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This report is based on work by the U.S. Geological Survey, in collaboration with the Desert Research Institute, and the State of Utah

A Report to Congress

Water Resources of the Basin and Range Carbonate-Rock Aquifer System, White Pine County, Nevada, and Adjacent Areas in Nevada and Utah—Draft Report

Open-File Report 2007–1156

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

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DIRK KEMPTHORNE, Secretary

U.S. Geological Survey
Mark D. Myers, Director

U.S. Geological Survey, Reston, Virginia: 2007

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Foreword

Water demands from the lower Colorado River system are increasing with the rapidly growing population of the southwestern United States. To decrease dependence on this over allocated surface-water resource and to help provide for the projected increase in population and associated water supply in the Las Vegas area, water purveyors in southern Nevada have proposed to utilize the ground-water resources of rural basins in eastern and central Nevada. Municipal, land management, and regulatory agencies have expressed concerns about potential impacts from increased ground-water pumping on local and regional water quantity and quality, with particular concern on water-rights issues and on the future availability of water to support springflow and native vegetation. Before concerns on potential impacts to pumping can be addressed, municipal and regulatory agencies have recognized the need for additional information and improved understanding of geologic features and hydrologic processes that control the rate and direction of ground-water flow in eastern and central Nevada.

In response to concerns about water availability and limited hydrogeologic information, Federal legislation (Section 131 of the Lincoln County Conservation, Recreation, and Development Act of 2004: PL 108-424) was enacted in December 2004 that directs the Secretary of the Interior, through the U.S. Geological Survey (USGS), the Desert Research Institute (DRI), and a designee from the State of Utah, to conduct a water-resources study of the basin-fill and carbonate-rock aquifers in White Pine County, Nevada, and smaller areas of adjacent counties in Nevada and Utah. The primary objectives of the Basin and Range Carbonate-rock aquifer system (BARCAS) study are to evaluate: (1) the extent, thickness, and hydrologic properties of aquifers, (2) the volume and quality of water stored in aquifers, (3) subsurface geologic structures controlling ground-water flow, (4) ground-water flow directions and gradients, and (5) distributions and rates of recharge and ground-water discharge. Geologic, hydrologic, and supplemental geochemical information will be integrated to determine basin and regional ground-water budgets.

Results of the study will be summarized in a USGS Scientific Investigations Report (SIR), to be prepared in cooperation with DRI and the State of Utah, and submitted to Congress by December 2007. The BARCAS study SIR is supported by USGS and DRI reports that document, in greater detail than the summary SIR, important components of and estimates made in support of the BARCAS study. These reports are varied in scope and include documentation of basic data including spring location and irrigated acreage, and interpretive studies of ground-water flow, recharge, evapotranspiration, and geology.

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[In pocket]

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Figures—Continued

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Conversion Factors and Datums

Conversion Factors

Multiply	By	To obtain
acre	4046.856	square meter (m^2)
acre	0.4047	hectare (ha)
calorie	4.184	joule (J)
calories per second per square foot	45.045	watt per square meter (W/m^2)
foot (ft)	0.3048	meter (m)
foot per second (ft/s)	0.3048	meter per second (m/s)
inch (in.)	25.4	millimeter (mm)
mile (mi)	1.609	kilometer (km)
mile per hour (mph)	0.44704	meter per second (m/s)
ounce, avoirdupois (oz)	28.35	gram (g)
square mile (mi^2)	2.58999	square kilometer (km^2)

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

$$°C=(°F-32)/1.8.$$

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

$$°F=(1.8\times°C)+32.$$

Datums

Vertical coordinate information is referenced to the North American Vertical Datum of 1929 (NVGD of 1929).

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

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

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Water Resources of the Basin and Range Carbonate-Rock Aquifer System, White Pine County, Nevada and Adjacent Areas in Nevada and Utah—DRAFT REPORT

By Alan H. Welch and Daniel J. Bright, Editors

Summary of Major Findings

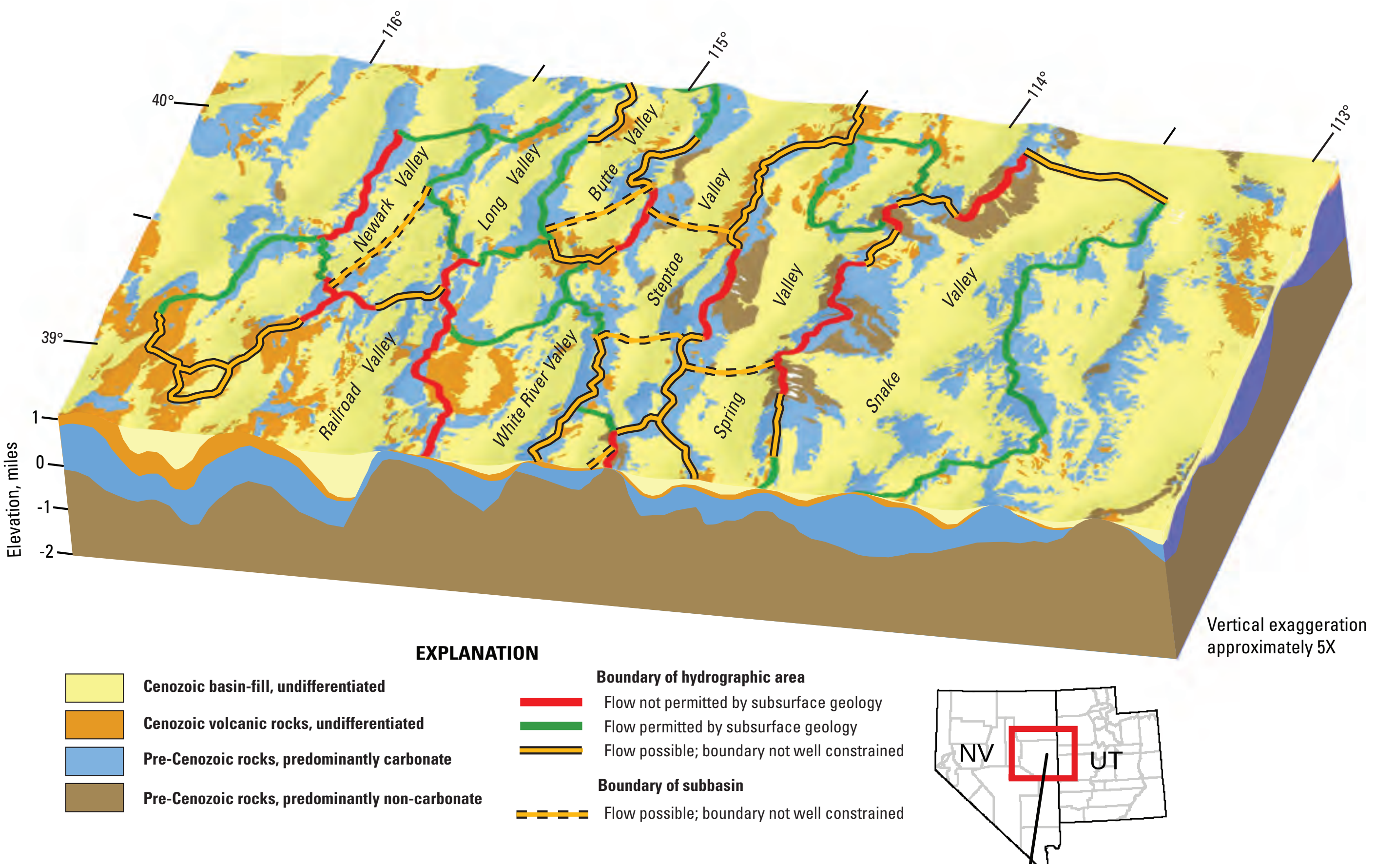
This report summarizes results of a water-resources study for White Pine County, Nevada, and adjacent areas in east-central Nevada and western Utah. The Basin and Range carbonate-rock aquifer system (BARCAS) study was initiated in December 2004 through Federal legislation (Section 131 of the Lincoln County Conservation, Recreation, and Development Act of 2004) directing the Secretary of the Interior to complete a water-resources study through the U.S. Geological Survey, Desert Research Institute, and State of Utah. The study was designed as a regional water-resource assessment, with particular emphasis on summarizing the hydrogeologic framework and hydrologic processes that influence ground-water resources.

The study area includes 13 hydrographic areas that cover most of White Pine County; in this report however, results for the northern and central parts of Little Smoky Valley were combined and presented as one hydrographic area. Hydrographic areas are the basic geographic units used by the State of Nevada and Utah and local agencies for water-resource planning and management, and are commonly defined on the basis of surface-water drainage areas. Hydrographic areas were further divided into subbasins that are separated by areas where bedrock is at or near the land surface. Subbasins represent subdivisions used in this study for estimating recharge, discharge, and water budget. Hydrographic areas represent the subdivision used for reporting summed and tabulated subbasin estimates.

Aquifer System

Most ground water in the study area flows through three types of aquifers—a shallow basin-fill aquifer, a deeper volcanic-rock aquifer, and an underlying carbonate-rock aquifer that forms the base of the ground-water flow system. Relatively impermeable basement rocks underlie the carbonate-rock aquifer throughout most of the study area. The basin-fill aquifer underlies every valley and is the primary source of ground water for the area. The thickness of basin fill beneath most valleys is about 6,600 feet; however, in Steptoe and Lake Valleys, it exceeds 13,000 feet. The volcanic-rock aquifer is thickest beneath the western and southern parts of the study area, extending laterally beneath the basin-fill aquifer and multiple hydrographic areas. Although some springs issue from volcanic rocks, these aquifers are not utilized as a significant source of water supply in the study area. Fractured, permeable carbonate rocks are regionally extensive, form many of the mountain ranges, and underlie the basin-fill and volcanic-rock aquifers throughout much of the study area. Ground water in the carbonate-rock aquifer discharges at perennial-flowing valley-floor springs and, because of the lateral continuity and relative high permeability of the carbonate rocks, most ground-water flow between adjacent valleys occurs through this aquifer. Although not a primary source of water supply in the study area, some ground water is pumped from the carbonate-rock aquifer for various uses.

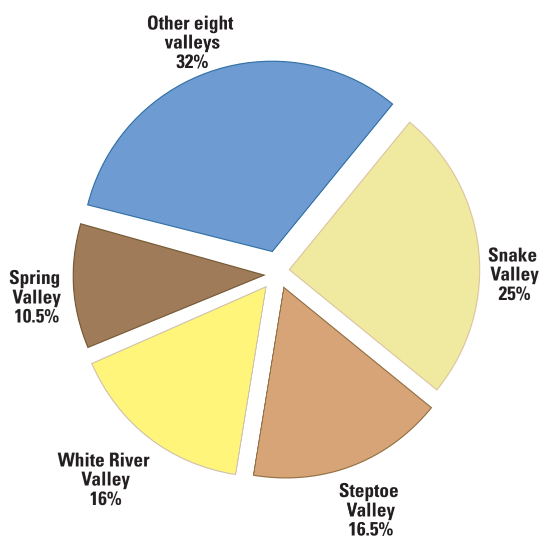
The distribution of aquifers and units of low permeability along hydrographic area boundaries is a primary control on ground-water flow between hydrographic areas. Ground-water flow across some hydrographic area boundaries may be negligible where carbonate or volcanic rocks are absent, or if the aggregate permeability of aquifers beneath a hydrographic area boundary is relatively low.



Perspective view of the primary aquifer systems.

Aquifer Storage

For equivalent volumes of aquifer material, the capacity of the basin-fill aquifer to store water is significantly greater than that of the carbonate-rock aquifer. For example, permeable deposits in the upper 100 ft of saturated basin-fill aquifer beneath valley floors throughout the study area store about 36 million acre-ft of water. In contrast, the upper 100 ft of saturated carbonate-rock aquifer beneath valley floors stores about 30,000 acre-ft of water, or about 3-orders of magnitude less than the basin-fill aquifer. About 75 percent of the water stored in the upper 100 ft of basin-fill and carbonate-rock aquifers occur in the four largest hydrographic areas—Snake, Steptoe, White River, and Spring Valleys. The evaluation of aquifer storage assumes ground-water is pumped from equivalent volumes of basin-fill and carbonate-rock aquifers, but does not consider the potential impacts to changes in storage caused by ground-water extractions, such as declining water levels in wells, decreasing spring discharge, diminished water quality, or loss of native vegetation.



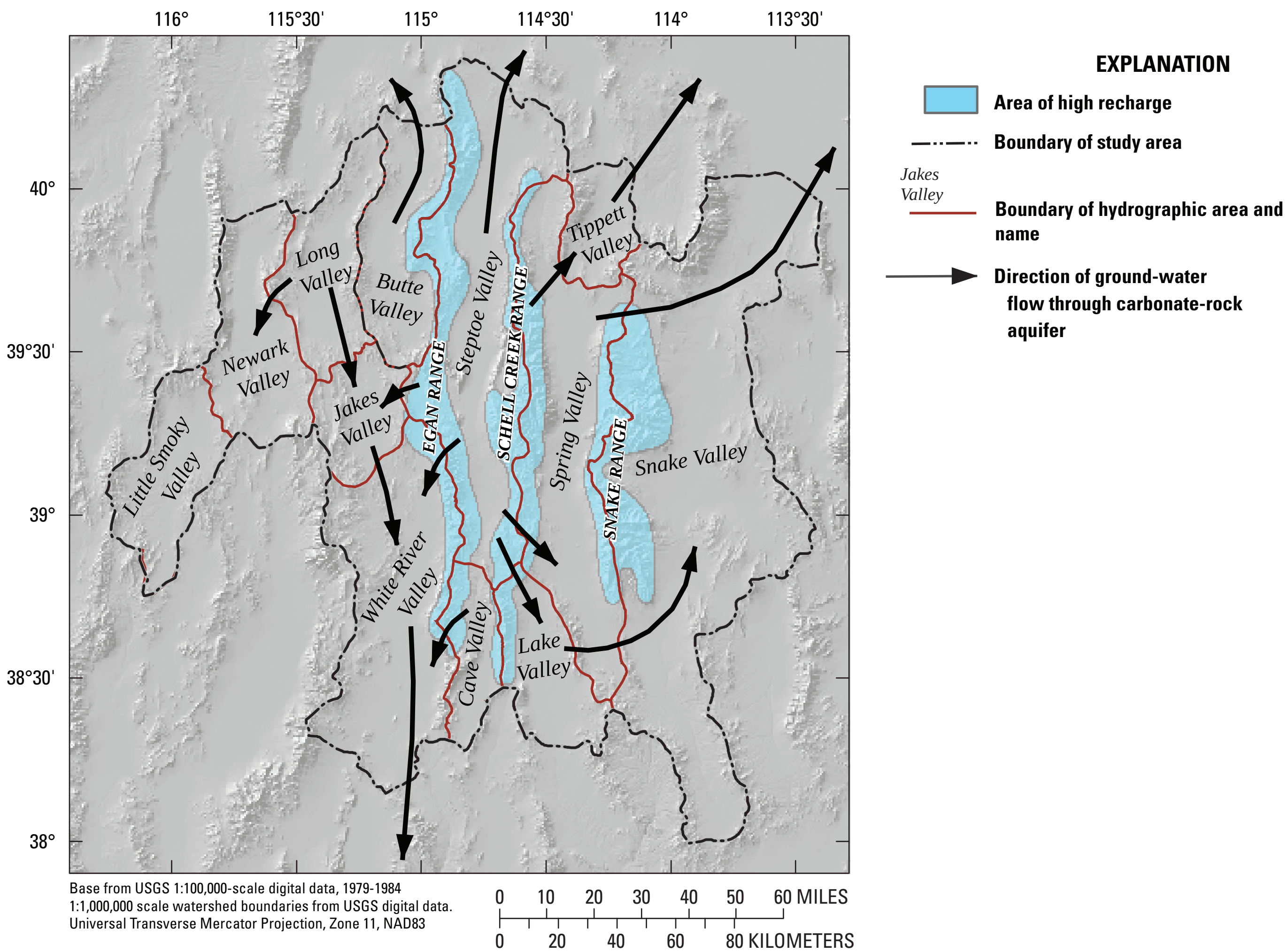
Percentage of water stored in basin-fill and carbonate aquifers.

Aquifer Water Quality

The inorganic chemical quality of ground water generally is acceptable for human consumption. No discernable patterns of poor water quality have been found except for chloride concentrations in some ground water in northern Snake Valley that exceed secondary drinking-water standards. Only a small number of analyses of anthropogenic organic compounds in ground water are available. No exceedances of drinking-water standards have been reported.

Regional Ground-Water Flow

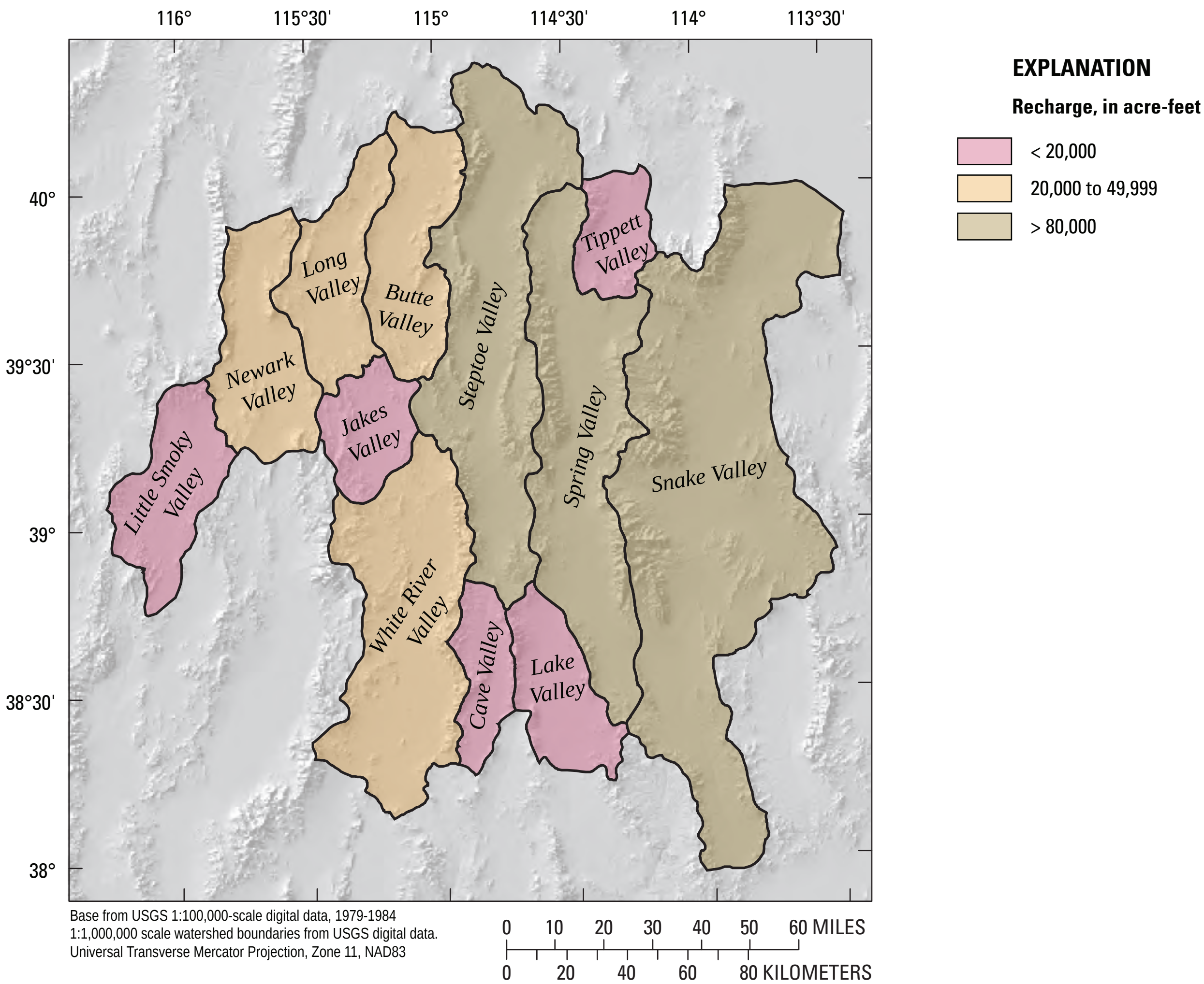
Carbonate rocks form much of the Egan, Schell Creek, and Snake Ranges, and the relatively high precipitation and recharge in these mountain ranges are the source for regional ground-water flow in the carbonate-rock aquifer. The Egan Range is the primary source area for northward ground-water flow through Butte Valley, and southward flow through Long, Jakes, and White River Valleys, where ground water exits the study area and flows toward the Colorado River. The Egan and Schell Creek Ranges are the primary source areas for ground water in Steptoe Valley, where the highest water-level altitudes in the basin fill are found in the study area. Ground water flows northward through Steptoe Valley and southeastward through southern Steptoe, Lake, Spring, and Snake Valleys. The Schell Creek and Snake Ranges are the primary source areas for northeastward ground-water flow through northern Spring, Tippett, and Snake Valleys. Ground water exits the study area from Snake and Tippett Valleys and flows northeastward toward a terminal discharge area in the Great Salt Lake Desert.



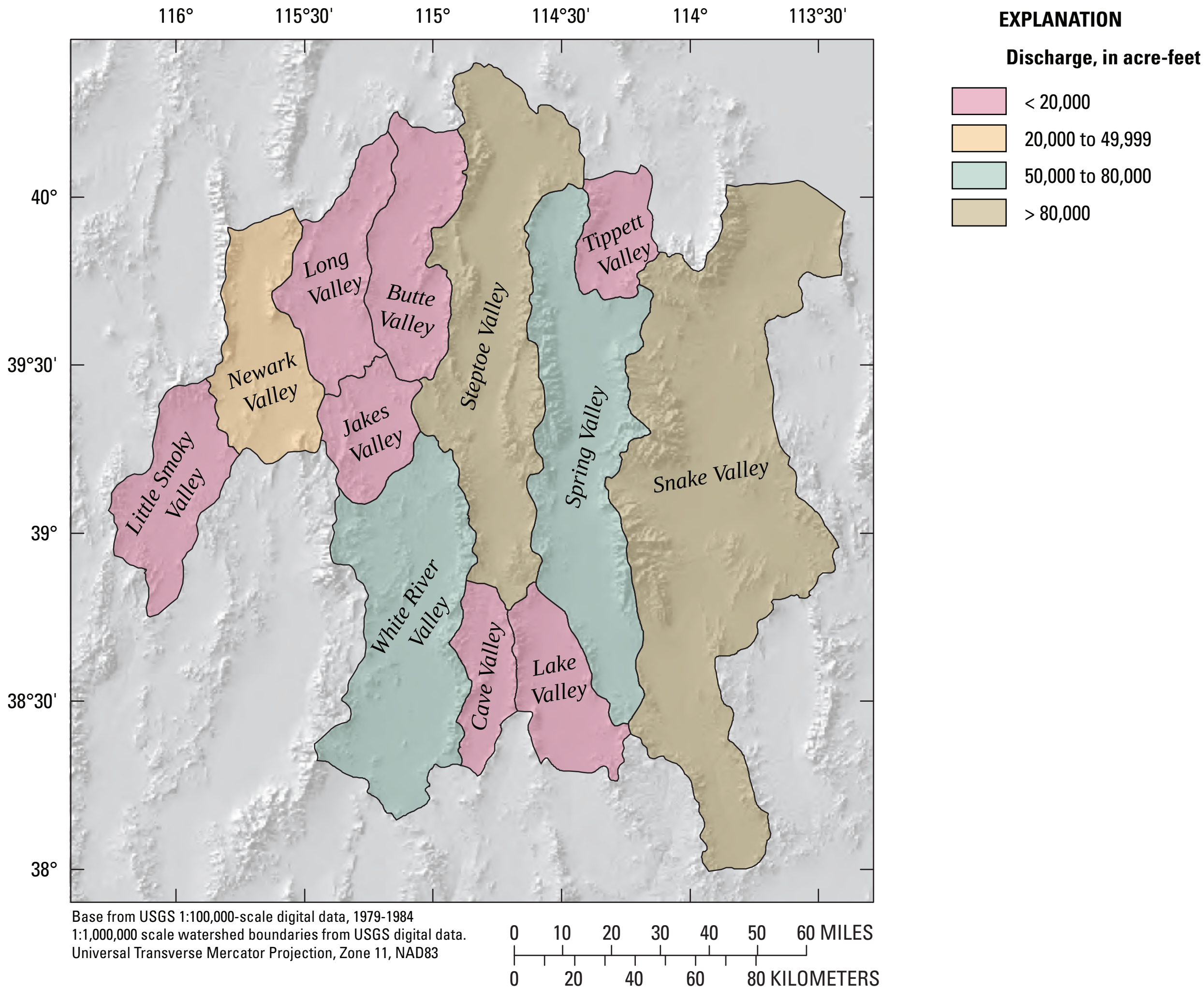
Regional ground-water flow through the carbonate-rock aquifer.

Basin Recharge and Discharge

The larger valleys in the study area, such as Steptoe, Snake, Spring, and White River Valleys, have the highest average annual ground-water recharge and discharge. The highest annual recharge occurs in Steptoe Valley (about 150,000 acre-ft) and Snake Valley (about 110,000 acre-ft). Estimated annual recharge for Steptoe Valleys is about 20,000 acre-ft higher than the highest previous estimate for this valley. The highest annual discharge occurs in Snake Valley (about 130,000 acre-ft) and Steptoe Valley (about 100,000 acre-ft). Estimated annual discharge for Snake Valley is significantly higher (about 45,000 acre-ft) than the highest previous estimate for this valley; estimated annual discharge for Steptoe Valley is within the range of previous estimates.



Average annual recharge to the ground-water system.

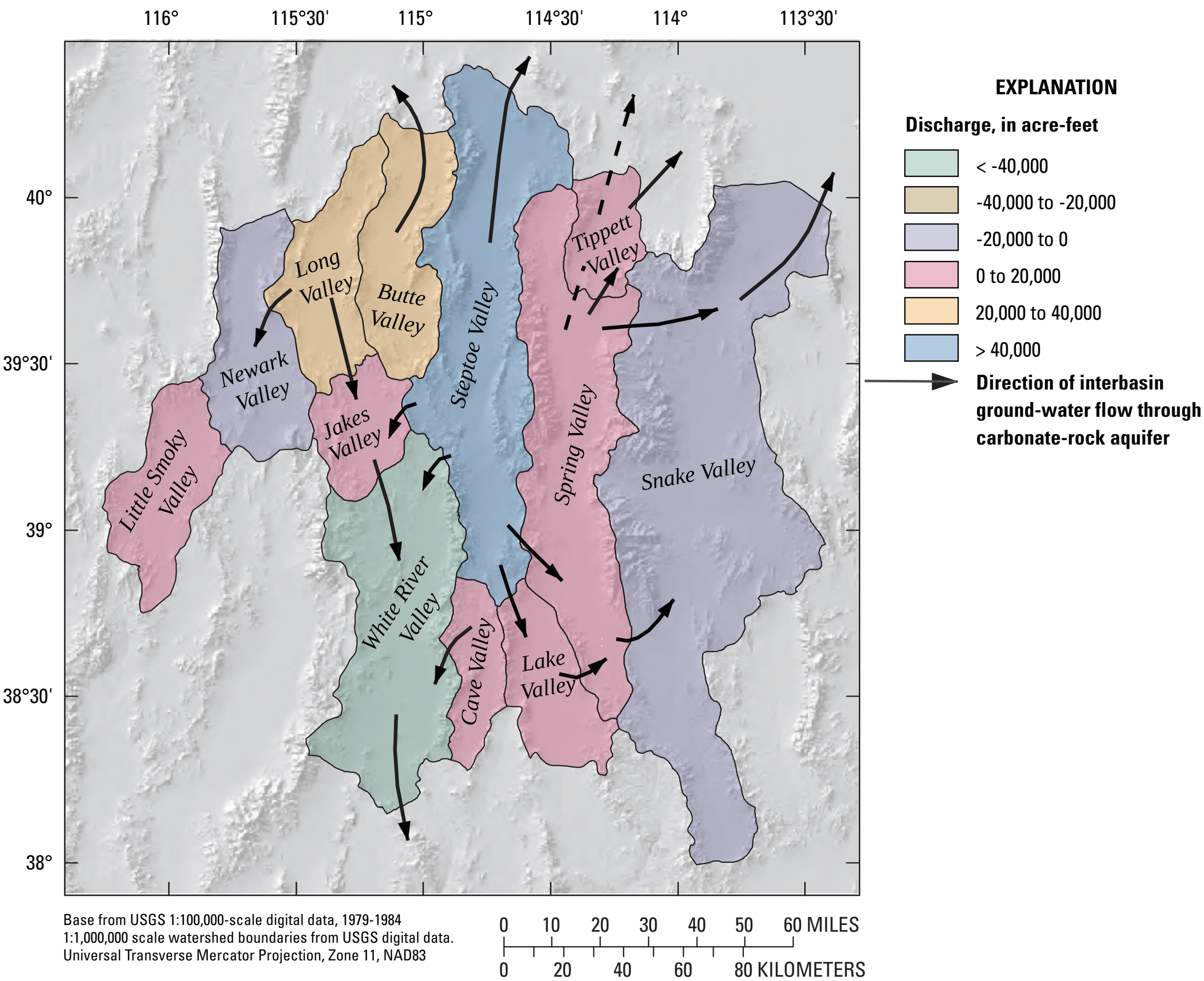


Average annual discharge from the ground-water system by evaporation and transpiration of vegetation.

Interbasin Ground-Water Flow

Differences in basin recharge and discharge provide a surplus or deficit of water that is balanced by ground-water flow entering or exiting a valley as inter-basin ground-water flow. For one-half of the hyrographic areas (6 of 12), recharge exceeds pre-development discharge by 10,000 acre-ft or more on an average annual basis. The high recharge in Steptoe Valley annually exceeds pre-development discharge by more than 50,000 acre-ft. The surplus of water in Steptoe Valley is the source of inter-basin ground-water flow to multiple valleys—to the north where ground water exits the study area, to the southeast toward Lake and southern Spring Valley, and to the west toward Jakes and northern White River Valleys. The latter two flow paths from southern and western Steptoe Valley have not been proposed in previous investigations. Flow from Steptoe Valley to other valleys suggest that parts of southern Steptoe and Lake Valleys may be included in the Colorado or Great Salt Lake Desert regional flow systems.

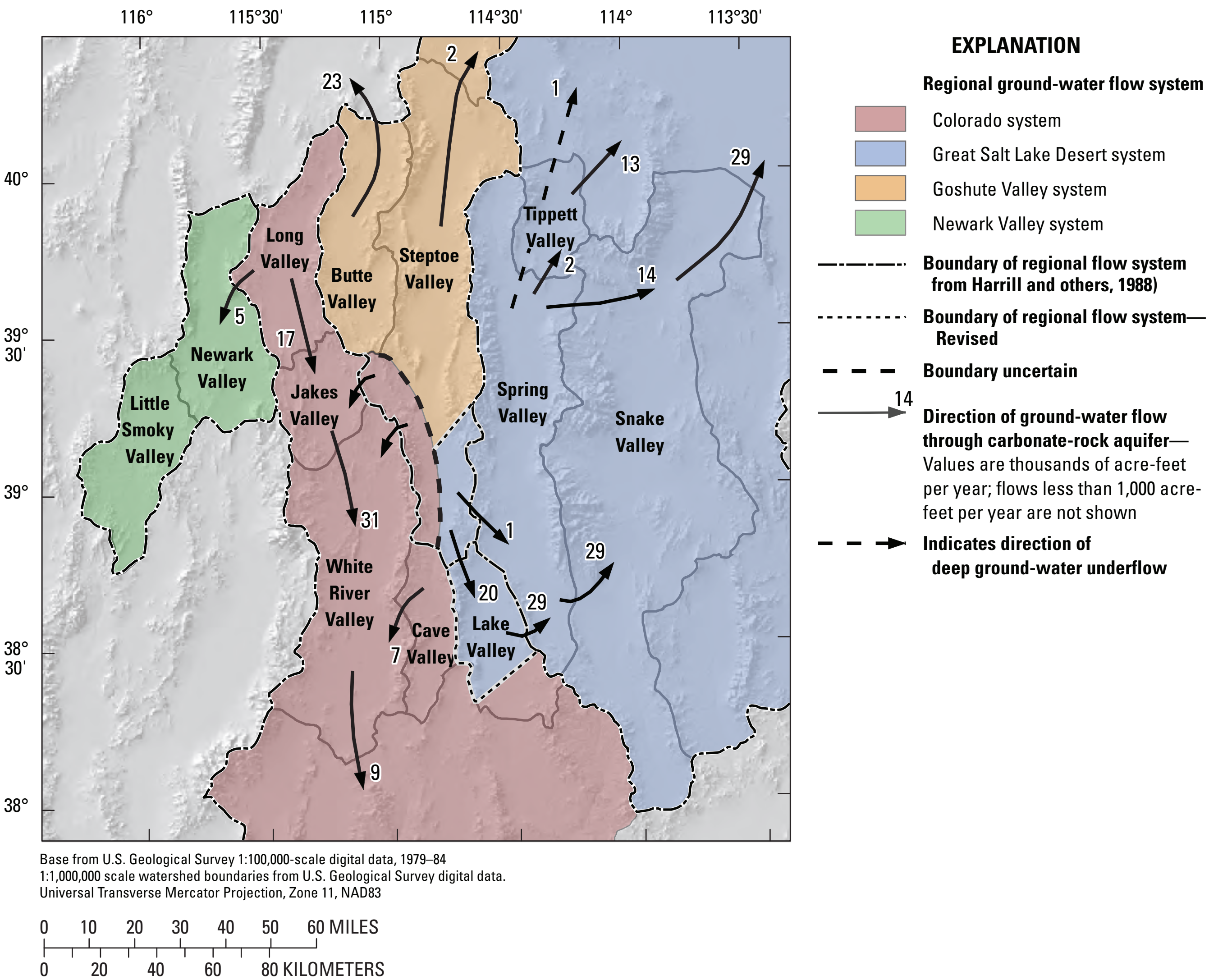
In contrast to Steptoe Valley, pre-development discharge annually exceeds the relatively low annual recharge in White River Valley by more than 40,000 acre-ft, indicating that water lost from evapotranspiration on the valley floor must be supported, in part, by subsurface inflow from adjacent valleys. The deficit of ground water in Whiter River Valley is balanced by inter-basin flow from Steptoe Valley to the northeast, Jakes Valley to the north, and Cave Valley to the east. Estimates of the magnitude of inter-basin flow differ from previous estimates for some hydrographic area boundaries. The largest differences are for estimated outflow from southern Steptoe Valley, where previous investigations proposed zero outflow, and for southern Spring Valley. The estimated 29,000 acre-ft/yr of ground-water flow from southern Spring Valley to Snake Valley is about twice the highest previous estimate.



Average annual recharge minus average annual ground-water discharge, and areas of inter-basin ground-water flow.

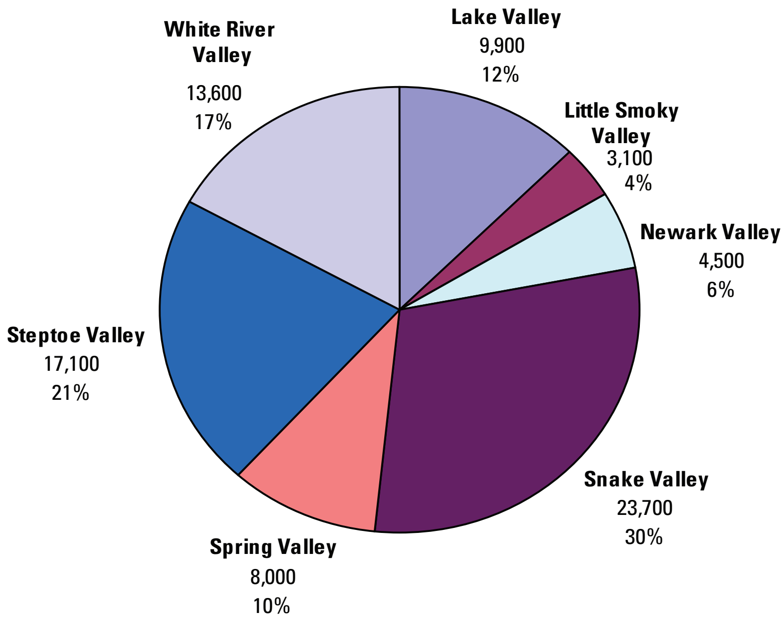
Regional Recharge and Discharge

For the entire study area, average annual recharge equals 530,000 acre-ft, and average annual ground-water discharge equals 440,000 acre-ft under pre-development conditions. The difference between recharge and discharge indicates that about 90,000 acre-ft of ground water exits the study area annually by subsurface outflow. Most ground-water flow likely exits the study area through Snake (29,000 acre-ft/yr), Butte (23,000 acre-ft/yr), Tippett (13,000 acre-ft/yr), and White River Valleys (9,000 acre-ft/yr).



Regional ground-water flow through the Colorado, Great Salt Lake Desert, and other regional flow systems.

The net amount of water removed by ground-water pumping was estimated to evaluate the significance of water withdrawals to ground-water discharge under pre-development conditions. Net ground-water pumpage represents the amount of water pumped from wells or diverted from regional springs minus excess water returned from mining, irrigation applications, or public supply that infiltrated and recharged the ground-water system. Of the 127,000 acre-ft of ground-water use in 2005, about 46,000 acre-ft returned to the aquifer system. The remaining 80,000 acre-ft nearly equals the estimated quantity of ground-water outflow from the study area (about 90,000 acre-ft/yr). On a regional scale, this condition suggests that long-term ground-water withdrawals equal to those estimated for 2005 could potentially capture much of the estimated average annual volume of ground water exiting the study area under pre-development conditions. These withdrawals also could, in some combination, reduce other discharge components such as inter-basin flow, spring discharge, or discharge by vegetation, or increase subsurface recharge from adjacent basins. However, actual reductions in ground-water outflow would be controlled by a number of factors, particularly, the spatial distribution of ground-water withdrawals, and the volume of ground-water removed from storage. For example, reductions in outflow would be less likely in Butte or Tippett Valleys where net pumpage was zero in 2005. Reductions in outflow would be more likely in sub-basins or hydrographic areas where net pumpage is nearly equal or greater than the estimated outflow, such as in Snake Valley where net pumpage was 24,000 acre-ft in 2005 and average annual ground-water outflow was estimated at 29,000 acre-ft. However, for ground-water withdrawals from the basin-fill aquifer, the relatively large volume of water stored in this aquifer likely will mitigate current or near-future reductions in the volume of ground-water outflow or other pre-development discharge components. Water-level measurements, water-use records, and data on pre-development discharge indicate that ground-water pumpage currently (2005) has not significantly altered evapotranspiration rates, the distribution of native vegetation, or regional springflow in the study area.



Percent distribution and volume of net regional ground-water pumpage from hydrographic areas

Although some uncertainty exists on estimated differences between annual recharge and pre-development discharge, a prevalence of hydrographic areas where recharge exceeds discharge and a significant quantity of subsurface outflow from the entire study area (90,000 acre-ft/yr) are not unexpected. Recharge estimates were model-derived; the accuracy of these estimates depends on the accuracy with which a number of hydrologic, atmospheric, and soil parameters were estimated. Estimates of pre-development discharge were derived through field measurements and, as a result of a more direct method of measurement, the uncertainty of estimated pre-development discharge is likely less than the uncertainty of estimated recharge. Future studies may reduce uncertainties of estimated recharge and discharge by evaluating a regional ground-water flow system bounded by ground-water divides, such as the Colorado or Great Salt Lake Desert regional flow systems. Evaluating entire regional flow systems provides the constraint that ground-water inflow and outflow across the study area boundary is minimal; therefore, cumulative recharge and pre-development discharge must balance for hydrographic areas within the regional flow system.

Introduction

A study initiated by Federal legislation (Lincoln County Conservation, Recreation, and Development Act of 2004; PL 108-424) directed the Secretary of the Interior, through the U.S. Geological Survey (USGS), the Desert Research Institute (DRI), and a designee from the State of Utah, to evaluate the basin-fill and carbonate-rock aquifers in White Pine County, Nevada, and adjacent areas in Nevada and Utah. This report is a draft that will be revised in response to a public comment period as required by the legislation. A final report will be transmitted to Congress no later than December 1, 2007. The congressionally mandated study is termed the Basin and Range carbonate-rock aquifer system (BARCAS) study, and was completed in cooperation with the Bureau of Land Management.

White Pine County in east-central Nevada (fig. 1) is a sparsely populated area, with less than 10,000 residents in 2006, most of which reside in and adjacent to the city of Ely, Nevada, the county seat (2001). The county contains typical basin and range topography—north-south trending valleys and mountains that range in altitude from 5,000 to 7,000 ft above sea level for valley floors, and above 10,000 ft for most mountain ranges. The mountain ranges are the principal source of recharge to four regional ground-water flow systems (fig. 1). Most ground water in White Pine County is used for irrigation and mining purposes. Lesser amounts of ground water are used for municipal and domestic purposes in and adjacent to the city of Ely.

Water purveyors in southern Nevada have proposed to use ground-water resources in White Pine County to help meet water needs associated with the projected increases in the population of Clark County in southern Nevada. As populations in southern Nevada and elsewhere in the Southwest continue to increase, the reliance on water from the Colorado River Basin becomes increasingly important, and the prospects of obtaining additional allotments of water from the Colorado River system, stipulated in the Colorado River Compact of 1922, are confounded by the legal and socio-political issues derived from the competition for those scarce resources by the seven Compact States. Alternatively, ground-water resources in rural basins north of Clark County, including basins in White Pine County, have been targeted as potential sources of imported water supply. Municipal and regulatory agencies have expressed concerns about potential impacts on water quantity and quality, water rights, sensitive wildlife habitats, and other beneficial uses from the proposed activities. As a first step in assessing potential impacts from ground-water development, agencies and stakeholders have recognized the need for additional hydrologic data and an improved understanding of hydrogeologic processes that control the rate and direction of ground-water flow in eastern and central Nevada.

Purpose and Scope

The purpose of this report is to summarize hydrogeologic factors affecting the occurrence and movement of ground water in the aquifer system of the study area. Ground-water resources were evaluated by focusing on the following hydrogeologic characteristics: (1) the extent, thickness, and hydrologic properties of aquifers, (2) subsurface geologic structures controlling ground-water flow, (3) ground-water flow directions and gradients, (4) the volume and quality of water stored in aquifers, and (5) the distribution and rates of recharge and discharge. Moreover, geologic, hydrologic, and supplemental geochemical information were evaluated to determine ground-water budgets in the study area. Finally, hydrogeologic characteristics were compiled and integrated to develop a three-dimensional hydrogeologic framework and conceptual understanding of ground-water flow in the study area.

Description of Study Area

The study area encompasses about 13,500 mi^2 and covers about 80 percent of White Pine County, Nevada, and smaller areas of adjacent counties in Nevada and Utah (fig. 1). White Pine County lies within the eastern part of the Great Basin—a unique internally drained physiographic feature of the Western United States. Basin and Range topography—north-south trending valleys and adjacent mountain ranges—dominates the region.

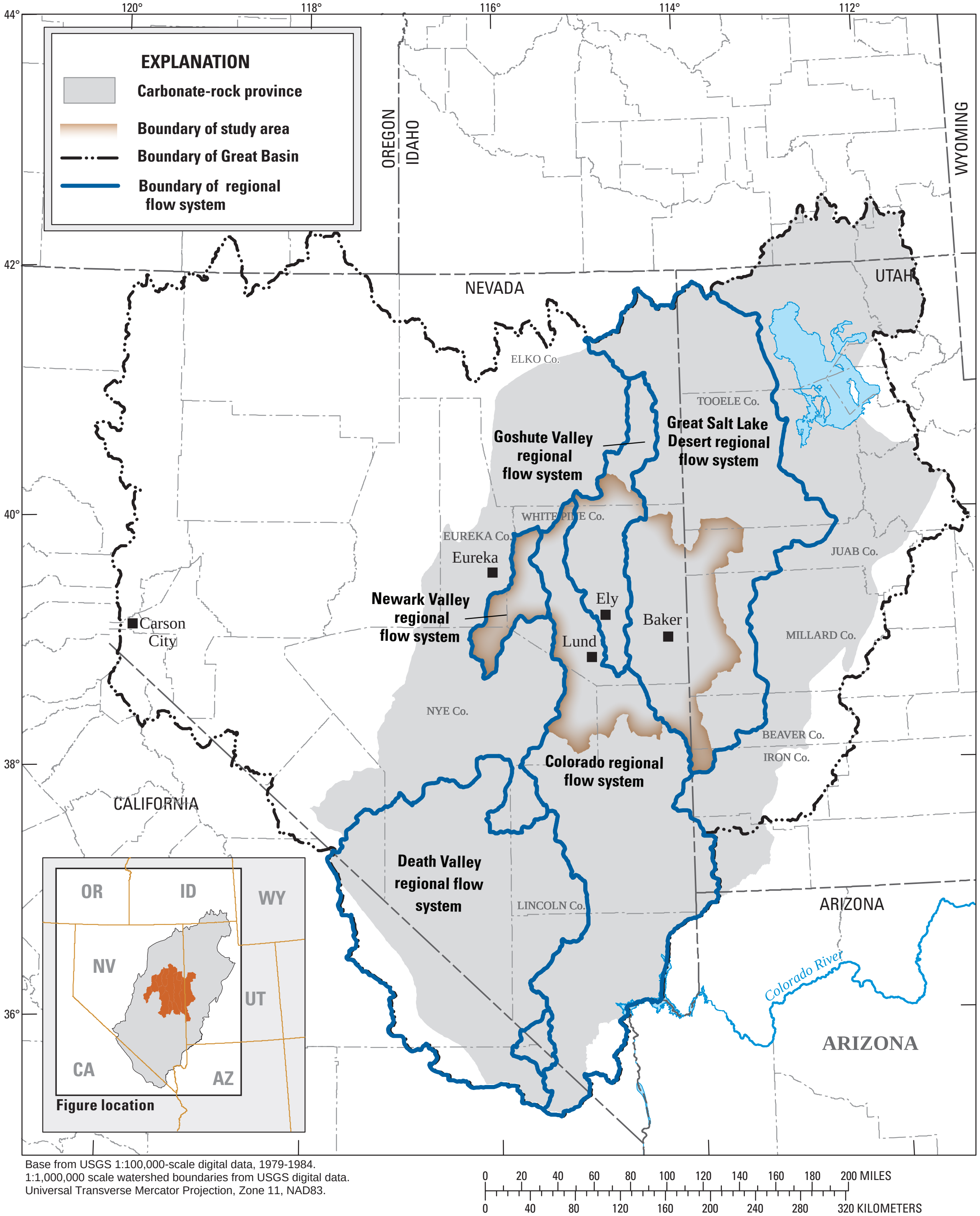


Figure 1. Carbonate-rock province, Basin and Range carbonate-rock aquifer system, and associated regional ground-water flow systems, Nevada and Utah.

The study area encompasses 13 hydrographic areas (HAs)[1] (pl. 4; fig. 2). Past studies have combined HAs to delineate intermediate or regional ground-water flow systems, primarily based on the direction of interbasin ground-water flow in the underlying carbonate-rock aquifer and the location of terminal discharge areas (Harrill and Prudic, 1998). Although most boundaries between HAs coincide with actual topographic basin divides, some are arbitrary divisions that have no basis in topography. In this report, HAs also are referred to as basins, and ground-water flow within these areas is referred to as intrabasin ground-water flow. Moreover, HAs were further divided into subbasins that are separated by areas where pre-Cenozoic rocks are at or near the land surface. For purposes of this report, areas that separate subbasins are referred to as intrabasin divides. Subbasins represent subdivisions used in this study for estimating recharge, discharge, and water budget. HAs represent the subdivision used for reporting summed and tabulated subbasin estimates. HAs within this report refer to formal HAs of Harrill and others (1988) with two exceptions: (1) ‘Little Smoky Valley’ refers to both HAs 155A and 155B, which are the northern and central parts of Harrill’s description of Little Smoky Valley, respectively, and (2) ‘Butte Valley’ refers only to HA 178B, which is the southern part of Harrill and others’ description of Butte Valley. For most figures and tables in this report, water-budget components were estimated for the northern and central parts of Little Smoky Valley, but were combined and reported as one value.

Precipitation in the study area provides recharge to four regional ground-water flow systems—the Newark Valley, Goshute Valley, Great Salt Lake Desert, and Colorado regional flow systems (fig. 1)—that headwater in White Pine County. These regional flow systems are characterized by flow across HA boundaries and discharge as warm springs. All these regional flow systems extend to areas outside of White Pine County. The Newark Valley and Goshute Valley flow systems are relatively small, internally drained flow systems. The remaining two flow systems terminate in areas hundreds of miles from their source area in White Pine County. The Great Salt Lake Desert regional flow system terminates at the Great Salt Lake, with intermediate discharge at Fish Springs in Juab County, Utah. The Colorado regional flow system terminates at Lake Mead and the Colorado River, with a principal intermediate discharge area at Muddy River Springs in Lincoln County, Nevada. In addition to these and other perennial valley-floor springs, numerous high-altitude ephemeral and perennial springs are found in the study area. Many of these perennial and ephemeral springs support native vegetation; some springs support protected aquatic or wildlife species, such as the Pahrump poolfish (*Empetrichthys latos*) in southeastern Spring Valley, and the White River spinedace (*Lepidomeda albivallis)* in White River Valley near Lund.

Regional ground-water flow in the study area primarily is controlled by carbonate rocks. Much of the carbonate-rock aquifer is fractured and these fractured rocks, where continuous, form a regional flow system that receives recharge in high-altitude mountain ranges in the study area where these rocks are exposed. Some water flows from the carbonate-rock aquifer into basin-fill aquifers. This regional discharge sustains many of the larger, perennial low-altitude springs in the study area. The basin-fill aquifers that overlie the carbonate-rock aquifer typically are more than 1,000-ft-thick deposits of volcanic rocks, gravel, sand, silt and clay (Harrill and Prudic, 1998). Basin-fill deposits locally can exceed 10,000 ft. Gravel and sand deposits yield water readily to wells and are the aquifers most commonly developed for agricultural, domestic, and municipal supply.

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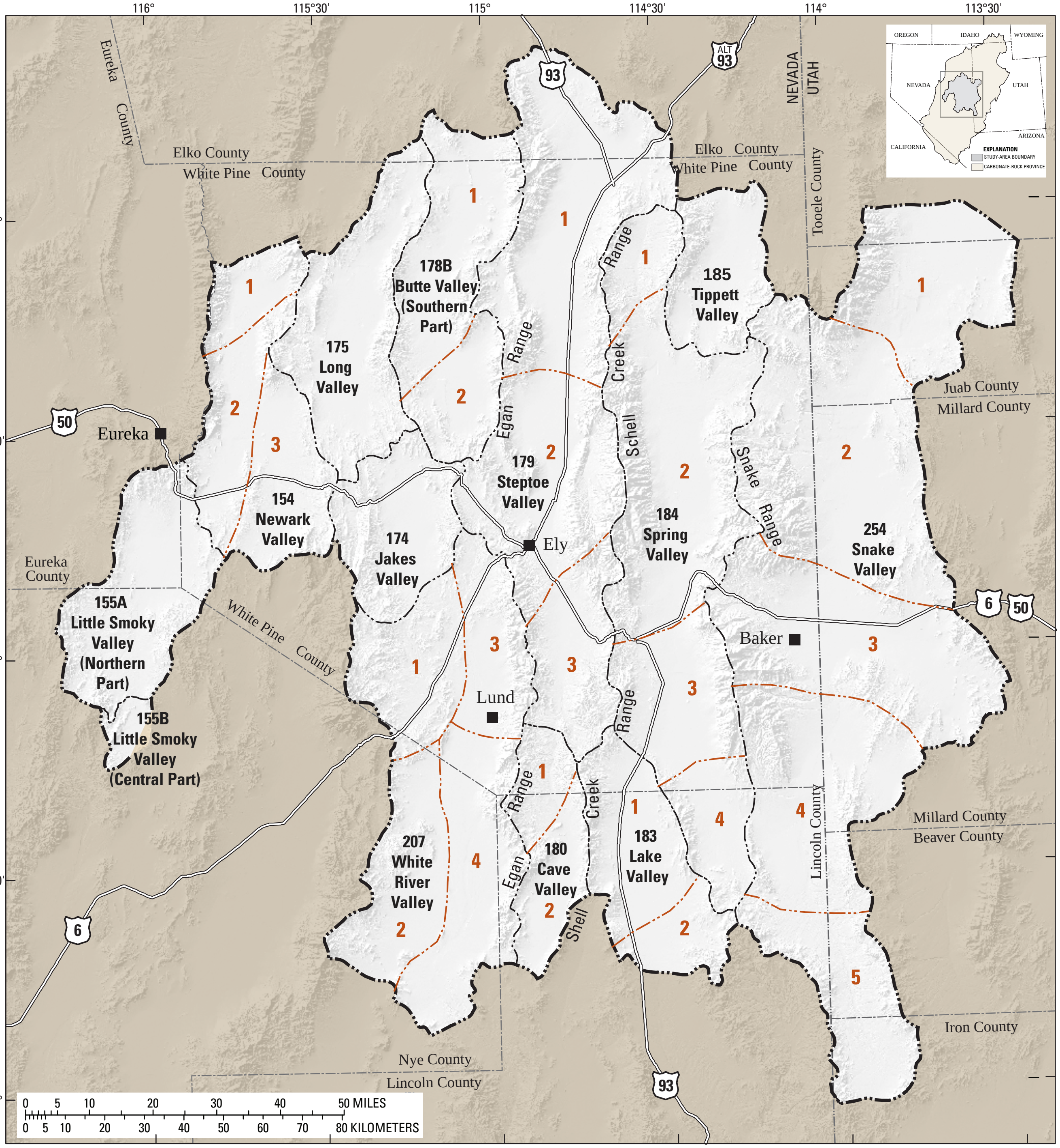
[1]Formal hydrographic areas in Nevada were delineated systematically by the U.S. Geological Survey and Nevada Division of Water Resources in the late 1960s (Cardinalli and others, 1968; Rush, 1968) for scientific and administrative purposes. The official hydrographic-area names, numbers, and geographic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



Base from U.S. Geological Survey digital data 1:100,000, 1978–89.
Universal Transverse Mercator Projection, Zone 11, NAD83.
Shaded relief base from 1:250,000-scale Digital Elevation Model;
sun illumination from northwest at 30 degrees above horizon.

EXPLANATION

Boundary of study area

174 Jakes Valley — Boundary of hydrographic area and name and number

4 — Boundary of subarea and No.

Figure 2. Hydrographic areas and subbasins, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Hydrogeologic Framework

By Donald S. Sweetkind, Lari A. Knochenmus, David A. Ponce, Alan R. Wallace, Daniel S. Scheirer, Janet T. Watt, and Russell W. Plume, U.S. Geological Survey

A variety of geologic and geophysical approaches have been used to improve the understanding of the hydrogeologic framework of the study area. Geologic map units and structures were compiled from digital versions of the Nevada (Stewart and Carlson, 1978; Raines and others, 2003) and Utah (Hintze and others, 2000) 1:500,000-scale State geologic maps. Drilling records and accompanying geophysical logs for oil and gas wells and exploration wells also were evaluated to understand down-hole lithology and stratigraphy, to estimate relative permeabilities of different rock types, and to augment the regional hydrogeologic framework. The new geologic data were integrated with existing information to develop a generalized hydrogeologic map (pl. 1) that portrays the configuration of rock units in the study area. The hydrogeologic map combines geologic units into hydrogeologic units (HGUs)—groupings of rock units that have reasonably similar hydrologic properties. HGU designations were based on lithologic, stratigraphic, and structural characteristics from published descriptions and from data collected during field mapping as part of the study. A generalized stratigraphic column and corresponding hydrogeologic unit designation for the study area are shown in figure 3.

Surface geophysical techniques were applied to take advantage of characteristic density, magnetic, electrical, and acoustic properties of different rocks in a way that provides additional insight into the subsurface geology. Detailed gravity, magnetic, electromagnetic, and seismic geophysical data (fig. 4) are used to identify faults, subsurface structure, and the interconnectivity of adjacent basins. The results of most of the geophysical investigations conducted for the BARCAS study are presented in Watt and Ponce (2007).

Geologic History

The geologic history of the eastern part of Nevada is preserved in rocks and geologic structures that span more than a billion years, ranging from Precambrian sedimentary rocks to widespread Quaternary alluvial deposits and active faults. The geologic framework that has resulted from the geologic events during this time profoundly affects ground-water flow. Thus, any water-resource assessment of the area must take into account the complex geologic history and consider the distribution of the diverse rocks types and geologic environments.

The geologic evolution of the study area since the end of Precambrian time may be subdivided into three general phases (Levy and Christie-Blick, 1989): (1) a late Precambrian to middle Paleozoic interval when dominantly marine sediments were deposited along a passive continental margin; (2) late Devonian to Eocene crustal shortening, compressive deformation, and changes in sedimentation patterns related to the accretion of exotic terrains along the western continental margin in western Nevada; and (3) middle to late Cenozoic extension, faulting, volcanism, and continental sedimentation. Within the context of this three-phase evolution, numerous tectonic events and accompanying changes in sedimentation patterns and igneous activity have occurred throughout geologic time in the study area (fig. 5). These tectonic-induced events have been summarized by De Courten (2003).

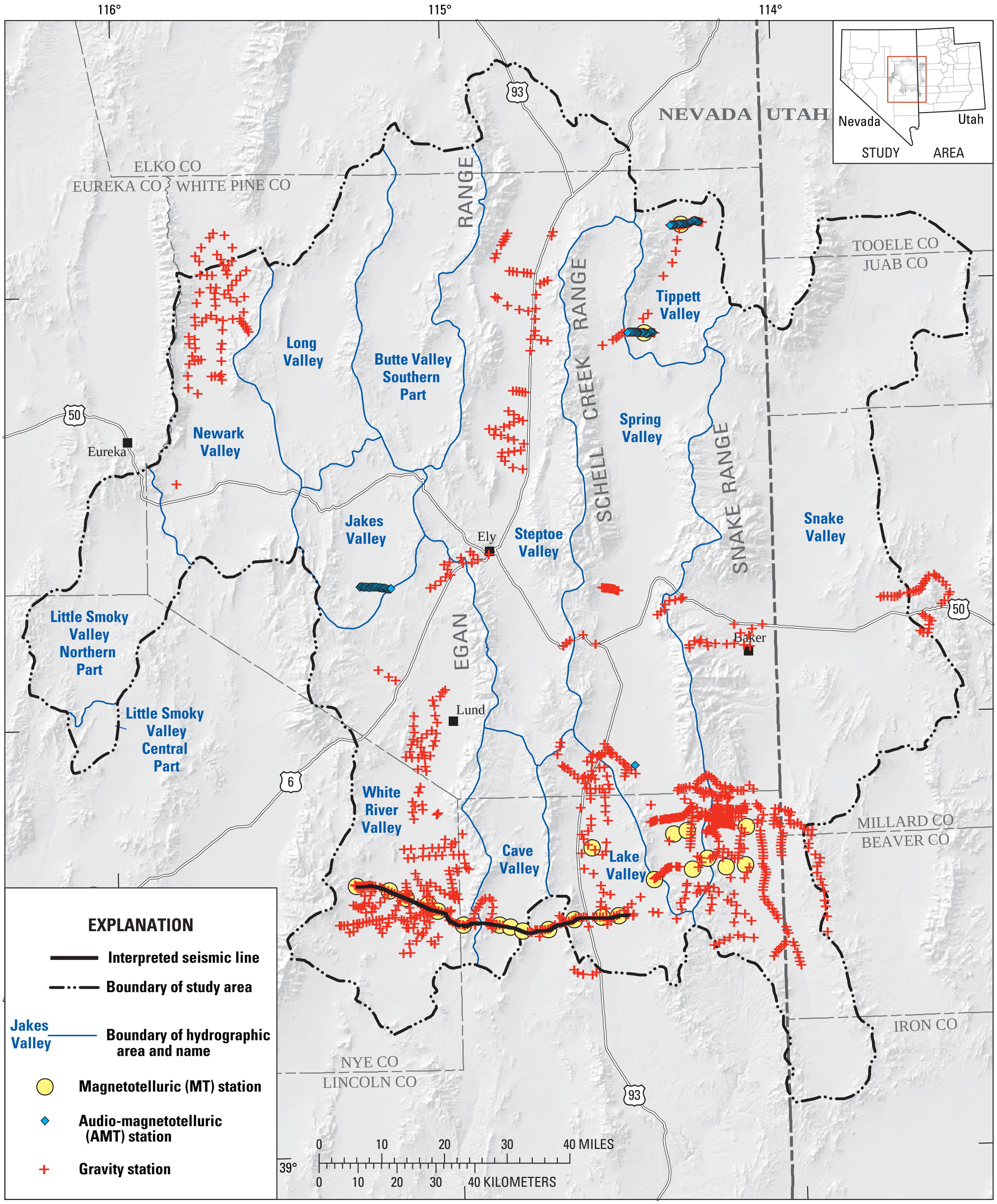
During the first phase of geologic events, from late Precambrian until middle Devonian time, the rocks in east-central Nevada were deposited in shallow to deep marine water in a stable continental shelf environment similar to that of modern-day Atlantic and Gulf Coast margins of the United States (Blakely, 1997; available at http://vishnu.glg.nau.edu/rcb/paleogeogwus.html). The stable shelf environment produced thick, extensive carbonate, quartzite, and shale deposits. Most of the widespread units of the older Paleozoic limestone and dolomite rocks (hydrogeologic unit LCU, pl. 1) were deposited in shallow water on a broad, stable continental shelf, known as a “carbonate platform” (Jackson, 1997; Cook and Corboy, 2004). To the west of the study area, correlative rocks were deposited on a gently sloping submarine surface that gradually deepened seaward of the platform (fig. 6). Sedimentary rocks accumulated to thicknesses of about 30,000 ft during this time (Kellogg, 1963; Stewart and Poole, 1974) and form the vast majority of the consolidated rocks exposed in the study area. These limestone and dolomite rocks have long been recognized as an aquifer in the Great Basin (Winograd and Thordarson, 1975; Bedinger and others, 1989; Dettinger and others, 1995; Harrill and Prudic, 1998). These rocks typically consist of an upper Precambrian and Lower Cambrian section of quartzite and shale, a Middle Cambrian to Lower Ordovician limestone section, a distinctive Middle Ordovician quartzite, and an Upper Ordovician to Middle Devonian dolomite section (Kellogg, 1963; Poole and others, 1992) (fig. 3).

Eon	Era	Period	Epoch
Phanerozoic	Cenozoic	Quaternary	Holocene
			Pleistocene
		Tertiary	Pliocene
			Miocene
			Oligocene
			Eocene
			Paleocene
	Mesozoic	Cretaceous	
		Jurassic	
		Triassic	
	Paleozoic	Permian	
		Pennsylvanian	
		Mississippian	
		Devonian	
		Silurian	
		Ordovician	
		Cambrian	
Proterozoic Eon*			
Archean Eon*			

* The Archean and Proterozoic Eons are major subdivisions of Precambrian time.

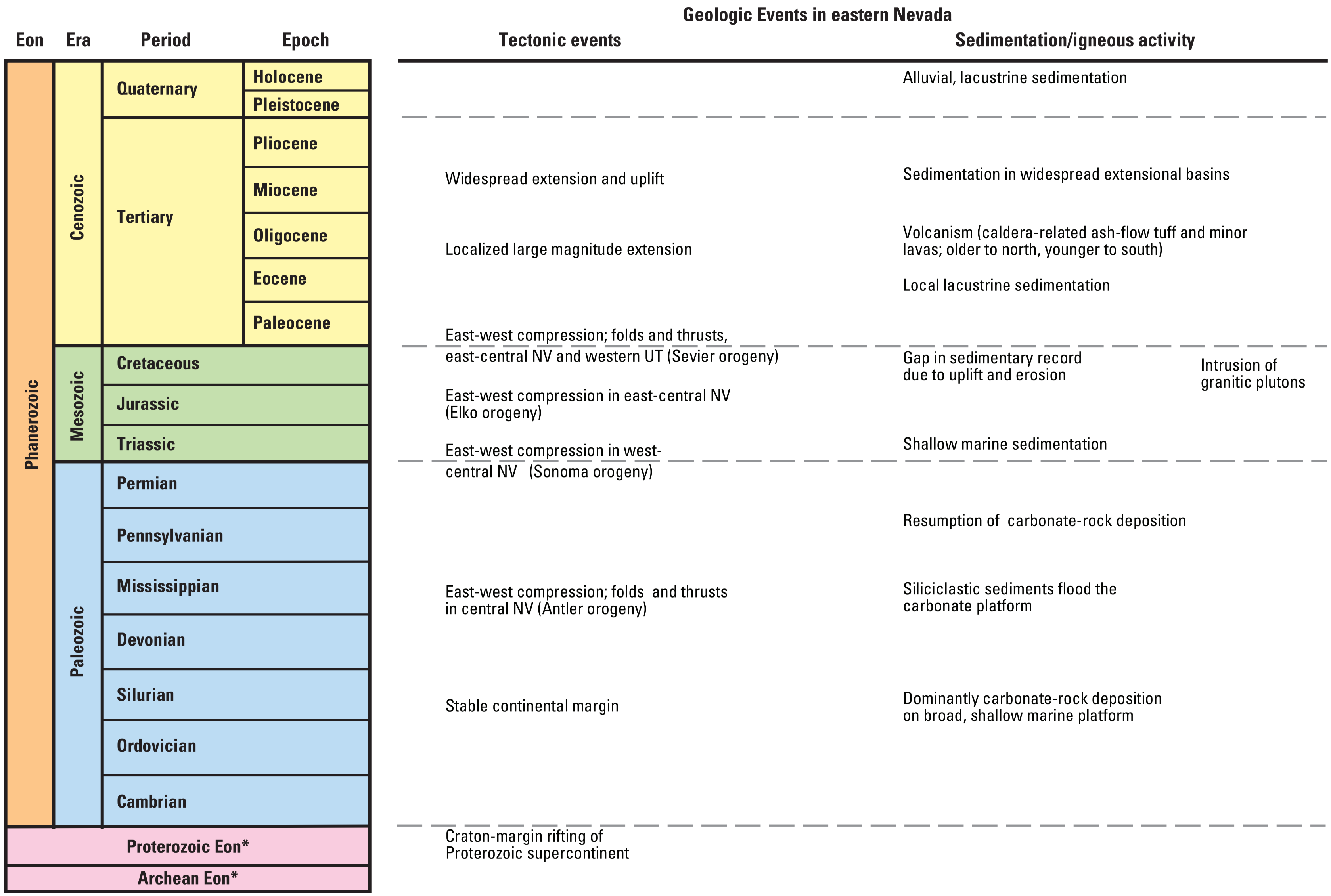
Hydrogeologic unit	Description of hydrogeologic unit	Examples of Nevada geologic formation names	Examples of Utah geologic formation names
FYSU	Fine-grained younger sedimentary rock unit; Holocene to Pliocene fine-grained playa and lake deposits of fine sand, silt, and clay.	Unconsolidated basin fill, includes playa, marsh, lake and alluvial-flat deposits.	Quaternary surficial deposits including Lake Bonneville deposits, marsh, salt and mudflat deposits.
CYSU	Coarse-grained younger sedimentary rock unit; Holocene to Pliocene alluvium, colluvium, and local fluvial deposits.	Unconsolidated basin fill, includes alluvial fan and stream channel deposits.	Quaternary and Pliocene Basin and Range valley-filling alluvial, and eolian deposits.
VFU	Volcanic flow unit; basalt, andesite and rhyolite lava flows.	Cenozoic basalt, andesite and rhyolite lava flows.	Miocene-Quaternary basalt, andesite and rhyolite lava flows.
VTU	Volcanic tuff unit; welded and nonwelded silicic ash-flow tuffs.	Shingle Pass Tuff, Cottonwood Wash Tuff, Lund Formation, and the Kalamazoo Tuff.	Oligocene Isom Formation and Needles Range Formation, Miocene volcanic rocks
OSU	Older sedimentary rock unit; consolidated Cenozoic (Eocene to Miocene) sedimentary rocks.	Includes Sheep Pass Formation (Eocene) and related units and unnamed tuffaceous sedimentary rocks	Various Paleocene and Eocene consolidated sedimentary rocks
MSU	Mesozoic sedimentary rock unit; includes limestone, sandstone and shale.	Moenkopi Formation, Thaynes Formation, and related rocks (Lower Triassic), in Butte Mountains.	Triassic Chinle, Moenkopi and Thaynes Formations (north of Confusion Range).
UCU	Upper carbonate-rock unit; predominantly limestone and silty limestone.	Ely Limestone (mostly Lower and Middle Pennsylvanian) and Lower Permian Arcturus Formation in White Pine County.	Pennsylvanian Ely Limestone and Permian Arcturus Formations.
USCU	Upper siliciclastic-rock unit; predominantly mudstone, sandstone, conglomerate, minor limestone.	Pilot Shale, Joana Limestone, Chainman Shale, and also Diamond Peak Formation in northern and western White Pine County	Mississippian Chainman Shale.
LCU	Lower carbonate-rock unit; predominantly limestone and dolomite with relatively thin sandstone and shale interbeds.	Cambrian Pioche Shale, Eldorado Dolomite, Geddes Limestone, Secret Canyon Shale, Hamburg Dolomite, Dunderberg Shale, and Windfall Formation; Ordovician Pogonip Group, Eureka Quartzite, and Ely Springs Dolomite; Silurian Laketown and Lone Mountain Dolomites; Devonian Sevy and Simonson Dolomites, Guilmette and Nevada Formations, and Devils Gate Limestone.	Cambrian Tripp Limestone, Wheeler Shale, Howell Limestone, Pioche Formation, Notch Peak Formation; Ordovician Ely Springs Dolomite, Eureka Quartzite, Lehman Formation, Kanosh Shale, Juab Limestone, Wah Wah Limestone, Fillmore Limestone, House Limestone; Silurian Laketown Dolomite; Devonian Pilot Shale, Guilmette Formation, Simonson Dolomite, Sevy Dolomite; Mississippian Ochre Mountain Limestone and Woodman Formation.
LSCU	Lower siliciclastic-rock unit; sandstones, siltstones and metamorphic equivalents.	Cambrian Prospect Mountain Quartzite, Proterozoic McCoy Creek Group and metamorphic rocks.	Cambrian Prospect Mountain Quartzite, Proterozoic McCoy Creek Group and metamorphic rocks.
IU	Intrsuive-rock unit; includes plutonic igneous rocks such as granite and granodiorite.	Jurassic through Oligocene intrusive rocks.	Jurassic through Oligocene intrusive rocks.

Figure 3. Generalized stratigraphic column and hydrogeologic units in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.



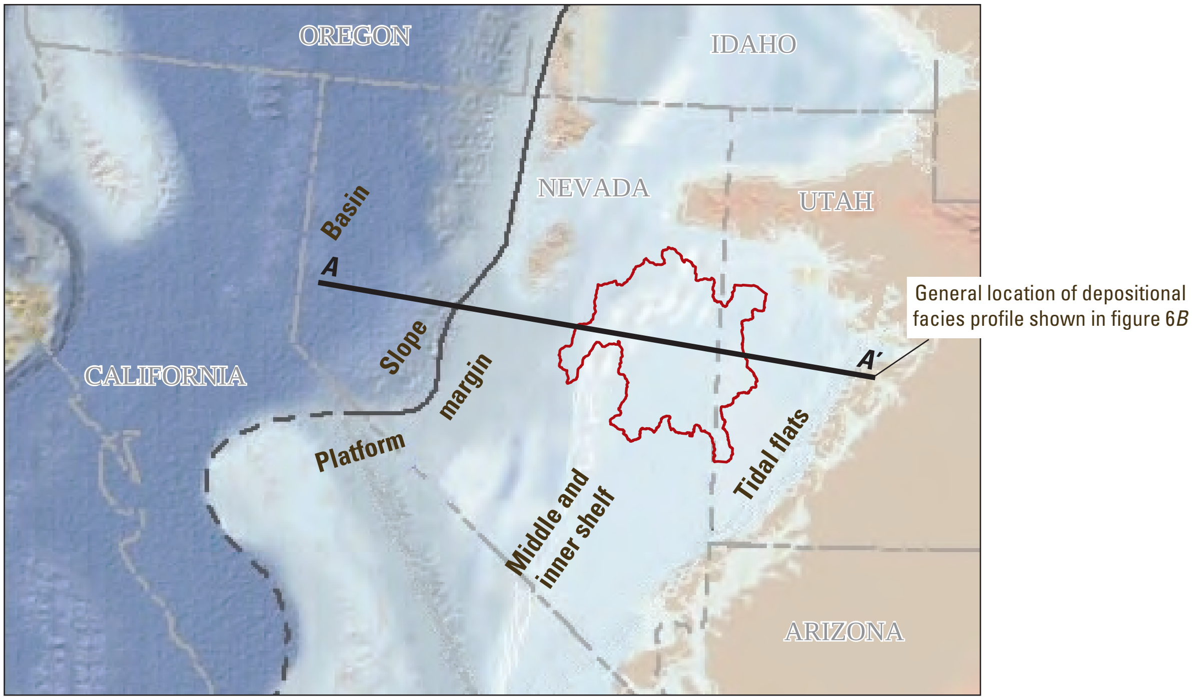
Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

Figure 4. Location of new geophysical data for the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2005–06.



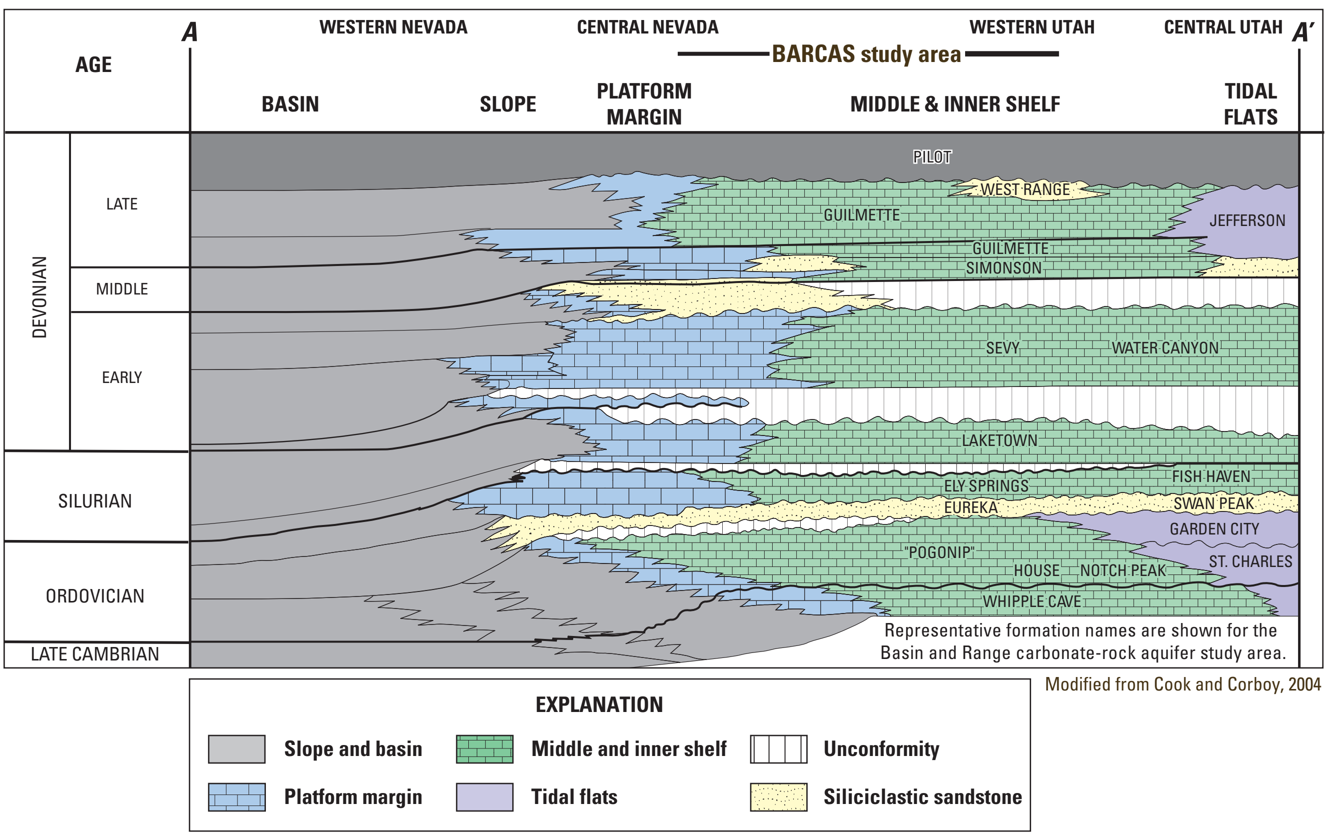
* The Archean and Proterozoic Eons are major subdivisions of Precambrian time.
** Million years before present

Figure 5. Major geologic events in eastern Nevada.



Modified from Blakely, 1997

A. Schematic representation of Silurian paleogeography



Modified from Cook and Corboy, 2004

Figure 6. Depositional facies and paleogeography, eastern Great Basin, Nevada and Utah.

From late Devonian to Eocene time, during the second major geologic phase of evolution, several episodes of east-directed compressive deformation that affected the central and western parts of Nevada and also influenced rocks in the study area (fig. 5). A Late Devonian to Early Mississippian compressive event, known as the Antler orogeny, interrupted deposition of carbonate rocks in the study area, resulting deposition a of thick sequence of siliciclastic rocks (Poole and Sandberg, 1977). Carbonate-shelf sedimentation resumed in Pennsylvanian and Permian time, again generating thick, widespread carbonate rocks in the study area. A late Jurassic through earliest Tertiary compressive event called the Sevier orogeny (fig. 5) resulted in the formation of regional-scale folds in the study area (Armstrong, 1968).

Starting in the middle to late Eocene through the remainder of the Tertiary period, extensional uplift and faulting, volcanism, and continental sedimentation characterized the third phase of in the geologic evolution of the study area (fig. 5) and adjacent areas in northern and eastern Nevada. During this time, modern basin-and-range landforms were created as a result of motion along both gently dipping and relatively high-angle faults, causing the relative rising of the ranges and sinking of adjacent basins. Generally accompanying the regional extension was the eruption of relatively large volumes of volcanic rocks, particularly ash-flow tuffs, that were deposited by caldera-forming eruptions during the Tertiary (Best and others, 1989). Caldera-forming eruptions from two major centers, the Indian Peak caldera complex and the Central Nevada caldera complex (pl. 1) resulted in deposition of volcanic rocks that extended across Nevada and Utah. Following Tertiary volcanism, unconsolidated sediments were deposited in the intermontane basins of the study area during the late Tertiary and Quaternary. These sedimentary deposits include Pliocene to Pleistocene-age fine-grained lake sediments (Reheis, 1999), and Quaternary age stream and alluvial-fan sediments of sand and gravel deposited along the basin margins, and changing to finer grained silt and clay sediments within playas along basin axes.

Structural Geology

East-central Nevada features structural domains that vary in style and intensity of deformation (Gans and Miller, 1983; Smith and others, 1991; Dettinger and Schaefer, 1996). Three principal structural domains are evident in the study area—compressional, extensional, and transverse (pl. 1). Compressional and extensional domains generally alternate spatially in the study area; for example, compressional domains represented by regional thrust belts or folds alternate with extensional domains of normal-faulted, highly attenuated stratigraphic sections (Gans and Miller, 1983). Transverse zones are regional scale, east-west structural alignments that generally perpendicular to the regional north-south alignment of mountain ranges and valleys. Salient structural features in the study area, including compressional thrust belts, large-magnitude extensional normal and detachment faults, and transverse zones, are shown on pl. 1.

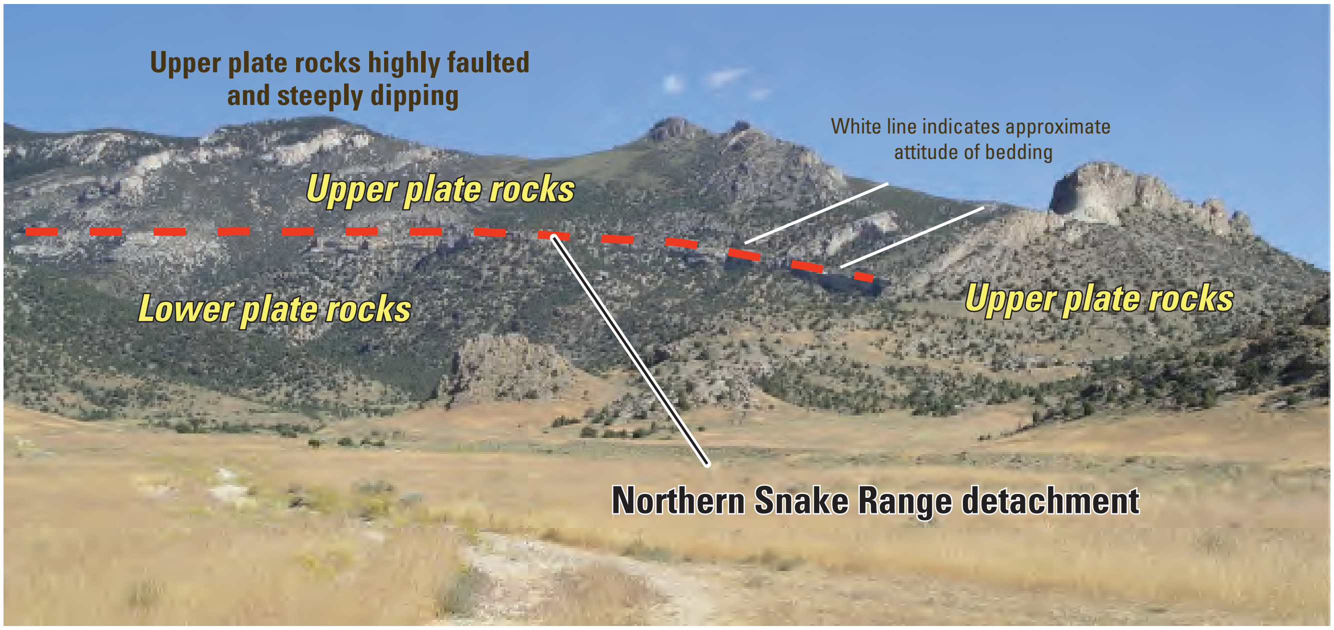
Thrust Belts

The only significant manifestation of the Mesozoic Sevier orogenic belt within the study area are two broad regional synclines, or downfolds, termed the Butte and Confusion Range synclinoria (Hose, 1977). These large folds are characterized by broadly sinuous but generally north-trending fold axes that preserve Triassic rocks and the entire underlying Paleozoic carbonate-rock section (pl. 1). The Butte synclinorium is present in the Maverick Springs Range and Butte Mountains, the central part of the Egan Range and the southern part of the Schell Creek Range (section *A-A*', pl. 1); the Confusion Range synclinorium is present in the Needle and Confusion Ranges of western Utah (section *B-B*', pl. 1).

Extension and Normal Faults

During Cenozoic time, north-south aligned mountain ranges of carbonate, siliciclastic, or metamorphic rocks were formed in the study area by episodes of structural extension. Structural extension was not uniform across the study area, but was segmented into domains of large-magnitude or relatively minor amounts of extension. Each domain generally is represented by specific HGUs that influence regional ground-water flow. The highly extended domains often have uplifted Precambrian to Cambrian siliciclastic rocks or metamorphic rocks of low permeability at or near the surface; whereas less-extended domains tend to preserve the entire thickness of Paleozoic carbonate rocks of higher permeability (pl. 1). Dettinger and Schaefer (1996) compared the structural setting and distribution of Paleozoic carbonate rocks with the location of regional ground-water flow systems within the carbonate-rock province. The two major ground-water flow systems in the study area, the Great Salt Lake Desert and the Colorado regional flow systems (fig. 1**)** were shown to correspond to areas with thick sections of Paleozoic carbonate rocks in parts of the study area that had been extended only slightly. However, the low-permeability siliciclastic rocks typically found in highly extended domains appear to completely disrupt carbonate-rock aquifer continuity and to partition ground-water flow into flow systems of limited lateral extent.

Within highly extended domains, extension was accomplished along gently to moderately dipping, large-offset extensional detachment faults. For example, in the northern Snake Range, an abrupt, gently dipping detachment fault brings low permeability granitic rocks and ductilely deformed and metamorphosed Cambrian and Precambrian quartzite, marble and pelitic schist to the surface (fig. 7; Miller and others, 1983). On the basis of seismic reflection



A. Low-angle detachment fault, eastern flank of northern Snake Range, Nevada.

B. Folded lower plate rocks, northern Snake Range, Nevada.

C. Brecciated upper plate rocks, northern Snake Range, Nevada.

Figure 7. Example of low-angle detachment, northern Snake Range, eastern Nevada.

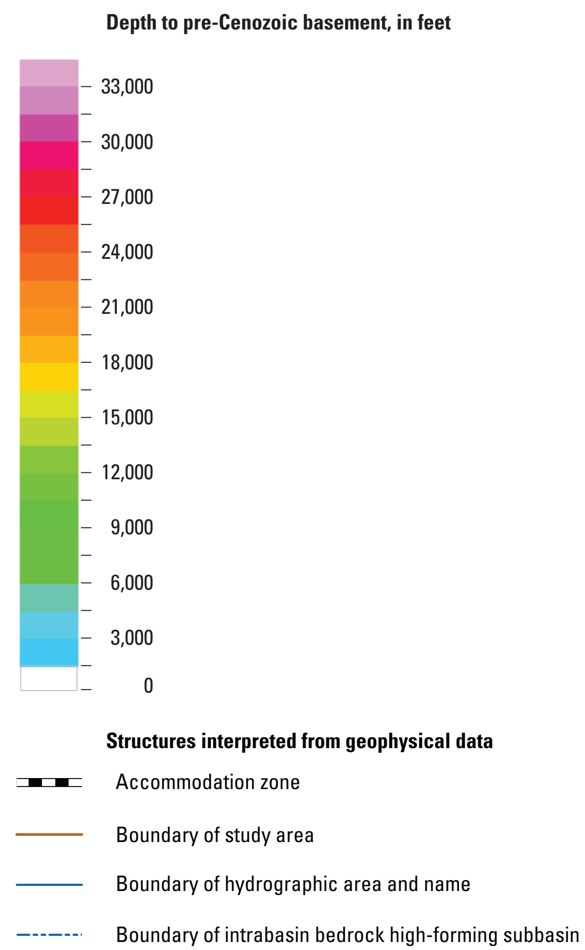
data, interpretive cross sections suggest that the moderately dipping detachment fault dips beneath Snake Valley (section *B-B*’, pl. 1) and beneath the Confusion Range to the east of the northern and southern Snake Range. Similar structures that bring low-permeability rocks to the surface exist in the southern Grant Range in northern Nye County (pl. 1) (Kleinhampl and Ziony, 1984; Lund and others, 1993) in the northern Egan and southern Cherry Creek Ranges (Armstrong, 1972; Gans and Miller, 1983) (section *A-A*’, pl. 1), and the Schell Creek Range (Dechert, 1967; Drewes, 1967; Armstrong, 1972).

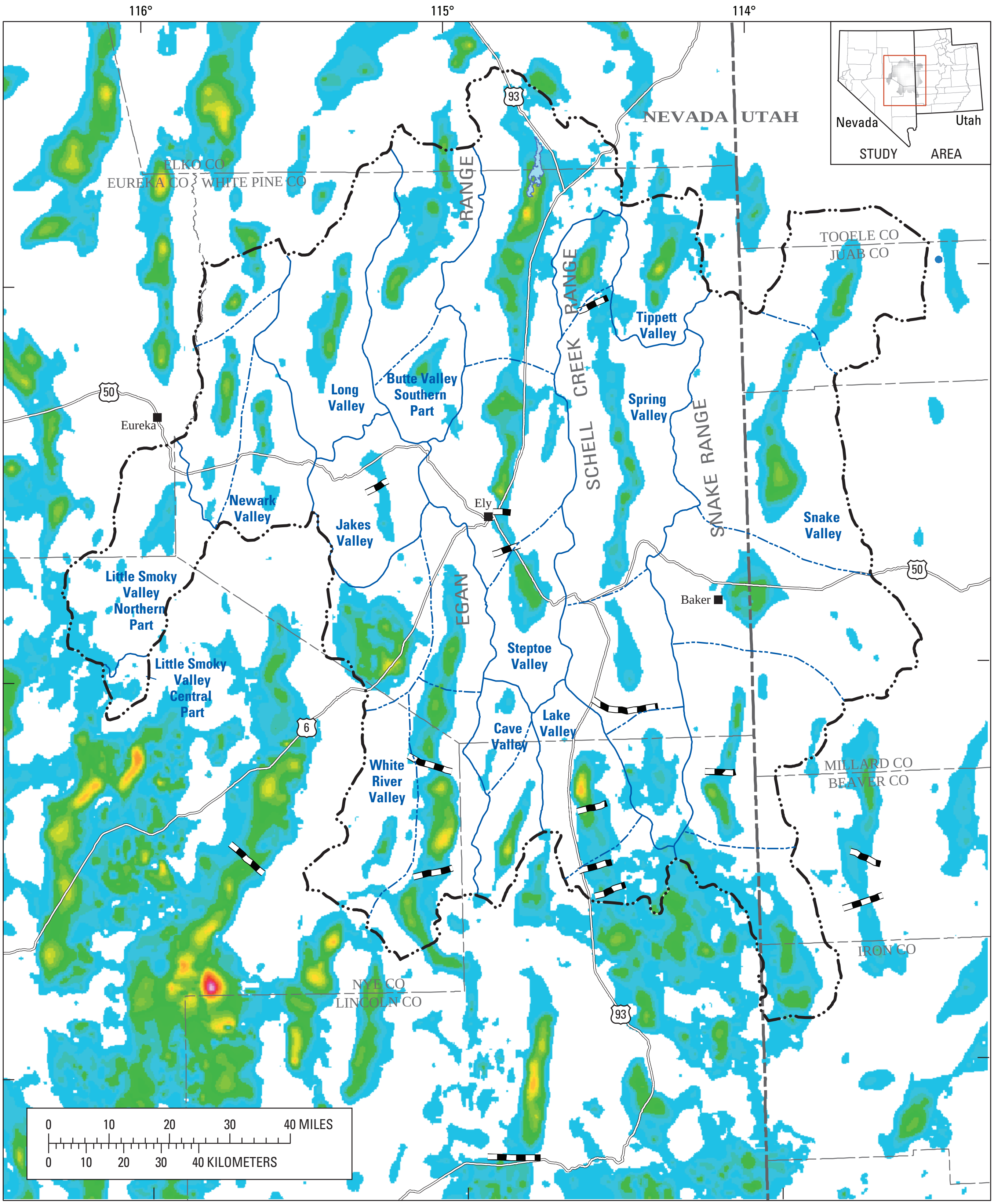
A second style of Tertiary extension is characterized by steeply dipping, range-bounding normal faults that produced elongate mountain ranges and have controlled the subsidence of intervening, down-faulted valleys (Zoback and others, 1981; Stewart, 1998). The range-bounding faults strike northeast and have displacements of several thousands of feet, typically juxtaposing the consolidated rocks within the range blocks against Cenozoic basin fill (Kleinhampl and Ziony, 1984). Basins commonly have a half-graben form in which the basin fill and basin floor are tilted toward a major fault on one side of the basin; this fault accommodates much of the extensional deformation and subsidence, producing a tilted, asymmetric basin (Stewart, 1998). Less commonly, basins have the form of a symmetric graben, with major faults bounding both sides of the basin. Symmetric grabens typically are located along the valley axis, with shallow pediments on either side. General relations between extensional range-bounding faults and associated asymmetric and symmetric grabens are annotated on cross section *C-C*’ on pl. 1. Geophysical data show that basins in the study area vary in their complexity of faulting and relative development (Saltus and Jachens, 1995; Dohrenwend and others, 1996). For example, in White River Valley, along the western part of seismic line ECN-01 (section *C-C*’, pl. 1), there are three east-dipping half-grabens increasing in size from west to east. These half-grabens are largely buried and are not evident from surface topography or bedrock outcrops. In contrast, Cave Valley is a single east-dipping half-graben, where the floor of the graben mimics the dip of the Paleozoic rocks on the west side of the basin and a steeply dipping fault zone bounds its eastern edge.

Analysis of regional gravity data provides the basis for assessing the thickness of the Cenozoic basin-fill deposits (fig. 8). Cross sections that incorporate the geophysical data portray the three-dimensional shape of pre-Cenozoic basement, the location of major basin-bounding structures, and the presence of significant intrabasin faults (fig. 9). The thickness of basin fill in the study area generally is about 6,600 ft; however, in some basins, such as Steptoe and Lake Valleys, the thickness of basin fill is more than 13,000 ft (fig. 8). With the exception of Steptoe Valley in the north, basins in the southern part of the study area contain thicker basin-fill deposits than basins in the northern part of the study area.

Gravity-derived models of pre-Cenozoic bedrock, integrated with seismic, aeromagnetic, and drilling data, indicate that many of the basins in the study area contain buried bedrock highs (sections *C-C*’ and *F-F*’, fig. 9). These bedrock highs represent intrabasin divides that separate most basins into two or more subbasins; geologically, they are referred to as accommodation zones (fig. 8) that developed in response to differential extension or tilting in different parts of the basin. In selected cases where the intrabasin divides are particularly shallow or distinctly separate deeper basins, these locations were chosen to subdivide hydrographic areas into subbasins (fig. 4). Subbasins do not necessarily represent individual ground-water basins, but merely areas separated by intrabasin divides where pre-Cenozoic bedrock has been uplifted and overlying basin-fill deposits are relatively thin. The geometry and structure of basins and associated subbasins in the study area are summarized in table 1.

EXPLANATION FOR FIGURE 8





Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

Caldera boundaries modified after Williams and others (1997), Loucks and others (1989), Raines and others (1996), Workman and others (2002), and Gans and others (1989).

Figure 8. Depth-to-bedrock map of the study area showing interpreted lineaments or features, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

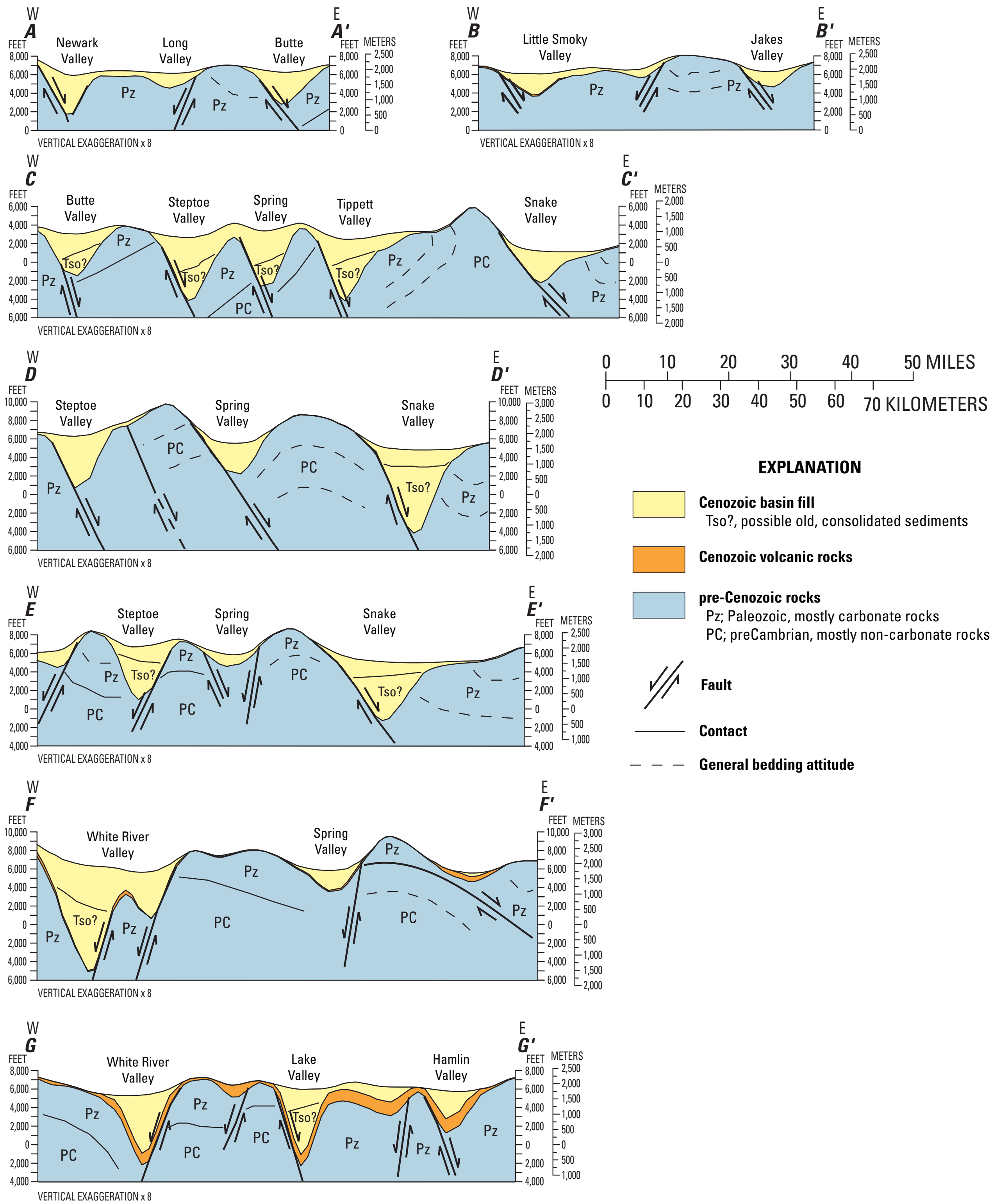


Figure 9. Modeled depth to pre-Cenozoic rocks and location of sections, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

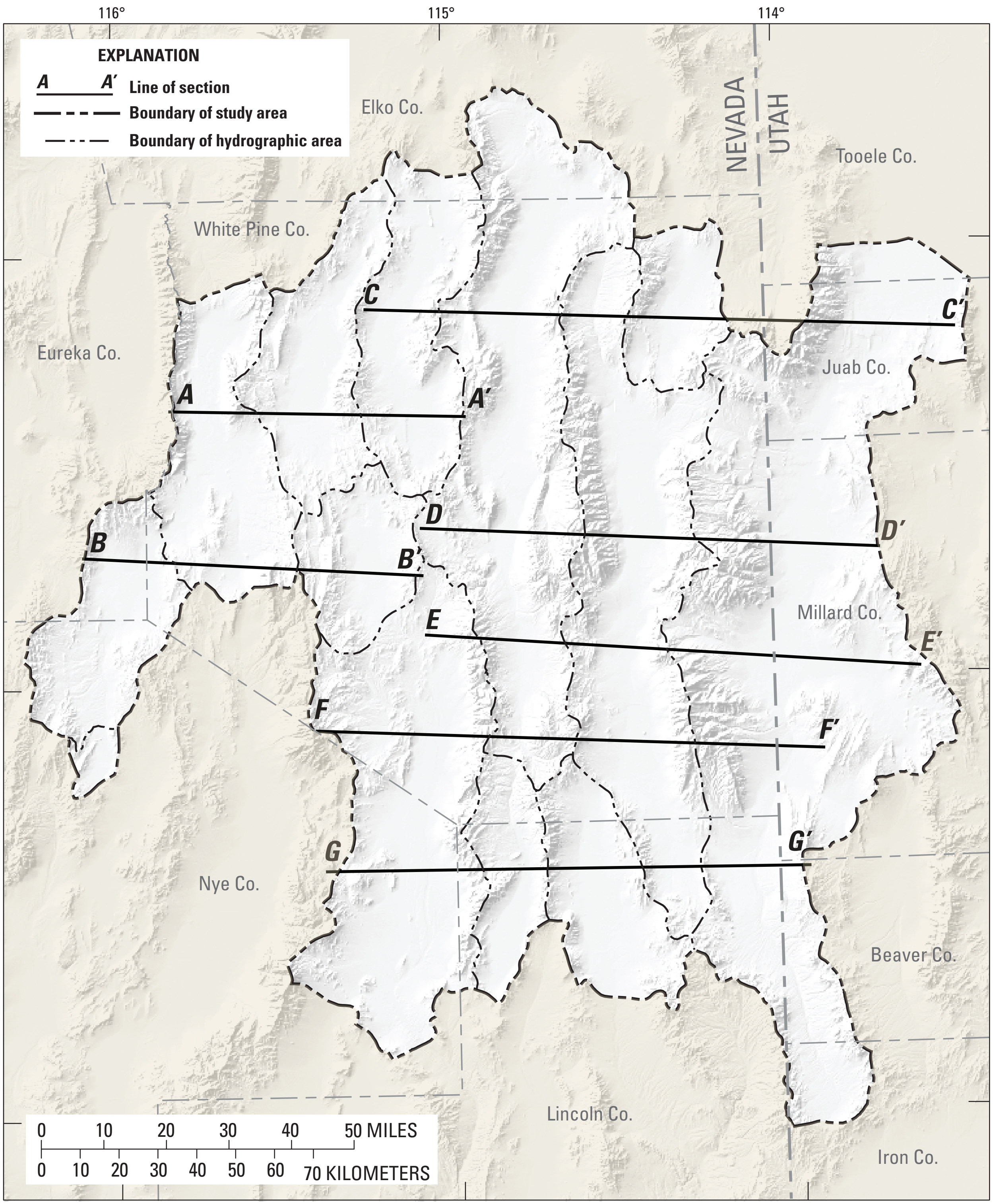


Figure 9. Continued.

Table 1. Basin structure of hydrographic areas in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[Abbreviations: km, kilometer; <, less than]

Hydrographic area name	Subbasin	Typical depth to bedrock, in miles	Maximum depth to bedrock, in mile	Basin geometry and structure
Butte Valley	1 2	<0.3 <0.3	2.5 1.9	Subbasin 1 is arcuate in shape and dips to the northwest, with depths increasing northwestward from <0.3 to 2.5 mi. Subbasin 2 dips to the north; basin depths increase northward from 0.6 to 1.9 mi.
Cave Valley	1 2	<0.3 <0.6	0.9 4.3	Subbasin 1 is very shallow; basin floor dips to the northwest, reaching a maximum depth of 1 mi, corroborated by data from a single drill hole (Hess and others, 2004). Subbasin 2 dips steeply to the southeast, reaching depths of 2.5–4.3 mi in the southeastern part. No late Cenozoic faults are present in subbasin 2.
Jakes Valley	One basin	<0.3	0.9	Single shallow, north-south-trending basin, largely <0.3 mi deep; small area in southeast as much as 1 mi deep. Basement depth confirmed by a single drill hole (Hess and others, 2004).
Lake Valley	1 2	<0.6 <0.6	3.7 3.1	Subbasin 1 as much as 3.7 mi deep, shallowing rapidly in an east-west direction to less than 0.6 mi. Basin-fill deposits dip gently to the east; small, west-dipping normal fault in the middle of the northern basin. Subbasin 2 ranges from 1.2 to 3 mi deep with major area of sediment deposition at the northern and southern ends. Aeromagnetic data suggest that basin is filled with highly magnetic volcanic rocks. Divide between the northern Lake Valley and Patterson Valley coincides with the interior of a caldera, with volcanic rocks on either side.
Little Smoky Valley	Northern part Central part	<0.3 <0.3	0.9 0.3	Northern part is west dipping and as much as 1.2 mi deep but generally <0.6 mi deep. A partially buried, north-trending basement high separates the northern and central parts.
Long Valley	One basin	<0.3	2.5	Gravity and drilling data (Hess and others, 2004) indicate that the basin is asymmetric to the east—basin depths range from <0.6 mi on west side to as much as 2.5 mi in east-central part of basin. Basin depths decrease to 1 mi in the north and <0.6 mi in the south. Pekarek (1988) reported that seismic data showed the valley to be as much as 8,000 ft deep.
Newark Valley	1 2 3	<0.3 <0.3	3.1 1.9	Two basins are separated by north-trending, shallowly buried basement ridge connecting southern Buck Mountain and northern Pancake Range. Eastern basin is between 0.6 and 1.9 mi deep; basin floor dips west so basin is deepest on the west side. Most of western basin is <0.6 mi deep; it can be further divided into three subbasins.
Snake Valley	1 2 3 4 5	<0.3 <0.3 <0.3 <0.3 0.3	4.3 1.9 1.9 3.1 1.9	Cenozoic basin fill generally <0.3 mi thick except for east of the Kern Mountains (basin fill >3 mi) and east of Sacramento Pass (basin fill >2 mi). Northern two subbasins predominantly filled with west-tilted Miocene synorogenic clastic sediments covered by thin late Cenozoic fill. Three southern basins interpreted from gravity, seismic (Alam, 1990), and drill-hole data (Hess and others, 2004). Subbasin 3 is gently west dipping and generally <0.6 mi deep; it is bounded on the west side by the Snake Range detachment and related hanging-wall normal faults. Subbasin 4 is a west-dipping half graben, bounded on west side by a listric east-dipping normal fault running along the base of the Limestone Hills. Basin as much as 3 mi deep on northwest, but <0.3 mi on the east. Southern basin dips gently to southwest, <0.3 mi on eastern margin.

Table 1. Basin structure of hydrographic areas in the Basin and Range carbonate-aquifer system study area, Nevada and Utah—Continued

[Abbreviations: km, kilometer; <, less than]

Hydrographic area name	Subbasin	Typical depth to bedrock, in miles	Maximum depth to bedrock, in mile	Basin geometry and structure
Spring Valley	1 2 3 4	<0.6 <0.3 <0.3 <0.3	3.1 2.5 1.2 1.2	Northern basin, west of the Antelope Range in northernmost Spring Valley, is elongate in a northeast-southwest direction, as much as 3 mi deep and separated from the rest of Spring Valley to the south by a ridge of Tertiary volcanic rocks. The north-central basin underlies most of northern Spring Valley north of U.S. Highway 50; it is confined to a narrow (3–6 mi-wide) zone near the center of the basin. An east-west trending structural high associated with Tertiary rhyolite exposed at Rattlesnake Knoll near U.S. Highway 50 separates the northern and central basins. Central subbasin is very shallow (<0.6 mi) except for a small circular area of sediment deposition at its southern margin as much as 1.2 mi deep. Southern subbasin lies between the Fortification Range (west) and the Limestone Hills (east) and is as much as 1.2 mi deep.
Steptoe Valley	1 2 3 4 5	<0.9 <0.6 <0.9 <0.3 <0.6	4.3 2.5 3.1 1.2 2.5	North of Ely Steptoe Valley is a graben, asymmetric to the west, composed of three main basins. The northern basin is segmented into three subbasins that deepen to the north and dip to the west towards the range front fault at the base of the Egan Range. The central basin located northwest of McGill is elongate in an east-west orientation, as much as 2.5 mi deep in the center and shallows to <1 mi to the north and south. The southern subbasin is narrow and elongate and as much as 3 mi deep at its southernmost margin. South of Ely, the northern basin is small and relatively shallow (<1.2 mi depth). The southern basin is as much as 2.5 mi deep, elongate in a northwest-southeast direction, and is east-dipping.
Tippett Valley	1 2	<0.6 <0.3	3.1 1.9	Northern basin is elongate in a NNE-SSW direction, is deepest at its southern end, and has a graben geometry. The southern basin is a shallow (<0.06 mi) west-dipping half-graben that is separated from the northern basin by a narrow buried basement high.
White River Valley	1 2 3 4	<0.6 <0.9 <0.6 <0.3	3.1 3.7 1.9 3.1	Based on seismic and drilling data and geologic mapping, the valley consists of three east-dipping, fault-bounded half grabens (Potter and others, 1991). These generally north-striking faults control scattered outcrops of Paleozoic bedrock that occur in the middle of White River Valley. The drilled depth to Paleozoic rocks ranges from 1,300 to more than 5,000 ft, depending on location. Within this structural framework, gravity data define three large basins (northern, central, and southern) and one small western basin. The northern basin is as much as 3 mi deep, elongate in a NNE-SSW direction and appears to dip gently to the east. The central basin is a steeply east-dipping half graben bounded on the east along the range front fault system at the base of the Egan Range (Potter and others, 1991). Basin depths reach as much as 3.7 mi along the fault, but shallow rapidly to <0.3 mi to the west. The western basin is a relatively shallow (<1.9 mi) east-dipping half graben. The southern basin is west-dipping and curves to the southeast forming an arcuate subbasin that follows the White River between the Seaman and North Pahroc Ranges south of Cave Valley. This subbasin is as much as 3 mi deep at its northern margin, but shallows southeastward to an average of 1.9 mi.

Transverse Zones

Transverse zones (Faulds and Varga, 1998) generally are regional scale, east-west-trending features that have been previously identified in the study area (Ekren and others, 1976; Rowley, 1998). Transverse zones segment subbasins, hydrographic areas, or larger regions into areas of different types, rates, or relative amounts of extension. Transverse zones commonly are oriented at a high angle to the long axes of current basins and ranges and, as a result, may influence the rate or direction of ground water flowing parallel to valley axes. The influence of such zones on ground-water flow patterns is largely unknown.

Hydrostratigraphy

HGUs have considerable lateral extent and similar physical characteristics that may be used to infer their capacity to transmit water. Material properties of basin fill and consolidated rock, therefore, were used as indicators of primary and secondary permeability, such as grain size and sorting, degree of compaction, rock lithology and competency, degree of fracturing, and extent of solution caverns or karstification.

The consolidated pre-Cenozoic rocks, Cenozoic sediments, and igneous rocks of the study area are subdivided into 11 HGUs (table 2; fig. 3). Pre-Cenozoic rocks and older Cenozoic rocks were classified as consolidated rocks (commonly referred to as bedrock) that may consist of limestone, dolomite, sandstone, siltstone, and shale. Consolidated pre-Cenozoic rocks are subdivided into HGUs based primarily on the degree to which the rocks fracture and, in the case of limestones and dolomites, the presence of solution openings. Proterozoic to Early Cambrian metamorphic and siliciclastic rocks, and Paleozoic siliciclastic rocks typically form the least permeable HGU within the consolidated, pre-Cenozoic rocks. Paleozoic carbonate rocks typically form the most permeable HGUs within the pre-Cenozoic consolidated rocks. These carbonate rocks extend throughout much of the subsurface in western Utah, central

Table 2. Description of hydrogeologic units of the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Hydrogeologic unit abbreviation for this study	Equivalent hydrogeologic unit abbreviation in the Death Valley ground-water flow system (Belcher, 2004)	Hydrogeologic unit name	Descripton of hydrogeologic unit
FYSU	ACU	Fine-grained younger sedimentary rock unit	Young Cenozoic lacustrine, playa and basin axis deposits
CYSU	AA	Coarse-grained younger sedimentary rock unit	Young Cenozoic alluvial and fluvial deposits
VFU	CHVU and BRU	Volcanic flow unit	Cenozoic basalt, andesite, dacite and rhyolite lava flows
VTU	TMVA, PVA, and CFPPA	Volcanic tuff unit	Cenozoic ash-flow tuffs
OSU	VSU	Older sedimentary rock unit	Consolidated Cenozoic sandstone and limestone
MSU	SCU	Mesozoic sedimentary rock unit	Mesozoic limestone, sandstone, and shale.
UCU	UCA	Upper carbonate-rock unit	Mississippian to Permian carbonate rocks
USCU	UCCU	Upper siliciclastic-rock unit	Mississippian siliciclastic rocks and some limestone
LCU	LCA	Lower carbonate-rock unit	Cambrian to Devonian predominantly carbonate rocks
LSCU	LCCU	Lower siliciclastic-rock unit	Cambrian and Precambrian siliciclastic rocks
IU	ICU	Intrusive Unit	Intrusive rocks such as granite and granodiorite, not divided by age

and southern Nevada, and eastern California (Dettinger, 1989; Harrill and Prudic, 1998), and crop out in many of the mountain ranges in the study area (pl. 1). Younger Cenozoic sediments were classified as basin-fill deposits that may consist of unconsolidated granular material such as sand, gravel, and clay. The unconsolidated Cenozoic basin fill is subdivided into HGUs based on grain size and sorting. Igneous rocks are subdivided on the degree to which the rocks fracture and, for the volcanic rocks, on the presence or absence of soft ashy material.

Pre-Cenozoic Sedimentary Rocks

The pre-Cenozoic sedimentary rocks of the study area are grouped into five HGUs: the lower siliciclastic-rock unit (LSCU), the lower carbonate-rock unit (LCU), the upper siliciclastic-rock unit (USCU), the upper carbonate-rock unit (UCU), and the Mesozoic sedimentary rock unit (MSU). This usage is similar to that established by Winograd and Thordarson (1975).

The lower siliciclastic-rock unit (LSCU) includes the oldest exposed sedimentary rocks in the study area, including the upper Precambrian McCoy Creek Group, which consists of more than 9,000 ft of siliceous and argillaceous metasediments and the Lower Cambrian Prospect Mountain Quartzite, which is as much as 4,500 ft thick of predominantly quartz-rich sandstone (fig. 10; Hose and others, 1976). Rocks of the LSCU are exposed in the Cherry Creek Range, the northern part of the Egan Range, the Schell Creek Range and the Snake Range (pl. 1 and fig. 10). Schists and marbles also are included in the LSCU, and these rocks form, in part, the lower plates of major extensional detachment faults in the Snake and Schell Creek Ranges.

The LSCU generally has low permeability throughout the eastern Great Basin (Winograd and Thordarson, 1975; Plume, 1996). Sandstones of the LSCU commonly are highly cemented, filling much of the original pore volume, and are overlain and underlain by a significant thickness of fine-grained shales, all of which contribute to the overall low permeability of this HGU. At shallow depths, rocks of the LSCU commonly are highly fractured (fig. 10) and can support small volumes of flow, such as at Strawberry Creek in the northeastern part of Great Basin National Park (Elliott and others, 2006). Schists and marbles of the LSCU that typically have schistose foliation lack a continuous fracture network. Based on the low permeability and capacity to transmit water, the top of the LSCU, for purposes of this report, represents the base of the ground-water flow.

The LCU represents a significant volume of carbonate rock that is prominently exposed in the mountain ranges in the study area (pl. 1), and is present beneath many of the valleys. The LCU includes Cambrian through Devonian limestones and dolomites with relatively minor interbedded siliciclastic rocks. A representative stratigraphic succession of the LCU in the study area typically consists of the following units, from lower (older) in the succession, to higher (younger) in the succession: a Middle Cambrian to Lower Ordovician limestone, silty limestone, siltstone, and shale section, a distinctive Middle Ordovician Eureka quartzite, an Upper Ordovician through Middle Devonian dolomite, and a limestone and minor dolomite of the Middle and Upper Devonian Guilmette Formation (fig. 11) (Kellogg, 1963; Poole and others, 1992).

The LCU, along with the carbonate-rock units of the UCU, forms a major high-permeability consolidated-rock unit in the Great Basin (Winograd and Thordarson, 1975; Bedinger and others, 1989; Dettinger and others, 1995; Harrill and Prudic, 1998). Carbonate rocks of the LCU and UCU have three distinct types of porosity that influence permeability and associated storage and movement of ground water—primary or intergranular porosity, fracture porosity, and vug or solution porosity. Lower Paleozoic carbonate rocks from southern Nevada have relatively low primary porosity (Winograd and Thordarson, 1975). Studies of ground-water flow within the carbonate-rock province (Winograd and Pearson, 1976; Dettinger and others, 1995; Harrill and Prudic, 1998) have continued to emphasize correspondence of faults and broad structural belts with zones of high transmissivity, presumably the result of the formation of fractures during deformation. Moreover, in their analyses of hydraulic property estimates for rocks equivalent to the LCU and UCU in the carbonate-rock province, Belcher and others (2001) concluded that extensive faulting and karst development significantly enhanced hydraulic conductivity. Fracture permeability may be enhanced if vertical fractures intersect horizontal fractures, creating a well-connected network of openings through which water can move. In addition, water can dissolve carbonate rocks to form solution openings that create additional pathways. For example, as a result of periodic declines in sea level during Paleozoic time, extensive areas of carbonate rock in east-central Nevada were exposed to the air and subsequent erosion. These intervals of erosion are represented in the sedimentary record as unconformities (fig. 6)—relatively long gaps in time when the carbonate platform was above sea level and conditions were favorable for erosion, dissolution, and development of solution caverns in the exposed carbonate rocks.

The Paleozoic carbonate rocks of the LCU are overlain by a sequence of Mississippian mudstone, siltstone, sandstone, and conglomerates that form the upper siliciclastic-rock unit (USCU). These rocks were formed by the muddy and sandy sediment influx associated with the Antler orogenic event and are represented by rocks of the Mississippian Chainman Shale, Diamond Peak Formation, and Scotty Wash Quartzite. This succession of sedimentary rocks is widely distributed across the study area and, where not structurally thinned, generally ranges in thickness from 1,000 to greater than 3,000 ft (Hose and others, 1976).

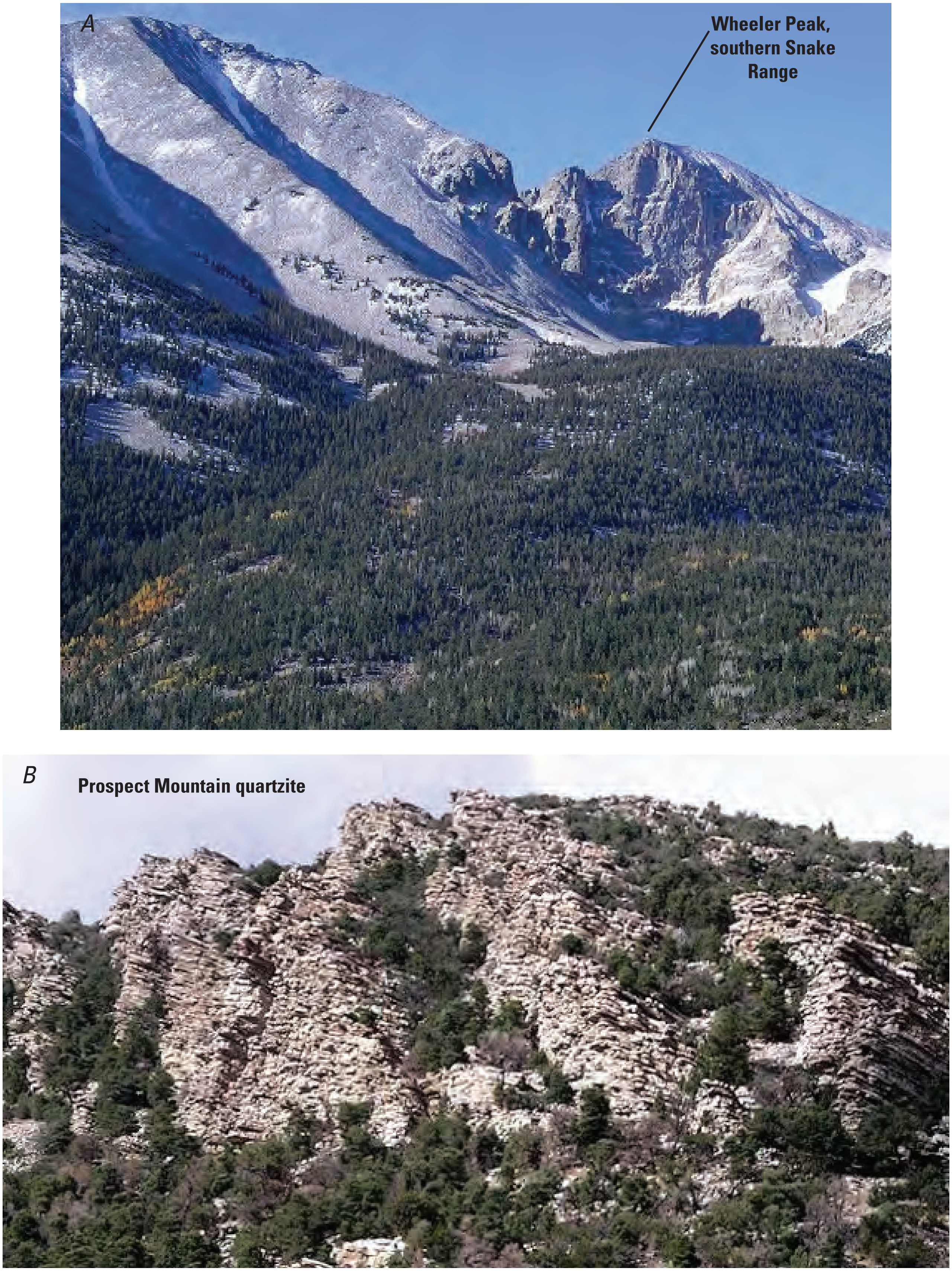


Figure 10. Lower Cambrian siliciclastic rocks, southern Snake Range, Nevada. Photographs taken by Donald S. Sweetkind, U.S. Geological Survey, (*A*) October 4, 2005; (*B*) September 10, 2004.

Figure 11. Lower Paleozoic sedimentary rocks, southern Egan Range, eastern Nevada. Photograph taken by Donald S. Sweetkind, U.S. Geological Survey, September 26, 2005.

The shaly siliciclastic rocks of the USCU are fine grained and of low permeability. Because of their low susceptibility to dissolution or fracturing, the USCU also lacks significant secondary permeability. The shaly rocks of the USCU yield in a ductile manner when deformed and deformation does not result in significant fracture openings through which water can flow. For example, in southern Nevada, steep hydraulic gradients at the Nevada Test Site are attributed to the low permeability of the Mississippian siliciclastic rocks (Winograd and Thordarson, 1975; D'Agnese and others, 1997); similar properties are expected for these rocks in the study area. The low porosity of the Chainman Shale in the study area has been tabulated (Plume, 1996) from data from oil and gas exploration wells. In the western part of the study area where the Chainman Shale grades laterally and upward into the coarser conglomeratic rocks of the Diamond Peak Formation, a number of exploration wells have penetrated this unit.

The upper carbonate-rock unit (UCU) are thick, widespread Pennsylvanian and Permian carbonate rocks that overlie the Mississippian rocks of the USCU. The rocks of the UCU were deposited during a resumption of upper Paleozoic carbonate-rock deposition in a stable shelf environment (Cook and Corboy, 2004). In the western and eastern parts of the study area that were less disturbed by subsequent structural extension, upper Paleozoic rocks dominate outcrops in ranges and at interbasin divides (pl. 1). Within these areas, the UCU includes as much as 4,000 ft of Ely Limestone and approximately 2,500 ft of Arcturus Group limestones and silty limestones (Hose and others, 1976). The UCU and LCU possess similar secondary fracture and solution permeability and, as a result, the UCU potentially is an important conduit for recharge and interbasin ground-water flow through ranges in the northwest part of White Pine County, in the central part of the Egan and Schell Creek Ranges, and in the Confusion Range in western Utah.

The Mesozoic sedimentary rock unit (MSU) is preserved in the cores of down-folded regional synclines and, therefore, is exposed only in isolated patches throughout the study area (pl. 1). Triassic rocks of the MSU consist of interbedded siltstone and limestone (Hose and others, 1976) that typically are relatively thin in exposure, about 150 ft thick in the Butte Mountains and slightly thicker in western Utah. Equivalent MSU rocks on the Colorado Plateau, southeast of the study area, are relatively permeable, but most exposures of the MSU in the study area are too small in lateral extent and shallow to be significant conduits for ground-water flow.

Cenozoic Basin-Fill Units

The Cenozoic sediments of the study area are grouped into three HGUs: the consolidated older sedimentary rock unit (OSU), and two unconsolidated units, the coarse-grained younger sedimentary rock unit (CYSU) and fine-grained younger sedimentary rock unit (FYSU) (table 1; fig. 3). The occurrence and lithologic characteristics of Cenozoic basin-fill deposits in the study area are summarized in table 3. Characteristics of the basin-fill deposits are described in terms of the abundance and type of volcanic rocks within the basin, and the presence or absence of sedimentary rocks or Pleistocene lake deposits (Reheis, 1999). Inferences regarding the character of the basin-fill deposits are made on the basis of surrounding geologic outcrops, information from oil and gas exploration wells (Hess and others, 2004), aeromagnetic data, and seismic data.

Table 3. Lithologic characteristics and occurrence of basin-fill deposits, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[**Abbreviations:** ft, foot; Ma, million years ago; mi, mile]

Hydrographic area name	Volcanic rocks	Sedimentary rocks and lake sediments
Butte Valley	Eocene lavas extensive at the south end of the valley (Feeley and Grunder, 1991), also along western basin margin, and in east-central part of basin (Gans and others, 1989). Surface and subsurface occurrences of these volcanic rocks are expressed as relatively high-amplitude magnetic anomalies.	Tertiary tuffaceous sedimentary rocks exposed in small areas at the southern and northern ends of the basin. A late Pleistocene lake existed in the central part of Butte Valley (Reheis, 1999).
Cave Valley	Oligocene volcanic extensively exposed in the Egan Range adjacent to the northern subbasin and at the southern end of the southern subbasin. However, none of the oil and gas wells in southern Cave Valley report encountering volcanic units below alluvium (Hess and others, 2004)	Subsurface data from oil and gas wells (Hess and others, 2004) include Miocene sediments and Eocene sediments, with no intervening volcanic rocks. Miocene sediments exposed on the east flank of the Egan Range are fluvial and tuffaceous, with a thickness of 2,000 ft (Kellogg, 1964). A Late Pleistocene lake existed in the southern part of the southern subbasin (Reheis, 1999).
Jakes Valley	Oligocene volcanic rocks extensive at the northeastern margin of the valley.	Pleistocene lake existed in the central part of the valley (Reheis, 1999).
Lake Valley	Tertiary volcanic rocks are extensively exposed in ranges flanking the valley and the northern margin of the Indian Peak caldera complex has been inferred to extend roughly west-southwest beneath Lake Valley (Best and others, 1989). Well data (Hess and others, 2004) and aeromagnetic data indicate that thick volcanic rocks are present at depth in the northern part of the valley but not in central Lake Valley.	Quaternary lacustrine deposits are exposed in the floor of the northern half of the valley. The northern part of Lake Valley contained a Pleistocene lake; none was present in the southern part (Patterson Valley) (Reheis, 1999). Late Miocene to Pliocene Panaca Formation is exposed in the southern half of the valley (Patterson Valley) (Phoenix, 1948); its presence in the northern half of the valley is unknown.
Little Smoky Valley (northern part)	Tertiary volcanic rocks are exposed locally along the eastern and southern margins of the valley; however, subsurface data from oil and gas exploration wells (Hess and others, 2004) indicate that there are no volcanic rocks within the basin fill.	Well data (Hess and others, 2004) indicate that the basin fill consists of Quaternary and Tertiary sediments. The northern half of the valley contained Pleistocene lakes (Reheis, 1999); the entire valley is covered by Quaternary sediments.
Little Smoky Valley (central part)	Tertiary volcanic rocks are exposed locally along the eastern and southern margins of the valley; however, subsurface data from oil and gas exploration wells (Hess and others, 2004) indicate that there are no volcanic rocks within the basin fill.	Well data (Hess and others, 2004) indicate that the basin fill consists of Quaternary and Tertiary sediments. The northern half of the valley contained Pleistocene lakes (Reheis, 1999); the entire valley is covered by Quaternary sediments.
Long Valley	Eocene-Oligocene volcanic rocks and small outcrops of tuffaceous Tertiary sedimentary rocks are exposed on the western side of the valley; but not on the eastern side. Data from oil and gas exploration wells (Hess and others, 2004) report depths to Oligocene volcanic rocks that range from 460 to 1,900 ft and have thicknesses of 194 to 2,434 ft, consistently thinning to the north from the center of the basin. The presence of these volcanic rocks is confirmed by aeromagnetic data.	Most of the valley contained Pleistocene lakes (Reheis, 1999).
Newark Valley	Oligocene to early Miocene (36–20 Ma) volcanic rocks and minor Miocene sediments that are likely ash rich are present at the southern end of the valley; oil and gas wells (Hess and others, 2004) provide no data regarding the presence or absence of volcanic rocks at depth.	Newark Valley contained Pleistocene lakes (Reheis, 1999) except in the southeastern arm of the valley to the east of the Pancake Range. Paleogene sediments are exposed at the northern end of the valley. Lithologic logs from oil and gas exploration wells in the valley (Hess and others, 2004) do not differentiate any of the Tertiary and Quaternary units, referring to the entire section as "valley fill."

Table 3. Lithologic characteristics and occurrence of basin-fill deposits, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah—Continued

[**Abbreviations:** ft, foot; Ma, million years ago; mi, mile]

Hydrographic area name	Volcanic rocks	Sedimentary rocks and lake sediments
Snake Valley	Volcanic rocks are absent in subbasins 1–3 and flanking ranges. Three wells (Hess and others, 2004) in subbasin 4 all penetrated volcanic rocks at depth. Drill-hole data and seismic data do not support the postulated existence of a source caldera for the Cottonwood Wash Tuff (Best and others, 1989). Subbasin 5 is primarily filled with volcanic rocks of the Indian Peak caldera complex Basin depths likely reflect a much thicker volcanic sequence in this area rather than a deeper post-volcanic basin.	West-dipping Miocene synorogenic sediments are exposed east of Sacramento Pass between the northern Snake and Kern Mountains; these sediments may be present at depth beneath Snake Valley. Lake Bonneville-related lacustrine sediments are present in the valley as far south as Baker. Three wells (Hess and others, 2004) in the subbasin 4 penetrated Quaternary and Tertiary sediments, underlain in two wells by thick sections of anhydrite. Alam (1990) divided the Quaternary and Tertiary units into three groups in southern Snake Valley, the oldest related to Miocene detachment (and containing the anhydrite) and the younger two related to ongoing and subsequent high-angle normal faulting and graben formation.
Spring Valley	In northern Spring Valley, basin fill includes thick Oligocene volcanic rocks, locally derived from the vicinity of the northern Schell Creek Range (Gans and others, 1989). A source area for the Kalamazoo Tuff (Gans and others, 1989) is inferred in the northern part of Spring Valley. A small outcrop of middle Tertiary rhyolite is present in the central part of the valley.	Spring Valley is covered by Quaternary sediments; a late Pleistocene lake covered most of the valley (Reheis, 1999). A drill hole near this seismic line (Hess and others, 2004) penetrated 3,600 ft of upper Cenozoic sediments, 1,230 ft of Oligocene volcanic rocks, and 870 ft of lower Tertiary (?) sediments.
Steptoe Valley	The basin fill in portion of Steptoe Valley north of Ely includes Oligocene volcanic rocks, locally derived from Kalamazoo Pass area (Gans and others, 1989).	Eocene and Oligocene volcanic and sedimentary rocks at depth in the valley dip much more steeply than the overlying Quaternary and Miocene-Pliocene sedimentary and volcanic rocks (Gans and Miller, 1983; Smith and others, 1991). Miocene sediments are exposed only at the northernmost end of the valley; they are fine-grained, ash-bearing lacustrine units with some siliciclastic interbeds. The valley did not contain a Pleistocene lakes (Reheis, 1999).
Tippett Valley	Oligocene volcanic rocks as much as 0.6 mi-thick likely present throughout basin (Gans and others, 1989). Younger basin-fill likely to be ash-rich, similar to exposed rocks near Ibapah to the northeast.	Most of the valley contained Pleistocene lakes (Reheis, 1999).
White River Valley	Oligocene volcanic rocks commonly intercepted by oil and gas wells (Hess and others, 2004). Seismic data indicate that volcanic rocks lie near floor of basin fill.	Cenozoic units reported from drilling include Quaternary alluvium, Miocene sediments, Oligocene volcanics, and Eocene sediments (Hess and others, 2004). Pre-Eocene units are present and variably thick in all wells; the Eocene Sheep Pass Formation commonly is present but not in all wells between the volcanic rocks and the Paleozoic bedrock. No late Cenozoic lake was present in the valley (Reheis, 1999).

Consolidated Cenozoic basin-fill rocks of the older sedimentary rock unit (OSU) range from late Eocene to Miocene in age and generally underlie the more recent basin-fill deposits. Eocene-age OSU rocks include fluvial and lacustrine limestone, sandstone, siltstone, and conglomerate and have only minor volcanogenic components compared with younger basin-filling rocks (fig. 12). Unlike the older Eocene-age rocks, Oligocene-age OSU rocks contain a major volcanogenic component, including relatively thin and areally restricted fluvial and lacustrine tuffaceous limestone, sandstone, and siltstone that are interbedded with volcanic tuff and ash (Stewart, 1980). Miocene- to Pliocene-age OSU rocks contain coarse sandstone and conglomerate, volcanic-rich sediment, lacustrine sediments, and tectonic landslide or megabreccia deposits (fig. 12). These deposits formed during synextensional faulting and uplift in the study area (fig. 5) that resulted in a characteristically tilted and highly faulted heterogeneous assemblage of rocks (fig. 13). Examples of

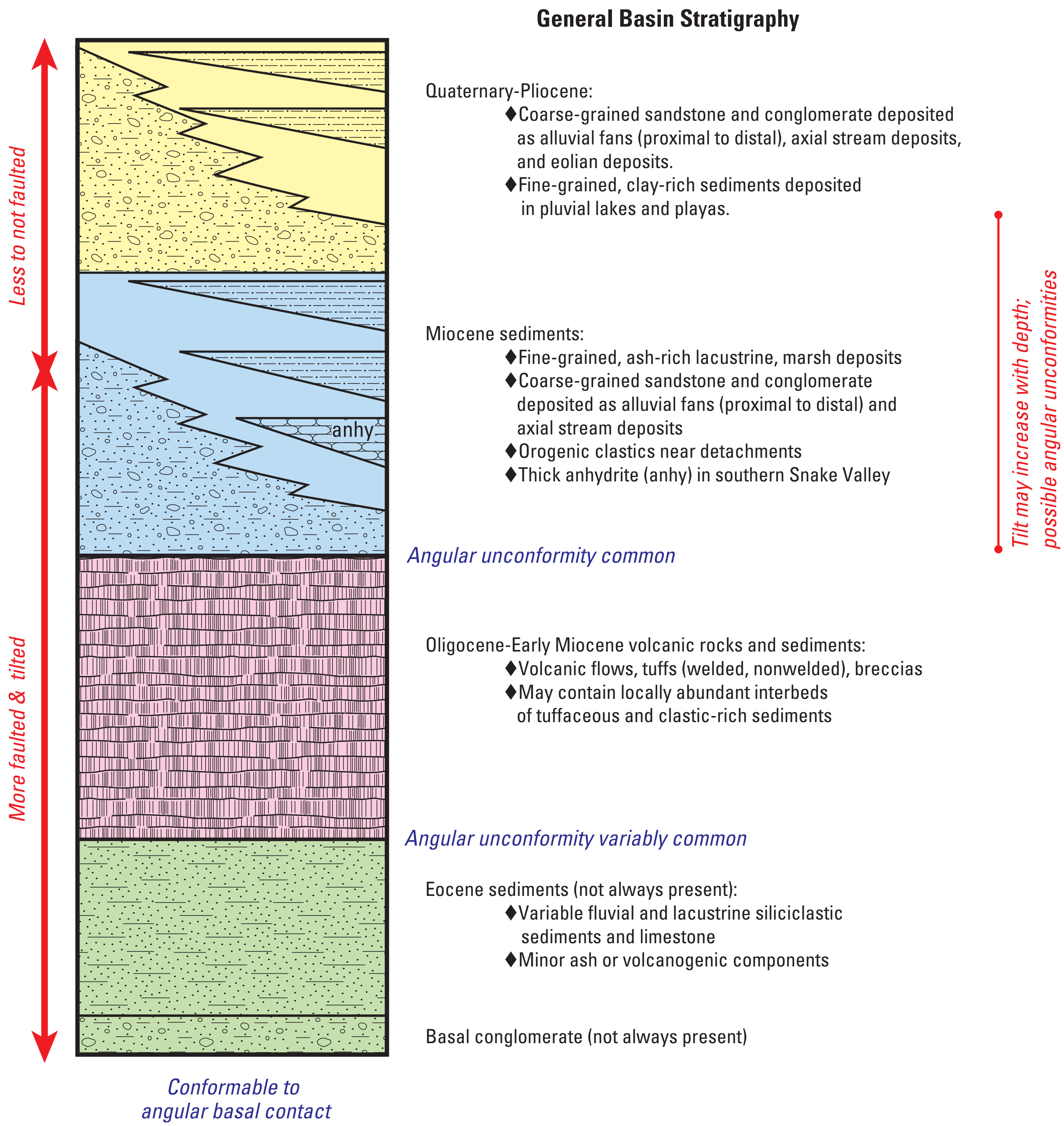
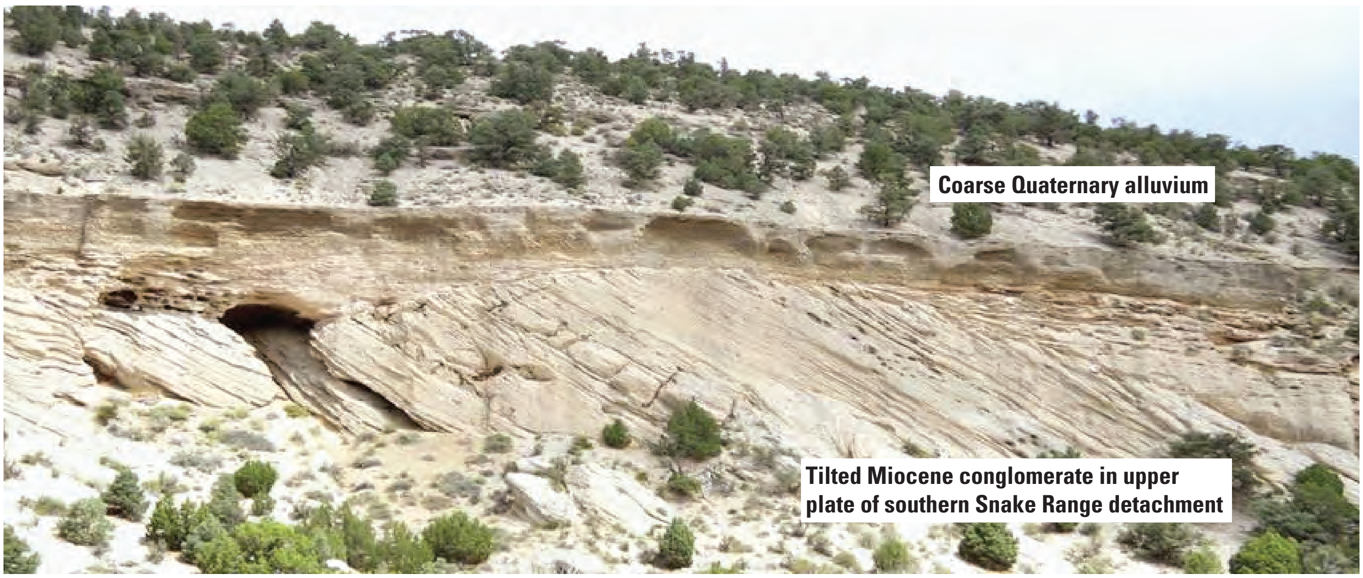


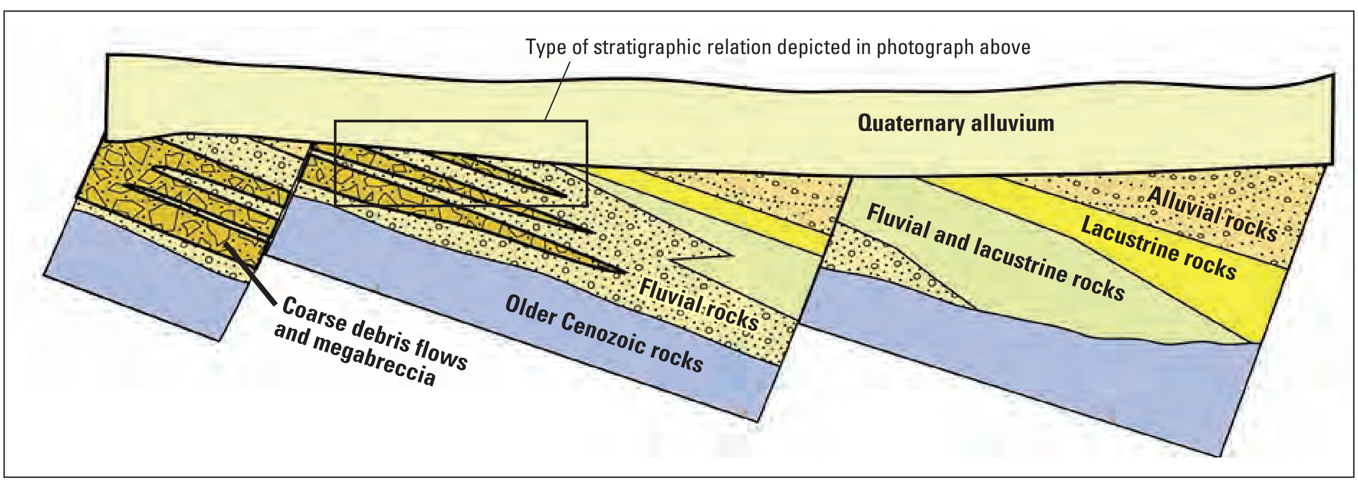
Figure 12. Generalized Cenozoic basin stratigraphy.

such synextensional basins include the sedimentary rocks in the Sacramento Pass area (Gans and Miller, 1983; Miller and others, 1999) between the northern and southern parts of the Snake Range, and the Horse Camp Formation in the northern part of the Grant Range and in Railroad Valley (Moores, 1968; Moores and others, 1968).

Analysis of rocks from southern Nevada that are similar to the OSU suggests that these consolidated rocks have significantly lower permeability than the overlying unconsolidated basin-fill deposits (Belcher and others, 2001) and could function as a low-permeability barrier between the overlying younger basin-fill and the underlying higher permeability pre-Cenozoic carbonate rocks. However, outcrops of Miocene- and Pliocene-age OSU rocks are not widespread, and probably were never thick. As a result, the lower permeability of this unit likely has minimal influence as a barrier to ground-water flow.



A. Synextensional Miocene sedimentary rocks, eastern flank of southern Snake Range, Nevada. Photograph taken by Donald S. Sweetkind, U.S. Geological Survey, September 10, 2004.



B. Schematic representation of stratigraphic variability in Cenozoic sedimentary basins.

Modified after Wallace (2005).

Figure 13. Local example and generalized stratigraphy of synextensional basins.

Holocene- to Pliocene-aged alluvium, colluvium and, in some valleys, fluvial deposits (Plume, 1996) form the unconsolidated coarse-grained younger sedimentary rock unit (CYSU). In general, these deposits predominantly consist of sandy gravel with interbedded gravelly sand, and sand. Where deposited as alluvial fans, the grain size of the CYSU gradually decreases from proximal to distal parts of the fan (Plume, 1996). Sediments of the CYSU are not commonly cemented, but are more indurated with increasing depth. These deposits, though discontinuous, are permeable aquifers, particularly alluvial fan and stream channel deposits (Belcher and others, 2001). However, in some areas, CYSU deposits may contain intercalated, less permeable finer grained sediments or volcanic ash. The fine-grained younger sedimentary rock unit (FYSU) consists of unconsolidated Holocene to Pliocene fine-grained playa and lake deposits that are widespread throughout the study area (Stewart, 1980). FYSU sediments were deposited along basin axes and, as a result, typically are mixtures of moderately to well stratified fine sand, silt, and clay of relatively low permeability and limited capacity to transmit water. Pliocene lacustrine and fluvial deposits consist of freshwater limestone, tuffaceous sandstone and siltstone, laminated clays, and water-lain tuffs and ash that include the Panaca and Muddy Creek Formations, and the White River lakebeds (Tschanz and Pampeyan, 1970). These deposits were formed by Quaternary lakes, such as Pleistocene Lake Bonneville and more local lakes in Antelope, Spring, Lake, Cave, and Jakes Valleys (Reheis, 1999).

Igneous Rocks

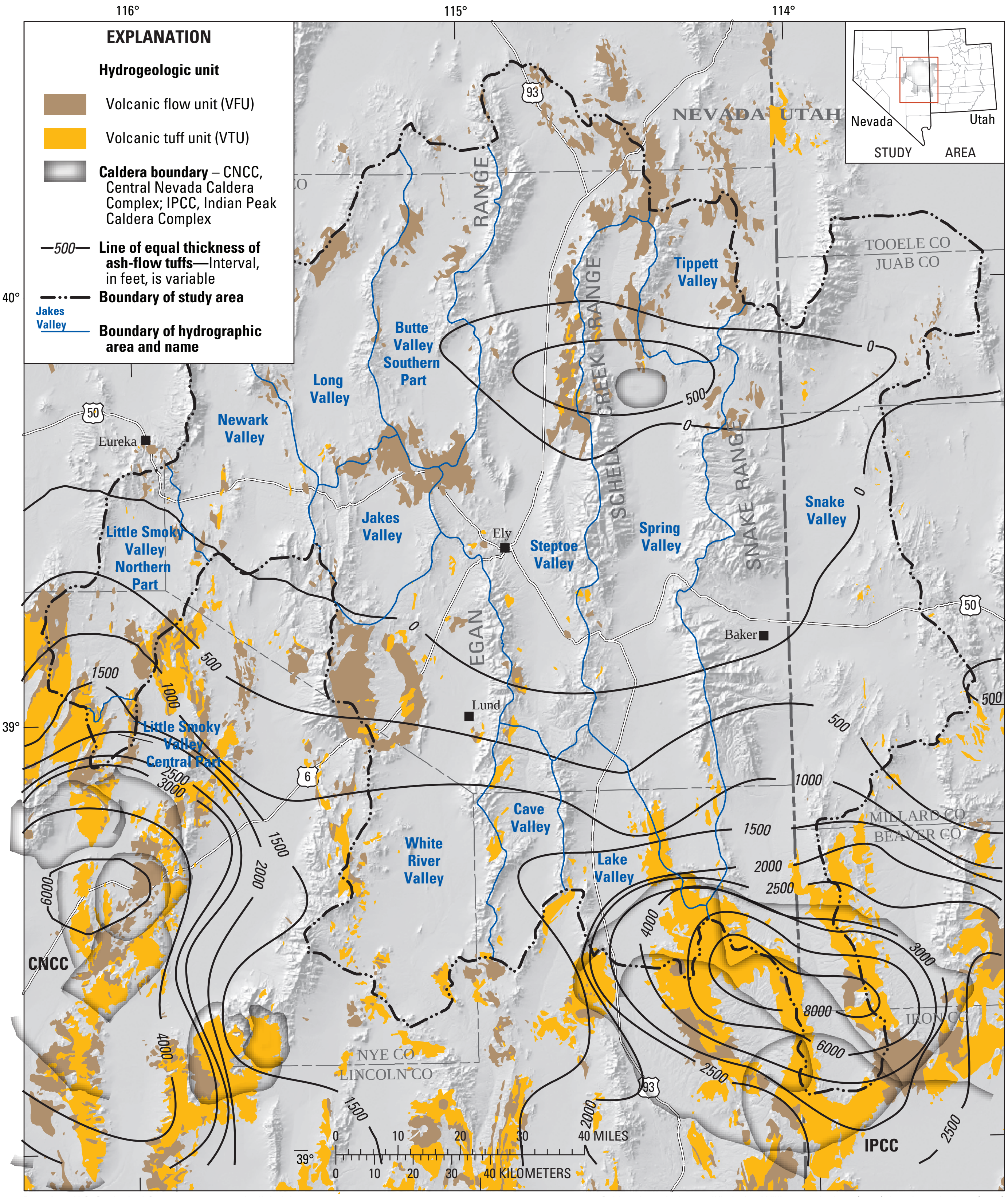
Igneous rocks in the study area consist of plutonic rocks and volcanic deposits that may be grouped into three primary HGUs—the intrusive rock unit (IU), volcanic tuff unit (VTU), and the volcanic flow unit (VFU) (table 2; fig. 3). The IU includes all Mesozoic and Cenozoic granitic plutonic rocks in the study area. The exposed or concealed plutonic rocks, typically granitic, are widely scattered, but most occur in the east and northeast parts of the study area (pl. 1). Geologic and aeromagnetic data indicate that plutonic rocks locally intrude the carbonate-rock units (LCU and UCU). Depending on how deeply the plutons are buried, granitic rocks may influence ground-water flow direction or magnitudes. Although small quantities of water may pass through these intrusive crystalline rocks where fractures or weathered zones exist, fractures in the IU typically are poorly connected. Where studied elsewhere, these rocks often impede ground-water flow (Winograd and Thordarson, 1975).

Volcanic rocks in the study area were divided into two principal HGUs (fig. 3), the volcanic tuff unit (VTU) and the volcanic flow unit (VFU). The use of these two HGUs follows the subdivision of volcanic rocks typically used on the State geologic maps. Rocks of the VTU include welded and nonwelded tuffaceous units of rhyolite-to-andesite composition; rocks of the VFU include basalt, andesite, and rhyolite lava flows. Relatively thick exposures of ash-flow tuffs occur in the southern and western parts of the study area (fig. 14), and these deposits also may be preserved in many of the intermontane valleys of the study area. The middle Tertiary volcanic rocks of east-central Nevada also include lavas and associated deposits that are a significant, though not especially voluminous, part of the geologic framework of this area.

In the southern parts of the study area, volcanic rocks, particularly densely welded tuffs of the VTU, are relatively thick and permeable over a considerable area. The thickness of the VTU is estimated to be greatest in the intra-caldera source areas for widely distributed ash-flow tuffs, such as in the Indian Peak caldera complex and in the Central Nevada caldera complex (fig. 14). In the northern half of the study area, the thickness of VTU is estimated to be relatively minor. Estimates of VTU thickness are based on an evaluation of volcanic rocks potentially preserved in down-faulted, Cenozoic graben valleys of east-central Nevada and west-central Utah. Fractured rhyolite-lava flows and moderately to densely welded ash-flow tuffs are the principal volcanic-rock aquifers. Rhyolite-lava flows (VFU) are laterally restricted, whereas welded ash-flow tuff sheets (VTU) are more widely distributed and may constitute a laterally continuous aquifer.

Aquifer Hydraulic Conductivity

Hydraulic properties can be highly non-uniform in many aquifer systems. Hydraulic conductivity is scale dependent and is affected by fracturing and chemical dissolution in the case of carbonate rocks. Consolidated rocks generally have a wider range of hydraulic conductivity compared to unconsolidated sediments. Estimates of hydraulic conductivity frequently are determined from aquifer tests in wells or boreholes. In fractured rock, at small scales on the order of inches to feet, contrasts in hydraulic conductivity result from the presence or absence of fractures. At larger scales, on the order of tens to hundreds of feet, contrasts in hydraulic conductivity arise from differences between zones of numerous, open, well-connected fractures and zones of sparse, tight, poorly connected fractures. Methods used to analyze aquifer tests that rely on simplifying assumptions is an additional complication. Violations of these assumptions may result in erroneous estimates for computed hydraulic properties (Belcher and others, 2001). Few aquifer tests have been completed in the study area and thus estimates of hydraulic properties are sparse. Because of limited data for the study area, estimates of hydraulic properties were compiled from aquifer tests in the Death Valley regional ground-water flow system (DVRFS; fig. 1; Belcher and others, 2001). Hydraulic properties for the DVRFS are considered to be representative of hydraulic properties in the study area because of similar rock types and HGUs (table 2).



Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from USGS digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

Caldera boundaries modified after Williams and others (1997), Loucks and others (1989), Raines and others (1996), Workman and others (2002), and Gans and others (1989).

Figure 14. Outcrop extent and inferred subsurface thickness of volcanic rocks, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Horizontal hydraulic conductivity (hereinafter referred to as hydraulic conductivity) values were grouped by HGU and statistically evaluated to determine the central tendency and range of values. Descriptive statistics, including the arithmetic and geometric means, median, and range of hydraulic conductivity for each HGU are shown in table 4. The arithmetic mean is the average value within the sampled data set. The geometric mean is the mean of the logarithms, transformed back to their original units, and commonly is used for positively skewed data (Helsel and Hirsch, 1992). The hydraulic conductivity was calculated by dividing estimates of aquifer transmissivity by the total saturated thickness of the aquifer material tested.

For the study area, the hydraulic conductivity for an HGU can span three to nine orders of magnitude. Carbonate and volcanic rocks typically are aquifers in the study area, however, where fractures and dissolution are largely non-existent, they are confining units. Grain size and sorting are important influences on hydraulic conductivity of the unconsolidated sediments (Belcher and others, 2001). The largest hydraulic conductivity values are associated with CYSU, VTU, UCU, and LCU. The arithmetic and geometric means are greater than or equal to 40 and 1 ft/d, respectively. The mean hydraulic conductivity of the VFU is an order of magnitude less than that for the VTU; whereas the geometric means only differ by a factor of 8 (table 4). The geometric mean of the hydraulic conductivity values of the MSU overlying the carbonate-rock aquifer, the USCU separating the upper and lower carbonate-rock aquifers, and the LSCU that underlies the carbonate-rock aquifer are a minimum of three orders of magnitude smaller than their adjacent aquifers; the LSCU that underlies the carbonate-rock aquifer has the lowest value (2.0×10^{-6} ft/d). The relatively greater hydraulic conductivity values for the FYSU, OSU, and VFU (values between those for aquifers and the aforementioned confining units) indicate that these HGUs may be semi-confining units. In some areas, these semi-confining units may be fractured to a sufficient degree to transmit water, although typically these units are not fractured and tend to retard ground-water flow.

Table 4. Hydraulic conductivity values for hydrogeologic units of the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[Description of hydrogeologic unit is given in table 2]

Major unit	Hydrogeologic unit abbreviation	Hydraulic conductivity, in feet per day					Count
		Arithmetic mean	Geometric mean	Minimum	Maximum	Median	
Cenozoic basin-fill sediments	FYSU	34	8	0.01	111	19	13
	CYSU	40	5	0.0002	431	10	43
	OSU	5	0.2	0.0001	21	0.4	15
Cenozoic volcanic rock	VFU	3	1	0.04	14	2	17
	VTU	51	8	0.09	179	37	9
Mesozoic sedimentary rock	MSU	0.07	0.006	0.0006	0.9	0.004	16
Paleozoic carbonate rock	UCU	145	1	0.0003	1,045	3	12
	USCU	0.4	0.06	0.0001	3	0.1	22
	LCU	169	4	0.009	2,704	4	45
	LSCU	0.8	0.000002	0.00000009	15	0.0000003	19
	IU	0.8	0.03	0.002	5	0.01	7

Hydrographic Area Boundaries and Intrabasin Divides

The distribution of aquifers and confining units along HA boundaries and intrabasin divides is a principal control on interbasin and intrabasin ground-water flow in the study area. The occurrence and juxtaposition of aquifers and confining units in these areas must be understood to assess the geologic controls on the relative potential for ground-water flow across these boundaries and divides. For example, ground-water flow across HA or subbasin boundaries may not be possible if one or more permeable HGUs are not present, or may not be likely if the minimum hydraulic conductivity of juxtaposed aquifers and confining units is relatively low.

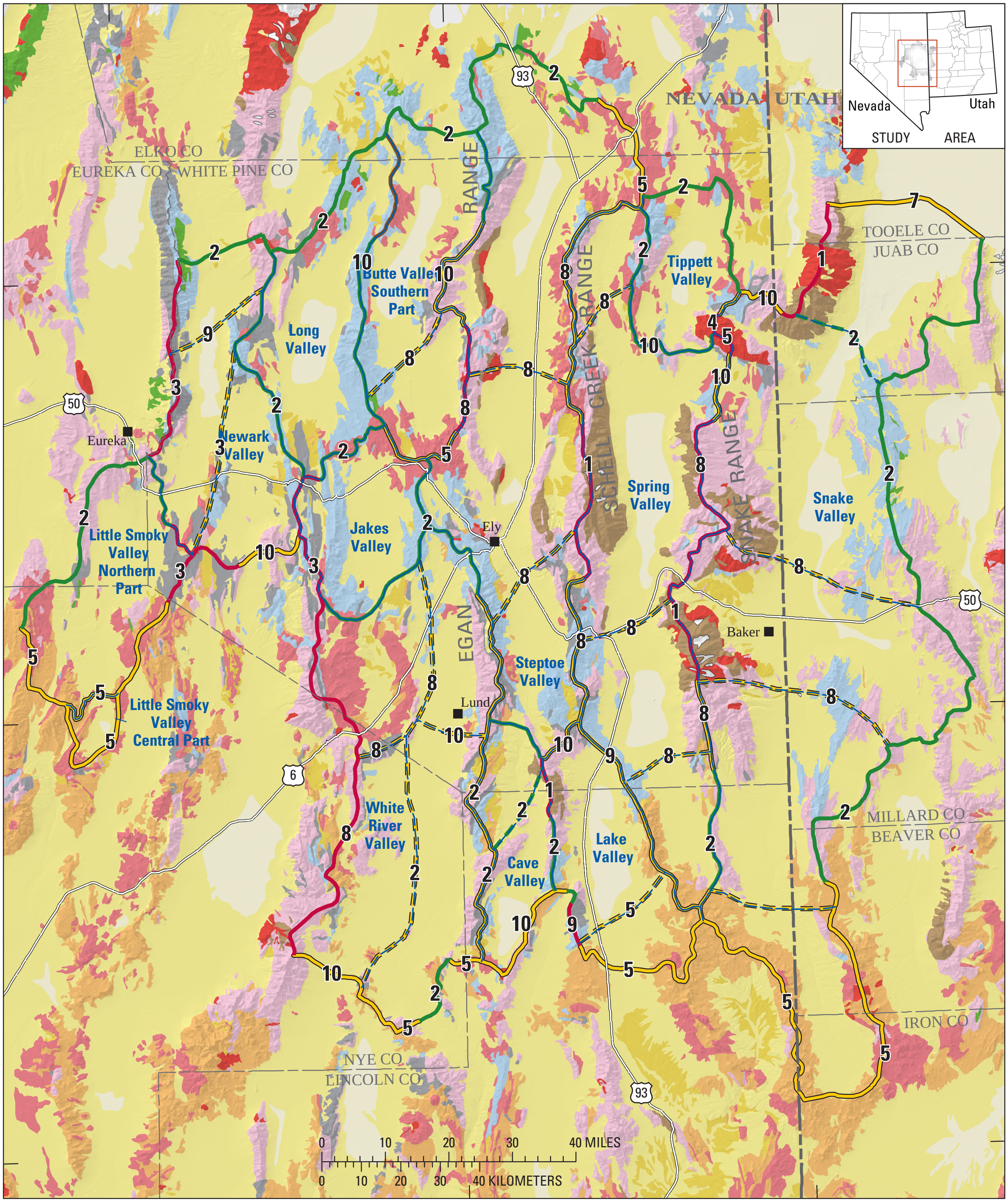
To assess the geologic controls on the potential for ground-water flow across HA boundaries and intrabasin divides, the stratigraphic and structural features described previously were integrated with subsurface geophysical data to categorize rocks into 1 of 10 general subsurface boundary conditions that are likely to result in differing ground-water flow characteristics. Each boundary condition represents the likely influence of one or more HGUs or structural conditions on ground-water flow along or across HA or intrabasin divides. The evaluation of boundary conditions primarily is based on the interpreted presence, juxtaposition, and average hydraulic properties of specific HGUs; degree of structural disruption is considered an important but secondary control. Each HA boundary and intrabasin divide was represented as a vertical, irregularly bending cross section. Relative differences in primary or secondary permeability and the mean hydraulic conductivity for HGUs were assumed to be constant along each boundary cross section. Structural disruption is considered as a boundary condition where closely spaced high-angle normal faults disrupt a relatively broad region and where carbonate-rock aquifers are highly faulted and disrupted in the upper plates of low-angle normal faults. Because few data are available, however, the categorization does not incorporate the effects of individual faults as distinct hydrologic entities. For example, the analysis omits potential effects of impermeable, clay-rich fault core zones, fractured and potentially more permeable zones that might lie outside of the fault core, or stratabound fractured intervals in volcanic or carbonate rocks. The occurrence of each subsurface boundary condition varies throughout the study area; for example, boundaries with LCU or UCU rocks occur in many HAs and subbasins; boundaries with FYSU or CYSU deposits are limited and absent in the study area, respectively. For each of the 10 subsurface boundary conditions, the potential for ground-water flow was evaluated in one of three ways (fig. 15)—(1) permeable rocks are likely to exist at depth such that ground-water flow likely is permitted by subsurface geology, (2) relatively impermeable rocks are likely to exist at depth such that ground-water flow likely is not permitted by subsurface geology, or (3) the subsurface geology beneath the boundary or divide is not well constrained or the nature of the subsurface framework is highly uncertain such that the geologic controls on ground-water flow are uncertain. The rationale for each of the 10 subsurface boundary conditions shown in figure 15 is described in the following paragraphs:

1. *Impermeable bedrock (LSCU) in subsurface*—Subsurface geologic conditions likely limit ground-water flow through HA boundaries identified as having impermeable bedrock in the subsurface. All these boundaries correspond to high-standing blocks of LSCU or its metamorphosed equivalent in the lower plate of detachment faults in the Snake, Schell Creek, Deep Creek, and Grant Ranges. In these areas, the LSCU is inferred to extend to great depths, with no aquifer units present.

2. *Thick permeable Paleozoic carbonate rocks (LCU or UCU) in subsurface*—Subsurface geology permits ground-water flow at HA boundaries or intrabasin divides identified as having relatively thick sections of permeable Paleozoic carbonate rocks (LCU or UCU) in the subsurface. Carbonate rocks with this boundary designation occur along the northwestern and eastern boundaries of the study, and in the Egan Range, Butte Mountains, White Pine Range, and southern Snake Range (pl. 1). Two of these boundaries are along the crest of the Egan Range in the center of the study area where Paleozoic carbonate rocks are exposed at the surface along the range front. The likelihood of flow across these boundaries is dependent on the altitude of the contact between the LCU and underlying LSCU relative to the ground-water table.

3. *Thick Chainman Shale (USCU) present in subsurface*—Subsurface geologic conditions likely limit ground-water flow crossing HA boundaries identified as having thick intervals of Chainman Shale (USCU) in the subsurface. All these boundaries are in the western part of the study area in the vicinity of the White Pine Range, the Pancake Range, and the Diamond Mountains. In many cases, the USCU dips steeply or is folded and as a result the subsurface extent of the USCU can be greater than the stratigraphic thickness of the Chainman Shale. Most of these boundaries were designated as subsurface geology that would not likely permit ground-water flow; however, one boundary corresponds to a buried bedrock high within Newark Valley where ground-water flow is designated as possible because the subsurface conditions are not well constrained. Because the LCA underlies this HGU, it is possible, given appropriate hydraulic head, that ground water could move across these boundaries through the underlying carbonate-rock aquifers.

4. *Pluton (IU) present in subsurface*—The HA boundary along the Kern Mountains (pl. 1) is underlain by plutonic igneous rocks (ICU) in the subsurface. Given that the igneous rocks are inferred to persist to great depths, ground-water flow likely does not cross this boundary.



Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

Figure 15. Characterized hydrographic area boundaries and surface geology, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

EXPLANATION FOR FIGURE 13

Hydrogeologic unit

- FYSU–Fine-grained younger sedimentary rock unit (primarily lacustrine and playa deposits)
- CYSU–Coarse-grained younger sedimentary rock unit (alluvi and fluvial deposits)
- OSU–Older sedimentary rock unit (consolidated Cenozoic ro
- VFU–Volcanic flow unit (basalt, andesite, dacite and rhyolite lava flows)
- VTU–Volcanic tuff unit (ash-flow tuffs)
- MSU–Mesozoic sedimentary rock unit
- UCU–Upper carbonate rock unit (Mississippian to Permian carbonate rocks)
- USCU–Upper siliciclastic rock unit (Mississippian siliciclasti rocks)
- LCU–Lower carbonate rock unit (Cambrian to Devonian predominantly carbonate rocks)
- LSCU–Lower siliciclastic rock unit (Early Cambrian and older siliciclastic rocks)
- IU–Intrusive unit

Boundary of hydrographic area and name

Jakes Valley

- Flow not permitted by subsurface geology
- Flow permitted by subsurface geology
- Flow possible; boundary not well constrained

Intrabasin bedrock high

- Flow permitted by subsurface geology
- Flow possible; boundary not well constrained

Explanation of numerical codes on boundary lines

Boundary code	Interpreted subsurface geologic unit
1	Impermeable bedrock (LSCU) in subsurface
2	Thick permeable Paleozoic carbonate rocks (LCU or UCU) in subsurface
3	Thick Chainman Shale (USCU) present in subsurfa
4	Pluton (IU) present in subsurface
5	Thick volcanic rocks (VTU or VFU) present in subsurface
6	Thick permeable basin fill (CYSU) in subsurface
7	Thick impermeable basin fill (FYSU) in subsurface
8	Permeable rocks (LCU or UCU) overlie shallow detachment fault
9	Thin Chainman Shale (USCU) present in subsurfac
10	Structural disruption may permit subsurface flow

5. *Thick volcanic rocks (VTU or VFU) present in subsurface*—Subsurface geologic conditions are characterized as uncertain across HA boundaries identified as having thick sections of Cenozoic volcanic rock (VFU or VTU) in the subsurface. Volcanic rocks with this boundary designation occur in the southeastern and southwestern part of the study area, near Lake Valley and Little Smoky Valley, respectively, and at the divide between Butte Valley and Jakes Valley. All these accumulations of volcanic rocks may have a wide range of aquifer properties and, as a result, the nature of these boundaries, and their influence on ground-water flow, remains uncertain without specific, more detailed information on hydraulic properties of volcanic HGUs.

6. *Thick permeable basin fill (CYSU) in subsurface*—In the study area, there were no HA boundaries or intrabasin divides categorized as underlain by a relatively thick section of permeable basin fill (CYSU).

7. *Thick impermeable basin fill (FYSU) in subsurface*—Subsurface geologic conditions are characterized as uncertain, along the HA boundary adjacent to the Great Salt Lake Desert in the far northeastern part of the study area. This part of the study area is underlain by thick, impermeable basin fill (FYSU) in the subsurface. The potential for ground-water flow across this boundary is uncertain because of the lack of specific subsurface information on the nature of the sedimentary section.

8. *Permeable rocks (LCU or UCU) that overlie a shallow detachment fault*—Ground-water flow is possible, but uncertain, across HA boundaries identified as having permeable carbonate rocks (LCU or UCU) overlying a shallow detachment fault. All these segments are associated with detachment faults in the Cherry Creek, Egan, Grant, Snake, and Schell Creek Ranges where the lower plate beneath the detachment faults may not be exposed but whose presence in the shallow subsurface reasonably is inferred. In these areas, the upper plate consists of highly faulted carbonate rocks that may have enhanced permeability caused by the structural disruption. However, ground-water flow likely is not permitted across four HA boundaries in the northern Snake Range, the Grant Range, and the northern Egan Range that correspond to well-exposed detachment faults and highly disrupted upper plate rocks. These boundaries mostly are in areas where the detachment fault must be projected some distance in the subsurface and are thus subject to greater uncertainty.

9. *Thin Chainman Shale (USCU) present in subsurface*—The geologic controls on the potential for ground-water flow varies across three HA boundaries identified as having thin intervals of Chainman Shale (UCU) in the subsurface. Ground-water flow likely is not permitted across the HA boundary at Grassy Pass, south of Dutch John Mountain on the west side of Lake Valley (pl. 1) because of the gentle northward dip of the Chainman Shale. Subsurface geologic conditions are less certain and flow is possible across the HA boundary along the Fortification Range and Lake Valley Summit at the northern and northeastern part of Lake Valley because the thickness and continuity of the Chainman Shale in this area are uncertain. Subsurface geologic conditions also are categorized as uncertain across the buried bedrock high that transects the northern part of Newark Valley. The bedrock high consists of structurally disrupted shales that may allow ground water to flow parallel to the general northern strike of these rocks.

10. *Structural disruption may permit subsurface flow*—Except for one boundary, the subsurface geologic conditions are categorized as uncertain across HA boundaries identified as having significant structural disruption, regardless of rock type. Several of these boundaries lie atop highly faulted and potentially permeable bedrock outcrops; however, the subsurface framework for these areas is uncertain. Structurally disrupted areas occur in the southern part of the Schell Creek Range to the north of Mount Grafton, to the south of the Kern Mountains, the Cherry Creek Range, and along the west side of the White Pine Range (pl. 1). Ground-water flow likely is permitted across the HA boundary between Spring and Tippett Valleys, where numerous north-striking faults may serve as conduits for ground-water flow.

Intrabasin divides represent locations where the basin-fill aquifer is interrupted by buried structural highs of pre-Cenozoic bedrock; however, these areas are not necessarily barriers to ground-water flow. The intrabasin divides were evaluated using the same rationale used to classify the HA boundaries. A much greater level of uncertainty exists in envisaging the subsurface geology and potential hydraulic effects across intrabasin divides (fig. 15). Except for one area, all intrabasin divides in the study area are interpreted as ground-water flow being possible across these divides, but uncertain because the subsurface geologic framework is not well constrained. Two of these intrabasin divides, in Lake Valley and in southern Snake Valley, were located at the buried northern margin of the Indian Peak caldera complex, even though the pre-Cenozoic surface does not show significant changes in topography. In these areas, relatively thick accumulations of volcanic rocks closer to the caldera likely influence ground-water flow differently than volcanic rocks interbedded with basin fill and farther away from the calderas. However, ground-water flow likely crosses an intrabasin divide near the northern part of Snake Valley (fig. 15) where carbonate rocks occur beneath the basin-fill aquifer.

Ground-Water Conditions

By Lari A. Knochenmus[1], Randell J. Laczniak[1], Michael T. Moreo[1], Donald S. Sweetkind[1], J.W. Wilson[1], James M. Thomas[2], Leigh Justet[1], Ronald L. Hershey[2], Sam Earman[2], and Brad F. Lyles

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Ground water in the study area is influenced by a combination of topography, climate, and geology. Ground water moves through permeable zones under the influence of hydraulic gradients from areas of recharge to areas of discharge, and this movement can be discussed in terms of local, intermediate, and regional flow systems (fig. 16). These ground-water terms are adopted from the terminology developed by Toth (1963) and Freeze and Cherry (1979), and are defined on the basis of depth of ground-water flow and length of the flow path. Local flow systems are characterized by relatively shallow and localized flow paths that terminate at upland springs. Local springs are low volume, tend to have temperatures similar to annual average ambient atmospheric conditions and have discharge that fluctuates according to the local precipitation. Intermediate flow systems include flow from upland recharge areas to discharge areas along the floor of the intermontane valley. Within intermediate-flow systems, springs typically discharge near the intersection of the alluvial fan and the valley floor near the range front. Intermediate-flow system springs often are of moderate volume and tend to have less-variable flow relative to local springs.

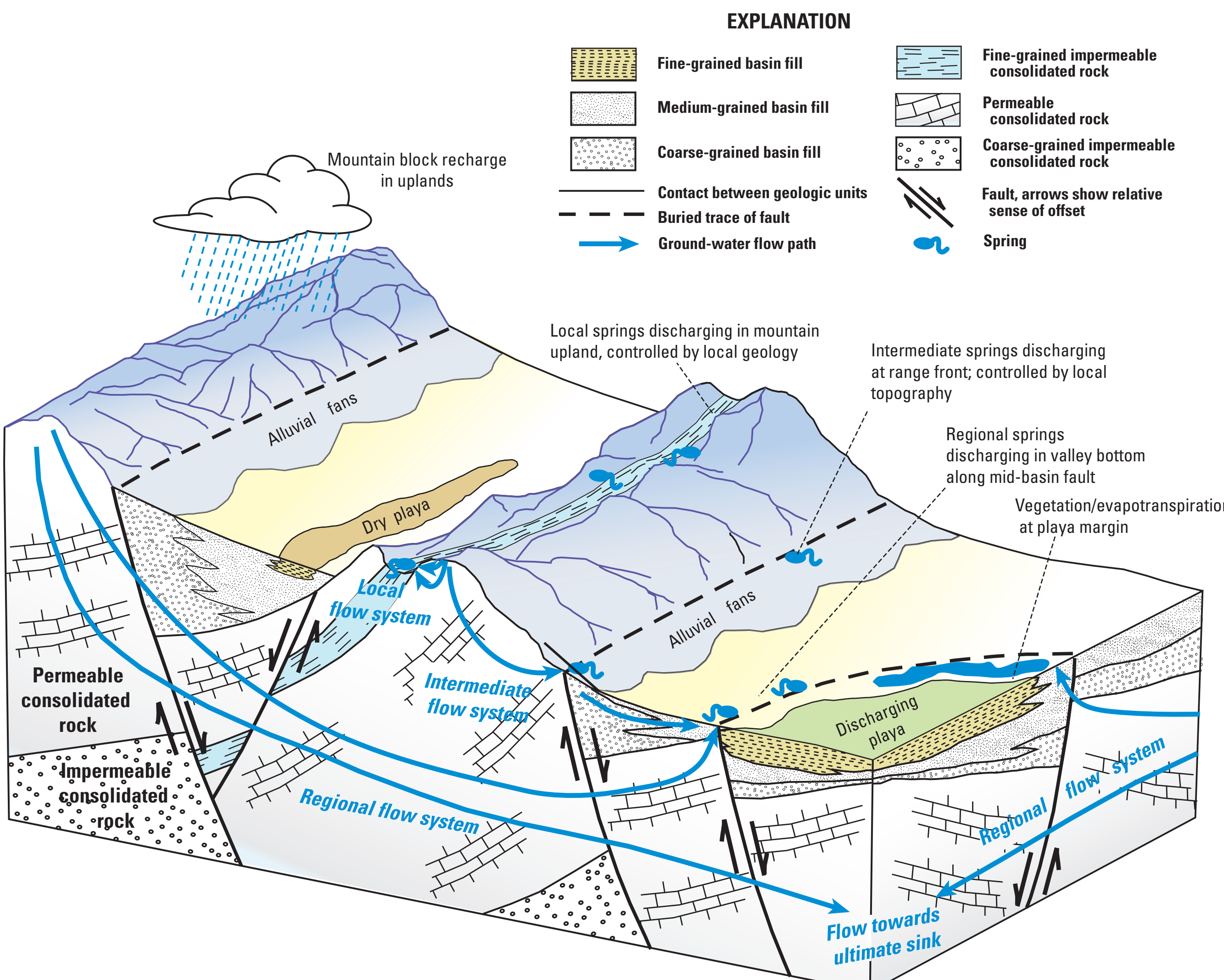


Figure 16. Conceptual ground-water flow systems.

Regional ground-water flow follows large-scale (tens to hundreds of miles) topographic gradients as water moves toward low altitudes in the region. Discharge from these regional flow systems manifests as large springs and, in some areas, extensive wetlands (Mendenhall, 1909). Meinzer (1911) recognized that certain large volume springs in the eastern Great Basin can not be supported by the available recharge from local surrounding mountain ranges, and that the flow from these springs must be supported in part from regional ground-water flow originating outside the basin. Based on chemistry, temperature, and other criteria, Mifflin (1968) identified some springs likely discharging interbasin flow, including Hot Creek in White River Valley and McGill Spring in Steptoe Valley. Regional ground-water flow is driven by hydraulic gradients that are continuous over long distances. Deep regional flow through basin-fill or consolidated bedrock aquifers is unconstrained by local topographic or drainage features. Under pre-development conditions, recharge to the regional ground-water flow system primarily originates in mountains and may travel beneath several basins and through multiple mountain ranges before reaching its ultimate discharge area.

Inputs to a ground-water system include direct recharge from precipitation, infiltration from lakes and streams, flow from an adjacent ground-water system, and recharge from human activities such as agricultural irrigation. Recharge is most prominent where water percolates into fractures in the bedrock of the mountain uplands and where streamflow infiltrates underlying or adjacent bedrock or alluvium at the range front or in the valleys (Harrill and Prudic, 1998).

Ground-water outputs from a basin include discharge from springs, discharge to streams and lakes, evapotranspiration (ET), flow across a ground-water flow system boundary to an adjacent system, and pumping for various uses. Activities such as ground-water pumping for agricultural uses and human consumption remove water from storage in a ground-water system and thereby reduce hydraulic heads, which are measured as ground-water levels in open wells. Ground-water pumping also can affect streams or springs in direct hydraulic connection with the ground-water system because declining ground-water levels can lead to increased recharge from streams and decreased springflow.

Areas of recharge and discharge were used as secondary data to develop water-level maps of hydraulic heads for shallow basin fill and deeper aquifers in the study area. Moreover, to better characterize these aquifers, water in storage was estimated for a representative volume of aquifer, and water-quality data were compiled and collected to assess water quality relative to primary and secondary drinking-water standards.

Ground-Water Flow

Ground-water flow was evaluated using a water-table map of the basin fill and a potentiometric-surface map of the regional carbonate-rock aquifer. The water table and potentiometric surface maps primarily were based on measured ground-water levels in wells. Water table and potentiometric-surface maps published in previous reports were used as secondary guides for developing these maps, particularly in areas where data were sparse (Mifflin, 1968; Hess and Mifflin, 1978; Garside and Schilling, 1979; Johnson, 1980; Pupacko and others, 1986; Thomas and others, 1986; and Bedinger and Harrill, 2005). Data used to develop the water-table and potentiometric-surface maps are summarized in Wilson (2007).

The water-table map was interpreted from water-level measurements for 299 wells completed in the basin-fill aquifer, and guided by geology, and known areas of recharge and natural ground-water discharge (pl. 2). Water-level altitudes above sea level ranged from less than 4,400 ft in northern Snake Valley to more than 6,800 ft in southern Steptoe Valley. Ground water in the basin fill generally flows from mountain fronts along the margin of valleys to the center of valley floors. Internally drained HAs, where water is lost by evaporative discharge, have closed, or nearly closed contours on the valley floors on plate 2. In some HAs, ground water in the basin fill flows parallel to the mountain front and toward the basin boundary, such as ground-water flow to the north in Steptoe and Snake Valleys and to the south in White River and Cave Valleys.

The potentiometric-surface map was developed using water levels measured in 119 wells (pl. 3). Because the number of wells completed in the deeper carbonate-rock aquifer are relatively sparse, the potentiometric-surface map of this aquifer represents a composite of water-level measurements for wells completed in basin fill (76 wells) and deeper geologic units including carbonate rocks (43 wells). Water levels measured in the basin fill wells were considered appropriate for mapping the potentiometric surface because there is regional hydraulic continuity between deep and shallow flow regimes (Bedinger and Harrill, 2005). Water-level altitudes ranged from less than 4,500 ft in northern Snake Valley to more than 6,500 ft in Steptoe Valley.

The source of ground water in the carbonate-rock aquifer within the study area is a relatively large recharge mound centered on the Snake, Schell Creek, and Egan Ranges (pl. 3). The recharge mound forms ground-water divides that separate the study area into multiple flow systems. Ground water in the carbonate-rock aquifer flows radially from these recharge areas to a number of HAs that form the headwaters of four regional flow systems. Ground water in west-central Steptoe Valley flows into Jakes and White River Valleys. Ground-water flow is toward the south in Long, Jakes, White

River, and Cave Valleys and is part of the Colorado regional flow system. Ground water in southern Steptoe Valley flows into Lake Valley and then moves east into Spring and Snake Valleys as part of the Great Salt Lake Desert regional flow system. Flow generally is toward the north-northeast in northern Steptoe, Tippett, and Snake Valleys. Although Butte Valley is considered part of the Goshute Valley regional flow system (Harrill and others, 1988), ground-water likely exits this valley to the north as part of the Ruby Valley flow system. Some regional ground water moves upward into overlying basin-fill sediments, such as in southern White River Valley and south-central Spring Valley, or is discharged from valley floor springs.

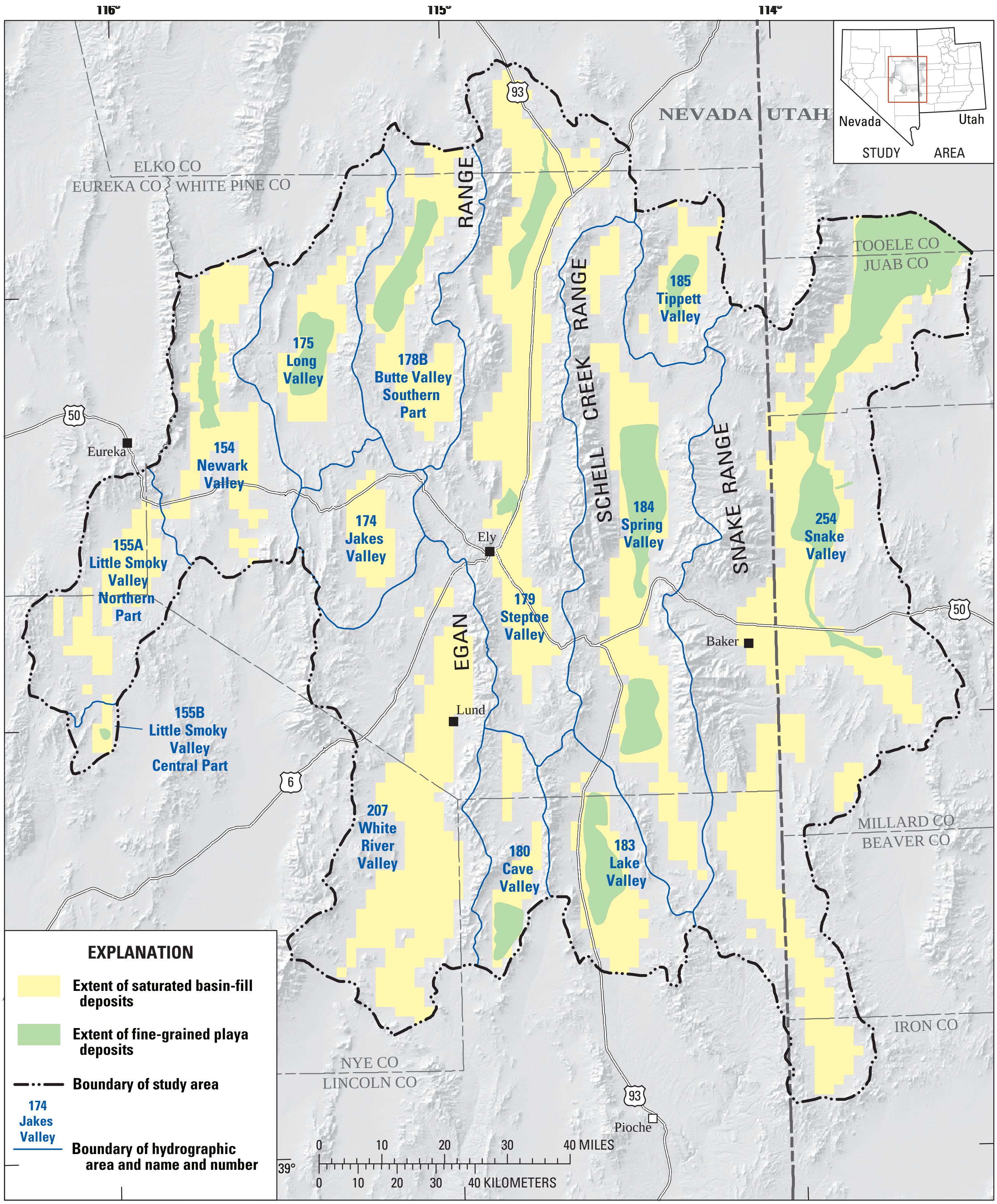
Volume of Water Stored in Aquifers

Water stored within aquifers becomes available as ground water is pumped and water levels decline. Water removed from storage by pumping commonly is referred to as "ground-water storage." When pumping ceases, water levels will not recover to previous levels if the amount of water removed is not replaced by an equal amount or if the declines may have altered the hydraulic or physical properties of the aquifer. The magnitude of water-level decline or recovery depends, in part, on the storage properties of the aquifer; that is, on whether ground water is unconfined (a water-table aquifer) or confined. Storage in a water-table aquifer represents the volume of water stored within the pore spaces of saturated unconsolidated sediment or rock that becomes available as the water table is lowered and the sediment drains. Under water-table conditions, storage is the product of the area of sediment or rock drained, the magnitude of the water-level decline in the drained area, and the specific yield of the drained sediment. Specific yield is limited by the porosity of the saturated sediment, but usually is less than the sediment porosity because some stored water is tightly bound to the sediment grains or the rock, preventing complete drainage of the pore water. For the study area, storage in the water-table aquifer is estimated as the water removed from basin-fill sediments under a specified decline in water level.

Storage in a confined aquifer represents the volume of water released as hydraulic head in the aquifer decreases, water expands, and sediment or rock material compresses. Under confined conditions, storage is the product of the area of confined aquifer where hydraulic heads are lowered, the magnitude of the hydraulic-head decline in the affected area, and the storage coefficient of the confined aquifer. In confined aquifers, the storage coefficient typically is between two to four orders of magnitude less than the specific yield.

Estimates of ground-water storage in water-table and confined aquifers in the study area are developed using the extent and thickness of basin-fill deposits, a specified water-level or hydraulic head decline, and estimates of specific yield or storage coefficient. The extent of saturated basin-fill deposits (fig. 17) is assumed to be equal to the area where basin-fill thickness exceeds 100 ft. The actual area of drainable basin fill is computed as the difference in area between saturated basin fill and fine-grained playa deposits (fig. 17 and appendix A). The subsurface extent of fine-grained playa deposits is assumed to be equivalent to the fine-grained marsh, playa, and alluvial-flat deposits delineated on the generalized geology map (pl. 1). The estimated acreage of drainable basin fill ranges from less than 100,000 acres for Cave, Jakes, Lake, Long, or Tippett Valleys to more than 350,000 acres for Snake, Steptoe, or White River Valleys. Snake Valley has the largest estimated acreage of drainable basin fill at nearly 600,000 acres (appendix A).

Ground-water storage estimates for each of the HAs in the study area are computed as the sum of the estimated unconfined and confined storage. A storage estimate including both unconfined and confined contributions accounts for potential pumping from the basin fill and carbonate-rock aquifers. Storage estimates (fig. 18 and appendix A) assume water-level and hydraulic-head declines of 100 ft, an average specific yield of 0.15, and an average storage coefficient of 0.0001. Storage estimates computed using these criteria range from less than 1 million acre-ft for Cave, Jakes, or Tippett Valleys to more than 3.5 million acre-ft for Snake, Spring, Steptoe, or White River Valleys. Storage estimates for the remaining HAs, Butte, Lake, Little Smoky, Long, and Newark Valleys, range from about 1.1 to 2.3 million acre-ft. Snake Valley has the largest estimated storage at nearly 9 million acre-ft. Unconfined storage accounts for more than 99 percent of the total storage estimated in any HA, whereas confined storage accounts for less than about 10,000 acre-ft of the total storage in any HA. Storage over the entire study area is estimated as described above, at about 36 million acre-ft, of which only about 30,000 acre-ft is contributed by storage from the confined system (appendix A). Storage, estimated by this procedure, is nearly linearly proportional to the decline in water level or hydraulic head and to the magnitude of the specific yield or storage coefficient. Water level and head declines of 100 ft were arbitrarily selected, but are considered reasonable to estimate ground-water storage and show linear relations between water-level declines and specific storage, and between head declines and storage coefficient. Estimates of storage do not account for any limiting geologic, hydrologic, or cultural factors, such as impermeable boundary conditions, recharge to basin fill or carbonate-rock aquifers, changes in water quality, or potential declines in springflow or water-level declines.



Base from U.S. Geological Survey digital data 1:100,000, 1978–89.
Universal Transverse Mercator Projection, Zone 11, NAD83.
Shaded relief base from 1:250,000-scale Digital Elevation Model;
sun illumination from northwest at 30 degrees above horizon.

Figure 17. Distribution of estimated extent of saturated basin-fill deposits and fine-grained playa deposits used to estimate storage in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

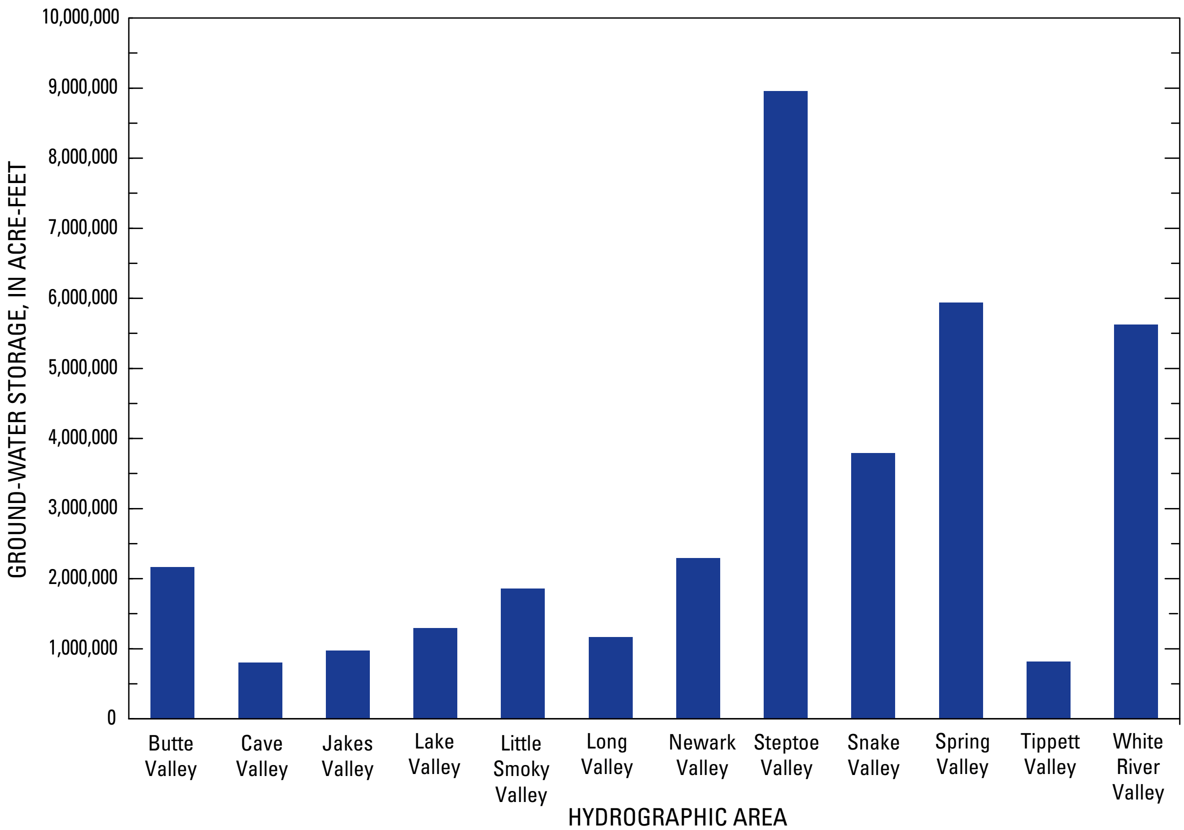


Figure 18. Ground-water storage estimates by hydrographic area based on a 100 ft lowering of water levels beneath valley floors, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Ground-Water Quality Relative to Drinking-Water Standards

Existing ground-water quality data were compiled from a number of sources for the study. These sources include the USGS National Water Information System (NWIS; http://waterdata.usgs.gov/nwis), Desert Research Institute data bases, and published reports (Bateman, 1976; Kirk and Campana, 1988; Pupacko and others, 1989). Additionally, geochemical samples were collected as part of the study from wells and springs in a number of HAs. Based on these data, and on a subset of constituents with health-based U.S. National primary drinking-water standards (U.S. Enviromental Protection Agency, 2004), ground water in the study area generally is of good quality (table 5). For chemical constituents with available analyses from more than 25 sampling sites, only arsenic and fluoride exceeded their primary standards at more than 1 site. Non-health related secondary drinking-water standards were exceeded more commonly than the primary standards. Values of pH were outside of the acceptable range of 6.5–8.5 at 21 of 179 sites. Chloride and sulfate exceeded their secondary standard at six and four sites, respectively. Except for chloride, an obvious spatial distribution of a constituent exceeding the primary or secondary standard was not apparent. Chloride exceeded the secondary standard at 7 of 10 sites in northern Snake Valley.

Only a small number of ground-water samples from the study area have been analyzed for anthropogenic organic compounds. Schaefer and others (2005) discuss the results of a broad range of organic constituents, including volatile compounds, and pesticides and their metabolites, in samples that included the study area. The study by Schaefer and others (2005) reports low concentrations of pesticides or their metabolites, and no volatile organic compounds were detected.

Table 5. Summary of exceedances of drinking-water standards for chemical constituents with available analyses from more than 25 sampling sites, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[Drinking-water standards: All values are in miligrams per liter except for pH, which is in standard units. –, no standard]

Constituent	Drinking-water standards		Number of sampling sites	
	Primary	Secondary	With constituent	Exceeding standard
Antimony	0.006	–	112	0
Arsenic	0.01	–	90	2
Barium	2	–	146	0
Beryllium	0.004	–	146	1
Cadmium	0.005	–	147	0
Chloride	–	250	179	6
Chromium	0.1	–	54	0
Copper	–	1	38	0
Fluoride	4	–	122	4
Iron	–	0.3	37	2
Manganese	–	0.05	48	2
pH	–	[1]6.5-8.5	179	21
Selenium	0.05	–	35	0
Sulfate	–	250	177	4
Thallium	0.002	–	112	0
Zinc	–	5	147	1

[1]Acceptable range for pH.

Ground–Water Budgets

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A basic way to evaluate the occurrence and movement of ground water in an aquifer system is to develop a water budget accounting for the aquifer system's inflows and outflows. Water budgets may be developed for aquifer systems of any size, and for this study, water budgets were developed at the subbasin, HA, and study-area scales. Previous estimates of water budgets for HAs in the study area are summarized and compared to water-budget estimates developed as part of the current study. Estimates of average annual recharge and ground-water discharge were developed at the subbasin scale for the study. These estimates were tabulated and summed to develop water budgets; additionally, water-budget estimates for HAs were summed to determine total average annual recharge and ground-water discharge for the entire study area. Differences in estimated recharge and ground-water discharge at subbasin and HA scales were used to evaluate intrabasin and interbasin ground-water flow, respectively.

Previous Ground-Water Recharge and Discharge Estimates

During the 1960s and 1970s, the USGS in cooperation with the State of Nevada, completed a series of reconnaissance studies to evaluate the ground-water resources of Nevada. The results of these studies were published in a series of reports describing the water resources of Nevada by HA. Each report provides estimates for some or all major water-budget components and most provide estimates of average annual recharge. The reconnaissance reports applied similar approaches for estimating recharge and discharge.

Estimates of recharge presented in reconnaissance reports typically were based on a method developed by Maxey and Eakin (1949) that has been applied to more than 200 basins in Nevada (table 6). The Maxey-Eakin method empirically relates recharge to annual precipitation by trial and error adjustments of the "recharge efficiencies" to generate a balance between estimated recharge and estimated discharge in 13 HAs in east-central Nevada (Maxey and Eakin, 1949; Dettinger, 1989). Recharge efficiency is the percentage of total precipitation in the recharge-source areas of a basin that becomes recharge on a long-term average basis (Dettinger, 1989). The method assumes that higher altitudes that receive greater precipitation have a greater percentage of precipitation that becomes recharge (Eakin, 1966). Five precipitation zones were defined by this method from the Hardman (1936) precipitation map of Nevada. Each of the five precipitation zones has an associated recharge efficiency. Recharge to a basin was estimated from the precipitation rate for each of the five zones, applying the associated recharge efficiency, and summing these values to obtain the total recharge rate.

Ground-water discharge typically was estimated using volumetric calculations of mean annual ET for areas of phreatophytic vegetation (table 7). In most of the HAs in Nevada, ground water is discharged by evaporation from free-water surfaces and soils, and transpiration by phreatophytes where the water table is at or near land surface (Eakin, 1962). ET estimates were based on maps delineating land-cover classes and coefficients relating the classes and ground-water discharge rates determined from pan-evaporation and lysimeter data. Ground-water discharge for an HA was estimated by computing the product of the ET rates and the corresponding area for a particular land cover, and integrating the products for all land cover classes in the HA. The volume of water used for irrigation and self-supply was small and usually neglected in water-budget computations. Springflow typically was not accounted for directly in the water budget but was accounted for indirectly in the ET estimate (Eakin, 1960). In some reconnaissance studies, however, ground-water discharges were not determined independently but were assumed to be equal to the Maxey-Eakin estimates of recharge.

Table 6. Estimates of annual ground-water recharge, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[**USGS authored reports** indicated in footnotes. **Recharge estimates** using two different methods are reported for Watson and others (1976) and Flint and others (2004). **Abbreviations:** USGS, U.S. Geological Survey; BCM, Basin Characterization Model; –, no estimate]

Hydrographic area name	Estimates of ground-water recharge, in thousands of acre-feet per year											
	USGS authored reports	Watson and others (1976)		Nicols (2000)	Epstein (2004)	Dettinger (1989)	Kirk and Campana (1990)	Thomas and others (2001)	Flint and others (2004)		Brothers and others (1993b,c, and 1994)	Current study, BCM
Butte Valley-southern	[1]15	16	14	69	29	12	–	–	22	18	–	35
Cave Valley	[3]14	9	8	–	15	–	11	20	10	9	[2]13	11
Jakes Valley	[4]17	–	–	39	14	–	18	24	11	8	–	16
Lake Valley	[5]13	9	9	–	24	–	–	41	15	12	–	13
Little Smoky Valley	[6]4	3	8	13	9	–	–	–	8	6	–	4
Long Valley	[7]10	7	12	48	22	–	5	31	16	14	–	25
Newark Valley	[8]18	13	14	49	29	–	–	–	18	15	–	21
Snake Valley	[9]103	–	–	–	–	–	–	–	93	82	[10]110	111
Spring Valley	[11]75	63	33	104	93	62	–	–	67	56	[12]72	93
Steptoe Valley	[13]85	75	45	132	101	–	–	–	111	94	–	154
Tippett Valley	[14]7	5	6	13	9	–	–	–	10	8	–	12
White River Valley	[4]38	–	–	–	42	–	35	62	35	31	–	35

[1]Glancy (1968).
[2]Brothers and others (1993c).
[3]Eakin (1962).
[4]Eakin (1966).
[5]Rush and Eakin (1963).
[6]Rush and Everett (1966).
[7]Eakin (1961).
[8]Eakin (1960).
[9]Hood and Rush (1965).
[10]Brothers and others (1993b).
[11]Rush and Kazmi (1965).
[12]Brothers and others (1994).
[13]Eakin and others (1967).
[14]Harrill (1971).

Since publication of the reconnaissance studies, various statistical, geochemical, and numerical methods have been used to reevaluate basin-wide recharge (table 6). These methods commonly are variations on the Maxey-Eakin method and are based on a different precipitation map and ground-water discharge estimates (Nichols, 2000), or on statistical analysis of Maxey-Eakin results for selected HAs (Watson and others, 1976; Epstein, 2004). Additional methods to estimate recharge include chloride-mass balance (Dettinger, 1989), deuterium-calibrated water accounting models (Kirk and Campana, 1990; Thomas and others, 2001), a water-budget accounting model (Flint and others, 2004), and numerical simulation (Brothers and others, 1993a, 1993b; Brothers and others, 1993c; Brothers and others, 1994). For the HAs included in the study, Nichols (2000) generally reports the highest recharge estimates; Watson and others (1976) generally report the lowest recharge, typically slightly lower than values reported in the reconnaissance reports.

For estimates of ground-water discharge (table 7), reported methods are variations on the Maxey-Eakin method of multiplying an ET rate by the associated area of phreatophytic vegetation. However, technological advances such as the utilization of micrometeorological and remote-sensing methods applied by Nichols (2000) have improved ground-based measurements of ET.

Table 7. Estimates of annual ground-water discharge, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[**USGS authored reports** indicated in footnotes. Qualitative discharge values in this table are presented as cited in the USGS reports for Cave and Jake Valleys. **Abbreviations:** USGS, U.S. Geological Survey; –, no estimate]

Hydrographic area name	Estimates of annual ground-water discharge, in thousands of acre-feet per year				
	USGS authored reports	Nichols (2000)	Thomas and others (2001)	Brothers and others (1993a, b, and 1994)	Current study
Butte Valley-southern	[1]110	45	–	–	12
Cave Valley	[2]0	–	5	[3]0	2
Jakes Valley	[4]0	1	1	–	1
Lake Valley	[5]9	–	24	–	6
Little Smoky Valley-northern	[6]2	6	–	–	4
Long Valley	[7]2	11	11	–	1
Newark Valley	[8]19	61	–	–	26
Snake Valley	[9]80	–	–	[10]87	132
Spring Valley	[11]70	90	–	[12]70	76
Steptoe Valley	[13]70	128	–	–	101
Tippett Valley	[14]0	3	–	–	2
White River Valley	[4]37	–	80	–	77

[1]Glancy (1968).
[2]Eakin (1962).
[3]Brothers and others (1993b).
[4]Eakin (1966).
[5]Rush and Eakin (1963).
[6]Rush and Everett (1966).
[7]Eakin (1961).
[8]Eakin (1960).
[9]Hood and Rush (1965).
[10]Brothers and others (1994).
[11]Rush and Kazmi (1965).
[12]Brothers and others (1993a).
[13]Eakin and others (1967).
[14]Harrill (1971)..

Ground-Water Recharge

The primary source of water recharging the ground water underlying the study area is precipitation originating in the high mountains that border the broad, elongated valleys characteristic of the region (fig. 19 and pl. 4). In general, the higher the mountain range, the greater the precipitation. The rate at which precipitation infiltrates through the surface and underlying rock to recharge the regional ground-water flow system depends on the permeability of the bedrock, local evapotranspiration, the permeability of the soil, and the amount of water stored in the soil. Because most bedrock in the region has low primary permeability, the rate of infiltration into mountain blocks is controlled by the rock's secondary permeability created by the fracturing of consolidated rock and enhanced by dissolution.

Water-Balance Method for Estimating Recharge

The distribution of ground-water recharge and first-order estimates of recharge rates were developed using a regional-scale model. The recharge-accounting model also was used to evaluate the processes, properties, and climatic factors that ultimately control recharge to the regional ground-water flow system. The model is an updated and refined version of the Basin Characterization Model (BCM) initially documented in Flint and others (2004).

The BCM is a mathematical deterministic water-balance method that integrates maps of geology, soils, vegetation, air temperature, slope, aspect, potential ET, and precipitation. The model uses many of these data sets and internal computations

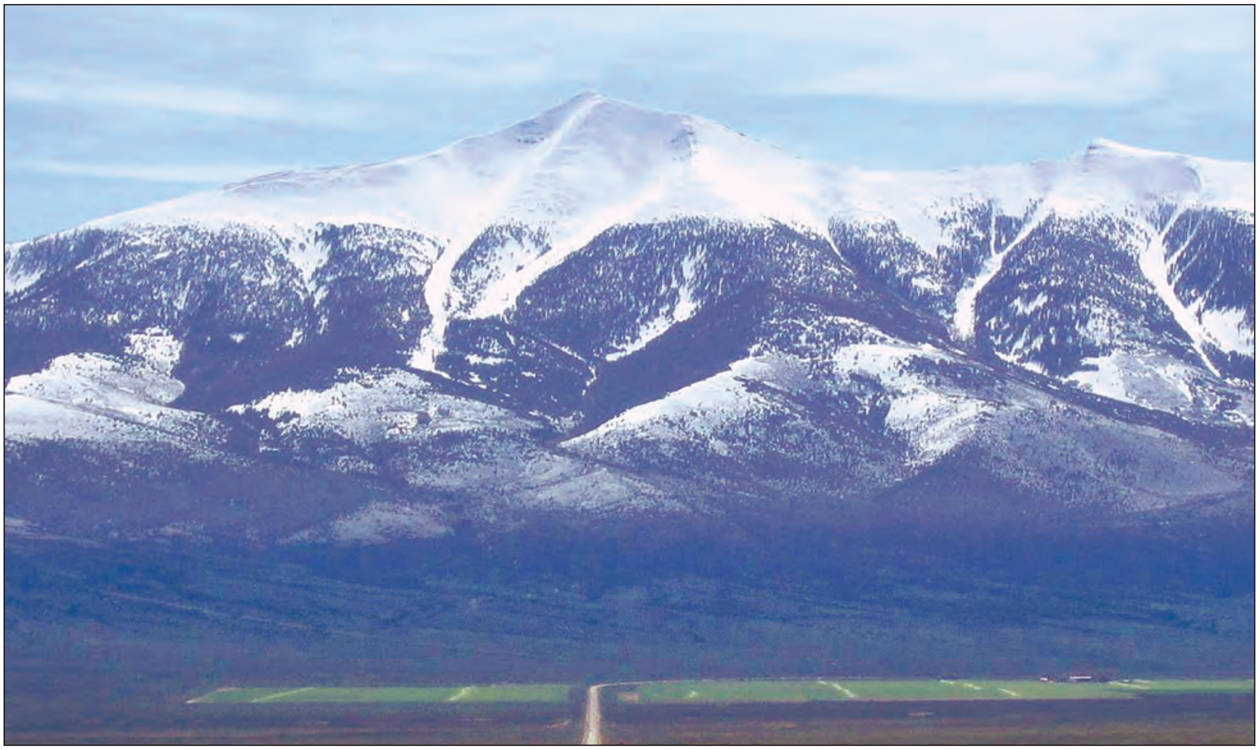
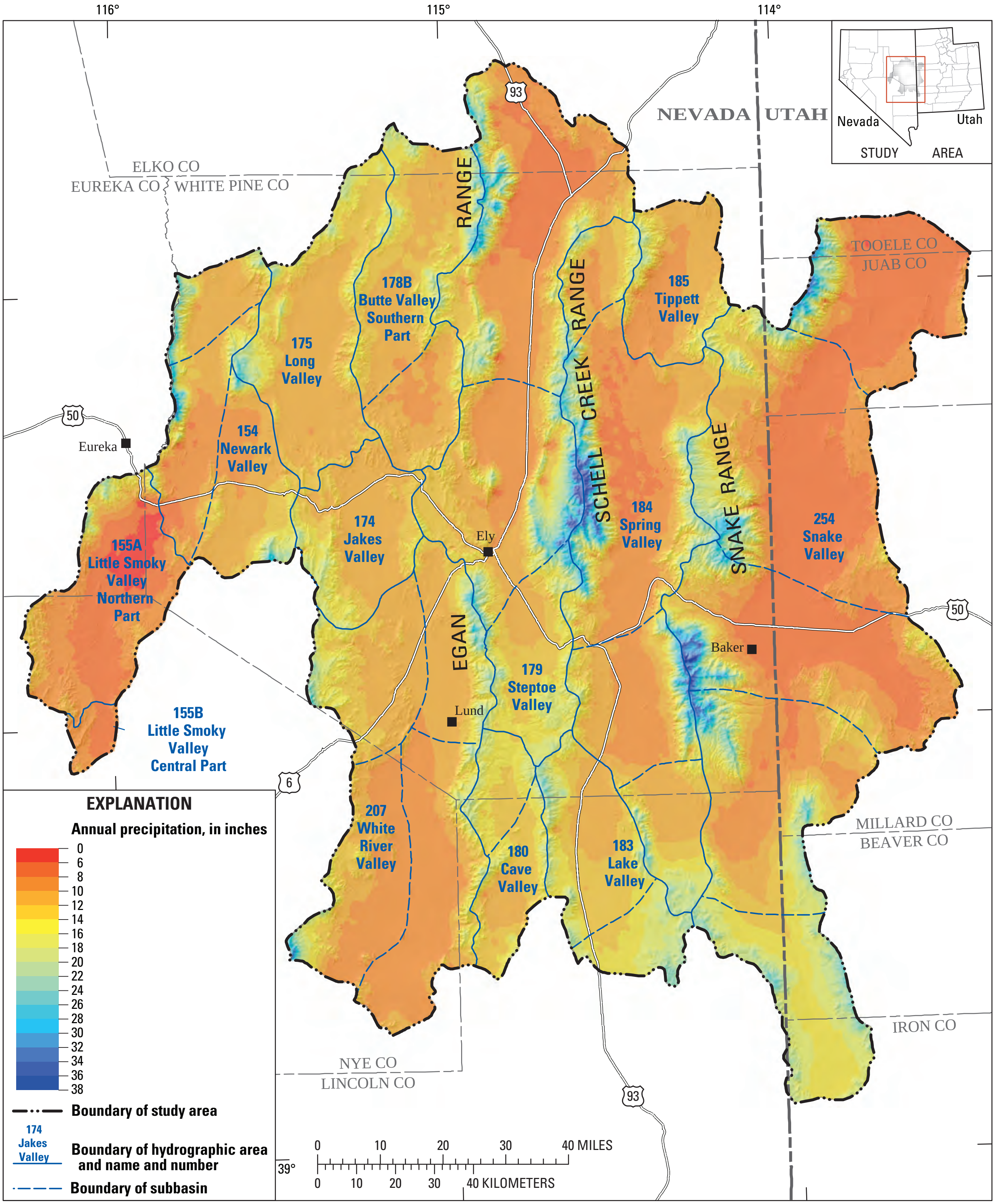


Figure 19. Precipitation (snowfall) on a typical bedrock highland flanking an alluvial valley in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah. Photograph taken May 17, 2005, of west side of 13,065-foot-high Wheeler Peak in southern Snake Range, Great Basin National Park. Mountain ranges accumulate winter snow. Snowmelt provides most of the infiltration recharging the local and regional aquifers of the Basin and Range carbonate-rock aquifer system study area. Agricultural fields seen in foreground. Photograph taken by Michael T. Moreo, U.S. Geological Survey, May 17, 2005.

to estimate the distribution of precipitation (fig. 20), snow accumulation and snowmelt, potential ET, soil-water storage, and bedrock permeability. Using digital elevation grid cells of 890 × 890 ft and spatially distributed estimates of monthly precipitation, monthly minimum and maximum air temperature, monthly potential ET, soil-water storage, and bedrock permeability, the BCM accounts for all water entering and leaving grid cells to determine areas where excess water is available, and whether this excess water is stored in the soil or infiltrates downward toward the underlying bedrock. Depending on the soil and bedrock permeability, the BCM partitions excess water either as in-place recharge or runoff. Runoff can evaporate or recharge along the mountain fronts or through stream channel sediments at some distance downstream of the mountain front.

Average annual potential recharge and runoff for each subbasin was estimated with the BCM in the 13 HAs of the study area (pl. 4). Based on 112 years of climate records the BCM simulations estimate about 476,000 acre-ft of potential in-place recharge and about 360,000 acre-ft of potential runoff (appendix A). Assuming that 15 percent of the potential runoff becomes regional ground-water recharge (Flint and Flint, 2007), about 530,000 acre-ft of the precipitation on average, annually recharges the ground-water flow system. The HAs contributing the greatest amount of ground-water recharge to the study area are Steptoe, Snake, Spring, and Butte Valleys (fig. 21). Spring, Steptoe, and Snake Valleys account for 68 percent of the ground-water recharge but only cover 54 percent of the study area. Except for Snake Valley, all other HAs are less than 1.3 million acres, and estimated annual recharge ranges between 4,000 acre-ft in Little Smoky Valley and 150,000 acre-ft in Steptoe Valley. Even though White River Valley is relatively large at more than 1 million acres (12 percent of the study area), estimated recharge is 35,000 acre-ft, which is 7 percent of total recharge. The 13 HAs in the study area averaged 0.06 ft/yr of recharge to the regional ground-water system. HAs that received more than 0.06 ft/yr of recharge are dominated by high permeability carbonate rock.



Base from U.S. Geological Survey digital data 1:100,000, 1978–89.
Universal Transverse Mercator Projection, Zone 11, NAD83.
Shaded relief base from 1:250,000-scale Digital Elevation Model;
sun illumination from northwest at 30 degrees above horizon.

Figure 20. Distribution of average annual precipitation in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 1971–2004.

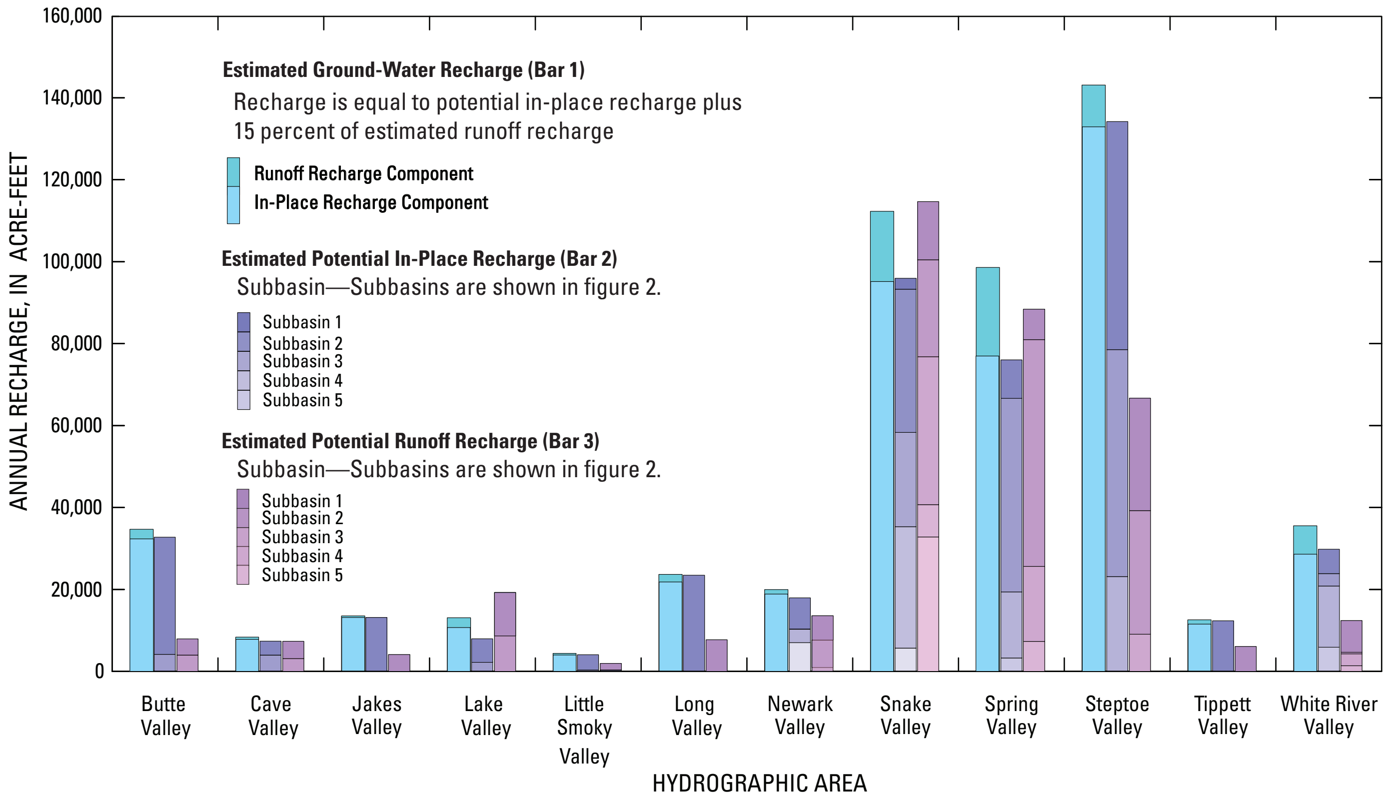


Figure 21. Mean annual ground-water discharge to hydrographic areas in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 1971–2004.

Average annual ground-water recharge ranges from less than 15,000 acre-ft for Cave, Lake, Little Smoky, and Tippett, Valleys to greater than 100,000 acre-ft for Snake and Steptoe Valleys (appendix A; fig. 21). Even though in-place recharge is the primary source of recharge for all HAs, some areas receive significantly high quantities of total runoff, and for a few basins, the quantity of total potential runoff is greater than the estimated annual ground-water recharge (fig. 21). For example, based on dissolved gas and stable-isotope samples collected from 15 sites in the valleys of Cave, Lake, Snake, Spring, and Steptoe HAs (Victor Heilweil, U.S. Geological Survey, written commun., 2007), the source of most ground-water recharge is in-place recharge at high altitudes. However, in Lake and Snake Valleys, the total potential runoff is estimated to be greater than the average annual ground-water recharge. The dominance of in-place recharge or runoff in an HA depends on a number of factors, including altitude, area, and the type of rock in the surrounding bedrock highlands.

The recharge estimates do not necessarily reflect the current short-term average recharge for the study area. Climate variability and the climate periods used in the analysis add uncertainty to the recharge estimates. Precipitation during 1970–2004 averaged 5 percent higher than during 1895–2006 for the study area (Flint and Flint, 2007). Precipitation increases ranged between 1 and 14 percent in Tippett and Little Smoky Valleys, respectively.

Long-term recharge during 1895–2006 was estimated by relating annual recharge to annual precipitation for each subbasin (Flint and Flint, 2007). The regression approach assumed that antecedent conditions from previous years do not affect annual recharge. This assumption is incorrect for predicting recharge in a particular year but should minimally affect an estimate of a 112-year average. The increase in recent (1970–2004) precipitation leads to a 10-percent greater estimate of recharge for the current climate (1970–2004) versus for the available long-term average precipitation data (1895–2006).

The uncertainties in the saturated hydraulic conductivity of bedrock was the greatest source of uncertainty for ground-water recharge estimates from the BCM because saturated hydraulic conductivity of bedrock partitions water between in-place recharge and runoff. Recharge from runoff ranges between 10 and 90 percent, which increases the uncertainty of ground-water recharge where runoff exceeds in-place recharge. The range of ground-water recharge exceeds 80 percent of the best estimate in Lake, Snake, and Spring Valleys where runoff exceeds in-place recharge (fig. 22).

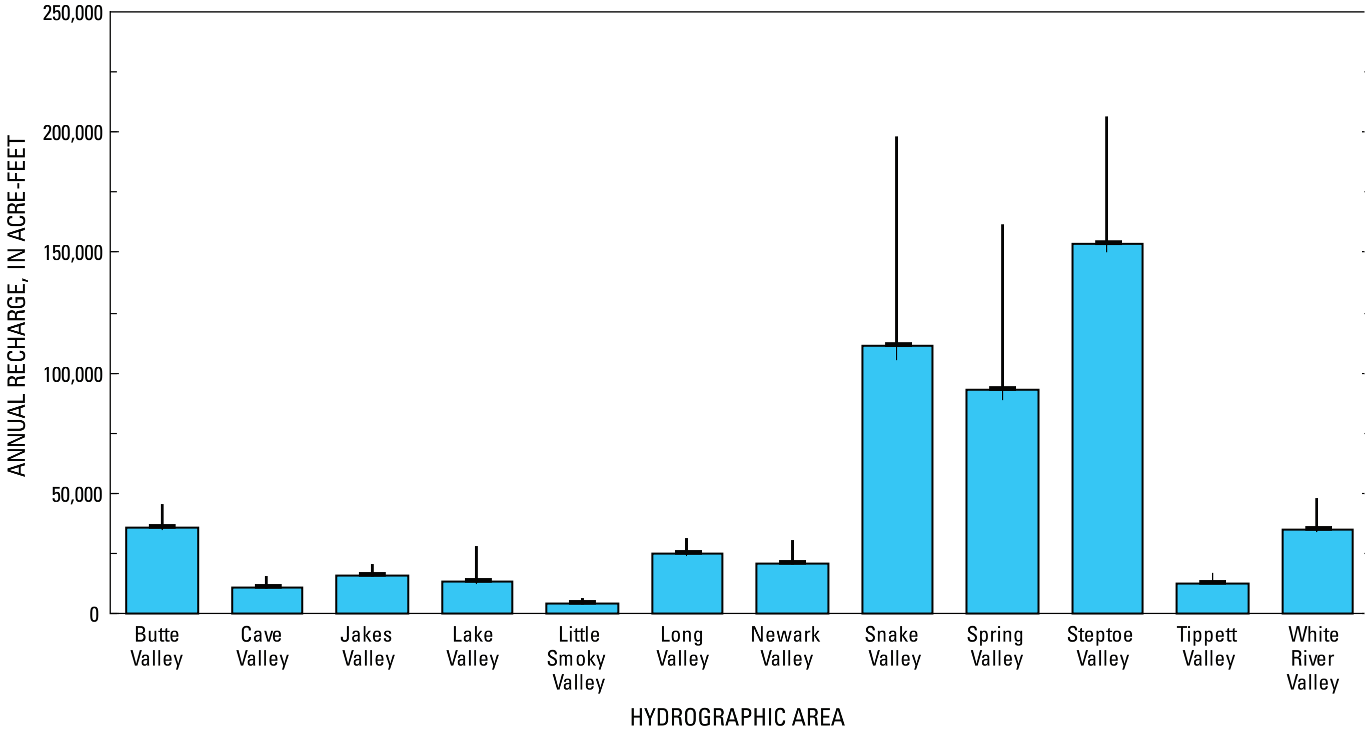


Figure 22. Uncertainty in ground-water recharge estimates by hydrographic area, Basin and Range carbonate-aquifer system study area, Nevada and Utah.

Comparison of Ground-Water Recharge Estimates

Most BCM recharge estimates are within the range of previous estimates (fig. 23). In Snake, Spring, Steptoe, and Tippett Valleys, the BCM recharge estimates are greater than or equal to the upper range of any previous values; for all other HAs, BCM recharge estimates are within the range of previous values. High recharge estimates for Snake, Spring, Steptoe, and Tippett Valleys may be the result of methodology. Unlike the BCM, previous investigations neglected the spatial variability in bedrock and soil permeability when determining recharge estimates. The highlands surrounding these HAs are dominated by high permeability carbonate bedrock that would facilitate in-place recharge.

The chloride mass-balance method for estimating ground-water recharge has been used in several studies in Nevada (Dettinger, 1989; Maurer and Berger, 1997; Russell and Minor, 2002). The method was applied to some HAs in the study area by Steven Mizelle (Desert Research Institute, written commun., 2007). Because of limitations to this methodology as discussed by Dettinger (1989), the chloride-mass balance estimated average annual recharge is expected to be more uncertain than estimates made using the water-balance method.

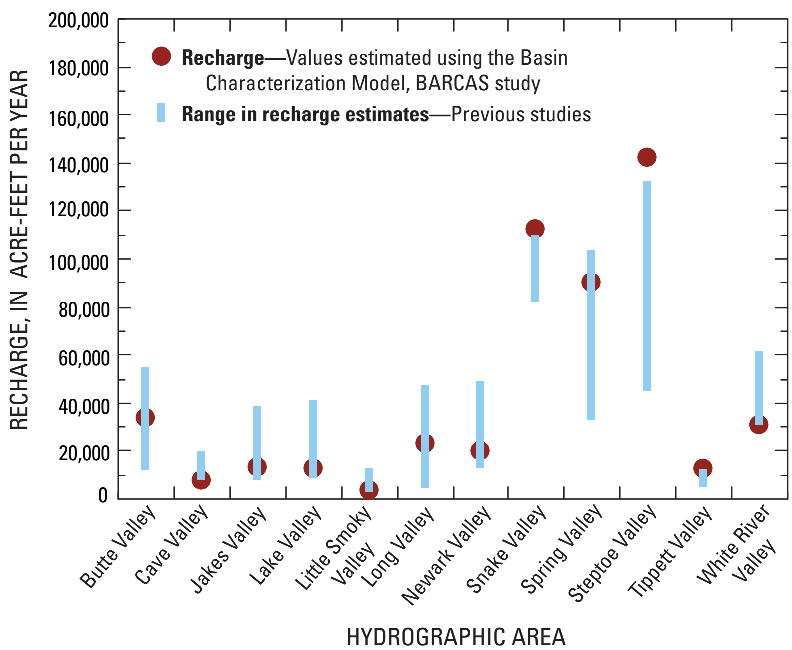


Figure 23. Comparison of ground-water recharge estimates, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Ground-Water Discharge

Ground water discharges from the study area through a combination of four primary processes—(1) spring and seep flow, (2) transpiration by local phreatophytic vegetation, (3) evaporation from soil and open water, and (4) subsurface outflow. Transpiration and evaporation are collectively referred to in this report as evapotranspiration (ET). Of these four processes, the first three occur at or near land surface directly from the discharge area and are the focus of this section. In addition to these pre-development discharges, water also is removed or discharged from the ground-water flow system through the pumping of wells.

Estimates of average annual ground-water discharge for the various HAs are based on estimates of average annual ET developed for ground-water discharge areas. Estimates of ground-water discharge represent pre-development conditions. Springflow is not considered a separate component of the total ground-water discharge. Water discharging from springs is either lost through ET or recharges shallow ground-water flow systems. The amount of springflow that is lost as ET is accounted for in the estimate of total ET. Including total springflow directly in the total discharge estimate would in effect be double accounting of this flow. Moreover, ET-based estimates of ground-water discharge do account for discharge contributed by upward diffuse flow from the underlying regional ground-water flow system. Average annual estimates of ground-water discharge do not account for ground water pumped for irrigation, public supply, and other uses. Ground water that exits in the study area as subsurface outflow is discussed in terms of the difference betwen estimated recharge and ground water ET.

Evapotranspiration

ET is the process that transfers water from land surface to the atmosphere both as evaporation from open water and soil and transpiration by plants. ET rates generally are affected by changes in the depth to the water table or in the moisture content of the soil. As water is removed by ET, the water table may decline and soils may dry. As water levels decline and soil moisture lessens, the vigor of phreatophytic vegetation may decrease. Conversely, as less water is removed, the water table rises and soils moisten, and the vigor of the phreatophytic vegetation may increase, although plants can be adversely affected by very shallow ground water, particularly if the water becomes saline. Changes in ET, the depth to the water table, and the extent and vigor of phreatophytic vegetation all are indicators of changes in water availability.

The volume of water lost to the atmosphere through ET can be computed as the product of the ET rate and the acreage of vegetation, open water, and moist soil that contribute to ET. Past ground-water resource assessments have used this calculation to estimate ET from many major discharge areas in Nevada and Utah (Maxey and Eakin, 1949; Eakin and Maxey, 1951; Eakin, 1960, 1961, 1962, 1966; Rush and Eakin, 1963; Hood and Rush, 1965; Rush and Kazmi, 1965; Eakin and others, 1951, 1967; Glancy, 1968; Laczniak and others, 1999, 2001; Nichols, 2000; Berger and others, 2001; Reiner and others, 2002). Using this calculation, average annual ET is computed for discharge areas within each subbasin and HA in the study area. The procedure used in this study delineates ET units based on similar vegetation and soil conditions, and computes annual ET for each ET unit with a natural discharge area. The average annual ET is estimated by summing annual ET computed for each of the ET units present. Average annual ET estimates for each ET unit are computed by multiplying the acreage of the unit by an appropriate ET rate based on the unit's vegetation and soil conditions. The associated acreage of each ET unit is determined through field mapping combined with an analysis of satellite imagery. ET rates were primarily estimated from rates given in the literature and from data collected at micrometeorological stations established primarily in shrubland vegetation in White River, Spring, and Snake Valleys (Moreo and others, 2007).

ET Units

Numerous studies have shown that the amount of water lost to the atmosphere from areas of ground-water discharge by evaporation and transpiration varies with vegetation type and density, and soil characteristics (Laczniak and others, 1999; 2001; 2006; Nichols, 2000; Berger and others, 2001; Reiner and others, 2002; DeMeo and others, 2003). In general, the more dense and healthy the vegetation and the wetter the soil, the greater is the ET. Many of these studies have used multi-spectral satellite imagery to identify and group areas of similar vegetation and soil conditions within major areas of ground-water discharge. Multi-spectral satellite imagery records digital numbers that represent the amount of incoming solar radiation reflected from the Earth's surface at different wavelengths within the electromagnetic spectrum (Anderson, 1976, p. 2; American Society of Photogrammetry, 1983, p. 23-25; Goetz and others, 1983, p. 576-581). Delineations based on these spectral groupings often are referred to as ET units in that they differentiate areas of differing ET.

ET was estimated from discharge areas in Nevada using Landsat Thematic Mapper (TM) imagery to map ET units in many of the more recent studies (Laczniak and others, 1999; Nichols, 2000; Berger and others, 2001; Nichols and VanDenburgh, 2001; Reiner and others, 2002; DeMeo and others, 2003). TM imagery has a resolution or pixel size of about 100 × 100 ft and includes six spectral bands. The moderate spatial and spectral resolution and the availability and cost of TM imagery are advantageous to mapping the different vegetation and soil conditions in ground-water discharge areas common to the study area. Ten ET units (appendix A) have been mapped from TM imagery in the study area (Smith and others, 2007). These ET units were selected to represent the different vegetation and soil conditions common to areas where ground water is lost to the atmosphere through ET. The characteristics of each ET unit differs—ranging from areas of no vegetation, such as open water, dry playa, and moist bare soil; to areas of denser vegetation often dominated by phreatophytic shrubs, grasses, rushes, and reeds. Three of the 10 ET units describe shrub dominated environments.

ET units were mapped using a modified-soil vegetation index, MSAVI (Qi and others, 1994), and a Tassled Cap transformation (Huang and others, 2002) and land classes (Kepner and others, 2005) (pl. 4). The acquisition time of the 2005 TM imagery used in this study coincides with the near-peak period of ET. SWReGAP (Southwest Regional Gap Analysis Project) used multiple years of imagery and multiple images in a year to delineate land classes. Because land classes delineated by SWReGAP are based on an analysis of multiple images acquired during different times of the year, these classes have characteristics similar to those of the identified ET units and are considered better estimators of ET-unit extents than delineations based solely on one TM image acquired in 2005. SWReGAP land classes included on the ET-unit map (pl. 4) are marshland, dry playa, and open water. Details of the mapping method and procedures can be found in Smith and others (2007).

Shrubland is the most prevalent ET unit in the study area (fig. 24**)**. Shrubland, defined as the combined acreage of sparse, moderately dense and dense desert shrubland, accounts for more than 80 percent of the acreage delineated as potentially contributing to ground-water discharge.

Prior to agricultural development, shrubland acreage was likely greater than accounted for in this study, considering that the ET units include irrigated cropland in areas likely to have been previously dominated by phreatophytic shrubland and riparian vegetation (Smith and others, 2007). Riparian vegetation, such as marshland, meadowland, and grassland, accounts for only about 6 percent of the ET-unit acreage in the study area; and open water accounts for only 0.1 percent.

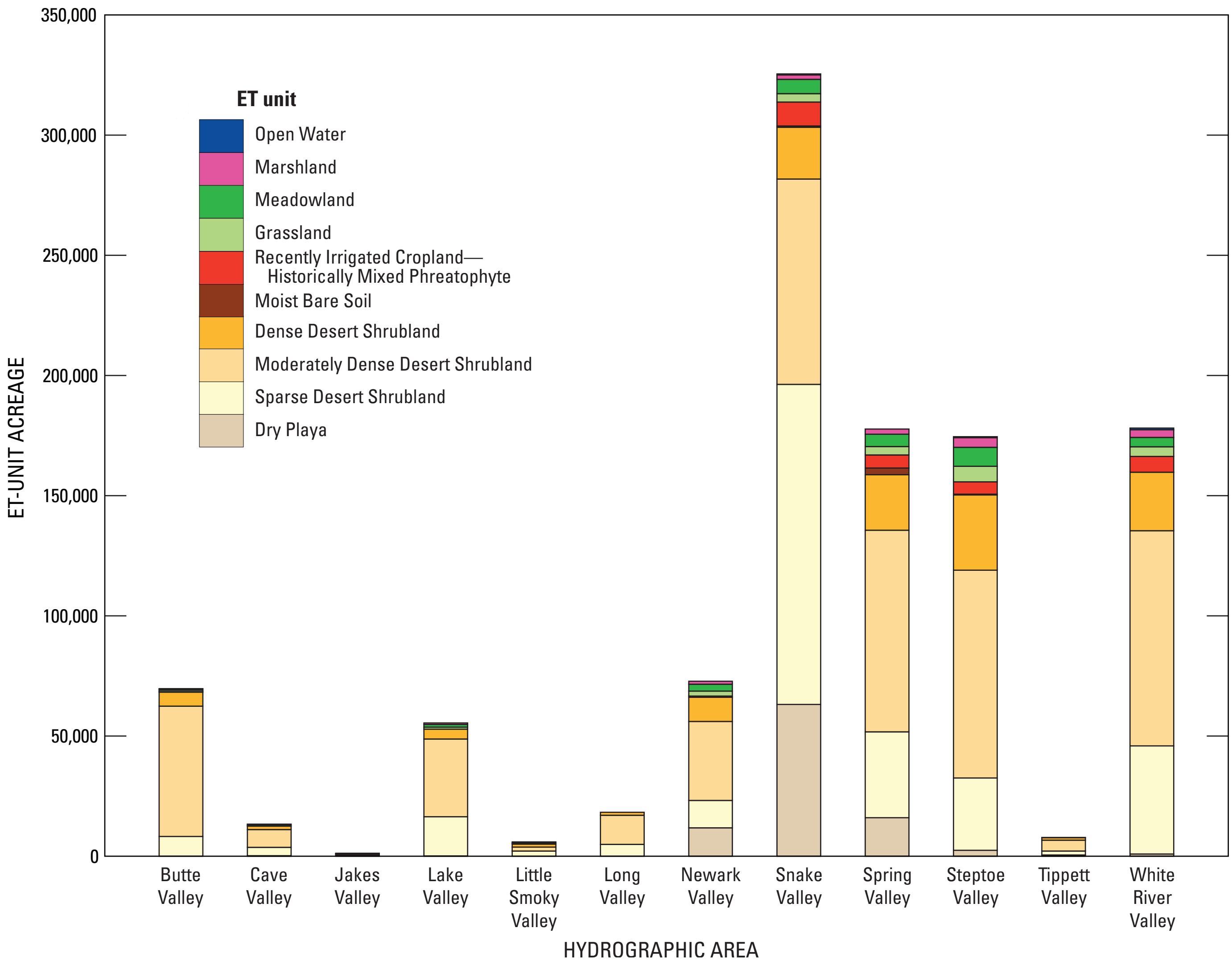


Figure 24. ET-unit acreage by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Shrubland accounts for more than 60 percent of the ET-unit acreage within every HA (fig. 24), but percentages of the different density shrubland units vary from valley to valley. For example, Tippett Valley has less sparse desert shrubland acreage than moderately dense shrubland, whereas in Snake Valley, sparse desert shrubland is the dominant ET unit. Other ET units account for no more than about 20 percent of the total ET-unit acreage in any HA. Dry playa is prevalent only in Newark, Snake, Spring, Steptoe, and Tippett Valleys (fig. 24). In Snake Valley, dry playa constitutes nearly 65,000 acres of the valley's ground-water discharge area.

HAs having the greatest ET-unit acreage are Newark, Snake, Spring, Steptoe, and White River Valleys. Only the latter four of these valleys have acreages exceeding 150,000 acres (fig. 24). Snake Valley has the greatest ET-unit acreage at nearly 330,000 acres. ET-unit acreage in Jakes, Little Smoky, and Tippett Valleys is less than 10,000 acres. Jakes Valley has the least ET-unit acreage at only 1,200 acres. In general, the larger the HA, the greater is the ET-unit acreage (pl. 4). The more densely vegetated ET units (meadowland and marshland) typically occur near springs and along major spring-drainage channels near the center of the valley floor. The less densely vegetated ET units, such as shrubland and grassland, typically occur along the outer edge of the discharge area or near the perimeter of the vegetation surrounding individual springs (pl. 4). For each HA, ET-unit acreage is shown by subbasin in figure 25. ET-unit acreages for the individual subbasins used to develop the ground-water discharge estimates are given in appendix A and described in Smith and others (2007).

Evapotranspiration Rates

Rates of ET from land and plant surfaces to the atmosphere are proportional to available solar energy. Available solar energy is the difference between incoming and outgoing long and shortwave radiation. This energy difference is defined as net radiation (R_n). Net radiation is absorbed at Earth's surface, and then is partitioned into energy that is transferred by heat conduction downward into the subsurface, by heat conduction or convection upward into the atmosphere, or is used to convert water from the solid or liquid to vapor phase (Brutsaert, 1982). The partitioning process, which is governed by the conservation of energy and described by the surface energy budget, can be expressed mathematically as:

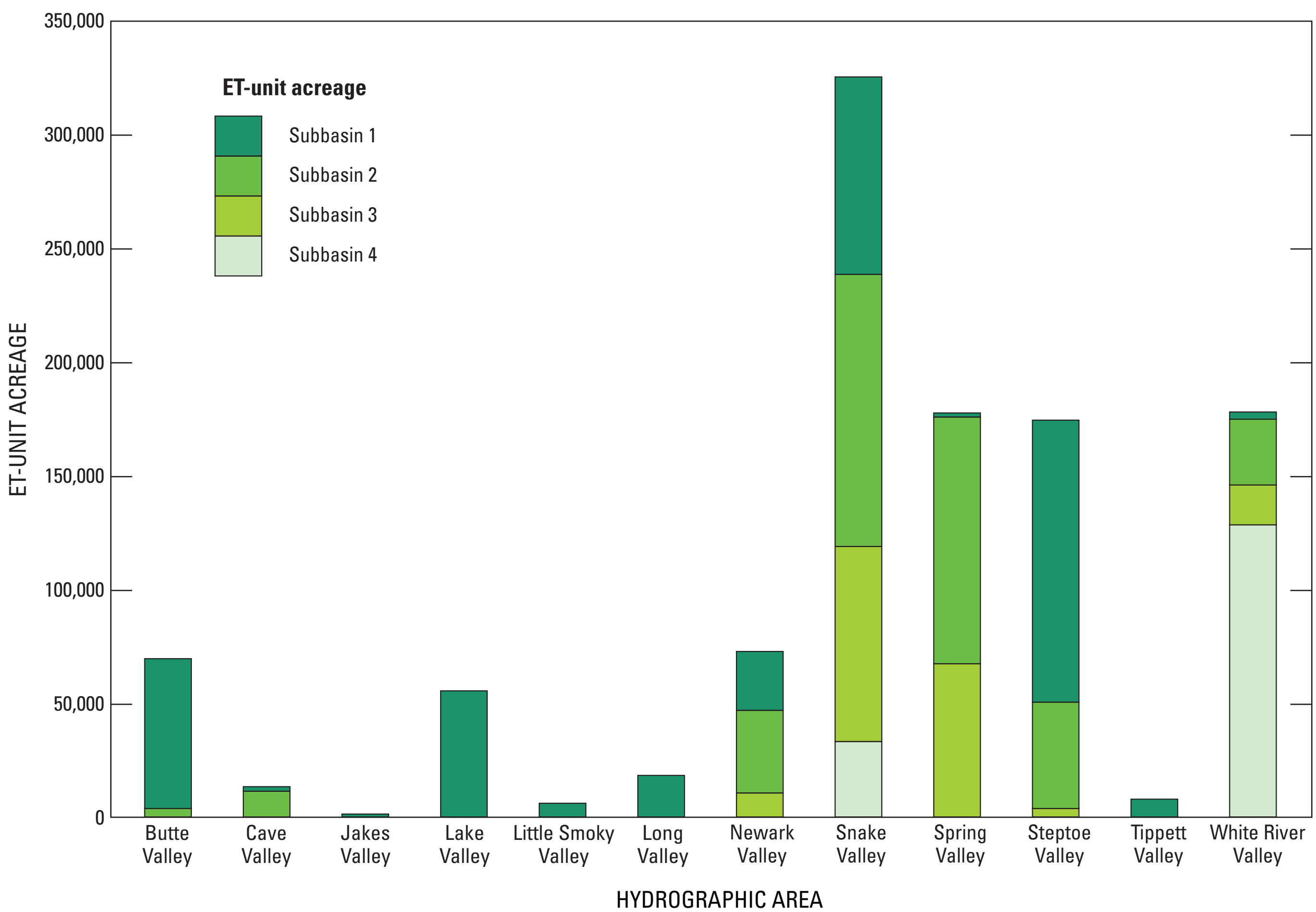


Figure 25. ET-unit acreage by hydrographic area and hydrographic-area subbasin in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

$$R_n = G + H + \lambda E, \quad (1)$$

where

R_n is net radiation (energy per area per time),

G is soil heat flux density (energy per area per time),

H is sensible heat flux density (energy per area per time), and

λE is latent heat flux density (energy per area per time).

The latent-heat flux component (λE) of the energy budget is the energy flux used for ET. Accordingly, ET can be calculated by subtracting the sensible heat (H) and soil heat (G) flux components of the energy budget from the net radiation (R_n, fig. 26). However, because this approach has been hampered historically by difficulties in measuring sensible-heat flux, a common solution to calculating ET has been the use of the Bowen ratio (Bowen, 1926). In simple terms, the Bowen ratio assumes that the proportionality between sensible and latent heat can be defined by the ratio between the temperature and vapor-pressure gradient. Because temperature and vapor pressure can be measured directly, the Bowen ratio can be substituted into the energy budget to solve for latent heat by directly using measurable parameters. Another technique used to estimate ET is the eddy-correlation method. Eddy correlation measures sensible- and latent-heat fluxes directly. Eddies are turbulent airflow caused by wind, the roughness of the Earth's surface, and convective heat flow at the boundary between the Earth's surface and the atmosphere (Kaimal and Finnigan, 1994).

A high-speed hygrometer and three-dimensional anemometer are used to measure sensible- and latent-heat fluxes carried by the turbulence in this boundary layer. These turbulent-type fluxes ($H + \lambda E$) can be compared to available energy (R_n–G) to assess the performance of the eddy-correlation system. Over the last 25 years, many of the estimates of ET made in Nevada and the surrounding area have been based on one of these two methods (Carman, 1989; Nichols, 1993; Nichols and Rapp, 1996; Stannard, 1997; Laczniak and others, 1999; Nichols, 2000; Berger and others, 2001; Reiner and others, 2002).

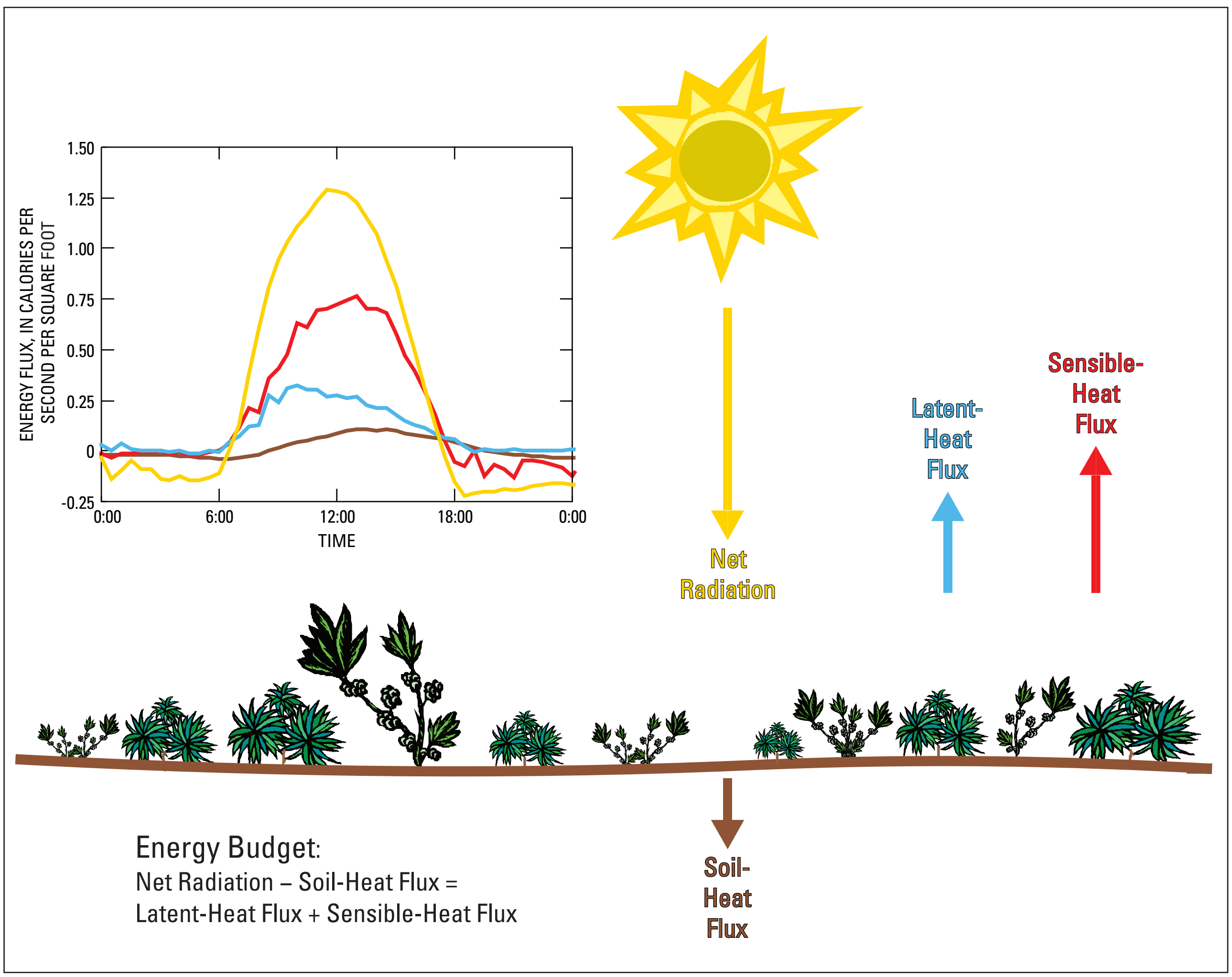


Figure 26. Surface energy processes and typical daily energy budget for shrubs, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

ET rates depend on vegetation type, vegetation density, soil type, soil moisture, and local micrometeorological factors (Duell, 1990; Nichols, 2000; Berger and others, 2001; Laczniak and others, 2001). ET rates for different plant communities and soil type and moisture conditions have been measured across the Western United States for more than a hundred years (Nichols, 2000). Many early groundwater discharge estimates made throughout Nevada relied on ET rates measured elsewhere in the Western United States. Reports published from the 1940s through the 1970s (Maxey and Eakin, 1949; Eakin and Maxey, 1951; Eakin, 1960, 1961, 1962; Hood and Rush, 1965; Rush and Kazmi, 1965; Eakin, 1966b; Eakin and others, 1967; Glancy, 1968) includes estimates of ET rates that were based on measurements made over vegetation and soil similar to that found throughout the study area (Lee, 1912; White, 1932; Young and Blaney, 1942). ET rates reported in the more recent literature (Nichols, 2000; Berger and others, 2001; Reiner and others, 2002; Cooper and others, 2006) were used to develop a range of average annual ET for each ET unit inclusive of the variations associated with the different vegetation and soil-moisture conditions making up the ET units delineated for the study area. Annual ET estimates developed from reported values vary from less than 1 ft over playa and sparse shrubland units to more than 5 ft from open water areas (fig. 27).

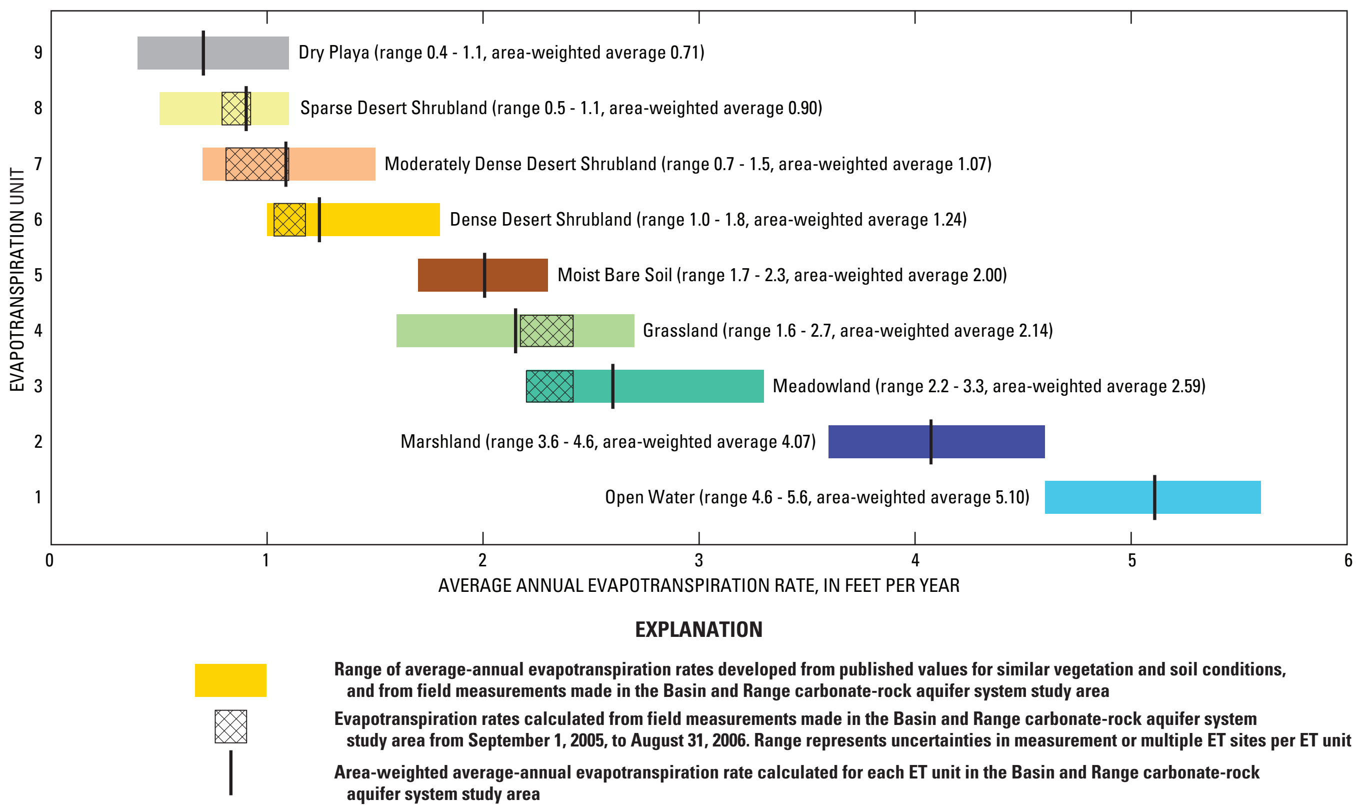


Figure 27. Estimated average annual ET-rate range for ET units identified, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Annual ET ranges for selected ET units were assessed and refined using field data collected at six eddy correlation sites deployed from September 1, 2005, to August 31, 2006. A typical site setup is illustrated in figure 28. Five of the six ET sites were located in the greasewood-dominated shrubland, and one was located in a grassland/meadowland area. Most of the sites were located purposely in shrubland to evaluate the effect of vegetation density on ET rates, and to better quantify ET rates for this dominant vegetation type. Daily ET for the grassland and meadowland ET site (SPV-3) was

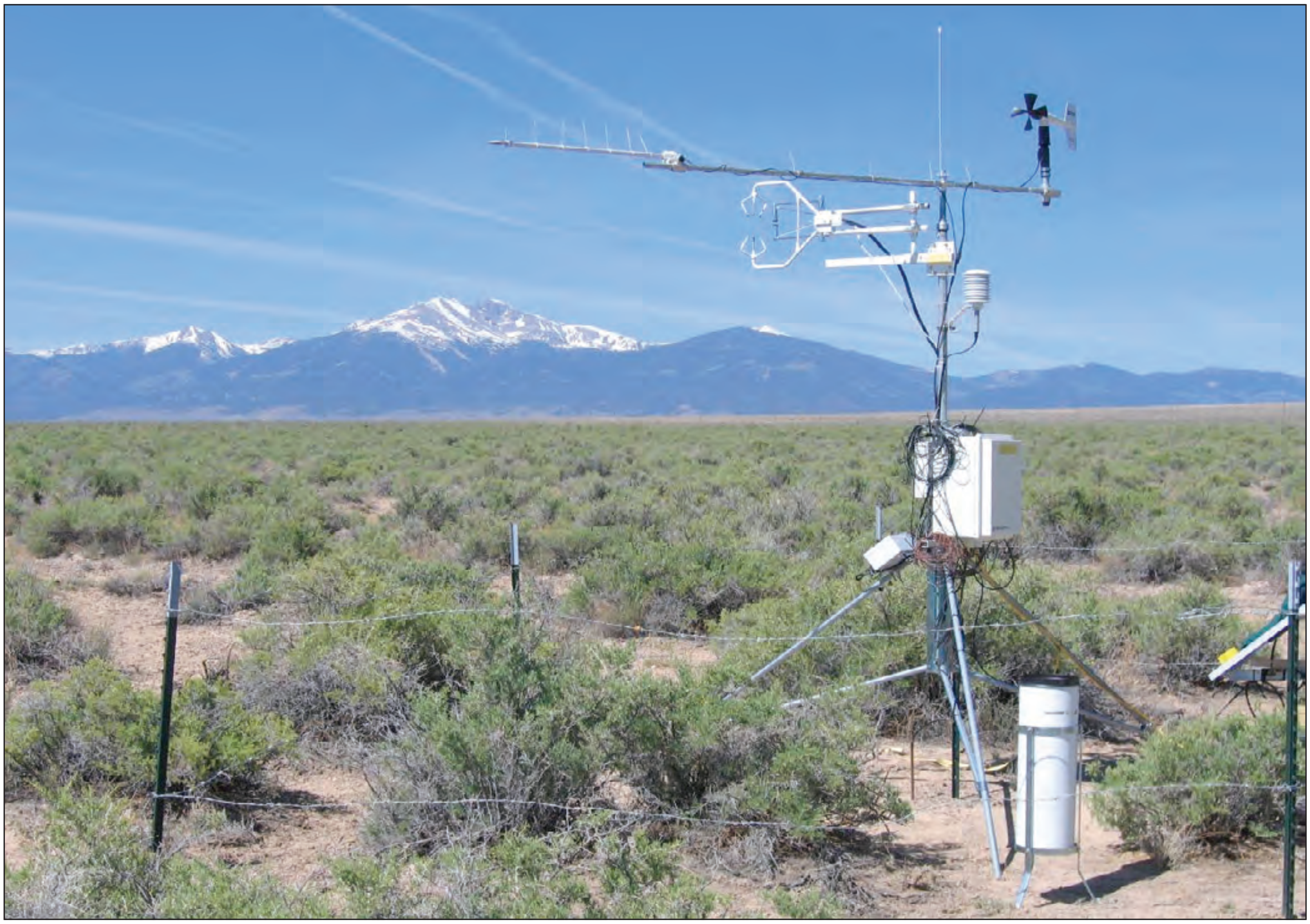


Figure 28. Eddy-correlation site used for measuring evapotranspiration in greasewood dominated shrubland in Snake Valley, Nevada. Northeast flank of southern Snake Range visible in background. Photograph taken by Michael T. Moreo, U.S. Geological Survey, June 1, 2006.

significantly greater than that for a shrubland ET site in Spring Valley (SPV-1) over the 1-year collection period (fig. 29). The SPV-3 ET site represents an environment where annual ET far exceeds annual precipitation, and where ground water rather than precipitation serves as the primary water source for local ET. The SPV-1 ET site represents a typical shrubland environment, where measured ET barely exceeds precipitation, indicating that precipitation rather than ground water is the primary source of water consumed by ET (Moreo and others, 2007). ET measured over the 1-year collection period ranged from about 10 in. in sparse shrubland to 27 in. in grassland and meadowland (figs. 27 and 30).

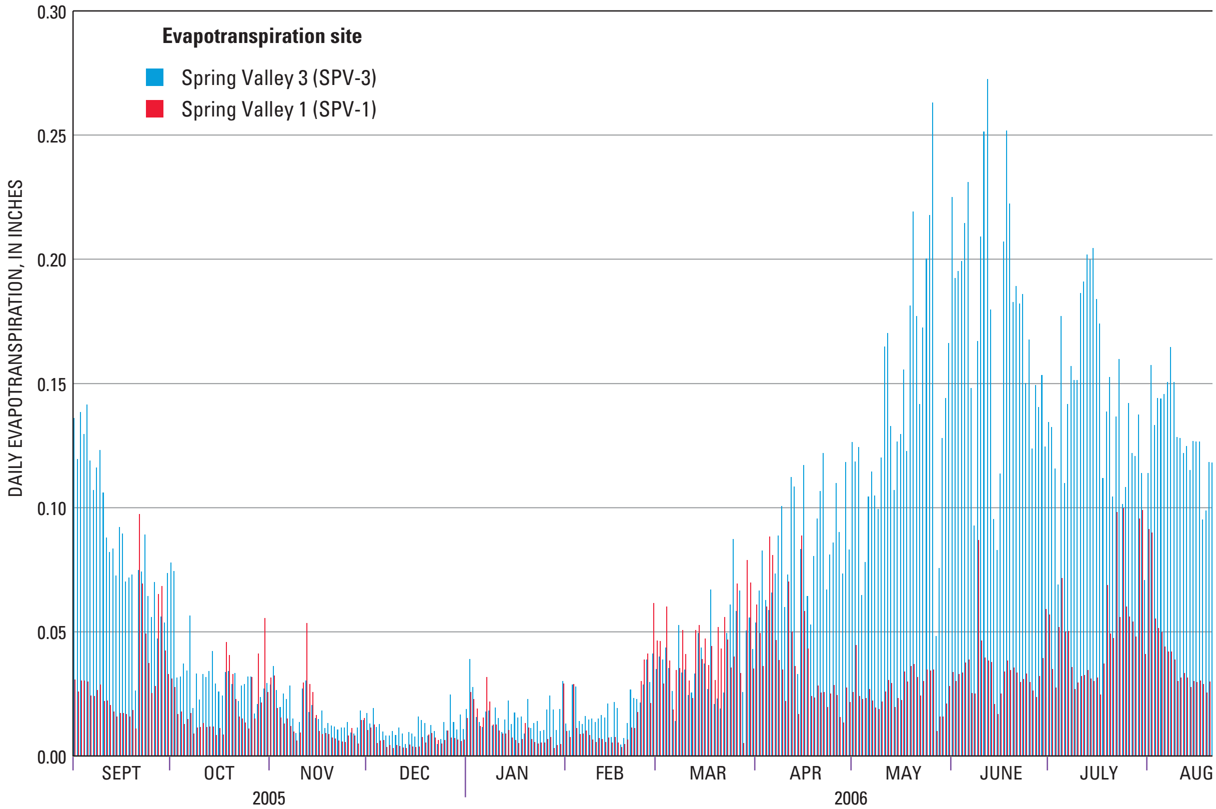


Figure 29. Daily ET from grassland/meadowland site (SPV-3) in Spring Valley, and a greasewood dominated shrubland site (SPV-1) also in Spring Valley, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, September 1, 2005, to August 31, 2006.

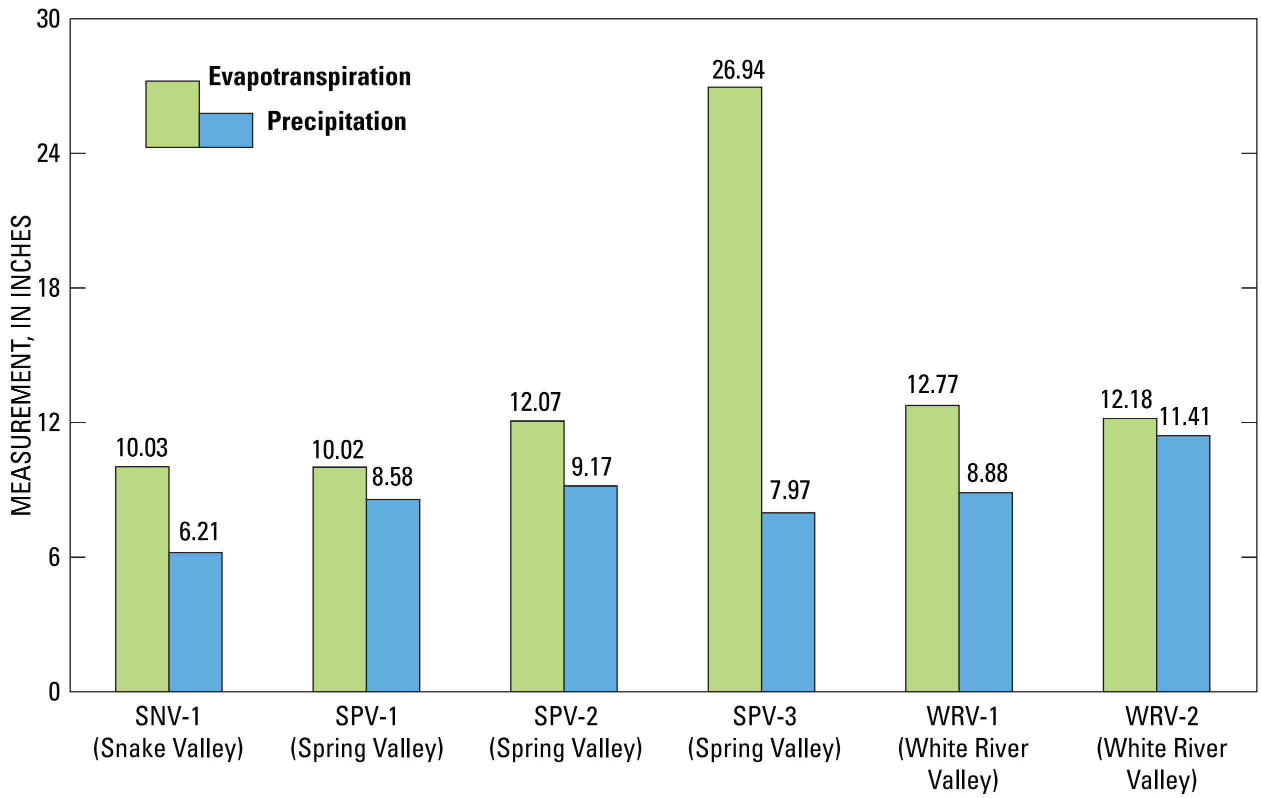


Figure 30. Total ET and precipitation measured at six ET sites in the Basin and Range carbonate-rock aquifer system, eastern Nevada and Utah, September 1, 2005, to August 31, 2006. All ET sites in greasewood dominated shrubland except SPV-3, which is in grassland/meadowland area.

Mean Annual Evapotranspiration

The average annual ET for a discharge area can be estimated volumetrically as the product of the ET rate and the area over which ET is occurring. ET rates used to estimate average annual ET were assumed representative of the pre-development, long-term rates occurring in the study area. Therefore, the ET rate used to represent acreages in the discharge area defined as recently irrigated cropland (Welborn and Moreo, 2007) was replaced with a mixed phreatophytic ET unit that was given an ET rate that equaled the area-weighted average ET rate for all other phreatophyte units delineated in the study area.

Total ET for a HA is estimated as the sum of estimated subbasin ET (fig. 31). Subbasin ET is estimated as the sum of ET for each ET unit within the subbasin. ET for each ET unit within a subbasin is computed as the product of the ET unit's ET rate and its acreage (fig. 32). A unit's ET rate is determined by linearly scaling the ET-rate range computed for the unit (fig. 27). Scaling within the range is done using the average modified soil adjusted vegetation index value (MSAVI) of the unit computed over the subbasin from the TM imagery. The scaling procedure assigns the highest average MSAVI value computed for any subbasin to the high value of the range and the lowest MSAVI value to the lowest value of the range.

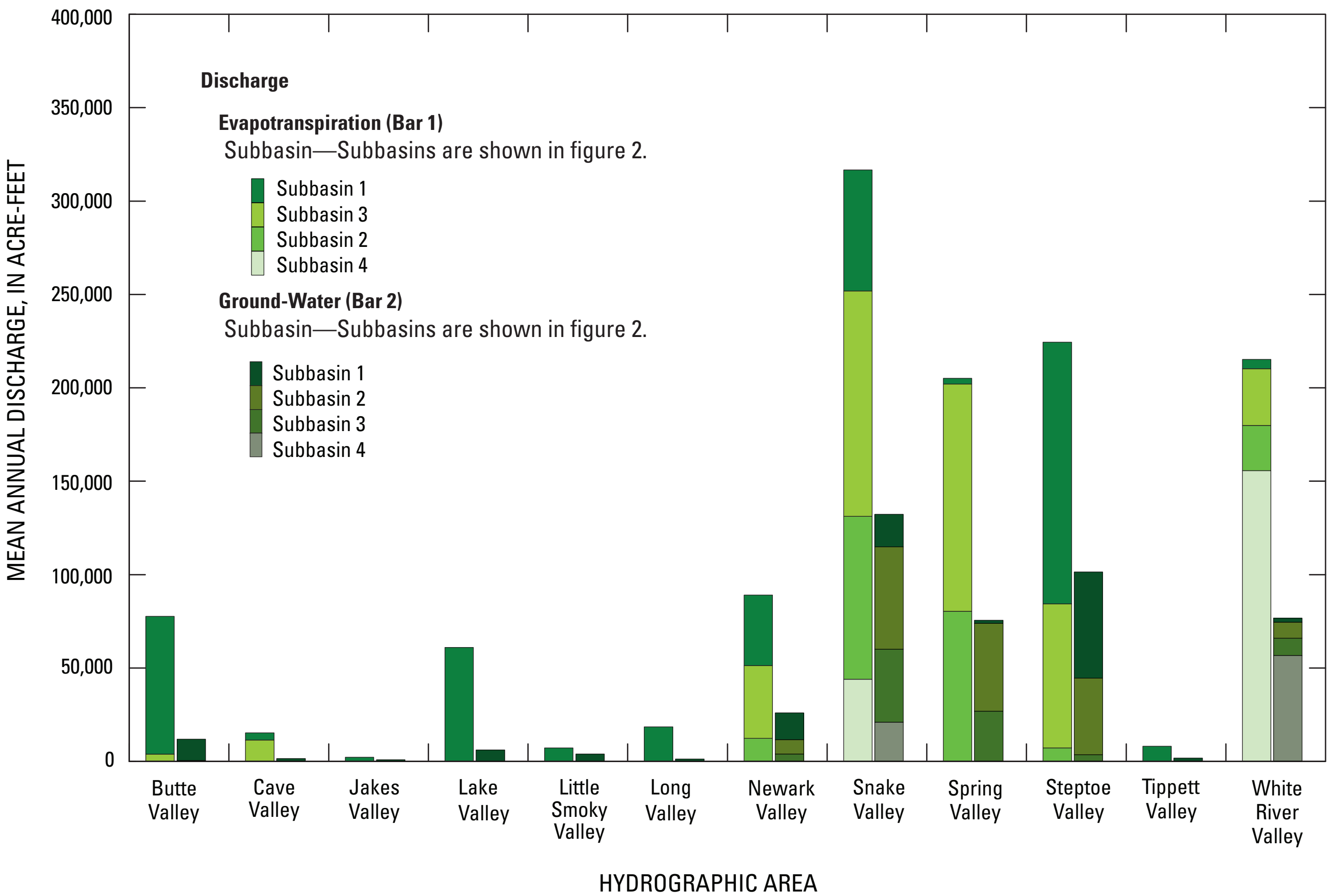


Figure 31. Estimates of mean annual evapotranspiration and ground-water discharge from hydrographic areas by subbasin, Basin and Range carbonate-rock aquifer system, Nevada and Utah.

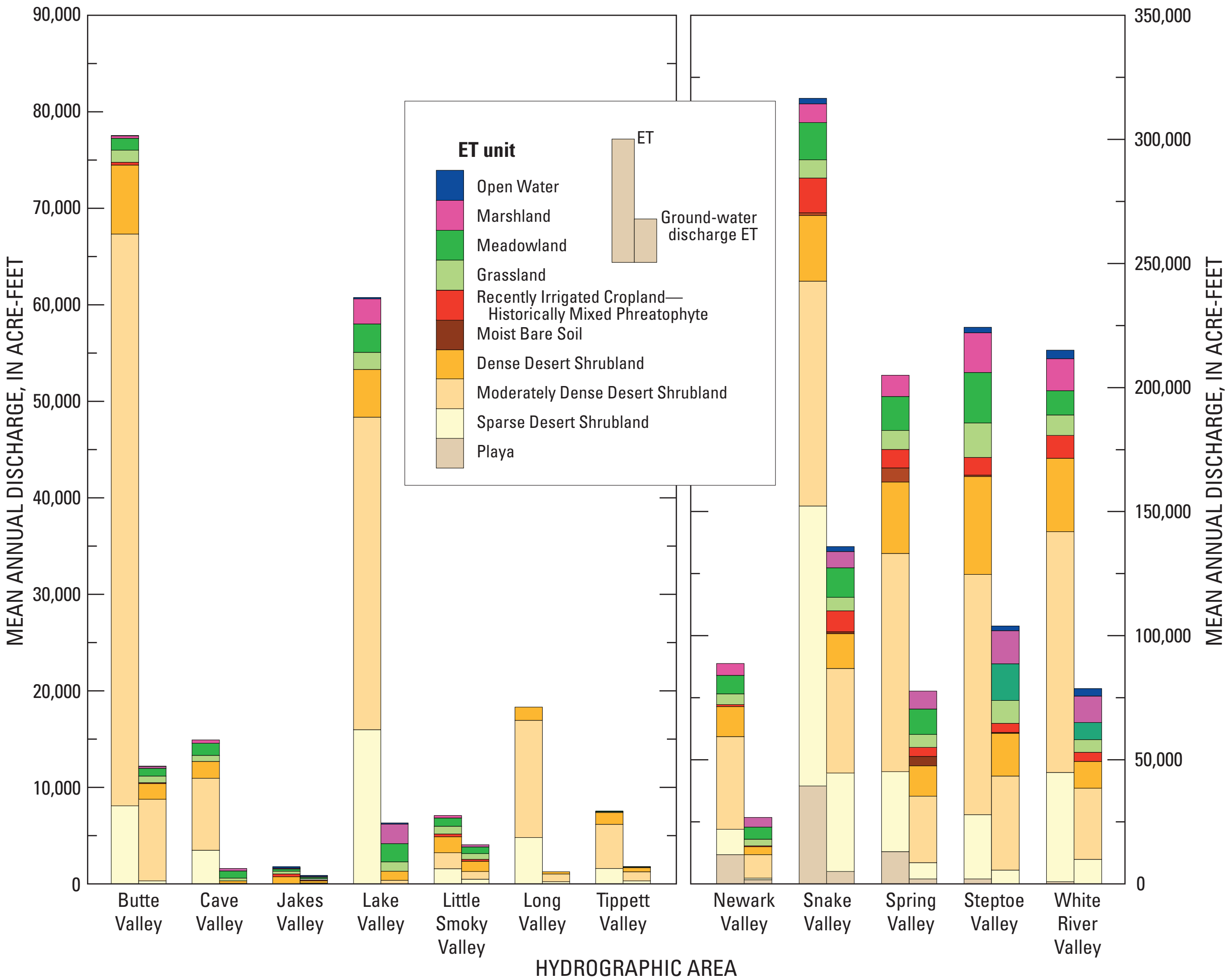


Figure 32. Estimates of mean annual evapotranspiration and ground-water discharge by ET unit from hydrographic areas, Basin and Range carbonate-rock aquifer system, Nevada and Utah.

Mean Annual Ground-Water Discharge

Precipitation directly on areas of ground-water discharge and surface-water run-on (overland flow to discharge areas), also contributes to the ET occurring at discharge areas. For this report, surface-water flow onto fine-grained playa sediments is assumed to evaporate and for the purpose of the water budget does not contribute to either ground-water recharge or discharge. In addition, precipitation falling on areas of ground-water discharge is assumed to be lost by ET rather than to contribute to ground-water recharge. These assumptions are considered reasonable for these semi-arid valleys of the study area.

The average annual precipitation falling directly on ET units was estimated from a map of mean annual precipitation generated from model simulations of monthly precipitation distributions used to estimate average annual recharge for the BARCAS area over the period 1970–2004 (Flint and Flint, 2007). Estimates of the average annual precipitation to discharge areas delineated within HAs range from about 6 in. in Little Smoky Valley to about 13 in. in Cave Valley (fig. 33, appendix A). In general, precipitation to discharge areas decreases from north to south. Contrarily, the highest annual precipitation occurs in Cave and Lake Valleys in the southern part of the study area. This anomaly is attributed to orographic effects that also contribute to higher annual precipitation in the southern subbasins of Snake and Steptoe Valleys.

Annual ground-water discharge from HAs equals the difference between annual ET and local precipitation, and ranges from only 860 acre-ft in Jakes Valley to 130,000 acre-ft in Snake Valley (fig. 31). Average annual ground-

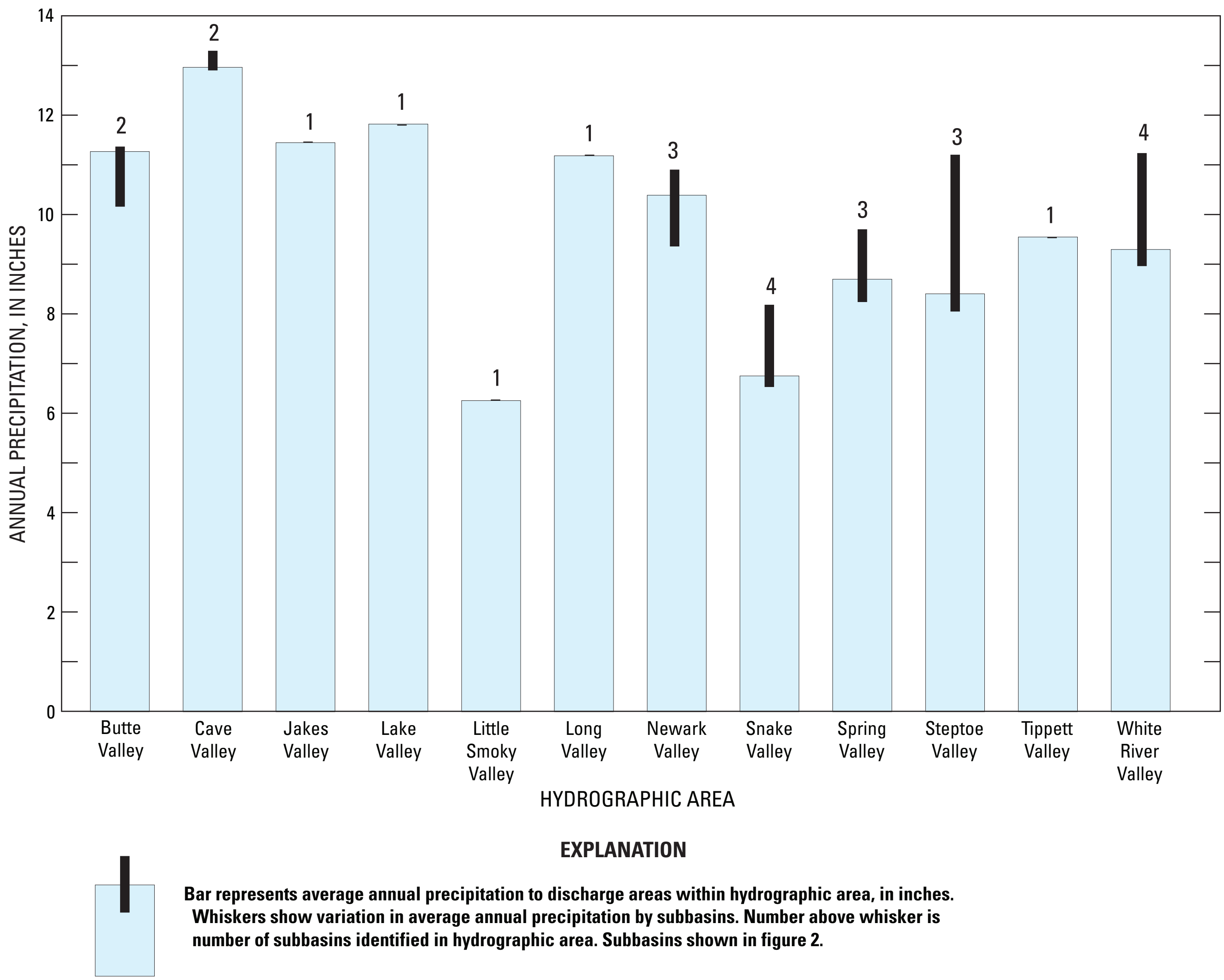


Figure 33. Average annual precipitation to discharge areas by hydrographic areas and by hydrographic-area subbasin, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

water discharge is estimated at more than 75,000 acre-ft in Snake, Spring, Steptoe, and White River Valleys, and at less than 10,000 acre-ft in Cave, Jakes, Lake, Little Smoky, Long, and Tippett Valleys. Combined ground-water discharge from Newark, Snake, Spring, Steptoe, and White River Valleys accounts for 95 percent of the estimated total annual discharge.

The proportion of the ET occurring as ground-water discharge generally decreases as the percentage of dry playa, sparse vegetation, or precipitation increases in the HA. That is, if the HA contains dominantly sparse phreatophytic vegetation or receives abundant precipitation, most of the ongoing ET is more likely to be supported by local precipitation rather than by regional ground water. For example, in Little Smoky Valley about 55 percent of the average annual ET is supported by regional ground-water discharge, whereas in Long Valley, only about 10 percent of the average annual ET is supported by regional ground-water discharge. The discharge area for Little Smoky Valley consists of shrubland and some meadowland and grassland, and receives only about 6.3 in. of precipitation annually. In contrast, Long Valley's discharge area consists wholly of shrubland and receives an average of about 11 in. of precipitation annually. The limited ground-water contribution to ET in Long Valley is a consequence of the valleys relatively high local precipitation.

Limitations and Considerations of Methodology

The overall accuracy of the ground-water discharge estimates given in this report depends on the validity of the assumptions made in calculating volumetric discharges; and on any errors in estimates of ET-unit acreage and rate, and in estimates of the direct precipitation falling on an ET unit. The primary assumptions affecting the accuracy of average annual discharge estimates are: (1) that contributions to ET other than by regional ground water can be removed by subtracting direct precipitation from the ET estimate, (2) that regional ground water is evaporated and transpired only from surfaces delineated as discharge areas, (3) that the spatial variation in ET from discharge areas of the study area can be adequately described using 10 ET units, (4) that the ET rates assigned to ET units adequately represent the average for that unit, (5) that estimates of mean annual precipitation used to compute mean annual ground-water discharge rates represent true long-term averages, and (6) that estimates represent pre-development conditions, and current pumping from the system has not yet significantly reduced phreatophyte acreage or local spring and seep flows. The potential error resulting from any of these assumptions is not expected to significantly alter estimates presented in this report. The potential error and relative certainty between HAs and subbasins has been further evaluated by an analysis of uncertainty described by Jianting Zhu (Desert Research Institute, written commun., 2007) (fig. 34).

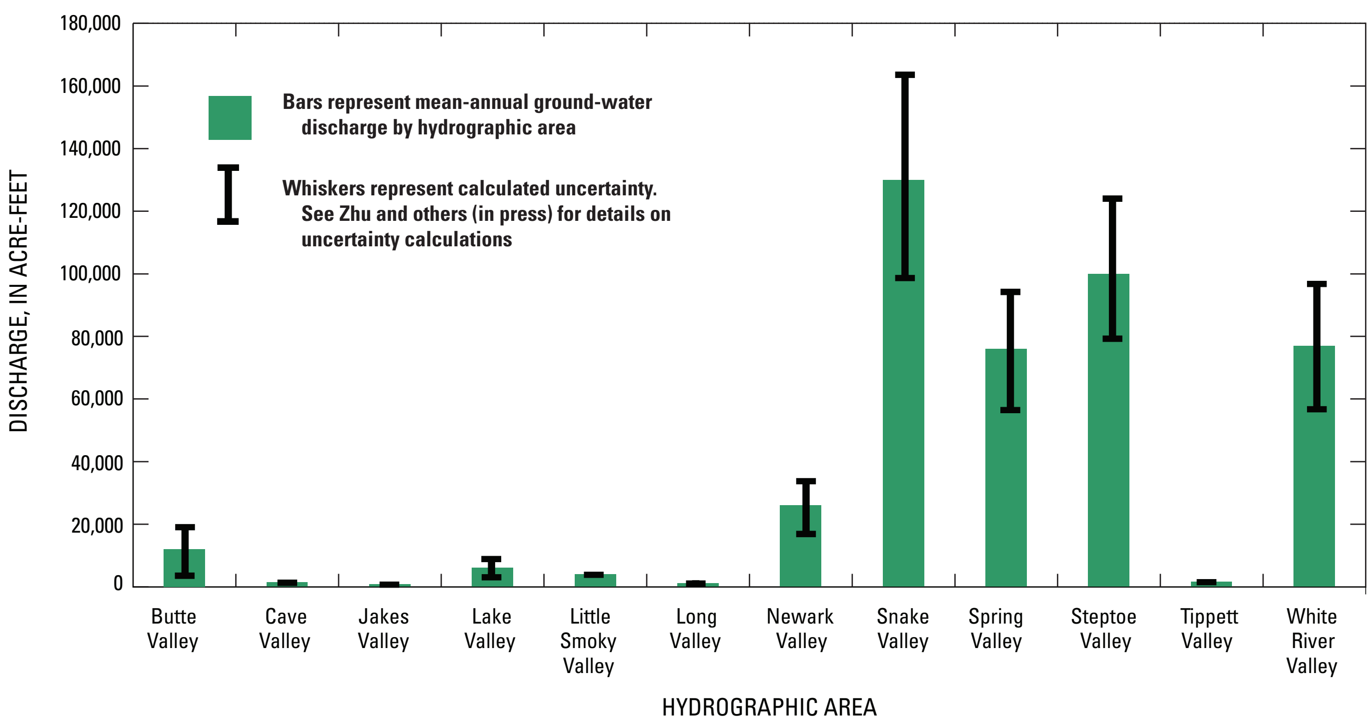


Figure 34. Uncertainty in ground-water discharge estimates by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Errors associated with estimates of ET-unit acreage largely depend on the quality and resolution of the multi-spectral imagery, on the appropriateness of the spectral technique used to delineate ET units, and on the accuracy of the boundaries used to depict the extent of phreatophytes in the study area. The MSAVI analysis of TM imagery used in this report, along with the inclusion of selected SWReGAP-delineated land classes, are assumed appropriate for identifying and delineating phreatophyte distributions. An assessment of the accuracy of the delineated ET units is included in Smith and others (2007). The uncertainties defined by their assessment were used to quantify uncertainty of discharge estimates in the analysis detailed in Jianting Zhu (Desert Research Institute, written commun., 2007).

Shrubland, grassland, meadowland, and moist bare soil ET units were developed from a single set of images acquired in July 2005. Changes in the local vegetation can result from seasonal or annual increases or decreases in precipitation. These changes affect the vigor of the local vegetation, soil-moisture conditions, and the depth to the water table. Although imagery acquired near summer solstice conditions is considered reasonable for mapping phreatophytes in the study area, delineations certainly could be improved by using multiple years of imagery and multiple images within years. The inclusion of multiple images would provide more confidence in acreage estimates intended to represent the long-term average ET rates. Errors in the ET rate are linked to any inaccuracies in reported values, and in potential errors associated with eddy-correlation measurements made in the study area. Uncertainty associated with the eddy-correlation technique, described in numerous publications and specifically addressed for this study in Moreo and others (2007), is expected to be less than about 10 percent. Because ET was computed from measurements made during only a 1-year period and at a limited number of ET sites, confidence in the degree to which these measurements are representative of average annual values and the average for an ET unit could be improved with additional temporal and spatial data.

Estimates of average annual ground-water discharge are intended to account only for that ground water lost to the atmosphere by ET, and are not inclusive of springflow diverted, evaporated, or transpired outside the discharge area, or subsurface outflow to adjacent basins. Without accurate measurements or estimates of these outflows, values given in this report should be considered minimum estimates of the total volume of ground water exiting an HA. In addition, estimates of average annual ground-water discharge are based on ET estimates minus an estimate of the precipitation falling directly on the discharge area. Estimates of ground-water discharge presented in this report are inclusive of any surface runoff and streamflow that infiltrates into the ground-water system from outside discharge areas.

Water Use

Ground water is pumped for farming, mining, ranching, light industry, and domestic and public supply. Pumpage is reported by water use, where each use describes the general application for which the water is used. Water uses were categorized as meeting irrigation and non-irrigation demands; the latter category includes public supply, domestic (self supplied), stock, and mining water use. Irrigation water use, the water-use class associated with the highest water consumption, is estimated for 2005 on the basis of irrigated acreages determined from multi-spectral satellite imagery and crop-application rates developed from climate data and known crop requirements. Non-irrigation water-use estimates were reported by county, state, and federal agencies responsible for regulating and planning current and future development.

Water withdrawn from wells or diverted from springs and mountain-front runoff in the study area is estimated for 2005 at 130,000 acre-ft (appendix A and fig. 35). Total water-use estimates for each HA range from less than 20 acre-ft in Cave and Tippett Valleys to 35,000 acre-ft in Snake Valley. Lake, Snake, Spring, Steptoe, and White River Valleys account for about 89 percent of the total water use from the study area. Public supply, domestic, mining, and stock use was significant only in Steptoe Valley, where it accounts for about 49 percent of total water demand. Combined stock and domestic uses accounted for less than 2 percent of total water demand.

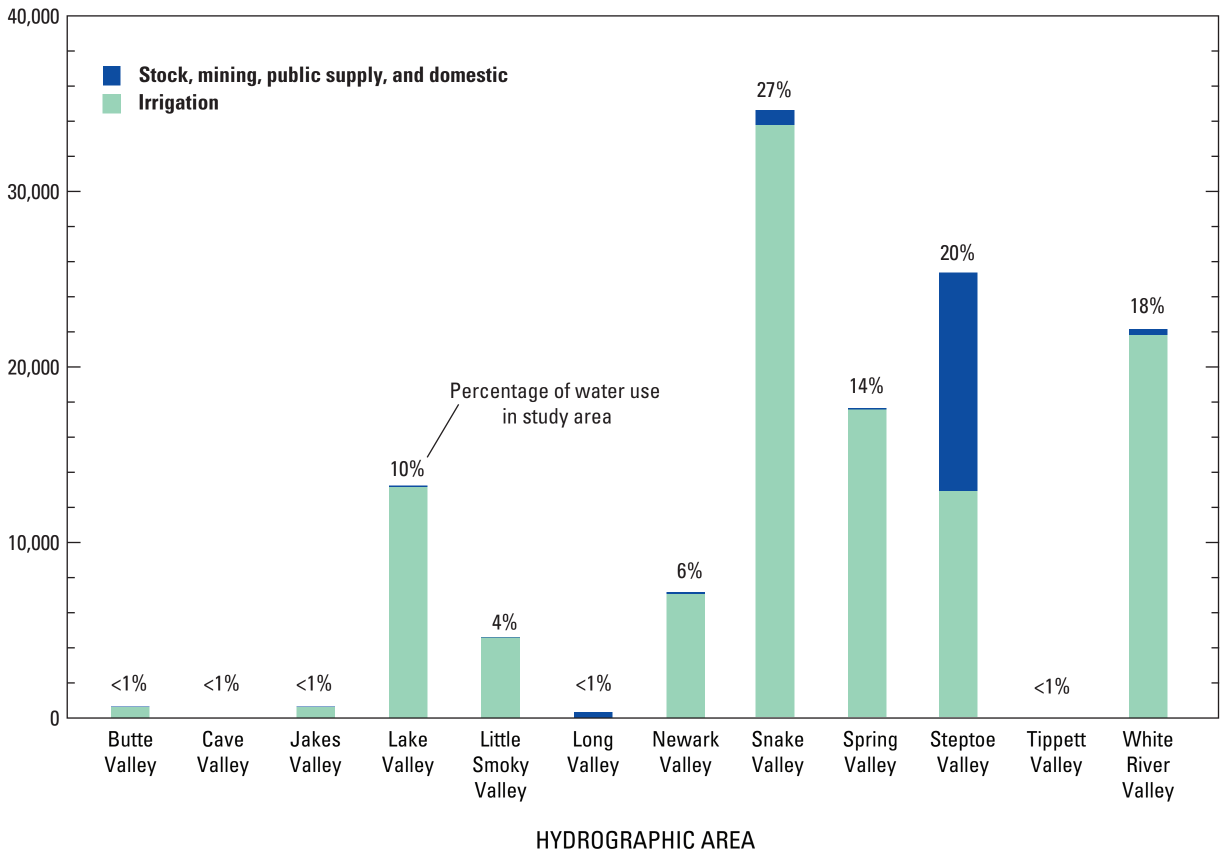
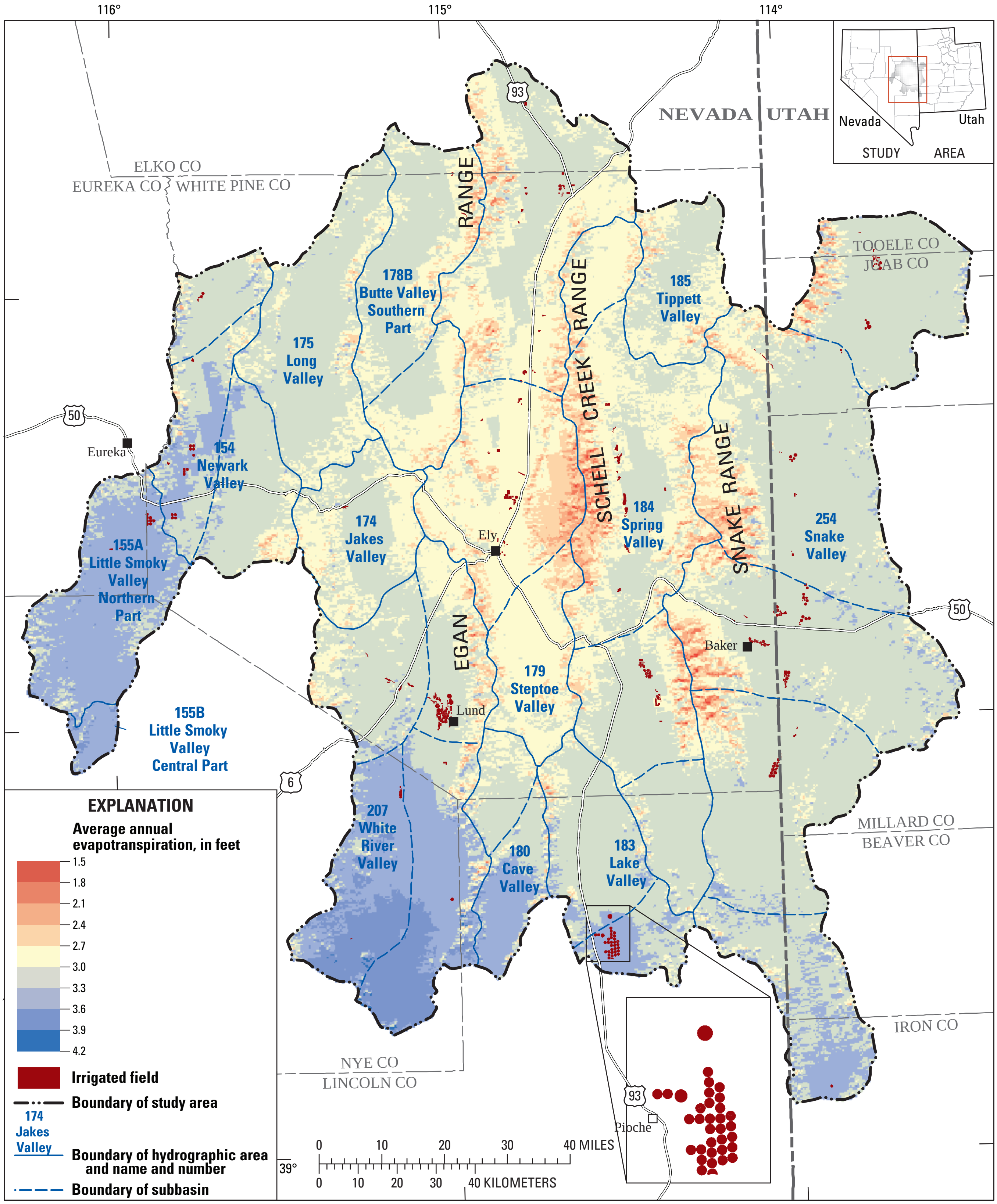


Figure 35. Water-use estimates by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2005.

Irrigation Water Use

Irrigated acreage was estimated from TM imagery using a procedure similar to that described in Moreo and others (2003). Details of the procedure are given in Welborn and Moreo (2007). About 600 irrigated fields were mapped for 2000, 2002, and 2005 (figs. 36 and 37). Actively irrigated fields identified from the 2005 TM imagery were assessed for accuracy by site visits made during the 2005 growing season. Less than 5 percent of the fields identified as active were determined to be inactive during the field inventory, and accordingly, were removed from the 2005 acreage inventory. Delineated acreage was compared to available Nevada Division of Water Resources (NDWR) crop inventories. Total irrigated acreage estimated by both methods agreed within about 13 percent (Welborn and Moreo, 2007). Irrigated acreage for 2005 totaled 32,000 acres, ranging from less than 200 acres in Butte, Cave, Jakes, Long, and Tippett Valleys to 9,200 acres in Snake Valley (appendix A, fig. 38). Irrigated acreage increased about 20 percent from 2000 to 2005. Cave, Long, and Tippett Valleys essentially had no active irrigation throughout this period.



Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

Figure 36. Distribution of average annual reference evapotranspiration (ETo) and extent of irrigated fields, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2005.

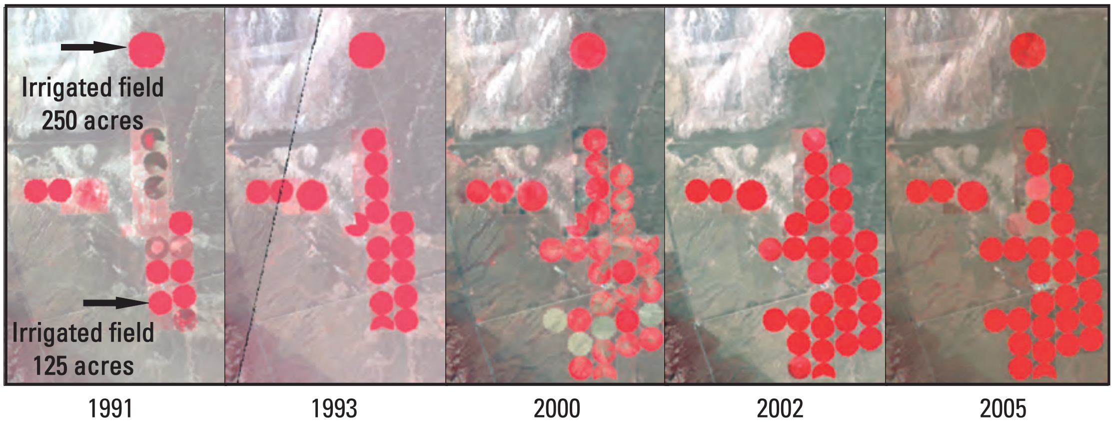


Figure 37. Irrigated fields in the Lake Valley delineated from TM imagery, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

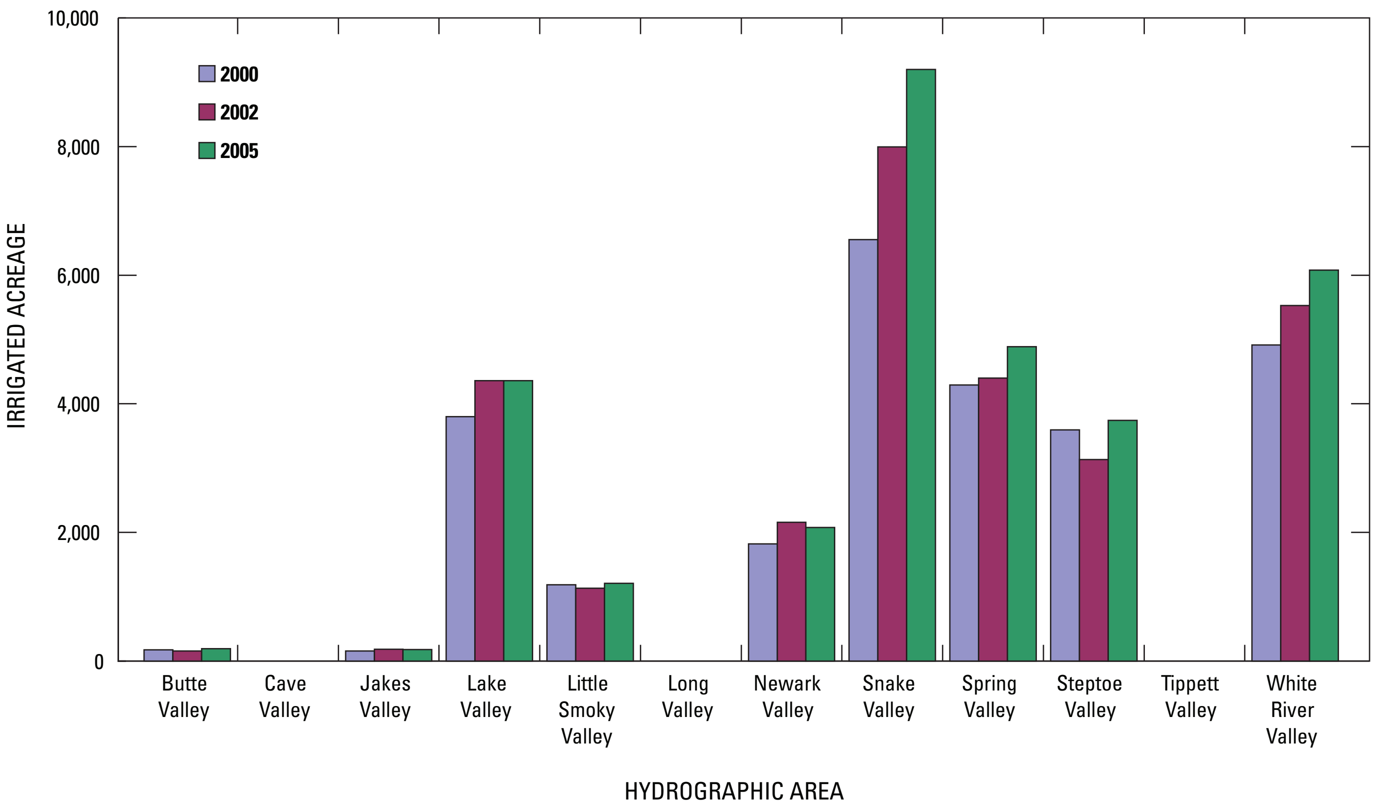


Figure 38. Estimates of irrigated acreage by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2000, 2002, and 2005.

The application rate, or the amount of water that needs to be applied to each field to obtain maximum crop yield, depends on the length of the growing season, climate, prevailing management practices, and crop type (U.S. Department of Agriculture, 1994). A range for the likely application rate of each field was developed from the equation:

$$AR = (ETc - Pe) \div Ep, \quad (2)$$

where

AR is application rate, in feet per year,

ETc is crop ET rate (also known as crop consumptive use), $ETc = ETo * Kc$, in feet per year,

ETo is reference crop ET, in feet per year,

Pe .is effective precipitation, in feet per year; and

Ep is project application efficiency, dimensionless.

ETc is estimated as the product of reference crop ET and the crop coefficient assuming standard conditions. Standard conditions assume optimal field, environmental, and management conditions (Allen and others, 1998). Estimates of consumptive use, based on the crop coefficient method, are used extensively throughout the world (http://www.fao.org/ag/agl/aglw/aquastat/water_use/index2.stm, accessed May 9, 2007). *ETo* is a measure of the evaporative power of the atmosphere and can be computed from solar radiation, temperature, wind speed, and humidity (Allen and others, 1998). *ETo* was estimated by extrapolating rates measured at more than 120 sites operated by the California Irrigation Management Information System (CIMIS; wwwcimis.water.ca.gov) into Nevada (Flint and Flint, 2007). The standardized Penman-Monteith reference equation is used by CIMIS to calculate *ETo* (Allen and others, 1998; Allen and others, 2005). *ETo* estimates for the study area average 2.8 ft/yr for the growing season (April-October) and 0.4 ft/yr for the non-growing season (Flint and Flint, 2007). *Kc* relates crop ET rate to the *ETo* rate, and depends on the growth and development of specific crops. CIMIS has developed *Kc* values specifically for calculating *ETc* as described above. For example, the average *Kc* is 1 for alfalfa during the growing season and for pastureland, as reported by Maurer and others (2006). The estimate for average annual crop consumptive use (*ETc*) in the study area is 2.9 ft/yr, and is in agreement with measured consumptive-use rates for alfalfa and pastureland given in Maurer and others (2006) for a similar climate.

Effective precipitation (*Pe*) is the amount of precipitation that remains in the root zone long enough to support crop growth. Factors such as precipitation amount, intensity, frequency and spatial distribution; topography and land slope; the depth, texture, and structure of the soil; depth to the water table; and water quality all affect *Pe* (1993). *Pe* is estimated to be 70 percent of the average annual precipitation (1993). *Pe* was estimated both for the growing and non-growing seasons because precipitation falling in the non-growing season increases the soil-water content, and any water retained in the root zone may be used for crop growth during the next growing season (1993). About two-thirds of average annual precipitation falls during the growing season Flint and Flint (2007).

Project application efficiency (*Ep*) is the ratio of the quantity of irrigation water stored in the root zone to quantities of water diverted or pumped, and varies with the irrigation method and irrigation system used. Irrigation-system inefficiencies result from surface runoff or infiltration past the root zone, direct evaporation from the air for sprinkler systems, and from water intercepted at soil and plant surfaces, wind drift, and conveyance losses. Application efficiency is difficult to estimate accurately because the efficiency of an irrigation system depends on many environmental factors and irrigator management decisions (U.S. Department of Agriculture, 1993). Because of these difficulties, *Ep* for the study area is estimated using standard published efficiency percentages (U.S. Department of Agriculture, 1993, 1997). Applying standard percentages, and field verifying irrigation methods and systems in the study area, *Ep* was estimated to range from 70 to 80 percent for center-pivot (continuously moving) sprinkler systems (fig. 39), from 55 to 70 percent for fixed and periodically moved sprinkler systems, and from 50 to 80 percent for the various types of flood irrigation systems. About 50 percent of irrigation applied in the study area is by center pivot sprinklers, about 30 percent by fixed and periodically moved sprinklers, and about 20 percent by flood irrigation.

Water withdrawn or diverted for irrigation is estimated as the product of irrigated acreage and an application rate estimated for each field. The average irrigation application rate for each HA ranged from 3.0 to 3.8 ft/yr. Higher application rates reflect higher *ETo* values, less efficient irrigation systems, lower effective precipitation amounts, or some combination thereof. The greatest average irrigation use estimated for 2005 is in Snake Valley at 34,000 acre-ft (fig. 35). Alfalfa and other hay production accounts for about 88 percent of the irrigated acreage. Pastureland accounts for about 10 percent, and corn, potatoes, and small grains for only about 2 percent of the total acreage irrigated. Uncertainties associated with estimating irrigation efficiencies for 2005 (appendix A and fig. 35) likely ranges from about 14 percent greater than or less than the estimates.

Irrigation return flow is that portion of the applied water that percolates beneath the root zone and ultimately returns to the ground-water flow system. Return flow is difficult to

Figure 39. Irrigation of a recently cut alfalfa field in Lake Valley, Nevada. Photograph taken by Michael T. Moreo, U.S. Geological Survey, September 26, 2006.

estimate because of the uncertainties in estimating application efficiency on a regional scale, travel time through the unsaturated zone, and the actual depth of the water table below the field. Stonestrom and others (2003) reports travel times on the order of several decades for 8–16 percent of applied irrigation water to return to the saturated zone in the Amargosa Desert in southern Nevada. Return flow rates probably differ between flood and sprinkler methods because sprinkler irrigation systems lose an estimated 10–15 percent of applied water directly to evaporation and wind drift (U.S. Department of Agriculture, 1993). Given these uncertainties and limited available data, an irrigation return flow estimate of 50 percent of water available for return flow is considered reasonable. For a hypothetical 125 acre field in Snake Valley planted in alfalfa and irrigated with a center-pivot sprinkler system with an Ep = 0.75. From equation 2, AR = (3.0 ft – 0.45 ft)/0.75 = 3.4 ft. The product of irrigated acreage (125 acres) and AR (3.4 ft) is 425 ac-ft. If 375 ac-ft (125 acres × 3.0 ft) is required by the crop, then 425 ac-ft needs to be withdrawn from the well to satisfy crop requirements because of irrigation system inefficiencies. Fifty percent of the unused portion of water withdrawn from the well (425 ac-ft withdrawn - 375 ac-ft = 50 ac-ft), or 25 ac-ft, is the estimated return flow.

Ground-water pumped from wells and diverted from valley springs accounts for an estimated 70 percent of irrigation water use in the study area during 2005, based primarily on field proximity to irrigation wells, springs, and natural and man-made drainage features, and where available—NDWR crop inventories (Welborn and Moreo, 2007). Perennial and intermittent streams sustained by upland springflow and snowmelt account for the remaining 30 percent of irrigation water use (Welborn and Moreo, 2007).

Non-Irrigation Water Use

Public supply, self-supplied domestic, stock, and mining water use account for only about 11 percent of total water use (appendix A). Public supply uses are metered and reported annually to the U.S. Environmental Protection Agency (USEPA) for inclusion in the Safe Drinking Water Information System (SDWIS) database (U.S. Environmental Protection Agency, 2004). Public supply estimates based on these records include water supplied by public water purveyors to households, commercial establishments, prisons, schools, and campgrounds (appendix A). An estimated 5,825 permanent residents and perhaps 3,812 primarily non-resident tourists were served by the public supply estimate in the study area.

Community populations served by public water supply systems were subtracted from the total HA population to determine the population of people served by self-supplied domestic water use. The entire population in the study area is estimated at 9,637 people (GeoLytics, 2001), and was assumed to use a self-supplied domestic water source. Therefore, self-supplied domestic use was estimated using this population and a water-use coefficient of 300 gallons per person per day (Nevada Department of Conservation and Natural Resources, 1999). For HAs with no public supply and relatively small populations (Butte, Cave, Jakes, Lake, Little Smoky, Long, and Tippett Valleys), annual domestic water use was assumed to equal 10 acre-ft. Stock water use was estimated as 0.32 percent (Nevada Department of Conservation and Natural Resources, 1999) of irrigation water use and this value was applied to all HAs: however, this estimate was modified by taking into account valleys with no irrigation or total livestock populations and locations of stock wells (U.S. Department of Agriculture, 1975, 2002). Mining water use typically is metered and reported annually to NDWR. Data obtained from NDWR indicate that mining water use was significant only in Steptoe Valley (appendix A).

Comparison of Ground-Water Discharge Estimates

Except for Snake Valley, ground-water discharge estimates for HAs are comparable with previous estimates, and generally fall below the median value of the range (fig. 40 and table 7) (Maxey and Eakin, 1949; Eakin and Maxey, 1951; Eakin, 1960, 1961, 1962; Hood and Rush, 1965; Rush and Kazmi, 1965; Eakin, 1966b; Eakin and others, 1967; Glancy, 1968; Nichols, 2000). The range defined from previous discharges for Snake Valley is based on two estimates (table 7). The variance in published discharge values primarily results from differences in methodology, but the overall range also is affected by the number and type of discharge estimates used to define the range. For example, some estimates for previous studies do not correct for precipitation and use total ET as their reported estimate of ground-water discharge, and some include pumping in their reported estimate of total ground-water discharge.

Estimated average annual ground-water discharge (440,000 acre-ft; appendix A) from the study area includes the quantity of spring discharge that infiltrates, recharges the shallow aquifer, and ultimately is evapotranspired by phreatophytic vegetation. The quantity of spring discharge that contributes to annual ground-water discharge cannot be measured directly, but generally can be estimated using available discharge measurements and assuming that discharge is log-normally distributed. A log-normal distribution for discharge has been observed in Florida (Scott and others, 2004) and Texas (Heitmuller and Reece, 2003). Available data for the study area show a cumulative annual discharge of about 170,000 acre-ft from 170 springs. However, diffuse discharge from ground water to phreatophytes occurs from a number of unmeasured springs and likely greatly exceeds the 170,000 acre-ft of measurable discharge from 170 springs. Most (150,000 acre-ft) of the measured discharge occurs at 60 springs at rates of 1 ft^3/s or greater. These springs are prominent landscape features, so it is unlikely that few if any springs discharging greater than 1 ft^3/s are not accounted for in this total. Using these data and a log-normal distribution for discharge, spring discharges would total 250,000 acre-ft if an additional 1,000 unmeasured springs and seeps existed. This general estimate of total spring discharge based on a log-normal distribution equals about one-half of the average annual ground-water discharge by phreatophytic vegetation (440,000 acre-ft).

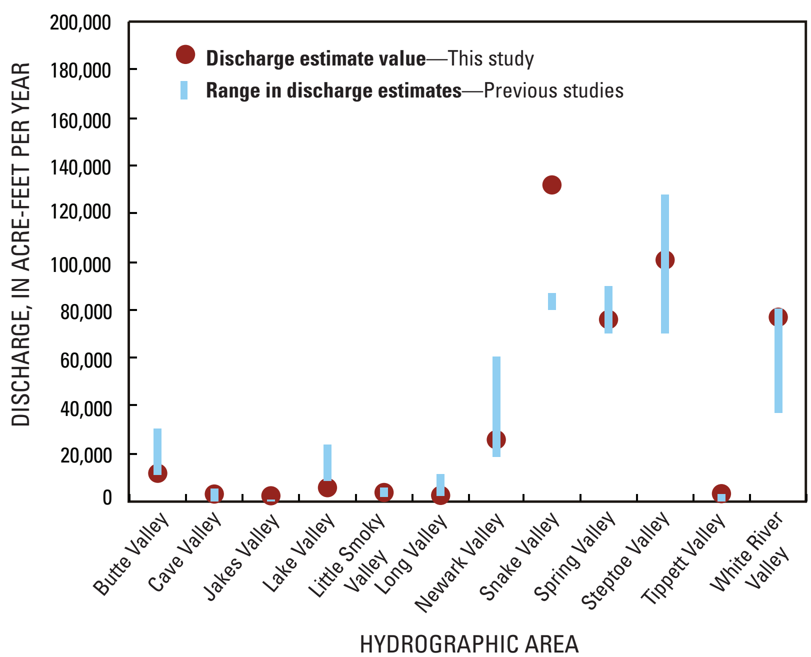


Figure 40. Range in pre-development ground-water discharge estimates developed from previous measurements.

Previous investigations estimated ground-water discharge from limited data and many estimates were not clearly defined. For example, early investigations estimated ground-water discharge in a basin (Maxey and Eakin, 1949) by delineating phreatophytic areas where depth to water was less than 50 ft and assuming average annual ground-water use of 0.1 ft (Jim Harrill, U.S. Geological Survey, retired, written commun., 2007). Nichols (1994) introduced new techniques for measuring evapotranspiration and quantifying ground-water discharge. Even with these advances, ground-water discharge estimates from Nichols (2000) were limited by early micrometeorological equipment, few annual estimates of evapotranspiration, underutilization of satellite imagery, and primitive remote-sensing technologies.

Annual ground-water discharge estimates were developed for this study using improved remote-sensing techniques for extrapolating calculated ET data, many measurements of local ET and precipitation, and more defined phreatophytic areas than used in previous estimates. Annual ET, precipitation, and ground-water discharge have been measured at 6 sites in the study area and more than 40 additional sites around Nevada since 1995 (Laczniak and others, 1999; Berger and others, 2001; Reiner and others, 2002; DeMeo and others, 2003; DeMeo and others, 2006; Laczniak and others, 2006; Maurer and others, 2006; Thodal and Tumbusch, 2006; Westenburg and others, 2006). Mapping phreatophytes in Nevada is continuously improving as more imagery becomes available and as the quality of imagery improves. Hundreds of additional satellite images have been analyzed since Nichols (2000) mapped phreatophytes with only two images. Additionally, unlike for earlier results, the uncertainty of annual ground-water discharge estimates for this study can be estimated because the uncertainty of each term can be quantified.

Interbasin Flow Estimates

Differences in average annual recharge and discharge provide a surplus or deficit of water for each HA that is balanced, for systems under pre-development conditions, by ground-water flow entering or exiting a basin (interbasin ground-water flow). For example, ground-water inflow may be significant to HAs where large spring discharges and phreatophytic areas can not be sustained by local recharge. Conversely, ground-water outflow may be significant from HAs where relatively deep water levels and small or non-existent phreatophytic areas have minimal potential for ground-water discharge by ET but generate excess recharge (Eakin, 1966b; Mifflin, 1968). For this study, a water surplus or deficit for each HA was balanced by interbasin ground-water inflow or outflow. This approach has been applied in previous studies on ground-water budgets for HAs in Nevada (Harrill and Prudic, 1998; Nichols, 2000).

For most HAs, average annual recharge exceeds ground-water discharge by 30 percent or more (tables 6 and 7). The high recharge in Steptoe Valley annually exceeds pre-development discharge by more than 40,000 acre-ft – the largest surplus of water for any HA – even though average annual discharge is about 70 percent of average annual recharge. An annual surplus of water also occurs in Butte and Long Valleys, where average recharge annually exceeds average discharge by more than 20,000 acre-ft. Except for Snake, Newark, and White River Valleys, average recharge exceeds average discharge by less than 14,000 acre-ft/yr for the remaining HAs. Even though these differences in water surplus are relatively small, for some HAs such as Cave, Long, Jakes, and Tippett Valleys, the percent difference between recharge and discharge may be relatively large. For these areas, average annual discharge is less than 20 percent of average annual recharge, indicating that most of the pre-development discharge from these valleys occurs as ground-water outflow.

In contrast to recharge-dominated HAs, pre-development discharge annually exceeds recharge in Newark, Snake, and White River Valleys. Newark and Snake Valleys are nearly balanced, with average annual recharge between 77 and 87 percent of average annual discharge, respectively. In White River Valley, however, the relatively low annual recharge is about 40 percent of average annual discharge, providing an annual water deficit of more than 40,000 acre-ft. The relatively large deficit in the pre-development estimates of recharge and discharge in White River Valley indicates that water discharging from springs and by evapotranspiration on the valley floor must be supported, in part, by subsurface inflow from adjacent valleys.

The potential for interbasin flow across HA boundaries is dependent on the magnitude of the surplus or deficit between average annual recharge and ground-water discharge, the transmissivity (the product of hydraulic conductivity and thickness) of aquifers along basin boundaries, and the hydraulic gradient of regional ground-water flow across basin boundaries. The magnitude of interbasin ground-water flow was estimated for all HAs in the study area using a water-budget accounting model, and these estimates were compared to estimates reported for previous studies, if available. For selected HA boundaries, estimates of the magnitude of interbasin flow were supported by evaluating transmissivity using the Darcy equation and by geochemical modeling.

Steady-State Water-Budget Accounting Model

A computer program described by Rosemary Carroll and Greg Pohll (Desert Research Institute, written commun., 2007) was used for the purpose of evaluating a water budget for the study area that includes intrabasin and interbasin ground-water flow. The model, which is describeda by Kevin Lundmark (Desert Research Institute, written commun., 2007) is a single-layer representation of the regional ground-water system that accounts for quantities of ground-water flow across intrabasin divides and HA boundaries using a simplified mass-balance mixing model that utilizes deuterium as a tracer. Deuterium values for ground-water recharge and regional ground-water flow systems were assigned to different parts of the study area based on measured values.

Within the study area, average annual recharge is greater than average annual discharge for 9 of the 13 HAs under pre-development conditions, indicating that a significant quantity of ground-water flows across intrabasin and interbasin boundaries. Intrabasin and interbasin ground-water flow, and flow to regions outside the study area, were: (1) constrained by the available volume of water (the difference between recharge and discharge estimates; pl. 4), (2) restricted to geologically and hydraulically suitable boundary segments, and (3) estimated using a deuterium-mixing model. Hydrogeologic restrictions to ground-water flow are indicated in figure 15. Hydraulic barriers to ground-water flow include relatively large areas of recharge creating mounds on the potentiometric surface and forming ground-water divides that separate the flow systems (pl. 3). The water-accounting model estimates quantities of ground-water inflow to, or outflow from, a HA but does not predict the location of ground-water flow across intrabasin or interbasin boundaries.

The accounting model was calibrated by approximately matching the simulated and measured deuterium concentrations and ground-water ET under pre-development conditions. For some HAs, model- predicted ground-water discharge rates were less than actual ground-water discharge rates estimated during this study. The differences were small, a few thousand acre-ft/yr or less, and are considered to be within the uncertainty associated with interbasin flow rates. The details of the model are described by Kevin Lundmark (Desert Research Institute, written commun., 2007).

Estimated intrabasin and interbasin flow rates are shown on pl. 3. The arrows on pl. 3 indicate only the direction of flow across the various boundaries and are not intended to suggest flow across a particular location within a boundary segement. Butte, Cave, Little Smoky, Long, and Steptoe Valleys receive no ground-water inflow and Newark and Tippett Valleys receive only small amounts of ground-water inflow. The remaining five HAs, Jakes, White River, Lake, Spring, and Snake, receive between 19,000 – 55,000 acre-ft/yr of ground-water inflow from adjacent HAs. Ground-water flow out of the study-area boundary includes about 2,000 acre-ft/yr toward the north, from Steptoe Valley to Goshute Valley, and about 42,000 acre-ft/yr toward the northeast from Tippett and Snake Valleys to the Great Salt Lake Desert regional flow system. About 9,000 acre-ft/yr of ground water exits the study area to the south from White River Valley, providing water to the lower part of the Colorado regional ground-water flow system. About 23,000 acre-ft/yr exits the northwest part of the study area from Butte Valley to the Ruby Valley regional flow system.

The model results represent a single solution that was obtained when the model was optimized to achieve a minimum difference between the simulated and observed deuterium concentrations and ground-water ET for the various HAs. However, model results are non-unique and other model simulations may yield similar residuals yet have significantly different flow patterns. Additionally, model-input deuterium values are sparse for several HAs, most notably Butte and Jakes Valleys. In addition to the uncertainty associated with a non-unique model and scarcity of deuterium data, the water-accounting model integrates data from multiple aspects of the study each with its own inherent uncertainty.

Hydrologic and Geochemical Constraints on Interbasin Flow Estimates

Hydrologic and geochemical assessments were completed to support interpretations of intrabasin ground-water flow rates and locations based on results of the water-accounting model and associated hydrogeologic evaluations. The quantity of interbasin ground-water flow at selected HA boundaries was assessed indirectly using the Darcy equation. Geochemical modeling was applied to assess whether representative changes occur in the isotopic or chemical compositions of ground-water flow along paths that cross interbasin boundaries. These assessments do not provide independent estimates of the quantity of ground-water flow crossing interbasin boundaries, but are considered secondary evidence to support the process of interbasin flow and provide general constraints on estimated flow rates.

Evaluation of Interbasin Flow Using Darcy's Law

Darcy's Law was used to indirectly evaluate interbasin flow rates estimated by the water-accounting model. The law describes the relation between volumetric discharge or flow rate, ground-water flow gradient, cross-sectional flow area, and aquifer hydraulic conductivity or transmissivity (Freeze and Cherry, 1979). Transmissivity was calculated by dividing interbasin flow by the product of the hydraulic gradient and effective width of the interbasin boundary segment and formulated as:

$$T = Kb = Q/(iW), \qquad (3)$$

where

T is the transmissivity, in feet squared per day,

K is the hydraulic conductivity, in feet per day,

b is the thickness of the aquifer units, in feet,

Q is the inter-basin ground-water flow, in cubic feet per day,

i is the hydraulic gradient, in foot per foot, and

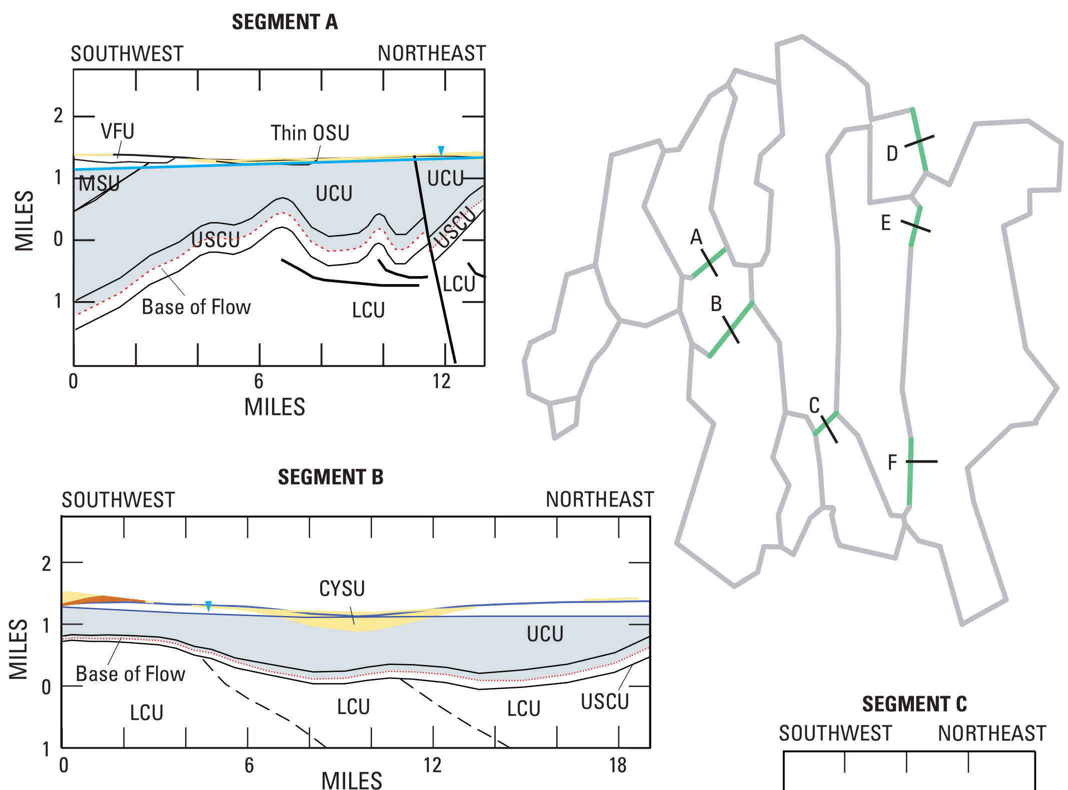
W is the effectie width of the aquifer units, in feet.

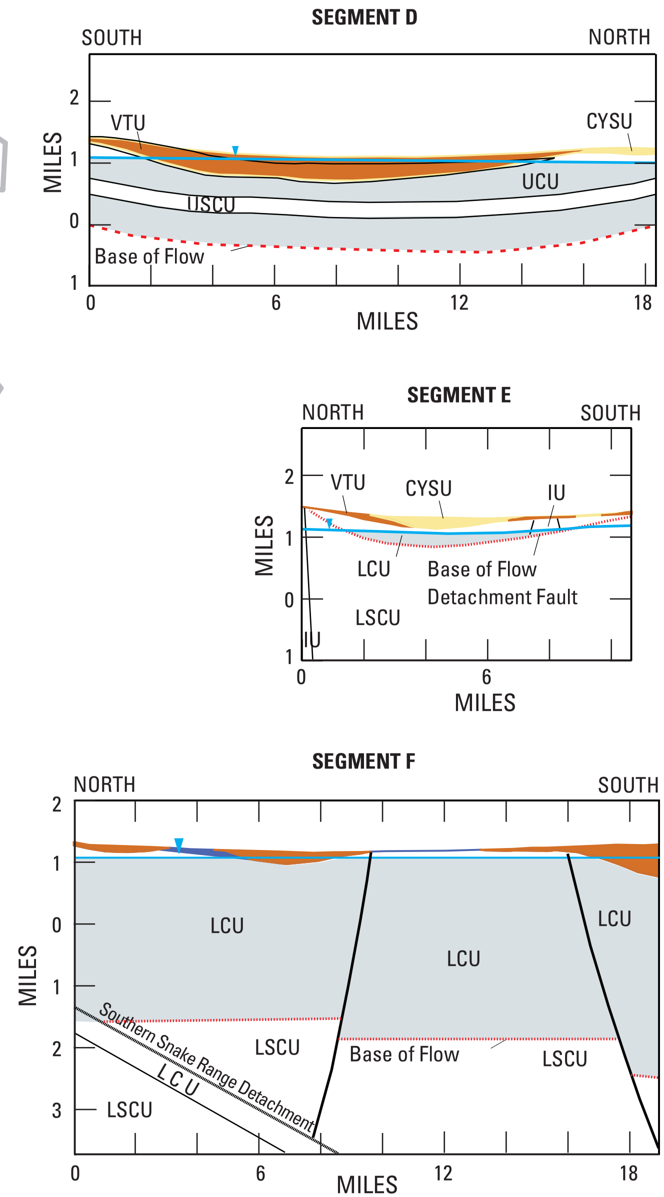
Transmissivity was estimated for six HA boundary segments and compared directly to aquifer test results. The interbasin flow values, cross-sectional areas, average thicknesses, hydraulic gradients, and corresponding hydraulic conductivity and transmissivity values for all boundary segments are shown in figure 41. Interbasin flow estimates from the water-accounting model (pl. 3) were used to calculate transmissivities. The hydraulic gradient across the HA boundary was estimated by calculating the ratio of the water-level difference and the distance between adjacent contour lines shown on the regional potentiometric-surface map (pl. 3). Aquifer widths were computed using cross sections extracted from a three-dimensional hydrogeologic framework model developed for this study (fig. 42).

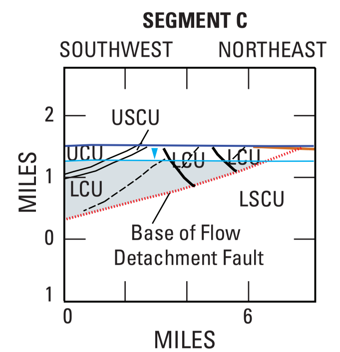
Transmissivities were estimated for two HA boundary segments in the western half of the study area (segments A and B, fig. 41). Aquifer units beneath the shared boundary of Jakes and Long Valleys (segment A, fig. 41) and the shared boundary of Jakes and White River Valleys (segment B, fig. 41) include the upper carbonate unit (UCU) and the permeable conglomerates of the Diamond Peak Formation found in the upper half of the upper siliciclastic confining unit (USCU). The base of the ground-water flow system is assumed to coincide with the base of the conglomerates within the USCU. Transmissivity estimates of 59,000 and 76,000 ft^2/d across segments A and B, respectively, are similar to estimates of Prudic and others (1995). The region used by Prudic and others (1995) is characterized as highly permeable, and nearby well data indicate that the carbonate rocks are characterized locally as uniformly high-porosity limestone.

Transmissivities were estimated for four HA boundary segments in the eastern half of the study area (segments C- F, fig. 41). The aquifer unit that underlies segments C, E, and F is the lower carbonate unit (LCU); whereas both the UCU and LCU aquifer units underlie segment D. The cross-sectional areas for boundary segments C and E are small (3 and 1 mi^2, respectively) due to relatively short boundary segment lengths and shallow depths to the base of the flow system. The base of the ground-water flow system is defined at the subsurface contact with a detachment fault and top of the LSCU. The cross-sectional area of boundary segment F is 53 mi^2 and the base of the flow system is relatively deep, coinciding with the top of the lower siliciclastic confining unit. The base of the flow system underlying segment D is unknown because each of the units, especially the UCU, likely contains numerous low-angle faults that may either disrupt the continuity of flow or promote brecciation of the rocks thereby increasing secondary permeability. The upper 0.6 mi of the LCU as well as the UCU are the aquifer units of interest underlying segment D. The transmissivities for segments C – F range from 1,400 to 5,100 ft^2/d. The apparent differences in transmissivity between segments A and B in the western half of the study area and segments C, D, E, and F in the eastern half of the study area may correspond to the westward thickening of the UCU and LCU carbonate units and the coarsening of the intervening siliciclastic unit (USCU).

The calculated transmissivities for the various boundary segements can be compared with values for carbonate rocks presented in Dettinger and others (1995) (fig. 43). Transmissivity values for the entire carbonate-rock province range from 10 to 250,000 ft^2/d. Based on aquifer tests at the seven wells located within or near the study area the range is from 200 to 17,000 ft^2/d (Dettinger and others, 1995). All estimated transmissivities fall within the limits for permeable carbonate units in the carbonate-rock province (fig. 43). This comparison suggests that the interbasin ground-water flow rates estimated using the water-accounting model are consistent with the hydrologic properties of the carbonate rocks underlying the six boundaries considered here.







Segment ID	Flow acre-ft/y	Area mi²	Thickness mi	Gradient ft/mi	k ft/d	T ft²/d
A	17,000	18	1.4	3	8	59,000
B	31,000	14	0.8	3	19	76,000
C	20,000	3	0.6	300	0.5	1,400
D [1]	13,000	18	1.0	50	0.4	1,800
E	14,000	1	0.1	50	7	5,100
F	29,000	53	2.8	40	0.4	5,100

[1]combined carbonate HGUs and excludes intervening confining unit.

Figure 41. Cross sections used to estimate transmissivities of hydrogeologic units.

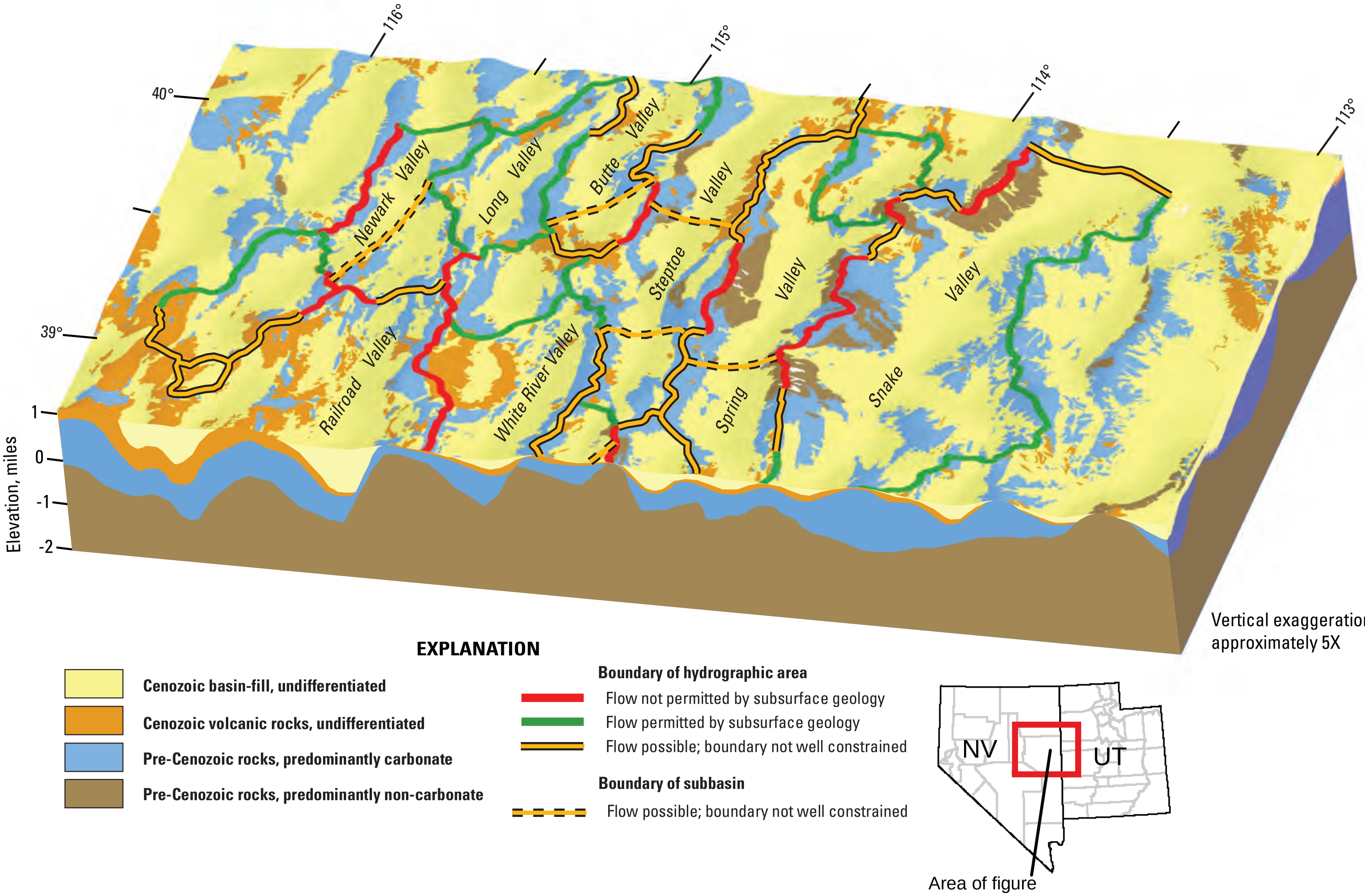


Figure 42. Perspective view of simplified three-dimensioinal geologic model of the study area.

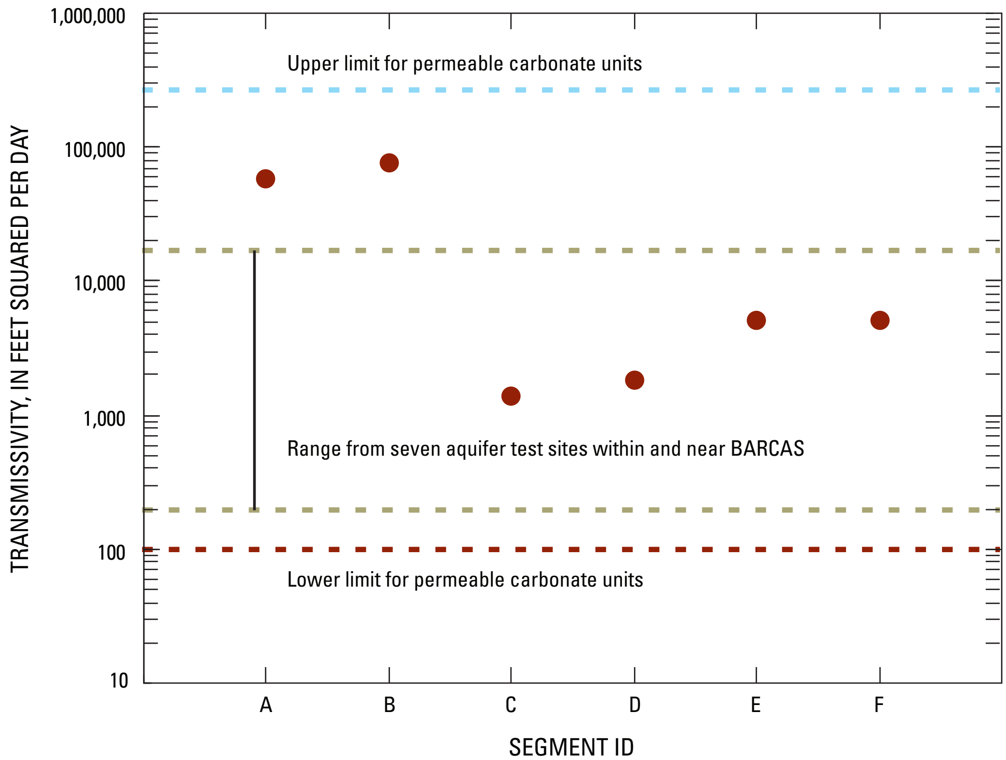


Figure 43. Transmissivity estimates for the boundary segments and published ranges in the carbonate-rock province. The upper and lower limits are based on data in Dettinger and others (1995).

Geochemical Modeling

Geochemical modeling was applied to support other evidence of interbasin and intrabasin ground-water flow in the study area. Geochemical process models can be used to evaluate potential ground-water flow across HA boundaries or intrabasin divides by determining whether measured or inferred changes in the isotopic or chemical compositions of ground water along these proposed flow paths are possible. Geochemical processes include the dissolution or precipitation of minerals, input and loss of gasses, and ion exchange. Ground water at the beginning of a flow path may be representative of water from a single source area or from a mixture of waters derived from multiple source areas. A geochemical model also may include calculations of ground-water travel times—the time elapsed for ground water to move along a flow path between two locations. Although results from a geochemical model may support ground-water flow along a particular path by matching known chemical and isotopic compositions of the ground water, modeling results are not unique. This non-uniqueness can lead to a range of possible models with a range of ground-water travel times.

Geochemical modeling focused on interbasin ground-water flow in the Spring Valley, Snake Valley, White River Valley, and Steptoe Valley HAs. These areas are the focus of the modeling because previous investigations (Harrill and Prudic, 1998; Nichols, 2000) concluded that ground water flows across boundaries between some of these HAs. Some geochemical data were available for the study area, including carbon isotope data needed to calculate ground-water travel times. Additional geochemical information was inferred where measurements were sparse or lacking. For some model evaluations, the isotopic and chemical composition of ground water was inferred from the mixing of known compositions of initial water and recharge water from upland springs.

Ground-water flow paths were evaluated and travel times calculated using the geochemical model NETPATH (Plummer and others, 1994); modeling results are summarized in table 8. After ground-water flow paths were selected, two geochemical models were evaluated using different mixing ratios of initial and recharge waters. The mixing amounts of initial and recharge waters are given as a range in percent in table 8. For example, percent initial and recharge waters used for the two

Table 8. Geochemical modeling results for inter-basin flow, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

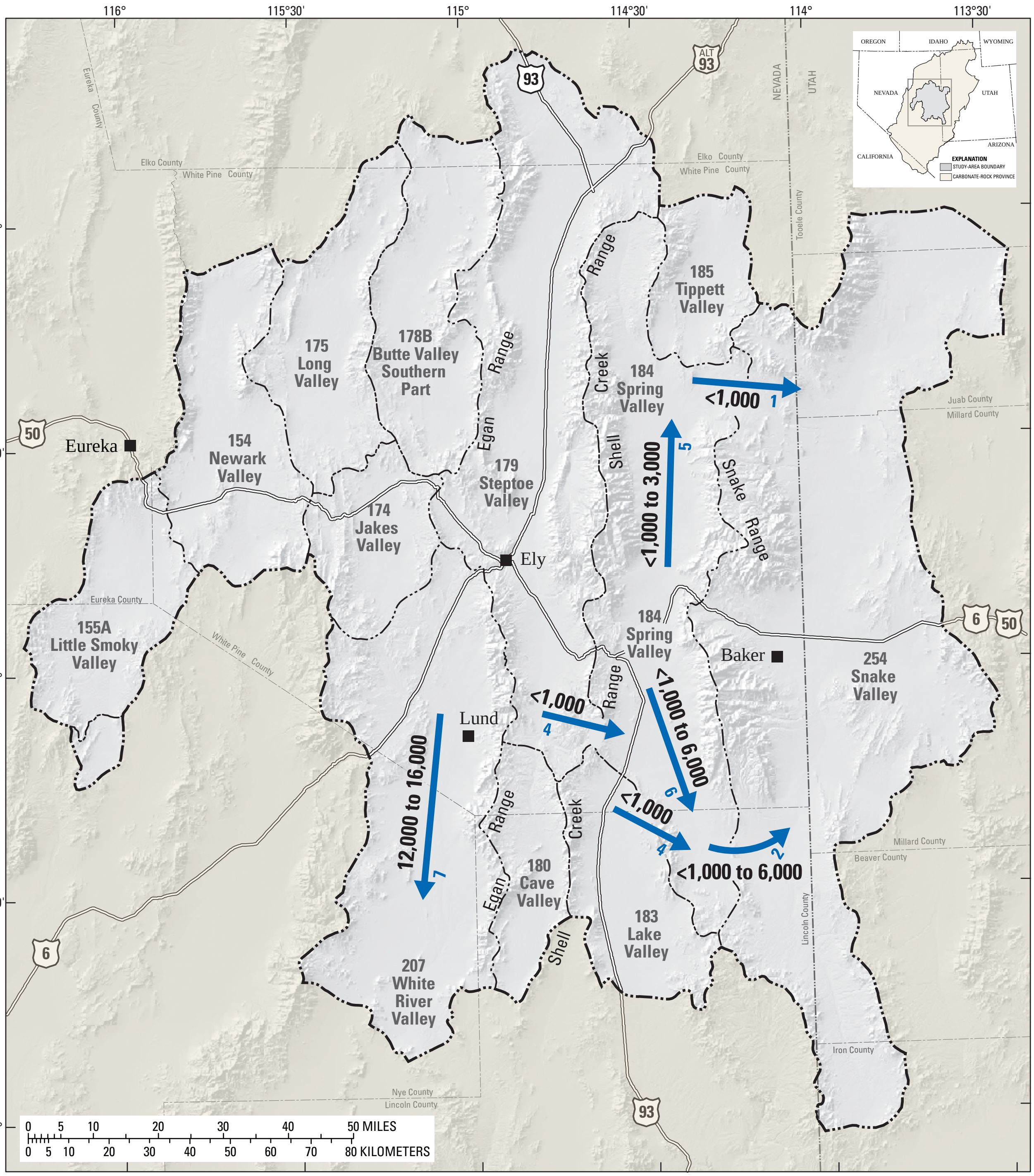
[**Flow path No.** matches corresponding number in figure 44. **Boundary or divide:** HA, hydrographic area; IB, intrabasin. **Geochemical model**, mixtures of initial and recharge waters, represents total mixture of initial and recharge waters for first (upper mixture) and second (lower mixture) model evaluations. **Initial water**, first point along selected ground-water flow path. **Recharge water** contributed from surrounding recharge areas. **Inorganic//Organic carbon travel time** represents time calculated from first (upper time) and second (lower time) model evaluations. **Abbreviations**: ft/yr, feet per year; <, less than; –, no data, NA, not applicable]

Flow path location and sites				Geochemical model – mixtures of water (percent)	Carbon travel time (years)		Inorganic ground-water flow velocity (ft/yr)	Geochemical model results
Flow path No.	Initial	Final	Boundary or divide	Initial – Recharge	Inorganic	Organic		
1	Northern Spring Valley	Northern Snake Valley	HA	0 – 100 30 – 70	<1,000 <1,000	2,000 4,000	100–200	Supports ground-water flow path
2	Southern Spring Valley	Southern Snake Valley	HA	0 – 100 100 – 0	<1,000 6,000	<1,000 2,000	20–100	Supports ground-water flow path
3	Southern Steptoe Valley	Southern Spring Valley	HA	70 – 30 100 – 0	<1,000 <1,000	NA	10–40	Supports ground-water flow path
4	Lake Valley	Southern Spring Valley	HA	95 – 5 100 – 0	<1,000 <1,000	<1,000 <1,000	50–60	Supports ground-water flow path
5	Southern part of northern Spring Valley	Northern part of northern Spring Valley	IB	0 – 100 60 – 40	<1,000 3,000	NA	40–150	Supports ground-water flow path
6	Central Spring Valley	Southern Spring Valley	IB	20 – 80 40 – 60	<1,000 6,000	6,000 6,000	10–200	Supports ground-water flow path
7	Central White River Valley	Southern White River Valley	IB	40 – 60 60 – 40	12,000 16,000	NA	10–20	Supports ground-water flow path
–	Southern Steptoe Valley	Lake Valley	HA	–	No model	NA	No model	No model
–	Cave Valley	Southern White River Valley	HA	–	No model	NA	No model	No model

model evaluations along the flow path from northern Spring Valley to northern Snake Valley was 30 and 70 percent in one evaluation, and zero and 100 percent in another evaluation. Details on chemical sampling and results of NETPATH model evaluations on geochemical reactions and calculated travel times were provided by Ron Hershey (Desert Research Institute, written commun., 2007).

Results of geochemical modeling support ground-water flow across selected HA boundaries, including ground water flowing (1) east from northern or southern Spring Valley into northern or southern Snake Valley, respectively, (2) southeast from southern Steptoe Valley to Spring Valley, and (3) southeast from Lake Valley to southern Spring Valley. Model results also support ground-water flow across selected intrabasin divides, including ground-water flowing north and south from central Spring Valley, and south from northern White River Valley into southern White River Valley. Moreover, chemical and isotopic data indicate that most of the ground water in Spring Valley originates as recharge in the surrounding Schell Creek and Snake Ranges, and that the Snake Range also is a major source of ground water in Snake Valley. However, geochemical model results for ground-water flow from southern Steptoe Valley to Lake Valley, and from Cave Valley to southern White River Valley were inconclusive because of sparse available chemical and isotopic data.

Ground-water travel times are presented for both dissolved inorganic carbon (DIC) and dissolved organic carbon (DOC) data to provide independent estimates of ground-water travel times (table 8 and fig. 44). Calculated ground-water ages using the DIC method represent an average ground-water travel time along a flow path; a DOC-calculated age reflects the average time elapsed since ground water was recharged. Thus, DOC ground-water ages should be the same or greater than DIC ages.



Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

EXPLANATION

Boundary of study area

174 Jakes Valley — **Boundary of hydrographic area and name and number**

1 — **Ground-water flow path**—Text by each arrow indicates approximately how long (in years) water takes to move along the flow path, as estimated using dissolved inorganic carbon-14 data

Figure 44. Ground-water travel times, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Ground-water ages calculated using the DOC method range from modern water (less than 1,000 years) in recharge areas to a maximum age of 16,000 years (Ron Hershey, Desert Research Institute, written commn., 2007). The oldest DOC ground-water ages are for ground water that has flowed through thick alluvial deposits of several thousand feet, in contrast to younger waters that have flowed primarily through fractured bedrock. In previous studies in east-central and southern Nevada, DOC-calculated ground-water ages for regional ground-water flow represent ground water discharging from bedrock outcrops or fractures from bedrock near the land surface (Thomas and others, 1996; Rose and others, 2002; Thomas and others, 2002). Because alluvial deposits in the study area likely contain some buried organic material with decayed carbon-14 that can be dissolved by ground water, the oldest DOC-calculated ground-water ages estimated for this study probably overestimate the actual age of the ground water. Generally, the DOC-calculated ages are in good agreement with DIC-calculated ages for younger ground waters, but significantly overestimate ages of the older ground waters. Ground-water flow velocities determined from travel times along potential interbasin and intrabasin flow paths range from 10 to 200 ft/yr.

Comparison of Interbasin Flow Estimates

No single report presents estimates of interbasin ground-water flow for all HAs included in the BARCAS study, but several previous studies have reported on ground-water flow for multiple basins in the study area, or have been completed for a single basin in the study area (table 9). Nichols (2000) and Thomas and others (2001) report interbasin flow estimates for 8 and 5 of the HAs in the study area, respectively. Locations, volumes, and directions of interbasin flow presented in Harrill and others (1998) were based on estimates compiled from reconnaissance reports, and generally represent evaluations of single HAs that also are included in the BARCAS study.

The interbasin flow estimates presented in Nichols (2000) assumed that (1) differences between recharge and discharge were equal to the interbasin ground-water flow into or out of the HA and (2) the system is in hydrologic equilibrium such that discharge combined with interbasin flow can be used as a surrogate for recharge (Nichols, 2000, p. C21). Excess or deficient recharge for a given HA was compensated by interbasin flow into or out of the area if these flows were proposed in earlier studies or were otherwise permissible, geologically and hydrologically. Nichols (2000) found that the interbasin flow volumes were consistent with, and tended to corroborate most of the boundaries defined by Harrill and others (1988).

Interbasin flow volumes and assumed ground-water flow directions were evaluated using a deuterium mass-balance model by Thomas and others (2001). Boundary conditions and input to their model were based on prominent geologic structure, stratigraphic continuity, and hydraulic gradients described in previous studies (Eakin, 1966; Thomas and others, 1986; Kirk and Campana, 1990; Dettinger and others, 1995; Thomas and others, 1996). Where recharge and ground-water inflow into a basin exceeded ET, excess ground water was assigned as subsurface flow to the next downgradient valley (Thomas and others, 2001).

Directions of interbasin flow presented in Harrill and others (1988), Nichols (2000), and Thomas and others (2001) are similar. The primary directions of flow in the study area are (1) from north (Long Valley) to south (White River Valley) in the Colorado regional flow system; and (2) toward the north-northeast from Steptoe, Tippett, and northern Snake Valleys in the Great Salt Lake Desert regional flow system. Interbasin flow in these reports also was described as flowing southwest to Railroad Valley, northwest to Clover and Ruby Valleys, and east from Spring Valley, through Snake Valley, and into western Utah. However, the magnitude of interbasin flow differs slightly among the reports. These differences were calculated by adding all of the inflows and subtracting all of the outflows, positive values indicate greater inflows and negative values indicate greater outflows (table 10).

Directions of interbasin ground-water flow also are similar to those reported by previous studies for the Colorado and Great Salt Lake Desert regional flow systems. However, based primarily on interpretations of HA boundary geology, regional ground-water flow, and water-accounting modeling, some interbasin flow directions discussed in this report differ from previous studies (fig. 45). For example, outflow from southern Steptoe Valley to Lake Valley, from southern Steptoe Valley to Spring Valley, and from Lake Valley to Spring Valley have not been posited or are of much greater rates compared with previous studies. Based on regional flow systems defined by Harrill and others (1988), these interbasin flow directions occur across the boundaries of the Goshute and Colorado regional flow systems (Steptoe to Lake Valleys), of the Goshute to Great Salt Lake Desert regional flow systems (Steptoe to Spring Valleys), and of the Colorado to the Great Salt Lake Desert regional flow systems (Lake to Spring Valleys).

Table 9. Historical estimates of annual interbasin flow from and to hydrographic areas, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[Shading indicates flow through valley fill. Locations of hydrographic areas shown in figure 2]

Hydrographic area name	Interbasin flow from and to hydrographic areas (HA), in acre-feet per year											
	Harrill and others (1988)				Nichols (2000)				Thomas and others (2001)			
	Inflow	From HA	Outflow	To HA	Inflow	From HA	Outflow	To HA	Inflow	From HA	Outflow	To HA
Butte Valley–southern	0	–	0	–	0	–	22,500	Cover Valley (177)	([1])		([1])	
							2,000	Ruby Valley (176)				
Cave Valley	0	–	14,000	White River Valley (207)	([1])		([1])		0	–	15,000	Pahroc Valley (208)
Jakes Valley	8,000	Long Valley (175)	25,000	White River Valley (207)	14,000	Long Valley (175)	51,200	White River Valley (207)	12,000	Long Valley (175)	35,000	White River Valley (207)
							700	Railroad Valley–northern (173B)				
Lake Valley	0	–	3,000	Patterson Valley (202)	([1])		([1])		0	–	17,000	Patterson Valley (202)
Little Smoky Valley–northern and central combined	4,000	Antelope Valley (151)	1,000	Newark Valley (154)	0	–	1,500	Newark Valley (154)	([1])		([1])	
	200	Stevens Basin (152)					5,500	Railroad Valley–northern (173B)				
Long Valley	0	–	8,000	Jakes Valley (174)	0	–	10,000	Newark Valley (154)	0	–	12,000	Jakes Valley (174)
			([2])	Newark Valley (154)			14,000	Jakes Valley (174)			8,000	Unspecified HA
							13,000	Railroad Valley–northern (173B)				
Newark Valley	1,000	Little Smoky–northern (155A)	([2])	Railroad Valley–northern (173B)	10,000	Long Valley (175)	0	–	([1])		([1])	
					1,500	Little Smoky–northern (155A)						
Snake Valley	4,000	Spring Valley (184)	42,000	Tule Valley (257)	14,000	Spring Valley (184)	([3])		([1])		([1])	
	11,000	Pine Valley (255)	([2])	Fish Springs Flat (258)	3,600	Tippett Valley (185)						
	8,500	Wah Wah Valley (256)	10,000	Great Salt Lake Desert–western (261A)								
Spring Valley	2,000	Tippett Valley (185)	4,000	Snake Valley (254)	0	–	14,000	Snake Valley (254)	([1])		([1])	
Steptoe Valley	([2])	Butte Valley–southern (178B)	Minor	Goshute Valley (187)	0	–	4,000	Goshute Valley (187)	([1])		([1])	
Tippett Valley	0	–	3,000	Antelope Valley–southern (186A)	0	–	6,000	Deep Creek Valley (253)	([1])		([1])	
			2,000	Great Salt Lake Desert–western (261A)			3,600	Snake Valley (254)				
			2,000	Spring Valley (184)								
White River Valley	25,000	Jakes Valley (174)	40,000	Pahroc Valley (208)	51,200	Jakes Valley (174)	([3])		35,000	Jakes Valley (174)	17,000	Pahroc Valley (208)
	14,000	Cave Valley (180)										

[1]Hydrographic area not evaluated.

[2]Flow volume not specified.

[3]Inflow but not outflow calculated for hydrographic area.

Table 10. Differences in historical annual inter-basin flow estimates, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[Shading indicates flow through valley fill. Negative numbers indicate more outflow than inflow; positive numers indicate more inflow than outflow]

Hydrographic area name	Total difference in interbasin flow, in acre-feet per year		
	Harrill and others (1988)	Nichols (2000)	Thomas and others (2001)
Butte Valley–southern	0	-27,500	([1])
Cave Valley	-14,000	([1])	-15,000
Jakes Valley	-17,000	-37,900	-23,000
Lake Valley	-3,000	([1])	-17,000
Little Smoky Valley–northern and central combined	3,200	-7,000	([1])
Long Valley	-8,000	-37,000	-20,000
Newark Valley	1,000	11,500	([1])
Snake Valley	-28,500	([2])	([1])
Spring Valley	-2,000	-14,000	([1])
Steptoe Valley	0	-4,000	([1])
Tippett Valley	-7,000	-9,600	([1])
White River Valley	-1,000	([2])	18,000

[1]Hydrographic area not evaluated.

[2]Inflow but not outflow calculated for hydrographic area.

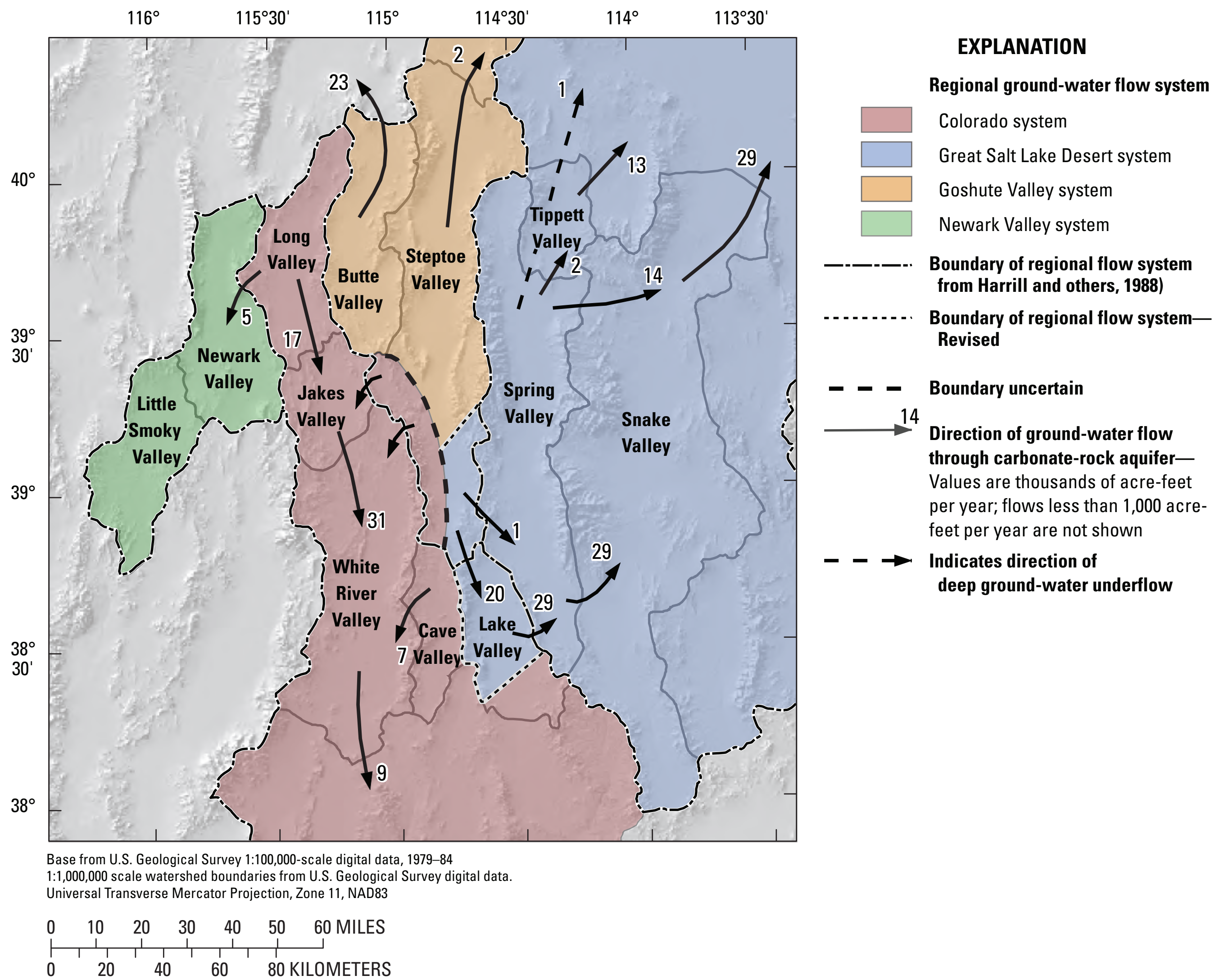


Figure 45. Regional ground-water flow exiting the study area through the Colorado, Great Salt Lake Desert, and other regional flow systems, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

BARCAS interbasin flow estimates are higher than previous estimates for some of the HAs, lower for other HAs, and only a few estimates fall within the range of previous estimates in the study area (fig. 46 and table 9). BARCAS inflow estimates are higher in Jakes, Lake, Snake, Spring, and White River Valleys than previous estimates; and in Newark Valley, estimated inflows are near the middle of the range of previous estimates. BARCAS ground-water outflow estimates are significantly higher than published estimates in Spring, and Steptoe Valleys, and slightly higher in Lake and Tippett Valleys; in Butte, Jakes, and Long Valleys the estimated outflows are within the range of published estimates and in Snake, Cave and White River Valleys the outflows are lower than published estimates (fig. 46).

Differences between estimates for this study and previous estimates primarily are attributed to variations in the applied methods. For example, some previous estimates neglected hydraulic connections between adjacent HAs. For instance, inflow from upgradient areas was not considered when constructing the water budgets. Additionally, recharge estimates for this study tend to be higher and discharge estimates tend to be lower than previous estimates for individual HAs. This larger difference between recharge and discharge components essentially increases the amount of available ground water from the study area to adjacent HAs. The greater outflows estimated are considered reasonable because the study area is a primary recharge area for the Colorado, Great Salt Lake Desert, and Goshute Valley regional ground-water flow systems (pl. 3).

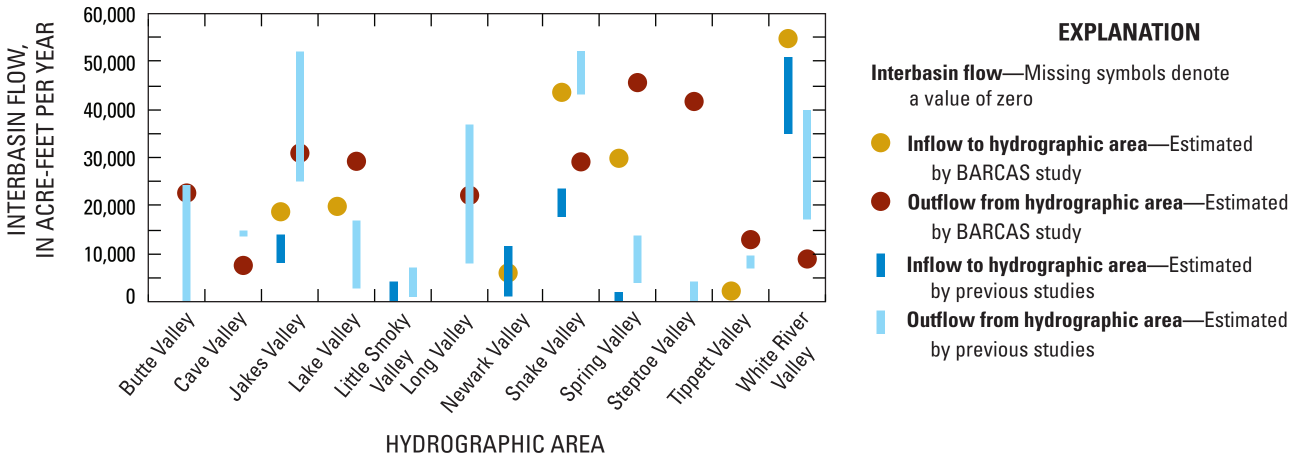


Figure 46. Comparison of interbasin ground-water flow estimates.

Regional Ground-Water Recharge and Discharge

Average annual recharge and ground-water discharge for HAs were summed and compared to evaluate the water budget for the study area, referred to in this report as the regional ground-water budget. Based on estimates for HAs, average annual ground-water recharge to the study area totals about 530,000 acre-ft, and average annual ground-water discharge totals about 440,000 (fig. 47). Assuming that these estimates represent pre-development conditions, the difference between estimated recharge and discharge indicates that about 90,000 acre-ft of ground water exits the study area annually as subsurface outflow. An outflow of this magnitude from the study area is not unexpected, considering that the area serves as the headwaters of two regional ground-water flow systems, the Colorado and Great Salt Lake Desert systems. Assuming that subsurface outflow supports these large regional flow systems, the likely major pathways for outflow are through Snake Valley to the northeast and White River Valley to the south (pl. 3). Ground-water outflow to the northeast from Tippett Valley also flows toward the terminal discharge area in the Great Salt Lake Desert flow system. Other major areas of ground-water outflow include the northern boundaries of Steptoe and Butte Valleys.

The net amount of water removed by ground-water pumping was estimated to evaluate the significance of water withdrawals to ground-water discharge under pre-development conditions. Net ground-water pumpage represents the estimated amount of ground-water pumped from wells or diverted from regional spring sources minus any water recharging the ground-water flow system as a result of water returned from mining, irrigation applications, or public supply. In making this estimate, local spring and surface runoff sources are assumed to account for 30 percent of the water-use estimates given in figure 35, and return flow as 50 percent of any unconsumed water. Net regional ground-water pumpage estimated for the HAs in the study area vary from near zero, primarily in unfarmed valleys, to nearly 24,000 acre-ft in Snake Valley; and in all HAs, are substantially less than the total water-use estimates (fig. 35). Only in Lake Valley is the net regional ground-water pumpage greater than the estimated average annual ground-water discharge under pre-development conditions. Net regional ground-water pumpage for the entire study area is estimated at about 80,000 acre-ft, or about 60 percent of the 2005 water-use estimate.

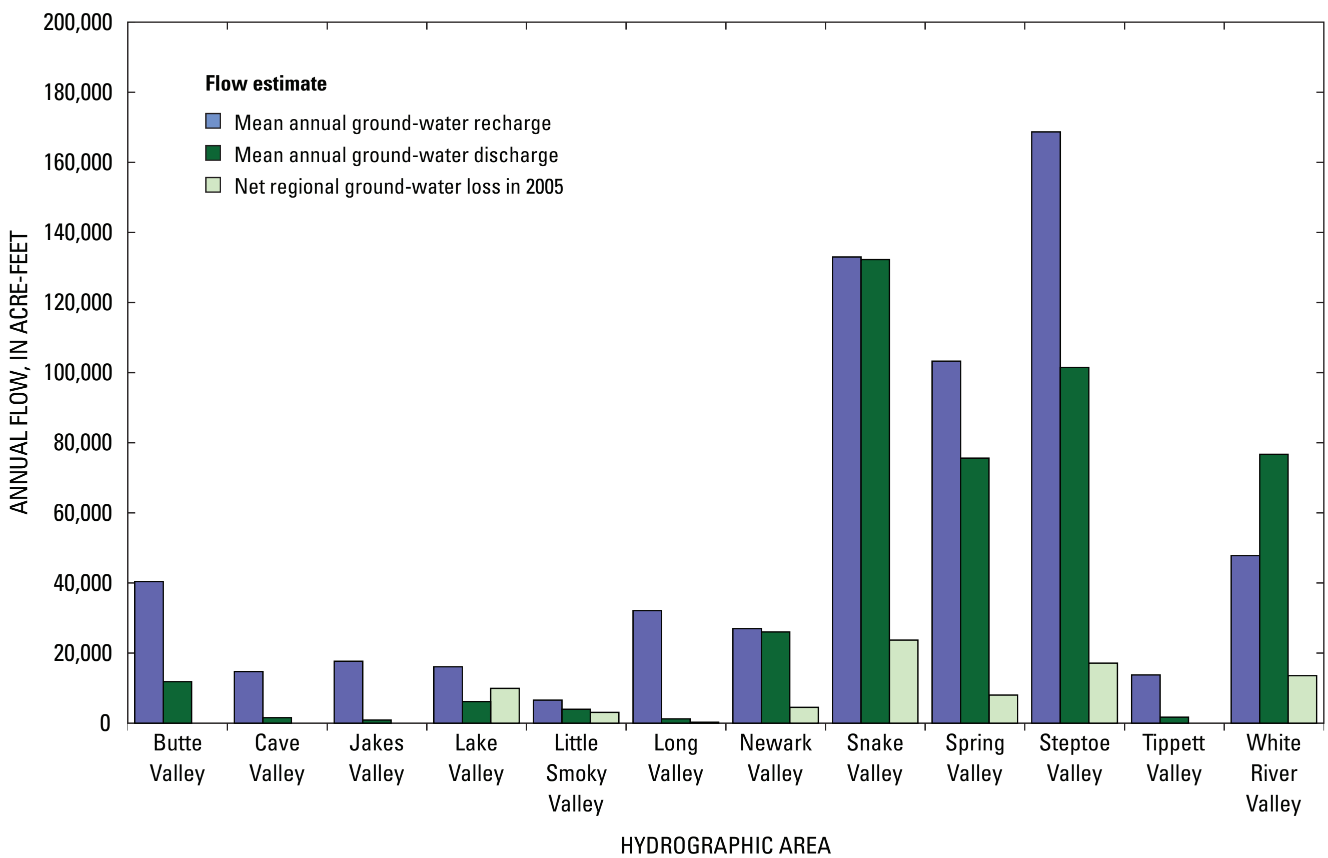


Figure 47. Annual estimates of average ground-water recharge and average ground-water discharge, and the 2005 net regional ground-water pumpage by hydrographic area in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

When including the estimated net ground-water pumpage for 2005 in the regional water budget, the recharge and discharge components of the ground-water budget are nearly balanced over the entire study area—average annual recharge (530,000 acre-ft) is approximately equal to average annual ground-water discharge under pre-development conditions (440,000 acre-ft) plus estimated net pumpage for 2005 (80,000 acre-ft). That is, the estimated net pumpage for 2005 is nearly equal to the estimated average annual ground-water outflow from the study area (90,000 acre-ft). On a regional scale, this condition suggests that long-term ground-water withdrawals of equal volume to those estimated for 2005 could potentially capture the estimated average annual volume of ground water exiting the study area. Moreover, this condition also could, in some combination, reduce subsurface outflow, reduce spring discharge, reduce phreatophytic discharge, or increase subsurface recharge from adjacent basins. However, actual reductions in the volume ground-water outflow, or in the volume of other pre-development discharge components such as interbasin flow, spring discharge, or evapotranspiration, would be controlled by a number of factors, particularly, the spatial distribution of ground-water withdrawals, and the volume of ground-water removed from storage. For example, reductions in outflow would be less likely in Butte or Tippett Valleys where net pumpage was zero in 2005 (fig. 47). Reductions in outflow would be more likely in subbasins having both high pumpage and relatively large outflow such as in Snake Valley where net pumpage was 24,000 acre-ft in 2005 and average annual ground-water outflow was estimated at 29,000 acre-ft. Additionally, the relatively large volume of water stored in the basin-fill aquifer (appendix A) would likely inhibit near-future reductions in ground-water outflow or in other pre-development discharge components if withdrawals are taken from the basin-fill aquifer. For example, water-level measurements show declines around major areas of pumping indicating that storage currently (2005) is a primary source of pumped ground water in the study area. Moreover, historical pumping has been periodic and often used only as a supplement to spring and surface sources, ground-water pumping in prior years was substantially less than that estimated in 2005, and much of the current pumping occurs outside major discharge areas. The conclusion being that ongoing pumping currently (2005) has not significantly altered ET rates, regional springflows, or distribution of native vegetation. Evaluation of the timing and location of potential reductions in pre-development ground-water discharge would be best accomplished through the application of a numerical ground-water flow model; however, the development of a regional model was beyond the scope of the current study.

Some uncertainty exists on estimated differences between average annual recharge and pre-development discharge. These estimates were made independently, and each methodology has inherent limitations and associated uncertainty. Recharge estimates were model-derived; the accuracy of these estimates depends on the accuracy with which a number of hydrologic, atmospheric and soil parameters were estimated. Estimates of pre-development discharge primarily were derived through field measurements and, as a result of a more direct method of measurement, the uncertainty of estimated pre-development discharge is likely less than the uncertainty of estimated recharge. Future studies may reduce uncertainties of estimated recharge and discharge by evaluating a regional ground-water flow system bounded by ground-water divides, such as the Colorado or Great Salt Lake Desert regional flow systems. Evaluating entire regional flow systems provides the constraint that ground-water inflow and outflow across the study area boundary is minimal; therefore, cumulative recharge and pre-development discharge must balance for HAs within the regional flow system.

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Harry E. Cook – Stratigraphy of the Great Basin

Edward A. DuBray – Volcanic and plutonic rocks of the Great Basin, geology of Seaman Range

Karen Lund – Structural geology of the Grant Range, stratigraphy of Proterozoic sedimentary rocks

Edward A. Mankinen – Gravity studies

Edwin H. McKee – Geology of the Great Basin

Connie J. Nutt – Structural geology in western and northern study area

F.G. Poole – Stratigraphy of the Great Basin

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Appendix A. Component Estimates of Recharge, Discharge, Water Use, and Aquifer Storage.

The spreadsheet distributed as part of this report is in Microsoft® Excel 2003 format. Appendix A data are available for download at URL: http://pubs.water.usgs.gov/ofr20071156.

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Glossary

Accommodation zone: A zone of geologic structures that typically cross-cuts a region and separates two areas of different type or amount of disruption or deformation.

Alluvial: Relating to, consisting of, or formed by sediment deposited by flowing water.

Anastomosing: Pertaining to a network of branching and rejoining fault or vein surfaces or surface traces.

Anastomosis: A form of network in which streams both branch out and reconnect.

Andesite: An igneous, volcanic rock. The mineral assembly typically is dominated by plagioclase plus pyroxene and/or hornblende.

Aquifer: Rock or sediment that is saturated and can transmit sufficient water to supply wells.

Argillaceous: Pertaining to, largely composed of, or containing clay-size particles or clay minerals

Ash-flow tuff: a volcanic rock consisting of ash and other volcanic detritus deposited from an explosive volcanic eruption. It is consolidated and sometimes densely compacted and fused.

Basement: In geology, an underlying complex that behaves as a unit mass and does not deform by folding. In geophysical studies, the term can refer to consolidated, older rocks that lie beneath young basin fill.

Breccia: Clastic rock made up of angular fragments of such size that an appreciable percentage of rock volume consists of particles of granule size or larger.

Caldera: Roughly circular, steep-sided volcanic basin with diameter at least three times depth. Results from very large magnitude, explosive volcanic eruptions.

Colluvium: Rock detritus and soil accumulated at the foot of a slope.

Confining Unit: The geologic layer of low permeability that is adjacent to an aquifer and retards flow into and out of the aquifer.

Detachment: Detachment structure of strata owing to deformation, resulting in independent styles of deformation in the rocks above and below. It is associated with faulting and structural removal of rock strata.

Deuterium: An isotope of hydrogen that has one proton and one neutron in its nucleus and that has twice the mass of ordinary hydrogen.

Domain: An areal subdivision based on shared geologic traits, such as type or intensity of faulting.

***en echelon*:** Said of geologic features that are in an overlapping or staggered arrangement, e.g., faults. Each is relatively short, but collectively they form a linear zone, in which the strike of the individual features is oblique to that of the zone as a whole.

Exotic: Applied to a boulder, block, or larger rock body unrelated to the rocks with which it is now associated, which has been moved from its place of origin by one of several processes. In plate tectonics, refers to land masses that were not originally part of the North American continent.

Facies: Assemblage of mineral, rock, or fossil features reflecting environment in which rock was formed. See sedimentary facies, metamorphic facies.

Foliation: Layering in some rocks caused by parallel alignment of minerals; textural feature of some metamorphic rocks. Produces rock cleavage.

Geosyncline: Refers to a basin in which thousands of feet of sediments have accumulated, with accompanying progressive sinking of basin floor. Common usage includes both accumulated sediments themselves and geometrical form of basin in which they are deposited.

Graben: Elongated, trench like, structural form bounded by parallel normal faults created when block that forms trench floor moves downward relative to blocks that form sides.

Great Basin: A unique internally drained physiographic feature of the western United States.

Highly attenuated domain: A region in which the stratigraphic section has been thinned as a result of tectonic processes, typically during extension, or stretching, of the earth's crust.

Hinterland: A subjective term referring to the relatively undisturbed terrain on the back of a folded mountain range.

Hydraulic head: Height above a datum plane (such as mean sea level) of the column of water that can be supported by the hydraulic pressure at a given point in a groundwater system.

Hydraulic conductivity: A coefficient of proportionality describing the rate at which water can move through a permeable medium such as an aquifer. Hydraulic conductivity is a function of both the intrinsic permeability of the porous medium and the kinematic viscosity of the water which flows through it.

Hydrogeologic unit: Any rock unit or zone which by virtue of its hydraulic properties has a distinct influence on the storage or movement of ground water.

Imbricate Structure: A tectonic structure displayed by a series of nearly parallel and overlapping minor thrust faults, high-angle reverse faults, or slides, and characterized by rock slices, sheets, plates, blocks, or wedges that are approx. equidistant and have the same displacement and that are all steeply inclined in the same direction.

Indurated: Said of a rock or soil hardened or consolidated by pressure, cementation, or heat.

Infiltration: Movement of water through the soil surface into the ground.

Karst: A type of topography that is formed on limestone and other rocks by dissolution and that is characterized by sinkholes, caves, and underground drainage.

Lacustrine: Related to lakes. For instance, lacustrine sediments refers to deposits formed beneath a lake.

Linear regression: A mathematical analysis that allows the examination of the relationship between a variable of interest and one or more explanatory variables. Of interest is the quantification of this relation into a model form to estimate or predict values for a variable based on knowledge of other variables, for which more data are available.

Listric fault: A curved downward-flattening fault, generally concave upward. Listric faults may be characterized by normal or reverse separation.

Lithosphere: Rigid outer layer of earth; includes crust and upper part of mantle.

Lysimeter: A device for measuring the infiltration of water through soils and for determining the soluble constituents removed in the drainage.

Magmatism: Of, pertaining to, or derived from magma. See also: igneous.

Metamorphic core complexes: a domelike exposure of metamorphic rocks exposed beneath a detachment fault; typically the result of large-magnitude extension, or stretching, of the earth's crust.

Metamorphosis: A process whereby rocks undergo physical or chemical changes or both to achieve equilibrium with conditions other than those under which they were originally formed. Agents of metamorphism are heat, pressure, and chemically active fluids.

Metasediment: A sediment or sedimentary rock that shows evidence of having been subjected to metamorphism.

Miogeosyncline: That part of a geosyncline in which volcanism is absent, generally located near craton.

Orogeny: Process by which mountain structures develop.

Orographic: Associated with or induced by the presence of mountains, such as orographic rainfall.

Permeability: For earth material, ability to transmit fluids.

Physiographic province: A region of which all parts are similar in geologic structure and which has consequently had a unified geomorphic history; a region whose pattern of relief features or landforms differs significantly from that of adjacent regions

Phreatophyte: A plant that obtains its water from the water table or the layer of soil just above it.

Physiography: Same as physical geography.

Playa: The lower part of an inland desert drainage basin that is periodically flooded.

Pluton: A body of igneous rock formed beneath earth surface by consolidation from magma. Sometimes extended to include bodies formed beneath surface by metasomatic replacement of older rock. A body of medium- to coarse-grained igneous rock that formed beneath the surface by crystallization of magma.

Potentiometric surface: Where based on water-level data for wells tapping the same elevation the surface is essentially a map of hydraulic head.

Quartzite: Metamorphic rock commonly formed by metamorphism of sandstone and composed of quartz.

Rhyolite: A volcanic rock rich in quartz and potassium feldspars that is the lava form of granite.

Schist: Metamorphic rock dominated by fibrous or platy minerals. Has schistose cleavage and is product of regional metamorphism.

Schistose: Said of a rock displaying schistosity.

Schistosity: The foliation in schist or other coarse-grained, crystalline rock due to the parallel, planar arrangement of mineral grains of the platy, prismatic, or ellipsoidal types, usually mica. It is considered by some to be a type of cleavage.

Silicic: In petrology, containing silica in dominant amount. Granite and rhyolite are typical silicic rocks. The synonymous terms “acid” and “acidic” are used almost as frequently as silicic.

Siliciclastic: A silica-rich sedimentary deposit.

Specific yield: The ratio of the volume of water that a given mass of saturated rock or soil will yield by gravity to the volume of that mass. This ratio is stated as a percentage.

Storage coefficient (also known as storativity): Specific storage, storativity, specific yield, and specific capacity are aquifer properties; they are measures of the ability of an aquifer to release groundwater from storage, due to a unit decline in hydraulic head. These properties are often determined in hydrogeology using an aquifer test.

Stratabound: Said of a mineral deposit confined to a single stratigraphic unit. The term can refer to a stratiform deposit, to variously oriented ore bodies contained within the unit, or to a deposit containing veinlets and alteration zones that may or may not be strictly conformable with bedding.

Stratigraphic: Pertaining to the composition, sequence, and correlation of stratified rocks

Stratigraphy: The science of rock strata. It is concerned not only with the original succession and age relations of rock strata but also with their form, distribution, lithologic composition, fossil content, geophysical and geochemical properties; indeed, with all characters and attributes of rocks as strata.

Subduction: Act of one tectonic unit’s descending under another. The process of one lithospheric plate descending beneath another.

Supercontinent: A hypothetical former large continent from which other continents are held to have broken off and drifted away.

Syncline: A configuration of folded, stratified rocks in which rocks dip downward from opposite directions to come together in a trough. Reverse of anticline. A fold in which the core contains the stratigraphically younger rocks; it is generally concave upward.

Synclinorium: A compound syncline; a closely folded belt, the broad general structure of which is synclinal. Plural – synclinoria.

Thrust: An overriding movement of one crustal unit over another, such as in thrust faulting.

Transmissivity: Rate of water movement through a unit width or thickness of aquifer. T is equal of hydraulic conductivity (K) times aquifer thickness. Transmissivity is essentially a measure of the aquifer’s ability to transmit water.

Transverse zone: Regional scale, east-west structural alignments that are generally perpendicular to the regional north-south alignment of mountain ranges and valleys. A zone of structures that typically cross-cuts a region and separates two areas of different type or amount of disruption or deformation.

Unconformity: Buried erosion surface separating two rock masses, older exposed to erosion for long interval of time before deposition of younger. If older rocks were deformed and not horizontal at time of subsequent deposition, surface of separation is angular unconformity. If older rocks remained essentially horizontal during erosion, surface separating them from younger rocks is called disconformity. Unconformity that develops between massive igneous or metamorphic rocks exposed to erosion and then covered by sedimentary rocks is called nonconformity.

Vug: Small unfilled cavity in rock, usually lined with crystalline layer of different composition from surrounding rock.

Water table: Surface of contact between the zone of saturation and the zone of aeration; that surface of a body of unconfined groundwater at which the pressure is equal to that of the atmosphere. Zeolite: A generic term for class of hydrated silicate minerals of aluminum and either sodium or calcium or both.

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Water Resources of the Basin and Range Carbonate-Rock Aquifer System, White Pine County, Nevada, and Adjacent Areas in Nevada and Utah: Draft Report

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