

Prepared in cooperation with the Bureau of Reclamation

Hydrogeologic Framework and Occurrence, Movement, and Chemical Characterization of Groundwater in Dixie Valley, West-Central Nevada

Scientific Investigations Report 2014–5152



Top: Dixie Valley Wash (dry), near the Stingaree and Dixie Valley border. Pictured looking downstream or to the north toward Dixie Valley with Chalk Mountain to the east. Photograph taken by Jena Huntington, U.S. Geological Survey, July 19, 2012.

Bottom: Dixie Valley playa depicting cracking and surface salts, viewed to the west toward the Stillwater Range. Photograph taken by Jena Huntington, U.S. Geological Survey, February 24, 2009.

Cover: Dixie Valley, as viewed southwest toward Stillwater Range. Dixie Valley Geothermal Plant in distance on left. Photograph taken by C. Amanda Garcia, U.S. Geological Survey, June 16, 2010.

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By Jena M. Huntington, C. Amanda Garcia, and Michael R. Rosen

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U.S. Geological Survey

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Contents

Abstract.....	1
Introduction.....	2
Purpose and Scope	4
Description of Study Area	4
Geologic Framework	6
Hydrogeologic Units.....	6
Carbonate and Clastic Sedimentary Rocks	6
Crystalline Rocks.....	6
Volcanic Rocks.....	9
Tertiary Sedimentary Rocks and Sediments	9
Basin-Fill Deposits	9
Playa Deposits.....	9
Structural Features.....	10
Hydrogeology.....	18
Aquifer Properties.....	18
Occurrence and Movement of Groundwater in Basin-Fill Deposits	18
Water-Level Change.....	24
Subsurface Flow	24
Chemical Composition of the Basin-Fill Aquifer	27
Quality Assurance	29
Chemical Signature	29
Stable Isotopes	31
Chemical Quality of Groundwater	33
Seasonal Changes in Groundwater Chemistry	37
Connection between Basin-Fill and Geothermal Aquifers	37
Geothermal and Non-Geothermal Indicators	40
Mixing of Geothermal and Basin-Fill Groundwater	40
Chemical Indicators	40
Geothermometry	43
Summary.....	49
Acknowledgments.....	51
References Cited.....	51
Appendix A. Stable Isotopic Data for Bulk Precipitation Sites Within Dixie Valley Study Area, May 2009–November 2011	57
Appendix B. Determination of Lithium Using Inductively Coupled Plasma-Optical Emission Spectrometry.....	58

Plate

[Available online only at <http://pubs.usgs.gov/sir/2014/5152/>.]

1. Potentiometric water-level elevation at selected wells and potentiometric contours of potentiometric water-level elevation, Dixie Valley, west-central Nevada, autumn 2009.

Figures

1. Map showing the flow system, including hydrographic area boundaries, and playa location, Dixie Valley, west-central Nevada3
2. Map showing hydrogeologic units, depth to bedrock, and known playa sediment depths, Dixie Valley, west-central Nevada7
3. Conceptual cross-sectional diagram showing inter-fingering of clay-dominated playa and basin-fill sediments, and photograph showing typical playa-deposits during well drilling in Dixie Valley, west-central Nevada10
4. Map showing data collection sites in Dixie Valley, west-central Nevada, 2009–1111
5. Map showing estimates of basin-fill transmissivity from single-well aquifer tests or specific-capacity calculations and water-temperature distributions in springs, Dixie Valley, west-central Nevada12
6. Graphs showing water-level change as indicated by intermittent, selected water-level measurements from eight wells in the northernmost basin, north of playa near Dixie Valley Geothermal Plant, just south of playa, and in southernmost basin, Dixie Valley, west-central Nevada, 1955–201126
7. Map showing locations of sites sampled for water quality and measured field groundwater temperature, Dixie Valley, west-central Nevada, 1959–201128
8. Trilinear diagram showing general water-quality conditions in northern and southern areas of Dixie Valley, west-central Nevada30
9. Graphs showing isotopic signature of groundwater, springs, and precipitation at selected sites in Dixie Valley, west-central Nevada, 2009–1132
10. Map showing fluoride concentrations in groundwater from sites and maximum contaminant level exceedances in the basin-fill aquifer, Dixie Valley, west-central Nevada35
11. Map showing total dissolved solids concentrations in groundwater from sites and secondary maximum contaminant level exceedance distributions in the basin-fill aquifer, Dixie Valley, west-central Nevada36
12. Map showing magnesium concentrations in groundwater from sites in Dixie Valley, west-central Nevada41
13. Graph showing relation between the molar ratio of magnesium to lithium and average geothermal silicon in geothermal groundwater samples, Dixie Valley, west-central Nevada42

Figures—Continued

14. Graph showing mixing between water high in silica (SiO ₂) and lithium (Li) with water high in magnesium and low in SiO ₂ and Li, and water temperature, Dixie Valley, west-central Nevada	43
15. Graph showing relation between boron and lithium concentrations, and water temperature in basin-fill groundwater samples collected in this study and in historical basin-fill and geothermal groundwater samples, Dixie Valley, west-central Nevada	44
16. Map showing lithium concentrations in groundwater from sites in Dixie Valley, west-central Nevada	45
17. Map showing locations of magnesium-lithium geothermometers and temperatures recorded in Dixie Valley, west-central Nevada	48

Tables

1. Hydrographic areas of the Dixie Valley flow system, west-central Nevada	5
2. Hydrogeologic unit characteristics, Dixie Valley, west-central Nevada	8
3. Well information, static water levels, and types of data collected in the Dixie Valley flow system, west-central Nevada	13
4. Basin-fill transmissivity estimated from single-well and multiple-well aquifer tests, specific-capacity estimates, slug tests, and well characteristics, Dixie Valley, west-central Nevada	16
5. Locations, elevations, and temperatures in springs and Quaternary faults in the Dixie Valley flow system, west-central Nevada	17
6. Water-quality data from sites sampled in Dixie Valley, west-central Nevada, 2009–11	20
7. Wells with seven or more water-level measurements, Dixie Valley, west-central Nevada, 1950–2011	25
8. Annual subsurface inflow estimates from adjacent basins through basin-fill deposits to Dixie Valley, west-central Nevada	25
9. Summary of drinking-water standards and exceedances for groundwater sites in the basin-fill aquifer, Dixie Valley, west-central Nevada	34
10. Selected chemical constituent concentration and percentage of change for samples collected in Dixie Valley, west-central Nevada, October 2009 and May 2010	38
11. Estimated reservoir temperatures based on magnesium-lithium geothermometer compared to sodium-potassium-calcium geothermometer and measured field groundwater temperatures, Dixie Valley, west-central Nevada	46

Conversion Factors, Datums, Township and Range, and Abbreviations and Acronyms

Conversion Factors

Inch/Pound to SI

Multiply	By	To obtain
Length		
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
Area		
acre	4,047	square meter (m ²)
square mile (mi ²)	2.590	square kilometer (km ²)
Volume		
acre-foot (acre-ft)	1,233	cubic meter (m ³)
Flow rate		
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year (m ³ /yr)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
gallon per day (gal/d)	0.003785	cubic meter per day (m ³ /d)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
Energy		
megawatt hour (MWh)	3,600,000,000	joule (J)
Hydraulic conductivity		
foot per day (ft/d)	0.3048	meter per day (m/d)
Hydraulic gradient		
foot per foot (ft/ft)	0.3048	meter per meter (m/m)
Transmissivity*		
foot squared per day (ft ² /d)	0.09290	meter squared per day (m ² /d)

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:

$$^{\circ}\text{F}=(1.8\times^{\circ}\text{C})+32.$$

Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft³/d)/ft²]ft. In this report, the mathematically reduced form, foot squared per day (ft²/d), is used for convenience.

Conversion Factors, Datums, Township and Range, and Abbreviations and Acronyms

Datums

Vertical coordinate information is referenced to North American Vertical Datum of 1988 (NAVD 88).

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

Elevation, as used in this report, refers to distance above the vertical datum.

Township and Range

A township is an approximately 6-mile square area of land containing 36 sections. The Public Land Survey System (PLSS) numbers townships in horizontal rows and ranges in vertical columns so that each 6 square-mile area has a unique township-range number.

Abbreviations and Acronyms

bls	below land surface
ICP-OES	Inductively Coupled Plasma-Optical Emission Spectrometry
LMWL	local meteoric water line
MCL	maximum contaminant level
PRISM	Parameter-elevation Regressions on Independent Slopes Model
r^2	coefficient of determination
Reclamation	Bureau of Reclamation
SMCL	secondary maximum contaminant level
TDS	total dissolved solids
USGS	U.S. Geological Survey

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Abstract

Dixie Valley, a primarily undeveloped basin in west-central Nevada, is being considered for groundwater exportation. Proposed pumping would occur from the basin-fill aquifer. In response to proposed exportation, the U.S. Geological Survey, in cooperation with the Bureau of Reclamation and Churchill County, conducted a study to improve the understanding of groundwater resources in Dixie Valley. The objective of this report is to characterize the hydrogeologic framework, the occurrence and movement of groundwater, the general water quality of the basin-fill aquifer, and the potential mixing between basin-fill and geothermal aquifers in Dixie Valley. Various types of geologic, hydrologic, and geochemical data were compiled from previous studies and collected in support of this study. Hydrogeologic units in Dixie Valley were defined to characterize rocks and sediments with similar lithologies and hydraulic properties influencing groundwater flow. Hydraulic properties of the basin-fill deposits were characterized by transmissivity estimated from aquifer tests and specific-capacity tests. Groundwater-level measurements and hydrogeologic-unit data were combined to create a potentiometric surface map and to characterize groundwater occurrence and movement. Subsurface inflow from adjacent valleys into Dixie Valley through the basin-fill aquifer was evaluated using hydraulic gradients and Darcy flux computations. The chemical signature and groundwater quality of the Dixie Valley basin-fill aquifer, and potential mixing between basin-fill and geothermal aquifers, were evaluated using chemical data collected from wells and springs during the current study and from previous investigations.

Dixie Valley is the terminus of the Dixie Valley flow system, which includes Pleasant, Jersey, Fairview, Stingaree, Cowkick, and Eastgate Valleys. The freshwater aquifer in the study area is composed of unconsolidated basin-fill deposits of Quaternary age. The basin-fill hydrogeologic

unit can be several orders of magnitude more transmissive than surrounding and underlying consolidated rocks and Dixie Valley playa deposits. Transmissivity estimates in the basin fill throughout Dixie Valley ranged from 30 to 45,500 feet squared per day; however, a single transmissivity value of 0.1 foot squared per day was estimated for playa deposits.

Groundwater generally flows from the mountain range uplands toward the central valley lowlands and eventually discharges near the playa edge. Potentiometric contours east and west of the playa indicate that groundwater is moving eastward from the Stillwater Range and westward from the Clan Alpine Mountains toward the playa. Similarly, groundwater flows from the southern and northern basin boundaries toward the basin center. Subsurface groundwater flow likely enters Dixie Valley from Fairview and Stingaree Valleys in the south and from Jersey and Pleasant Valleys in the north, but groundwater connections through basin-fill deposits were present only across the Fairview and Jersey Valley divides. Annual subsurface inflow from Fairview and Jersey Valleys ranges from 700 to 1,300 acre-feet per year and from 1,800 to 2,300 acre-feet per year, respectively. Groundwater flow between Dixie, Stingaree, and Pleasant Valleys could occur through less transmissive consolidated rocks, but only flow through basin fill was estimated in this study.

Groundwater in the playa is distinct from the freshwater, basin-fill aquifer. Groundwater mixing between basin-fill and playa groundwater systems is physically limited by transmissivity contrasts of about four orders of magnitude. Total dissolved solids in playa deposit groundwater are nearly 440 times greater than total dissolved solids in the basin-fill groundwater. These distinctive physical and chemical flow restrictions indicate that groundwater interaction between the basin fill and playa sediments was minimal during this study period (water years 2009–11).

Groundwater in Dixie Valley generally can be characterized as a sodium bicarbonate type, with greater proportions of chloride north of the Dixie Valley playa, and greater proportions of sulfate south of the playa. Analysis of major ion water chemistry data sampled during the study period indicates that groundwater north and south of Township 22N differ chemically. Dixie Valley groundwater quality is marginal when compared with national primary and secondary drinking-water standards. Arsenic and fluoride concentrations exceed primary drinking water standards, and total dissolved solids and manganese concentrations exceed secondary drinking water standards in samples collected during this study. High concentrations of boron and tungsten also were observed.

Chemical comparisons between basin-fill and geothermal aquifer water indicate that most basin-fill groundwater sampled could contain 10–20 percent geothermal water. Geothermal indicators such as high temperature, lithium, boron, chloride, and silica suggest that mixing occurs in many wells that tap the basin-fill aquifer, particularly on the north, south, and west sides of the basin. Magnesium-lithium geothermometers indicate that some basin-fill aquifer water sampled for the current study likely originates from water that was heated above background mountain-block recharge temperatures (between 3 and 15 degrees Celsius), highlighting the influence of mixing with warm water that was possibly derived from geothermal sources.

Introduction

Quantification of water resources is essential in the arid Southwest, where demand for water is increasing because of population growth. One of the driest states in the United States, Nevada, increased in population by 35 percent between 2000 and 2010 (U.S. Census Bureau, 2011). Most of this population increase occurred in urban areas. Finite water resources in urban areas have led to the implementation of large-scale water importation projects, which commonly include applications for additional, new water appropriations. In order to sustain continued growth in the Fallon urban area of the Carson Desert, Churchill County is investigating additional water resources. Dixie Valley, primarily an undeveloped basin neighboring the Carson Desert to the east (fig. 1), is considered a potential resource for water importation to the Carson Desert (Public Law 110-161, Section 208). Proposed groundwater pumping would be from the fresh basin-fill aquifer.

A large amount of research and exploration has been done in Dixie Valley since the 1960s. The first hydrologic report for the valley and basin-fill aquifers (Cohen and Everett, 1963) included water-budget estimates (recharge, groundwater evapotranspiration, and subsurface inflow), a limited groundwater quality evaluation, and a map showing water-level elevations. This reconnaissance-level report described

multiple groundwater systems in Dixie Valley including a freshwater basin-fill system, a brine water playa-sediment system, and a deep, variable-quality geothermal system with elevated temperatures. In the 1990s, Harrill and Hines (1995) updated the 1963 basin-fill groundwater budget by incorporating groundwater evapotranspiration estimates based on mapped distributions of phreatophyte types and densities, and precipitation across the valley.

Much research completed in Dixie Valley has focused on describing geologic structure. A compilation report by Thompson and others (1967) documented the geologic structure in Dixie Valley using seismic refraction, aeromagnetic measurements, and mapping or interpretation of aerial photographs. In the early 1980s, Schaefer (1983) did a gravity survey across the valley to evaluate the depth to bedrock or base of basin-fill sediments. Mankhemthong and others (2008) gathered detailed gravity data in southern Dixie Valley and made interpretations of bedrock thicknesses.

The developed geothermal aquifer in northern Dixie Valley, near the Stillwater Range (fig. 1), is one of the most studied geothermal resources in the United States (Blackwell and others, 2007). Many studies have concentrated on describing the geologic structure of the geothermal field. Wilt and Goldstein (1983) did an aero-electromagnetic survey mapping areas of high-resistivity (low electrical conductivity) rock zones to help locate geothermal waters. Barton and others (1997) used flow, temperature, and imaging logs from boreholes extending into the geothermal reservoir to locate hydraulically conductive fractures in bedrock adjacent to the Stillwater Fault located along the eastern face of the Stillwater Range. Additional work was done by Blackwell and others (1999) and Smith and others (2001) to characterize the complex faulting associated with the main Stillwater Fault, including gravity surveys, seismic lines, field mapping, and aerial photo analyses. In 2002, another aero-electromagnetic survey conducted over a larger (than the geothermal field) area of Dixie Valley (U.S. Geological Survey, and Pearson, deRidder, and Johnson, Inc., 2002) was used to locate shallow faults not visible at land surface (Grauch, 2002). Caine and others (2010) further investigated geothermal fluid flow using field mapping and rock characterization along fault lines. Many other studies have focused on the Dixie Valley geothermal field, some of which are listed on the Geothermal Resources Council Web site, at <https://www.geothermal-library.org/>.

The Dixie Valley Geothermal Power Plant, in northwestern Dixie Valley (fig. 1), produces the largest amount of geothermal energy from a single plant in the State of Nevada (Lowell Price, Nevada Division of Minerals, oral commun., 2012). Pumping from the geothermal aquifer for power production began in 1985, and pressure augmentation pumping from the basin-fill aquifer for reinjection above the geothermal aquifer began in 1997; pumping from both aquifers continues today (2014).

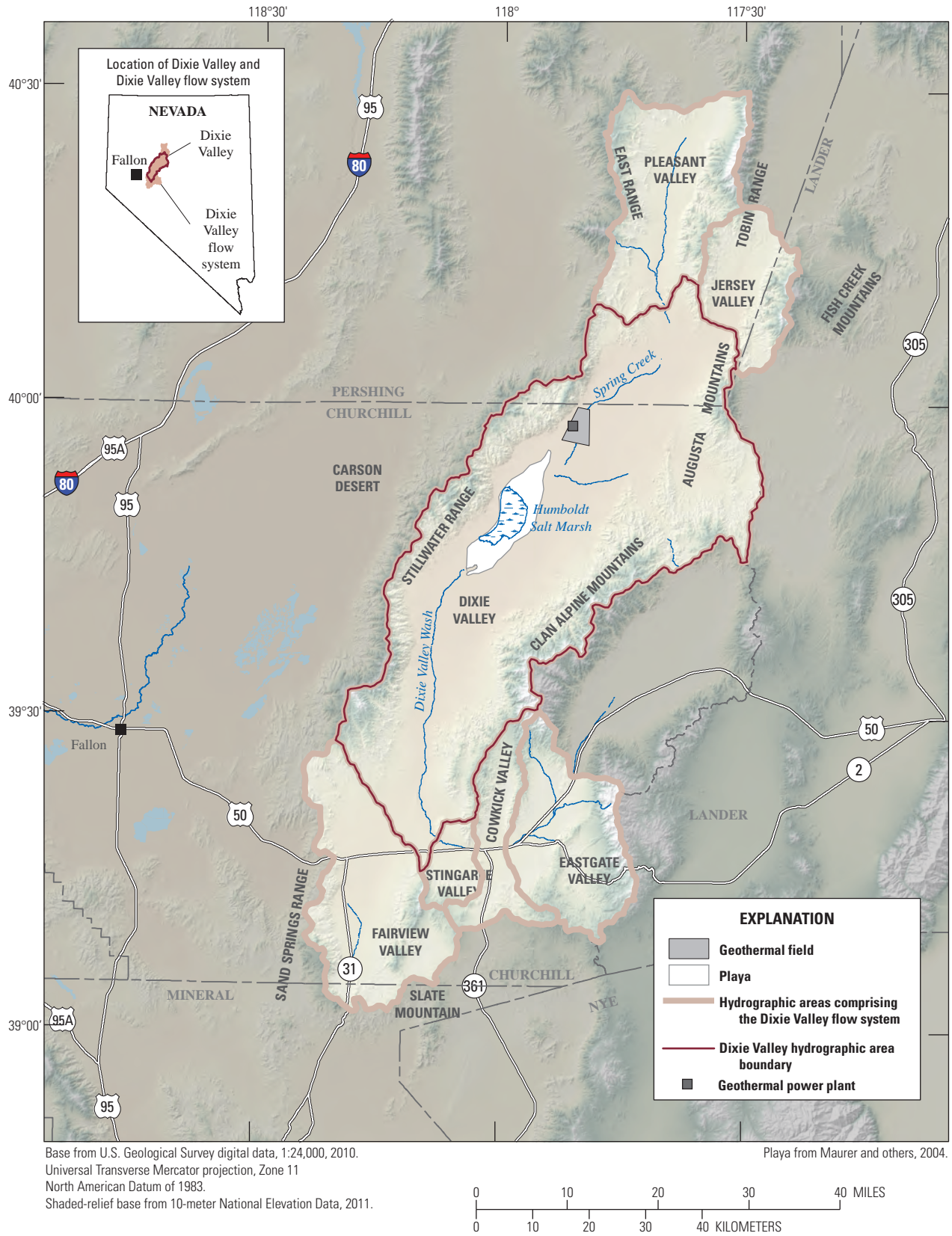


Figure 1. Flow system, including hydrographic area boundaries, and playa location, Dixie Valley, west-central Nevada.

During water years¹ 2010–11, average annual geothermal groundwater withdrawals totaled about 21,400 acre-ft (23,600 and 19,200 acre-ft in 2010 and 2011, respectively), with an average water temperature of about 160 °C. About 487,000 megawatt hours of energy were produced annually from this water. About 15,000 acre-ft were re-injected annually, resulting in an average annual consumptive use of geothermal water of about 6,600 acre-ft (about 9,600 and 3,600 acre-ft in 2010 and 2011, respectively) (Lowell Price, Nevada Division of Minerals, written commun., 2012). Little is known about potential interactions between the different geothermal systems and the basin-fill aquifer, although one study (Bruton and others, 1997) indicated that there could be 15–25 percent geothermal water in the basin-fill aquifer, and another study (Nimz and others, 1999) indicated that geothermal water might have been derived from local recharge. Several investigations have sampled for water chemistry and hydrogen and oxygen isotopes from basin-fill aquifers, springs, and geothermal wells (Cohen and Everett, 1963; Garside, 1994; Nimz and others, 1999; Stamates, 2001; Goff and others, 2002; Lahontan GeoScience, Inc., 2004) to characterize potability and the hydrologic relations between basin-fill and geothermal aquifers. A more comprehensive investigation of the basin-fill aquifer and potential interactions with geothermal systems was needed to understand how basin-fill groundwater exportation might affect geothermal resources in Dixie Valley.

With competing water exportation and geothermal development interests, an improved understanding of the groundwater resources is necessary before additional groundwater development in Dixie Valley can reasonably be considered. County and State water-resource managers will need this information to make informed decisions regarding the proposed inter-basin water transfer. To address these needs and concerns, the U.S. Geological Survey (USGS), in cooperation with the Bureau of Reclamation (Reclamation) and Churchill County, evaluated the groundwater resources of Dixie Valley in west-central Nevada. A companion USGS report by Garcia and others (2014) describes a refined estimate of groundwater discharge from evapotranspiration in Dixie Valley.

Purpose and Scope

This report characterizes the occurrence, movement, and chemistry of the basin-fill aquifer by evaluating the hydrogeologic features controlling groundwater resources

¹A “water year” is the 12-month period from October 1, for any given year, through September 30 of the following year and is designated by the calendar year in which it ends.

²Formal hydrographic areas in Nevada were delineated systematically by the U.S. Geological Survey and Nevada Division of Water Resources in the late 1960s for scientific and administrative purposes (Cardinalli and others, 1968; Rush, 1968). The official hydrographic-area names, numbers, and geographic boundaries continue to be used in U.S. Geological Survey scientific reports and Nevada Division of Water Resources administrative activities.

in Dixie Valley. Groundwater resources were evaluated by delineating the hydrogeologic framework, estimating aquifer properties, describing groundwater-flow directions and gradients in the valley, characterizing the general quality of groundwater from the principal basin-fill aquifer, and evaluating potential interactions between the basin-fill and geothermal aquifers. This report provides a reassessment of the groundwater flow regime. The hydrogeologic framework was based on geologic and hydrogeologic studies completed during the past 50 years. Aquifer properties were estimated from a series of aquifer and specific capacity tests. Groundwater-flow directions and gradients were derived from water-level elevation measurements made primarily during this study, whereas changes in water-level elevation were evaluated from data collected since the 1950s. The chemical quality of the basin-fill aquifers was characterized using water-quality data from previously published datasets (samples collected from 1959 to 2004) and sites sampled for this study during water years 2009–11.

Description of Study Area

The Dixie Valley flow system covers an area of about 2,380 mi² (Nevada Division of Water Resources, 2012) in west-central Nevada, mainly is in Churchill County (with smaller parts in Pershing, Lander, and Mineral Counties), and consists of seven hydrographic areas²: Pleasant Valley, Jersey Valley, Fairview Valley, Dixie Valley, Stingaree Valley, Cowkick Valley, and Eastgate Valley ([fig. 1](#); [table 1](#)). Dixie Valley is the terminus of the basin-fill groundwater flow system and the focus of this report. Surface water in six of the hydrographic areas (except Fairview Valley) drains into Dixie Valley. Fairview Valley is drained internally or is a closed basin, meaning no surface water leaves the valley. Groundwater flow from all hydrographic areas historically was thought to drain into Dixie Valley through basin-fill sediments (Cohen and Everett, 1963, p. 3).

Dixie Valley is a northeast-trending structural trough (Schaefer, 1983) covering an area of 1,303 mi² (Cardinalli and others, 1968; Rush, 1968; [table 1](#)). The valley is bordered on the west and northwest by the Stillwater Range, and on the east and southeast by the Augusta and Clan Alpine Mountains ([fig. 1](#)). The southern end of Dixie Valley is bounded by a low range of hills between the Sand Springs Range and Slate Mountain, which form the boundary with Fairview Valley. An alluvial divide between the East and Tobin Ranges creates the northwestern boundary with Pleasant Valley, and a second alluvial divide connecting the Tobin Range to the Augusta Mountains forms the northeastern boundary with Jersey Valley. Each alluvial divide has an elevation of about 4,900 ft (Cohen and Everett, 1963). Land-surface elevations in the study area range from 9,900 ft in the Clan Alpine Mountains to 3,380 ft on the Humboldt Salt Marsh, which is in the center of the Dixie Valley playa. The Dixie Valley playa represents the lowest point in northern Nevada (Cohen and Everett, 1963).

Table 1. Hydrographic areas of the Dixie Valley flow system, west-central Nevada.

[Areas are shown in [figure 1](#) and designated by the U.S. Geological Survey and Nevada Division of Water Resources for scientific and administrative purposes (Cardinalli and others, 1968; Rush, 1968)]

Name	Identification No.	Area	
		(square miles)	(acres)
Dixie Valley	128	1,303	833,920
Pleasant Valley	130	285	182,400
Jersey Valley	132	142	90,880
Fairview Valley	124	285	182,400
Stingaree Valley	125	43	27,520
Cowkick Valley	126	110	70,400
Eastgate Valley	127	216	138,240
Totals		2,384	1,525,760

The climate in the mountains of Dixie Valley is categorized as sub-humid with orographic effects similar to those of the Sierra Nevada Mountains but of a lesser magnitude, whereas climate in the valley lowlands is arid (Cohen and Everett, 1963, p. 5). Therefore, annual precipitation varies substantially with elevation in Dixie Valley and the adjoining ranges. In the Stillwater Range and Clan Alpine Mountains, annual precipitation averaged 13.3 and 14.5 in., respectively during October 2009–September 2011 (Chris Mahannah, Mahannah and Associates, LLC, written commun., 2011). The 30-year average annual precipitation from the Parameter-elevation Regressions on Independent Slopes Model (PRISM) was estimated as 10.6 and 10.7 in., respectively (1981–2010; PRISM Climate Group, 2004). On the playa, annual precipitation averaged 5.2 in. during this study (October 2009–September 2011; Garcia and others, 2014). Precipitation predominantly occurs in the winter as snow in the mountains, and as a mix of snow and rain on the valley floor. Lesser amounts of precipitation fall on the valley floor as rain from isolated thunderstorms in the spring, summer, and autumn. Seasonal temperatures in valley lowlands range from about -10 to 40 °C (Garcia and others, 2014).

Only ephemeral streams fed by spring snowmelt or infrequent, major storms, or intermittent streams fed by springs, are present in Dixie Valley ([fig. 1](#)). All streams flow toward the Dixie Valley playa, although little surface water (less than 500 acre-feet per year [acre-ft/yr]) actually is discharged onto the playa (Interflow Hydrology, Inc. and Mahannah and Associates, LLC, 2013). Most streamflow that originates from the mountain front is lost to infiltration or evapotranspiration before reaching the playa (Cohen and Everett, 1963; Interflow Hydrology, Inc. and Mahannah and Associates, LLC, 2013). Dixie Valley Wash is the predominant stream in the valley that enters Dixie Valley from Stingaree Valley to the south, and flows ephemerally about 40 mi north until it discharges onto the Dixie Valley playa ([fig. 1](#)). Spring Creek flows ephemerally from Pleasant Valley southward into Dixie Valley.

Dixie Valley has many cool and geothermal springs that discharge in the Clan Alpine Mountains and Stillwater Range, along the mountain front, and on the valley floor, several of which are adjacent to the Dixie Valley playa. Current discharge rates and temperatures of valley-floor springs range from less than 1 to 300 gallons per minute (Interflow Hydrology, Inc. and Mahannah and Associates, LLC, 2012a), and from 4 to more than 60 °C. Vegetation surrounding spring discharge areas on the valley floor is composed mostly of thick grasses with some trees and reeds, indicating that most discharge is lost to evapotranspiration.

Groundwater in Dixie Valley is present in several aquifer systems. Groundwater persists primarily within the fresh basin-fill aquifer (depths of as much as nearly 10,000 ft below land surface [bls]) and the saline playa confining unit. Deep groundwater (maximum estimated depth of 20,000 ft bls; Blackwell and others, 2003; McKenna and Blackwell, 2004; Wisian and Blackwell, 2004) has been found in at least 12 geothermal aquifer systems within consolidated rock (Benoit, 2011).

Dixie Valley is sparsely populated, with less than 100 residents (Katy Rossiter, U.S. Census Bureau, written commun., 2011). Consequently, domestic groundwater development from basin-fill deposits is negligible, equating to less than 25 acre-ft/yr based on the Nevada domestic self-supplied, per-capita use of 206 gallons per day (gal/d) (Kenny and others, 2009). Current and historical use of basin-fill groundwater includes livestock, agricultural irrigation, domestic, and geothermal development. Cattle are grazed throughout much of the valley rangeland, and alfalfa is cultivated for seed and feed. Currently irrigated agriculture covers about 1,360 acres of Dixie Valley. However, agriculture historically covered more than three times as much of the current agricultural landscape, much of which was in the historical agricultural settlement (1940–80s) just south of the playa. Many artesian wells were drilled to sustain irrigation and domestic requirements in this settlement area.

Large-scale pumping of the basin-fill aquifer to augment the pressure of the geothermal reservoir began in 1997 (Benoit and others, 2000). The Dixie Valley geothermal power plant in northwestern Dixie Valley pumps an average of 2,100 acre-ft/yr from the basin-fill aquifer surrounding the plant (period of record, 2009–11; Nevada Division of Minerals, written commun., 2012), all of which subsequently is re-injected above the deeper (500–9,700 ft bls) geothermal aquifer. Just north of the geothermal well field, subsidence rates of as much as 4.1 in/yr between 1992 and 1997 (Foxall and Vasco, 2003) and 1.8 in/yr between 2006 and 2008 (John Bell, Nevada Bureau of Mines and Geology, written commun., 2010) were documented in interferometric synthetic aperture radar data. If these rates are applied at 4.1 in/yr from 1992 to 2001 and at 1.8 in/yr from 2001 to 2011, total subsidence in the affected area could be nearly 59 in. Additional geothermal exploration currently is being done in several locations throughout Dixie Valley, as well as in Jersey Valley.

Geologic Framework

Dixie Valley is a complex asymmetric graben bounded by high-angle normal faults. The bedrock basin is composed of subbasins separated by bedrock highs in which basin-fill deposits of Quaternary age have accumulated. The bedrock beneath the basin fill and adjacent mountains is composed of clastic sedimentary and crystalline rocks ranging in age from Triassic to Tertiary and mostly extrusive volcanic rocks of Tertiary age (Willden and Speed, 1974). Hydrogeologic features of Dixie Valley that partly control groundwater flow include lithology and sediments through which flow occurs, lithology and sediment thicknesses and hydraulic properties, and the structural components of the valley.

Hydrogeologic Units

Maurer and others (2004) delineated 12 hydrogeologic units in Dixie Valley based on the 1:500,000-scale geologic map of Nevada (Stewart and Carlson, 1978). In this study, these 12 hydrogeologic units were grouped into 6 hydrogeologic units based on lithology (fig. 2; table 2). Rocks and sediments with similar lithologies generally were assumed to have similar hydraulic properties affecting groundwater flow. The six hydrogeologic units that either transmit or impede groundwater flow, from oldest to youngest, include: (1) carbonate and clastic sedimentary rocks of Triassic-to-Jurassic age; (2) crystalline rocks of Triassic-to-Tertiary age; (3) volcanic rocks of Tertiary age; (4) sedimentary rocks and sediments of Tertiary age; (5) basin-fill deposits of Quaternary age; and (6) playa deposits of Quaternary age.

The primary hydraulic properties controlling groundwater movement include porosity, hydraulic conductivity, and transmissivity. Porosity controls the volume of water that can be stored within a rock or unconsolidated sediment. Primary porosity is a measure of the open pore space within the crystalline structure and (or) sediment grains of a rock, and is directly affected by grain sorting (Lohman and others, 1972, p. 4). With the exception of extrusive igneous or volcanic rocks, the primary porosity of unconsolidated sediments generally is greater than that of consolidated rocks because pore spaces between sediment grains are more numerous and more interconnected than pores in consolidated rock (Maurer and others, 2004, p. 4). Volcanic rocks can develop secondary porosity from open spaces along fractures and joints, or from dissolution of rocks along fractures and joints (Lohman, 1979, p. 4).

Hydraulic conductivity and transmissivity describe the rate at which water moves through an aquifer (Lohman and others, 1972, p. 4). The hydraulic conductivity of basin-fill deposits generally is related to primary porosity and can vary by six orders of magnitude (Maurer and others, 2004, p. 6). The hydraulic conductivity of consolidated rocks depends mostly on the degree of fracturing (secondary porosity) and

respective fracture permeability, and can vary by 12 orders of magnitude (Maurer and others, 2004, p. 6). Transmissivity describes the rate of groundwater movement through a section of aquifer and is expressed as the product of local hydraulic conductivity and saturated unit aquifer thickness (Lohman, 1979, p. 6). Transmissivity is a large-scale or “bulk” aquifer property, whereas hydraulic conductivity is a property that can vary between and within rock and sediment types. Therefore, it is possible for an aquifer to have a low transmissivity even if it contains fractions of material with high hydraulic conductivity.

Carbonate and Clastic Sedimentary Rocks

Carbonate and clastic sedimentary rocks in the mountains adjacent to Dixie Valley are present in the Stillwater Range and the Clan Alpine and Augusta Mountains, and are exposed mostly from the central to northern parts of the mountain block (fig. 2). Rock types are composed of limestone, dolomite, sandstone, siltstone, conglomerate, and shale ranging in age from Early Triassic to Middle Jurassic (Stewart and Carlson, 1978) (table 2). Many of these rocks have undergone some low-grade metamorphism (Page, 1965).

The hydraulic conductivity of the combined carbonate and clastic sedimentary unit likely is variable owing to the range of lithologies present. Carbonate rocks can have greater hydraulic conductivity due to fracturing and where groundwater dissolution has widened fractures through time (Plume, 1996, p. B11). As a result, hydraulic conductivity can range from 3×10^{-5} to 3,300 ft/d (Maurer and others, 2004, table 2). Hydraulic conductivities of clastic sedimentary rocks typically are low (2×10^{-9} –20 ft/d) owing to either cementation (calcium carbonate or silica) or some degree of metamorphism (Maurer and others, 2004, table 2). When clastic rocks are interbedded or in contact with rock units of greater hydraulic conductivity (such as carbonate rocks or basin-fill deposits), they often are considered barriers to groundwater flow (Davis and DeWiest, 1966, p. 347; D’Agnese and others, 1997, p. 20; Harrill and Prudic, 1998, p. A19).

Crystalline Rocks

Few crystalline rocks are present in Dixie Valley relative to other rock types (fig. 2). Granitic rocks, such as quartz monzonite, of Triassic-to-Tertiary age, and intrusive rhyolites of Tertiary age, are present mostly near the southern part of Dixie Valley, in the Stillwater Range and the Clan Alpine Mountains. Gabbroic complexes of Early and Middle Jurassic age (Humboldt lopolith) are exposed mostly in the Stillwater Range and include gabbro, basalt, and associated quartz sandstone (Stewart, 1980, p. 71).

These crystalline rocks likely extend to great depths and have hydraulic conductivity ranging from 7×10^{-8} to 30 ft/d (Maurer and others, 2004, table 2). The hydraulic conductivity of rhyolite likely is higher than that of other crystalline rocks owing to fracturing.

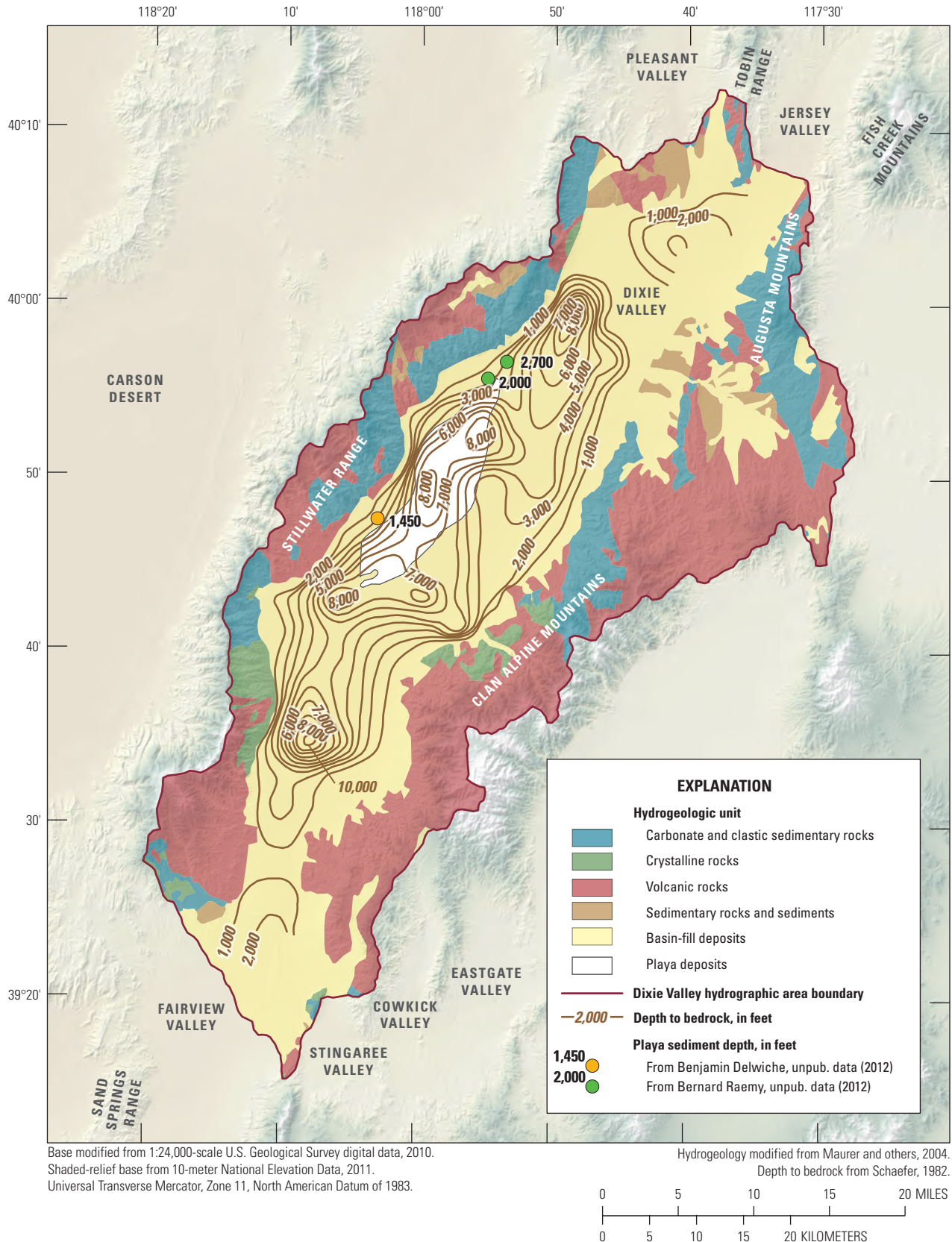


Figure 2. Hydrogeologic units, depth to bedrock, and known playa sediment depths, Dixie Valley, west-central Nevada.

Table 2. Hydrogeologic unit characteristics, Dixie Valley, west-central Nevada.

Geologic age	Rock or stratigraphic unit	Lithology	Water-bearing characteristics
Carbonate and clastic sedimentary rocks			
Triassic	Tobin Formation Dixie Valley Formation Augusta Mountain Formation Cane Springs Formation	Sandstone, siltstone, mudstone, conglomerate, limestone, shale, and dolomite. Each formation overlies the next.	Generally impedes the movement of groundwater. Hydraulic conductivity of carbonate rocks ranges from 3×10^{-5} to 3.3×10^3 feet per day (ft/d) owing to secondary porosity; clastic rocks range from 2×10^{-9} to 20 ft/d
Lower Jurassic to Upper Triassic	Sedimentary rocks	Shale and siltstone	
Crystalline rocks			
Triassic, Jurassic, Cretaceous, and Tertiary	Humboldt lopolith	Gabbroic complexes consisting of gabbro and basalt	Impedes the movement of groundwater. Hydraulic conductivity ranges from 7×10^{-8} to 30 ft/d.
	Granitic rocks Intrusive rocks	Granodiorite, felsite, quartz monzonite, intrusive rhyolite to rhyodacite	
Volcanic rocks			
Tertiary	Volcanic rocks	Lava flows of rhyolitic, basaltic, and andesitic composition. Welded and non-welded ash-flow tuffs of rhyolitic and rhyodacitic composition.	Typically impedes the movement of groundwater. Hydraulic conductivity ranges from 3×10^{-7} to 1.3×10^3 ft/d.
Sedimentary rocks and sediments			
Tertiary	Sedimentary rocks and sediments	Fine-grained semi-consolidated sediments of limestone, diatomite, tuffaceous shale, sandstone, and pebble conglomerate interbedded with basalt and tuff.	Generally impedes the movement of groundwater. Hydraulic conductivity ranges from 2×10^{-4} to 20 ft/d.
Basin-fill deposits			
Quaternary	Alluvial fan and basin lowland deposits with some areas of pediment gravels	Unconsolidated deposits of silt, sand, gravel, cobbles, and boulders in alluvial fans. Interbedded clay, silt, sand, and gravel in basin lowlands.	Comprise shallow water-table aquifers and shallow to deeper confined aquifers. Hydraulic conductivity ranges from 2×10^{-4} to 2.2×10^3 ft/d.
Playa deposits			
Quaternary	Playa deposits, includes Humboldt Salt Marsh	Mostly clay and minor silt in lowest central part of the valley.	Comprise confined aquifers to unknown depths. Acts as barrier to groundwater flow. Hydraulic conductivity about 7×10^{-4} ft/d.

Volcanic Rocks

Volcanic rocks of Tertiary age are prevalent regionally and constitute a large part of exposed consolidated rocks in the mountain blocks surrounding Dixie Valley ([fig. 2](#)). Andesitic and basalt flow outcrops are scattered and laterally discontinuous, whereas those for rhyolitic, ash-flow, and welded tuffs are more extensive. Volcanic rocks, mostly basalt, also underlie basin-fill deposits throughout much of the valley (Wilt and Goldstein, 1983, p. 3–4; Abbott and others, 2001, fig. 10; Mankhemthong and others, 2008; Ben Delwiche, Ormat Technologies Inc., written commun., 2012). Caldera structures are present in the Stillwater Range and the Clan Alpine Mountains. (Ludington and others, 1996). The hydraulic conductivity of volcanic rocks can vary greatly (3×10^{-7} –1,300 ft/d), depending on the degree of jointing as the lava (or ash) flow solidified and was subsequently fractured by tectonic activity (Davis and DeWiest, 1996, p. 337; Maurer and others, 2004, table 2).

Tertiary Sedimentary Rocks and Sediments

Tertiary sedimentary rocks and sediments are composed of semi-consolidated sedimentary rocks of lacustrine or fluvial origin interbedded with volcanic basalt flows and tuffs. Sedimentary deposits are described as limestone, diatomite and opalized diatomite, tuffaceous shale, sandstone, claystone, and pebble conglomerate (Willden and Speed, 1974, p. 28; Johnson, 1977, p. 36; Stewart, 1980, p. 89–93). These sedimentary rocks and sediments are exposed near the northern end of Dixie Valley in the Clan Alpine Mountains and the Stillwater Range, but also underlie the younger basin-fill deposits in some areas of the basin (Wilt and Goldstein, 1983, p. 3–4; Plume, 1996, p. B15; Abbott and others, 2001, fig. 10; Mankhemthong, 2008; Ben Delwiche, Ormat Technologies Inc., written commun., 2012). Tertiary sediments are grouped separately from unconsolidated Quaternary basin fill sediments because typically they are more fine-grained and less permeable (Maurer and others, 2004, p. 7). The hydraulic conductivity of Tertiary sedimentary rocks and sediments can range from 2×10^{-4} to 20 ft/d (Maurer and others, 2004, table 2).

Basin-Fill Deposits

Basin-fill deposits in Dixie Valley are composed of heterogeneous mixtures of gravel, sand, silt, and clay sediments that have been eroded from the surrounding mountain ranges and deposited on the basin floor. Younger and older basin-fill deposits of Quaternary age were mapped previously by Willden and Speed (1974, pl. 1). Because fresh, basin-fill aquifers in Dixie Valley are present in younger and older deposits, no distinction will be made between these two units in this report and both will be referred to as “basin fill.”

Basin-fill deposits cover 48 percent (about 627 mi²) of the basin-floor surface area and extend to varying depths (Maurer and others, 2004). Gravity profiles, which are composed of individual gravity measurements made by a gravimeter, indicate that basin-fill deposits generally are shallow near the mountain fronts, becoming thick near the basin center (Schaefer, 1983; Mankhemthong and others, 2008). Profiles also show several deep sub-basins along the basin center, with maximum depths of as much as 10,000 ft bls ([fig. 2](#); Schaefer, 1983).

Playa Deposits

A series of lakes covered much of northwestern Nevada during the Pleistocene Epoch (Morrison, 1964; Reheis, 1999) and Lake Dixie occupied Dixie Valley. Lake Dixie, has since evaporated, leaving behind thick Quaternary lake-bottom sediments constituting the modern playa ([fig. 2](#)). Playa deposits are composed mostly of clay and silt (Soil Survey Staff, 2012), cover 5 percent (68 mi²) of the basin-floor surface area, and extend to an unknown depth. Shallow well logs from the southern (Nevada Division of Water Resources well log number 108770) and northern (Nevada Division of Water Resources well log number 61383) playa areas noted the persistence of lacustrine clay to depths of 50 and 500 ft bls, respectively, which are synonymous with total drilled depth. Deeper geothermal well logs document that lacustrine clay persists to about 1,450 ft bls near the west-central edge of the playa (Ben Delwiche, Ormat Technologies Inc., written commun., 2012) and to about 2,700 ft bls near the northern edge of the playa (Bernard Raemy, Terra-Gen Power, LLC, written commun., 2012) ([fig. 2](#)).

The playa-basin fill interface likely is composed of interfingering deposits ([fig. 3](#)). Clay and basin-fill interfingering is supported by many flowing artesian wells (each to differing degrees) clustered just south of the playa ([fig. 4](#); [table 3](#), wells 30–44, 46). Drillers’ logs indicate that subsurface sediments adjacent to most of the flowing-artesian wells in this area are composed of alternating gravel and sand layers beneath or between interfingering clay lenses that create semi-confined conditions.

The hydraulic conductivity of shallow playa deposits adjacent to well 27 was estimated at about 6.5×10^{-4} ft/d from a slug test ([fig. 5](#); U.S. Geological Survey, 2010). Well 27 penetrates about 9.6 ft of playa deposits and is screened from about 5.6 to 9.6 ft bls ([table 4](#)). Because the well was pumped dry, only water-level recovery data were used for the slug-test analysis. Hydraulic conductivity was estimated using an Excel® spreadsheet program (Halford and Kuniansky, 2002) that uses the Bouwer and Rice (1976) method. A transmissivity estimate of 0.10 ft²/d was calculated by multiplying the hydraulic conductivity by an assumed aquifer thickness of 200 ft ([table 4](#)). Although the playa clay is much thicker, the upper 200 ft was assumed to be the active part of the playa aquifer where groundwater flow is likely to occur.

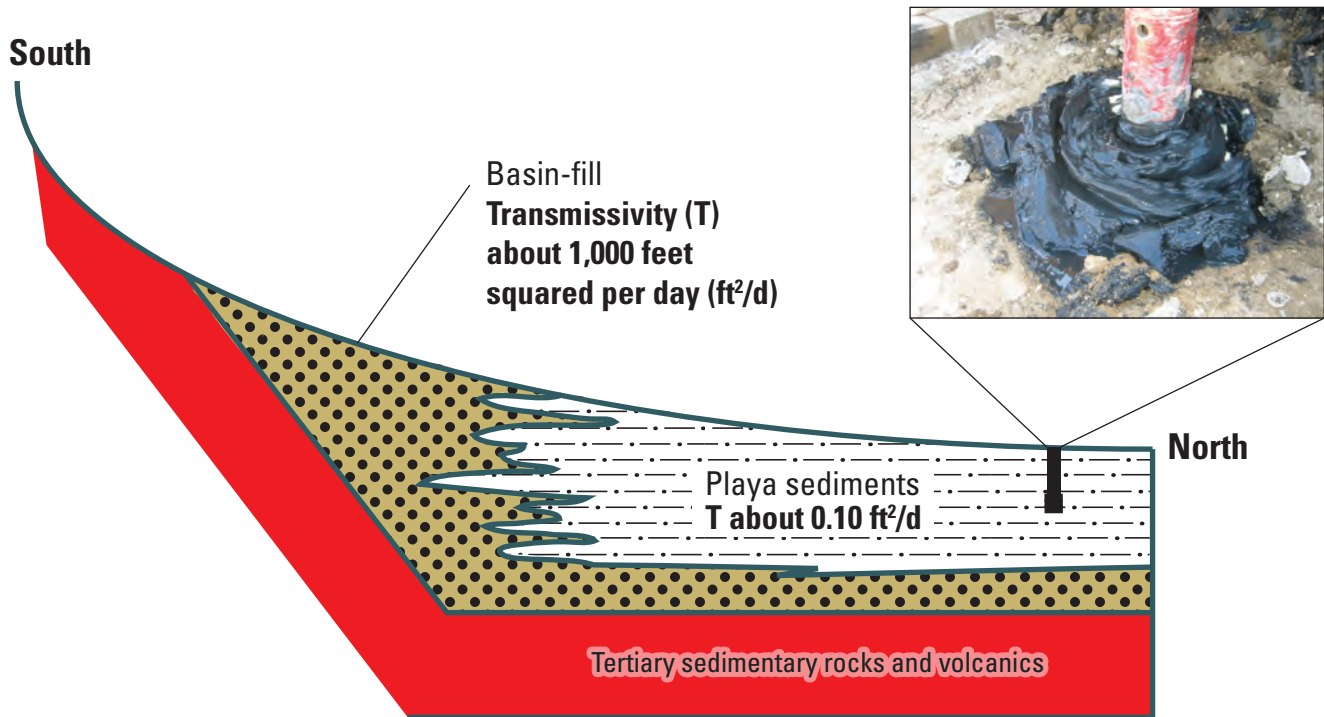


Figure 3. Conceptual cross-sectional diagram showing inter-fingering of clay-dominated playa and basin-fill sediments, and photograph showing typical playa-deposits during well drilling in Dixie Valley, west-central Nevada. Photograph associated with Nevada Division of Water Resources well log number 108770.

Structural Features

Faults can be important features in controlling groundwater flow. A fault can facilitate groundwater flow by acting as a conduit or preferential flow path, or it can impede groundwater flow where hydrogeologic units of differing hydraulic conductivity become juxtaposed (Plume, 2009, p. 12). Quaternary fault locations in Dixie Valley (Great Basin Center for Geothermal Energy, 2004) are shown in [figure 5](#).

Dixie Valley is bounded on the east and west sides by normal faults (Great Basin Center for Geothermal Energy, 2004). The main range-bounding fault is the Stillwater Fault ([fig. 5](#)), which is still active and is at the base of the Stillwater Range along with a series of smaller subsidiary faults not visible at the land surface (Schaefer, 1983, p. 11; Smith and others, 2001). A severe earthquake occurred along the Stillwater Range in 1954, creating scarps of 10 ft on average on the west side of Dixie Valley (Romney, 1957). The valley floor became down-tilted to the west and sediments were thought to compact as a result of the earthquake ground movement (Zones, 1957, p. 395). This sediment compaction caused water levels to rise temporarily (by less than 1 to as much as 4.6 ft) in wells south of the playa and caused spring

discharge to increase south of the playa and along the western basin margin (Zones, 1957, p. 395–396).

Several geothermal studies have mapped and characterized faults in Dixie Valley smaller in scale than the Stillwater Fault; however, the extent to which basin-fill groundwater flow is affected locally by these faults is mostly unknown. Linear spring alignment, as well as scarps, fissures, and distinctive vegetation patterns, have been used to identify fault locations in the valley (Smith and others, 2001). There are 46 documented springs in Dixie Valley ([fig. 5](#); [table 5](#)), 17 of which are within about 0.2 mi of a Quaternary fault. Of these 17 springs, 10 have been measured for water temperature. Cold water temperatures (less than 20 °C) were measured in four of the fault-adjacent springs, and warm temperatures (between 20 and 50 °C) were measured in five of these springs. Spring water temperatures greater than 20 °C likely suggest geothermal mixing (see section, “[Connection Between Basin-Fill and Geothermal Aquifers](#)”). The remaining spring, Dixie Valley Hot Springs (S28, [table 5](#); [fig. 5](#)), was hot (greater than 50 °C). Although many fault-adjacent springs were warm-to-hot, faults generally do not act as significant conduits for transporting geothermal water to land surface.

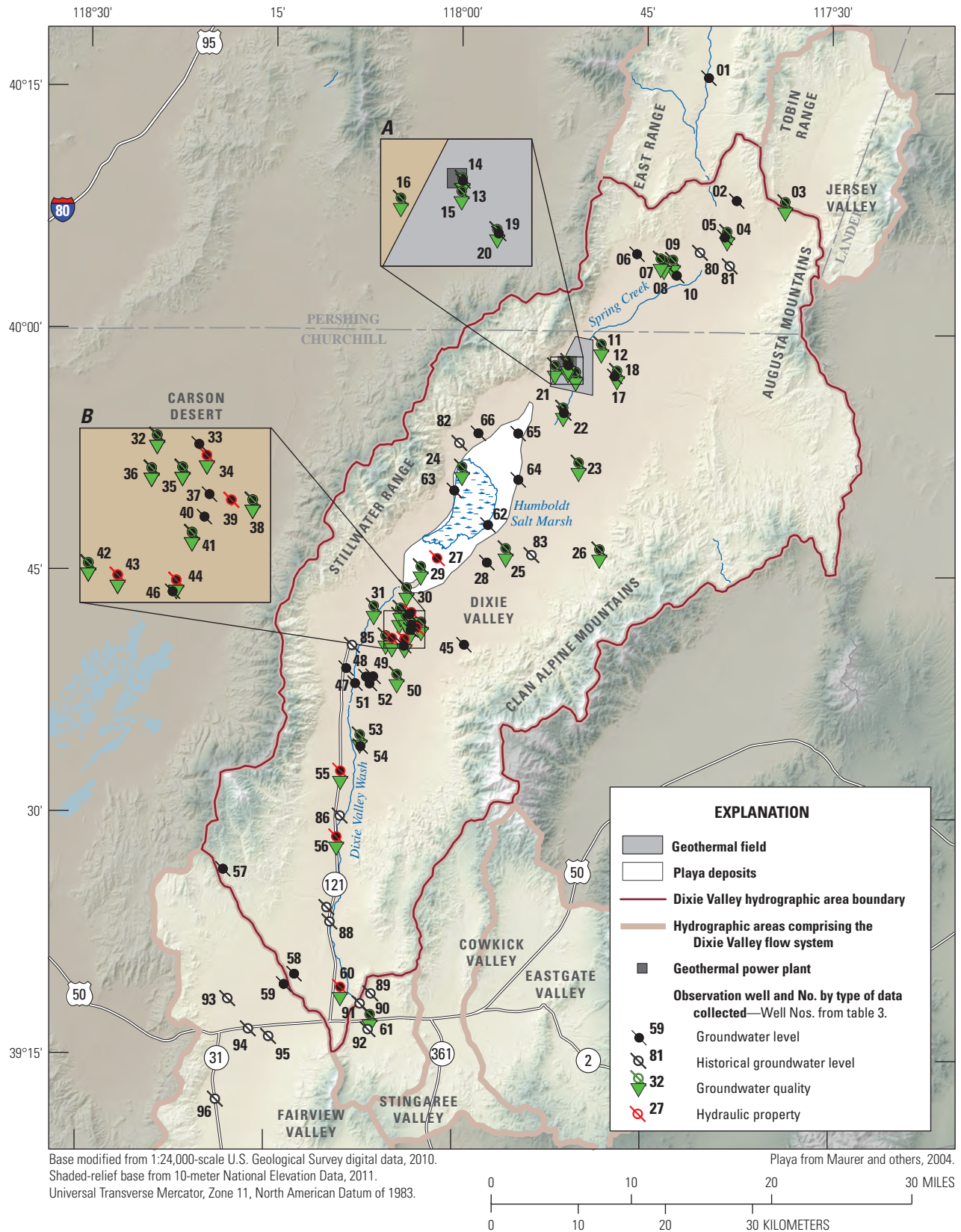


Figure 4. Data collection sites in Dixie Valley, west-central Nevada, 2009–11. Types of data collected include groundwater level (WL), historical groundwater level (HWL), groundwater quality (QW), and hydraulic properties (HP).

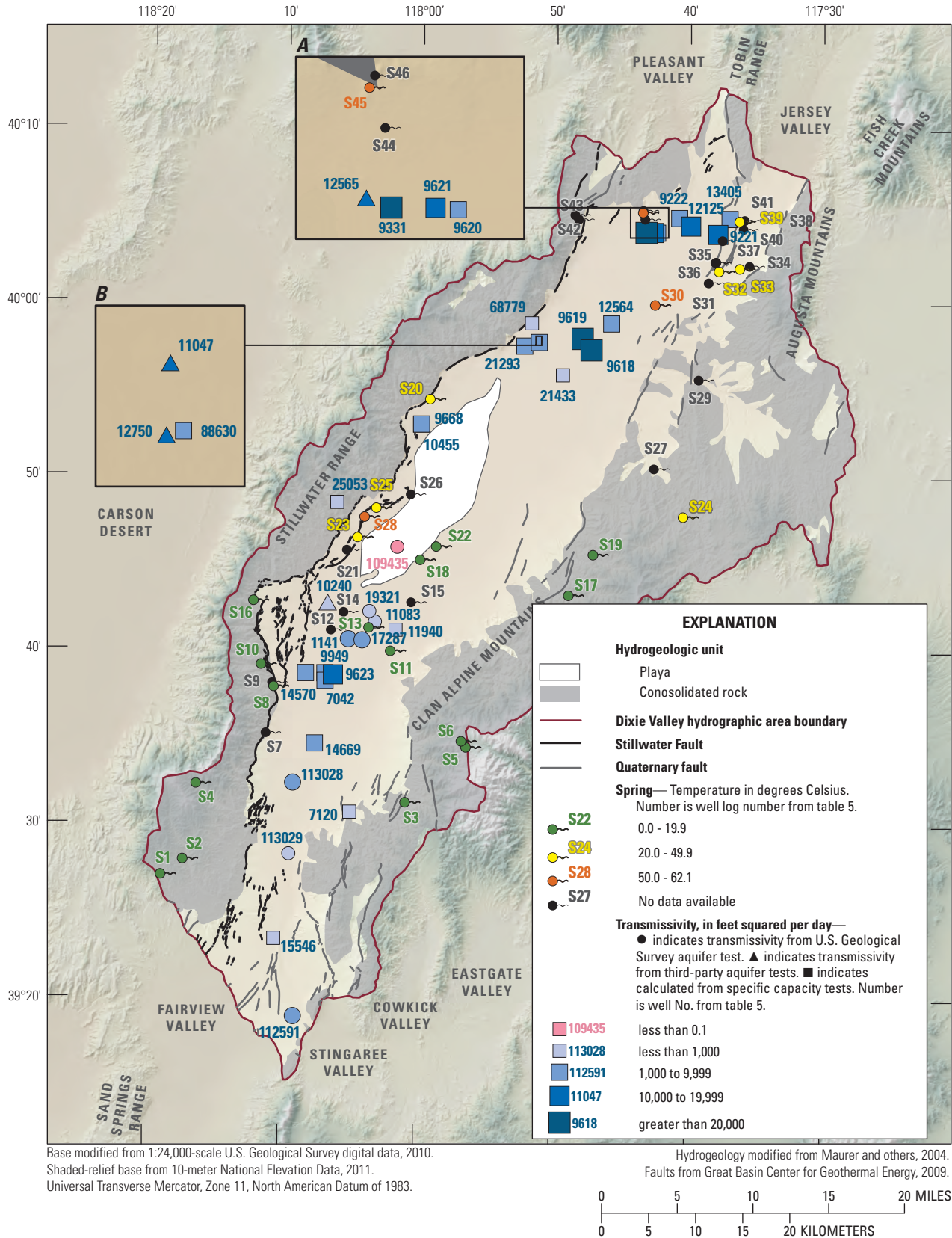


Figure 5. Estimates of basin-fill transmissivity from single-well aquifer tests or specific-capacity calculations and water-temperature distributions in springs, Dixie Valley, west-central Nevada.

Table 3. Well information, static water levels, and types of data collected in the Dixie Valley flow system, west-central Nevada.

[Well No.: Locations of wells are shown in figure 4. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Available data types:** WL, water levels; HWL, historical water level, measured prior to this study; QW, water quality; HP, hydraulic properties. **Land surface and water-level elevation:** Datum is North American Vertical Datum of 1988. **Abbreviations:** ft, foot; in, inch; bls, below land surface; BLM, Bureau of Land Management; USGS, U.S. Geological Survey; n/a, not applicable; <, less than; >, greater than; -, no data]

Well No.	USGS standard identification No.	Project site name	Latitude	Longitude	Land surface elevation (ft)	Total well depth (ft bls)	Casing		Available data types	Water-level	
							Diameter (in.)	Perforation (ft bls)		Measurement date	Elevation (ft)
01	400133117395702	Pleasant Valley Windmill Well	40 15 53.99	117 40 04.13	4,289	20	-	-	WL	06-28-11	4,273
02	400817117374301	Northwest Dixie Well	40 08 18.11	117 37 43.68	3,876	-	-	-	WL	06-28-11	3,617
03	400753117334401	USBLM McCoy Well	40 07 54.15	117 33 47.75	3,883	200	6	154-200	WL, QW	10-29-09	3,752
04	400600117380001	Little McCoy Ranch	40 06 01.92	117 38 28.87	3,702	114	6	81-114	WL, QW	10-29-09	3,612
05	400603117383601	Little Mc Ag Well	40 06 03	117 38 36	3,708	400	16	200-400	WL	10-29-09	3,612
06	395004117455601	Seven Devil Stock Well	40 04 57.17	117 45 44.43	3,712	-	-	-	WL	06-28-11	3,469
07	400421117434701	Seven Devil	40 04 20.67	117 43 49.65	3,576	134	8	104-134	WL, QW	10-29-09	3,483
08	400419117433401	Seven Devils Ag Well #3	40 04 19.3	117 43 33.83	3,580	270	16	90-270	WL, QW	10-29-09	3,489
09	400416117424401	Ag Well #1	40 04 16.55	117 42 50.50	3,592	350	16	100-235	WL, QW	10-29-09	3,490
10	400338117422801	Ag Well #4	40 03 38.01	117 42 31.23	3,592	354	12	114-354	WL	10-29-09	3,497
11	395901117483402	Terra Gen I-A	39 59 01.00	117 48 33.92	3,454	50	4	20-30	WL	10-29-09	3,436
12	395901117483401	Terra Gen I-B	39 59 01.00	117 48 33.92	3,454	70	4	50-60	WL, QW	10-29-09	3,436
13	395759117511301	Goeringer Well	39 57 58.65	117 51 13.24	3,471	430	16	50-430	WL	10-21-09	3,409
14	395800117510901	73W-7	39 57 57.07	117 51 13.10	3,470	300	17	80-300	WL, QW	10-22-09	3,430
15	395747117510901	Blackwater	39 57 46.26	117 51 13.93	3,468	160	6	140-160	WL, QW	10-22-09	3,419
16	395740117521401	Irrigation	39 57 40.50	117 52 17.21	3,556	300	16	0-165	WL, QW	10-23-09	3,422
17	395722117471601	West pair Ag well	39 57 22.12	117 47 16.83	3,504	330	16	100-320	WL, QW	10-30-09	3,443
18	395723117472601	East pair Ag well	39 57 21.89	117 47 27.47	3,499	400	16	50-400	WL	10-30-09	3,438
19	395716117503602	Terra Gen II-A	39 57 16.14	117 50 36.38	3,435	50	4	20-30	WL, QW	10-23-09	3,420
20	395716117503601	Terra Gen II-B	39 57 16.14	117 50 36.38	3,435	70	4	50-60	WL, QW	10-23-09	3,421
21	395503117513602	Terra Gen III-A	39 55 03.21	117 51 35.66	3,406	50	4	20-30	WL	10-23-09	1,3,408
22	395503117513601	Terra Gen III-B	39 55 03.21	117 51 35.66	3,406	70	4	50-60	WL, QW	10-23-09	1,3,411
23	395139117501801	DVMV	39 51 39.26	117 50 18.31	3,444	41	2	31-41	WL, QW	11-19-09	3,431
24	395123117594802	DVPLW shallow	39 51 19.75	117 59 40.21	3,377	2.5	2.5	1.5-2.5	QW	11-04-09	1,3,376
25	394618117560601	East central well	39 46 18.03	117 56 06.34	3,428	29.5	2	19.5-29.5	WL, QW	11-04-09	3,423
26	394617117483401	BLM Bernice Cr at mtn fr	39 46 17.15	117 48 34.29	4,383	139	8	-	WL, QW	10-30-09	4,276
27	394559118013701	DVPL2	39 45 59.21	118 01 36.60	3,382	9.6	1	5.6-9.6	HP	11-03-09	3,381
28	394545117573602	DVDV-2	39 45 45.05	117 57 36.37	3,430	19	1	17-19	WL	11-05-09	3,414
29	394508118025503	DVPL1 composite	39 45 07.79	118 02 54.63	3,385	10	1	4-10	QW	11-03-09	3,382
30	394348118040201	DVSU	39 43 48.38	118 04 02.15	3,386	24.5	2	19.5-24.5	WL, QW	11-02-09	1,3,384
31	394239118063901	Knitille Well	39 42 38.86	118 06 39.20	3,413	213	16	183-213	WL, QW	10-22-09	1,3,433
32	392748118042501	North Settle	39 42 32.22	118 04 30.12	3,406	72	-	-	QW	n/a	1,3,406
33	394229118034701	Navy 16C	39 42 29.48	118 03 46.80	3,404	210	16.5	105-155; 200-210	WL	10-28-09	1,3,417
34	394217118033801	Navy 31B	39 42 16.75	118 03 37.90	3,410	194	10.75	134-154; 174-194	WL, QW, HP	10-22-09	1,3,420

Table 3. Well information, static water levels, and types of data collected in the Dixie Valley flow system, west-central Nevada.—Continued

[Well No.: Locations of wells are shown in figure 4. Latitude and longitude: Shown in degrees, minutes, and seconds. Available data types: WL, water levels; HWL, historical water level, measured prior to this study; QW, water quality; HP, hydraulic properties. Land surface and water-level elevation: Datum is North American Vertical Datum of 1988. Abbreviations: ft, foot; in, inch; bls, below land surface; BLM, Bureau of Land Management; USGS, U.S. Geological Survey; n/a, not applicable; <, less than; >, greater than; -, no data]

Well No.	USGS standard identification No.	Project site name	Latitude	Longitude	Land surface elevation (ft)	Total well depth (ft bls)	Casing		Available data types	Water-level	
							Diameter (in.)	Perforation (ft bls)		Measurement date	Elevation (ft)
35	394205118035001 16E		39 42 06.97	118 04 03.05	3,411	158	3	None	WL, QW	10-22-09	13,440
36	392708118042301 17L		39 42 06.18	118 04 35.65	3,413	-	-	-	WL, QW	10-28-09	13,427
37	394150118033501 Navy 35A		39 41 49.61	118 03 34.78	3,416	-	-	-	WL	10-28-09	13,439
38	394142118025101 Navy 33A		39 41 41.52	118 02 50.76	3,420	260	4	200–260	WL, QW	10-22-09	13,444
39	394141118030901 34D		39 41 40.98	118 03 08.88	3,419	160	-	-	HP	n/a	>13,419
40	394131118034001 Navy 47B		39 41 31.48	118 03 39.81	3,424	292	-	-	WL	10-28-09	13,449
41	394114118034001 56A		39 41 14.77	118 03 52.77	3,430	200	6.6	175–195	WL, QW	10-22-09	13,437
42	394048118053001 81H		39 40 49.11	118 05 40.14	3,444	171	6	161–171	WL, QW	10-21-09	13,454
43	392602118050101 84B		39 40 39.82	118 05 10.10	3,444	150	14.3	25–145	WL, QW, HP	10-21-09	13,453
44	394037118040901 Navy 86A		39 40 36.52	118 04 08.97	3,445	184	16	None	WL, QW, HP	10-22-09	13,456
45	392646117574001 Grover Point Well		39 40 38.55	117 59 22.39	3,531	118	6	-	WL	10-30-09	3,459
46	394036118035001 Navy 89A		39 40 32.15	118 04 12.39	3,444	255	16	45–145	WL	10-28-09	13,460
47	393920118084001 Dixie Valley Cemetery Well		39 39 06.51	118 08 48.39	3,510	112	11	-	WL	04-14-10	3,480
48	393837118071301 Navy 101C		39 38 36.46	118 07 12.77	3,507	155	16	70–145	WL	10-20-09	3,476
49	393838118063601 102A		39 38 36.71	118 06 39.48	3,503	162	16	84–154	WL	10-28-09	3,477
50	393840118050001 USBLM Corral		39 38 26.00	118 04 44.52	3,494	45	-	-	WL, QW	11-06-09	3,460
51	393810118080001 100A		39 38 10.93	118 08 02.88	3,520	-	16	-	WL	10-20-09	3,486
52	393810118065201 106A		39 38 10.67	118 06 55.23	3,514	200	12	35–172	WL	10-28-09	3,479
53	393420118074001 108B		39 34 39.56	118 07 39.02	3,632	390	16	155–385	WL, QW	10-20-09	3,503
54	393414118072901 Pump Jack Well		39 34 14.1	118 07 33.74	3,648	6	6	-	WL	05-25-10	3,503
55	393224118091001 Nr Horse Creek Rd		39 32 24.04	118 09 10.39	3,814	500	8	400–500	WL, QW, HP	01-23-11	3,519
56	392819118092501 Dixie HWY S Nr Marker 13		39 28 19.16	118 09 25.46	3,892	519	8	399–519	WL, QW, HP	12-17-10	3,544
57	392635118182501 BLM La Plata Windmill		39 26 34.63	118 18 25.32	5,727	70	6	50–70	WL	05-25-10	5,708
58	390520118125801 USBLM North Well		39 20 08.23	118 12 40.42	4,223	364	6.6	303–363.5	WL	05-25-10	3,922
59	391927118132301 USN Centroid		39 19 26.90	118 13 26.71	4,243	400	10.8	360–420; 460–530	WL	10-21-09	3,927
60	391900118085801 Wonder Well		39 18 59.96	118 08 58.35	4,349	500	8	400–500	WL, QW, HP	12-06-10	3,925
61	390234118070701 State of Nevada		39 17 19.81	118 06 34.67	4,391	288	8	-	WL, QW	10-20-09	4,138
62	394805117573501 PLGW4 deep ²		39 48 05.15	117 57 35.08	3,381	9	2	4–9	WL	10-30-09	3,380
63	395012118001701 PLGW1 deep ²		39 50 12.06	118 00 17.23	3,381	9	2	7–9	WL	10-29-09	3,380
64	395054117551001 PLGW3 deep ²		39 50 54.36	117 55 10.08	3,387	9	2	6–9	WL	10-30-09	3,385

Table 3. Well information, static water levels, and types of data collected in the Dixie Valley flow system, west-central Nevada.—Continued

[Well No.: Locations of wells are shown in figure 4. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Available data types:** WL, water levels; HWL, historical water level, measured prior to this study; QW, water quality; HP, hydraulic properties. **Land surface and water-level elevation:** Datum is North American Vertical Datum of 1988. **Abbreviations:** ft, foot; in, inch; bls, below land surface; BLM, Bureau of Land Management; USGS, U.S. Geological Survey; n/a, not applicable; <, less than; >, greater than; -, no data]

Well No.	USGS standard identification No.	Project site name	Latitude	Longitude	Land surface elevation (ft)	Total well depth (ft bls)	Casing		Available data types	Water-level	
							Diameter (in.)	Perforation (ft bls)		Measurement date	Elevation (ft)
65	395345117551301	PLGW5 deep ²	39 53 45.46	117 55 12.76	3,389	9	2	7-9	WL	05-12-10	3,387
66	395348117582301	PLGW2 deep ²	39 53 47.72	117 58 23.40	3,383	9	2	4-8	WL	10-29-09	3,382
80	400508117403401	North 3500 Well	40 05 08	117 40 34	3,660	257	n/a	n/a	HWL	11-17-82	3,566
81	400416117381101	North 3600 Well	40 04 16	117 38 11	3,663	256	n/a	n/a	HWL	04-12-83	3,608
82	395310117595001	West valley well	39 53 10	117 59 50	3,480	-	n/a	n/a	HWL	03-31-82	3,446
83	393143117541401	East upland well	39 46 15	117 54 00	3,562	181	n/a	n/a	HWL	08-23-51	3,437
85	392551118083801	USN	39 40 33	118 08 19	3,500	32	n/a	n/a	HWL	04-17-63	3,475
86	391511118092401	Unnamed	39 29 58	118 09 10	3,817	329	n/a	n/a	HWL	04-01-82	3,530
87	392418118100501	Eleven Mile Well	39 24 18	118 10 05	4,118	-	n/a	n/a	HWL	03-30-82	3,752
88	392323118095001	South Valley Well	39 23 23	118 09 50	4,100	475	n/a	n/a	HWL	11-21-08	3,888
89	391859118063101	Chalk Mtn Well	39 18 59	118 06 31	4,384	-	n/a	n/a	HWL	07-19-12	4,137
90	n/a	Dixie Wash Abandoned Mine	39 18 21.65	118 07 20.94	4,327	136	n/a	n/a	HWL	07-19-12	³ <4,191
92	391645118064001	Stingaree valley south	39 16 45	118 06 40	4,500	650	n/a	n/a	HWL	03-30-82	4,136
93	391833118175201	Fairview west	39 18 33	118 17 52	4,242	390	n/a	n/a	HWL	03-30-82	3,936
94	390152118163201	Fairview vly central	39 16 44	118 16 15	4,154	280	n/a	n/a	HWL	03-30-82	3,927
95	391620118143001	Fairview valley east-central	39 16 12.8	118 14 38.2	4,154	435	n/a	n/a	HWL	03-27-08	3,935
96	382901118190201	Fairview south upland	39 12 29	118 18 34	4,300	699	n/a	n/a	HWL	10-21-09	4,023

¹Artesian well.

² Sites are completed in playa sediments even though they may appear outside the playa boundary from Maurer and others, 2004.

³Abandoned vertical mine shaft, dry bedrock present at about ~136 feet below land surface.

Table 4. Basin-fill transmissivity estimated from single-well and multiple-well aquifer tests, specific-capacity estimates, slug tests, and well characteristics, Dixie Valley, west-central Nevada.

[Well No.: Locations of wells are shown in [figure 5](#). NDWR well log: Can be viewed by searching for log number at: <http://water.nv.gov/data/welllog/>. Colors correspond to transmissivity estimates from [figure 5](#). Land surface, total well depth, screened interval, and water levels are referenced to North American Vertical Datum of 1988 (NAVD 88); data from National Elevation Datasets (<http://ned.usgs.gov/>). Flow rate: Values shown for multiple-well aquifer tests and single-well aquifer tests in flowing wells. Abbreviations: NDWR, Nevada Division of Water Resources; ft bls, foot below land surface; gal/min, gallon per minute; ft²/d, foot squared per day; >, greater than; n/a, not applicable; –, no data]

Well No.	NDWR well log No.	Land surface elevation (ft)	Total well depth elevation (ft)	Screened interval elevation (ft)	Static water level elevation (ft)	Date of test	Flow rate (gal/min)	Transmissivity (ft ² /d)
Transmissivity from multiple-well aquifer testing								
08	12565	3,580	3,310	3,490–3,310	3,485	4-24 to 5-21-2012	n/a	¹ 10,400
13	11047	3,471	3,041	3,421–3,041	3,411	2-28 to 2-29-2012	n/a	¹ 14,800
15	12750	3,468	3,308	3,328–3,308	3,416	2-28 to 2-29-2012	n/a	¹ 16,900
Transmissivity from single-well aquifer testing								
56	113029	3,892	3,373	3,493–3,373	3,546	5-2 to 5-27-2011	n/a	700
60	112591	4,349	3,849	3,949–3,849	3,917	5-16 to 5-19-2011	n/a	2,500
55	113028	3,814	3,314	3,414–3,314	3,021	7-5 to 7-7-2012	n/a	6,000
Transmissivity from single-well aquifer testing in flowing wells								
34	19321	3,410	3,216	3,276–3,216	>3,410	7-29-2010	7	400
39	11083	3,419	3,256	3,339–3,289	>3,419	7-29-2010	25	900
44	17287	3,445	3,261	–	>3,445	7-29-2010	27	1,000
43	1141	3,444	3,289	3,419–3,299	>3,444	7-29-2010	27	1,400
31	10240	3,413	3,200	3,230–3,200	>3,413	5-17 to 5-18-2012	35	¹ 800
Transmissivity from single-well slug testing								
27	109435	3,382	3,372	3,376–3,372	3,381	7-13 to 7-15-2011	n/a	² 0.10
Transmissivity from drillers' log specific capacity								
n/a	25053	5,283	4,965	5,009–4,969	5,272	08-19-1983	0.2	30
n/a	15546	4,168	3,686	3,903–3,763	3,956	04-18-1976	0.23	30
n/a	21433	3,343	3,078	3,228–3,078	3,323	07-05-1980	0.63	100
n/a	11940	3,433	3,270	3,390–3,270	–	02-04-1971	0.71	110
n/a	68779	3,611	3,111	3,271–3,111	3,461	08-06-1997	1.33	240
n/a	7120	4,286	4,086	4,136–4,086	–	03-15-1963	2.78	560
n/a	21293	3,536	3,371	3,535–3,371	3,411	06-01-1974	5.45	1,200
n/a	13405	3,704	3,304	3,604–3,304	3,664	06-29-1973	11.56	2,900
n/a	14669	3,630	3,240	3,475–3,245	3,504	04-14-1975	14.29	2,700
n/a	14570	3,506	3,005	3,437–3,232	–	11-16-1974	15.25	4,000
n/a	12564	3,503	3,283	3,433–3,283	3,463	08-20-1972	15.56	4,000
n/a	9620	3,597	3,312	3,497–3,322	3,505	06-17-1966	17	4,500
n/a	9949	3,501	3,346	3,431–3,356	3,471	02-29-1968	22.73	6,400
n/a	9222	3,657	3,418	3,574–3,424	3,593	01-15-1965	23.68	6,700
n/a	7042	3,510	3,255	3,450–3,255	3,475	02-17-1963	24.5	7,000
n/a	9668	3,447	3,312	3,402–3,312	3,412	07-22-1967	25	7,100
n/a	88630	3,462	3,162	3,382–3,162	3,403	11-20-2002	28.33	8,200
n/a	10455	3,447	3,199	3,410–3,208	3,410	01-17-1969	30.3	8,900
n/a	12125	3,654	3,304	3,594–3,304	3,589	02-19-1972	47.78	15,000
n/a	9623	3,499	3,224	3,399–3,234	3,481	06-29-1967	51.47	16,400
n/a	9221	3,671	3,433	3,556–3,437	3,621	04-19-1965	54.46	17,600
n/a	9621	3,592	3,347	3,492–3,357	3,500	07-29-1966	58.33	19,000
n/a	9619	3,472	3,142	3,372–3,172	3,427	05-17-1967	100	35,500
n/a	9618	3,487	3,157	3,387–3,167	3,442	06-11-1967	121.0	44,300
n/a	9331	3,580	3,335	3,480–3,340	3,488	07-20-1966	123.81	45,500
Average (excluding playa)								8,700

¹Transmissivity value based on aquifer test by Interflow Hydrology, Inc. and Mahannah and Associates, LLC.

²Sites are completed in playa sediments even though they may appear outside the playa sediment boundary by Maurer and others (2004) depicted in [figure 4](#).

Table 5. Locations, elevations, and temperatures in springs and Quaternary faults in the Dixie Valley flow system, west-central Nevada.

[Site No.: Locations of sites are shown in [figure 5](#). **Latitude and longitude:** Shown in degrees, minutes, seconds. **Spring elevation:** In feet above North American Vertical Datum of 1988 (NAVD 88). **Abbreviations:** °C, degrees Celsius; ft, foot; USGS, U.S. Geological Survey; –, no data]

Site No.	USGS standard identification No.	Project site name	Latitude	Longitude	Elevation (ft)	Temperature (°C)
S1	392704118185201	La Plata Spring	39 27 03.93	118 18 52.46	5,940	7.8
S2	392758118171601	Burnt Cabin Spring	39 27 57.78	118 17 15.61	5,920	8
S3	393119118005001	Horse Ck Ranch Spg	39 31 19.47	118 00 49.59	5,194	13.8
S4	393219118162001	Spring	39 32 18.87	118 16 20.09	6,630	9.5
S5	393431117562101	Spring	39 34 30.85	117 56 21.23	7,600	7.1
S6	393452117564201	Spring	39 34 51.68	117 56 41.82	7,640	8.5
S7	393515118111301	Spring	39 35 15.04	118 11 12.62	4,190	–
S8	393755118103901	Willow Spring	39 37 54.95	118 10 39.38	3,963	15.3
S9	393808118104601	Mud Spring	39 38 08.28	118 10 46.32	4,006	–
S10	393911118113601	Spring	39 39 11.27	118 11 36.45	4,320	7.7
S11	394000118020101	Eastern Settlement spring	39 40 00.43	118 02 01.33	3,471	6.9
S12	394111118062801	Spring	39 41 11.14	118 06 27.7	3,438	–
S13	394120118033901	Settlement spring	39 41 20.23	118 03 38.9	3,422	4.3
S14	394213118053201	Spring	39 42 13.09	118 05 31.69	3,412	–
S15	394248118003101	Spring	39 42 48.41	118 00 31.09	3,422	–
S16	394251118121401	Alameda Canyon Spg	39 42 50.57	118 12 14.17	4,570	8.7
S17	394317117485001	Spring	39 43 16.51	117 48 49.53	5,560	11.6
S18	394515117595201	Southeast playa spring	39 45 14.89	117 59 52.12	3,386	10.3
S19	394537117470101	Spring	39 45 36.75	117 47 01.34	4,910	12.3
S20	395429117591501	PW spring	39 54 28.54	117 59 15.35	3,399	22.4
S21	394547118052001	Spring	39 45 46.93	118 05 19.72	3,416	–
S22	394601117584101	Buckbrush Spring	39 46 01.26	117 58 41.32	3,389	18.6
S23	394631118043301	Cold Spring	39 46 31.31	118 04 32.83	3,425	29
S24	394749117402101	Spring	39 47 48.53	117 40 20.89	5,120	28.8
S25	394813118031101	Northern meadows spring	39 48 13.09	118 03 10.94	3,425	26
S26	394900118003801	Seep	39 49 00	118 00 37.89	3,386	–
S27	395034117423201	Spring	39 50 33.93	117 42 32.19	4,370	–
S28	395322118040101	Dixie Valley Hot Springs	39 53 22	118 04 01	3,418	58.5
S29	395541117391501	Hole in the Wall Spg	39 55 40.9	117 39 14.97	3,908	–
S30	395959117423101	Hyder Hot Springs	39 59 59	117 42 31	3,575	60.9
S31	400137117382001	Spring	40 01 16	117 38 30	4,100	–
S32	400155117374801	North-east spring	40 01 55.39	117 37 47.77	3,740	21.8
S33	400205117361101	Hot Sp, Lower Ranch	40 02 05	117 36 11	3,970	30.2
S34	400211117350801	Spring	40 02 14	117 35 27	3,650	–
S35	400226117380401	Spring	40 02 25.95	117 38 03.64	3,665	–
S36	400227117376001	Spring	40 02 27.18	117 37 59.57	3,685	–
S37	400341117373201	Spring	40 03 41.33	117 37 32.02	3,668	–
S38	400447117361301	S-92 Springs	40 04 47	117 36 13	3,728	–
S39	400448117361301	McCoy Spring	40 04 48	117 36 13	3,725	44.8
S40	400452117355001	Spring	40 04 21	117 35 55	3,760	–
S41	400452117355002	Spring	40 04 52	117 35 50	3,760	–
S42	400452118481701	Spring	40 04 52.27	117 48 16.89	4,669	–
S43	400507117483301	Spring	40 05 07.19	117 48 33.4	4,823	–
S44	400517117430801	Sow Spring	40 04 54	117 43 16	3,780	–
S45	400517117432801	Seven Devil Hot spring	40 05 17	117 43 28	3,700	62.1
S46	400524117432401	Spring	40 05 24	117 43 24	3,752	–

Hydrogeology

The Dixie Valley flow system is hydrologically closed. In Dixie Valley, the terminus of the flow system, most groundwater, surface water, and precipitation that enter the basin remain there until naturally discharged by springs or evapotranspiration. Limited groundwater pumping for agriculture, stock watering, and geothermal supplementation also occurs in the valley. Groundwater in basin-fill deposits of Dixie Valley is present mostly under unconfined conditions. Discontinuous confined groundwater conditions are present along the perimeter of the playa where lacustrine clay and basin-fill deposits are inter-fingered (fig. 3), most notably just south of the playa, but also north of the playa (table 3, wells 21 and 22).

Aquifer Properties

Hydraulic conductivity of basin-fill deposits generally grades from higher values along the mountain front to lower values near the basin center. As sediments are eroded from mountain ranges and deposited in the valley, coarse-grained sediments generally filter out and deposit along alluvial fans and finer-grained material is deposited in the center of the basin (Plume, 1996, p. B15–B17). Hydraulic conductivity of basin-fill deposits can vary by many orders of magnitude (2×10^{-4} to 2,200 ft/d), depending on the location and composition of the deposits in the basin (Maurer and others, 2004).

Hydraulic properties of the basin fill were characterized by transmissivity estimated from single- and multiple-well aquifer tests and specific-capacity data (fig. 5; table 4). Reliable estimates of transmissivity can be made using single-well aquifer tests, even when the aquifer thickness is not well known (Halford and others, 2006, p. 469). Single-well aquifer tests were completed in three non-flowing wells (55, 56, and 60) in southern Dixie Valley and in four flowing wells (34, 39, 43, and 44) in south-central Dixie Valley south of the playa (table 4; fig. 5). Drawdown data from each test were analyzed using an Excel® spreadsheet program (Halford and Kuniansky, 2002) using the Cooper and Jacob (1946) method for non-flowing wells and the Jacob and Lohman (1952) method for flowing wells.

Transmissivity estimates from the non-flowing wells in southern Dixie Valley ranged from 700 to 6,000 ft²/d with an average of about 3,100 ft²/d (table 4; U.S. Geological Survey, 2010). These wells have a depth of about 500 ft and are screened across at least 100 ft of unconfined basin-fill aquifer. The low transmissivity of 700 ft²/d estimated at well 56 represents a screened interval composed predominantly of gravel, with 5–10 percent clay. The high transmissivity value of 6,000 ft²/d estimated at well 55 represents a screened interval of clean gravels and sand. Transmissivity estimates from five flowing wells ranged from 400 to 1,400 ft²/d (table 4; U.S. Geological Survey, 2010; Interflow

Hydrology, Inc., and Mahannah and Associates, LLC, 2012b). These wells penetrate similar depths (less than 200 ft) and are generally open to more than 60 ft of basin-fill material consisting of alternating mixtures of sand, gravel, and clay. Additional multiple well tests were completed in northern Dixie Valley (table 4, wells 08, 13, and 15) (Interflow Hydrology, Inc., and Mahannah and Associates, LLC, 2012b). The test results indicate transmissivity ranges from 10,400 to 16,900 ft²/d in northern Dixie Valley (wells 08, 13, and 15; table 4).

The specific capacity of a well, which commonly is included on the well drillers' log, is often used to estimate transmissivity when aquifer test data are sparse (Lohman and others, 1972, p. 52). Where specific-capacity data were available, an empirical log-based regression equation that relates specific capacity to transmissivity was used to estimate transmissivity (Maurer, 2011, p. 25). Transmissivities ranging from 30 to 45,500 ft²/d were estimated for 25 wells using specific-capacity data (fig. 5; table 4). Transmissivity estimates derived from aquifer tests generally are more accurate and reliable than those derived from specific-capacity data. Because aquifer test data were not available to validate specific-capacity-derived transmissivity estimates for the same well, these estimates should be considered only as an approximation of transmissivity. However, transmissivity estimated from well driller's log 11940 south of the playa (110 ft²/d) was within the same order of magnitude of transmissivity estimated from a flowing well test from well 39 (driller's log 11083; 900 ft²/d), 1.4 mi to the west of 11940 (table 4; fig. 5). In some areas of Dixie Valley, specific-capacity-derived transmissivity estimates varied by one or more orders of magnitude among wells with similar construction and near each other. For example, in the northern part of Dixie Valley (fig. 5, box A), transmissivity estimates from four wells within 1 mi of each other ranged from 4,500 to 45,500 ft²/d (well log numbers 12565, 9620, 9621, and 9331 from table 4). These transmissivity differences were primarily controlled by differing amounts of fine-grained material in the formation adjacent to screened intervals.

Overall, basin-fill transmissivity estimates from Dixie Valley averaged 8,700 ft²/d (table 4, north and south of the playa averaged about 11,000 and 5,400 ft²/d, respectively), and are within the range of basin-fill estimates determined across the State of Nevada. Basin-fill transmissivity estimates from 106 aquifer and slug tests completed throughout Nevada ranged from 20 to 90,000 ft²/d and averaged 8,600 ft²/d (U.S. Geological Survey, 2010).

Occurrence and Movement of Groundwater in Basin-Fill Deposits

The occurrence and movement of groundwater in Dixie Valley is interpreted using a potentiometric water-surface map (pl. 1). The potentiometric map represents the elevation to which water will rise in a well casing (Lohman and others,

1972). In many instances, the water table is the potentiometric water level. When wells are screened in confining units, such as silts and clays, the potentiometric water level may rise above the depth at which groundwater is present in the aquifer in response to pressures beneath confining units. Potentiometric surface contours were developed from depth-to-water measurements and land-surface elevation from 54 wells within or near the boundary of the Dixie Valley study area (table 3; pl. 1). Most depth-to-water measurements were made during autumn 2009 by USGS personnel and Mahannah and Associates, LLC; additional wells were measured between winter 2010 and summer 2011 as permissions were granted and new wells were drilled. Depths to groundwater range from nearly 30 ft above land surface just south of the playa (artesian conditions, well 35) to more than 400 ft bls in the southernmost part of the valley (well 60). Some wells south of the playa penetrate both the water-table aquifer and one or more confined aquifers within the upper 250 ft of alternating gravel and clay; therefore, some water levels could represent a composite groundwater elevation rather than the water table. The potentiometric surface map (pl. 1) represents a current (about 2009) potentiometric surface of the Dixie Valley groundwater system, which is mostly unconfined.

The potentiometric water-level elevation contours depict the top of the groundwater surface and indicate general directions of groundwater flow in Dixie Valley (pl. 1). Historical and current groundwater pumping in the basin seem to have minimally affected regional water level elevations and groundwater-flow paths. Historical water levels were used to help guide the position and shape of water-level elevation contours in areas where current data were limited (table 3; pl. 1). In addition to measurements made during this study, 14 wells, 26 springs and 1 abandoned vertical mine shaft with historical water-level measurements from 1951 to 2009 were used to guide the construction of contour lines (table 3; pl. 1).

Groundwater generally moves radially from the mountain-range uplands toward the central valley lowlands, and eventually discharges near or along the playa edge (pl. 1). Groundwater flow generally is parallel with the topographic gradient from the southernmost and northernmost extents of the valley toward the playa. Flow patterns east and west of the playa indicate that groundwater is moving eastward from the Stillwater Range and westward from the Clan Alpine Mountains toward the playa. Harrill and Hines (1995, p. 10) hypothesized that the playa hydraulically separates groundwater north and south of Township 22N (pl. 1). The north-south and east-west trending flow patterns support this hypothesis and suggest that mixing between northern and southern groundwater systems in Dixie Valley likely is minimal.

Groundwater movement between the basin-fill aquifer and playa sediment is hydraulically constrained owing to physical and chemical differences between systems.

Groundwater flow in the playa and exchange with the basin-fill aquifer is physically limited because the playa sediment transmissivity is nearly 10,000 times less than that measured in the surrounding basin fill (table 4, well 27, screened in playa sediment), compared to the average of wells 34, 39, 43, and 44 (all screened in basin-fill). It should be noted this finding is based on a transmissivity estimate from a slug test from only one well screened down to 10 feet in depth in playa deposits, however playa sediments are assumed to become less transmissive with depth due to overburden pressure. Chemically, the groundwater interface between the basin fill and playa is analogous to coastal aquifers. Density contrasts between freshwater and saltwater systems in coastal aquifers suppress mixing (Barlow, 2003, p. 8–10). Groundwater density measurements were determined for many wells in this study using a standard equation that combines water temperature, total dissolved solids (TDS) concentration, and pressure. Basin-fill groundwater density averaged about 1.0 g/mL, whereas playa groundwater density averaged 1.16 g/mL and ranged from 1.14 to 1.21 g/mL (from sites 24, 27, and 29). Concentrations of TDS in the playa groundwater range from about 184,000 to 310,000 mg/L (average 247,000 mg/L from sites 24 and 29), which is 5–9 times greater than the TDS of seawater (about 35,000 mg/L). Playa groundwater TDS concentrations measured from well 62–66 ranged from 62,000 to 294,000 mg/L (average 212,000 mg/L; Mahannah and Associates, LLC, written commun., 2013). The high TDS of playa groundwater classifies it as a brine (Drever, 1982). Concentrations of TDS in the basin-fill aquifer ranges from 98 to 3,580 mg/L (average 637 mg/L, determined from all wells except 24 and 29 [table 6]). A sharp TDS contrast of 683 and 184,000 mg/L is shown across less than 2 mi separating well 30, located adjacent to the southern playa, and well 29 (fig. 4), in the southern playa, respectively. Substantial measured differences in transmissivity, density, and TDS between the basin fill and playa aquifer systems create an even sharper freshwater-saltwater interface than in coastal regions. Some groundwater exchange likely occurs between the basin fill and playa; however, this exchange occurs at substantially lower rates than groundwater flow within the basin-fill aquifer.

Fresh groundwater in closed-desert basins is known to discharge near playa edges (Toth, 1962; Duffy and Al-Hassan, 1988, p. 1678). In addition to numerous adjacent springs, the Dixie Valley playa is surrounded by phreatophytic vegetation on all sides. These phreatophytes, composed predominantly of greasewood (*Sarcobatus vermiculatus* [Hooker] Torrey), are sustained by groundwater and deep soil water (within about 60 ft bls; Robinson, 1958) in addition to precipitation during the growing season. Discharge of groundwater by springs and phreatophytic vegetation surrounding the playa likely constitutes most fresh groundwater discharge from the flow system. Discharge upgradient from and along the edges of the playa provides additional evidence of minimal mixing between fresh groundwater and the playa brine.

Table 6. Water-quality data from sites sampled in Dixie Valley, west-central Nevada, 2009–11.

[Site No.: Locations of sites are shown in [figure 7](#); “L.” from Lahontan GeoScience, Inc. (2004), table 2; “C.” from Goff and others (2002), table 1; “B” from Great Basin Center for Geothermal Energy (GBCGE) (2009); “S” spring. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Abbreviations:** BLM, Bureau of Land Management; ft bls, foot below land surface; °C, degrees Celsius; mg/L, milligram per liter; µg/L, microgram per liter; n/a, not applicable; –, no data]

Site No.	USGS standard identification No.	Project site name	Data source	Site type	Total well depth (ft bls)	Latitude	Longitude	Sample date	Temperature (°C)	Total dissolved solids (mg/L)	Arsenic (µg/L)	Fluoride (mg/L)	Lithium (µg/L)	Magnesium (mg/L)	Boron (mg/L)
Northern sites															
'03	400753117334401	USBLM McCoy Well	This study	Well	200	40 07 54.15	117 33 47.75	10-23-09	19.3	477	14	0.55	31.4	13.7	0.3
'04	400600117380001	Little McCoy Ranch	This study	Well	114	40 06 01.92	117 38 28.87	10-29-09	17.5	910	9	2.24	488	31	2.1
'07	400421117434701	Seven Devil	This study	Well	134	40 04 20.67	117 43 49.65	11-03-09	25.4	1,070	13.8	4.65	604	24.8	1.7
'08	400419117433401	Seven Devils Ag Well #3	This study	Well	270	40 04 19.30	117 43 33.83	11-03-09	22.3	1,060	9.4	4.22	553	24.1	1.5
'09	400416117424401	Ag Well #1	This study	Well	350	40 04 16.55	117 42 50.50	05-26-10	28.9	915	11.8	0.83	231	34.8	1.0
'12	395901117483401	Terra Gen I-B	This study	Well	70	39 59 01.00	117 48 33.92	11-02-09	15.5	638	23.1	0.74	99.1	23.8	0.9
'14	395800117510901	73W-7	This study	Well	300	39 57 57.07	117 51 13.10	05-12-10	28.2	1,070	9.9	1.13	2597	50	2.9
'15	395747117510901	Blackwater	This study	Well	160	39 57 46.26	117 51 13.93	11-04-09	21.7	1,060	6.9	1.01	637	41.2	2.6
'16	395740117521401	Irrigation	This study	Well	300	39 57 40.50	117 52 17.21	11-04-09	34.4	613	18	0.76	154	27.1	0.6
'17	395722117471601	West pair Ag well	This study	Well	330	39 57 22.12	117 47 16.83	05-25-10	19	784	11.2	1.08	192	23.1	1.2
'19	395716117503602	Terra Gen II-A	This study	Well	50	39 57 16.14	117 50 36.38	10-22-09	15.2	3,580	31.8	1.15	844	95.5	2.6
'20	395716117503601	Terra Gen II-B	This study	Well	70	39 57 16.14	117 50 36.38	10-22-09	15.5	808	21.9	1.16	235	37.2	1.0
'22	395503117513601	Terra Gen III-B	This study	Well	70	39 55 03.21	117 51 35.66	11-02-09	15.5	776	40	1.41	235	22.2	1.1
'23	395139117501801	DVMV	This study	Well	41	39 51 39.26	117 50 18.31	11-19-09	15.4	658	10	0.28	31	44.4	0.3
24	395123117594802	DVPW Shallow	This study	Well	2.5	39 51 19.75	117 59 40.21	11-04-09	–	310,000	–	20.5	–	<1.20	–
'S20	395429117591501	PW spring	This study	Spring	n/a	39 54 28.54	117 59 15.55	10-23-09	22.4	2,450	35.8	0.75	270.8	51.3	1.2
'S23	394631118043301	Cold Spring	This study	Spring	n/a	39 46 31.31	118 04 32.83	10-27-09	29	829	5.3	10.2	2379	0.587	1.1
'S25	394813118031101	Northern meadows spring	This study	Spring	n/a	39 48 13.09	118 03 10.94	10-30-09	26	565	34.7	12.1	2219	1.34	1.1
'S28	395322118040101	Dixie Valley hot springs	This study	Spring	n/a	39 47 41.83	118 04 04.96	10-23-09	58.5	n/a	n/a	n/a	2306	0.141	1.0
'S30	395959117423101	Hyder Hot Springs	This study	Spring	n/a	39 59 58.69	117 42 34.43	10-25-09	60.9	n/a	n/a	n/a	21,086	9.23	4.4
'S32	400155117374801	North-east spring	This study	Spring	n/a	40 01 55.39	117 37 47.77	10-30-09	21.8	597	7.5	3.44	2256	14.4	1.1
'S33	400205117361101	Hot Sp, Lower Ranch	This study	Spring	n/a	40 02 04.70	117 36 14.44	10-24-09	30.2	563	6.9	3.32	203	13	1.2
'S39	400448117361301	McCoy Spring	This study	Spring	n/a	40 04 47.69	117 36 16.42	10-24-09	44.8	894	12	1.66	2167	29.4	0.8
'S45	400517117432801	Seven Devil Hot spring	This study	Spring	n/a	40 05 16.69	117 43 31.43	10-24-09	62.1	n/a	n/a	n/a	2557	20.7	1.4
'L200	–	LGS sample	Lahontan Geoscience, Inc., 2004	Well	40	39 48 05.67	117 54 59.54	02-13-04	16.7	n/a	11	1.1	n/a	n/a	–
'L233	–	LGS sample DV259	Lahontan Geoscience, Inc., 2004	Well	–	39 49 21.10	117 54 07.24	02-13-04	18	390	6	0.26	n/a	n/a	–
'L235	–	DV262	Lahontan Geoscience, Inc., 2004	Well	67	39 51 31.59	117 48 51.28	02-10-04	–	780	32	0.85	n/a	n/a	–
'L285	–	DV2320	Lahontan Geoscience, Inc., 2004	Well	55	39 55 50.52	117 47 40.78	02-09-04	15	1,100	5.6	<0.10	n/a	n/a	–
'G49	–	Hole in the Wall #2 Well	Goff and others, 2002	Well	–	39 53 19.68	117 33 22.32	11-04-97	13.7	400	1	0.49	30	5.62	0.4
'G111	–	62-21 Well	Goff and others, 2002	Well	–	39 55 58.08	117 49 11.28	10-28-98	11.5	2,011	300	6.60	420	1.04	1.6
'G106	–	Stu's Seep	Goff and others, 2002	Spring	n/a	39 56 42.00	117 55 54.48	04-30-98	17.2	1,196	10	0.58	70	47.2	0.3
'G210	–	Dead Travertine Spring, Road Seep	Goff and others, 2002	Spring	n/a	39 56 44.52	117 53 49.92	05-08-99	19	220	4.2	2.46	228	90.4	1.5
'G1	–	Domestic Well	Goff and others, 2002	Well	–	39 57 56.88	117 51 20.16	10-31-97	29.2	220	1,650	0.78	420	28.8	1.7

Table 6. Water-quality data from sites sampled in Dixie Valley, west-central Nevada, 2009–11.—Continued

[Site No.: Locations of sites are shown in [figure 7](#); “L” from Lahontan GeoScience, Inc. (2004), table 2; “C” from Goff and others (2002), table 1; “B” from Great Basin Center for Geothermal Energy (GBCGE) (2009); “S” spring. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Abbreviations:** BLM, Bureau of Land Management; ft bls, foot below land surface; °C, degrees Celsius; mg/L, milligram per liter; µg/L, microgram per liter; n/a, not applicable; –, no data]

Site No.	USGS standard identification No.	Project site name	Data source	Site type	Total well depth (ft bls)	Latitude	Longitude	Sample date	Temperature (°C)	Total dissolved solids (mg/L)	Arsenic (µg/L)	Fluoride (mg/L)	Lithium (µg/L)	Magnesium (mg/L)	Boron (mg/L)
Northern sites—Continued															
G39	–	Goerenger Well	Goff and others, 2002	Well	–	39 58 09.48	117 51 32.76	05-04-99	27.7	260	12	1.21	930	47.8	2.8
G59	–	45-W-5 Well	Goff and others, 2002	Well	–	39 58 12.72	117 51 29.52	11-05-97	26.4	715	2	1.95	680	3.02	1.6
G122	–	97-2 Well	Goff and others, 2002	Well	–	39 58 55.56	117 50 26.16	05-05-98	19.7	1,689	23	3.73	1,450	47.5	6.7
G123	–	32-6 Well	Goff and others, 2002	Well	–	39 59 54.24	117 50 46.68	05-06-98	32	910	2.4	1.55	70	64.1	1.3
G47	–	Sou Hot Springs	Goff and others, 2002	Spring	n/a	40 05 20.40	117 43 26.40	05-04-98	72	n/a	n/a	n/a	10	20.7	1.3
G57	–	Bolivia Artesian Well	Goff and others, 2002	Well	–	39 59 51.72	117 54 56.52	11-05-97	28.8	n/a	n/a	n/a	160	66.8	0.4
G3	–	45-33 Well Archived	Goff and others, 2002	Well	–	39 59 22.92	117 49 33.60	08-21-86	–	n/a	n/a	n/a	2,630	0.04	8.0
G122-1	–	Dixie Jack Gradient Well #1	Goff and others, 2002	Well	–	39 59 18.24	117 50 33.72	05-17-98	49	n/a	n/a	n/a	165	5.3	7.1
G53	–	46-32 Well (Stu's Well)	Goff and others, 2002	Well	–	39 59 17.16	117 50 36.24	11-05-97	155	n/a	n/a	n/a	20	0.04	0.1
G56	–	Dead Travertine Spring, Upper	Goff and others, 2002	Spring	n/a	39 59 16.44	117 53 34.44	05-07-99	19.2	n/a	n/a	n/a	2,050	76	1.4
G122-7	–	Dixie Jack Gradient Well #7	Goff and others, 2002	Well	–	39 59 14.64	117 50 54.24	05-14-98	55	n/a	n/a	n/a	90	0.18	1.2
G25	–	27-33 Well	Goff and others, 2002	Well	–	39 59 12.84	117 49 50.52	04-29-98	174	n/a	n/a	n/a	2,270	0.01	11.4
G54	–	27-32 Well (Dick's Well)	Goff and others, 2002	Well	–	39 59 10.32	117 50 54.24	11-05-97	144	n/a	n/a	n/a	10	0.01	0.1
G29	–	37-33 Well	Goff and others, 2002	Well	–	39 59 09.96	117 49 52.32	04-28-98	165	n/a	n/a	n/a	2,210	0.01	9.0
G122-4	–	Dixie Jack Gradient Well #4	Goff and others, 2002	Well	–	39 59 06.72	117 50 46.68	05-20-98	77	n/a	n/a	n/a	1,170	0.57	8.1
G30	–	28-33 Well	Goff and others, 2002	Well	–	39 59 06.72	117 49 54.48	04-28-98	165	n/a	n/a	n/a	2,210	0.01	9.2
G168	–	38-32 Well	Goff and others, 2002	Well	–	39 59 03.48	117 50 49.20	10-26-98	87.7	n/a	n/a	n/a	980	1.08	7.3
G20	–	82A-7 Well	Goff and others, 2002	Well	–	39 58 03.00	117 51 17.28	05-05-99	–	n/a	n/a	n/a	2,250	0.01	11.8
G11	–	73-7 Well	Goff and others, 2002	Well	–	39 57 55.80	117 51 27.36	05-04-99	–	n/a	n/a	n/a	2,210	0.01	0.0
G23	–	73B-7 Well	Goff and others, 2002	Well	–	39 57 54.72	117 51 19.80	04-29-98	174	n/a	n/a	n/a	2,270	0.01	11.4
G86	–	63-7 Well	Goff and others, 2002	Well	–	39 57 54.00	117 51 28.80	10-22-98	166	n/a	n/a	n/a	2,210	0.01	11.7
G13	–	84-7 Well	Goff and others, 2002	Well	–	39 57 50.04	117 51 29.52	10-29-97	–	n/a	n/a	n/a	2,450	0.01	11.5
G14	–	74-7 Well	Goff and others, 2002	Well	–	39 57 49.32	117 51 28.80	05-04-99	–	n/a	n/a	n/a	2,320	0.33	11.5
G8	–	76-7 Well (V104 Separator)	Goff and others, 2002	Well	–	39 57 32.40	117 51 25.20	04-28-98	168	n/a	n/a	n/a	2,280	0.01	9.4
G104	–	66-21 Well	Goff and others, 2002	Well	–	39 55 51.96	117 55 42.24	04-30-98	57.4	n/a	n/a	n/a	489	0.35	5.8
G102	–	45-14 Well	Goff and others, 2002	Well	–	39 51 57.24	118 00 17.64	04-30-98	123.5	n/a	n/a	n/a	70	0.02	0.5
G69	–	Dixie Hot Spring	Goff and others, 2002	Spring	n/a	39 48 01.44	118 03 33.12	05-05-98	83.5	n/a	n/a	n/a	420	0.22	1.0
G51	–	Old Man Main Spring	Goff and others, 2002	Spring	n/a	39 48 00.00	117 39 57.60	11-04-97	10.8	n/a	n/a	n/a	80	1.38	0.7
B2	–	JS Ranch Well, McCoy Ranch	GBCGE, 2009	Well	–	40 05 06.76	117 36 39.06	07-23-63	21	–	–	1.9	–	0.02	1.1
B3	–	Brinkerhoff Well, Dixie Landing Strip	GBCGE, 2009	Well	–	40 05 09.35	117 40 40.30	04-05-98	16.8	–	5.2	0.57	310	0.05	1.1

Table 6. Water-quality data from sites sampled in Dixie Valley, west-central Nevada, 2009–11.—Continued

[Site No.: Locations of sites are shown in [figure 7](#); “L” from Lahontan GeoScience, Inc. (2004), table 2; “G” from Goff and others (2002), table 1; “B” from Great Basin Center for Geothermal Energy (GBCGE) (2009); “S” spring. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Abbreviations:** BLM, Bureau of Land Management; ft bis, foot below land surface; °C, degrees Celsius; mg/L, milligram per liter; µg/L, microgram per liter; n/a, not applicable; –, no data]

Site No.	USGS standard identification No.	Project site name	Data source	Site type	Total well depth (ft bis)	Latitude	Longitude	Sample date	Temperature (°C)	Total dissolved solids (mg/L)	Arsenic (µg/L)	Fluoride (mg/L)	Lithium (µg/L)	Magnesium (mg/L)	Boron (mg/L)
Northern sites—Continued															
B4	–	Seven Devils Hot Spring, Sou Hills-5	GBCGE, 2009	Spring	n/a	40 05 20.04	117 43 30.22	11-03-97	72.6	n/a	n/a	n/a	720	0.02	1.3
B5	–	Seven Devils Hot Spring, Sou Hills-2	GBCGE, 2009	Spring	n/a	40 05 14.35	117 43 32.74	10-01-59	76.7	n/a	n/a	n/a	670	0.02	1.2
B6	–	North Injection Well DF 45-14, Boyer Ranch	GBCGE, 2009	Well	–	39 58 37.25	117 51 01.29	04-30-98	107	n/a	n/a	n/a	2,500	0.02	11.4
B7	–	Geothermal Plant Brine, Boyer Ranch	GBCGE, 2009	Well	–	39 58 00.17	117 51 25.77	04-29-98	41	n/a	n/a	n/a	261	0.15	0.4
B8	–	Geothermal Separator V105, Boyer Ranch	GBCGE, 2009	Well	–	39 57 58.37	117 51 16.77	04-29-98	–	n/a	n/a	n/a	232	–	0.1
B9	–	Injection Well Lamb 1, Boyer Ranch	GBCGE, 2009	Well	–	39 57 12.71	117 52 12.22	10-24-96	107	n/a	n/a	n/a	252	0.01	12.6
B10	–	South Injection Well 32-1, Boyer Ranch	GBCGE, 2009	Well	–	39 57 11.93	117 51 49.90	10-31-97	102	n/a	n/a	n/a	253	–	12.0
B11	–	Dixie Valley Well, Boyer Ranch	GBCGE, 2009	Well	–	39 56 59.69	117 51 33.44	05-29-80	–	n/a	n/a	n/a	–	0.03	7.9
B12	–	South Injection Well 65-1, Boyer Ranch	GBCGE, 2009	Well	–	39 56 48.53	117 51 42.69	10-31-97	108	n/a	n/a	n/a	2,610	0.01	12.1
B13	–	Geothermal Well DF 66-21, Boyer Ranch	GBCGE, 2009	Well	–	39 56 21.26	117 49 23.38	05-01-98	75.5	n/a	n/a	n/a	490	0.41	5.5
B14	–	Dixie Hot Springs, Dixie Meadows	GBCGE, 2009	Spring	n/a	39 48 01.14	118 03 36.59	06-19-73	72	n/a	n/a	n/a	380	0.02	0.9
Southern sites															
'25	394618117560601	DVHVb	This study	Well	29.5	39 46 18.03	117 56 06.34	11-04-09	15.3	478	41.3	1.19	30.8	11.4	0.623
'26	394617117483401	BLM Bernice Cr at mtn fr	This study	Well	139	39 46 17.15	117 48 34.29	11-05-09	16.7	658	1.4	0.29	25.4	59.4	0.307
29	394508118025503	DVPI Composite	This study	Well	10	39 45 07.79	118 02 54.63	11-03-09	21.4	184,000	13,800	26.3	104	1.4	112
'30	394348118040201	DVLV	This study	Well	24.5	39 43 48.38	118 04 02.15	11-02-09	15.8	683	78.4	5.97	6.6	0.527	0.57
'31	394239118063901	Knittle Well	This study	Well	213	39 42 38.86	118 06 39.20	10-22-09	15.4	298	13	0.91	32.2	2.67	0.201
'32	392748118042501	NWIS 2	This study	Well	72	39 42 32.22	118 04 30.12	10-23-09	–	308	15.3	1.49	36.3	2.59	0.304
'34	394217118033801	Navy 31B	This study	Well	194	39 42 16.75	118 03 37.90	10-22-09	16.2	306	20.9	5.95	46.9	1.45	0.338
'35	394205118035001	16E	This study	Well	158	39 42 06.97	118 04 03.05	10-22-09	17.4	312	17.2	5.89	61.7	1.05	0.352
'36	392708118042301	17L	This study	Well	–	39 42 06.18	118 04 35.65	05-11-10	15	310	19.4	3.48	36.1	2.12	0.334
'38	394142118025101	Navy 33A	This study	Well	260	39 41 41.52	118 02 50.76	10-22-09	20	342	21.2	5.64	51.5	1.12	0.38
'41	394114118034001	56A	This study	Well	200	39 41 14.77	118 03 52.77	10-22-09	19.4	282	19.4	7.39	70.6	0.47	0.295
'42	394048118053001	81H	This study	Well	171	39 40 49.11	118 05 40.14	10-21-09	19.4	306	17.7	5.31	28.4	1.99	0.36

Table 6. Water-quality data from sites sampled in Dixie Valley, west-central Nevada, 2009–11.—Continued

[Site No.: Locations of sites are shown in [figure 7](#); “L” from Lahontan GeoScience, Inc. (2004), table 2; “C” from Goff and others (2002), table 1; “B” from Great Basin Center for Geothermal Energy (GBCGE) (2009); “S” spring. **Latitude and longitude:** Shown in degrees, minutes, and seconds. **Abbreviations:** BLM, Bureau of Land Management; ft bls, foot below land surface; °C, degrees Celsius; mg/L, milligram per liter; µg/L, microgram per liter; n/a, not applicable; —, no data]

Site No.	USGS standard identification No.	Project site name	Data source	Site type	Total well depth (ft bls)	Latitude	Longitude	Sample date	Temperature (°C)	Total dissolved solids (mg/L)	Arsenic (µg/L)	Fluoride (mg/L)	Lithium (µg/L)	Magnesium (mg/L)	Boron (mg/L)
Southern sites—Continued															
I43	392602118050101 84B		This study	Well	150	39 40 39.82	118 05 10.10	10-22-09	19.9	304	26.5	7.53	35.4	1.55	0.349
I44	394037118040901 Navy 86A		This study	Well	184	39 40 36.52	118 04 08.97	10-22-09	20.5	278	23.2	7.58	66.9	0.44	0.29
I50	393840118050001 USBLM Corral		This study	Well	45	39 38 26.00	118 04 44.52	05-27-10	18	98	0.14	4.38	² 30.5	0.55	0.092
I53	393420118074001 108B		This study	Well	390	39 34 39.56	118 07 39.02	05-24-10	—	316	68.6	8.24	43.7	0.58	0.384
I55	393224118091001 Nr Horse Creek Rd		This study	Well	500	39 32 24.04	118 09 10.39	06-02-11	41.4	318	29.7	6.17	35.1	1.27	0.365
I56	392819118092501 Dixie HWY S Nr Marker 13		This study	Well	519	39 28 19.16	118 09 25.46	05-25-11	26.3	331	17.1	4.06	44.4	2.48	0.367
I60	391900118085801 Wonder Well		This study	Well	500	39 18 59.96	118 08 58.35	05-18-11	22.2	469	7	0.74	36.7	17	0.348
61	390234118070701 State of Nevada		This study	Well	288	39 17 19.81	118 06 34.67	05-24-10	—	194	3.6	0.4	15.9	1.68	0.154
I53	393119118005001 Horse Ck Ranch Spg		This study	Spring	n/a	39 21 19.47	118 00 49.59	10-30-09	13.8	102	1.3	0.20	² 5.44	1.91	0.084
I58	393755118103901 Willow Spring		This study	Spring	n/a	39 37 54.95	118 10 39.38	10-27-09	15.3	289	1.2	0.41	² 16.0	6.94	0.248
I511	394000118020101 Eastern Settlement spring		This study	Spring	n/a	39 40 00.43	118 02 01.33	10-28-09	6.9	637	16.5	10.4	² 116	5.17	0.333
I513	394120118033901 Settlement spring		This study	Spring	n/a	39 41 20.23	118 03 38.90	10-28-09	4.3	419	12.0	7.7	² 88.2	1.97	0.282
I516	394251118121401 Alameda Canyon Spg		This study	Spring	n/a	39 42 50.57	118 12 14.17	10-27-09	8.7	1,790	15.6	1.56	² 77.8	33.2	1.24
I518	394515117595201 Southeast playa spring		This study	Spring	n/a	39 45 14.89	117 59 52.12	10-25-09	10.3	2,880	517	49.8	² 43.2	2.57	4.22
I522	394601117584101 Buckbrush spring		This study	Spring	n/a	39 46 01.26	117 58 41.32	10-25-09	18.6	307	0.74	3.92	² 45.0	1.16	0.324
I4	— North Well	Lahontan GeoScience, Inc., 2004		Well	—	39 20 02.56	118 12 36.29	01-30-04	26.6	380	20	1.60	n/a	n/a	—
I4	— NV Paving/DV15	Lahontan GeoScience, Inc., 2004		Well	300	39 23 25.18	118 09 52.11	01-23-04	20	360	5.6	0.41	n/a	n/a	—
I4	— Parcel 098/DV98	Lahontan GeoScience, Inc., 2004		Well	501	39 38 30.95	118 08 33.75	1983	<20	324	5	1.21	n/a	n/a	0.2
I4	— Persinger/DV150	Lahontan GeoScience, Inc., 2004		Well	163	39 40 52.52	118 02 33.39	1981	<20	309	11	6.16	n/a	n/a	0.61
I4	— A1-A6 DRI	Lahontan GeoScience, Inc., 2004		Well	—	39 42 1.33	117 58 02.73	06-25-79	20	353	—	—	n/a	n/a	—
I4	— Shaw Well	Goff and others, 2002		Well	—	39 40 52.32	118 03 01.08	11-07-97	19.2	336	29	7.21	60	0.74	0.32
I4	— Lofthouse Spring	Goff and others, 2002		Spring	n/a	39 44 01.32	117 49 52.32	07-05-98	14.8	n/a	n/a	n/a	10	15.6	0.25
I4	— Not-So-OK Spring	Goff and others, 2002		Spring	n/a	39 43 09.12	117 48 51.48	10-27-98	10	n/a	n/a	n/a	10	13.9	0.17
I4	— Upper Cherry Spring	Goff and others, 2002		Spring	n/a	39 34 53.40	117 56 38.04	10-28-98	7.4	n/a	n/a	n/a	10	5.44	0.1
I4	— E.H. Stark Well, Cattle Road	GBCGE, 2009		Well	—	39 38 10.46	118 06 54.97	03-01-73	22.8	—	—	6	—	0.002	5.48

¹Samples used in current study analysis; remaining samples are “temperature hot” and likely derived from geothermal sources or did not have enough information to water type, but were used in lithium and magnesium geothermal analysis.

²Lithium concentration corrected for erroneous laboratory method ([appendix B](#)).

³Estimate based on cation and anion calculation.

Water-Level Change

Groundwater levels in Dixie Valley have remained generally constant since the 1950s, but some localized change has occurred as a result of withdrawals for irrigation, livestock, domestic use, and augmentation of geothermal reservoir pressure. A comparison between the potentiometric surface map delineated in this study and the historical water-level elevation map from Cohen and Everett (1963) indicate that water levels have remained generally constant. In mostly undeveloped areas, water level differences between water-level maps were mostly attributed to a more spatially extensive dataset used in this study; however, in historically developed areas differences in water levels were reflective of the localized development. Generally undeveloped areas lacked sufficient data for evaluation. Mean decadal water levels for 19 wells are summarized in [table 7](#) and represent areas of groundwater development. The earliest water-level measurement was made in mid-1951, and the most recent measurement was made in mid-2011. Each of the 19 wells had a minimum of 7 measurements spanning at least three decades; however, the average number of total measurements for any one well was 31. Sufficient measurements were available to evaluate water-level change for eight wells across various time periods ([fig. 6](#); [table 7](#)). Although water-level measurements representative of differing time periods are limited, some observations are provided here.

Water-level change measured in wells north of the playa is likely due to irrigation and geothermal pumping (wells 3–5, 7–10, and 15–20; [fig. 4](#)). Some water levels in the northernmost wells of Dixie Valley showed rising and declining patterns from the 1950s to 2011, corresponding to changes in agricultural pumping. For example, water levels in wells near the border of Dixie and Jersey Valleys (wells 3–5; [fig. 4](#)) declined from the late 1960s to early 1980s by about 15 ft, and then rose by as much as 5 ft between measurements made in the 1980s and the 2000s ([fig. 6A](#)). The declining and subsequent rising patterns likely correspond to the cessation of agricultural pumping. Water levels in wells between the playa and the Churchill-Pershing County line (wells 15–20; [fig. 4](#)) declined by about 3–6 ft between the 1980s and early 2010s (wells 17, 18, [table 7](#); [fig. 6B](#)). Large-scale pumping to augment geothermal injection began in 1997 (Benoit and others, 2000). Several wells clustered near the Dixie Valley Geothermal Power plant are used to augment and monitor the injection of basin-fill water above the geothermal aquifer (wells not shown in [fig. 6B](#)). Wells farther east of the geothermal plant (wells 17 and 18) are used for irrigated agriculture.

Water levels measured in wells south of the playa, where the basin-fill aquifer is semi-confined (wells 45, 47, 50–52, [fig. 4](#); [table 7](#)), have fluctuated by less than 4 ft since the 1950s and generally have risen by 0.7–3.0 ft between the 1980s and early 2010s (for example, see wells 47 and 50, [fig. 6C](#)). Many wells in this area were drilled to support an agricultural settlement that was present from the 1940s to mid-1980s. Between 1985 and 1987, the U.S. Navy purchased private land in this settlement area through a congressionally approved buyout of 12,000 acres (Misrach, 1990) to use as a supersonic testing ground. This action, in turn, resulted in the dismantling of most of the settlement homesteads, leaving many artesian wells uncapped and flowing for more than 20 years. Between 2002 and 2010, the U.S. Navy capped or sealed 65 wells, 45 of which were artesian (Gary Cottle, U.S. Navy, written commun., 2012). Fifteen of the 23 remaining wells in the area surveyed for this study are artesian ([table 3](#), wells 30–52, noted by footnote; [fig. 4](#)), with water levels ranging from about 0.4 ft bls to nearly 32 ft above land surface (October 2009–August 2011). Artesian well depths range from 24.5 to 292 ft bls (average about 180 ft bls), whereas non-artesian wells range in depth from 45 to 200 ft bls (wells 45, 47–52; average about 130 ft bls). Rising water levels in non-artesian wells likely indicate aquifer recovery from reduced agricultural pumping and capping of flowing wells following the U.S. Navy land acquisition.

Water-level data from wells in southern Dixie Valley (south of latitude 39°35'0"N, [fig. 4](#); [table 3](#)) are sparse. Well 88 has a lengthy water-level record and shows a decline in water-levels of about 3.5 ft from the late 1980s to the late 2000s ([fig. 6D](#)). The water-level decline is possibly from groundwater withdrawals for livestock watering.

Subsurface Flow

Groundwater in basin-fill sediments moves northward from Fairview Valley and southward from Jersey Valley into Dixie Valley as subsurface inflow. Groundwater historically was thought to move southward from Pleasant Valley to Dixie Valley, as well as from Eastgate Valley to Cowkick Valley, from Cowkick Valley to Stingaree Valley, and, finally, as a combined flow entering Dixie Valley (Cohen and Everett, 1963, p. 17). Previous subsurface-inflow estimates from adjacent basins were based on residual differences between estimated recharge and discharge (Cohen and Everett, 1963; Harrill and Hines, 1995). In this study, subsurface inflow from adjacent basins through the basin-fill aquifer was estimated using Darcian flux calculations, which incorporate local aquifer transmissivity, current hydraulic gradients, and effective flow width near basin divides ([table 8](#)).

Table 7. Wells with seven or more water-level measurements, Dixie Valley, west-central Nevada, 1950–2011.[Site No.: Locations of sites are shown in [figure 4](#). ft bls, foot below land surface; –, no data]

Site No.	Number of water levels measured within decade							Average decadal water level (ft bls)						
	1950s	1960s	1970s	1980s	1990s	2000s	2010–11	1950s	1960s	1970s	1980s	1990s	2000s	2010–11
¹ 03	–	–	–	3	–	1	6	–	–	–	136.0	–	131.0	130.8
¹ 04	–	2	5	7	–	1	7	–	76.5	84.0	91.7	–	89.6	89.1
¹ 05	–	–	–	5	–	1	6	–	–	–	96.2	–	96.0	95.1
08	–	–	1	–	–	2	6	–	–	67.0	–	–	91.3	93.6
09	–	1	–	6	–	1	–	–	92.0	–	101.2	–	102.1	–
10	–	–	–	6	–	1	–	–	–	–	92.4	–	94.9	–
15	–	–	1	3	–	2	1	–	–	40.0	31.2	–	49.8	56.8
16	–	–	2	–	–	2	6	–	–	125.0	–	–	133.9	134.2
¹ 17	–	–	–	5	–	1	7	–	–	–	58.7	–	61.1	61.3
¹ 18	–	–	1	4	–	1	1	–	–	57.0	54.1	–	61.4	59.9
19	–	–	–	1	–	2	6	–	–	–	6.0	–	15.0	14.5
20	–	–	–	1	–	2	6	–	–	–	6.0	–	14.3	14.1
45	1	–	–	2	–	2	6	72.9	–	–	73.3	–	73.3	72.2
¹ 47	15	13	9	6	6	1	6	28.3	29.7	31.5	33.2	32.1	30.6	30.1
¹ 50	18	14	8	6	6	2	1	34.0	33.8	34.7	35.1	34.8	34.4	34.4
51	–	–	1	2	–	1	8	–	–	38.0	37.3	–	34.7	34.9
52	–	–	–	1	–	1	6	–	–	–	37.6	–	34.3	34.4
53	–	1	–	3	–	1	6	–	126.0	–	131.4	–	128.8	128.9
¹ 88	–	–	–	4	7	65	–	–	–	208.9	211.4	210.6	211.2	–

¹Site used in groundwater-level trend analysis ([fig. 6](#)).**Table 8.** Annual subsurface inflow estimates from adjacent basins through basin-fill deposits to Dixie Valley, west-central Nevada.

[All values in acre-feet per year]

Annual subsurface inflow estimates through basin-fill sediments to Dixie Valley			
	Cohen and Everett (1963)	Harrill and Hines (1995)	This study
Fairview Valley	500	2,300	700–1,300
Eastgate, Cowkick, Stinagree Valleys	5,600	6,300	– ¹
Jersey Valley	500	1,100	1,800–2,300
Pleasant Valley	800	1,100	– ²
Total	7,400	10,800	2,500–3,600

¹Not estimated; water table below basin-fill sediments at basin boundary.²Not estimated; no appreciable connection of basin-fill sediments between valleys.

Harrill and Hines (1995, p. 11) also hypothesized that groundwater might flow from Eastgate, Cowkick, and Stinagree Valleys to Dixie Valley, through fractured volcanic rocks beneath the Clan Alpine Mountains, and enter Dixie Valley somewhere in the northern part of the basin. This hypothesis was prompted by discrepancies between recharge and discharge estimates in northern and southern areas of Dixie Valley. There is no conclusive evidence to date (2014) supporting groundwater movement through more than 20 mi of fractured volcanic rocks beneath the mountain block. Subsurface inflow volumes reported in this study only represent flow through basin-fill deposits.

Subsurface inflow to Dixie Valley is the predominant form of groundwater discharge in Fairview Valley. Groundwater flows from Fairview Valley to Dixie Valley through a wide section of basin-fill deposits (about 13 mi) across a slight hydraulic gradient (9.3×10^{-4} ft/ft between sites 59 and 58). Using basin-fill transmissivity estimated from both the aquifer test for well 60 ([table 4](#)) and the specific-capacity data from well 59 (Nevada driller's log 9665, specific capacity = 6 gpm/ft, estimated transmissivity = 1,360 ft²/d) to estimate a range, the annual subsurface inflow from Fairview Valley to Dixie Valley ranges from 700 to 1,300 acre-ft/yr ([table 8](#)). This range falls between estimates by Cohen and Everett (1963) and Harrill and Hines (1995).

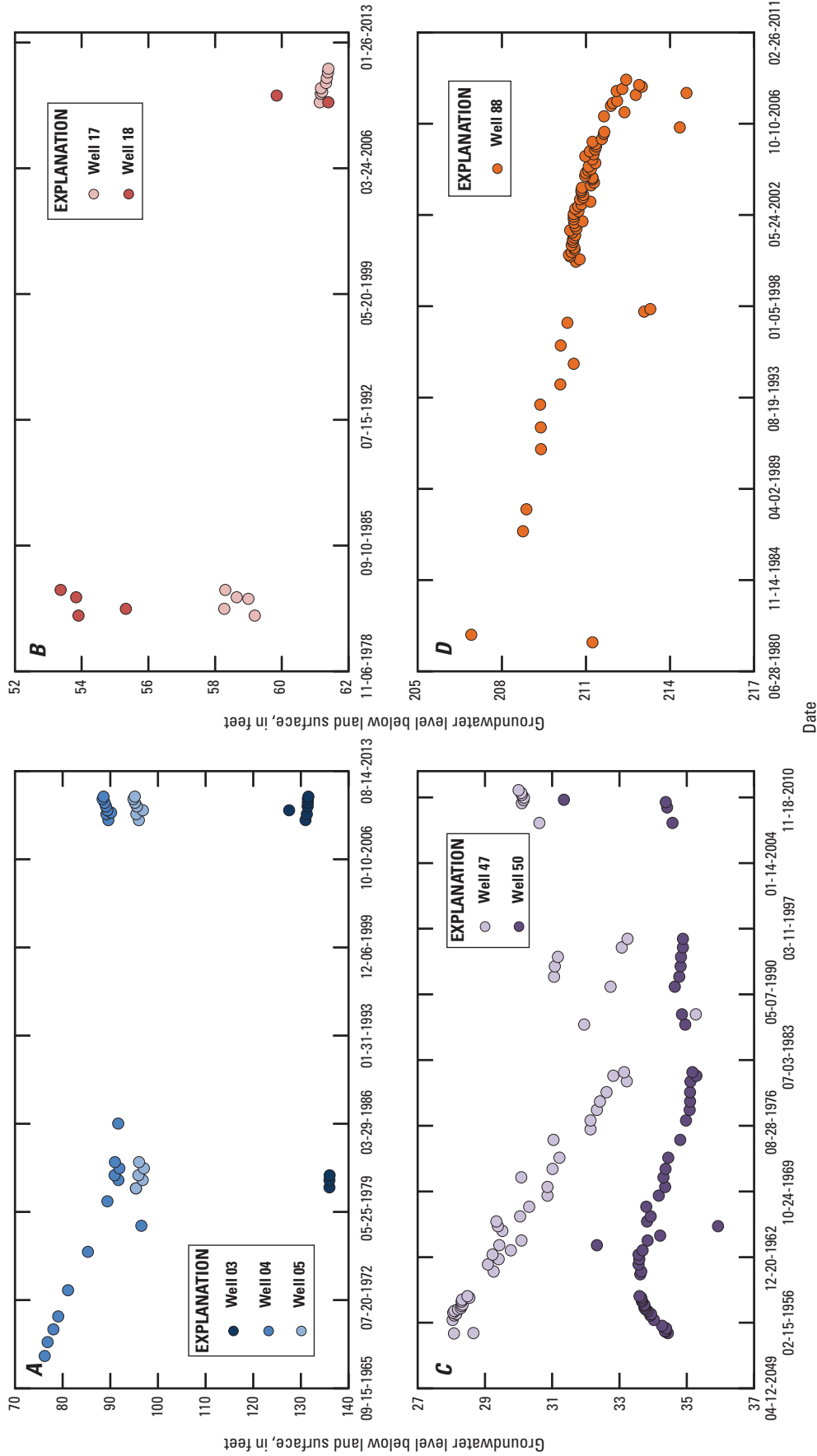


Figure 6. Water-level change as indicated by intermittent, selected water-level measurements from eight wells (A) in the northernmost basin, (B) north of playa near Dixie Valley Geothermal Plant, (C) just south of playa, and (D) in southernmost basin, Dixie Valley, west-central Nevada, 1955–2011. Well locations are shown in [figure 4](#).

The basin-fill deposits between Stingaree Valley and Dixie Valley are constricted (less than 3.2 mi) with a steep hydraulic gradient (1.4×10^{-2} ft/ft between sites 61 and 60) as a result of consolidated rock outcrops within the basin boundary area (pl. 1). Groundwater possibly could enter Dixie Valley through four separate sections of basin fill between consolidated rock outcrops. However, based on available groundwater data and topographic gradients, the most likely location for groundwater to enter Dixie Valley is through the basin-fill section containing Dixie Valley Wash (pl. 1). The three remaining basin-fill sections to the northeast and southwest are topographically higher, likely have higher bedrock elevations and thinner basin fill, and, therefore, are less likely to have groundwater flow across the boundary. Andesite tailings adjacent to a dry, abandoned mine shaft immediately northeast of Dixie Valley Wash (site 90) indicate bedrock is present near the mine base at about 136 ft bls. The presence of bedrock above the water table indicates that subsurface flow between the two boundaries is not occurring through overlying basin-fill deposits (pl. 1). Therefore, a Darcian flux calculation was not made at this basin boundary because the transmissivity of the volcanic rocks is unknown. However, a hydraulic gradient is present across the basin boundary, suggesting that some volume of groundwater might be moving from basin fill in Stingaree Valley, through the shallow volcanic rock dividing Stingaree Valley and Dixie Valley, and then reentering the basin fill in Dixie Valley.

Subsurface inflow to Dixie Valley from Jersey Valley occurs through a 2.5–3.1 mi section of basin-fill deposits with a 5.7×10^{-3} ft/ft hydraulic gradient (between sites 3 and 4), but limited aquifer transmissivity data are available in the area. The closest transmissivity estimate, 2,900 ft²/d (NDWR well log 13405 [table 4]), is based on specific-capacity data from a well site more than 4 mi to the southwest of the basin divide. Potential subsurface inflow estimated from the variable basin width ranges from 1,800 to 2,300 acre-ft/yr, and is at least 700 acre-ft/yr greater than previous estimates (table 8).

Groundwater flow from Pleasant Valley to Dixie Valley was not estimated in this study because there is no appreciable connection of basin-fill sediments between the two valleys. Near the Pleasant Valley southern hydraulic terminus, there is only a 1,750-ft-wide section of limestone and dolomite (Natchez Pass Formation; Johnson, 1977, p. 16). A hydraulic gradient is present between the basin fill in each respective valley (pl. 1), indicating that groundwater could flow through carbonate rocks at the divide, but a Darcian flux calculation was not made as transmissivity of the carbonate rocks is not known. Previous estimates of subsurface inflow from Pleasant Valley range from 800 to 1,100 acre-ft/yr (table 8) (Cohen and Everett, 1963; Harrill and Hines, 1995). However, precipitation data used by Harrill and Hines (1995) to estimate recharge in Pleasant Valley, and subsequently subsurface flow to Dixie Valley, might have been skewed (Jim Harrill, USGS retired, oral commun., 2010), resulting in an underestimate of groundwater moving between the two basins.

Chemical Composition of the Basin-Fill Aquifer

Water-quality data were used to evaluate the chemical signature and quality of groundwater in the Dixie Valley basin-fill aquifer. Major ion and stable isotope data were used to describe the chemical signature, to identify possible differences in general chemistries between basin-fill and playa groundwater, and to interpret relative amounts of meteoric water possibly recharging the basin-fill aquifer. Changes in seasonal groundwater chemistry were also evaluated using selected trace metals and nitrogen. The chemical quality of groundwater was evaluated relative to established water-quality standards.

The groundwater signature and quality of the Dixie Valley basin-fill aquifer were characterized using groundwater-quality data from 52 wells and 15 valley springs (fig. 7; table 6). Groundwater quality data were collected from 32 wells and 13 valley springs between October 2009 and June 2011. Water-quality sampling followed U.S. Geological Survey (2006) field procedures to ensure comparability between samples. Non-artesian wells were purged by removing three well volumes of water prior to sampling. This ensured the sampling of aquifer water rather than borehole water. Groundwater flowing from artesian wells (table 3) was assumed to be representative of the aquifer; therefore, purging was deemed unnecessary. Samples were analyzed by two USGS laboratories: National Water Quality Laboratory, Lakewood, Colorado, and Reston Stable Isotope Laboratory, Reston, Virginia. Most wells were sampled in autumn 2009; an 11-well subset of these wells was sampled again in spring 2010 to investigate seasonal water-quality changes. Three of the 32 wells were drilled in southern Dixie Valley during winter 2010–11 and were sampled during aquifer testing in May–June 2011. Twenty-two additional samples reported in previous investigations were incorporated in the analysis to expand the spatial characterization of water quality in the basin-fill aquifer. These additional samples included nine well samples collected by Lahontan GeoScience, Inc. (2004), eight well and two spring samples collected by Goff and others (2002), and three well samples from Great Basin Center for Geothermal Energy (2004).

Limitations related to using multiple data sources included the amount and quality of supplemental data describing well construction and sampling and analysis techniques. Well construction information related to historical samples, including depth drilled and interval screened, rarely was provided (table 6; Goff and others, 2002; Great Basin Center for Geothermal Energy, 2004; Lahontan GeoScience, Inc., 2004). A lack of construction information precluded evaluation of chemical gradients and composition differences between neighboring wells, as these differences could be related to groundwater sampling depth. Another limitation was that the same parameters measured in historical samples were not always measured in this study, preventing direct comparison of portions of the datasets.

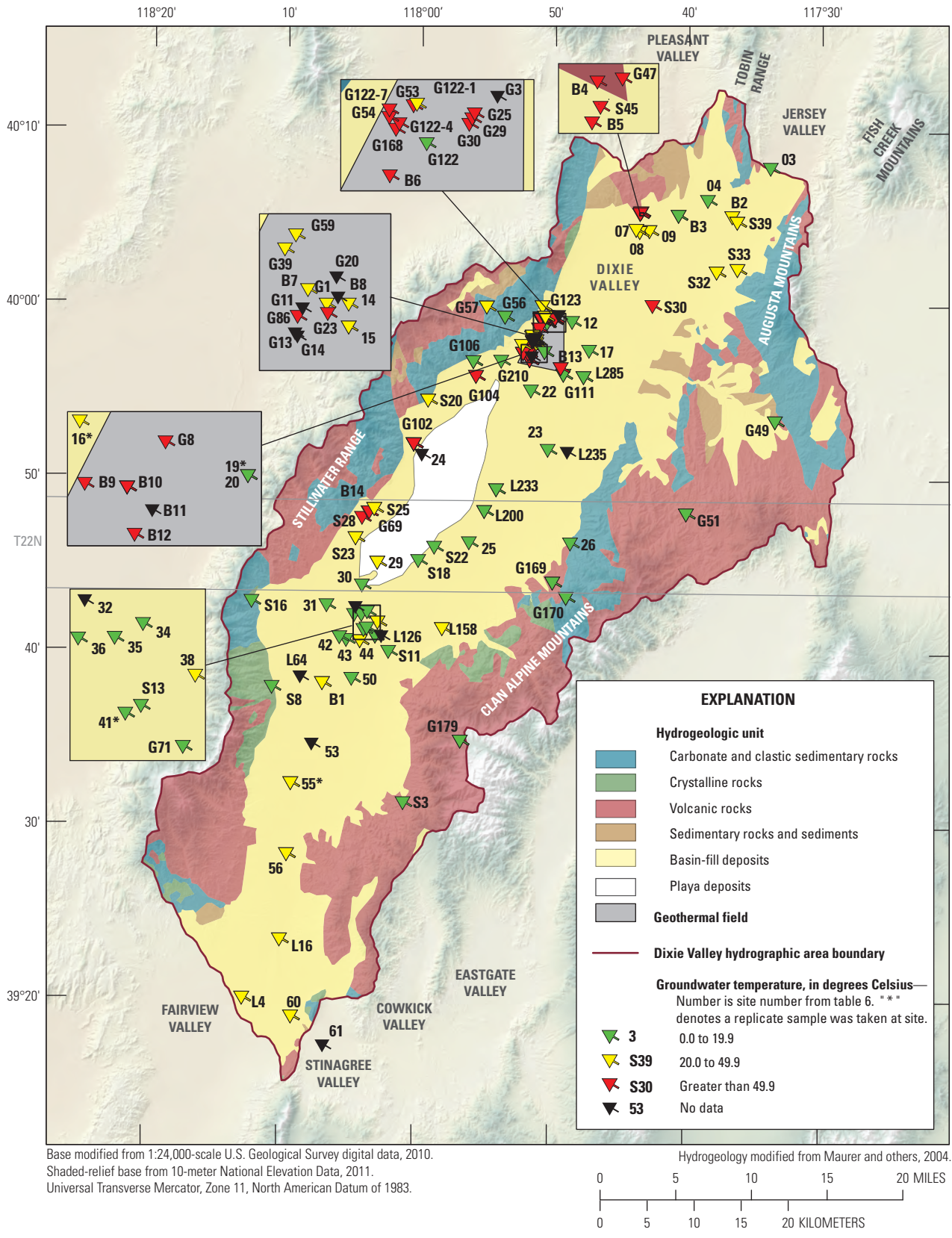


Figure 7. Locations of sites sampled for water quality and measured field groundwater temperature, Dixie Valley, west-central Nevada, 1959–2011.

Quality Assurance

Laboratory equipment and field blanks were run for quality assurance on pumps and sampling lines prior to and during sampling to ensure that sampling equipment did not contaminate samples. Internal components of the pump used in this study to purge and sample water from wells are made of stainless steel, which can corrode and release trace metals (Wilde, 2004), even after diligent decontamination. Although no corrosion was observed on the pump, several trace metals generally at or near laboratory reporting-limit concentrations were measured in laboratory equipment blanks, including aluminum (5.8 µg/L), manganese (0.43–1.96 µg/L), barium (0.15 µg/L), arsenic (0.06–0.19 µg/L), and antimony (0.26 µg/L). Field equipment blanks were used to evaluate possible contamination resulting from processing groundwater samples in the field. Although concentrations of several trace metals (iron, cobalt, manganese, magnesium, lithium, strontium, and arsenic) measured in field blanks were greater than laboratory reporting limits (within 1 percent), these concentrations were within an average of 2 percent of those in environmental samples.

Sequential replicate samples were collected to quantify variability associated with the collection and processing of a sample (U.S. Geological Survey, 2006). Replicate samples were collected from wells 16, 19, 41, and 55 (fig. 7); replicate and environmental sample concentrations generally were similar (coefficient of variation average of less than 10 percent).

Chemical Signature

The chemical signature of groundwater in Dixie Valley varies spatially. As water moves through an aquifer, minerals from encountered rocks and sediments dissolve and precipitate leaving a record in the groundwater chemistry. This record is represented mainly in the major-ion composition of groundwater, which can be used to evaluate groundwater source areas and flow characteristics. Major ion chemistry in groundwater representing the upper 500 ft of the basin-fill aquifer was used to further evaluate the playa as a flow barrier separating northern and southern groundwater with a water typing and statistical comparison using the Mann-Whitney rank sum test (Bradley, 1968).

Differences in water chemistry between the northern and southern areas of Dixie Valley support the hypothesis that the playa acts as a groundwater divide. General groundwater ionic composition is shown in figure 8, where data points represent a relative proportion of major ions. Groundwater in northern Dixie Valley generally can be characterized as a sodium bicarbonate-chloride type (Na-HCO₃-Cl; fig. 8). Most samples contain 40–80 percent sodium (little potassium [K] present relative to sodium) and 15–50 percent calcium. Magnesium (Mg) ranges from 10 to 30 percent for most northern groundwater. Bicarbonate and chloride

generally co-dominate the anion composition of northern groundwater, with most samples composed of 20–50 percent and 25–60 percent bicarbonate and chloride, respectively. Sulfate (SO₄) also was present in these samples, contributing from 20 to 40 percent of the anion composition (fig. 8). Groundwater south of Township 22N can be characterized as a sodium bicarbonate-sulfate type (Na-HCO₃-SO₄) (fig. 8). Most southern samples were dominated strongly by sodium (Na; 70–100 percent), with varying calcium contributions (Ca; 0–45 percent). Anion compositions in southern groundwater contain mixtures of bicarbonate (HCO₃; 40–65 percent) and SO₄ (25–50 percent) (fig. 8). Chloride was also present in these samples, contributing to about 15–40 percent of the anion composition.

A distinguishing characteristic between southern and northern groundwater is the lower contribution of magnesium in southern groundwater relative to that in the north. Southern groundwater is nearly 100 percent depleted of magnesium, whereas magnesium contributions in northern groundwater generally range from 10 to 30 percent. High concentrations of magnesium in groundwater indicate the presence of either magnesium carbonates (dolomite, hydromagnesite, or magnesite) or mafic (basalt) rock within the groundwater flow path (Hounslow, 1995). Large volumes of basalt are present on the east and west side of the basin, but basalt is more prevalent as outcrops on the east side of the basin in the Clan Alpine Mountains. Corresponding with more prevalent basalt outcrops in the eastern Clan Alpine Mountains, groundwater sampled east of the playa was enriched in magnesium relative to the rest of the basin. Water in southern wells 26 and 60 has much greater concentrations of magnesium than most southern wells on the valley floor (table 6). High magnesium concentrations in the upgradient groundwater-flow path in southern Dixie Valley indicate that groundwater is losing magnesium as it approaches the playa. Loss of magnesium could be the result of the incorporation of magnesium into clay-rich sediments (see section, “[Mixing of Geothermal and Basin-Fill Groundwater](#)”).

In addition to water typing, statistical comparisons of major ions indicate that northern and southern waters differ significantly. Mann-Whitney rank sum tests (Bradley, 1968) indicate all major ions, including dissolved solids, differed significantly between samples collected north and south of Township 22N (all p-values <0.007, 99-percent confidence level). These results indicate that mixing between groundwater north and south of the playa is minimal.

Several sites in northern Dixie Valley plot as outliers in figure 8. Northern sites 23 and G123 contained about 40 percent magnesium (10–30 percent higher than other northern samples). The higher magnesium in well 23 could result from groundwater flowing from the magnesium-rich dolomite and basalt present in the Clan Alpine Mountains, whereas magnesium in well G123 could be from the magnesium-rich basalt present in the gabbroic complexes in the Stillwater Range (fig. 7; rock types are listed in table 2).

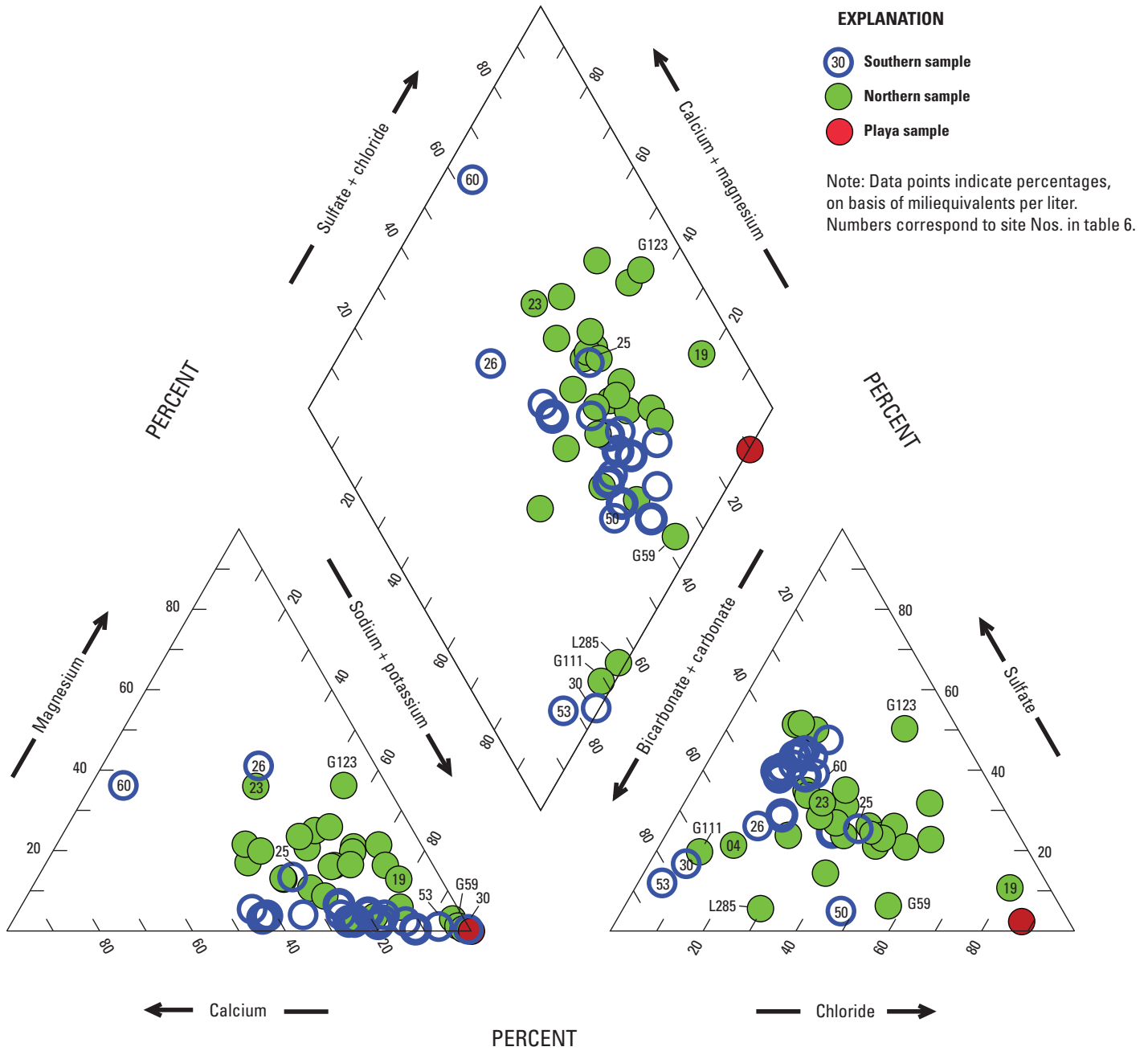


Figure 8. Trilinear diagram showing general water-quality conditions in northern and southern areas of Dixie Valley, west-central Nevada. Relative concentrations of cations and anions are presented in lower left and right triangles, respectively; relative concentrations are projected onto the central diamond to show combined major-ion chemistry.

Several characteristically different groundwater anion compositions were detected in samples or groups of samples in northern Dixie Valley (A = wells L285 and G59; B = well 19). The samples collected from wells L285 and G59 had the lowest sulfate contribution. Groundwater samples collected from well 19 contained the highest chloride and lowest bicarbonate-sulfate concentrations of any other sample. Differences between these three samples and those of neighboring wells all near or in the geothermal field could be related to the depth of groundwater sampled (depths of wells L285 and 19 were 55 and 50 ft bls, respectively), but a lack of construction information for many wells results in inconclusive comparisons.

Several outlying samples also were present in the southern sites (fig. 8). Samples from wells 26 and 60, which are on the alluvial fan of the Clan Alpine Mountains and the Stingaree Valley boundary, respectively, contained substantially less sodium (contributions less than 35 percent) than other wells, and groundwater in well 60 had much more calcium (contribution of about 60 percent). These differences likely result from the upgradient position of these wells in the groundwater flow path near the Dixie Valley basin boundaries (fig. 7). Samples collected from wells 30 and 53 had the greatest bicarbonate contributions when compared with all samples (northern and southern sites). The sample from well 50 had higher chloride contributions and came from a shallower well depth (less than 50 ft) than most southern wells.

Stable Isotopes

Precipitation is altered by local meteoric processes (that is, evaporative enrichment) before infiltrating to become groundwater (Clark and Fritz, 1997). Therefore, the stable isotopic signature of oxygen-18 and deuterium in precipitation can aid in groundwater characterization. Samples were collected (90 total samples from 13 sites) for isotopic evaluation of mountain-block and valley-floor precipitation between May 2009 and May 2011 (appendix A). Mountain-block precipitation samples (42 samples from 9 sites, collected each spring and autumn) were collected by Mahannah and Associates, LLC, and Interflow Hydrology, Inc. Subterranean samplers, similar to those used by Kormos (2005) and Huntington and Niswonger (2012), were used for the collection of samples representative of groundwater recharge. Samples of valley floor precipitation (48 samples from 4 sites, collected quarterly) were collected by USGS personal from bulk precipitation collectors. Mineral oil was added to precipitation collectors to prevent evaporation of accumulated water.

Stable isotopes of oxygen-18 and deuterium were evaluated using water samples collected from 32 wells and 12 valley-floor springs (table 3, denoted by footnote as site sampled for current study) and precipitation from 4 valley-floor and 9 mountain-block locations (Interflow Hydrology, Inc., written commun., 2012) (appendix A; fig. 9). Precipitation signatures of oxygen-18 and deuterium were used to compute

the local meteoric water line (LMWL). Values ranged from -16.8 to 5.0 and -128.0 to 15.0 per mil, respectively, at valley-floor locations, and from -16.8 to -7.9 and -126.0 to -61.6 per mil, respectively, at mountain-block locations. The variation in measurements from more negative to more positive values reflects evaporation of precipitation during warmer months. Average annual signatures of oxygen-18 and deuterium were -11.2 and -89.1 per mil at valley floor locations (2010–late 2011) and -14.1 and -106.0 per mil at mountain block locations (2009–early 2011).

Most mountain-block precipitation falls between October and April of each year (cool season). During water year 2010, about 76 percent of total annual mountain-block precipitation (Mahannah and Associates, LLC, and Interflow Hydrology, Inc., written commun., 2012) occurred between October and April; during water year 2011, cool-season precipitation increased to 83 percent of the annual total. Cool-season oxygen-18 and deuterium signatures of mountain-block precipitation averaged -15.5 and -117.9 per mil, respectively, in water years 2010–11 (fig. 9). Warm-season oxygen-18 and deuterium signatures of mountain-block precipitation averaged -14.1 and -104.3 per mil, respectively, in water year 2010 only (no water year 2011 warm-season data available). To account for these differences in contributions from warm- and cool-season precipitation, isotopic signatures were weighted by the relative contribution of seasonal precipitation volumes, and the results were summed, thereby calculating an annual volume-weighted isotopic signature at each site, similar to Izbicki (2004). Average volume-weighted, mountain-block precipitation signatures of oxygen-18 and deuterium (over water year 2010) were -15.3 and -116.0 per mil, respectively (fig. 9).

Isotopic signatures of both cold and warm basin-fill groundwater (sampled from 43 sites) provide insight on recharge sources and groundwater mixing. The mean oxygen-18 and deuterium signature for all basin-fill groundwater sites was -15.9 and -125.7 per mil, respectively. This signature was about 0.4 and 7.8 per mil more negative (lighter) in oxygen-18 and deuterium, respectively, than the mean winter mountain-block precipitation signatures. This indicates that most groundwater sampled in this study likely was recharged under cooler climate conditions (Rademacher and others, 2002). Mean geothermal (temperatures greater than 50 °C) groundwater signatures of oxygen-18 and deuterium were -11.4 and -126.2 per mil. Geothermal samples exhibited similar deuterium signatures as basin-fill groundwater, but were about 4.5 per mil more enriched in oxygen-18 than basin-fill groundwater and about 5.4 per mil more enriched than the LMWL for a similar deuterium value. Most basin-fill (cool and warm) groundwater samples analyzed lie parallel to the LMWL, but are slightly (about 0.5–1.5 per mil, or an average of 1 per mil) more enriched in oxygen-18 relative to deuterium when compared to the LMWL. This apparent shift in oxygen-18 signatures of basin-fill groundwater from the LMWL likely is evidence of mixing between basin-fill and geothermal groundwater (fig. 9).

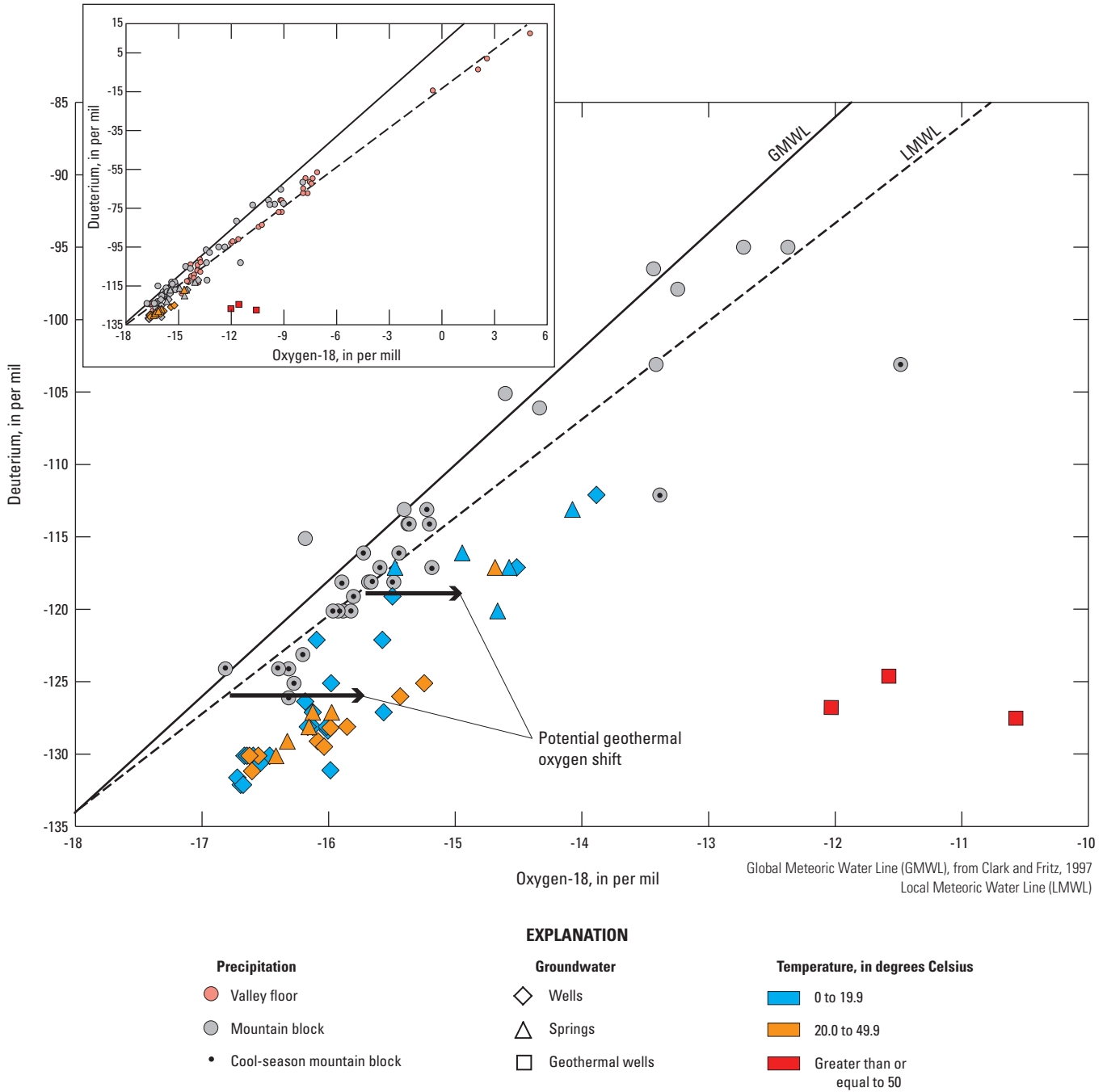


Figure 9. Isotopic signature of groundwater, springs, and precipitation at selected sites in Dixie Valley, west-central Nevada, 2009–11. Small graph shows full dataset, large graph highlights groundwater and spring datasets in relation to cool season precipitation.

Chemical Quality of Groundwater

Dixie Valley groundwater quality was evaluated with respect to national primary and secondary drinking-water standards (table 9; U.S. Environmental Protection Agency, 2009). Primary standards (maximum contaminant levels [MCLs]) have been established by the U.S. Environmental Protection Agency for constituents that pose potential health risks to humans. Secondary standards generally are non-enforceable guidelines designed to ensure water quality with desirable cosmetic or aesthetic characteristics such as taste and odor (U.S. Environmental Protection Agency, 2009, p. 6); however, the Nevada Division of Environmental Protection enforces secondary standards (Nevada Administrative Code 445A.455).

Arsenic concentrations in samples from 41 of 64 sites exceeded the primary drinking water standard (10 µg/L; U.S. Environmental Protection Agency, 2009) (table 9). Arsenic is present in groundwater throughout Dixie Valley as a result of groundwater interaction with the volcanic rocks in surrounding mountain blocks and volcanic-rock-derived basin-fill sediments. Arsenic concentrations in volcanic rocks are highly variable, but high concentrations are common in basalt, rhyolite, and gabbro (Boyle and Jonasson, 1973, p. 256). High arsenic concentrations generally were measured near the playa fringe or near the geothermal plant from shallow wells (less than 100 ft deep, where depth data are available), whereas lower concentrations were measured in or near mountain fronts. The highest arsenic concentration measured in the basin-fill aquifer was near the geothermal field (1,650 µg/L, site G1, table 6).

Fluoride concentration exceeded the MCL of 4 mg/L (U.S. Environmental Protection Agency, 2009) at 17 of 45 sites and exceeded the Nevada secondary standard of 2 mg/L (Nevada Administrative Code 445A.455) at 21 of 45 sites (table 9). Most fluoride exceedances occurred at southern sites (fig. 10). The source of fluoride is likely natural and derived from groundwater interaction with volcanic and igneous rocks. Fluoride occurs in biotite, which is present in granite and rhyolite (Edmunds and Smedley, 2013), both of which constitute most of the mountain block in the southern parts of the Stillwater Range and Clan Alpine Mountains (Willden and Speed, 1974) (fig. 10). Fluorspar, a fluoride-containing mineral, historically was mined in the southeastern Stillwater Range in southern Dixie Valley, as well as at the Wonder mine in the southwestern Clan Alpine Mountains (Willden and Speed, 1974, p. 50).

The national secondary drinking water standard (SMCL) for TDS (500 mg/L; U.S. Environmental Protection Agency, 2009) was exceeded in samples collected from 35 of 65 sites (mostly in northern Dixie Valley) (table 9; fig. 11). The State of Nevada enforces a secondary drinking water standard of 1,000 mg/L (exceeded in 15 of 65 sites, fig. 11). A large volume of carbonate rock is present in the northern areas of

the Stillwater Range and the Clan Alpine Mountains, which may contribute to high TDS concentrations (fig. 11). The highest TDS concentrations, ranging from 563 to 3,580 mg/L, were from samples in wells screened in shallow intervals (average bottom screened interval of 130 ft bls), whereas samples with TDS concentrations below the secondary standard were from wells screened in deep intervals (average bottom screened interval of 253 ft bls). High concentrations of TDS also could be derived from wind-transported salts deposited in mountain-block recharge areas or from downward percolation of evaporated lake water during the Pleistocene or early Holocene. Wind-blown salts originate from the Dixie Valley playa or the Carson Desert playa just west of the Stillwater Range (fig. 1). Both playas are remnants of Pleistocene lakes where high concentrations of evaporites accumulated during more than 10,000 years of drying. Predominant wind patterns from the west and southwest (Garcia and others, 2014) likely guide the depositional pattern, leading to greater concentrations in the north.

The SMCL for manganese (0.05 mg/L) was exceeded in samples collected from 15 of 62 sites (table 9), with exceedance locations distributed across the valley. Noticeable effects of manganese concentrations in groundwater greater than the SCML include a bitter metallic taste and dark color that can stain (U.S. Environmental Protection Agency, 1992). High manganese concentrations can be associated with high iron concentrations; high iron concentrations typically occur under reducing conditions characterized by low dissolved oxygen and nitrate (McMahon and Chapelle, 2008). Of the 62 wells sampled for manganese, 20 also had dissolved oxygen, iron, and nitrate data. Groundwater samples that had manganese and iron concentrations greater than their respective secondary drinking water standards had dissolved oxygen concentrations ranging from 0.1 to 3.1 mg/L and nitrate + nitrite (as N) concentrations ranging from 0.016 to 0.034 mg/L.

Boron does not have a national drinking water standard, but can cause adverse health effects to humans (U.S. Environmental Protection Agency, 2008) and can be highly toxic to plants (Eaton, 1935). The World Health Organization suggests a provisional guideline of 0.5 mg/L (World Health Organization, 2003) for states without a boron standard. The World Health Organization guideline was exceeded in 33 of 60 sites (table 9), most of which are adjacent to the playa or in northern Dixie Valley. Boron sources in groundwater can be from weathering of sedimentary rocks such as shale and borate minerals formed in association with evaporites (Williams and Hervig, 2002). In the 1870s, northern playa salt deposits were mined for borax and ulexite, both boron-containing minerals (Vanderburg, 1940, p. 43). Although no chemical analysis for boron was done on playa salts as part of this study, boron concentrations in shallow playa groundwater (well 29) ranged from 85.4 to 112 mg/L.

Table 9. Summary of drinking-water standards and exceedances for groundwater sites in the basin-fill aquifer, Dixie Valley, west-central Nevada.

[Exceedances are summarized based on location north or south of Township 22N (fig. 7). Percentage above the drinking-water standard for playa well 29 also is shown for comparison. **Constituent:** National drinking water standards are available for mercury (0.002 mg/L), nitrate (10 mg/L as N), and nitrite (1 mg/L as N). Mercury was not sampled, only total nitrogen (nitrate + nitrite, as N) was sampled, and relative quantities of nitrate and nitrite cannot be determined. **National drinking-water standards:** All values are in milligrams per liter (mg/L) except for pH, which is in standard units; primary standards from U.S. Environmental Protection Agency (2009); secondary standards from U.S. Environmental Protection Agency (1992). –, no standard; n/a, not applicable]

Constituent	National drinking-water standards		Number of sites sampled	Number of sites exceeding standard			Concentration at well 29, playa (percent of standard)
	Primary	Secondary		Total	Northern sites	Southern sites	
Aluminum	–	0.05–0.2	42	2	1	1	900
Antimony	0.006	–	31	16	6	10	383
Arsenic	0.01	–	64	41	19	22	120,500
Barium	2	–	54	0	n/a	n/a	n/a
Beryllium	0.004	–	30	0	n/a	n/a	n/a
Cadmium	0.005	–	30	0	n/a	n/a	n/a
Chloride	–	250	67	11	10	1	43,800
Chromium	0.1	–	30	0	n/a	n/a	n/a
Copper	1.3	1	55	0	n/a	n/a	n/a
Fluoride	4	¹ 2	45	17	3	14	576
Iron	–	0.3	65	20	10	10	339
Lead	0.015	–	52	0	n/a	n/a	n/a
Manganese	–	0.05	62	15	11	4	117
pH	–	² 6.5–8.5	67	3	2	1	109
Selenium	0.05	–	31	0	n/a	n/a	n/a
Silver	–	0.1	30	0	n/a	n/a	n/a
Sulfate	–	250	67	7	6	1	1,674
Thallium	0.002	–	30	0	n/a	n/a	100
Total dissolved solids	–	500 (³ 1,000)	65	35	30	5	35,700
Uranium	0.03	–	32	0	n/a	n/a	1,455
Zinc	–	5	49	0	n/a	n/a	n/a
Boron ⁴	–	⁵ 0.5	60	33	28	5	19,740

¹Nevada specific secondary drinking water standard. Number of sites exceeding the Nevada secondary standard of 2.0 mg/L equals 21 (U.S. Environmental Protection Agency, 1992). Number of northern and southern sites exceeding standard equals 6 and 15, respectively.

²Acceptable range for pH.

³The State of Nevada enforces a secondary drinking water standard of 1,000 mg/L.

⁴Not regulated by U.S. Environmental Protection Agency, but has known health effects.

⁵Provisional guideline value (World Health Organization, 2003); no maximum contaminant level for boron, but can cause adverse health effects (U.S. Environmental Protection Agency, 2008).

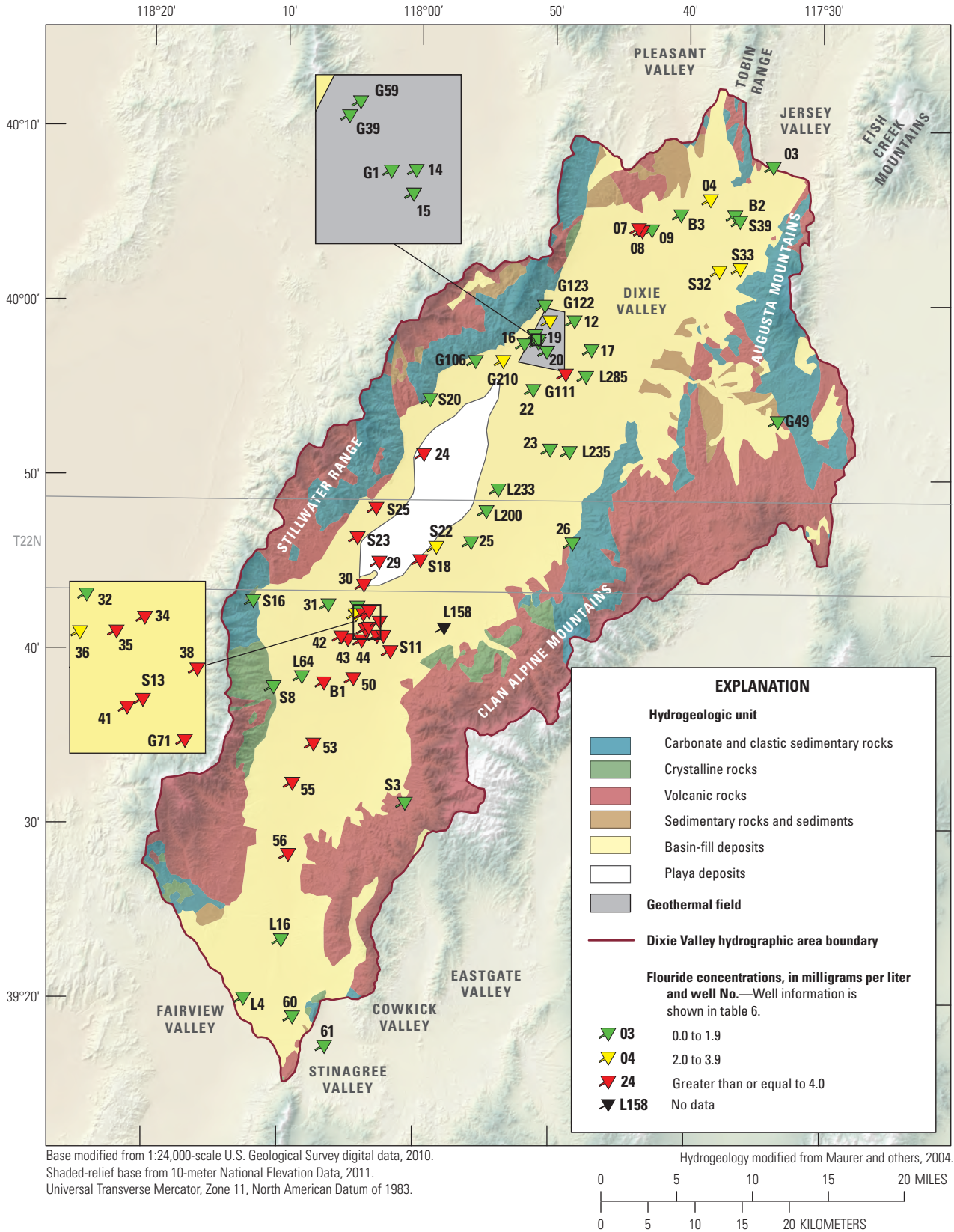


Figure 10. Fluoride concentrations in groundwater from sites and maximum contaminant level (MCL) exceedances in the basin-fill aquifer, Dixie Valley, west-central Nevada. National MCL is 4 milligrams per liter (U.S. Environmental Protection Agency, 2009), Nevada State MCL is 2 milligrams per liter (Nevada Administrative code 445A.455).

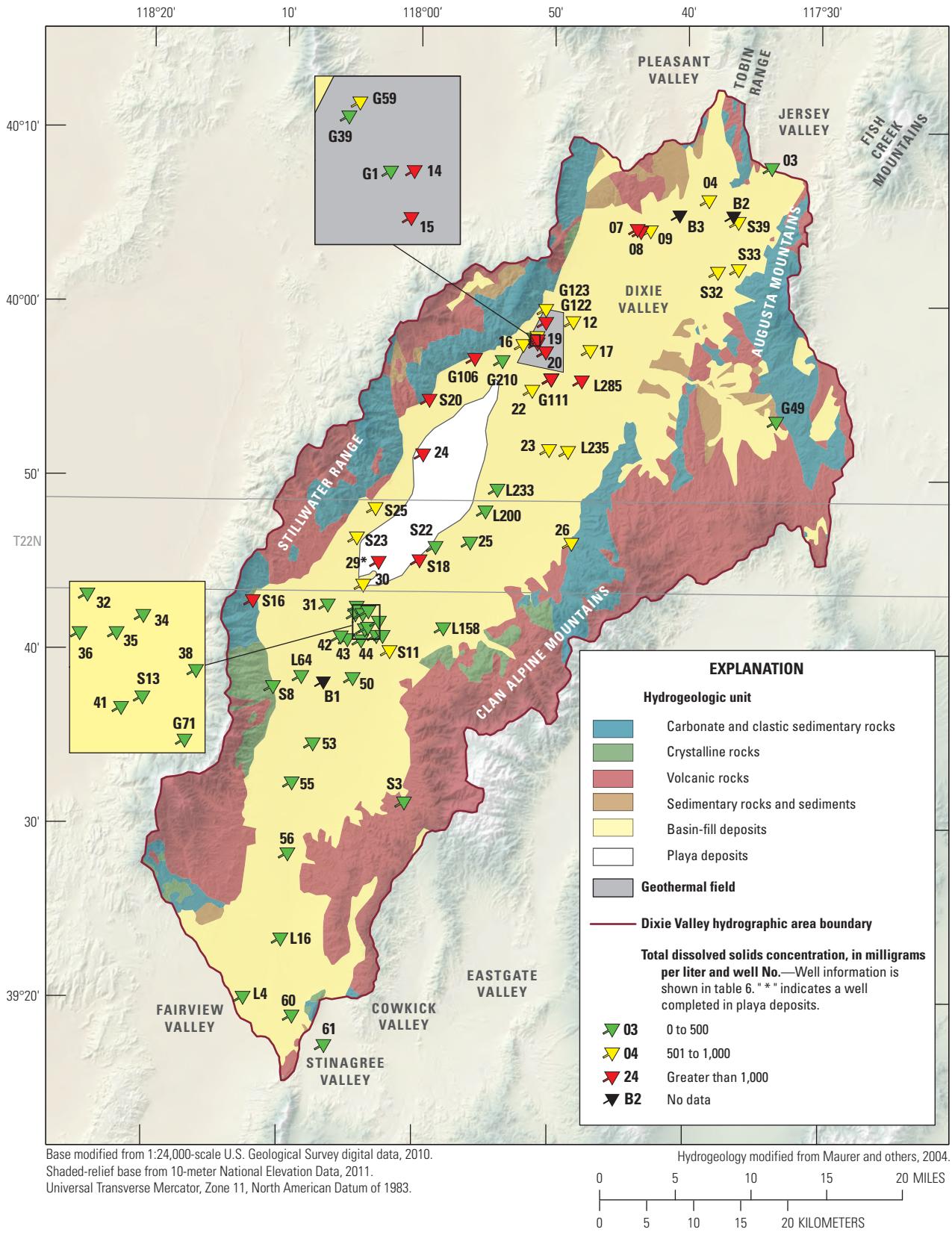


Figure 11. Total dissolved solids concentrations in groundwater from sites and secondary maximum contaminant level (SCML) exceedance distributions in the basin-fill aquifer, Dixie Valley, west-central Nevada. National SMCL is 500 milligrams per liter (U.S. Environmental Protection Agency, 2009), Nevada State SMCL is 1,000 milligrams per liter (Nevada Administrative code 445A.455).

Tungsten does not have a national drinking water standard, but is an emerging contaminant of concern because of potential health effects to humans (National Institute for Occupational Safety and Health, 2005; U.S. Environmental Protection Agency, 2010). Concentrations of tungsten from 32 sites in the Dixie Valley basin fill aquifer range from 0.04 to 126 $\mu\text{g/L}$, with a median concentration of 7.0 $\mu\text{g/L}$. For comparison purposes, tungsten concentrations from 171 groundwater sites throughout Nevada ranged from less than 0.5 to 677 $\mu\text{g/L}$, with a median concentration = 0.75 $\mu\text{g/L}$ (U.S. Geological Survey, 2014). The median tungsten concentrations in Dixie Valley groundwater is more than nine times greater than the median concentrations in the rest of the state. Previous studies indicate that high soil pH can increase the solubility of tungsten, causing it to leach more readily into groundwater (Agency for Toxic Substances and Disease Registry, 2005; Association of State and Territorial Solid Waste Management Officials, 2008). Tungsten concentrations greater than 25 $\mu\text{g/L}$ in Dixie Valley were measured in 25 percent of samples, and 75 percent of those had pH values of greater than 8.0 and were south of the playa. In the remainder of Nevada, 16 percent of samples had tungsten concentrations greater than 25 $\mu\text{g/L}$ and 81 percent of those samples had a pH of 8.0 or greater.

Groundwater in playa deposits is chemically distinct from the fresh basin-fill aquifer (fig. 8), and 13 of 21 constituents analyzed for drinking water standards exceeded the standard, often by more than 300 percent (table 9). For all constituents of playa groundwater exceeding standards, concentrations of those constituents were greater than concentrations from all other sites sampled except iron, lead, manganese, and pH. As would be expected at the terminus of a groundwater-flow system, the greatest exceedances typically occurred with increasing proximity to the playa because of the continued enrichment as groundwater moves along the flow path. For example, arsenic concentrations in groundwater collected from wells 26 (mountain front), 25 (mid-valley), and 29 (playa) (table 6; fig. 7) increase from 1.6 to 40.6 to 12,050 $\mu\text{g/L}$, respectively. With respect to national drinking water standards, groundwater approaching and in playa sediments would require considerable treatment before it would be considered potable.

Seasonal Changes in Groundwater Chemistry

Seasonal changes in groundwater chemistry were evaluated from 11 wells initially sampled in autumn 2009, and again in spring 2010. Constituent concentrations were considered to vary seasonally if the relative difference in autumn and spring concentrations was equal to or greater than 15 percent (table 10). About two-thirds of the constituents

analyzed changed by less than 15 percent between sampling periods, whereas one-third either increased or decreased by more than 15 percent.

Iron and manganese concentrations decreased (by an average of about 70 and 90 percent, respectively) at about one-half of the sites sampled, whereas lithium concentrations increased (by an average of about 40 percent) at 10 of the 11 sites (table 10). Increases in lithium corresponding to decreases in magnesium could suggest an influx of geothermal water into the basin-fill aquifer (see section, “[Connection between Basin-Fill and Geothermal Aquifers](#)”). At well 30, however, corresponding decreases in arsenic concentrations and water temperature indicate geothermal mixing is unlikely because geothermal water typically has high arsenic concentrations and warm temperatures.

Fluctuations of trace metals concentrations (including arsenic, fluoride, lithium, manganese, and tungsten) in groundwater samples collected from well 26 on the Clan Alpine Mountains alluvial fan (table 10; fig. 7) were larger than would be expected from such an upgradient site. Greater arsenic and fluoride concentrations in spring 2010 samples than in autumn 2009 samples could represent interactions between winter mountain-front recharge and volcanic rocks containing arsenic and fluoride. A tungsten mining prospect just south of well 26 and in the Clan Alpine Mountains (Willden and Speed, 1974, p. 58) could be the source of tungsten with a higher springtime water table possibly releasing loosely bound ions from clays and other sediments, thereby enriching the local groundwater.

Connection between Basin-Fill and Geothermal Aquifers

Multiple geothermal groundwater reservoirs are present beneath the basin-fill aquifer in Dixie Valley to estimated depths of 20,000 ft bls (Blackwell and others, 2003; McKenna and Blackwell, 2004; Wisian and Blackwell, 2004). In addition to the established geothermal field north of the playa, many other geothermal features exist in other areas of Dixie Valley including many warm and hot springs (fig. 5) and wells (fig. 7) in northern Dixie Valley, the Dixie Valley hot spring complex (fig. 5; S23, S25, and S28) along the Stillwater Fault west of the playa, and warm groundwater south of the playa (in the vicinity of site 55; fig. 7). In its basic composition, geothermal water is heated meteoric water that has undergone distinctive chemical changes caused by water-rock interactions induced by temperatures of as much as 300 °C (Ellis and Mahon, 1977; Giggenbach, 1992). These chemical changes make geothermal water chemically distinct from cold meteoric-derived groundwater, and, therefore, allow the evaluation of mixing between hot and relatively cold meteoric-derived groundwater.

Table 10. Selected chemical constituent concentration and percentage of change for samples collected in Dixie Valley, west-central Nevada, October 2009 and May 2010.

[Relative percent increase or decrease in constituent concentrations between autumn and spring sampling periods was used to classify constituents as constant (less than plus or minus 15 percent) or changing (greater than or equal to plus or minus 15 percent). Values highlighted in green are greater than or equal to plus 15 percent, and values highlighted in blue are greater than or equal to minus 15 percent. **Abbreviations:** °C, degrees Celsius; mg/L, milligram per liter; µg/L, microgram per liter; –, no data]

Chemical constituent	Unit of measure	Northern well No.									
		04		07		19		22		23	
		2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change
Temperature	°C	17.5	4	25.4	-27	15.2	1	15.5	5	15.4	–
pH	pH	7.4	1	7.4	0	7.8	-3	7.8	0	7.6	1
TDS	mg/L	910	-2	1,070	1	3,580	18	776	-1	658	-0
Calcium	mg/L	64.5	-5	102	5	112	-4	39.2	-4	57.1	3
Magnesium	mg/L	31	-12	24.8	-3	95.5	-2	22.2	-5	44.4	0
Sodium	mg/L	201	-7	166	1	1,020	7	157	-8	76.5	4
Potassium	mg/L	26.2	-2	34.5	-1	28.3	-1	19.8	-5	7.7	4
Bicarbonate	mg/L	544	3	333	-7	340	6	284	1	240	-2
Sulfate	mg/L	151	-4	374	2	337	9	156	1	155	1
Chloride	mg/L	84.6	-9	74.8	2	1,730	20	123	2	107	0
Arsenic	µg/L	9	17	13.8	12	31.8	13	40	-13	10	11
Boron	µg/L	2,130	-5	1,740	-49	2,570	17	1,050	4	297	-1
Bromine	mg/L	0.2	-12	0.1	0	1.1	0	0.22	-15	0.2	-6
Fluoride	mg/L	2.2	-6	4.7	0	1.2	9	1.41	3	0.3	10
Iron	µg/L	73	-92	11	43	1,050	-3	6	0	6	-40
Lithium	µg/L	488	21	604	-8	844	54	235	40	31	35
Manganese	µg/L	5.9	-68	2.9	32	742	1	45.2	-4	2.6	-115
Nitrogen	mg/L	4	-39	0.4	123	–	–	0.03	11	0.4	-3
Silica	mg/L	59.7	1	73.7	-1	58.5	-2	82.2	-3	64	0
Tungsten	µg/L	1.5	-31	18.4	9	8.6	14	31.4	-12	0.8	0
Uranium	µg/L	2.8	1	0.9	-11	0.2	10	4.6	1	6.0	1

Chemical constituent	Unit of measure	Southern well No.											
		25		26		30		31		38		42	
		2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change	2009 concentration	2010 percent change
Temperature	°C	15.3	-3	16.7	-4	15.8	-22	15.4	-3	20	-3	19.4	2
pH	pH	7.9	1	7.8	-3	9.8	0	8.2	-1	7.3	13	8.4	-5
TDS	mg/L	478	1	658	0	683	-1	298	1	342	-5	306	6
Calcium	mg/L	44.1	2	58.9	-1	1.0	0	34.6	-3	20.2	-1	20.4	2
Magnesium	mg/L	11.4	3	59.4	-5	0.5	-16	2.7	-7	1.1	-6	2.0	0
Sodium	mg/L	84	4	90	-4	224	3	43	-7	66.9	2	61.8	-1
Potassium	mg/L	5.0	0	2.2	-3	1.96	-5	3.9	-2	4.6	3	3.9	4
Bicarbonate	mg/L	145	-7	390	0	202	24	–	–	–	–	–	–
Sulfate	mg/L	81.9	2	145	0	85.2	0	50	0	74.4	0	76.6	-1
Chloride	mg/L	95.8	4	76.6	-1	28.7	1	28.6	6	22.3	2	24	2
Arsenic	µg/L	41.3	-4	1.4	19	78.4	-29	13	0	21.2	3	17.7	7
Boron	µg/L	623	4	307	1	570	7	201	-3	380	-5	360.0	-4
Bromine	mg/L	0.2	-6	0.1	0	0.1	0	0.1	0	0.1	-15	0.1	0
Fluoride	mg/L	1.2	2	0.3	27	6.0	0	0.91	-2	5.6	-4	5.3	-4
Iron	µg/L	3	67	12	8	1,910	-19	22	-59	39.0	-147	3	29
Lithium	µg/L	30.8	40	25.4	45	6.6	41	32.2	39	51.5	30	28.4	41
Manganese	µg/L	0.3	29	2.8	83	66.4	-12	0.8	-67	1.3	-171	0.3	-40
Nitrogen	mg/L	0.5	-6	0.22	-5	0.02	0	0.8	0	0.1	1	0.3	-1
Silica	mg/L	53.5	0	15.9	-1	67.1	-5	54.7	0	67.3	-2	55.7	0
Tungsten	µg/L	12.3	-3	0.02	100	55.5	6	3	-3	27.5	12	6.4	21
Uranium	µg/L	4.0	4	4.27	-4	0.2	0	3.7	3	0.8	10	0.5	9

The Dixie Valley geothermal field ([fig. 1](#)) was studied extensively in the late 1990s and early 2000s to evaluate the geothermal resource (for example, Goff and others, 2002) and to understand losses in fluid pressures caused by geothermal water extraction (Benoit and others, 2000). One goal of this research was to determine the extent to which the geothermal and cold-basin-fill aquifers were connected. One conclusion, based on limited isotopic and chemical information, indicated that the geothermal and basin-fill water originated almost exclusively from downward percolation of Pleistocene Lake Dixie water as the climate warmed, and not from mountain recharge (Nimz and others, 1999; Gregory Nimz, Lawrence Livermore National Laboratory [retired], written commun., 2012). Carbon-14 ages of 12–20 thousand years for the geothermal and basin-fill waters, respectively, appear to be consistent with this interpretation. Stable hydrogen and oxygen isotopic data, radiochemical chlorine-36 data, and comparisons of the chemical composition of geothermal and basin-fill aquifer water indicated almost no mountain-front recharge to either aquifer and that the chemical composition of the basin-fill groundwater was derived from water-rock interaction with underlying Mesozoic carbonates (Nimz and others, 1999). However, in contrast to the results of Nimz and others (1999), a study by Campana and others (1986) comparing stable isotopic signatures and chloride concentrations concluded that geothermal and basin-fill groundwater in Dixie Valley has not evaporated extensively, and likely originates from snowmelt and mountain-block recharge. Simulation results from a mixing model developed by Bruton and others (1997) indicated that the basin-fill aquifer could contain as much as 15 to 25 percent geothermal water in the geothermal field area north of the Dixie Valley playa ([fig. 7](#)).

With the availability of additional, more spatially extensive groundwater chemistry data representative of the basin-fill aquifer, a reassessment of the interaction between geothermal and basin-fill aquifer water was possible. The purpose of this reevaluation was to determine if the interaction of shallow basin-fill groundwater with the deeper geothermal aquifer fits the previously determined mixing model of Bruton and others (1997) in areas of Dixie Valley outside the established geothermal field.

In order to contrast the geothermal and basin-fill aquifers, selected geothermal data published in Goff and others (2002) and the Great Basin Center for Geothermal Energy (2009) were compared to recent basin-fill groundwater samples collected for this study. Groundwater samples were divided into three classes based on field temperature measurements for the purposes of this study ([fig. 7](#)). Groundwater with a temperature of less than 20 °C was considered cold-basin-fill groundwater based on generally observed groundwater

temperatures in Nevada. Water with a field temperature of 20–50 °C was considered warm groundwater, or potentially influenced by geothermal water; groundwater with a field temperature of greater than 50 °C was considered geothermal water, or heavily influenced by geothermal water. The upper limit of 50 °C was selected to describe geothermal water based on a study by Schaefer and others (2005) that concluded groundwater temperatures of as much as roughly 50 °C could be attributed to travel along deep, long flow paths that did not require geothermal heating. Schaefer and others (2005) assumed that to reach temperatures of greater than 50 °C required geothermal input based on the geothermal gradients in the region. Although water temperatures in the shallow basin-fill aquifer sampled in this study were less than 50 °C, the wells sampled were much shallower, and the flow paths were much shorter than those sampled by Schaefer and others (2005). Therefore, water temperatures between 20 and 50 °C likely indicate mixing between geothermal and meteoric-derived groundwater, and not heating derived from deep, long flow paths. The basin-fill aquifer sites sampled for this study also were divided into these groundwater designations. The warmest basin-fill groundwater sampled during this study was 41.4 °C (site 55).

Warm groundwater temperature designations may not necessarily represent influence from a geothermal source. For example, groundwater collected from the playa was 21.4 °C ([table 6](#), well 29) and, therefore, is considered warm water. Warm playa water temperatures are likely due to measurements near the playa surface in November, after warming all summer and prior to much cooling in the autumn and winter, rather than mixing with geothermal water. Nevertheless, these temperature designations allow the establishment of a simple gradient between water likely to be less influenced by mixing with geothermal water and water more likely to be mixed with geothermal water.

As previously stated in section, “[Chemical Composition of the Basin-Fill Aquifer](#),” the limitations of using multiple data sources include the type of supplemental data available such as well construction and techniques used during sample collection and analysis (Goff and others, 2002; Great Basin Center for Geothermal Energy, 2009). First, information on well construction is needed to evaluate chemical gradients and composition differences between neighboring wells because these differences could be related to groundwater sampling depth. Second, documentation of any losses of gas or water during the collection of samples from geothermal wells and springs by degassing or steam is important when comparing to data from cold water because loss of water vapor concentrates the sample; this documentation was not always apparent in the sample site descriptions. Finally, the same parameters measured in this study were not always measured in early studies, making comparisons among datasets difficult.

Geothermal and Non-Geothermal Indicators

Several chemicals have been used previously to indicate groundwater heating under geothermal conditions (Kharaka and Mariner, 1989). These chemicals are lithium, boron, silica, rubidium, and antimony. Conditions often associated with geothermal conditions are groundwater temperatures greater than 50 °C and low magnesium concentrations. Under these circumstances, relatively high groundwater temperatures can create an environment where magnesium has a lower solubility and reacts to form clay minerals (Gunnlaugsson, 2008). Dixie Valley geothermal water has low concentrations of magnesium, ranging from below detection to less than 0.1 mg/L, likely owing to the reaction of magnesium with host rocks to create altered minerals such as magnesium-silicates and clays. Dixie Valley geothermal groundwater also has high sodium and chloride concentrations, but evaporated, or old (thousands of years) basin-fill groundwater also will have high concentrations of sodium and chloride that are not caused by high temperatures. Ratios of magnesium, sodium, and potassium can be used to determine the presence of geothermal groundwater and to calculate the geothermal reservoir temperature from which the water was derived (Kharaka and Mariner, 1989). In this study, groundwater temperature and concentrations of lithium ([appendix B](#)), boron, and magnesium were used as geothermal indicators; silica (silicon dioxide, SiO₂) also was used.

Just as some chemicals are indicators of geothermally heated water, other chemicals are indicative of non-geothermal water. High concentrations of magnesium is an indicator of non-geothermal groundwater and can result from either magnesium carbonates (dolomite, hydromagnesite, or magnesite) or mafic (basalt) rock in the groundwater-flow path (Hounslow, 1995). Groundwater sampled east of the Dixie Valley playa was enriched in magnesium relative to the rest of the basin ([fig. 12](#)), likely owing to basalt in the Clan Alpine Mountains. Several sites near the geothermal plant indicate high magnesium concentrations and although well depth information is not available for all sites, most high magnesium concentrations are associated with shallow basin-fill wells. High sulfate and bromide concentrations also might indicate non-geothermal waters, as well as waters that have low TDS.

High silica concentrations often are indicative of geothermal mixing with cold groundwater, but the presence of both crystalline and non-crystalline silica in Dixie Valley confounds the use of this indicator. For example, concentrations of silica greater than 30 mg/L can be indicative of geothermal water in cold groundwater. At typical geothermal reservoir temperatures (200–300 °C), silica is highly soluble and concentrations in geothermal water can exceed 500 mg/L. Dixie Valley geothermal water has silica concentrations ranging from 100 to 700 mg/L. Quartz (crystalline silica) is very resistant to chemical weathering at non-geothermal temperatures (less than 50 °C) and usually has concentrations of less than 10 mg/L (Hounslow, 1995).

Non-crystalline or poorly crystalline silica is present in volcanic deposits such as tuffs in Dixie Valley, which are less stable and are more easily weathered. Therefore, cold groundwater that has flowed through tuff deposits might have concentrations of silica greater than 30 mg/L. Silica concentrations for 86 percent of groundwater samples collected during this study were greater than 30 mg/L. Therefore, high silica concentrations were not necessarily indicative of geothermal mixing with basin-fill groundwater.

Mixing of Geothermal and Basin-Fill Groundwater

Chemical Indicators

In order to determine the extent of mixing between geothermal and basin-fill aquifer groundwater, end-members (waters that are not mixed) were established. Dixie Valley geothermal water generally is distinct in chemical composition from cool basin-fill groundwater. Water collected from geothermal production wells (as provided by Goff and others [2002], and Zehner and others [2006]) was sufficient to be used as the geothermal end-member. The main chemical indicators of geothermal water used in this study were high lithium, silica, and boron.

Basin-fill groundwater (the other end-member), derived from either Pleistocene recharge from Lake Dixie or mountain-front recharge, is less clearly defined because of varying lithologies around the basin. Although water-rock interaction with basalts, granites, limestone and other less-abundant rock types lead to a mixture of weathering products, diagenetic alterations, and complex ions in solution, certain weathering characteristics of some lithologies make it feasible to determine a general basin-fill end-member.

The molar ratio of magnesium to lithium was used to characterize geothermal groundwater in Dixie Valley. Data used to evaluate the geothermal nature of groundwater showed a strong correlation between low magnesium-to-lithium ratios and high geothermal silicon concentrations, which indicates that geothermal fluids are sufficiently identified by the magnesium-to-lithium ratio ([fig. 13](#)). Groundwater samples collected for this study show a similar pattern between magnesium-to-lithium ratios and dissolved-silica concentrations ([fig. 14](#)). Most non-geothermal groundwater has higher magnesium-to-lithium ratios than warm or hot water. Cooler basin-fill aquifer water shows the least mixing, but some samples from this study and from historical analyses show mixing.

High concentrations of boron and lithium in groundwater are mostly derived from geothermal heating (Giggenbach, 1992; Hounslow, 1995). The relation of boron and lithium between cold groundwater and geothermal fluids is shown in [figure 15](#). Even the lowest concentrations of boron and lithium generally occur along a mixing line between concentrated geothermal fluids and dilute cold groundwater ([fig. 15](#), inset).

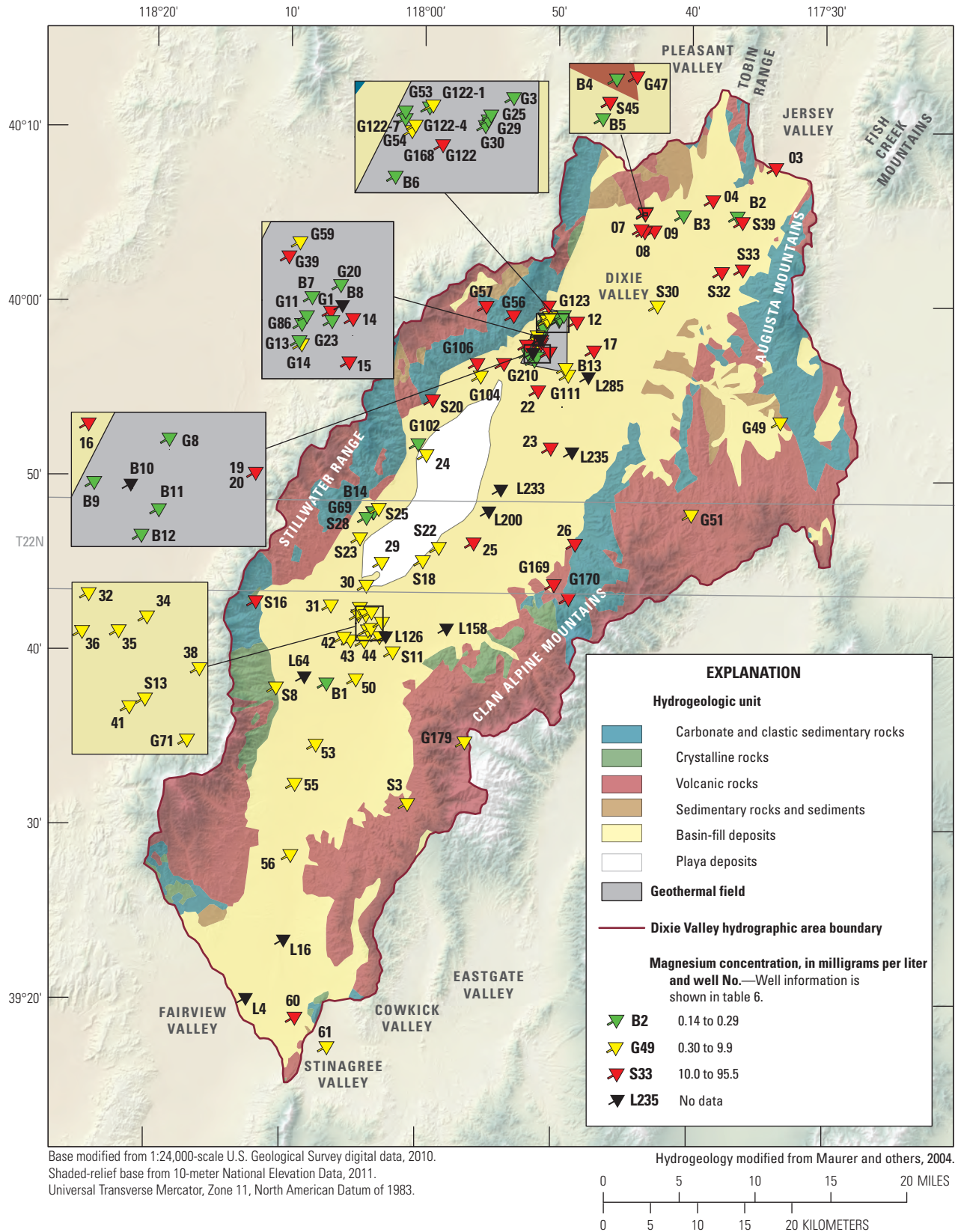


Figure 12. Magnesium concentrations in groundwater from sites in Dixie Valley, west-central Nevada. Magnesium concentrations are higher on east side of basin than in other areas.

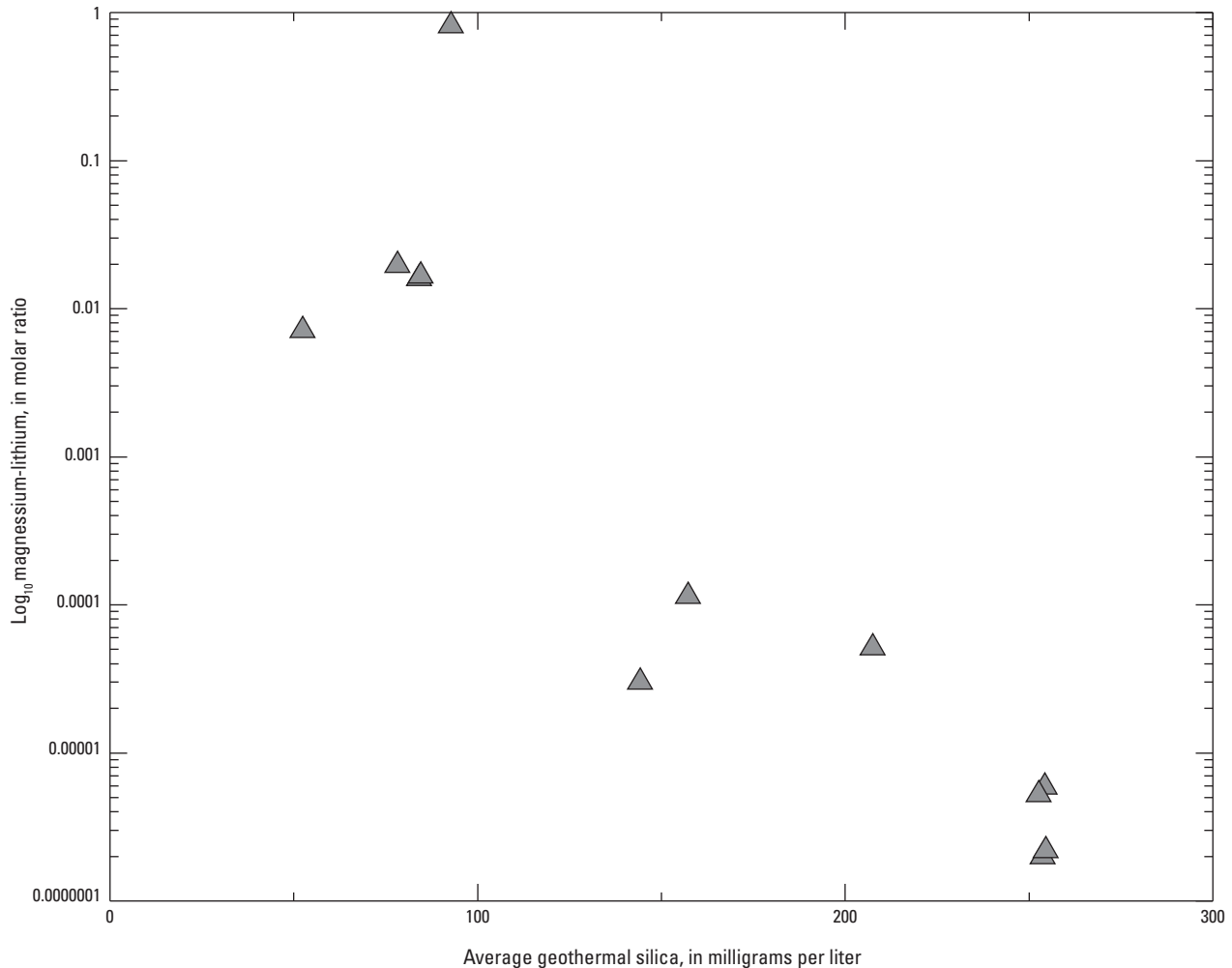


Figure 13. Relation between the molar ratio of magnesium to lithium and average geothermal silicon in geothermal groundwater samples, Dixie Valley, west-central Nevada.

The correlation, (coefficient of determination, $r^2=0.56$) estimated from a sample size of 138 paired boron and lithium analyses is significant at the 1-percent level ($p\text{-value} < 0.0001$). The relatively consistent increase of lithium and boron indicates that there is a connection between the geothermal and cold groundwater systems. Variations in the boron and lithium correlation can be attributed to many factors that may include, but are not limited to, differences in groundwater chemistry from various rock-water interaction rates with different lithologies throughout the basin, differences in rainfall and snowmelt during the sampling period (more than 20 years), and differences in laboratory analyses methods through time.

Determining the amount of mixing between geothermal water and cold basin-fill groundwater is difficult to quantify because of a lack of distinct end-member solutions; however, by using different lines of evidence, it is possible to show relatively consistent amounts of mixing. Lithium concentrations in the geothermal production wells range

from about 350 $\mu\text{g/L}$ to as much as 2,820 $\mu\text{g/L}$ (Goff and others, 2002), although the average lithium concentration of geothermal wells is 2,245 $\mu\text{g/L}$. The average lithium concentration of the cold basin-fill groundwater (water that had a temperature when sampled of less than 20 °C and is not influenced by nearby geothermal sources) is 51 $\mu\text{g/L}$. If linear mixing is used and it is assumed that any lithium in a water sample greater than 51 $\mu\text{g/L}$ is derived from geothermal fluids, every additional 100 $\mu\text{g/L}$ of lithium would represent 4.5 percent mixing with a geothermal fluid. Twenty-seven of 51 (or about 53 percent of) basin-fill wells and springs sampled had lithium concentrations greater than 51 $\mu\text{g/L}$, indicating that mixing of as much as 46 percent was possible, although average mixing was 13 percent. This is a somewhat crude method for determining mixing between the cold groundwater and geothermal fluids, but because end-member solutions are not well constrained, these approximations are the best estimates.

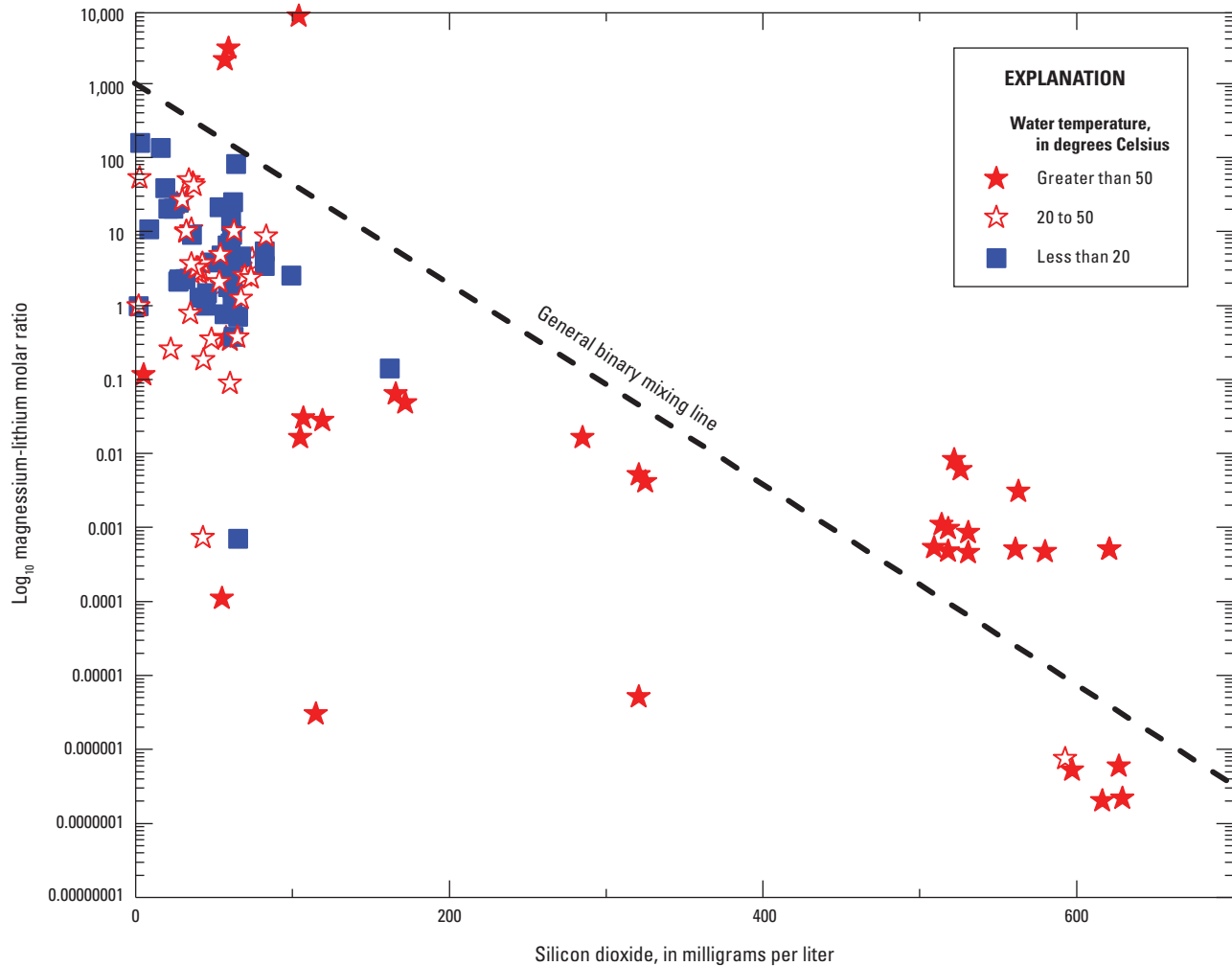


Figure 14. Mixing between water high in silica (SiO_2) and lithium (Li) with water high in magnesium (Mg) and low in SiO_2 and Li, and water temperature, Dixie Valley, west-central Nevada. Water samples include current basin-fill groundwater and historical basin-fill and geothermal groundwater.

The highest concentrations of geothermally derived constituents (lithium, silica, boron, and temperature) generally are present in the northern, western, and southern areas of the basin, near known faults or geothermal features (lithium, [fig. 16](#); temperature, [fig. 7](#)). High magnesium concentrations derived from the weathering of volcanic rocks mostly are present in the eastern and northern parts of the basin ([fig. 12](#)). Groundwater in the southern and northern parts of the basin likely is mixed with local geothermal water present in these areas, and not from mixing with geothermal water transported through the basin-fill aquifer from near the Dixie Valley Geothermal Power Plant. Some of the spatial variations in concentrations might be due to differences in well depth between the samples.

Geothermometry

Several methods have been developed to determine the temperature from which geothermal fluids or heated groundwater originated. These methods (geothermometry) usually are used to estimate reservoir temperatures of deep geothermal aquifers, but these methods also can be used to estimate how hot the water was originally at depth for warm groundwater sampled near the surface. These methods use formulas for key geothermal indicator elements based on feldspar equilibrium (sodium, potassium, and calcium). Other methods involve silica equilibrium and (or) magnesium and lithium empirical equations (Kharaka and Mariner, 1989).

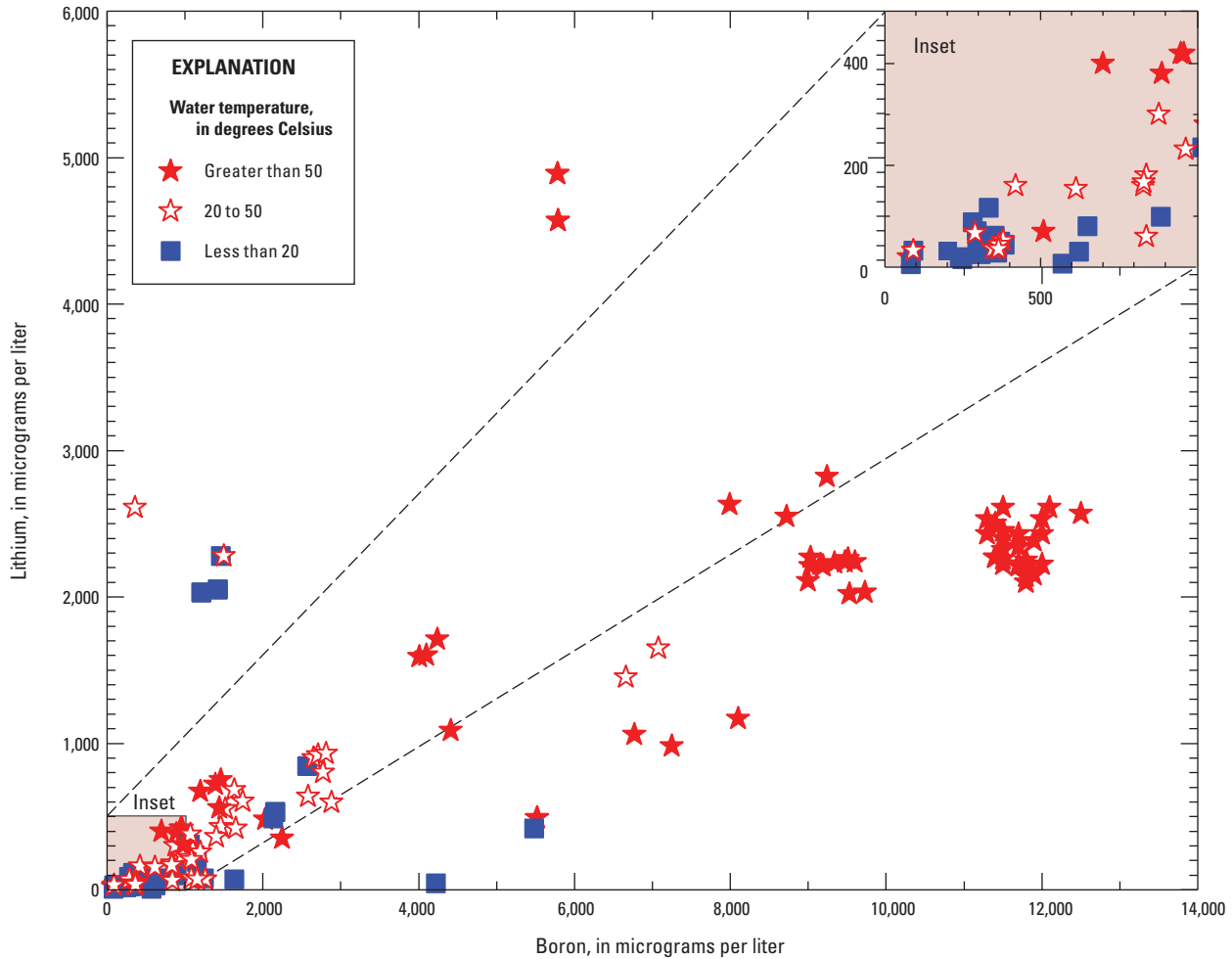


Figure 15. Relation between boron and lithium concentrations, and water temperature in basin-fill groundwater samples collected in this study and in historical basin-fill and geothermal groundwater samples, Dixie Valley, west-central Nevada. Coefficient of determination (r^2) equals 0.56, and p -value is less than 0.0001.

Geothermometry methods also can be used to determine sources of heating (long flow paths compared to geothermal sources) in groundwater (Schaefer and others, 2005). When geothermometry is used, it is assumed that the chemistry of the water is in equilibrium with the heating that occurred at depth. For geothermal fluids, this generally is true because the fluids have little opportunity to interact with host rock because many geothermal fluids travel through fractures in the rock rather than through interconnected pores. For heated basin-fill aquifer water, longer flow paths, longer residence time in the aquifer, and greater contact with rock materials in matrix pores potentially allow for non-equilibrium dissolution or precipitation of some elements and compounds used in geothermometry. This means that multiple geothermometers should be used and care should be taken when evaluating temperatures calculated from equilibrium-based

geothermometers in basin-fill aquifers (Land and MacPherson, 1992). For example, various parts of the basin in Dixie Valley have undergone different chemical reactions in the groundwater because of the heterogeneity of the aquifer rocks in the basin. Therefore, a geothermometer that might be useful in one part of the basin may not be useful everywhere in the basin. Because of disequilibrium reactions involving feldspar that can occur in alluvial basins, the magnesium-lithium geothermometer was selected as the most useful approach for this study (Land and MacPherson, 1992). However, comparison with other geothermometers were made to check the results and to evaluate areas that may have been affected by additions or subtractions of lithium or magnesium because of water-rock interactions involving clays or volcanic rocks in the basin (table 11).

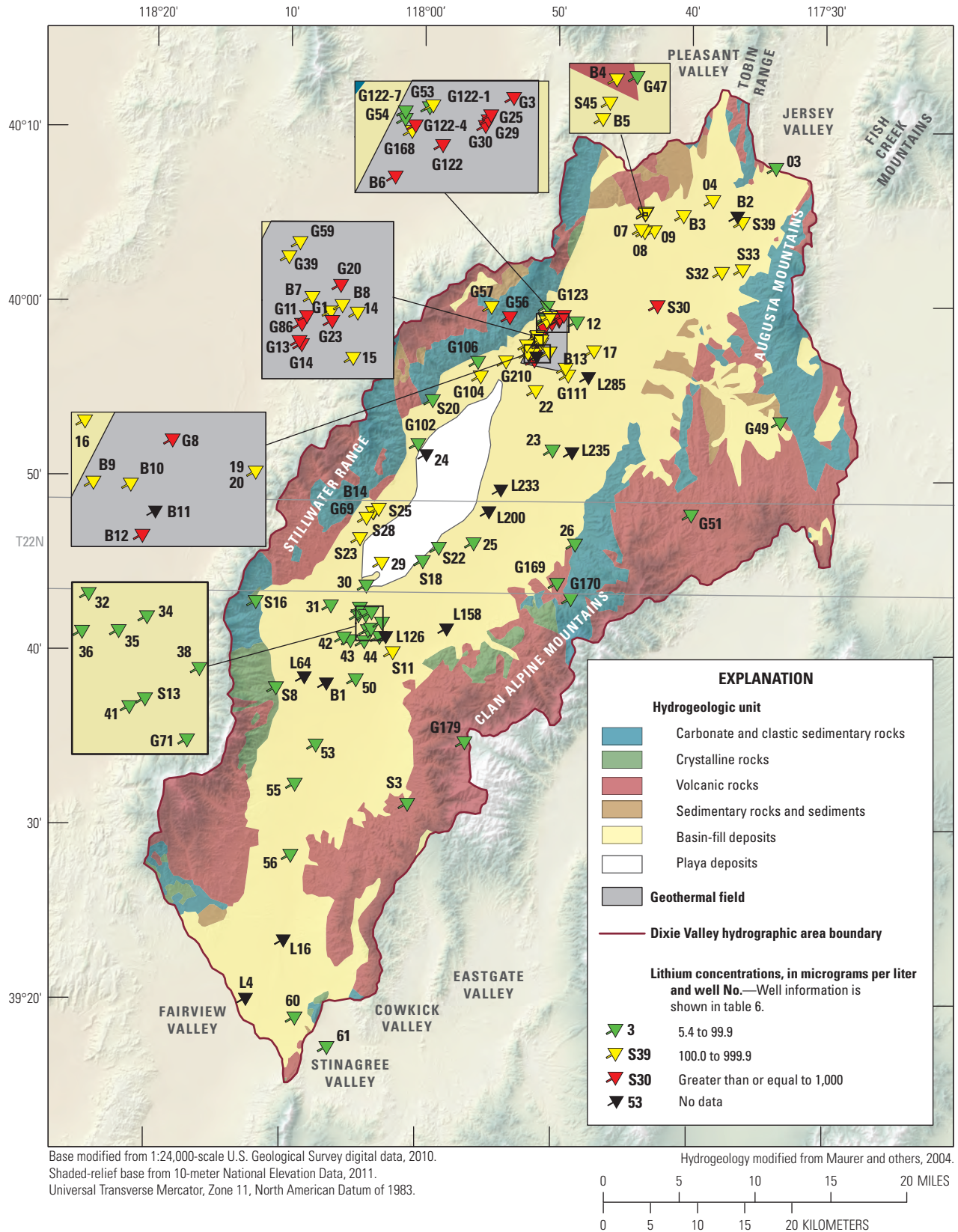


Figure 16. Lithium concentrations in groundwater from sites in Dixie Valley, west-central Nevada. Highest lithium concentrations in groundwater sampled from western, northern, and southern areas of basin.

This magnesium-lithium geothermometry method uses an empirical equation derived from the observed concentration of magnesium and lithium at different temperatures in basin-fill and geothermal basins (Kharaka and Mariner, 1989; Land and MacPherson, 1992). The equation used for this study was used to describe temperatures from geothermal basins worldwide (Kharaka and Mariner, 1989), was applied to basin-fill aquifers along the U.S. Gulf Coast, and is (results shown in [fig. 17](#)):

$$T = \frac{2200}{(5.47 + \log[\sqrt{Mg / Li}])} - 273.15 \quad (1)$$

where

- T* is the calculated temperature in degrees Celsius;
- Mg* is the magnesium concentration of the fluid in milligrams per kilogram (mg/kg); and
- Li* is the concentration of lithium in the fluid in mg/kg.

In Dixie Valley, the magnesium-lithium geothermometer (equation 1) generally is in agreement (within 10 percent) with calculated reservoir temperatures using sodium-potassium-calcium geothermometry for geothermal wells ([table 11](#)). However, the magnesium-lithium geothermometer shows reservoir temperatures averaging about 120 percent less than the equilibrium-based geothermometers using sodium-potassium-calcium geothermometry for basin-fill wells (range is from 732 percent less to 82 percent more; [table 11](#)). This may be owing to: (1) non-equilibrium conditions in the basin-fill aquifer, (2) possible addition of magnesium to the basin from weathering of basalts and dolomite that could affect the empirical relation between magnesium and lithium, or (3) high concentrations of silica (greater than 50 mg/L) in cold groundwater from dissolution of diatoms or tuff deposits. Additionally, evapotranspiration of the shallow groundwater and wind-blown evaporites from the playa and nearby Carson Sink could increase sodium and potassium concentrations so they are not in equilibrium with feldspar reactions. These factors might explain the differences in separate geothermometer-derived reservoir temperatures calculated for groundwater in the cold basin-fill aquifer.

Table 11. Estimated reservoir temperatures based on magnesium-lithium geothermometer compared to sodium-potassium-calcium geothermometer and measured field groundwater temperatures, Dixie Valley, west-central Nevada.

[**Site No.:** Locations of sites are shown in [figure 17](#). **Geothermal mixing:** Zero percent mixed indicates no mixing with geothermal aquifer. **Abbreviations:** °C, degrees Celsius; NA, sodium; K, potassium, Ca, calcium; n/a, not applicable; –, no data]

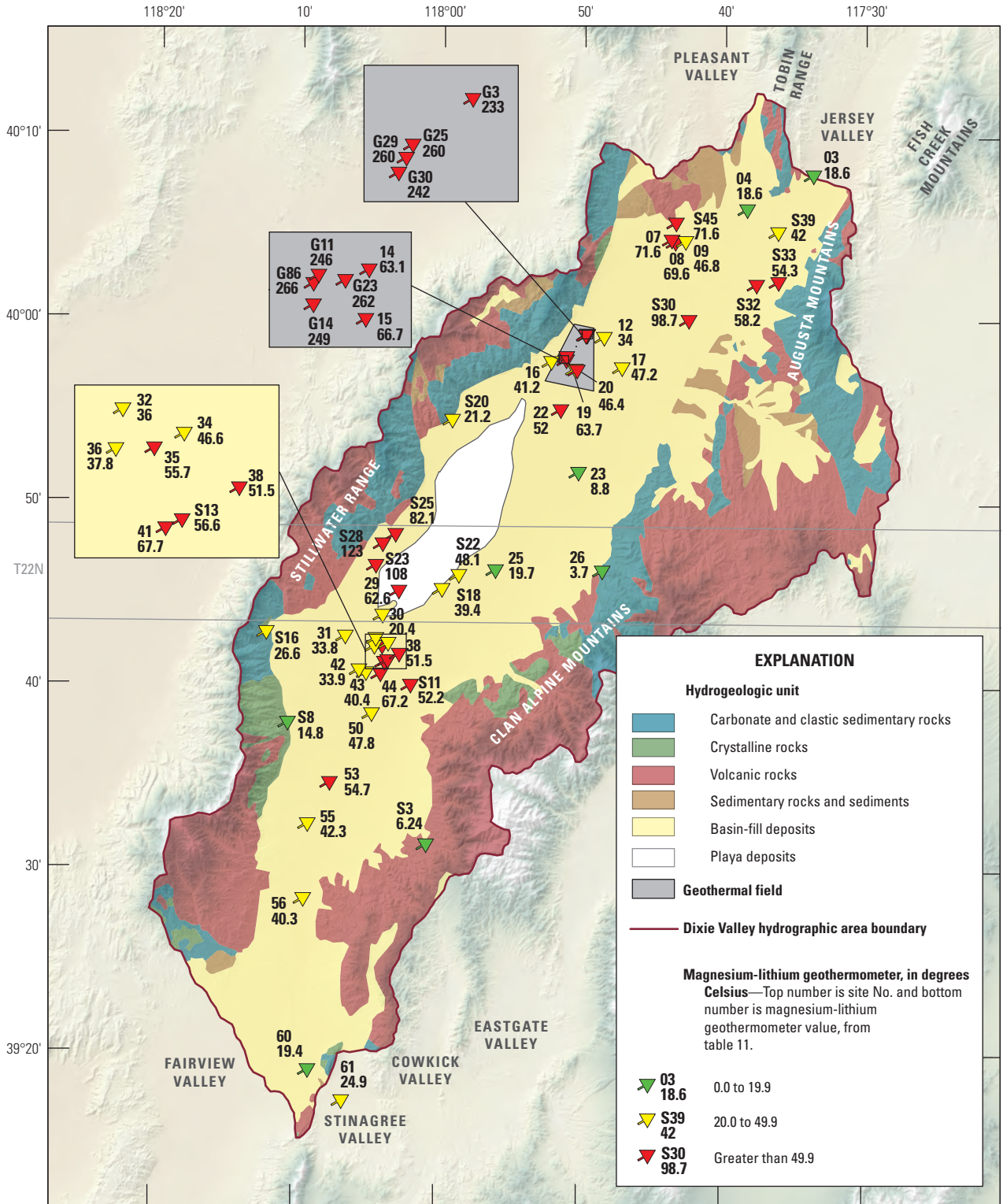
Site No.	Groundwater temperature (°C)	Geothermometer			Geothermal mixing (percent)
		Magnesium-lithium (°C)	Sodium-potassium-calcium (°C)	Difference (percent)	
Basin-fill sites					
09	15.5	46.8	83.9	-79	12.4
12	28.2	34	66.6	-96	2.3
14	21.7	63.1	170	-169	16.4
15	34.4	66.7	172	-158	12.8
16	19	41.2	91	-121	8.8
17	15.2	47.2	174	-269	12.7
19	15.5	63.7	132	-107	19.1
20	15.5	46.4	85.6	-84	12.3
22	15.4	52	186	-258	14.5
23	–	8.8	65.41	-643	0.0
25	15.3	19.7	57.9	-194	1.7
26	16.7	3.7	30.8	-732	0.0
29	21.4	62.6	126	-101	16.3
30	15.8	20.4	97.5	-378	1.8
31	15.4	33.8	49.1	-45	7.3
32	–	36	57.7	-60	0.0

Table 11. Estimated reservoir temperatures based on magnesium-lithium geothermometer compared to sodium-potassium-calcium geothermometer and measured field groundwater temperatures, Dixie Valley, west-central Nevada.—Continued

[Site No.: Locations of sites are shown in [figure 17](#). **Geothermal mixing:** Zero percent mixed indicates no mixing with geothermal aquifer. **Abbreviations:** °C, degrees Celsius; NA, sodium; K, potassium, Ca, calcium; n/a, not applicable; –, no data]

Site No.	Groundwater temperature (°C)	Geothermometer			Geothermal mixing (percent)
		Magnesium-lithium (°C)	Sodium-potassium-calcium (°C)	Difference (percent)	
Basin-fill sites—Continued					
34	16.2	46.6	61.9	-33	12.1
35	17.4	55.7	62.9	-13	15.2
36	15	37.8	60.0	-59	9.0
38	20	51.5	69.6	-35	12.5
41	19.4	67.7	63	7	19.2
42	19.4	33.9	63	-86	5.8
43	19.9	40.4	69.5	-72	8.1
44	20.5	67.2	65.6	2	18.5
50	18	47.8	64.6	-35	11.8
53	–	54.7	82.7	-51	0.0
55	41.4	42.3	72.1	-70	0.4
56	26.3	40.3	62.8	-56	5.6
60	22.2	19.4	29.1	-50	0.0
61	–	24.9	19.5	22	0.0
S3	13.8	6.24	21.8	-249	0.0
S8	15.3	14.8	2.66	82	0.0
S11	6.9	52.2	208	-298	18.0
S13	4.3	56.6	49.5	13	20.8
S16	8.7	26.6	19.5	27	7.1
S18	10.3	39.4	122	-210	11.5
S20	22.4	21.2	74	-249	0.0
S22	18.6	48.1	99.5	-107	11.7
S23	29	108	66.5	38	31.3
S25	26	82.1	48.9	40	22.3
S28	58.5	123	123	0	25.6
S30	60.9	98.7	164	-66	15.0
S32	21.8	58.2	96.6	-66	14.4
S33	30.2	54.3	98.7	-82	9.6
S39	44.8	42	72.8	-73	0.0
S45	62.1	71.6	60.9	15	3.8
Average	23	47	87	-121	10
Selected geothermal wells¹					
G25	174	260	235	10	n/a
G30	165	242	238	2	n/a
G29	165	260	237	9	n/a
G3	–	233	254	-8	n/a
G86	166	266	235	12	n/a
G11	–	246	232	6	n/a
G23	174	262	234	11	n/a
G14	–	249	230	8	n/a
Average	169	252	237	6	n/a

¹Calculated Na-K-Ca temperatures from Goff and others (2002).



Base modified from 1:24,000-scale U.S. Geological Survey digital data, 2010.
 Shaded-relief base from 10-meter National Elevation Data, 2011.
 Universal Transverse Mercator, Zone 11, North American Datum of 1983.

Hydrogeology modified from Maurer and others, 2004.

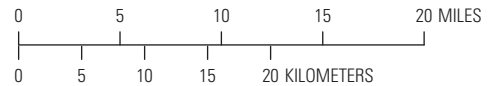


Figure 17. Locations of magnesium-lithium geothermometers and temperatures recorded in Dixie Valley, west-central Nevada.

The use of the magnesium-lithium geothermometry represents a conservative estimate of basin-fill aquifer heating because temperatures calculated using this method are lower than temperature from all the other equilibrium-based geothermometers. Therefore, percentages of mixing between geothermal and cold basin-fill groundwater could be higher than estimated here.

Groundwater that initially may have had high magnesium concentrations from a cold-water source, or high lithium concentrations from geothermal water could show lower concentrations of these constituents because of water-rock interaction with the playa aquifer sediments rich in smectite clay. For example, a combination of low magnesium concentrations (less than 0.3 mg/L) and low lithium concentrations (about 100 µg/L) in groundwater collected in the shallow subsurface of the playa is uncharacteristic of evaporated, cold basin-fill water and (or) geothermally mixed water, and indicate that magnesium and lithium can be lost to clays without geothermal heating. Although playa water is warm (about 20 °C), temperatures are not high enough to lower magnesium solubility. Loss of magnesium and lithium from solution most likely is due to interactions with smectite clay minerals present in the playa sediments. X-ray diffraction analyses confirm the presence of high-purity smectite clays in the less-than-0.03 micron fraction (D. Webster, U.S. Geological Survey, written commun., 2012) that can readily sorb lithium at exchange sites (Williams and Hervig, 2002). Smectite clays also use magnesium in their crystal structure (Deer and others, 1966). Therefore, in areas where aquifer water interacts with smectite clay-rich playa sediments, the use of magnesium and lithium constituents to investigate geothermal mixing could produce erroneous results.

Given the complications of using the magnesium-lithium geothermometry for the samples in this study, the field measured groundwater temperatures ranged from as low as 3.7 °C in a mountain front well (site 26) to more than 105 °C in two springs near known geothermal areas (S23 and S28, [table 11](#); [fig. 17](#)). Most basin-fill aquifer water sampled in this study seems to come from water that was heated above background mountain-front recharge temperatures (between 3 and 15 °C), indicating the influence of mixing with warm, possibly geothermally derived water.

Mixing of geothermal and cold basin-fill groundwater also was calculated by comparing the calculated magnesium-lithium geothermometry reservoir temperature to the field groundwater temperature of the well or spring when sampled ([table 11](#)). The difference in temperature divided by the average geothermal production well temperatures (252 °C, calculated by magnesium-lithium geothermometry in [table 11](#)) indicates that an average of about 10 percent mixing occurs between the geothermal and basin-fill aquifers, with a range of 0 percent mixing to more than 31 percent in some springs.

Confirmation of geothermal mixing with basin-fill groundwater comes from several lines of evidence: (1) high

concentrations of geothermally derived trace elements such as lithium, silica, and boron; (2) water temperatures greater than mountain-front recharge temperatures; (3) results of geothermometry calculations using multiple methods; and (4) although not discussed here, augmentation of geothermal pressures by reinjecting cold basin-fill groundwater above the geothermal reservoir. This reinjection has reduced the chloride content of the geothermal production bores, indicating a connection between the overlying aquifer and the geothermal reservoir (Benoit and others, 2000).

There is no single method for estimating geothermal mixing that is considered ideal in a natural environment. Methods presented in this study, including groundwater temperature, geothermal and non-geothermal indicators (lithium, silica, and boron), and magnesium-lithium geothermometry, all have complicating factors. However, each method points to the same general conclusion that most basin-fill groundwater sampled throughout Dixie Valley during this study contains some fraction of water derived from a warmer source, indicating that groundwater could contain as much as 46 percent geothermal water, with the average mixing estimated at 10–12 percent—findings similar to previous findings by Bruton and others (1997). Samples collected from wells in the north, south, and west sides of the basin show greater mixing than those on the east side of the Dixie Valley.

Summary

Dixie Valley primarily is an undeveloped basin in west-central Nevada, and the terminus of the Dixie Valley flow system, which includes Pleasant, Jersey, Fairview, Stingaree, Eastgate, and Cowkick Valleys. Churchill County, Nevada, is investigating additional water resources in Dixie Valley and has proposed to pump and transport groundwater from the fresh basin-fill aquifer. This study provides an improved understanding of groundwater resources in Dixie Valley by characterizing the hydrogeologic framework, the occurrence and movement of groundwater, the general chemical quality of the basin-fill aquifer, and the potential mixing between basin-fill and geothermal aquifers in Dixie Valley.

Rocks and sediments in Dixie Valley were grouped into six hydrogeologic units based on similar lithologies and hydraulic properties affecting groundwater flow: (1) the playa deposits, (2) basin-fill deposits, (3) carbonate and clastic sedimentary rocks, (4) crystalline rocks, (5) tertiary sedimentary rocks and sediments, and (6) volcanic rocks. Consolidated rock units include carbonate and clastic sedimentary rocks of Triassic-to Jurassic-age, crystalline rocks of Triassic-to-Tertiary age, sedimentary rocks and sediments of Tertiary age, and volcanic rocks of Tertiary age. Unconsolidated rock units include basin-fill and playa deposits of Quaternary age.

The basin-fill hydrogeologic unit can be several orders of magnitude more transmissive than the surrounding and underlying consolidated rocks and playa deposits. The hydraulic properties of the basin-fill deposits were estimated from single- and multi-well aquifer tests and specific-capacity tests. Transmissivity estimated from aquifer tests done on flowing wells screened in basin-fill just south of the playa ranged from 400 to 1,400 feet squared per day (ft²/d) (average of 925 ft²/d). Estimates of transmissivity from three single-well aquifer tests done on southern Dixie Valley wells screened in basin-fill ranged from 700 to 6,000 ft²/d (average of 3,100 ft²/d). Transmissivity estimated using specific-capacity data from 25 well sites throughout Dixie Valley and screened in basin-fill ranged from 30 to 45,500 ft²/d. Values estimated from wells north of Township 22N averaged 225 percent more than values from wells south of Township 22N. Although transmissivity estimates from aquifer tests generally are more accurate and reliable than estimates from specific capacity, the aquifer-test- and specific-capacity-derived estimates from neighboring wells typically were similar (within the same order of magnitude).

A potentiometric surface map developed from current and historic depth-to-groundwater and land-surface elevation measurements was used to evaluate the occurrence and movement of groundwater in Dixie Valley. Groundwater generally flows from the mountain range uplands toward the central valley lowlands and eventually discharges near the playa edge. Potentiometric contours east and west of the playa indicate that groundwater is moving eastward from the Stillwater Range and westward from the Clan Alpine Mountains toward the playa. Similarly, groundwater flows from the southern and northern basin boundaries toward the Dixie Valley playa. North-south- and east-west-trending flow patterns support a previous hypothesis that the playa hydraulically separates groundwater north and south of Township 22N.

In addition to hydraulically separating groundwater flow between the northern and southern systems, groundwater in the Dixie Valley playa is chemically distinct from the freshwater basin-fill aquifer. Groundwater mixing between basin-fill and playa groundwater systems is likely physically impeded by transmissivity contrasts of about four orders of magnitude. Density contrasts between basin-fill and playa groundwater averaged 0.16 milligram per liter (mg/L), resulting in chemical flow barriers. Total dissolved solids concentrations in the playa groundwater were nearly 440 times greater than concentrations in the basin-fill aquifer. It should be noted that this finding is based on two wells screened in the playa deposits at depth of 10 feet or less. These large differences between physical and chemical characteristics suggest that groundwater interaction between the basin-fill and playa sediments was minimal during this study; however, some interaction likely occurs over longer time periods.

Groundwater levels in Dixie Valley have remained generally constant since the 1950s. Observed changes are mostly in areas of historical localized development as a result of withdrawals for irrigation, livestock, and domestic use, and from augmentation of geothermal reservoir pressure. Water levels north of the Dixie Valley playa have risen and declined over time (-15 to 5 ft), and likely correspond to changes in agricultural and geothermal pumping. South of the playa, where the basin-fill aquifer is semi-confined, water levels have increased by an average of 1.5–3 ft since the 1980s. Rising water levels likely are a result of the capping and sealing of local flowing wells and aquifer recovery from reduced agricultural pumping following U.S. Navy land acquisitions in the 1980s. Water-level data are sparse in southern Dixie Valley (south of latitude 39°35'0"N); however, one well showed a water-level decline of about 3.5 ft from the late 1980s to the late 2000s, possibly from groundwater withdrawals for livestock watering.

Estimates of subsurface inflow to Dixie Valley from adjacent basins were calculated using a Darcian flux through the basin-fill aquifer. Annual subsurface inflow from Fairview and Jersey Valleys ranges from 700 to 1,300 acre-feet per year and from 1,800 to 2,300 acre-feet per year, respectively. Basin-fill deposits at the divide between Dixie and Stingaree Valleys are shallow and dry, indicating that the water table is in the deeper, less transmissive consolidated rock. Therefore, subsurface inflow was not estimated because transmissivity and water-level data within the volcanic rock unit were largely unknown. Although groundwater likely flows from Pleasant Valley to Dixie Valley, the divide is composed of carbonate rocks; therefore, subsurface flow was not estimated in this study owing to a lack of hydraulic information within the rock unit.

The chemical characteristics of groundwater in Dixie Valley were evaluated using data collected from wells and springs. Differences in magnesium concentrations and co-dominant anion composition in groundwater samples collected from the north and south valley areas indicate that groundwater north and south of Township 22N differs chemically. Groundwater in northern Dixie Valley generally is a sodium bicarbonate-chloride type, whereas groundwater south of Township 22 N is a sodium bicarbonate-sulfate type. All major ions, including dissolved solids, differed significantly between samples collected north and south of the playa.

Some constituents in groundwater samples collected from Dixie Valley exceeded established drinking-water quality criteria. Primary drinking water standards were exceeded for arsenic (0.01 mg/L; 41 of 64 sites) and fluoride (4 mg/L; 17 of 45 sites), and secondary drinking water standards were exceeded for total dissolved solids (500 mg/L; 35 of 65 sites) and manganese (0.05 mg/L; 15 of 62 sites). Boron and tungsten do not have national drinking water standards but

can cause adverse health effects. Although a drinking water criterion was unavailable for boron, the provisional boron guideline of 0.5 mg/L was exceeded in 33 of 60 sites; no guideline currently exists for tungsten.

Several methods of chemical comparisons between basin-fill and geothermal aquifer water indicate that most basin-fill groundwater sampled generally contains 10–12 percent geothermal water—a range similar to that of previous findings. Stable isotopes of oxygen-18 and deuterium, and geothermal indicators such as high temperature, lithium, boron, chloride, and silica indicate that mixing occurs in many wells that tap the basin-fill aquifer, particularly on the north, south, and west sides of the basin. A horizontal positive shift in oxygen-18 signatures of cool and warm groundwater away from the local meteoric water line and toward geothermal groundwater highlights potential mixing between basin-fill and geothermal aquifer waters. The mixing of lithium end-members indicates that many concentrations of lithium in cold groundwater are less than 51 micrograms per liter, and suggests only 1 or 2 percent mixing with geothermal water. Twenty-seven of the 51 cold samples analyzed (near or north of the geothermal field) show as much as 46 percent mixing with geothermal water. Magnesium-lithium geothermometers indicate that most basin-fill aquifer water sampled for this study apparently originates from water that was heated above background mountain-block recharge temperatures (between 3 and 15 degrees Celsius), indicating the influence of mixing with warm water, likely derived from geothermal sources.

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Appendix A. Stable Isotopic Data for Bulk Precipitation Sites Within Dixie Valley Study Area, May 2009–November 2011

Appendix A is a Microsoft® Excel file and can be downloaded at <http://pubs.usgs.gov/sir/2014/5152>.

Appendix B. Determination of Lithium Using Inductively Coupled Plasma-Optical Emission Spectrometry

Prior to 2003, the U.S. Geological Survey National Water Quality Laboratory measured dissolved and whole-water recoverable lithium using radial-view Inductively Coupled Plasma-Optical Emission Spectrometry (ICP-OES). With the implementation of updated instrumentation, lithium subsequently has been measured using axial-view ICP-OES. Initial prove-out studies indicated that there was no significant difference between results obtained using the axial and radial views for samples with a wide range of dissolved-solids concentrations. Recently (2013), some analyses of samples from the Branch of Quality Systems Inorganic Blind Sample Program have shown a positive bias in lithium results. Such bias can occur as a result of ionization interferences associated with samples with relatively high concentrations of alkali (that is, sodium and potassium) and (or) alkali earth elements (that is, calcium and magnesium). This ionization interference is reduced substantially when using the radial view because of the shorter path length of the light emission measured.

The shorter path length in the radial view affects the method detection limit. The reporting limits for lithium in dissolved and whole-water samples analyzed using the axial view are 0.05 and 0.04 micrograms per liter ($\mu\text{g/L}$), respectively, whereas the corresponding reporting limits for lithium using the radial view are estimated at 0.1 $\mu\text{g/L}$.

A procedure for correcting axial-view lithium concentrations for bias associated with ionization interference was established by simulating the axial and radial lithium results relative to alkali and alkali earth concentrations for more than 100 dissolved and whole-water samples. The lithium concentrations in the model samples ranged from 0.5 to more than 6,000 $\mu\text{g/L}$, whereas concentrations

of concomitant alkali and alkali earth elements ranged from 0.5 to more than 19,000 mg/L. Linear least squares regression of the ratio of radial-lithium concentration to axial-lithium concentration as a function of \log_{10} (calcium [Ca]+magnesium [Mg]+sodium [Na]+potassium [K] concentration) resulted in an equation that can be used to convert axial-lithium results to radial-lithium results if calcium, magnesium, sodium and potassium concentrations are known. The regression equation is:

$$\text{(axial-Li / radial-Li) ratio} = 0.3062 \times \log_{10}(\text{Ca} + \text{Mg} + \text{Na} + \text{K, in mg/L}) + 0.7748 \quad (\text{B1})$$

After determining the (axial-Li/radial-Li) ratio for a particular sample, the radial-Li concentration is calculated by dividing the previously measured axial-Li concentration by the (axial-Li/radial-Li) ratio.

The accuracy of the correction equation for more than 900 dissolved and whole-water samples is shown in [figure B1](#). Results for lithium from measured radial-ICP-OES are compared to calculated radial-lithium results from equation B1. For the dissolved samples tested, there is a positive bias of about 6 percent between the measured and calculated radial-lithium concentrations; for whole-water samples, there is a negative bias of about 6 percent. This bias signifies the overall accuracy of the correction for the range of lithium, alkali, and alkali earth concentrations in the samples used in the evaluation. Perfect correlation between the measured and calculated results would give a regression line with a slope equal to 1.0.

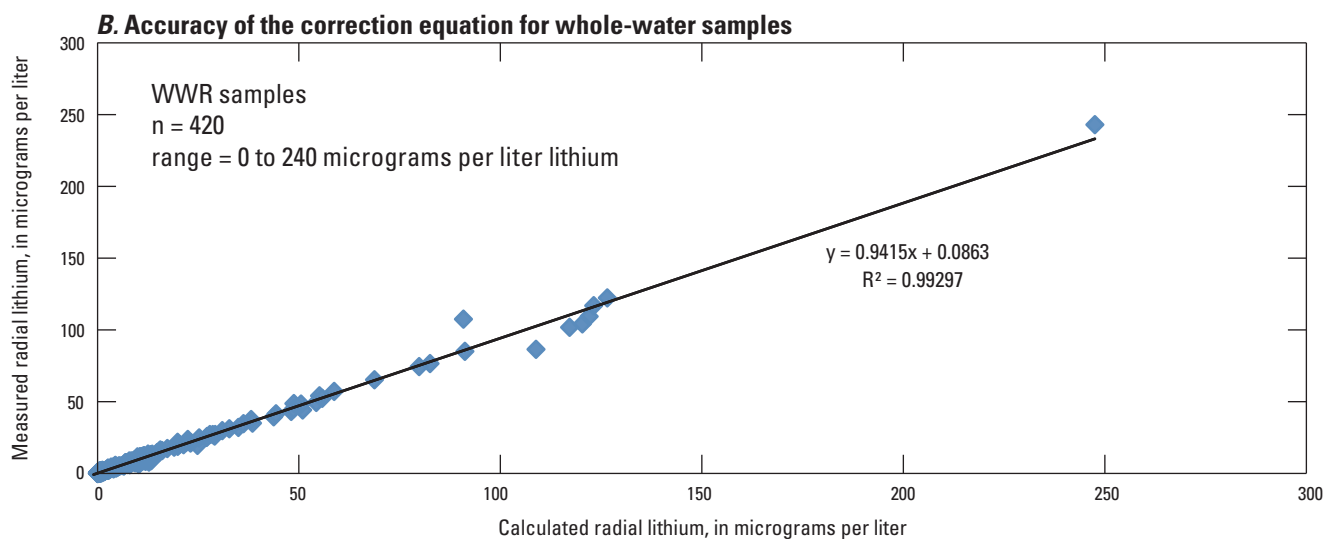
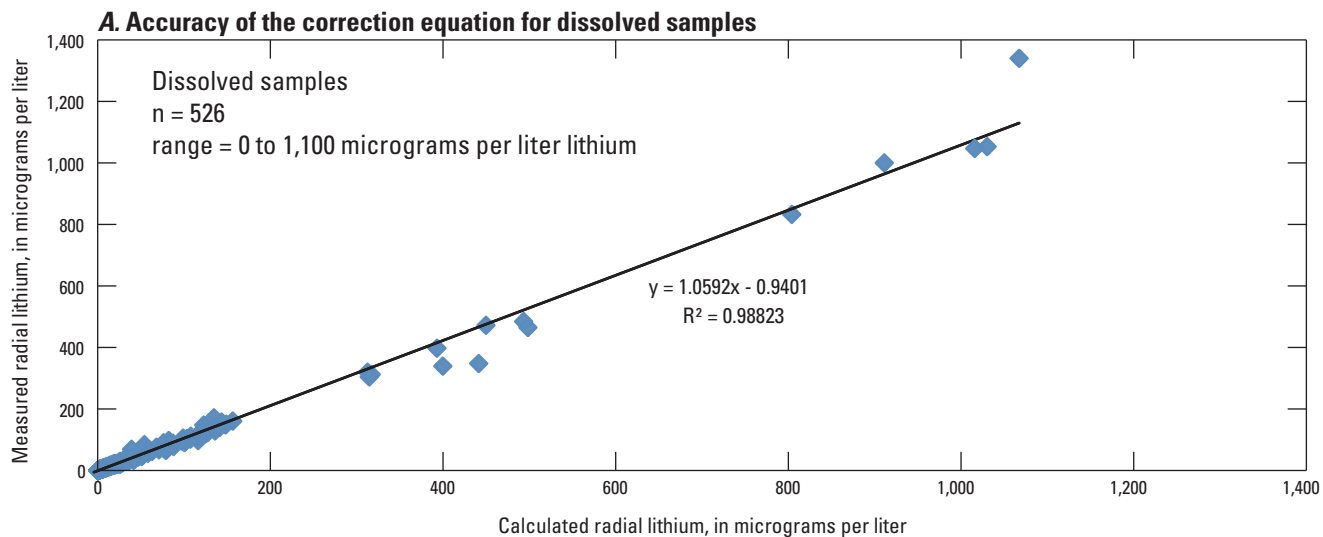


Figure B1. Accuracy of correction equation for (A) dissolved and (B) whole-water samples, Dixie Valley, west-central Nevada.



Top: Flowing artesian well, site 44, in the historic Dixie Valley settlement, viewed toward the north. Photograph taken by Chris Mahannah, Mahannah and Associates, LLC, June 9, 2009.

Bottom: Eastern part of historic Dixie Valley settlement as viewed to the east toward the Clan Alpine Mountains. Photograph taken by Jena Huntington, U.S. Geological Survey, October 26, 2011.

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