and 1.2×10^{-3}, were assumed to give a reasonable test of how results might change. The model was rerun using these adjusted storage values, and figures 16 through 18 show how various key budget components change throughout the simulation, compared to the results obtained using the original storage values.

Figure 16 shows how regional spring discharge varies in response to changing storage-coefficient estimates. In general, storage-coefficient values for the upper layer have little effect on simulated spring discharge. At any given time, the smaller storage coefficients cause less discharge from the drains, whereas larger storage values for the upper layer allow for more

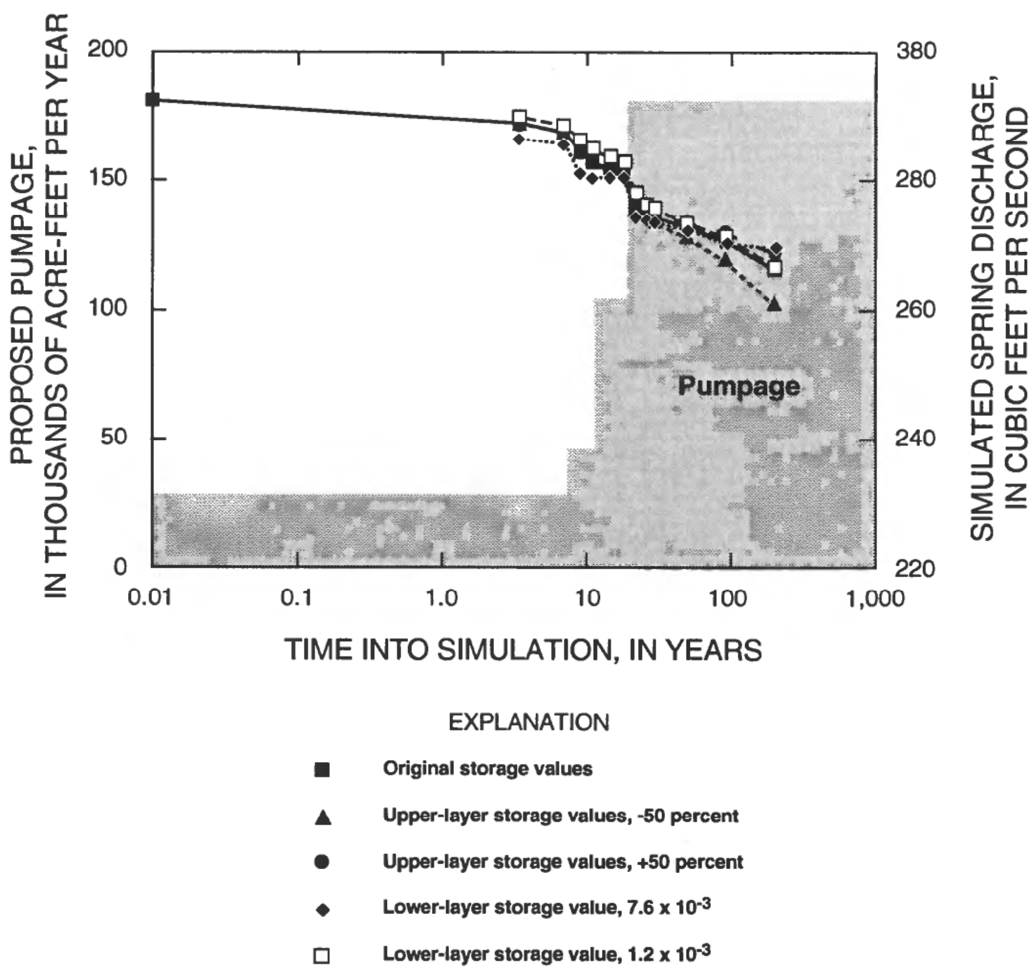

Figure 16. Changes in total model-simulated spring discharge with selected storage values and changing pumpage, east-central and southern Nevada. (All simulated spring discharge totals for the several values converged to a simulated total spring discharge of 234 cubic feet per second in the steady-state simulation.)

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Figure 22 shows the simulated drawdowns at the selected cell in Northern Railroad Valley in both layers. The upper layer demonstrates a difference in drawdowns of about 40 ft after about 100 years into the simulation. The lower layer shows a difference of about 50 ft after the same time period.

Figure 23 shows the simulated drawdowns at the selected cell near Moapa for both layers. The upper layer shows a difference of about 0.02 ft in the simulated drawdowns and the lower layer shows about a 2-ft difference, after about 100 years into the simulation.

Overall, the model appears to be relatively insensitive to variations in aquifer storage coefficients. Changes in these values elicit only minor changes in evapotranspiration, spring discharge, movement of ground water out of storage, and variations in simulated drawdowns. Changes in pumping—location and rate—have a greater influence on model results.

Ultimate Source of Pumped Water

The simulation of pumping ground water in east-central and southern Nevada illustrates several concepts discussed by Theis (1940). The ultimate source of pumped ground water in an aquifer system is an increase in recharge, a decrease of natural discharge, or removal of ground water from storage. As was stated succinctly by Theis (p. 280), "All water discharged by wells is balanced by a loss of water somewhere."

The boundaries for this simulation do not allow additional water to be made available to the ground-water system of the Great Basin; pumpage will not increase precipitation and, hence, recharge. If wells were placed near some of the bounding surface-water bodies, some additional water would recharge the local ground water to make up any deficit caused by pumping. But throughout the study area, additional water from these sources is not available.

The previous discussion of how pumping in the study area affects ET and spring discharge suggests that much of the ground water pumped would be derived from these sources. Since ET is dependent on shallow water levels to support vegetation, once water levels decline sufficiently, ET would cease. Simulated spring discharge is also affected by the proposed pumping in the sense that ground-water flow to the spring is intercepted by the expanded cones of depression of the wells.

The last source of water available to the proposed pumping is from ground water in storage. Figure 24 illustrates the change in various ground-water model budget components as the simulation progresses. Also shown is a series of figures illustrating the source of water pumped in the simulation. Early in the simulation, the major source of pumped water is from ground-water storage (83 percent at 9 years into the simulation). As the simulation progresses, less and less water is removed from storage and the remainder of the pumped water comes from reduction in ET and spring discharge. The final stage of this progression is the steady-state simulation, where none of the pumped water is from storage, 77 percent is from what had been used by ET, and 23 percent is from reduction of springflow. This represents a simulated equilibrium within the ground-water system.

Limitations and Uses of the Model

Simulations of the proposed pumpage show that many aspects of the ground-water systems in the Great Basin may be affected. The simulations were based on a computer model of regional ground-water flow that greatly simplifies the complex distribution of geology and, consequently, the hydraulic properties of many of the rocks in the Great Basin. As the authors of the original model state, "Simulation results are based on assuming recharge to the province is known with the distribution of transmissivities simulated to match the general distribution of water levels and estimates of discharge. However, water levels in consolidated rocks are generally unknown and estimates of recharge and discharge are known only approximately" (Prudic and others, 1993, p. 91).

The adequacy of the model in simulating the effects of the proposed pumping will remain untested until actual pumping stresses have been in place long enough to cause measurable effects within the system. This would allow for calibration of transient simulations that was not possible with the previous model.