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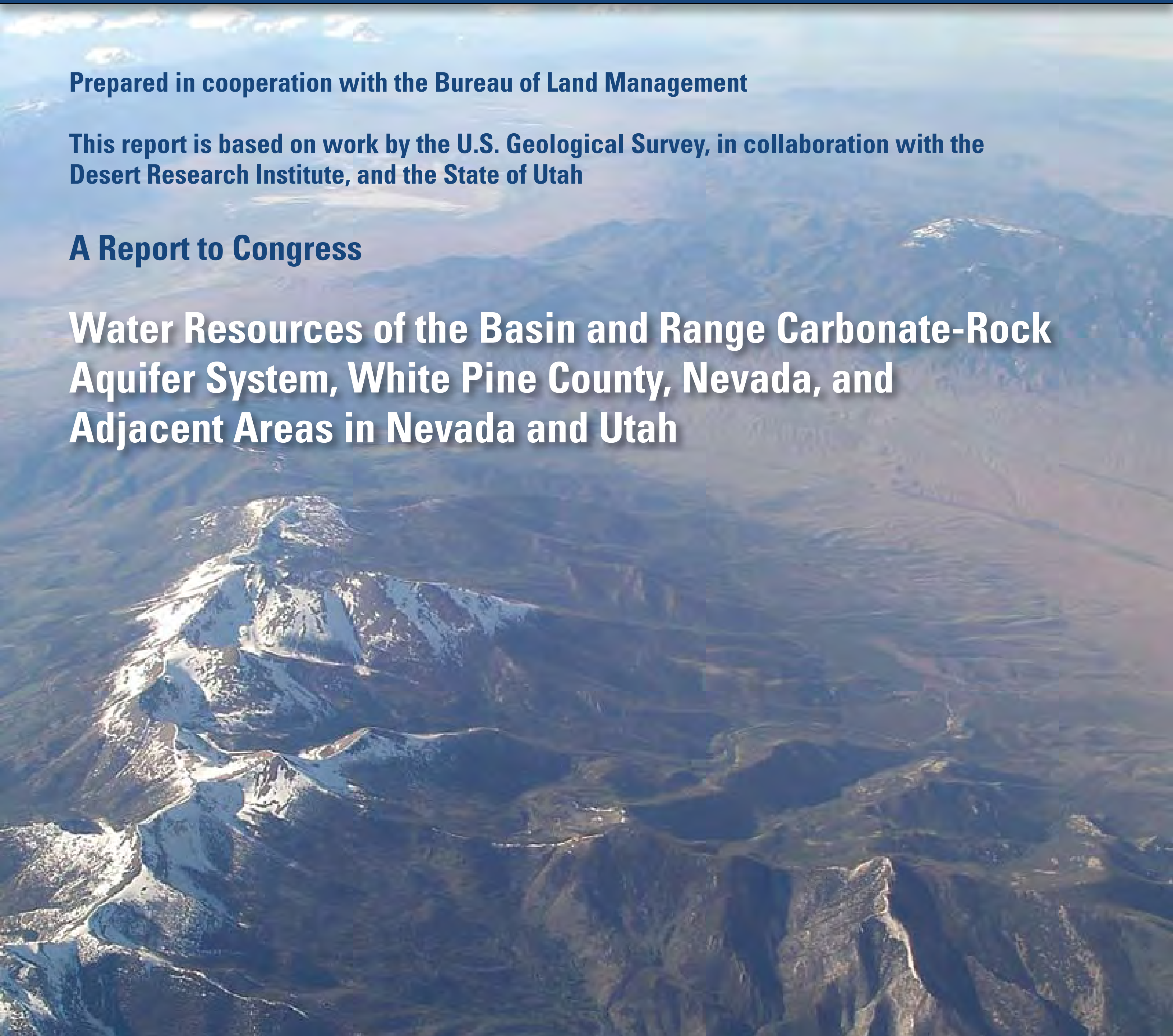
This report is based on work by the U.S. Geological Survey, in collaboration with the Desert Research Institute, and the State of Utah

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



Scientific Investigations Report 2007–5261

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

DRI Desert Research Institute

UTAH LIFE ELEVATED

Cover: View from above and to the northwest of the Snake Range, Nevada. Photograph taken by Donald S. Sweetkind, U.S. Geological Survey, June 19, 2005.

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

By Alan H. Welch, Daniel J. Bright, and Lari A. Knochenmus, Editors

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

U.S. Department of the Interior
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 301(e) 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 geochemical information are integrated to determine basin and regional ground-water budgets.

Results of the study are summarized in a USGS Scientific Investigations Report (SIR), prepared in collaboration with DRI and the State of Utah, and in cooperation with the Bureau of Land Management. The report was submitted to Congress in 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 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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Figures—Continued

Figures—Continued

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Conversion Factors, Datums, and Acronyms

Conversion Factors

Multiply	By	To obtain
acre	4,047	square meter (m^2)
acre	0.4047	hectare (ha)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
acre-feet per year (acre-ft/yr)	1,233	cubic meter per year (m^3/yr)
calorie	4.184	joule (J)
calories per second per square foot	45.045	watt per square meter (W/m^2)
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
foot (ft)	0.3048	meter (m)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per foot (ft/ft)	0.3048	meter per meter (m/m)
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
foot squared per day (ft^2/d)	0.0929	meter squared per day (m^2/d)
inch (in.)	25.4	millimeter (mm)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	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:

$$°F=(1.8×°C)+32.$$

Datums

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

Conversion Factors, Datums, and Acronyms—Continued

Acronyms

Acronym	Definition
AMT	audio-magnetotelluric
AZMET	Arizona Meteorological Network
BARCAS	Basin and Range Carbonate-rock aquifer system
BCM	Basin Characterization Model
CIMIS	California Irrigation Management Information System
CYSU	coarse-grained younger sedimentary rock unit
DRI	Desert Research Institute
DVRFS	Death Valley Regional Flow System
ET	evapotranspiration
FYSU	fine-grained younger sedimentary rock unit
HA	hydrographic area
HGU	hydrogeologic unit
IU	intrusive-rock unit
LCU	lower carbonate-rock unit
LSCU	lower siliciclastic-rock unit
MODIS	Moderate Resolution Imaging Spectroradiometer
MSAVI	modified soil-adjusted vegetation index
MSU	Mesozoic sedimentary rock unit
MT	magnetotelluric
MX	wells drilled as part of the U.S. Air Force missile-siting investigation
NDWR	Nevada Division of Water Resources
NV	Nevada
OSU	older sedimentary rock unit
PL	public law
PRISM	Parameter-Elevation Regressions on Independent Slopes Model
SDWIS	Safe Drinking Water Information System
SIR	Scientific Investigations Report
SPV	Spring Valley
SWReGAP	Southwest Regional Gap Analysis Project
TM	Thematic Mapper
UCU	upper carbonate-rock unit
USCU	upper siliciclastic-rock unit
USEPA	U.S. Environmental Protection Agency
USGS	U.S. Geological Survey
VFU	volcanic flow unit
VTU	volcanic tuff unit

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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

By Alan H. Welch, Daniel J. Bright, and Lari A. Knochenmus, Editors

Summary of Major Findings

Introduction

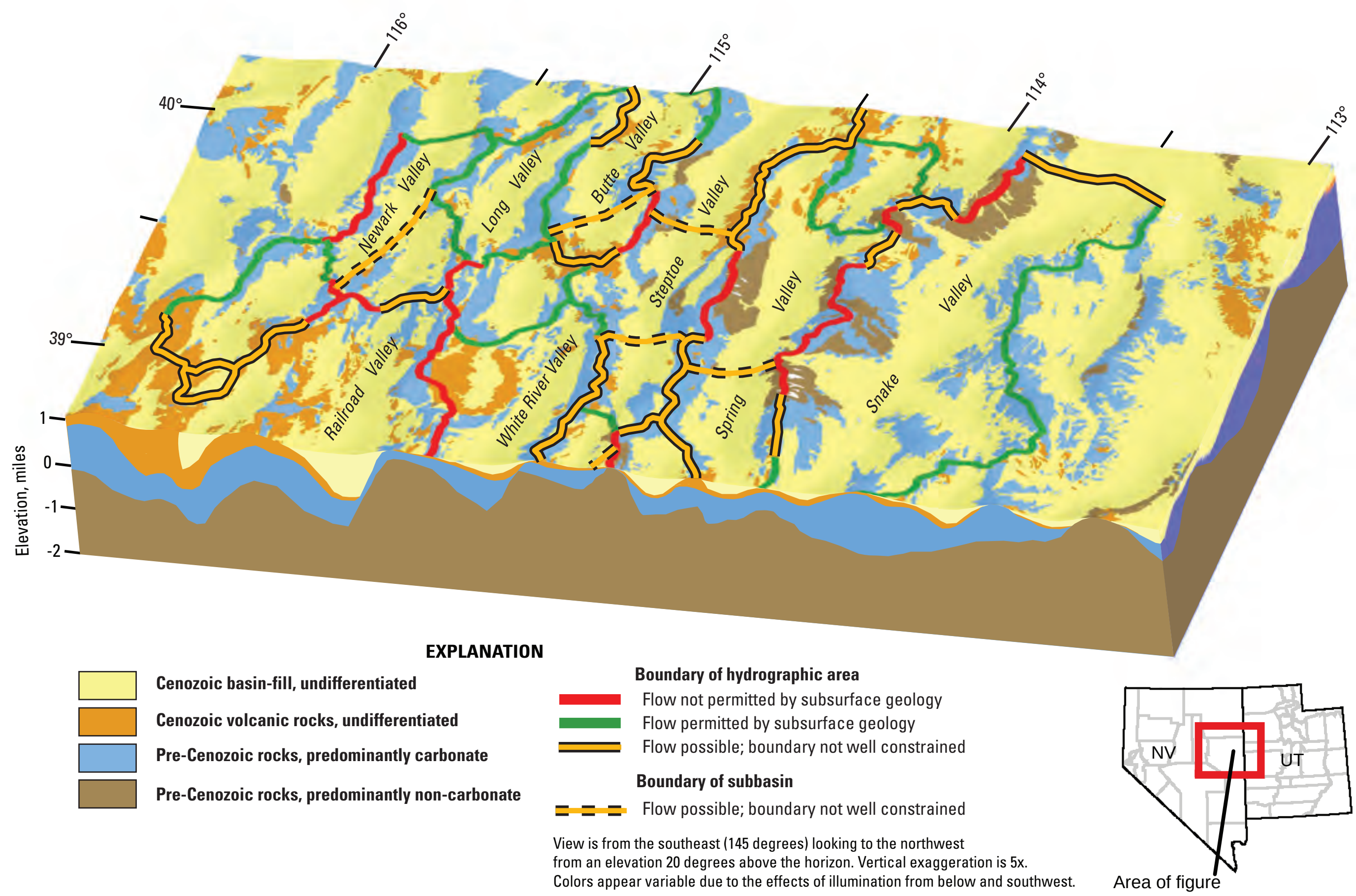
This report summarizes results of a water-resources study for White Pine County, Nevada, and adjacent areas in east-central Nevada and western Utah. The Basin and Range carbonate-rock aquifer system (BARCAS) study was initiated in December 2004 through Federal legislation (Section 301(e) of the Lincoln County Conservation, Recreation, and Development Act of 2004; PL108-424) 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 are the subdivisions used in this study for estimating recharge, discharge, and water budget. Hydrographic areas are 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. Typical thicknesses of basin fill range from 0.3 to 0.9 miles; maximum thicknesses of basin fill range from about 1 mile to more than 3 miles. The volcanic-rock aquifer is thickest beneath the western and southern parts of the study area, extending laterally beneath the basin-fill aquifer in 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 controls ground-water flow between hydrographic areas. Ground-water flow across some hydrographic area boundaries is negligible where carbonate or volcanic rocks are absent, or if the aggregate permeability of aquifers beneath a hydrographic area boundary is relatively low.



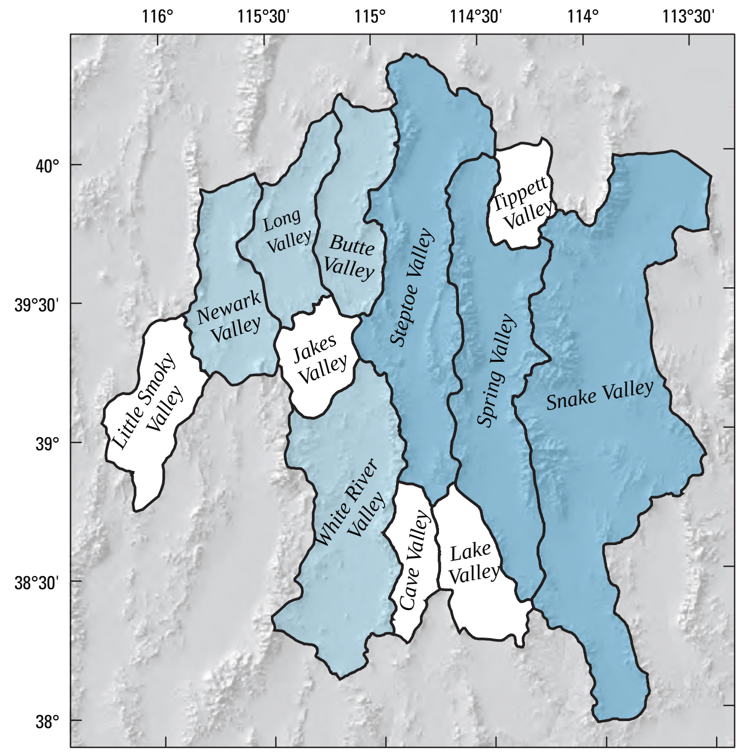
Perspective view of the primary aquifer systems.

Aquifer Water Quality

Based on a subset of chemical constituents having National primary and secondary drinking-water standards, the inorganic chemical quality of ground water generally is acceptable for human consumption. For chemical constituents with available analyses from more than 25 sampling sites, only arsenic (2 sites) and fluoride (4 sites) exceeded their primary standards at more than 1 site. Secondary drinking-water standards were exceeded more often than the primary standards. A small number of analyses of anthropogenic organic compounds in ground water are available, and from these analyses only low concentrations of pesticides or their metabolites are reported, and no volatile organic compounds were detected.

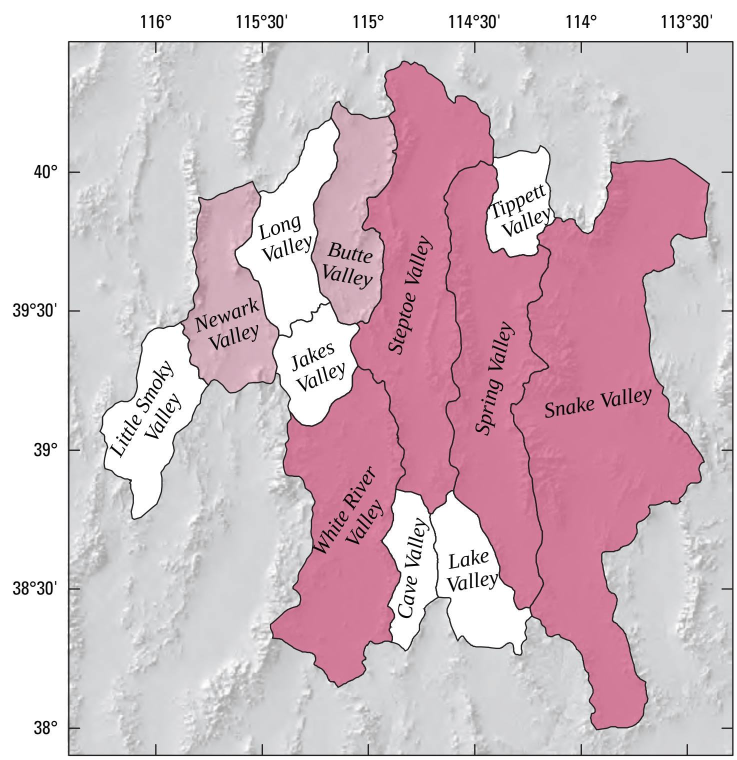
Basin Recharge and Discharge

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 154,000 acre-ft) and Snake Valley (about 111,000 acre-ft). Estimated annual recharge for Steptoe Valley is about 20,000 acre-ft higher than any previous estimate for this valley. The highest annual discharge occurs in Snake Valley (about 132,000 acre-ft) and Steptoe Valley (about 101,000 acre-ft). Estimated annual discharge for Snake Valley is significantly higher (about 45,000 acre-ft) than any previous estimate and the estimated annual discharge for Steptoe Valley is within the range of previous estimates.



Shaded relief base from 1:250,000-scale National Elevation Data; sun illumination from northwest at 30 degrees above horizon.
1:1,000,000 scale watershed boundaries from USGS digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83

Average annual recharge to the ground-water system.

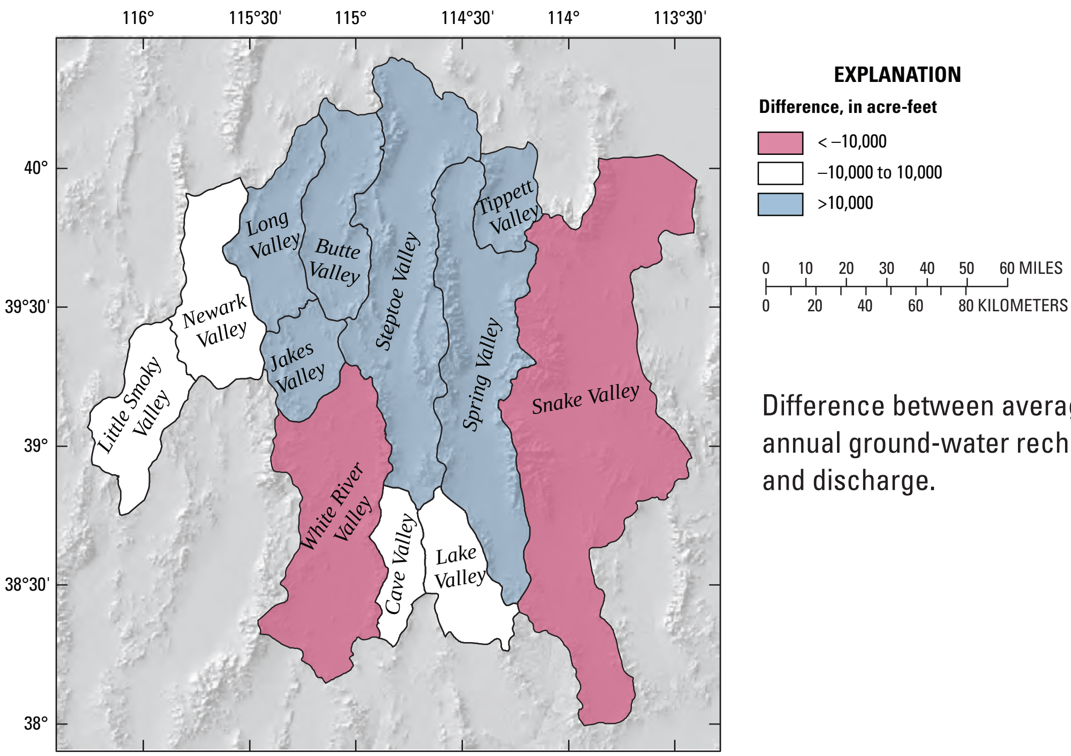


Shaded relief base from 1:250,000-scale National Elevation Data; sun illumination from northwest at 30 degrees above horizon.
1:1,000,000 scale watershed boundaries from USGS digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83

Average annual discharge from the ground-water system.

Interbasin Ground-Water Flow

Differences in basin recharge and discharge provide a surplus or deficit of water that under equilibrium conditions is balanced by ground-water flow entering or exiting a valley as interbasin ground-water flow. Recharge exceeds pre-development discharge by 10,000 acre-ft or more on an average annual basis in almost one-half of the hydrographic areas (5 of 12). Recharge in Steptoe Valley annually exceeds pre-development discharge by about 53,000 acre-ft. The surplus of water in Steptoe Valley is the source of interbasin ground-water flow to multiple valleys. In contrast to Steptoe Valley, pre-development discharge annually exceeds the relatively low annual recharge in White River Valley by about 41,000 acre-ft, indicating that ground water lost to the atmosphere on the valley floor must be supported, in part, by subsurface inflow from adjacent valleys such as Steptoe Valley to the northeast, Jakes Valley to the north, and Cave Valley to the east. Estimates of the magnitude of interbasin flow through some hydrographic area boundaries differ from previous estimates. The largest differences are for the outflow estimated for southern Steptoe Valley, where previous investigations proposed zero outflow, and for southern Spring Valley. The estimated interbasin ground-water flow from southern Spring Valley to Snake Valley is about twice as high as any previous estimate.

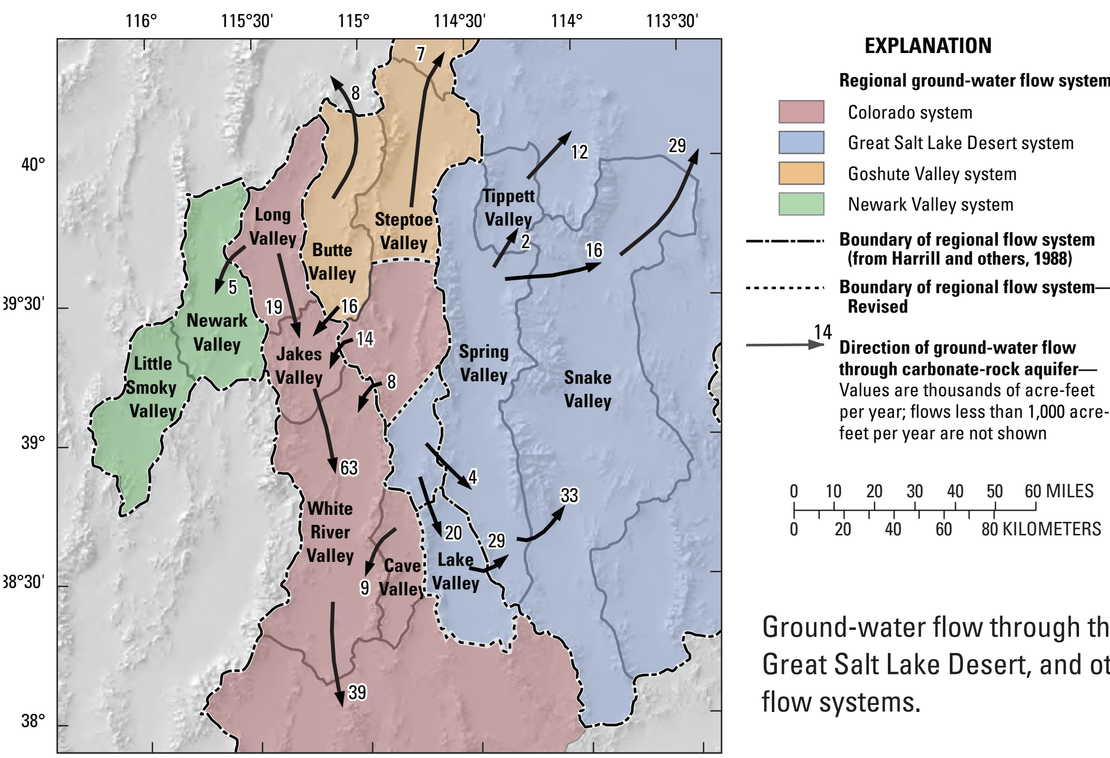


Shaded relief base from 1:250,000-scale National Elevation Data; sun illumination from northwest at 30 degrees above horizon.
1:1,000,000 scale watershed boundaries from USGS digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83

Difference between average annual ground-water recharge and discharge.

Regional Ground-Water Flow

Carbonate rocks comprise much of the Egan, Schell Creek, and Snake Ranges, and the relatively high precipitation and recharge in these mountain ranges create a large mound that is a primary source of recharge to the ground-water flow systems in the basin-fill, volcanic-rock, and carbonate-rock aquifers of the study area. 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. 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 and carbonate-rock aquifers are found in the study area. Ground-water outflow from northern Steptoe Valley is toward the northeast and exits the study area. Ground-water outflow from central Steptoe Valley is to Jakes and northern White River Valleys; and outflow from southern Steptoe Valley is to Lake and southern Spring Valleys. The latter two flow paths from central and southern Steptoe Valley have not been identified in previous investigations. Southeasterly flow from southern Steptoe Valley suggests that southern Steptoe and Lake Valleys may be part of the Great Salt Lake Desert regional flow system. Southwesterly flow from central Steptoe Valley suggests that central Steptoe Valley may be part of the Colorado regional flow system. 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. Most ground-water flow likely exits the study area through Snake (29,000 acre-ft/yr), Butte (8,000 acre-ft/yr), Tippett (12,000 acre-ft/yr), and White River Valleys (39,000 acre-ft/yr).



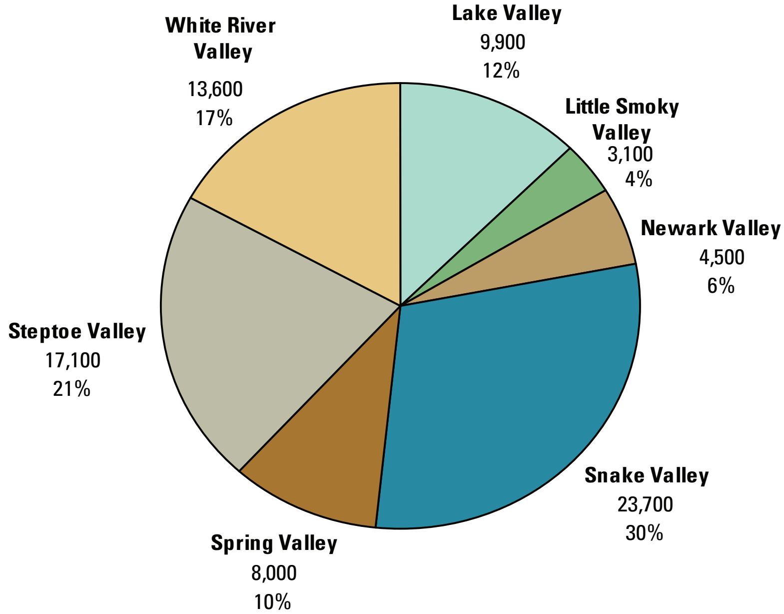
Shaded relief base from 1:250,000-scale National Elevation Data; sun illumination from northwest at 30 degrees above horizon.
1:1,000,000 scale watershed boundaries from USGS digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83

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

Regional Water Budgets

Average annual recharge equals 530,000 acre-ft, and average annual ground-water discharge equals 440,000 acre-ft for the entire study area 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.

The net amount of regional ground water removed from the study area was estimated to evaluate the significance of the ground-water withdrawn to ground-water discharged under pre-development conditions. Net regional ground-water use for the study area is about 80,000 acre-ft and includes 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. The net ground-water use of 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 the long-term use of ground water at a rate of 80,000 acre-ft could 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, decrease other discharge components such as interbasin flow, spring discharge, or discharge by vegetation, or increase subsurface recharge from adjacent basins. However, actual decreases 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, decreases in outflow would be less likely in Butte or Tippett Valleys where net ground-water use was zero in 2005. Decreases in outflow would be more likely in subbasins or hydrographic areas where net ground-water use is nearly equal to or greater than the estimated outflow, such as in Snake Valley where net ground-water use 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 decreases 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 pumping has not significantly altered local evapotranspiration rates, the distribution of native vegetation, or regional springflow in the study area.



Percentage and volume, in acre-feet, of net regional ground-water use by hydrographic area.

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. The final report was transmitted to Congress in December 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. The area within the county is characterized by 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. Precipitation on the mountain ranges is the principal source of recharge to four regional ground-water flow systems in the study area (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.

The Colorado River system is currently the principal source of water supply for southern Nevada. The prospect of obtaining additional allotments of water from the Colorado River system are confounded by the legal and socio-political issues derived from the competition for those scarce resources. Proposed ground-water development is based, in part, on concerns that water from the Colorado River Basin is not a resource capable of supporting future growth in southern Nevada and elsewhere in the southwestern United States due to a persistent drought in the Basin. Water purveyors in southern Nevada have proposed to develop in-state ground-water resources in rural basins north of Clark County, including basins in White Pine County, Nevada. Municipal and regulatory agencies have expressed concerns about potential impacts on water quantity and quality, existing water rights, sensitive wildlife habitats, and other beneficial uses from developing these ground-water resources. As a first step in assessing the potential impacts of any proposed large-scale 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 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 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 half 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.

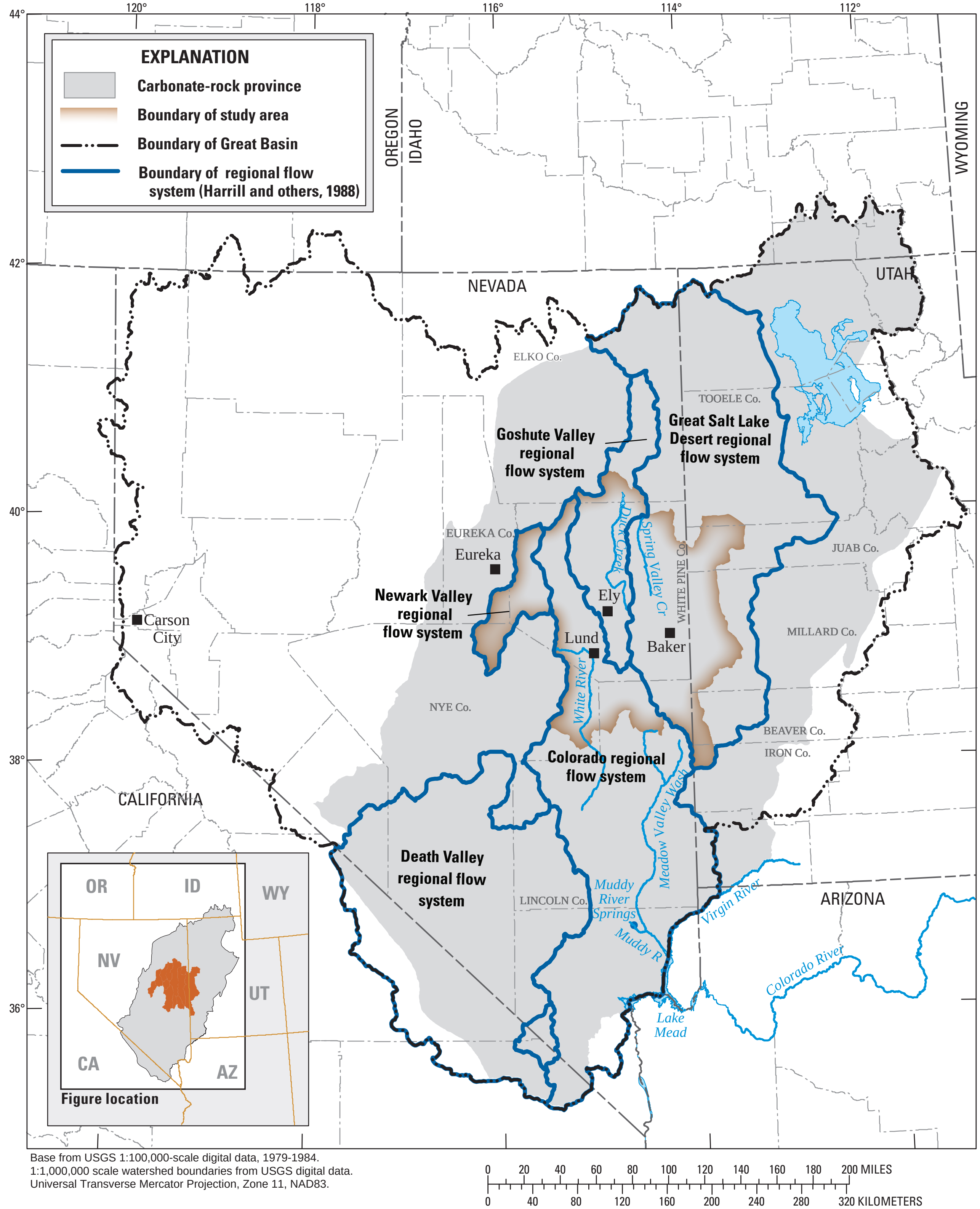


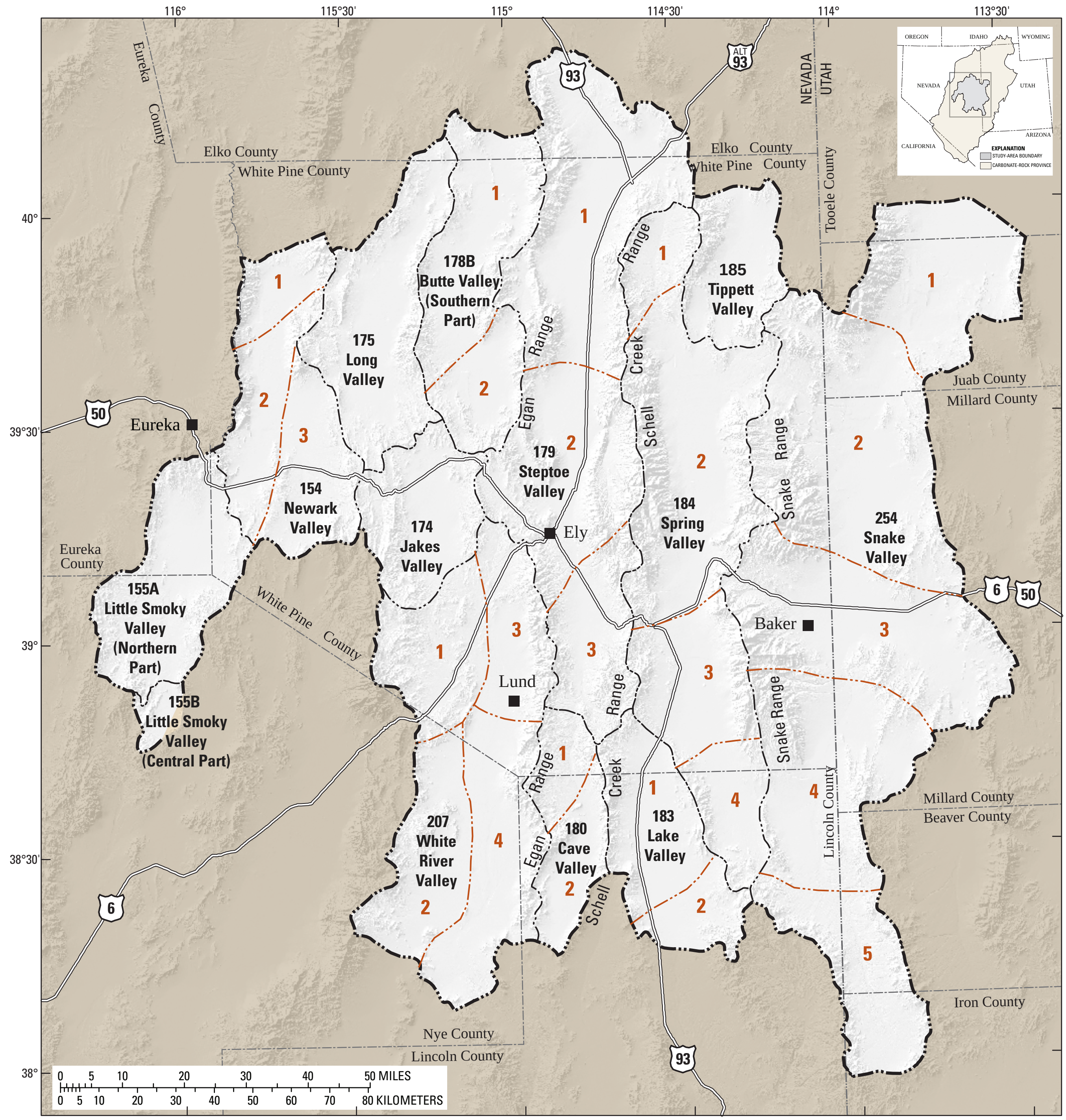
Figure 1. Carbonate-rock province, Basin and Range carbonate-rock aquifer system study area, and selected regional ground-water flow systems, Nevada and Utah.

The study area encompasses 13 hydrographic areas (HAs)[1] (fig. 2). For most figures and tables in this report, water-budget components were estimated independently for the northern and central parts of Little Smoky Valley, and then were combined and reported as one value. 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 major recharge and terminal discharge areas (Harrill and Prudic, 1998). Although most boundaries between HAs coincide with topographic basin divides, some are arbitrary divisions that have no topographic basis. In this report, HAs also are referred to as basins, and ground-water flow within a basin 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 are the subdivision used to estimate recharge and discharge in this study. HAs are the subdivision used to report summed and tabulated subbasin water budgets. 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 and others' 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.

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. As perceived by Harrill and others (1988), the Newark Valley and Goshute Valley flow systems are relatively small, internally drained flow systems, whereas the Great Salt Lake Desert and Colorado 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; and 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 through the 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 in thickness. Gravel and sand deposits yield water readily to wells and form the aquifers most commonly developed for agricultural, domestic, and municipal water supply.

[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 subbasin and number

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 hydrogeologic framework defines the physical geometry and rock types in the subsurface through which water flows. 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 geologic units were grouped into hydrogeologic units (HGUs)—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.

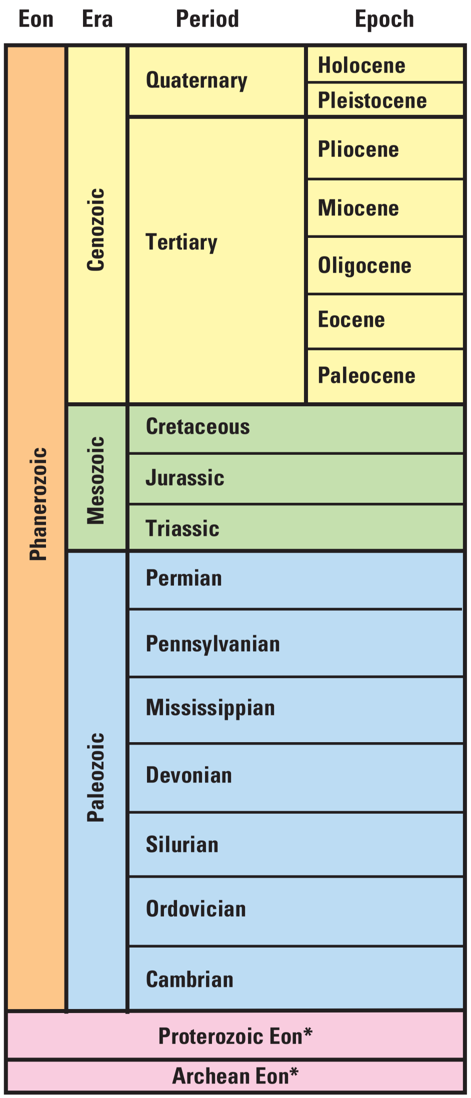
Surface geophysical techniques were applied to take advantage of characteristic density, magnetic, electrical, and acoustic properties of different rocks in a way that provides additional insight into the subsurface geology. Detailed gravity, magnetic, electromagnetic, and seismic geophysical data (fig. 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).

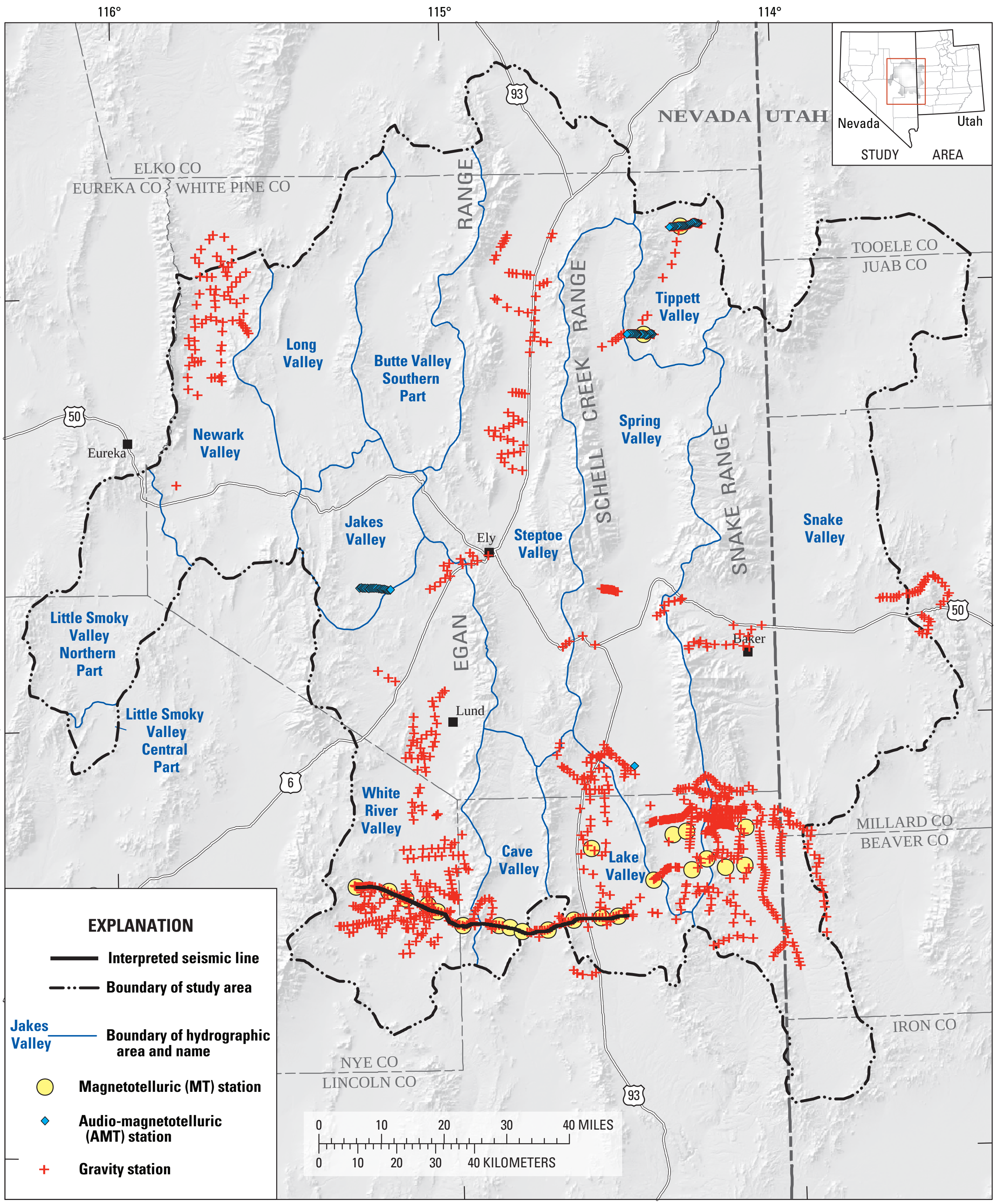
During the first phase of geologic evolution, from late Precambrian until middle Devonian time, the rocks in east-central Nevada were deposited in shallow to deep marine water in a stable continental shelf environment similar to that of modern-day Atlantic and Gulf Coast margins of the United States (Blakely, 1997; available at http://jan.ucc.nau.edu/~rcb7/paleogeogwus.html). The stable shelf environment produced thick and laterally 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).



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

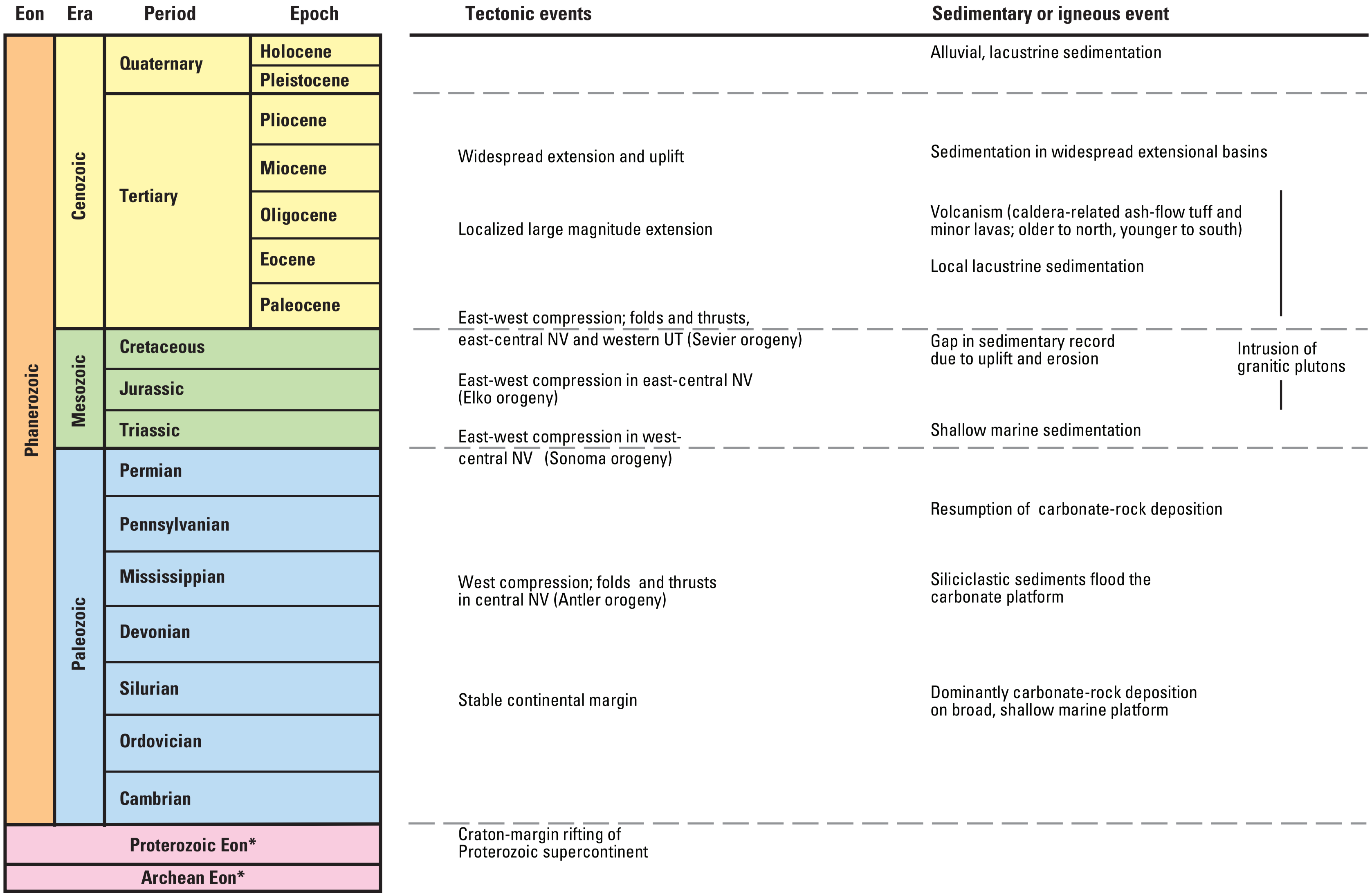
Hydrogeologic unit	Description of hydrogeologic unit	Examples of Nevada geologic formation names	Examples of Utah geologic formation names
FYSU	Fine-grained younger sedimentary rock unit; Holocene to Pliocene fine-grained playa and lake deposits of fine sand, silt, and clay.	Unconsolidated basin fill, 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 Tripple 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	Intrusive-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.



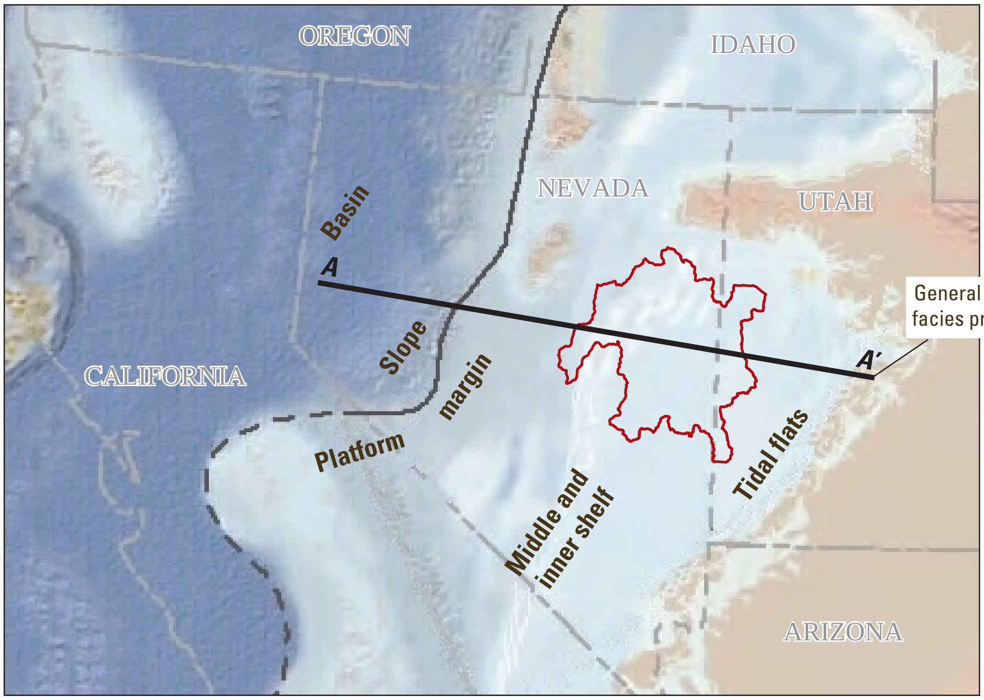
Base from U.S. Geological Survey 1:100,000-scale digital data, 1979–84.
1:1,000,000 scale watershed boundaries from U.S. Geological Survey digital data.
Universal Transverse Mercator Projection, Zone 11, NAD83.

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



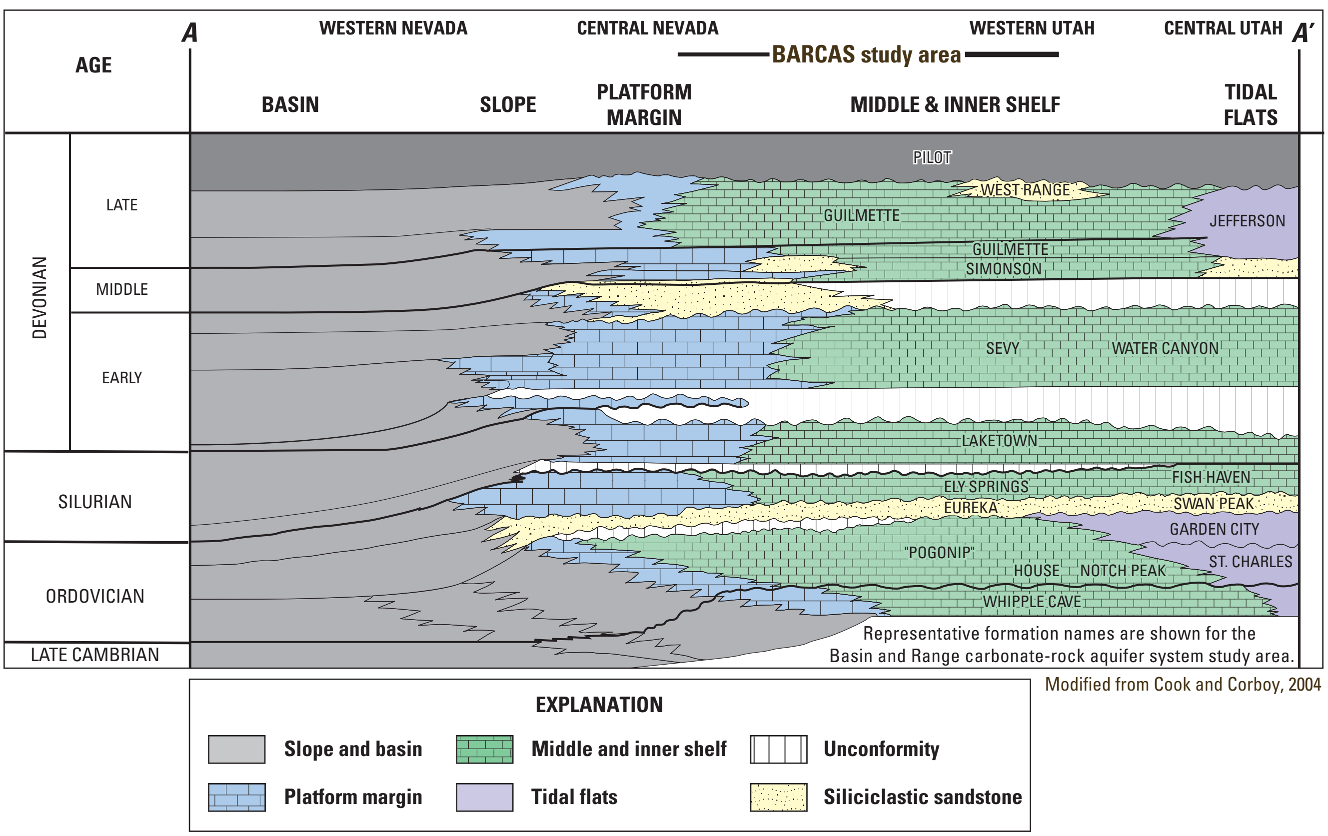
* The Archean and Proterozoic Eons are major subdivisions of Precambrian time.

Figure 5. Major geologic events in eastern Nevada.



Modified from Blakely, 1997

A. Schematic representation of Silurian paleogeography



Modified from Cook and Corboy, 2004

B. Depositional facies profile

Figure 6. Depositional facies and paleogeography, 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 primarily 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 carbonate sedimentation and resulted in the deposition of a thick sequence of siliciclastic rocks (Poole and Sandberg, 1977). Carbonate-shelf sedimentation resumed in Pennsylvanian and Permian time, creating a thick, widespread carbonate sequence 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 geologic evolution in 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 extend over parts of 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 fine-grained lake sediments (Reheis, 1999), and Quaternary 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 are perpendicular to the regional north-south alignment of mountain ranges and valleys. Prominent structural features in the study area, including compressional thrust belts, large-magnitude extensional normal and detachment faults, and transverse zones, are shown on plate 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 either 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 resulting in ground-water 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). Based on seismic reflection 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).

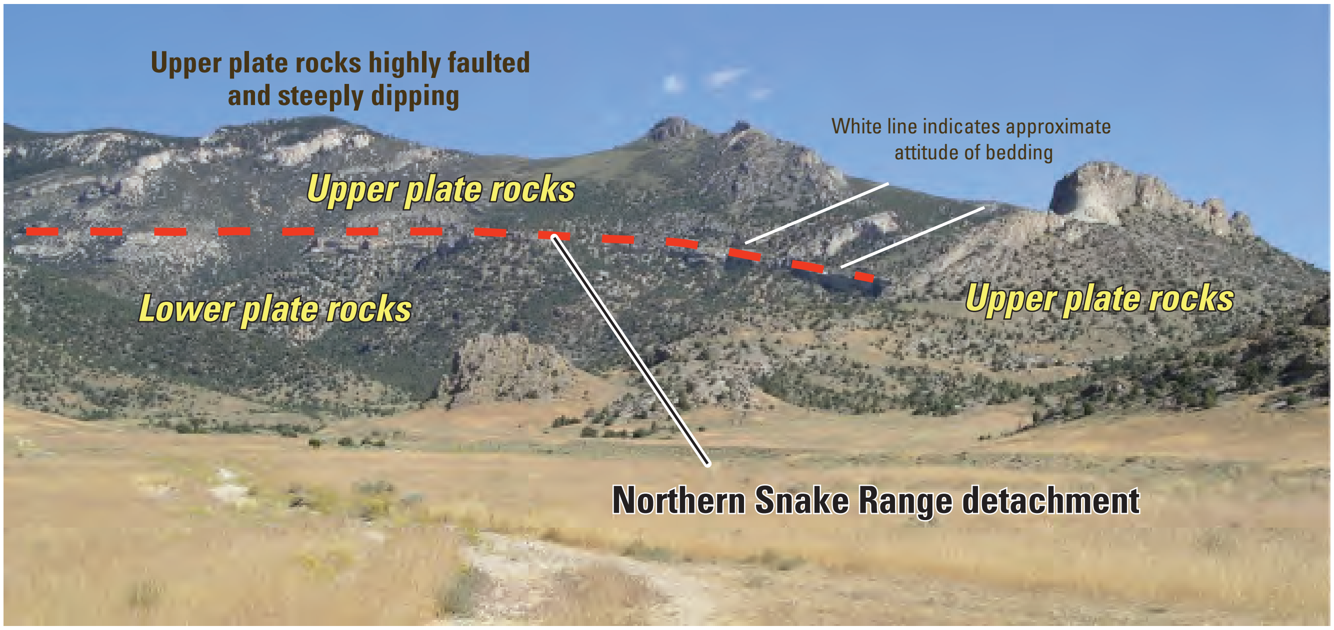
A second style of Tertiary extension is characterized by steeply dipping, range-bounding normal faults that produced elongate mountain ranges and have controlled the subsidence of intervening, down-faulted valleys (Zoback and others, 1981; Stewart, 1998). The range-bounding faults strike northeast and 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. The general relation between extensional range-bounding faults and resulting asymmetric or symmetric grabens is annotated on section *C-C*' shown on plate 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.

Regional gravity data were used to assess 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). Typical thicknesses of the basin fill range from 0.3 to 0.9 mi; maximum thicknesses of basin fill range from about 1 mi to more than 3 mi (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 (fig. 8); geologically, they are referred to as accommodation zones 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. 2). 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.

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.

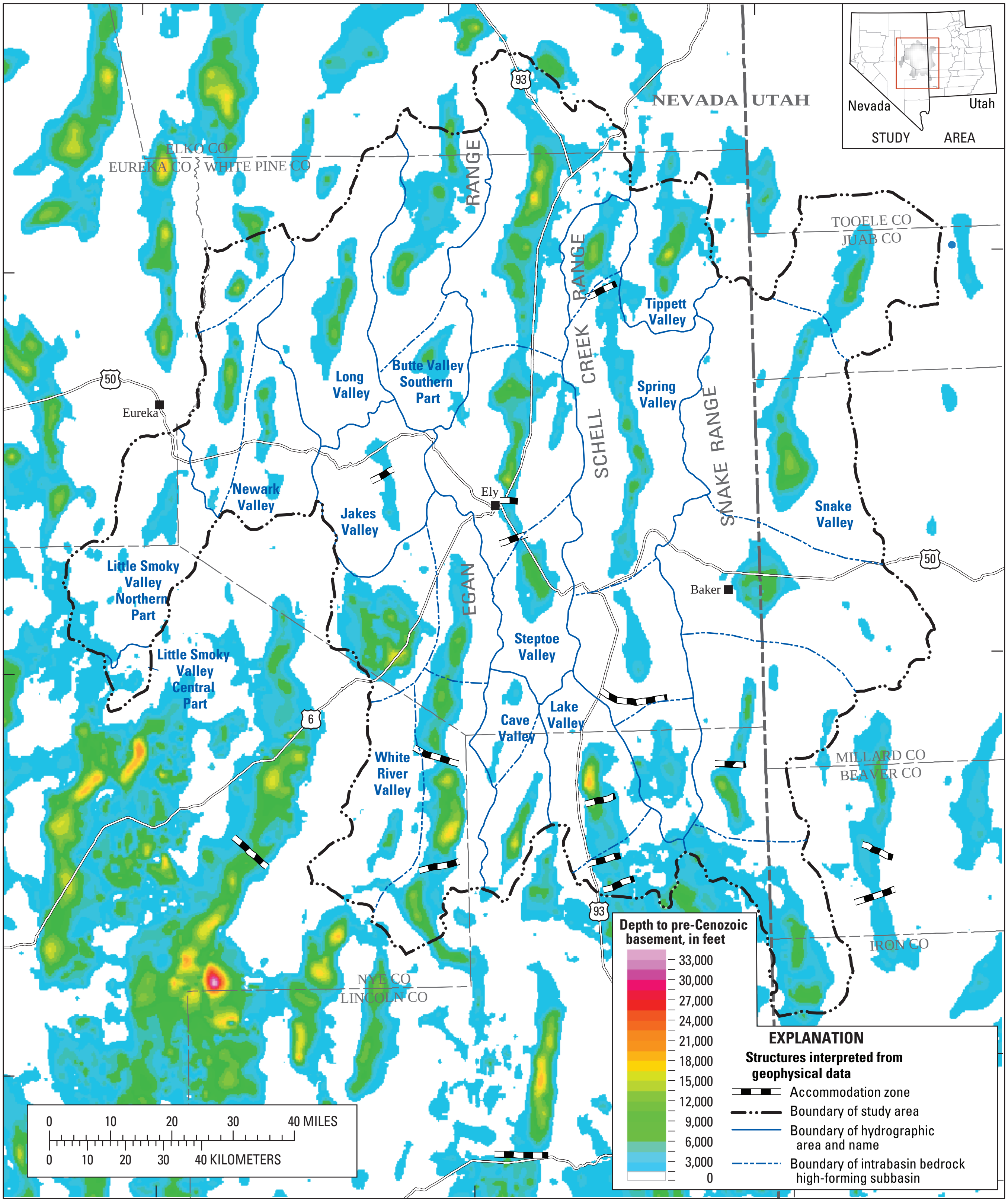


A. Low-angle detachment fault, eastern flank of northern Snake Range, Nevada.

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.



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.

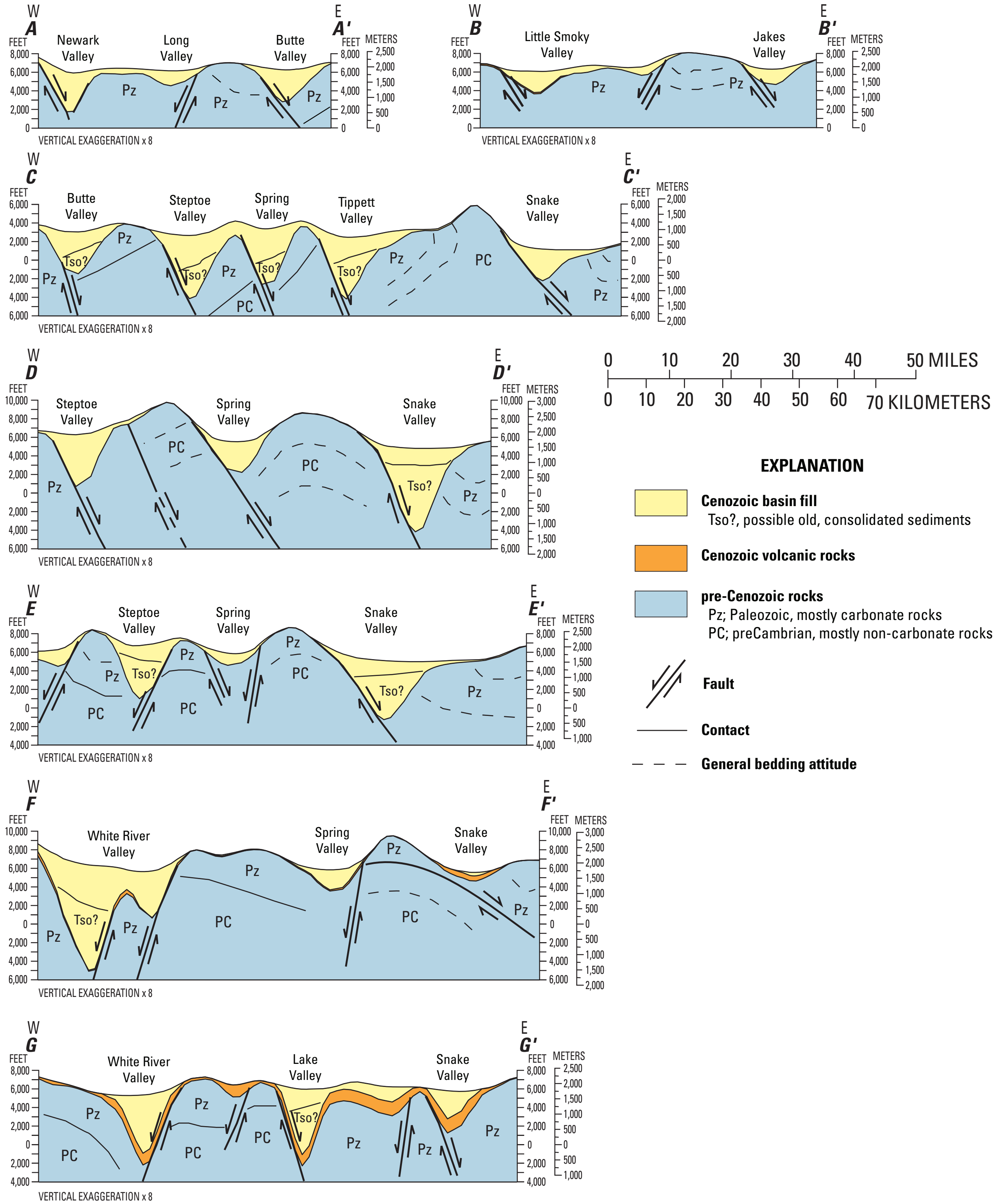


Figure 9. Modeled depth to pre-Cenozoic rocks and location of sections, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

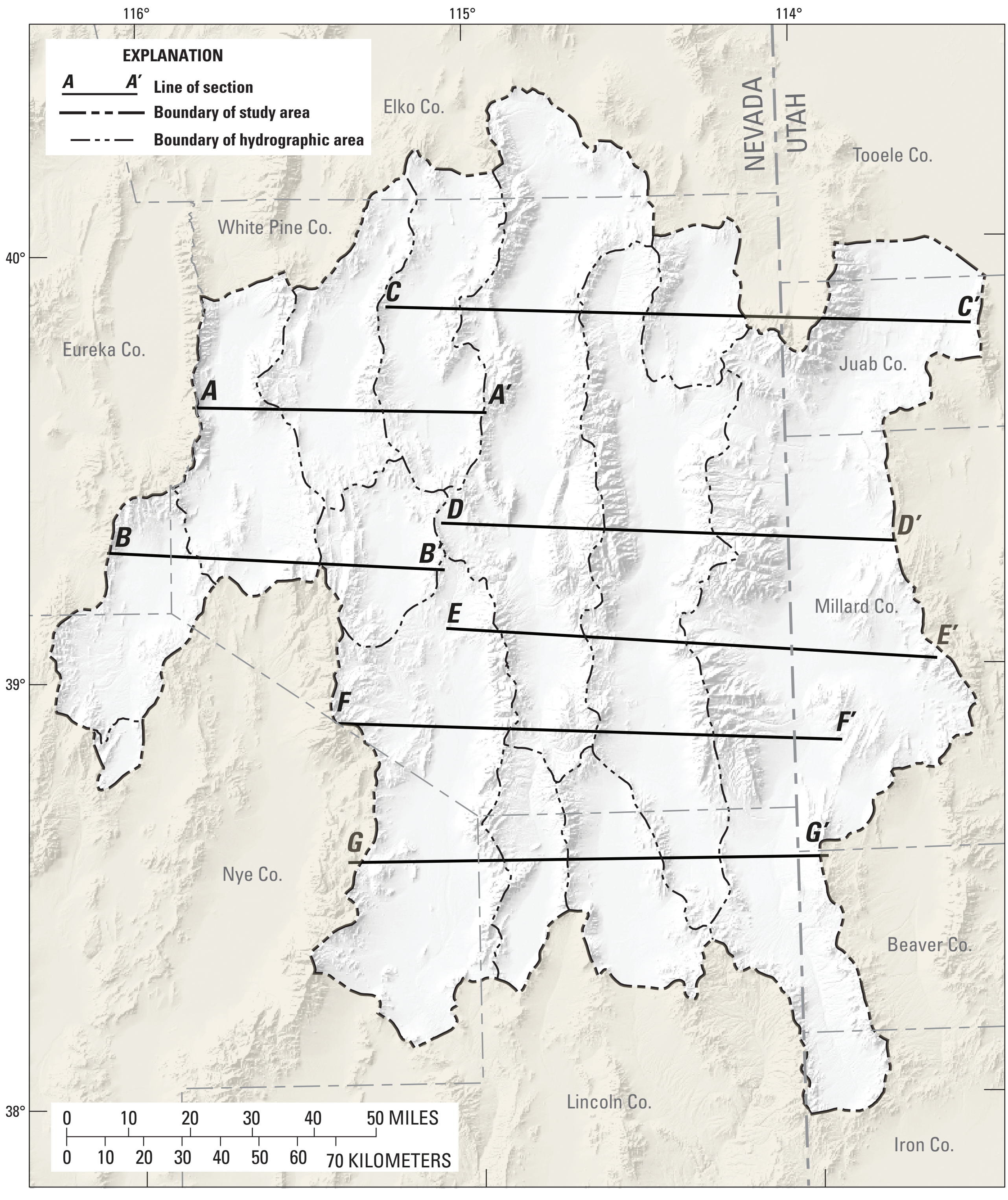


Figure 9. Continued.

Hydrostratigraphy

Hydrogeologic units (HGUs) are defined as having 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 1; 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 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.

Table 1. 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	Description 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-rock Unit	Intrusive rocks such as granite and granodiorite, not divided by age

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 influxes 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).

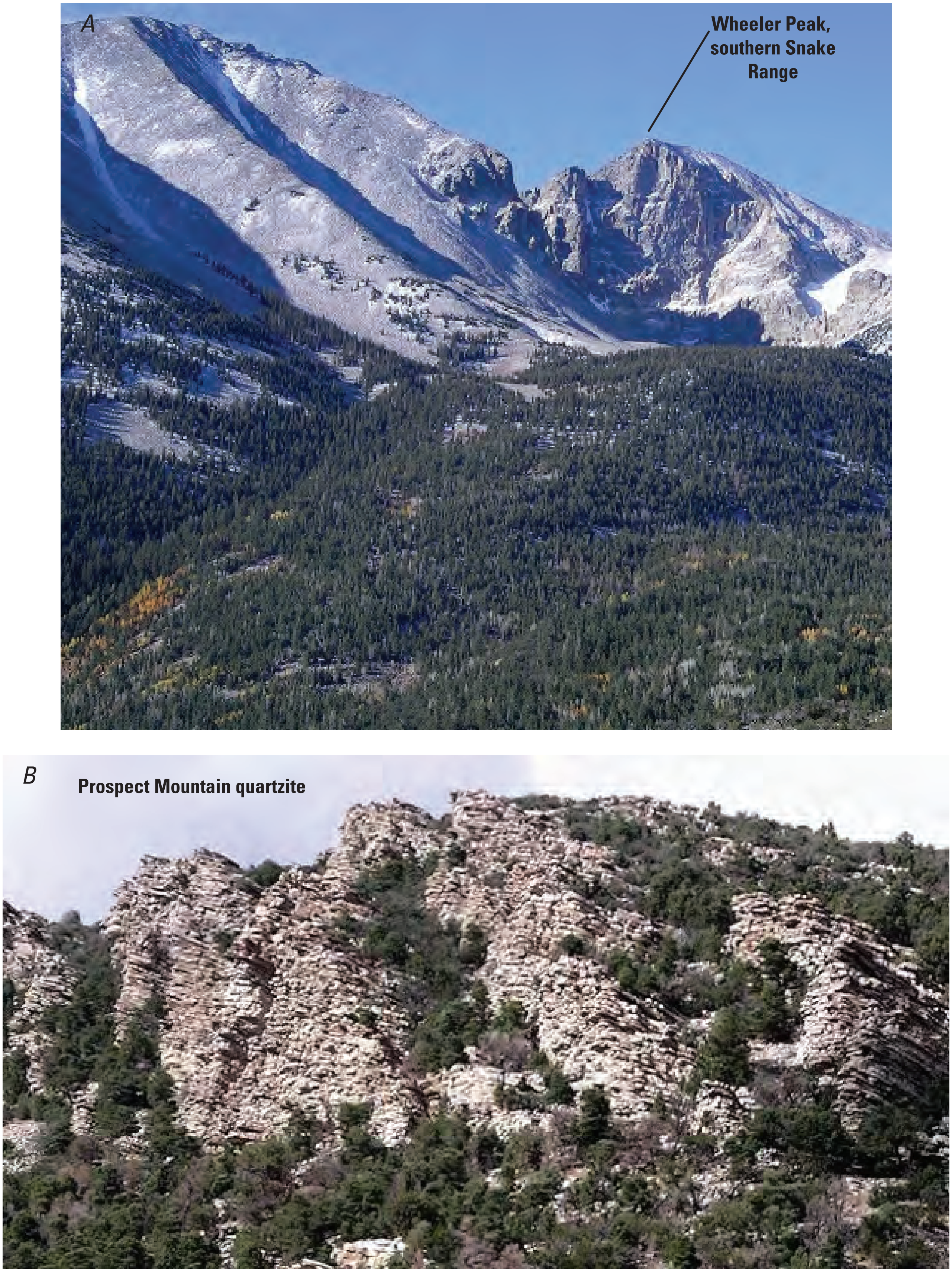


Figure 10. Lower Cambrian siliciclastic rocks, southern Snake Range, Nevada. Photographs taken by Donald S. Sweetkind, U.S. Geological Survey, (*A*) October 4, 2005; (*B*) September 10, 2004.

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) consists of thick, widespread Pennsylvanian and Permian rocks that overlie the Mississippian rocks of the USCU. 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 2. 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 2. 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 2. 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 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 penetrated 3,600 ft of upper Cenozoic sediments, 1,230 ft of Oligocene volcanic rocks, and 870 ft of lower Tertiary (?) sediments (Hess and others, 2004).
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 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 rocks, Oligocene 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 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

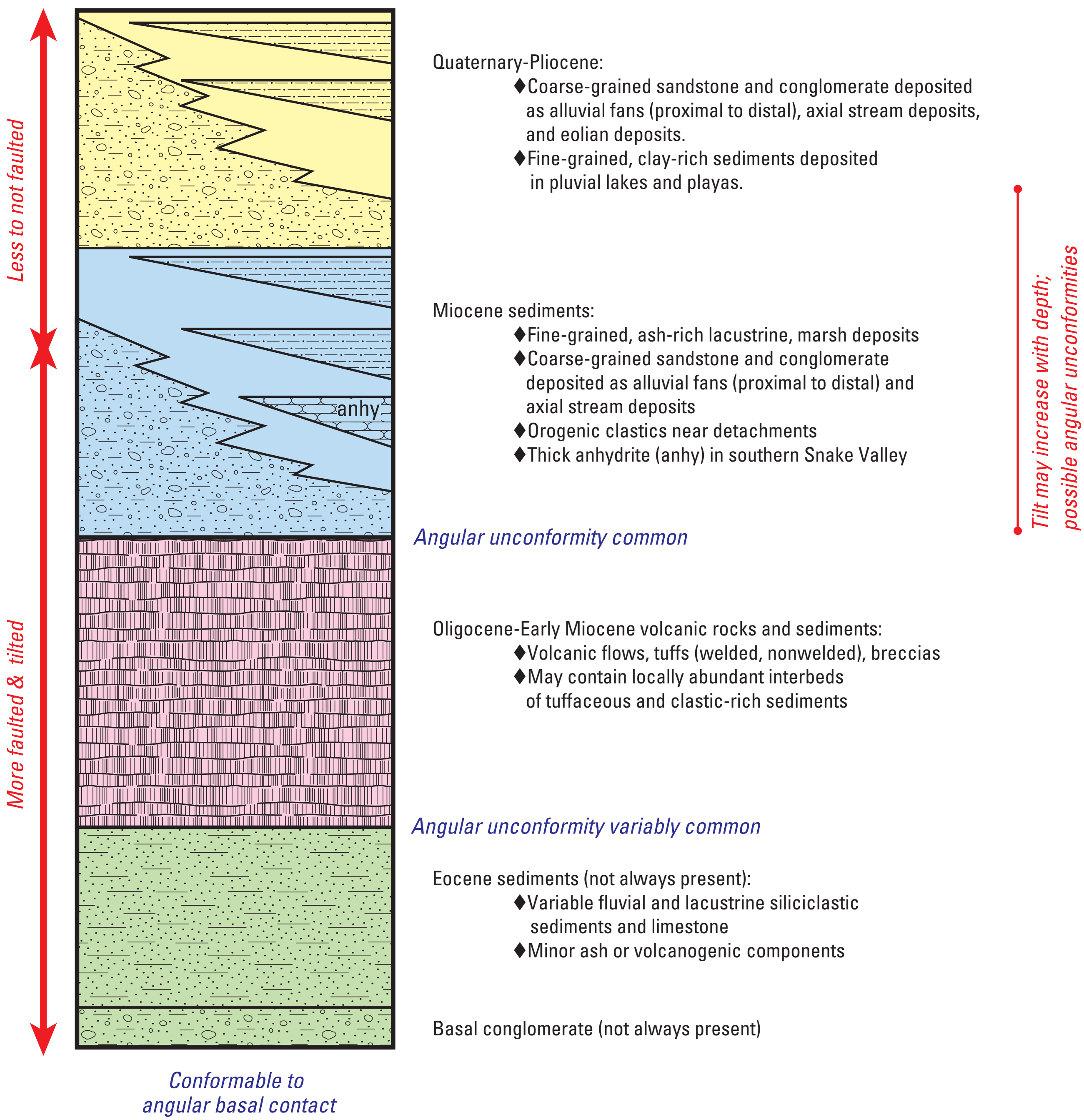
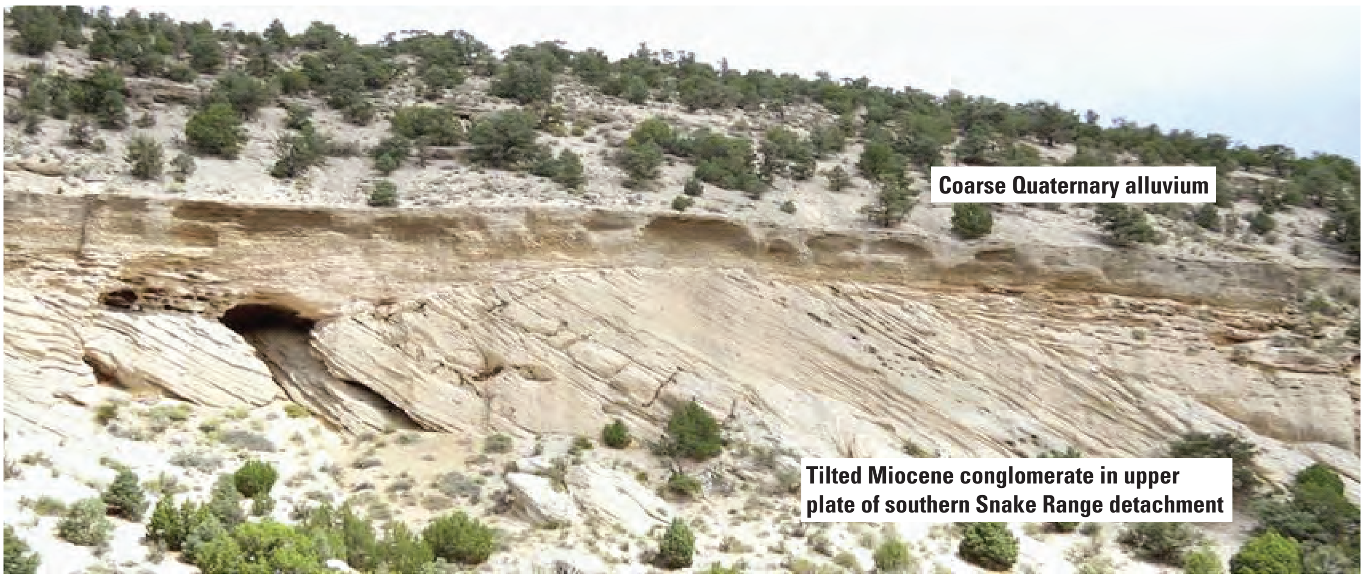


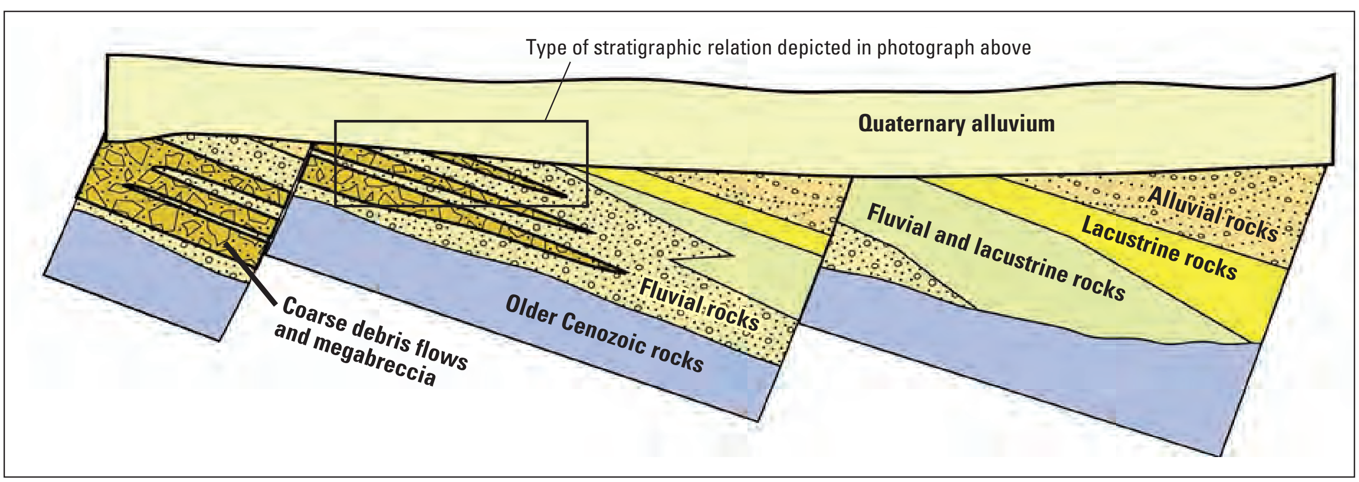
Figure 12. Generalized Cenozoic basin stratigraphy.

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 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 groundwater 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 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 increasingly indurated with 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 1; 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).

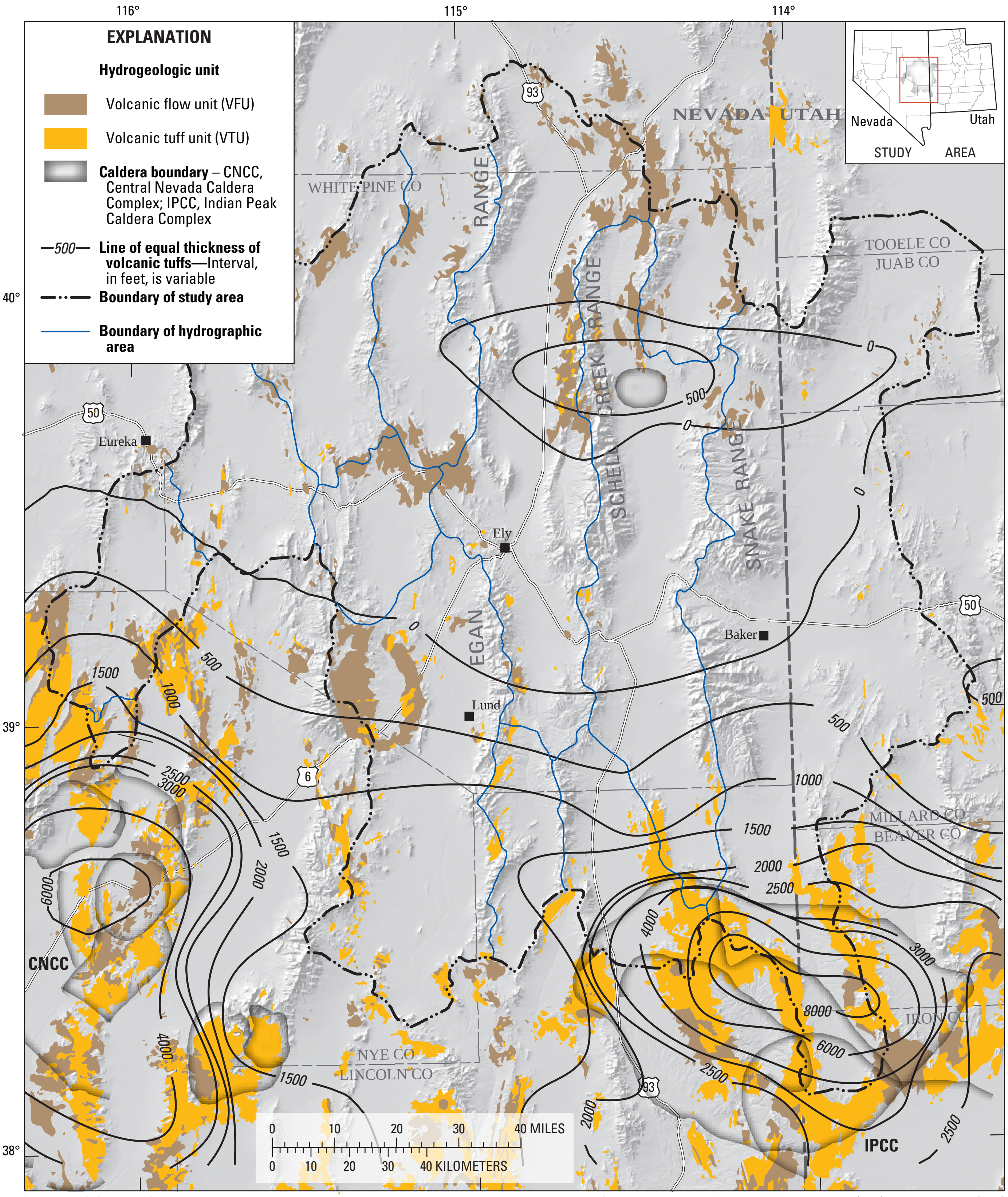
Volcanic rocks in the study area were divided into two principal HGUs (fig. 3), the volcanic tuff unit (VTU) and the volcanic flow unit (VFU). The use of these two HGUs follows the subdivision of volcanic rocks typically used on the State geologic maps. Rocks of the VTU include welded and nonwelded tuffaceous 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.

Distribution of Hydrogeologic Units Forming Aquifers and Lower Permeability Units

The hydrogeologic units in the BARCAS study area form three distinct aquifer systems composed of alternating more permeable and less permeable units. The three general types of aquifer materials are: basin-fill alluvium (CYSU), some volcanic rocks (VTU), and carbonate bedrock (LCU and UCU). Each of these units may include one or more water-bearing zones but are stratigraphically and structurally heterogeneous resulting in a highly variable ability to store and transmit water. The intervening lower permeability units, FYSU, OSU, VFU, USCU, and LSCU, separate the three aquifer systems.

The basin-fill aquifer occurs in each hydrographic area and subbasin, extending across most intrabasin divides and some hydrographic area boundaries. The lateral extent of the HGUs that form this aquifer vary, but in most basins, the coarser grained CYSU deposits occur near the mountain front and along drainages, the finer grained FYSU occur along valley axes. The consolidated OSU deposits typically underlie these younger basin-fill deposits and, in the southern part of the study area, contain significant quantities of volcanic ash and tuff. The volcanic aquifer primarily occurs in the western and southern parts of the study area, extending laterally beneath the basin-fill aquifer and multiple hydrographic areas.



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.

Of the two HGUs that form this aquifer, the lateral extent and thickness of the VTU typically are greater than that of the VFU. The carbonate aquifer is the most laterally extensive aquifer in the study area, underlying the basin-fill or volcanic aquifers in most hydrographic areas. The upper part of this aquifer is composed of UCU rocks and the lower part is composed of LCU rocks; these HGUs are separated by rocks of the USCU. Rocks of the LSCU underlie the carbonate aquifer. MSU and IU rocks are widely scattered throughout the study area, and generally are of lower permeability and limited aerial extent. However, depending on the depth of IU plutons, these rocks may intrude into overlying carbonate, volcanic, or basin-fill aquifers and influence the direction or magnitude of ground-water flow.

Relative differences in hydraulic properties were used to delineate aquifers from confining or semi-confining HGUs in the BARCAS study area. These evaluations primarily were based on relative differences in permeability determined from HGU material properties, or on estimates of hydraulic conductivity, a quantitatively derived parameter that serves as a measure of permeability (Todd, 1980). For the BARCAS study, differences in hydraulic conductivity also were used, in part, to evaluate the potential for ground-water flow across hydrographic area boundaries and intrabasin divides. Differences in hydraulic properties along these boundaries and divides typically are the result of structural disruption that may cause, for example, the juxtaposition of aquifers and lower permeability units, or uplifted bedrock areas where the saturated thickness of overlying aquifers is thinned. Relative differences in hydraulic conductivity of HGUs and the distribution of these HGUs along boundaries and divides are, therefore, important controls on intrabasin and interbasin ground-water flow.

Hydraulic Conductivity of Hydrogeologic Units

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 1).

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 3. The arithmetic mean is the average value within the sampled dataset. 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 3). 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 3. Hydraulic conductivity values for hydrogeologic units of the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

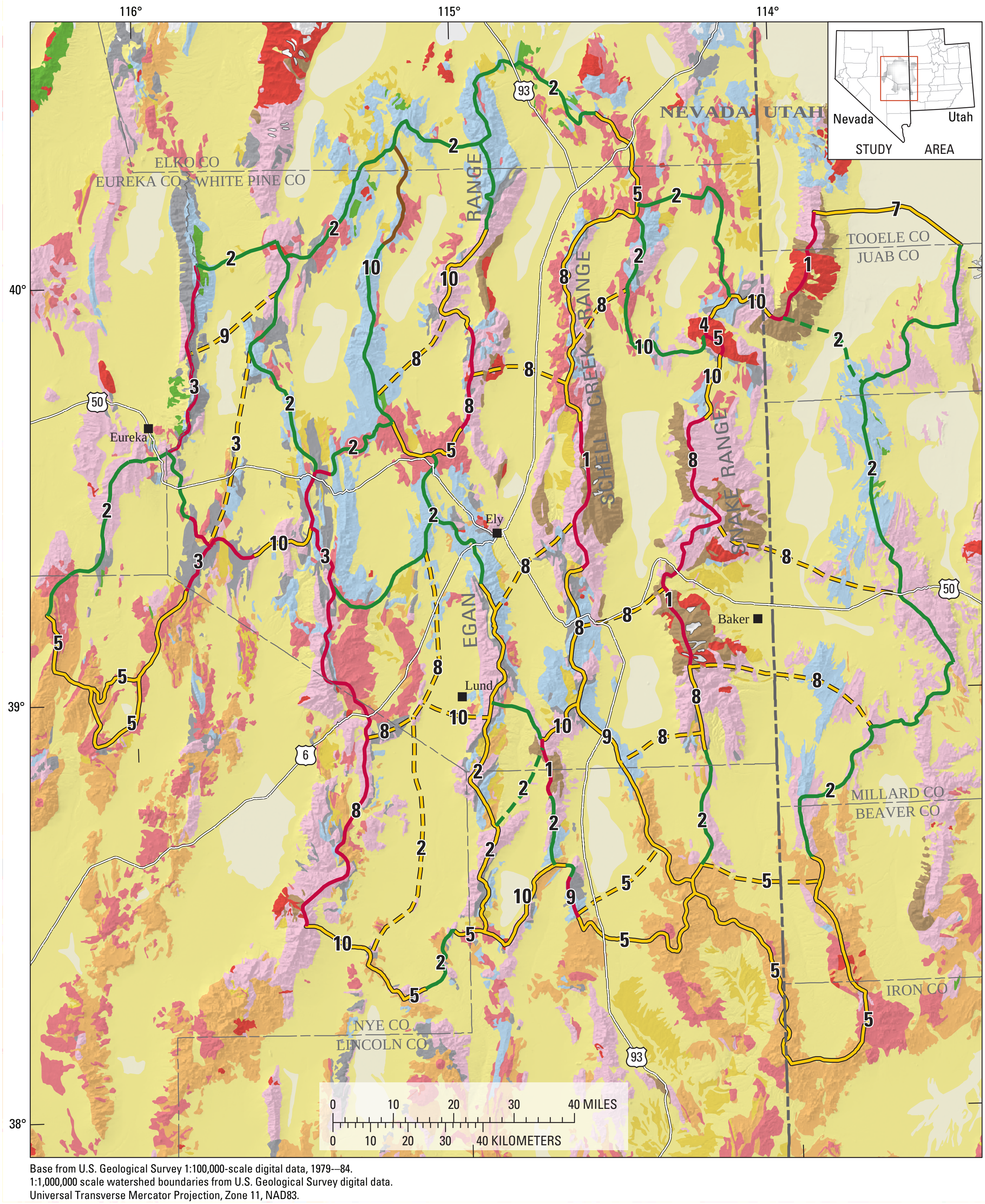
[Description of hydrogeologic unit is given in table 1]

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 hydraulic connection of aquifers and confining units across 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 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.



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 15

Hydrogeologic unit

FYSU–Fine-grained younger sedimentary rock unit (primarily lacustrine and playa deposits)

CYSU–Coarse-grained younger sedimentary rock unit (alluvial and fluvial deposits)

OSU–Older sedimentary rock unit (consolidated Cenozoic rocks)

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 siliciclastic 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 subsurface
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 subsurface
10	Structural disruption may permit subsurface flow

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 LCU 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 (IU) in the subsurface. Given that the igneous rocks are inferred to persist to great depths, ground-water flow likely does not cross this boundary.

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 (USCU) 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

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The ground-water flow system in the study area is influenced by a combination of topography, climate, and geology. Driven by the hydraulic gradient, ground water moves through permeable zones from areas of recharge to areas of discharge. The ground-water flow system includes flow paths of three distinct scales—local, intermediate, and regional (fig. 16). These terms are adapted from Toth (1963) and Freeze and Cherry (1979), and were defined by the depth of ground-water circulation and length of the flow path. Local flow systems are characterized by relatively shallow and localized flow paths that terminate at upland springs. These 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.

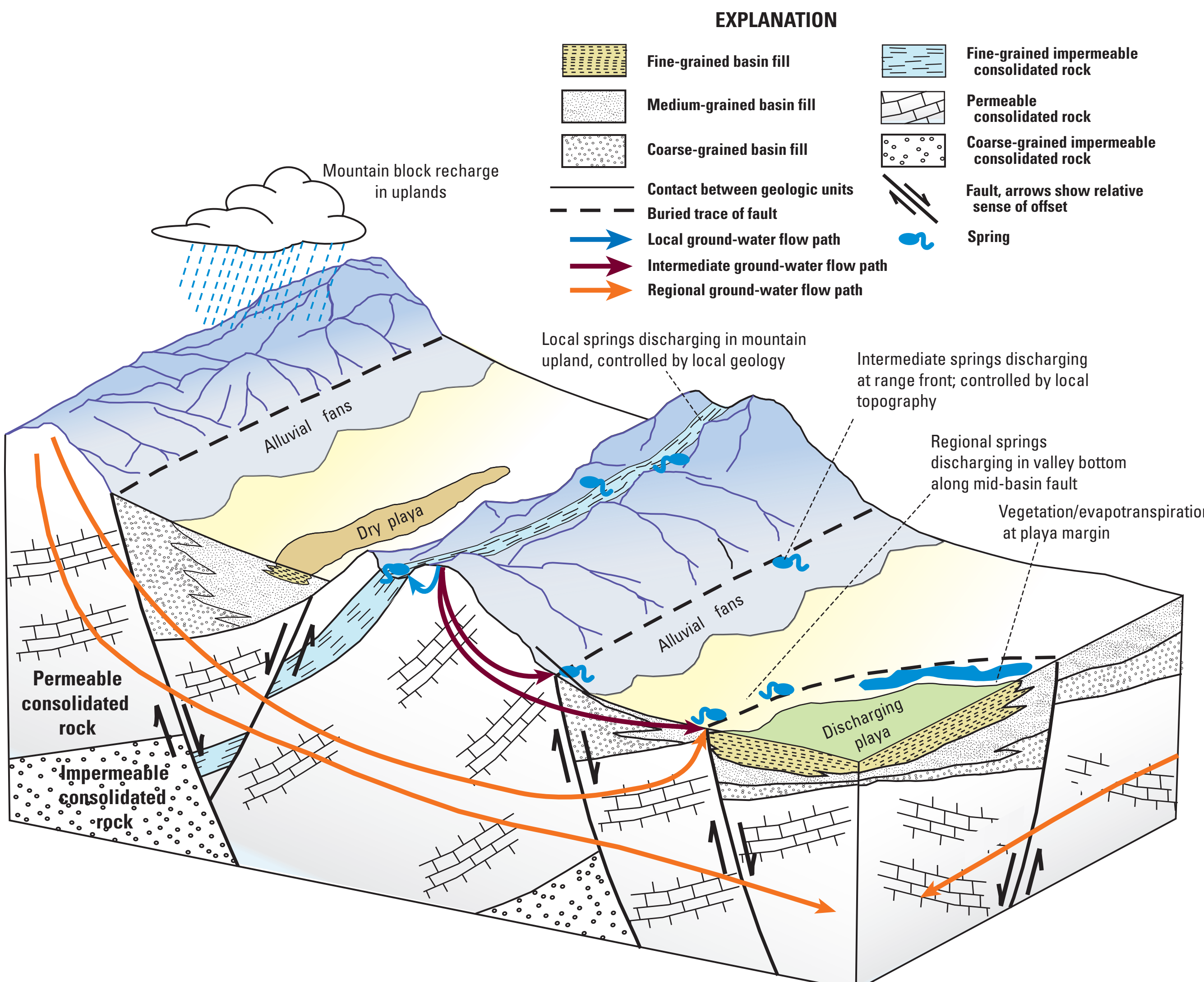


Figure 16. Conceptual ground-water flow systems.

Intermediate flow systems include flow from upland recharge areas to discharge areas along the floor of the intermontane valley. Within intermediate-flow systems, ground-water discharge from springs typically occurs near the intersection of the alluvial fan and the valley floor near the range front and on the adjacent valley floor. Intermediate-flow system springs often are of moderate volume and tend to have less variable flow relative to local springs. Regional ground-water flow is driven by hydraulic gradients that continue over long distances (tens to hundreds of miles). Deep regional flow through basin-fill or consolidated bedrock aquifers is less constrained 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 mountain ranges before reaching its ultimate discharge area. Discharge from these regional flow systems manifests as large springs and, in some areas, extensive wetlands (Mendenhall, 1909).

Under steady-state conditions, ground-water inputs and ground-water outputs are equal and storage is constant. Ground-water input to a basin includes recharge from precipitation and infiltration from hydraulically connected lakes and streams. Ground-water output from a basin includes discharge from springs and hydraulically connected lakes and streams, and evapotranspiration (ET). Early on numerous scientists recognized that many individual basins, particularly in the Great Basin Province were not closed systems and that subsurface inflow and outflow to basins must be considered (Meinzer, 1911; Eakin, 1966; Harrill and others, 1988; Prudic and others, 1995; Harrill and Prudic, 1998; Nichols, 2000). Excess recharge relative to discharge for individual basins was an important factor in recognizing the existence of flow across basin boundaries. Additionally large volume springs in the study area could not be supported entirely by the local recharge from the adjacent mountain ranges, and therefore must be supplied in part from subsurface ground-water flow originating outside the basin. Based on chemistry, temperature, and other criteria, Mifflin (1968) identified selected springs that likely are discharge points from the regional aquifer system.

Typically ground-water pumping initially removes water from storage. This transition from steady-state to transient conditions is recognized by lowering of water levels in wells, declines in spring flow, and, where the ground-water system is hydraulically connected to surface-water bodies, can lead to increased recharge from streams or loss of baseflow. To better characterize the aquifers in White Pine County, water in storage was estimated for a representative volume of aquifer, and water-quality data were compiled and collected to assess the quality of ground water relative to primary and secondary drinking-water standards.

Ground-Water Flow

Ground-water flow patterns in the basin-fill and carbonate-rock aquifer systems can be inferred from the water-table and potentiometric-surface maps, respectively. A spatially interpolated contour map of the ground-water potential (potentiometric surface) is a visual representation of a surface connecting points of equal altitude to which water will rise in tightly cased wells that tap a confined aquifer system (Lohman, 1979). The water table is a particular potentiometric surface; the pressure is atmospheric (Lohman and others, 1970). Water-table and potentiometric-surface maps were constructed to exemplify aquifer system scales, hydraulic barriers, and gradients that control the direction and relative rates of ground-water flow. Ground water generally flows from areas of recharge (high heads) to areas of discharge (low heads) in a direction perpendicular to the water-level contours. The potentiometric-surface map was used to evaluate the permissible locations for flow between HAs and provided hydraulic gradient information needed to assess the volume of ground water flowing across basin boundaries.

The water-table and potentiometric-surface maps primarily were based on measured ground-water levels in wells. Data used to construct the water-table and potentiometric-surface maps shown on plates 2 and 3, respectively, are summarized in Wilson (2007, appendix A). In areas where few control points were available, published water-table and potentiometric-surface maps were used to guide map construction (Mifflin, 1968; Hess and Mifflin, 1978; Garside and Schilling, 1979; Johnson, 1980; Pupacko and others, 1989; Thomas and others, 1986; and Bedinger and Harrill, 2005). Geologic information and delineated recharge and discharge areas also aided in map construction.

Ground water in the basin-fill aquifer generally flows from recharge areas (high heads) at the intersection of the mountain front with the valley margin to discharge areas (lower heads) on the 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. Ground water can exit a basin as subsurface flow to downgradient basins where hydraulic continuity in the basin fill exists between HAs or where ground-water in the basin-fill aquifer flows downward into the underlying carbonate (Thomas and others, 1986). Hydraulic continuity among several basins is depicted by open water-level contours on the water-table map. The water-table map was constructed by contouring the water-level data from 299 wells completed in the basin-fill aquifer at 100-ft intervals (pl. 2). Water-level altitudes ranged from slightly more than 6,800 ft to slightly less than 4,400 ft above sea level in southern Steptoe Valley and in northern Snake Valley, respectively.

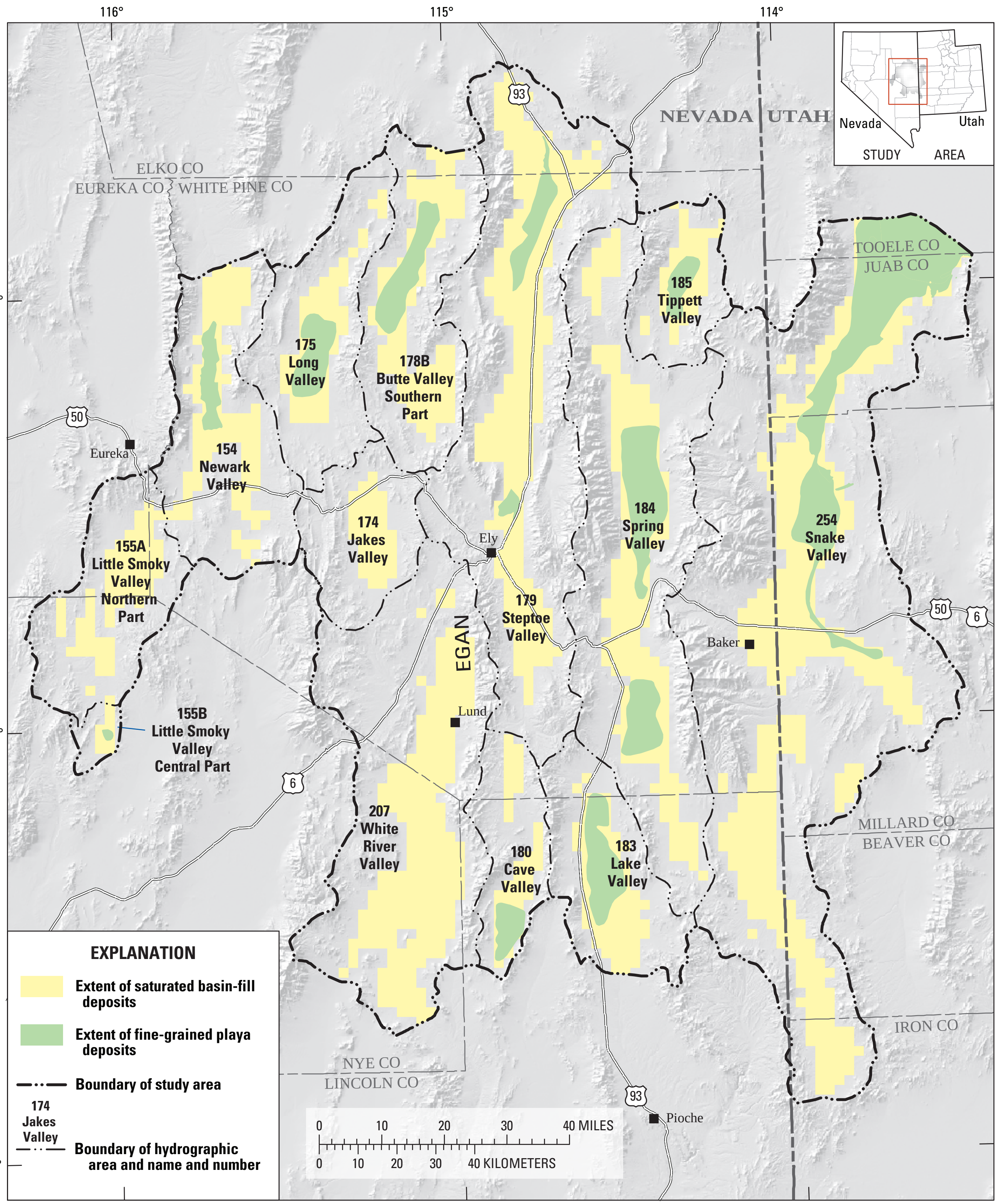
The potentiometric-surface map was constructed from water levels in wells completed in basin fill and underlying carbonate rock. In many places, basin fill and carbonate rocks are hydraulically connected resulting in a single continuous ground-water flow system (Thomas and others, 1986). The following general guidelines used for identifying regional head for mapping regional potential are from Bedinger and Harrill (2005). Regional hydraulic head can be represented by shallow water levels in large areas of low topographic relief and virtually no recharge. Regional hydraulic head is at or above shallow water levels in areas of local, intermediate, and terminal discharge by ET in basins, at regional spring heads, and areas where ground water is discharging to major surface-water bodies. Regional hydraulic head is below the altitude of non-discharging dry playas, lower than the water table in areas of recharge, and lower than local spring heads. Using these guidelines, it was considered acceptable to map selected water-level data from wells completed in the basin fill where suitable data from the carbonate aquifer are scarce or lacking. Due to the scarcity of wells completed in the carbonate rocks, the control points used to construct the map are less precise and a large contour interval (500 ft) was selected for representing the potentiometric surface of the carbonate-rock aquifer. In locations where multiple wells are completed at differing depths, such as at MX well sites, the vertical gradients generally are less than 200 ft (Tumbusch and Schaefer, 1996, tables 1-2). The potentiometric-surface map was constructed by contouring water-level data from 119 wells, 76 basin fill wells, and 43 carbonate-rock and other consolidated-rock wells (pl. 3). Water-level altitudes ranged from slightly more than 6,500 ft to slightly less than 4,500 ft above sea level in Steptoe Valley and in northern Snake Valley, respectively.

The regional ground-water recharge area for the carbonate-rock aquifer is a relatively large recharge mound over Steptoe, Butte, Long, and Jakes Valleys; small, high mounds are centered on the Schell Creek, and Egan Ranges (pl. 3). This large recharge mound comprises the headwaters of four regional flow systems—Great Salt Lake Desert, Goshute Valley, Colorado, and Newark Valley (fig. 1). 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 unconfined and confined aquifers becomes available as ground water is pumped and water levels decline. 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 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. Water is stored within the pore spaces of saturated unconsolidated sediment or rock in a water-table aquifer and 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. Water stored within confined aquifers becomes available 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.

The volume of water stored in unconfined and confined aquifers was computed using the extent of basin-fill deposits, a water-level decline of 100 ft, and a storage term (the specific yield of the basin-fill aquifer is 0.15; the storage coefficient of the carbonate-rock aquifer is 0.001). A water-level decline of 100 ft was arbitrarily selected, but likely is a reasonable limit for widespread lowering of the ground-water surface for a valley. The area used to calculate storage is the region where the thickness of the basin fill is equal to or greater than 100 ft. This area is assumed to reasonably approximate the acreage of saturated basin fill. In calculating the unconfined storage, the saturated basin-fill area was reduced by removing the acreage of 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 shown on the geologic map (pl. 1). This small area is assumed to reasonably approximate the acreage of the drainable basin fill. 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).



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.

Ground-water storage is estimated for each HA and is the contribution of unconfined and confined storage (fig. 18 and appendix A). Estimates range from less than 1 million acre-ft for Cave, Jakes, or Tippett Valleys to more than 5 million acre-ft for Snake, Steptoe, or White River Valleys. Storage estimates for the remaining HAs range from more than 1 to less than 4 million acre-ft. Snake Valley has the largest estimated storage at nearly 9 million acre-ft. 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. About 36 million acre-ft of water is stored in a 100 ft of saturated basin-fill aquifer beneath all valley floors. In contrast, only about 300,000 acre-ft of water is stored in a 100 ft of saturated carbonate-rock aquifer for a slightly larger area, or about 2 orders of magnitude less than the basin-fill aquifer. Confined storage contributes less than 100,000 acre-ft to the total storage of any HA. Estimates of storage do not consider the effects of any limiting geologic, hydrologic, or cultural factors, such as impermeable or low permeability lithologies, recharge to basin fill or carbonate-rock aquifers, declining water levels in wells, decreasing spring flow, diminished water quality, or loss of native vegetation.

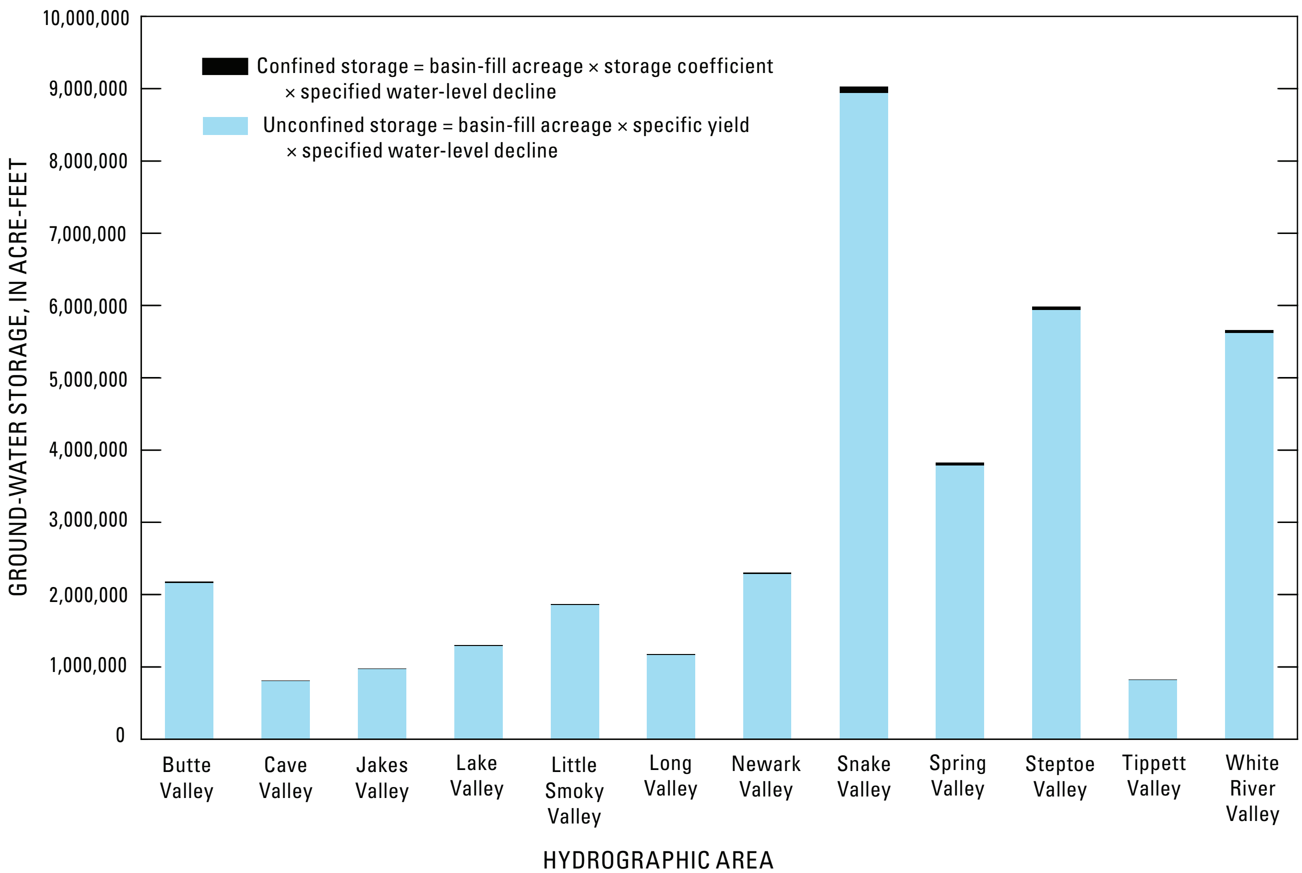


Figure 18. Ground-water storage estimates by hydrographic area based on a 100-foot lowering of water levels beneath valley floors, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

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 databases, 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 a subset of chemical constituents having National primary and secondary drinking-water standards (U.S. Environmental Protection Agency, 2004), ground water in the study area generally is of good quality (table 4). Primary standards regulate constituents that are believed to pose a risk to human health if consumed above a certain threshold. Secondary standards regulate water-quality parameters that are not believed to pose a risk to human health, but can have undesirable aesthetic, cosmetic, or technical effects (Hershey and others, 2007). 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. Secondary drinking-water standards were exceeded more often than the primary standards but exceedances were not common. Values of pH were outside of the acceptable range of 6.5–8.5 (secondary standard) at 21 of 179 sites. Chloride concentrations exceeded their secondary standard at 6 of 179 sites. Sulfate concentrations exceeded their secondary standard at 4 of 177 sites.

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 from wells located in 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 4. 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 milligrams per liter except for pH, which is in standard units. –, no standard]

Constituent	Drinking-water standards		Number of sampling sites	
	Primary	Secondary	Constituent detected	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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[1]U.S. Geological Survey
[2]Desert Research Institute

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 (recharge) and outflows (discharge). 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 for this study. Average annual recharge and ground-water discharge were estimated at the subbasin scale to develop a water budget for each HA. In addition, recharge and discharge estimates were summed for the entire study area and used to develop a study-area water budget. 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 all applied similar approaches for estimating recharge and discharge.

Annual recharge has been estimated for the 12 HAs in the BARCAS study area and published in numerous reports (table 5). Estimates of recharge presented in reconnaissance reports typically were based on a method developed by Maxey and Eakin (1949). The method originally was developed to estimate the recharge to 13 HAs in east-central Nevada and 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 (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 receive greater precipitation and 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. Recharge efficiencies, determined by balancing recharge and discharge, were associated with each of the five precipitation zones. 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. The method has been applied to more than 200 basins in Nevada.

Ground-water discharge typically has been estimated using a volumetric calculation of ET from major areas of phreatophytic vegetation (table 6). 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). Ground-water discharge estimates are based on maps that delineate distinct groupings of phreatophytes and moist soils in ground-water discharge areas and coefficients relating these groupings to ET or ground-water discharge rates. ET rates were determined from pan evaporation and lysimeter data, and ground-water discharge rates from ET rates were adjusted downward to remove the local precipitation component. Ground-water discharge for an HA was estimated by computing the product of the ground-water discharge rates and the corresponding area for a particular vegetation or soil moisture grouping, and integrating the products for all groupings in the HA. At the time of most reconnaissance estimates, the volume of water used for irrigation and self-supply was small and often was ignored in water-budget computations. Spring flow typically was not accounted for directly in the water budget but was indirectly accounted for because the total ET estimated or measured from a discharge area includes spring flow (Eakin, 1960). In some reconnaissance studies, ground-water discharge was not determined independently but was assumed to be equal to the Maxey-Eakin estimate of recharge.

Table 5. Estimates of annual ground-water recharge, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

[**USGS authored reports** indicated in bold 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)		Nichols (2000)	Epstein (2004)	Dettinger (1989)	Kirk and Campana (1990)	Thomas and others (2001)	Flint and others (2004)		Brothers and others (1993a,b, 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 (1993a).

[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 5). These methods commonly are variations on the Maxey-Eakin method and often have relied on a different precipitation map, more recent estimates of ground-water discharge (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 recharge-accounting model (Flint and others, 2004), and numerical simulation (Brothers and others, 1993a, 1993b; Brothers and others, 1994). For HAs in the study area, 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 6), reported methods are variations on the Maxey-Eakin method of multiplying a ground-water discharge rate by the associated area of phreatophytic vegetation. However, technological advances such as the utilization of micrometeorological and remote-sensing methods have improved ground-based measurements and area-wide estimates of ET (Nichols, 2000).

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

[**USGS authored reports** indicated in bold 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]11	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 (1993a).
[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)**.

Ground-Water Recharge

The primary source of water recharging 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 is 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, recharge-accounting model. This recharge model provides a means for evaluating and comparing the processes, properties, and climatic factors that ultimately control the potential for recharge under differing hydrologic conditions (Flint and others, 2004). The Basin Characterization Model (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.

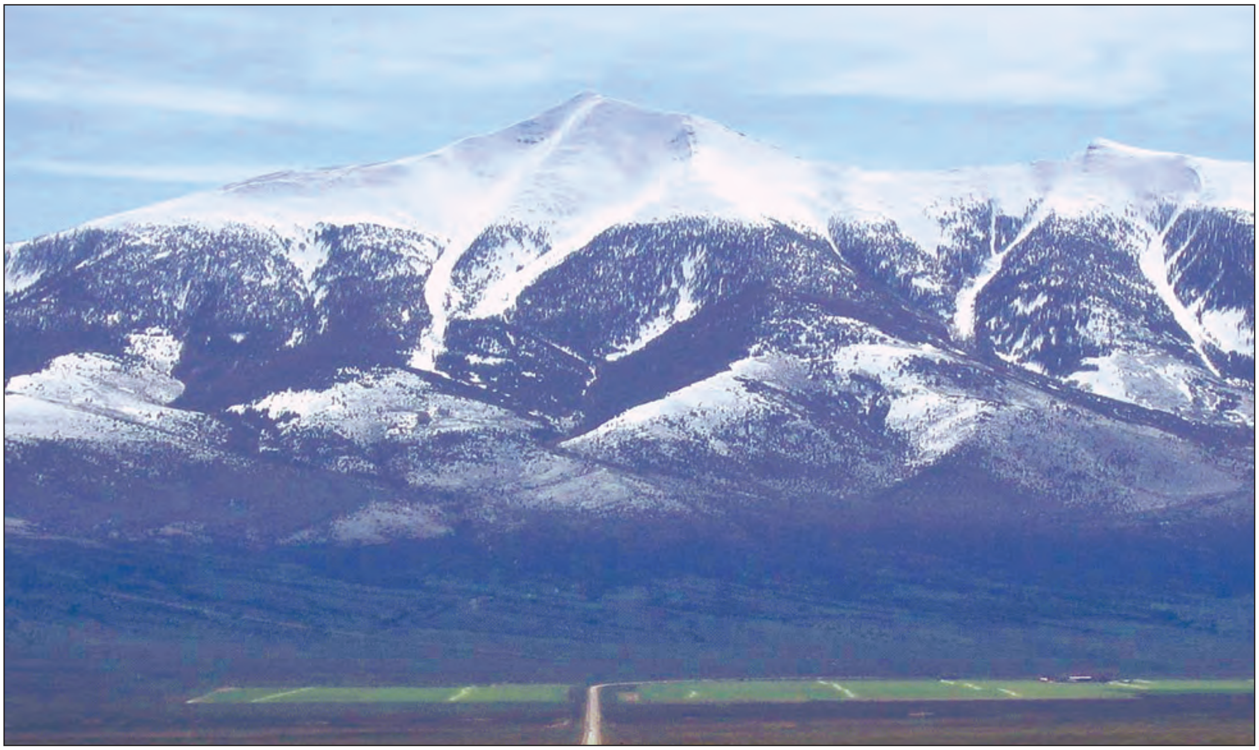


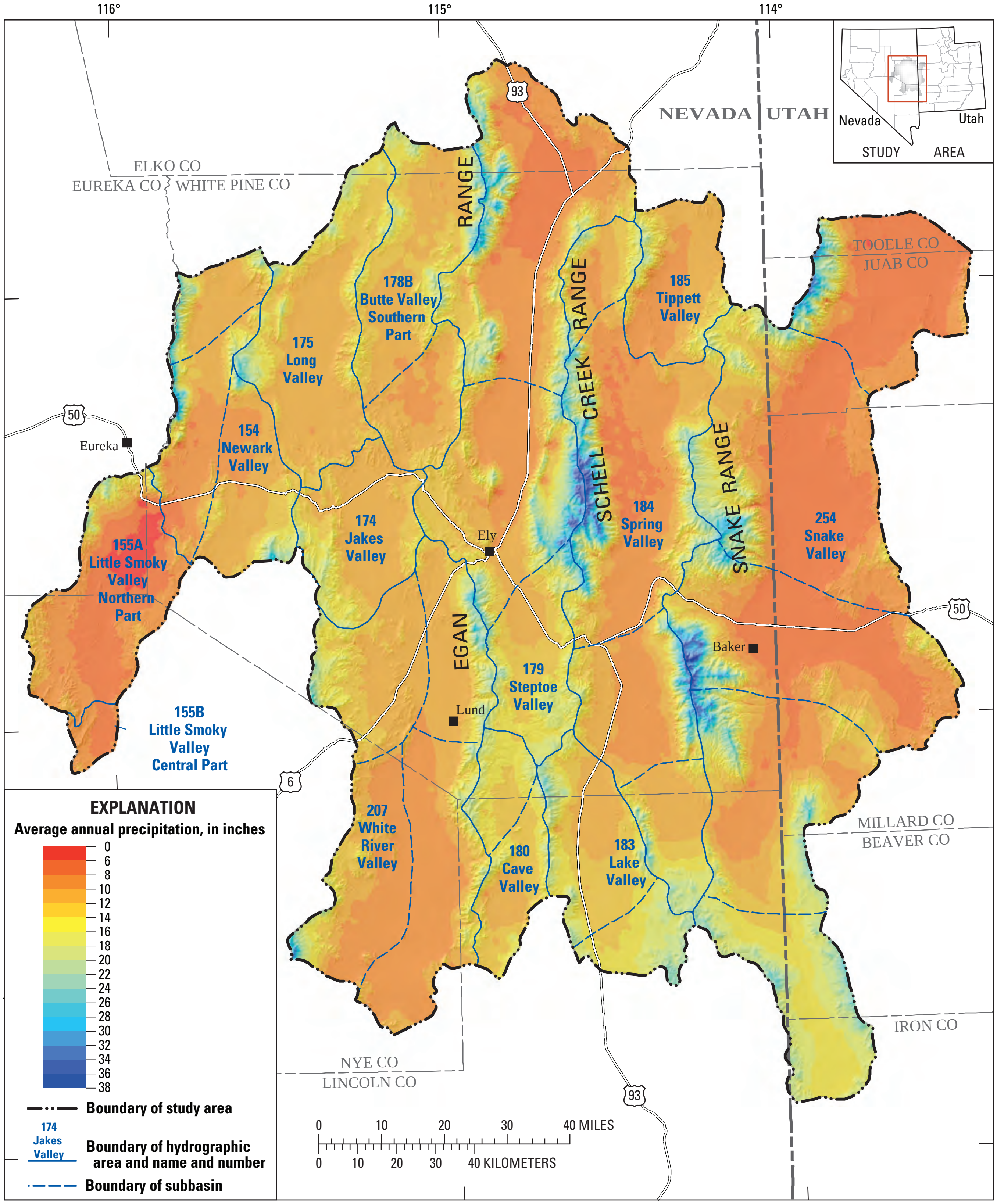
Figure 19. Precipitation (snowfall) on a typical bedrock highland flanking an alluvial valley in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah. West side of 13,063-foot-high Wheeler Peak in southern Snake Range, Great Basin National Park, Nevada. Photograph taken by Michael T. Moreo, U.S. Geological Survey, May 17, 2005.

The model is an updated and refined version of the BCM initially documented in Flint and others (2004) and was applied in this study to estimate potential annual in-place recharge and recharge from runoff for 1970–2004. Details of the updated and refined BCM as applied to the BARCAS study area can be found in Flint and Flint (2007).

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 to estimate the distribution of precipitation (fig. 20), snow accumulation and snowmelt, potential ET, soil-water storage, and bedrock permeability. The distribution of precipitation was estimated using the Parameter-Elevation Regressions on Independent Slopes Model (PRISM; Daly and others, 1994). PRISM estimated precipitation was compared to measured precipitation at 155 stations in Nevada and Utah. Annual measured precipitation for these stations averaged 12 in. and was about 1 in. less than PRISM estimates. Differences between measured precipitation and PRISM estimates had a standard deviation of 4 in. Therefore errors resulting from using PRISM to distribute precipitation in the BCM were considered negligible.

The accuracy of other BCM-estimated parameter distributions also was evaluated, such as snow accumulation and snowmelt, runoff, and potential ET. Snow accumulation and snowmelt models were calibrated to mid-winter and late spring Moderate Resolution Imaging Spectroradiometer (MODIS) satellite data for 2000–2004 (Flint and Flint, 2007). BCM-simulated runoff and potential ET have been calibrated to measured data in previous studies (Flint and Flint, 2007). For example, simulated runoff was within 5 percent of measured discharge from the nine Arizona basins with between 10 and 35 years of streamflow records. Simulated potential ET was calibrated to 204 California Irrigation Management Information System (CIMIS) sites in California and 26 Arizona Meteorological Network (AZMET) sites in Arizona in 2006.

The BCM precipitation and recharge relation for 1970–2004 were extrapolated using regression analysis to estimate long-term average recharge for 1895–2006. Recharge during the 1895–2006 period was assumed to be representative of long-term recharge to the BARCAS area and differed from BCM results because annual precipitation for 1895–2006 was 5 percent less than for 1970–2004 (Flint and Flint, 2007). The long-term average recharge for 1895–2006 was estimated for



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, 1970–2004.

each subbasin by relating annual recharge computed by the BCM for 1970–2004 to annual precipitation using a threshold-limited, power function (Flint and Flint, 2007). The threshold-limited, power function largely interpolated values because more than 98 percent of the annual precipitation volumes for 1895–2006 were within the range that was observed during the 1970–2004 precipitation period. This approach assumes that antecedent conditions from previous years do not affect annual recharge. This assumption may be inappropriate for predicting recharge in any given year but should minimally affect the estimate of a 112-year average.

Long-term recharge was calculated as the combination of in-place recharge and 15 percent of the potential runoff for each subbasin in the 12 HAs of the study area (pl. 4). Total long-term recharge is estimated at about 477,000 acre-ft and potential runoff at about 361,000 acre-ft (appendix A). Assuming that 15 percent of the potential runoff becomes regional ground-water recharge, 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 (68 percent) are Steptoe, Snake, and Spring Valleys (fig. 21). The ground-water recharge in these HAs averages more than 100,000 acre-ft annually. The estimated annual recharge for all other HAs ranges from about 4,000 acre-ft in Little Smoky Valley to 35,000 acre-ft in White River and Butte Valleys. Average annual ground-water recharge is less than 15,000 acre-ft for Cave, Lake, Little Smoky, and Tippett Valleys (appendix A; fig. 21). Even though White River Valley is relatively large at more than 1 million acres (12 percent of the study area), this HA only contributes 7 percent of the total recharge. The 12 HAs in the study area average 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.

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. Even though in-place recharge is the primary recharge source 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). BCM results indicate that the source of most ground-water recharge is in-place recharge at high altitudes in Cave, Lake, Snake, Spring, and Steptoe Valleys. This conclusion is supported by dissolved gas and stable-isotope data collected during this study from 15 sites (Victor Heilweil, U.S. Geological Survey, written commun., 2007). In contrast, BCM results indicate that the estimated total potential runoff is greater than the average annual ground-water recharge in Lake and Snake Valleys. Climate variability and the climate periods used in the analysis add uncertainty to the recharge estimates.

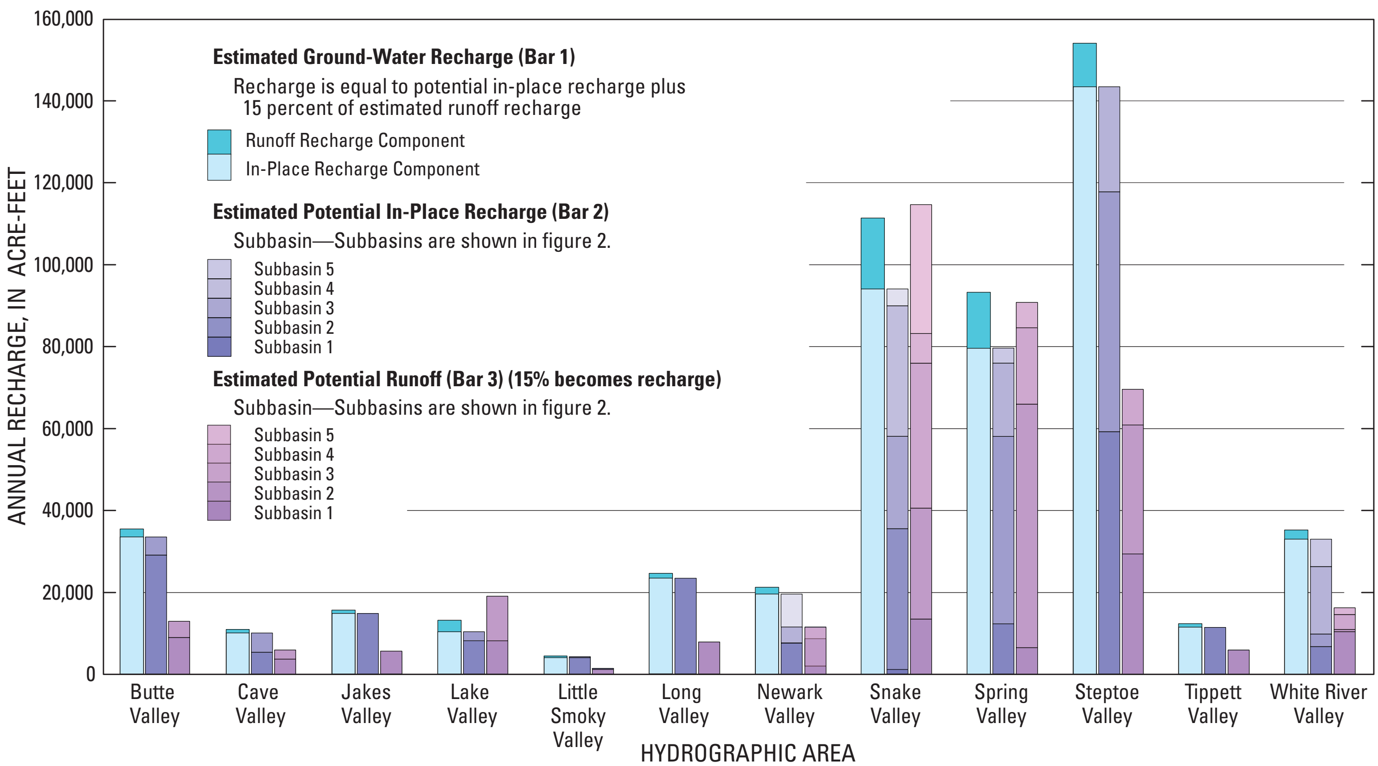


Figure 21. Mean annual ground-water recharge to hydrographic areas in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 1895–2006.

Limitations and Considerations of Methodology

The accuracy of recharge estimates given in this report depends primarily on the validity of assumed hydrologic processes and on the associated uncertainty of input parameters to the BCM. Although the overall accuracy in estimates of volumetric recharge may be difficult to quantify, the estimates presented in this report are considered reasonable because the BCM is a physically based method and therefore scientifically defensible. The BCM incorporates spatially distributed moisture accumulation, soils, and geology that directly affect recharge magnitude and distribution. Hydraulic properties and geologic distributions undoubtedly are not perfect, but the BCM is conceptually correct.

BCM recharge estimates were derived by assuming that net infiltration is equal to in-place recharge and that topographic boundaries coincide with ground-water divides. Actual conditions may differ from these assumptions for some areas but the effect on average annual, regional recharge estimates likely would be minimal because the primary recharge areas occur along ranges entirely within the study area.

Data are limited for some input parameters used in the BCM. As a result, the uncertainties associated with these parameters may be a significant source of potential error for estimates of ground-water recharge, particularly for two parameters—the saturated hydraulic conductivity of bedrock and the associated volume of runoff that becomes recharge. The hydraulic conductivity estimates for bedrock are uncertain because of limited data on hydraulic properties in recharge areas, particularly on the properties and spatial distributions of fractures, faults, and fault gouge. Hydraulic conductivities used in the BCM span up to five orders of magnitude for the least permeable units (Flint and Flint, 2007, table 2). BCM-estimated recharge is relatively sensitive to changes in saturated hydraulic conductivity of bedrock because this parameter determines the partitioning of water between in-place recharge and runoff (Flint and Flint, 2007). Although the portion of runoff that becomes recharge varies significantly across Nevada (Flint and others, 2004), an assumed value of 15 percent is considered reasonable for areas of central Nevada dominated by in-place recharge; however, the uncertainty likely is greater in runoff-dominated areas. Previous investigations used percentages of recharge from runoff ranging from a low value of 10 percent in the Death Valley regional flow system in southern Nevada to a high value of 90 percent for the Humboldt regional flow system in northern Nevada (Flint and Flint, 2007). These percentages were therefore chosen as the endpoints representing the range of uncertainty for the recharge estimates shown in figure 22 (gray bars)—10 percent for the low end and 90 percent for the high end. As a result, the uncertainty in recharge estimates increases for basins such as Lake, Snake, and Spring Valleys where the potential runoff recharge exceeds the potential in-place recharge (figs. 21 and 22).

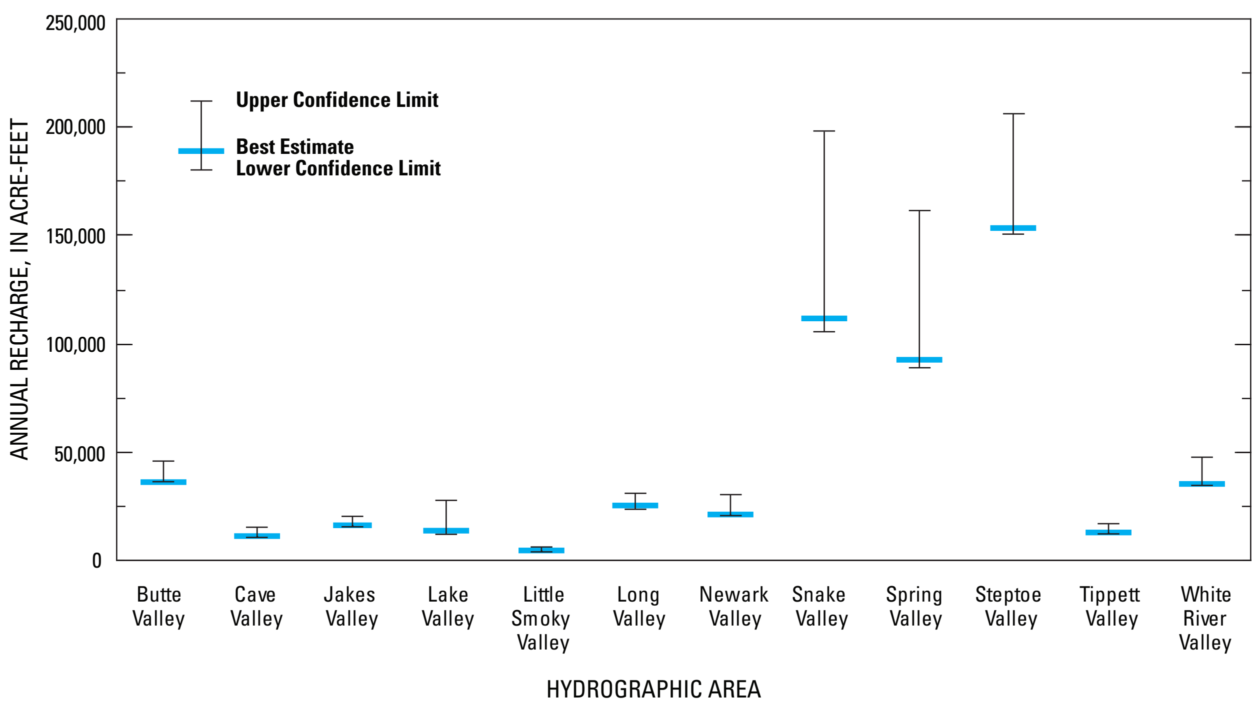


Figure 22. Uncertainty in ground-water recharge estimates by hydrographic area, Basin and Range carbonate-aquifer system study area, Nevada and Utah.

Comparison of Ground-Water Recharge Estimates

The long-term average recharge estimates are within the range of previous estimates except in Snake, Steptoe, and Tippett Valleys (fig. 23). In these three HAs, the recharge estimates are greater than or equal to the upper limit of the range of previous values likely resulting from the prevalence of high permeability carbonate bedrock in the highlands surrounding these HAs. Unlike the adjusted BCM estimates, most of the alternative methods used to estimate recharge neglect the effects of spatial variability in bedrock and soil permeability.

One such alternative method, used in several studies in Nevada, is the chloride mass-balance method for estimating ground-water recharge (Dettinger, 1989; Maurer and Berger, 1997; Russell and Minor, 2002; Mizell and others, 2007). Mizell and others (2007) recognized that the chloride mass-balance analysis provides a reconnaissance level estimate of recharge for 9 of 12 HAs in the study area. Recharge estimates could not be made in three HAs (Long, Tippett, and Newark Valleys) due to the lack of available chloride data. 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.

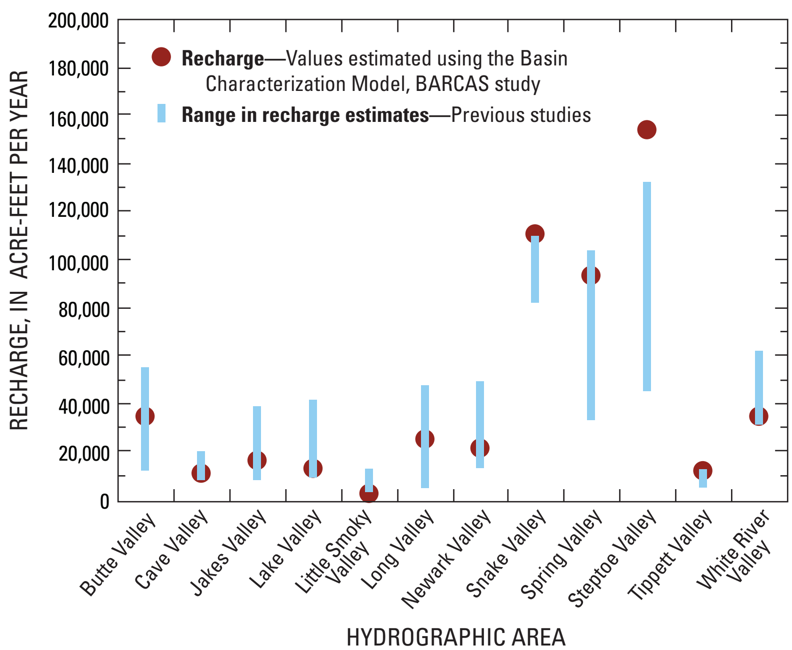


Figure 23. Comparison of ground-water recharge estimates, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Ground-Water Discharge

Ground water discharges naturally 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 natural discharge processes, water also can be 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 each of the major ground-water discharge areas. Estimates of ground-water discharge represent pre-development conditions. Springflow does not need to be accounted for directly in an ET-based estimate of ground-water discharge because water discharging from springs is either lost through ET or recharges shallow ground-water flow systems where it later is transpired by the local phreatophytes. 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 lateral inflow or 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 exiting an HA or the study area as subsurface outflow is discussed in terms of the difference between the estimated mean annual recharge and mean annual ground-water discharge.

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 declines and soils dry. As water levels decline and soil moisture lessens, the vigor of phreatophytic vegetation decreases. Conversely, as less water is removed, the water table rises and soils moisten, and the vigor of the phreatophytic vegetation often increases. 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 water to the ET process. 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, 1967; Glancy, 1968; Laczniak and others, 1999, 2001; Nichols, 2000; Berger and others, 2001; Reiner and others, 2002). Using this calculation, an average annual estimate of ET is computed by summing ET from all areas discharging ground water. The procedure, as applied in this study, delineates groupings of similar vegetation and soil-moisture conditions (ET units) within major discharge areas, and computes the annual ET from each ET unit within a discharge area. Average annual ET for an area, such as a subbasin or an HA, is estimated by summing the 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 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 that are intended to differentiate areas of similar ET often are referred to as ET units.

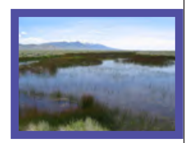
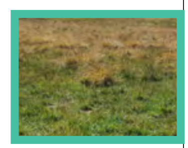
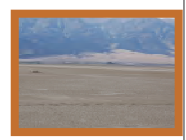
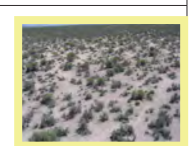
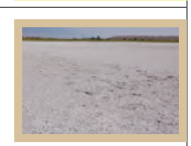
More recent studies have estimated ET from many discharge areas in Nevada using Landsat Thematic Mapper (TM) imagery to map ET units (Laczniak and others, 1999; Nichols, 2000; Berger and others, 2001; Nichols and VanDenburgh, 2001; Reiner and others, 2002; DeMeo and others, 2003). TM imagery has a resolution 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 (table 7) 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 in the study area where ground water is lost by ET to the atmosphere. 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 ten ET units describe shrub dominated environments.

The method used to delineate ET units and their spatial distributions varied based on attributes specific to the ET unit of interest. For example, ET units whose spatial distribution varies significantly from year to year because of changes in climatic conditions are best delineated as an average distribution from multiple years of imagery. An example of a temporally varying ET unit is open water. The distribution of open water increases and decreases with changes in annual precipitation, whereas the distribution of shrubland is scarcely influenced by these same changes. Shrubland, grassland, meadowland, and moist bare soil ET units were delineated using modified soil-adjusted vegetation indices (MSAVI, Qi and others, 1994) and a tasseled cap transformation (Huang and others, 2002) computed from a single TM image (Smith and others, 2007). Dry playa, marshland, and open water ET units were delineated using a published land cover map (Southwest Regional Gap Analysis Program, SWReGAP) that was based on multiple dates of TM imagery (Kepner and others, 2005). And lastly, recently irrigated lands were delineated from multiple dates of TM imagery (Welborn and Moreo, 2007). The acquisition time of the TM images used in this analysis generally coincides with the near-peak period of ET around the summer solstice. The ET units delineated by the different techniques were combined in each ground-water discharge area and together form the ET unit map shown on plate 4. Additional details and the assessment of the overall accuracy of the mapped ET units can be found in Smith and others (2007).

Shrubland is the most prevalent ET unit in the study area (fig. 24, pl. 4). Shrubland, defined as the combined acreage of sparse, moderately dense and dense desert shrubland, occupies more than 80 percent of the acreage delineated as potentially contributing to ground-water discharge (Smith and others, 2007).

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 populated with phreatophytic shrubs and riparian vegetation (Smith and others, 2007).

Table 7. Evapotranspiration (ET) units identified, delineated, and mapped for different vegetation and soil conditions in potential areas of ground-water discharge in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, September 2005–August 2006.

ET-unit name	ET-unit description	Photograph
Xerophytic	Area of no substantial ground-water evaporation. Area dominated by bare dry soil and/or sparse, non-phreatophytic vegetation.	
Open Water	Area of open water including reservoirs, ponds, and spring pools.	
Marshland	Area dominated by dense wetland vegetation, primarily tall reeds and rushes, and some grasses. Vegetation cover typically is greater than 50 percent. Open water is present but typically less than 25 percent. Perennially flooded. Water at or very near surface. Depth to water typically is less than 1 foot.	
Meadowland	Area dominated by short, dense perennial grasses, primarily marsh and meadow grasses. Unit includes occasional desert shrubs and trees, primarily Rocky Mountain junipers and cottonwoods. Vegetation cover typically is greater than 50 percent. Soil typically is moist except in later summer and autumn. Depth to water table typically is less than 5 feet.	
Grassland	Area dominated by short, sparse, perennial grasses, including salt grass, and sod and pasture grasses typically a mix of vegetation types. Unit includes sparse desert shrubs and occasional trees, primarily Rocky Mountain junipers or cottonwoods. Vegetation cover is between 10 and 100 percent. Soil typically is damp to dry. Depth to water table typically is less than 8 feet.	
Moist Bare Soil	Area dominated by moist playa. Near surface soil is damp throughout much of the year. Water table is near or below land surface. Depth to water typically is less than 10 feet.	
Dense Desert Shrubland	Area dominated by sparse desert shrubs, including greasewood, rabbitbrush, shadscale, big sagebrush, and saltbush. Shrubs typically are mixed. Vegetation cover typically is greater than 25 percent. Depth to water can range from about 3 to 50 feet.	
Moderately Dense Desert Shrubland	Area dominated by sparse desert shrubs, including greasewood, rabbitbrush, shadscale, big sagebrush, and saltbush. Shrubs typically are mixed. Vegetation cover typically ranges from 10 to 30 percent. Depth to water can range from about 3 to 50 feet.	
Sparse Desert Shrubland	Area dominated by sparse desert shrubs, including greasewood, rabbitbrush, shadscale, big sagebrush, and saltbush. Shrubs typically are mixed. Vegetation cover typically ranges from 5 to 15 percent. Depth to water can range from about 3 to 50 feet.	
Dry Playa	Area dominated by dry playa. Soil typically dry year round. Water table below land surface. Depth to water typically is greater than 10 feet. This unit may not contribute to ground-water discharge.	
Recently Irrigated Cropland—Historically Mixed Phreatophyte	Area dominated by irrigated cropland. Soil moisture varies with irrigation practice. Water table is below land surface. Depth to water table typically is greater than 5 feet. Prior to irrigation, the unit likely was dominated by sparse desert shrubs to grassland.	

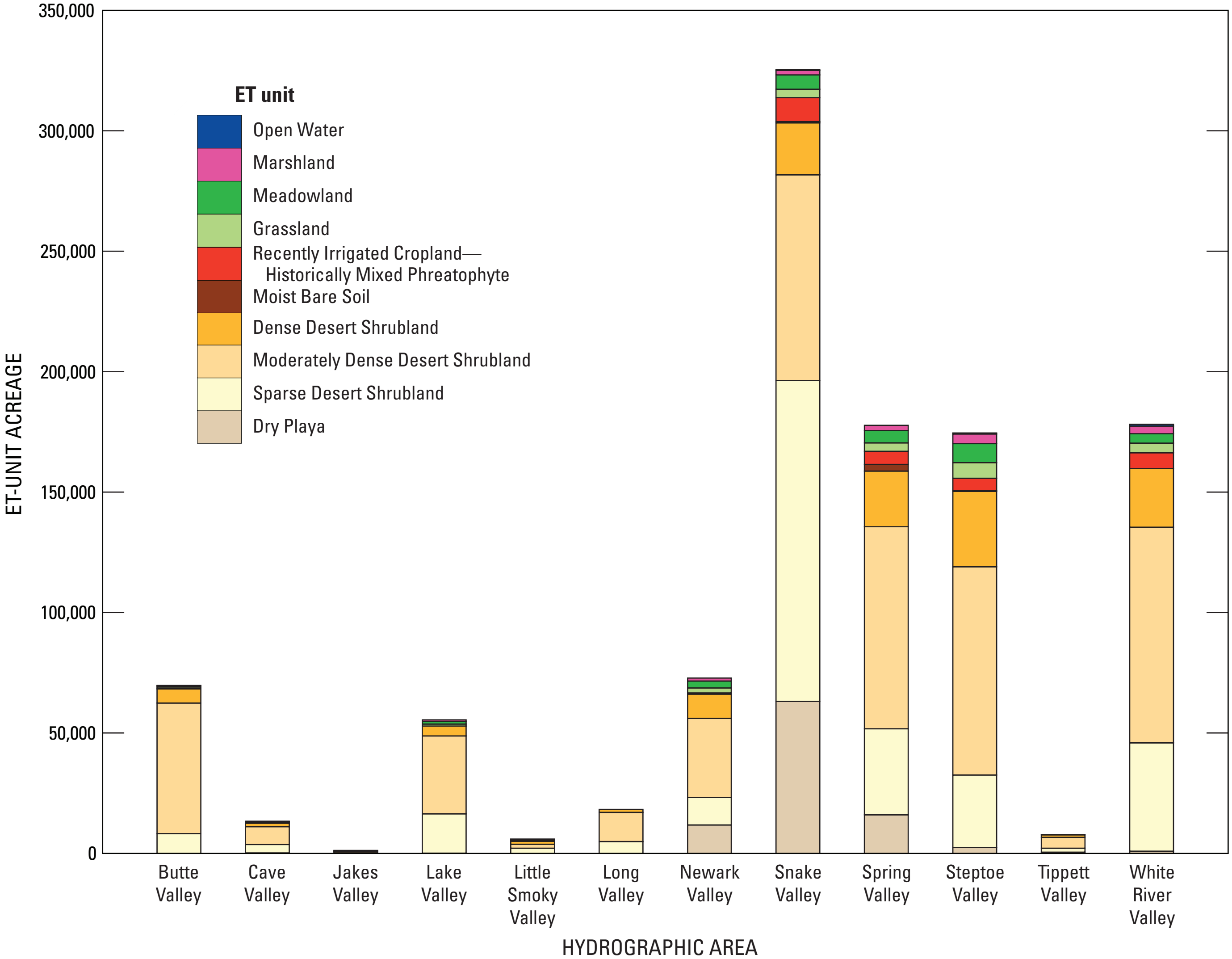


Figure 24. ET-unit acreage by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Riparian vegetation, such as marshland, meadowland, and grassland, occupies only about 6 percent and open water occupies less than 0.1 percent of the ET-unit acreage in the study area.

Shrubland occupies 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, Steptoe 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 occupy no more than about 20 percent of the total ET-unit acreage in any HA. Dry playa is prevalent only in Newark, Snake, and Spring Valleys (fig. 24). In Snake Valley, dry playa occupies nearly 65,000 acres of the valley's ground-water discharge area.

HAs having ET-unit acreage exceeding 150,000 acres are Snake, Spring, Steptoe, and White River Valleys (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 by subbasin is shown in figure 25. ET-unit acreages for individual subbasins used to develop the ground-water discharge estimates are given in appendix A and described in Smith and others (2007).

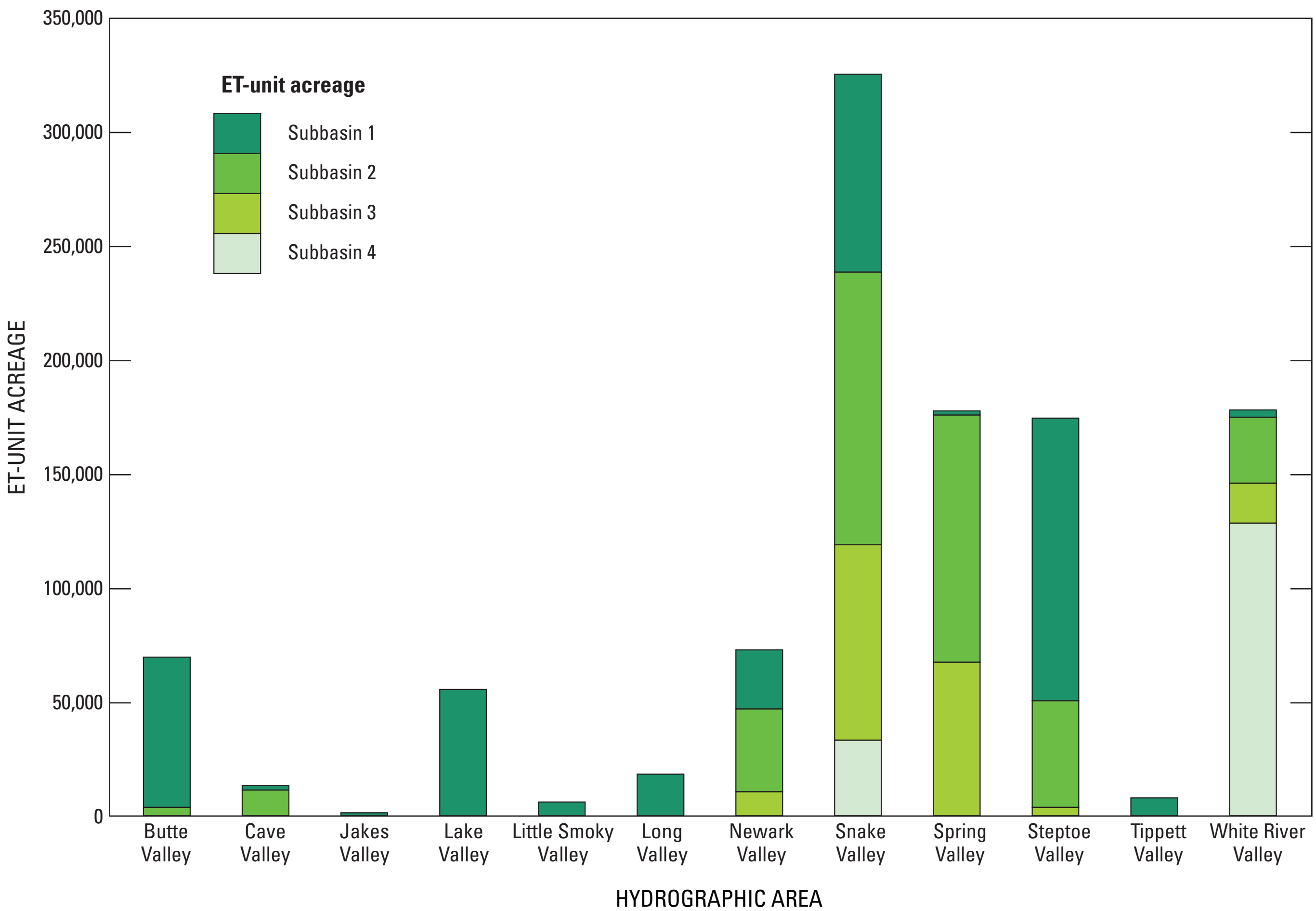


Figure 25. ET-unit acreage by hydrographic area and hydrographic-area subbasin in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Evapotranspiration Rates

The rate at which water evaporates from the earth's surface and is transpired by plants is referred to as the ET rate. The ET rate is driven by the 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, fig. 26). Net radiation is absorbed at the 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:

$$R_n = G + H + \lambda E, \tag{1}$$

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

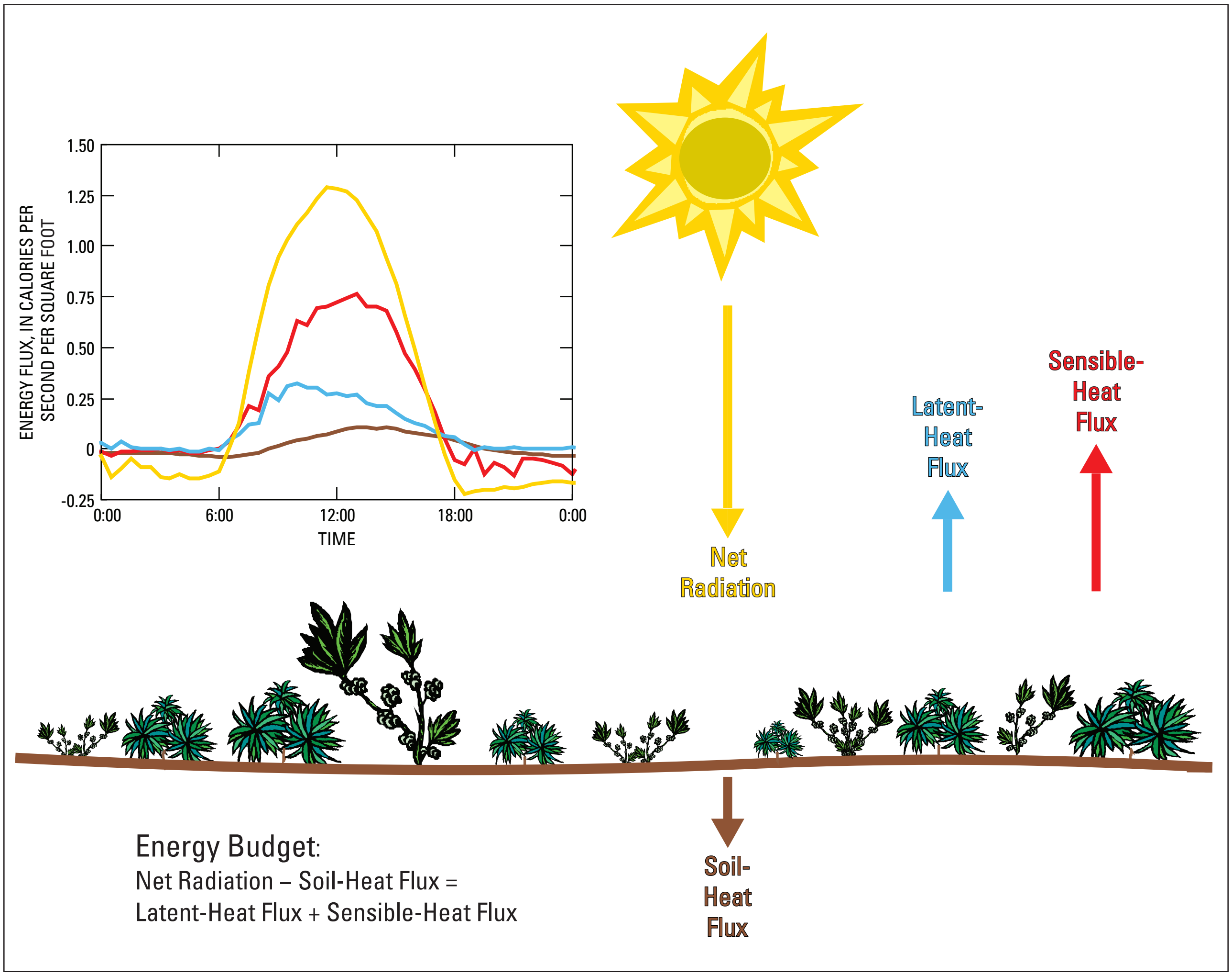


Figure 26. Surface energy processes and typical daily energy budget for shrubs, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

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).

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 ground-water 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, 1966; Eakin and others, 1967; Glancy, 1968) include 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).

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. The majority of the sites were located in shrubland to evaluate the effect of vegetation density on ET rates, and to better quantify ET rates for this dominant vegetation type. Daily and annual ET for the grassland/meadowland ET site (SPV-3) was significantly greater than that for shrubland ET sites in Spring Valley over the 1-year collection period (figs. 29 and 30). 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 supporting 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. at the grassland/meadowland ET site (fig. 30).

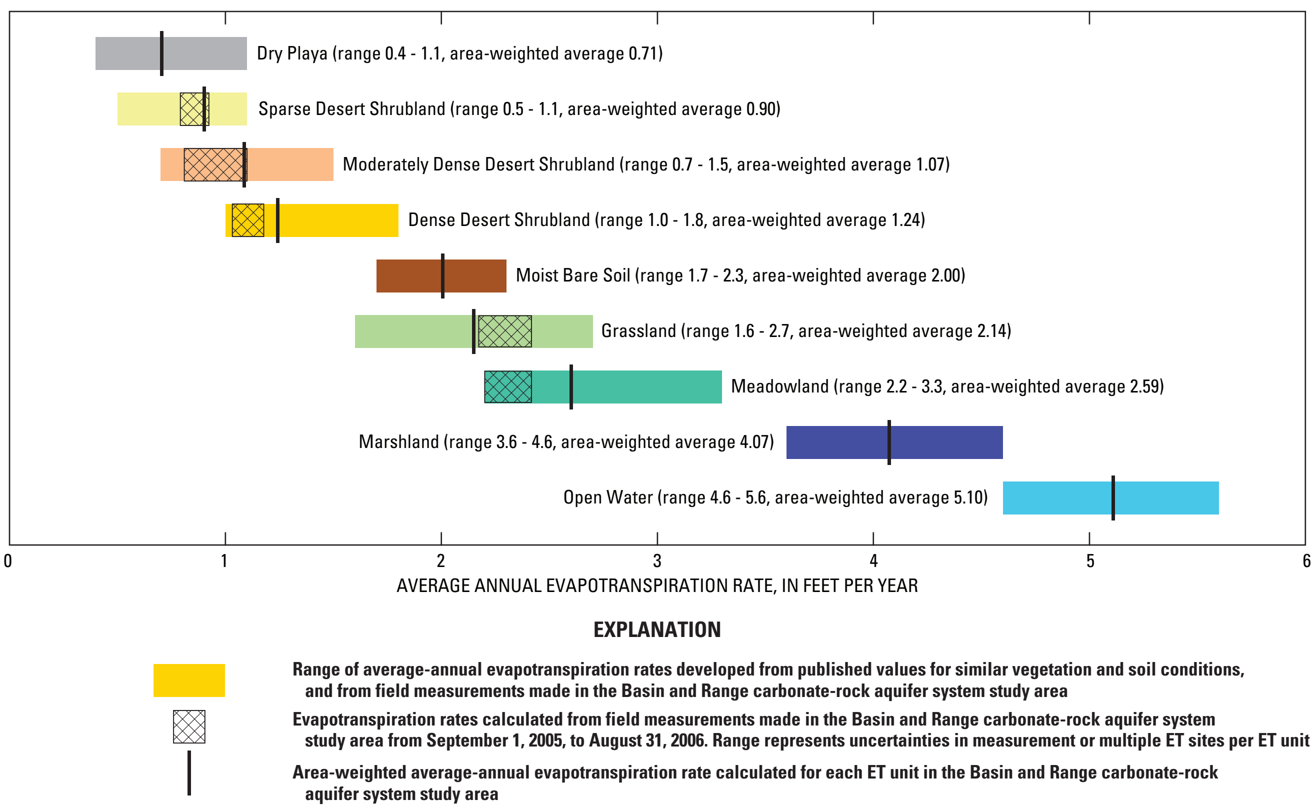


Figure 27. Estimated average annual ET-rate range for ET units identified, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

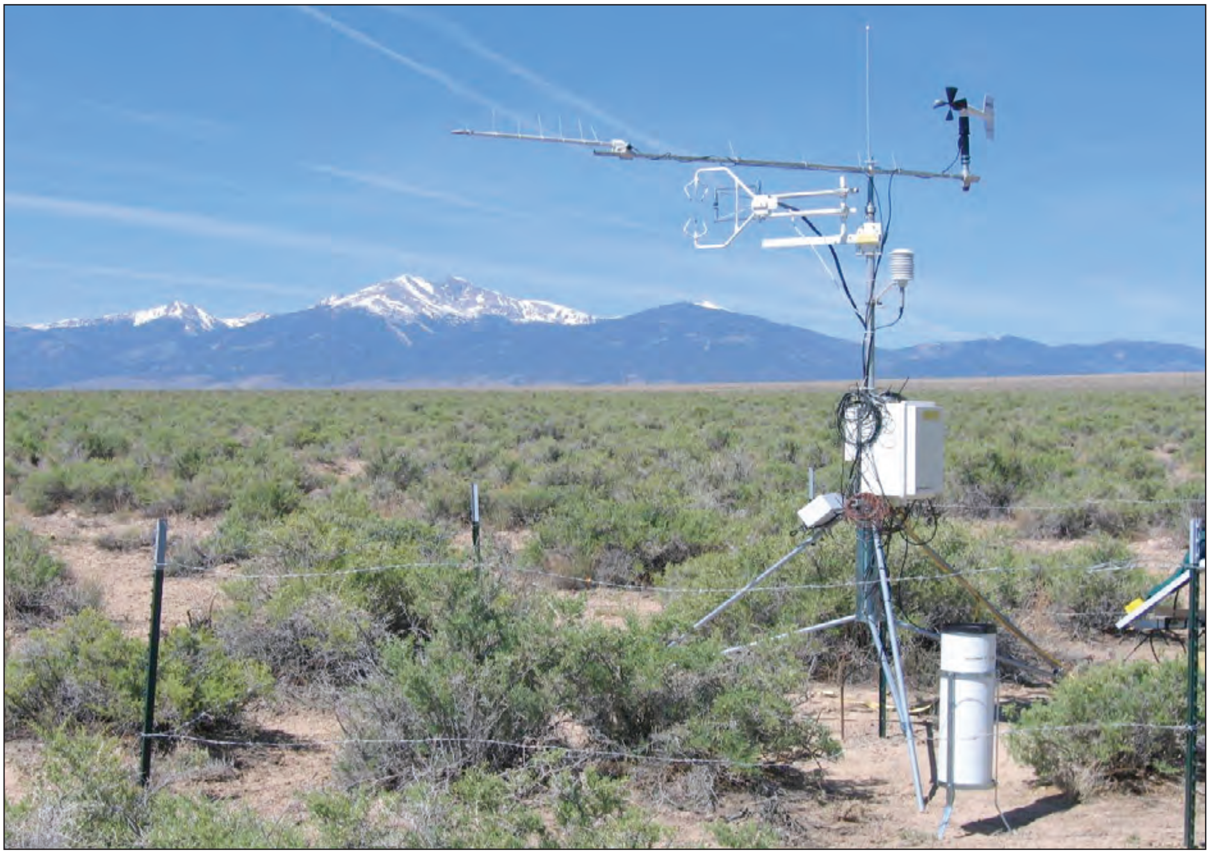


Figure 28. Eddy-correlation site used for measuring evapotranspiration in greasewood dominated shrubland in Snake Valley, Nevada. Northeast flank of southern Snake Range visible in background. Photograph taken by Michael T. Moreo, U.S. Geological Survey, June 1, 2006.

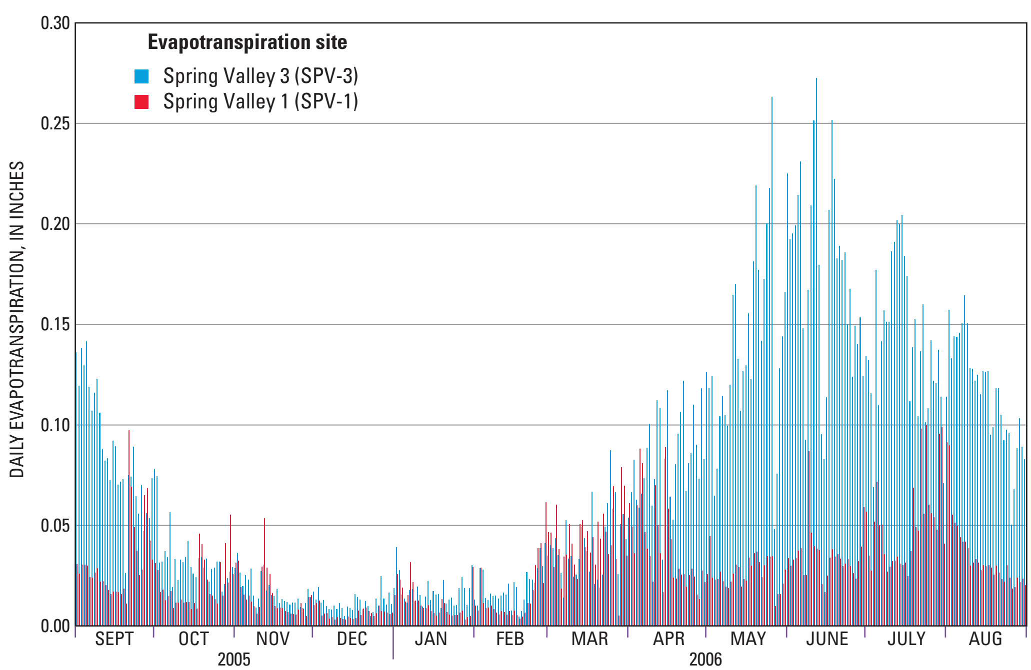


Figure 29. Daily ET from grassland/meadowland site (SPV-3) in Spring Valley, and a greasewood dominated shrubland site (SPV-1) also in Spring Valley, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, September 1, 2005, to August 31, 2006.

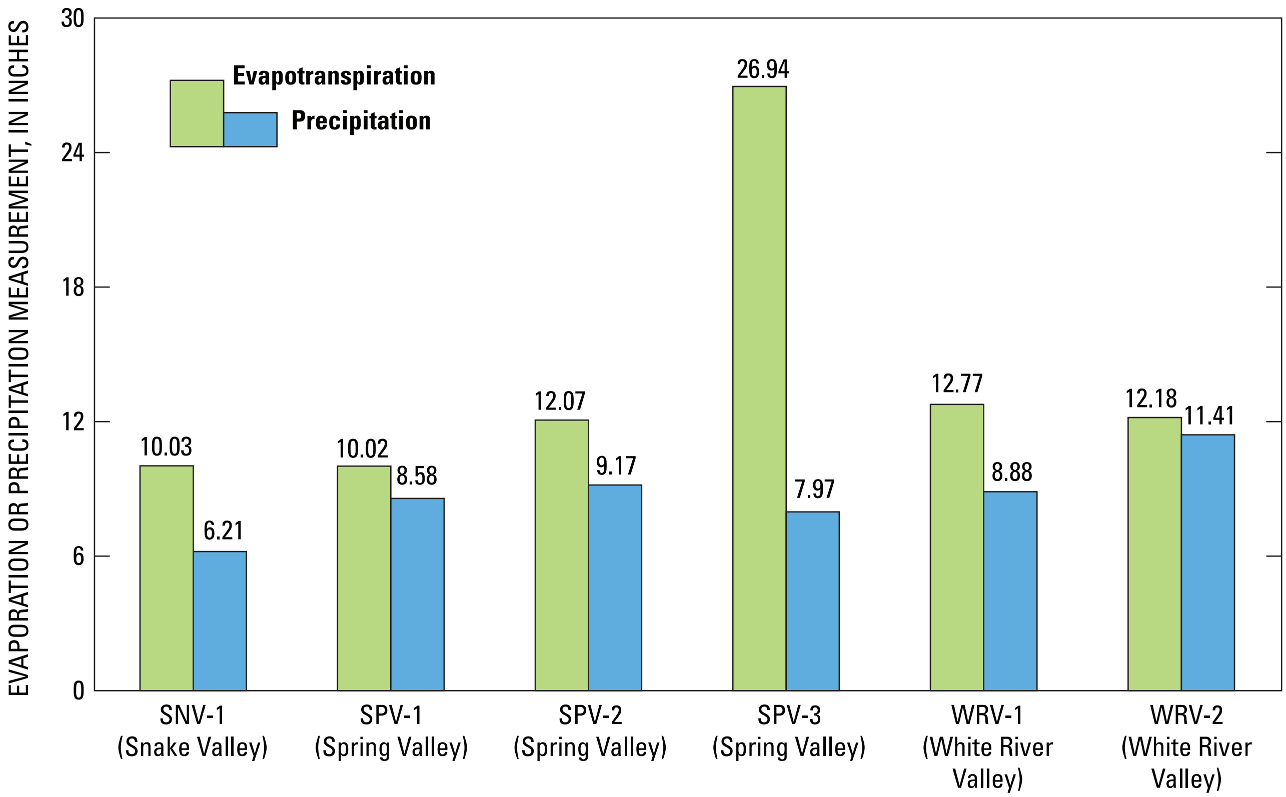


Figure 30. Total ET and precipitation measured at six ET sites in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, September 1, 2005, to August 31, 2006. All ET sites in greasewood dominated shrubland except SPV-3, which is in grassland/meadowland area.

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 assigned an ET rate that equaled the area-weighted average ET rate for all other phreatophyte units delineated in the study area.

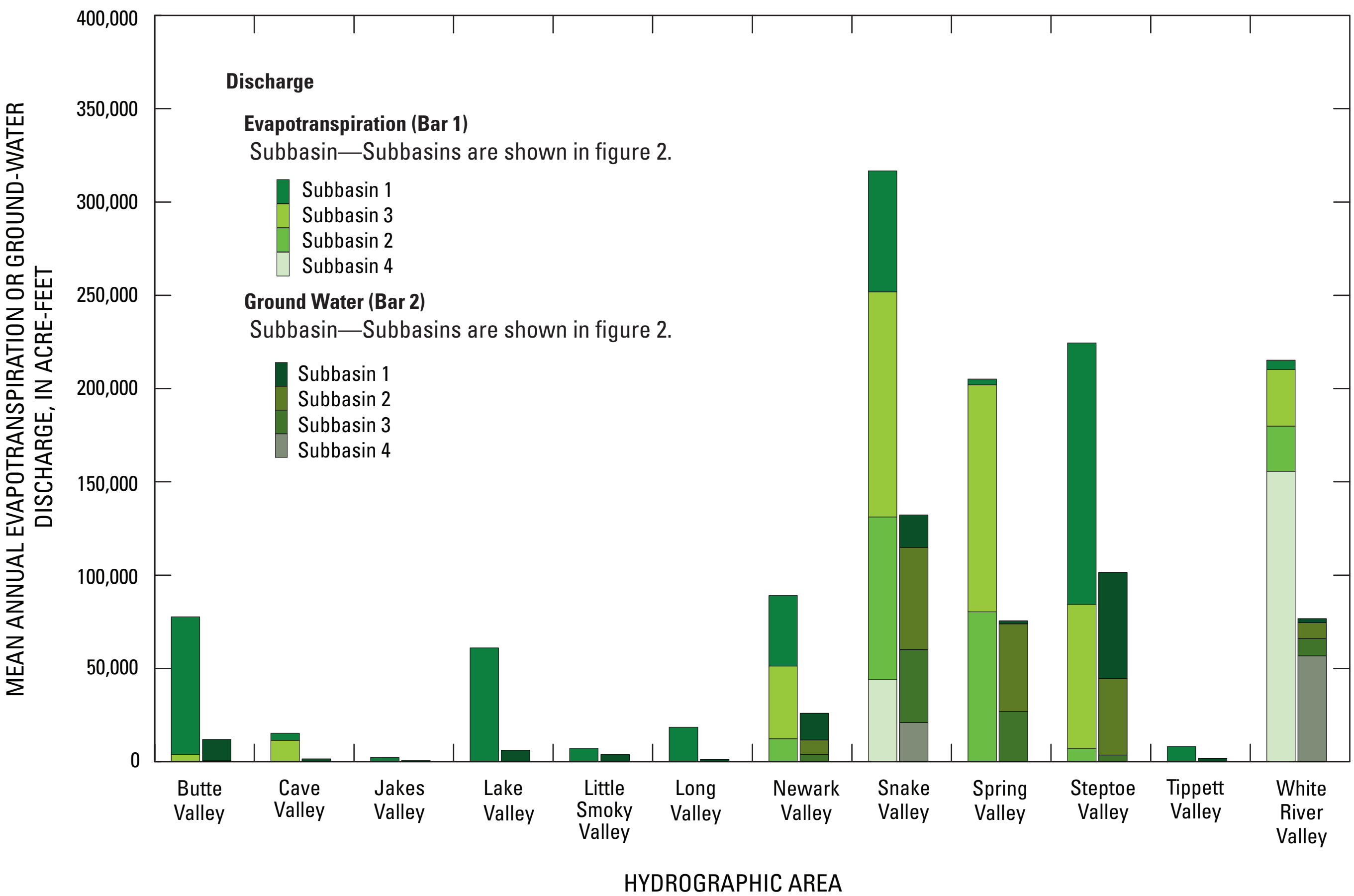


Figure 31. Estimates of mean annual evapotranspiration and ground-water discharge from hydrographic areas by subbasin, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Total ET estimated for a HA is the sum of its subbasin ET estimates (fig. 31). The subbasin ET estimate is the sum of ET estimates for each ET-unit. The ET estimate for an ET-unit is computed as the product of its ET rate and acreage (fig. 32). An ET-unit's ET rate is determined by linearly scaling the ET-rate range computed for the unit (fig. 27, appendix A). Scaling within the range was done using the average MSAVI value for the unit computed over the subbasin from 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. Details on the calculation and distribution of the MSAVI values used to scale ET rates are given in Smith and others (2007).

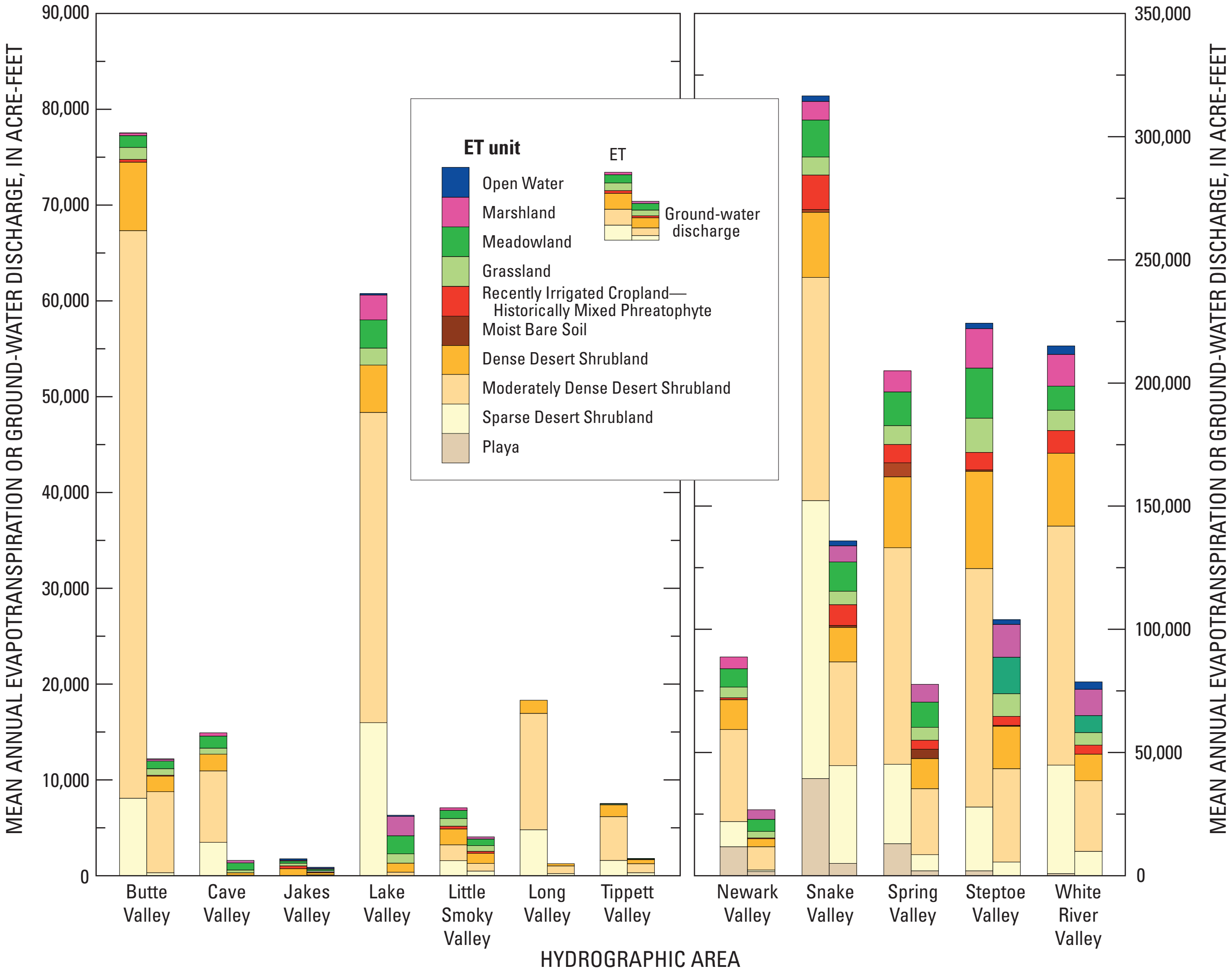
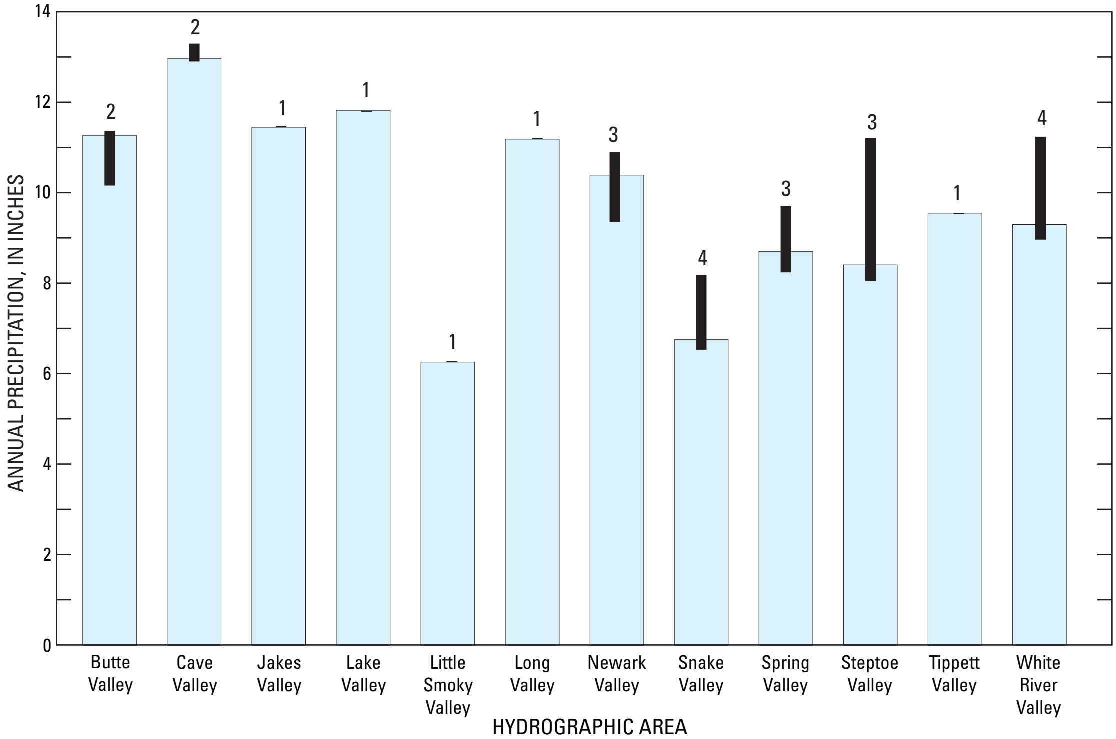


Figure 32. Estimates of mean annual evapotranspiration and ground-water discharge by ET unit from hydrographic areas, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

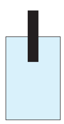
Mean Annual Ground-Water Discharge

Annual ground-water discharge from HAs is computed as the difference between annual ET and local precipitation. Precipitation falling directly on ground-water discharge areas and surface-water run-on (overland flow) to discharge areas contribute to ET. Most precipitation falling directly on areas of ground-water discharge ultimately is lost by local ET, and therefore, is assumed not to contribute to ground-water recharge. In addition, most if not all surface-water flow onto fine-grained playa sediments evaporates, and for the purpose of the water budget is assumed not to contribute to either ground-water recharge or discharge. 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 study 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



EXPLANATION



Blue bar represents average annual precipitation to discharge areas within hydrographic area, in inches. Black bars show variation in average annual precipitation by subbasins. Number above the black bar is number of subbasins identified in hydrographic area. Subbasins shown in figure 2.

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

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. These effects also contribute to higher annual precipitation in the southern subbasins of Snake and Steptoe Valleys.

Annual ground-water discharge from HAs is 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-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 local 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:

- Contributions to ET other than by regional ground water can be removed by subtracting direct precipitation from the ET estimate,
- Regional ground water is evaporated and transpired only from surfaces delineated as discharge areas,
- Spatial variation in ET from discharge areas of the study area can be adequately described using 10 ET units,
- ET rates assigned to ET units adequately represent the average for that unit,
- Estimates of mean annual precipitation used to compute mean annual ground-water discharge rates represent true long-term averages, and
- 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 and was further evaluated in a detailed statistical analysis by Zhu and others (2007). Their analysis computes uncertainty stochastically using Monte Carlo simulations and compares differences in the uncertainty range computed for HA and subbasin discharge estimates. Standard deviations were used to define the uncertainty ranges shown in figure 34 for each HA discharge estimates.

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 for purposes of this report. 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 by Zhu and others (2007) to quantify the uncertainty associated with of the discharge estimates given in this report.

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

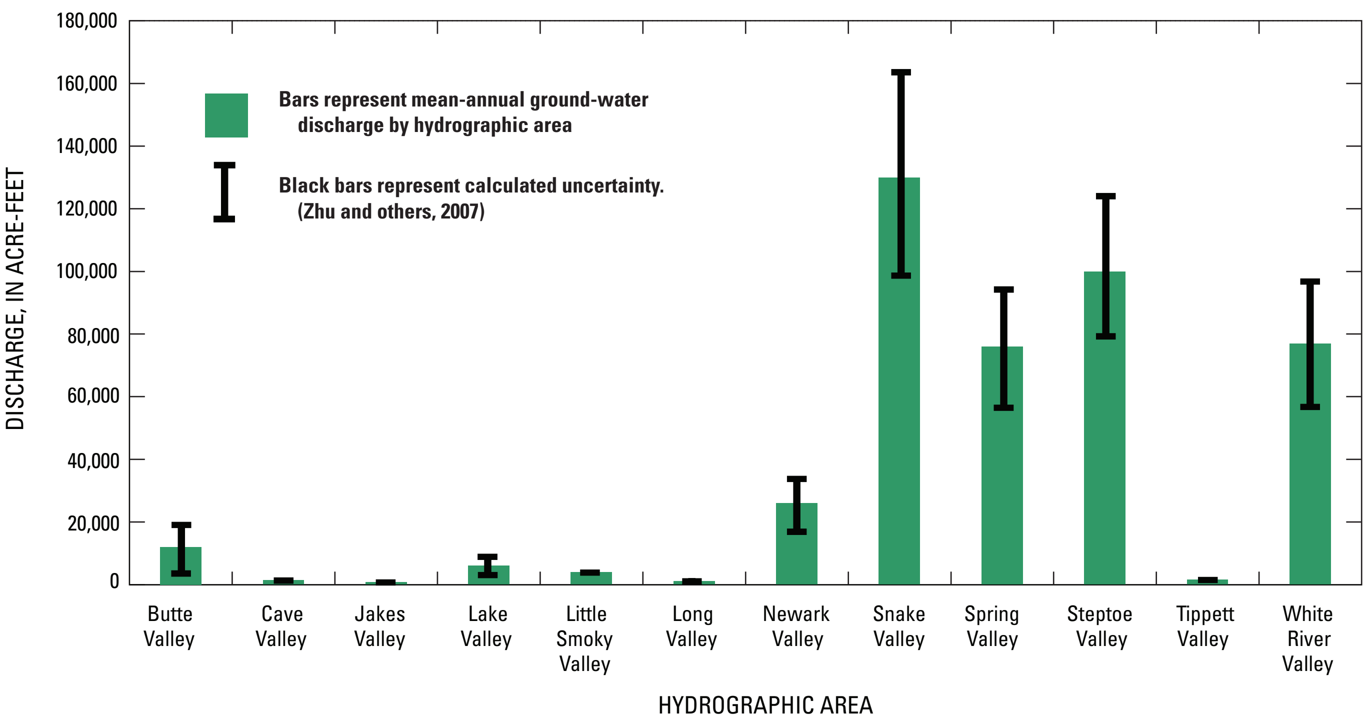


Figure 34. Uncertainty in ground-water discharge estimates by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

imagery acquired near the summer solstice 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 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 represent average annual values and the average value over an entire ET unit would be improved with additional temporal and spatial data.

Estimates of average annual ground-water discharge are intended to account only for that ground water being lost to the atmosphere by ET, and are not inclusive of any springflow that leaves the discharge area by means other then by ET, or any 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. Because ground-water discharge is estimated from ET, annual estimates of HA and subbasin ground-water discharge presented in this report include the surface runoff and streamflow that enters the ground-water flow system from areas outside of a discharge area.

Water Use

Water is used for farming, mining, ranching, light industry, and domestic and public supply and 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 (89 percent of the total water demand), is estimated for 2005 on the basis of irrigated acreage delineated from multi-spectral satellite imagery and crop-application rates developed from climate data and known crop requirements. Estimates of non-irrigation water use 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 at 126,000 acre-ft in 2005 (appendix A). 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 (fig. 35). Lake, Snake, Spring, Steptoe, and White River Valleys account for about 89 percent of the total water demand from the study area. Public supply, domestic, mining, and stock use were significant only in Steptoe Valley, where these uses accounted for about one-half of the total water demand. Combined stock and domestic uses accounted for less than 2 percent of total water demand.

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). More than 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 9 percent in 2005 (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.

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, 1993). A range for the likely application rate of each field was developed from the equation:

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

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,

Kc is crop coefficient, dimensionless

Pe is effective precipitation, in feet per year; and

Ep is project application efficiency, dimensionless.

Crop consumptive use is estimated as the product of reference crop ET and a 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

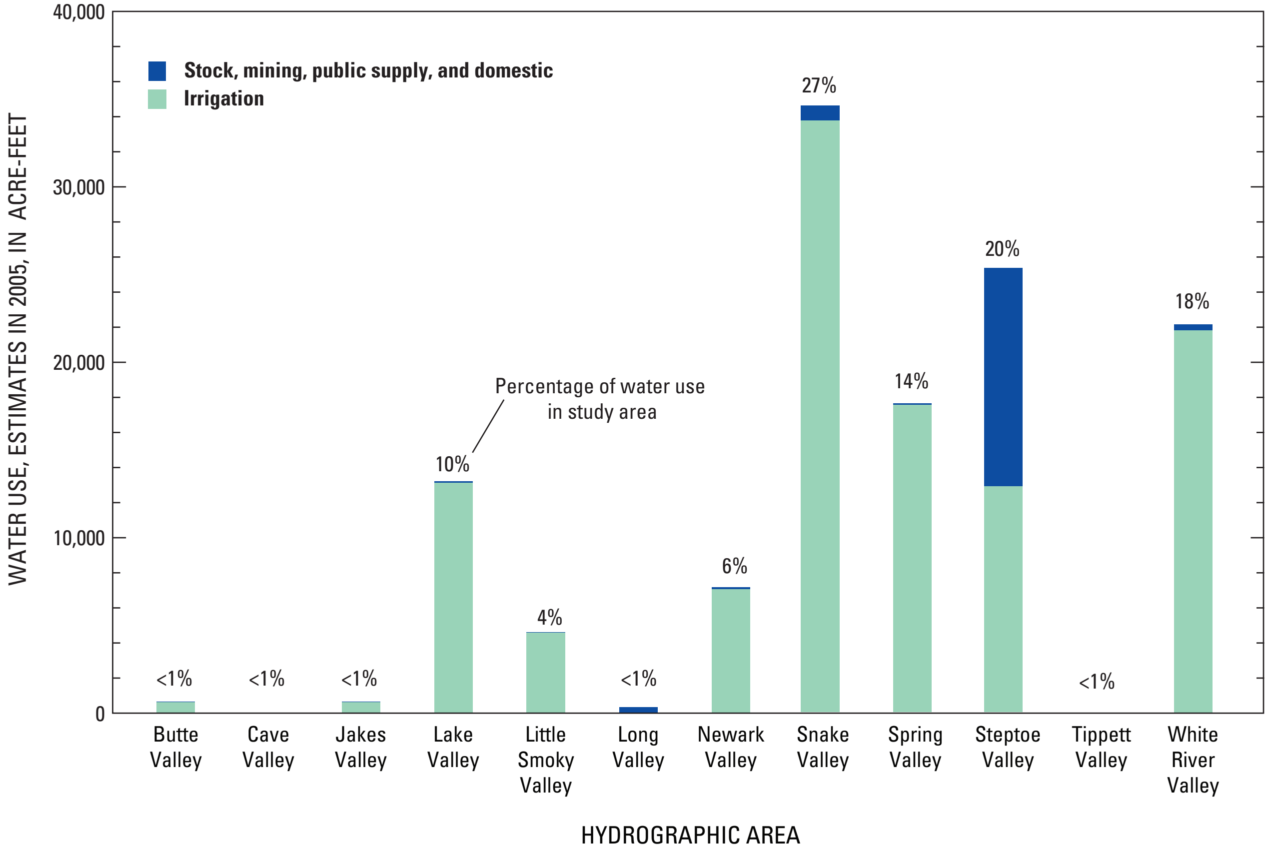
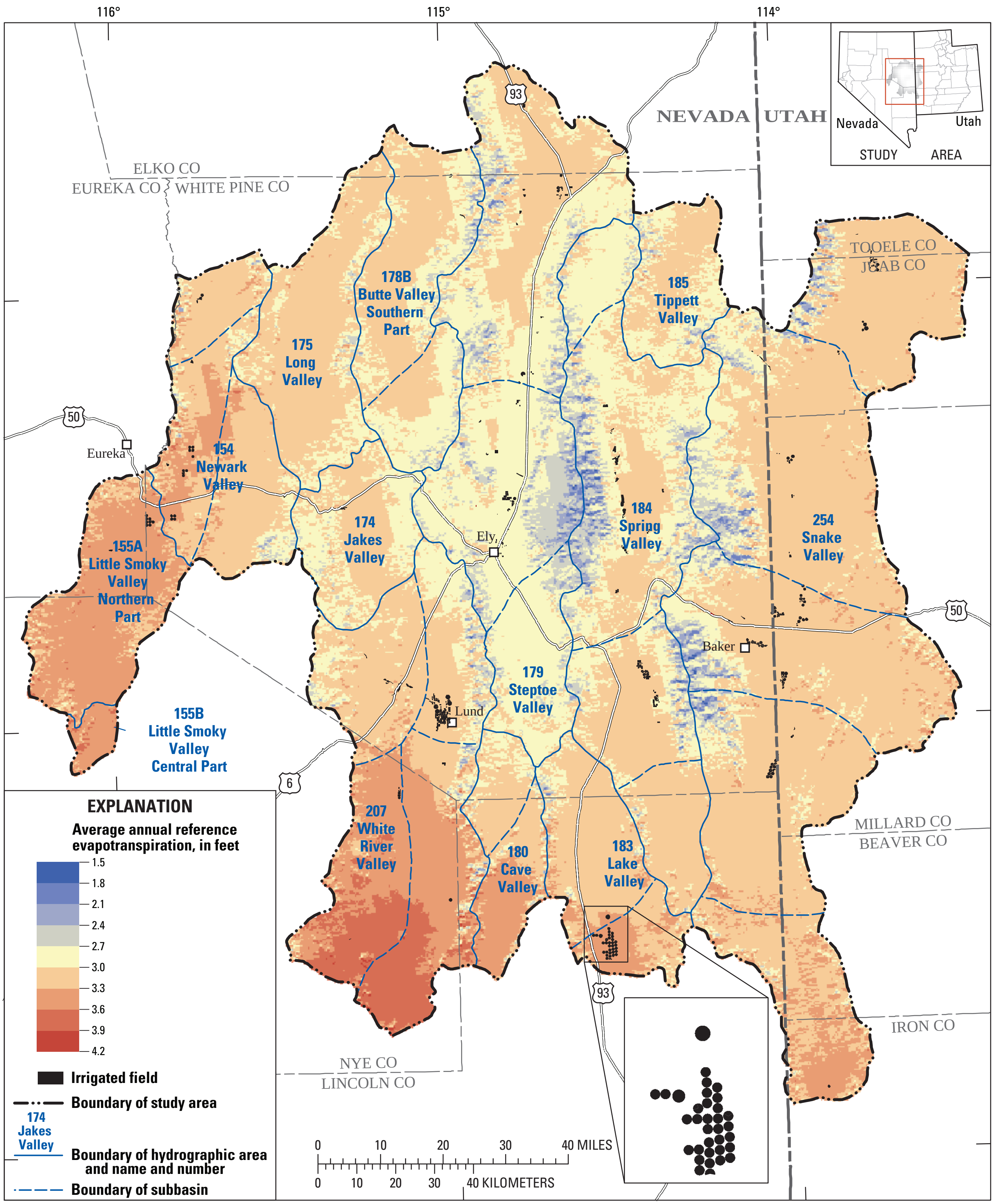


Figure 35. Water-use estimates by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2005.

crop coefficient method, are used extensively throughout the world (http://www.fao.org/nr/water/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 for Nevada using more than 240 weather sites operated by the California Irrigation Management Information System (CIMIS; http://wwwcimis.water.ca.gov) and 26 weather sites operated in Arizona by the Arizona Meteorological Network (AZMET; http://ag.arizona.edu/azmet/) (Flint and Flint, 2007). The standardized Penman-Monteith reference equation is used by CIMIS to calculate *ETo* (Allen and others, 1998, 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; Welborn and Moreo, 2007).

Kc relates crop consumptive use to the *ETo* rate, and depends on the growth and development of specific crops. CIMIS has developed *Kc* values specifically for calculating crop consumptive use as described above. For example, the average *Kc* is 1 for alfalfa during the growing season. Estimates of average crop consumptive use (ET*c*) for each HA, ranging from 2.78 to 3.08 ft/yr, are in agreement with measured consumptive-use rates for alfalfa and pastureland given in Maurer and others (2006) for a similar climate. Alfalfa and other hay production accounts for about 88 percent of the irrigated acreage in the study area. Pastureland accounts for about 10 percent, and corn, potatoes, and small grains account for about 2 percent of the total acreage irrigated in 2005.

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* (U.S. Department of Agriculture, 1993). *Pe* is estimated to be 70 percent of the average annual precipitation and 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 could be used for crop growth during the next growing season (U.S. Department of Agriculture, 1993). About two-thirds of the average annual precipitation falls during the growing season (Flint and Flint, 2007; Welborn and Moreo, 2007).



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.

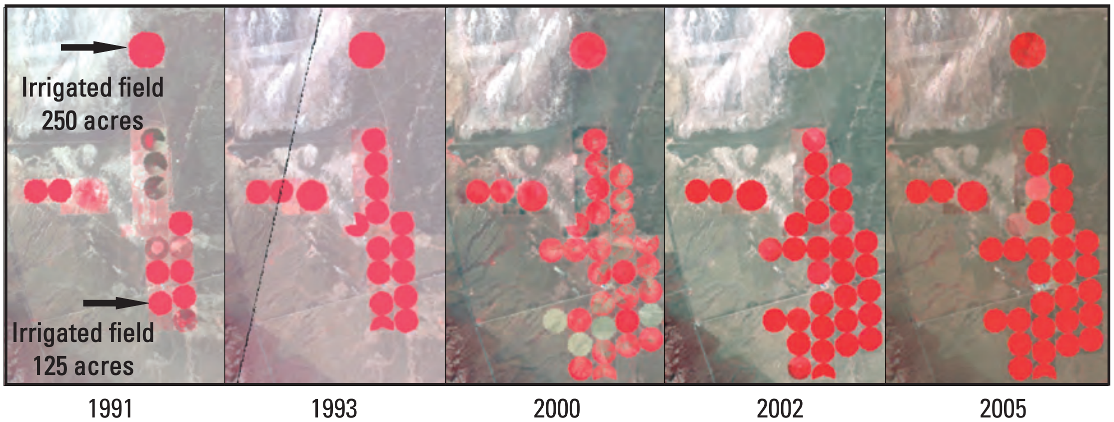


Figure 37. Thematic mapper imagery showing irrigated fields in Lake Valley, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah. See inset in figure 36.

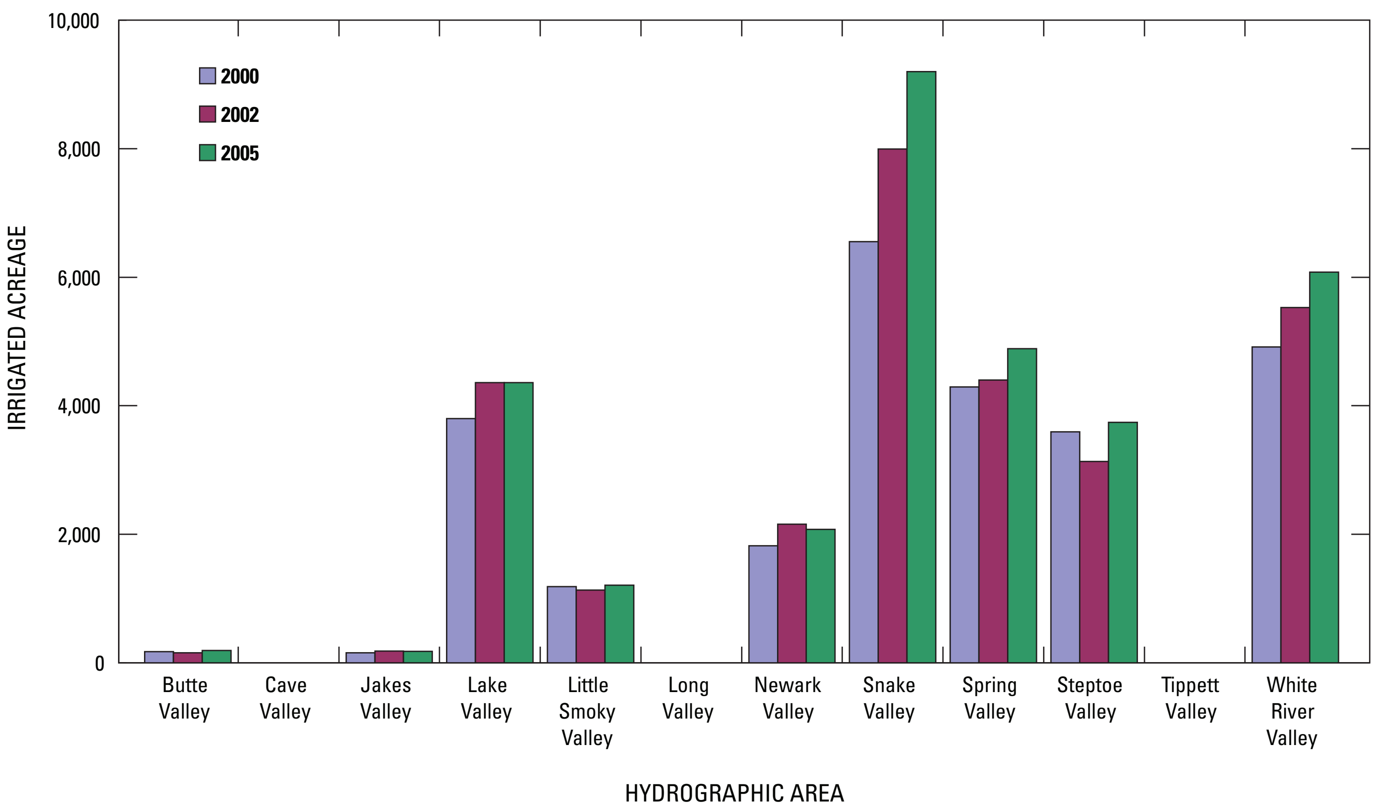


Figure 38. Estimates of irrigated acreage by hydrographic area, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah, 2000, 2002, and 2005.

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 into the atmosphere, 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 is highly dependent on 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). After 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. Fifty-three percent of irrigation applied in the study area is by center pivot sprinklers, 25 percent by fixed and periodically moved sprinklers, and 22 percent by flood irrigation (Welborn and Moreo, 2007).

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 (appendix A). Higher application rates reflect higher *ETo* rates, lower *Pe* rates, less efficient irrigation systems, or some combination thereof. The highest irrigation water use estimated for 2005 was in Snake Valley at 34,000 acre-ft (fig. 35). The uncertainty associated with irrigation water-use estimates is about plus or minus 15 percent based on the range of irrigation system efficiencies.

Irrigation return flow is that portion of the applied water that percolates beneath the root zone and ultimately returns

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.

to the ground-water flow system. Return flow is difficult to 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) report 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 example, applying equation 2 to a hypothetical 125-acre alfalfa field in Snake Valley irrigated with a center-pivot sprinkler system with $Ep = 0.75$, $ETc = 3.0$, and $Pe = 0.45$ ft results in an *AR* (application rate) of 3.4 ft. The product of irrigated acreage (125 acres) and *AR* (3.4 ft) is 425 acre-ft. If 375 acre-ft (125 acres × 3.0 ft) is required by the crop, then 425 acre-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 acre-ft – 375 acre-ft = 50 acre-ft), or 25 acre-ft, is the estimated return flow.

Ground-water pumped from wells and diverted from valley springs accounts for an estimated 70 percent of the water used for irrigation in 2005. The percentage is based primarily on field proximity to irrigation wells, springs, and natural and manmade drainage features, and where available—NDWR crop inventories (Welborn and Moreo, 2007). Perennial and intermittent streams sustained by upland springflow and above-average snowmelt account for the remaining 30 percent of the irrigation water applied in 2005.

Non-Irrigation Water Use

Public supply, self-supplied domestic, stock, and mining water use account for only about 11 percent of total water demand (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 include water supplied by public water purveyors to households, commercial establishments, prisons, schools, and campgrounds. Of the 9,637 people estimated to live in the study area (GeoLytics, 2001), an estimated 5,825 permanent residents and an unspecified non-resident population (primarily tourists) were served by public supply (appendix A). Community populations served by public water supply systems were subtracted from the total population and the remaining population of 3,812 people was assumed to used a self-supplied domestic water system. The self-supplied domestic use was estimated using this population (3,812 people) and a water-use coefficient of 300 gallons per person per day was applied (Nevada Department of Conservation and Natural Resources, 1999). Hydrographic areas with very small populations were assumed to use 10 acre-ft for domestic use (appendix A). Stock water use for most HAs was estimated as 0.32 percent of irrigation water use (Nevada Department of Conservation and Natural Resources, 1999): however, this estimate was modified to account for valleys having stock wells but no irrigation or total livestock populations (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 are less than the median value of the range (fig. 40 and table 6) (Maxey and Eakin, 1949; Eakin and Maxey, 1951; Eakin, 1960, 1961, 1962; Hood and Rush, 1965; Rush and Kazmi, 1965; Eakin, 1966; Eakin and others, 1967; Glancy, 1968; Nichols, 2000). The range in previous estimate values defined for Snake Valley is based on only two estimates (table 6). Differences in published discharge values primarily result 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 of the estimates in previous studies do not correct for precipitation and use total ET as their reported estimate of ground-water discharge, and others include pumping in their estimate of total ground-water discharge.

Previous investigations estimated ground-water discharge from limited data and many estimates are not clearly defined. For example, early investigations estimated ground-water discharge in a basin (Maxey and Eakin, 1949) simply 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 less accurate micrometeorological equipment, few annual estimates of evapotranspiration, higher cost of satellite imagery, and less advanced remote-sensing technologies.

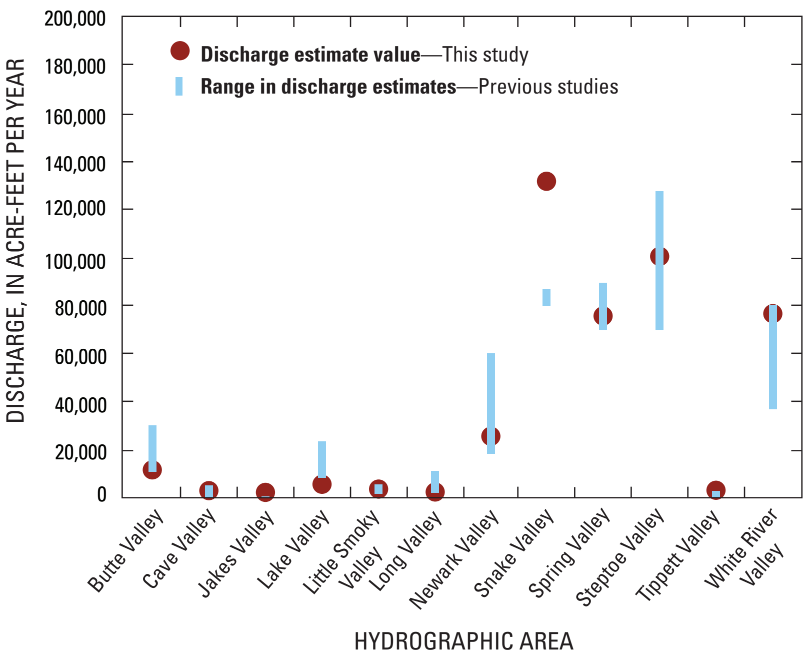


Figure 40. Comparison of pre-development ground-water discharge estimates, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

Annual ground-water discharge estimates were developed for this study using more advanced remote-sensing techniques for identifying and classifying vegetation, many more measurements of local ET and precipitation, and more robust and accurate micrometeorological equipment resulting in more accurate delineations of ground-water discharge areas and improved estimates of local ET rates. 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. Additionally, unlike for earlier results, the uncertainty of annual ground-water discharge can be estimated because the uncertainty associated with each component term can be better 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 recharge is high and relatively deep water levels and small or non-existent phreatophytic areas result in less local ET (Eakin, 1966; 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, the estimated average annual recharge exceeds the estimated ground-water discharge by 20 percent or more (tables 5 and 6). The high recharge in Steptoe Valley annually exceeds pre-development discharge by more than 50,000 acre-ft, the largest surplus of water for any HA. Large annual water surpluses 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, the annual recharge exceeds annual discharge in the remaining HAs, ranging from less than 1,000 to 18,000 acre-ft. Even though these surpluses are relatively small, the percent difference between recharge and discharge can be quite large. In Cave, Long, Jakes, and Tippett Valleys, the annual discharge is 20 percent or less than the annual recharge, indicating that most of the pre-development discharge exits these valleys as subsurface outflow to adjacent valleys.

In contrast to recharge-dominated HAs, pre-development discharge annually exceeds recharge in Newark, Snake, and White River Valleys. The average annual recharge is about 20 percent less than the average annual discharge in Newark and Snake Valleys. In White River Valley, the annual recharge is less than one-half of average annual discharge, resulting in an average annual water deficit of more than 40,000 acre-ft. This relatively large deficit 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 documented by Rosemary Carroll and Greg Pohll (Desert Research Institute, written commun., 2007) was used to evaluate a water budget for the study area that included intrabasin and interbasin ground-water flow. The model, which is described by Lundmark and others (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 conservative tracer. Deuterium values representing ground-water recharge and regional ground-water flow systems were based on existing and new data collected as part of the current study. These values were used as model input for recharge areas and to help calibrate the mixing model. A complete description of the spatial distribution and calculated average values for deuterium data model input and calibration can be found in Lundmark and others (2007).

Under pre-development conditions, the average annual recharge is greater than average annual discharge for 9 of the 12 HAs in the study area, indicating that a significant quantity of ground water must flow 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. Geologic barriers to ground-water flow are shown 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 that ground-water flows 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-feet per year or less, and are considered to be within the uncertainty associated with interbasin flow rates. The details of the model are described by Lundmark and others (2007).

Model estimated interbasin flow rates and the general direction of flow across HA boundary segments are shown in figure 41. Butte, Cave, Little Smoky, Long, and Steptoe Valleys receive no ground-water inflow; Newark and Tippett Valleys receive only small amounts of ground-water inflow. The remaining five HAs, Jakes, White River, Lake, Spring, and Snake, receive ground-water inflow from adjacent HAs ranging from 20,000 to 80,000 acre-ft/yr. Ground-water flow out of the study-area boundary includes about 7,000 acre-ft/yr toward the north, from Steptoe Valley to Goshute Valley, and about 40,000 acre-ft/yr toward the northeast from Tippett and Snake Valleys to the Great Salt Lake Desert regional flow system. About 39,000 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 flow system. About 8,000 acre-ft/yr exits the northwestern 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 measured 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 information from multiple aspects of the study, including recharge and discharge estimates, and water-level data, each with its own inherent uncertainty.

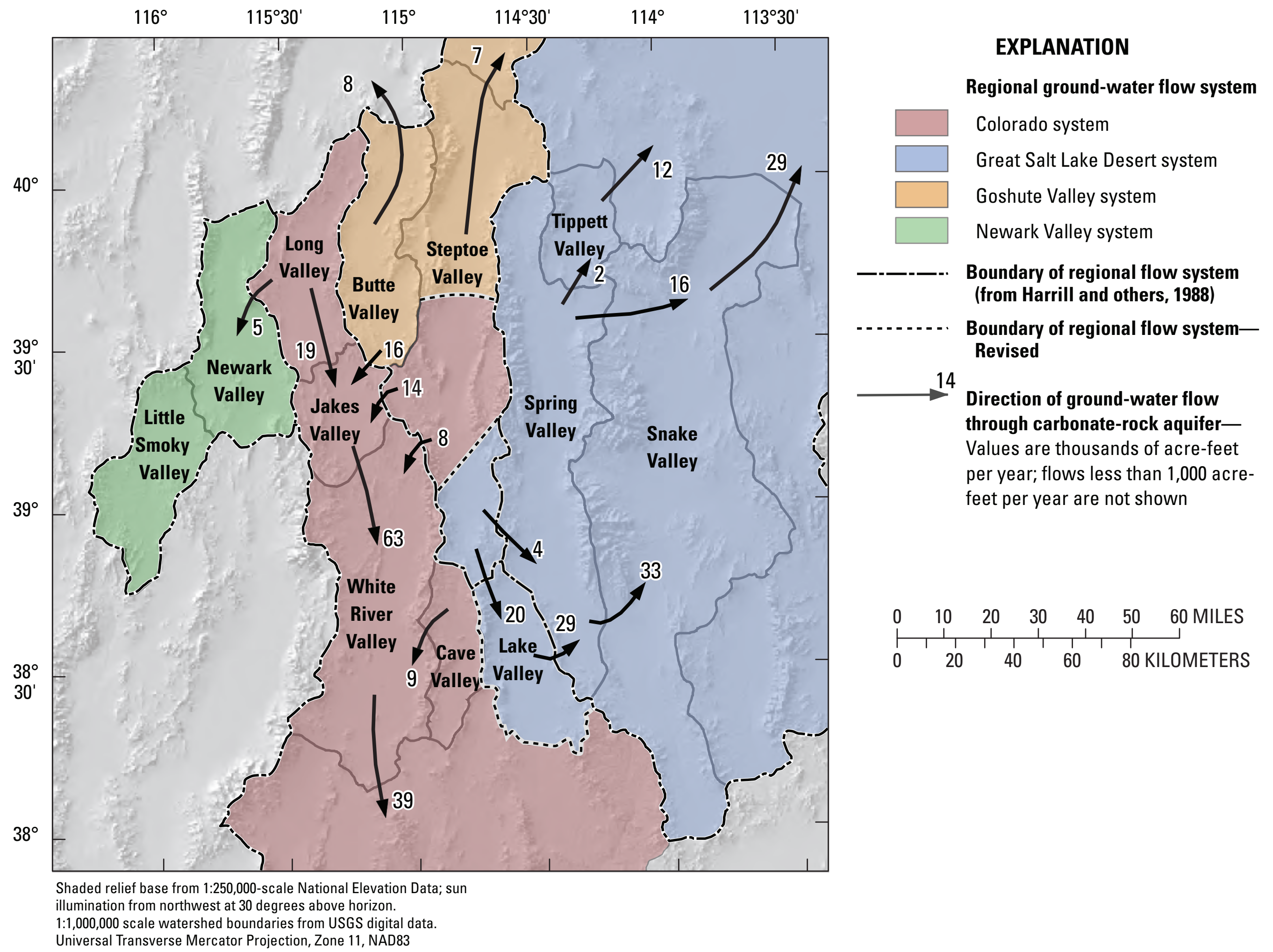


Figure 41. Regional ground-water flow through the Colorado, Great Salt Lake Desert, and other regional flow systems, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

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 or refute 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:

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

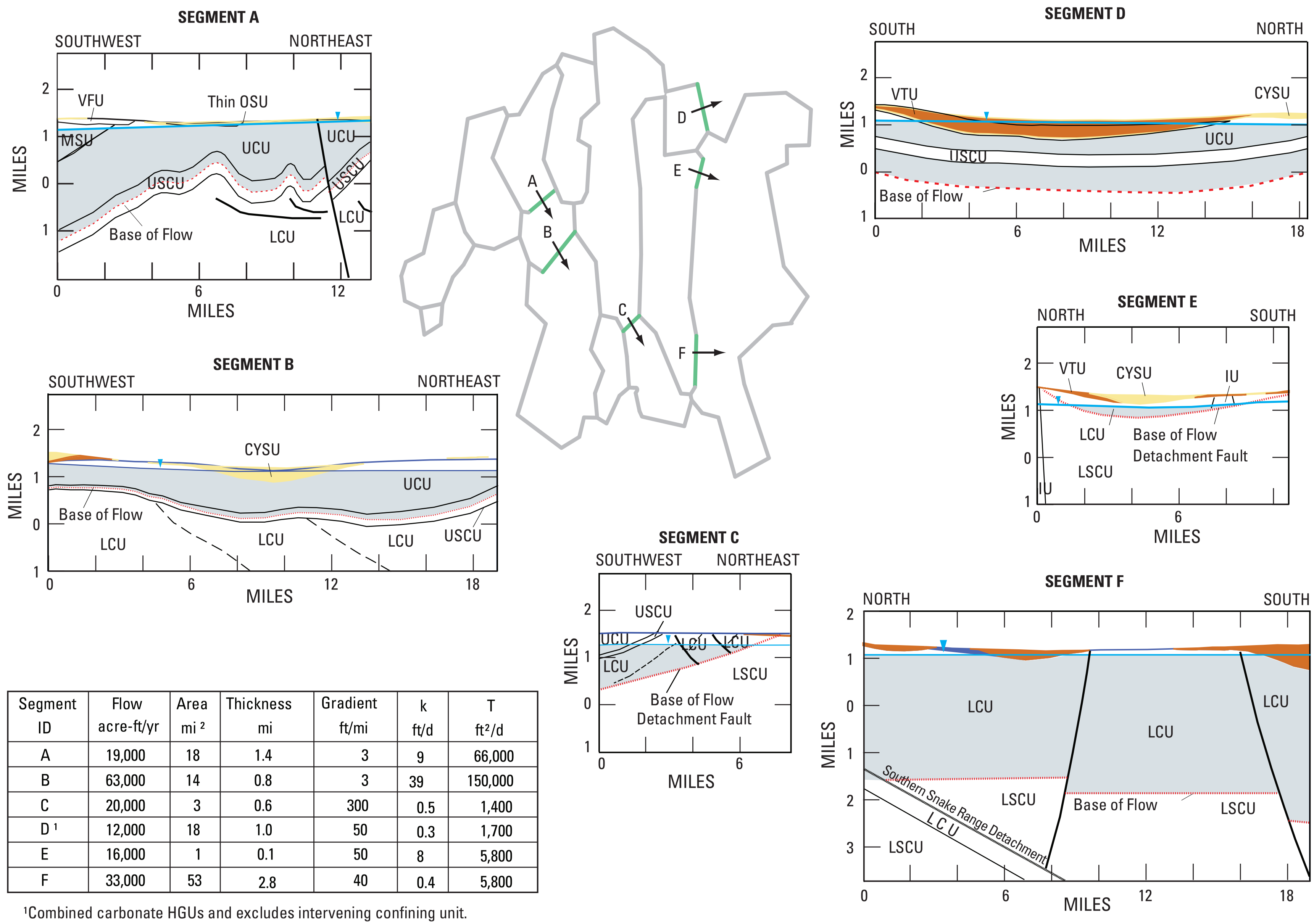
where

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 interbasin ground-water flow, in cubic feet per day,
i is the hydraulic gradient, in foot per foot, and
W is the effective width of the aquifer units, in feet

Transmissivity was estimated for six HA boundary segments and compared directly to aquifer test results given in Dettinger and others (1995). 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 42. Using equation 3, interbasin flow estimates from the water-accounting model (fig. 41) 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 from cross sections extracted from a three-dimensional hydrogeologic framework model developed for this study (fig. 43).

Transmissivities were estimated for two HA boundary segments in the western half of the study area (segments A and B, fig. 42). Aquifer units beneath the shared boundary of Jakes and Long Valleys (segment A, fig. 42) and the shared boundary of Jakes and White River Valleys (segment B, fig. 42) 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 66,000 and 150,000 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.

Transmissivities were estimated for four HA boundary segments in the eastern half of the study area (segments C–F, fig. 42). 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 5,800 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).



Segment ID	Flow acre-ft/yr	Area mi²	Thickness mi	Gradient ft/mi	k ft/d	T ft²/d
A	19,000	18	1.4	3	9	66,000
B	63,000	14	0.8	3	39	150,000
C	20,000	3	0.6	300	0.5	1,400
D[1]	12,000	18	1.0	50	0.3	1,700
E	16,000	1	0.1	50	8	5,800
F	33,000	53	2.8	40	0.4	5,800

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

Figure 42. Cross sections used to estimate transmissivities of hydrogeologic units.

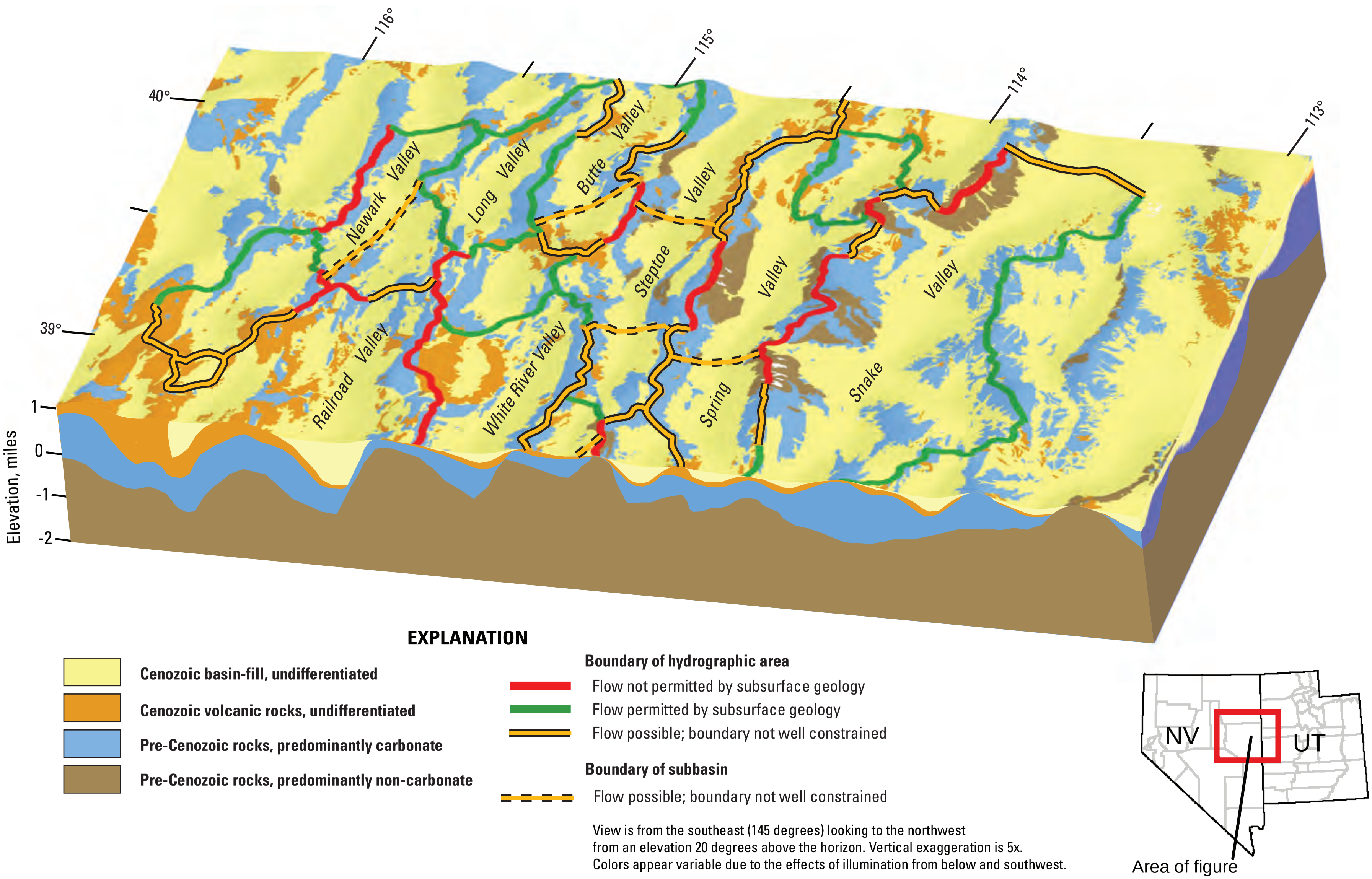


Figure 43. Perspective view of simplified three-dimensional geologic model of the study area.

The calculated transmissivities for the various boundary segments can be compared with values for carbonate rocks presented in Dettinger and others (1995) (fig. 44). 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. 44). 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.

Geochemical Modeling

Geochemical modeling was applied to support other evidence of interbasin and intrabasin ground-water flow in the study area. 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. 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 and are limited by knowledge of minerals and gases present in the aquifer and available geochemical data along potential flow paths (Hershey and others, 2007).

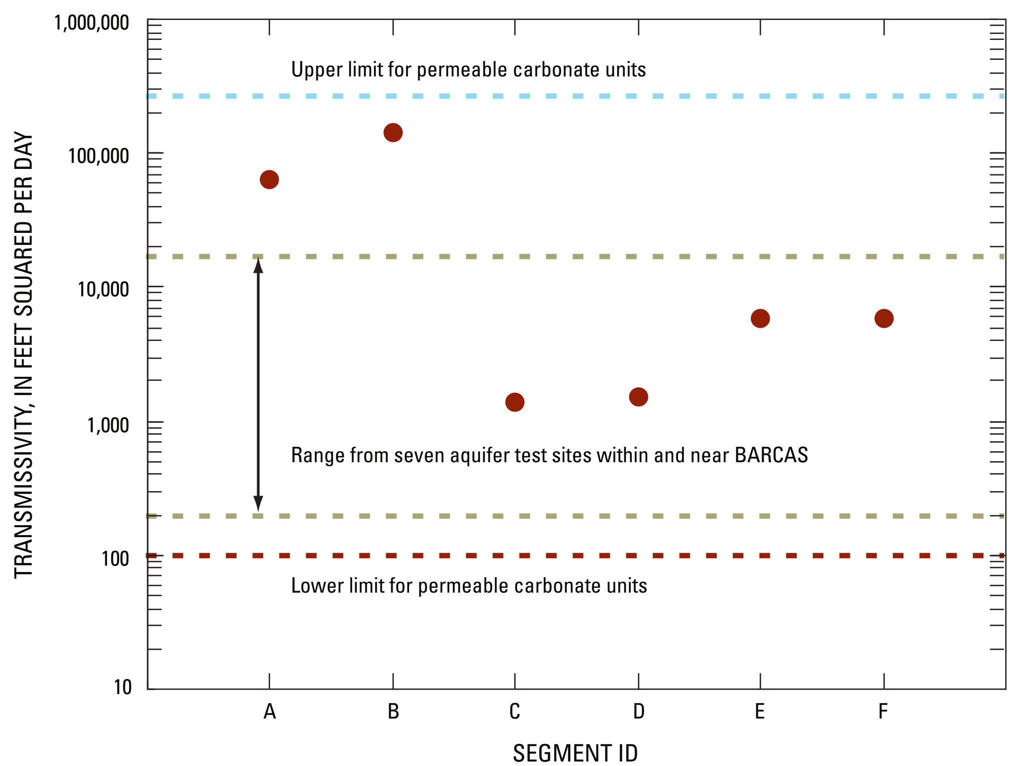


Figure 44. Transmissivity estimates for the boundary segments and published ranges in the carbonate-rock province. Upper and lower limits are based on data in Dettinger and others (1995). Boundary segments are shown in figure 42.

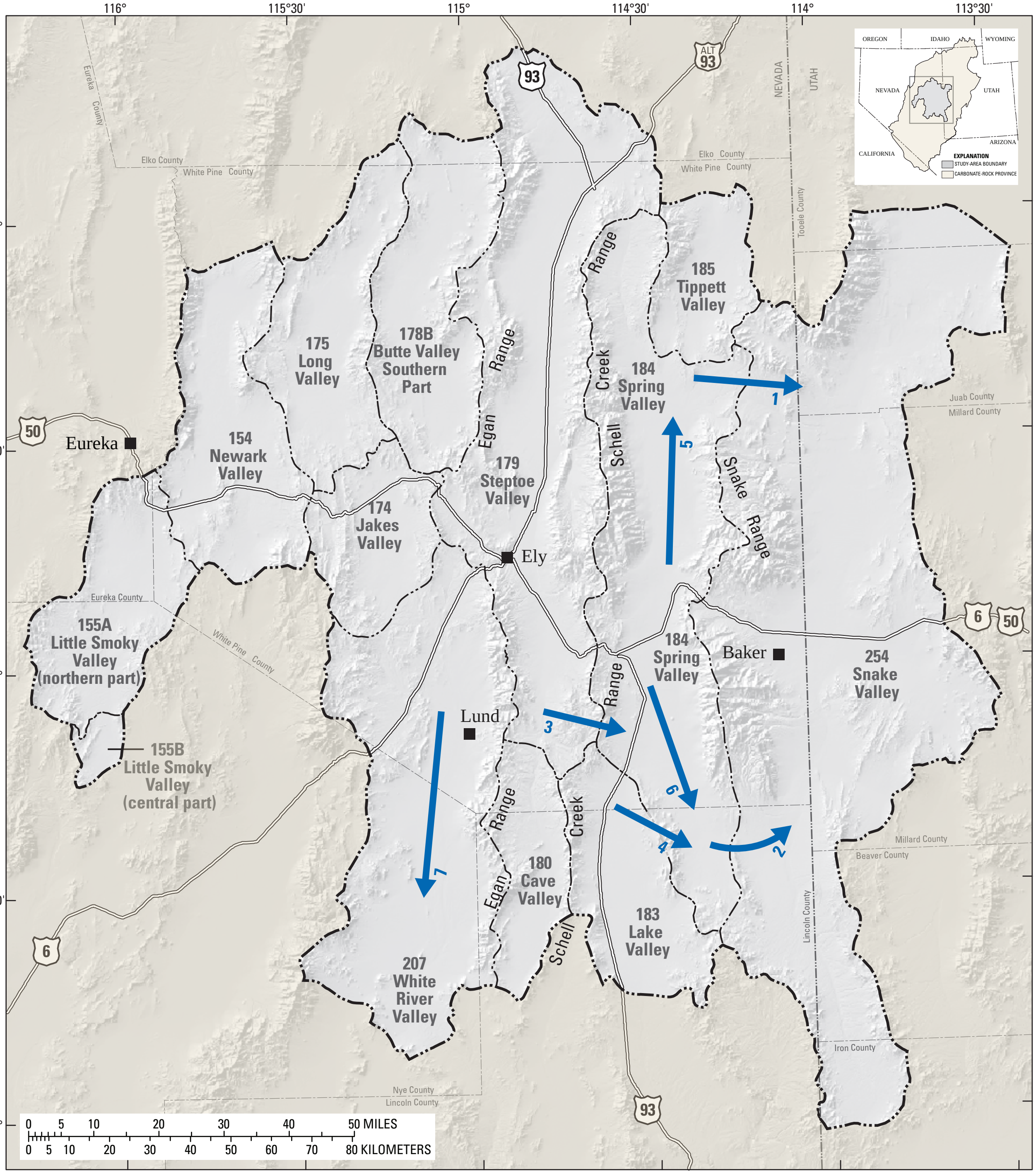
Ground-water flow paths that had reasonable major-ion and conservative tracer mixing relations were modeled geochemically using the computer program NETPATH (Plummer and others, 1994). The reasonableness of major-ion and conservative tracer mixing was assessed graphically (Hershey and others, 2007). NETPATH was used to interpret net water-rock mass-balance reactions between initial and final water compositions along a proposed ground-water flow path. Geochemical modeling results including ratios of initial and recharge waters (given as a range in percent) are summarized in table 8. Numerous valid water-rock reaction models for many of the flow paths were possible; for example, percentage of initial and recharge waters used for the two model evaluations along the flow path from northern Spring Valley to northern Snake Valley was 30 and 70 percent in the first evaluation (upper mixture, table 8), and 0 and 100 percent in the second evaluation (lower mixture, table 8). "Initial water" is water in springs or wells found at the beginning of the main flow path; "recharge water" is water from subsurface inflow of higher altitude springs tributary to the main flow path that potentially can be mixed with initial water. Details on chemical sampling, geochemical data, and results of NETPATH model evaluations on geochemical reactions and calculated travel times are provided in Hershey and others (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 (fig. 45). 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 (table 8). 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. Geochemical model could not be validated for other basins in the BARCAS study area due to lack of available chemical and isotopic data (Hershey and others, 2007). All geochemical models supporting ground-water flow across HA boundaries required some portion of local recharge along the flow paths. A detailed description of NETPATH-modeling results, including associated ground-water flow travel times and velocities can be found in Hershey and others (2007).

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

[**Flow path No.** matches corresponding number in figure 45. **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]

Flow path location and sites				Geochemical model – mixtures of water (percent)	Geochemical model results
Flow path No.	Initial	Final	Boundary or divide	Initial – Recharge	
1	Northern Spring Valley	Northern Snake Valley	HA	0 – 100 30 – 70	Supports ground-water flow path
2	Southern Spring Valley	Southern Snake Valley	HA	0 – 100 100 – 0	Supports ground-water flow path
3	Southern Steptoe Valley	Southern Spring Valley	HA	70 – 30 100 – 0	Supports ground-water flow path
4	Lake Valley	Southern Spring Valley	HA	95 – 5 100 – 0	Supports ground-water flow path
5	Southern part of northern Spring Valley	Northern part of northern Spring Valley	IB	0 – 100 60 – 40	Supports ground-water flow path
6	Central Spring Valley	Southern Spring Valley	IB	20 – 80 40 – 60	Supports ground-water flow path
7	Central White River Valley	Southern White River Valley	IB	40 – 60 60 – 40	Supports ground-water flow path



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**

Ground-water flow path—Blue number matches corresponding number in table 8

Figure 45. Ground-water flow paths, Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

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. Harrill and others (1988) present reconnaissance-level estimates for all HAs in the study area. Their estimates were based on locations, volumes, and directions of interbasin flow compiled from reconnaissance reports that generally represent evaluations of single HAs.

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 also were evaluated by Thomas and others (2001) using a deuterium mass-balance model. The deuterium data used to construct their mass-balance models are primarily historical data from DRI and USGS reports and databases. Estimated recharge to a valley and inflow from adjacent valleys were validated using deuterium data from regional springs and wells. 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). The model developed by Thomas and others (2001) is similar to the accounting model used in the current study, in that both models are modified versions of a deuterium mass-balance model originally developed by Campana (1975). However, these two models differ in a number of aspects, including their conceptualization of the aquifer system, model input values of recharge and hydraulic head, and spatial distribution and concentrations of deuterium data used to calibrate each model.

Directions of interbasin flow presented in Harrill and others (1988), Nichols (2000), and Thomas and others (2001) are in general agreement. However, the magnitude of interbasin flow differs slightly between reports. 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.

In general, interbasin ground-water flow directions described in this report are similar to those reported in 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 gradients, and water-accounting modeling, some interbasin flow directions discussed in this report differ from previous studies (fig. 41). 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 postulated 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).

In most cases, BARCAS interbasin flow estimates are higher than previously reported with only a few estimates falling within the range defined by previous estimates in the study area (fig. 46). BARCAS inflow estimates are higher in Jakes, Lake, Tippett, 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 moderately higher in Lake and Jakes Valleys. In Butte, Long, and White River Valleys, the estimated outflows are within the range of published estimates, and in Snake and Cave Valleys, the outflows are lower than published estimates (fig. 46). A Monte Carlo uncertainty analysis of the interbasin inflows and outflows was conducted as part of the steady-state water-budget accounting modeling effort. Results of the analysis are based on thousands of simulations and identify the 95-percent confidence interval for net interbasin inflow to and outflow from each of the hydrographic areas as well as for flow between subbasins. The 95-percent confidence intervals (error bars) for net interbasin flow are shown in figure 46.

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.

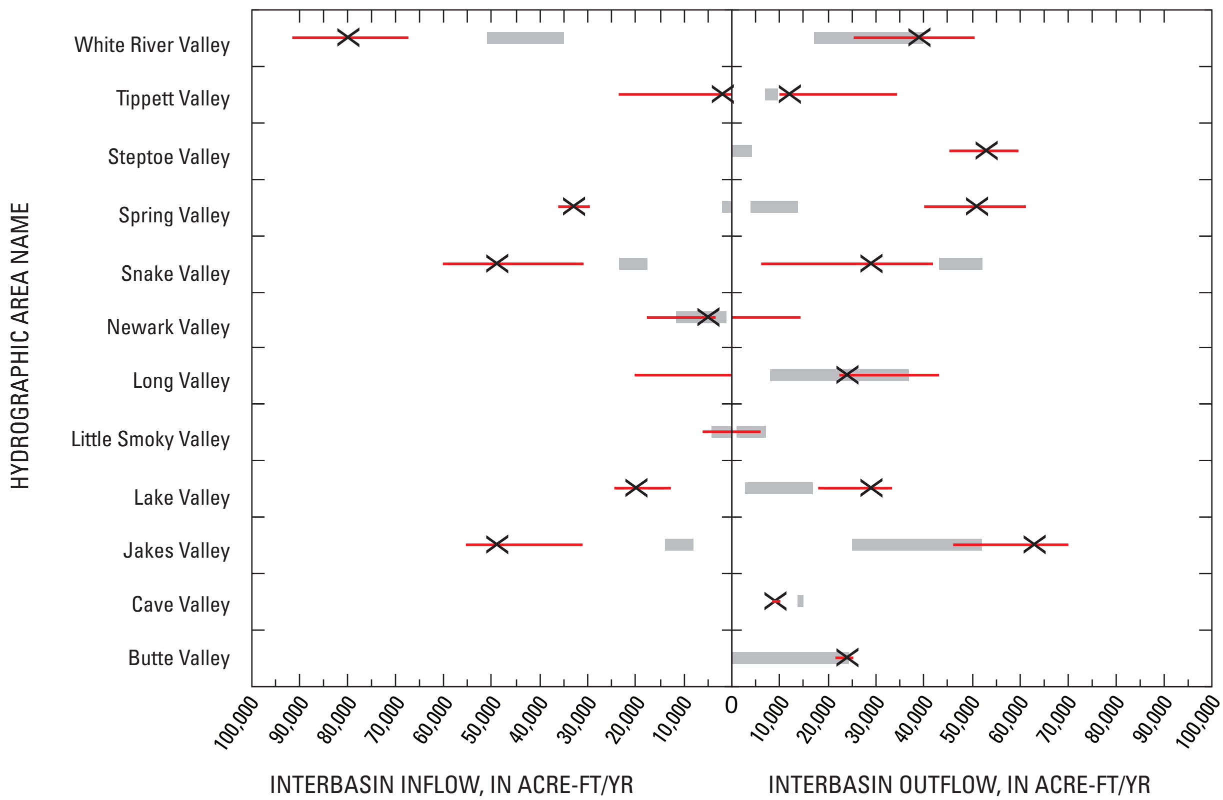
[Shading indicates flow through valley fill. Locations of hydrographic areas shown in figure 2]

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.



EXPLANATION

Interbasin flow—Missing symbols denote a value of zero

× Interbasin flow by hydrographic area—Estimated by BARCAS study

— Error bars show 95 percent confidence interval—Monte Carlo Uncertainty Analysis

▬ Range of interbasin flow by hydrographic area—Estimated by previous studies

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

Interbasin flow estimates may have a substantial amount of associated uncertainty because the accounting model integrates data from multiple aspects of the BARCAS study. This uncertainty is an accumulation of the associated uncertainty from the representation of the regional potentiometric surface, interpreted hydrogeologic boundary classifications, and recharge and discharge estimates. Moreover, deuterium data used in the model are relatively sparse in parts of the study area (Lundmark and others, 2007). The potential error and relative uncertainty has been evaluated and described in detail by Lundmark and others (2007).

Differences between estimates given in this study and in previous reports primarily are attributed to variations in the applied methods. For example, some previous estimates neglected hydraulic connections between adjacent HAs and, as a result, inflow from upgradient areas was not considered when constructing the water budgets. Additionally, discharge estimates from previous studies tend to equal the low end of the range for many of the individual HAs. This larger difference between recharge and discharge components suggests that larger volumes of ground water may be available for interbasin flow from recharge-dominated HAs to adjacent HAs. The greater estimated outflow from some HAs is considered reasonable because the study area is a primary recharge area for the Colorado, Great Salt Lake Desert, and Goshute Valley regional flow systems (pl. 3 and fig. 41).

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 acre-ft (appendix A). 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 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 (fig. 41). 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 (fig. 41).

The net amount of regional ground water removed from the study area was estimated to evaluate the significance of the ground water withdrawn to ground water discharged under pre-development conditions. Net ground-water use 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 use estimated for each of the HAs in the study area ranged 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 (figs. 35 and 47). Net regional ground-water pumpage is greater only in Lake Valley than the estimated average annual ground-water discharge under pre-development conditions. Net regional ground-water use for the entire study area is estimated at about 80,000 acre-ft, or about 60 percent of the 2005 water-use estimate.

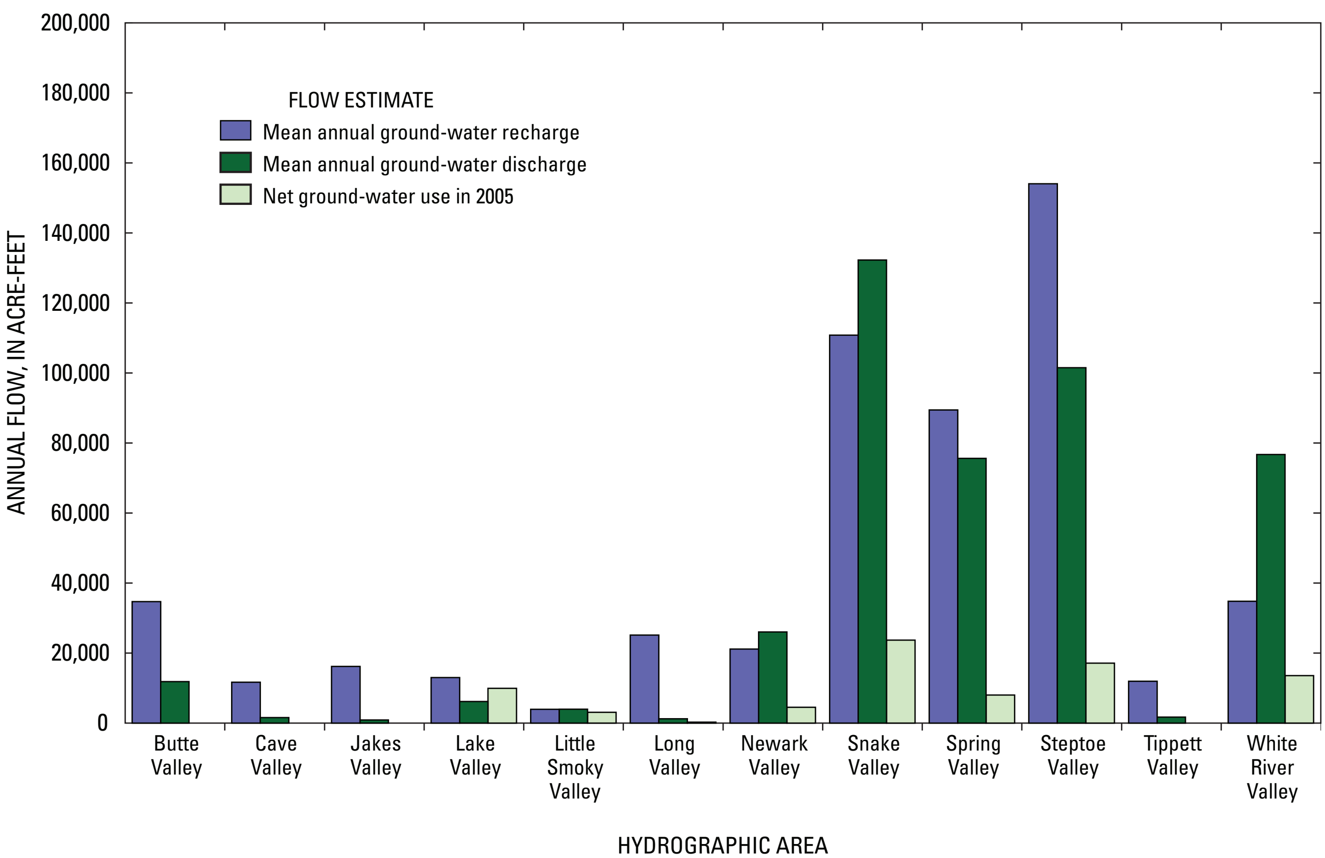


Figure 47. Estimates of mean annual ground-water recharge, mean annual ground-water discharge, and the 2005 net regional ground-water use by hydrographic area in the Basin and Range carbonate-rock aquifer system study area, Nevada and Utah.

When including the estimated net ground-water use 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 ground-water use for 2005 (80,000 acre-ft). That is, the estimated net 2005 ground-water use 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 the long-term use of ground water at about the 2005 estimate could capture the estimated average annual volume of ground water exiting the study area. Moreover, this condition also could, in some combination, decrease subsurface outflow, decrease spring discharge, decrease phreatophytic discharge, or increase subsurface recharge from adjacent basins. However, actual decreases in the volume of 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, decreases in outflow would be less likely in Butte or Tippett Valleys where net ground-water use was zero in 2005 (fig. 47). Decreases in outflow would be more likely in subbasins having both high pumping and relatively large outflow such as in Snake Valley where net ground-water use was 24,000 acre-ft in 2005 and average annual ground-water outflow was estimated at 29,000 acre-ft (fig. 47). Additionally, the relatively large volume of water stored in the basin-fill aquifer (appendix A) would likely inhibit near-future decreases 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. Ongoing pumping has not yet significantly altered ET rates, regional springflows, or distribution of native vegetation. Evaluation of the timing and location of potential decreases 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 likely is 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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Sue Beard – Structural geology of the Grant Range, stratigraphy of Cenozoic sedimentary basins

Susan G. Buto – Geographic information systems

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.usgs.gov/sir/2007/5261.

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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.

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 and resulting 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.

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.

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.

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.

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 ground-water system.

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.

Indurated: 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 relation 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.

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

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

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.

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

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.

Physiography: Same as physical geography.

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

Pluton: A body of medium- to coarse-grained igneous rock formed beneath Earth's surface by crystallization of magma. This term also can be defined as including bodies formed beneath the surface by metasomatic replacement of older rock.

Potentiometric surface: Where based on water-level data for wells tapping the same altitude, 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. Rock has schistose cleavage and is product of regional metamorphism.

Schistose: A rock displaying 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 ground water from storage, due to a unit decline in hydraulic head. These properties are often determined in hydrogeology using an aquifer test.

Stratabound: 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.

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 ground water at which the pressure is equal to that of the atmosphere.

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