Prepared in cooperation with the Douglas County, Nevada

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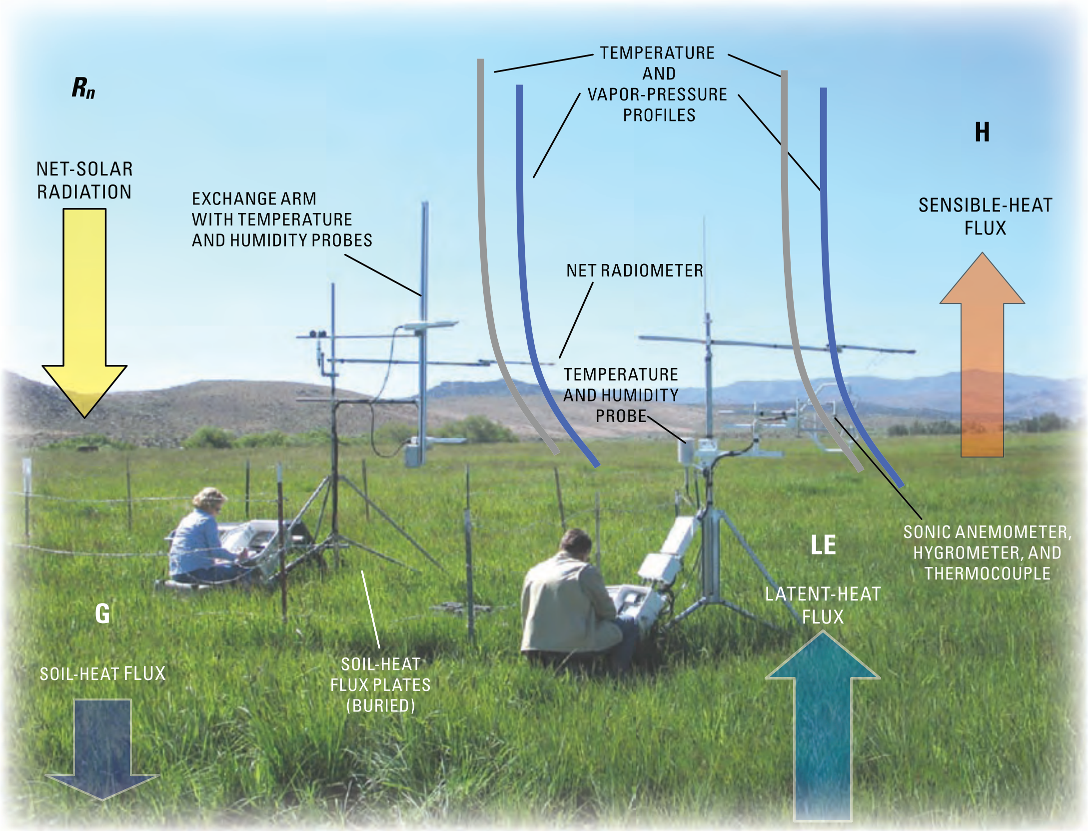

Scientific Investigations Report 2005–5288

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

Cover: Photograph of instrumentation used to collect micrometerological data for estimating evapotranspiration showing relative direction of energy-budget components during daytime. (Photograph taken by Douglas K. Maurer, USGS Nevada WSC, on 04/27/04.)

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Conversion Factors

Multiply	By	To obtain
acre	4,047	square meter
	0.4047	square hectometer
	0.004047	square kilometer
acre-foot (acre-ft)	1,233	cubic meter
	0.001233	cubic hectometer
cubic foot per second per mile (ft^3/s)/mi	0.0176	cubic meter per second per kilometer
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
inch (in.)	2.54	centimeter
	25.4	millimeter
inch per year (in/yr)	2.54	centimeter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	259.0	hectare
	2.590	square kilometer

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

$$°F=(1.8\times°C)+32.$$

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

$$°C=(°F-32)/1.8.$$

Datums

Vertical coordinate information is referenced to the North American Vertical Datum of 1929 (NAVD 29). Horizontal coordinate information is referenced to the North American Datum of 1927 (NAD 27).

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

Water-Quality Measurement Abbreviations

Acronym	Definition
g	gram
mL	milliliter
mg/L	milligram per liter

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Abstract

Rapid growth and development in Carson Valley is causing concern over the continued availability of water resources to sustain such growth into the future. A study to address concerns over water resources and to update estimates of water-budget components in Carson Valley was begun in 2003 by the U.S. Geological Survey, in cooperation with Douglas County, Nevada. This report summarizes micrometeorologic, soil-chloride, and streambed-temperature data collected in Carson Valley from April 2003 through November 2004. Using these data, estimates of rates of discharge by evapotranspiration (ET), rates of recharge from precipitation in areas of native vegetation on the eastern and northern sides of the valley, and rates of recharge and discharge from streamflow infiltration and seepage on the valley floor were calculated. These rates can be used to develop updated water budgets for Carson Valley and to evaluate potential effects of land- and water-use changes on the valley's water budget.

Data from eight ET stations provided estimates of annual ET during water year 2004, the sixth consecutive year of a drought with average or below average precipitation since 1999. Estimated annual ET from flood-irrigated alfalfa where the water table was from 3 to 6 feet below land surface was 3.1 feet. A similar amount of ET, 3.0 feet, was estimated from flood-irrigated alfalfa where the water table was about 40 feet below land surface. Estimated annual ET from flood-irrigated pasture ranged from 2.8 to 3.2 feet where the water table ranged from 2 to 5 feet below land surface, and was 4.4 feet where the water table was within 2 feet from land surface. Annual ET estimated from nonirrigated pasture was 1.7 feet. Annual ET estimated from native vegetation was 1.9 feet from stands of rabbitbrush and greasewood near the northern end of the valley, and 1.5 feet from stands of native bitterbrush and sagebrush covering alluvial fans along the western side of the valley. Uncertainty in most ET estimates is about 12 percent, but ranged from +30 and +50 percent to –20 and –40 percent for nonirrigated pasture and native bitterbrush and sagebrush. Estimated rates for water year 2004 likely are less than those during years of average, or above average precipitation when the water table would be closer to land surface.

Test holes drilled in areas of native vegetation on the northern and eastern sides of Carson Valley had high concentrations of soil chloride at depths ranging from 4 to 18 feet below land surface at six locations on the eastern side of the valley. The high chloride concentrations indicate that modern-day precipitation at the six locations does not percolate deeper than the root zone of native vegetation. Estimates of the time required to accumulate the measured amount of chloride to depths of about 30 feet below land surface at the six test holes ranged from about 3,000 to 12,000 years.

Low concentrations of soil chloride in two test holes on the northern end of Carson Valley and in a test hole on the eastern side of Fish Spring Flat indicate that a small amount of recharge from modern-day precipitation is taking place. Estimated annual recharge from precipitation at the two locations was 0.03 and 0.04 foot on the northern end of the valley and 0.02 foot on the eastern side of Fish Spring Flat. Uncertainty in the estimated recharge rates was about ±0.01 foot. Estimates of the time required to accumulate the measured amount of chloride to depths of about 30 feet below land surface at the three test holes ranged from about 100 to 700 years. The two test holes near the northern end of the valley are in gravel and eolian sand deposits and recharge from precipitation may be taking place at similar rates in other areas with gravel and eolian sand deposits. Based on results from other test holes, recharge at the rate estimated for the test hole on the eastern side of Fish Spring Flat is not likely applicable to a large area.

Data from 37 sites on the floor of Carson Valley where streambed temperatures were measured indicate that stream sites generally were gaining or neutral on the western-most side of the valley and north of Muller Lane, and sites were losing flow on the eastern side and southern end of the valley. Estimated rates of surface-water infiltration to the water table at losing sites range from about 1 to 4 feet per day, and estimated rates of ground-water seepage to streams at gaining sites may range from 0.1 to 1 foot per day. The greater infiltration rates may not be applicable to stream reaches longer than about 1 mile and application of the rates should be made with caution. Estimated seepage rates appear consistent with measured gains in streamflow.

Introduction

Rapid growth and development in Carson Valley is causing concern over the continued availability of water resources to sustain such growth into the future. As growth continues, ground-water pumping will increase and land presently used for agriculture will be urbanized. The effects of these changes on the valley's water budget are uncertain, and the changes may affect flow in the Carson River which, in turn, may affect water users dependent on river flow downstream of Carson Valley (fig. 1).

In the early 1980s, the U.S. Geological Survey (USGS) estimated water-budget components for Carson Valley (Maurer, 1986). Major water-budget components include inflow from precipitation and infiltration of streamflow, and outflow from evapotranspiration (ET) and ground-water seepage to the Carson River. To address concerns over water resources in Carson Valley, a cooperative study between USGS and Douglas County, Nevada, began in February 2003 to provide updated estimates of major components of the water budget. As part of the study, reports have been published providing updated estimates of the areal distribution of annual precipitation in Carson Valley (Maurer and Halford, 2004), and updated estimates of annual and monthly streamflow tributary to the floor of Carson Valley (Maurer and others, 2004).

Since the USGS study in the early 1980s, new methods have been routinely applied to estimate ET from plants using micrometeorologic measurements (Duell, 1990; Nichols, 1992), to estimate recharge from precipitation using the chloride concentration of pore water in soil profiles (Allison and others, 1994; Phillips, 1994), and to estimate streamflow gains and losses using temperature measurements (Constantz and others, 2001; 2002). These new methods were used in Carson Valley to develop updated estimates of the major water-budget components so the effects of land- and water-use changes on the valley's water budget can be evaluated.

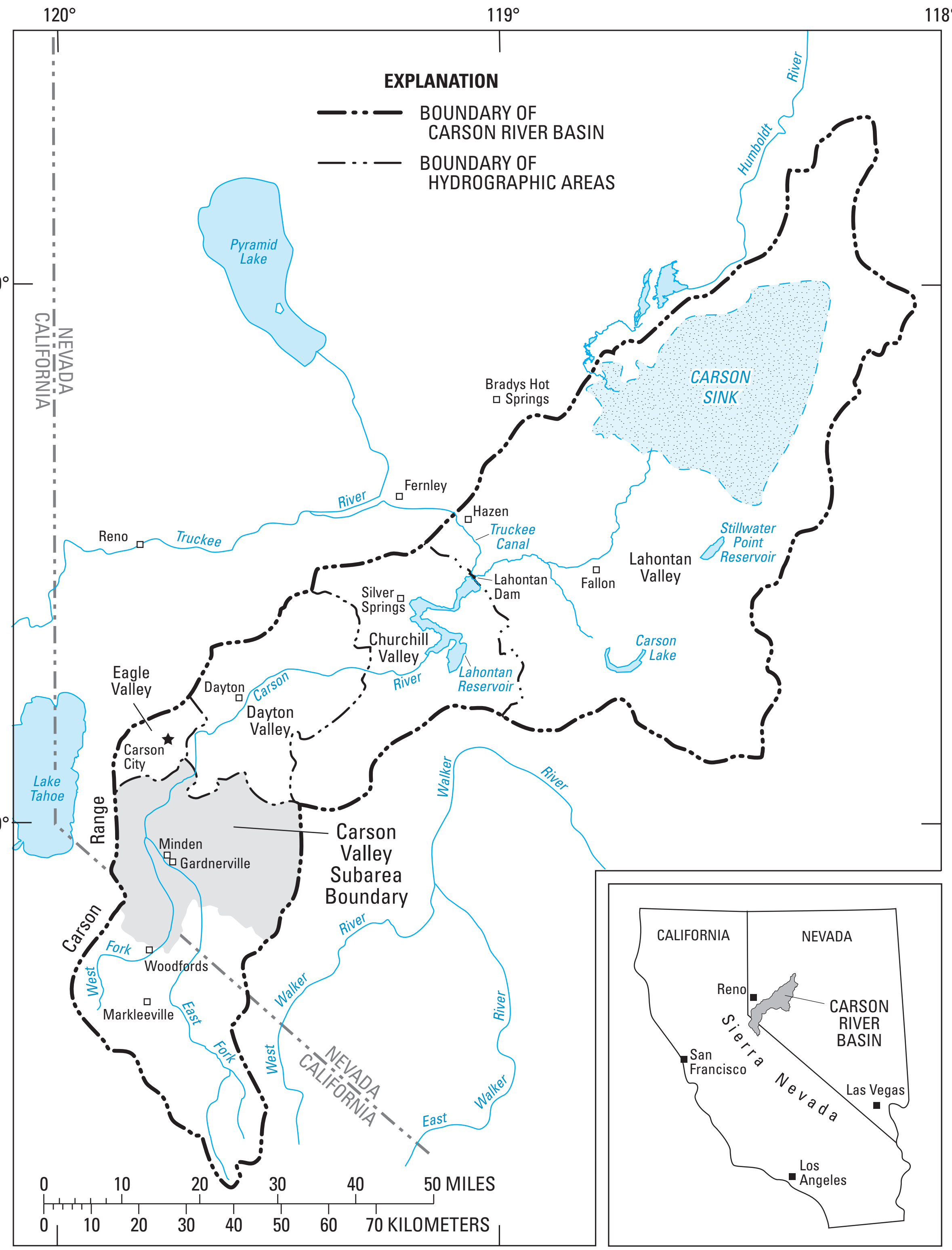
Purpose and Scope

This report presents and summarizes micrometeorologic, soil-chloride, and streambed-temperature data collected in Carson Valley between April 2003 and November 2004, and presents estimated rates of ET, infiltration of precipitation, and streamflow gain and loss determined using these measurements. Micrometeorologic data were collected at eight sites over various periods from mid-April 2003 through November 2004 and were used to estimate annual and monthly rates of ET from flood-irrigated alfalfa fields, flood-irrigated and nonirrigated pastures, and nonirrigated stands of native vegetation. Soil-chloride samples were collected from nine test holes as deep as 30 ft in areas of native vegetation and used to estimate rates of recharge from precipitation on the northern and eastern sides of the valley. Streambed-temperature data were collected at 37 sites on the Carson River, irrigation canals, and ditches and were used to identify locations of gaining and losing streamflow. The data also were used to estimate rates of surface-water infiltration to the water table and rates of ground-water seepage to streamflow at selected sites.

Geographic Setting

Carson Valley is in Douglas County, Nevada, south of Carson City, Nevada's capital (fig. 2). The southern end of the valley floor extends about 3 mi into Alpine County, California. The floor of the valley is oval-shaped, about 20 mi long and 8 mi wide, and slopes from about 5,000 ft above sea level at the southern end to about 4,600 ft at the northern end. The Carson Range of the Sierra Nevada rises abruptly from the valley floor on its western side with mountain peaks ranging from 9,000 to 11,000 ft, whereas on the eastern side, the Pine Nut Mountains rise gradually to peaks ranging from 8,000 to 9,000 ft (fig. 2).

The valley floor is covered with native pasture grasses, and crop lands of primarily alfalfa, and phreatophytes such as greasewood, rabbitbrush, and big sage near the northern end of the valley. In 1997, about 38,000 acres in Douglas County were irrigated and 26,000 acres were designated as cropland (U.S. Department of Agriculture, 2004). On the western side of the valley, bitterbrush and sagebrush cover steep alluvial fans, and manzanita and ponderosa pine cover the slopes of the Carson Range. Alluvial fans and foothills on the eastern side of the valley are covered with sparse, low-lying sagebrush, whereas pinyon and juniper are more prevalent in higher altitudes of the Pine Nut Mountains.



Base from U.S. Geological Survey digital data, 1:100,000, 1988
Universal Transverse Mercator projection, zone 11.
North American Datum of 1983 (NAD 83).
Shaded-relief from 30-meter Digital Elevation Model.
Sun illumination from northwest at 30 degrees above horizon.

Figure 1. Location of Carson River Basin and Carson Valley Hydrographic Area, Nevada and California.

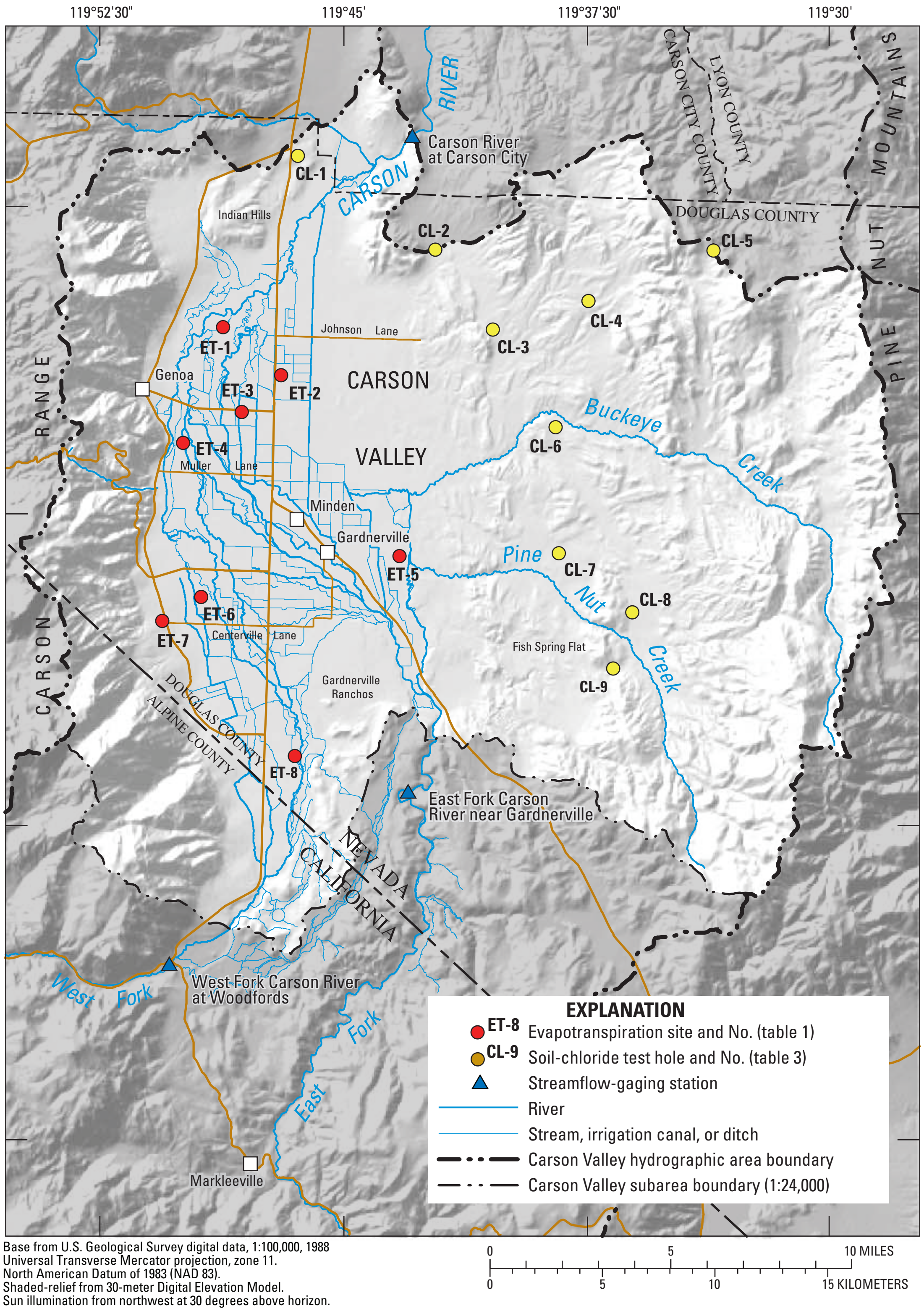


Figure 2. Location of Carson Valley subarea, evapotranspiration sites, and soil-chloride test holes, Nevada and California.

The major towns in the valley are Minden and Gardnerville with populations in 2000 of 2,800 and 3,400, respectively (U.S. Census Bureau, 2003; fig. 2). Subdivisions of the Gardnerville Ranchos to the south and Johnson Lane and Indian Hills to the north are growing rapidly, with populations in 2000 of 11,000, 4,800, and 4,400, respectively (U.S. Census Bureau, 2003). In addition, development is increasing along the eastern and western margins of the valley, and on the valley floor on land that historically has been agricultural. Douglas County's population, as a whole, has grown from about 28,000 in 1990 to 41,000 in 2000, an increase of 49 percent (Economic Research Service, 2003).

A major geographic feature of Carson Valley is the Carson River. The East and West Forks of the Carson River enter from the southeast and southwest corners of the valley, respectively, and flow northward to join near Genoa. The combined flow of the Carson River continues north to leave Carson Valley about 5 mi southeast of Carson City (fig. 2). Flow of the East and West Forks of the Carson River is diverted across the valley floor through a network of canals and ditches for flood irrigation of pasture grasses and crops.

For purposes of this study, a subarea of the entire Carson Valley Hydrographic Area[1] was delineated (figs. 1 and 2) to include only those parts of the Hydrographic Area connected by permeable materials capable of transmitting ground water to the floor of Carson Valley. The only difference between the hydrographic area and the subarea is along the southern boundary, where the headwaters of the East and West Forks of the Carson River have been excluded. Bedrock underlies the subarea boundary where the East and West Forks of the Carson River enter the valley, and ground-water inflow is minimal through very thin fluvial sediments underlying the river channels (fig. 3). Ground-water flow across the alluvial fan west of the West Fork of the Carson River likely is parallel to the subarea boundary with minimal flow across the boundary. The subarea used for this report, as shown in figure 2, covers 253,570 acres, or about 396 mi^2.

Geologic Setting

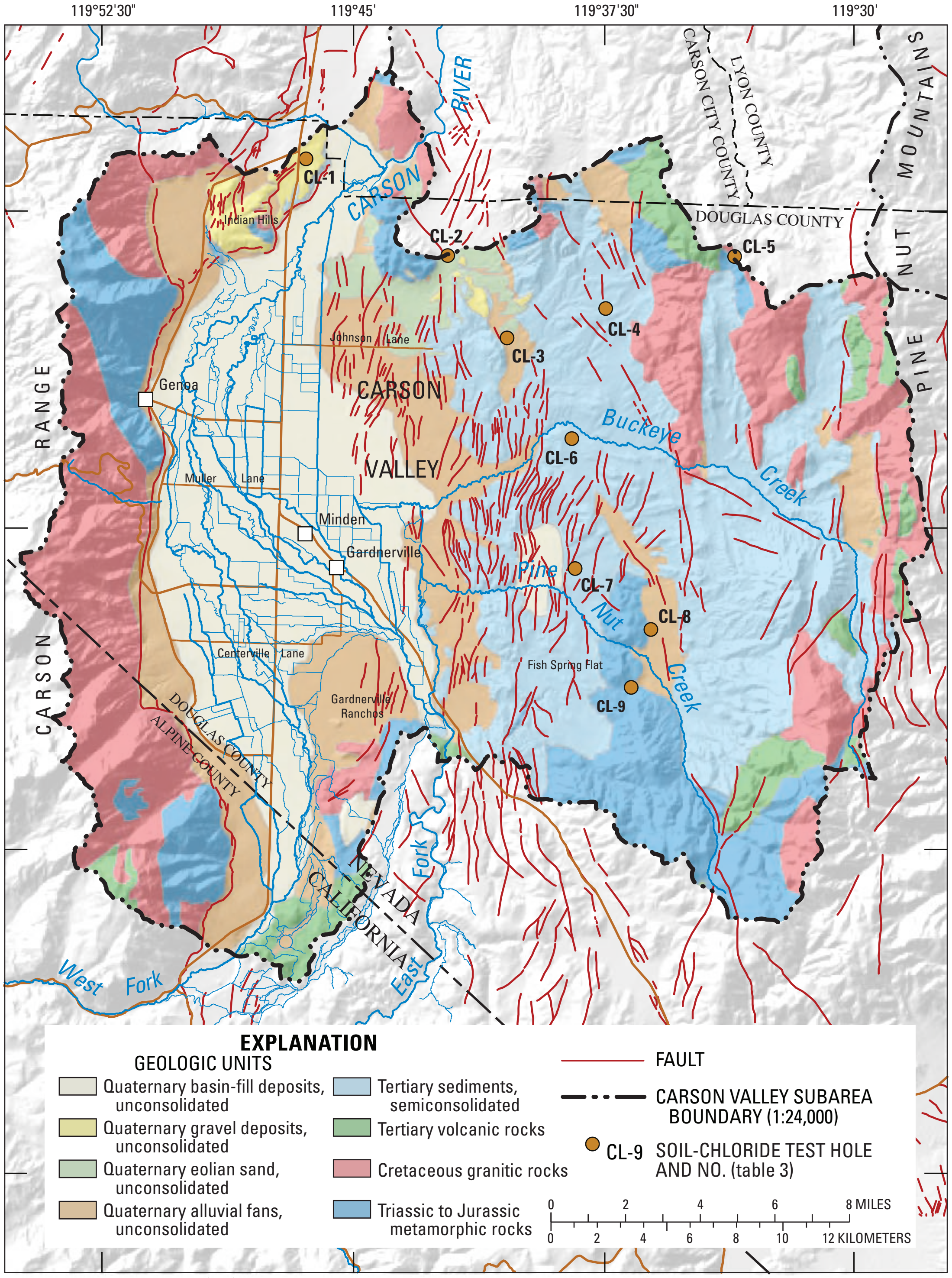
The granitic magma of the Sierra Nevada pluton was intruded 63–138 Ma (million years ago) during the Cretaceous Period into sedimentary and volcanic rocks deposited 138–240 Ma during the Triassic and Jurassic Periods. The resulting granodioritic, metavolcanic, and metasedimentary rocks form the bulk of the Carson Range and the Pine Nut Mountains (fig. 3), and underlie the floor of Carson Valley (Moore, 1969, p. 18; Pease, 1980, p. 2). Basin and range faulting, which produced much of the present topography in Carson Valley, took place from 7 to 10 Ma (Muntean, 2001, p. 9), uplifting the Carson Range and the Pine Nut Mountains, and down dropping the floor of Carson Valley.

Prior to, and contemporaneous with the faulting, volcanic rocks and sediments were deposited during the Tertiary Period, 1.6–66 Ma. Through time, the sediments have become semiconsolidated. Volcanic rocks are exposed primarily on the extreme northeastern and southeastern ends of the valley (fig. 3). The semiconsolidated sediments are exposed primarily on the eastern side of the valley, but dip towards the west and probably are present beneath the entire valley. The semiconsolidated Tertiary sediments vary in their degree of compaction (Pease, 1980, p. 14), and vary in lithology from fine-grained and tuffaceous siltstone with isolated lenses of sandstone and conglomerate, to primarily sandstone and conglomerate (Muntean, 2001, p. 18–31). The coarser grained Tertiary sediments are exposed primarily on the southeastern part of the valley at the base of the Pine Nut Mountains (Muntean, 2001, p. 19). The aggregate thickness of the Tertiary sediments is estimated to exceed 3,000 ft (Muntean, 2001, pl. 5).

Throughout the Quaternary Period (present day–2 Ma), unconsolidated sediments (fig. 3) have been deposited on the valley floor by the Carson River and tributary streams surrounding the valley. These sediments generally are well-sorted sand and gravel, interbedded with fine-grained silt and clay from overbank flood deposits. Unconsolidated sediments deposited by tributary streams are coarse- to fine-grained, poorly sorted deposits, which form alluvial fans at the base of the mountain blocks.

The mountain blocks bounding Carson Valley are west-tilted structural blocks (Stewart, 1980, p. 113), with the valley occupying the down-dropped western edge of the Pine Nut Mountains block (Moore, 1969, p. 18). A steep, well-defined normal fault creates a 5,000 ft escarpment along the Carson Range on the west, whereas a diffuse fault zone on the eastern side of the valley divides the Pine Nut Mountains block into several smaller blocks (fig. 3). Continued westward tilting is shown by recent faulting along the base of the Carson Range (Pease, 1980, p. 15) and by displacement of the Carson River to the extreme western side of the valley floor (Moore, 1969, p. 18). A gravity survey by Maurer (1985) indicates the depth to the top of consolidated bedrock beneath the western half of Carson Valley is as much as 5,000 ft.

[1]The U.S. Geological Survey and Nevada Division of Water Resources systematically delineated formal hydrographic areas in Nevada in the late 1960s for scientific and administrative purposes (Cardinalli and others, 1968). The official hydrographic-area names, numbers, and geographic boundaries continue to be used in U.S. Geological Survey scientific reports and Nevada Division of Water Resources administrative proceedings and reports. Hydrographic-area boundaries generally coincide with drainage-area boundaries.



Base geology from Stewart and Carlson (1978) at 1:250,000 scale.
Datum is North American Datum of 1983 (NAD 83).
Modified from geologic mapping by Stewart (1999) at 1:100,000 scale;
by Armin and John (1983), Stewart and Noble (1979) at 1:62,500 scale;
and by dePolo and others (2000), Garside and Rigby (1998), and
Muntean (2001) at 1:24,000 scale.

Figure 3. Geologic units and faults in Carson Valley subarea, and location of soil-chloride test holes relative to geologic units, Nevada and California.

Hydrologic Setting

Carson Valley lies in the rainshadow of the Sierra Nevada, with precipitation at the town of Minden, averaging 8.4 in/yr for 1971–2000 (National Oceanic and Atmospheric Administration, 2002, p. 12). For the same period, precipitation averaged about 40 in/yr at the top of the Carson Range to the west, and precipitation averaged 15–18 in/yr near the top of the Pine Nut Mountains to the east (Maurer and Halford, 2004, p. 35). Precipitation over most of the northern and eastern sides of Carson Valley was estimated to average 10–15 in/yr for 1971–2000 (Maurer and Halford, 2004, p. 33–34). Precipitation data for the eastern side of Carson Valley, about 7 mi east of Minden, indicates annual precipitation ranged from about 16 in. in 1995 to about 5 in. from 1999 through 2002 (period of record 1991–2002; Fish Springs RAWS site, Western Region Climate Center, written commun., 2003). Monthly precipitation at that location ranges from a maximum of about 1 in. from December through March to a minimum of about 0.3 in. from July through September (period of record 1991–2002; Fish Springs RAWS site, Western Region Climate Center, written commun., 2003).

Since 1999, annual precipitation near Minden, which generally is representative of conditions throughout the subarea, has been less than average, with 2004 being the sixth consecutive year of average or below average precipitation. Annual precipitation was slightly less than average in 2001, but was considerably less than average from 2002 through 2004 (fig. 4).

The hydrology of Carson Valley is dominated by flow of the Carson River. Average annual inflow (1990–2002) from the East Fork Carson River near Gardnerville, Nevada, was 257,000 acre-ft and from the West Fork Carson River at Woodfords, California, was 75,150 acre-ft (Maurer and others, 2004, p. 14), for a total of about 332,000 acre-ft. Average annual outflow of the mainstem Carson River was 287,300 acre-ft for that period (Maurer and others, 2004, p. 14). Thirteen perennial streams drain the Carson Range and are tributary to the floor of Carson Valley (Maurer and others, 2004), whereas only two perennial streams, Buckeye and Pine Nut Creeks, drain the Pine Nut Mountains (fig. 2).

Infiltration of surface water from streams, ditches, and flood-irrigated fields maintains a shallow water table beneath much of the valley floor where depth to ground water is less than 5 ft (Maurer and Peltz, 1994, sheet 2). Depth to water beneath alluvial fans on the western side of the valley increases to more than 200 ft within 1 mi of the valley floor, whereas depth to water on the eastern side of the valley reaches 200 ft about 3 mi from the valley floor (Maurer and Peltz, 1994, sheet 2).

Ground water flows from the west and east towards the Carson River and then northward (Berger and Medina, 1999). Along the main axis of the valley, water-level gradients range from about 100 ft/mi in the southwestern part of the valley to about 5 ft/mi in the central and northern parts of the valley (calculated from Maurer, 1986, p. 18). Beneath alluvial fans on the western side of the valley, the gradient generally is eastward at about 100 ft/mi, whereas on the eastern side of the valley the gradient generally is westward and ranges from 20 to 100 ft/mi (calculated from Maurer, 1986, p. 18).

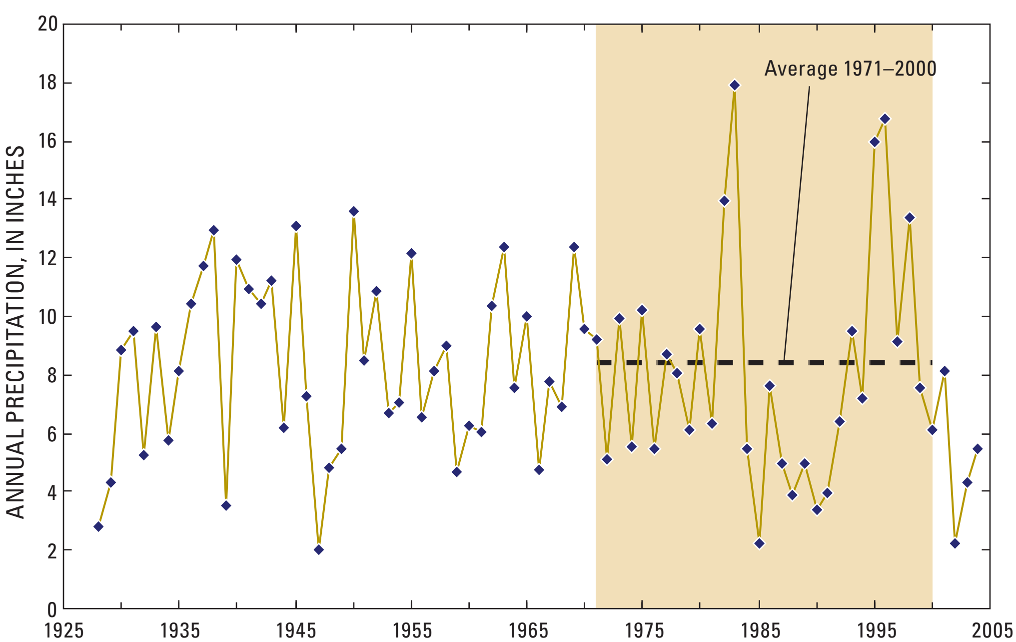


Figure 4. Annual precipitation at Minden, Nevada, for period of record 1928–2004, and average annual precipitation for 1971–2000, Carson Valley, Nevada and California.

Consolidated granitic and metamorphic bedrock surrounding and underlying Carson Valley is relatively impermeable to ground-water flow, although some wells produce sufficient water from fractures for domestic use. In semiconsolidated Tertiary sediments, lenses of sand and gravel are the primary water-bearing features, and probably transmit most ground water that moves through the unit. Unconsolidated sediments that form alluvial fans surrounding the valley floor and underlie the flood plain of the Carson River are the principal aquifers in Carson Valley (Maurer, 1986, p. 17).

Estimated Rates

Evapotranspiration

ET is the process in which water is evaporated and changes from a liquid to a vapor and discharged to the atmosphere. ET is the sum of direct evaporation of free water and transpiration by plants. Transpiration is the process whereby shallow ground water or water retained in the subsurface as soil moisture is adsorbed by plant roots, moves through the plant, and is evaporated from leaves to the atmosphere. ET in Carson Valley includes evaporation from bare soil, evaporation of precipitation and open water, and transpiration by plants. The source of water transpired by plants includes precipitation, applied irrigation water, soil moisture, and ground water. Estimates of daily, monthly, and annual ET rates presented in this report are for the total ET supplied by all these sources of water. Estimates of ET rates for the various plant communities in Carson Valley are useful for evaluating the effects of changes in land use and developing updated water budgets in Carson Valley.

Methods Used

Significant and measurable amounts of energy are required for water to be evaporated. Instrumentation used for this study measured components of the energy budget to solve an energy-budget equation and compute estimates of ET. The energy-budget equation partitions energy into four principal flux components: (1) net-solar radiation, (2) soil-heat flux, (3) sensible-heat flux, and (4) latent-heat flux. The energy-budget equation can be written as modified from Brutsaert (1982, p. 2):

$$R_n - G = H + LE, \qquad (1)$$

where

R_n is the net-solar radiation, in energy per area per time,

G is the soil-heat flux, in energy per area per time,

H is the sensible-heat flux, in energy per area per time, and

LE is the latent-heat flux, in energy per area per time.

Flux components that make up the energy budget are illustrated in figure 5 with their typical daytime directions. The primary source of heat energy is incoming solar radiation. The balance between incoming and outgoing solar radiation at the surface of the Earth is called net radiation (R_n). Net radiation is positive during the daytime when the incoming solar and long-wave radiation exceeds outgoing radiation transmitted back to the atmosphere. Soil-heat flux (G) is the energy exchange through the soil and is considered positive during the daytime when moving downward. Net radiation and soil-heat flux are measured using net radiometers mounted above the plant canopy and heat-flux plates buried 0.2 ft beneath the soil surface. The energy difference between net radiation and soil-heat flux is the energy available to drive the sensible-heat flux (H) and latent-heat flux (LE) at the Earth's surface. Sensible-heat flux is the amount of energy that heats the air and is positive during the daytime. Latent-heat flux is the energy consumed for ET and is positive when vapor is transferred upward. The directions of energy movement reverse at nighttime when net radiation, soil-heat flux, and sensible-heat flux become negative, and latent-heat flux is approximately zero.

In this study, two standard methods were used to solve the energy-budget equation and estimate ET, the Bowen-ratio method (Bowen, 1926), and the eddy-correlation method (Swinbank, 1951). Details of the application of these methods are given by Nichols (1992),

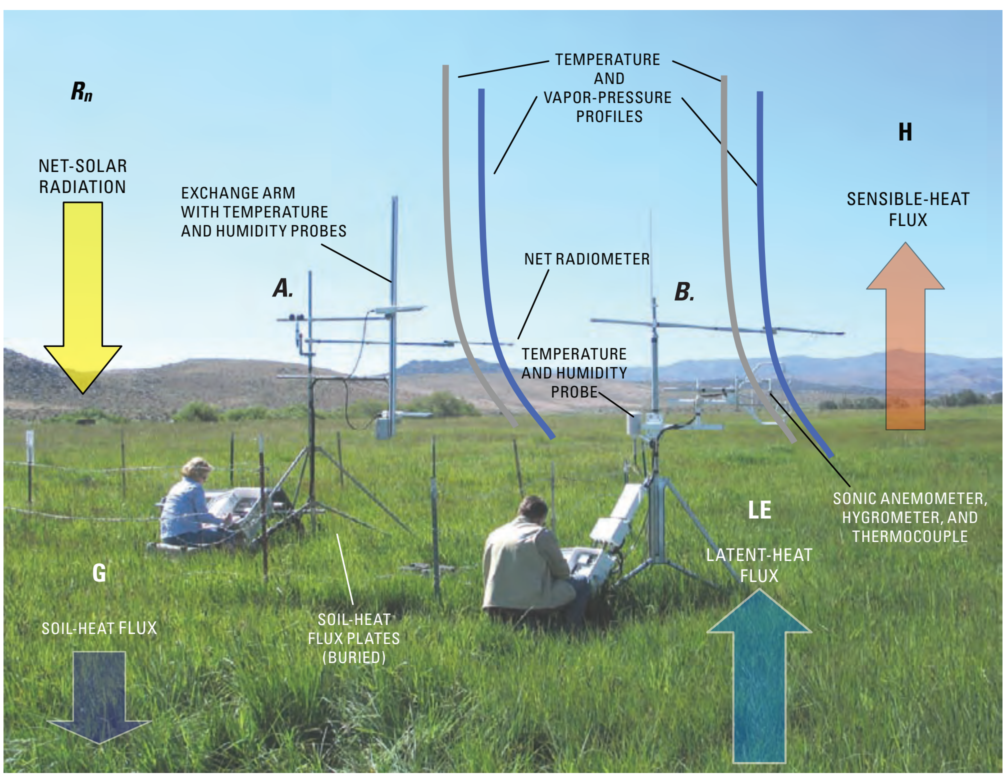


Figure 5. Instrumentation used to collect micrometeorological data for estimating evapotranspiration showing relative direction of energy-budget components during daytime.
Instruments shown are (*A*) Bowen-ratio instrumentation and (*B*) eddy-correlation instrumentation.

Laczniak and others (1999), and Berger and others (2001). The two methods require different types of instrumentation (fig. 5) to collect the micrometeorological data necessary to determine the flux components of the energy budget near land surface. The energy budget is solved for LE to obtain an estimate of the energy consumed by ET, which then can be converted to the rate of ET.

The Bowen-ratio method uses instrument stations with sensors at two heights above the plant canopy to measure vertical differences in air temperature and vapor pressure, along with sensors to measure net radiation and soil-heat flux. The vertical differences in air temperature and vapor pressure are used to approximate the ratio of H to LE (known as the Bowen ratio). The measured difference between net radiation, R_n, and soil-heat flux, G, are then used with the Bowen ratio to solve the energy-budget equation (Bowen, 1926; Nichols, 1992; Laczniak and others, 1999; Berger and others, 2001). The positions of the air-temperature and vapor-pressure sensors are reversed every 10 minutes to cancel any instrumentation bias and obtain 20-minute values for ET computations. The Bowen-ratio instrumentation is capable of obtaining data in adverse weather conditions. For this reason, along with instrument availability, four Bowen-ratio stations were installed in spring 2003 and remained in operation, except for periods of instrument malfunction, until autumn 2004 at sites ET-1, ET-2, ET-3, and ET-8 (fig. 2, table 1).

The eddy-correlation method uses instrument stations with a sonic anemometer, a hygrometer, and a thermocouple (fig. 5) to measure rapid changes in vertical wind speed, vapor density, and temperature, respectively. The measurements of changes in vertical wind speed and air temperature are used to calculate sensible-heat flux. Measurements of changes in vertical wind speed and vapor density are used to calculate the energy consumed by ET (latent-heat flux). The instrumentation for the eddy-correlation method is much more susceptible to malfunction under rainy or windy conditions, but is relatively portable compared to the Bowen-ratio instrumentation. For these reasons, the eddy-correlation stations were not operated during winter and were moved between various types of plant communities at sites ET-4, ET-5, ET-6, and ET-7 (fig. 2, table 1). One-dimensional (1-D) sonic anemometers were used at eddy-correlation sites in 2003, but required frequent maintenance to minimize equipment malfunction. In 2004, a more robust three-dimensional (3-D) sonic anemometer was used to collect eddy-correlation data at sites ET-4 and ET-8 (fig. 2, table 1).

Estimates of ET rates were made at eight sites on and near the valley floor (fig. 2). The sites were selected to obtain estimates for the major types of vegetation present in Carson Valley: native phreatophytes (plants that use ground water) such as rabbitbrush and greasewood; native nonphreatophytes such as bitterbrush and sagebrush that cover alluvial fans on the western side of Carson Valley; and irrigated crops such

Table 1. Location, instrumentation type, period of record, descriptions of vegetation, and depth to water table at eight evapotranspiration sites, Carson Valley, Nevada and California.

[Site locations are shown in figure 2. **Latitude and Longitude:** Geographic coordinates referenced to North American Datum of 1927 (NAD 27), in degrees, minutes, and seconds. Depth to water table is shown in feet below land surface]

Site No.	West latitude	North longitude	Instrumentation type	Period of record	Description of vegetation
ET-1	39°01’47”	119°48’21”	Bowen ratio	04/17/03–10/25/04	Rabbitbrush/greasewood, depth to water table is 3–5 feet.
ET-2	39°00’40”	119°46’32”	Bowen ratio	04/25/03–11/04/04	Flood-irrigated alfalfa, depth to water table is 3–6 feet.
ET-3	38°59’45”	119°47’43”	Bowen ratio	05/01/03–11/04/04	Flood-irrigated pasture, depth to water table is 2–5 feet.
ET-4	38°58’58”	119°49’30”	Eddy correlation, 3-dimensional	06/08/04–11/01/04	Flood-irrigated pasture, depth to water table is 3–4 feet.
ET-5	38°56’25”	119°42’43”	Eddy correlation, 1-dimensional	08/12/03–10/28/03; 04/21/04–10/28/04	Flood-irrigated alfalfa, depth to water table is 40 feet.
ET-6	38°55’17”	119°48’47”	Eddy correlation, 1-dimensional	06/08/04–10/16/04	Non-irrigated pasture, depth to water table is 6–7 feet.
ET-7	38°54’41”	119°49’58”	Eddy correlation, 1-dimensional	05/16/03– 11/04/03	Bitterbrush/sagebrush, depth to water table is 60 feet.
ET-8	38°51’31”	119°45’45”	Bowen ratio	06/17/03–11/16/04	Flood-irrigated pasture, depth to water table is 0–2 feet.
			Eddy correlation, 1-dimensional	04/02/04– 06/06/04	Flood-irrigated pasture, depth to water table is 0–2 feet.
			Eddy correlation, 3-dimensional	03/30/04–06/07/04	Flood-irrigated pasture, depth to water table is 0–2 feet.

as alfalfa and pasture grasses. ET from low-lying and sparse sagebrush communities covering large areas on the northern and eastern sides of Carson Valley was not measured because such plants do not tap the water table, and recharge from precipitation was estimated from soil-chloride profiles in those areas. Depth to water is known to vary beneath areas of irrigated crops, thus, instrumentation sites for irrigated crops were selected in areas having different depths to water because this has been shown to affect ET rates (Nichols, 2000, p. 9). The exact placement of instrumentation sites was limited to locations where permission to install the equipment could be obtained. Although fairly large areas of crops on the eastern side of Carson Valley are irrigated with sprinkler systems, permission to install in sprinkler-irrigated areas was not obtained and only flood-irrigated areas were instrumented.

The instrumentation used at each site and the type of vegetation are shown in figures 6*A*–6*H*. Long-term Bowen-ratio stations were installed at one site with native vegetation (site ET-1) and at three sites on agricultural land (sites ET-2, ET-3, and ET-8). The four long-term Bowen-ratio sites were: (1) ET-1 in a stand of rabbitbrush, sagebrush, greasewood, and mixed grasses where the water table was 3–5 ft below land surface on the northern end of the valley; (2) ET-2 in a flood-irrigated alfalfa field where the water table was 3–6 ft below land surface; (3) ET-3 in a flood-irrigated field of pasture grass where the water table was 2–5 ft below land surface; and (4) ET-8 in a flood-irrigated field of pasture grass where the water table was 0–2 ft below land surface. At site ET-8, land surface often was saturated during site visits and the local rancher used the term "subirrigated" for the pasture even though it routinely was flood irrigated. Plant density at site ET-1 was estimated to be 73 percent and essentially was 100 percent at the pasture and alfalfa sites. The Bowen-ratio stations obtained continuous record except for relatively short periods when equipment malfunctioned due to failed exchange mechanisms for the temperature and relative-humidity sensors, damage from cattle that breached the fenced enclosure at site ET-3, and a fire at site ET-8.

Two 1-D eddy-correlation stations at sites ET-7 and ET-5 were installed in 2003 and operated for periods of 3–5 months to obtain data from different vegetation communities (fig. 2, table 1). Site ET-7 was operated from mid-May to November 2003 in a stand of bitterbrush (5–7 ft tall), sagebrush, and mixed grasses where the water table was about 60 ft below land surface. Plant density at site ET-7 was estimated to be 35 percent. Site ET-5 was operated from early August through October 2003 in a flood-irrigated alfalfa field where the water table was about 40 ft below land surface. At sites ET-5 and ET-7, the water table likely is below the reach of plant roots; thus, ground water probably is not consumed by transpiration at these sites.

In 2004, a 1-D and a 3-D eddy-correlation station were colocated with the Bowen-ratio station at site ET-8 from early April through May to obtain data with which to compare ET estimates from the three types of stations. From April through October 2004, a 1-D eddy-correlation station was reestablished at site ET-5. In early June, the two eddy-correlation stations from site ET-8 were moved to sites ET-4 and ET-6 and operated through October and mid-October 2004, respectively. Site ET-4 is a flood-irrigated field of pasture grasses where the water table was 3–4 ft below land surface, and site ET-6 is a nonirrigated field of pasture grasses where the water table was 6–7 ft below land surface. Plant density at sites ET-4 and ET-6 essentially was 100 percent.

ET data were collected in Carson Valley from mid-April 2003 through early November 2004 (table 1). To provide an annual estimate of ET during water year 2004, two methods were used to estimate daily ET during periods of equipment malfunction or when the stations were being moved. The first method used equations developed by simple linear regression between measured daily net radiation and the corresponding natural log of daily ET at each site. In the regressions, daily net radiation was the independent variable and natural log of daily ET was the dependent variable (fig. 7). The equations were used to generate daily ET estimates from Bowen-ratio stations at sites ET-1 and ET-8 (January 2004) where net radiation was available, but measurements of one or more flux components were missing because of equipment failure.

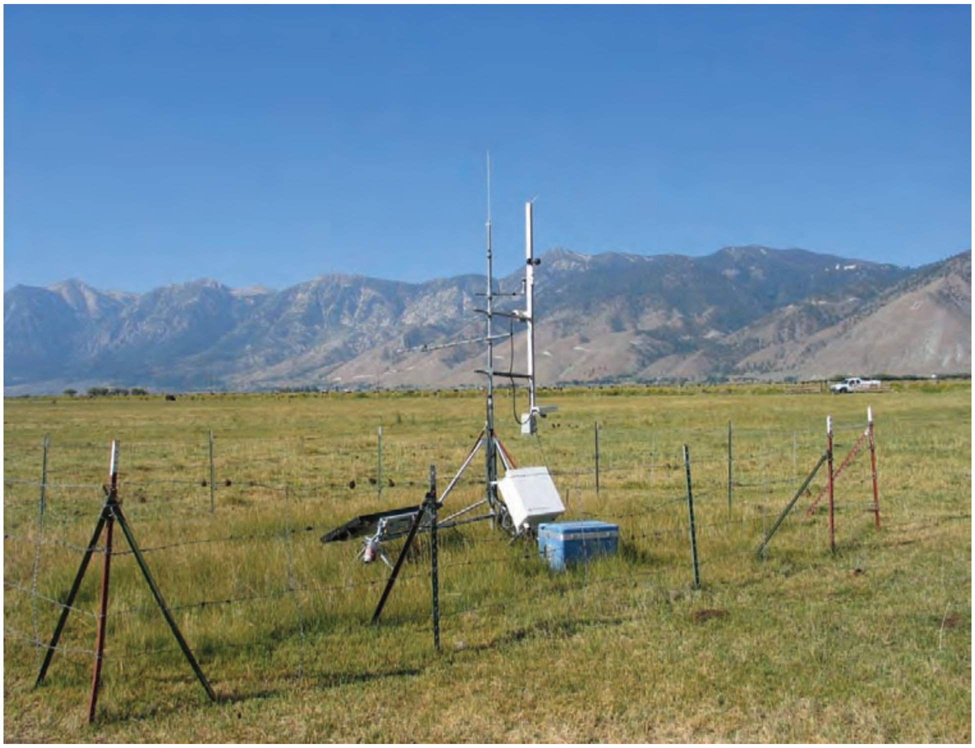
The second method to estimate daily ET was used for Bowen-ratio sites ET-2, ET-3, and ET-8 when net radiation was not available, and for all eddy-correlation sites ET-4–ET-7. This method used equations developed by simple linear regression between measured daily net radiation at each site and the average of daily net radiation recorded at all functioning ET stations in Carson Valley. In the regressions, measured daily net radiation was the dependent variable, average daily net radiation was the independent variable, and the equations are in the form of a linear function (fig. 8). The equations shown in figure 8 were used to estimate daily net radiation for periods of equipment malfunction and when the eddy-correlation stations were moved. The estimates of daily net radiation were then used with equations from figure 7 to estimate daily ET. In this report, daily ET estimates calculated using both methods are called predicted daily ET values.

The coefficient of determination (R^2) for equations in figure 7 are measures of the amount of variance in the data explained by the regression and range from 77 to 88 percent for the Bowen-ratio sites and from 54 to 73 percent for the eddy-correlation sites. R^2 for equations, in figure 8, ranges from 96 to 97 percent for the Bowen-ratio sites and from 79 to 98 percent for the eddy-correlation sites.

A. Bowen-ratio station at site ET-1 in stand of rabbitbrush and greasewood. Depth to the water table is from 3 to 5 feet below land surface. Looking northward towards Indian Hills.

B. Bowen-ratio station at site ET-2 in stand of flood-irrigated alfalfa. Depth to the water table is from 3 to 6 feet below land surface. Looking westward towards the Carson Range.

Figure 6. Instrumentation used at each site and the type of vegetation, Carson Valley, Nevada and California.



C. Bowen-ratio station at site ET-3 in stand of flood-irrigated pasture. Depth to the water table is from 2 to 5 feet below land surface. Looking southwestward towards the Carson Range.

D. Three-dimensional eddy-correlation station at site ET-4 in stand of flood-irrigated pasture. Depth to the water table is from 3 to 4 feet below land surface. Looking southwestward towards the Carson Range.

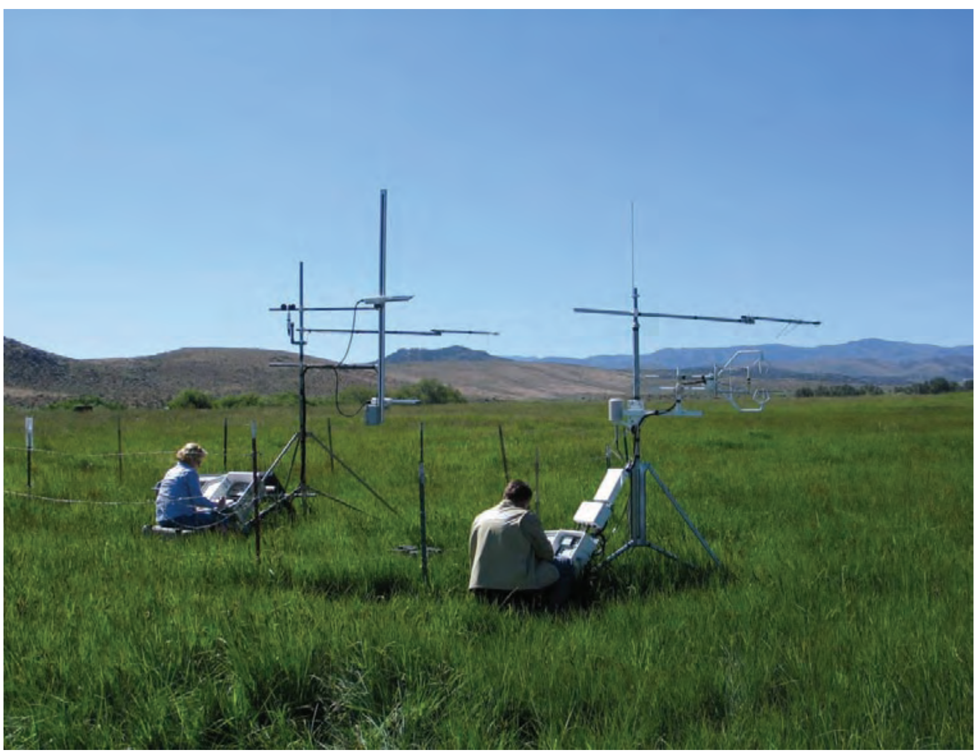
Figure 6.—Continued.

E. One-dimensional eddy-correlation station at site ET-5 in stand of flood-irrigated alfalfa. Depth to the water table is about 40 feet below land surface. Looking westward towards the Carson Range.

F. One-dimensional eddy-correlation station at site ET-6 in stand of non-irrigated pasture. Depth to the water table is from 6 to 7 feet below land surface. Looking westward towards the Carson Range.

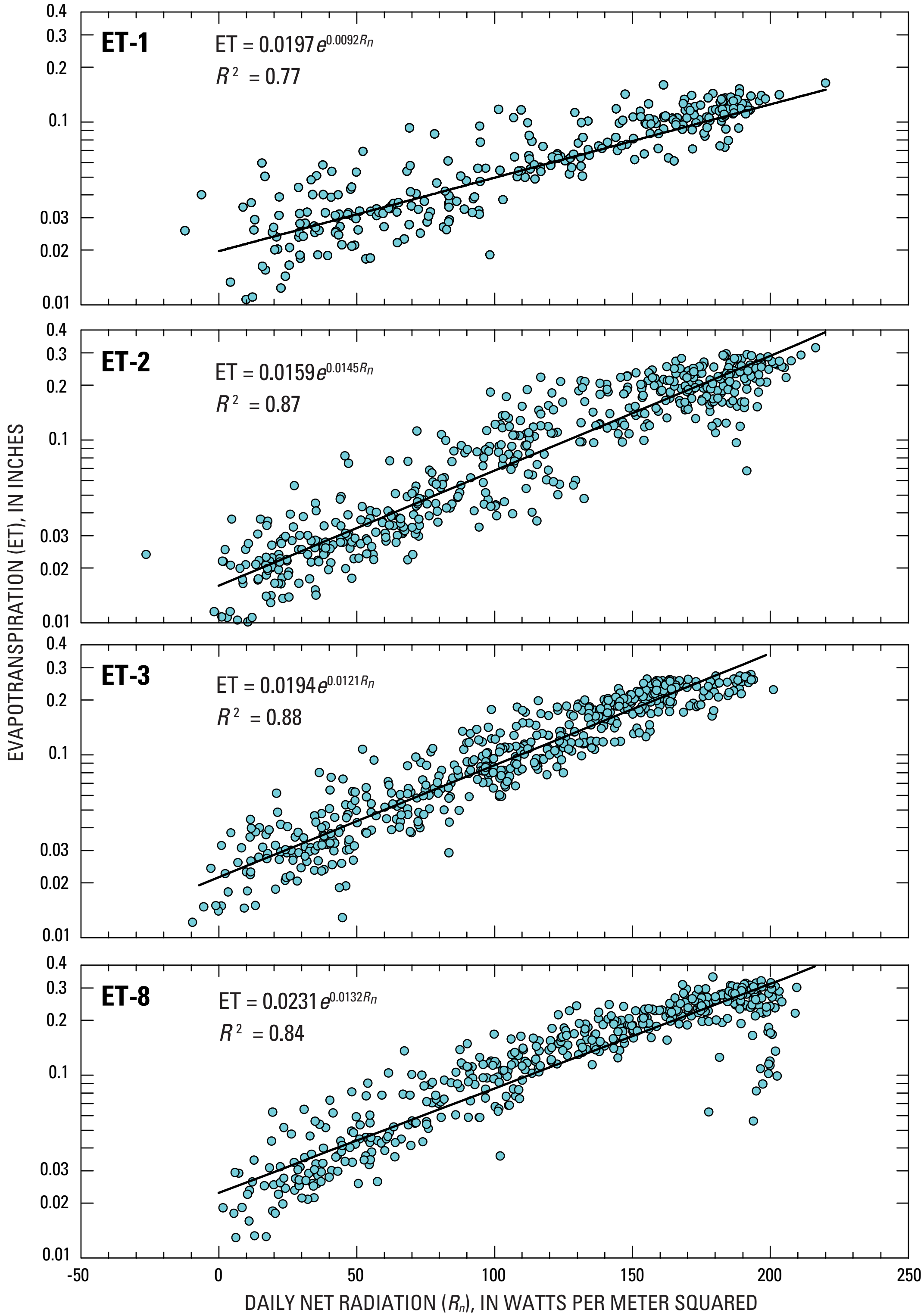
Figure 6.—Continued.

G. One-dimensional eddy-correlation station at site ET-7 in stand of sagebrush and bitterbrush. Depth to the water table is about 60 feet below land surface. Looking westward towards the Carson Range.



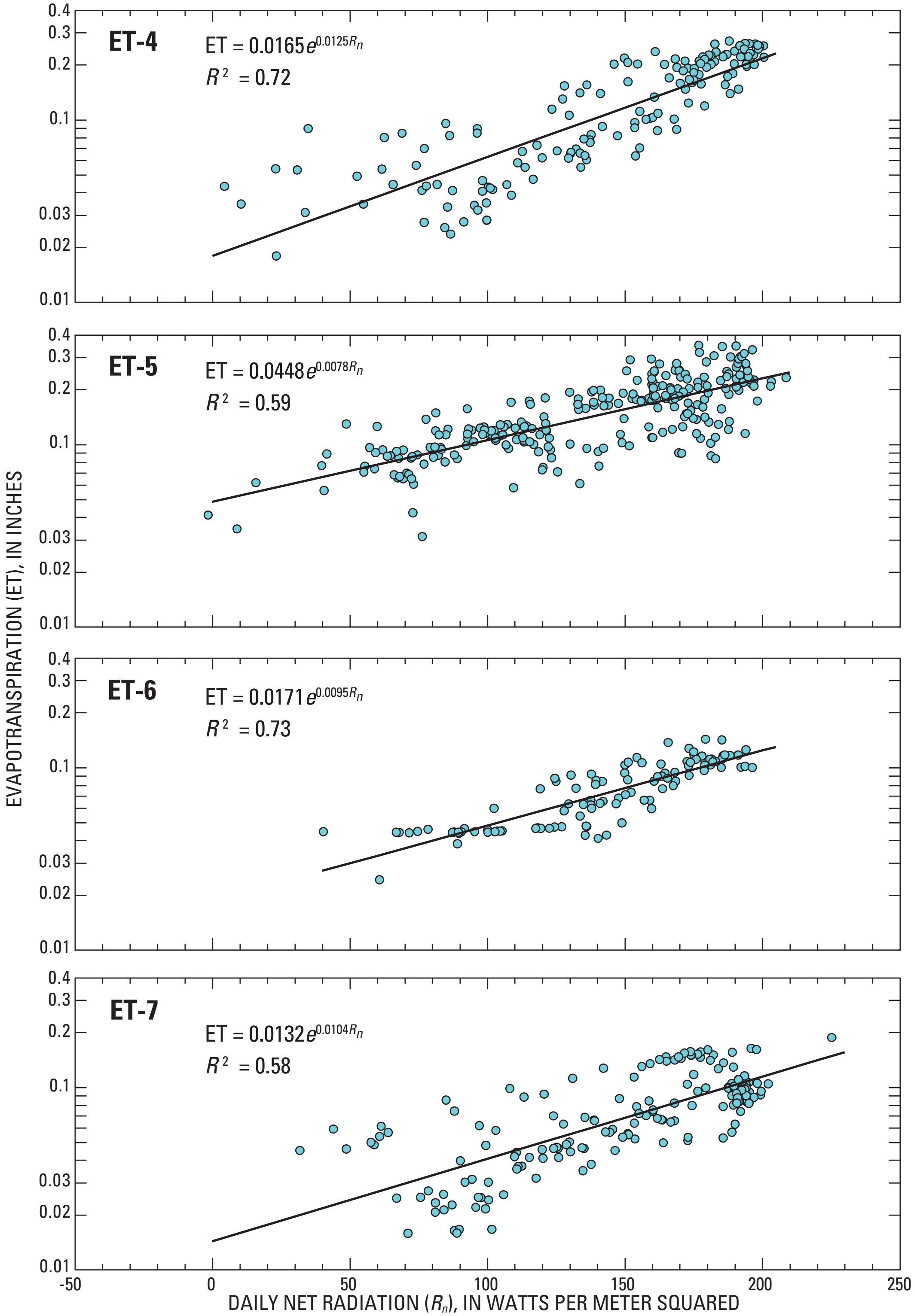
H. Bowen-ratio station on left and three-dimensional eddy correlation station on right at site ET-8 in stand of flood-irrigated pasture. Depth to the water table is from 0 to 2 feet below land surface. Looking southward.

Figure 6.—Continued.



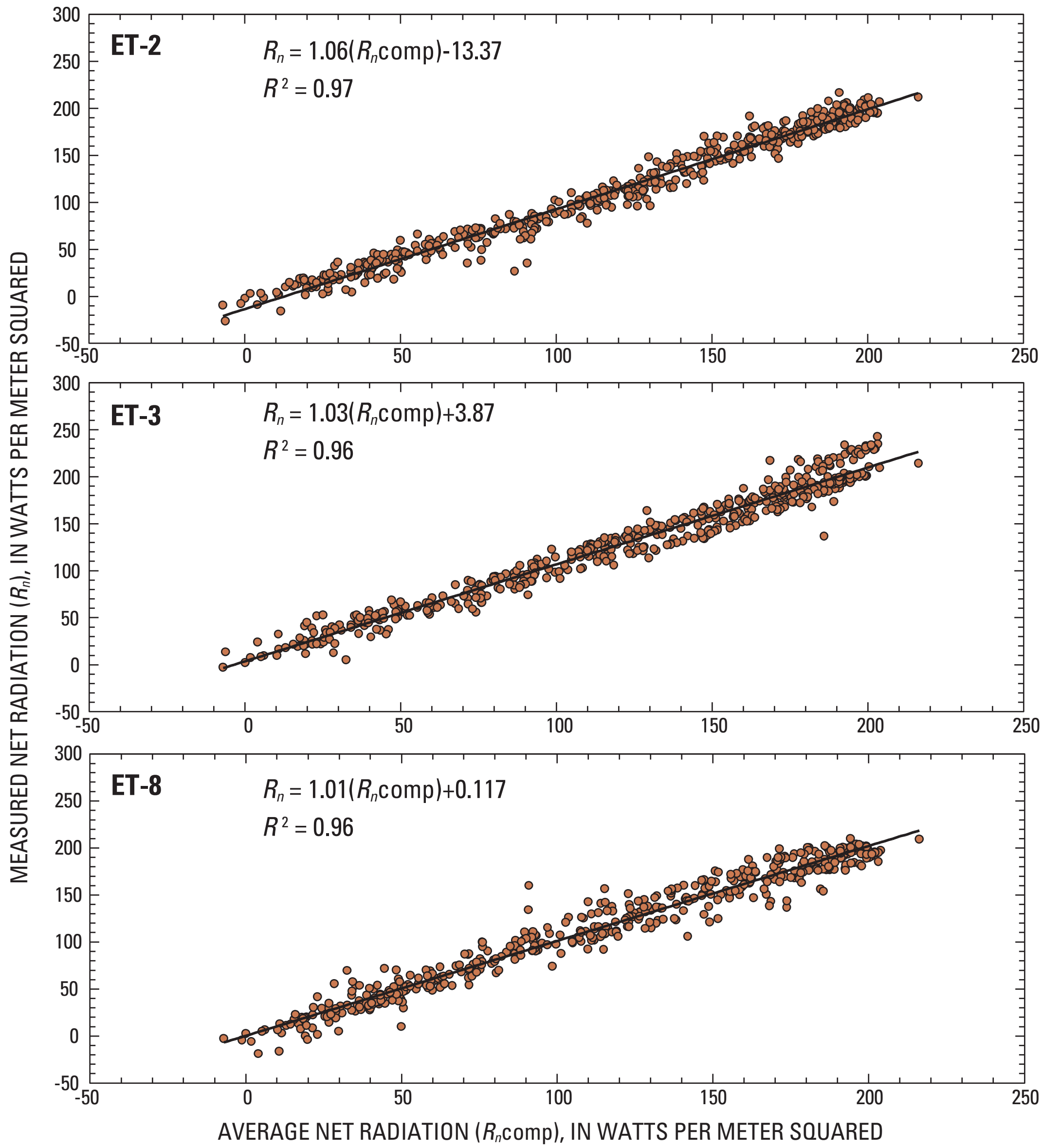
A. Bowen-ratio sites.

Figure 7. Relation between daily net radiation and natural log of evapotranspiration at Bowen-ratio and eddy-correlation sites, Carson Valley, Nevada and California.
R^2 is coefficient of determination.



B. Eddy-correlation sites.

Figure 7.—Continued.

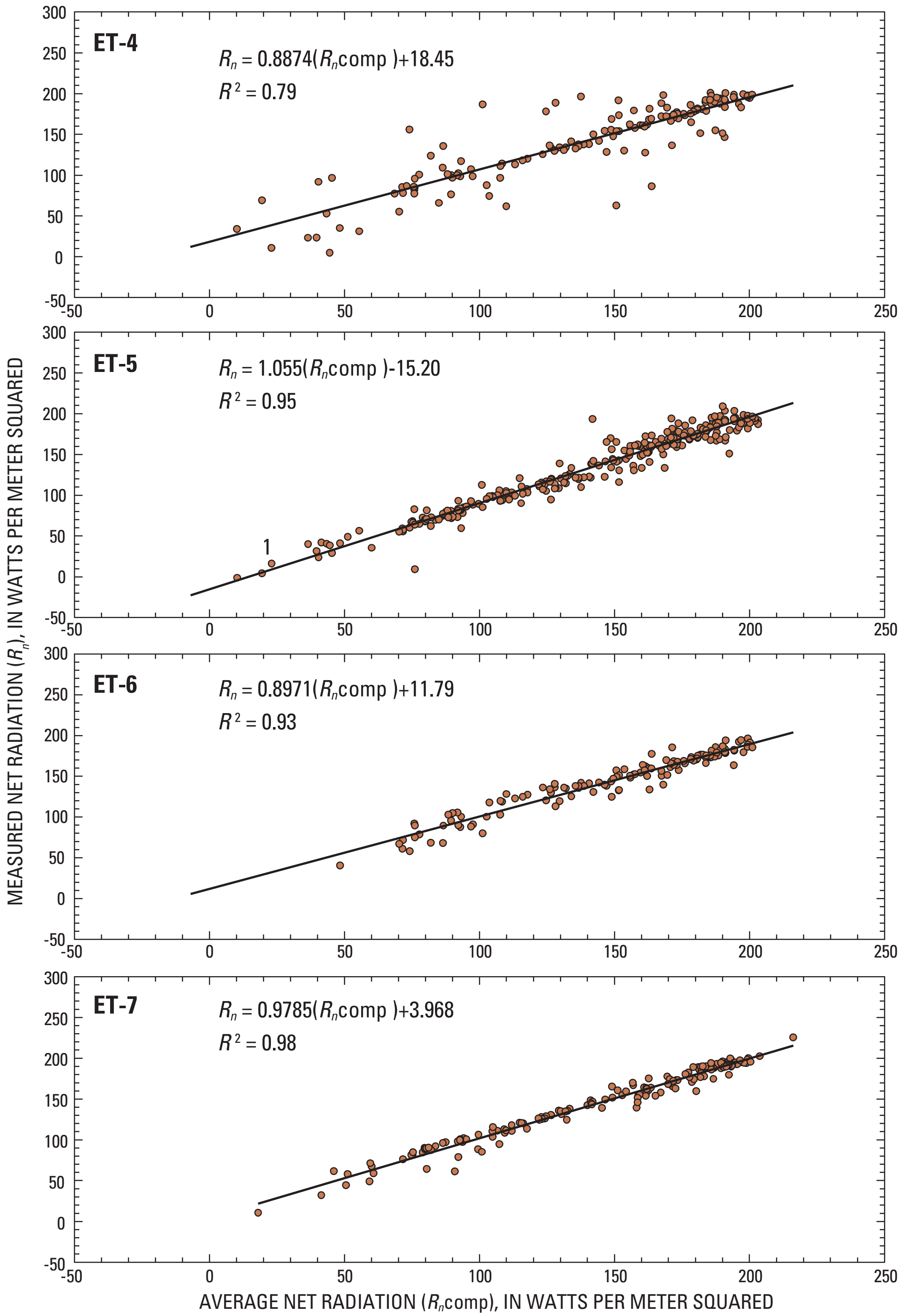


A. Bowen-ratio sites.

Figure 8. Relation between measured daily net radiation and average daily net radiation at Bowen-ratio sites ET-2, ET-3, and ET-8, and eddy-correlation sites, Carson Valley, Nevada and California. R^2 is coefficient of determination.

A natural-log transformation of ET was necessary because the variance in daily ET increases as net radiation increases. Variations in the natural log of daily ET generally are explained reasonably well by variations in daily net radiation. Residual plots for all sites, except site ET-8, indicate that the residuals are distributed randomly and show no distinct patterns (fig. 9). The residual plot for site ET-8 exhibits a typical ‘horn’ pattern signifying heteroscedasticity or nonconstant variance (Helsel and Hirsch, 1992, p. 231). Bias that may be associated with application of the equation for site ET-8 is negligible because of the very short periods of predicted ET estimates (fig. 10*A*).

The two methods provided reasonable values of daily ET for periods of missing data (no large increases or decreases in predicted values compared to measured values) at all sites except during winter months at site ET-5 on irrigated alfalfa. Predicted daily ET at site ET-5 during winter months was unreasonably high. For this reason, average daily ET rates measured at the other alfalfa site, ET-2, were substituted for each month with missing data at site ET-5 (November 2003–April 2004) to obtain estimates of daily ET for water year 2004 (appendix A). The substitution of values from site ET-2 during winter months had little effect on estimates of the total annual ET at site ET-5, because ET rates were low during winter compared to summer. Daily rates of estimated and predicted ET for all sites are listed in appendix A.



B. Eddy-correlation sites.

Figure 8.—Continued.

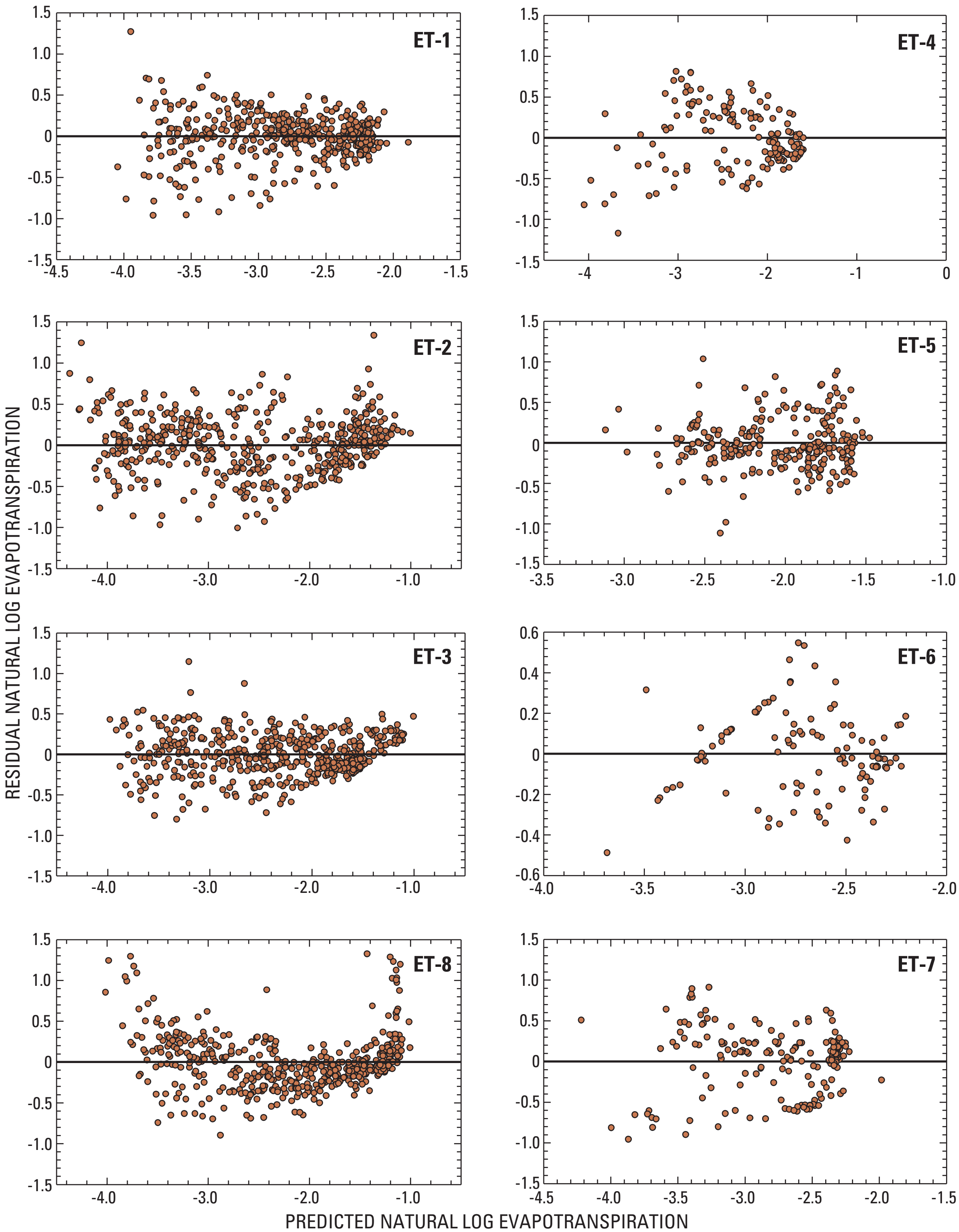
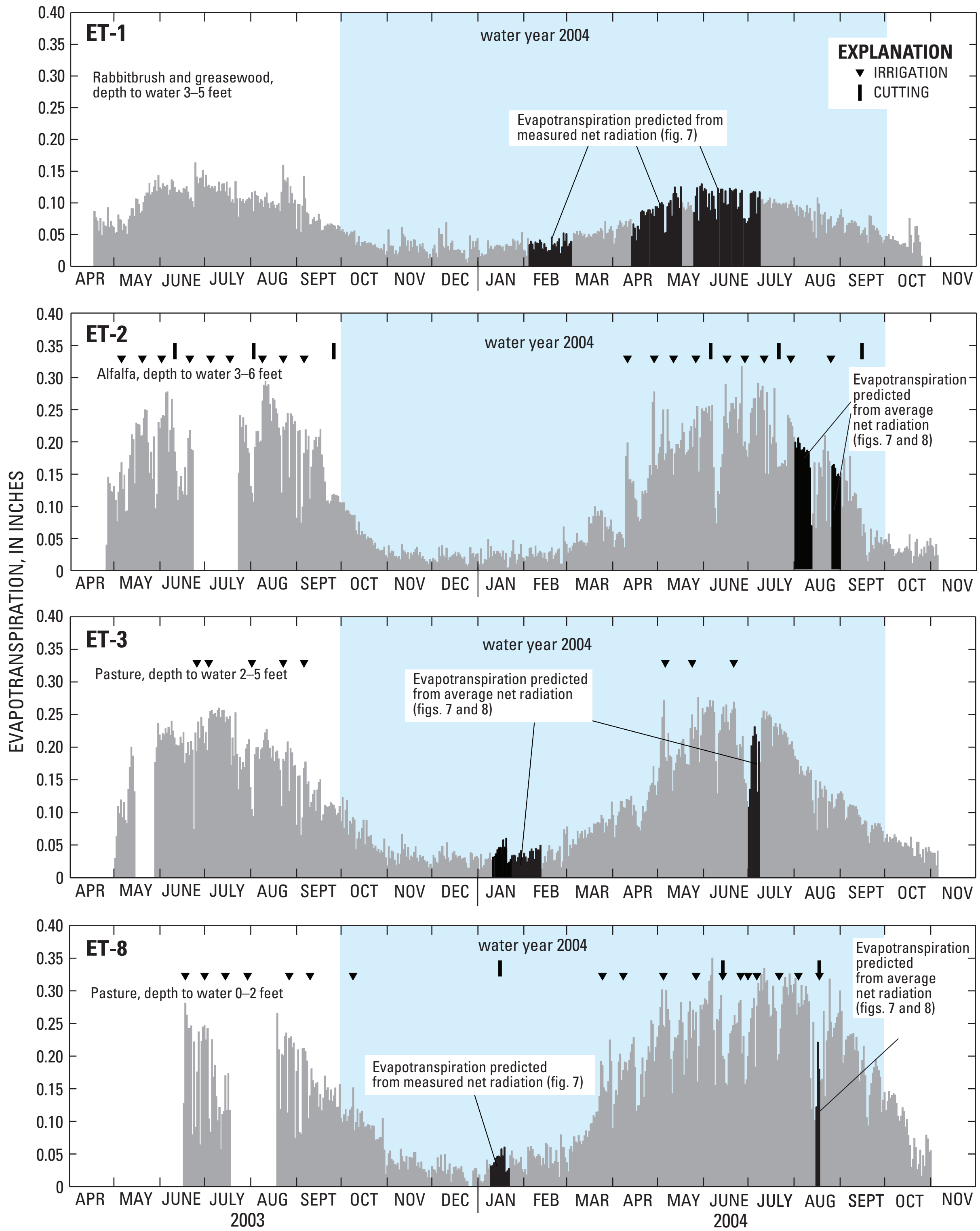
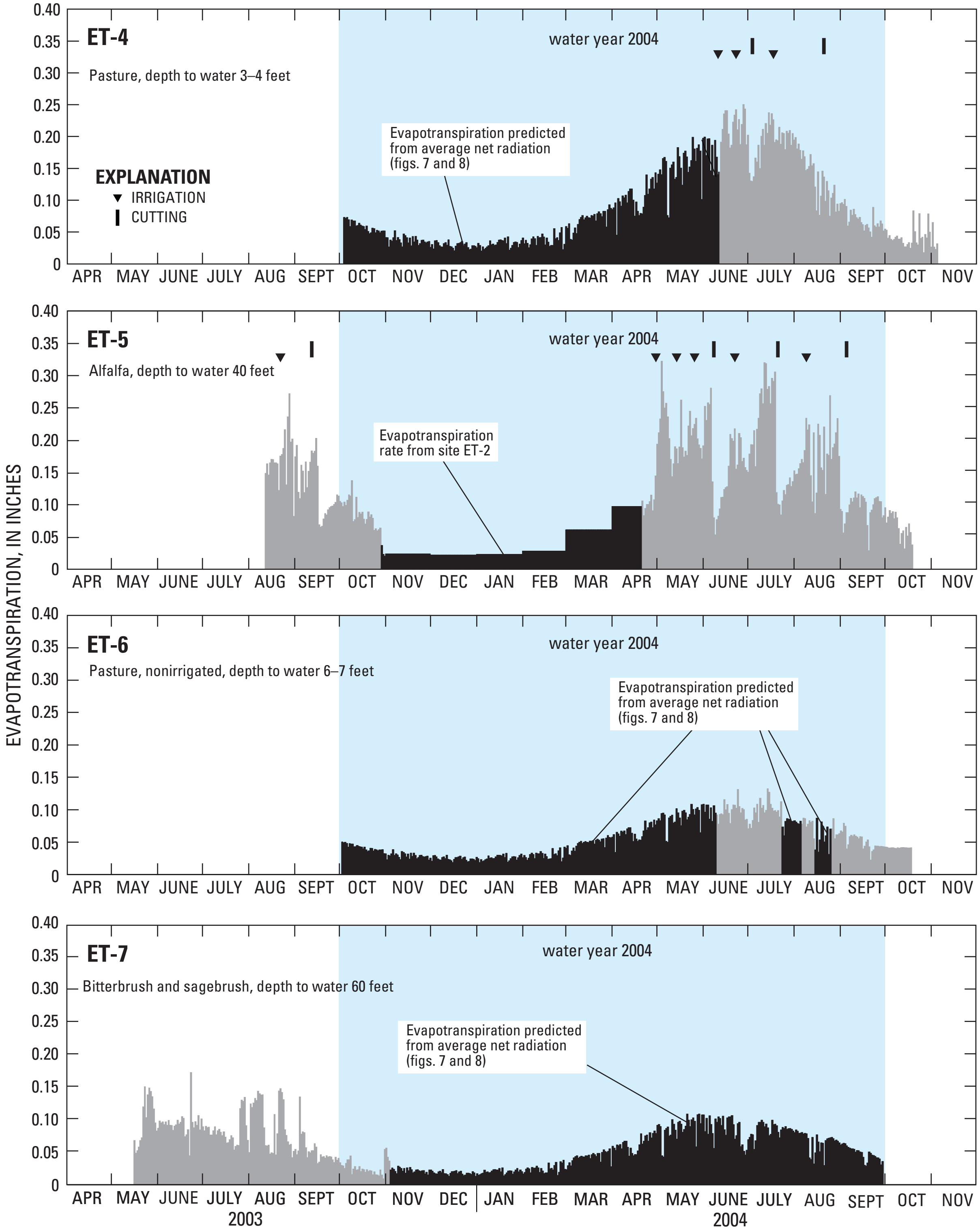


Figure 9. Residual and predicted natural log of evapotranspiration, Carson Valley, Nevada and California.



A. Bowen-ratio sites.

Figure 10. Estimated and predicted daily evapotranspiration rates for Bowen-ratio and eddy-correlation sites, Carson Valley, Nevada and California.
Depth to water is given in feet below land surface.



B. Eddy-correlation sites.

Figure 10.—Continued.

Results

Daily and monthly ET computed for the sites are listed in appendix A. Plots of daily estimated and predicted ET from April 2003 through November 2004 are shown in figure 10. To obtain the annual ET estimate for each site, the daily estimated and predicted ET was summed for water year 2004 (October 1, 2003–September 30, 2004; table 2).

The more complete data sets for the Bowen-ratio sites (fig. 10*A*) clearly indicate the seasonal variation in ET, with daily winter ET being less than 0.05 in. and daily summer ET being greater than 0.25 in. Variations in daily ET are caused by localized cloud cover and precipitation, and by the timing of irrigation and crop cuttings at the pasture and alfalfa sites (ET-2, ET-4, ET-5, and ET-8; fig. 10). The greatest daily ET, 0.35 in., was estimated at site ET-8 between irrigation applications in early June 2004.

Application of irrigation water and crop cuttings produces cyclic patterns in daily ET. These patterns are most pronounced at site ET-5, where the depth to water is about 40 ft below land surface and the only source of water for ET is applied irrigation water and precipitation (fig. 10*B*). Precipitation during the summer 2004 growing season near site ET-5 was 1.2 in. at Minden (National Climate Data Center, written commun., 2005). The average daily ET at site ET-5 for the 5 days prior to the June 8, 2004, crop cutting was 0.023 in. and averaged 0.08 in. for the following 5 days (fig. 10*B*, appendix A). Daily ET steadily increased following each cutting, with greater increases in daily ET following irrigation. In contrast, at sites ET-2, ET-4, and ET-8, where depth to water is about 5 ft or less, changes in daily ET resulting from irrigation and cuttings are much less pronounced (fig. 10*A*). At these sites, ET rates also decrease abruptly after cutting; however, ET recovers to precutting rates relatively rapidly because the shallow water table provides a constant source of readily available water. On average, about 7–10 in. of water is consumed by ET between crop cuttings.

ET from native vegetation and nonirrigated pasture sites is considerably less than ET from irrigated sites. The lowest annual ET in 2004, only 1.5 ft (table 2), was at site ET-7 in a stand of bitterbrush and sagebrush where the depth of water is about 60 ft. The annual ET for nonirrigated pasture at site ET-6 was slightly greater at 1.7 ft. Annual ET from rabbitbrush and greasewood at site ET-1 was 1.9 ft. During winter, daily ET rates at this native-vegetation site generally were greater than daily rates at cropland sites irrigated during summer. In comparison, annual ET from irrigated sites is close to 3.0 ft, except at site ET-8 where the annual ET was 4.4 ft. At site ET-8, the water table is 2 ft or less below land surface and ET likely was higher because of the shallow water table.

Estimates of ET made during this study are similar to most previous estimates. Estimates of annual ET from stands of rabbitbrush and greasewood were 1.3 ft in Ruby Valley, Nev. (Berger and others, 2001, p. 16), and from 0.8 to 2.0 ft in Owens Valley, Calif. (Nichols, 2000, p. 7). Annual ET from irrigated crops, often called crop consumptive use, were estimated for Carson Valley by Ball (1970), Guitgens and Mahannah (1972), Spane (1977), Pennington (1980), and Maurer (1986). Pennington (1980, p. 46–53) used meteorological data collected at three locations in Carson Valley, 1974–77. The data were used to evaluate seven different empirical methods developed by the Soil Conservation Service and to estimate crop-consumptive use ranging from 2.8 to 3.7 ft. Using methods similar to that of Pennington (1980), Guitgens and Mahannah (1972, p. 12) estimated potential ET rates to range from 2.9 to 4.9 ft for study plots on pasture grasses and alfalfa in 1972 and 1973, respectively; Spane (1977, p. 89 and 91) estimated annual rates ranging from 3.5 to 4.0 ft for 1973 and 1974, respectively; and Ball (1970, p. 41) estimated a rate of 2.5 ft. Maurer (1986, p. 42) estimated an annual rate of 3.5 ft for irrigated lands.

Table 2. Monthly and annual evapotranspiration at eight sites, Carson Valley, Nevada and California, water year 2004.

[Site locations are shown in figure 2. Data were summarized from appendix A]

Site No.	Vegetation type	Oct. 2003	Nov. 2003	Dec. 2003	Jan. 2004	Feb. 2004	Mar. 2004	Apr. 2004	May 2004	June 2004	July 2004	Aug. 2004	Sept. 2004	Total ET, water year 2004 (inches)	Total ET, water year 2004 (feet)
ET-1	Rabbitbrush/greasewood	1.20	0.94	0.96	0.94	1.00	1.56	2.12	3.06	3.13	3.13	2.43	1.81	22.3	1.9
ET-2	Flood-irrigated alfalfa	1.83	.73	.69	.72	.80	1.90	3.59	6.32	6.44	6.78	4.86	2.46	37.1	3.1
ET-3	Flood-irrigated pasture	2.01	.98	.97	.97	1.12	2.25	3.31	6.34	6.62	6.71	4.21	2.63	38.1	3.2
ET-4	Flood-irrigated pasture	1.72	1.05	.88	.95	1.12	2.20	3.16	4.84	5.86	5.99	3.88	1.93	33.6	2.8
ET-5	Flood-irrigated alfalfa	2.63	.73	.69	.72	.80	1.90	2.98	6.47	4.96	6.11	5.50	2.88	36.4	3.0
ET-6	Nonirrigated pasture	1.25	.86	.75	.80	.89	1.52	1.99	2.78	2.89	2.91	2.11	1.47	20.2	1.7
ET-7	Bitterbrush/sagebrush	.75	.76	.57	.61	.70	1.32	1.83	2.71	2.64	2.59	2.09	1.46	18.0	1.5
ET-8	Flood-irrigated pasture	2.93	1.09	.77	1.05	1.31	3.21	5.38	7.45	7.29	8.63	7.37	5.72	52.2	4.4

ET estimates for plant communities in Carson Valley may not be applicable to drier areas in Nevada. Pennington (1980) showed that crop-consumptive use increased in drier areas of Nevada from about 4.1 ft near Fallon to greater than 5.8 ft near Pahrump.

Uncertainty in Evapotranspiration Estimates

ET estimates made using data from Bowen-ratio stations generally are greater than estimates from eddy-correlation stations (Dugas and others, 1991, p. 13; Tomlinson, 1996, p. 55). A comparison of ET estimates using the Bowen-ratio and eddy-correlation methods from this study is available for the period when three stations were colocated at site ET-8 for 65 days from April to June 2004 (table 1; Tumbusch and Johnson, 2005). For periods of common data collection, ET estimates from the Bowen-ratio station were 11 to 15 percent greater than those from the eddy-correlation stations. Tumbusch and Johnson (2005) concluded that estimates from the Bowen-ratio station likely overestimated actual ET because the method did not account for energy lost as heat storage in irrigation water moving across the field. The field was irrigated four times during the 65-day period (fig. 10, table 1) and, if irrigation water was applied for about 2 days during each irrigation, surface water was present for about 8 days, or about 12 percent of the period of data collection. Thus, the presence of flood-irrigation water in the fields may account for the differences in the amount of ET estimated by the two methods.

Dugas and others (1991, p. 13) and Tomlinson (1996, p. 55) also showed that eddy-correlation ET estimates were less than the Bowen-ratio estimates, but with discrepancies of 20–30 percent between the two estimates. The study by Dugas and others (1991) was for irrigated wheat, whereas the study by Tomlinson (1996) was for grasslands. Contrary to results from this study, Dugas and others (1991, p. 16) concluded that ET values from the eddy-correlation stations likely underestimated the actual ET. Thus, it is not clear which ET estimate may be most representative of actual ET for the data collected in Carson Valley.

Weighing lysimeters may provide a more accurate measure of ET than micrometeorological methods if conditions within the lysimeters are representative of those outside the lysimeter. Tomlinson (1996, p. 34) reported that, when conditions in the lysimeters are representative of those outside the lysimeters, annual estimates of ET made using Bowen-ratio instruments were about 5 percent greater than those made using weighing lysimeters. Tomlinson (1995, p. 15) also analyzed the error of individual instruments used to calculate ET with micrometeorological methods and concluded that the cumulative error of all instruments may be about 12 percent.

For purposes of this study, estimates of ET from Bowen-ratio and eddy-correlation stations are assumed to be comparable and likely accurate to within about 12 percent of the actual ET for sites where measured daily ET is available for most of water year 2004 (sites ET-1, ET-2, ET-3, ET-5, and ET-8). For sites where daily ET was predicted for a large part of water year 2004 (sites ET-4, ET-6, and ET-7), the uncertainty was estimated from the 95-percent confidence interval for the daily values predicted by equations in figure 7. The uncertainty ranged from +30 to –20 percent for sites ET-4 and ET-6, and from +50 to –40 percent for site ET-7.

ET measured in water year 2004 may be less than what would be measured in a year with average, or above average precipitation. Water year 2004 was considerably drier than average, being the sixth consecutive year of a drought. Precipitation in 2004 at Minden was 5.5 in. (National Climate Data Center, written commun., 2005), almost 3 in., or 36 percent less than the 30-year average of 8.4 in. for 1971–2000 (fig. 4). In average or wet years, depth to water beneath the sites likely would be less than that in 2004. Because greater ET was measured at site ET-8 where the depth to water was less than at the other sites, the depth to water is an important factor controlling the ET rate.

Recharge from Precipitation on the Northern and Eastern Sides of Carson Valley

For this study, recharge from precipitation on the northern and eastern sides of Carson Valley was estimated using soil-chloride data from test holes. The northern and eastern sides of Carson Valley lying above the valley floor cover a large area, about 90,000 acres (Maurer and others, 2004, p. 14). Because of the large area, recharge from precipitation in this part of the valley may be a significant component in the valley's water budget. Additional work using other methods is underway currently (2005) to evaluate recharge from precipitation in other parts of the valley.

Concentration of chloride in pore water in the unsaturated zone (soils above the water table) has been used by many workers to estimate recharge in arid and semi-arid environments (Allison and Hughes, 1983; Scanlon, 1991; Phillips, 1994; Prudic, 1994; Prych, 1998; Stonestrom and others, 2003). Distribution of chloride in soil profiles develops as small quantities of chloride are deposited onto the land surface by precipitation (wet fall) and dust (dry fall), and move downward with infiltrating precipitation. As the water moves downward, it is partly lost to ET, and, assuming none discharges to streams, the remainder percolates to the water table as recharge. Chloride is not taken up through the roots of most plants and becomes concentrated by ET near the base of the root zone (Allison and others, 1994, p. 8).

Ideally, the steady-state profile of chloride concentration with depth will show concentrations increasing from land surface to a maximum concentration near the base of the root zone (Wood, 1999, p. 3). Beneath the point of maximum concentration but above the water table, chloride concentrations decrease to a relatively constant value that is related to the rate of recharge. In areas of active recharge, the chloride concentrations in pore water within the profile and at the peak are low, generally less than 100 mg/L. Whereas, in areas where most precipitation is discharged by ET and recharge is minimal, chloride concentrations are high, often exceeding 1,000 mg/L (Stonestrom and others, 2003, fig. 6; Scanlon, 2004, p. 245).

A fundamental assumption for using soil-chloride profiles to estimate recharge from precipitation is that precipitation is the only source of chloride to the pore water. Application of irrigation water or overland flow of precipitation onto or away from a site are potential sources of, or sinks for chloride. Such sources or sinks should be negligible for the method to provide realistic recharge estimates (Wood, 1999, p. 2). On the steeply sloping alluvial fans along the western side of Carson Valley, overland flow can take place during infrequent periods of intense rainfall, thus the method cannot be applied reasonably. On the valley floor, large volumes of surface water are used for flood irrigation. Although the method also can be used to calculate recharge rates of surface water applied for irrigation (Wood, 1999), the shallow depth to ground water beneath most of the valley floor and infiltration of surface water likely has flushed pore-water chloride into the ground water. For these reasons, estimates of recharge from precipitation using soil-chloride profiles on the western alluvial fans and on the valley floor where flood irrigation is applied were not attempted. On the northern and eastern sides of Carson Valley, numerous hills with relatively flat tops, and broad drainage divides between ephemeral watersheds provide locations where overland flow is negligible and the soil-chloride method may be applied.

Methods Used

Test holes were drilled at nine locations from which samples were collected and analyzed (fig. 2, table 3, appendix B). Locations for the test holes were selected at broad topographic divides and the flat tops of low-lying hills to provide a geographic distribution of data over the northern and eastern sides of the valley.

The holes were drilled using the ODEX hammer method with an Ingersoll Rand Th-75 drill rig (Hammermeister and others, 1985). Cores were collected every 2 ft using a 4-in. inside-diameter core barrel. The core barrel was 24 in. long with a 4-in. core tip and fitted with four 6-in. long aluminum sleeves. The core barrel was driven 2 ft by a percussive air hammer and removed from the test hole. Sleeves were extracted from the core barrel using a hydraulic ram and immediately capped and taped to retain the soil and in-situ soil moisture. Nominal 6-in. flush-thread ODEX casing was then driven to the bottom of the cored interval to hold the test hole open as drilling advanced. At some test holes, cores could not be obtained because of gravelly zones (appendix B). These zones were drilled using the ODEX casing and a percussive air hammer, and coring resumed once finer-grained sediments were encountered. When drilling reached about 30 ft below land surface, or when conditions prevented further drilling, the ODEX casing was removed from the test hole and the test hole was back-filled with native material.

The core samples were analyzed in the laboratory to determine gravimetric and volumetric water content and soil-bulk density using procedures from Donahue and others (1977). Soil-water extractions were collected from selected intervals for chloride analysis. The core samples were weighed in the lab prior to removal of caps and tape to determine the wet weight, then the caps and tape were removed and retained, and the core sleeve and sediment were oven dried at 125ºC for about 1 week. After drying, the core sleeve and sediment

Table 3. Location, altitude, test-hole depth, and description of vegetation at nine soil-chloride test holes, Carson Valley, Nevada.

[Test hole locations are shown in figure 2. **Latitude and longitude:** Geographic coordinates referenced to North American Datum of 1927 (NAD 27), in degrees, minutes, and seconds. **Land-surface altitude** is referenced to National Geodetic Vertical Datum of 1929 (NGVD 29), estimated from U.S. Geological Survey 1:24,000-scale topographic maps]

Test hole No.	West latitude	North longitude	Land-surface altitude	Test hole depth (feet)	Description of vegetation
CL-1	39°05’57”	119°46’13”	4,930	22.7	Sagebrush and bitterbrush is 4–5 feet tall.
CL-2	39°03’48”	119°41’54”	5,122	30.8	Sagebrush and bitterbrush is 5–6 feet tall.
CL-3	39°01’55”	119°40’03”	5,265	28.3	Sagebrush is 1–2 feet tall.
CL-4	39°02’20”	119°37’08”	5,727	25.0	Sagebrush is 1–2 feet tall.
CL-5	39°03’58”	119°33’19”	6,900	12.5	Sagebrush and Pinyon pine.
CL-6	38°59’37”	119°38’02”	5,280	29.8	Sagebrush is 1 foot tall.
CL-7	38°56’35”	119°37’49”	5,420	30.0	Sagebrush is 3–4 feet tall.
CL-8	38°55’13”	119°35’31”	5,640	30.4	Bare soil.
CL-9	38°53’52”	119°36’03”	5,700	30.2	Sagebrush and Pinyon pine.

along with the original caps and tape were reweighed to determine the dry weight. Dry net weight was calculated as the dry weight minus the weight of the core sleeve, caps, and tape. Gravimetric water content was calculated as the difference between wet and dry weight, divided by the dry net weight. Soil-bulk density was calculated as the dry net weight divided by the volume of the core sleeve. Volumetric water content was calculated as the gravimetric water content multiplied by the soil-bulk density.

Core material was selected for extraction at about 1-ft intervals at each test hole. Soil-water extractions were obtained by mixing about 200 g of oven-dried sediment from selected cores with a similar weight of deionized water. Exact weights of sediment and water were measured to determine the water-to-soil ratio. The soil-water mixtures were shaken periodically for 48 hours, after which the supernatant liquid was decanted into a syringe. The syringe then was placed in a mechanical press and the supernatant liquid was filtered through a 0.45 microfilter into 20 mL glass vials. Replicate samples were collected for each extraction. Samples were analyzed for chloride concentration using an ion chromatograph in Menlo Park, Calif. (K.C. Akstin, U.S. Geological Survey, written commun., March 2004). Pore-water concentrations were calculated by multiplying the chloride concentration of the supernatant liquid by the water-to-soil ratio, dividing that by the gravimetric water content, and multiplying the result by the density of pore water (1 mg/cm^3).

The rate of recharge was calculated from the chloride concentration in pore water below the root zone and above the water table, and the chloride concentration in precipitation (including wet fall and dry fall) using the equation from Allison and others (1985):

$$q_w = (C_p P) / C_{pw}, \qquad (2)$$

where

q_w is the recharge rate, in inches per year;

C_p is the chloride concentration of precipitation, in milligrams per liter;

P is the rate of precipitation, in inches per year; and

C_{pw} is the average chloride concentration of pore water below the root zone, in milligrams per liter.

Important assumptions made in applying equation 2 are that (1) land surface is neither aggrading nor degrading; (2) atmospheric deposition is the only source of chloride and is constant through time; (3) chloride moves with water (negligible dispersion) steadily and uniformly downward; and (4) pore-water chloride concentrations below the root zone are in equilibrium with the chloride flux at land surface. In other words, there are no sources or sinks for chloride and chloride is not accumulating between land surface to depths below the root zone (Prudic, 1994, p. 10; Wood, 1999).

Assumption 1 is met because soil samples were collected from test holes drilled on the flat tops of topographic divides or hills where the land surface cannot aggrade and degradation is slow relative to the recharge process. Assumption 2 that chloride deposition occurs at a constant rate is seldom met in the real world. For example, precipitation is not constant through time. In Carson Valley, about 70 percent of the annual precipitation falls from November through March and only about 10 percent falls from July through September (Owenby and Ezell, 1992, p. 15). Chloride from dry-fall deposition accumulates on land surface during summer months and is incorporated into early winter precipitation (Dettinger, 1989, p. 60–61). This, coupled with ET during summer, likely causes a spike in the chloride concentration as the accumulated chloride infiltrates. Such a process could produce peaks in the chloride profile, followed by troughs as more dilute water enters the soil later in winter and early spring. Multiple peaks also could be caused by geologic heterogeneities or macropores that allow fast percolation rates through a small part of the sediments (Wood, 1999, p. 2).

The validity of assumptions 3 and 4 can be tested by plotting cumulative water content against cumulative chloride content. If assumptions 3 and 4 are valid, such plots should indicate relatively linear increases (Stonestrom and others, 2003, p. 14–15). For each depth interval sampled, cumulative chloride was calculated by summing the product of chloride concentration and water content, and cumulative water content was calculated by summing the volumetric water content. Values for individual samples were assumed to represent the entire interval defined by the midpoints between samples and the top or bottom of the holes.

Estimating recharge from precipitation using equation 2 requires estimates of annual precipitation at the data-collection sites. Annual precipitation at the data-collection sites was estimated based on their altitude using the following relation from Maurer and Halford (2004, p. 29):

Average annual precipitation, in inches
= 0.0027 × altitude, in feet – 5.3646.

Results

Vegetation in the area where soil-chloride test holes were drilled consists of low-lying and sparse sagebrush. Lithologic descriptions of soils penetrated by the test holes (appendix B) indicate that only sites CL-1 and CL-2 are comprised of unconsolidated sediments for the entire depth drilled. At the remaining sites, soils are semiconsolidated at depths ranging from 1.5 to 22 ft. Test-hole CL-5, near the crest of the Pine Nut Mountains, penetrated semiconsolidated sediments at 1.5 ft and weathered bedrock at 4.7 ft; drilling was suspended at a total depth of 12.5 ft. The semiconsolidated soils likely are Tertiary sediments exposed at land surface at sites CL-4, CL-6, and CL-7 and, based on the descriptions, present at shallow depths at sites CL-3, CL-5, CL-8, and CL-9 (fig. 3).

The semiconsolidated sediments are relatively fine-grained consisting primarily of silt or clay with mixtures of sand and gravel except at test holes CL-7 and CL-9. At test hole CL-7, the semiconsolidated sediments are coarse grained to a depth of about 25 ft, and at test hole CL-9 become coarse grained at a depth of about 22 ft.

Plots of chloride concentration in pore water against depth below land surface (fig. 11) show chloride concentrations peak at 3–4 ft below land surface at sites CL-1, CL-2, and CL-5, whereas peaks at the remaining sites occur between depths of about 5 to 18 ft. The depth of the peak concentration indicates the depth to which plant roots likely extend at each site; chloride in pore water is concentrated by ET above the depths of the peaks. Peak concentrations are lowest at sites CL-1, CL-2, and CL-7, about 25, 140, and 260 mg/L, respectively, whereas peak concentrations at the remaining sites ranged from about 560 to almost 5,000 mg/L. Recharge rates in the Chihuahuan Desert in Texas approach zero for concentrations greater than about 100 mg/L (Scanlon, 2004, p. 245). Thus, high chloride concentrations at the peaks for test holes CL-3 through CL-6 and CL-8 and CL-9 indicate that ET from native plants is effectively capturing most modern-day precipitation before it can percolate past the root zone to the water table.

Cumulative water and cumulative chloride contents were plotted against depth below land surface, and against each other for each test hole to evaluate the subsurface movement of water and chloride (figs. 12–14). Linear increases in cumulative water with depth are seen at test holes CL-5–CL-9 (fig. 12), which indicate a constant water content with depth at these test holes. At test holes CL-1–CL-4, changes in slope indicate changes in water content at various depths in the profile. At test hole CL-4, the slope changes at about 10 ft, which corresponds to a change from semiconsolidated silt, sand, and gravel to stiff, semiconsolidated clay where more water may be held in the sediment pores (appendix B). At this site, chloride concentrations also decrease at depths greater than 10 ft (fig. 11) indicating that at depths less than 10 ft, ET from plants may be depleting the water content and increasing the chloride concentrations in the root zone. Cumulative water increases at test holes CL-1 and CL-2 at depths greater than about 14 and 18 ft, respectively, and conversely, at test hole CL-3 cumulative water decreases at a depth of about 8 ft. There are no apparent changes in lithology or changes in chloride concentrations at these depths in the three test holes, and the causes for the changes in slope are unknown.

Test holes CL-1, CL-2, and CL-7 (fig. 13) show relatively linear increases in cumulative chloride with depth, indicating chloride is being transported downward at a uniform rate throughout the depth of the test hole. Large changes in slope are seen at one or more points in the remaining test holes, indicating chloride is being concentrated near the base of the root zone where plants extract water, or that the chloride flux at land surface has changed. Changes of the chloride flux at land surface may be caused by changes in the volume of infiltrating precipitation. In test hole CL-3, the peak in chloride occurs at 12 ft (fig. 11) and cumulative chloride greatly increases from a depth of 5 to 12 ft (fig. 13), where very hard semiconsolidated sediments were encountered (appendix B). Plant roots may not be able to penetrate the semiconsolidated sediments at depths greater than 12 ft, extracting percolating water and causing chloride to accumulate in the soil from depths of 5 to 12 ft. Changes in slope at test holes CL-4 and CL-5 are similar near the chloride peaks (fig. 11) at depths of about 10 and 5 ft, respectively, which correlates with lithologic changes in the test holes (appendix B). Changes in cumulative chloride slope in test holes CL-6–CL-9 do not correspond to changes in lithology, but approximately correspond to depths having high chloride concentrations (figs. 11 and 13). Chloride concentrated at the base of the root zone by ET likely is the cause of these changes in slope and rapid increases in cumulative chloride.

A linear increase in cumulative water with cumulative chloride indicates that the data fits the assumption that pore-water chloride concentrations below the root zone are in equilibrium with the chloride flux at land surface (Stonestrom and others, 2003, p. 15). Relatively linear increases are shown at test holes CL-1, CL-2, and CL-7 (fig. 14). At the remaining test holes, a change in slope is evident at one or more depths, indicating that chloride is accumulating near the base of the root zone.

The time required for accumulation of the mass of chloride above a particular depth within a test hole, or throughout each test hole, can be estimated from the cumulative mass of chloride above the depth of interest, divided by the rate of chloride deposition at land surface (table 4; Phillips and others, 1988, p. 1882; Prudic, 1994, p. 10). The rate of chloride deposition is determined by multiplying the average annual precipitation at each site by the average chloride concentration of precipitation. Average annual precipitation at each site was determined from a relation between altitude and precipitation developed by Maurer and Halford (2004, p. 29) for the eastern side of Carson Valley. The average chloride concentration of precipitation used was 0.5 mg/L, as determined by Feth and others (1964, p. 35) from 79 snow samples collected in the Sierra Nevada, representing both wet and dry deposition.

The estimated time required for chloride accumulation to the depths of peak chloride concentrations is from 16 to 17 years at test holes CL-1 and Cl-2; 110 years at test hole CL-4; 175 years at test hole CL-9; 600 years at test hole CL-7; and more than 1,000 years at test holes CL-3, CL-5, CL-6, and CL-8 (table 4). The time required for chloride to accumulate at the bottom of the test holes is about 100 years at test holes CL-1 and CL-2; about 700 years at test hole CL-7, and ranges from 1,400 to almost 12,000 years at the remaining test holes (table 4).

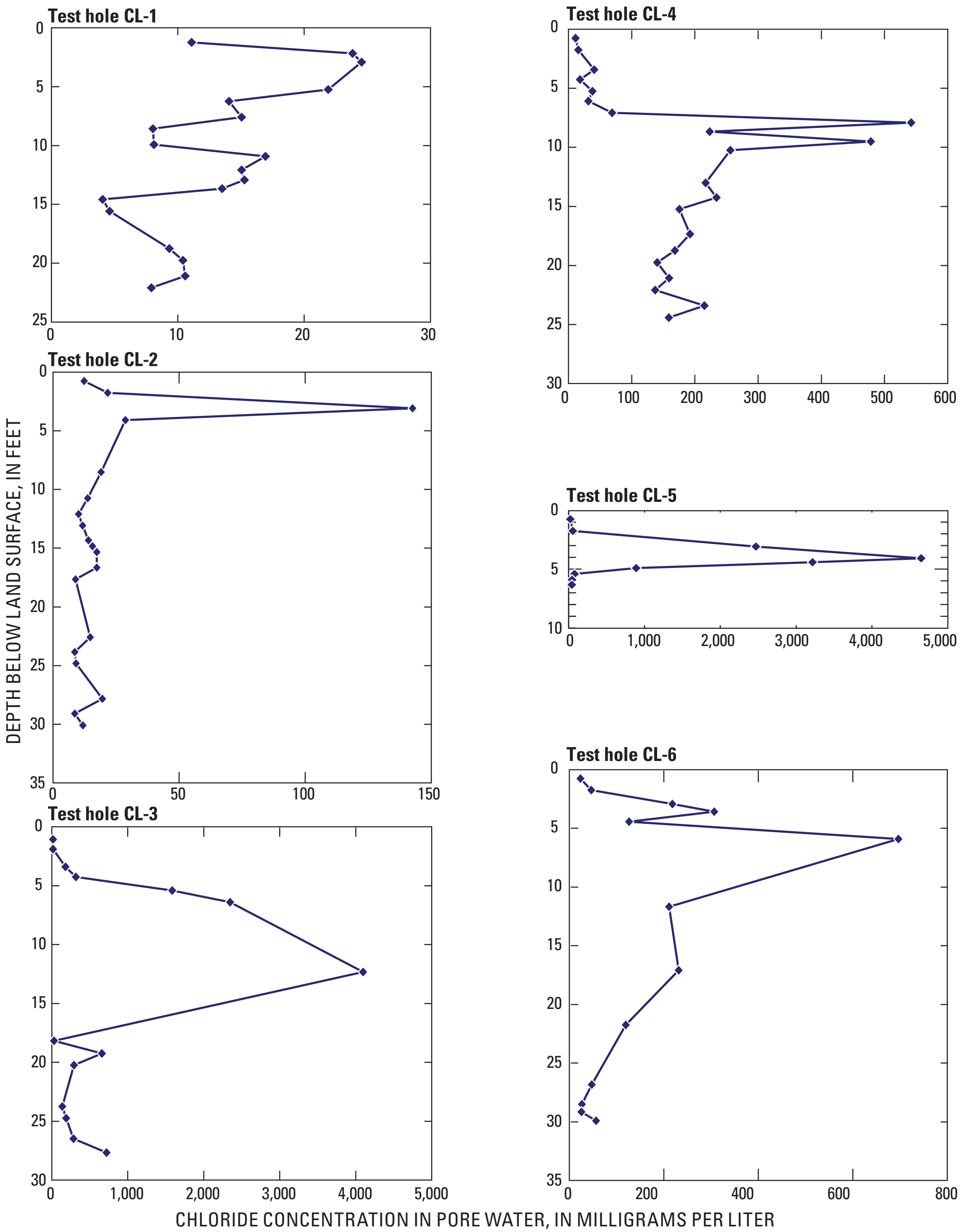


Figure 11. Relation between chloride concentration in pore water and depth for nine test holes in Carson Valley, Nevada and California.

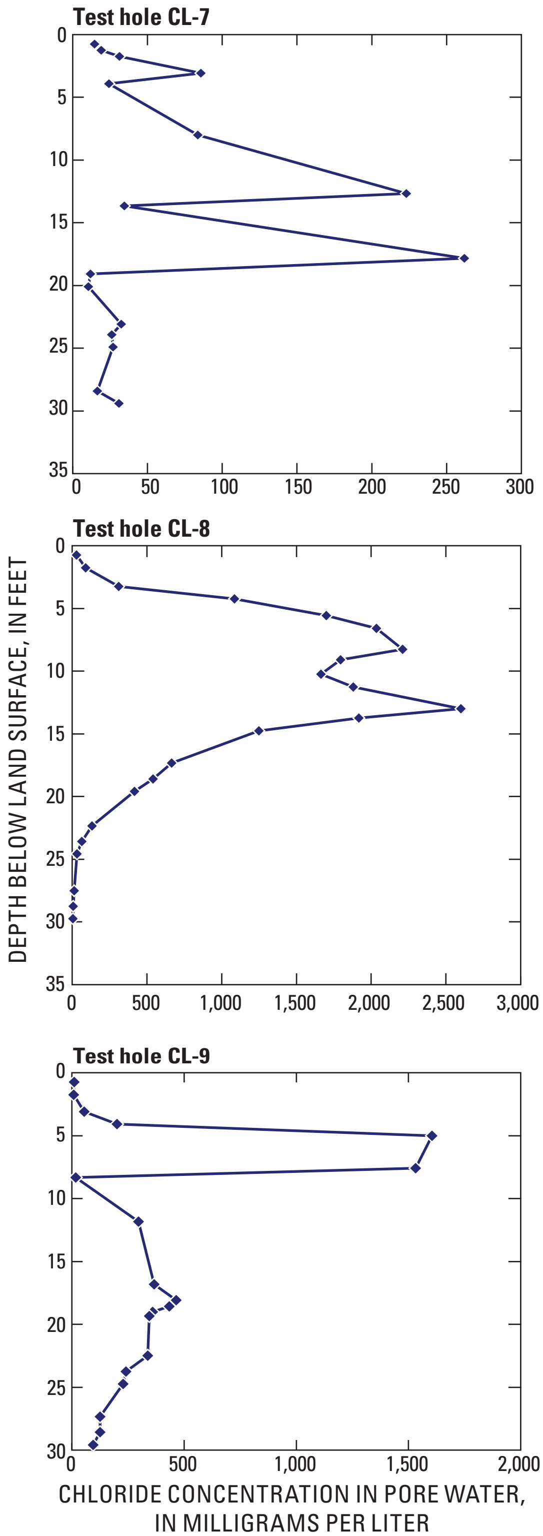


Figure 11.—Continued.

Other studies in the southwestern United States estimated that 10,000–15,000 years were needed to account for the measured accumulation of chloride in similar profiles (Scanlon, 2004, p. 245). Large peak chloride concentrations have been interpreted to signify a lack of recharge from modern-day precipitation and low chloride concentrations at depths greater than the peak concentrations have been attributed to recharge that occurred in wet climates near the end of the Pleistocene Epoch (more than 10,000 years ago; Phillips and others, 2004, p. 285; Scanlon, 2004, p. 244–245).

In Carson Valley, the estimated times for accumulation indicate that recharge from modern-day precipitation is taking place only at test holes CL-1 and CL-2, and possibly at test hole CL-7, if modern-day precipitation is defined as taking place over the past several hundred years. Modern-day precipitation at the other six test holes was consumed by ET and has not infiltrated beyond the base of the root zone. Accordingly, modern-day precipitation likely has not infiltrated to the water table to become recharge over much of the eastern side of Carson Valley where annual precipitation, vegetation, and lithology are similar to those at the six sites.

The long times required for accumulation of chloride to the base of the test holes (about 30 ft below land surface) indicate that water that began infiltrating during the Pleistocene Epoch still could be slowly draining through unsaturated sediments beneath the root zone to recharge the water table. Estimates of the rate of recharge from the Pleistocene Epoch could be made if the rates of precipitation and chloride deposition were known for that time period. However, the rates of chloride deposition for that time period are not well known and such estimates would be highly speculative (Stonestrom and others, 2003, p. 16).

The rate of recharge from modern-day precipitation at test holes CL-1, CL-2, and CL-7 can be estimated using the average annual precipitation at the sites (table 4), the chloride concentration of precipitation (0.5 mg/L; Feth and others, 1964), and the average concentration of chloride in pore water near the bottom of the test holes (table 4, footnotes 1–3). Average chloride concentrations in pore water near the bottom of test holes at these three sites were 7.8 mg/L at test hole CL-1, 11.6 mg/L at test hole CL-2, and 22.2 mg/L at test hole CL-7 (table 4). Corresponding annual rates of recharge were 0.04 ft at test hole CL-1, 0.03 ft at test hole CL-2, and 0.02 ft at test hole CL-7 (table 4).

Test hole CL-1 is in gravel deposits that cap the Indian Hills area, and test hole CL-2 is in eolian sand deposits that cover areas north of the Johnson Lane area (figs. 2 and 3, appendix B). At these locations precipitation infiltrates through the coarse-grained and well-sorted sediments to depths below the root zone. Recharge from modern-day precipitation may be taking place at similar rates in other areas where these gravel and eolian sand deposits are mapped. Test hole CL-7 is in Tertiary sediments near the eastern edge of basin-fill deposits in the Fish Spring Flat area (figs. 2 and 3). Based on the results from the other test holes, the recharge rate estimated for test hole CL-7 is unlikely taking place over a large area.

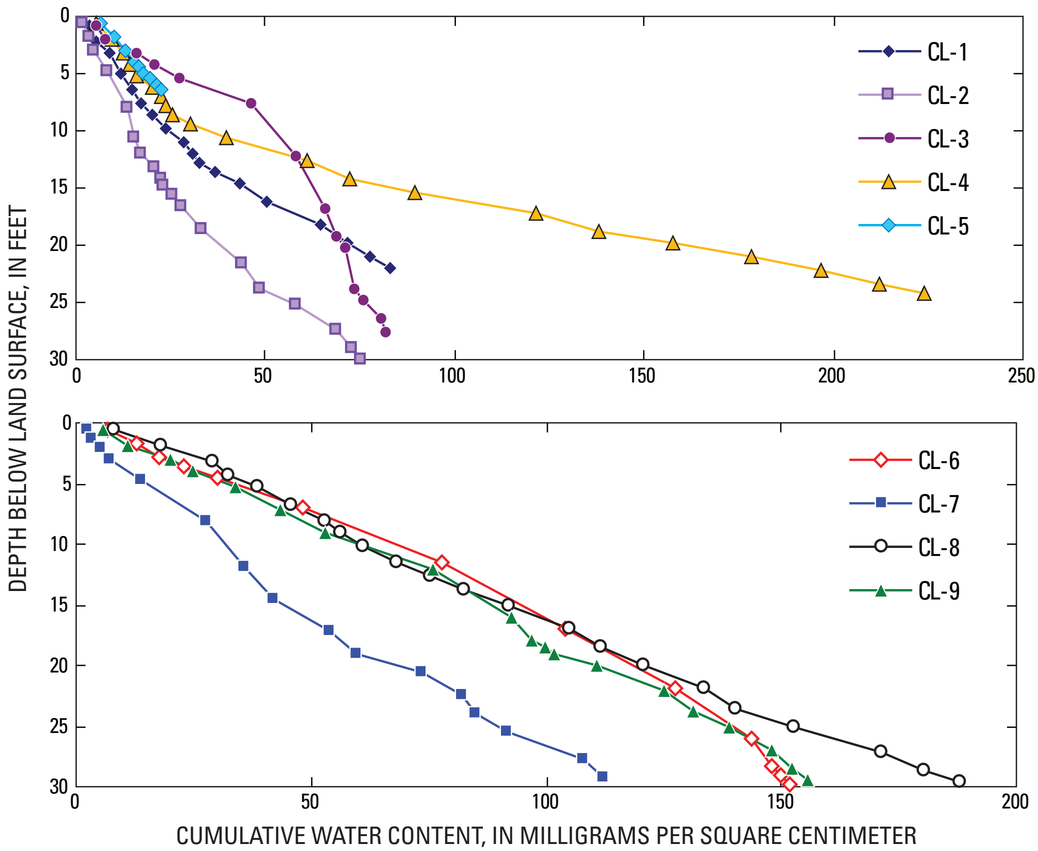


Figure 12. Relation between cumulative water content and depth for nine test holes in Carson Valley, Nevada and California.

Uncertainty in Estimated Recharge Rates

Uncertainty in estimated recharge rates is related directly to uncertainties in annual precipitation and chloride concentration in precipitation. Uncertainty associated with the estimate of annual precipitation made using the altitude relation is about 15 percent (Maurer and Halford, 2004, p. 37). Other soil-chloride studies in Nevada used a value of about 0.4 mg/L for the chloride concentration in precipitation (Dettinger, 1989, p. 63; Berger and others, 1997, p. 46; Maurer and Thodal, 2000, p. 24).

Values from 0.4 to 0.6 mg/L represent a reasonable range of chloride concentrations in bulk precipitation in Nevada. Average chloride concentration for precipitation at 74 sites sampled in Nevada was 0.4 mg/L (Dettinger, 1989, p. 62). Average chloride concentration for bulk precipitation at eight sites in Nevada was 0.6 mg/L, but the value could be overestimated because of localized remobilization of dry fall between periods of precipitation at the sampling sites (Dettinger, 1989, p. 62). The chloride concentration of bulk precipitation ranged from 0.07 to 1.3 mg/L (average 0.38 mg/L) in 24 samples collected from five sites north of Reno, Nev., 1992–93 (Berger and others, 1997, p. 46). The lowest values were measured in winter months, and highest values were measured in June and September. As pointed out by Dettinger (1989), chloride deposited as dry fall during summer becomes incorporated into early winter precipitation.

Applying the minimum estimate of precipitation and a chloride concentration of 0.4 mg/L to equation 2 provides a minimum estimate of the recharge rate. Whereas, applying the maximum estimate of precipitation and a chloride concentration of 0.6 mg/L provides a maximum estimate (table 5). The resulting range in annual rates is ±0.01 ft of the estimated recharge at each test hole (CL-1, 0.03–0.05 ft; CL-2, 0.02–0.04 ft; and CL-7, 0.01–0.03 ft).

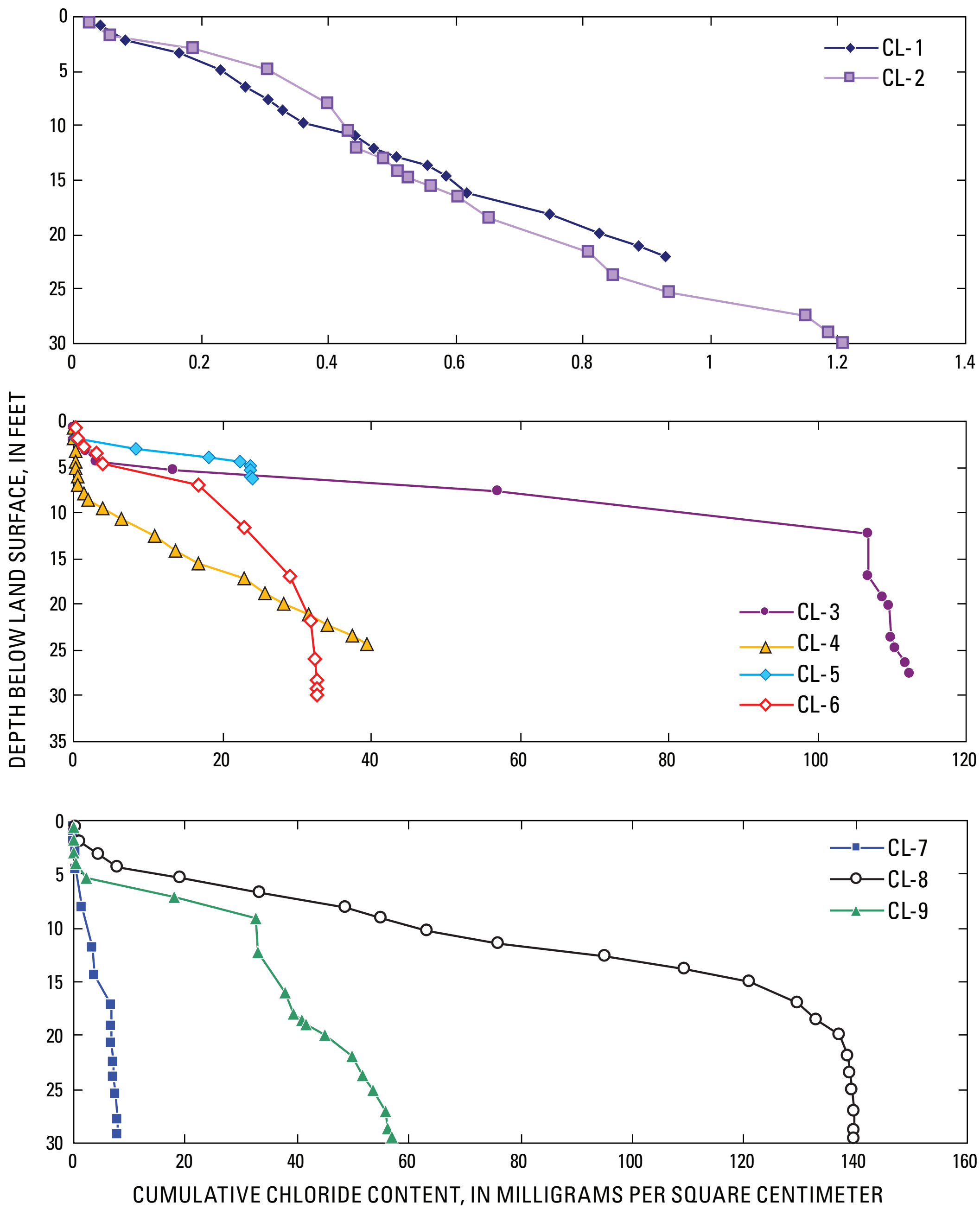


Figure 13. Relation between cumulative chloride content and depth for nine test holes in Carson Valley, Nevada and California.

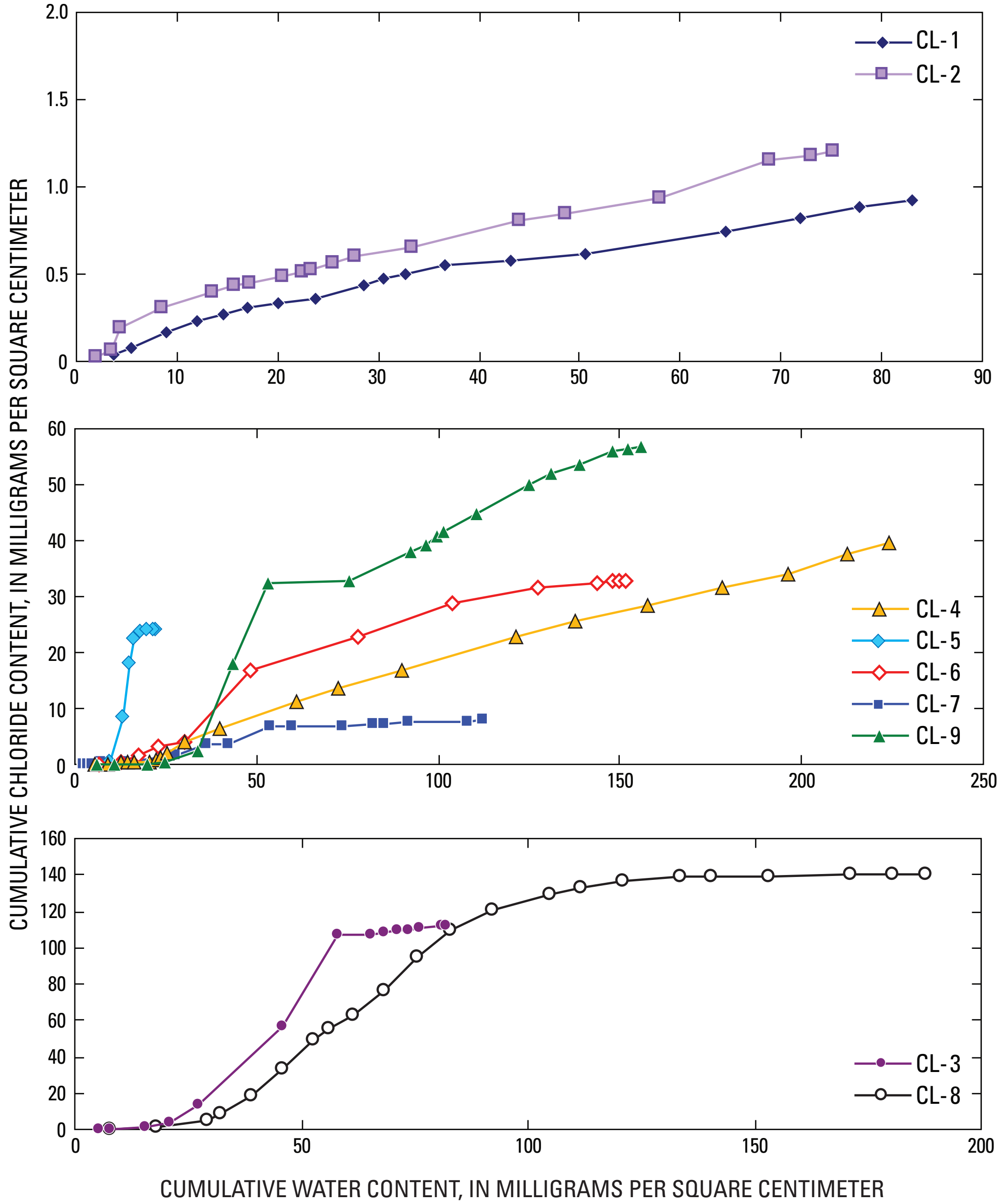


Figure 14. Relation between cumulative water content and cumulative chloride content for nine test holes in Carson Valley, Nevada and California.

Table 4. Cumulative chloride, time required for accumulation, and estimated recharge at soil-chloride test holes, Carson Valley, Nevada and California.

[Test hole locations are shown in figure 2. **Average annual precipitation:** Estimated using equation developed by Maurer and Halford (2004, p. 29); average annual precipitation, in inches, equals altitude of test hole multiplied by 0.0027 minus 5.3646. **Annual rate of chloride deposition:** Chloride concentrations of precipitation (0.5 mg/L) from Feth and others (1964) multiplied by average annual precipitation at test hole, in meters. **Cumulative chloride content to chloride peak**: From figure 11 multiplied by 10 to obtain grams per square meter. **Years to accumulate**: Cumulative chloride divided by annual rate of chloride deposition. **Estimated annual recharge from precipitation**: Calculated from equation 1 in text; recharge is equal to annual precipitation multiplied by chloride concentration of precipitation, divided by average chloride concentration of pore water below root zone. **Abbreviations:** ft, foot; g/m^2, gram per square meter; in., inch; mg/L, milligram per liter; n, indicates no significant recharge]

Test hole No.	Average annual precipitation (in.)	Annual rate of chloride deposition (g/m^2)	Cumulative chloride content to chloride peak (g/m^2)	Years to accumulate (rounded)	Cumulative chloride content to bottom of test hole (g/m^2)	Years to accumulate (rounded)	Average chloride concentration in pore water below root zone (mg/L)	Estimated annual recharge rate from precipitation (ft)
CL-1	7.95	0.101	1.7	16	9.3	90	[1]7.8	0.04
CL-2	8.46	.107	1.9	17	12.1	110	[2]11.6	.03
CL-3	8.85	.112	1,070	9,500	1,125	10,000	n	n
CL-4	10.1	.128	14.0	110	394	3,070	n	n
CL-5	13.3	.168	180	1,070	240	1,420	n	n
CL-6	8.89	.113	166	1,480	328	2,900	n	n
CL-7	8.70	.111	67	600	78	710	[3]22.2	.02
CL-8	9.27	.118	953	8,100	1,400	11,900	n	n
CL-9	10.0	.127	22	175	568	4,460	n	n

[1]Average chloride concentration of pore water is from 14 to 22 ft.

[2]Average chloride concentration of pore water is from 17 to 30 ft.

[3]Average chloride concentration or pore water is from 19 to 29 ft.

Table 5. Uncertainty in estimates of recharge rates at three test holes, Carson Valley, Nevada and California.

[Test hole locations are shown in figure 2. **Average annual precipitation:** Estimated using equation developed by Maurer and Halford (2004, p. 29); average annual precipitation, in inches, equals altitude of test hole multiplied by 0.0027 minus 5.3646. **Potential range in annual precipitation:** From Maurer and Halford (2004, p. 37). **Potential range in chloride concentration of precipitation**: Chloride concentration of precipitation: value of 0.6 mg/L from eight bulk-precipitation sites in Nevada (Dettinger, 1989, p. 62), value of 0.4 mg/L from 24 samples from five sites near Reno, Nevada (Berger and others, 1997, p. 46). **Estimated annual recharge rate from precipitation:** From table 4, calculated from equation 2 in text; recharge is equal to annual precipitation multiplied by chloride concentration of precipitation, divided by average chloride concentration of pore water below root zone, divided by 12 to obtain annual recharge in feet. Values used from column 2 and 0.5 mg/L for chloride concentration of precipitation. **Minimum annual recharge rate**: Calculated from equation 2 in text, using minimum annual precipitation and chloride concentration and dividing by 12 to obtain minimum annual recharge in feet. **Maximum annual recharge rate:** Calculated from equation 2 in text, using maximum annual precipitation and chloride concentration and dividing by 12 to obtain maximum annual recharge in feet. **Abbreviations:** in., inch; mg/L, milligram per liter; ft, foot]

Test hole No.	Average annual precipitation (in.)	Potential range in annual precipitation, ±15 percent (in.)	Potential range in chloride concentration of precipitation (mg/L)	Estimated annual recharge rate from precipitation (ft)	Annual recharge rate (ft)	
					Minimum	Maximum
CL-1	7.95	6.75–9.14	0.4–0.6	0.04	[1]0.03	[1]0.05
CL-2	8.46	7.19–9.73	.4–.6	.03	[2].02	[2].04
CL-7	8.70	7.40–10.01	.4–.6	.02	[3].01	[3].03

[1]Calculated using average chloride concentration of pore water from 14 to 22 feet of 7.8 mg/L.

[2]Calculated using average chloride concentration of pore water from 17 to 30 feet of 11.6 mg/L.

[3]Calculated using average chloride concentration of pore water from 19 to 20 feet of 22.2 mg/L.

Streamflow Infiltration and Seepage

If the streambed is sufficiently permeable to allow flow, streams lose flow and the infiltrating surface water recharges the underlying aquifer when the altitude of the water table adjacent to the stream is lower than the stream's stage. Conversely, when the altitude of the water table is higher than the stream's stage, ground-water discharge from the aquifer to the stream takes place and streams gain flow. Locations of streamflow gain and loss, and estimates of their rates, are useful for evaluating the effects of changes in land use and developing updated water budgets.

Methods Used

Temperature data can be used to determine whether streams are gaining or losing flow. Solar heating of stream water during the day and its cooling over night creates diurnal changes in temperature that can be used to trace the movement of water through the streambed. The difference between stream and streambed temperature, and the timing and amplitude of the diurnal temperature signal in sediments beneath the stream can be used to identify where streams gain and lose flow. Diurnal changes in temperature beneath gaining streams are relatively small, whereas those beneath losing streams follow diurnal temperature changes in the streamflow (Constantz and Stonestrom, 2003, p. 2 and 4).

Temperatures at depths of 1.6 and 3.3 ft beneath streambeds were measured using thermocouple probes. Each probe was made of stainless-steel pipe, 3.3-ft long with a 0.25-in. inside diameter. The bottom of the pipe was pounded and ground to a point, and a coupler was threaded to the top of the pipe. Thermocouples (type T, copper-Constantan wires) were placed at the bottom and midway points in the pipe, and silicone sealant was injected through holes drilled in the pipe between and above the thermocouples. Sealant also was applied to the thermocouple tips to ensure electrical insulation from water.

To place the probes into the streambed, thermocouple wires were run through an additional length of pipe threaded into the top of the probe and a hollow, sliding pounder was used to drive the probes into the streambed until the top of the probe was level with the streambed. The upper pipe was then removed and the thermocouple wires were connected to a data recorder. An additional thermocouple was connected to the recorder, placed in the stream, and weighted to remain in place on the streambed. An aluminum cover was placed over the wiring panel to reduce temperature gradients across the panel. Thermocouple temperatures were obtained from differential measurements referenced to the temperature of the wiring panel in the data recorder, which was recorded at the same intervals as the thermocouple data. Thermocouple temperatures were recorded every 5 seconds and averaged over 5-minute periods. The data recorder was housed in a plastic enclosure and secured on the streambank to stakes or the trunks of nearby willows.

Temperatures recorded from thermocouple probes immersed in a water bath were within 0.01ºC of temperatures determined by a laboratory reference thermometer. Tests of the thermocouple probes compared to thermocouples without a surrounding pipe were made to determine the precision of the recorded temperatures. Thermocouples in pipes were placed in the streambed about 1 ft from thermocouples placed without a surrounding pipe. Over a period of 5 days with data recorded at 5-minute intervals, temperatures recorded by the thermocouples at 1.6 ft beneath the streambed differed by an average of 0.17ºC with a standard deviation of 0.05ºC. Temperatures recorded by the thermocouples at 3.3 ft beneath the streambed differed by an average of 0.25ºC with a standard deviation of 0.15ºC.

Temperature data were collected at 37 sites throughout Carson Valley to identify gaining and losing reaches and estimate infiltration and seepage rates (fig. 15). Site locations were selected along roadways to facilitate access and were limited to those where permission to install the equipment could be obtained. Rocky streambeds in the East Fork of the Carson River and in numerous ditches between the East Fork and Highway 88 south of Minden prevented installation of the thermocouple probes in those areas. However, rocky streambeds along the West Fork of the Carson River did not prevent installation of thermocouples at three sites (ST-35–ST-37).

Streambed temperatures were recorded for periods of 3 to 20 days during May to early November 2003 and May to mid-October 2004 (table 6). Temperature data were not collected during winter months because the daily change in temperature likely would not be large enough for data analysis. Data were collected repeatedly at some sites to determine the variability of temperature fluctuations from year to year, and during an irrigation season. Repeated temperature data were collected within about 10 ft of the initial location.

Appendix C lists the 5-minute temperatures recorded at each site. Preliminary graphs of the recorded stream and streambed temperatures along with recorded data-logger panel temperatures allowed detection of problems with the collected data. Declining stream stage at sites ST-2 and ST-33 caused exposure of the stream thermocouple so the recorded stream temperatures were approximately the same as the air temperatures recorded at the data-logger panel (appendix D, figs. D1 and D3). Similarly, flow in the ditch stopped after only a few days of data collection at sites ST-6, ST-33, and ST-36 and recorded temperatures were approximately the same as air temperatures (appendix D, figs. D2 and D3). At site ST-34, ground water pumped from a nearby well into the ditch caused a marked decrease in stream temperatures for various days during data collection (appendix D, fig. D2).

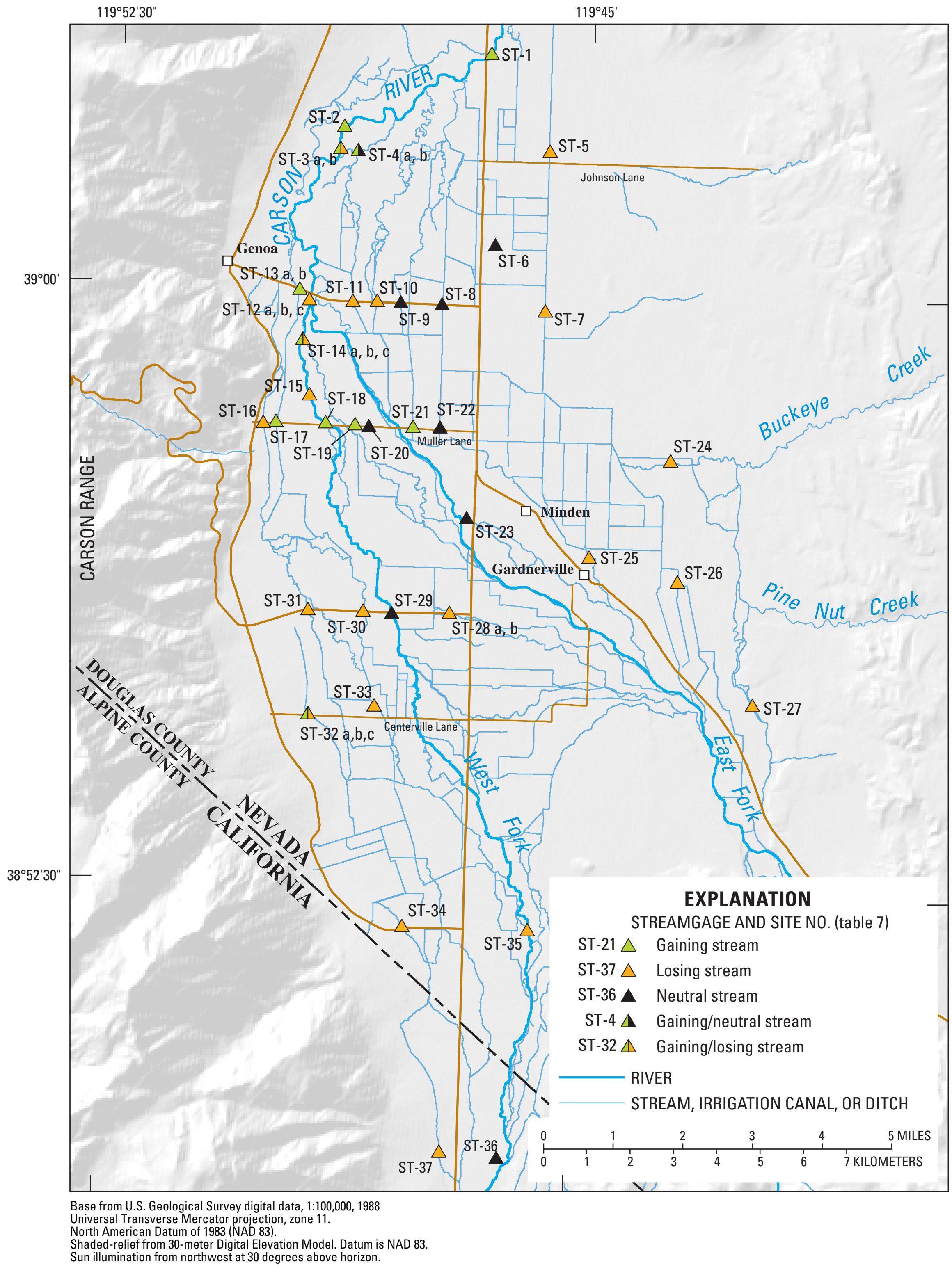


Figure 15. Location of streambed temperature sites and gaining, losing, or neutral conditions at 37 sites in Carson Valley, Nevada and California.

Table 6. Location and description of streambed temperature sites, Carson Valley, Nevada.

[Site locations are shown in figure 15. **Latitude and Longitude:** Geographic coordinates referenced to North American Datum of 1927 (NAD 27), in degrees, minutes, and seconds. **Abbreviations:** ft, foot; ft^3/s, cubic foot per second; nm, not measured]

Site No.	Stream	West latitude	North longitude	Date In	Date Out	Stream width (ft)	Water depth at probe (ft)	Probe distance from bank (ft)	Sediment type and comments
ST-1	Carson River	39°03’02”	119°46’38”	07/09/03	07/16/03	50	0.8	15	Sand and silt, probe pushed by hand.
ST-2	Carson River	39°02’05”	119°48’57”	07/09/03	07/16/03	60	.8	10	Silt and gravel, ponded water.
ST-3a	Carson River	39°01’48”	119°49’00”	07/09/03	07/16/03	90	1.22	3	Fine, probe pushed by hand, ponded water.
ST-3b	Carson River	39°01’48”	119°49’00”	05/21/04	06/01/04	90	1.22	3	Fine sand.
ST-4a	Unnamed ditch	39°01’47”	119°48’44”	07/21/03	07/24/03	15	.5	5	Silt and large gravel.
ST-4b	Unnamed ditch	39°01’47”	119°48’44”	05/24/04	06/01/04	15	1	5	Armored, hard then gravelly.
ST-5	Heyburn Ditch	39°01’50”	119°45’40”	06/17/03	06/24/03	13	2.05	5	Firm packed sand and gravel.
ST-6	Unnamed ditch	39°00’39”	119°46’29”	06/03/03	06/13/03	6	.58	3	Gravel and sand.
ST-7	Heyburn Ditch	38°59’50”	119°45’39”	08/05/04	08/20/04	10	.9	5	Very firm silt and sand, 1 ft standing water.
ST-8	Unnamed ditch	38°59’53”	119°47’19”	09/04/03	09/11/03	6	.8	3	Soft to 1 ft, then firm sand and gravel.
ST-9	Middle Ditch	38°59’54”	119°47’58”	09/11/03	09/18/03	8	.8	4	Soft sand, probed pushed by hand.
ST-10	Williams Slough	38°59’54”	119°48’21”	09/04/03	09/11/03	9	1.4	4.5	Fine and soft.
ST-11	Unnamed ditch	38°59’54”	119°48’45”	08/28/03	09/04/03	10	1	5	Soft silt and sand.
ST-12a	Carson River	38°59’54”	119°49’26”	07/24/03	08/07/03	14	1	4	Sand and gravel.
ST-12b	Carson River	38°59’54”	119°49’26”	06/01/04	06/14/04	45	2	4	Soft, fine sediment.
ST-12c	Carson River	38°59’54”	119°49’26”	09/07/04	09/23/04	15	nm	4	Firm sand.
ST-13a	Unnamed ditch	39°00’02”	119°49’36”	08/21/03	08/28/03	35	2	3	Fine and soft.
ST-13b	Unnamed ditch	39°00’02”	119°49’36”	06/14/04	06/22/04	35	1.5	3	Fine.
ST-14a	W. Fork Carson River	38°59’24”	119°49’31”	08/07/03	08/15/03	45	1	12	Very fine sand, weedy.
ST-14b	W. Fork Carson River	38°59’24”	119°49’31”	06/01/04	06/08/04	50	1.5	12	Sandy.
ST-14c	W. Fork Carson River	38°59’24”	119°49’31”	09/07/04	09/27/04	15	1	7.5	Sand and gravel.
ST-15	Unnamed ditch	38°58’43”	119°49’23”	07/24/03	08/07/03	4	.8	2	Soft silt.
ST-16	Brockliss Slough	38°58’21”	119°50’06”	09/29/03	10/08/03	12	1.5	3	Sand and silt.
ST-17	Unnamed ditch	38°58’22”	119°49’55”	08/15/03	08/21/03	20	1	5	Sand to 2.5 ft, then firm gravel.
ST-18	W. Fork Carson River	38°58’23”	119°49’07”	08/07/03	08/15/03	50	1.5	3	Soft silt, hard near 3.3 ft.
ST-19	Unnamed ditch	38°58’22”	119°48’39”	09/18/03	09/29/03	4	.5	2	Fine and soft sediment.
ST-20	Unnamed ditch	38°58’21”	119°48’25”	09/29/03	10/08/03	15	1	7.5	Fine grained, soft, channel dry on 10/08/03.
ST-21	Unnamed ditch	38°58’21”	119°47’43”	09/17/03	09/28/03	15	.7	7.5	Soft sand and gravel.
ST-22	Rosser Ditch	38°58’21”	119°47’17”	06/18/03	06/24/03	7	2.6	3.5	Soft.
ST-23	Cottonwood Slough	38°57’14”	119°46’49”	07/02/03	07/07/03	10	.75	5	Mucky top 1 ft then packed gravel.
ST-24	Buckeye Creek	38°58’00”	119°43’35”	09/30/04	10/06/04	9	.4	4	Sand and cobbles.
ST-25	Martin Slough	38°56’47”	119°44’50”	09/23/04	09/30/04	10	1	2	Fine sand, 2 ft^3/s on 9/23, just wet on 9/30.
ST-26	Virginia Ditch	38°56’30”	119°43’25”	09/23/04	10/14/04	3	.4	1.5	Fine sand.
ST-27	Allerman Canal	38°55’00”	119°42’10”	05/24/04	05/28/04	20	2	2	Fine sand.
ST-28a	Henningson Slough	38°56’03”	119°47’03”	10/17/03	11/03/03	3.5	.8	1.8	Firm packed sand and gravel.
ST-28b	Henningson Slough	38°56’03”	119°47’03”	06/09/04	06/14/04	4	1.5	2	Firm sand and gravel.
ST-29	W. Fork Carson River	38°56’02”	119°47’58”	10/17/03	11/03/03	25	1	5	Soft first 1.7 ft then packed gravels.
ST-30	Brockliss Slough	38°56’02”	119°48’25”	10/08/03	10/17/03	20	2	3	Fine sand and silt, firm.
ST-31	Big Ditch	38°56’02”	119°49’18”	10/08/03	10/17/03	20	3	3	Soft sand, near left bank.
ST-32a	Park and Bull Slough	38°54’44”	119°49’15”	07/02/03	07/07/03	8	.45	5	Cobbly on surface then gravels.
ST-32b	Park and Bull Slough	38°54’44”	119°49’15”	06/14/04	06/22/04	18	1.5	5	Downstream 10 ft from last years site.
ST-32c	Park and Bull Slough	38°54’44”	119°49’15”	09/07/04	09/23/04	10	2.5	5	Sand and gravel.
ST-33	Brocklis Slough	38°54’52”	119°48’12”	06/29/04	07/07/04	40	2	5	Soft.
ST-34	Fredericksburg Ditch	38°52’07”	119°47’39”	07/07/04	07/16/04	5	1	2	Firm sand some cobbles.
ST-35	W. Fork Carson River	38°52’07”	119°45’39”	07/07/04	07/16/04	30	1	5	Fine sand and boulders
ST-36	Unnamed ditch	38°49’16”	119°46’02”	07/22/04	08/05/04	2.5	.6	1.3	Sand, cobbles, boulders.
ST-37	Fredericksburg Ditch	38°49’19”	119°46’57”	07/22/04	08/05/04	6	.5	2.5	Firm sand and cobbles.

Preliminary graphs also indicated that recorded stream and streambed temperatures were affected by rapid changes in panel temperature of the data logger. Rapid decreases in panel temperature caused abrupt decreases in recorded stream and streambed temperatures, and the reverse occurred during rapid temperature increases (fig. 16, appendix C). In addition, high frequency (5–20 minutes) low amplitude (0.5–2 degrees) fluctuations also were present as noise in the recorded data.

Corrections for the effects of rapid changes in panel temperatures of the data logger were applied to the recorded temperatures of the stream and the streambed at a depth of 1.6 ft, using temperatures recorded at 3.3 ft beneath the streambed. A polynomial equation was derived using decimal days as the independent variable and the temperature at 3.3 ft beneath the streambed at 4 a.m. as the dependent variable. During early morning hours, the data-logger panel temperatures were relatively constant. The difference between the recorded temperature at 3.3 ft and the temperature predicted at the same time by the polynomial equation was assumed to be caused by changing panel temperatures. The difference was then subtracted from the temperatures recorded in the stream and at a depth of 1.6 ft beneath the streambed to obtain adjusted temperature values. This procedure accounts for long-term variations in temperature during the period of data collection.

The method used to adjust the recorded temperatures assumes that: (1) errors caused by changing panel temperatures were the same for all thermocouples, and (2) diurnal-temperature fluctuations at a depth of 3.3 ft beneath the streambed were caused by temperature fluctuations of the data-logger panel and were not caused by flow through the streambed. Thus, the correction will have little effect where streams are gaining flow, but may result in under predicting the flow of water and heat through the streambed in places where streams are losing flow. The effects of rapid changes in panel temperature could have been reduced if the data logger had been insulated from solar radiation by burying the enclosure, however, this was not practical because the recorders were frequently moved and roots along the channel edges made digging difficult.

Adjusting the recorded temperature of the stream and at 1.6 ft beneath the streambed was limited to the period used to derive the polynomial equation (fig. 16*C*; appendix D). The smooth line for the temperature at 3.3 ft beneath the streambed represents the polynomial equation used to determine adjustments for the temperatures of the stream and at 1.6 ft beneath the streambed. Any diurnal signal at 3.3 ft beneath the streambed is lost, however, the adjusted temperatures are considered more representative of actual temperatures than unadjusted temperatures (fig. 16*C*). To further reduce noise in the adjusted temperatures, running 30-minute averages of the adjusted stream and streambed temperatures were calculated to produce the plots in figure 16*C* and appendix D.

At site ST-34, ground water, having a relatively constant lower temperature than the stream, was observed being pumped into the ditch a short distance upstream of the data-collection site on 4 days during the period of data collection (July 9, 12, 14, and 15; appendix D, fig. D2). Clear diurnal fluctuations in temperature were measured at a depth of 1.6 ft beneath the streambed on July 7–9, 2004, indicating losing conditions at the site. Diurnal fluctuations in streambed temperature greatly decreased when ground water was pumped into the ditch and cooler ground water was infiltrating through the streambed. Stream temperature was nearly constant on July 14–15. During this period, streambed temperature lost its diurnal character and gradually approached stream temperature. This would be expected in a losing stream, and indicates the adjusted temperatures are reasonably accurate. The accuracy of adjusted temperatures at other sites is unknown, but based on results from site ST-34, the adjustment is reasonable for estimating gaining and losing conditions.

High-frequency temperature fluctuations of about 2ºC are still present in some of the adjusted data, for example sites ST-11 and ST-27 (appendix D). For most data sets, the effects are minimal, and for sites ST-11 and ST-27 the adjusted streambed temperatures appear to respond to changes in stream temperature. The cause of this noise likely is an instrumentation artifact that was not corrected by the panel-temperature adjustment.

Other noise in the data could be the result of many streambed sites located near flood-irrigated fields. Infiltration of applied water beneath the fields could produce subsurface temperature changes that are not in phase with normal diurnal solar heating and cooling of the stream. In addition, fluctuations in stream temperature beyond those caused by solar heating could be caused by surface runoff from irrigated fields entering the stream upstream of the site. Surface runoff can be warmed to temperatures greater than those in the stream causing unpredictable temperature changes. These factors may be the cause for noise inherent in some of the data.

Differences in temperature between the stream and streambed, the presence of diurnal temperature fluctuations in the streambed, and the propagation or lack of propagation of long-term (longer than diurnal) temperature changes from the stream to the streambed were used to identify gaining, losing, or neutral conditions at the 37 streambed-temperature sites. The identification of conditions are representative of relatively short periods of data collection at each site, and conditions may change over time.

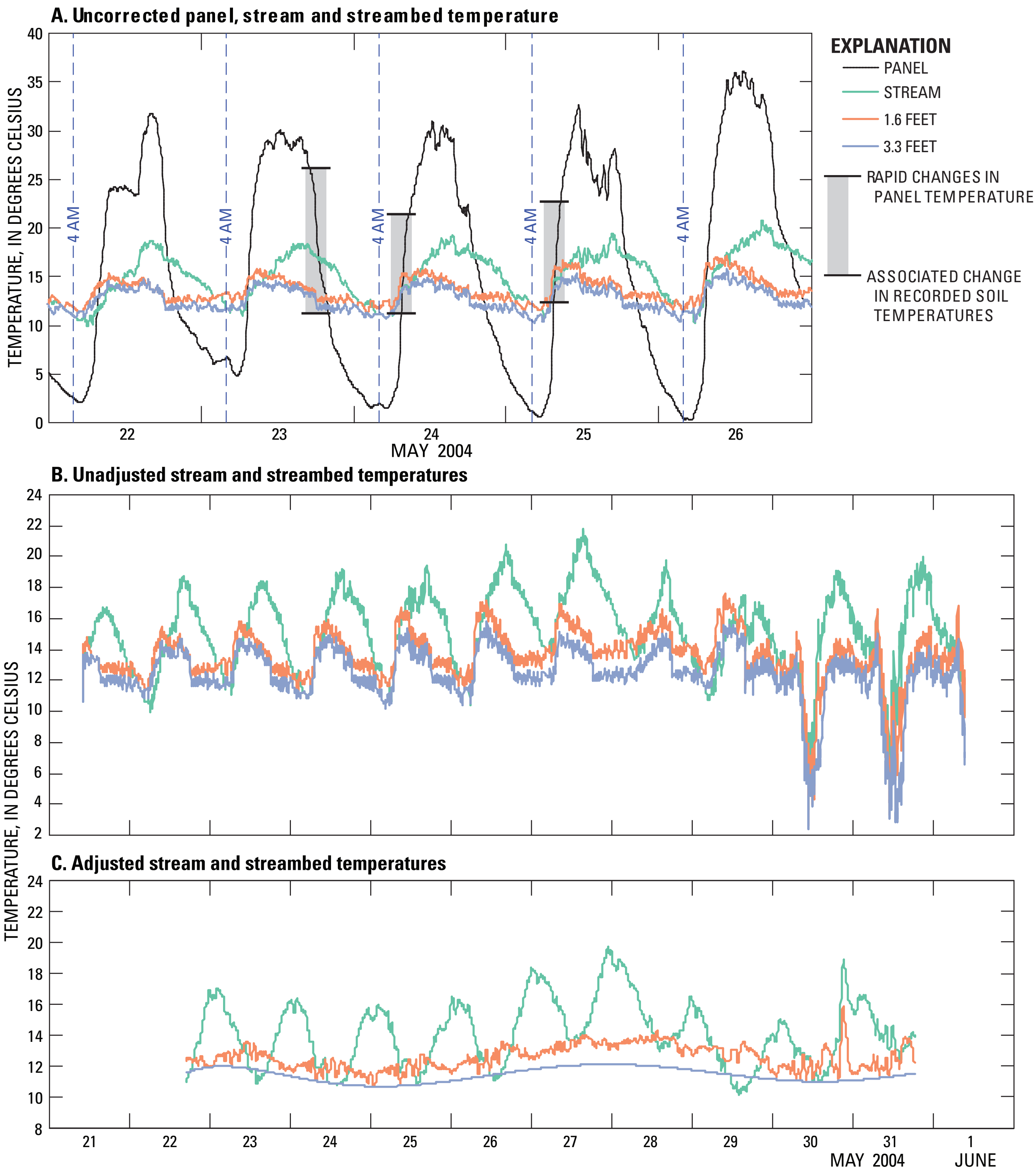


Figure 16. Effects of rapid changes in panel temperature on stream and streambed temperature, and unadjusted and adjusted stream and streambed temperatures at site ST-3b, Carson Valley, Nevada and California.

Changes in stream stage during the periods of data collection and seasonal temperature variations may affect the determination of gaining or losing conditions. Changes in stream stage may be responsible for changes in the amplitude of diurnal streambed-temperature fluctuations during the period of data collection at some sites. However, the effects of changes in stream stage on streambed temperature cannot be evaluated with the existing data, because stream stage was not recorded during the periods of temperature data collection. Changes in streambed temperature can be caused by heat conduction alone without movement of water. Lapham (1989, p. 8) showed that seasonal changes in stream temperature caused temperatures of saturated coarse-grained sediment at a depth of 10 ft below the streambed to vary annually as much as 5ºC, even though no water flowed through the streambed. Under conditions of no water flow through the streambed, streambed temperatures versus stream temperatures are most different in May and November and are most similar from late June through late August (Lapham, 1989, p. 8).

At the relatively shallow depths of the thermocouple probes (1–3 ft), streambed temperatures likely reach equilibrium with stream temperatures faster when moving water transmits heat than when heat transfer is by conduction alone. At losing stream reaches in western Washington, streambed temperatures at depths of 1–3 ft become cooler than stream temperatures as streams begin to warm in early March, and become warmer than stream temperatures as streams cool in early October (Simonds and others, 2004, p. 32 and 43).

Application of 2-D models, such as VS2DH (Healy and Ronan, 1996), to the temperature data could be used to provide estimates of infiltration rates at each site. Niswonger and Prudic (2003, p. 87) noted, however, that several types of data are needed for application of VS2DH to obtain reliable estimates of streambed infiltration and the hydraulic conductivity of streambed sediments. These data, which include ground-water levels adjacent to the streambed, vertical hydraulic gradients within the aquifer, and variations in stream stage are not available for most sites.

The difference in time (phase difference) between diurnal-temperature peaks in the stream and streambed at 1.6 ft was used to estimate infiltration rates for losing stream reaches. The difference in time between diurnal peaks and the amplitude of the diurnal fluctuations were determined using measurements made on plots of the data at expanded time and temperature scales (periods used for the measurements shown in appendix D). An equation by Constantz and Thomas (1996, p. 3598) provides accurate estimates of infiltration rates where heat transport is dominated by advection for rates as low as 0.8 ft/d:

$$Q = V_T(\bullet_s c_s / \bullet_w c_w) \,. \qquad (3)$$

where

Q is the infiltration rate, in feet per day,

V_T is the vertical velocity of the temperature peak, and

$\bullet_s c_s$ and $\bullet_w c_w$ are the products of density and specific heat capacity (volumetric heat capacities) for wet sediments (signified by the subscripts 's') and water (signified by the subscripts 'w'), respectively.

The vertical velocity of the temperature peak is calculated as:

$$V_T = 1.6 \text{ ft}/dT \,, \qquad (4)$$

where

1.6 is the depth of the thermocouple, in feet, and

dT is the difference in time between peak temperatures of the stream and streambed, in decimal days.

Equation 3 was used to estimate infiltration rates for losing stream sites from the difference in time, in decimal days, between the peak stream temperatures and peak streambed temperatures at a depth of 1.6 ft. Values of 2.2×10^6 J/m^3 ºC for the volumetric heat capacity of sediments and 4.2×10^6 J/m^3 ºC for the volumetric heat capacity of water were used in the calculations (Stonestrom and Blasch, 2003, p. 76).

Results

During the period of data collection, from May to early November, average stream temperatures at each site were highest, approaching or exceeding 20ºC, in July and August and lowest, approaching or less than 10ºC, in May and June and in September and October (table 7). Streambed temperatures follow the same pattern with the exceptions of sites ST-1, ST-17, and ST-18 that were 10–12ºC in July and August. Streambed temperatures generally were less than stream temperatures except during September and October, when stream temperatures became cooler than streambed temperatures (table 7). Negative-temperature differences, when the stream temperature is cooler than streambed temperature, generally were smallest (less than 1ºC) in late September and early October (sites ST-8, ST-11, ST-12c, and ST-16) and generally more negative (greater than 3ºC) in late October and early November (sites ST-29 through ST-31). Thus, small temperature differences measured in late September and early October likely are caused by the annual cycle in stream and streambed temperatures previously discussed.

An exception to this pattern is site ST-5, where streambed temperatures were slightly warmer than stream temperatures in mid-June 2003. The warm streambed temperatures compared to the stream temperatures likely were caused by streamflow infiltration during a warm period prior to data collection and a cool period during actual data collection. In late May and early June, mean daytime air temperatures were greater than 25ºC, however, mean daytime air temperatures had decreased to about 9ºC by the end of the data-collection period on June 24, 2003 (Warren Hibbard, National Weather Service observer, written commun., 2005).

The Carson River and irrigation ditches generally gained flow at sites on the westernmost side and northern end of the valley floor, generally north of Muller Lane (fig. 15). The Carson River and irrigation ditches at sites on the southern end and eastern side of the valley floor generally lost flow to ground water. The distribution of gaining, losing, and neutral sites represents conditions for 2003 and 2004 and may change during wetter or drier periods.

Gaining Sites

Strongly gaining streams are shown by data from sites ST-1, ST-17, and ST-18 (table 7, appendix D, fig. D1), where temperatures beneath the streambed were less than the coolest temperatures in the stream. Little to no diurnal-temperature fluctuation was measured at a depth of 1.6 ft beneath the streambed, and the temperatures at depths of 1.6 and 3.3 ft are nearly equivalent. Data at these sites were collected in July and August when infiltration of warmer streamflow would increase the temperature beneath the streambed and cause diurnal fluctuations. Ground water of relatively constant temperature moving from the water table into the stream caused temperatures at depths of 1.6 and 3.3 ft to be nearly equivalent. Lapham (1989, p. 13) shows that annual temperature fluctuations beneath streambeds are near zero when upward ground-water flow velocities, or seepage rates, are 1 ft/d or greater.

Temperature data collected at sites ST-2, ST-3a, ST-4a, ST-13b, ST-14a, ST-14b, and ST-32a indicated less strongly gaining streams (table 7, appendix D, fig. D1). At these sites, streambed temperatures were greater at depths of 1.6 ft than at 3.3 ft, streambed temperatures were less than stream temperatures, and there was little to no diurnal fluctuation in temperature. Ground water at 1.6 ft may have been warmed by conduction of heat through the streambed, but was kept cooler than the stream by mixing with cooler ground water moving upward from depths of 3.3 ft or greater. At site ST-32b (appendix D, fig. D2), streambed temperatures also were less than stream temperatures, however, relatively large, albeit noisy, diurnal-temperature fluctuations were recorded at 1.6 ft. For this reason, the stream reach at site ST-32b likely is a losing reach (table 7).

Gaining sites had large positive temperature differences, from 2.6 to greater than 7ºC, indicating streambed temperatures were cooler than stream temperatures (table 7). Data at most of these sites were collected during July and August when stream and streambed temperatures had been warming for 2–3 months and temperatures likely are in equilibrium (Lapham, 1989, p. 8).

At sites ST-13a, ST-19, and ST-21 (table 7, appendix D, fig. D1), relatively large diurnal and long-term (longer than diurnal) fluctuations in the stream did not propagate downward into the streambed, indicating gaining conditions and a lack of streamflow infiltration. At sites ST-13a and ST-21, changes of 5–6ºC in stream temperature on 1 or 2 days did not propagate into the streambed. At site ST-19, streambed temperatures continually decreased, whereas stream temperatures increased September 20–25, and decreased after September 25.

Losing Sites

Propagation of diurnal stream-temperature fluctuations downward through the streambed were considered indicative of losing stream reaches. At some sites, the diurnal fluctuations were well defined at a depth of 1.6 ft beneath the streambed and had relatively high amplitude, such as sites ST-5, ST-7, ST-12b, ST-16, ST-24, ST-28a, ST-34, and ST-37 (table 7, appendix D, fig. D2). The amplitude of the diurnal streambed-temperature fluctuations at a depth of 1.6 ft, as a percentage of the stream-temperature fluctuations ranged from 20 to 48 percent (table 7). The temperature differences between the stream and streambed at these losing reaches measured during summer months are small, less than 0.8ºC. Small temperature differences measured during summer months provided additional evidence that streambed temperatures were in approximate equilibrium with infiltrating streamflow and that infiltration was relatively rapid. The small, negative temperature differences measured at sites ST-16, ST-24, and ST-28a in September and October may result from long-term seasonal cycles in stream and streambed temperatures previously discussed.

Relatively high-amplitude diurnal fluctuations at 1.6 ft beneath the streambed at sites ST-31, ST-32b, and ST-35 (table 7, appendix D, fig. D2) are somewhat masked by noise in the temperature signal, however, streams likely are losing flow at these sites. The amplitude of diurnal streambed-temperature fluctuations, as a percentage of the stream-temperature fluctuations, ranged from 27 to 36 percent (table 7).

Relatively low amplitude yet well-defined diurnal fluctuations at a depth of 1.6 ft beneath the streambed generally following long-term changes in stream temperatures are shown at sites ST-10, ST-12a, ST-14c, ST-15, and ST-26 (appendix D, fig. D2). These sites likely are losing flow and have streambed-temperature fluctuations with amplitudes ranging from 12 to 27 percent of the stream fluctuations (table 7). At sites ST-3b, ST-11, ST-12c, ST-25, ST-27, ST-28b, ST-30, ST-32c, and ST-33, streambed temperatures show small diurnal fluctuations and generally follow stream temperatures (appendix D, fig. D2). However, these sites also are likely losing flow. Noise in the temperature signal does not allow accurate determination of the relative amplitude and phase difference of the fluctuations at these sites.

Table 7. Temperature data, estimated infiltration rates at stream sites, and estimated gain/loss conditions at stream sites, Carson Valley, Nevada and California.

[Site locations are shown in figure 15; a, b, and c, indicate multiple measurement dates at same site. **Average stream temperature:** Includes only time stream probe was in water. **Average temperature difference:** Includes only time stream probe was in water; positive difference indicates streambed temperature cooler than stream temperature, negative difference indicates streambed temperature warmer than stream temperature. **Estimated infiltration rate:** Infiltration rate (*Q*) = [1.6 ft divided by phase difference (decimal days)] time ratio of volumetric heat capacity of wet sediments to water (0.5238; Constantz and Thomas, 1996, p. 3589). **Estimated gain/loss conditions**: G, gaining; L, losing; N, neutral. **Abbreviations**: °C, degrees Celsius; ft, foot; ft/d, foot per day; nd, not determined where diurnal streambed fluctuation is small]

Site No.	Stream	Date		Average stream temperature (°C)	Average streambed temperature at 1.6 ft depth (°C)	Average streambed temperature at 3.3 ft depth (°C)	Average temperature difference between stream and streambed at 1.6 ft depth (°C/ft)	Diurnal amplitude in stream temperature (°C)	Diurnal amplitude in streambed temperature at 1.6 ft depth (°C)	Ratio of diurnal streambed to stream temperature amplitude (percent)	Approximate time difference between stream and streambed temperature peaks (decimal days)	Estimated infiltration rate using equation 3 in text (ft/d)	Estimated gain/loss conditions
		In	Out										
ST-1	Carson River	07/09/03	07/16/03	18.8	11.5	11.4	7.3	10.2	nd	nd	nd	nd	G
ST-2	Carson River	07/09/03	07/16/03	20.0	16.3	14.2	3.8	5.0	nd	nd	nd	nd	G
ST-3a	Carson River	07/09/03	07/16/03	23.4	20.6	17.7	2.9	6.4	nd	nd	nd	nd	G
ST-3b	Carson River	05/12/04	06/01/04	14.1	12.5	11.4	1.6	6.1	nd	nd	nd	nd	L
ST-4a	Unnamed ditch	07/21/03	07/24/03	25.9	21.7	17.6	4.2	14.3	nd	nd	nd	nd	G
ST-4b	Unnamed ditch	05/24/04	06/01/04	17.9	15.7	13.5	2.2	8.6	nd	nd	nd	nd	N
ST-5	Heyburn Ditch	06/17/03	06/24/03	16.0	16.4	15.6	-0.4	5.8	2.8	48	0.36	2.3	L
ST-6	Unnamed ditch	06/03/03	06/13/03	15.7	15.4	14.0	0.3	6.5	nd	nd	nd	nd	N
ST-7	Heyburn Ditch	08/05/04	08/20/04	18.9	18.7	18.1	0.3	6.4	1.7	27	0.37	2.3	L
ST-8	Unnamed ditch	09/04/03	09/11/03	17.2	18.0	17.3	-0.8	10.1	nd	nd	nd	nd	N
ST-9	Middle Ditch	09/11/03	09/18/03	13.9	15.9	16.8	-2.0	16.9	nd	nd	nd	nd	N
ST-10	Williams Slough	09/04/03	09/11/03	14.9	15.9	16.6	-1.0	13.5	1.8	13	0.42	2.0	L
ST-11	Unnamed ditch	08/28/03	09/04/03	17.3	17.5	16.6	-0.2	12.0	nd	nd	nd	nd	L
ST-12a	Carson River	07/24/03	08/07/03	22.6	20.3	18.1	2.3	7.4	2.0	27	0.24	3.5	L
ST-12b	Carson River	06/01/04	06/14/04	15.9	13.3	12.4	2.6	7.0	1.5	21	0.19	4.4	L
ST-12c	Carson River	09/07/04	09/23/04	15.0	15.1	15.0	-0.1	13.9	1.3	9	0.64	1.3	L
ST-13a	Unnamed ditch	08/21/03	08/28/03	20.7	20.4	19.6	0.4	8.5	nd	nd	nd	nd	G
ST-13b	Unnamed ditch	06/14/04	06/22/04	19.5	16.8	14.4	2.7	5.6	nd	nd	nd	nd	G
ST-14a	W. Fork Carson River	08/07/03	08/15/03	21.6	19.1	17.1	2.6	8.1	nd	nd	nd	nd	G
ST-14b	W. Fork Carson River	06/01/04	06/08/04	19.3	16.0	13.4	3.4	6.5	nd	nd	nd	nd	G
ST-14c	W. Fork Carson River	09/07/04	09/27/04	15.7	14.1	12.9	1.5	13.1	1.6	12	0.90	0.9	L
ST-15	Unnamed ditch	07/24/03	08/07/03	21.6	20.8	18.3	0.8	7.6	1.2	16	0.77	1.1	L
ST-16	Brockliss Slough	09/29/03	10/08/03	14.9	15.3	14.9	-0.4	6.6	1.3	20	0.77	1.1	L
ST-17	Unnamed ditch	08/15/03	08/21/03	18.9	12.3	12.3	6.6	11.4	nd	nd	nd	nd	G
ST-18	W. Fork Carson River	08/07/03	08/15/03	16.6	10.3	10.2	6.3	7.4	nd	nd	nd	nd	G
ST-19	Unnamed ditch	09/18/03	09/29/03	16.3	16.2	16.9	0.1	7.2	nd	nd	nd	nd	G
ST-20	Unnamed ditch	09/29/03	10/08/03	13.5	14.8	15.1	-1.3	25.2	nd	nd	nd	nd	N
ST-21	Unnamed ditch	09/17/03	09/28/03	14.9	16.0	17.9	-1.1	5.7	nd	nd	nd	nd	G

Table 7. Temperature data, estimated infiltration rates at stream sites, and estimated gain/loss conditions at stream sites, Carson Valley, Nevada and California—Continued.

[Site locations are shown in figure 15; a, b, and c, indicate multiple measurement dates at same site. **Average stream temperature:** Includes only time stream probe was in water. **Average temperature difference:** Includes only time stream probe was in water; positive difference indicates streambed temperature cooler than stream temperature, negative difference indicates streambed temperature warmer than stream temperature. **Estimated infiltration rate:** Infiltration rate (Q) = [1.6 ft divided by phase difference (decimal days)] time ratio of volumetric heat capacity of wet sediments to water (0.5238; Constantz and Thomas, 1996, p. 3589). **Estimated gain/loss conditions**: G, gaining; L, losing; N, neutral. **Abbreviations**: °C, degrees Celsius; ft, foot; ft/d, foot per day; nd, not determined where diurnal streambed fluctuation is small]

Site No.	Stream	Date		Average stream temperature (°C)	Average streambed temperature at 1.6 ft depth (°C)	Average streambed temperature at 3.3 ft depth (°C)	Average temperature difference between stream and streambed at 1.6 ft depth (°C/ft)	Diurnal amplitude in stream temperature (°C)	Diurnal amplitude in streambed temperature at 1.6 ft depth (°C)	Ratio of diurnal streambed to stream temperature amplitude (percent)	Approximate time difference between stream and streambed temperature peaks (decimal days)	Estimated infiltration rate using equation 3 in text (ft/d)	Estimated gain/loss conditions
		In	Out										
ST-22	Rosser Ditch	06/18/03	06/24/03	18.2	19.2	18.1	-0.9	4.0	nd	nd	nd	nd	N
ST-23	Cottonwood Slough	07/02/03	07/07/03	19.2	17.6	15.9	1.6	12.8	nd	nd	nd	nd	N
ST-24	Buckeye Creek	09/30/04	10/06/04	12.0	12.8	13.8	-0.8	5.4	1.6	30	0.30	2.8	L
ST-25	Martin Slough	09/23/04	09/30/04	13.7	13.8	15.0	-0.1	11.4	nd	nd	nd	nd	L
ST-26	Virginia Ditch	09/23/04	10/14/04	12.5	13.3	14.2	-0.8	12.1	1.6	13	0.46	1.8	L
ST-27	Allerman Canal	05/24/04	05/28/04	12.2	11.0	10.2	1.2	4.3	nd	nd	nd	nd	L
ST-28a	Henningson Slough	10/17/03	11/03/03	8.2	8.6	10.0	-0.4	10.5	2.8	27	0.33	2.5	L
ST-28b	Henningson Slough	06/09/04	06/14/04	13.7	12.9	11.9	0.9	10.9	1.3	12	0.37	2.3	L
ST-29	W. Fork Carson River	10/17/03	11/03/03	9.5	13.2	12.3	-3.7	3.7	nd	nd	nd	nd	N
ST-30	Brockliss Slough	10/08/03	10/17/03	11.2	12.6	13.8	-1.4	3.9	0.9	23	0.63	1.3	L
ST-31	Big Ditch	10/08/03	10/17/03	9.9	12.3	13.2	-2.4	4.7	1.7	36	1.40	0.6	L
ST-32a	Park and Bull Slough	07/02/03	07/07/03	19.9	17.2	15.5	2.6	6.1	nd	nd	nd	nd	G
ST-32b	Park and Bull Slough	06/14/04	06/22/04	19.8	17.3	15.2	2.5	7.4	2.0	27	0.28	3.0	L
ST-32c	Park and Bull Slough	09/07/04	09/23/04	13.5	14.4	14.2	-0.9	6.9	1.6	23	0.40	2.1	L
ST-33	Brockliss Slough	06/29/04	07/07/04	19.2	18.6	18.3	0.7	7.4	nd	nd	nd	nd	L
ST-34	Fredericksburg Ditch	07/07/04	07/16/04	16.5	16.3	17.0	0.1	12.2	3.7	30	0.44	1.9	L
ST-35	W. Fork Carson River	07/07/04	07/16/04	17.2	14.8	13.4	2.4	6.4	2.2	34	0.34	2.5	L
ST-36	Unnamed ditch	07/22/04	08/05/04	21.0	19.1	18.1	1.9	5.3	nd	nd	nd	nd	N
ST-37	Fredericksburg Ditch	07/22/04	08/05/04	18.9	18.8	18.2	0.1	7.4	2.0	27	0.42	2.0	L

Neutral Sites

Sites ST-4b, ST-6, ST-8, ST-9, ST-20, ST-22, ST-23, ST-29, and ST-36 (table 7, appendix D, fig. D3) do not appear to be gaining or losing significant amounts of flow. At these sites, streambed temperatures at 1.6 ft beneath the streambed have little to no diurnal-temperature fluctuations, but generally follow long-term changes in stream temperatures. The streambed at these sites likely is warmed by conduction from the stream, but infiltration is not sufficient to produce diurnal changes in temperature.

Estimated Infiltration and Seepage Rates

Infiltration rates at losing stream reaches are proportional to the amplitude of the streambed-temperature fluctuation relative to the amplitude of the stream-temperature fluctuation (Lapham, 1989, p. 12). When infiltration rates are high, the amplitude of the diurnal-temperature fluctuation of the streambed closely mimics the diurnal-temperature fluctuation of the stream. Under such conditions, the difference in time between temperature peaks in the stream and streambed is inversely proportional to the infiltration rate (Constantz and Thomas, 1996, equation 3). It follows that sites with high infiltration rates have small differences in time between diurnal stream- and streambed-temperature peaks, and relatively large ratios in the amplitude of streambed to stream diurnal temperatures.

At sites where data were collected during more than one period (sites ST-12a, ST-12b, ST-12c, ST-28a, ST-28b, ST-32b, and ST-32c; table 7), small differences in time between stream- and streambed-temperature peaks generally correspond to large ratios in streambed to stream amplitudes. When different sites were compared, some had large differences in time and large amplitude ratios (table 7; sites ST-30 and ST-31), whereas others had small differences in time and small amplitude ratios (sites ST-12b and ST-26). The relation between the ratio of streambed- to stream-temperature amplitude and the phase difference likely is a complex function of the hydraulic setting and geometry, along with the hydraulic and thermal properties of the streambed sediments at each site.

Time differences ranging from 0.19 to 1.4 days (table 7) were used in equations 3 and 4 to estimate infiltration rates for losing stream sites. Estimated infiltration rates ranged from 0.6 to 4.4 ft/d.

Rates of streamflow loss per mile of stream were estimated by multiplying the estimated infiltration rates by the measured stream widths at each site, then multiplying by 5,280 ft, and dividing by 86,400 seconds per day to provide a loss rate in cubic feet per second per mile at each site (table 8). This method of calculating stream-loss rates assumes that infiltration rates and stream widths are constant over a 1-mi reach of stream. The resulting rates range from less than 0.6 (ft^3/s)/mi for the smaller ditches (sites ST-15, ST-26, ST-28a, ST-28b, and ST-34) to 12.1 (ft^3/s)/mi for site ST-12b (Carson River near Genoa) in early June 2004.

Stream stage, width, and flow at site ST-12b were high in early June 2004, and flow of the Carson River near Genoa was about 300 ft^3/s (http://nevada.usgs.gov/ADR/wy04/sw/10310407_2004_sw.pdf). During this period, a loss rate of about 12 (ft^3/s)/mi may be reasonable if it occurs for only 1 or 2 mi. In August 2003, when flow in the river at the site averaged 30 ft^3/s (Stockton and others, 2004, p. 202), a loss rate of 3 (ft^3/s)/mi was estimated (site ST-12a; table 8). The rates indicate a much greater percentage of flow was lost in August 2003 (10 percent) than in June 2004 (2 percent). This indicates that rates may be variable over time at a particular

Table 8. Loss and gain rates for selected stream sites, Carson Valley, Nevada and California.

[Site locations are shown in figure 15. **Abbreviations:** ft/d, foot per day; ft, foot; (ft^3/s)mi, cubic foot per second per mile]

Site No.	Estimated infiltration or seepage rate (ft/d)	Stream width (ft)	Estimated rate of loss or gain (ft^3/s)/mi
Infiltration – Losing sites			
ST-5	2.3	13	1.8
ST-7	2.3	10	1.4
ST-10	2	9	1.1
ST-12a	3.5	14	3.0
ST-12b	4.4	45	12.1
ST-12c	1.3	15	1.2
ST-14c	.9	15	.8
ST-15	1.1	4	.3
ST-16	1.1	12	.8
ST-24	2.8	9	1.5
ST-26	1.8	3	.3
ST-28a	2.5	3.5	.5
ST-28b	2.3	4	.6
ST-30	1.3	20	1.6
ST-31	.6	20	.7
ST-32b	3	18	3.3
ST-32c	2.1	10	1.3
ST-34	1.9	5	.6
ST-35	2.5	30	4.6
ST-37	2	6	.7
Seepage – Gaining sites			
ST-1 [1]	1	50	3.1
ST-17 [1]	1	20	1.2
ST-18 [1]	1	50	3.1
ST-1 [1]	.1	50	.3
ST-17 [1]	.1	20	.1
ST-18 [1]	.1	50	.3

[1] Calculated using an application rate of 0.1 ft/d (Lapham, 1989). All calculations for gaining sites from Lapham, 1989

site, and that application of some of the larger rates, from 3 to 12 (ft^3/s)/mi, to reaches of 1 or more miles may yield infiltration rates that exceed the actual flow of the stream, particularly during periods of low flow. Thus, some of the larger estimates of the infiltration rate may be high, or may be applicable only to relatively short stream reaches.

An estimated seepage rate of 1 ft/d, as determined from model simulations by Lapham (1989, p. 13) for gaining stream sites with no annual streambed-temperature fluctuations, was applied to sites ST-1, ST-17, and ST-18 (table 8). The resulting rates of streamflow gain are 1–3 (ft^3/s)/mi of channel. Seepage rates as low as 0.1 ft/d also were shown by Lapham (1989, p. 13) to produce minimal annual streambed-temperature fluctuations. An application rate of 0.1 ft/d would decrease the rates of streamflow gain by an order of magnitude from 0.1 to 0.3 (ft^3/s)/mi (table 8).

Uncertainty in the Distribution of Gaining and Losing Sites and in Infiltration and Seepage Rates

Identification of sites as gaining and losing is somewhat uncertain because of the uncertainty of the complete removal of panel-temperature effects in the adjusted stream and streambed temperatures, the relatively low amplitude and poorly defined diurnal fluctuations for some losing sites, and the potential for changing conditions over time. Data collected in different years, and in different seasons of the same year, indicate that conditions may change at a particular site.

At sites ST-3a, ST-3b, ST-32a, and ST-32b conditions changed from gaining in the summer of 2003 to losing in the spring of 2004 (table 7). Similarly, conditions changed from gaining in the spring of 2004 to losing in late summer of 2004 at sites ST-14b, ST-14c, ST-32b, and ST-32c. Such changes likely are caused by changes in stream stage relative to the altitude of the adjacent water table. Stream stages generally are highest in spring months and decline over the summer on the main stems of the Carson River. Water-table altitudes also are highest in spring after recharge from winter precipitation and widespread application of surface water for irrigation, and decline over summer months.

Stream stage in irrigation ditches varies on weekly, daily, or hourly time scales as surface water is diverted through the ditches and stage is adjusted at control structures to supply water to different fields. Small differences between stream stage and the water-table altitude can cause changes from gaining or losing conditions because of the low slope of land surface across the valley floor and the low gradient of the water table. Finally, drought conditions were present in 2003 and 2004, and it is uncertain how the distribution of gaining and losing reaches might change during periods of average or above average precipitation.

The equation used to estimate infiltration rates at the stream sites assumes that downward flow is purely vertical; however, streamflow lost through the streambed likely also moves in a lateral direction away from the stream. Thus, equation 3 may underestimate infiltration rates, but the magnitude of the error is unknown. In addition, infiltration rates are estimated for particular points on the stream and conditions may change upstream or downstream of the site, and at a particular site, conditions can change over time. Thus, the estimated infiltration rates should be considered point measurements and should be applied only with caution to stream reaches greater than 1 mi. Application of infiltration rates to distances greater than 1 mi can predict streamflow losses that exceed the actual flow of a stream.

Estimates of seepage rates ranging from 0.1 to 1.0 ft/d result in streamflow gain rates ranging from 0.1 to 3 (ft^3/s)/mi. For comparison, a streamflow gain of about 10 ft^3/s was measured in the 10-mi reach between the Genoa and Carson City gaging stations (fig. 2) in November 2003 when flow of the Carson River was 90–100 ft^3/s. This results in a gain rate of about 1.0 (ft^3/s)/mi, or one-third that was calculated using a seepage rate of 1.0 ft/d in table 8 for site ST-1. Thus, a more reasonable seepage rate for that reach may be about 0.3 ft/d. However, the combined uncertainty of the individual measurements at flow rates of 90–100 ft^3/s is approximately equal to the measured difference in flow. Seepage gains from 0.1 to 1.0 (ft^3/s)/mi probably are reasonable bounds for gaining streams in Carson Valley.

Summary and Conclusions

Rapid growth and development in Carson Valley is causing concern over the continued availability of water resources as land presently used for agriculture is converted to residential and commercial use. Demand for ground water likely will increase in order to supply these areas. The effects of these changes on ground-water flow and flow in the Carson River are uncertain. The flow of the Carson River downstream of Carson Valley is important to water users dependent on water in the river for many varied uses.

The U.S. Geological Survey, in cooperation with Douglas County, Nevada, began a study in February 2003 to update estimates of Carson Valley's water-budget components. This report presents and summarizes micrometeorologic, soil-chloride, and streambed-temperature data collected as part of the study and presents updated estimates of the rate of evapotranspiration (ET), recharge from precipitation, and streamflow infiltration and seepage. Micrometeorologic data were used to estimate annual and monthly ET rates for water year 2004 from irrigated pasture and alfalfa, and stands of native vegetation including greasewood, rabbitbrush, bitterbrush, and sagebrush. Soil-chloride data were used to estimate rates of recharge from precipitation on the northern and eastern sides of the valley. Streambed-temperature data were used to identify gaining and losing stream reaches and estimate rates of gain and loss for selected sites on the Carson River and irrigation canals and ditches. This information can be used to update estimates of the major water-budget components in Carson Valley, and to evaluate the effects of land- and water-use changes on the water budget.

ET rates from native vegetation and nonirrigated land during water year 2004 were considerably less than ET rates on land that was regularly irrigated. A stand of bitterbrush and sagebrush on an alluvial fan on the western side of Carson Valley where the depth to water is about 60 ft had the lowest ET rate of 1.5 ft/yr. Estimated ET from nonirrigated pasture was only slightly higher at 1.7 ft/yr. A stand of rabbitbrush and greasewood on the northern end of the valley had an estimated ET rate of 1.9 ft/yr. For comparison, the ET rate from most irrigated lands, both alfalfa and pasture, was about 3 ft/yr, but was greater than 4 ft/yr from irrigated pasture where the water table was 2 ft or less from land surface.

ET rates measured in water year 2004 probably are less than what would be measured in a year with average, or above average precipitation because water year 2004 was considerably drier than average. Water year 2004 was the sixth consecutive year of a drought with average or below average precipitation. During average or wet years, the water table beneath the sites likely would be shallower than the water table during 2004. The greatest ET rates were measured at site ET-8, where the depth to water was less than at the other sites; thus, the depth to water is an important factor controlling the ET rate.

The estimated uncertainty in ET rates was about 12 percent for sites ET-1, ET-2, ET-3, ET-5, and ET-8 where measured daily ET was available for most of water year 2004. The estimated uncertainty in the ET rates for sites ET-4 and ET-6 were +30 percent and –20 percent, and for site ET-7 was +50 percent and –40 percent.

Data from six soil-chloride test holes in areas of native vegetation on the northern and eastern sides of Carson Valley indicate that recharge from modern-day precipitation in these areas is not taking place because the precipitation that infiltrates below land surface is lost to ET by native plants. High concentrations of soil chloride at depths ranging from 4 to 18 ft below land surface in six test holes on the eastern side of Carson Valley indicate that modern-day precipitation does not percolate deeper than the roots of native vegetation. The presence of fine-grained semiconsolidated sediments at some test holes appears to limit the depth to which plant roots may penetrate. Estimates of the time required to accumulate the amount of chloride to depths of about 30 ft below land surface at the six test holes range from 3,000 to 12,000 years.

Data from two soil-chloride test holes near the northern end of the valley and one test hole on the eastern side of Fish Spring Flat indicate that a small amount of recharge from modern-day precipitation is taking place. Low concentrations of soil chloride in the three test holes indicate annual recharge from precipitation is 0.03 and 0.04 ft at the northern test holes, and is 0.02 ft on the eastern side of Fish Spring Flat. Estimates of the time required to accumulate the amount of chloride to depths of about 30 ft below land surface at the three test holes range from about 100 to 700 years. The uncertainty in the estimated recharge rates is about ±0.01 ft.

The two test holes near the northern end of the valley are in gravel and eolian sand deposits where precipitation infiltrates through the coarse-grained and well sorted sediments to depths below the root zone. Recharge from modern-day precipitation may be taking place at similar rates in other areas where gravel and eolian sand deposits are mapped. Based on results from the other test holes, it is unlikely that the recharge rate estimated for the test hole on the eastern side of Fish Spring Flat is applicable to a large area.

Data from 37 streambed-temperature sites indicate that the Carson River and irrigation ditches generally gain flow from ground water on the extreme western side of the valley and north of Muller Lane, and generally lose flow over the remainder of valley. Estimated infiltration rates at losing sites ranged from 1 to 4 ft/d, and estimated seepage rates at gaining sites may be about 1 ft/d. Extrapolating the estimated loss rates for a site to reaches more than 1 mi should be made with caution. An estimated seepage rate of 0.3 ft/d for the Carson River in the northern part of Carson Valley indicates a gain in streamflow from ground water that compares well with measured gain in streamflow during November 2003.

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Appendix A. Estimated and predicted daily monthly evapotranspiration for eight evapotranspiration sites, Carson Valley, Nevada and California, April 2003–November 2004.

Daily rates of estimated and predicted ET for all sites are listed in appendix A. These data are available in an Excel data base for download at URL: http://pubs.water.usgs.gov/sir20055288/appendix/appA.xls.

Appendix B. Lithologic description of soil penetrated by nine soil-chloride test holes, Carson Valley, Nevada and California.

[Test hole locations are shown in figure 2. Description in approximate order of sediment amount, by volume. Color codes from standard rock-color chart]

Test hole No.	Depth below land surface (feet)	Lithologic description
CL-1	0–2.7	Sand (fine), silt, and gravel (fine); pale brown (5YR5/2).
	2.7–6.0	Sand (fine), silt, and gravel (coarse, up to 1-inch diameter); moderate yellowish brown (10YR5/4).
	6.0–8.3	Sand (coarse), silt, and gravel (fine to medium); moderate yellowish brown (10YR5/4).
	8.3–12.5	Rounded gravel (coarse), sand (fine to coarse), and angular gravel (decomposed granite); grayish orange (10YR7/4).
	12.5–14.3	Sand (fine to coarse), angular gravel (decomposed granite), and rounded gravel (coarse, 1- to 3-inch diameter); grayish orange (10YR7/4).
	14.3–18.0	Angular gravel (decomposed granite), sand (coarse), and rounded gravel (medium); grayish orange (10YR7/4).
	18.0–22.7	Sand (fine) and silt; grayish orange (10YR7/4).
CL-2	0–4.7	Sand (fine to coarse) and silt with occasional fine gravel; pale yellowish brown (10YR6/2).
	4.7–8.0	No sample retained.
	8.0–14.6	Sand (fine to medium) and silt; grayish orange (10YR7/4).
	14.6–18.0	Sand (fine) and silt; grayish orange (10YR7/4).
	18.0–22.1	No sample retained.
	22.1–30.8	Sand (fine to coarse) and silt with occasional fine gravel; pale yellowish brown (10YR7/4).
CL-3	0–12.1	Semiconsolidated silt, clay, and rounded gravel (fine to medium); yellowish gray (5Y7/2).
	12.1–18.3	Very hard, semiconsolidated sand (fine) and silt; yellowish gray (5Y7/2).
	18.3–20.5	Semiconsolidated sand (fine) and silt with occasional fine gravel; yellowish gray (5Y7/2).
	20.5–25.3	Semiconsolidated sand (fine), silt, and clay with occasional fine gravel; yellowish gray (5Y7/2).
	25.3–28.3	Semiconsolidated sand (fine), silt, and clay with occasional fine gravel; pale yellowish brown (10YR6/2).
CL-4	0–3.3	Silt, rounded gravel (fine to medium), and sand (fine to coarse); pale yellowish brown (10YR6/2).
	3.3–6.5	Semiconsolidated silt and sand (fine to medium), caliche near 56 inches (4.7 feet); yellowish gray (5Y8/1).
	6.5–9.6	Semiconsolidated sand (coarse), gravel (fine to coarse), and silt; grayish orange (10YR7/4).
	9.6–25.0	Semiconsolidated clay and silt (stiff and plastic), fractures coated with yellowish oxide; pale yellowish brown (10YR6/2).
CL-5	0–1.5	Silt, sand (fine to coarse), and rounded gravel (fine to medium); pale yellowish brown (10YR6/2) to grayish orange (10YR7/4).
	1.5–2.3	Semiconsolidated silt and clay; grayish orange (10YR7/4).
	2.3–4.7	Semiconsolidated silt, clay, and gravel (fine to medium); pale yellowish brown (10YR6/2), grayish orange (10YR7/4), and very pale orange (10YR8/2).
	4.7–12.5	Weathered or altered volcanic bedrock?, cuttings are crumbly with abundant phenocrysts; very pale orange (10YR8/2) to grayish orange (10YR7/4).
CL-6	0–2.5	Silt, clay, sand (medium to coarse), and occasional fine gravel; pale reddish brown (10R4/2) to pale red (10R6/2).
	2.5–4.7	Semiconsolidated silt, clay, sand, and rounded gravel (coarse); grayish orange (10YR7/4).
	4.7–27.3	Semiconsolidated, rounded gravel (fine to coarse, up to 3-inch diameter, with caliche coatings near 12.2 feet), silt, clay, and sand; dark yellowish orange (10YR6/6) to grayish orange (10YR7/4).
	27.3–29.8	Semiconsolidated silt, clay, sand (fine to coarse), and rounded gravel (fine); pale yellowish brown (10YR6/2).
CL-7	0–2.3	Silt, sand (fine to coarse), and rounded gravel (fine to medium); pale yellowish brown (10YR6/2).
	2.3–18.3	Semiconsolidated, rounded gravel (fine to coarse), silt, and sand (fine to coarse); pale yellowish brown (10YR6/2).
	18.3–25.2	Semiconsolidated sand (fine to coarse), rounded gravel (coarse), and silt; moderate yellowish brown (10YR5/4).
	25.2–28.7	Semiconsolidated clay, silt, sand (fine to coarse), and rounded gravel (fine); dark yellowish brown (10YR4/2).
	28.7–30.0	Semiconsolidated clay, silt, sand (coarse), and gravel (fine to medium); moderate yellowish brown (10YR5/4).

Appendix B. Lithologic description of soil penetrated by nine soil-chloride test holes, Carson Valley, Nevada and California.—Continued

[Test hole locations are shown in figure 2. Description in approximate order of sediment amount, by volume. Color codes from standard rock-color chart]

Test hole No.	Depth below land surface (feet)	Lithologic description
CL-8	0–3.5	Silt, clay, and sand (fine); dark yellowish brown (10YR4/2).
	3.5–5.3	Semiconsolidated silt, clay, and sand (fine); grayish orange (10YR7/4).
	5.3–7.7	Semiconsolidated silt, sand (fine), and clay; moderate yellowish brown (10YR5/4) to grayish orange (10YR7/4).
	7.7–11.8	Semiconsolidated silt, sand (fine), and clay; dark yellowish brown (10YR4/2) to pale yellowish brown (10YR6/2).
	11.8–17.8	Semiconsolidated silt, clay, and sand (fine); pale yellowish brown (10YR6/2).
	17.8–24.8	Semiconsolidated silt, clay, sand (fine), and rounded gravel (fine to medium); grayish orange (10YR7/4).
	24.8–30.4	Semiconsolidated silt, clay, and sand (fine); pale yellowish brown (10YR6/2).
CL-9	0–3.3	Silt, sand (fine) and rounded gravel (medium); pale yellowish brown (10YR6/2).
	3.3–14.3	Sand (coarse), silt, angular gravel (decomposed granite), and rounded gravel (coarse, up to 2-inch diameter); moderate yellowish brown (10YR5/4) to pale yellowish brown (10YR6/2).
	14.3–19.5	Sand (fine to coarse), and silt, with rounded gravel below 226 inches (18.8 feet); pale yellowish brown (10YR6/2).
	19.5–22.0	No sample retained.
	22.0–25.3	Semiconsolidated sand (fine) and silt; pale yellowish brown (10YR6/2).
	25.3–30.2	Rounded gravel (fine to coarse, up to 3-inch diameter), sand (fine to coarse), and silt; pale yellowish brown (10YR6/2).

Appendix C. Panel, stream, and streambed temperature data for 37 sites, Carson Valley, Nevada and California.

Appendix C lists the 5-minute temperatures recorded at each of the 37 sites. These data are available in an Excel data base for download at URL: http://pubs.water.usgs.gov/sir20055288/appendix/appC.xls.

Appendix D. Adjusted stream and streambed temperatures at gaining sites, losing sites, and neutral sites, and periods used to determine the amplitude of stream and streambed temperature fluctuations, and time difference between peaks for selected sites, Carson Valley, Nevada and California.

GAINING SITES

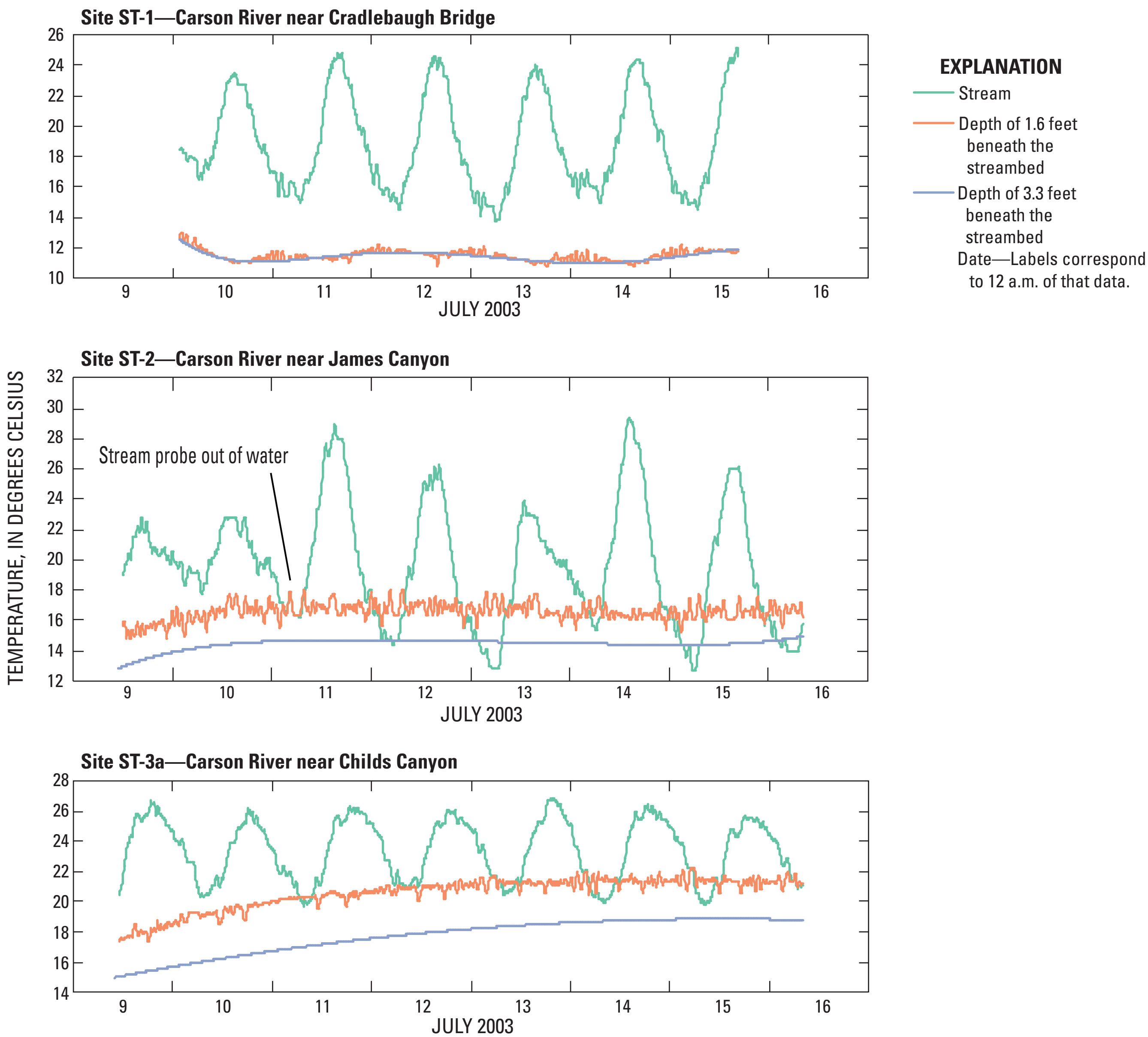


Figure D1. Adjusted stream and streambed temperatures at gaining sites, Carson Valley, Nevada and California.

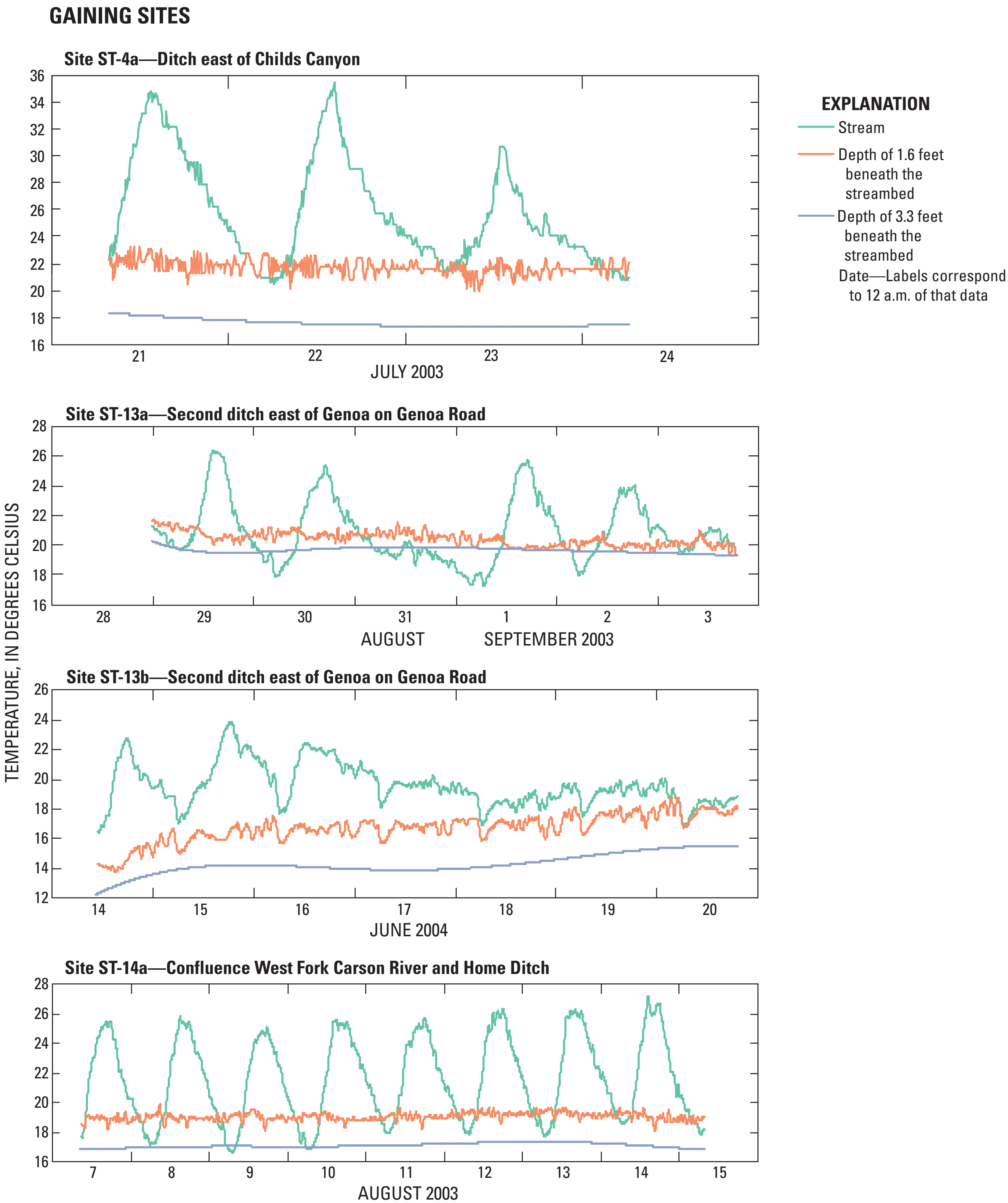
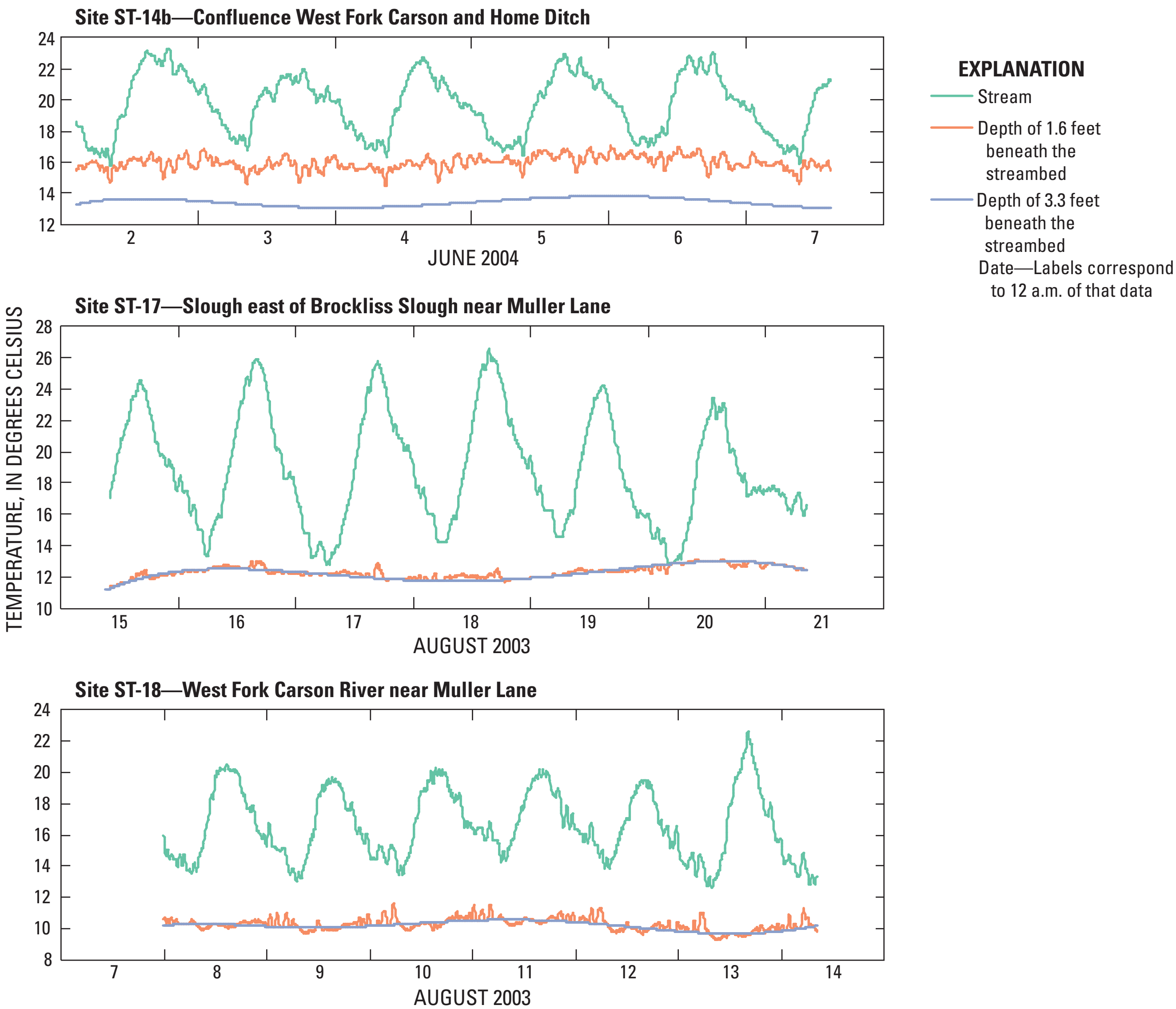


Figure D1.—Continued.

GAINING SITES



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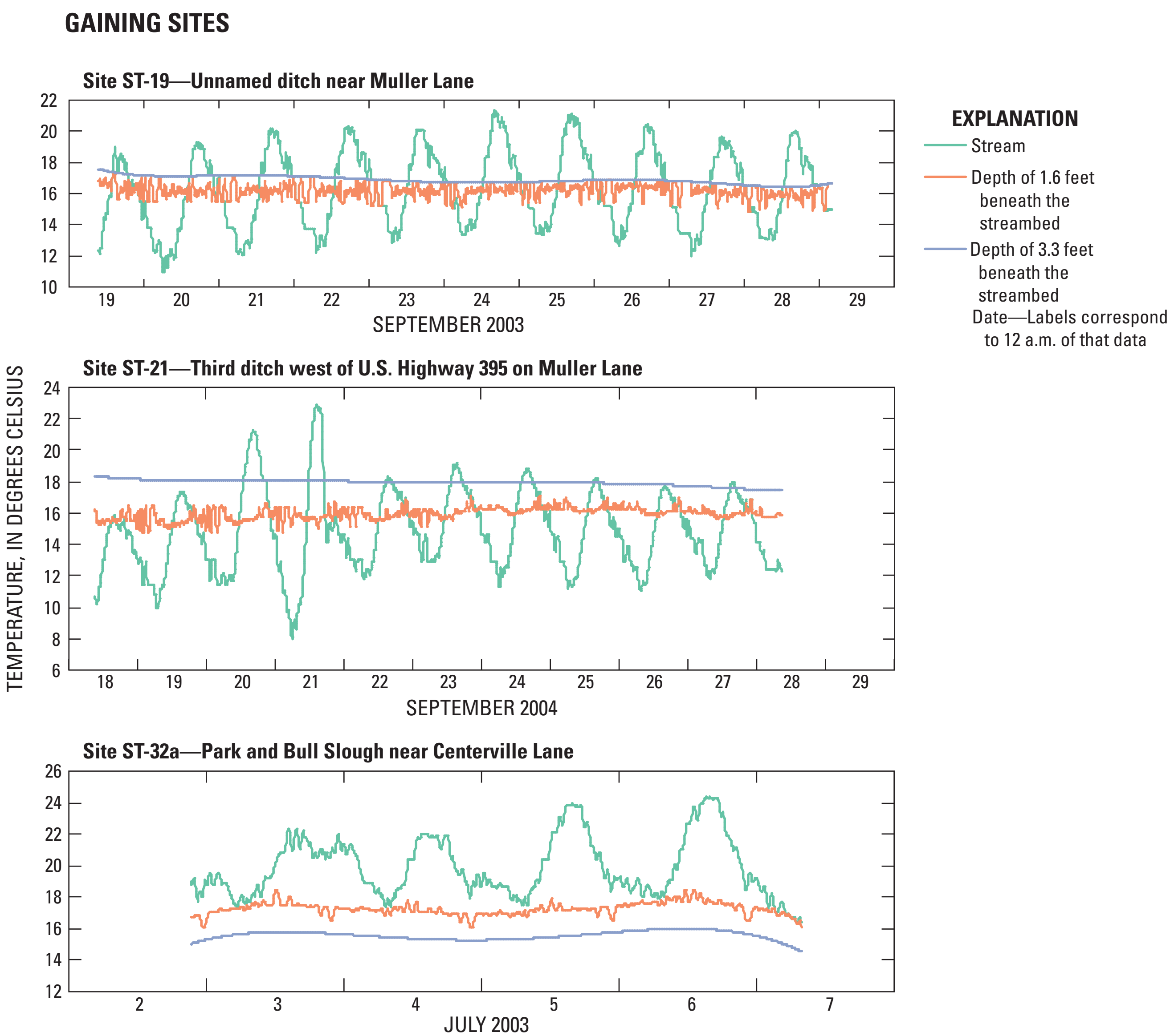


Figure D1.—Continued.

LOSING SITES

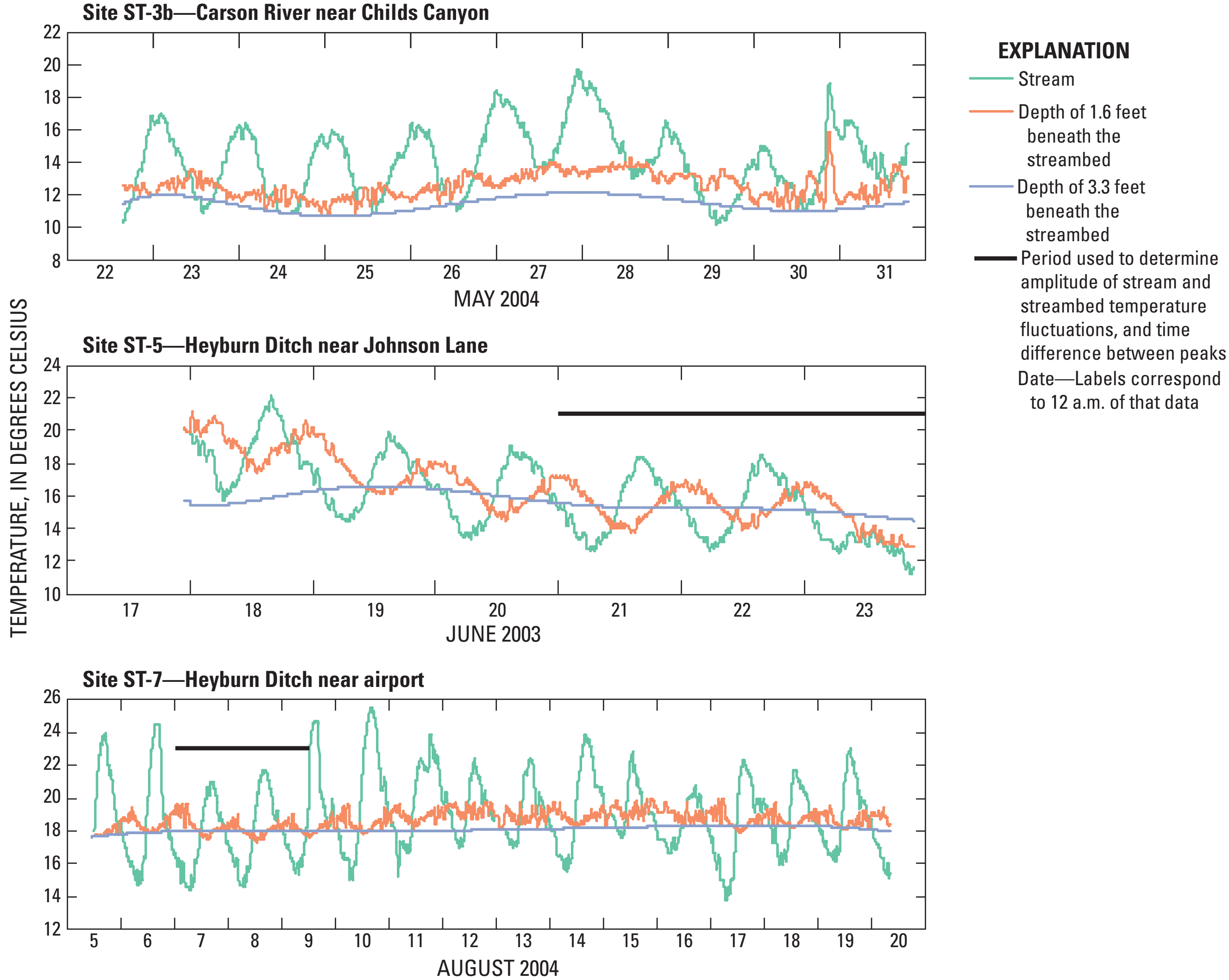
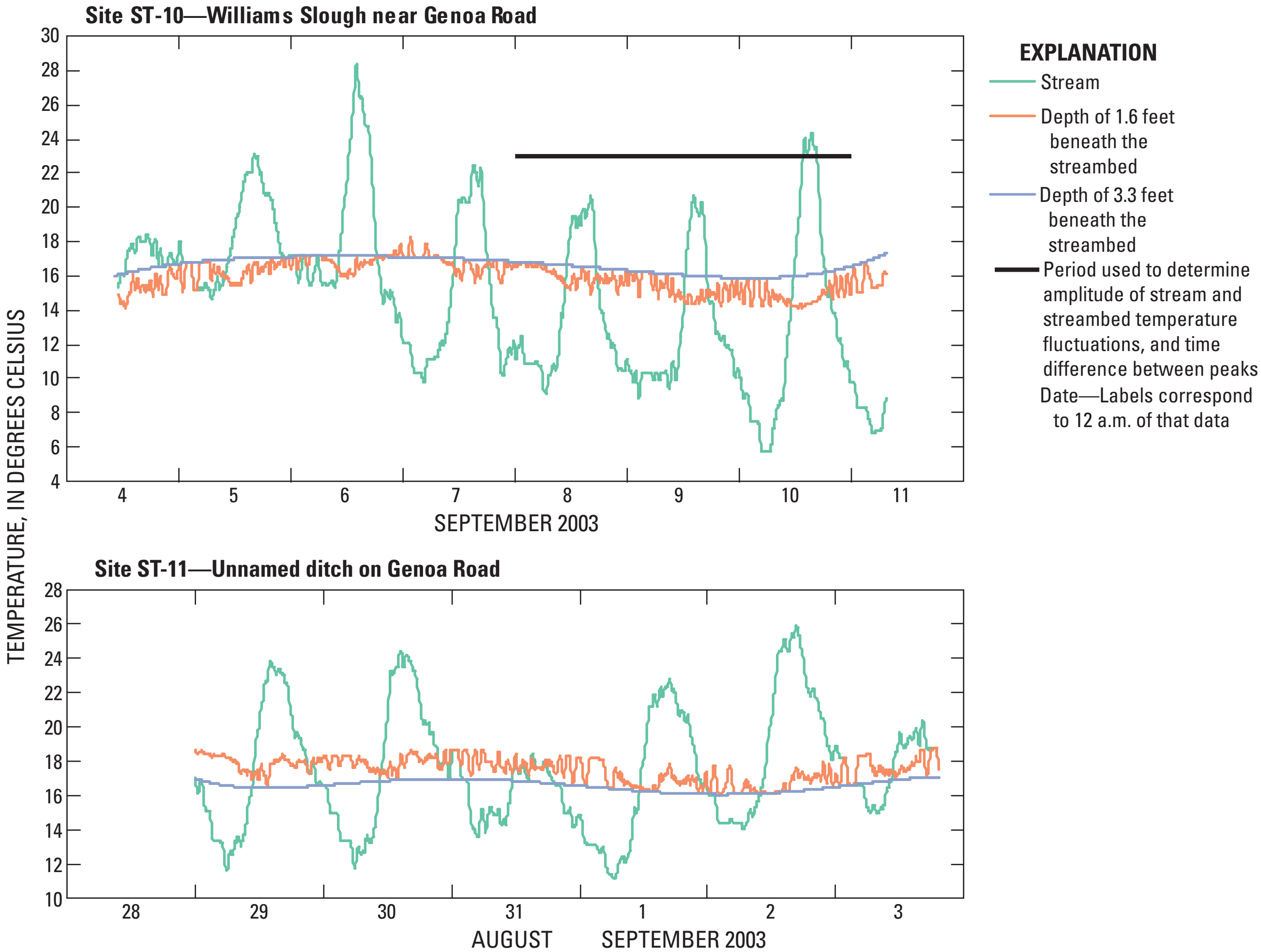


Figure D2. Adjusted stream and streambed temperatures at losing sites and periods used to determine the amplitude of stream and streambed temperature fluctuations, and time difference between peaks for selected sites, Carson Valley, Nevada and California.

LOSING SITES



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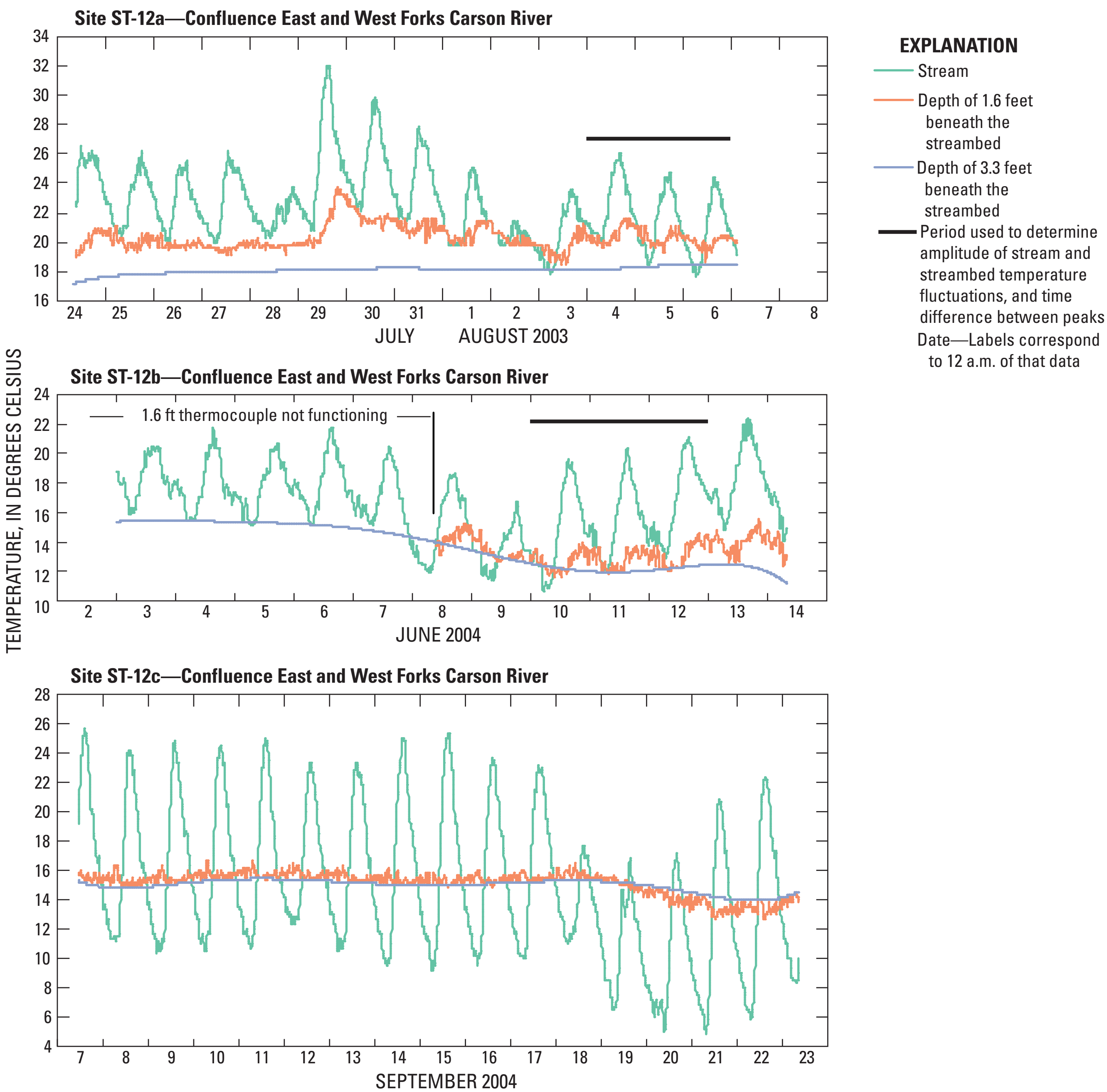

Figure D2.—Continued.

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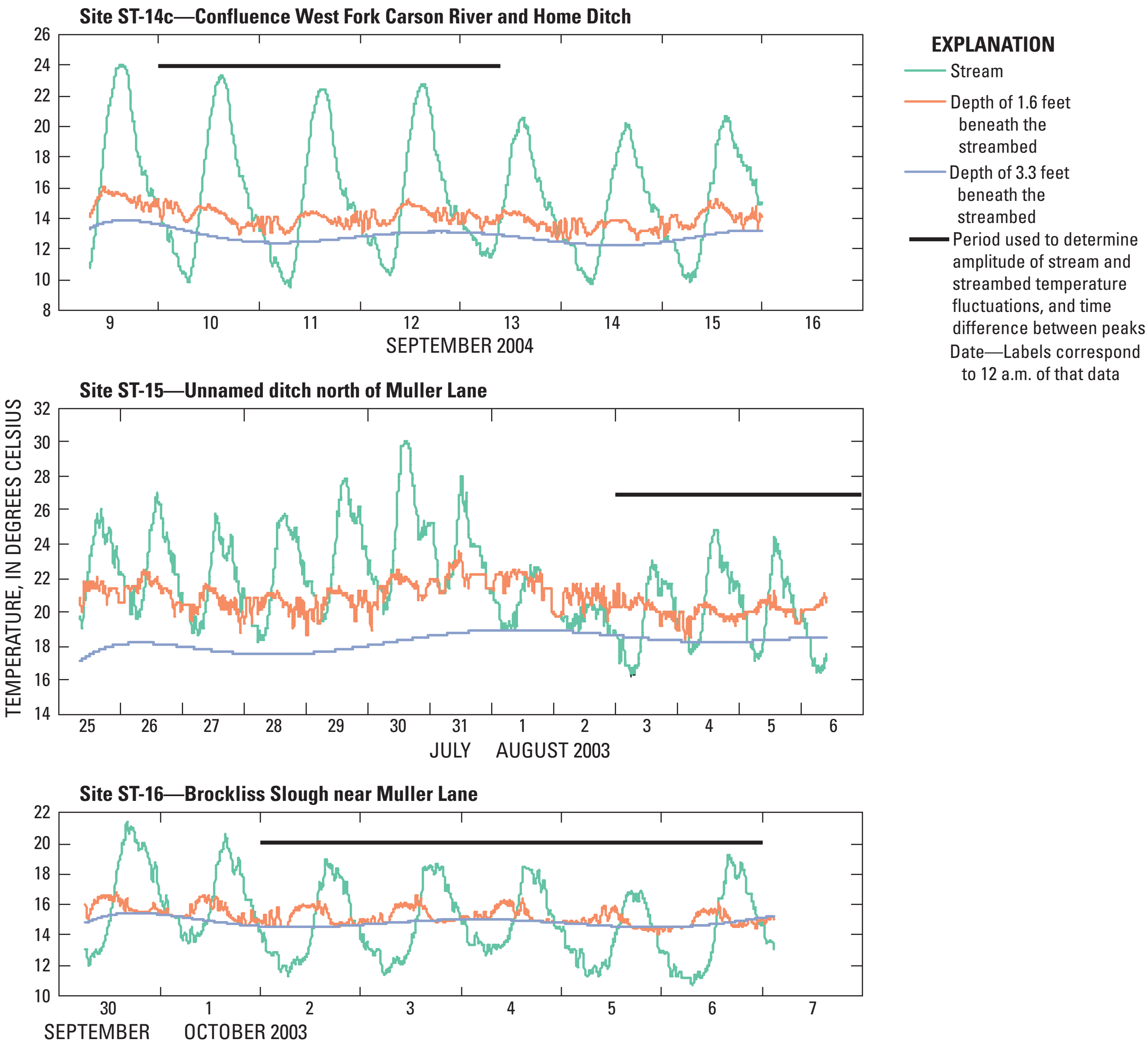

Figure D2.—Continued.

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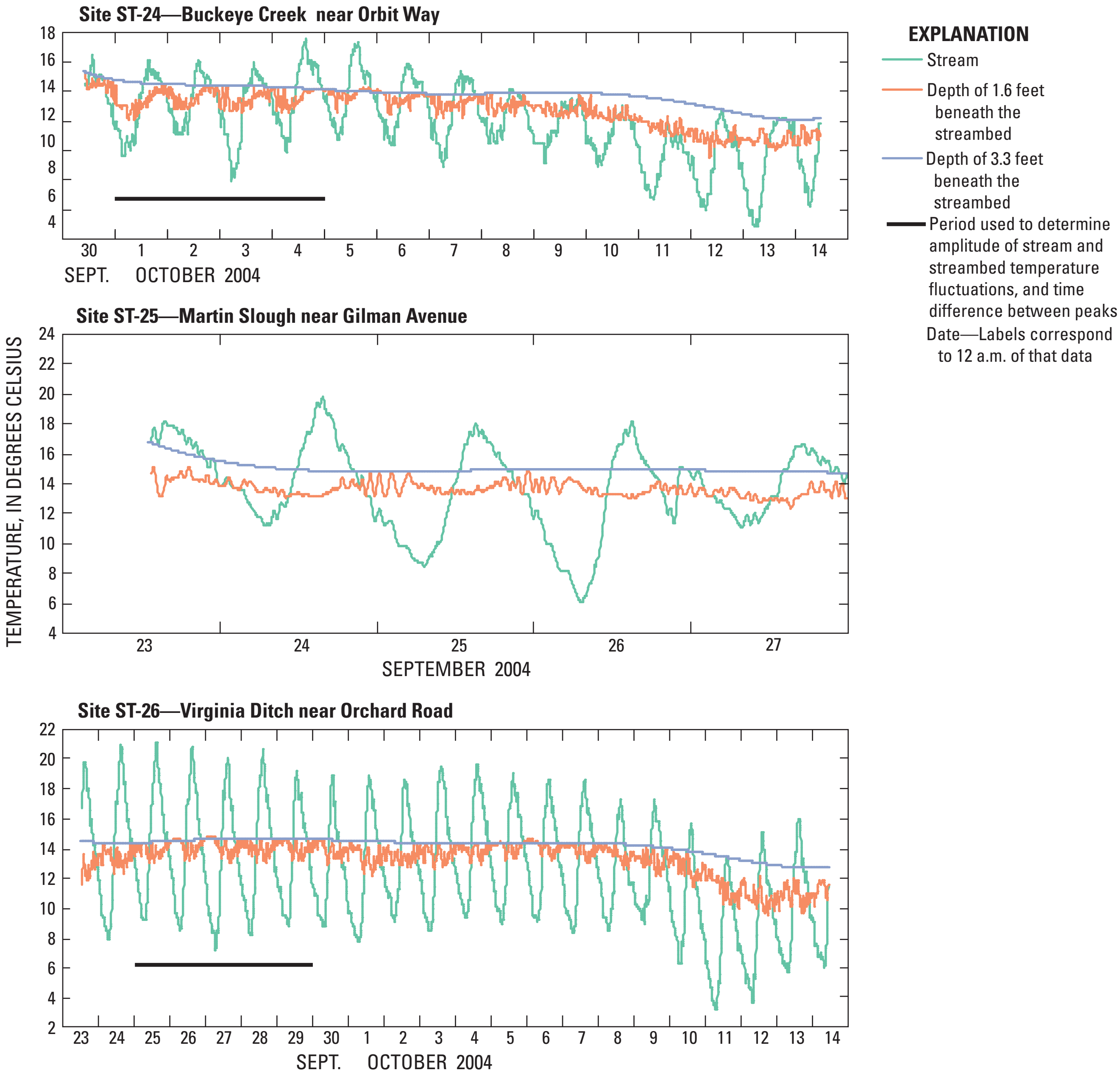

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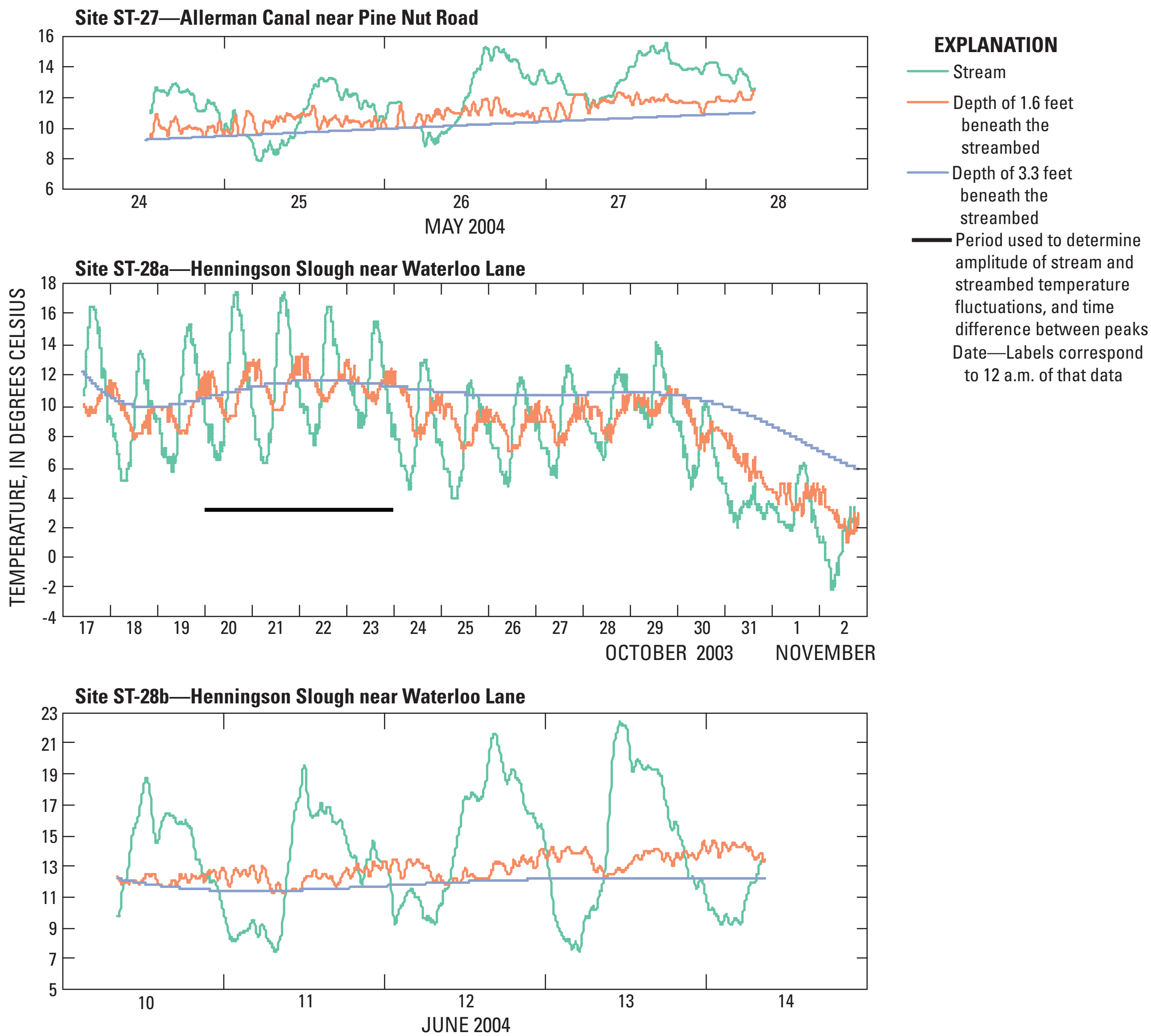

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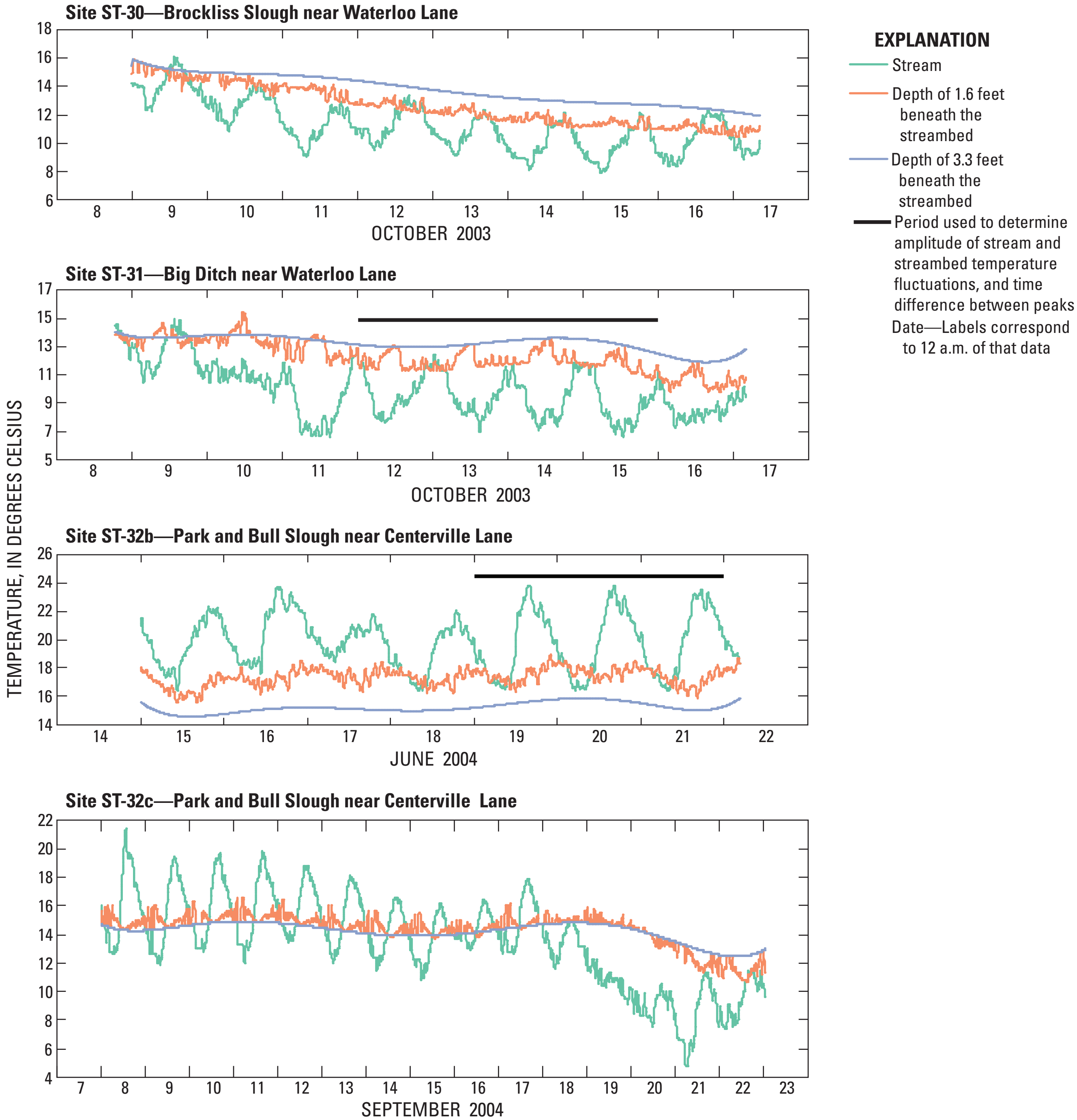

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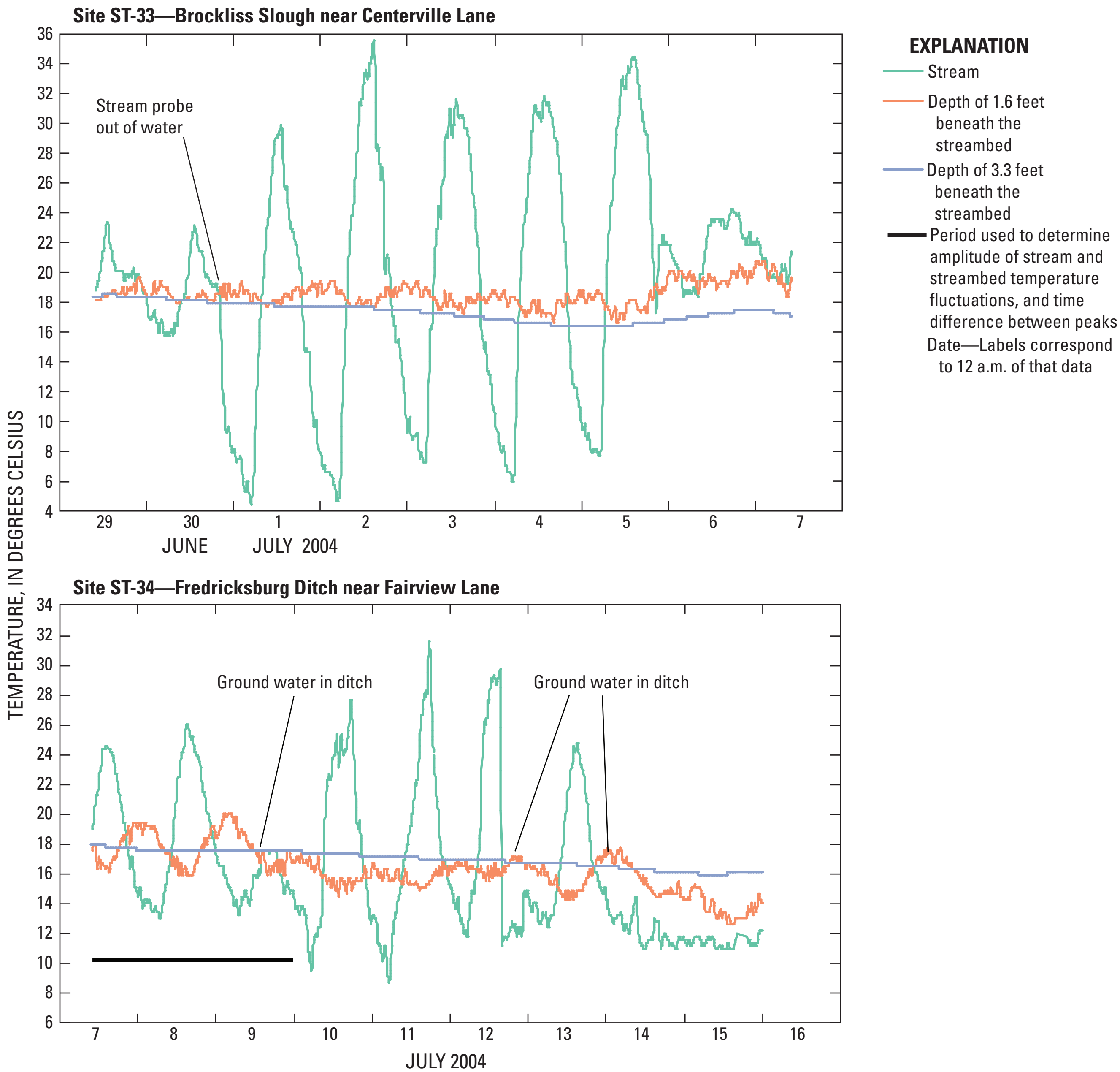

Figure D2.—Continued.

LOSING SITES

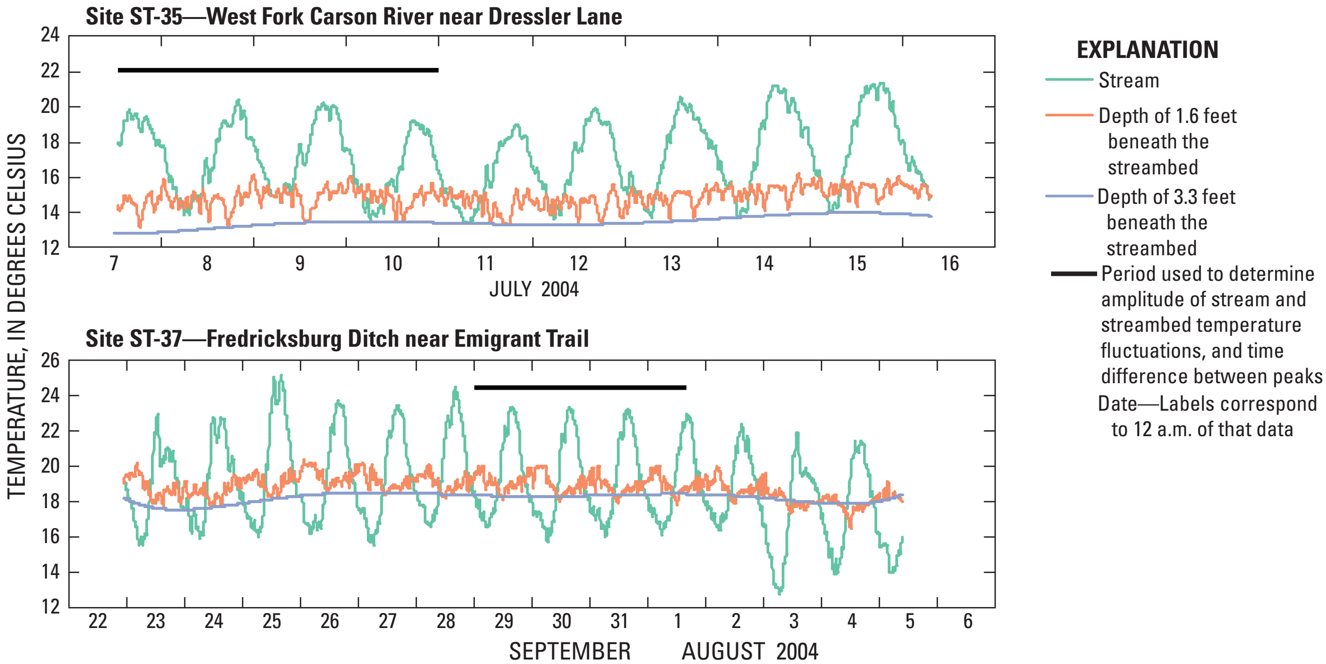

Figure D2.—Continued.

NEUTRAL SITES

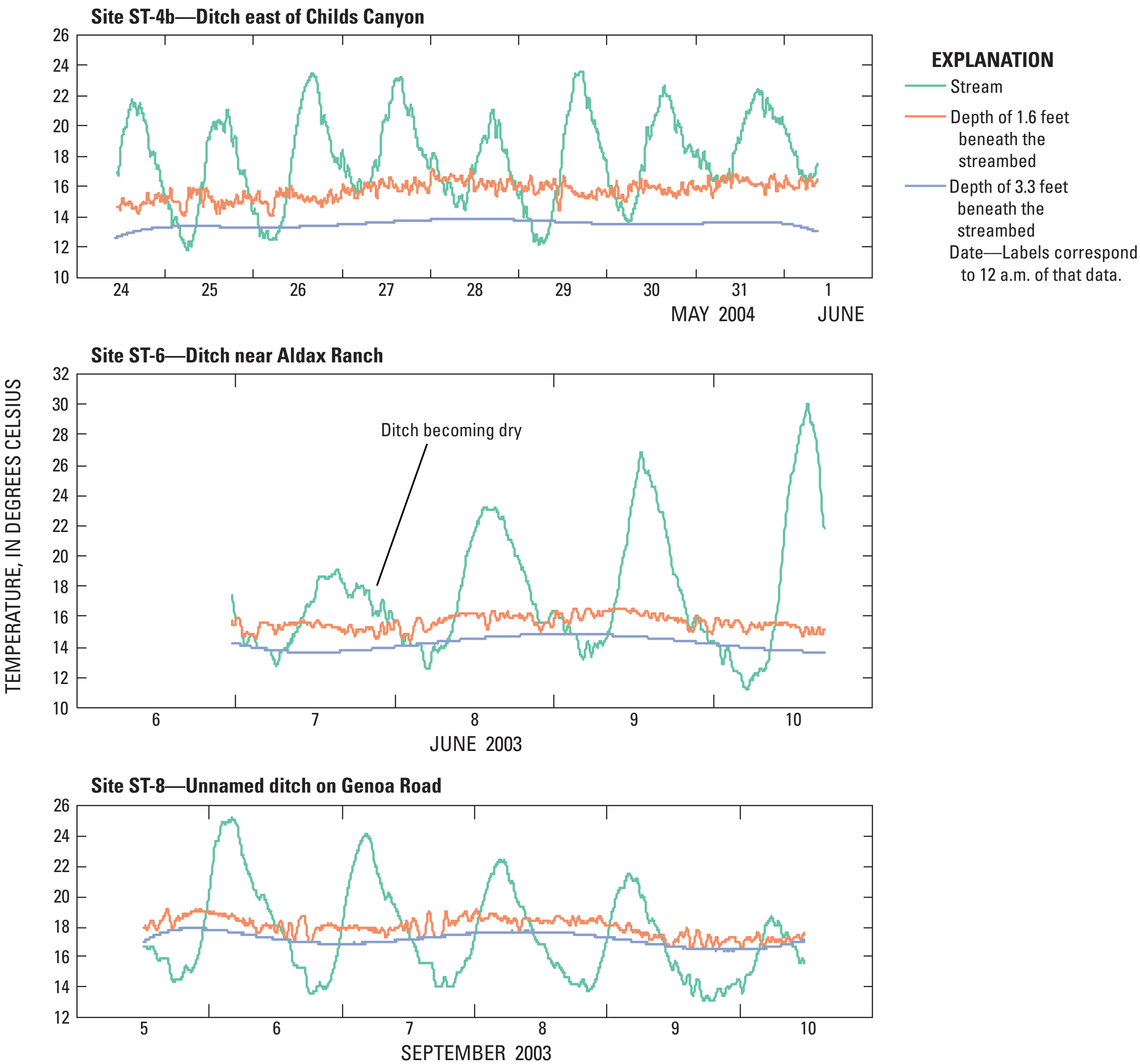

Figure D3. Adjusted stream and streambed temperatures at neutral sites, Carson Valley, Nevada and California.

NEUTRAL SITES

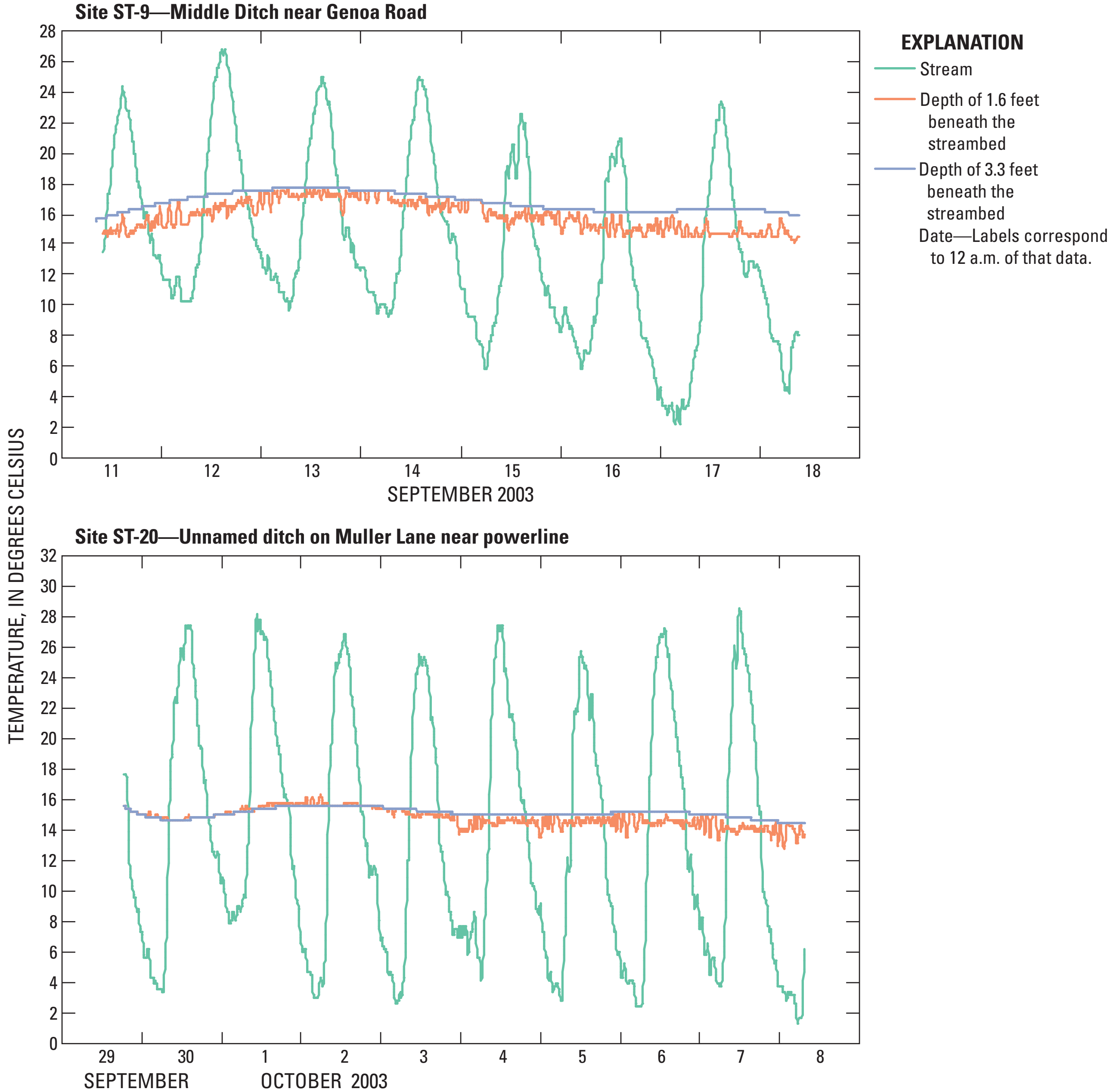

Figure D3.—Continued.

NEUTRAL SITES

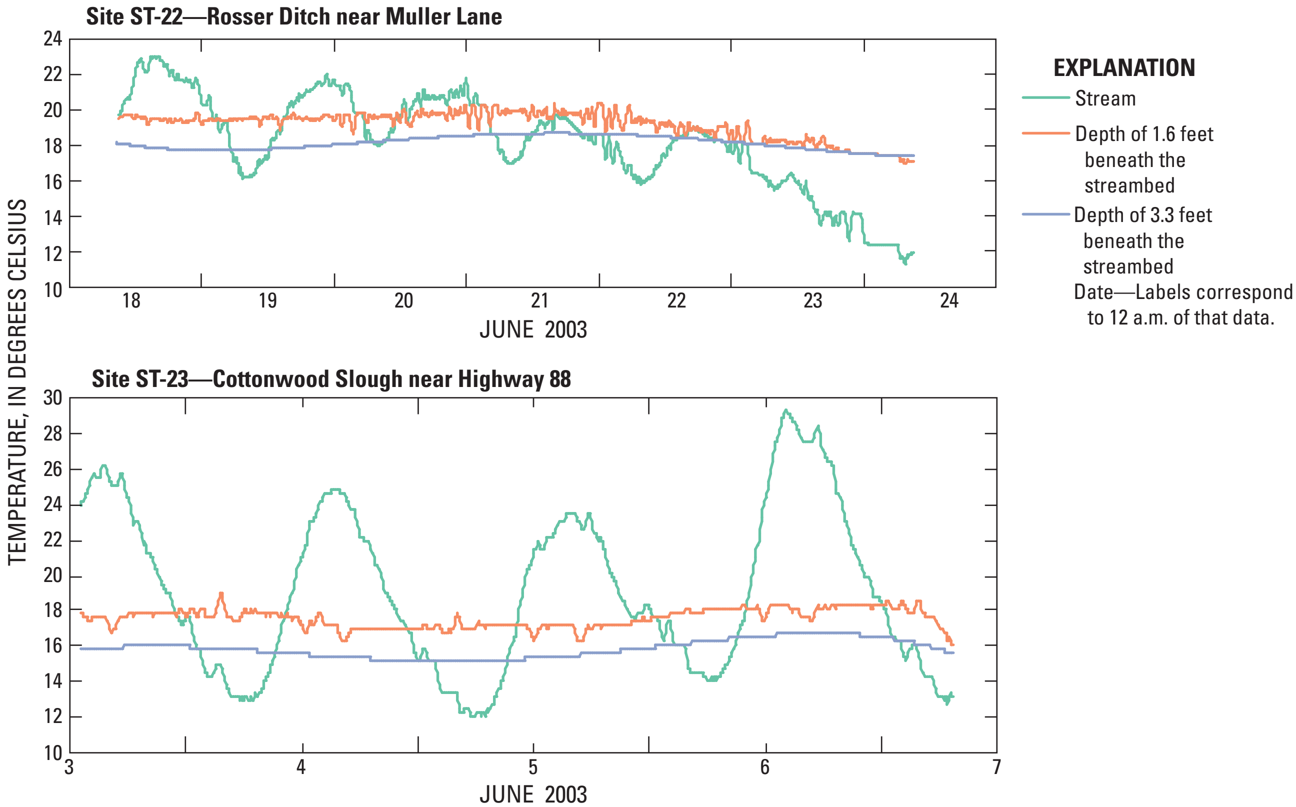

Figure D3.—Continued.

NEUTRAL SITES

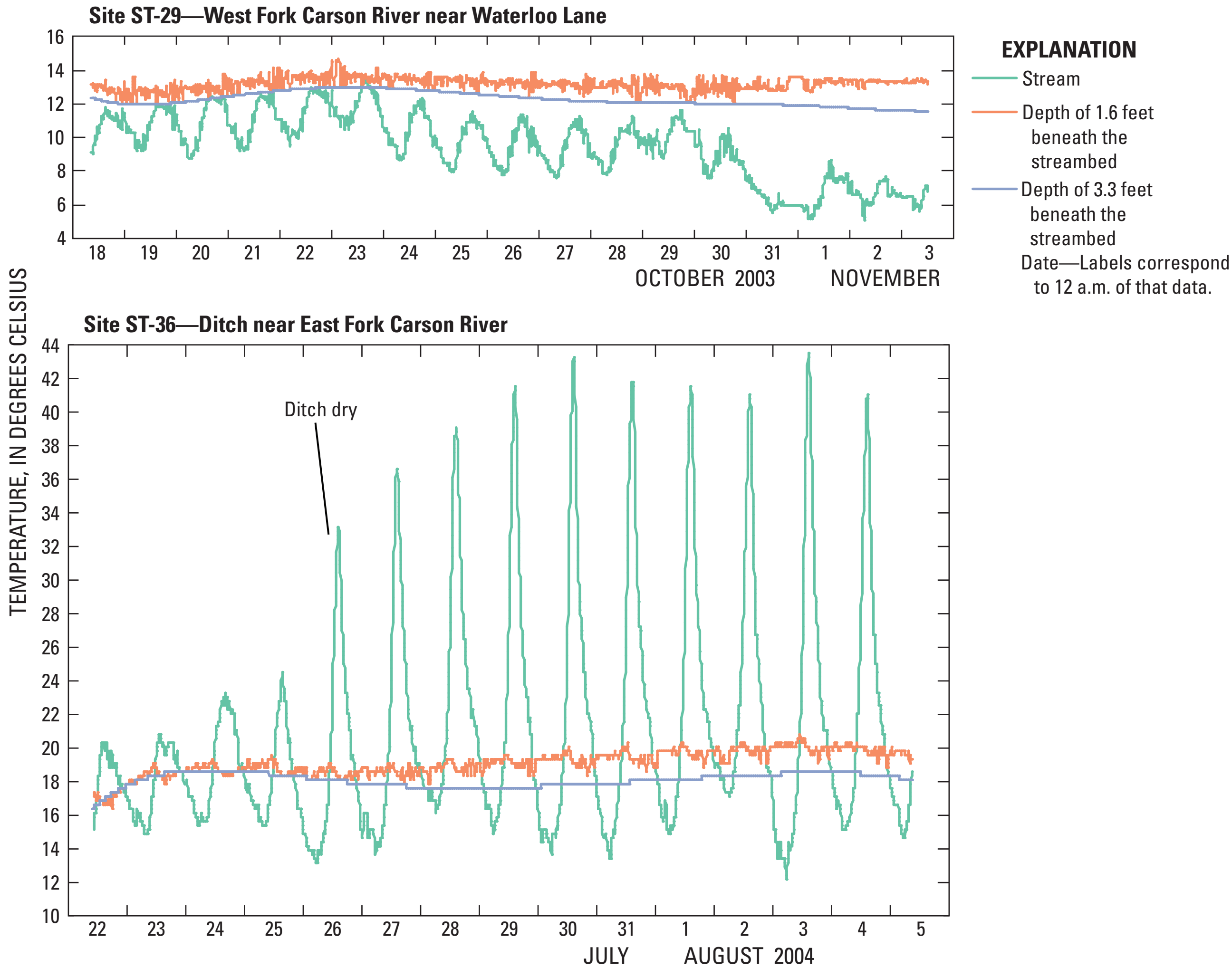

Figure D3.—Continued.

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Maurer and others

Rates of Evapotranspiration, Recharge from Precipitation Beneath Selected Areas of Native Vegetation, and Streamflow Gain and Loss in Carson Valley, Douglas County, Nevada, and Alpine County, California

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