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Hydrogeologic and Geochemical Characterization of Groundwater Resources in Deep Creek Valley and Adjacent Areas, Juab and Tooele Counties, Utah, and Elko and White Pine Counties, Nevada

Scientific Investigations Report 2015–5097

U.S. Department of the Interior
U.S. Geological Survey

Windmill over an abandoned stock well on the Goshute Indian Reservation looking east with the Deep Creek Range in the background.

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By Philip M. Gardner and Melissa D. Masbruch

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SALLY JEWELL, Secretary

U.S. Geological Survey
Suzette M. Kimball, Acting Director

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Contents

Figures

Tables

Conversion Factors, Datums, and Water-Quality Units

Inch/Pound to SI

Multiply	By	To obtain
	Length	
inch (in)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
acre	0.004047	square kilometer (km^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)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per year (ft/y)	0.3048	meter per year (m/y)

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

°F=(1.8×°C)+32

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929 (NGVD29).

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

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

Concentrations of chemical constituents in water are reported in milligrams per liter (mg/L), micrograms per liter (µg/L), and in milliequivalents per liter. Milligrams per liter and micrograms per liter are units expressing the concentration of chemical constituents in solution as weight (grams) of solute per unit volume (liter) of water. A liter of water is assumed to weigh 1 kilogram, except for brines or water at high temperatures because of changes in the density of the water. For concentrations less than 7,000 mg/L or 7,000,000 µg/L, the numerical value is the same as for concentrations in parts per million or parts per billion, respectively. Milliequivalents per liter are units expressing concentrations that are chemically equivalent in terms of atomic or molecular weight and electrical charge.

Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25 °C).

Concentrations of dissolved gases are reported in cubic centimeters of gas at standard temperature and pressure per gram of water (ccSTP/g). Tritium concentration is reported in tritium units (TU) where one TU is equivalent to one molecule of tritiated water ($^3H^1HO$) in 10^{18} molecules of non-tritiated water (^1H_2O).TU or 3.2 picocuries per liter. Carbon-14 activity is reported as percent modern carbon (pMC). Stable-isotope ratios are reported as delta (δ) values, which are parts per thousand or permil (‰) difference(s) from a standard.

List of Acronyms and Abbreviations

BARCAS	Basin and Range carbonate-rock aquifer system
BCM	Basin Characterization Model
BLM	Bureau of Land Management
BP	before present
DIC	dissolved inorganic carbon
DRI	Desert Research Institute
EPA	Environmental Protection Agency
ET	evapotranspiration
GBCAAS	Great Basin carbonate and alluvial aquifer system
GIS	geographic information systems
GMWL	Global Meteoric Water Line
HA	hydrographic area
HGU	hydrogeologic unit
LBFAU	lower basin-fill aquifer unit
LCAU	lower carbonate aquifer unit
LMWL	Local Meteoric Water Line
NAIP	National Agricultural Imagery Program
NCCU	non-carbonate confining unit
NWIS	National Water Information System
PRISM	Parameter-elevation Regression on Independent Slopes Model
SNWA	Southern Nevada Water Authority
TU	tritium unit
UCAU	upper carbonate aquifer unit
UGS	Utah Geological Survey
USCU	upper siliciclastic confining unit
USGS	U.S. Geological Survey
UTWSC	Utah Water Science Center
VSMOW	Vienna Standard Mean Ocean Water
VU	volcanic unit

Abbreviations related to modeling noble-gas recharge temperatures

A	a parameter used in the CE model defined as the dimensionless ratio of the total volume of trapped (moist) air at the pressure and temperature of the free atmosphere to the volume of water beneath the water table
CE	closed-system equilibration
F	a parameter used in the CE model defined as a fractionation factor for partial dissolution of trapped air bubbles
H_r	recharge altitude
H_{rmax}	maximum recharge altitude
H_{rmin}	minimum recharge altitude
NGT	noble-gas recharge temperature
NGT_{avg}	average noble-gas recharge temperature
NGT_{max}	maximum noble-gas recharge temperature
NGT_{min}	minimum noble-gas recharge temperature
T_r	recharge temperature

Hydrogeologic and Geochemical Characterization of Groundwater Resources in Deep Creek Valley and Adjacent Areas in Juab and Tooele Counties, Utah, and Elko and White Pine Counties, Nevada

By Philip M. Gardner and Melissa D. Masbruch

Abstract

The water resources of Deep Creek Valley were assessed during 2012–13 with an emphasis on better understanding the groundwater flow system and groundwater budget. Surface-water resources are limited in Deep Creek Valley and are generally used for agriculture. Groundwater is the predominant water source for most other uses and to supplement irrigation. Most groundwater withdrawal in Deep Creek Valley occurs from the unconsolidated basin-fill deposits, in which conditions are generally unconfined near the mountain front and confined in the lower-altitude parts of the valley. Productive aquifers are also present in fractured bedrock that occurs along the valley margins and beneath the basin-fill deposits. The consolidated-rock and basin-fill aquifers are hydraulically connected in many areas with much of the recharge occurring in the consolidated-rock mountain blocks and most of the discharge occurring from the lower-altitude basin-fill deposits.

Average annual recharge to the Deep Creek Valley hydrographic area was estimated to be between 19,000 and 29,000 acre-feet. Groundwater recharge occurs mostly from the infiltration of precipitation and snowmelt at high altitudes. Additional, but limited recharge occurs from the infiltration of runoff from precipitation near the mountain front, infiltration along stream channels, and possible subsurface inflow from adjacent hydrographic areas. Groundwater moves from areas of recharge to springs and streams in the mountains, and to evapotranspiration areas, springs, streams, and wells in the basins. Discharge may also occur as subsurface groundwater outflow to adjacent hydrographic areas. Average annual discharge from the Deep Creek Valley hydrographic area was estimated to be between 21,000 and 22,000 acre-feet, with the largest portion of discharge occurring as evapotranspiration.

Groundwater samples were collected from 10 sites for geochemical analysis. Dissolved-solids concentrations ranged from 126 to 475 milligrams per liter, and none of the sites sampled during this study had dissolved-solids concentrations that exceeded the Environmental Protection Agency secondary standard for drinking water of 500 milligrams per liter. Tritium concentrations from 1.6 to 10.1 tritium units at 3 of the 10 sample sites indicate the presence of modern (less than 60 years old) groundwater, and apparent tritium/helium-3 ages calculated for these sites ranged from 7 to 29 years. The other seven sample sites had tritium concentrations less than or equal to 0.4 tritium units and are assumed to be premodern. Adjusted minimum radiocarbon ages of these seven pre-modern water samples ranged from 1,000 to 8,000 years with the ages of at least four of the samples being more than 3,000 years. Noble-gas recharge temperatures indicate that groundwater sampled along the valley axis recharged at both mountain and valley altitudes, providing evidence for both mountain-block and mountain-front recharge.

Water-level altitude contours and groundwater ages indicate the potential for a long flow path from southwest to northeast between northern Spring and Deep Creek Valleys through Tippett Valley. Although information gathered during this study is insufficient to conclude whether or not groundwater travels along this interbasin flow path, dissolved sulfate and chloride data indicate that a small fraction of the lower altitude, northern Deep Creek Valley discharge may be sourced from these areas. Despite the uncertainty due to limited data collection points, a hydraulic connection between northern Spring Valley, Tippett Valley, and Deep Creek Valley appears likely, and potential regional effects resulting from future groundwater withdrawals in northern Spring Valley warrant ongoing monitoring of groundwater levels across this area.

Introduction

Deep Creek Valley is located in Tooele and Juab Counties, Utah, and Elko and White Pine Counties, Nevada, along the state border about 120 miles (mi) southwest of Salt Lake City, Utah (fig. 1). The Confederated Tribes of the Goshute Indian Reservation encompasses about 112,900 acres in the southern end of the valley, including part of the headwaters of the Deep Creek watershed and lands on the divide between Deep Creek Valley and Tippett Valley to the west. Few perennial streams flow into the basin, and surface-water resources are limited. Groundwater resources that sustain streams, springs, wetlands, and the local agricultural economy are also limited.

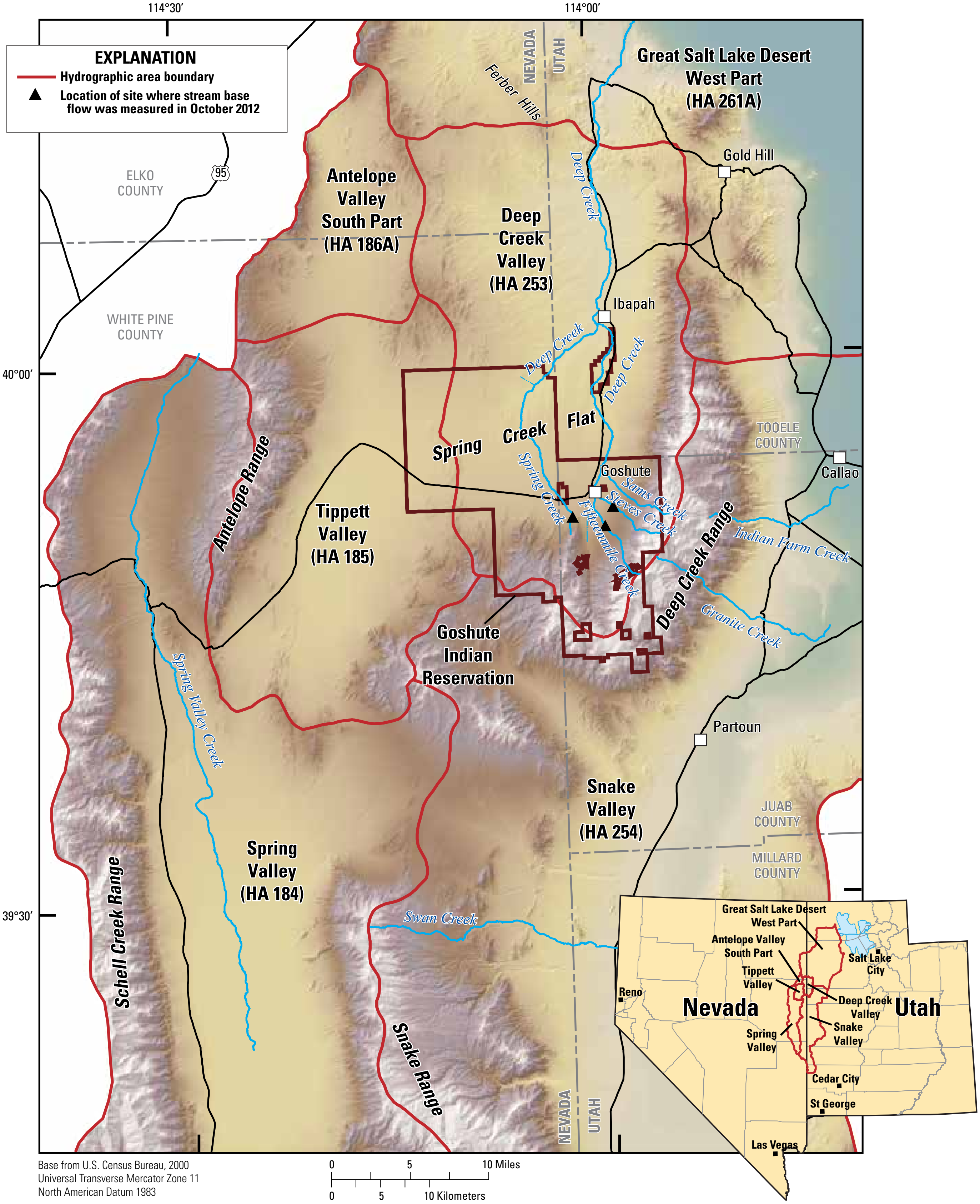

Figure 1. Location of Deep Creek Valley and adjacent hydrographic areas, Utah and Nevada.

Minimal growth has occurred within Deep Creek Valley during the past 40 years; however, both Snake Valley to the east-southeast and Spring Valley to the southwest have been targeted for significant groundwater development. Present concerns include the Southern Nevada Water Authority (SNWA) groundwater project, which proposes withdrawals of more than 60,000 acre-feet (acre-ft) of groundwater from Spring Valley over the next two decades. Generalized hydrogeology in the study area indicates possible interbasin hydraulic connection from Spring Valley through Tippett Valley to Deep Creek Valley and warrants a better understanding of the groundwater resources of the region to assess the potential effect of the SNWA project on the Deep Creek Valley aquifer(s).

An initial hydrologic reconnaissance of Deep Creek Valley was completed by Hood and Waddell (1969). Results for this study were incorporated into a hydrologic reconnaissance of the southern Great Salt Lake Desert (Gates and Kruer, 1981). Since these studies, water levels have been measured annually in selected wells within Deep Creek Valley as part of the U.S. Geological Survey (USGS) Utah Water Science Center (UTWSC) annual groundwater monitoring program (Burden and others, 2013).

Cooperative efforts are improving the understanding of the groundwater resources through recent and ongoing investigations in east-central Nevada and western Utah. The Basin and Range carbonate-rock aquifer system (BARCAS) study (Welch and others, 2007), conducted by the USGS and the Desert Research Institute (DRI), provides a regional assessment of groundwater in 13 hydrographic areas adjacent to Deep Creek Valley and the Goshute Reservation, and includes revised estimates of interbasin groundwater flow. Prudic and Glancy (2009) completed a detailed assessment of the sources of water to Cave Springs, located in Great Basin National Park on the western side of Snake Valley. Heilweil and Brooks (2011) completed a groundwater availability study of the Great Basin carbonate and alluvial aquifer system, including the entire carbonate-rock terrain from Death Valley, California, in the southwest, to Cache Valley, Utah in the northeast. Halford and Plume (2011), in cooperation with the National Park Service, refined and recalibrated the Great Basin Regional Aquifer-System Analysis numerical model (Prudic and others, 1995) to assess the hydrologic effects of developing groundwater in Snake Valley on the water resources of Great Basin National Park. Both the Utah Geological Survey (UGS) and the SNWA have drilled an extensive network of monitoring wells in western Utah and eastern Nevada, extending from near Fish Springs to southern Snake Valley (Hamlin Valley) and Spring Valley. In cooperation with the UGS, SNWA, and the DRI, the UTWSC has also collected and compiled a large geochemical dataset from wells and springs in the region (Hershey and others, 2007; Acheampong and others, 2009; Gardner and Heilweil, 2014; Utah Department of Natural Resources, 2014). In addition, with financial support from SNWA, the UTWSC maintained a water-level monitoring network of 76 wells in western Utah between the Snake Range in Snake Valley and Fish Springs National Wildlife refuge in Fish Springs Flat. Water levels in these wells were measured quarterly from September 2008 through September 2014 and are accessible from the USGS National Water Information System (NWIS) database at *http://waterdata.usgs.gov/nwis.*

Purpose and Scope

The purpose of this report is to reassess the groundwater hydrology and refine the conceptual model of the groundwater flow system in the southern end of Deep Creek Valley and eastern Tippett Valley that encompasses the Goshute Reservation. A five-point approach was used to meet the objectives of this study including (1) development of a new potentiometric surface map to assess direction of groundwater movement; (2) compilation of historical information on groundwater withdrawals, water-level changes, and stream and spring discharge measurements to provide a baseline of conditions for interpretation of future changes; (3) estimation of updated recharge and discharge rates using the Basin Characterization Model (BCM), a geographic information systems (GIS) based evapotranspiration analysis, and information regarding water-related land use; (4) verification of the hydrogeologic framework and development of hydrogeologic cross sections through areas with potential interbasin flow; and (5) an assessment of geochemical and environmental tracers in groundwater from wells and springs to determine groundwater ages, flow paths, and recharge sources. Results of the study will provide hydrologic data intended to better quantify current hydrologic conditions in the Goshute Reservation area, and to assess potential effects of groundwater withdrawals on groundwater and surface-water resources.

Physical Characteristics of the Study Area

Deep Creek Valley (fig. 1) is located in the Basin and Range physiographic province (Fenneman, 1931) and exhibits geologic and topographic characteristics typical of the region. The study area is part of the Great Basin carbonate and alluvial aquifer system (GBCAAS), which comprises aquifers and confining units in unconsolidated basin-fill and volcanic deposits in the basins, carbonate, and other bedrock units in the mountain ranges separating the basins (Heilweil and others, 2011). In some areas of the GBCAAS, aquifers are hydraulically connected between basins. Basins within the study area are divided on the basis of hydrographic area (HA) boundaries (Harrill and others, 1988), which generally coincide with topographic basin divides. This study was focused on Deep Creek Valley (HA 253) but also extends into Tippett Valley (HA 185), Antelope Valley South Part (HA 186A), and the northern part of Spring Valley (HA 184) (fig. 1). Deep Creek Valley is approximately 450 square miles (mi^2). The basin-fill occupies an area of about 315 mi^2 at altitudes between about 5,000 and 8,000 feet (ft), surrounded by the contributing mountain watersheds that make up the remaining

125 mi^2. Deep Creek Valley is bounded by the Deep Creek Range to the east and south, an unnamed area of volcanic badlands and low hills to the west, and the Ferber Hills to the north (fig. 1).

Altitudes in Deep Creek Valley range from about 5,000 ft at the north end of the valley to over 12,000 ft in the highest parts of the Deep Creek Range. The highest points in the study area are found in the central part of the Deep Creek Range (fig. 1). The lowest point in the study area is at the northern end of the HA where Deep Creek drains out of the HA to the north.

Population and Land Use

Deep Creek Valley is sparsely populated. The only town is Ibapah, which had a population of 128 in the 2000 census (U.S. Census Bureau, 2000) and has declined since 1960 when the population was 213 (Hood and Waddell, 1969). The Goshute Reservation had a resident population of 105 in the 2000 census (U.S. Census Bureau, 2000).

Land use within Deep Creek Valley includes irrigated and nonirrigated farmland and pasture used for agriculture and livestock grazing, unincorporated residential areas, and mining. The adjacent mountain areas remain mostly undeveloped and are used primarily for recreation, mining, and grazing.

Precipitation

The average annual precipitation (1981–2010) estimated from Parameter-elevation Regression on Independent Slopes Model (PRISM) data (Daly and others, 2008) in Deep Creek Valley ranges from about 9 inches (in.) in the basin to about 30 in. at the higher altitudes of the Deep Creek Range (fig. 2). Most precipitation occurs during the winter and early spring months as snowfall and the least occurs during July and August. Four weather stations located in or near Deep Creek Valley with long-term records of precipitation illustrate the variation in annual precipitation (fig. 3). Although the Gold Hill station has a shorter period of record than the other stations (1966–1990), all four stations recorded extended periods (greater than about 5 years) of below average precipitation (1972–1976) and above average precipitation (1980–1986). The other three stations also recorded an extended period of below average precipitation from 1953–1962, which corresponds to the southwestern regional drought lasting from about 1953–1965 (U.S. Geological Survey, 1991). It is estimated that all of the precipitation that falls at valley altitudes (below about 6,000 ft) is consumed by evapotranspiration (ET). Precipitation falling at higher altitudes generally exceeds the amount consumed by ET and becomes either direct infiltration into the mountain block or runoff in streams draining the mountains.

Streamflow

Most streams in Deep Creek Valley are intermittent and flow only in response to periods of snowmelt or intense rainfall. Much of the water in these intermittent streams is lost to ET or infiltration on the alluvial slopes of the valley. Small amounts of streamflow occasionally reach Deep and Spring Creeks in the valley, where most of this water is diverted for irrigation. A small amount of water intermittently flows out of the valley in Deep Creek toward Great Salt Lake Desert (fig. 1).

Four perennial streams, Spring, Fifteenmile, Sams, and Steves Creeks, drain the mountains on the southeast side of Deep Creek Valley. High flow in these streams occurs during the spring months when they collect runoff from the melting high-altitude snowpack in the Deep Creek Range. These streams are sustained year round by groundwater discharge either from mountain springs or as base flow directly to the stream.

There is limited historical measurement data to estimate streamflow. Fifteenmile, Sams, and Steves Creeks were measured on an almost monthly basis from April 1964 to September 1967 (Hood and Waddell, 1969; fig. 4). Additionally, average daily streamflow for Deep Creek near Goshute, Utah (USGS gage 10172893; data available at *http://waterdata.usgs.gov/nwis/inventory/?site_no=10172893&agency_cd=USGS&)* was measured from April 1964 to September 1968, and annual peak streamflow for Deep Creek near Ibapah (USGS gage 10172895; data available at *http://waterdata.usgs.gov/nwis/inventory/?site_no=10172895&agency_cd=USGS&)* was measured from 1959 to 1968 (fig. 5).

Average annual (1971–2000) streamflow was estimated for Fifteenmile, Sams, and Steves Creeks (table 1) using regional regression equations that were developed to predict mean annual streamflow at ungaged sites in Utah (Wilkowske and others, 2008). The regression equations used to calculate annual streamflow are based only on basin size and average annual precipitation, and some of these parameters for these creeks were outside of the suggested range used in the development of the regression equations. These estimates, therefore, are an extrapolation with unknown errors.

Table 1. Average annual streamflow, 1971–2000, for perennial streams in Deep Creek Valley, Utah and Nevada.

[Streamflow estimated using the U.S. Geological Survey's StreamStats program, available at *http://water.usgs.gov/osw/streamstats/*. Estimates are extrapolations with unknown errors. Abbreviations: mi^2, square miles; ft^3/s, cubic feet per second]

Stream name	Drainage area (mi^2)	Average annual streamflow (ft^3/s)
Fifteenmile Creek	6.09	4.6
Sams Creek	2.50	2.5
Steves Creek	1.97	2.1

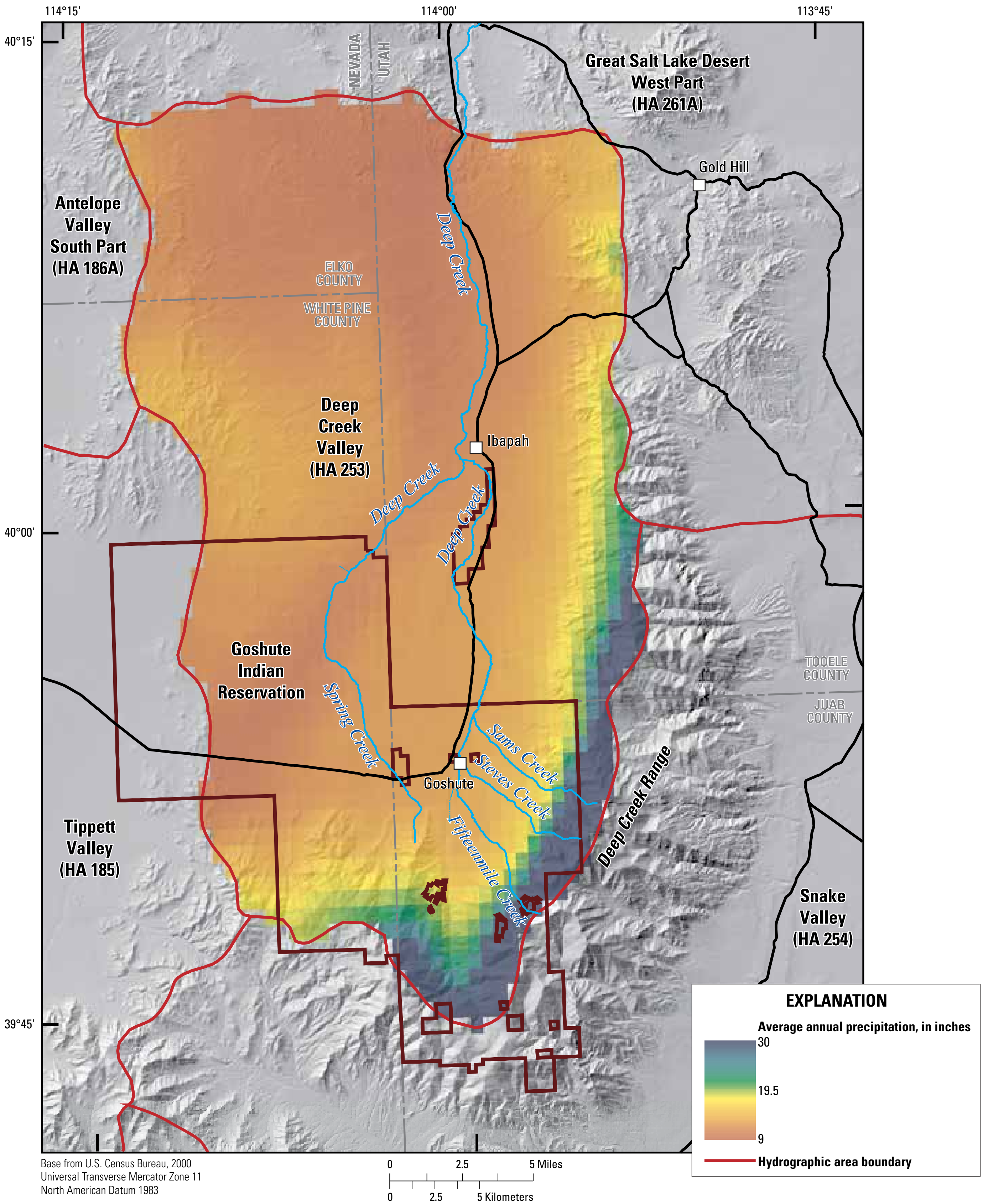

Figure 2. Average annual precipitation, 1981–2010, in Deep Creek Valley, Utah and Nevada.

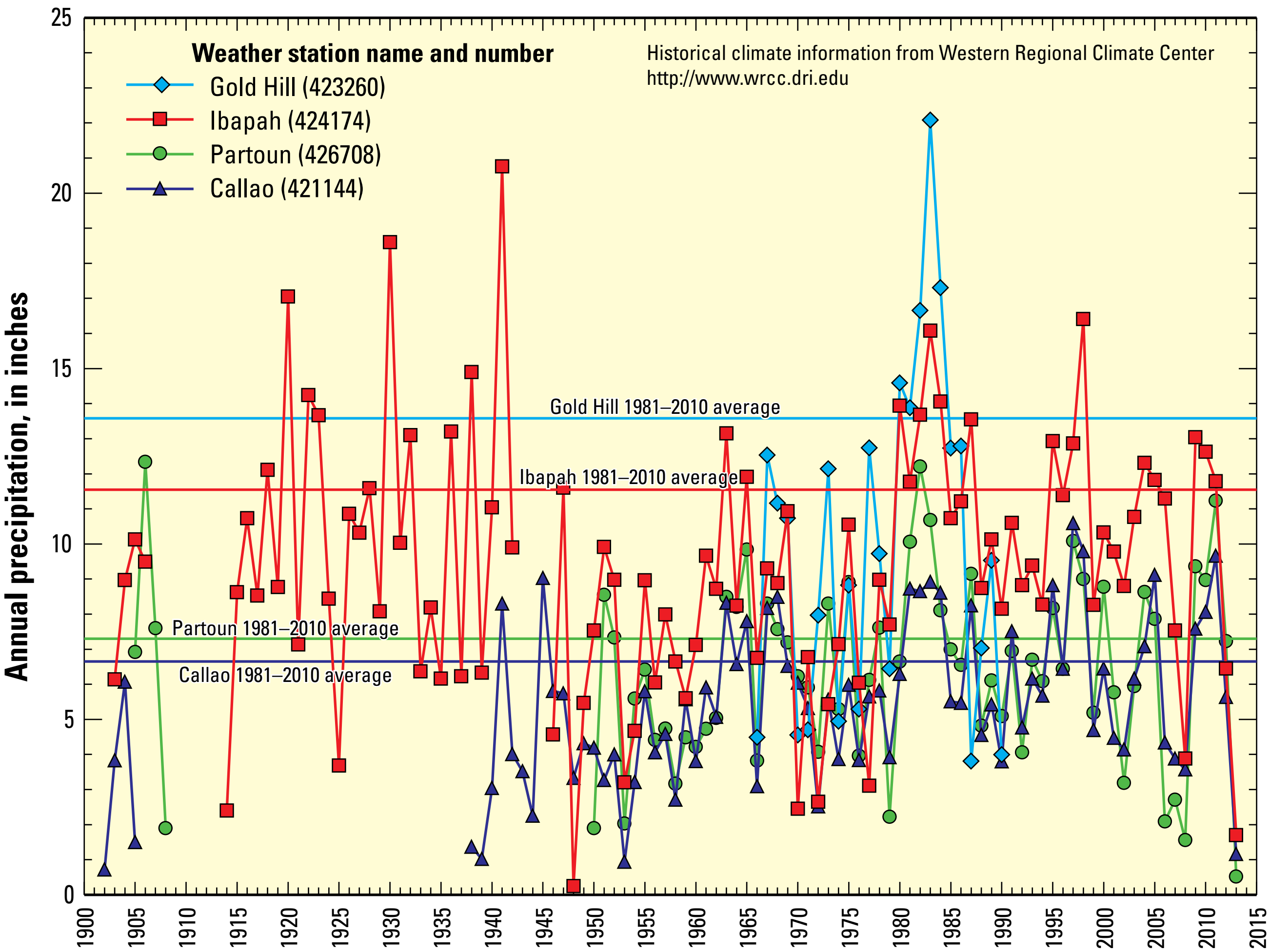

Figure 3. Long-term annual precipitation recorded at four weather stations in or near Deep Creek Valley, Utah and Nevada.

Base-flow measurements were made at Spring, Fifteenmile, and Steves Creeks in October 2012 (table 2 and fig. 1). Base flow measured in 2012 in Fifteenmile and Steves Creeks was approximately 50 percent of the base flow measured in the mid-1960s.

Table 2. Base-flow measurements and comparison to historical base flow for perennial streams in Deep Creek Valley, Utah and Nevada.

[Abbreviations: ft^3/s, cubic feet per second; NA, not applicable]

Stream name	Date	2012 base flow (ft^3/s)	1964–1966 base flow (ft^3/s)[1]
Spring Creek	10/11/12	2.72	NA
Fifteenmile Creek	10/11/12	1.26	3.02
Steves Creek	10/11/12	0.90	1.59
Sams Creek	NA	NA	0.56

[1]Average of October 1964, 1965, and 1966 measurements from Hood and Waddell (1969, table 11).

Geology

Deep Creek Valley is a large, north-south trending, internally drained basin that is defined by a series of narrow, normal-fault-bounded bedrock mountain ranges and adjoining low hills that surround a broad, gently sloping valley floor, typical of the Basin and Range physiographic province (Fenneman, 1931). Bedrock in the mountains and hills surrounding and within Deep Creek Valley is characterized by a thick section of complexly faulted and folded Precambrian- through Permian-age metasedimentary and sedimentary rocks intersected by and capped with Tertiary- to Holocene-age igneous and sedimentary rocks (Hood and Waddell, 1969). The basin fill in Deep Creek Valley includes a range of semi-consolidated to unconsolidated sediments eroded from the surrounding mountains as a result of weathering processes, including glaciation during the Pleistocene Epoch (Hood and Waddell, 1969). The unconsolidated basin fill contains the principal aquifers in Deep Creek Valley (Hood and Waddell, 1969).

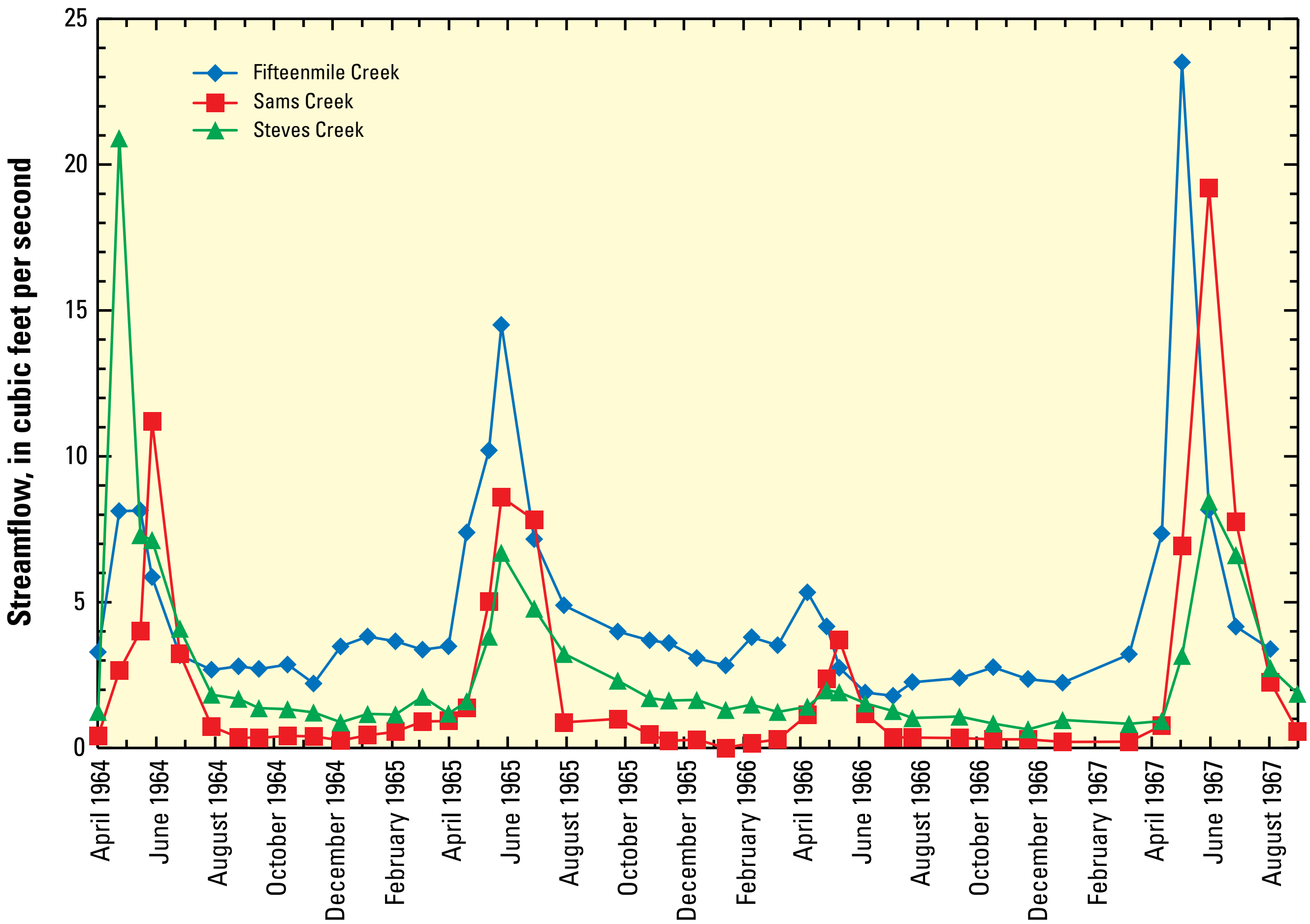

Figure 4. Streamflow measured in Fifteenmile, Sams, and Steves Creeks for 1964–1967, Deep Creek Valley, Utah and Nevada.

The early tectonic history of Deep Creek Valley is recorded by exposed Precambrian- to Permian-age strata that were deposited initially across a broad subsiding marine platform and later within the rapidly subsiding Oquirrh Basin. Rocks deposited during this period include (1) Precambrian- to Early Cambrian-age quartzite, shale, and conglomerate; (2) a thick sequence of Middle Cambrian- to Mississippian-age limestone, dolomite, sandstone, shale, and quartzite; and (3) Pennsylvanian- to Early Permian-age limestone, sandstone, and quartzite. These rocks were folded and faulted by predominantly eastward thrust faulting and compression during the Late Jurassic- to Eocene-age Sevier orogenic event (Armstrong, 1968; DeCelles and Coogan, 2006).

During the Eocene Epoch, crustal shortening was replaced by roughly east-west extension and significant regional volcanism (Constenius, 1996; Constenius and others, 2003). Early extension, localized along north-south striking normal faults, controlled the formation of narrow, rapidly subsiding basins into which sediment from surrounding uplands and nearby volcanic centers was deposited. Basin-fill deposits include (1) Eocene- or Oligocene-age intrusive and extrusive volcanics and pyroclastics, (2) consolidated to semiconsolidated Eocene- to Pliocene-age tuffaceous sedimentary rocks and interbedded pyroclastics, and (3) unconsolidated latest Pleistocene- to Holocene-age alluvial, colluvial, and glacial outwash deposits (Hood and Waddell, 1969). Extension remains the predominant tectonic force in the area, but has varied in magnitude, style, and extent during Eocene to Holocene time.

Groundwater Hydrology

The groundwater system in the study area consists of water in unconsolidated deposits in the basins and water in consolidated rock underlying the basins and in the adjacent mountain blocks. The consolidated-rock and basin-fill aquifers are well connected hydraulically (Sweetkind and others, 2011b), with most of the recharge occurring in the consolidated-rock mountain blocks and most of the discharge occurring from the lower altitude basin-fill deposits. Additionally, there is a possible interbasin hydraulic connection from Spring Valley through Tippett Valley to Deep Creek Valley.

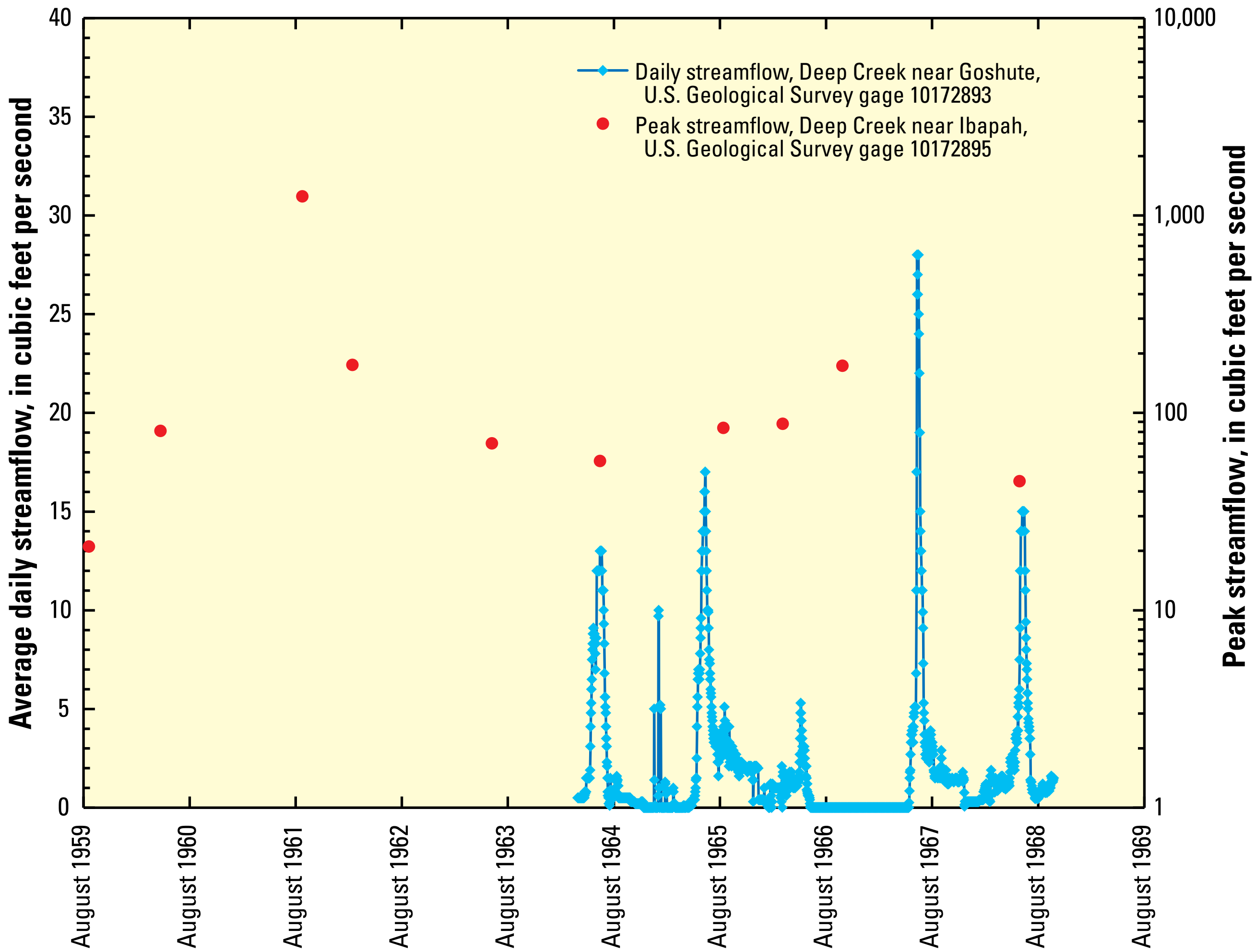

Figure 5. Average daily streamflow for 1964–1968 and peak streamflow for 1959–1968, measured in Deep Creek, Deep Creek Valley, Utah and Nevada.

Groundwater is the primary source of drinking water in Deep Creek Valley and also is used for irrigation and stock watering. Aquifers are present in both bedrock and unconsolidated basin-fill deposits. The majority of wells in the study area are completed within the basin-fill deposits because of the ease of drilling, accessibility, and proximity to populated areas.

Hydrogeology

The geologic units in the study area both store and convey groundwater and place basic controls on regional groundwater movement. As part of the GBCAAS study, a three-dimensional hydrogeologic framework of the eastern Great Basin was constructed (Cederberg and others, 2011; Sweetkind and others, 2011a). The GBCAAS study area is inclusive of the current study area; therefore, this same hydrogeologic framework was used in the current study. The framework was constructed using data from a variety of sources, including geologic maps and cross sections, drillhole data, geophysical models, and stratigraphic surfaces created for other three-dimensional hydrogeologic frameworks within the GBCAAS study area. The framework was developed using a 1-mi^2 grid cell size.

The geologic complexity, scale of the study area, and potential for groundwater to reside in and travel through multiple geologic units necessitate generalization of the aquifer system. In the hydrogeologic framework developed for the GBCAAS, the consolidated pre-Cenozoic-age rocks, Cenozoic sediments, and igneous rocks in the study area were subdivided into nine hydrogeologic units (HGUs) (Sweetkind and others, 2011a). An HGU has considerable lateral extent and reasonably distinct physical characteristics that may be used to infer the capacity of a sediment or rock to transmit water. The definition of HGUs is important for conceptualizing the hydrogeologic system and construction of a geologic framework for describing the groundwater flow system.

Of the nine HGUs defined in the hydrogeologic framework developed for the GBCAAS, seven exist in the current study area (figs. 6 and 7). The HGUs that exist in the current study

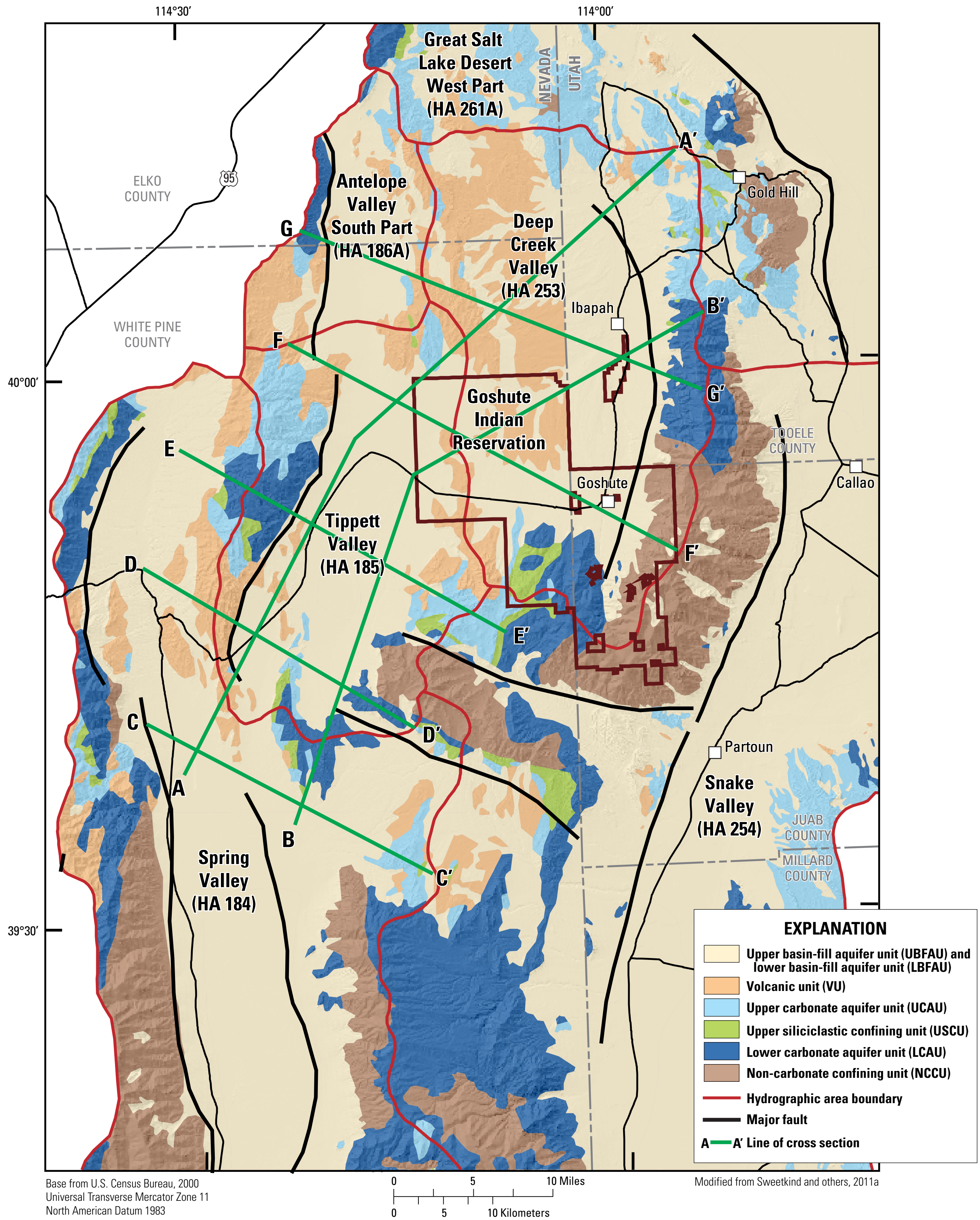

Figure 6. Surficial extent of hydrogeologic units, major faults, and lines of cross section, in Deep Creek Valley and adjacent areas, Utah and Nevada.

area are (1) a non-carbonate confining unit (NCCU) representing low- to moderate-permeability Precambrian-age siliciclastic formations as well as intrusive igneous rocks that are locally exposed in mountain ranges, and underlie parts of the study area; (2) a lower carbonate aquifer unit (LCAU) representing a thick succession of predominantly moderate- to high-permeability Cambrian through Devonian-age carbonate rocks that are locally exposed in the mountain ranges, and present beneath most of the valleys within the study area; (3) an upper siliciclastic confining unit (USCU) representing low-permeability Mississippian-age siliciclastic rocks, predominantly shales, that are limited in extent within the study area; (4) an upper carbonate aquifer unit (UCAU) representing a thick succession of low- to high-permeability Pennsylvanian- and Permian-age carbonate rocks that are locally exposed in the mountain ranges and exist beneath some of the valleys within the study area; (5) a volcanic unit (VU) representing large volumes of low- to high-permeability Cenozoic-age volcanic rocks that are locally exposed in the mountain ranges and exist beneath some of the valleys within the study area; (6) a lower basin-fill aquifer unit (LBFAU) representing the lower (deepest) one-third of the Cenozoic basin-fill sediments, including moderate- to high-permeability volcanic rocks buried within the basin fill and consolidated older basin-fill sediments; and (7) an upper basin-fill aquifer unit (UBFAU) representing the upper (shallowest) two-thirds of the Cenozoic basin-fill sediments, including a wide variety of low- to moderate-permeability basin-fill sediments (Sweetkind and others, 2011a).

Hydraulic Properties

Hydraulic properties describe the ability of a groundwater system to transmit and store water. The distribution of these properties in the study area is variable and depends on the depositional environment of sediments in the basin-fill aquifer and confining units, and on the degree of structural deformation, fracturing, and (or) chemical dissolution in the bedrock aquifers and confining units.

Sweetkind and others (2011a) estimated thickness and hydraulic properties of the HGUs in the GBCAAS study area (table 3). These were taken from studies by Belcher and others (2001, 2002) that analyzed and compiled estimates of transmissivity, hydraulic conductivity, storage coefficients, and anisotropy ratios for HGUs within the Death Valley regional groundwater flow system. HGUs within the Death Valley area are similar in origin and analogous to those within the current study area.

Occurrence and Movement of Groundwater

Groundwater in Deep Creek Valley occurs in both unconsolidated basin-fill and in consolidated-rock aquifers under confined and unconfined conditions. Within the basin fill, unconfined or water-table conditions generally exist along the valley margins within alluvial fan and colluvial deposits, and confined conditions generally exist in the lowest parts of the valley. Groundwater moves under confined conditions where lacustrine and fluvial deposits have created zones of permeable material mixed with semicontinuous to continuous layers of low-permeability clay or silt. Although unconfined groundwater movement occurs within most bedrock mountain areas, structural geologic features and variations in lithology likely result in localized areas of confined conditions.

Groundwater recharge occurs mostly from the infiltration of precipitation at high altitudes (Welch and others, 2007; San Juan and others, 2010; Masbruch and others, 2011). Much of this recharge occurs in the form of snowmelt. Additional, but limited recharge occurs from the infiltration of runoff from precipitation near the mountain front, and infiltration along stream channels (Hevesi and others, 2003; Flint and Flint, 2007a, b; Flint and others, 2011; Masbruch and others, 2011). There also may be recharge from applied irrigation; however, most of this applied water likely evaporates or is consumed by crops before reaching the water table. Groundwater moves from areas of recharge to springs and streams in the mountains, and to evapotranspiration areas, springs, streams, and wells in the basins.

Table 3. Hydraulic properties of hydrogeologic units from the Great Basin carbonate and alluvial aquifer system study area.

[Modified from Belcher and others, 2001, 2002; and Sweetkind and others, 2011a. Abbreviations: UBFAU, upper basin-fill aquifer unit; LBFAU, lower basin-fill aquifer unit; LCAU, lower carbonate aquifer unit; NCCU, non-carbonate confining unit; UCAU, upper carbonate aquifer unit; USCU, upper siliciclastic confining unit; VU, volcanic unit; >, greater than; NC, not calculated]

Major hydrogeologic unit	Hydrogeologic unit abbreviation	Hydraulic conductivity (feet per day)			
		Arithmetic mean	Geometric mean	Minimum	Maximum
Cenozoic basin-fill sediments	UBFAU and LBFAU	31	4	0.0001	431
Cenozoic volcanic rock	VU	20	3	0.04	179
Upper Paleozoic carbonate rock	UCAU	62	0.4	0.0003	1,045
Upper Paleozoic siliciclastic confining rock	USCU	0.4	0.06	0.0001	3
Lower Paleozoic carbonate rock	LCAU	169	4	0.009	2,704
Non-carbonate confining rock	NCCU	0.8	0.008	0.00000009	15

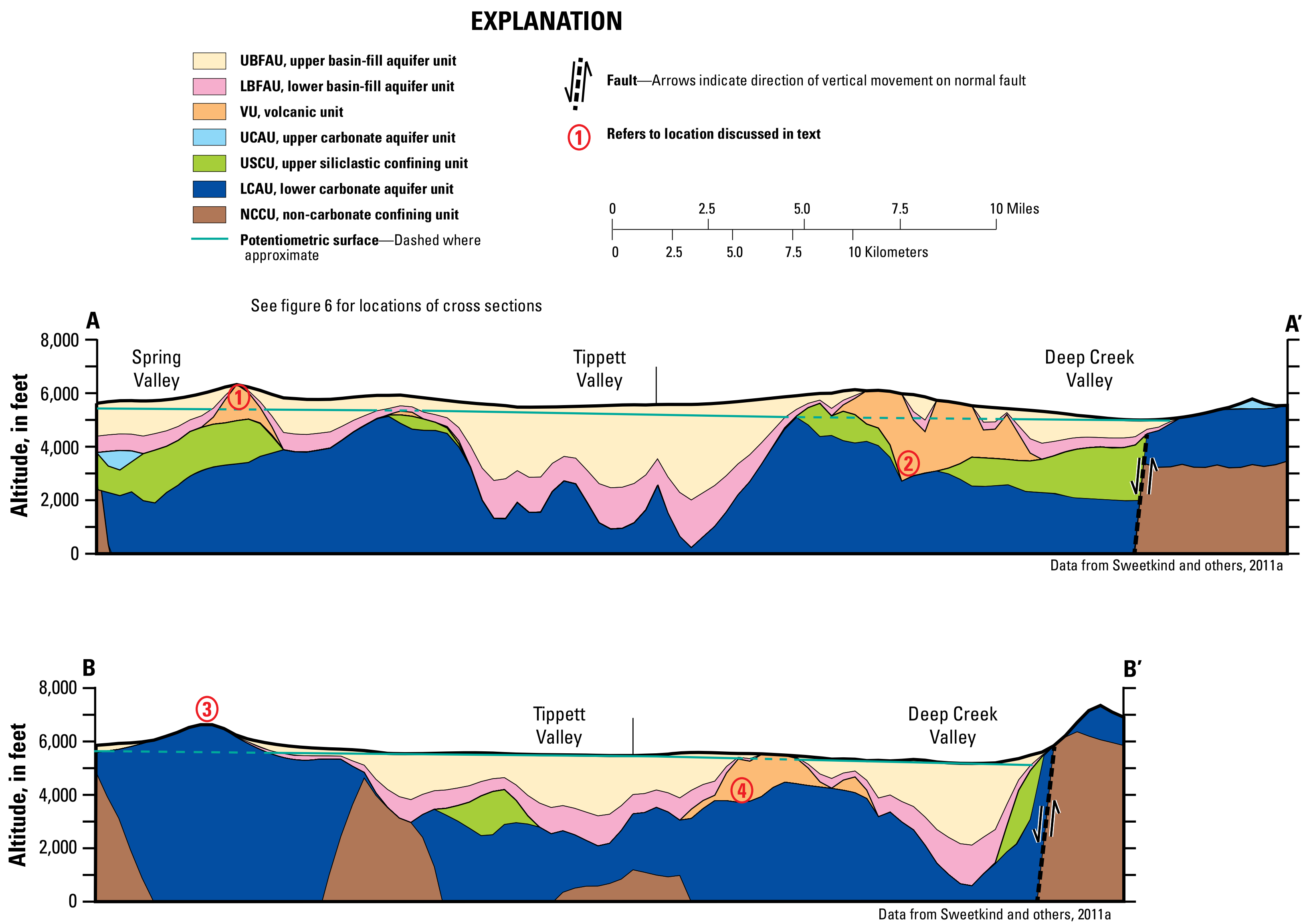

Figure 7. Hydrogeologic cross sections (*A-A′* to *G-G′*) for selected locations in Deep Creek Valley and adjacent areas, Utah and Nevada.

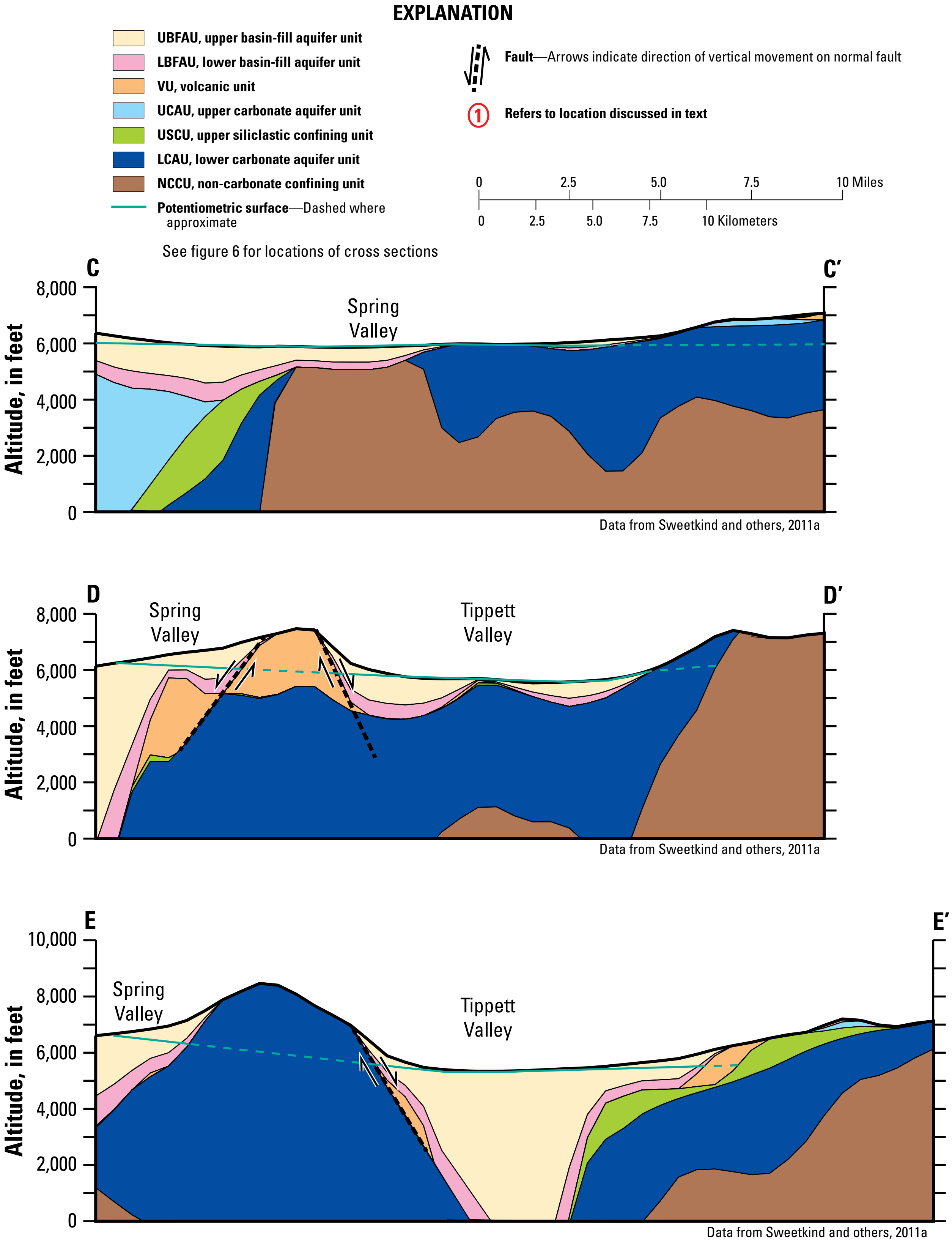

Figure 7. Hydrogeologic cross sections (*A-A′* to *G-G′*) for selected locations in Deep Creek Valley and adjacent areas, Utah and Nevada. —Continued

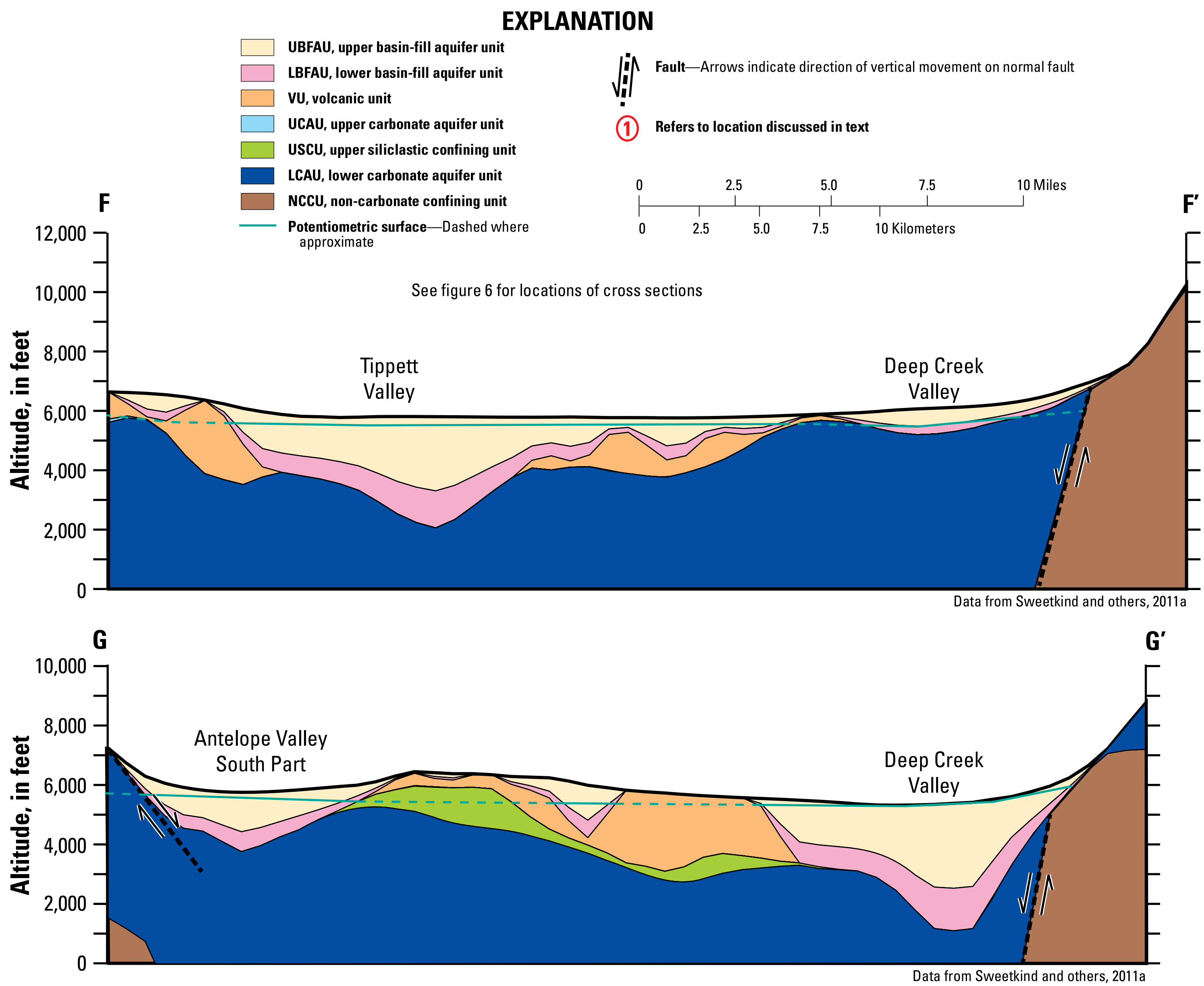

Figure 7. Hydrogeologic cross sections (*A-A′* to *G-G′*) for selected locations in Deep Creek Valley and adjacent areas, Utah and Nevada. —Continued

A water-level surface map was developed for the study area from (1) water-level measurements made at 28 wells during February and March of 2012, and (2) historical water levels at 13 wells measured between 1960 and 2010, to show general directions of groundwater movement (fig. 8, table A1–1). Groundwater movement is generally from the mountains toward the center of the basins. Groundwater in Deep Creek Valley moves from the south and east toward Deep Creek along the central axis of the basin, and then flows northward. A groundwater divide is present in northern Spring Valley, indicating the potential for groundwater recharge in northern Spring Valley to either move south into the central part of the valley, or move north into Tippett Valley and toward Deep Creek Valley (fig. 8). The likelihood of hydraulic connection across these HA boundaries based on geology (Sweetkind and others, 2011b) is also illustrated on figure 8.

Groundwater Budget

Development of a groundwater budget is important in understanding the occurrence and movement of groundwater in the flow system, and in evaluating the balance between flow into and out of the system. The primary components of the groundwater budget are recharge from precipitation (including direct infiltration and infiltration of runoff at lower altitudes), infiltration of mountain stream base flow, and infiltration of unconsumed water applied for irrigation. Discharge components of the budget include evapotranspiration of groundwater (ETg), springs, mountain streams, and well withdrawals. Recharge or discharge as subsurface (lateral) flow into or out of an HA or the study area also may be occurring.

The current study considers all forms of recharge to and discharge from the groundwater system, including the surrounding mountains. This is illustrated by considering the fate of recharge from direct infiltration of mountain precipitation and subsurface inflow from adjacent areas to permeable consolidated rock of the mountain block (fig. 9, R1 and R3). Part of this recharge moves directly through the subsurface from the mountain block into the adjacent unconsolidated basin fill. Another part of this recharge becomes groundwater discharge to mountain streams and springs (fig. 9, D1). A fraction of this mountain-block groundwater discharge is consumptively lost to evapotranspiration, both in the mountains and as this water enters the valley in streams, and a fraction of the remaining water in the streams, combined with surface-water runoff becomes recharge to the unconsolidated basin fill (fig. 9, R2). This water ultimately discharges in the valley lowlands as evapotranspiration or to basin-fill springs and streams (fig. 9, D2 and D3), wells (fig. 9, D4), or subsurface outflow (fig. 9, D5).

The groundwater budget for the current study is compiled from estimates of average annual recharge to and discharge from the Deep Creek Valley HA (table 4). Although records used to construct the individual budget components vary somewhat in their temporal length, the budget is intended to represent average conditions over approximately the last half century. Previous studies of western basins in Utah generally developed groundwater budgets that focused only on the basin fill (valley) part of the basin (Hood and Waddell, 1969) where groundwater was being developed as a resource. The groundwater budget compiled for Deep Creek Valley uses annual net recharge and discharge of the complete groundwater system, including the bedrock of the surrounding mountains and underlying the basin fill.

Table 4. Average annual groundwater budget for Deep Creek Valley, Utah and Nevada.

[Average annual volume in acre-feet per year. Abbreviations: NA, not applicable]

Recharge	Utah	Nevada	Total	Uncertainty (percent)[1]
Direct infiltration of precipitation (in-place recharge)	11,000	5,000	16,000	50
Infiltration of runoff (includes unconsumed surface-water irrigation)	870	280	1,100	50
Mountain stream base flow	390	NA	390	NA
Unconsumed irrigation from well withdrawals	negligible	negligible	negligible	NA
Subsurface inflow from Tippett Valley[2]	NA	2,000 to 12,000	2,000 to 12,000	unknown
Total:			**19,000 to 29,000**	**50**
Discharge				
Groundwater evapotranspiration	12,500	1,500	14,000	30
Mountain springs and streams (base flow)[3]	3,900	0	3,900	8
Well withdrawals[4]	600	negligible	600	unknown
Subsurface outflow to Great Salt Lake Desert[5]	2,700 to 3,000	0	2,700 to 3,000	unknown
Total:			**21,000 to 22,000**	**30**

[1]Uncertainty estimates are explained in appendix 2.

[2]Based on estimates from Scott and others (1971) and Welch and others (2007).

[3]Based on base-flow measurements made in October 2012 for Spring, Fifteenmile, and Steves Creeks, and average October 1964–1966 streamflow for Sams Creek.

[4]Based on 1966 estimates from Hood and Waddell (1969).

[5]Based on estimates from Scott and others (1971) and Harrill and others (1988).

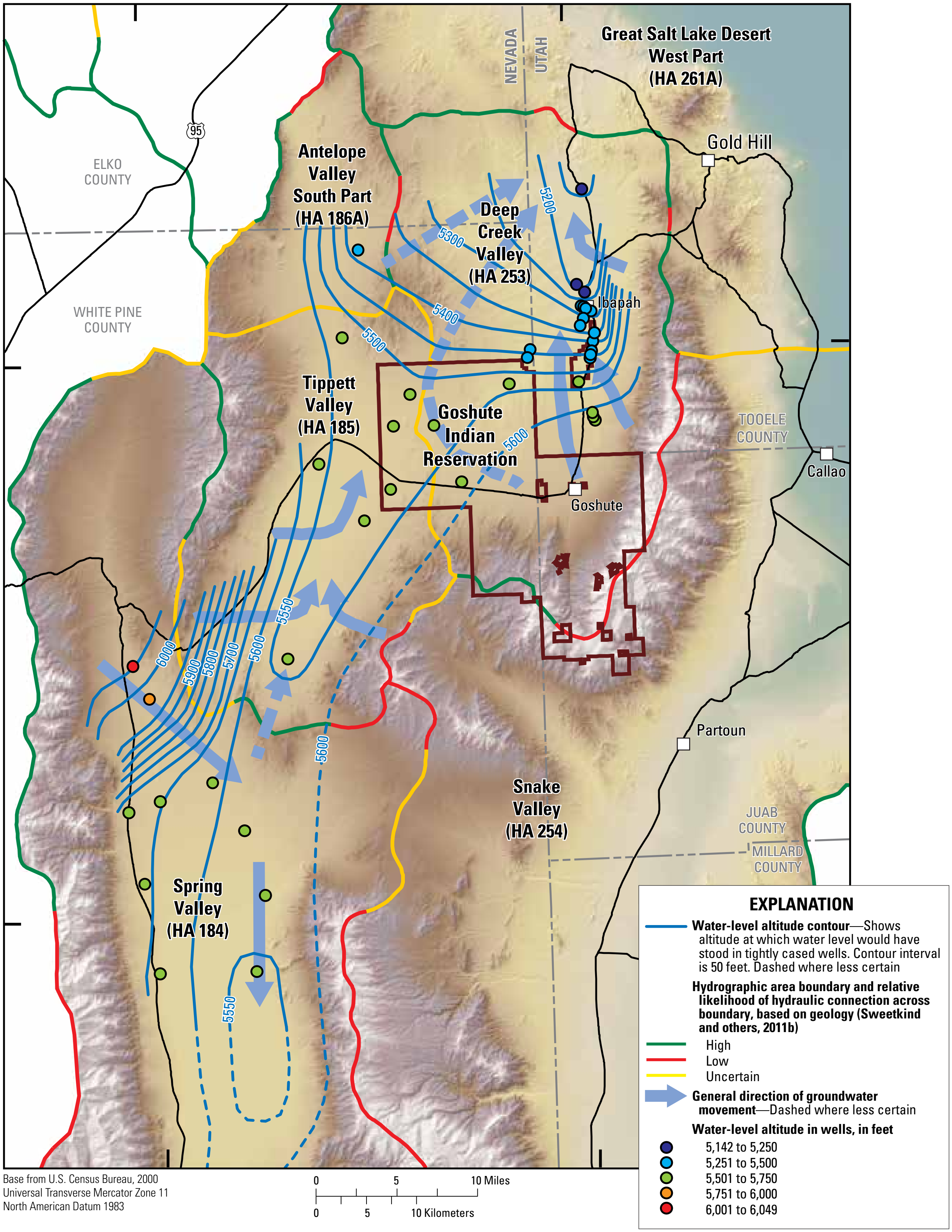

Figure 8. Regional water-level (potentiometric) surface map and general direction of groundwater movement for Deep Creek Valley and adjacent areas, Utah and Nevada.

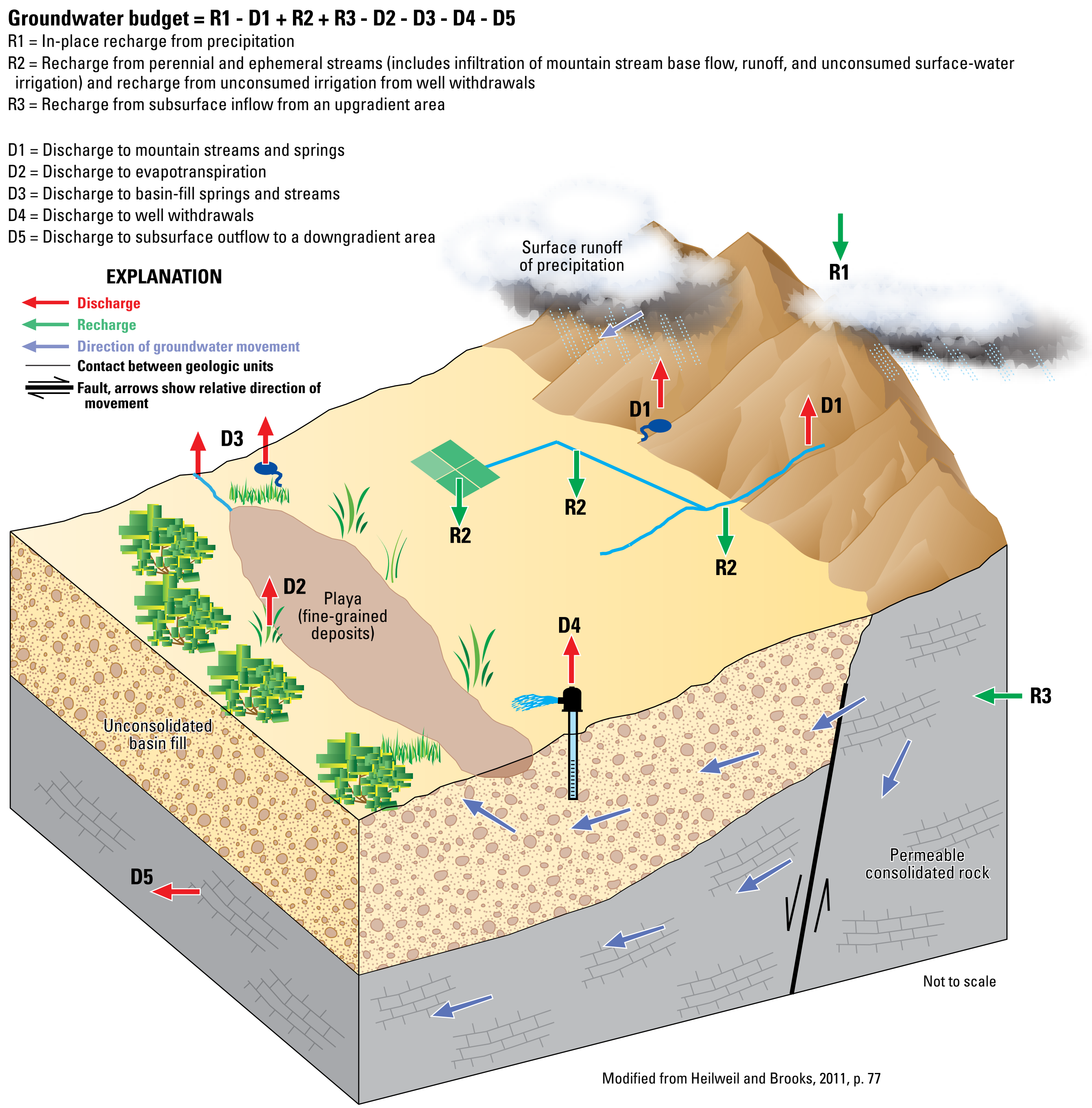

Figure 9. Conceptualization of groundwater budget components and budget calculation for Deep Creek Valley, Utah and Nevada.

The groundwater budget presented in table 4 lists the budget components for the Deep Creek Valley groundwater system by state and for the entire HA. Average annual recharge to and discharge from Deep Creek Valley are estimated to be 19,000 to 29,000 and 21,000 to 22,000 acre-ft, respectively. The groundwater budget shows an imbalance because of uncertainty in the individual budget estimates, the largest of which is subsurface inflow from Tippett Valley. Groundwater budget uncertainty is discussed in more detail in appendix 2.

Recharge

Precipitation within the study area is the primary source of groundwater recharge. The majority of precipitation comes as winter snowfall on the mountain ranges, with lesser amounts falling as rain. Infiltration of precipitation and snowmelt to the mountain block provides (1) discharge to mountain springs and base flow to mountain streams; (2) discharge to ETg, springs, streams, and wells in the adjacent basin; and (3) flow that follows deeper and longer flow paths to regional discharge locations, including large springs and areas of ETg in basins not directly adjacent to the mountain block. The majority of groundwater recharge within the study area occurs in the higher altitude mountain ranges as direct infiltration of precipitation (in-place recharge).

During the 1960s and 1970s, the USGS, in cooperation with the states of Utah and Nevada, completed a series of reconnaissance studies to evaluate the groundwater resources in these states. Generally, these studies developed groundwater budgets focused on the basin-fill (valley) portion of each HA where groundwater was being developed as a resource. Estimates of recharge from precipitation in these reports were based on a method developed by Maxey and Eakin (1949) that was calibrated to estimated groundwater discharge in the valleys, and provided estimates of "net" recharge to the unconsolidated basin-fill aquifer based on precipitation zones. These earlier methods did not consider groundwater discharge within the mountain block such as stream base flow and spring discharge, nor the subsequent recharge of part of this water as infiltration of runoff to unconsolidated basin-fill deposits.

In recent years, a new class of spatially distributed recharge estimation techniques utilizing water-balance methods have been developed that provide estimates for "total" recharge from precipitation in a watershed or HA (Leavesley and others, 1983; Hevesi and others, 2003; Flint and Flint, 2007a, b; Flint and others, 2011; Masbruch and others, 2011). Because these newer estimates include the partial loss of in-place recharge as groundwater discharge in the mountains to streams and springs, not considered in the earlier Maxey-Eakin method of estimating recharge, these newer spatially distributed recharge methods often yield higher recharge estimates. Consequently, these newer spatially distributed recharge estimates may cause over-appropriations of water rights if the consumptive losses of groundwater discharge in the mountains are not also considered.

Recharge to the Deep Creek Valley groundwater system is almost entirely by direct infiltration of snowmelt and rainfall that occurs in the mountains surrounding Deep Creek Valley. The amount and distribution of recharge controls water levels and groundwater movement throughout much of the HA. A small but significant amount of recharge also occurs from the infiltration of unconsumed irrigation water sourced from mountain streams and springs, and possibly from subsurface groundwater inflow from Tippett Valley.

Precipitation

A regional-scale water-balance method, known as the Basin Characterization Model (BCM; Flint and Flint, 2007a) developed for the GBCAAS study, was used to provide estimates of annual recharge from direct infiltration of precipitation (in-place recharge) and runoff. The BCM is a distributed-parameter water-balance accounting model used to identify areas having climatic and geologic conditions that allow for precipitation to become potential in-place recharge or runoff, and to provide estimates of each (Flint and others, 2011; Masbruch and others, 2011). The BCM in-place recharge is calculated as the volume of water per time that percolates through the soil zone past the root zone and becomes net infiltration to consolidated rock or unconsolidated deposits. Runoff is the volume of water per time that runs off the surface, and may (1) infiltrate the subsurface, (2) undergo evapotranspiration farther downslope, or (3) become streamflow that can, in turn, recharge the unconsolidated deposits from infiltration beneath the stream channels, irrigation canals, and (or) fields irrigated with surface water (Masbruch and others, 2011, p. 79). The BCM does not track or route runoff. The BCM calculations are made on a 270-m grid for each water year from 1940 to 2006. This 67-year period was selected because it encompassed the most up-to-date BCM recharge and runoff estimates available for this part of the Great Basin and because limited climatic data are available prior to the 1940s. Flint and others (2011) and Masbruch and others (2011) provide a more complete description of the BCM developed for the GBCAAS study.

In-place recharge is calculated at the location as it occurs in the BCM (fig. 10). The highest amounts of in-place recharge occur in the Deep Creek Range, with lesser amounts occurring on the western side of the HA. The average annual (1940–2006) in-place recharge calculated by the BCM in Deep Creek Valley is 16,000 acre-ft (table 4).

During 1940–2006, annual BCM in-place recharge ranged from 82 acre-ft in 1972, to 42,000 acre-ft in 1941 (fig. 11). Compared to precipitation, in-place recharge has larger annual variations. It is higher during very wet years and greatly diminished during very dry years (Gates, 2007; Masbruch and others, 2011, fig. D–3), mainly due to evapotranspiration in the recharge areas (mountains). During wet periods more water is available than is needed by vegetation, and during dry periods vegetation tries to maintain its usual rate of evapotranspiration. As a result, there is more groundwater recharge during wet periods and less during dry periods than would be estimated from a simple ratio of annual average precipitation to average annual 1940–2006 precipitation (Masbruch and others, 2011).

The BCM calculates runoff where it originates (fig. 12). The highest amounts of runoff originate in the Deep Creek Range, especially in areas where the NCCU is at or near the surface (fig. 6). Because the BCM does not route runoff, runoff that originates at higher altitudes likely becomes streamflow and recharges areas along the mountain front that contain unconsolidated basin-fill deposits (alluvial fans) or farther down the valley where surface water is used for irrigation. The average annual (1940–2006) total runoff calculated by the BCM in Deep Creek Valley is 11,000 acre-ft. The amount of runoff that infiltrates the subsurface and recharges the groundwater system is typically calculated as a percentage of the total BCM runoff. For this study, it was assumed that 10 percent of the total runoff calculated by the BCM infiltrates the subsurface (Masbruch and others, 2011, p. 86); this includes recharge from the infiltration of unconsumed surface-water irrigation (table 4). Thus, the estimated average annual infiltration from runoff is 1,100 acre-ft.

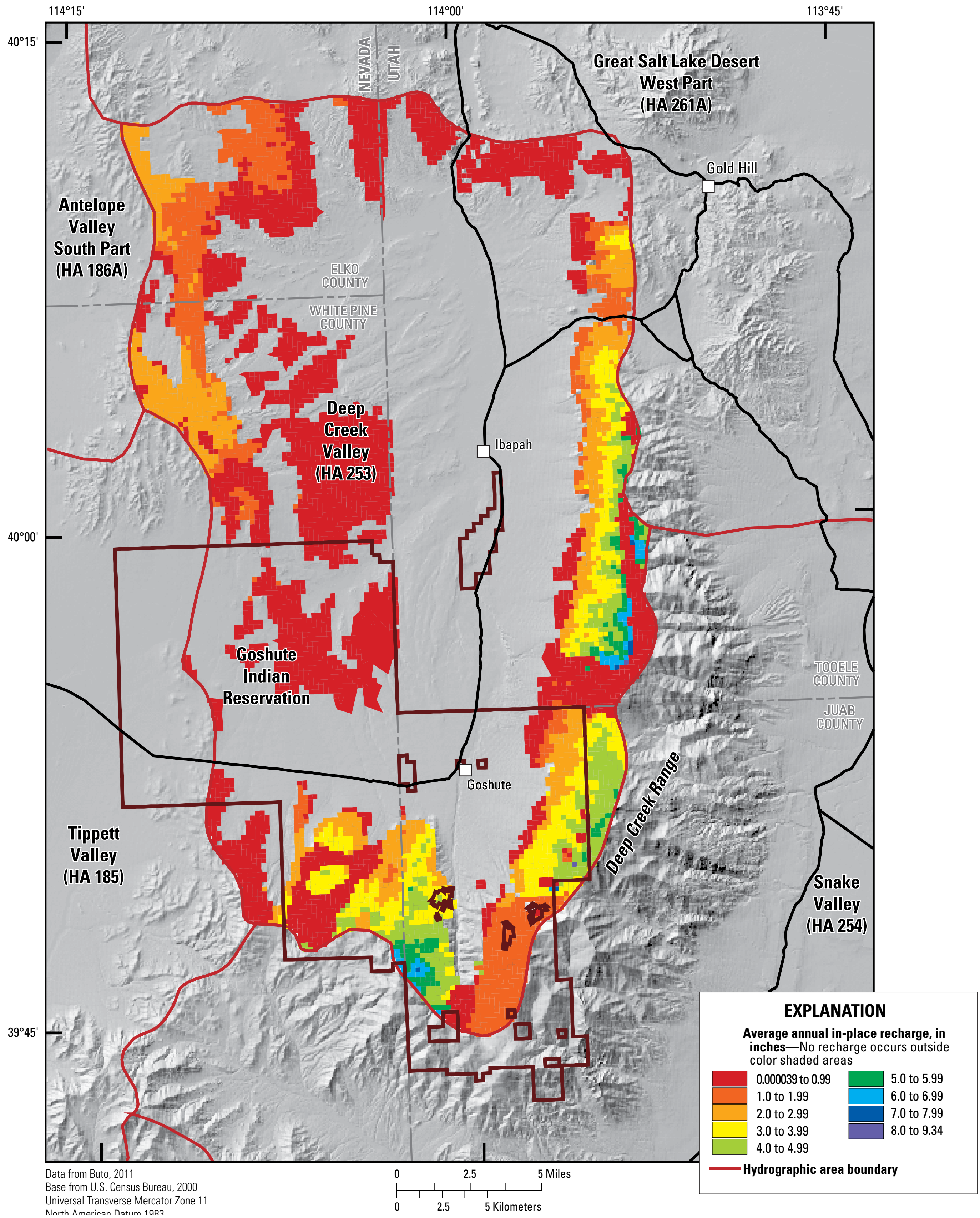

Figure 10. Distribution of average annual (1940–2006) in-place recharge estimated by the Basin Characterization Model for Deep Creek Valley, Utah and Nevada.

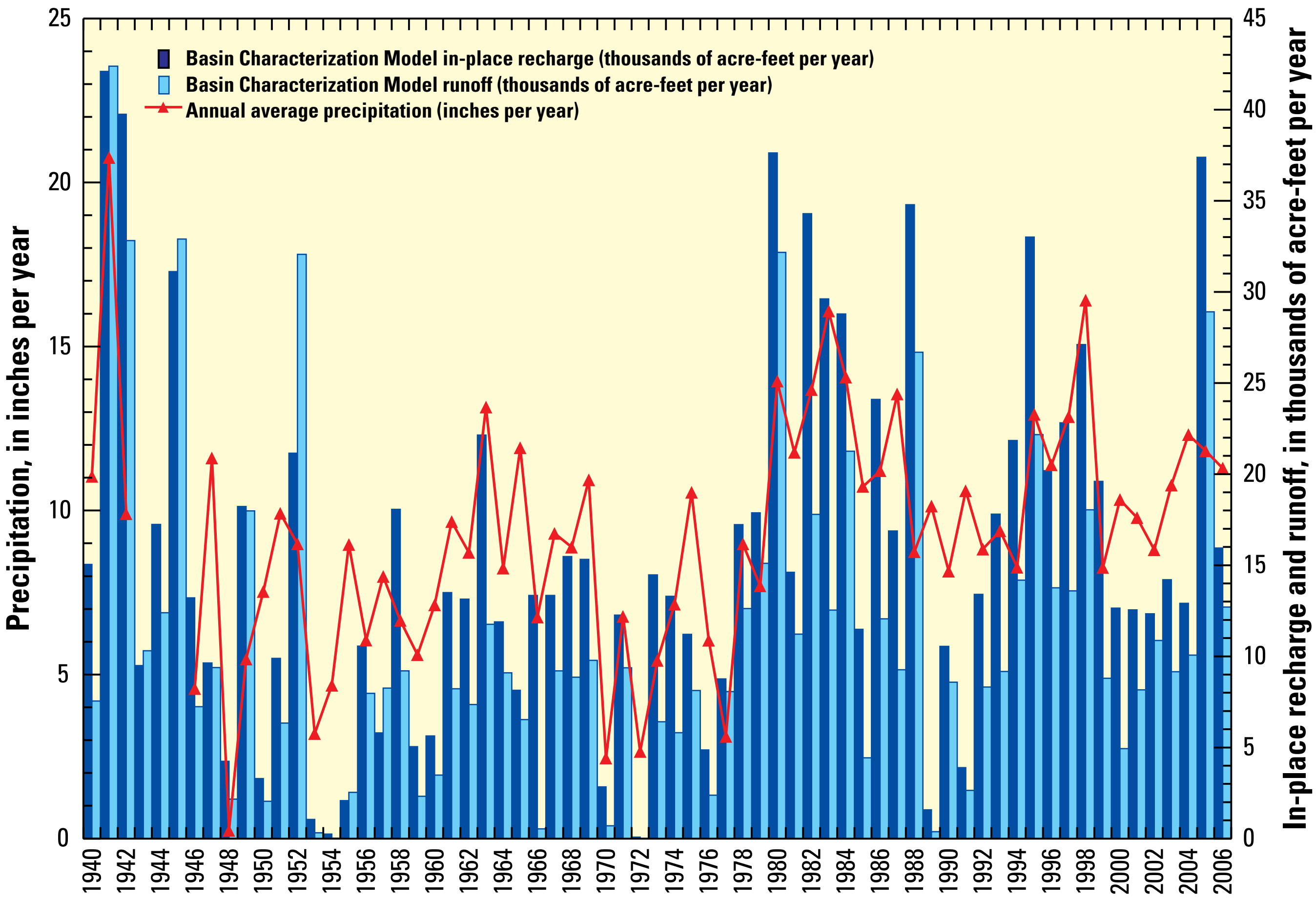

Figure 11. Annual average precipitation (at Ibapah, Utah) and Basin Characterization Model estimates of in-place recharge and runoff (1940–2006) for Deep Creek Valley, Utah and Nevada.

Streamflow at the mountain front also includes base flow. This water originates as in-place recharge in the mountains and then discharges to mountain streams. Similar to runoff, a portion of this base flow subsequently recharges the basin-fill deposits as infiltration beneath stream channels, irrigation canals, or fields irrigated with surface water (Masbruch and others, 2011, p. 79). It was also assumed that 10 percent of the mountain stream base flow infiltrates the subsurface and becomes recharge (Masbruch and others, 2011, p. 92) (table 4); this recharge is also likely to occur in areas along the mountain front that contain unconsolidated basin-fill deposits. The other 90 percent of runoff and mountain stream base flow is assumed to be consumptively lost to evapotranspiration before it can infiltrate into the aquifer (Hevesi and others, 2003; San Juan and others, 2010; Masbruch and others, 2011). Estimates of recharge from mountain stream base flow for the Deep Creek Valley HA were calculated by using the 2012 base-flow values for Spring, Fifteenmile, and Steves Creeks, and the average October 1964–1966 streamflow for Sams Creek (table 2). Total base flow is about 3,900 acre-ft/yr. Recharge from mountain stream base flow in Deep Creek Valley, therefore, is about 390 acre-ft/yr.

Unconsumed Irrigation from Well Withdrawals

Most well withdrawals in the study area are used for irrigation and it is assumed that part of these withdrawals recharges the aquifer system as infiltration of unconsumed irrigation water applied to fields. Irrigation return flow studies in the Amargosa Desert, CA (Stonestrom and others, 2003) and the Milford Area, UT (Susong, 1995) show that recharge from irrigation on sprinkler-irrigated fields ranges from 8 to 16 percent of the applied irrigation, and recharge on flood-irrigated fields can be as high as 50 percent of the applied irrigation (Susong, 1995).

In Deep Creek Valley, well withdrawals for irrigation have only been estimated for 1966 (Hood and Waddell, 1969), and were 600 acre-ft. It does not appear that there has been a substantial amount of development in Deep Creek Valley since the 1960s, so it was assumed that well withdrawals are not likely to be much larger than 600 acre-ft/yr. Additionally, most of the groundwater withdrawn for irrigation is applied in areas of groundwater discharge and is highly unlikely to recharge the groundwater system. Estimates of recharge from unconsumed irrigation water from well withdrawals in Deep Creek Valley, therefore, are assumed to be negligible.

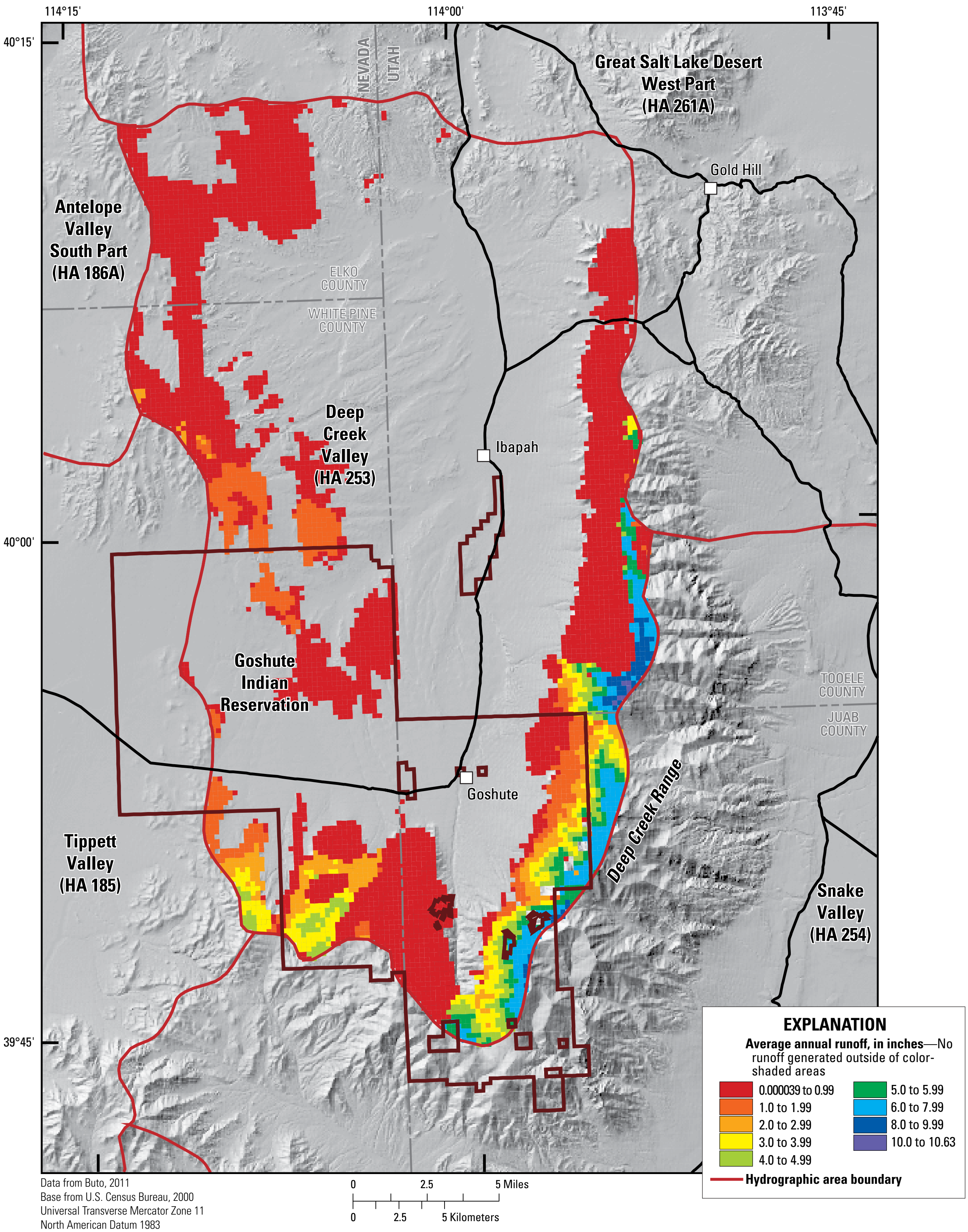

Figure 12. Distribution of average annual (1940–2006) runoff estimated by the Basin Characterization Model for Deep Creek Valley, Utah and Nevada.

Subsurface Inflow from Tippett Valley

The water-level map developed for the study area indicates that groundwater may enter and leave the study area via subsurface inflow and outflow along segments of the Deep Creek Valley HA boundary. Cross sections from Spring Valley to Deep Creek Valley show possible connections between permeable units (LCAU, LBFAU, UBFAU, and possibly VU), allowing groundwater flow from northern Spring Valley through Tippett Valley to Deep Creek Valley (fig. 7, cross sections *A-A′* and *B-B′*). Previous studies have estimated subsurface flow by a variety of methods, with little to no indication of the uncertainties associated with these estimates which can vary widely due to differences in the methods used to calculate them. Previous estimates of subsurface inflow into Deep Creek Valley from Tippett Valley ranged from 2,000 to 12,000 acre-ft/yr (Scott and others, 1971; Welch and others, 2007). More knowledge of groundwater discharge and groundwater levels in this area may reduce the uncertainty of these estimates.

Discharge

Discharge from the groundwater system in Deep Creek Valley occurs by ETg, as discharge to springs, as discharge to mountain streams (base flow), as well withdrawals, and as subsurface outflow to Great Salt Lake Desert. All ETg is assumed to occur in the valley. Much of the groundwater that discharges from significant mountain springs or as base flow to mountain streams is captured and diverted to be used for public supply or irrigation. Groundwater that discharges from springs in the valley is eventually consumed by ETg and, therefore, is included in the estimate for that component of the groundwater budget. Groundwater withdrawn by wells is used for irrigation, public and domestic supply, and stock watering.

Evapotranspiration

Discharge of groundwater by ETg is the combination of groundwater consumed (transpired) by plants with roots that extend to the shallow water table and direct evaporation from areas of open water or soils that are wetted by shallow groundwater. The total volume of water (both groundwater and surface water) discharged by evapotranspiration (total ET) can be calculated as the product of the rate at which water is transferred from the land to the atmosphere (ET rate, in feet per year) and the acreage of the vegetation, open water, and soils that transfer this water. ETg, the fraction of total ET made up of groundwater, is calculated by subtracting precipitation and delivered irrigation water from the total ET. Average annual ETg in Deep Creek Valley is estimated to be about 14,000 acre-ft (table 4).

As a first step in calculating ETg, total ET was calculated for the area where groundwater is shallow enough to be transmitted by ETg. This area was subdivided according to similar ET-related characteristics such as vegetation type, density, and land cover, and representative ET rates were applied to each zone. Average annual precipitation (1981–2010) was determined from PRISM model data (Daly and others, 2008) that was resampled to 10-m resolution. Precipitation was subtracted from the total ET to arrive at an annual estimate for ETg.

The outer boundary of the ET area delineated in this study approximates the extent of the phreatophytic vegetation (including areas of moist bare soil) where groundwater may be transferred to the atmosphere by ET. Results of ET studies in areas of Nevada and California (Nichols, 2000; Berger and others, 2001) indicate that most ETg occurs when the water table is within 15 to 20 ft of land surface and that phreatophytes commonly grow in areas where the depth to water is within about 40 ft of land surface (DeMeo and others, 2008). The boundary used to calculate ETg in this study was modified from one delineated for large-scale ET areas in the Great Basin (Medina, 2005) to coincide with the area where groundwater is shallow according to the water-level surface maps developed during this study. After refinement using 1-m resolution National Agricultural Imagery Program (NAIP) Digital Orthorectified Aerial Images from 2006 (U.S. Department of Agriculture, 2006) and field verification, the final ETg area boundary approximately corresponded to where groundwater is within about 25 ft of land surface.

Prior to assigning ET rates, the ET area boundary was subdivided into smaller zones (ET units) on the basis of vegetation and land-cover characteristics determined by using existing land-cover and land-use data. An ET unit is an area of similar vegetation or land-cover characteristics that is assigned one ET rate. Southwest Regional Gap Analysis Program (SWReGAP) land-cover data (U.S. Environmental Protection Agency, 2007) and Utah water-related land-use survey data from 1993 (Utah Department of Natural Resources, 2004) were used to identify subareas of common vegetation or land cover. These data provided subarea boundaries that could be verified or slightly modified using the NAIP imagery and grouped into ET units. The Utah water-related land-use survey boundaries, considered more accurate than the SWReGAP data because they are field mapped and updated approximately once per decade, were used preferentially in developed areas. Evapotranspiration rates reported in recent literature (Nichols, 2000; Berger and others, 2001; Reiner and others, 2002; Cooper and others, 2006; Moreo and others, 2007; Welch and others, 2007) for vegetation types, land cover, and climate similar to those in the study area were assigned to seven ET units in Deep Creek Valley (fig. 13). Much of the area where ETg is occurring in Deep Creek Valley is undeveloped and covered by the ET units designated as moderately dense to dense desert shrubland. The ET unit that occupies most of the developed area is pasture/range. Table 5 contains the values used in the ET calculations.

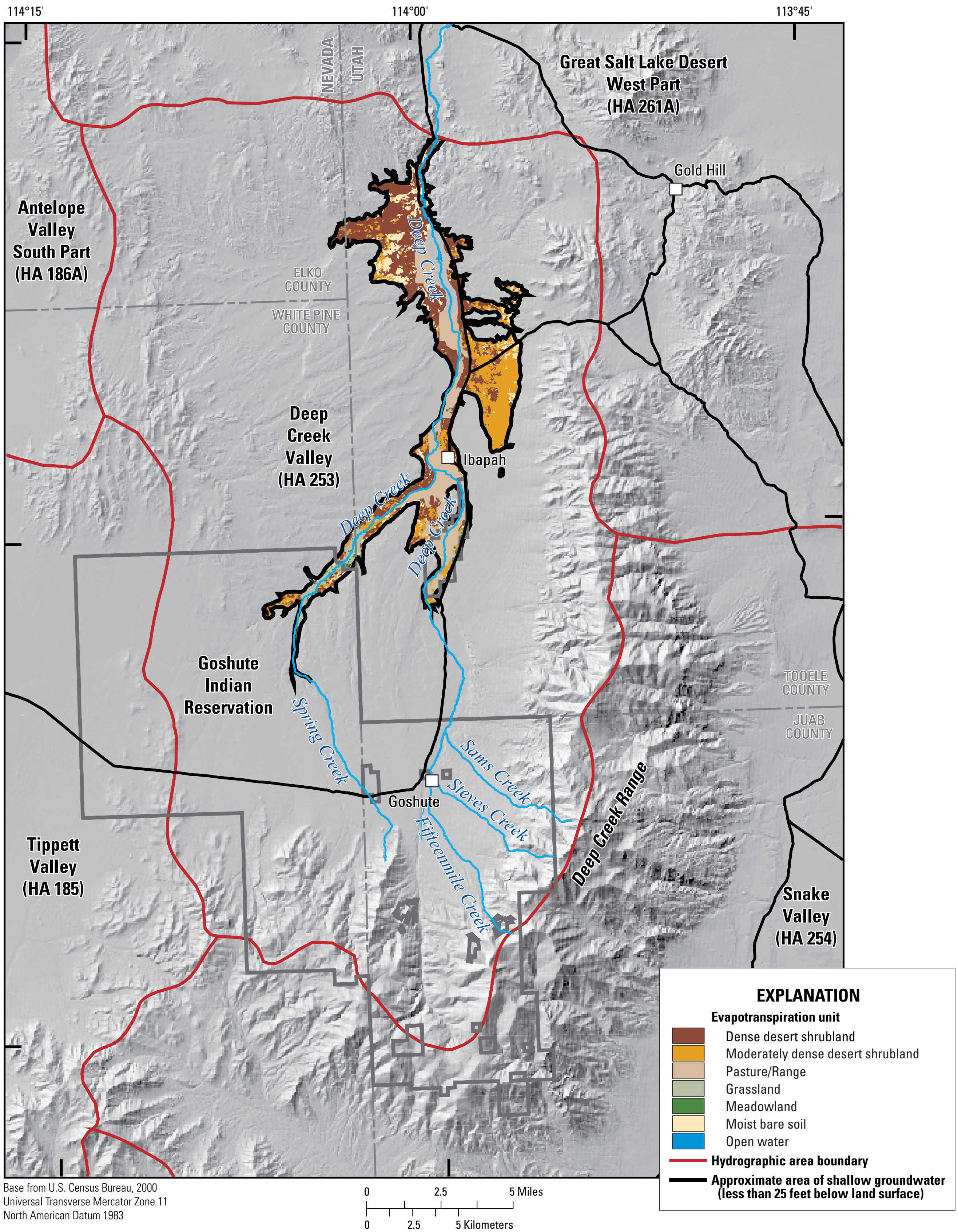

Figure 13. Location and classification of evapotranspiration units used to calculate average annual groundwater evapotranspiration in Deep Creek Valley, Utah and Nevada.

Agricultural fields exist within the ET area. Water used for irrigation in these areas is either surface water diverted from nearby mountain streams or groundwater withdrawn from wells. The contributions of both of these sources are accounted for elsewhere in the groundwater budget. Ten percent of surface water applied for irrigation becomes recharge and is added to the groundwater budget; the remainder is consumed as ET by crops. Groundwater from wells that is applied to crops is subtracted from the budget as the "well withdrawal" component of discharge. For these reasons, the ET estimated for these fields was omitted from the total calculated ET. Agricultural fields classified as "sub-irrigated" were not omitted from the ETg calculation.

Mountain Springs and Base Flow to Mountain Streams

Discharge from significant mountain springs enters stream channels in their respective drainages and is included (as part of the base-flow component) in the gaged or estimated annual streamflow for the individual streams listed in table 2. Estimates were made as described in the "Streamflow" subsection of this report. Estimates of discharge to mountain stream base flow for the Deep Creek Valley HA were calculated using the base-flow measured in 2012 for Spring, Fifteenmile, and Steves Creeks, and average October 1964–1966 streamflow for Sams Creek (table 2). Groundwater that discharges from mountain springs or as base flow to mountain streams is estimated to be about 3,900 acre-ft/yr (table 4).

Well Withdrawals

In Deep Creek Valley, well withdrawals have only been estimated for 1966 (Hood and Waddell, 1969), and were 600 acre-ft (table 4). It does not appear that there has been a substantial amount of development in Deep Creek Valley since the 1960s, so it was assumed that well withdrawals are not likely to be much larger than 600 acre-ft/yr.

Subsurface Outflow to Great Salt Lake Desert

The water-level surface map (fig. 8) indicates that groundwater in Deep Creek Valley discharges northward toward the Great Salt Lake Desert. Previous estimates of subsurface outflow from Deep Creek Valley to Great Salt Lake Desert ranged from 2,700 to 3,000 acre-ft/yr (Scott and others, 1971; Harrill and others, 1988).

Water-Level Fluctuations

Water levels in wells fluctuate in response to imbalances between groundwater recharge and discharge. Water levels rise when recharge exceeds discharge for a period of time and decline when the opposite occurs. Variations in recharge and discharge are driven by natural and anthropogenic (human-induced) processes. Examples of natural processes are recharge from the infiltration of precipitation and evapotranspiration of groundwater in a marsh or wetland. The infiltration of unconsumed water applied to irrigate crops or groundwater withdrawn by wells are examples of anthropogenic processes. Long-term water-level changes were examined for selected wells within or adjacent to the Deep Creek Valley HA (fig. 14). Long-term water-level fluctuations are presented for six wells along the drainage axis of Deep Creek Valley, with depths ranging from 58 to 506 ft, where repeated measurements have been made for various periods of time to illustrate the groundwater system's response to interannual variations in recharge and discharge (fig. 15). Long-term water-level data for these six wells were filtered to include only spring season water-level measurements for each year. Water levels in these wells fluctuate to varying degrees in response to climate-driven variations in recharge. Water levels in these wells do not show long-term water-level declines, indicating that past and current groundwater withdrawals have not adversely affected aquifer conditions to date in Deep Creek Valley. Water levels are

Table 5. Evapotranspiration unit rates and areas used to calculate average annual evapotranspiration of groundwater in Deep Creek Valley, Utah and Nevada.

[Abbreviations: ET, evapotranspiration; ft/yr, feet per year; acre-ft/yr, acre-feet per year]

ET unit	ET rate, area-weighted average (ft/yr)	Range of ET rate (ft/yr)	Acres			Total ET (acre-ft/yr)		
			Utah	Nevada	Total	Utah	Nevada	Total
Dense desert shrubland[1]	1.24	1.0–1.8	6,969	586	7,555	8,641	727	9,368
Moderately dense desert shrubland[1]	1.07	0.7–1.5	4,612	387	4,999	4,935	414	5,349
Pasture/Range[2]	1.97	0.8–3.1	4,082	0	4,082	8,041	0	8,041
Grassland[1]	2.14	1.6–2.7	295	617	912	631	1,320	1,951
Meadowland[1]	2.59	2.2–3.3	87	75	162	225	194	419
Moist bare soil[1]	2.00	1.7–2.3	2,534	159	2,693	5,068	318	5,386
Open water[1]	5.10	4.6–5.6	90	0	90	459	0	459
			Sum of total ET			28,000	2,973	30,973
			-Precipitation over ET area			15,512	1,518	17,030
			= ET from groundwater (rounded)			12,500	1,500	14,000

[1]ET rates and ranges for these ET units are summarized in Welch and others (2007), p. 56.

[2]ET rates and ranges for this ET unit from Utah State University (1994), table 25.

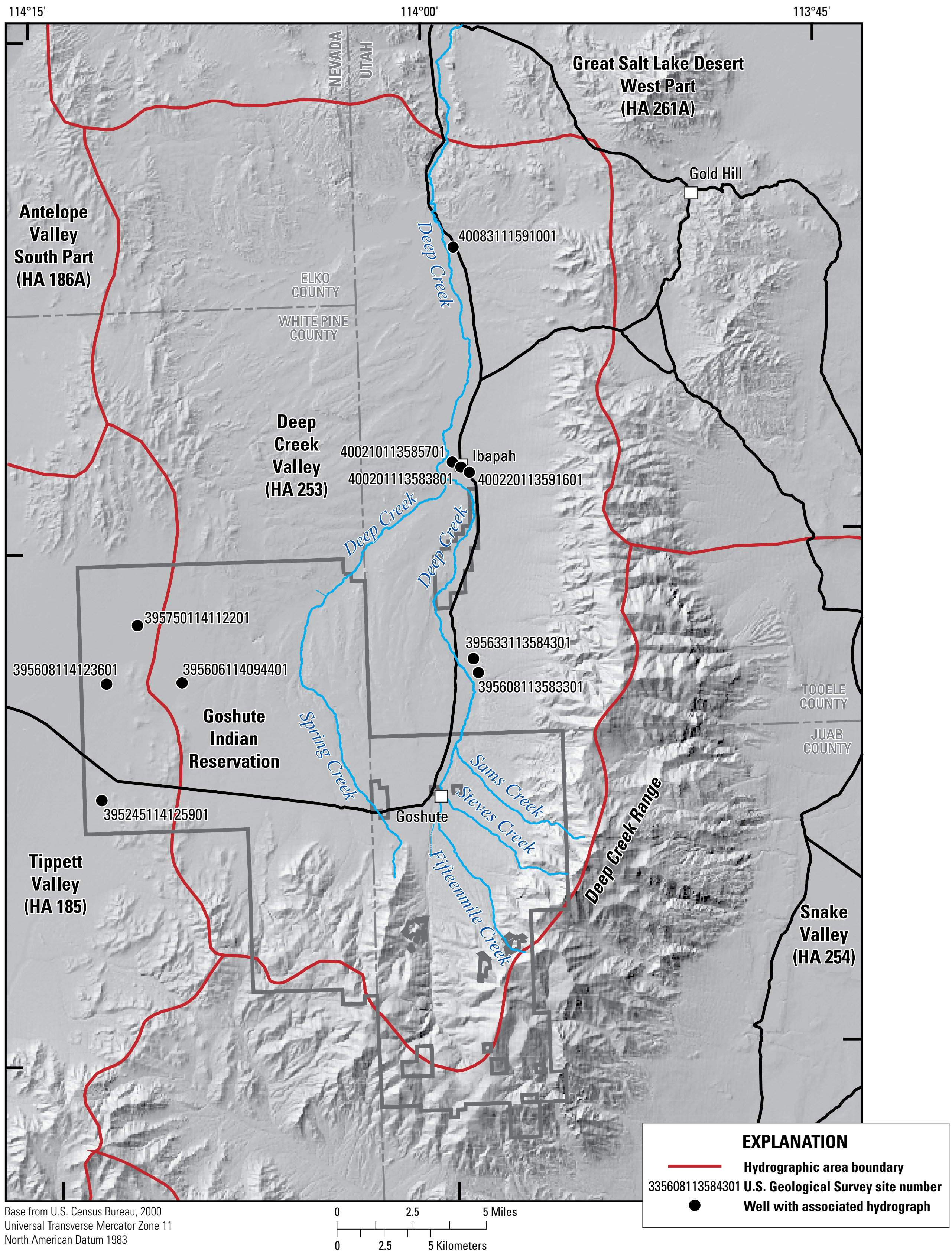

Figure 14. Location of selected wells with long-term and discrete water-level data, Deep Creek Valley and adjacent areas, Utah and Nevada.

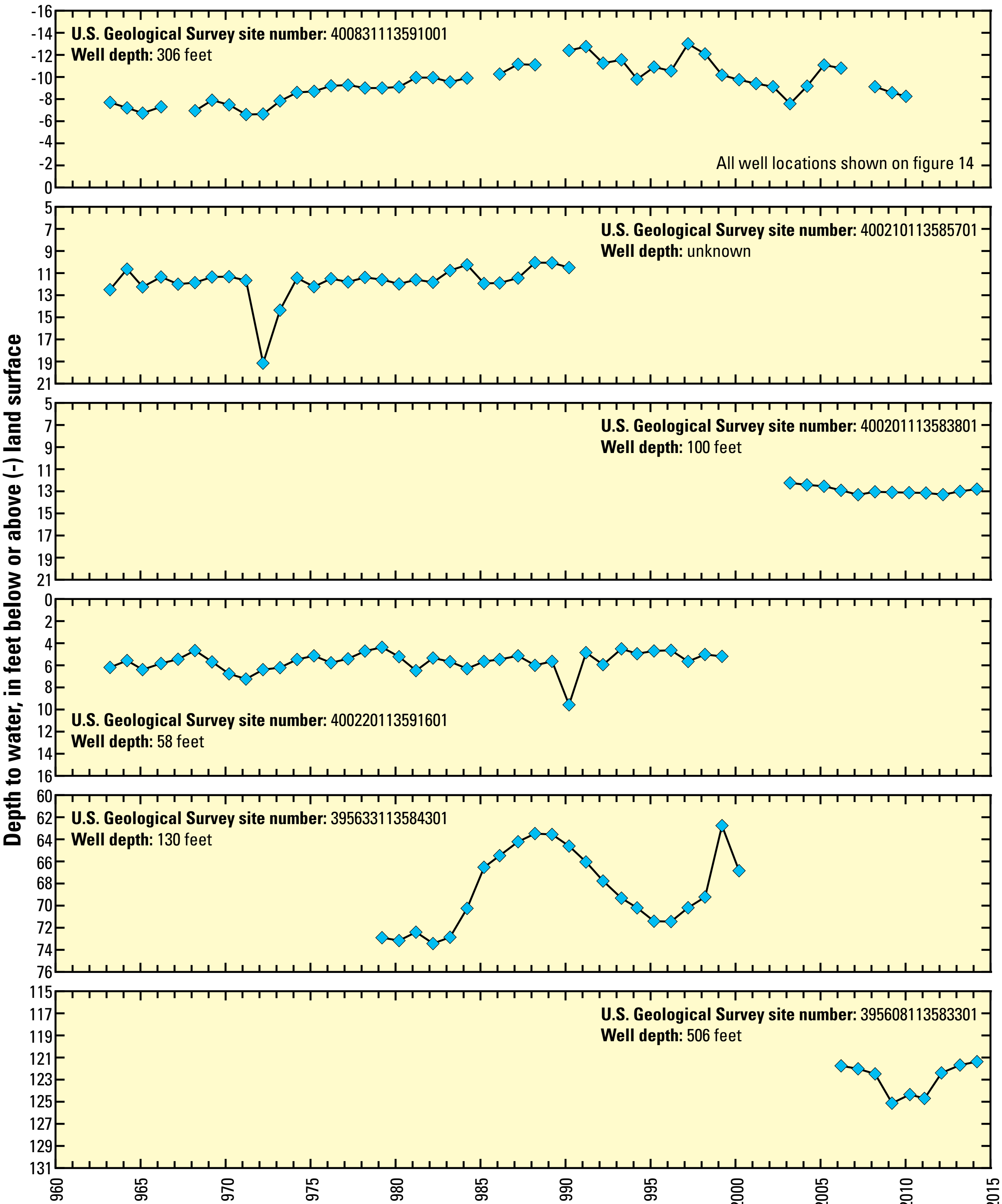

Figure 15. Long-term water-level fluctuations in selected wells along the drainage axis of Deep Creek Valley, Utah and Nevada.

also presented for four stock wells, with only a few discrete water-level measurements each, along the boundary between Deep Creek and Tippett Valleys (fig. 16). Water levels in three of these wells were between about 8 and 12 ft higher in 1984 than in 1969, possibly due to a significant recharge event during the three consecutive wet years of 1982–1984 (fig. 11). Since that time, records generally indicate little net water-level change or slight water-level rises in these wells. The fourth well showed a water-level decline between 1969 and 1984 with subsequent recovery by 2012. All water-level data are available through the USGS NWIS database (*http://waterdata.usgs.gov/nwis)*.

Groundwater Geochemistry

Water samples were collected from 10 sites in the Deep Creek Valley study area that included domestic, stock, and unused flowing wells and four perennial springs (tables 6 and A1–2). The water samples were analyzed for major ions, nutrients, and selected trace metals to characterize general chemistry and patterns of water quality. Water samples also were analyzed for a suite of environmental tracers that included the stable isotopes of oxygen, hydrogen, and carbon, dissolved noble gases, and radioactive isotopes of carbon (^{14}C) and hydrogen (tritium, ^{3}H). These environmental tracers were used to investigate sources of recharge, groundwater flow paths, ages, and travel times, and to support the development of a conceptual model of the basin-wide groundwater system. All groundwater geochemical data are available through the USGS NWIS database (*http://waterdata.usgs.gov/nwis*).

Sample Collection and Analysis

Field parameters and geochemical samples were collected from six wells and four springs (fig. 17; tables 6 and A1–2). Water samples were collected from most wells using either a portable submersible pump or a well's dedicated pump. Samples were collected from springs and flowing wells under natural free-flowing conditions. Wells requiring pumping were purged of a minimum of three casing volumes of water prior to sample collection, and water was collected from an outlet as close to the wellhead as possible.

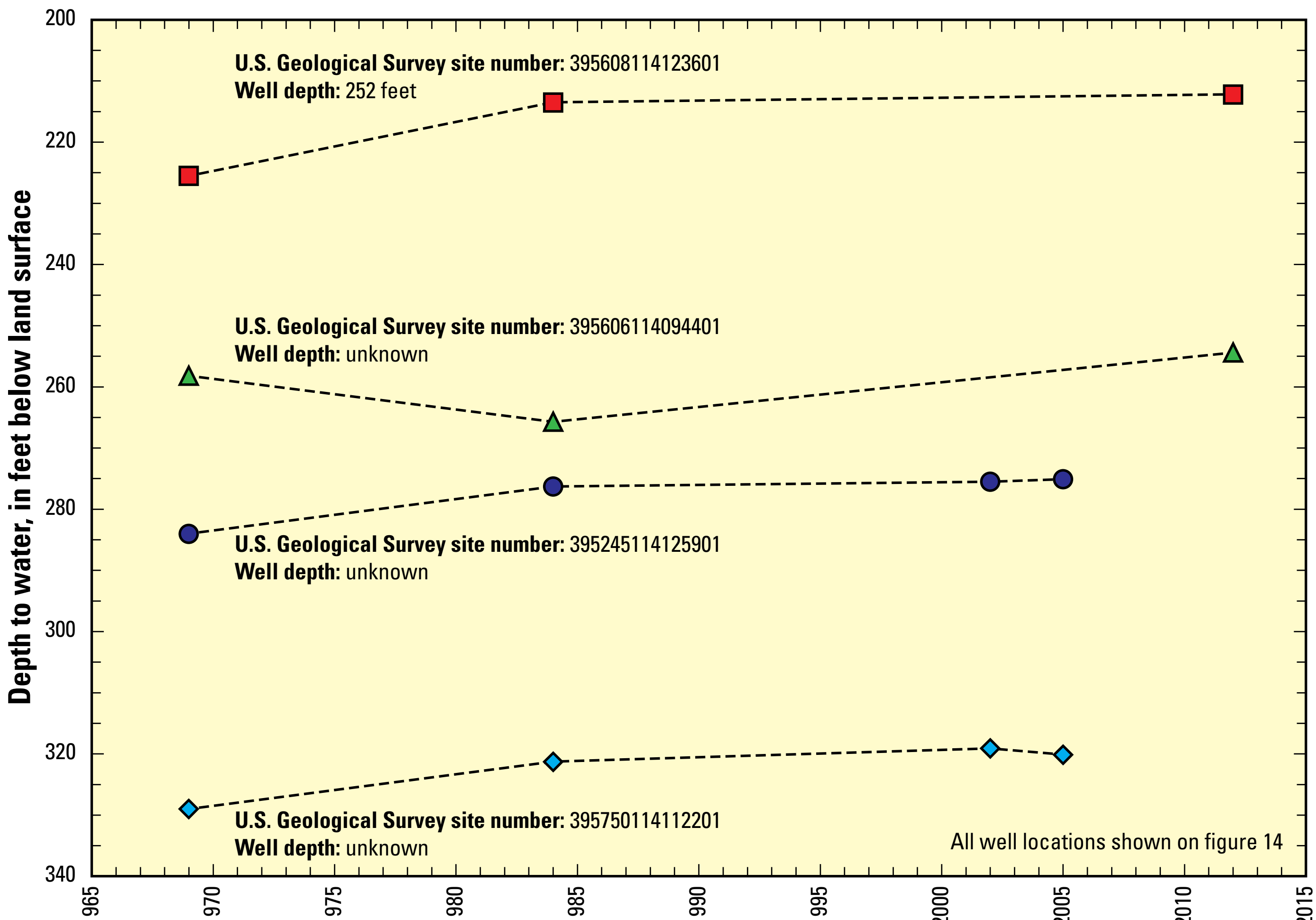

Figure 16. Discrete water-level data for selected wells in Deep Creek Valley and adjacent areas, Utah and Nevada.

Table 6. Measured field parameters and dissolved concentrations of major ions, nutrients, and selected metals for groundwater sampled during May 2012 in Deep Creek Valley and adjacent areas, Utah and Nevada.

[Site ID: see figure 17 for locations and table A1-2 for additional information. Value shown in red exceeds the Environmental Protection Agency maximum contaminant level or secondary standard. Abbreviations: USGS, U.S. Geological Survey; °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter; µg/L, micrograms per liter; <, less than; —, no data]

Site ID	USGS site number	Water temperature, °C	Specific conductance, µS/cm	pH, standard units	Dissolved oxygen, mg/L	Dissolved solids, mg/L	Alkalinity, mg/L as $CaCO_3$	Bicarbonate, mg/L as HCO_3	Bromide, mg/L as Br	Calcium, mg/L as Ca	Chloride, mg/L as Cl	Fluoride, mg/L as F	Iron, µg/L as Fe
1	400543114145101	23.1	614	7.5	5.7	370	115	141	0.283	39.2	64.3	0.29	<3.2
2	400216113591701	11.6	258	7.5	4.3	193	97	141	0.043	19.5	11.6	0.36	13.1
3	394554114003501	6.2	216	6.9	8.1	135	105	128	0.021	35.5	3.98	0.09	34.2
4	395112114014501	11.1	313	7.0	8.3	185	156	190	0.029	47.5	5.10	0.07	11.8
5	400024113582701	12.7	178	7.2	7.8	126	76	92	0.029	19.9	7.10	0.19	10.6
6	394149114302201	10.1	419	6.9	2.0	267	179	218	0.108	54.7	17.3	0.13	38.4
7	395815114043401	13.2	693	6.9	7.6	475	165	201	0.060	98.7	18.2	0.13	4.9
8	395237114222501	21.1	503	7.0	3.4	285	238	290	0.052	55.1	6.74	0.07	3.6
9	400831113591001	12.0	542	7.2	5.9	319	191	232	0.101	49.2	45.1	0.12	<3.2
10	394422114205201	14.5	382	7.7	5.6	255	126	154	0.098	29.3	17.9	0.53	10.4
[1]11	395935113584501	20.2	205	8.0	5.4	—	103	106	0.03	28.3	9.00	0.20	330
[1]12	—	16.6	248	7.9	6.4	—	100	117	0.03	30.6	10.4	0.22	<50
Environmental Protection Agency maximum contaminant level (MCL)													
Environmental Protection Agency secondary standard						500					250	2	300

[1]Data for Sample IDs 11 (Goshute Lower Community Well) and 12 (Goshute #2 Well) from Hershey and others, 2007, *http://www.dri.edu/images/stories/research/projects/BARCAS/barcas.pdf*.

Table 6. Measured field parameters and dissolved concentrations of major ions, nutrients, and selected metals for groundwater sampled during May 2012 in Deep Creek Valley and adjacent areas, Utah and Nevada.—Continued

[Site ID: see figure 17 for locations and table A1-2 for additional information. Value shown in red exceeds the Environmental Protection Agency maximum contaminant level or secondary standard. Abbreviations: USGS, U.S. Geological Survey; °C, degrees Celsius; µS/cm, microsiemens per centimeter at 25 degrees Celsius; mg/L, milligrams per liter; µg/L, micrograms per liter; <, less than; —, no data]

Site ID	USGS site number	Magnesium, mg/L as Mg	Manganese, µg/L as Mn	Potassium, mg/L as K	Silica, mg/L as SiO_2	Sodium, mg/L as Na	Sulfate, mg/L as SO_4	Nitrate plus nitrite, mg/L as N	Ortho-phosphate, mg/L as P	Arsenic, µg/L as As	Molybde-num, µg/L as Mo	Selenium, µg/L as Se	Uranium, µg/L as U
1	400543114145101	9.27	1.64	4.04	27.9	74.2	73.0	1.96	0.018	11.5	5.38	1.3	3.06
2	400216113591701	11.0	4.20	2.89	47.7	18.1	10.5	0.415	0.020	4.7	1.11	0.23	6.47
3	394554114003501	5.35	0.56	0.51	13.8	5.93	5.62	0.186	0.010	0.23	0.080	0.07	0.244
4	395112114014501	11.7	0.19	0.64	12.3	6.60	5.60	0.447	0.014	0.67	0.131	0.12	0.391
5	400024113582701	4.26	<0.16	0.95	29.1	12.0	5.59	0.366	0.028	0.70	0.822	0.18	6.33
6	394149114302201	13.6	20.9	2.81	28.1	20.8	21.6	0.072	0.011	3.8	2.27	0.29	2.34
7	395815114043401	26.5	0.30	2.44	28.2	14.5	187.0	0.169	0.013	2.1	0.370	0.37	1.02
8	395237114222501	30.7	<0.16	0.99	12.5	7.90	25.2	0.624	0.008	1.4	1.27	0.54	3.33
9	400831113591001	26.7	<0.16	1.63	23.1	29.3	26.5	0.805	0.013	3.0	0.812	1.1	2.89
10	394422114205201	17.7	4.65	3.32	39.6	26.4	41.4	0.562	0.019	8.8	3.60	0.98	1.08
[1]11	395935113584501	7.41	2.20	1.41	29.9	16.2	9.1	—	—	1	<2	<1	9.93
[1]12	—	9.93	0.4	1.73	32.4	18.5	10.6	—	—	<1	<2	<1	10.9
Environmental Protection Agency maximum contaminant level (MCL)										10		50	30
Environmental Protection Agency secondary standard			50				250	10					

[1]Data for Sample IDs 11 (Goshute Lower Community Well) and 12 (Goshute #2 Well) from Hershey and others, 2007, *http://www.dri.edu/images/stories/research/projects/BARCAS/barcas.pdf*.

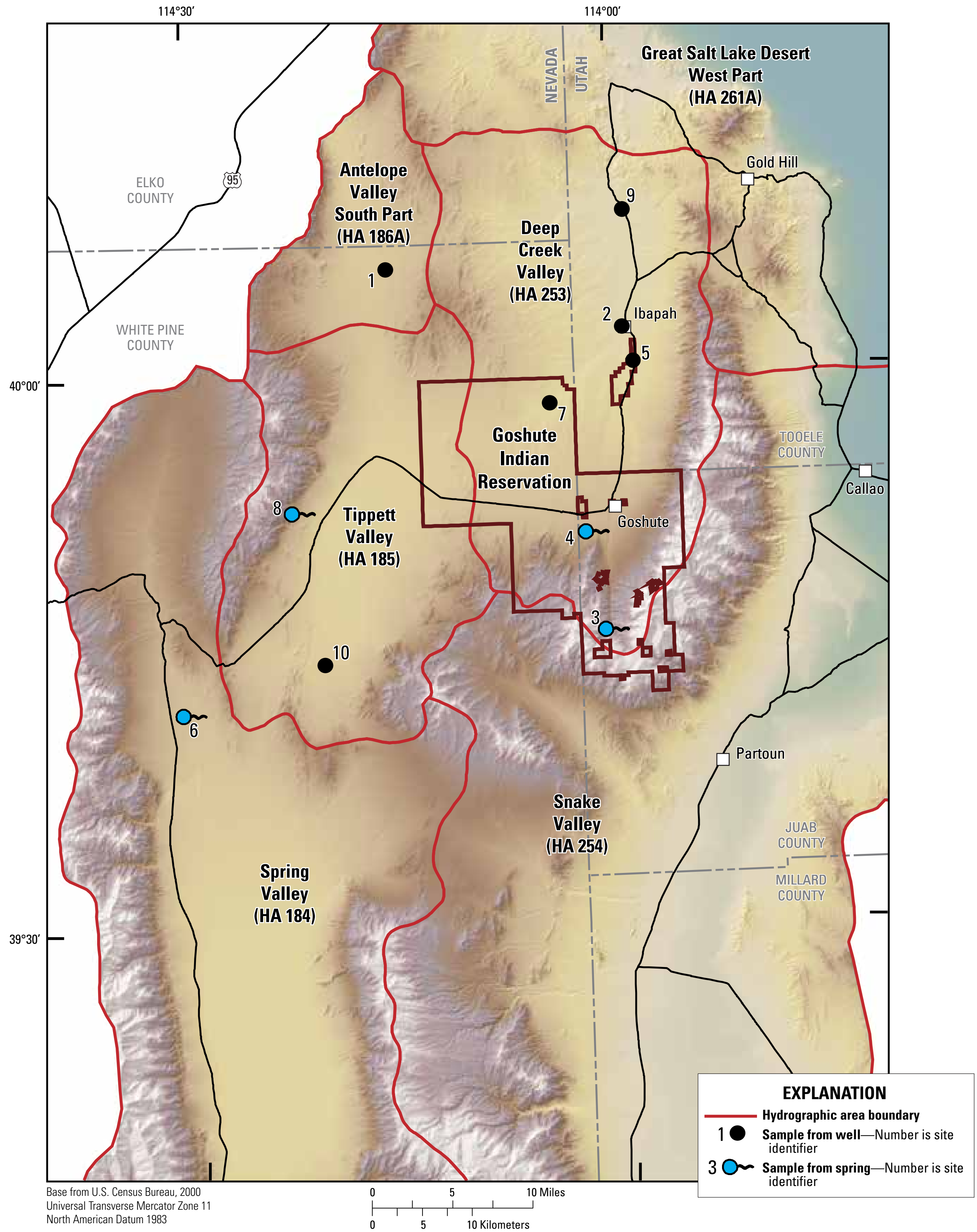

Figure 17. Location of 10 wells and springs sampled during this study in Deep Creek Valley and adjacent areas, Utah and Nevada.

Field parameters measured during water-sample collection included temperature, specific conductance, pH, dissolved oxygen, and total dissolved-gas pressure. These parameters were measured using a calibrated multimeter probe following USGS protocols (Wilde and Radtke, 1998). Samples for dissolved major ions and nutrients were filtered with a 0.45-micrometer (µm) filter and the cation subsample was preserved with nitric acid. Dissolved major-ion and nutrient analyses were done by the USGS National Water Quality Laboratory in Denver, Colorado.

Unfiltered samples for stable isotopes of oxygen and hydrogen were collected in 60-milliliter (mL) glass containers, sealed with polyseal caps to leave no air space, and analyzed by the USGS Stable Isotope Laboratory in Reston, Virginia. The 2-standard deviation uncertainty of oxygen and hydrogen isotopic measurements is 0.2 and 2 permil, respectively. Unfiltered samples for tritium (^3H) were collected in 500-mL or 1-liter (L) polyethylene bottles, capped with no air space, and analyzed by the University of Utah Dissolved Gas Service Center in Salt Lake City. The detection limit of ^3H is reported to be 0.1 tritium units (TU), and the analytical precision was generally 0.1 TU but as high as 0.8 TU. Samples for carbon-14 (^{14}C) and stable isotopes of carbon ($\delta^{13}C$) were filtered (0.45 µm) and collected in 500-mL or 1-L glass bottles. The bottles were filled from the bottom and allowed to overflow for several volumes in order to rinse the bottles while minimizing contact with the air, sealed with polyseal caps, and analyzed by the National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at the Woods Hole Oceanographic Institution in Woods Hole, Massachusetts. Analytical error for ^{14}C was less than 0.5 percent modern carbon (pMC) and for $\delta^{13}C$ was 0.3 permil or better.

Dissolved noble-gas samples were collected either as water samples sealed in copper tubes as described by Stute and Schlosser (2000) or as gas samples collected with diffusion samplers similar to those described by Sanford and others (1996) and Gardner and Solomon (2009). The copper tube method consists of attaching a 76-centimeter (cm) long section of 1-cm diameter copper tubing to a sampling port at the wellhead, allowing the tube to flush with well water until all air bubbles have been evacuated, then sealing both ends with clamps. The diffusion sampler method was used at wells and springs where either in-situ placement or uninterrupted flow using a flow-through chamber was possible for a minimum of 24 hours. The diffusion sampler is constructed of 30-millimeter (mm) diameter copper tubing and a semipermeable gas diffusion membrane. The flow-through chamber is an airtight chamber connected to a discharge point at the wellhead, allowing water to flow through the chamber and past the membrane. After 24 hours, when the gases in the diffusion sampler are assumed to be in equilibration with the dissolved gases in the sample water, the sampler is removed from the well or spring and immediately sealed. All dissolved-gas samples were analyzed by the University of Utah Dissolved Gas Service Center using both quadrupole and sector-field mass spectrometers. Analytical uncertainties (1-standard deviation) for ^3He, helium-4 (^4He), neon-20 (^{20}Ne), argon-40 (^{40}Ar), krypton-84 (^{84}Kr), and xenon-129 (^{129}Xe) were 2, 2, 3, 3, 5, and 5 percent, respectively.

Geochemical Methods

Environmental tracers can be used in developing and refining conceptual models of groundwater flow systems, and in determining sources of recharge, rates of movement, and ages of groundwater. They also help to refine groundwater flow paths originally delineated using water-level surface maps.

Oxygen-18 and Deuterium

The stable isotopes of water were used to better understand recharge sources to the groundwater basin. Most water molecules are made up of hydrogen (^1H) and oxygen-16 (^{16}O). However, some water molecules (less than 1 percent) contain the heavier isotopes of deuterium (^2H or D) and oxygen-18 (^{18}O). “Heavier” refers to the condition when there are additional neutrons in the nucleus of the hydrogen or oxygen atom, thereby increasing the mass or atomic weight of the water molecule.

Stable isotopes are analyzed by measuring the ratio of the heavier, less abundant isotope to the lighter, more abundant isotope and are reported as differences relative to a known standard. The isotope ratios are reported as delta (δ) values expressed as parts per thousand (permil). The δ value for an isotope ratio, R, is determined by

$$\delta R = (R_{sample}/R_{standard} - 1) \times 1{,}000 \qquad (1)$$

where

δR is the δ value for a specific isotope in the sample (^2H or ^{18}O),

R_{sample} is the ratio of the less abundant isotope to the common isotope for a specific element in the sample, and

$R_{standard}$ is the ratio of the less abundant isotope to the common isotope for the same element in the reference standard. The reference standard used in this report is Vienna Standard Mean Ocean Water (VSMOW) (Craig, 1961b; Coplen, 1994).

A positive δR value indicates that the sample is enriched in the heavier isotope with respect to the standard. A negative δR value indicates that the sample is depleted in the heavier isotope with respect to the standard. Heavier isotopes are more difficult to evaporate and easier to condense; for example, water in a lake or stream contains more heavy isotopes than the water vapor evaporated from the lake or stream. Because of this effect, water vapor in the atmosphere that condenses and falls out as precipitation will become progressively more depleted in the heavier isotopes at cooler temperatures and at higher altitudes. The proportional depletion of ^2H and ^{18}O

results in isotopic compositions of precipitation (and groundwater sourced from precipitation) that plot along a trend referred to as a meteoric water line when the δD is plotted against the $\delta^{18}O$. Because isotopic composition is affected by temperature, cooler (or high-altitude) precipitation values generally plot lower on the trend line and warmer (or low-altitude) precipitation values generally plot higher on the trend line (Mazor, 1991).

The trend line for worldwide precipitation defines the Global Meteoric Water Line (GMWL) and is described by the equation:

$$\delta D = 8(\delta^{18}O) + d \qquad (2)$$

where

d is defined as the D excess (Dansgaard, 1964). The mean global value for *d* in freshwater is 10 (Craig, 1961a).

Depending on conditions and sources of precipitation, isotopic data from specific areas may plot along a trend line that is above or below the GMWL referred to as a local meteoric water line (LMWL). In addition to temperature, isotopic composition is also affected by evaporation, particularly during irrigation or from open-water bodies. Evaporation creates preferential enrichment in ^{18}O relative to D, resulting in a shift from, and a slope less than, the LMWL or the GMWL. Groundwater with "evaporated" stable isotope compositions can often be identified as containing recharge from distinct sources such as lakes and irrigation canals.

Tritium and Helium

Tritium and helium isotopes were used in this study to examine the age of groundwater samples. Tritium (^3H) is a radioactive isotope of hydrogen with a half-life of 12.32 years that decays to tritiogenic helium-3 ($^3He_{trit}$). Tritium is present in water as part of the water molecule, whereas its decay product, $^3He_{trit}$, exists as a noble gas dissolved in water. Concentrations of ^3H and $^3He_{trit}$ can be used to determine the apparent age of groundwater that is less than about 60 years old. During the 1950s and 1960s, large amounts of ^3H were released into the atmosphere and introduced into the hydrologic cycle by above-ground thermonuclear weapons testing. As a result, ^3H concentrations in precipitation in the northern hemisphere during 1963–64 peaked at three orders of magnitude above natural concentrations (Michel, 1989). Comparison of reconstructed initial ^3H concentrations with atmospheric concentration data is a tool that can distinguish between groundwater recharged before or after the beginning of weapons testing in the mid-1950s. By using the concentrations of both ^3H and its decay product, $^3He_{trit}$, the age of groundwater (time elapsed since recharge) can be refined to an apparent recharge year. These ages are referred to as "apparent" because they can differ from the true mean age of the sample if it contains a mixture of water of different ages. Mixtures of modern (post-mid-1950s recharge) and pre-modern (pre-mid-1950s recharge) water typically have apparent $^3H/^3He_{trit}$ ages that represent the age of the young fraction of the sample because dilution with pre-modern water will leave the ratio of ^3H to $^3He_{trit}$ virtually unchanged. Details of this groundwater dating method are presented in Solomon and Cook (2000).

Tritium concentrations typically are reported in tritium units (TU), where one TU is equivalent to one molecule of tritiated water ($^3H^1HO$) in 10^{18} molecules of non-tritiated water (^1H_2O). In a sample of pre-modern groundwater, ^3H will have decayed from background "pre-bomb" concentrations of about 6 to 8 TU to less than 0.3 TU, which is approaching the analytical detection limit. Samples collected during this study having concentrations of 0.4 TU or less (accounting for a typical analytical uncertainty of 0.1 TU) are interpreted to contain no modern water, whereas samples having concentrations more than 1 TU are interpreted to contain more than a small fraction of modern water. Apparent $^3H/^3He_{trit}$ ages were computed for samples having concentrations of more than 0.4 TU.

In addition to ^3He derived from ^3H decay, groundwater also accumulates dissolved helium as it is produced from the radioactive decay of naturally occurring uranium- and thorium-series elements in aquifer solids ("crustal He") and from the upward advection and (or) diffusion of primordial helium from the mantle ("mantle He"). Crustal- and mantle-sourced He are collectively referred to as "terrigenic He" (He_{terr}) (Solomon, 2000). Crustal- and mantle-sourced He are distinguishable by their relative abundance of ^3He and ^4He isotopes. These values are generally expressed as a $^3He/^4He$ ratio (R) relative to the atmospheric $^3He/^4He$ ratio (R_a). Because crustal He has an R/R_a value of approximately 0.02 and mantle He has an R/R_a value of approximately 10–30, the R/R_a of a water sample provides information on the relative amount of crustal versus mantle sources of He_{terr}. Modern groundwater has an R/R_a value approximately equal to 1, indicating that it contains atmospheric solubility concentrations of He. In most aquifers, crustal He makes up the majority of the He_{terr}. Where this is the case, the R/R_a value of groundwater will fall below 1 as it acquires He_{terr} from time spent in contact with aquifer materials. Because He_{terr} concentrations generally increase with increasing residence time, dissolved $^4He_{terr}$ concentrations have been used as a semiquantitative tool for dating groundwater with ages from 10^3 to more than 10^6 years (Mazor and Bosch, 1992; Solomon, 2000). No attempts were made to accurately date groundwater in this study using $^4He_{terr}$, because crustal He_{terr} production rates are highly variable and substantial additional data would have been required to constrain these rates within the study area. Solomon (2000) reported average crustal ^4He production rates ranging from 0.28 to 2.4 micro cubic centimeters per cubic meters per year at standard temperature and pressure (μccSTP $m^{-3}yr^{-1}$). At these rates, groundwater should not acquire significant concentrations of $^4He_{terr}$ (more than about $2x10^{-8}$ ccSTP/g) until it has been in contact with aquifer materials for more than about 1,000 years. Even without precise knowledge of local ^4He production rates, ^4He concentrations in excess of atmospheric solubility are useful as qualitative measures of groundwater age.

Carbon-14

Carbon-14 (^{14}C) is a naturally occurring radioactive isotope of carbon that can be used to determine the apparent age of groundwater on time scales ranging from several hundred to more than 30,000 years. The method of radiocarbon dating is based on the radioactive decay of ^{14}C. In this study, the ^{14}C activity (its effective concentration) of dissolved inorganic carbon (DIC) was used to estimate the age of groundwater determined to be "pre-modern" by ^{3}H and $^{3}He_{trit}$. Unadjusted ages were calculated from non-normalized ^{14}C activities of DIC using the Libby half-life (5,568 years), assuming an initial ^{14}C activity of 100 percent modern carbon (pMC). Kalin (2000) provides a comprehensive review of the radiocarbon groundwater dating method.

Carbon-14 is produced in the upper atmosphere and is rapidly oxidized to carbon-14 dioxide ($^{14}CO_2$). Materials such as plants and water that utilize or react with atmospheric carbon dioxide (CO_2) will have a ^{14}C activity (A_o) equal to that in the atmosphere (Pearson and White, 1967). By convention, the modern pre-1950 (before nuclear weapons testing) activity of atmospheric ^{14}C is 100 pMC. Carbon-14 generally enters the hydrologic cycle through any of four predominant pathways: (1) dissolution of atmospheric CO_2 into rain water and surface water, (2) plant respired CO_2 in the soil zone that dissolves into water, (3) oxidation of organic material in soil that dissolves into water, and (4) dissolution of minerals containing geologically young carbon.

The DIC in precipitation presumably has a ^{14}C activity in equilibrium with atmospheric CO_2. As precipitation infiltrates the subsurface, its ^{14}C activity is modified by carbon exchange with soil-zone CO_2 and minerals in the unsaturated zone until it enters the saturated zone. After entering the saturated zone, interaction with soil-zone carbon ceases and the ^{14}C in the DIC decays with time. The radiocarbon age of groundwater refers to the time that has elapsed since this water was isolated from carbon in the unsaturated zone.

In addition to radioactive decay, the ^{14}C activity of groundwater in the saturated zone can be affected by additions of and (or) reactions with carbon-bearing minerals and organic phases. Four processes are of particular interest with respect to ^{14}C dating of groundwater: (1) dissolution of carbonate minerals such as limestone can increase the concentration of DIC having 0 pMC, thus decreasing the ^{14}C activity (Plummer and Sprinkle, 2001), (2) oxidation with older organic matter having 0 pMC can increase the concentration of DIC having low pMC, also decreasing the ^{14}C activity (Aravena and others, 1995), (3) sorption of calcium and magnesium ions to mineral surfaces may cause dissolution of carbonate minerals having 0 pMC, thus decreasing the ^{14}C activity (Plummer and others, 1990), and (4) carbonate mineral recrystallization (dissolution and subsequent precipitation of the same mass of carbonate mineral), which results in an isotope effect (Kendall and Caldwell, 1998), causing groundwater DIC to have a higher stable carbon isotope ratio ($\delta^{13}C$) and a lower ^{14}C activity (Parkhurst and Plummer, 1983).

These processes can greatly decrease the ^{14}C activity of groundwater. For example, in carbonate terrains such as the mountains surrounding Deep Creek Valley, modern carbon in groundwater may be diluted with dissolved ^{14}C-free carbonate minerals to the extent that very young groundwater may have ^{14}C activities as low as 50 pMC (Clark and Fritz, 1997). Thus, a correction is required to account for reaction effects on ^{14}C activity and obtain accurate radiocarbon ages. This is accomplished through a variety of models that attempt to quantify the processes described above to determine the ^{14}C activity of DIC derived from atmospheric CO_2 at the point of recharge—after the water passes through the unsaturated zone and prior to any reactions occurring within the aquifer. Several models exist to correct ^{14}C activity for the effects of the processes listed above. The most widely used formula-based models of this type are the Ingerson and Pearson (1964), Tamers (1975), and Fontes and Garnier (1979) models.

Ingerson and Pearson (1964) use a carbonate dissolution model to estimate initial ^{14}C activity (A_o) of groundwater DIC from $\delta^{13}C$ data for the inorganic carbon system, assuming that all DIC is derived from soil zone CO_2 and solid carbonates (Plummer and others, 1994). Disadvantages of the model are that it requires input that can be difficult to obtain and must often be assumed, such as the $\delta^{13}C$ of soil CO_2, and that it does not consider the effects of geochemical reactions other than mineral dissolution, particularly isotope exchange reactions. The Tamers (1975) model is a mass-balance model that considers only carbonate reaction with CO_2 gas and is based on chemical concentrations rather than $\delta^{13}C$ (Plummer and others, 1994). The dissolution of carbonate minerals dilutes ^{14}C activity by the reaction of dissolved CO_2 with solid carbonate to form bicarbonate (HCO_3^-). This model does not correct for the effects of isotope exchange. The Fontes and Garnier (1979) model is a hybrid of the Ingerson and Pearson (1964) and Tamers (1975) models, combining both chemical and isotopic data to correct for reaction effects on ^{14}C activity.

Dissolved Noble Gases

Dissolved noble-gas samples (^{20}Ne, ^{40}Ar, ^{84}Kr, and ^{129}Xe) were used to determine noble-gas recharge temperatures (NGTs, assumed to equal the temperature of groundwater recharge as it crosses the water table) as an indicator of mountain versus valley recharge. Noble gases dissolved in groundwater are primarily of atmospheric origin, and their concentrations are a function of their solubility (with the possible addition of excess air) at the temperature, pressure, and salinity conditions present as recharge crosses the water table. Most noble gases are geochemically inert and, unlike physical temperatures and age tracers (^{14}C, $^{4}He_{terr}$, $^{3}H/^{3}He_{trit}$) that change with time, noble-gas concentrations and, therefore, groundwater NGTs, should be preserved along the length of a groundwater flow path. A complete review of how these gases are used as groundwater tracers is included in Stute and Schlosser (2000).

For this study, the noble-gas concentrations were interpreted using the closed-system equilibration (CE) model (Aeschbach-Hertig and others, 2000; Kipfer and others, 2002). In addition to recharge temperature, the CE model also calculates the dimensionless ratio of the total volume of trapped (moist) air at the pressure and temperature of the free atmosphere to the volume of water beneath the water table (A) and a fractionation factor for partial dissolution of trapped air bubbles (F). Recharge altitude (the proxy for barometric pressure) is an unknown parameter, a typical situation in locations with a high topographic gradient. Because recharge temperature (T_r) and recharge altitude (H_r) are correlated, a range of NGTs (assumed to equal T_r) was estimated for each sample, as described by Manning and Solomon (2003) and Manning (2011). This method uses a minimum recharge altitude (H_{rmin}), typically that of the sample site, to calculate a maximum noble-gas recharge temperature (NGT_{max}). Conversely, the maximum possible recharge (water-table) altitude (H_{rmax}) is used to calculate a minimum noble-gas recharge temperature (NGT_{min}). For this study, H_{rmax} was estimated to be 10,500 ft based on the altitude of the highest observed springs within the study area. The recharge parameters (NGT, A, F) were evaluated using this range of recharge altitudes with a standard Newton inversion technique to minimize the error-weighted misfit (χ^2) between measured and modeled dissolved-gas concentrations (Aeschbach-Hertig and others, 1999; Manning and Solomon, 2003). A χ^2 probability threshold of 3.84, based on four measured gases and three recharge parameters ($P > 0.05$), was used to define good model fits for NGT, A, and F. Uncertainty in NGTs due to noble-gas measurement precision is generally 0.5 to 1.5 °C (Manning and Solomon, 2003; Manning, 2009; Masbruch and others, 2012). This method also uses a local H_r-T_r relationship (solution zone), derived using shallow groundwater temperatures and mountain-block NGTs for a large part of the eastern Great Basin, to further constrain the range of the NGT-H_r pairs and to define an average noble-gas recharge temperature (NGT_{avg}). Derivation of the H_r-T_r solution zone for the study area and additional details related to reported NGTs are described by Gardner and Heilweil (2014).

Results

Major Ions, Nutrients, and Selected Trace Metals

Dissolved major-ion, nutrient, and selected trace metal concentrations in groundwater samples from six wells and four springs (table 6 and fig. 18) were used to evaluate groundwater source areas and flow paths and to describe general water-quality conditions within and near Deep Creek Valley. Groundwater quality is very good throughout the study area. The concentration of dissolved solids for the 10 groundwater sites sampled during this study ranged from 126 to 475 milligrams per liter (mg/L) (table 6), and none of the sites sampled during this study had dissolved-solids concentrations that exceeded the Environmental Protection Agency (EPA) secondary standard of 500 mg/L for drinking water (U.S. Environmental Protection Agency, 2014). With the exception of slightly elevated arsenic in a stock well located west of northern Deep Creek Valley in Antelope Valley (site 1), none of the sampled wells or springs contained concentrations of dissolved solids that exceeded EPA maximum contaminant levels (MCLs) or secondary standards (table 6).

Groundwater with less than 200 mg/L dissolved-solids concentrations (sites 2, 3, 4, and 5) is found both in and near headwater recharge areas and discharge areas along the drainage axis of Deep Creek Valley (fig. 18). Groundwater sampled by the USGS and DRI in 2005 from two municipal wells on the Goshute Reservation (Goshute Lower Community Well and Goshute #2 Well) as part of the BARCAS study (Hershey and others, 2007) shows similar chemical characteristics and general high quality as other waters (sites 2, 3, 4, and 5) sampled along the drainage axis of Deep Creek Valley (table 6). Samples from sites 1 and 7 have the highest dissolved-solids concentrations (370 and 475 mg/L, respectively) and contain considerably more magnesium and sulfate combined than upgradient samples in the study area (table 6). Both of these samples were collected from sites in areas west of central Deep Creek Valley, with exposures of volcanic bedrock that could provide the source of these constituents. Groundwater from site 1 also contains considerably more sodium than any of the other samples.

The principal dissolved constituents in most samples within the study area are calcium, magnesium, and bicarbonate, all of which are directly derived from dissolution of the carbonate rocks and alluvium of eroded carbonate rocks that are abundant throughout the region. Stiff diagrams illustrate the subtle differences in water types across the study area (fig. 18). Groundwater sampled from wells on the east side of Deep Creek Valley (for example, sites 2 and 5 and one previous sample reported in Hood and Waddell (1969) shown on fig. 18) appears distinct from groundwater in upgradient areas to the south and west (sites 1, 7, 8, and 10) where groundwater contains more dissolved solids, particularly sulfate, magnesium, and sodium. Groundwater from sites 2 and 5 is recharged directly from the adjacent high-altitude portion of the Deep Creek Range, either as in-place mountain-block recharge or as recharge from runoff. Groundwater sampled from a flowing well at the northern end of Deep Creek Valley (site 9) had approximately twice the dissolved-solids concentrations as groundwater from sites 2 and 5. This could be the result of greater mineral dissolution because of a longer travel time through the aquifer or a mixture of upgradient low dissolved-solids waters from the south and east with higher dissolved-solids waters from the west. Two surface-water samples in northern Deep Creek Valley (fig. 18) have higher dissolved-solids concentrations and notably higher magnesium and sulfate concentrations than in nearby and upgradient groundwater samples, indicating that they contain a significant fraction of groundwater from the western part of the valley (or HA).

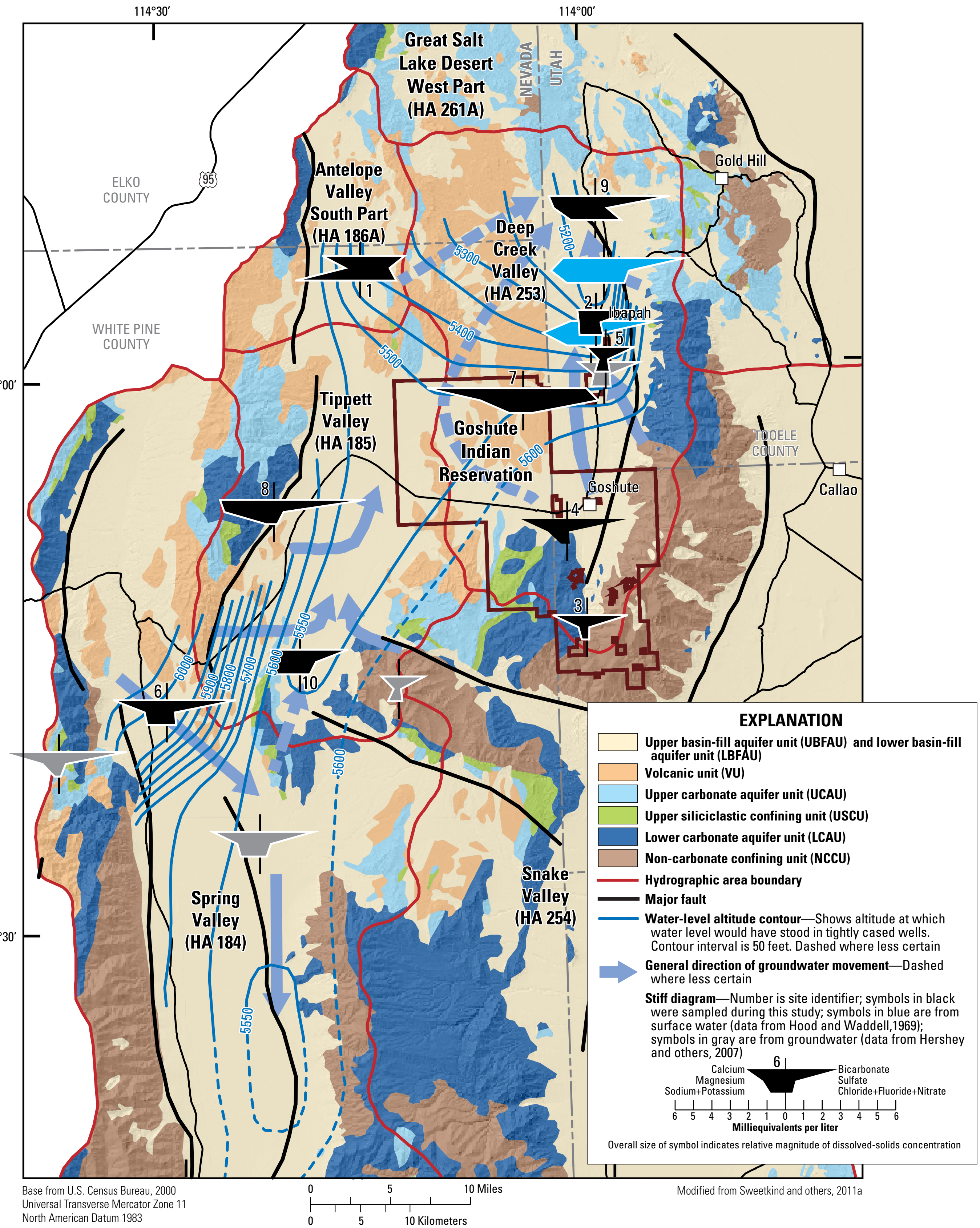

Figure 18. Stiff diagrams showing major-ion composition of groundwater from hydrogeologic units in Deep Creek Valley and adjacent areas, Utah and Nevada.

The spatial patterns of major-ion chemistry can be explained when the directions of groundwater movement shown by the water-level contour map are considered. For example, groundwater recharging the carbonate mountain block to the south and west of Spring Creek (site 4) moves northwest through the Spring Creek Flat area and then flows toward the northeast, acquiring solutes as it moves through volcanic bedrock, before re-joining and possibly mixing with low dissolved-solids groundwater (for example, sites 2 and 5) along Deep Creek in the northern end of Deep Creek Valley. Calcium-bicarbonate-type groundwaters from northern Spring Valley (site 6 and the two previous samples reported in Hood and Waddell, 1969) are not likely to move northward through Tippett Valley into Deep Creek Valley. Instead, water-level contours indicate these waters move toward the south side of the groundwater divide in northern Spring Valley.

Oxygen-18 and Deuterium

The stable-isotope composition of samples collected as part of this study are plotted along with those from 129 groundwater samples collected throughout the eastern Great Basin (Gardner and Heilweil, 2014, supplementary material) near the Utah-Nevada border for comparison (fig. 19). Stable-isotope compositions of all groundwater samples collected during this study plot between the GMWL (Craig, 1961a) and a regional arid-zone meteoric water line (Welch and Preissler, 1986), indicating that they are waters of meteoric origin (table 7 and fig. 19). Stable-isotope compositions for the greater eastern Great Basin samples range from -98 to -126 permil and from -12.3 to -16.8 permil for δD and $\delta^{18}O$, respectively. In general, samples sourced from precipitation falling at higher altitudes and (or) during the winter should be isotopically lighter (more negative values) and plot lower and farther to the left along the global and arid-zone meteoric water lines, whereas samples sourced from precipitation falling at lower altitudes and (or) during the summer should be isotopically heavier (less negative values) and plot higher and farther to the right (fig. 19). Waters with more negative values (isotopically lighter) are said to be more “depleted” because they contain fewer of the heavy stable isotopes. Stable-isotope compositions for sites 2–10 are tightly grouped with values ranging from -122 to -125 permil and from -15.8 to -16.6 permil for δD and $\delta^{18}O$, respectively (table 7). These samples are comparable to the isotopically lightest (most depleted) samples from the greater eastern Great Basin sample set and represent groundwater sourced mostly from nearby high-altitude winter precipitation. Samples from two municipal wells located on the Goshute Reservation (Goshute Lower Community Well and Goshute #2 Well) and sampled by the USGS and DRI in 2005 as part of the BARCAS study (Hershey and others, 2007), have stable-isotope values of -122 and -16.0 permil for δD and $\delta^{18}O$, respectively, indicating that they also are likely sourced from the same nearby high-altitude winter precipitation (table 7).

The stable-isotope composition for water from site 1 plots alone, with δD and $\delta^{18}O$ values of -134 and -17.1, respectively (fig. 19 and table 7). In addition to altitude and seasonal controls, there is a latitudinal effect on the stable-isotope composition of precipitation as well. Smith and others (2002) showed the gradational depletion of heavy stable-isotope compositions in modern winter recharge from south to north across the interior Great Basin. Their findings showed that modern recharge with values similar to the sample from site 1 are likely to originate from less than about 30 mi to the north and west of the current study area, agreeing with the general direction of groundwater flow indicated by potentiometric contours (fig. 8). It is also possible that groundwater in this area is Pleistocene in age and recharged during a cooler climate, resulting in more depleted isotope values.

The light (depleted) stable-isotope compositions in groundwater samples do not distinguish between recharge that occurred within the mountains (in-place mountain block recharge) and recharge of high-altitude mountain precipitation that occurs as infiltration of runoff at lower valley altitudes (mountain-front recharge). It is possible that the heaviest (isotopically) of these samples (for example, site 10) might contain a fraction of recharge derived from low altitude, likely monsoonal, precipitation. The generally light (depleted) stable-isotope compositions in the 10 Deep Creek study samples, however, are a strong indicator that most groundwater in Deep Creek Valley and the surrounding areas is sourced from high-altitude winter precipitation.

Tritium and Helium

Groundwater ^{3}H concentrations range from below detection (about 0.1 TU) to 10.1 TU and clearly identifiy “modern” water at 3 of the 10 sample sites (table 7). Terrigenic helium-4 ($^{4}He_{terr}$) concentrations range from 5.91×10^{-10} to 1.18×10^{-7} ccSTP/g and R/R_a values range from 0.24 to 1.14. The high end of the $^{4}He_{terr}$ concentrations and low end of the R/R_a values clearly indicate that some of the samples contain water that is far too old to be accurately dated using ^{3}H.

The three samples with ^{3}H values greater than 1 TU (sites 3, 4, and 8) are from springs located either within mountain recharge areas or along the mountain front of these recharge areas (fig. 20). Samples from sites 3 and 4, containing 10.1 and 8.1 TU of ^{3}H, respectively, represent groundwater from springs with short subsurface residence times in the Deep Creek Range. These samples have relatively low $^{4}He_{terr}$ concentrations and R/R_a values greater than 1, indicating that they are modern water. Site 8 is a spring discharging at the eastern foot of the Antelope Range near the western boundary of Tippett Valley. This sample has a ^{3}H concentration of 1.6 TU, a $^{4}He_{terr}$ concentration of 2.09×10^{-8} ccSTP/g, and an R/R_a value of 0.84. This combination of ^{3}H greater than 1 TU with relatively elevated $^{4}He_{terr}$ and an R/R_a significantly below 1 identifies this spring as a mixture of modern and pre-modern groundwater.

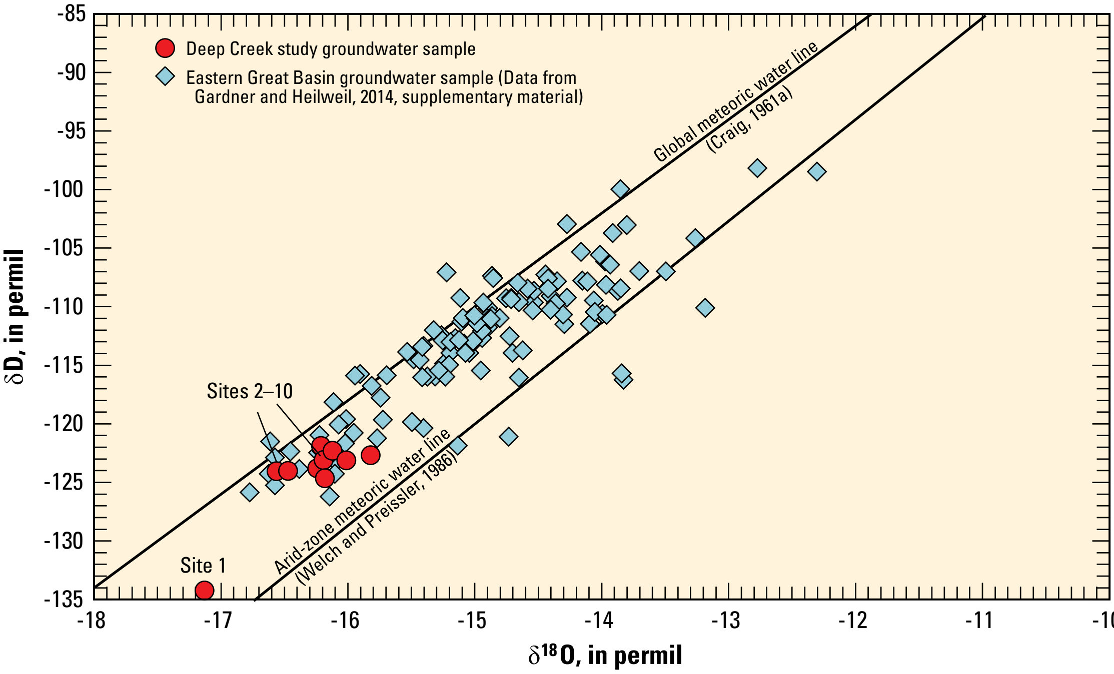

Figure 19. Stable-isotope values for groundwater in Deep Creek Valley and adjacent areas, Utah and Nevada, in reference to the global meteoric water line and an arid-zone meteoric water line.

Ratios of ^{3}H to $^{3}He_{trit}$ were used to further refine modern groundwater ages (table 7). Samples from sites 3 and 4 have apparent $^{3}H/^{3}He_{trit}$ ages of 7 and 8 years, respectively. The modern component of the mixed water from site 8 has an apparent $^{3}H/^{3}He_{trit}$ age of 29 years. The remaining groundwater samples were collected from valley-altitude wells and springs, contain 0.4 TU or less of ^{3}H, and clearly consist of pre-modern water.

Carbon-14

Carbon-14 activity measured from DIC in groundwater samples from the 10 sites within and surrounding Deep Creek Valley ranged from 15 to 82 pMC (table 7 and fig. 21). Samples from two springs (sites 3 and 4) that had been determined to discharge only modern water, on the basis of ^{3}H and $^{3}He_{trit}$ values, had ^{14}C activities of 82 and 72 pMC, respectively. These springs are in southern Deep Creek Valley and represent groundwater with short subsurface residence times in the Deep Creek Range. Water from a spring discharging at the eastern foot of the Antelope Range near the western boundary of Tippett Valley (site 8) had a ^{14}C activity of 44 pMC that is interpreted to be a mixture of modern and pre-modern water based on ^{3}H and $^{3}He_{trit}$ concentrations. Samples from one flowing well in Deep Creek Valley (site 5) and a valley-altitude spring in northern Spring Valley (site 6), were determined to be pre-modern with respect to ^{3}H, and had ^{14}C activities of 65 and 70 pMC, respectively, indicating that they may not be significantly older than pre-1950s water. The remaining five samples had ^{14}C activities ranging from 15 to 29 pMC, indicating that groundwater throughout much of the study area may be thousands of years old or more.

Radiocarbon ages were calculated for the eight samples (sites 1, 2, 5, 6, 7, 8, 9, and 10) that were determined to be pre-modern or contain pre-modern water based on ^{3}H and ^{4}He concentrations. Table 7 lists unadjusted and adjusted radiocarbon ages for these samples. Adjusted ages were determined using the formula-based adjustment models of Ingerson and Pearson (1964), Tamers (1975), and Fontes and Garnier (1979). The adjusted ages were converted to calendar years before present (BP) using the Fairbanks 0107 calibration curve (Fairbanks and others, 2005). In the event that a particular model resulted in an unreasonable (negative) age, the adjusted age was designated as either modern (recharged after the mid-1950s), pre-modern (recharged prior to the mid-1950s), or a mixture of modern and pre-modern water with respect to ^{14}C based on evaluating other age-related tracers. Because the Tamers model does not rely on isotopic data to correct for reaction effects on ^{14}C activity, the younger Ingerson and Pearson (I&P) and Fontes and Garnier (F&G) (table 7) adjusted ages are considered conservative and likely more representative of the true age. Sensitivity analysis using the adjustment models indicates that the uncertainty in adjusted radiocarbon ages is on the order of several thousand years, making pre-modern waters with adjusted ^{14}C ages of less than 2,000 years difficult to categorize. Adjusted radiocarbon ages for each of these models are compared to unadjusted ages in table 7. The radiocarbon-age adjustment models require ^{14}C and $\delta^{13}C$ values of

Table 7. Stable- and radio-isotope data used to estimate ages of groundwater sampled during May 2012 in Deep Creek Valley and adjacent areas, Utah and Nevada.

[Site ID: see figure 17 for locations and table A1-2 for additional information. Pre-modern, groundwater that recharged prior to the mid-1950s; Modern, groundwater that recharged after the mid-1950s; Modern mixture, sample that contains a mixture of pre-modern and modern groundwater. Abbreviations: USGS, U.S. Geological Survey; ^{18}O, oxygen-18; D, deuterium (hydrogen-2); TU, tritium units; ccSTP/g, cubic centimeters per gram of water at standard temperature and pressure; BP, before present; pMC, percent modern carbon; ^{13}C, carbon-13; I&P, Ingerson and Pearson (1964) model; F&G, Fontes and Garnier (1979) model; —, no data]

Site ID	USGS site number	$\delta^{18}O$[1], permil	δD[1], permil	Tritium (^3H) and precision, TU	R/R_a[2]	Measured Helium-4 (^4He), ccSTP/g	[3]Terrigenic Helium-4 ($^4He_{terr}$), ccSTP/g	[3]Tritiogenic Helium-3 ($^3He_{trit}$), TU	Apparent $^3H/^3He_{trit}$ age, years BP	Carbon-14 (^{14}C), pMC	[1]$\delta^{13}C$, permil	Unadjusted ^{14}C age, thousands of years BP	^{14}C age I&P, thousands of years BP	^{14}C age Tamers, thousands of years BP	^{14}C age F&G, thousands of years BP
1	400543114145101	-17.1	-134	0.0 +/- 0.1	0.24	1.56×10^{-7}	1.18×10^{-7}	—	Pre-modern	17	-7.7	14	7	10	6
2	400216113591701	-16.6	-124	0.4 +/- 0.1	0.34	1.46×10^{-7}	1.01×10^{-7}	—	Pre-modern	29	-9.4	10	3	6	3
3	394554114003501	-16.5	-124	10.1 +/- 0.8	1.14	4.57×10^{-8}	1.27×10^{-9}	4.3	7.0	82	-10	Modern	Modern	Modern	Modern
4	395112114014501	-16.2	-124	8.1 +/- 0.6	1.14	3.57×10^{-8}	3.20×10^{-9}	4.6	8.0	72	-9.7	Modern	Modern	Modern	Modern
5	400024113582701	-16.2	-122	0.0 +/- 0.1	0.91	4.93×10^{-8}	8.99×10^{-9}	2.5	Pre-modern	65	-12	3	Pre-modern	Pre-modern	Pre-modern
6	394149114302201	-16.2	-123	0.1 +/- 0.1	0.88	5.15×10^{-8}	1.07×10^{-8}	2.4	Pre-modern	70	-9.9	3	Pre-modern	Pre-modern	Pre-modern
7	395815114043401	-16.2	-125	0.1+/- 0.1	0.58	8.79×10^{-8}	4.10×10^{-8}	1.7	Pre-modern	24	-6.6	11	2	8	1
8	395237114222501	-16.1	-122	1.6 +/- 0.1	0.84	5.83×10^{-8}	2.09×10^{-8}	6.3	29.0	44	-8.4	7	Modern mixture	3	Modern mixture
9	400831113591001	-16.0	-123	0.0 +/- 0.1	0.80	4.93×10^{-8}	6.65×10^{-9}	—	Pre-modern	15	-7.6	15	8	13	8
10	394422114205201	-15.8	-123	0.0 +/- 0.1	1.00	4.22×10^{-8}	5.91×10^{-10}	—	Pre-modern	20	-6.1	13	3	9	3
[4]11	395935113584501	-16.0	-122	—	—	—	—	—	—	63	-9.1	—	—	—	—
[4]12	—	-16.0	-122	—	—	—	—	—	—	—	—	—	—	—	—

[1]Delta values (**δ**) are defined in equation 1.

[2]R is the $^3He/^4He$ ratio of the sample, and Ra is the $^3He/^4He$ ratio of air (1.384×10^{-6}).

[3]Interpreted value derived using the Closed-Equilibrium dissolved-gas model (Aeschbach-Hertig and others, 2000; Kipfer and others, 2002).

[4]Data for Site IDs 11 (Goshute Lower Community Well) and 12 (Goshute #2 Well) from Hershey and others, 2007, *http://www.dri.edu/images/stories/research/projects/BARCAS/barcas.pdf*.

soil zone CO_2 and carbonate minerals that are available to react with or add to the DIC of the groundwater. In all cases, ^{14}C activity was assumed to be 100 pMC and soil zone CO_2 was assumed to have a $\delta^{13}C$ value of -22 permil, the average value for C3 plant-dominated terranes in Utah (Hart and others, 2010). Carbonate minerals were assumed to have 0 pMC because of their age (middle Cambrian to Permian) and a $\delta^{13}C$ value of 0 permil, approximately the worldwide average for marine limestones (Keith and Weber, 1964).

Unadjusted radiocarbon ages range from 3,000 to 15,000 years for water determined to be pre-modern based on tritium (table 7). The largest age adjustments (difference between unadjusted and adjusted radiocarbon ages using the various adjustment models) ranged from 3,000 to 10,000 years, yielding minimum adjusted radiocarbon age estimates of less than 1,000 to 8,000 years. None of the I&P and F&G adjusted ages are older than 8,000 years, and only one of the Tamers adjusted ages is older than 10,000 years, indicating that all of the samples are likely Holocene in age, and that these waters recharged during similar climatic conditions (temperature and precipitation) to those that exist today. Sites 5 and 6 are a flowing well in Deep Creek Valley and a spring in northern Spring Valley that clearly discharge pre-modern groundwater as determined by their lack of ^3H, elevated $^4He_{terr}$, and an R/R_a of less than 1. Radiocarbon age-adjustment models, however, resulted in negative ages indicating that these pre-modern waters are late Holocene in age (between about 50 and 2,000 years old), a range difficult to date with either ^3H or ^{14}C methods. Carbon-14 ages support the ^3H and He isotope data that identifies site 8 as a spring containing a mixture of modern and pre-modern groundwater.

Noble-Gas Recharge Temperatures

Dissolved noble-gas concentrations and NGTs are presented for the 10 sample sites in table 8. The range of possible NGT values calculated for each of the 10 sites is shown on figure 22, in which the left and right points for each sample represent NGT_{min} and NGT_{max}, respectively. Because NGTs represent estimates of recharge temperature (the water table temperature at the location of recharge), they are compared to valley water-table temperatures to identify areas where groundwater consists of mountain rather than valley recharge. Domenico and Schwartz (1998) noted that shallow water-table temperatures (and thus, recharge temperatures, T_r) are generally close to, but slightly warmer (about 1 to 2 °C) than, the mean annual air temperature at the land surface for typical water-table depths of less than 65 ft. Because air temperatures and thus, water-table temperatures, decrease with increasing altitude, modern (or Holocene) mountain recharge should have T_r values that are cooler than the temperature of the water table in adjacent valleys. Groundwater temperatures ranging from 11.2 to 15.4 °C from 30 valley wells with water

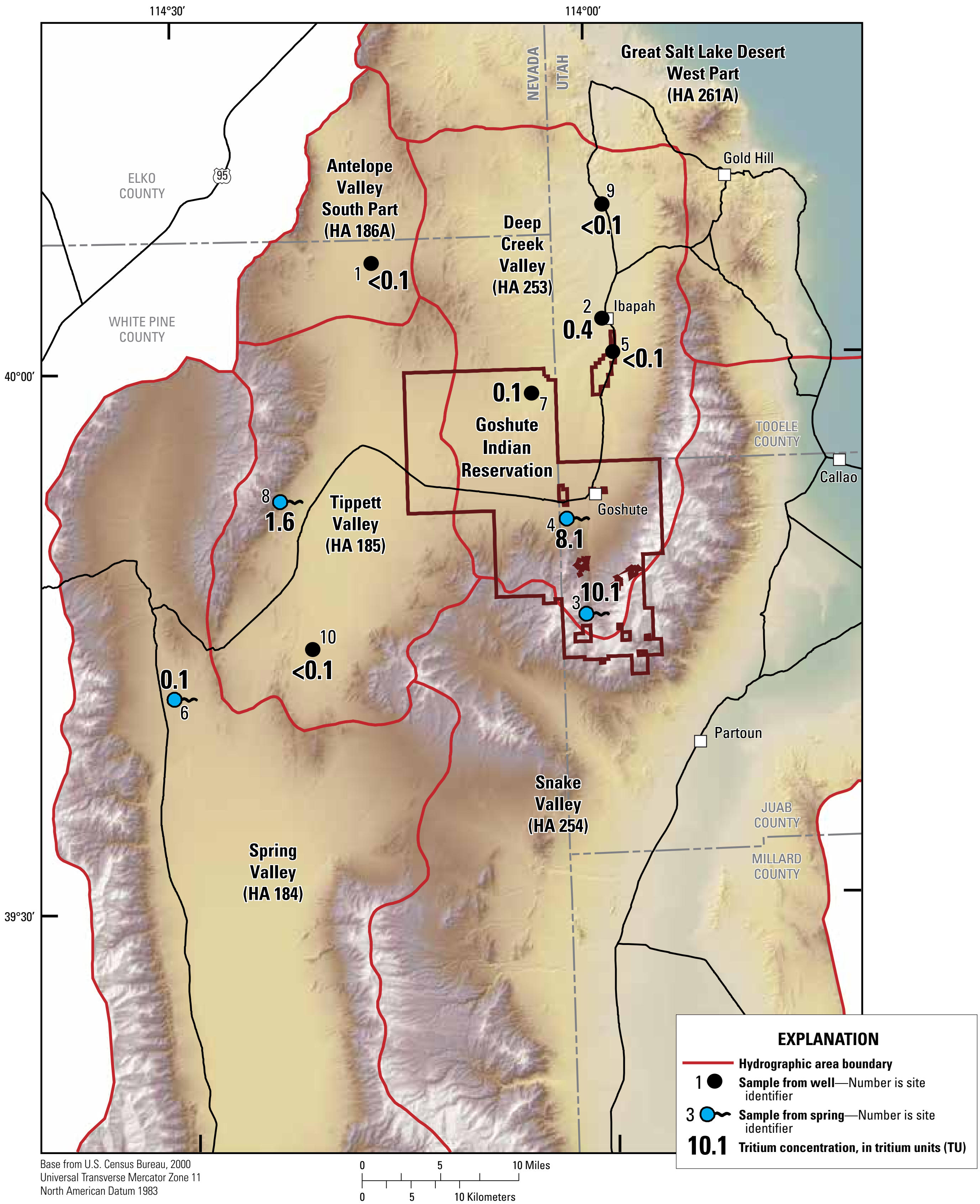

Figure 20. Tritium concentrations in groundwater in Deep Creek Valley and adjacent areas, Utah and Nevada.

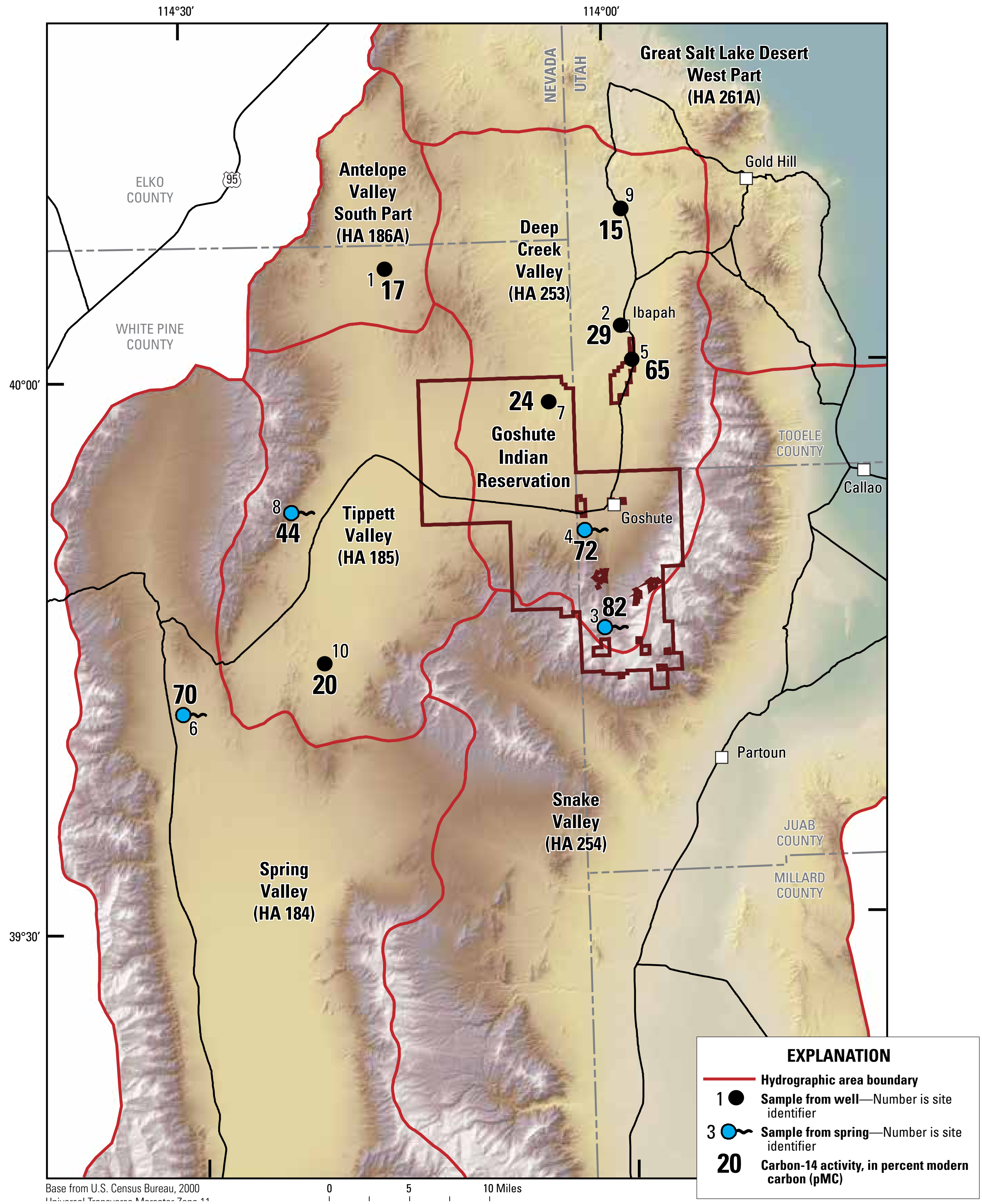

Figure 21. Carbon-14 activities in groundwater in Deep Creek Valley and adjacent areas, Utah and Nevada.

depths less than 230 ft, and temperatures ranging from 4.3 to 14.6 °C of 65 springs at all altitudes in nearby parts of the eastern Great Basin, verify that water-table temperatures on average are about 3 °C warmer than mean annual air temperatures in this region (fig. 23).

Values of NGT_{min} and NGT_{max} for the study area range from 1.4 to 12.9 °C and from 4.0 to 17.8 °C, respectively (fig. 22 and table 8). Average dissolved-gas recharge temperatures (NGT_{avg}) are cooler than valley recharge temperatures for six of the samples (sites 2, 3, 4, 6, 8, and 9) and within the range of valley recharge temperatures for four of the samples (sites 1, 5, 7, and 10) (fig. 22). NGT_{avg}, the mid-point for each sample displayed on fig. 22, is calculated by assuming that the recharge altitude (H_r) is equal to the median altitude within the watershed. The NGT_{avg} values are less than measured water-table temperatures for all but two samples (sites 6 and 7), which was expected because recharge should have occurred at a higher altitude than sample collection. The NGT_{avg} values from sites 2, 3, and 4 are all less than 4.5 °C, indicating that these waters all originated as mountain recharge. This is expected for sites 3 and 4, springs located within or at the mountain front in the southern Deep Creek Range. Site 2 is a 216-ft deep alluvial well along the drainage axis of Deep Creek Valley near areas of groundwater discharge by springs and ET. The low NGT_{avg} from site 2 is evidence that direct infiltration to mountain bedrock in parts of the Deep Creek Range moves into the adjacent basin-fill aquifer through the subsurface. Samples from sites 6, 8, and 9 have somewhat warmer NGT_{avg}. Groundwater from these sites might represent integrated mixtures of groundwater flow paths with a range of recharge temperatures or simply reflect the fact that they are located downgradient of lower-altitude mountain recharge areas. Samples from sites 1, 7, and 10 have NGT_{avg} values of 14.6, 15.4, and 13.5 °C, respectively, which clearly indicate recharge at valley altitudes. Groundwater from these sites likely recharged as infiltration of runoff through coarse stream-bed or alluvial fan material upgradient of each of the sample sites. Site 5, a flowing well along the drainage axis of Deep Creek Valley on a part of the Goshute Indian Reservation, has an NGT_{avg} of only 11.5 °C, which is very close to the transition between mountain and valley recharge (fig. 22). Discharge from this flowing well may represent mountain front recharge originating at the foot of the Deep Creek Range or a mixture of mountain front recharge and direct mountain block recharge originating higher in the Deep Creek Range.

Dissolved-gas data and associated NGTs support a conceptual model where most groundwater in the alluvial aquifer of eastern Deep Creek Valley, and along Deep Creek, consists of modern or Holocene recharge that originated as infiltration of precipitation in the Deep Creek Range or of runoff along the mountain front. Furthermore, these data also agree well with previous studies (Hood and Waddell, 1969) showing recharge in the mountains as the largest source of water to the basin-fill aquifers adjacent to the Deep Creek Range.

Table 8. Dissolved-gas concentrations and related noble-gas temperature data for groundwater sampled in Deep Creek Valley and adjacent areas, Utah and Nevada.

[Site ID: see figure 17 for locations and table A1-2 for additional information. Dissolved-gas sample collection method: CT, copper tube; DS, diffusion sampler. Abbreviations: USGS, U.S. Geological Survey; mmHg, millimeters of mercury; ccSTP/g, cubic centimeters at standard temperature and pressure per gram of water; H_{rs}, standard recharge altitude, which is the altitude at which the sample was collected; NGT_{rs}, standard noble-gas temperature calculated using H_{rs}; A, dimensionless ratio of the total volume of trapped (moist) air at the pressure and temperature of the free atmosphere to the volume of water; F, fractionation factor for partial dissolution of trapped air bubbles; $\Sigma\chi^2$, sum of error-weighted misfit for each of the noble gases; H_{rmin}, minimum recharge altitude; NGT_{max}, maximum noble-gas recharge temperature; H_{ravg}, average recharge altitude; NGT_{avg}, average noble-gas recharge temperature; H_{rmax}, maximum recharge altitude; NGT_{min}, minimum noble-gas recharge temperature; °C, degrees Celsius]

Site ID	USGS site number	Sample method	Dissolved-gas pressure, mmHg	Dissolved noble-gas concentrations: Helium-4 (^4He), ccSTP/g	Neon-20 (^{20}Ne), ccSTP/g	Argon-40 (^{40}Ar), ccSTP/g	Krypton-84 (^{84}Kr), ccSTP/g	Xenon-129 (^{129}Xe), ccSTP/g	Modeled recharge parameters: H_{rs}, feet	NGT_{rs}, °C	A	F	$\Sigma\chi^2$	H_{rmin}, feet	NGT_{max}, °C	H_{ravg}, feet	NGT_{avg}, °C	H_{rmax}, feet	NGT_{min}, °C
1	400543114145101	CT	657	1.56×10^{-7}	3.32×10^{-8}	2.75×10^{-4}	1.43×10^{-7}	2.28×10^{-9}	5,988	16.8	0.017	0.93	1.20	5,988	16.8	8,243	14.6	10,500	12.3
2	400216113591701	CT	623	1.46×10^{-7}	4.40×10^{-8}	3.49×10^{-4}	1.75×10^{-7}	3.32×10^{-9}	5,280	7.4	0.001	0.00	2.28	5,280	7.4	7,889	4.4	10,500	2.0
3	394554114003501	DS	625	4.57×10^{-8}	1.72×10^{-7}	3.52×10^{-4}	4.74×10^{-8}	3.32×10^{-9}	7,560	4.0	0.003	0.33	0.03	7,560	4.0	9,029	2.7	10,500	1.4
4	395112114014501	DS	613	3.57×10^{-8}	1.33×10^{-7}	3.14×10^{-4}	4.39×10^{-8}	3.10×10^{-9}	6,335	6.0	0.020	1.19	0.01	6,335	6.0	8,417	4.2	10,500	2.3
5	400024113582701	DS	656	4.93×10^{-8}	1.59×10^{-7}	3.14×10^{-4}	3.39×10^{-8}	2.59×10^{-9}	5,380	12.4	0.100	0.94	0.04	5,380	12.4	6,790	11.5	8,200	9.5
6	394149114302201	DS	613	5.15×10^{-8}	1.73×10^{-7}	3.55×10^{-4}	4.50×10^{-8}	2.88×10^{-9}	5,980	11.0	0.100	0.82	0.49	5,980	11.0	7,091	10.3	8,200	9.6
7	395815114043401	DS	662	8.79×10^{-8}	1.76×10^{-7}	2.98×10^{-4}	3.72×10^{-8}	2.27×10^{-9}	5,540	17.8	0.016	0.67	0.17	5,540	17.8	8,020	15.4	10,500	12.9
8	395237114222501	DS	653	5.83×10^{-8}	1.45×10^{-7}	2.91×10^{-4}	4.06×10^{-8}	2.54×10^{-9}	6,240	11.7	0.000	0.00	1.35	6,240	11.7	8,369	10.5	8,150	9.5
9	400831113591001	CT	591	4.93×10^{-8}	6.05×10^{-8}	3.18×10^{-4}	1.67×10^{-7}	4.17×10^{-9}	5,134	11.6	0.001	0.00	1.30	5,134	11.6	7,817	8.9	9,800	6.2
10	394422114205201	CT	624	4.22×10^{-8}	3.55×10^{-8}	3.08×10^{-4}	1.64×10^{-7}	2.61×10^{-9}	5,775	15.9	0.017	0.82	3.80	5,775	15.9	8,136	13.5	10,500	11.2

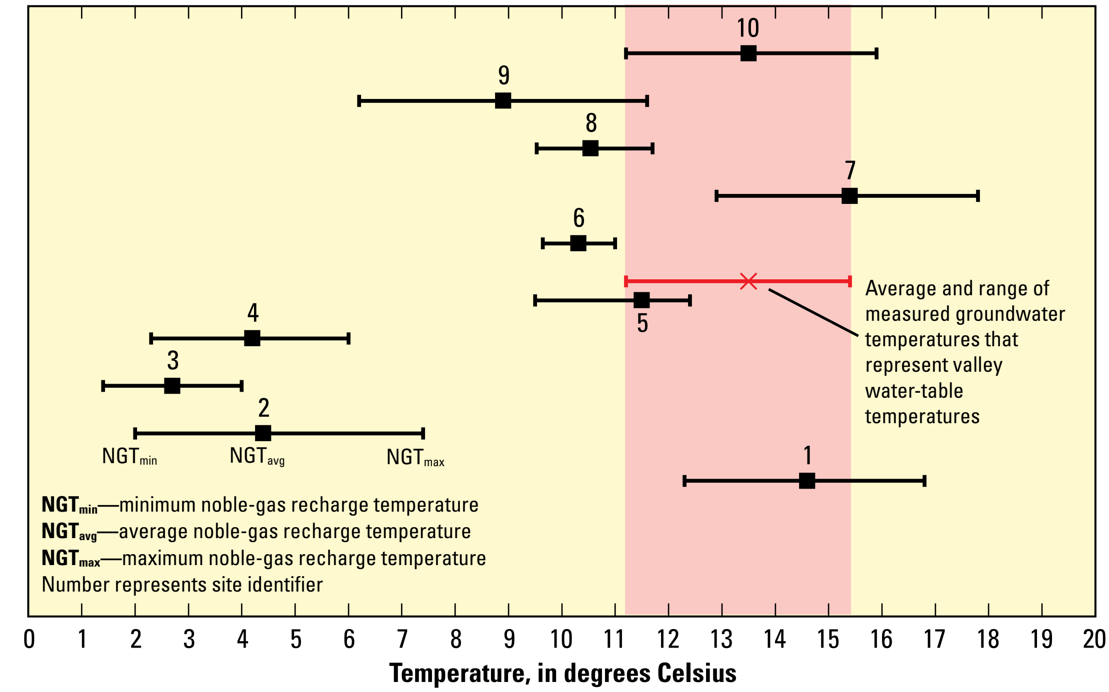

Figure 22. Noble-gas recharge temperatures for groundwater in Deep Creek Valley and adjacent areas, Utah and Nevada, compared to regional valley water-table temperatures.

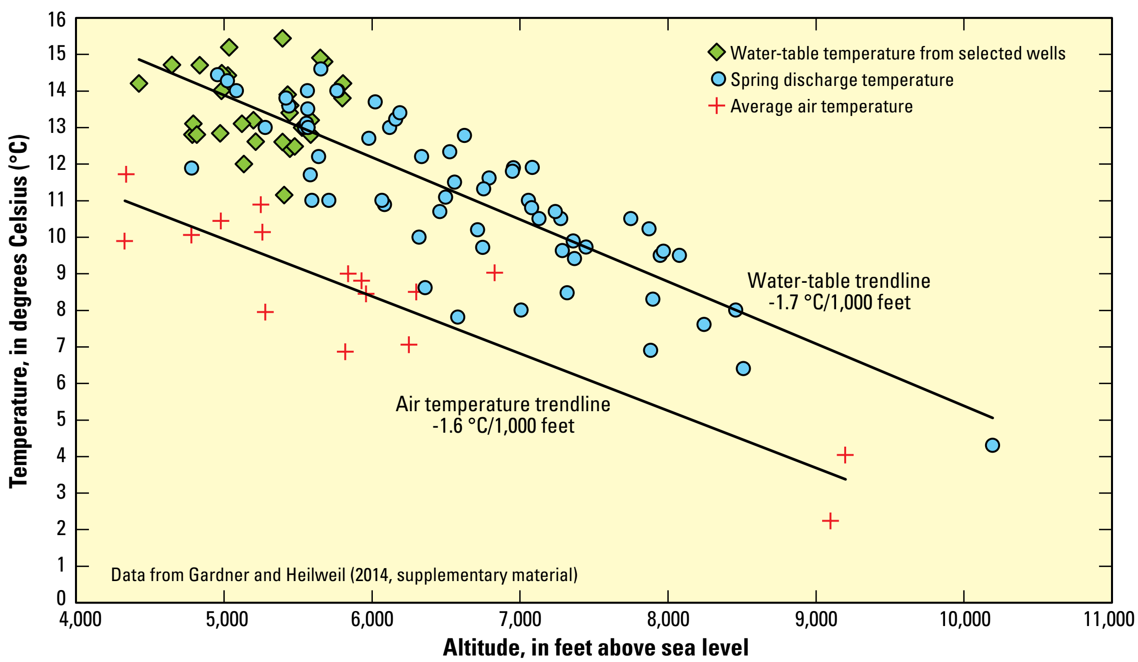

Figure 23. Air and water-table temperatures versus altitude for selected wells and springs in the eastern Great Basin, Utah and Nevada.

Discussion

The principal basin-fill aquifer in Deep Creek Valley is similar to basin-fill aquifers throughout the eastern Great Basin in many ways. Precipitation rates are generally too low to provide groundwater recharge at low altitudes. Instead, recharge originates as infiltration of precipitation that falls on mountains surrounding the valley. Where the mountains are composed of permeable bedrock, recharge enters the aquifer system as direct infiltration of precipitation. In mountain areas with less permeable bedrock, precipitation and snowmelt form runoff that gathers in stream channels and infiltrates permeable basin-fill deposits near the mountain front. Both forms of infiltration become groundwater recharge that then moves downgradient toward the valley axis where continuous discharge occurs by evapotranspiration and springs. The principal basin-fill aquifer in Deep Creek Valley is the source of water to all valley springs, wells, and wetlands. The sustainability (or resilience) of these resources is a primary concern to the Goshute Tribe. Water-level altitudes and geochemical and hydrogeologic data were evaluated during this study to provide a more detailed understanding of the sources of recharge, groundwater flow paths and travel times, and areas of potential hydraulic connection within Deep Creek Valley and between neighboring basins.

Principal Aquifer Characteristics

Water-level altitude (or potentiometric) contours indicate that groundwater moves into the principal basin-fill aquifer in Deep Creek Valley from the east and south, and into the northern part of Deep Creek Valley, from the southwest and west (figs. 8 and 18). Groundwater moving from the east and south originates as mountain-block or mountain-front recharge from the Deep Creek Range. Samples collected during this study (sites 2 and 5) and by Hood and Waddell (1969) show that portions of the aquifer to the south and east of Deep Creek contain calcium-bicarbonate type groundwater with low dissolved solids, a chemical signature obtained from the mineral dissolution of the carbonate rocks that are abundant in the Deep Creek Range and beneath Deep Creek Valley as the UCAU and LCAU. Also, NGTs indicate that groundwater sampled along the valley axis recharged at both mountain and valley altitudes, providing evidence for both mountain-block and mountain-front recharge.

Potentiometric contours imply groundwater flow from western Spring Creek Flat (near site 7), and from the directions of Antelope Valley (near site 1) and Tippett Valley (near sites 8 and 10) towards the valley axis of Deep Creek Valley. It is not likely that significant recharge occurs from precipitation falling directly on the low hills to the west of Deep Creek because all of the stable isotope samples are depleted in heavy isotopes, indicating that all groundwater in the study area originates as high-altitude precipitation. Although information gathered during this study is insufficient to conclude whether or not groundwater does travel along these extended and possibly interbasin flow paths, it does suggest that a small fraction of the lower altitude, northern Deep Creek Valley discharge may be sourced from these areas.

Groundwater originating from either Spring Creek Flat or from southwest and west of Deep Creek Valley is likely to contact volcanic rocks or sediments derived from volcanic rocks of the VU that are prevalent across the area of low hills and badlands within the northwestern quarter of the Deep Creek Valley HA. Dissolution of minerals sourced from these rocks can explain the increase in dissolved sulfate and chloride in groundwater sampled at sites 7 and 9 and in surface water sampled by Hood and Waddell (1969) (fig. 18), indicating that some Deep Creek Valley groundwater is sourced from these areas. The NGTs for Deep Creek discharge-area groundwater (sites 2, 5, and 9), however, are significantly cooler than most water sampled from upgradient areas to the southwest and west (sites 1, 7, and 10) (table 8), indicating that Deep Creek discharge-area groundwater can contain no more than a small fraction of groundwater sourced from these distant areas. Furthermore, flow-weighted average groundwater age estimates at several flowing wells (sites 5, 7, and 9) in Deep Creek Valley indicate that travel times of groundwater discharging from the Deep Creek principal basin-fill aquifer are between 1,000 and 8,000 years. If substantial amounts of the Deep Creek basin-fill groundwater were sourced from the basins to the west and southwest, where recharge rates are lower and flow paths are longer, travel times would likely be longer, similar to that in basins to the south and east of Deep Creek Valley, where groundwater is Pleistocene (greater than 10,000 years) in age.

Potential Interbasin Connection

Potentiometric contours and geologic data indicate a clear hydraulic separation between Deep Creek Valley and Snake Valley to the east and south; as a result, future withdrawals in Snake Valley are not likely to affect water resources in Deep Creek Valley. Rates of recharge and runoff over the Deep Creek Range are high relative to the valleys on either side (figs. 10 and 12) of the range, causing the groundwater level beneath or immediately adjacent to the range to remain higher than the groundwater level beneath the valleys on both sides; this results in a groundwater divide between Snake and Deep Creek Valleys. Additionally, a thick sequence of low-permeability bedrock (NCCU along the east and USCU in the south) exists at or below the surface along most of the boundary between Deep Creek Valley and Snake Valley, acting as a hydraulic barrier between the basins (figs. 6 and 7, cross sections *B-B′*, *E-E′*, *F-F′*, and *G-G′*).

Hydraulic connectivity between Spring Valley, Tippett Valley, and Deep Creek Valley is a pre-requisite for pumping in Spring Valley to have an effect on groundwater conditions in Deep Creek Valley more than 20 mi to the northeast. The hydrogeologic framework of the GBCAAS as described by Sweetkind and others (2011a) and Cederberg and others (2011), was used to construct cross sections through the current study area and to evaluate the potential for hydraulic connectivity between these basins. Hydraulic connection is possible wherever permeable units are adjacent to one another and not interrupted by relatively impermeable units or geologic structures. The UBFAU, LBFAU, UCAU, and LCAU are generally permeable enough to transmit groundwater whereas the USCU and NCCU generally act as confining units, hydraulically separating portions of local and regional aquifers where they are present. The VU is reported to have a highly variable permeability (Sweetkind and others, 2011a), functioning as an aquifer in parts of the Great Basin and as a confining unit in others. Only discontinuous Quaternary faults occur between Spring Valley and Deep Creek Valley.

Cross sections *A-A′* and *B-B′* (fig. 7) show the regional HGUs at depth along the line for which a hydraulic connection between Spring Valley, Tippett Valley, and Deep Creek Valley is assessed. In section *A-A′*, saturated conditions in basin-fill aquifers exist at an altitude of over 5,750 ft in northern Spring Valley, slope down to around 5,450 ft in northeastern Tippett Valley, and are as low as about 5,150 ft in northern Deep Creek Valley. Saturated basin-fill deposits between Spring and Tippett Valleys are separated by only a short section of VU overlying USCU between the basins (fig. 7, point 1). Saturated basin-fill deposits between Tippett and Deep Creek Valleys along this section are separated by USCU at depths up to 2,000 ft below the potentiometric surface with a small break in the USCU at this depth (fig. 7, point 2) providing a potential hydraulic connection through the VU. In both cases, the hydraulic connection depends on the uncertain permeability of the VU. Furthermore, a connection between Tippett and Deep Creek Valleys depends on the accuracy of the USCU thickness at depth.

In cross section *B-B′*, saturated conditions in basin-fill aquifers exist at an altitude of about 5,600 ft in northern Spring Valley, slope down to about 5,525 ft in eastern Tippett Valley, and are as low as about 5,300 ft in central Deep Creek Valley. Saturated basin-fill deposits between Spring and Tippett Valleys are separated by only the relatively permeable LCAU (fig. 7, point 3). Saturated basin-fill deposits between Tippett and Deep Creek Valleys along this section are separated by VU at depths of up to about 2,000 ft below the potentiometric surface. Below 2,000 ft, the saturated basin-fill deposits in Tippett and Deep Creek Valleys are separated by a thick section of LCAU (fig. 7, point 4). Water-level records from monitoring wells with multiple depth completions and monitoring wells screened at various depths located near one another in Snake Valley indicate that basin-fill and underlying carbonate aquifers are in good hydraulic connection (Gardner and others, 2011; Utah Department of Natural Resources, 2014). If the same is true in the current study area, the relation between aquifer units in section *B-B′* indicates that a hydraulic connection between northern Spring and Deep Creek Valleys through Tippett Valley is likely.

Potentiometric contours and groundwater ages also indicate the potential for a long flow path between northern Spring and Deep Creek Valleys through Tippett Valley (fig. 24). Potentiometric contours indicate a groundwater divide near the central axis of northern Spring Valley and a continuous gradient in a northeastward direction from this point to the northernmost part of Deep Creek Valley. Furthermore, north of the Spring Valley-Tippett Valley divide a 20-mi long area of very low hydraulic gradient ($3x10^{-4}$ ft/ft, from southeast to northwest along the central axis of Tippett Valley) exists that may indicate high permeability associated with a thick section of basin-fill deposits along the central axis of the valley (fig. 7, *A-A′* and *E-E′*).

Geochemical data are generally inconclusive with regard to delineating interbasin flow paths in the study area. The number of sites sampled is limited, especially given the large multiple-basin area, and samples collected from the Deep Creek Valley discharge area must be interpreted as mixtures of groundwater from varying recharge sources with unknown mixing ratios. Groundwater ages interpreted from multiple age-related geochemical tracers, however, do generally increase in a down-gradient direction from pre-modern to late Holocene (sites 6, 8, and 10) to early Holocene (site 9) (fig. 24). Despite the uncertainty because of limited data collection points, especially across much of Tippett and western Deep Creek Valleys, a hydraulic connection between northern Spring Valley and Deep Creek Valley, through Tippett Valley, appears likely, and potential regional effects resulting from future groundwater withdrawals in northern Spring Valley warrant ongoing monitoring of groundwater levels in appropriate areas.

Deep Creek Valley aquifer conditions have remained unaffected for the most part over approximately the past half-century under current (2014) and past levels of groundwater withdrawal. Annual water levels are currently being monitored at three wells in Deep Creek Valley: USGS sites 400219113591901, 400201113583801, and 395608113583301 (fig. 25 and table A1-1). Continued annual water-level monitoring of these wells in eastern Deep Creek Valley should be sufficient to indicate if water levels decline as a result of increased groundwater withdrawals within the HA in the future. Water-level records for these wells and all wells and springs in the Utah active water-level network can be accessed at *http://groundwaterwatch.usgs.gov/statemap.asp?sc=49&sa=UT.*

Effects on Deep Creek Valley groundwater resources resulting from future groundwater withdrawals in Spring Valley (by SNWA or other large-scale future development) will take substantially longer to occur than effects from nearby groundwater withdrawals. It is not possible with current information and existing (steady-state) numerical models to estimate the time required after the onset of withdrawals for water-level declines to be observed in Deep Creek Valley resulting from a lowering of the water table in Spring Valley. By the time an effect is observed over this distance (greater than 30 mi), the volume of water removed from the groundwater system would be so large that an immediate reduction in withdrawals would not prevent further decline in water levels or spring discharge. For this reason, it is recommended that a water-level monitoring network be established between Deep Creek Valley and northern Spring Valley to provide early indications of potential effects.

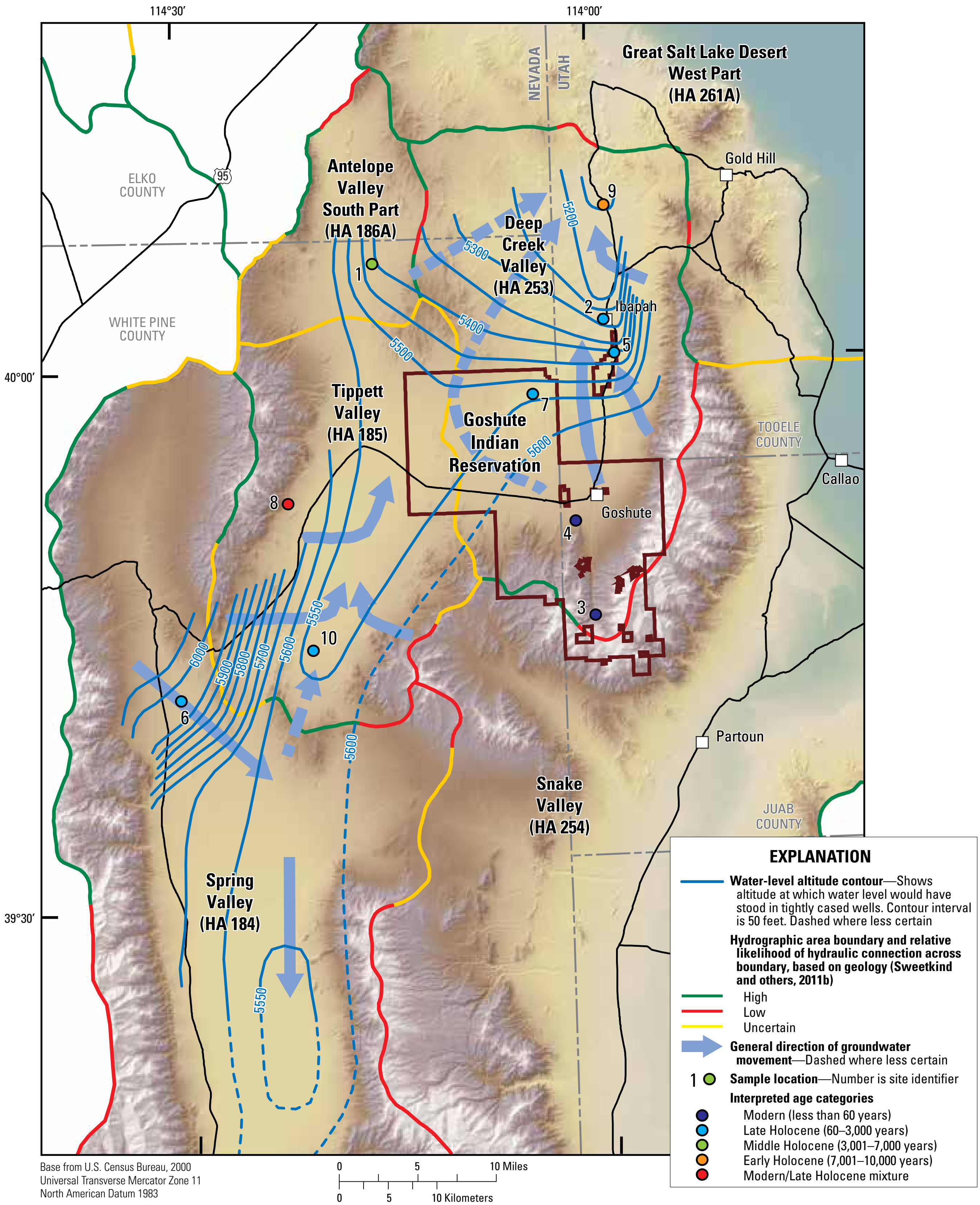

Figure 24. Interpreted age categories for groundwater sampled in Deep Creek Valley and adjacent areas, Utah and Nevada.

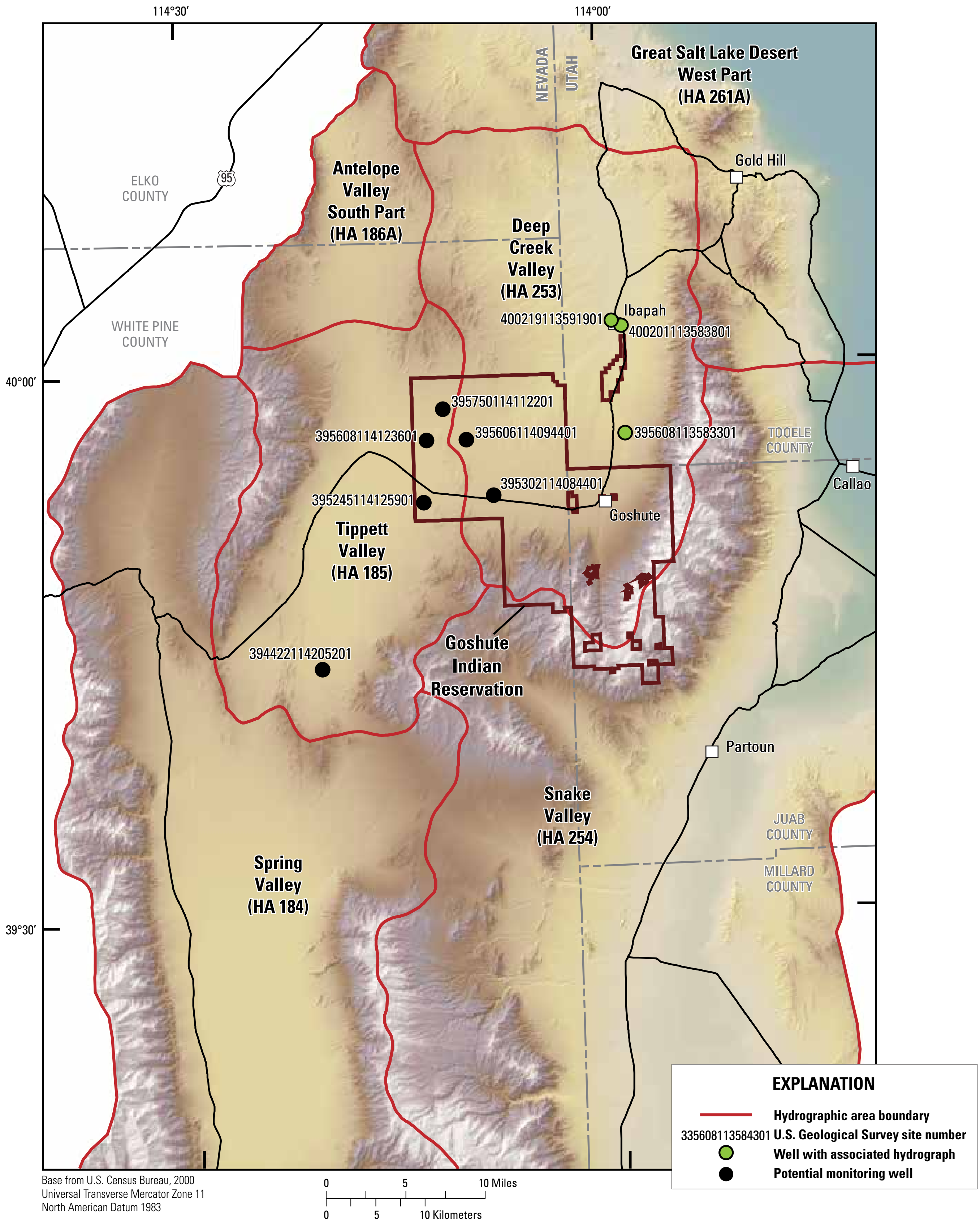

Figure 25. Location of current and potential monitoring wells in Tippet and Deep Creek Valleys, Utah and Nevada.

During this study, six existing wells were identified that could serve as an effective water-level monitoring network (fig. 25) to provide early indications of future Spring Valley groundwater withdrawals having far-reaching effects. One of these wells (USGS site number 394422114205201) is an active, 350-ft deep Bureau of Land Management (BLM) stock well in southern Tippett Valley less than 4 mi north of Spring Valley, with water levels in the basin-fill aquifer more than 290 ft below land surface. The other five wells (USGS site numbers 395608114123601, 395245114125901, 395750114112201, 395606114094401, and 395302114084401), known locally as the Goshute Windmill wells, are abandoned stock wells with undetermined depths and water levels in basin-fill deposits between 210 and 400 ft below land surface. These wells are located along the low drainage divide between Tippett Valley and Deep Creek Valley between about 16 and 33 mi northeast of Spring Valley. If a minimum of three of the Goshute Windmill wells were reconditioned and paired with the BLM stock well in southern Tippett Valley, they could provide an optimal network of wells to monitor for long-term groundwater withdrawal related water-level declines. Additionally, groundwater sampled from the Goshute Windmill wells for the same suite of environmental tracers collected at the 10 other sites during this study would fill an important spatial data gap that currently exists between sites sampled during May 2012.

Summary

The water resources of Deep Creek Valley in eastern Nevada and western Utah, were assessed during 2012–13 with an emphasis on better understanding the groundwater flow system and groundwater budget. Water resources are limited in Deep Creek Valley, with few perennial streams entering the valley from the mountains and one stream, Deep Creek, flowing through and exiting the valley. The limited surface-water resources generally are used for agriculture, leaving groundwater to supplement irrigation and as the predominant water source for most other uses. The principal source of groundwater in Deep Creek Valley is from the unconsolidated basin-fill deposits, in which conditions are generally unconfined near the mountain front and confined in the lower-altitude parts of the valley. Productive aquifers are also present in bedrock. The consolidated-rock and basin-fill aquifers are hydraulically connected in many areas, with much of the recharge occurring in the consolidated-rock mountain blocks and most of the discharge occurring within the lower-altitude basin-fill deposits. Additionally, there is a possible interbasin hydraulic connection from Spring Valley through Tippett Valley to Deep Creek Valley. Although the majority of wells in the study area are completed within the basin-fill deposits, bedrock wells within the study area are increasingly being developed.

Groundwater recharge in the study area occurs mostly from the infiltration of precipitation at high altitudes. Much of this recharge occurs as snowmelt. Additional, but limited recharge occurs from the infiltration of runoff from precipitation near the mountain front, infiltration along stream channels, and possible subsurface inflow from adjacent hydrographic areas. Average annual recharge to the Deep Creek Valley hydrographic area was estimated to be between 19,000 and 29,000 acre-ft. Groundwater moves from areas of recharge to springs and streams in the mountains, and to evapotranspiration areas, springs, streams, and wells in the basins. Discharge may also occur as subsurface groundwater outflow to adjacent hydrographic areas. Average annual discharge from the Deep Creek Valley hydrographic area was estimated to be between 21,000 and 22,000 acre-ft, with the largest portion of discharge occurring as evapotranspiration. The groundwater budget shows an imbalance because of uncertainty in the individual budget estimates, the largest of which is subsurface inflow from Tippett Valley.

Groundwater samples were collected from 10 sites and analyzed for major ions, nutrients, and selected trace metals to characterize general geochemistry and patterns of water quality. Dissolved-solids concentrations ranged from 126 to 475 mg/L. With the exception of the slightly elevated arsenic in a stock well located west of northern Deep Creek Valley in Antelope Valley, none of the sampled wells or springs contained concentrations of dissolved solutes that exceeded Environmental Protection Agency maximum contaminant levels or secondary standards.

The groundwater samples were also analyzed for a suite of environmental tracers that included the stable isotopes of oxygen, hydrogen, and carbon, dissolved noble gases, and the radioactive isotopes of carbon (^{14}C) and hydrogen (tritium, ^{3}H) to investigate sources of recharge, groundwater flow paths, ages, and travel times. Stable-isotope ratios of oxygen and deuterium and noble-gas recharge temperature data indicate that most groundwater in the basin-fill aquifer of eastern Deep Creek Valley, and along Deep Creek, consists of recharge that originated as infiltration of precipitation within the Deep Creek Range or of runoff along the mountain front. Concentrations of ^{3}H between 1.6 and 10 TU indicate the presence of modern (less than 60 years old) groundwater in samples from 3 of the 10 sample sites. Apparent $^{3}H/^{3}He$ ages, calculated for these three sites, ranged from 7 to 29 years. Adjusted minimum radiocarbon ages of pre-modern water samples ranged from 1,000 to 8,000 years with the ages of at least four of the samples being more than 3,000 years.

Potentiometric contours indicate groundwater flow from western Spring Creek Flat, and from the directions of Antelope and Tippett Valleys into Deep Creek Valley. Noble-gas recharge temperatures for groundwater discharging in Deep Creek Valley, however, are substantially cooler than most water sampled from upgradient areas to the southwest and west, indicating that recharge to the basin-fill aquifer is predominantly from the Deep Creek Range. Furthermore, flow-weighted average groundwater age estimates at several flowing wells in Deep Creek Valley indicate subsurface travel times between 1,000 and 8,000 years. If significant

fractions of Deep Creek basin-fill groundwater were sourced from the basins to the west and southwest, where recharge rates are lower and flow paths are longer, travel times would likely be longer. Although information gathered during this study is insufficient to conclude whether or not groundwater does travel along these extended and possibly interbasin flow paths, dissolved sulfate and chloride data indicate that a small fraction of the lower-altitude, northern Deep Creek Valley discharge may be sourced from these areas.

Water-level altitude (potentiometric) contours and groundwater ages indicate the potential for a long flow path between northern Spring and Deep Creek Valleys through Tippett Valley. Potentiometric contours indicate a groundwater divide near the central axis of northern Spring Valley and a continuous gradient in a northeastward direction from this point to northernmost Deep Creek Valley. Groundwater ages generally increase in a downgradient direction from pre-modern to late Holocene to early Holocene. A hydraulic connection between northern Spring Valley, Tippett Valley, and Deep Creek Valley appears likely, and potential regional effects resulting from future groundwater withdrawals in northern Spring Valley warrant ongoing monitoring of groundwater levels in appropriate areas.

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Appendix 1. Data Tables

Table A1-1. Selected attributes of wells and water levels measured in wells used in constructing the water-level surface map for Deep Creek Valley and adjacent areas, Utah and Nevada.

[Data, except where indicated, is from the U.S. Geological Survey's National Water Information System (NWIS) database. Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83). Abbreviations: USGS, U.S. Geological Survey; —, no information; >, greater than; <, less than]

USGS site number	USGS site name	Latitude (decimal degrees)	Longitude (decimal degrees)	Well depth (feet)	Altitude of land surface (feet)	Measurement date	Depth to water below or above (-) land surface (feet)	Water-level altitude (feet)
400003114025901	193 N25 E70 34ACCC1	40.00094444	-114.04975000	90	5,453	3/6/2012	15.12	5,438
395815114043401	193 N24 E70 09CBAB1	39.97072222	-114.07605556	—	5,541	3/6/2012	-4.78	5,546
400831113591001	(C- 8-19) 9dad- 1	40.14361111	-113.98341667	306	5,134	3/11/2010	-8.25	5,142
400329113593601	(C- 9-19) 9caa- 1	40.05816667	-113.99319444	540	5,233	3/8/2012	42.84	5,190
395608113583301	(C-10-19)22cdc- 1	39.93549420	-113.97666740	506	5,726	3/14/2012	122.39	5,604
395620113584001	(C-10-19)22cbd- 1	39.93869444	-113.97869444	300	5,689	8/22/2002	87	5,602
395633113584301	(C-10-19)22bcd- 1	39.94243856	-113.97944540	130	5,672	3/30/2000	66.84	5,605
400201113583801	(C- 9-19)22bac- 1	40.03344444	-113.97738889	100	5,315	3/5/2012	13.30	5,302
400219113591901	(C- 9-19)16dca- 2	40.03863889	-113.98863889	—	5,272	3/6/2012	7.05	5,265
—	[1](C- 9-19)16ddd- 1	40.03750000	-113.98555556	—	5,282	9/15/1995	10	5,272
400210113585701	(C- 9-19)22bbb- 1	40.03604846	-113.98333600	—	5,289	3/26/1990	10.5	5,279
400135113591001	(C- 9-19)21dac- 1	40.02725000	-113.98647222	50	5,300	3/8/2012	1.17	5,299
400125113592501	(C- 9-19)28aba- 1	40.02088889	-113.99050000	70	5,320	3/8/2012	10.50	5,310
400024113582701	(C- 9-19)34baa- 1	40.00650000	-113.97627778	147	5,380	8/1/1960	-19	5,399
400056113582701	(C- 9-19)27caa- 1	40.01394444	-113.97425000	45	5,367	3/5/2012	17.41	5,350
395929113584901	(C-10-19) 3bbd- 1	39.99147222	-113.98033333	—	5,469	3/5/2012	12.65	5,456
395952113584101	(C- 9-19)34cac- 1	39.99780556	-113.97800000	—	5,440	3/5/2012	4.87	5,435
400543114145101	186A N26 E68 25ACCC1 USBLM BLACK HILLS WELL	40.09533333	-114.26408333	592	5,886	11/6/1985	497	5,389
400110114154101	185 N25 E68 26B 1 USBLM	40.01697222	-114.26925000	448	5,906	3/7/2012	378.03	5,528
395106114150601	185 N23 E68 23DDBB1 USBLM	39.85169444	-114.24986111	—	5,787	3/5/2012	>268.5, Dry	>5,518
395433114173701	185 N23 E68 04B 1 USBLM	39.90400000	-114.30108333	175	5,673	10/22/1969	115.18	5,558
395245114125901	185 N23 E69 07DCBD1 GOSHUTE RESERVATION	39.87899720	-114.21811390	—	5,804	7/19/2005	275.09	5,529
395607114095201	193 N24 E69 27 1 G NE	39.93525000	-114.16488889	260	5,784	3/5/2012	254.45	5,530
395302114084401	193 N23 E69 11 1 G E END	39.88388889	-114.13516667	<500	5,965	2/7/2012	397.20	5,568
395608114123601	185 N24 E69 19DDCD1 Goshute Reservation	39.93572222	-114.21280556	252	5,741	3/5/2012	212.23	5,529
394422114205201	185 N22 E67 36DBAC1 USBLM	39.72950000	-114.34405556	350	5,840	3/7/2012	306.18	5,534
392703114230501	184 N18 E67 01CCAA1	39.44966386	-114.39056130	42	5,587	3/6/2012	40.35	5,547
393211114320701	184 N19 E66 11B 1	39.53080556	-114.51727778	400	5,698	2/14/2012	37.61	5,661
393059114221501	184 N19 E67 13AAAC1	39.51799688	-114.37806200	53	5,614	3/6/2012	48.56	5,565
393442114231801	184 N20 E67 26ABBD1 USBLM	39.57632936	-114.40028630	130	5,705	3/6/2012	122.35	5,583
393729114265401	184 N20 E67 09AABD1 DOUTRE	39.62016667	-114.43555560	280	5,787	3/6/2012	178.25	5,609
393617114295001	184 N20 E66 13BADA1	39.60475000	-114.49727778	296	5,775	2/15/2012	124.03	5,651
393544114320301	184 N20 E66 15DCBC1	39.59547800	-114.53410300	420	5,960	9/4/2009	315	5,645
—	Willow Spring Well[1]	39.69700300	-114.50609400	35	5,990	5/5/2010	8	5,982
394407114320401	184 N21 E66 04B 1	39.72715975	-114.52446060	—	6,070	7/16/1964	21.4	6,049
392700114300901	184A N18 E66 01CCAA1 KEEGAN SPRING	39.45002778	-114.50252778	22	5,625	5/8/2010	1.5	5,624
395750114113201	185 N24 E69 17 1 G NW	39.96388889	-114.19222220	340	5,850	3/5/2012	319.62	5,530
400303113590301	(C- 9-19)16aaa- 1	40.05083333	-113.98402778	—	5,236	3/8/2012	7.07	5,229
395814113594101	(C-10-19) 9caa- 1	39.97052778	-113.99477778	—	5,551	3/6/2012	10.65	5,540
395935113584501	(C- 9-19)34ccd- 1	39.99444444	-113.97950000	80	5,453	3/6/2012	4.89	5,448
395937114031301	193 N25 E70 34DCD1	39.99372222	-114.05363889	—	5,479	3/6/2012	3.47	5,476

[1]Not in the U.S. Geological Survey's National Water Information System (NWIS) database.

Table A1-2. Selected attributes of groundwater sites sampled during the spring of 2012 in Deep Creek Valley and adjacent areas, Utah and Nevada.

[Site ID: see figure 17 for locations. Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83). Abbreviations: USGS, U.S. Geological Survey; UBFAU, upper basin-fill aquifer unit; LBFAU, lower basin-fill aquifer unit; LCAU, lower carbonate aquifer unit; NCCU, non-carbonate confining unit; —, no information]

Site ID	USGS site number	USGS site name	Site type	Latitude (decimal degrees)	Longitude (decimal degrees)	Well depth (feet)	Depth to top and bottom of openings (feet)	Altitude of land surface (feet)	Hydrogeo-logic unit	Sample date
1	400543114145101	186A N26 E68 25ACCC1	Stock well	40.09533333	-114.26408333	590	550–580	5,988	LBFAU[1]	5/16/2012
2	400216113591701	(C- 9-19)16ddc- 2	Domestic well	40.03786100	-113.98797200	216	186–216	5,280	UBFAU[1]	5/18/2012
3	394554114003501	(C-12-19)20aac-S1	Spring	39.76494400	-114.00969400	—	—	7,560	NCCU	5/14/2012
4	395112114014501	(C-11-19)19cad-S1	Spring	39.85333300	-114.02919400	—	—	6,335	LCAU	5/14/2012
5	400024113582701	(C- 9-19)34baa- 1	Unused flowing well	40.00650000	-113.97627778	147	112–147	5,380	UBFAU[1]	5/17/2012
6	394149114302201	184 N21 E66 15DBDD1	Spring	39.69672200	-114.50697200	—	—	5,980	UBFAU	5/15/2012
7	395815114043401	193 N24 E70 09CBAB1	Unused flowing well	39.97072200	-114.07605600	—	—	5,540	UBFAU[1]	5/17/2012
8	395237114222501	185 N23 E67 14BA 2	Spring	39.87694400	-114.37352800	—	—	6,240	LCAU	5/16/2012
9	400831113591001	(C- 8-19) 9dad- 1	Unused flowing well	40.14363300	-113.98342800	306	—	5,134	UBFAU[1]	5/17/2012
10	394422114205201	185 N22 E67 36DBAC1	Stock well	39.73938400	-114.34862100	350	295–350	5,775	UBFAU[1]	5/15/2012

[1]Not determined from driller's log but rather assumed based on location or depth.

Appendix 2. Groundwater Budget Uncertainty

The groundwater budget values in table 4 are estimates based on models, assumptions, correlations, or regressions that are fundamentally derived from representative measurements often made at only a few points in time. As a result, these estimates have an associated uncertainty that is difficult to quantify but important to acknowledge. An attempt has been made in the current study to quantify these uncertainties. Each of the groundwater budget components in table 4, along with the total recharge and discharge, are presented with an uncertainty value (expressed as a percentage of the component value) that is intended to convey the possible range that the actual value might vary. Often, budget components are derived from several variables, and the uncertainty reported in table 4 is that of the variable with the largest contribution to the total uncertainty. The list that follows briefly explains how each of the uncertainties was derived.

Recharge Components

(1) The uncertainty in the infiltration of precipitation is based on a sensitivity analysis of the Basin Characterization Model in-place recharge by the authors of the model and documented in Flint and others (2011) and Masbruch and others (2011).
(2) The uncertainty in the subsurface inflow estimates is unknown because the methods by which these estimates were made were not documented in the reports from which these estimates were taken.

Discharge Components

(1) The uncertainty in the evapotranspiration of groundwater is based on an error analysis that assumes that the range of evapotranspiration rates are either (i) the high and low measured rates for different vegetation types reported in recent literature (Nichols, 2000; Berger and others, 2001; Reiner and others, 2002; Cooper and others, 2006; Moreo and others, 2007; Welch and others, 2007), or (ii) plus or minus 1 standard deviation for rates reported as consumptive use of irrigated crops (Utah State University, 1994). This analysis also takes into account that the uncertainty in precipitation that is subtracted from total evapotranspiration to obtain the estimate of groundwater evapotranspiration is plus or minus 15 percent (Jeton and others, 2006).
(2) The uncertainty in the discharge to mountain springs and stream base flow is the assumed measurement error.
(3) The uncertainty in the well discharge estimate is unknown.
(4) The uncertainty in the subsurface outflow estimate is unknown because the methods by which these estimates were made were not documented in the reports from which these estimates were taken.

Gardner and Masbruch—**Hydrogeologic and Geochemical Characterization of Groundwater Resources in Deep Creek Valley and Adjacent Areas, Juab and Tooele Counties, Utah, and Elko and White Pine Counties, Nevada**—SIR 2015–5097

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