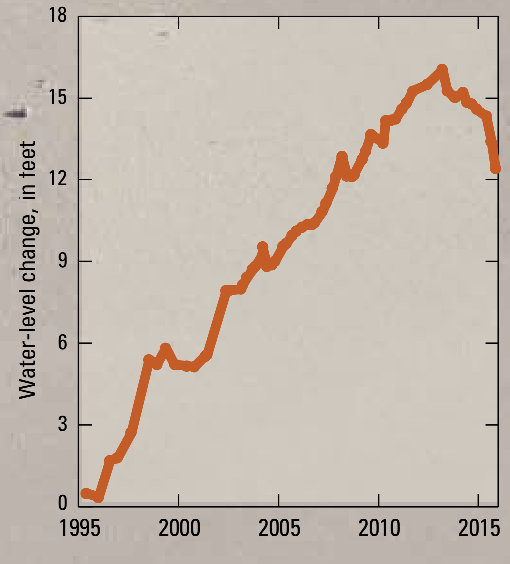
Prepared in cooperation with the Department of Energy, National Nuclear Security Administration Nevada Site Office, Office of Environmental Management under Interagency Agreement, DE-NA0001654

Conceptual Framework and Trend Analysis of Water-Level Responses to Hydrologic Stresses, Pahute Mesa–Oasis Valley Groundwater Basin, Nevada, 1966–2016



Scientific Investigations Report 2018–5064

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Cover: Underground emplacement borehole U-19bh in Pahute Mesa, Nevada. The emplacement borehole is a 96-inch diameter borehole that was drilled to a depth of 2,148 ft below land surface in 1991. The original intended use of the borehole was to test a nuclear device underground; however, the United States signing of the Comprehensive Nuclear Test Ban Treaty in 1992 prohibited nuclear testing from occurring in this borehole. Water levels have been measured in borehole U-19bh from 1991 to present (2018) as part of a long-term water-level monitoring network. The hydrograph of water levels in borehole U-19bh shows a rising trend in response to episodic recharge from multiple wet winters between 1995 and 2016. Photograph by Steven R. Reiner, U.S. Geological Survey, November 18, 2015.

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Conversion Factors

U.S. customary units to International System of Units

Multiply	By	To obtain
	Length	
inch (in.)	2.54	centimeter (cm)
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square foot (ft^2)	929.0	square centimeter (cm^2)
square foot (ft^2)	0.09290	square meter (m^2)
	Volume	
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m^3)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
	Flow rate	
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year (m^3/yr)
acre-foot per month (acre-ft/mo)	14,802	cubic meter per year (m^3/yr)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

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 1983 (NAD 83).

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

Supplemental Information

Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness $[(ft^3/d)/ft^2]ft$. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Kiloton (kt): Equal to the power of 1,000 tons of TNT and megaton (MT): TNT equivalent, explosive energy equal to 4.184 petajoules.

Abbreviations

BWSD	Beatty Water and Sanitation District
LOWESS	Locally Weighted Scatterplot Smoothing
NNSS	Nevada National Security Site
NOAA	National Oceanic and Atmospheric Administration
NWIS	National Water Information System
PMOV	Pahute Mesa–Oasis Valley
USGS	U.S. Geological Survey
WLM	Water-level model

Conceptual Framework and Trend Analysis of Water-Level Responses to Hydrologic Stresses, Pahute Mesa–Oasis Valley Groundwater Basin, Nevada, 1966–2016

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Abstract

This report identifies water-level trends in wells and provides a conceptual framework that explains the hydrologic stresses and factors causing the trends in the Pahute Mesa–Oasis Valley (PMOV) groundwater basin, southern Nevada. Water levels in 79 wells were analyzed for trends between 1966 and 2016. The magnitude and duration of water-level responses to hydrologic stresses were analyzed graphically, statistically, and with water-level models.

The conceptual framework consists of multiple stress-specific conceptual models to explain water-level responses to the following hydrologic stresses: recharge, evapotranspiration, pumping, nuclear testing, and wellbore equilibration. Dominant hydrologic stresses affecting water-level trends in each well were used to categorize trends as nonstatic, transient, or steady state.

Nonstatic water levels are affected by wellbore equilibration. Water-level hydrographs of four wells open to volcanic tuffs are dominated by wellbore equilibration, where hydraulic conductivities range between 0.01 and 2×10^{-4} feet per day and the period of water-level recovery spans from less than 1 to more than 20 years.

Transient trends resulted from nearby nuclear testing and (or) pumping. Long-term water-level responses to nuclear testing have occurred in five wells, and the responses differ depending on whether the well is near or far from the point of detonation. Well *Beatty Wash Terrace* is the only study area well affected by municipal pumping for the town of Beatty, Nevada. Water levels in six Pahute Mesa wells have been affected by groundwater pumping from water-supply well *U-20 WW* and potentially have been affected by nearby nuclear testing.

Steady-state trends reflect departures from long-term average water levels caused by short-term variability in recharge and evapotranspiration. A conceptual model of episodic recharge and steady aquifer discharge is presented to reconcile the definition of steady state (no change with time) with naturally fluctuating groundwater levels. An assumed century-scale period of steady state was tested by determining whether the conceptual model can explain naturally occurring, rising water levels in the study area. Graphical and statistical analyses indicate that 43 of the 62 wells with steady-state trends have upward trends from 1995 to 2016. The conceptual model shows that the study area has been in a relatively wet period from 1968 to 2016, and both conceptual and water-level model results indicate that the observed rising trends can be explained by episodic recharge from multiple wet winters over the last several decades. These rising trends are considered short-term (decadal) fluctuations within the long-term (century-scale) period of steady-state equilibrium.

Steady-state trends were categorized into eleven geographic areas based on water-level responses to episodic recharge and other factors affecting the trends. In each geographic area, water levels respond similarly to recharge, where trends were influenced by transmissivity, unsaturated zone depth, and (or) proximity to recharge areas.

Groundwater recharge is temporally and spatially variable in the study area. Recharge responses to the 1995, 1998, and 2005 winters were ubiquitous. Recharge responses to the 2000 winter were observed in wells below an altitude of 4,800 feet. Recharge responses to the 2001 and 2010 winters were observed in wells within the Cactus Range, Yucca Mountain, northern Oasis Valley, Fortymile Wash, and Amargosa Narrows. Recharge responses to the 2011 winter were observed only in wells within Pahute, Buckboard, and Rainier Mesas.

The conceptual framework of water-level responses to hydrologic stresses and trend analyses provide a comprehensive understanding of the PMOV basin and vicinity. The trend analysis links water-level fluctuations in wells to hydrologic stresses and potential factors causing the trends. Transient and steady-state trend categorizations can be used to determine the appropriate water-level data for groundwater studies.

Introduction

The Pahute Mesa–Oasis Valley (PMOV) groundwater basin is in Nye County, southern Nevada (fig. 1). The PMOV basin incorporates historic underground nuclear testing areas in Pahute Mesa, which is part of the Nevada National Security Site (NNSS). A total of 85 nuclear tests were detonated underground within the PMOV basin between 1965 and 1992 (U.S. Department of Energy, 2015), where 77 of 85 tests were detonated near or below the water table and potentially have introduced test-generated contaminants into the groundwater system (Laczniak and others, 1996). Because of potential public health concerns posed by nuclear testing, the U.S. Department of Energy and other Federal and State agencies are interested in the rate and movement of radionuclide contaminants in groundwater migrating beyond the NNSS boundary (U.S. Department of Energy, 2009). Radionuclides are migrating downgradient toward Oasis Valley, near the community of Beatty, Nevada (Pawloski and others, 2010; Fenelon and others, 2016; Russell and others, 2017). Accurate conceptualization of the groundwater-flow system will aid current and future studies of groundwater flow and contaminant transport.

Analysis of water-level trends in well hydrographs can be used to better understand and conceptualize a groundwater-flow system. A water-level trend reflects the summation of all natural and anthropogenic hydrologic stresses acting on the aquifer at the well location. Common hydrologic stresses in the PMOV basin include recharge, evapotranspiration, pumping, nuclear testing, and water-level equilibration following localized disturbances in the wellbore.

A trend analysis interprets water-level fluctuations to provide a conceptualization of the groundwater-flow system for future groundwater studies. Trend analysis results can be used to guide groundwater studies on the use of specific water-level data in the development of potentiometric maps, and computation of vertical hydraulic gradients and groundwater-flow paths. Trend analysis interpretations also can be used in steady-state and transient groundwater-flow models. For example, water-level trends are analyzed to determine whether water levels only are affected by climatic conditions, such as recharge, or whether water levels also are affected by pumping or nuclear testing. A groundwater modeler can use the results of the trend analysis to determine appropriate spatial and (or) temporal boundary conditions. A trend analysis also provides information on the appropriate water-level data to use in a steady state or transient numerical model. Trend analysis results (and their implementation in numerical models) aid in the understanding of groundwater-flow systems and the forecast of radionuclide transport rates and directions.

The U.S. Geological Survey (USGS), in cooperation with the Department of Energy, completed a study to identify groundwater-level trends and provide a set of conceptual models to explain the trends within and near the PMOV basin. The conceptual models form a framework for understanding how groundwater levels respond to natural (climatic, barometric, tidal) and anthropogenic (pumping, nuclear testing) hydrologic stresses.

Purpose and Scope

This report documents results of a trend analysis for groundwater levels within and near the PMOV basin. The study objectives were to (1) identify trends in groundwater levels in wells and (2) provide a conceptual framework that explains the hydrologic stresses and factors (or potential factors) causing the trends. Common stresses evaluated include precipitation-derived recharge, groundwater evapotranspiration, groundwater pumping, nuclear testing, and water-level equilibration following localized disturbances in the wellbore.

The conceptual framework consists of stress-specific conceptual models that explain water-level responses to each hydrologic stress. Conceptual models were developed to explain nonstatic, transient, and steady state water-level trends. Definitions for these trend categorizations are provided in detail within this report.

Quantitative and qualitative methods were used to analyze trends. Graphical and statistical methods analyzed groundwater levels for upward or downward trends, whereas water-level models identified potential hydrologic stresses causing the trends. Temporal and spatial variability in water-level trends was linked to factors (or potential factors) affecting the trends, such as distance to recharge areas, transmissivity, and unsaturated zone depth. The trend analysis included 79 wells, where water-level data were collected from 1966 to 2016. All pertinent data and models are published in Jackson (2018).

Description of Study Area

The study area encompasses the PMOV groundwater basin and areas within about 5 miles (mi) of the basin boundary, except in the Yucca Mountain area where the study area extends farther south (fig. 1). The PMOV basin is part of the Death Valley regional groundwater-flow system within the Great Basin physiographic province (Harrill and Prudic, 1998). Dominant topographic features bounding the PMOV basin include: Cactus Range and Cactus Flat to the north; the Belted Range to the east; Timber Mountain and Bare Mountain to the south; and Bullfrog Hills, Sarcobatus Flat, and Black Mountain to the west. Beatty, Nevada in southern Oasis Valley is near the terminus of the groundwater basin. Farther south in the Alkali Flat–Furnace Creek Ranch groundwater basin are Yucca Mountain, Shoshone Mountain, and Jackass Flats (fig. 2). Altitudes in the study area range from about 3,300 feet (ft) near Beatty to about 8,300 ft in the Kawich and Belted Ranges (fig. 1).

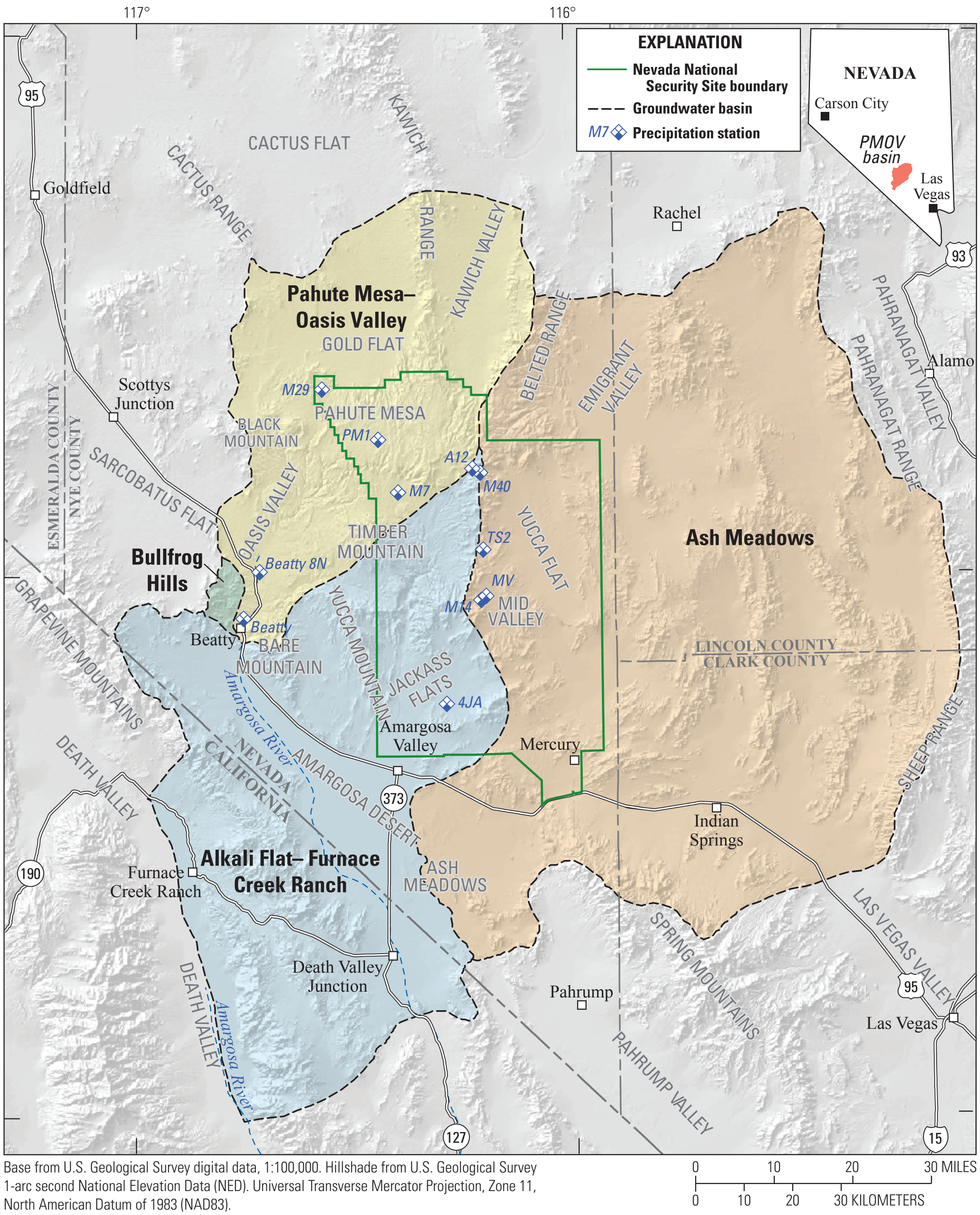


Figure 1. Geographic features and groundwater basins surrounding the Pahute Mesa–Oasis Valley groundwater basin, Nevada and California. Modified from Fenelon and others (2016).

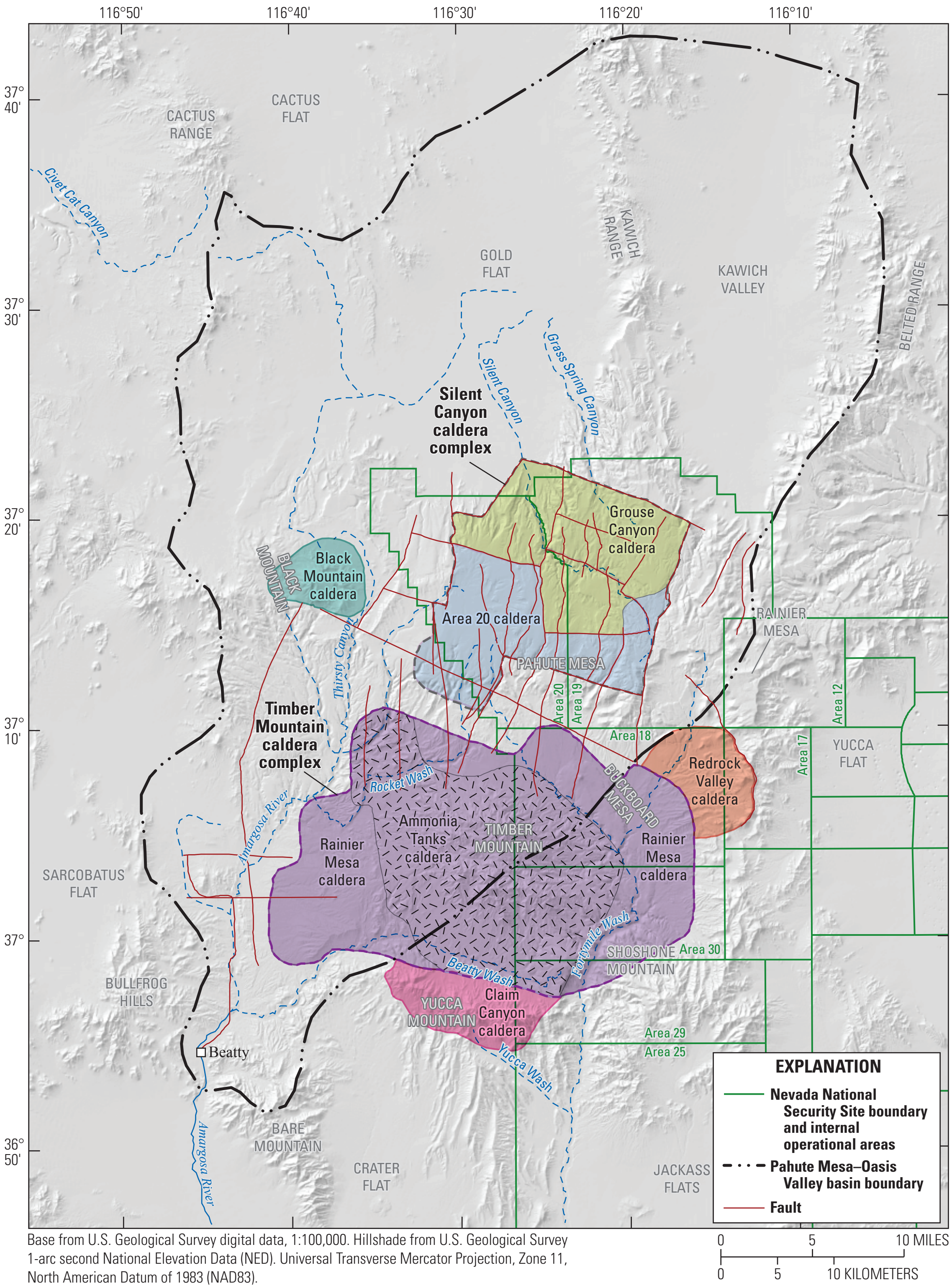


Figure 2. Physiography and geologic structures in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

Hydrogeology

Geologic processes during the Cenozoic Era dominate the hydrogeologic framework of the study area. From the mid-to-late Tertiary Period, the Basin and Range structural province began to form by crustal extension, resulting in the formation of low- and high-angle, northwest- and northeast-striking normal and strike-slip faults (fig. 2; Guth, 1981; Wernicke and others, 1988). Concurrent with crustal extension, multiple volcanic eruptions formed an extensive caldera complex of partially overlapping caldera vents, known as the southwest Nevada volcanic field (Christiansen and others, 1977; Byers and others, 1989). Basin and Range crustal extension continued throughout and after volcanic episodes, causing lateral translation, tilting, and vertical offset of geologic units in the study area. Vertical offsets along faults have displaced rocks by more than 1,000 ft in Pahute Mesa (U.S. Department of Energy, 2010). From late Tertiary to present day, gravel, sand, silt, and clay have been deposited within modern alluvial basins.

The southwest Nevada volcanic field was formed by successive volcanic eruptions that occurred during the Miocene Epoch between 16 and 8 million years ago (Byers, Carr, Orkild, and others, 1976; Sawyer and others, 1990; 1994). The volcanic eruptions deposited thick sequences of rhyolitic, andesitic, and dacitic lava flows as well as welded and nonwelded tuffs into at least seven large calderas (fig. 2; Laczniak and others, 1996). Two extensive caldera complexes include (1) the Silent Canyon caldera complex at Pahute Mesa, formed by the partial overlap of the Grouse Canyon and Area 20 calderas (Sawyer and others, 1994) and (2) the Timber Mountain caldera complex, formed by the partial overlap of the Ammonia Tanks and Rainier Mesa calderas (Byers, Carr, Christiansen, and others, 1976; Byers, Carr, Orkild, and others, 1976; Sawyer and others, 1994). Three additional calderas include: the (1) Black Mountain caldera; (2) Claim Canyon caldera; and (3) Redrock Valley caldera (fig. 2; Hildenbrand and others, 2006; National Security Technologies, LLC, 2007). Volcanic deposits are more than 10,000 ft thick within the caldera margins centered on Pahute Mesa, Black Mountain, and Timber Mountain (Blankennagel and Weir, 1973).

Cenozoic volcanic rocks form the principal aquifers and confining units in the study area. Rhyolitic-to-dacitic lava flows and moderately to densely welded ash-flow tuffs form local and regional volcanic aquifers. Lava flows and densely welded ash-flow tuffs are characterized by high fracture permeability and form volcanic aquifers when fracturing causes a laterally extensive and hydraulically connected fracture network. Lava flows typically form local aquifers because lavas are deposited within the caldera complex, whereas welded ash-flow tuffs typically form regional aquifers because welded tuffs are deposited within and outside the caldera complex. Partially welded ash-flow tuffs, and nonwelded ash-flow and ash-fall tuffs typically form confining units because these rocks are susceptible to mineral alteration to zeolites, which reduces permeability (Blankennagel and Weir, 1973; Winograd and Thordarson, 1975). Furthermore, partially welded and nonwelded tuffs are characterized by low fracture permeability because, compared to lava flows and densely welded tuffs, these rocks are less susceptible to strain fracture and, if fractured, are more likely to reseal (Fenelon, 2000).

Cenozoic valley-fill sediments form secondary aquifers in the study area. Valley-fill aquifers of unconsolidated gravel and sand are characterized by high porosity and permeability, but do not contain significant volumes of groundwater because of limited saturated thickness. In Jackass Flats, the saturated thickness in valley-fill aquifers is limited because, even though alluvial deposits have local thicknesses of more than 1,000 ft, the water table is more than 700 ft below land surface (Fenelon and others, 2010). In Oasis Valley, the saturated thickness in valley-fill aquifers is limited to less than 500 ft.

Paleozoic carbonate rocks form aquifers predominantly outside the PMOV basin. Carbonate rocks crop out in Rainier Mesa and underlie parts of Oasis Valley, Buckboard Mesa, the Belted Range, Yucca and Jackass Flats, and Bare, Shoshone, and Yucca Mountains (Laczniak and others, 1996). Within and directly adjacent to the study area, carbonate-rock aquifers typically are localized and not hydraulically well connected (Belcher and Sweetkind, 2010; Fenelon and others, 2010). Carbonate-rock aquifers are unconformably overlain by valley-fill and volcanic rocks (Fenelon and others, 2010).

Siliciclastic and granitic rocks form confining units in the study area. A siliciclastic confining unit forms a hydrologic barrier near the southern part of the PMOV basin, where water is forced to the surface in springs and seeps at Oasis Valley (Laczniak and others, 1996). The siliciclastic confining unit has a maximum thickness of about 6,500 ft, and consists of Mississippian silica-cemented conglomerates, sandstones, siltstones, and shale (Laczniak and others, 1996). Minor granitic intrusions of low permeability formed to the north of Rainier Mesa during the Cretaceous period (Hodges and Walker, 1992).

Source and Movement of Groundwater

Sources of groundwater recharge in the study area include infiltration of precipitation on volcanic highlands and infiltration of surface runoff downgradient of highland areas into alluvial deposits (Blankennagel and Weir, 1973). Groundwater recharge occurs as precipitation infiltrates permeable rocks and percolates below the root zone to the water table, either through an interconnected network of fractures or the rock matrix. Greater amounts of precipitation occur at higher altitudes; consequently, greater amounts of recharge typically occur in highland areas. An exception to this conceptualization occurs where low-permeability rocks underlie highland areas. In these areas, precipitation from snowmelt (or high-intensity rainfall) flows downgradient and infiltrates into adjacent alluvial-fan deposits.

Potential groundwater recharge occurs in volcanic highland areas such as Pahute and Rainier Mesas, Black, Timber, Shoshone, Yucca, and Bare Mountains, and the Kawich and Belted Ranges (fig. 3). Localized low-permeability rocks in Rainier Mesa create perched and semi-perched water tables thousands of feet above the regional water table, where perched groundwater slowly moves laterally and vertically to the regional water table (Laczniak and others, 1996). Localized low-permeability rocks in Bare Mountain also have created a semi-perched groundwater system (Fenelon and others, 2016). The Belted and Kawich Ranges also are composed of low-permeability siliciclastic rocks and likely have semi-perched systems (Fenelon and others, 2016).

Groundwater flow in the study area moves in a south-southwest direction (Fenelon and others, 2016). In the PMOV basin, groundwater moves from areas of recharge, such as Pahute Mesa, and discharge in Oasis Valley (fig. 3). Southeast of the PMOV basin boundary, groundwater moves from Rainier Mesa, Shoshone Mountain, and Yucca Mountain to the south-southwest toward the Amargosa Desert.

In southern Nevada, most recharge is derived from precipitation that occurs during the winter season (Winograd and others, 1998). Precipitation, in the form of rain or snow, typically occurs from late autumn to early spring, herein termed the winter season (Fenelon and others, 2010). Recharge is limited or nonexistent during the summer because high temperatures and growing plants drive the process of evapotranspiration. Most, if not all, of the precipitation that infiltrates the soil zone during the summer is lost to evapotranspiration (Smith and others, 2017).

Infiltration losses along the Amargosa River and its major tributaries during ephemeral flows produce small amounts of recharge (Claassen, 1985; Savard, 1998; Stonestrom and others, 2007). The Amargosa River has perennial reaches maintained by groundwater discharge in Oasis Valley near Beatty. Downgradient of Beatty in the Amargosa Desert, the Amargosa River is an ephemeral channel that is dry greater than 98 percent of the time (Stonestrom and others, 2007). Thirsty Canyon, Rocket Wash, Beatty Wash, and Fortymile Wash are major ephemeral tributaries to the Amargosa River that drain the southern half of the PMOV basin (fig. 3). Greater recharge likely occurs in these tributaries, compared to the Amargosa River, because their channels are confined between steep valley walls that limit the floodplain, are near upland recharge areas, and are underlain by coarse-grained sediments (Claassen, 1985). Therefore, these tributaries likely receive focused recharge during ephemeral flows from the many small tributaries draining the surrounding highlands.

Study Methods

Trends were analyzed using water-level data from 79 wells in or adjacent to the PMOV basin (fig. 3). Wells adjacent to the basin boundary were used to supplement interpretations of water-level trends in the basin.

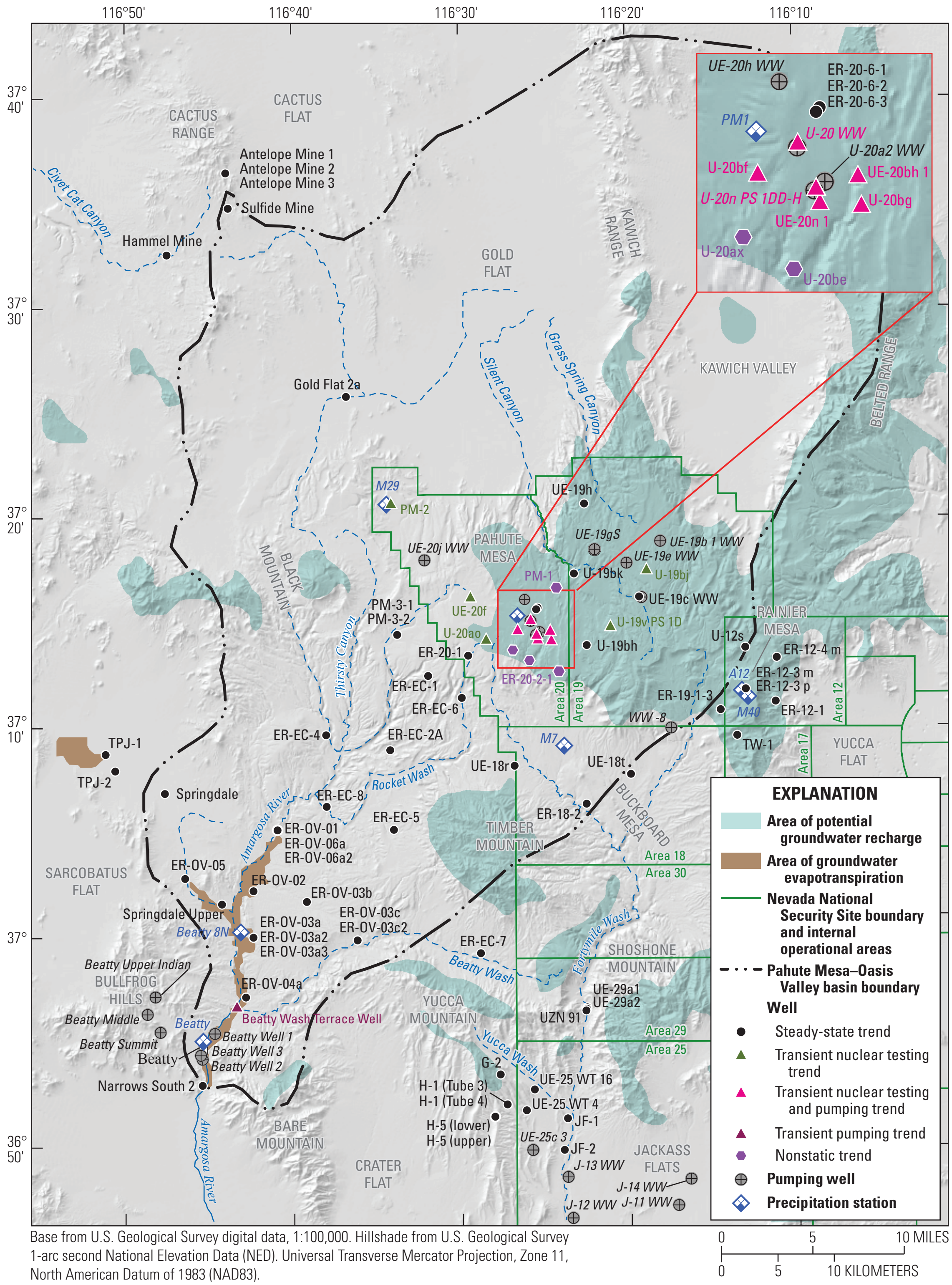
Graphical, statistical, and numerical methods were used to analyze trends. Graphical and statistical methods analyzed trends for variability and for upward or downward trends. Water-level models (WLMs) were used to differentiate hydrologic stresses affecting a trend and quantify the magnitude of the effect of each stress on the trend. Stresses evaluated include recharge, evapotranspiration, pumping, nuclear testing, and water-level equilibration following localized disturbances in the wellbore. Winter precipitation data greater than a defined threshold were used as a proxy for recharge and correlated to rising water levels. Groundwater-withdrawal data were used to determine if water levels in study area wells were affected by pumping. The WLMs and compiled water-level, precipitation, and groundwater-withdrawal data are published in Jackson (2018).

Data Compilation

Water-level, precipitation, groundwater-withdrawal, earthquake, and nuclear-testing data were compiled for use in the water-level trend analysis and development of the conceptual framework. The timing, magnitude, and location of earthquakes were compiled from the USGS Earthquake Hazards Program (http://earthquake.usgs.gov/). Characteristics of nuclear tests in the study area were compiled from U.S. Department of Energy (2015).

Water-Level Data

Water-level measurements were retrieved from the USGS National Water Information System (NWIS) database (http://waterdata.usgs.gov/nwis). These water levels were measured quarterly and are termed “periodic” measurements. Periodic water levels compiled for the trend analyses are current through March 2016. Individual water levels were flagged as representative of either steady-state, transient, or nonstatic conditions; a few levels were considered suspect or had insufficient supporting information to determine the general condition. A suspect water level indicates the water level is anomalous or in error and cannot be attributed to any known hydrologic cause. Steady-state water levels represent natural hydrologic conditions in the groundwater-flow system.



Base from U.S. Geological Survey digital data, 1:100,000. Hillshade from U.S. Geological Survey 1-arc second National Elevation Data (NED). Universal Transverse Mercator Projection, Zone 11, North American Datum of 1983 (NAD83).

Figure 3. Location of wells used in water-level trend analysis, pumping wells, precipitation monitoring stations, areas of potential groundwater recharge, groundwater evapotranspiration areas, and physiographic and hydrographic features, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada

Transient water levels represent non-steady hydrologic conditions that result from perturbations such as groundwater pumping and (or) nuclear testing. Nonstatic water levels do not represent hydrologic conditions in the formation open to the well. Instead, water levels in the well are equilibrating to formation water levels following localized disturbances in the wellbore, such as drilling, bailing, or pumping of the well. Nonstatic conditions typically occur when wells are open to a low transmissivity formation. Months to years may be necessary for the water level in the well to come into equilibration with the formation water level because the well-bore storage volume is large relative to the rate of water filling the wellbore. Flags of water-level data for all study area wells are provided in Jackson (2018).

Water-level trends in study area wells are categorized based on the dominant hydrologic condition (steady state, transient, or nonstatic) affecting water levels in a well. Well-site and construction information are provided for wells with nonstatic and transient trends (table 1), and steady-state trends (table 2). Short well names listed in the tables are used in place of USGS well names in the text, tables, and figures of this report for brevity. For clarity, well names are italicized in this report, whereas borehole names are not italicized.

Water-level measurements from two well completions in borehole ER-EC-6 were combined. From 2011 through 2015, water levels were measured in the shallow piezometer, *ER-EC-6 shallow*. Prior to this, from 2000 to 2011, water levels were measured in well *ER-EC-6*, which represents a composite hydraulic head of the shallow completion and deeper completions. For the trend analysis, water-level measurements in piezometer *ER-EC-6 shallow* were appended to the *ER-EC-6* water-level record.

Precipitation Data

Three precipitation indexes were created to represent long-term precipitation trends for three geographic areas: Beatty, Rainier Mesa, and Pahute Mesa. Three indexes were used to represent spatially variable precipitation (and recharge) patterns within the study area. Precipitation data from multiple monitoring stations were needed to construct long-term (more than 40 years) precipitation indexes. Precipitation stations that began operation in the 1960s and 1970s were deactivated and replaced with new stations in 2000 or 2011. The new precipitation stations are located between 0.5 and 10 mi from deactivated stations.

Location and altitude information of all precipitation stations used to construct long-term precipitation-index records are provided in table 3. Total monthly precipitation, which includes rain and snow, were compiled for these precipitation stations from the Western Regional Climate Center, Community Environmental Monitoring Program, and the National Oceanic and Atmospheric Administration (NOAA). For clarity, precipitation station names are italicized in this report.

Long-term precipitation-index records were constructed using data compiled from four or more monitoring stations. Data overlap between deactivated and active precipitation stations in Beatty allows for the use of statistical correlation methods between stations to construct a long-term precipitation-index record. However, the lack of data overlap between deactivated and active precipitation stations in Rainier Mesa and in Pahute Mesa preclude direct comparison of data and require proxy comparisons to nearby stations. Supplementary precipitation stations were used to fill data gaps and spatially correlate precipitation between deactivated and active precipitation stations. Precipitation data and analyses used to construct long-term precipitation records are provided in Jackson (2018).

The Beatty precipitation-index record was constructed using precipitation data from the *Beatty 8N* and *Beatty* monitoring stations (fig. 1). Deactivated *Beatty 8N* (1973–2008) and active *Beatty* (2000–2015) monitoring stations are 6.5 mi apart, differ in altitude less than 200 ft, and have 8 years of data overlap (2000–2008) (table 3). Spearman's rho correlation coefficient was used to measure the strength of association between precipitation data at *Beatty 8N* and *Beatty*. The precipitation stations are highly correlated with a Spearman's rho correlation coefficient of 0.90. Furthermore, average percentage differences in monthly precipitation between the *Beatty 8N* and *Beatty* monitoring stations is about 3 percent, which is near the measurement error of about 2 percent (Garcia and others, 2014). A strong Spearman's rho correlation coefficient (0.90) and small average percentage differences in monthly precipitation indicate the *Beatty* station precipitation data can be appended to precipitation data from the *Beatty 8N* station, with no adjustments to either record.

The Rainier Mesa precipitation-index record was constructed using precipitation data from the *A12* and *M40* monitoring stations (fig. 1). Stations *A12* (1959–2011) and *M40* (2011–2015) are about 0.5 mi apart and differ in altitude by 35 ft. The NOAA deactivated station *A12* in early September 2011 and set up station *M40* in late September 2011 (table 3). Statistical correlations cannot be used to measure the strength of association between *A12* and *M40* precipitation records because data do not overlap. No nearby stations have continuous records from 1959 to 2015; however, precipitation data from supplementary Mid Valley stations (*MV* and *M14*) can be combined to form a continuous record and used for statistical correlations between Rainier Mesa stations.

Table 1. Location and completion information for wells dominated by nonstatic or transient water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**USGS site identification No.:** Unique, 15-digit, U.S. Geological Survey (USGS) site identification number.
USGS well name: U.S. Geological Survey well name.
Short name: Reduced form of well name used in report text and figures for brevity. Well locations are shown in figure 3.
Latitude: Latitude, in decimal degrees; referenced to the North American Datum of 1983.
Longitude: Longitude, in decimal degrees; referenced to the North American Datum of 1983.
Land-surface altitude: Altitude of land surface at well site, referenced to feet (ft) above National Geodetic Vertical Datum of 1929 (NGVD 29).
Depth drilled: Total depth of well, in feet below land surface.
Open interval: Area of well that is open to aquifer and where, if saturated, groundwater may enter the well. Open interval typically consists of open borehole and (or) well screen, including gravel pack. Where multiple open intervals occur in a well, altitudes are to top of the uppermost interval and bottom of the lowermost interval. Altitude referenced to feet above NGVD 29.
Primary hydrogeologic unit: Primary water-bearing hydrogeologic unit within the open interval that likely is contributing water to the well: AA, alluvial aquifer; LFA, lava-flow aquifer; TCU, tuff confining unit; VOLC, volcanic undifferentiated; WTA, welded-tuff aquifer]

USGS site identification No.	USGS well name	Short name	Latitude	Longitude	Land-surface altitude	Depth drilled	Open interval		Primary hydrogeologic unit
							Top	Bottom	
Wells with water levels dominated by nonstatic trends									
371246116240101	ER-20- 2-1	ER-20-2-1	37.21277	-116.40120	6,705	2,524	2,293	2,524	TCU
371350116264701	U -20ax	U-20ax	37.23057	-116.44721	6,536	2,200	62	2,200	TCU
371332116254101	U -20be	U-20be	37.22558	-116.42902	6,492	2,220	53	2,220	VOLC
371649116242102	PM- 1 (7543-7858 ft)	PM-1	37.28021	-116.40678	6,558	7,858	7,543	7,858	WTA
Wells with transient water levels potentially affected by nuclear testing									
372042116340501	PM- 2	PM-2	37.34494	-116.56898	5,592	8,788	2,506	8,788	VOLC
371736116184701	U -19bj	U-19bj	37.29315	-116.31380	7,035	2,153	57	2,153	LFA
371453116205751	U -19v PS 1D	U-19v PS 1D	37.24791	-116.35000	6,842	4,113	3,875	3,885	TCU
371416116282201	U -20ao	U-20ao	37.23775	-116.47364	6,279	2,150	20	2,150	LFA
371617116291701	UE-20f (4456-13686 ft)	UE-20f	37.27135	-116.48882	6,116	13,686	4,456	13,686	VOLC
Wells with transient water levels potentially affected by nuclear testing and groundwater pumping									
371505116254501	U -20 WW (cased)	U-20 WW	37.25136	-116.43017	6,468	3,268	65	3,268	LFA
371444116263001	U -20bf	U-20bf	37.24538	-116.44252	6,522	2,250	48	2,250	WTA
371414116242901	U -20bg	U-20bg	37.23708	-116.40888	6,567	2,200	58	2,200	WTA
371425116252401	U -20n PS 1DD-H (3025 ft)	U-20n PS 1DD-H	37.24022	-116.42421	6,468	4,520	2,417	4,290	LFA
371442116243301	UE-20bh 1	UE-20bh 1	37.24491	-116.41006	6,637	2,810	1,941	2,810	LFA
371425116251902	UE-20n 1 (2834 ft)	UE-20n 1	37.24025	-116.42282	6,461	3,300	2,308	2,834	LFA
Wells with transient water levels affected by groundwater pumping									
365640116431501	Beatty Wash Terrace Well	Beatty Wash Terrace	36.94439	-116.72173	3,450	75	50	75	AA

Table 2. Location and completion information for wells dominated by steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**USGS site identification No.:** Unique, 15-digit, U.S. Geological Survey (USGS) site identification number.
USGS well name: U.S. Geological Survey well name.
Short name: Reduced form of well name used in report text and figures for brevity. Well locations are shown in figure 3.
Latitude: Latitude, in decimal degrees; referenced to the North American Datum of 1983.
Longitude: Longitude, in decimal degrees; referenced to the North American Datum of 1983.
Land-surface altitude: Altitude of land surface at well site, referenced to feet above National Geodetic Vertical Datum of 1929 (NGVD 29).
Depth drilled: Total depth of well, in feet below land surface. An “e” indicates estimated depth and “–” indicates unknown depth..
Open interval: Area of well that is open to aquifer and where, if saturated, groundwater may enter the well. Open interval typically consists of open borehole and (or) well screen, including gravel pack. Where multiple open intervals occur in a well, altitudes are to top of the uppermost interval and bottom of the lowermost interval. Altitude referenced to feet (ft) above NGVD 29.
Primary hydrogeologic unit(s): Primary water-bearing hydrogeologic unit(s) within the open interval that likely are contributing water to the well: AA, alluvial aquifer; CA, carbonate aquifer; GCU, granite confining unit; LFA, lava-flow aquifer; TCU, tuff confining unit; VOLC, volcanic undifferentiated; WTA, welded-tuff aquifer]

USGS site identification No.	USGS well name	Short name	Latitude	Longitude	Land-surface altitude	Depth drilled	Open interval		Primary hydrogeologic unit(s)
							Top	Bottom	
370648116473001	BLM Springdale	Springdale	37.11355	-116.79312	4,035	e117	–	–	WTA, LFA
371106116110401	ER-12-1 (1641–1846 ft)	ER-12-1	37.18486	-116.18509	5,817	3,588	1,641	1,846	CA
371142116125102	ER-12-3 main	ER-12-3 m	37.19497	-116.21499	7,391	4,908	2,447	4,908	CA
371142116125101	ER-12-3 piezometer	ER-12-3 p	37.19497	-116.21499	7,391	4,908	55	2,210	TCU
371311116105902	ER-12-4 main	ER-12-4 m	37.21958	-116.18402	6,884	3,715	2,501	3,715	CA
370615116222401	ER-18-2	ER-18-2	37.10393	-116.37376	5,437	2,500	1,351	2,500	WTA
371043116142103	ER-19-1-3 (shallow)	ER-19-1-3	37.17847	-116.24002	6,140	3,595	1,301	1,422	TCU
371321116292301	ER-20- 1	ER-20-1	37.22240	-116.49240	6,181	2,065	1,940	2,065	WTA
371537116251501	ER-20- 6-1 (3-in string)	ER-20-6-1	37.26015	-116.42166	6,475	3,200	2,437	2,947	LFA
371536116251601	ER-20- 6-2 (3-in string)	ER-20-6-2	37.25988	-116.42206	6,475	3,200	2,414	2,945	LFA
371533116251801	ER-20- 6-3 (3-in string)	ER-20-6-3	37.25914	-116.42243	6,466	3,200	2,436	2,807	LFA
371223116314701	ER-EC- 1	ER-EC-1	37.20626	-116.53063	6,026	5,000	2,258	4,895	LFA, WTA
370852116340502	ER-EC- 2A (1635–2236 ft)	ER-EC-2A	37.14494	-116.56827	4,902	4,974	1,635	2,236	TCU
370935116375302	ER-EC- 4 (952–2295 ft)	ER-EC-4	37.15881	-116.63195	4,760	3,487	952	2,296	LFA
370504116335201	ER-EC- 5	ER-EC-5	37.08450	-116.56545	5,077	2,500	1,169	2,500	WTA
371120116294802	ER-EC- 6 (1581–3820 ft)	ER-EC-6	37.18872	-116.49758	5,604	5,000	1,581	3,820	LFA, WTA
371120116294805	ER-EC- 6 shallow	ER-EC-6 shallow	37.18872	-116.49758	5,604	5,000	1,507	1,948	LFA, WTA
365910116284401	ER-EC- 7	ER-EC-7	36.98490	-116.47858	4,805	1,386	895	1,386	LFA
370610116375301	ER-EC- 8	ER-EC-8	37.10279	-116.63217	4,334	2,000	632	2,000	WTA
370504116404901	ER-OV-01	ER-OV-01	37.08439	-116.68117	4,007	180	142	180	WTA
370210116421501	ER-OV-02	ER-OV-02	37.03605	-116.70506	3,880	200	160	200	AA
365956116421601	ER-OV-03a	ER-OV-03a	36.99883	-116.70534	3,841	251	198	251	WTA
365956116421602	ER-OV-03a2	ER-OV-03a2	36.99883	-116.70534	3,841	821	560	655	TCU
365956116421603	ER-OV-03a3	ER-OV-03a3	36.99883	-116.70534	3,841	821	88	160	WTA
370139116390501	ER-OV-03b	ER-OV-03b	37.02744	-116.65228	4,233	400	272	400	AA
365948116360401	ER-OV-03c	ER-OV-03c	36.99661	-116.60200	4,188	542	496	542	WTA
365948116360402	ER-OV-03c2	ER-OV-03c2	36.99661	-116.60200	4,189	321	270	321	WTA
365705116424201	ER-OV-04a	ER-OV-04a	36.95133	-116.71256	3,488	151	89	151	AA
370246116461901	ER-OV-05	ER-OV-05	37.04605	-116.77284	3,935	200	136	200	AA

Table 2. Location and completion information for wells dominated by steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. —Continued

USGS site identification No.	USGS well name	Short name	Latitude	Longitude	Land-surface altitude	Depth drilled	Open interval		Primary hydrogeologic unit(s)
							Top	Bottom	
370504116404902	ER-OV-06a	ER-OV-06a	37.08439	-116.68117	4,008	536	488	536	WTA
370504116404903	ER-OV-06a2	ER-OV-06a2	37.08439	-116.68117	4,007	71	44	65	LFA
372543116363502	Gold Flat 2a	Gold Flat 2a	37.42849	-116.61061	5,230	380	250	360	VOLC
373228116472001	Hammel Mine Well	Hammel Mine	37.54105	-116.78978	5,540	e240	–	–	VOLC
365253116450801	Narrows South Well 2	Narrows South 2	36.88134	-116.75311	3,180	120	0	120	AA
370753116502701	NDOT TPJ-2	TPJ-2	37.13140	-116.84174	4,005	e123	–	–	AA
371421116333703	PM- 3-1 (1919–2144 ft)	PM-3-1	37.23902	-116.56107	5,823	3,019	1,872	2,192	WTA
371421116333704	PM- 3-2 (1442–1667 ft)	PM-3-2	37.23902	-116.56107	5,823	3,019	1,379	1,687	TCU
370131116440801	Springdale Upper Well	Springdale Upper	37.02522	-116.73645	3,775	e91	–	–	AA
373622116434601	TTR Antelope Mine 1	Antelope Mine 1	37.60551	-116.73017	6,350	–	–	–	VOLC
373622116434701	TTR Antelope Mine 2	Antelope Mine 2	37.60608	-116.73039	6,356	–	–	–	VOLC
373623116434701	TTR Antelope Mine 3	Antelope Mine 3	37.60617	-116.73050	6,362	–	–	–	VOLC
373446116433301	TTR Sulfide Mine	Sulfide Mine	37.57932	-116.72686	6,130	120	–	–	VOLC
370929116132311	TW- 1 (1615–4206 ft)	TW-1	37.15813	-116.22380	6,156	4,206	1,615	4,206	WTA, CA
371342116125102	U -12s (1480 ft)	U-12s	37.22829	-116.21669	6,794	1,596	12	1,480	GCU
371349116222001	U -19bh	U-19bh	37.23014	-116.37318	6,768	2,148	72	2,148	TCU
371714116230301	U -19bk	U-19bk	37.28729	-116.38509	6,670	2,198	57	2,198	LFA
370806116264001	UE-18r	UE-18r	37.13470	-116.44560	5,538	5,004	1,629	5,004	WTA
370741116194501	UE-18t	UE-18t	37.12809	-116.33003	5,201	2,600	120	2,600	WTA
371608116191002	UE-19c WW	UE-19c WW	37.26872	-116.32036	7,033	8,489	2,421	8,489	LFA
372034116222504	UE-19h (recompleted)	UE-19h	37.34275	-116.37445	6,780	3,705	2,050	2,287	LFA
365140116260301	UE-25 WT 4	UE-25 WT 4	36.86113	-116.43497	3,836	1,580	50	1,580	LFA
364945116235001	UE-25 WT 13 (JF-2)	JF-2	36.82865	-116.39845	3,388	1,160	222	1,160	WTA
365116116233801	UE-25 WT 15 (JF-1)	JF-1	36.85447	-116.39477	3,554	1,360	127	1,360	WTA
365239116253401	UE-25 WT 16	UE-25 WT 16	36.87735	-116.42687	3,971	1,710	108	1,710	LFA
365624116222901	UE-29 UZN 91	UZN 91	36.94001	-116.37552	3,949	94	89	94	WTA
365629116222601	UE-29a 1 HTH	UE-29a1	36.94126	-116.37472	3,984	215	35	215	LFA
365629116222602	UE-29a 2 HTH	UE-29a2	36.94130	-116.37482	3,985	1,383	285	1,383	LFA
370840116510101	USBLM TPJ-1	TPJ-1	37.14485	-116.85122	3,991	107	–	–	AA
365322116273501	USW G- 2	G-2	36.88951	-116.46065	5,097	6,006	795	2,598	TCU
365157116271204	USW H- 1 HTH (Tube 3)	H-1 (Tube 3)	36.86595	-116.45428	4,274	6,000	2,349	2,510	WTA
365157116271205	USW H- 1 HTH (Tube 4)	H-1 (Tube 4)	36.86595	-116.45428	4,274	6,000	335	2,208	WTA
365122116275503	USW H- 5 HTH (lower)	H-5 (lower)	36.85603	-116.46619	4,852	4,000	3,580	4,000	LFA
365122116275502	USW H- 5 HTH (upper)	H-5 (upper)	36.85603	-116.46619	4,852	4,000	311	3,580	WTA

Table 3. Location and altitude information for precipitation monitoring stations used to construct long-term precipitation-index records, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**Precipitation station:** Precipitation monitoring station name. Precipitation station locations are shown in figure 1.
Index: Precipitation index representing long-term precipitation in a geographic area.
Latitude: Latitude, in decimal degrees; referenced to the North American Datum of 1983.
Longitude: Longitude, in decimal degrees; referenced to the North American Datum of 1983.
Land-surface altitude: Altitude of land surface at precipitation monitoring site, referenced to feet above National Geodetic Vertical Datum of 1929.
Period of record: Beginning and end date, in month and year, for the period of data collection at precipitation monitoring station.
Reporting agency: WRCC, Western Regional Climate Center; DRI/DOE, Desert Research Institute/U.S. Department of Energy National Nuclear Security Administration Nevada Field Office (NNSA/NFO); ARL/SORD, National Oceanic and Atmospheric Administration (NOAA) Air Resources Laboratory (ARL)/Special Operations and Research Division (SORD)]

Precipitation station	Index	Latitude	Longitude	Land-surface altitude	Period of record	Reporting agency
Beatty 8N	Beatty	36.99500	-116.71889	3,550	January 1973 to July 2008	WRCC[1]
Beatty	Beatty	36.91667	-116.75000	3,311	January 2000 to December 2015	DRI/DOE[2]
A12	Rainier Mesa	37.19111	-116.21528	7,490	March 1959 to September 2011	ARL/SORD[3]
M40	Rainier Mesa	37.18517	-116.20689	7,525	September 2011 November 2015	ARL/SORD[3]
PM1	Pahute Mesa	37.24889	-116.43750	6,550	January 1964 to August 2011	ARL/SORD[3]
M7	Pahute Mesa	37.15136	-116.39561	5,451	September 2011 January 2016	ARL/SORD[3]
Supplementary precipitation stations						
4JA		36.78472	-116.28889	3,422	January 1959 to July 2011	ARL/SORD[3]
MV		36.97250	-116.17194	4,660	September 1964 to December 2011	ARL/SORD[3]
M14		36.96756	-116.18136	4,716	October 2011 to January 2016	ARL/SORD[3]
M29		37.34486	-116.56892	5,585	October 2011 to January 2016	ARL/SORD[3]
TS2		37.05306	-116.19139	4,980	January 1960 to August 2011	ARL/SORD[3]

[1]Beatty 8N monitoring station data were retrieved from the WRCC (site identifier 260718, http://www.wrcc.dri.edu).

[2]Beatty monitoring station data were retrieved from the Community Environmental Monitoring Program (CEMP), which is maintained by NNSA/NFO and DRI (site identifier Beatty, Nevada; http://www.cemp.dri.edu).

[3]Monitoring station data were retrieved from NOAA ARL/SORD (Soulé, 2006; http://www.sord.nv.doe.gov/)

Precipitation data from *MV* and *M14* can be combined with no adjustments to either record because these stations are about 0.6 mi apart, differ in altitude less than 56 ft, and have strong correlation coefficients. Data from stations *A12* and *M40* were correlated to nearby stations *MV* (1964–2011) and *M14* (2011–2016) (table 3), about 15 mi from the Rainier Mesa precipitation stations. Strong Spearman's rho correlation coefficients of 0.81 and 0.88 were computed between *A12* and *MV* and between *M40* and *M14*, respectively. The *A12* and *M40* precipitation records were combined because these stations are in close proximity, occur at similar altitudes, and have strong consistent Spearman's rho correlation coefficients with nearby stations.

Data gaps in Beatty (*Beatty 8N*) and Rainier Mesa (*A12*) precipitation stations were estimated using regression analysis. *Beatty 8N* is missing precipitation data in 1984 and 1999, whereas *A12* is missing precipitation data from 1995 to 1997. Monthly precipitation in these data gaps was estimated by regressing precipitation data from stations *Beatty 8N* and *A12* with precipitation data from all other nearby stations (table 3) to find two precipitation stations with the best correlation. Jackass Flats (*4JA*) and Pahute Mesa (*PM1*) stations best correlate with *Beatty 8N*, whereas Pahute Mesa (*PM1*) and Tippipah Springs (*TS2*) stations best correlate with *A12*. Within the data gaps at stations *Beatty 8N* and *A12*, monthly precipitation was estimated using the method of Dunne and Leopold (1978) (see Jackson [2018] for details).

The Pahute Mesa precipitation-index record was constructed using precipitation data from the *PM1* and *M7* monitoring stations (fig. 1). Pahute Mesa station *PM1* (1964 to August 2011) is about 7 mi from station *M7* (September 2011 to 2016). Large distances between precipitation stations preclude directly combining the record from *PM1* with *M7*. To combine precipitation records from these stations, a correction factor was applied to account for spatial variability in precipitation. Precipitation data from *PM1* and *M7* were regressed to precipitation data from the other two precipitation indexes (table 3). Precipitation from the Rainier Mesa index best correlates with Pahute Mesa stations with correlation coefficients of 0.53 and 0.89 for *PM1* and *M7*, respectively. The ratio of correlation coefficients was used as a correction factor to scale and append precipitation data from station *M7* to the *PM1* precipitation record (see Jackson [2018] for details).

Groundwater Withdrawal Data

Groundwater withdrawal data were compiled for water-supply wells in the study area through 2015 (table 4). Water-supply wells are in Pahute Mesa, Jackass Flats, Bullfrog Hills, and Beatty, Nevada. Groundwater-withdrawal data used for NNSS operational activities in Pahute Mesa and Jackass Flats were compiled from Elliott and Moreo (2011) and U.S. Geological Survey (2017). Municipal withdrawals from wells in the town of Beatty and nearby Bullfrog Hills were compiled from data reported by the Beatty Water and Sanitation District (BWSD), published in Jackson (2018). Domestic groundwater withdrawals near Beatty were computed using a domestic use estimate of 0.5 acre-ft per year (acre-ft/yr) for individual households (Geter, 2015). Location and construction information for all water-supply wells are provided in table 4.

Episodic Recharge

Water levels fluctuate naturally in response to episodic recharge events. Episodic recharge is precipitation-derived groundwater recharge during brief intermittent periods and can be observed as distinct water-level rises in well hydrographs (see fig. 4 as an example). In the study area, most episodic recharge is derived during the winter months (October–March) from greater-than-average precipitation when evapotranspiration approaches zero (Winograd and others, 1998; Hershey and others, 2008).

Precipitation thresholds can be used as a proxy for episodic recharge events. French and others (1996) successfully used threshold precipitation during winter months as a proxy for potential recharge events at the NNSS. French and others (1996) defined a threshold precipitation event as precipitation that infiltrates at least 3 ft into the subsurface and has the potential to become recharge. In this study, winter precipitation greater than a defined threshold is assumed to fully replenish root-zone water storage and recharge the groundwater system (Smith and others, 2017).

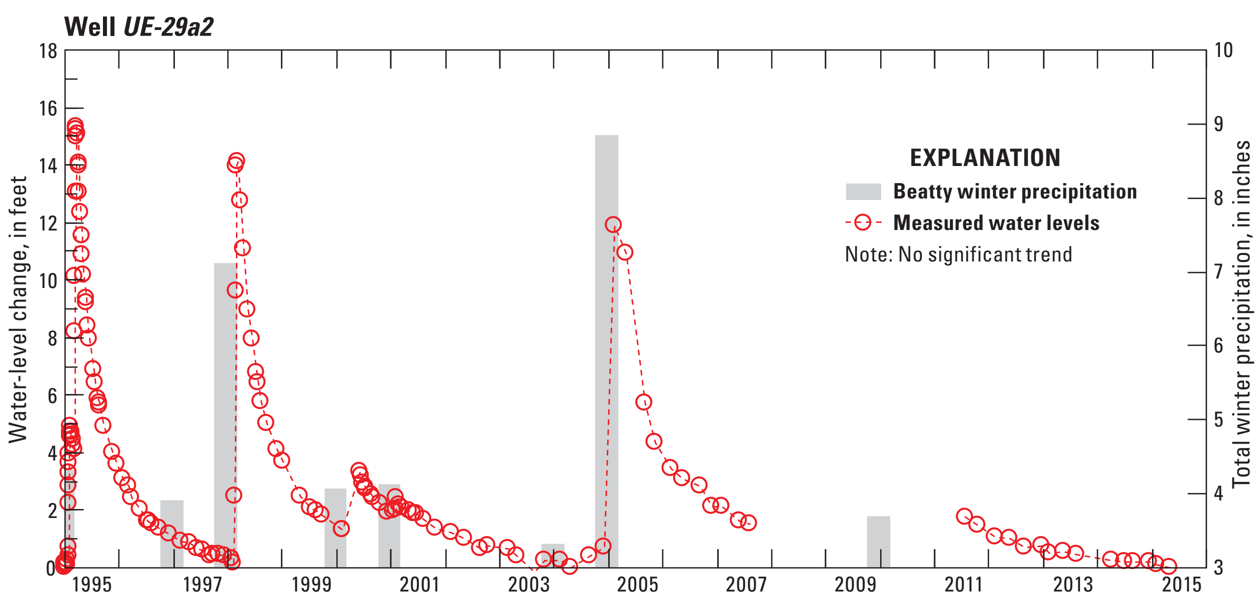


Figure 4. Comparison of water levels in well *UE-29a2* to total winter precipitation from the Beatty precipitation index, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, October–March 1995–2015.

Table 4. Location and completion information for water-supply wells in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**USGS site identification No.:** Unique, 15-digit, U.S. Geological Survey (USGS) site identification number.
USGS well name: U.S. Geological Survey well name.
Short name: Reduced form of well name used in report text and figures for brevity. Well locations are shown in figure 3.
Latitude: Latitude, in decimal degrees; referenced to the North American Datum of 1983.
Longitude: Longitude, in decimal degrees; referenced to the North American Datum of 1983.
Land-surface altitude: Altitude of land surface at well site, referenced to feet (ft) above National Geodetic Vertical Datum of 1929 (NGVD 29).
Depth drilled: Total depth of well, in feet below land surface.
Open interval: Area of well that is open to aquifer and where, if saturated, groundwater may enter the well. Open interval typically consists of open borehole and (or) well screen, including gravel pack. Where multiple open intervals occur in a well, altitudes are to top of the uppermost interval and bottom of the lowermost interval. Altitude referenced to feet above NGVD 29.
Total water pumped (1975–2015): Total groundwater withdrawals from water-supply wells between 1975 and 2015, in acre-feet]

USGS site identification No.	USGS well name	Short name	Latitude	Longitude	Land-surface altitude	Depth drilled	Open interval		Total water pumped (1975–2015)
							Top	Bottom	
365619116483901	Beatty Middle Well	Beatty Middle	36.93967	-116.81339	4,110	700	100	700	38
365527116475301	Beatty Summit Well	Beatty Summit	36.92439	-116.79978	3,882	700	65	700	893
365709116481101	Beatty Upper Indian Well	Beatty Upper Indian	36.95161	-116.80423	4,240	693	55	693	1,130
365524116444001	Beatty Well No. 1	Beatty Well 1	36.91550	-116.76089	3,365	200	50	200	498
365409116452301	Beatty Well No. 2	Beatty Well 2	36.90633	-116.75867	3,300	195	70	195	206
365420116453001	Beatty Well No. 3	Beatty Well 3	36.90495	-116.75784	3,290	300	70	300	281
364706116170601	J -11 WW	J-11 WW	36.78509	-116.28588	3,443	1,327	16	1,327	113
364554116232401	J -12 WW	J-12 WW	36.76489	-116.39097	3,128	1,139	16	1,139	1,979
364554116232400	J -12 WW (885 ft)	J-12 WW (885 ft)	36.76489	-116.39097	3,128	885	16	885	358
364828116234001	J -13 WW	J-13 WW	36.80803	-116.39546	3,318	3,488	435	3,488	4,059
364821116162201	J -14 WW	J-14 WW	36.80591	-116.27366	3,610	1,775	120	1,775	11.5
371505116254501	U -20 WW (cased)	U-20 WW	37.25136	-116.43017	6,468	3,268	65	3,268	2,382
371434116251601	U -20a 2 WW	U-20a2 WW	37.24259	-116.42213	6,472	4,500	860	4,500	398
371425116252401	U -20n PS 1DD-H (3025 ft)	U-20n PS 1DD-H	37.24022	-116.42421	6,468	4,520	2,417	4,290	17.8
371433116251301	U -20n PS 1DD-H (4309 ft)	U-20n PS 1DD-H	37.24022	-116.42421	6,468	4,520	2,417	4,290	10.7
371425116252403	U -20n PS 1DD-H (recompl)	U-20n PS 1DD-H	37.24022	-116.42421	6,468	4,520	2,417	4,290	0.3
371852116175701	UE-19b 1 WW	UE-19b 1 WW	37.31453	-116.30017	6,802	4,500	2,190	4,500	21.0
371608116191002	UE-19c WW	UE-19c WW	37.26872	-116.32036	7,033	8,489	2,421	8,489	3,220
371750116195901	UE-19e WW	UE-19e WW	37.29705	-116.33407	6,919	6,005	2,475	6,005	189
371830116215300	UE-19gS (2650–4508 ft)	UE-19gS	37.30822	-116.36551	6,719	4,508	2,650	4,508	0.8
371830116215303	UE-19gS (2650–7500 ft)	UE-19gS	37.30822	-116.36551	6,719	7,500	2,650	7,500	408
371618116260201	UE-20h WW	UE-20h WW	37.27178	-116.43473	6,557	7,207	2,518	7,207	67
371801116320301	UE-20j WW	UE-20j WW	37.30023	-116.53507	5,903	5,690	1,740	5,690	179
364947116254501	UE-25c 3	UE-25c 3	36.82925	-116.42989	3,714	3,000	1,368	3,000	68
364947116254505	UE-25c 3 (2285–2667 ft)	UE-25c 3	36.82925	-116.42989	3,714	3,000	2,285	2,667	357
364947116254504	UE-25c 3 (2286–2879 ft)	UE-25c 3	36.82925	-116.42989	3,714	3,000	2,286	2,879	27.7
370956116172101	WW- 8 (30–2031 ft)	WW-8	37.16554	-116.29003	5,695	5,490	30	2,031	5,723
370956116172133	WW- 8 (2031–5490 ft)	WW-8	37.16554	-116.29003	5,695	5,490	2,031	5,490	0.5

Selecting Precipitation Indexes

Water-level fluctuations in well hydrographs were compared to greater-than-average winter precipitation data to select the most appropriate precipitation-index record for each well. A precipitation index was selected based on the temporal pattern, not magnitude, of greater-than-average winter precipitation. That is, greater-than-average winter precipitation is a proxy for recharge, indicating the timing, not magnitude, of precipitation contributing to recharge. The selected precipitation-index record for a hydrograph best characterizes observed water-level responses to recharge.

Selecting Precipitation Thresholds

For each precipitation-index record, precipitation was summed during the winter months (October 1–March 1) and a threshold amount of winter precipitation was specified. Thresholds were used to determine years where total winter precipitation likely was sufficient to generate a recharge response. Observed water-level responses to recharge were used to select the thresholds used in the analysis. Total winter precipitation amounts for the Rainier Mesa, Pahute Mesa, and Beatty precipitation indexes are shown in figure 5, where the long-term average is the average of total winter precipitation from 1973–2015.

Precipitation thresholds were computed as percent-of-average winter precipitation (fig. 5), which was computed as the ratio of the threshold winter precipitation to the long-term average winter precipitation, expressed as a percentage. For example, long-term average winter precipitation is 6.2 inches (in.) for the Rainier Mesa precipitation index. A threshold amount of 12.4 in. is two-times the long-term winter average, where 12.4 in. (threshold)/6.2 in. (long-term average) × 100 percent = 200 percent.

For each well, a precipitation index and precipitation threshold were selected based on water-level responses to recharge. Selected precipitation thresholds ranged between 115 and 160 percent of the long-term average. A low and high threshold were selected for each precipitation index. The low threshold was used for wells with shallow (less than 350 ft) unsaturated zone depths that are in close proximity to recharge areas or ephemeral channels. The high threshold was used for wells where the unsaturated zone is thick or the well is distant from recharge areas. Increasing the threshold reduces the number of winters that are classified as “wet” in the precipitation record. One threshold could not be specified for all wells because wells open to aquifers with shallow water tables typically respond to most years with greater-than-average winter precipitation, whereas wells open to aquifers with deep water tables typically only respond to the wettest winters. Low and high thresholds were used for the Rainier Mesa (115 and 130 percent) and Beatty (125 and 160 percent) precipitation indexes, whereas only a high threshold (120 percent) was needed for the Pahute Mesa precipitation index.

Analysis Methods

Graphical and Statistical Methods

Graphical and statistical methods were used to assess whether steady-state water levels have an upward, downward, or no significant trend over the period of analysis from 1995 to 2016. The period 1995–2016 was selected to provide a consistent record for analyzing steady-state water-level data. Water-level data from 62 wells (table 2) were analyzed graphically and statistically for trends. Graphical analysis included smoothing water-level data using Locally Weighted Scatterplot Smoothing (LOWESS) and statistical analysis included use of the Mann-Kendall trend test and Kendall’s tau correlation coefficient.

The Mann-Kendall trend test and Kendall’s tau correlation coefficient were used to assess water-level trends statistically. The Mann-Kendall trend test (Mann, 1945; Kendall, 1975) was used to assess the presence of a monotonic upward or downward water-level trend at a well statistically. The Mann-Kendall trend test is a non-parametric method, which means that there is no requirement for water-level data to be either normally distributed or have a linear trend. Kendall’s tau correlation coefficient was computed to measure the strength of the monotonic trend in steady-state water levels. Kendall’s tau values range between -1 and 1. A Kendall’s tau value equal to 0 indicates no monotonic trend, a 1 indicates a strong rising trend, and a -1 indicates a strong declining trend.

Locally Weighted Scatterplot Smoothing was used to smooth steady-state water-level data to graphically detect a water-level trend. LOWESS is a non-parametric regression method that is especially helpful to visually detect water-level trends in data with relatively large data scatter. LOWESS curves were fit to water-level data because many water-level fluctuations have sinusoidal patterns that cannot be captured accurately with linear and other monotonic trend lines. LOWESS curves also were used to quantify the magnitude of water-level change with time at each well. The magnitude of water-level change at each well was computed from the difference between the maximum and minimum water-level value on the LOWESS curve. The magnitude of water-level change was not computed using the first and last water-level measurement from the analysis period because many of the trends are neither linear nor monotonic. The magnitude of water-level change was used to quantitatively compare water-level trends at different wells.

Graphical methods were used in conjunction with statistical correlations to circumvent statistically significant water-level trends that are not meaningful. This can be explained with the following example. Well *UE-29a2* has a Kendall’s tau of -0.3 and Mann-Kendall trend test results indicate a statistically significant downward trend at the 99-percent confidence level. Graphical analysis shows that well *UE-29a2* does not have a long-term declining

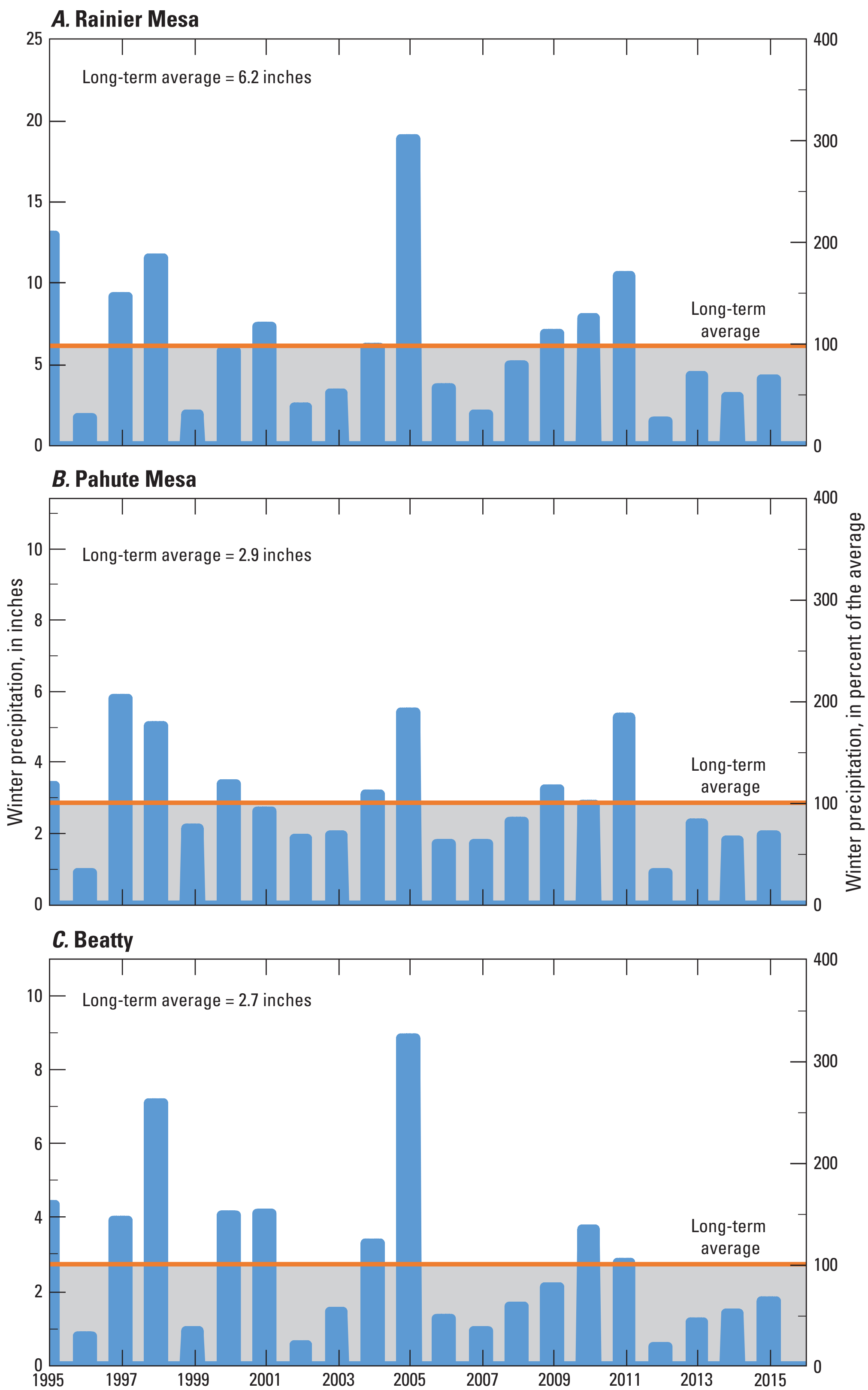


Figure 5. Total winter precipitation and percent-of-average winter precipitation for precipitation indexes at selected sites, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, October–March 1995–2015.

trend. Instead, long-term water levels have no significant trend (neither upward nor downward), as observed by a net water-level change of zero between 1995 and 2016 (fig. 4). Water levels have short-term water-level rises in response to focused recharge along Fortymile Wash (figs. 3 and 4). To preclude statistically significant water-level trends that are not meaningful, water-level trends are considered significant if the following criteria are true: (1) the Mann-Kendall trend test identifies a monotonic upward or downward trend within a confidence level of 99 percent (p-value less than 0.01); (2) Kendall's tau is greater than 0.26; and (3) the maximum change in water level on the LOWESS curve is greater than or equal to 0.2 ft, where the maximum change occurs over a period of more than 7 years.

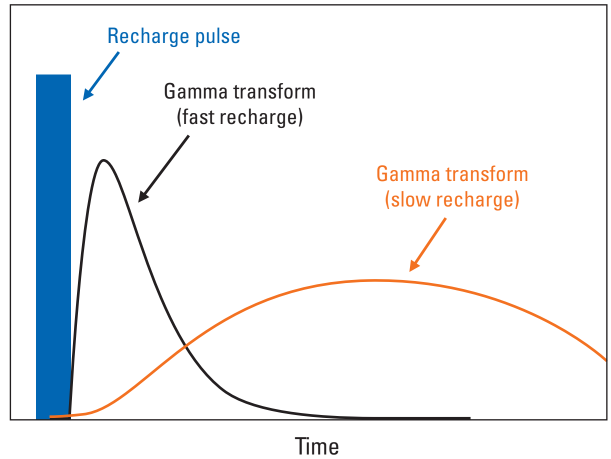


Figure 6. Gamma transforms of water-level responses to precipitation-derived recharge.

Water-Level Modeling

Steady-state, transient, and nonstatic water-level trends were modeled analytically to identify potential hydrologic stresses causing the trends. Potential stresses include recharge, evapotranspiration, pumping, and wellbore equilibration. WLMs were used to differentiate stresses affecting water-level trends and quantify the effect of each stress on the trends. WLMs were generated using SeriesSEE, a Microsoft Excel® add-in (Halford and others, 2012). All WLM analyses discussed in this report are published in Jackson (2018).

A WLM is an analytical model that fits a synthetic curve to measured water levels. The synthetic curve is the sum of one or more time-series components that likely explain the water-level fluctuations in the trend. Time-series components include (1) recharge response simulated with Gamma transforms of winter precipitation above a threshold, (2) evapotranspiration response simulated with Fourier transforms of water levels that were affected only by evapotranspiration in a background well, (3) pumping drawdown simulated with Theis transforms of pumping schedules from water-supply wells, and (4) wellbore equilibration simulated with Bouwer and Rice transforms.

Recharge Response

The Gamma transform (Halford and others, 2012) was used to simulate a water-level response to recharge using precipitation above a threshold during wet winters as a proxy for recharge. The transform accounts for the behavior of recharge with respect to unsaturated zone thickness (O'Reilly, 2004). As the unsaturated zone increases, the timing of recharge is lagged and the magnitude of recharge is attenuated. Recharge was transformed into a time series using the Gamma probability distribution function (fig. 6). The amplitude, scale, and shape of the Gamma transform were adjusted to match the synthetic curve to measured water levels (Halford and others, 2012).

If one Gamma transform poorly matched synthetic and measured water levels and the unsaturated zone is thick (greater than 1,000 ft), then two Gamma transforms were used to simulate recharge. Conceptually, two transforms represent fast and slow recharge pathways in a dual-porosity system. Hydraulically connected fracture networks provide fast recharge pathways and the rock matrix or disconnected fracture networks cause slow diffuse recharge in a thick unsaturated zone.

The purpose of the Gamma transform in a WLM is to show if water-level responses in a well can be explained by episodic recharge. When using one or two Gamma transforms, the fitting parameters are non-unique, but are constrained by the timing and magnitude of recharge pulses. Only hydrographs with water-level rises and declines consistent with the timing and magnitude of recharge events can be fit with the Gamma transform. A lack of fit between synthetic and measured water levels suggests that water-level fluctuations in a well cannot be explained by recharge.

Evapotranspiration Response

Groundwater evapotranspiration was simulated using continuous (hourly) water-level data from a background well (*Spring Meadows Rd Well*) affected only by evapotranspiration. *Spring Meadows Rd Well* is located in a discharge area within the Ash Meadows groundwater basin (fig. 1), but is considered a suitable surrogate for the response to groundwater evapotranspiration in Oasis Valley. Continuous water-level data exist for the well from 1996 to 1997 (USGS site 362536116211801; https://nwis.waterdata.usgs.gov/nwis/).

Monthly average water levels were computed from the data and duplicated for each year from 1997 to 2016 to generate a long-term evapotranspiration record (fig. 7). The long-term evapotranspiration time series was used as a water-level component of the synthetic curve. The amplitude and phase of the long-term time series were adjusted to match the synthetic curve to measured water levels.

Pumping Drawdown

The Theis (1935) analytical solution solves for water-level change, or drawdown, at a specified time and pumping rate. To solve for water-level changes based on pumping rates that vary with time, multiple Theis (1935) solutions are superimposed within a WLM. The superposition of Theis (1935) solutions is termed a Theis transform (Halford and others, 2012). A Theis transform was used to transform monthly or yearly groundwater withdrawals from pumping wells to water-level responses. Short-term (hourly or daily) pumping schedules were not needed because water levels were measured every 1 to 3 months and short-term changes in pumping are attenuated between the pumping and observation wells by the aquifer system (Garcia and others, 2011). Transforming pumping schedules with superimposed Theis (1935) solutions works exceptionally well regardless of aquifer medium or hydrogeologic complexity (Garcia and others, 2013).

The Theis (1935) solution has three parameters: radial distance, transmissivity, and storativity. The radial distance between the pumping and observation well is known. Transmissivity and storativity parameters are adjusted to match the synthetic curve to measured water-level changes. The WLM has been shown to work well even when the Theis (1935) solution's simplifying assumptions—that is, radial flow in an aquifer with infinite extent, uniform thickness, and isotropic, homogenous hydraulic properties—are completely violated (Garcia and others, 2013). Transmissivity and storativity are used solely as fitting parameters in the WLM and values should not be reported as meaningful (Halford and others, 2012).

Theis transforms were used to determine if measured water levels were affected by pumping. WLMs that included or excluded Theis transforms were compared to determine if water levels likely were affected by pumping. If the inclusion of a Theis transform improved the fit between measured and synthetic water levels, then water levels likely were affected by pumping. An improvement between measured and synthetic water levels was defined as a reduction in root-mean-square error of 0.10 ft or more. This definition is arbitrary and based on an observed goodness of fit. Theis transforms also were used to demonstrate that some water-level responses were not the result of pumping.

Wellbore Equilibration

Wellbore equilibration is observed as either a steep exponential rise or decline in water levels that results from water in the wellbore equilibrating with water in the formation open to the well. Wellbore equilibration is observed in wells open to low-permeability units, where the low permeability is evidenced by water levels either slowly rising or declining in response to drilling, bailing, or other localized disturbances to the wellbore. Following the localized disturbance, water-level recovery to reach equilibrium conditions (between the well and formation) typically takes months to years (Halford and others, 2005).

The Bouwer and Rice (1976) analytical solution was used to simulate wellbore equilibration following localized disturbances to the wellbore. By rearranging the Bouwer and Rice solution, water-level recovery is simulated as:

$$BR(t) = \Delta WL \cdot 10^{-Kt}, \tag{1}$$

where

ΔWL is the difference between the initial water-level measurement and the equilibrium water level;

K is the hydraulic conductivity;

t is time; and

$BR(t)$ is the Bouwer and Rice transform.

Formation hydraulic conductivity (K) was estimated by analyzing water-level recovery following the localized disturbance as a single-well slug test (Bouwer and Rice, 1976). This method assumes that the disturbance to the wellbore is instantaneous compared to the long recovery period (Halford and others, 2005). The Bouwer and Rice transform is a time-series component of the synthetic curve, where the transform is the superposition of multiple Bouwer and Rice (1976) solutions with time.

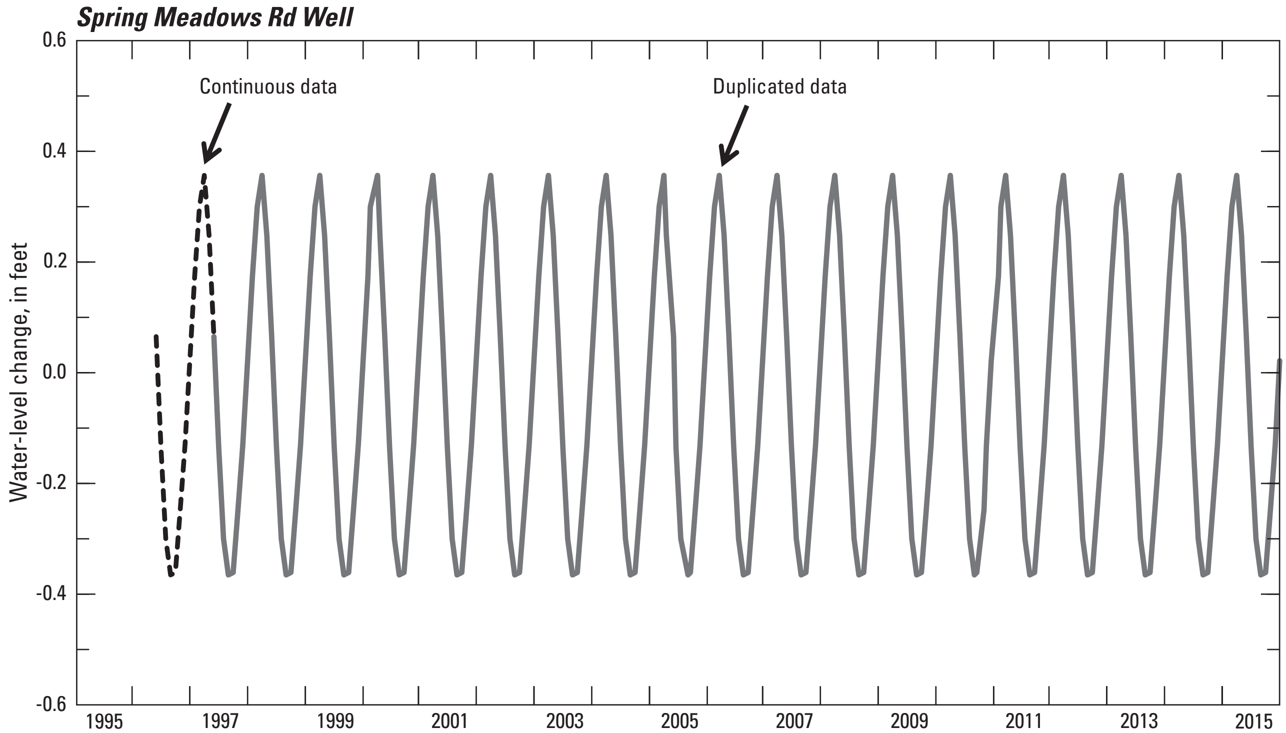


Figure 7. Continuous water-level record at *Spring Meadows Rd Well*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Continuous water-level data include 1996–97 and duplicated data include 1997–2016.

Conceptual Framework of Water-Level Responses to Hydrologic Stresses

A conceptual framework is presented to explain water-level trends. A water-level trend reflects the summation of all hydrologic stresses acting on the aquifer or in the wellbore at the well location. The conceptual framework consists of multiple stress-specific conceptual models to explain the water-level response to a hydrologic stress.

Conceptual models were developed to explain hydrologic stresses causing nonstatic, transient, and steady-state water-level trends. Nonstatic trends in wells are dominated by wellbore equilibration. Transient trends in wells are dominated by anthropogenic stresses, such as groundwater pumping or nuclear detonations. Steady-state trends in wells are dominated by natural stresses. Natural hydrologic stresses include recharge, evapotranspiration, barometric pressure, earth tides, and earthquakes. Water-level responses to each hydrologic stress are summarized in figure 8.

Nonstatic Trends

Nonstatic water levels are defined in this study as water levels that do not represent hydrologic conditions in the surrounding aquifer system. Instead, nonstatic levels occur when water in the borehole is equilibrating to groundwater in the formation open to the well. This is called wellbore equilibration. Nonstatic levels typically are observed in wells open to low-permeability units, where water levels inside the well are equilibrating following localized disturbances in the wellbore. Localized disturbances include drilling operations, well development and testing, slug injection, dewatering the borehole, or bailing the well for sampling. Nonstatic water levels also are observed when large well losses occur in a well during pumping.

Recognizing nonstatic water levels is important to avoid misinterpreting a water-level trend. Nonstatic water levels represent a localized hydrologic condition in the wellbore. Thus, nonstatic water levels are not representative of hydrologic stresses affecting the groundwater system and will have no measurable effect on water levels away from the affected well.

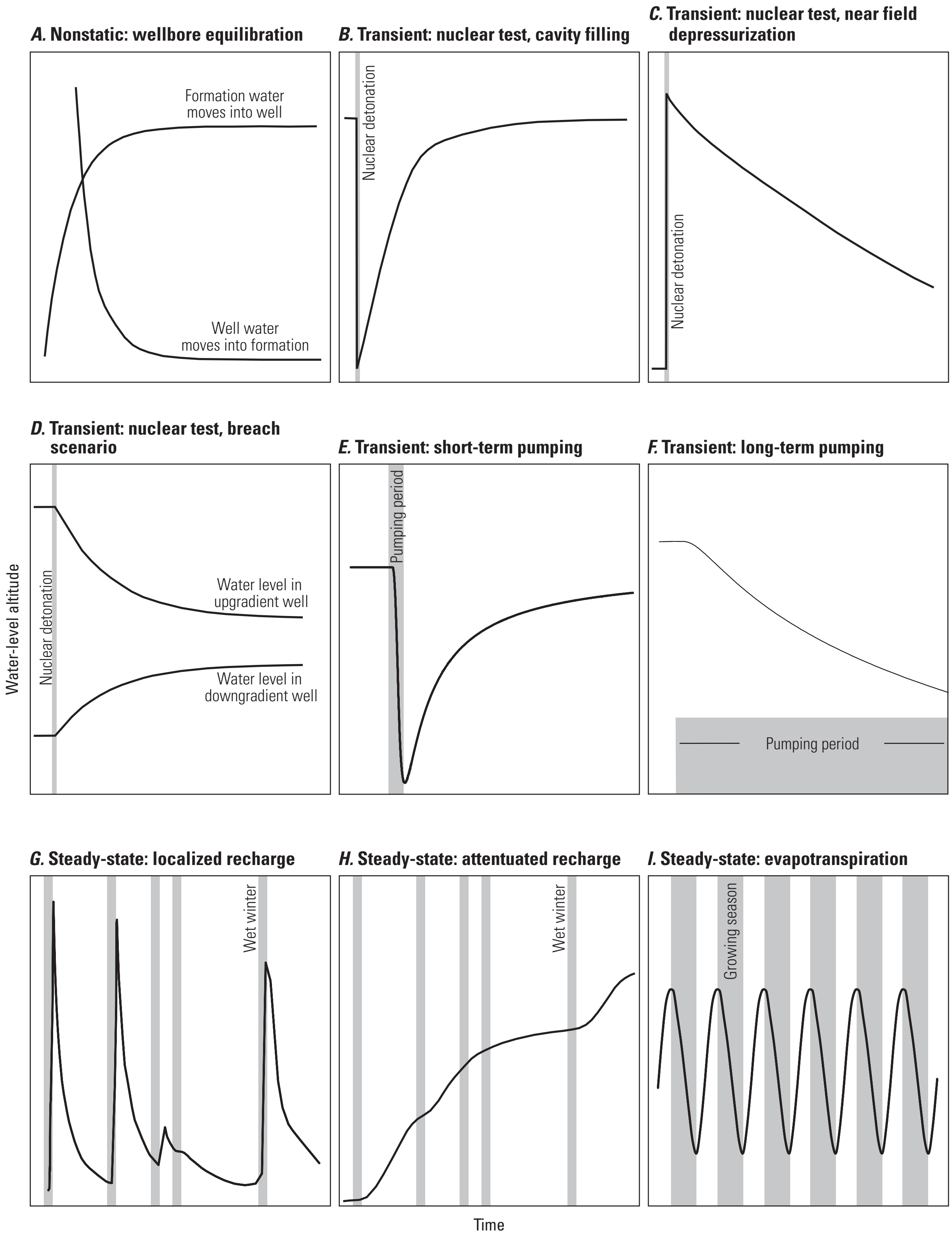


Figure 8. Conceptualized water-level responses to nonstatic, transient, and steady-state stresses.

Factors Affecting Trends under Non-Pumping and Pumping Conditions

Properties of the formation and the well cause nonstatic water levels to occur under non-pumping and pumping conditions. Under equilibrating conditions (post-pumping or non-pumping), the rate of water-level recovery in a well is controlled by formation transmissivity and wellbore storage (Bouwer and Rice, 1976). Formation transmissivity controls the rate of movement of water into or out of the well. Wellbore storage can be significant for large diameter wells open to low-transmissivity formations, where water levels in the well have not equilibrated with the formation. Wells drilled for underground nuclear tests, denoted by a “U-“ at the beginning of the well name, have large diameters (8 ft) and long (1,000 ft) open holes. The volume of wellbore storage in these wells can be significant (several hundred to several tens of thousands of gallons). Wellbore storage prolongs the rate of water-level recovery because more time is required for water levels to equilibrate with the formation when the wellbore either is filled or evacuated following a localized disturbance.

Under pumping conditions, water levels in the well are affected by the transmissivity and storativity of the formation, and frictional well losses. Nonstatic conditions arise during pumping from frictional well losses when changes in the pumping rate cause large (more than 100 ft) drawdowns in the well. The large drawdowns are a result of how the well was constructed, and are not a property of the formation.

Wellbore Equilibration

Wellbore equilibration occurs in wells open to low-permeability formations, where estimated transmissivities are less than 1 foot squared per day (ft^2/d). An example is presented to explain the wellbore equilibration process and is followed by a discussion of how the timing of water-level recovery is estimated.

This example discusses the wellbore equilibration process in a post-drilling scenario. Following the recent drilling of a well, water-level recovery may take days, months, years, or decades to equilibrate to the formation(s) open to the well because water is injected into and (or) evacuated from the formation during well construction. After the well is completed, the water level in the well is no longer in equilibrium with the “static” hydraulic head or water level in the formation. The imbalance in water levels between the well and formation occurs because, following drilling, the water level in the well was left in a state of non-equilibrium with the formation. Either excess drilling fluid was left in the wellbore, which must move into the formation to equilibrate, or water was evacuated from the wellbore and formation water must fill the well to equilibrate. As water in the formation moves either into or out of the well, the movement is limited by the transmissivity of the formation, where low transmissivity slows recovery.

Prior to drilling a well, the static water level in the formation is unknown. However, the Bouwer and Rice (1976) solution can be used to estimate formation hydraulic conductivity and approximate the timing of static conditions, where static means the water levels are representative of formation conditions. Halford and others (2005) used this method to estimate formation hydraulic conductivity in low-transmissivity formations. The Bouwer and Rice (1976) solution can be fit to the period of water-level equilibration between the well and formation, which is observed as a steep exponential rise or decline in water levels (fig. 8*A*), in order to estimate the time necessary to reach static conditions.

Transient Trends

In this study, transient water levels represent non-steady hydrologic conditions in the groundwater system that result from anthropogenic stresses, such as groundwater pumping and (or) nuclear testing. Conceptual models are presented to explain how nuclear testing and pumping affect water levels in wells. The conceptual model of groundwater pumping is followed by a discussion of historical and current groundwater withdrawals in the study area. Groundwater has been withdrawn near Beatty, Nevada, and within the NNSS on Pahute Mesa and Jackass Flats (fig. 3; table 4).

Underground Nuclear Testing

Between 1961 and 1992, 85 nuclear tests were detonated beneath Pahute Mesa and 62 tests were detonated beneath Rainier Mesa (U.S. Department of Energy, 2015). At Pahute Mesa, large-yield nuclear devices (greater than 200 kilotons) were detonated in volcanic rocks near or below the water table, where the thick unsaturated zone was used to contain the explosive force and prevent the release of radioactive byproducts into the atmosphere (Laczniak and others, 1996). At Rainier Mesa, most of the nuclear tests were small-yield nuclear devices (less than 20 kilotons) that were detonated in a network of tunnels in low-permeability volcanic rocks (Townsend and others, 2007).

Nuclear detonations in Pahute Mesa have induced thousands of earthquakes in the study area (Hamilton and others, 1972; Rogers and others, 1977). Hamilton and others (1972) reported that large-yield nuclear devices (greater than 200 kilotons) initiated earthquake sequences, or aftershocks, lasting between 10 and 70 days, where each earthquake sequence had between tens and thousands of earthquakes with magnitudes between 2 and 5. For example, the 1.15-megaton BENHAM nuclear test induced 2,012 earthquakes over 70 days (Hamilton and others, 1972; U.S. Department of Energy, 2015). The number of earthquake aftershocks following a nuclear test decreased as nuclear testing progressed (Rogers and others, 1977).

The spatial distribution of earthquake aftershock sequences appears to be controlled by faults. Rogers and others (1977) noted that aftershocks appear to occur on north-south striking vertical faults. Hamilton and others (1972) observed dip-slip movement along normal faults in northern Pahute Mesa and right-lateral strike-slip movement along faults in southern Pahute Mesa. The epicenters of earthquake aftershocks were within about 8 mi from the point of detonation and occurred deeper (between 2.5 and 4 mi) than the burial depths (about 0.6 mi) of the nuclear tests (Hamilton and others, 1972).

Near-Cavity Response

The magnitude and duration of water-level responses to a nuclear test are dependent on rock hydraulic properties, distance from the nuclear test, and the magnitude of the seismic event (earthquake) produced by the test. When a nuclear device is detonated, a cavity approximated by a spherical volume is formed as the surrounding rock is vaporized and melted (Laczniak and others, 1996). Cavity size primarily is dependent on the yield of the nuclear test, where cavity radii range from about 100 to 325 ft for nuclear tests with specified yields in Pahute Mesa (Zavarin, 2014). Beyond the cavity, a shock wave produced by the nuclear detonation propagates outward from the cavity, generating new radial fractures and causing slip along pre-existing faults (Dickey, 1968). The formation and extent of fracturing is dependent on the yield of the nuclear device.

In wells located within tens of miles from the point of detonation, short-term and long-term water-level responses have been observed. Short-term (less than 1 day), small-magnitude oscillations in water levels have been detected in continuous water-level data and are due to nuclear-test induced earthquakes (Dudley and others, 1971). These short-term oscillations were not observed in study area wells because water levels were measured infrequently. Long-term water-level responses to nuclear testing occur in wells open to low-permeability units or wells drilled into the cavity after detonation (table 1).

Near-cavity water-level responses to nuclear testing differ between the cavity and fractures. The vaporization and ejection of groundwater from the cavity during the nuclear explosion causes the slow backfilling of water into the cavity, which is observed as a water-level rise with time (fig. 8*B*; Laczniak and others, 1996). Near the cavity, the detonation may compress the matrix of low-permeability rocks, causing highly elevated water levels that slowly re-equilibrate to the regional hydraulic head over many years (fig. 8*C*; Halford and others, 2005).

Breach Scenario

A breach scenario occurs when a nuclear detonation fractures a confining unit that separates two aquifers with different hydraulic heads. The breach in the confining unit causes a hydraulic connection between the aquifers, allowing groundwater to flow to the aquifer with lower hydraulic head. Over time, groundwater levels equilibrate to a new composite hydraulic head. This scenario is analogous to a well screened across multiple aquifers with large vertical head gradients, where the well hydraulically connects the aquifers. The breach scenario was first proposed by Carle and others (2008) to explain an anomalous water-level trend in well *UE-2ce*, adjacent to the NASH nuclear test in Yucca Flat.

A vertical breach scenario can hydraulically connect heads in a shallow and deep aquifer. For example, consider a shallow and deep aquifer separated by a confining unit, where the vertical head gradient between aquifers is upward (fig. 9*A*). If a nearby nuclear test breaches the confining unit because of chimney formation (Laczniak and others, 1996) or fracturing, then groundwater will move from the deep to the shallow aquifer (fig. 9*B*). The potentiometric surfaces of the two aquifers in the area of the breach will converge (fig. 8*D*) as the aquifer systems equilibrate to a post-test static condition (fig. 9*C*). On Pahute Mesa, a vertical breach could hydraulically connect shallow, semi-perched groundwater with deeper groundwater at a lower hydraulic head. Semi-perched groundwater occurs throughout Pahute Mesa and, locally, causes elevated hydraulic heads in shallow aquifers (Brikowski and others, 1993; Gardner and Brikowski, 1993; O'Hagan and Laczniak, 1996).

The breach scenario also could occur laterally at Pahute Mesa. Potentiometric maps of Pahute Mesa show alternating gentle and steep horizontal gradients along a groundwater-flow path (O'Hagan and Laczniak, 1996; Fenelon and others, 2010). An area where a steep gradient separates two gentle-gradient areas is analogous to a stream channel with a sequence of pools (gentle gradients) and riffles (steep gradients). A lateral breach scenario could dewater a groundwater "pool" upgradient of a steep hydraulic gradient created by a low-permeability feature. For example, consider a nuclear test that fractures low-permeability rocks and creates enhanced flow between an upgradient "pool" and downgradient "pool." The new enhanced flow path creates a short-term, non-equilibrium condition where the upgradient pool is lowered as it equilibrates with the downgradient pool. Ultimately, as with the vertical breach scenario, a post-test static condition would be obtained between the upgradient and downgradient pools (fig. 8*D*).

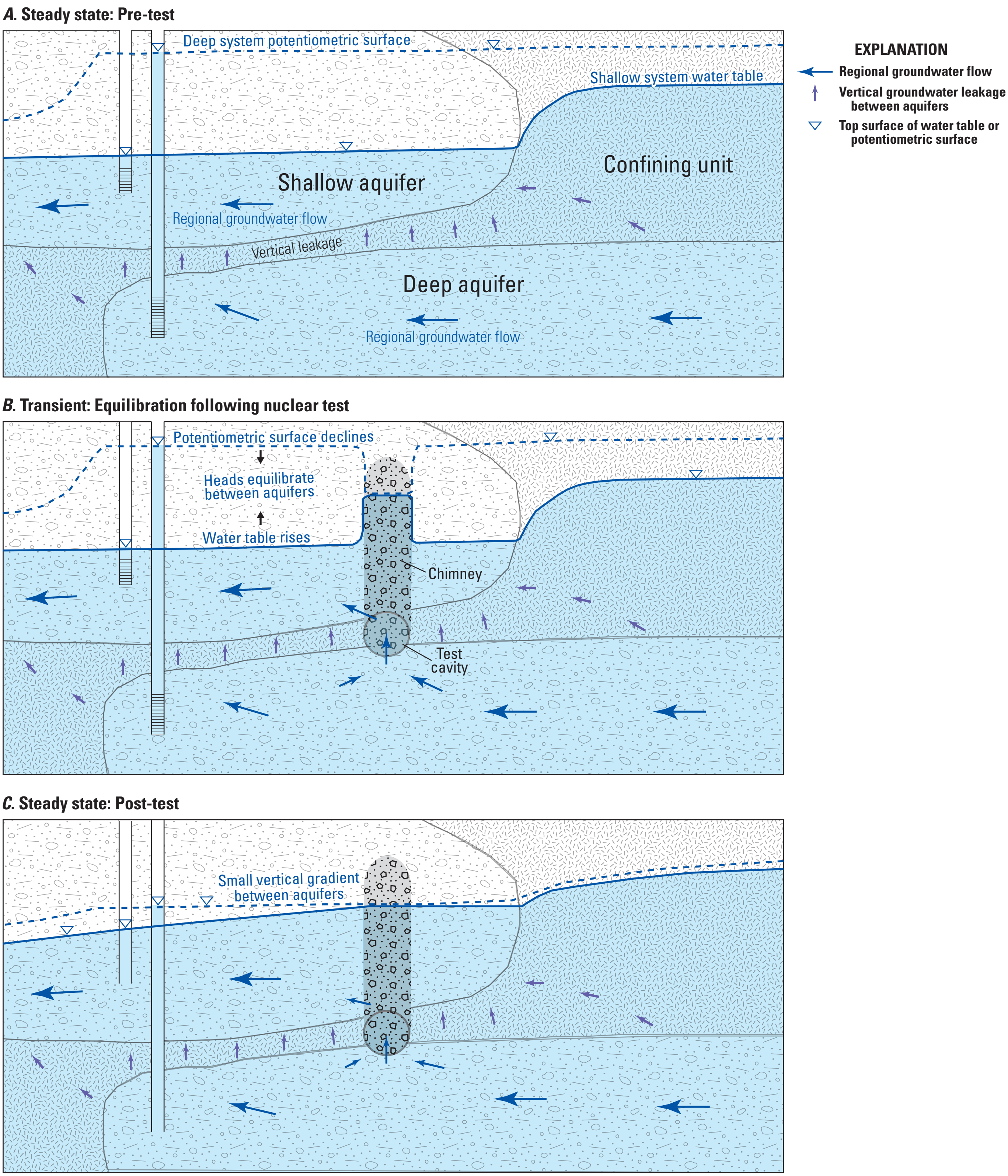


Figure 9. Generalized sections of a vertical breach scenario. (*A*) Steady-state conditions in groundwater system before nuclear testing. (*B*) Transient conditions as water levels equilibrate following testing. (*C*) Steady-state conditions in groundwater system after nuclear testing.

Groundwater Pumping

In a typical pumping scenario, groundwater is withdrawn from a pumping well and water levels decline in the well. If the formation being stressed by pumping is transmissive or is hydraulically connected to more transmissive formations, then the pumping signal readily will propagate outward. Drawdowns may be observed at nearby wells (termed observation wells) because of the pumping stress applied to the formation by the pumping well.

The magnitude of drawdown at an observation well is proportional to the pumping rate at the pumping well. For example, consider a well that is pumped continuously at a rate of 100 gal/min from 2004-05 and at a rate of 50 gal/min from 2009–10. If drawdown is measured at an observation well from both pumping periods, the magnitude of drawdown in the observation well should be greater during the 2004–05 period compared to the 2009–10 period.

In a "short-term" pumping scenario, water levels decline during pumping and recover to formation conditions post-pumping (fig. 8*E*). This means that prior to pumping, water levels in the well are assumed static, and post-pumping, water levels equilibrate to the static water level in the formation. A lack of full recovery to pre-pumping conditions indicates that another process has affected or currently is affecting water levels in the formation.

In a "long-term" pumping scenario, water levels decline during pumping and recovery is not observed (fig. 8*F*). Pumping continues over many years to decades. Water levels continue to decline in the aquifer being stressed until a source of water is captured, such as a nearby spring.

Beatty, Nevada

Groundwater withdrawals have occurred from six wells in the Beatty, Nevada, area: three wells in Bullfrog Hills and three wells in Beatty (fig. 3, table 4). The primary groundwater user (BWSD) uses the six wells for water supply in the town of Beatty (Reiner and others, 2002). Total annual withdrawals from the six wells declined 76 percent (from 410 to 98 acre-ft) from 1996 to 2015 (fig. 10) because of a population decline in Beatty in response to the decommissioning of nearby Barrick Bullfrog Mine in 1998. About 3,100 acre-ft of groundwater was withdrawn between 1996 and 2015 (fig. 10).

Withdrawals from domestic wells were minimal and excluded from water-level analyses. Twenty-two wells near Beatty were identified as supplying water to local homes and ranches for domestic use. Geter (2015) estimated an annual withdrawal of 0.5 acre-ft per domestic well. Using this annual withdrawal estimate, total withdrawals from the 22 domestic wells between 1996 and 2015 is 220 acre-ft, or about 7 percent of total withdrawals in the Beatty area.

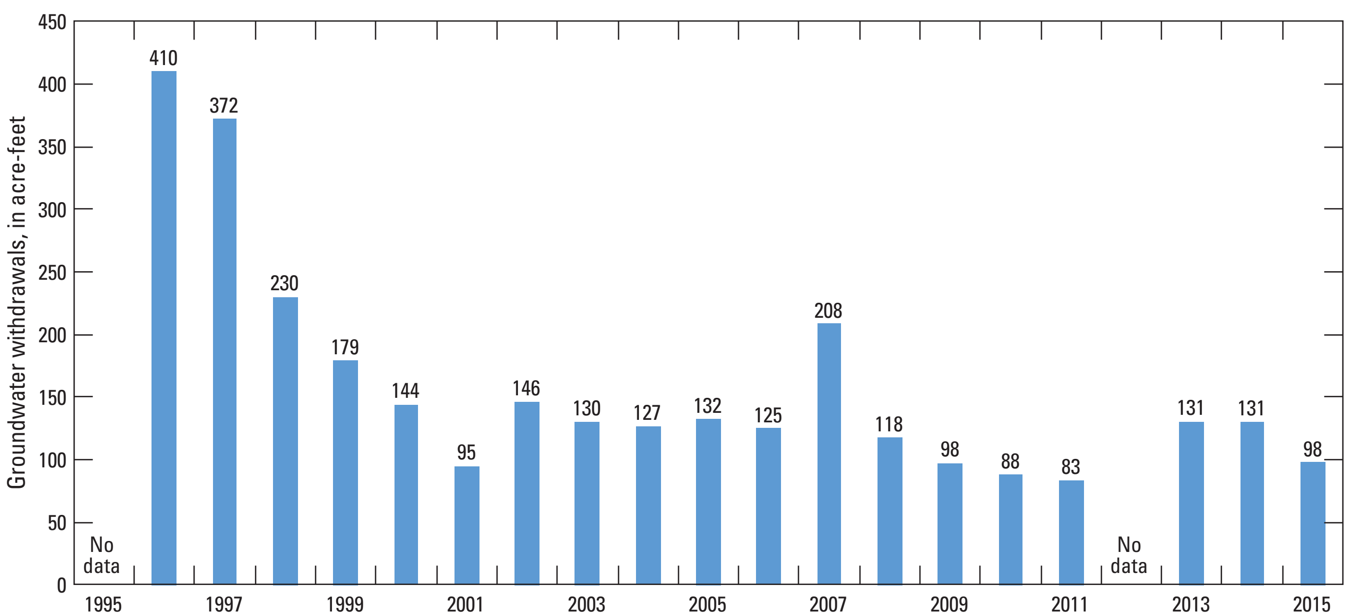


Figure 10. Total annual withdrawals from water-supply wells in and near Beatty, Nevada, between 1995 and 2015.

Nevada National Security Site

In NNSS operational areas within the study area, groundwater was withdrawn from 22 wells (table 4) to support nuclear-testing and other operations, road and drill pad construction for the drilling of wells, and municipal water supply for the onsite workforce. Groundwater withdrawals began in 1957 and have continued to present (2018). Between 1957 and 2015, about 19,600 acre-ft of groundwater was withdrawn from the NNSS in the study area. Large withdrawals in the mid-1960s were required to support the Nuclear Rocket Development Station in Jackass Flats (Young, 1972), whereas large withdrawals from the mid-1980s to early-1990s supported nuclear-testing operations (fig. 11). Groundwater withdrawals have declined since the early-1990s, which coincides with the United States signing the Comprehensive Nuclear Test Ban Treaty—an international agreement that bans the testing of all nuclear explosives (United Nations, 1996). Annual withdrawals from 2000 to 2015 were substantially less than previous periods and provided support primarily for environmental drilling operations and aquifer testing.

Seventeen withdrawal wells are in Pahute Mesa. These wells were pumped for water supply to support NNSS activities. Most groundwater in the study area was withdrawn from well *WW-8* (table 4), which has been pumped annually since 1963. Large withdrawals also occurred from the intermittent pumping of wells *U-20 WW* and *UE-19c WW* between 1975 and 2009.

Eight withdrawal wells are in Jackass Flats. Wells *J-11 WW*, *J-12 WW*, *J-12 WW (885 ft), J-13 WW*, and *J-14 WW* were pumped for water supply in Jackass Flats and three wells within borehole UE-25c 3 were pumped for 1 or 2 years between 1995 and 1997 as part of an aquifer test.

Steady-State Trends

Water levels in study area wells fluctuate with time in response to naturally occurring short-term and long-term hydrologic stresses. Barometric pressure, earth tides, earthquakes, and evapotranspiration cause short-term (hourly to daily) or seasonal (monthly) water-level fluctuations. Long-term (annual to decadal) water-level fluctuations largely are due to temporal variations in recharge.

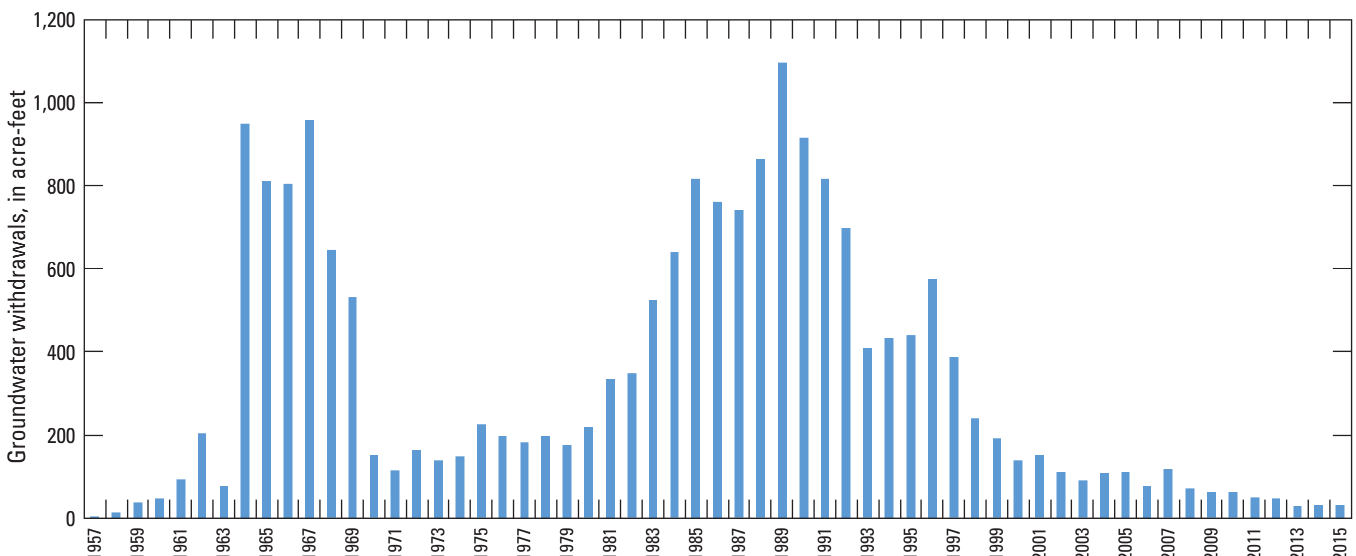


Figure 11. Total annual withdrawals on the Nevada National Security Site, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, 1957–2015.

In this study, steady-state water levels represent natural hydrologic conditions in the groundwater-flow system. This means that steady-state water levels are affected only by naturally occurring hydrologic stresses. Natural stresses cause hourly-to-decadal, steady-state water-level fluctuations in the study area. By definition, however, steady-state water levels do not change with time.

A conceptual model is presented to reconcile the definition of "steady state" as an unchanging condition with steady-state water levels that fluctuate with time. The conceptual model assumes that steady-state water levels are in a state of dynamic equilibrium. Dynamic equilibrium recognizes that water levels are not stationary, but fluctuate with time because of short-term and long-term natural stresses. However, steady-state water levels are assumed to be stable or unchanging over a defined steady-state timescale.

Conceptual Model of Recharge and Discharge

A conceptual model was used to define the steady-state timescale and to explain the cause of long-term, steady-state, water-level trends. Natural stresses that cause these long-term trends are groundwater recharge and aquifer discharge. A hypothetical water-level record was constructed to explain steady-state water-level trends based on recharge patterns in the study area.

The conceptual model for steady-state trends in the PMOV basin is that water levels remain steady over a defined steady-state timescale, but may fluctuate over shorter timescales because of varying natural hydrologic stresses. For the conceptual model discussion, the "long-term" period is the steady-state timescale. In a steady-state groundwater system, long-term water levels are in a state of dynamic equilibrium, where long-term cumulative recharge is balanced by long-term cumulative discharge and the net change in long-term cumulative storage is zero.

Water-level data indicate that the steady-state timescale is more than 25 years. The longest consistent water-level records (1990s to 2016) show that water levels have been rising in the study area and throughout southern Nevada (Fenelon and Moreo, 2002; Elliott and Fenelon, 2010; Jackson, 2018). Therefore, the steady-state timescale, where water levels remain constant, must be more than 25 years. The conceptual model described herein assumed a timescale of about a century. The timescale is tested using a precipitation data set from 1900 to 2016 to determine if recent rising water-level trends can be explained within the context of long-term steady-state conditions.

Dominant, naturally occurring hydrologic stresses affecting steady-state water levels in the study area are groundwater discharge and precipitation-derived recharge. Groundwater discharge in Oasis Valley occurs primarily from springs or seeps, which have nearly constant rates of annual discharge (Reiner and others, 2002). Aquifer discharge is constant with time because it is controlled by the nearly constant regional hydraulic gradient and unchanging hydraulic properties of the groundwater system. On the contrary, precipitation-derived recharge varies temporally and spatially, and is the primary cause of annual to decadal water-level changes in wells.

The conceptual model used a long-term (century-scale) precipitation record to represent recharge patterns in the study area. Precipitation data from the Beatty, Rainier Mesa, and Pahute Mesa precipitation indexes could not be used because these records only span between 42 and 56 years (table 3). Therefore, the conceptual model used monthly precipitation data compiled from the south-central Nevada precipitation index (Western Regional Climate Center, 2016). This index is useful because it is representative of the entire study area and has a long-term precipitation record from 1900 to 2016. Even though the precipitation index does not represent the true magnitude of precipitation in the study area, the relative distribution of dry and wet years is assumed similar.

Hypothetical recharge was determined by applying a threshold to winter (October to March) precipitation data in the south-central Nevada precipitation index, where winter precipitation above the threshold was assumed to recharge the groundwater system. In the study area, most recharge is derived during the winter months from greater-than-average precipitation when evapotranspiration approaches zero (Winograd and others, 1998; Hershey and others, 2008). In a typical winter, which has average or less-than-average precipitation, snowmelt infiltrates the root zone, and mostly or entirely contributes to the soil-moisture reservoir that was depleted by evapotranspiration during the previous summer. Little to no recharge occurs during a typical winter. Following winters with greater-than-average precipitation (wet winters), snowmelt infiltrates the root zone, wetting the soil sufficiently to exceed its field capacity and allow percolation downward to recharge the groundwater system (Smith and others, 2017). The conceptual model is simplified in that it does not consider soil moisture dynamics; specifically, the effect of annual changes in root-zone water storage between wet and dry years.

Potential recharge is calculated as the excess winter precipitation after applying a threshold. The long-term average of winter precipitation from 1900 to 2016 is 4.8 in. for the south-central Nevada precipitation index. A threshold of 125 percent of the long-term average (6 in.) was selected. The threshold used is a reasonable estimate, but could be somewhat lower or higher. For example, potential recharge is shown in fig. 12*A*, which is calculated by applying a threshold of 125 percent of the long-term average. A similar long-term recharge pattern is obtained using a threshold of 100 percent (the long-term average) or 150 percent of the long-term average. The recharge pattern in fig. 12*A* shows that about 70 percent of the recharge during the 117-year period occurred after 1968, indicating that the study area has been in a wet period from 1968 to 2016.

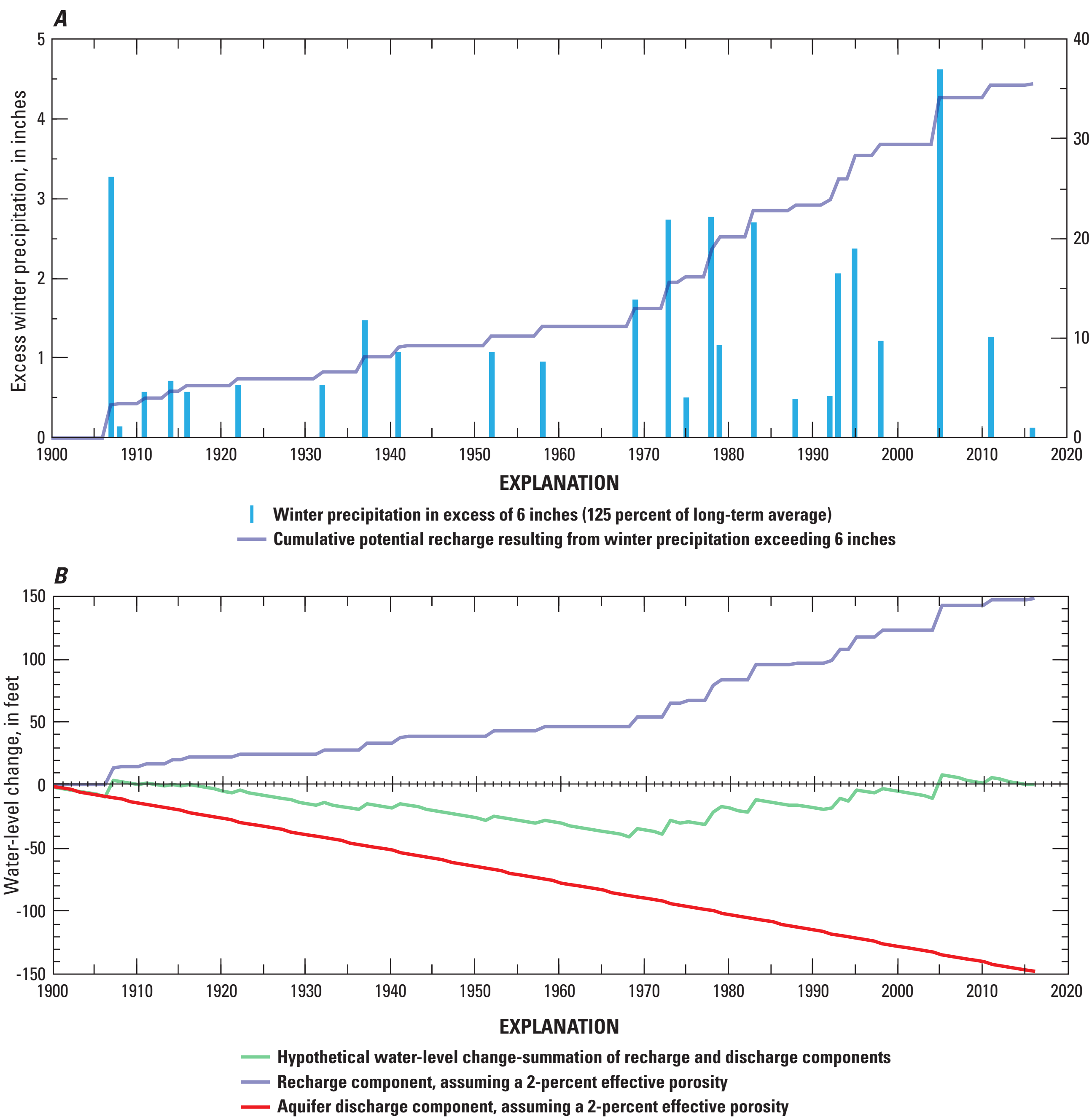


Figure 12. Steady-state conceptual model, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, 1900–2016. (*A*) Excess winter precipitation and cumulative potential recharge. (*B*) Hypothetical water-level change, potential recharge, and aquifer discharge.

A hypothetical long-term water-level record was computed assuming steady-state conditions over the 117-year record; that is, cumulative recharge equals cumulative aquifer discharge. The hypothetical water-level record is the sum of potential recharge and aquifer discharge (fig. 12*B*). The aquifer-discharge rate is assumed steady and was computed as total potential recharge from 1900 to 2016 divided by the 117-year period. Potential recharge and aquifer discharge were converted to water-level change by dividing rates (in feet per year) by an assumed fractured-rock effective porosity of 2 percent.

The hypothetical water-level record of recharge and aquifer discharge can explain the observed rising water-level trends in the study area using an assumed century-scale steady-state period. Consistent with the concept of long-term steady-state conditions, the net hypothetical water-level change for the period of record is zero (fig. 12*B*). However, the hypothetical water-level change shows a declining trend from 1900 to 1968 and a rising trend from 1968 to 2016 (fig. 12*B*). Assuming that steady-state conditions occur on a century timescale, measured water-level trends in the study area from 1995 to 2016 should be upward because the study area was in a relatively wet period since 1968. The magnitude and exact pattern of water-level trends is expected to differ between the hypothetical water-level record and water-level records in study area wells because of differing rock porosities, recharge rates, and aquifer discharge rates.

Reconciling Modern Recharge with Old Groundwater Ages

The conceptual model to explain steady-state water-level trends assumes that the groundwater system receives modern recharge. Deep unsaturated zone depths between 1,000 and 3,000 ft in Pahute and Rainier Mesas and Yucca Mountain do not preclude modern recharge. The following discussion reconciles a conceptual model of assumed modern recharge with mean regional groundwater ages from the end of the Pleistocene Epoch. Groundwater isotopic data from previous studies are presented. Stable isotopes of hydrogen (δD) and oxygen ($\delta^{18}O$) are used to interpret recharge sources (Gat, 1980). Elevated concentrations of “bomb-pulse” tracers, such as tritium (^{3}H), carbon-14 (^{14}C), and chloride-36 (^{36}Cl), are event markers that correlate to atmospheric thermonuclear weapons testing during the 1950s and 1960s (Bentley and others, 1982; Nydal and Lovseth, 1983). Groundwater with high concentrations of bomb-pulse tracers indicate the groundwater system has received recharge within 60 years. A measure of the carbon isotope values, $\delta^{13}C$ and ^{14}C, can further refine recharge source interpretation and groundwater age and mixing (Rose and others, 2006).

Numerous chemical and isotopic analyses have been done in the study area to characterize hydrochemical facies, source areas, groundwater ages, and flow paths (Schoff and Moore, 1964; Blankennagel and Weir, 1973; Winograd and Thordarson, 1975; Winograd and Pearson, 1976; White, 1979; White and others, 1980; Claassen, 1986; White and Chuma, 1987; Thomas and others, 1996; 2002; Kwicklis and others, 2005; Rose and others, 2006; Hershey and others, 2008). These studies indicate that the *primary source* of regional groundwater in the study area is derived from a colder climatic period (end of the Pleistocene).

Previous geochemical analyses report that modern recharge occurs throughout the study area. Modern groundwater is defined as groundwater less than 1,000 years old. Rose and others (2006) estimated average groundwater age ranges using radiocarbon, and determined that modern recharge occurs in Pahute Mesa (wells *UE-19h* and *U-20 WW*), Oasis Valley (wells *ER-OV-01* and *ER-OV-03a*), and Fortymile Wash (wells *UE-29a1*, *UE-29a2*, and *UZN 91*). Rose and others (2006) estimated that modern recharge could account for as much as 16 percent of the groundwater discharged in Oasis Valley, whereas Kwicklis and others (2005) estimated that modern recharge could be no more than 20 percent. Kwicklis and others (2005) also used bomb-pulse signatures to indicate modern recharge in Beatty Wash and Fortymile Wash. Therefore, downgradient of Pahute Mesa, modern recharge is derived from focused recharge along the Thirsty Canyon, Beatty Wash, and Amargosa River drainages, whereas modern recharge in Fortymile Canyon is derived from focused recharge in Fortymile Wash (Thomas and others, 2002; Rose and others, 2006).

Isotopic data suggest mixing of modern groundwater with regional (older) groundwater in areas where the water table is deep. Well *WW-8* has an average depth-to-water of 1,080 ft; however, high ^{14}C and low $\delta^{18}C$ values, compared to other Pahute Mesa wells, and heavy δD and $\delta^{18}O$ values indicate regional groundwater mixing with modern recharge (Rose and others, 2006). Wells *J-12 WW* and *J-13 WW* in Jackass Flats have average depth-to-waters of 740 and 930 ft, respectively; however, ^{14}C, δD, and $\delta^{18}O$ values indicate mixing with young groundwater (Rose and others, 2006). Similarly, well *ER-EC-7* near Timber Mountain has an average depth-to-water of 746 ft, but heavy δD and $\delta^{18}O$ values and ^{14}C values indicate a modern recharge signature, possibly from focused recharge along Beatty Wash (Rose and others, 2006).

Chemical and isotopic studies of springs, water in wells, and discharge from tunnels indicate modern recharge occurs in Rainier Mesa. Clebsch (1961) measured bomb-pulse tritium concentrations in Whiterock Spring (about 3.5 mi northeast of well *ER-12-1*) and E-Tunnel discharge at Rainier Mesa, and estimated groundwater ages between 0.8 and 6 years. Russell and others (1987) reported that N-Tunnel discharge responds rapidly to winter recharge events and used δD and $\delta^{18}O$ to determine that the water is of recent meteoric origin. Norris and others (1990) detected high concentrations of a bomb-pulse tracer (^{36}Cl) centered on a fault intersecting G-Tunnel at Rainier Mesa, and attributed the ^{36}Cl to preferential flow of modern recharge along the fault.

Previous geochemical studies of the study area indicate mixing of small volumes of modern recharge with older groundwater in the regional flow system. A conceptual model that attributes rapid water-level rises to specific wet winters does not necessarily require large volumes of recharge to cause the rises. Water-level rises in wells following a wet winter can result from small amounts of groundwater percolating to the water table and filling poorly connected fractures in a rock unit with low effective porosity.

Natural Hydrologic Stresses

Natural hydrologic stresses include recharge, evapotranspiration, barometric pressure, earth tides, and earthquakes. Water-level responses to each natural stress are described.

Recharge

Recharge causes short-term or long-term water-level rises in wells. Short-term water-level rises and declines (fig. 8*G*) occur in response to focused recharge, such as in an ephemeral channel following a streamflow event. The water-level response is highly sensitive to recharge events and the entire rise occurs rapidly (within months) after an event. The response appears similar to a flood peak and recession curve on a streamflow hydrograph. A long-term water-level rise (fig. 8*H*) occurs in response to recharge through a thick unsaturated zone. The attenuated response may occur within several months to several years following the recharge event. The rising limb of the hydrograph can last several years before the water level begins declining. If multiple recharge events are "stacked together", such that a recharge event occurs every few years, the hydrograph will continue to rise as the recession part of the curve is interrupted by each new recharge pulse.

Evapotranspiration

Evapotranspiration causes seasonal water-level fluctuations in groundwater discharge areas. Groundwater discharge areas occur within southern Oasis Valley and Sarcobatus Flat (fig. 3). Evapotranspiration occurs each year, primarily during the growing season (about March through September). Evapotranspiration in Oasis Valley causes water levels to fluctuate as much as 7 ft annually and 0.2 ft daily (Reiner and others, 2002). Water levels with responses to evapotranspiration have a diagnostic cyclical pattern, where water levels decline in the growing season and recover during the late autumn and winter (fig. 8*I*).

Barometric Pressure and Earth Tides

Water levels measured quarterly in the study area can be affected by barometric pressure and earth tides. Barometric pressure and earth tides cause water levels to fluctuate in wells open to confined aquifers with low storage, such as the fractured volcanic-rock aquifers in Pahute Mesa (Fenelon, 2000) because, in aquifers with low storage, the aquifer skeleton expands or contracts in response to barometric or tidal forcing. Barometric pressure and earth tides typically induce daily to seasonal cyclic water-level changes of less than 1 ft (Fenelon, 2000; Halford and others, 2012). These short-term water-level fluctuations in study area wells do not affect analyses of trends that span multiple years and, therefore, were not analyzed for this study.

Earthquakes Induced by Tectonic Activity

Responses to natural earthquakes were not observed in measured periodic water levels of study area wells. Some water levels in the study area do respond to earthquakes, but these short-term responses have been observed only in continuous (transducer) water-level data. For example, continuous water-level monitoring in well *UE-18r* captured a short-term response to the 7.9 magnitude earthquake that occurred southeast of Little Sitkin Island, Alaska, on June 23, 2014 (Navarro, 2015). Wells screened in moderate to high transmissivity lithologic units, such as well *UE-18r*, equilibrate within minutes to days following an earthquake. Water-level fluctuations from earthquakes range from small (a few inches) to moderate (a few feet) (Fenelon, 2000). Water-level records based on quarterly tape-down measurements are not likely to show water-level responses to earthquakes because measurements are infrequent.

Trend Analysis of Groundwater Levels

Water-level trends in study area wells are categorized as either nonstatic, transient, or steady state (tables 1 and 2). The categorization of a water-level trend as nonstatic, transient, or steady state does not imply that the entire water-level record at a well is representative of that categorization. For example, the initial water-level record at a well may show a 1-year recovery response following well completion and development, but water-level data over the next 15 years are representative of steady-state conditions. In this case, the water-level trend in the well is classified as steady state and the short-term nonstatic trend is removed from the steady-state analysis. Therefore, the categorization of water-level trends is based on the dominant water-level response in wells. The only exception is wells with water-level responses to nuclear testing. If any part of a water-level record is (or potentially is) affected by nuclear testing, then the water-level trend is categorized as transient and subcategorized as nuclear testing or combined nuclear testing and pumping effects. The reason for this exception is that nuclear testing effects on water levels are a unique phenomenon important to groundwater studies on the NNSS.

Only wells within the PMOV basin were selected for nonstatic and transient trend analyses, and the entire water-level record was used in the analyses. That is, the period of analysis for nonstatic and transient trends ranged from 1 to 41 years between 1966 and 2016. The focus of the steady-state trend analysis is on water levels measured between January 1995 and March 2016, a period when most wells have a consistent record. Wells selected for the steady-state trend analysis each have at least 30 steady-state water-level measurements spanning 10 or more years between 1995 and 2016.

Nonstatic Trends

Water-level trends are classified as nonstatic if the water-level record is dominated by wellbore equilibration, where the equilibration period spans from months to decades. Some wells classified as transient or steady state have nonstatic water levels within the water-level record;

however, water levels equilibrated to the formation and the wells have sufficient water-level data to be used in transient or steady-state trend analyses. Four water-level hydrographs are dominated by nonstatic conditions, whereas 19 water-level hydrographs have transient or steady-state trends with short-term periods of nonstatic water levels.

Wells Dominated by Wellbore Equilibration

Water-level hydrographs from four wells in the PMOV basin (fig. 3; table 1) are dominated by wellbore equilibration following localized disturbances to the wellbore. Nonstatic water levels were analyzed using Bouwer and Rice (1976) transforms to estimate formation hydraulic conductivity and to determine the period of water-level recovery to the static water level. The static water level represents steady-state or transient conditions in the regional groundwater system.

The early period of water levels in well *U-20ax* are dominated by wellbore equilibration (fig. 13). Water levels initially declined about 85 ft following well drilling. The well was bailed on January 25, 1988, which caused the water level to drop an additional 60 ft. A WLM was used to simulate wellbore equilibration from drilling and bailing recovery with Bouwer and Rice transforms. A hydraulic conductivity of 0.003 ft/d was estimated for the volcanic tuff open to the well. WLM results show that water levels recovered to steady state by September 1988. This result was used to flag water levels as steady state from September 1988 through 1992.

Water levels in well *U-20be* show a discernible wellbore equilibration response to drilling (fig. 14). A Bouwer and Rice transform was used to simulate wellbore equilibration by fitting the transform to measured water levels. Results indicate that nonstatic conditions occur from June 1989 to August 1990. The long recovery period is a result of the large-diameter (8 ft) well, which required more than 20,000 gallons of water to drain out of the well annulus, and the low hydraulic conductivity (0.01 ft/d) of the bedded tuffs open to the well. Water levels recovered by late 1990, with later water levels representing steady-state conditions.

Water levels in well *ER-20-2-1* are dominated by wellbore equilibration and have declined about 4 ft from 1996 to 2016 following well construction (fig. 15). A hydraulic conductivity of 2×10^{-4} ft/d was estimated for the zeolitic bedded and nonwelded tuffs open to the well by fitting a Bouwer and Rice transform to the water-level data. Water levels in 2014 approximate steady-state conditions, based on Bouwer and Rice transform results that indicate levels are almost fully recovered.

Well *PM-1* is dominated by nonstatic water-level responses. The well was bailed for annual water sampling between 1989 and 2011 (Townsend and Grossman, 2003; fig. 16). Bailing between 5 and 10 liters of water from the 10-in. diameter well casing caused measured water-level declines of between 0.2 and 1.8 ft. The slow water-level recovery indicates the ash-flow tuff open to well *PM-1* has low permeability (Elliott and Fenelon, 2010). The last water-level measurement in 2015 is flagged as steady state; however, this water level approximates steady-state conditions because water levels are still recovering.

Short-Term Nonstatic Effects

Five of the 12 wells dominated by transient water-level trends (table 1) have nonstatic water-level measurements. Initial water levels in wells *PM-2, U-19bj*, *U-20ao*, and *U-20bf* were equilibrating to the formation following well construction. The dominant water-level trends in these wells are classified as transient because the water levels have been affected by nuclear testing. Well *U-20 WW* has 11 water levels flagged as nonstatic between 1994 and 2008, when the well was pumping from 140 to 335 gal/min (Jackson, 2018). The large (between 350 and 750 ft) drawdown and recovery responses represent well-loss effects.

Fourteen of the 62 wells dominated by steady-state water-level trends (table 2) have nonstatic water-level measurements. Nonstatic levels were measured in wells *ER-12-3 m*, *ER-12-4 m*, *ER-18-2*, *ER-EC-7*, *JF-1*, *PM-3-1*, *TW-1*, and *U-12s* when small volumes of water were withdrawn for well development or purging the wells for water-quality sampling (Elliott and Fenelon, 2010). Initial water levels in wells *ER-12-3 p*, *ER-12-4 m*, *ER-18-2*, *U-19bh*, *UE-18t*, *UE-29a2*, *UZN 91*, and *G-2* were equilibrating to the formation following well construction (Jackson, 2018). Water levels in these wells ultimately equilibrated to the surrounding aquifer so that the long-term trends are representative of steady-state conditions. For example, water levels in well *U-19bh* show a discernible wellbore equilibration response after the well was dewatered in 1991 (fig. 17). The nearly 1-year recovery period in the well was a result of the large diameter (8 ft) of the well and the low hydraulic conductivity (0.008 ft/d) of the partially zeolitized, nonwelded ash-flow tuff open to the well. Fitting a Bouwer and Rice transform to measured water levels indicates that water levels in the well from June 1992 to 2016 do not represent well recovery, but rather are representative of steady-state conditions. These steady-state water levels were used in the steady-state trend analysis.

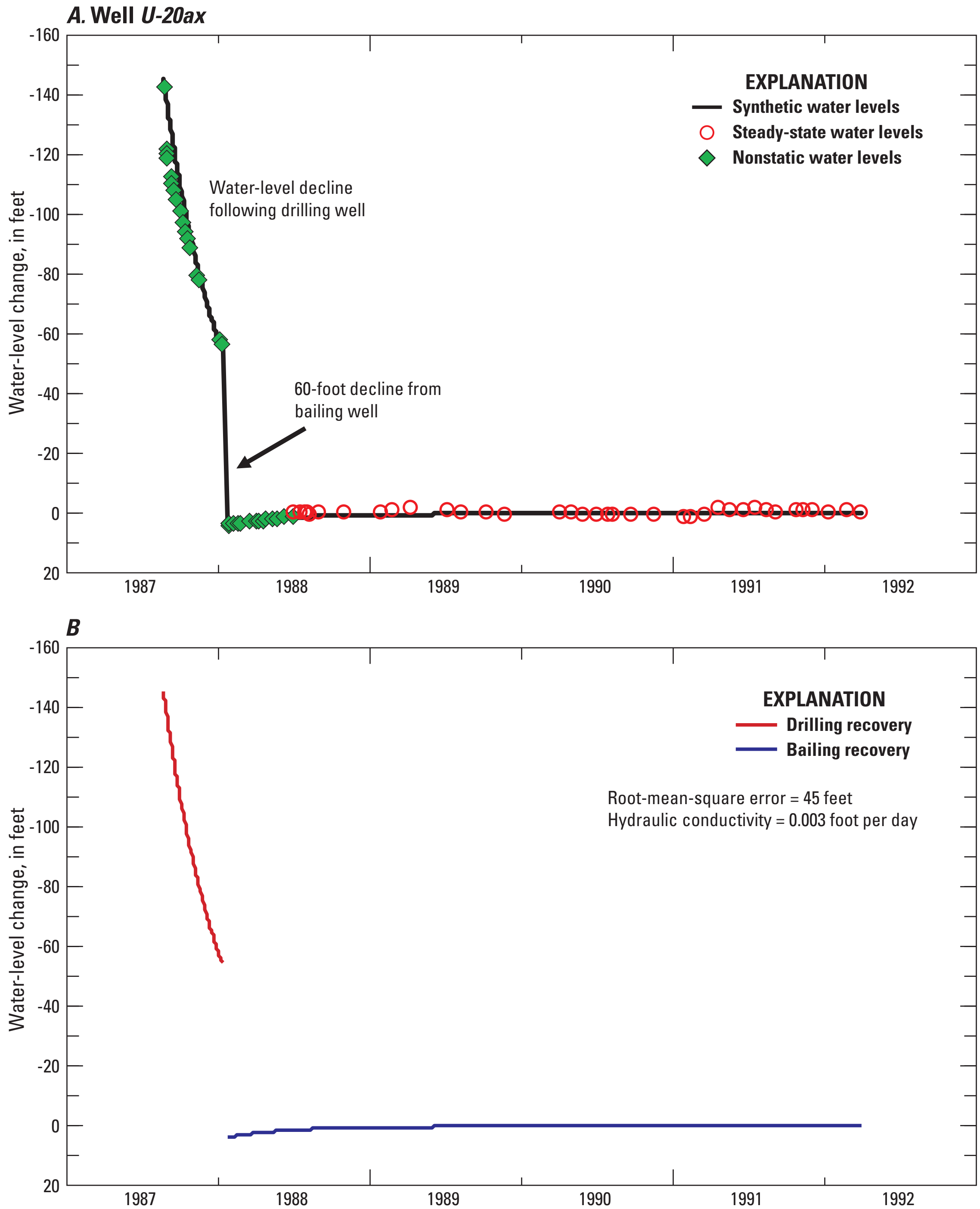


Figure 13. Water-level model results for well *U-20ax*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. (*A*) Comparison of synthetic, nonstatic, and steady-state water levels and (*B*) two components of synthetic curve: Bouwer and Rice transforms simulating wellbore equilibration following drilling and bailing recovery.

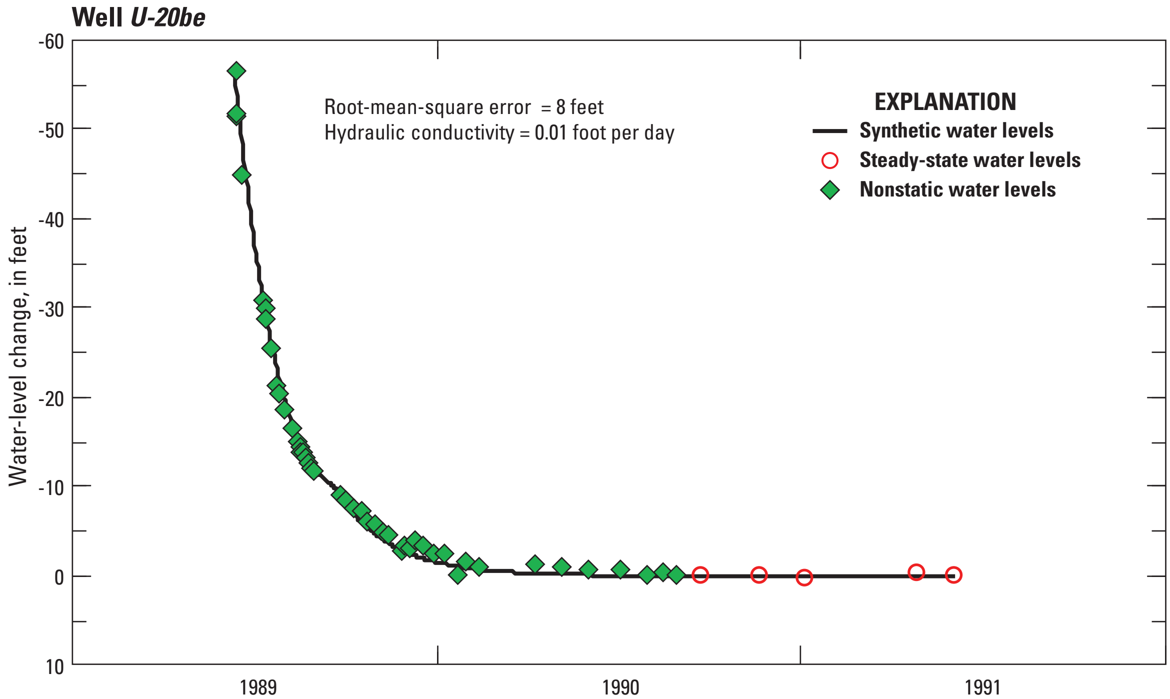


Figure 14. Comparison of nonstatic and steady-state water levels to synthetic water levels (Bouwer and Rice transform) in well *U-20be*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

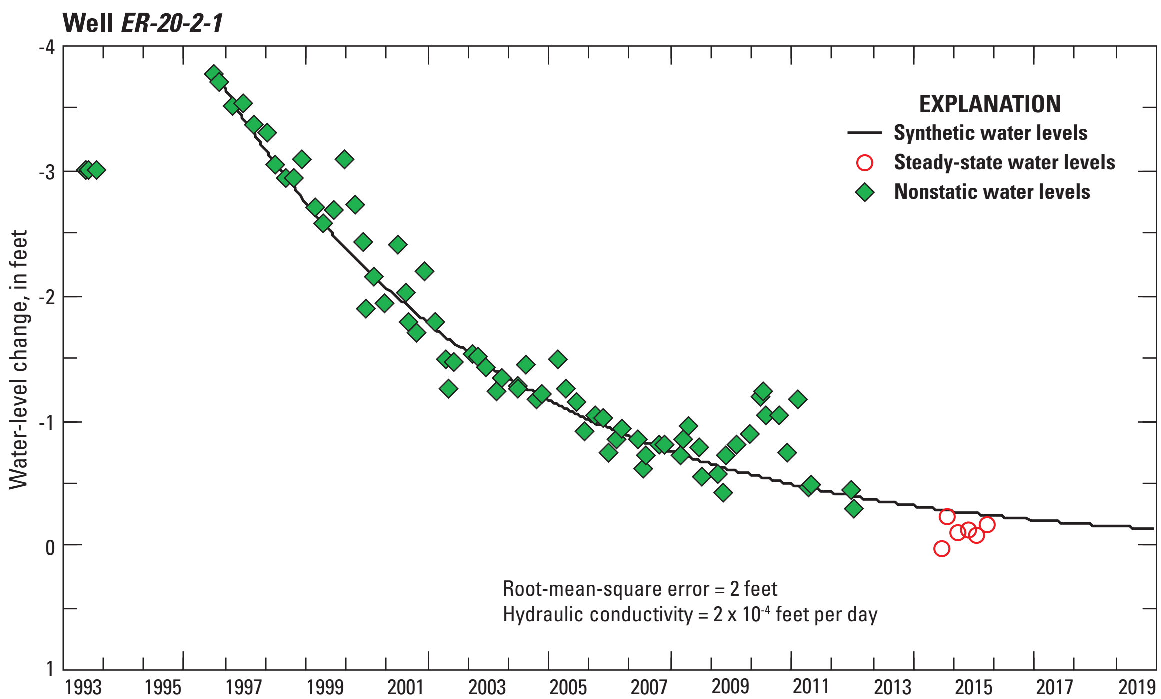


Figure 15. Comparison of nonstatic and steady-state water levels to synthetic water levels (Bouwer and Rice transform) in well *ER-20-2-1*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

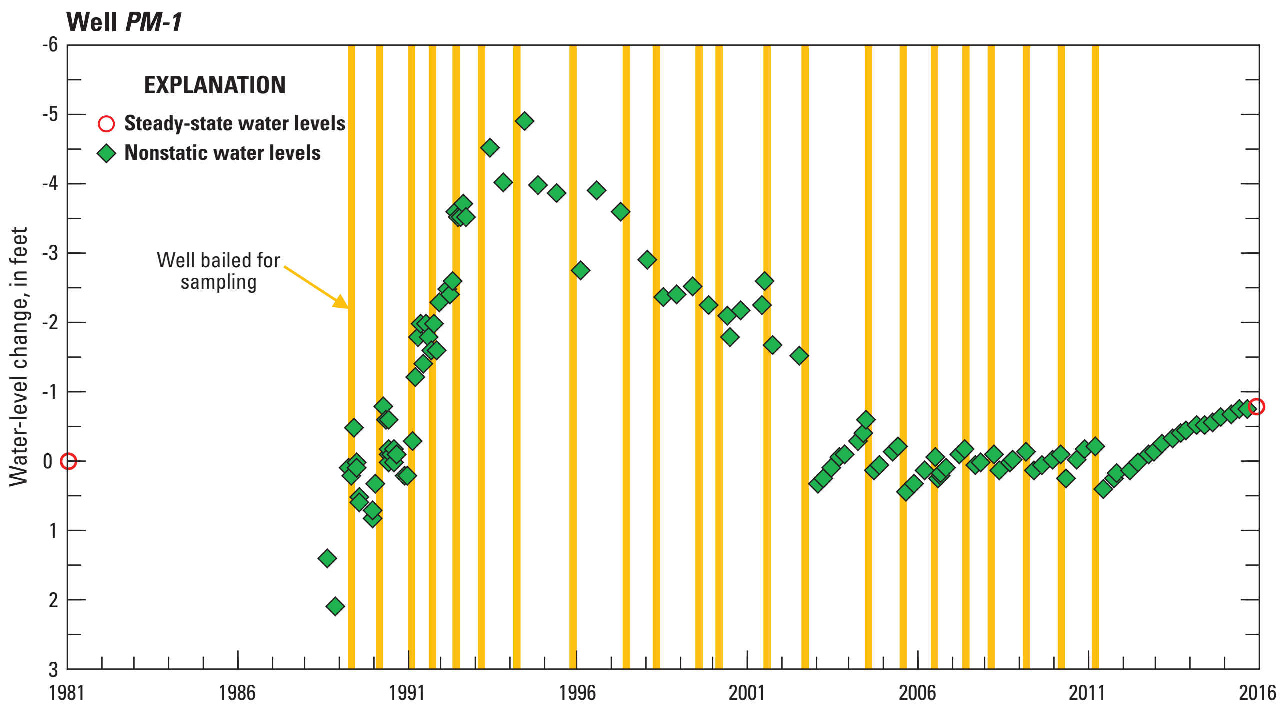


Figure 16. Nonstatic and steady-state water levels in well *PM-1*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

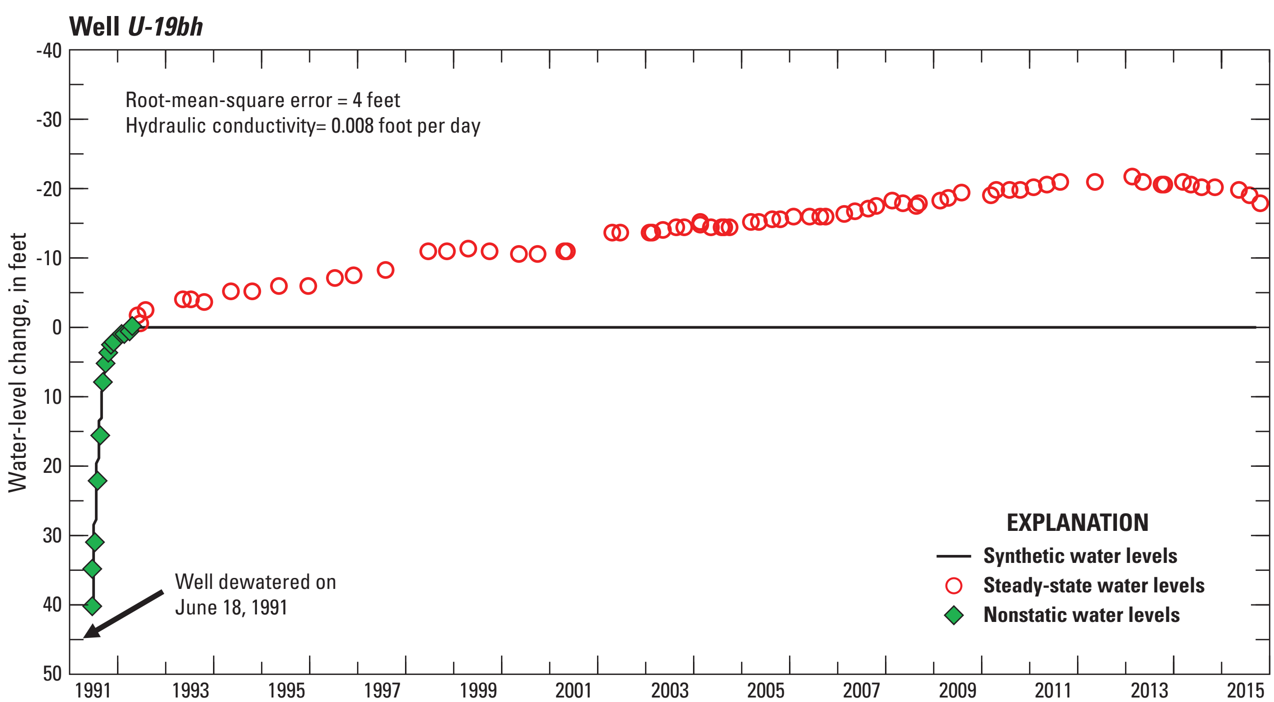


Figure 17. Comparison of nonstatic and steady-state water levels to synthetic water levels (Bouwer and Rice transform) in well *U-19bh*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

Transient Trends

Transient water levels were analyzed to estimate the magnitude of water-level responses to nuclear testing and groundwater pumping. Water-level hydrographs from 12 wells (fig. 3; table 1) have transient trends. Eleven wells potentially have been affected by nuclear testing: five of these wells were affected only by testing and six may have been affected by pumping and nuclear testing. Potential nuclear-test effects on water levels are discussed qualitatively. Water levels in a few study area wells have been affected by groundwater withdrawals from Pahute Mesa, Beatty, and Jackass Flats. Water levels with long-term pumping responses were simulated with WLMs to quantify the drawdown.

Nuclear Testing

Five wells have potential water-level responses only to underground nuclear testing. All five wells are located in Pahute Mesa, in NNSS operational areas 19 and 20 (fig. 3; table 1). Three of the wells potentially were affected by deep (more than 1,000 ft) nearby underground nuclear tests, one well potentially was affected by a nearby shallow crater test, and one well has a declining trend that potentially is attributed to a breach scenario in response to a nuclear test.

Wells *UE-20f*, *U-19v PS 1D*, and *U-20ao* potentially have been affected by deep underground nuclear tests (table 1). Water levels in well *UE-20f* rose after the nearby BENHAM (1.15 megaton) nuclear test was detonated on December 19, 1968 (U.S. Department of Energy, 2015; fig. 18*A*). Water levels in well *UE-20f* were elevated more than 50 ft above pre-test water levels following the BENHAM nuclear test, detonated about 3 mi away. Eight years after the test, water levels were 4.6 ft higher than pre-test levels and still recovering. Well *U-19v PS 1D* is a post-shot test hole drilled into the ALMENDRO (200–1,000 kiloton) nuclear-test cavity shortly after the nuclear detonation on June 6, 1973 (U.S. Department of Energy, 2015; fig. 18*B*). Water levels in the well rose about 860 ft from September 1973 to 2009 because of the slow backfilling of water into the cavity after groundwater was ejected or vaporized from the cavity during the nuclear explosion. Water levels in well *U-20ao* rose after the SALUT (20–150 kiloton) nuclear test was detonated 1.1 mi away on June 12, 1985 (U.S. Department of Energy, 2015; fig. 18*C*). The measured water level in well *U-20ao* was elevated about 39 ft above pre-test levels 15 days after the test.

Water levels in well *PM-2* potentially have been affected by a shallow crater nuclear test. Well *PM-2* is 860 ft northwest of the SCHOONER (30 kiloton) crater test, which was detonated 365 ft below land surface on December 8, 1968 (U.S. Department of Energy, 2015). The 12-ft water-level decline in well *PM-2*, measured 6 months after the test, potentially was caused by the ejection of groundwater following the nuclear explosion (fig. 18*D*). Discussion of nonstatic trends from 1966 to 1968 and from 1983 to 2006, and leakage of water down the well annulus is provided in appendix 1.

Water levels in well *U-19bj* potentially have been affected by a breach caused by a nearby nuclear test. Water levels in the well have a declining trend that cannot be explained exclusively by wellbore equilibration following well construction (fig. 19). Comparison of measured water levels to a Bouwer and Rice transform indicates that the initial water-level decline between August 1992 and August 1997 is recovery following well construction. However, between 1998 and 2016, water levels show a nearly linear decline that is not a nonstatic trend as defined in this report, but might be attributed to groundwater equilibration between aquifers from nearby nuclear testing. For example, a nuclear test may have breached a confining unit between aquifers (breach scenario) prior to the construction of well *U-19bj*, and water levels are equilibrating to a new static condition. Eight nuclear tests were detonated within 1 mi of well *U-19bj* between 1976 and 1991, and range in announced yield between 20 and 1,000 kilotons (U.S. Department of Energy, 2015)

Groundwater Pumping Effects

Water levels in wells have been affected by long-term or short-term pumping. One well has long-term (more than 20 years) water-level changes from groundwater pumping, whereas five wells have short-term (less than 3 years) water-level changes from groundwater pumping. These wells also have water-level responses to natural hydrologic stresses. However, only water-level trends in wells dominated by long-term pumping are categorized as transient (pumping).

Long-Term Effects

Water levels in well *Beatty Wash Terrace* likely are affected by pumping in *Beatty Well 1*, about 2 mi to the southwest (fig. 3). Both wells are screened in shallow basin fill within a groundwater discharge area. Water levels in well *Beatty Wash Terrace* have strong recharge responses from the 1998 and 2005 winters superimposed on seasonal evapotranspiration responses. Water levels also have a statistically significant declining trend that suggests water levels may be affected by nearby pumping.

To determine if water levels are affected by pumping, a WLM was used to simulate recharge, evapotranspiration, and pumping responses. A good fit was achieved between measured (transient) and synthetic water levels (fig. 20*A*) using these three inputs. A water-level decline of about 1 ft was estimated at well *Beatty Wash Terrace* because of pumping from *Beatty Well 1* (see pumping drawdown curve in fig. 20*B*).

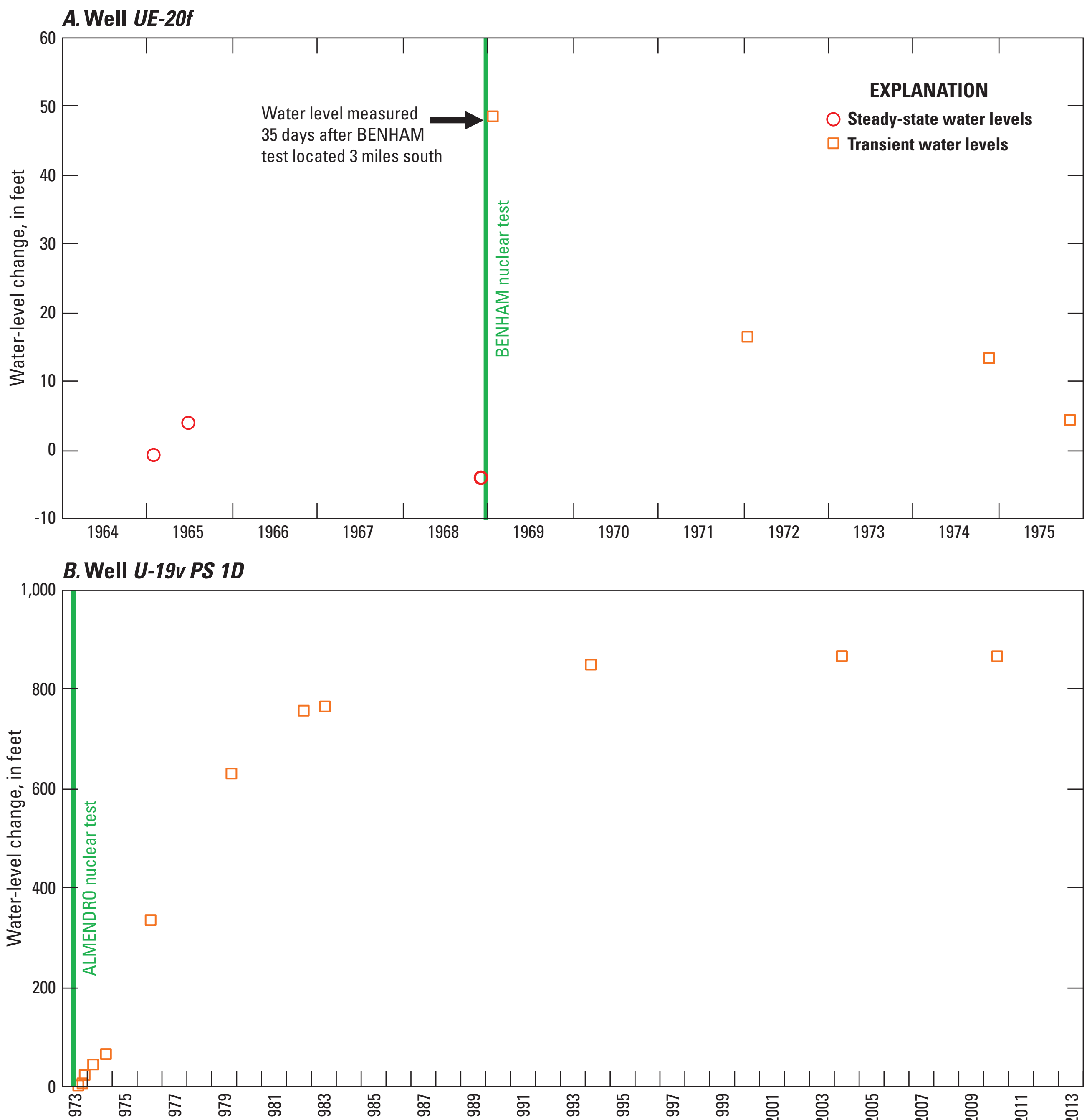


Figure 18. Measured water levels in wells (*A*) *UE-20f*, (*B*) *U-19v PS 1D*, (*C*) *U-20ao*, and (*D*) *PM-2*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

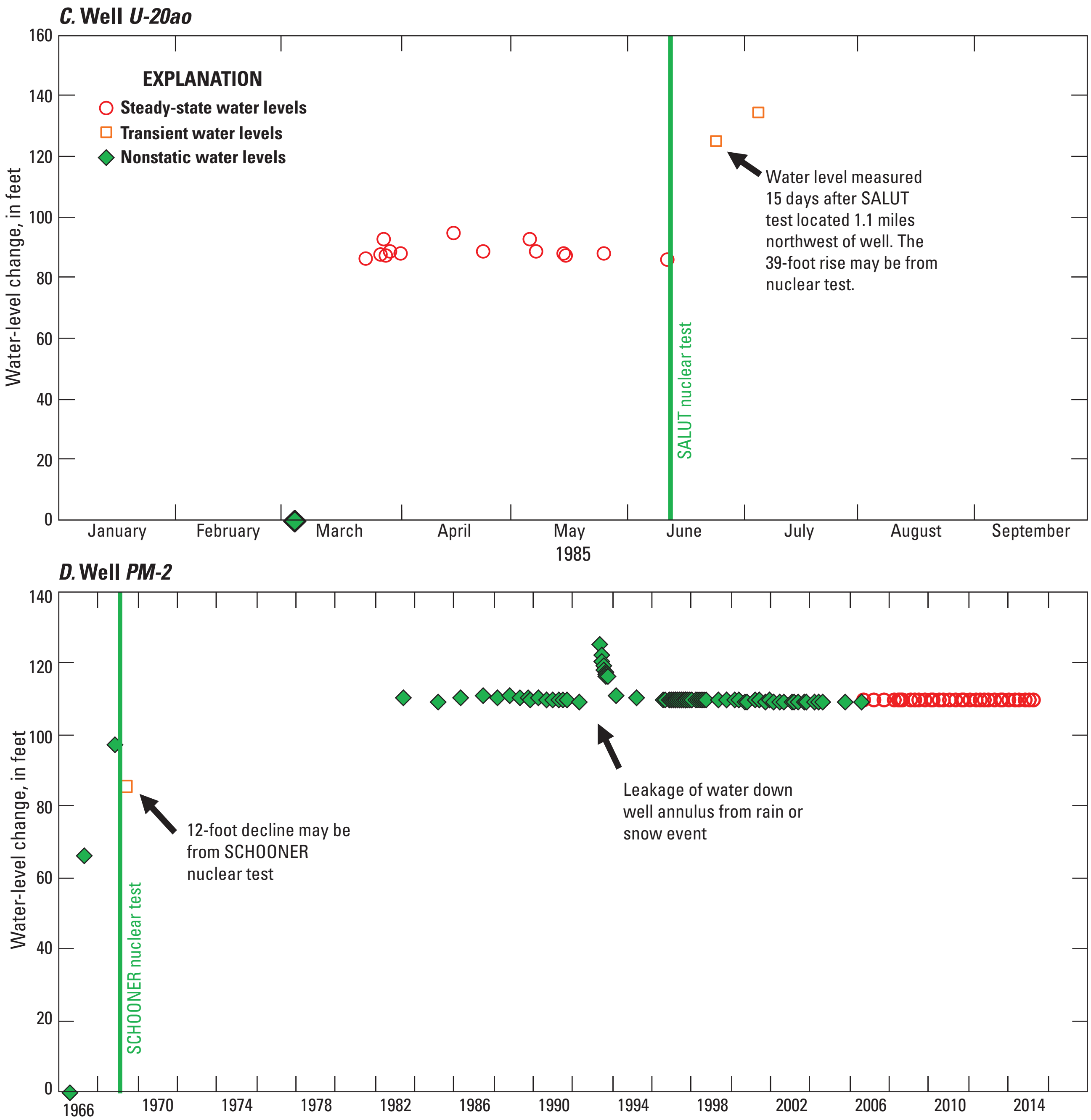


Figure 18.—Continued.

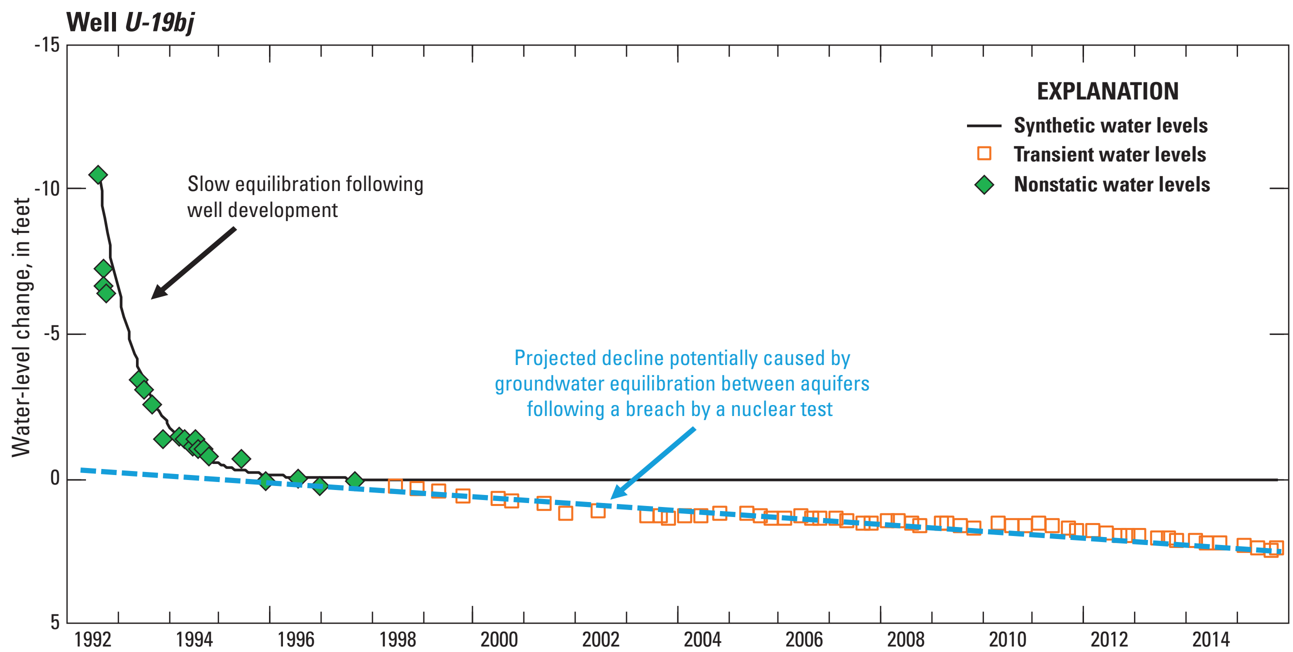


Figure 19. Comparison of measured water levels to synthetic water levels (Bouwer and Rice transform) in well *U-19bj*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

Well *Beatty Wash Terrace* is the only study area well with measured water levels affected by pumping in Beatty. No other study wells have water-level hydrographs dominated only by groundwater pumping.

Short-Term Effects

Five wells have between 3 and 20 water-level measurements that show short-term water-level changes from groundwater pumping (Jackson, 2018). Water levels in all five wells are dominated by natural hydrologic stresses; therefore, levels representing periods of pumping were excluded from the steady-state trend analysis. Short-term transient effects resulted from pumping during aquifer tests or from water-supply wells. Between June and October 2000, water levels in well *ER-EC-8* were affected by well development, aquifer testing, and recovery in the well (Elliott and Fenelon, 2010). Well *UE-19c WW* was used as a water-supply well for the NNSS between 1975 and 1994, and three water-level measurements in the well between 1984 and 1992 show short-term changes from pumping (Jackson, 2018). Nine water-level measurements in well *G-2* indicate that at least 11 ft of drawdown occurred when the well was pumped during a single-well aquifer test in 1996. Between 1995 and 1997, aquifer testing in wells within borehole UE-25c 3 (Geldon and others, 2002) likely induced water-level declines in wells *UE-25 WT 4* and *UE-25 WT 16* between 1996 and 1998.

Combined Nuclear Testing and Pumping Effects

In eastern Pahute Mesa, water-level fluctuations in six wells are attributed to nearby nuclear testing and pumping from water-supply well *U-20 WW* (fig. 21). Wells *U-20 WW*, *UE-20bh 1*, *UE-20n 1*, *U-20n PS 1DD-H*, *U-20bg*, and *U-20bf* have similar water-level trends (fig. 22). Previous studies have documented that these wells are affected by pumping in well *U-20 WW* (Fenelon, 2000; Garcia and others, 2011). However, water-level measurements from 1985 to 2000 show a declining trend of more than 10 ft that cannot be attributed to pumping from well *U-20 WW*. The declining trend potentially was caused by a lateral or vertical breach from a nearby nuclear test, which permanently lowered the hydraulic head in the aquifer open to the wells.

WLM results indicate that the water-level trends in wells near *U-20 WW* (fig. 22) cannot be explained by pumping alone. A WLM was used to simulate water levels in well *U-20 WW* because this well has the longest water-level record and is representative of water levels in nearby wells, which have similar trends (fig. 22). A synthetic water-level curve was fit to measured water levels, where the synthetic curve is the sum of Theis transforms simulating pumping in well *U-20 WW* from 1985 to 2016 (fig. 23). Synthetic water levels approximate measured water levels from 1996 to 2016, indicating that pumping can explain most of the water-level response; however, synthetic and measured water levels have

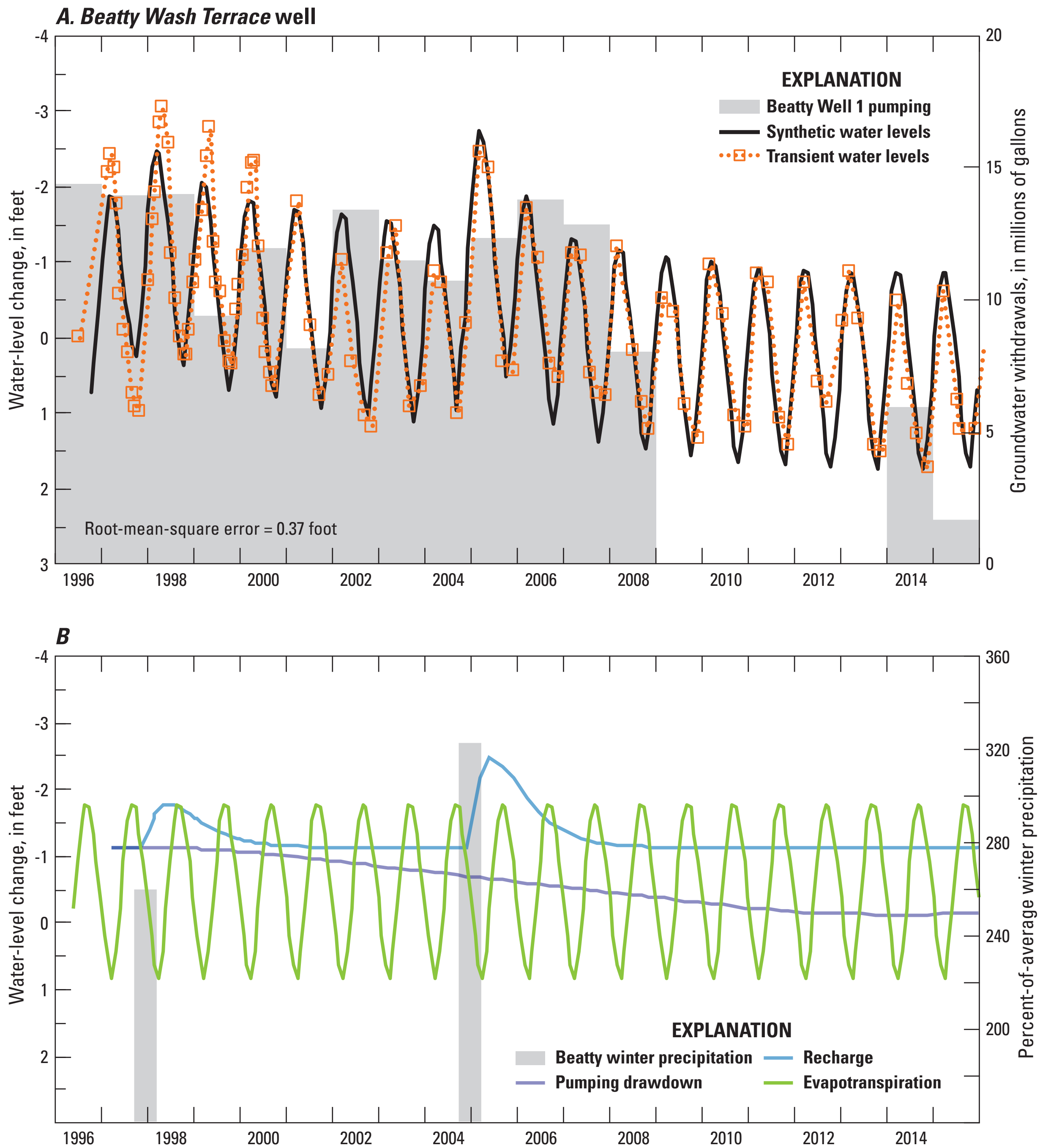


Figure 20. Water-level model results for well *Beatty Wash Terrace*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. (*A*) Comparison of transient and synthetic water levels and (*B*) components of synthetic curve.

a poor fit from 1991 to 1992. The poor fit occurs because measured water levels prior to pumping (1982–85) are about 15 ft higher than water levels in 2016, which are minimally affected by pumping. This suggests that another stress caused water levels in well *U-20 WW* and in the five nearby wells (fig. 22) to decline from 1985 to 2000.

Effects from a nearby nuclear detonation potentially caused most of the water-level decline between 1985 and 1996 in well *U-20 WW* and nearby wells. Measured water levels in wells *U-20 WW* and *U-20n PS 1DD-H* did not decline from July 1982 to September 1985 (fig. 22). However, water levels were declining by December 1986, indicating that the

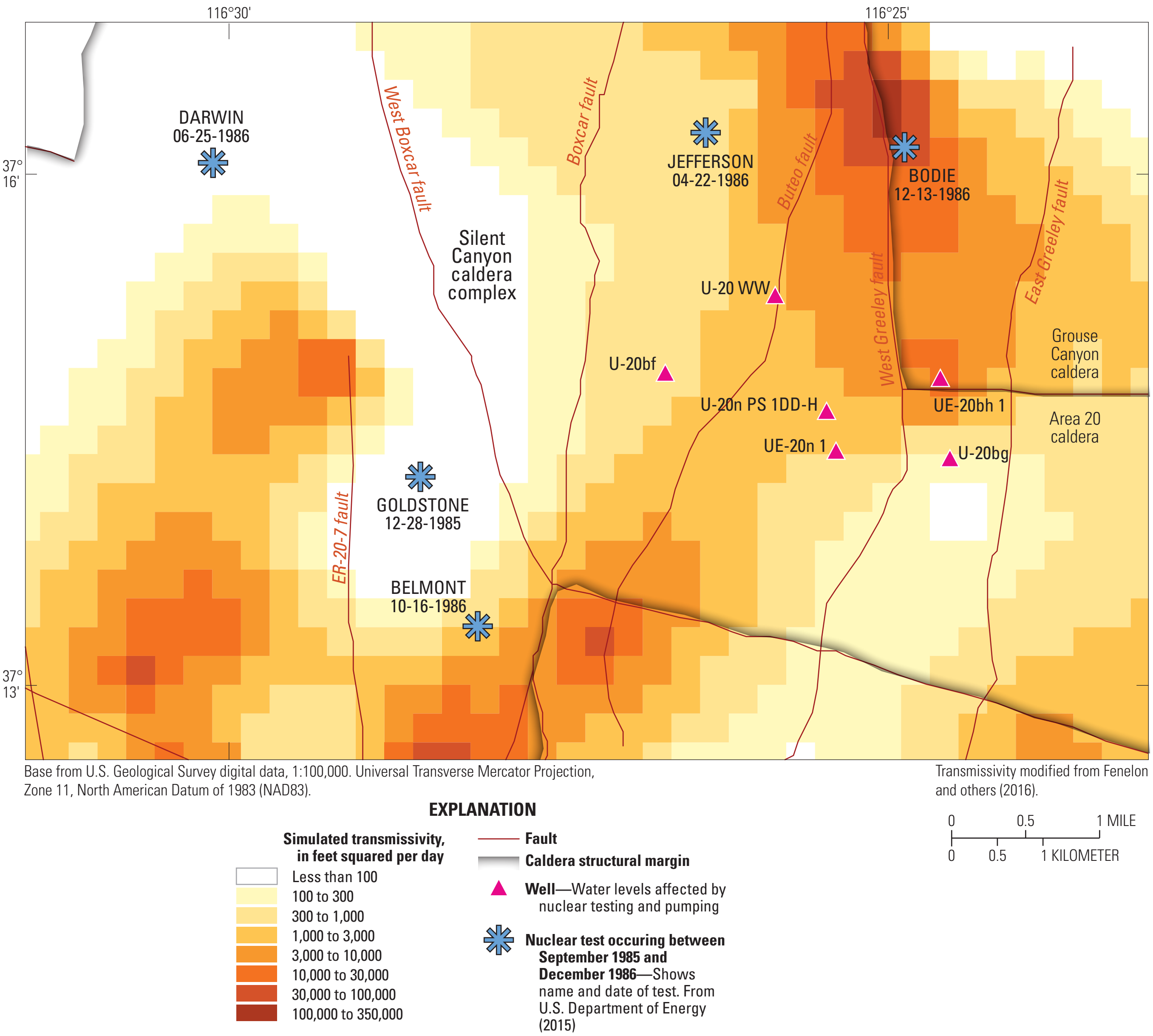


Figure 21. Location of wells, nuclear tests, geologic structures, and simulated transmissivity near well *U-20 WW*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

declining trend began between September 1985 and December 1986. Five nuclear tests were detonated during this period: GOLDSTONE (December 28, 1985), JEFFERSON (April 22, 1986), DARWIN (June 25, 1986), BELMONT (October 16, 1986), and BODIE (December 13, 1986) (fig. 21). These nuclear tests were within 5 mi of the center of affected wells, where burial depths ranged from 1,800 to 2,100 ft and announced yields were between 20 and 150 kilotons (U.S. Department of Energy, 2015).

Hydrogeology, hydraulic properties, and water levels provide evidence to support a vertical or lateral breach scenario as the cause of the declining trend. For example, a chimney created by one of the aforementioned nuclear tests may have breached a confining unit, causing a shallow and deep aquifer to become hydraulically connected (fig. 9). The vertical gradient is upward between shallow and deep units in borehole U-20a2 WW, located near well *UE-20n 1* (fig. 21). Water levels in shallow packer intervals from borehole U-20a2 WW were 20 to 30 ft lower than water levels in deeper packer intervals (Elliott and Fenelon, 2010).

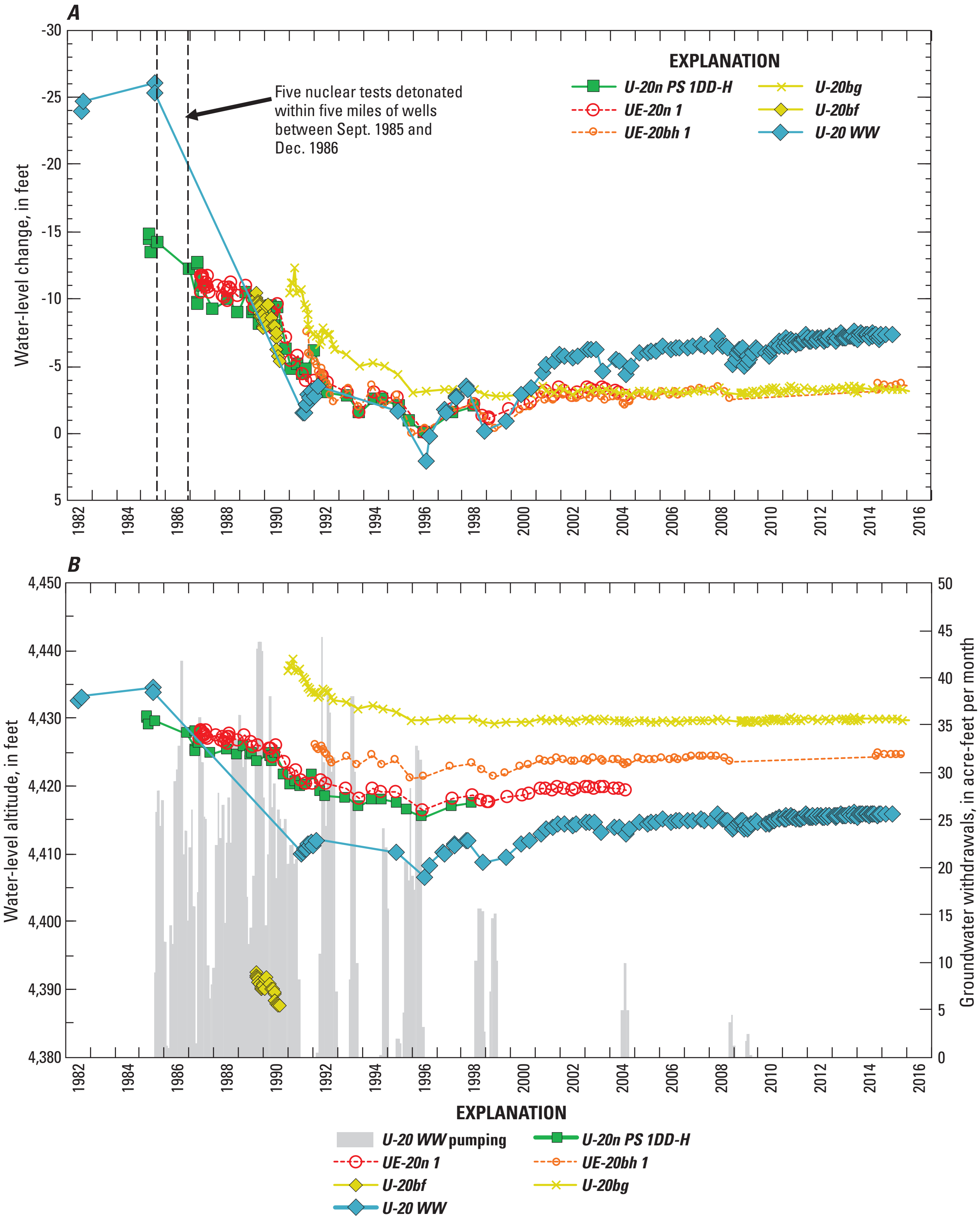


Figure 22. Comparison of (*A*) water-level change and (*B*) water-level altitudes in wells *U-20WW*, *UE-20bh 1*, *UE-20n 1*, *U-20n PS 1DD-H*, *U-20bg*, and *U-20bf*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

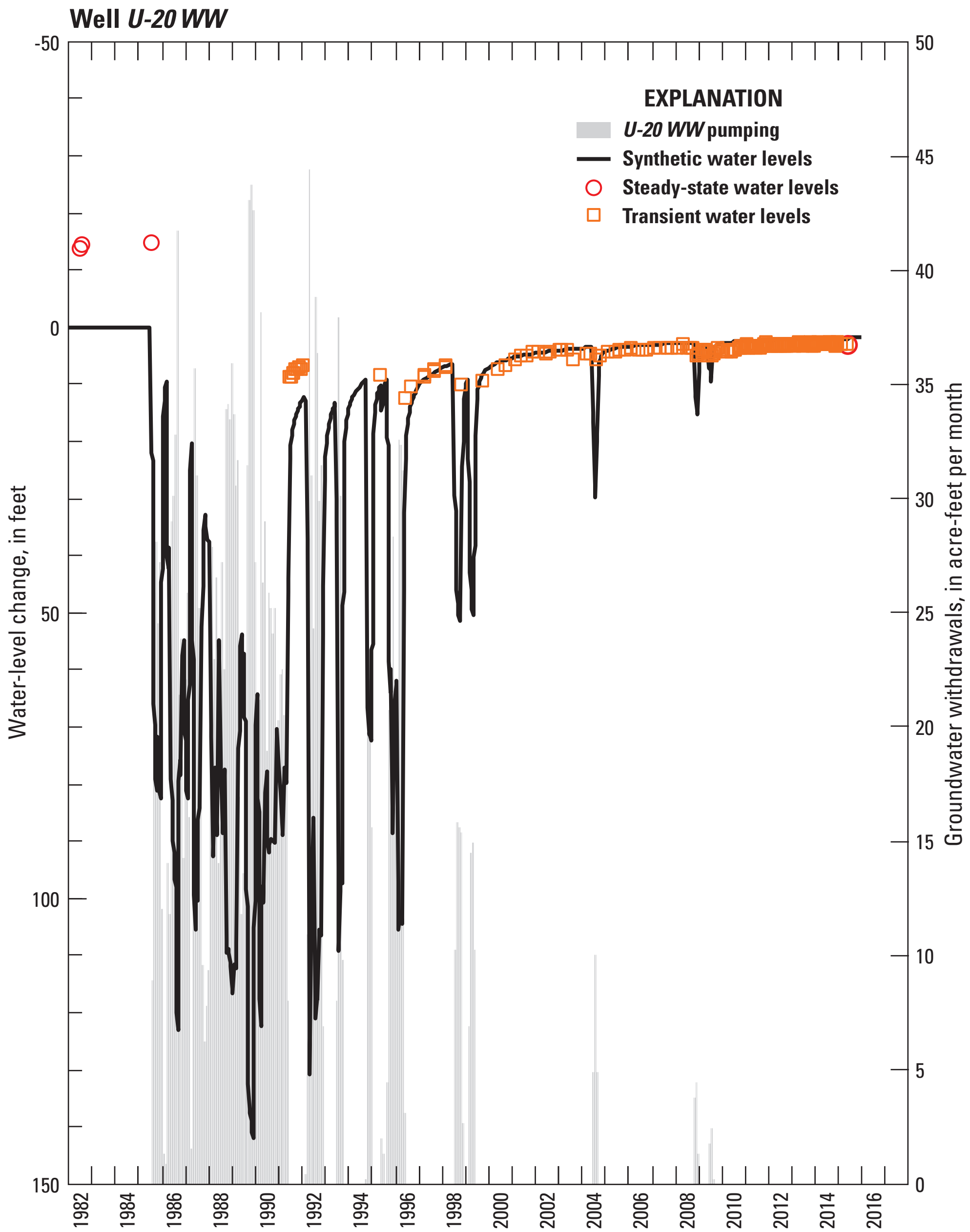


Figure 23. Comparison of measured and synthetic water levels in well *U-20 WW*, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. The synthetic curve simulates Theis transforms of pumping at well *U-20 WW* from 1985 to 2016.

A vertical breach in an area where there is a vertical head difference could provide a path for upward movement of groundwater from the deep to the shallow aquifer. After heads equilibrate to the breach, they would be permanently lowered and part of a new steady-state equilibrium. Alternatively, a nuclear test may have fractured low-permeability rocks in the area or induced offset along a pre-existing fault that forms a hydraulic barrier, causing a preferred pathway through the barrier. In this case, water levels that were elevated upgradient of the hydraulic barrier prior to nuclear testing would be permanently lowered after the test, similar to the water-level trends in well *U-20 WW*.

Whether the breach was vertical or lateral, water levels were permanently lowered in the area. This is consistent with recent, stable, water-level trends in well *U-20 WW* and nearby wells, which did not fully recover to pre-pumping conditions (fig. 22). The latest (2016) water-level measurements in these wells have approximately equilibrated to a new steady-state condition following permanent dewatering of a groundwater system near these wells.

Steady-State Trends

Water-level hydrographs from 62 wells are dominated by natural hydrologic stresses and are used in the steady-state trend analysis (fig. 3; table 2). Of the 62 wells, 35 are in the PMOV groundwater basin. The remaining 27 wells are within 15 mi of the basin boundary near Cactus Range to the northwest, Sarcobatus Flat to the west, Timber Mountain and Yucca Mountain to the south, and Rainier Mesa to the east. Wells outside the basin were used to support interpretations of steady-state trends within the basin.

Graphical, statistical, and numerical methods were used to analyze steady-state trends from 1995 to 2016. The period 1995 to 2016 was selected to provide a consistent record for analyzing steady-state trends. Graphical analyses were used to estimate the magnitude of water-level responses to variations in natural hydrologic stresses. Statistical analyses were used to determine whether steady-state water levels exhibit statistically significant trends. WLMs were used to differentiate natural stresses affecting trends and to demonstrate how winter precipitation thresholds can be used as a proxy for recharge.

Steady-state trends were grouped into geographic areas based on natural hydrologic stresses affecting water levels, such as recharge and evapotranspiration, and potential factors causing the trends. Potential factors considered include unsaturated zone thickness, transmissivity, and proximity of wells to recharge sources.

Graphical and Statistical Analysis

Graphical (LOWESS) and statistical (Mann-Kendall) analyses indicate that 43 of the 62 wells with steady-state trends in the study area have significant upward trends, 6 have downward trends, and 13 have no trend (table 5). For wells with upward trends, the maximum magnitude of smoothed water-level change ranged from 0.2 ft at well *Springdale* to 44.1 ft at well *U-12s*. For wells with downward trends, the maximum magnitude of smoothed water-level change ranged from 0.8 ft at well *ER-20-1* to 13.3 ft at the *Antelope Mine 2* well.

The spatial distribution in water-level trends from 1995 to 2016 is shown in fig. 24. Wells with significant upward trends occur throughout the study area. Wells with significant downward trends occur near the Cactus Range, in southern Oasis Valley, and in Pahute Mesa in NNSS Area 20. Wells with no significant trend occur in Gold Flat, Oasis Valley, near Rocket Wash, and along Fortymile Wash.

Maximum magnitudes of smoothed water-level change are greatest in high-altitude areas dominated by recharge and smallest in low-altitude areas dominated by groundwater discharge (table 5; fig. 24). Maximum magnitudes of smoothed water-level change are greater than 10 ft for wells in or near recharge areas, such as the Cactus Range, Rainier Mesa, Pahute Mesa in NNSS Area 19, and along Fortymile Wash. Most wells with a maximum magnitude of smoothed water-level change of less than 0.5 ft are in or near areas of groundwater discharge in Oasis Valley and Sarcobatus Flat.

Statistically significant upward and downward trends, as well as the absence of a trend, provide insight into the PMOV groundwater-flow system. Steady-state trends from 1995 to 2016 represent relatively short-term fluctuations within the longer period of steady state, as discussed in section, "Conceptual Model of Recharge and Discharge." Significant upward trends are the result of groundwater recharge exceeding aquifer discharge during the period of trend analysis. Significant downward trends typically occur because, even though water levels in these wells respond to recharge from wet winters (see section, "Trends by Geographic Area"), aquifer discharge exceeds recharge from 1995 to 2016. The absence of a water-level trend can indicate that recharge and discharge are in balance or that a well is screened in a low-permeability unit that is not well connected to the regional flow system.

Simulating Trends Using Precipitation Thresholds as a Proxy for Recharge

Water-level trends in four wells were simulated using WLMs to test the hypothesis that most steady-state trends in the PMOV basin can be explained by groundwater recharge. WLMs were generated to simulate trends in four wells that are representative of trends in Pahute Mesa, Rainier Mesa, Timber Mountain, and southern Oasis Valley. Water-level trends in each well were modeled using winter precipitation data from one of three precipitation indexes: Beatty, Pahute Mesa, or Rainier Mesa (table 3).

WLMs simulated water-level trends using transforms of recharge and discharge components. Gamma transforms simulate water-level responses to episodic recharge by transforming total winter precipitation above a specified threshold to water-level changes. Aquifer discharge is assumed constant and was simulated as a linear decline.

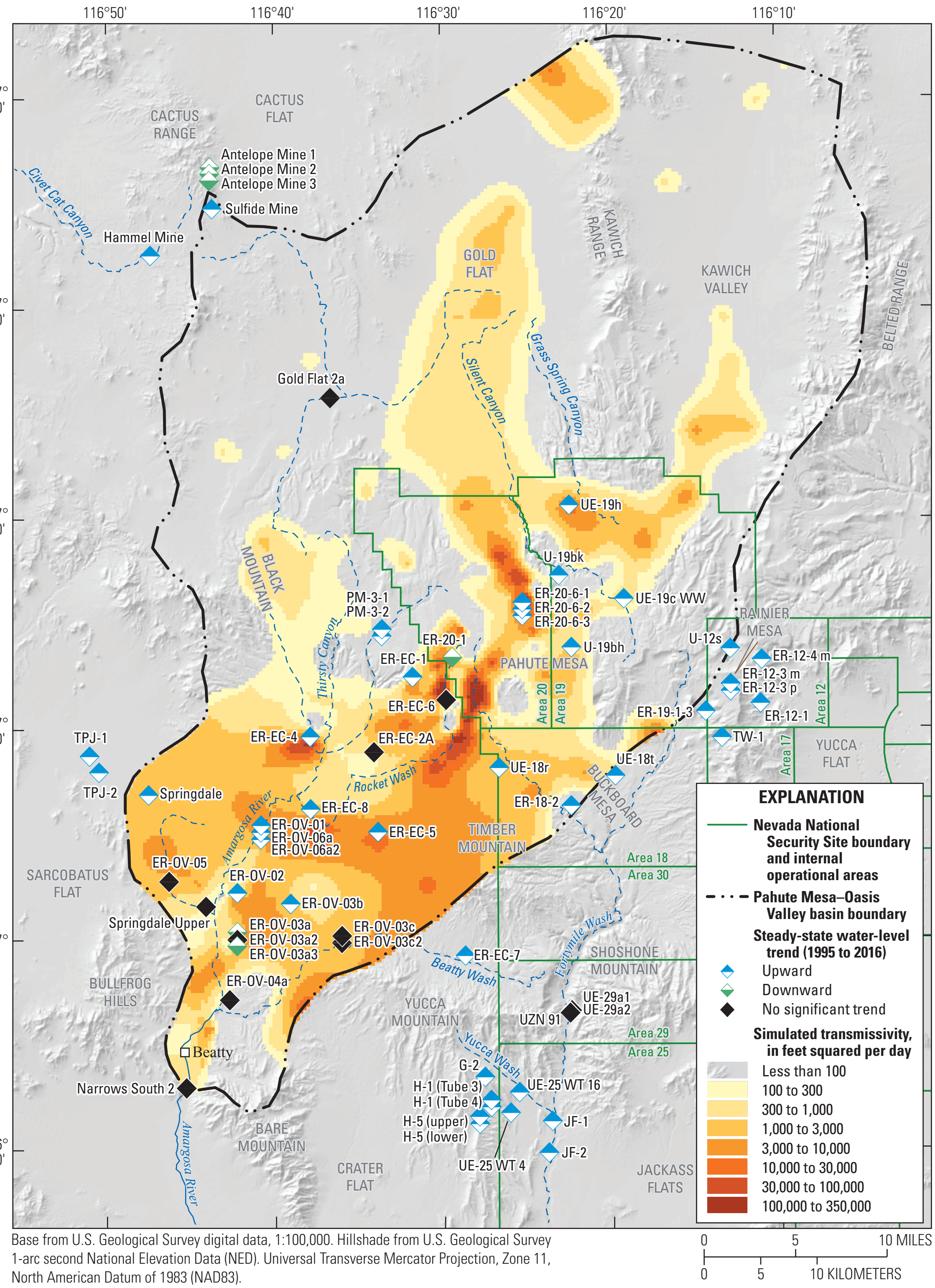


Figure 24. Statistical analysis results for steady-state water-level trends from 1995 to 2016 and simulated transmissivity in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Simulated transmissivity from Fenelon and others (2016).

Table 5. Analysis of steady-state water-level trends, using the Mann-Kendall test, for selected wells in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**Short name:** Reduced form of well name used in report text and figures for brevity. Full well name provided in table 2.
Period of record analyzed: Beginning and end year used to analyze water-level data for each well. Maximum period of record spans from January 1, 1995 to March 1, 2016. Wells with water-level data prior to January 1, 1995, were truncated.
Number of observations: Number of water-level measurements in the period of record analyzed.
Level of significance (p): Probability that water-level changes are due to chance rather than a trend; p-values less than (<) 0.001 are highly significant; p-values less than 0.01 are considered statistically significant.
Kendall's tau: A measure of the strength of the monotonic trend in water levels for the period of analysis. Kendall's tau values range between -1 and 1, where tau equal to 0 indicates no monotonic trend, tau equal to 1 indicates a strong rising trend, and tau equal to -1 indicates a strong declining trend.
Maximum change in smoothed water level: A measure of the absolute change in water level, in feet (ft), over the period of analysis. "Water-level change" is the maximum change, computed as the difference between the maximum and minimum water-level values on the smoothed (LOWESS) curve. "Time of maximum change" is the difference, in years, between the maximum and minimum water-level values on the smoothed curve.
Statistically significant trend: Considered significant if (1) the level of significance (p-value) is less than 0.01; (2) Kendall's tau is greater than 0.26; (3) the maximum change in smoothed water level is greater than or equal to 0.2 ft; and (4) the maximum change in smoothed water level occurs over a period of more than 7 years; up, water-level rising; down, water-level declining; none, no monotonic trend for the analyzed period of record]

Short name	Period of record analyzed	Number of observations	Level of significance (p)	Kendall's tau	Maximum change in smoothed water level		Statistically significant trend
					Water-level change	Time of maximum change	
Springdale	1997–2016	70	<0.001	0.73	0.2	16	Up
ER-12-1	1995–2015	86	<0.001	0.75	15.8	14	Up
ER-12-3 m	2005–2016	54	<0.001	0.67	8.6	8	Up
ER-12-3 p	2005–2016	41	<0.001	0.82	1.6	10	Up
ER-12-4 m	2006–2015	37	<0.001	0.87	4.7	10	Up
ER-18-2	2001–2016	60	<0.001	0.85	2.1	16	Up
ER-19-1-3	1995–2016	71	<0.001	0.88	4.1	21	Up
ER-20-1	1995–2015	79	<0.001	-0.43	0.8	17	Down
ER-20-6-1	1999–2015	102	<0.001	0.42	1.8	14	Up
ER-20-6-2	2000–2016	64	<0.001	0.54	1.2	11	Up
ER-20-6-3	1999–2016	57	<0.001	0.57	1.8	13	Up
ER-EC-1	1999–2015	95	<0.001	0.43	0.5	9	Up
ER-EC-2A	2000–2016	84	0.006	-0.20	0.2	14	None
ER-EC-4	2000–2014	64	<0.001	0.52	0.4	10	Up
ER-EC-5	1999–2016	76	<0.001	0.54	1.3	13	Up
ER-EC-6[1]	2000–2016	98	0.030	0.15	0.3	6	None
ER-EC-7	1999–2016	71	<0.001	0.61	1.5	12	Up
ER-EC-8	1999–2016	85	<0.001	0.73	0.8	15	Up
ER-OV-01	1997–2016	81	<0.001	0.74	0.6	17	Up
ER-OV-02	1997–2016	81	<0.001	0.43	0.4	12	Up
ER-OV-03a	1997–2016	81	<0.001	-0.82	4.9	18	Down
ER-OV-03a2	1997–2016	76	0.005	-0.22	0.5	3	None
ER-OV-03a3	1997–2016	80	<0.001	-0.82	5.0	18	Down
ER-OV-03b	1997–2016	77	<0.001	0.59	0.7	14	Up
ER-OV-03c	1997–2016	81	0.021	0.17	0.4	13	None
ER-OV-03c2	1997–2016	86	<0.001	0.25	0.5	14	None
ER-OV-04a	1997–2016	80	0.047	0.15	0.4	9	None
ER-OV-05	1997–2016	79	<0.001	0.35	0.2	3	None
ER-OV-06a	1997–2016	92	<0.001	0.72	0.6	15	Up
ER-OV-06a2	1997–2016	76	<0.001	0.63	0.4	18	Up
Gold Flat 2a	1995–2015	32	0.758	-0.04	0.5	3	None
Hammel Mine	1995–2014	28	<0.001	0.74	0.4	14	Up
Narrows South 2	1999–2009	53	0.141	-0.14	1.5	1	None
TPJ-2	1995–2015	57	<0.001	0.70	0.3	17	Up
PM-3-1	1995–2015	75	<0.001	0.86	2.2	20	Up
PM-3-2	1995–2015	133	<0.001	0.67	1.9	20	Up
Springdale Upper	1996–2016	120	0.128	-0.09	0.4	17	None

Table 5. Analysis of steady-state water-level trends, using the Mann-Kendall test, for selected wells in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.—Continued

Short name	Period of record analyzed	Number of observations	Level of significance (p)	Kendall's tau	Maximum change in smoothed water level		Statistically significant trend
					Water-level change	Time of maximum change	
Antelope Mine 1	1995–2015	41	<0.001	-0.45	10.0	10	Down
Antelope Mine 2	1995–2015	41	<0.001	-0.46	13.3	9	Down
Antelope Mine 3	1995–2015	41	<0.001	-0.45	13.2	10	Down
Sulfide Mine	1995–2015	41	<0.001	0.37	8.2	8	Up
TW-1	1995–2016	75	<0.001	0.95	4.1	21	Up
U-12s	1995–2015	60	<0.001	0.77	44.1	17	Up
U-19bh	1996–2015	64	<0.001	0.88	15.1	18	Up
U-19bk	1995–2016	79	<0.001	0.73	1.6	19	Up
UE-18r	1995–2013	50	<0.001	0.80	1.6	15	Up
UE-18t	1995-2016	61	<0.001	0.93	3.2	21	Up
UE-19c WW	1995–2015	67	<0.001	0.58	0.7	12	Up
UE-19h	1995–2015	82	<0.001	0.39	0.5	19	Up
UE-25 WT 4	1995–2016	64	<0.001	0.70	1.3	21	Up
JF-2	1995–2016	182	<0.001	0.76	1.6	18	Up
JF-1	1995–2016	229	<0.001	0.78	1.7	12	Up
UE-25 WT 16	1995–2016	84	<0.001	0.75	10.6	19	Up
UZN 91	1995–2016	150	<0.001	-0.26	17.5	< 1	None
UE-29a1	1995–2016	144	<0.001	-0.51	21.9	< 1	None
UE-29a2	1995–2016	152	<0.001	-0.30	15.0	< 1	None
TPJ-1	1995–2015	57	<0.001	0.39	0.3	14	Up
G-2	1998–2016	61	<0.001	0.79	4.2	17	Up
H-1 (Tube 3)	1995–2016	68	0.001	0.27	2.2	21	Up
H-1 (Tube 4)	1995–2016	82	<0.001	0.69	2.5	21	Up
H-5 (lower)	1995–2016	75	<0.001	0.73	2.3	21	Up
H-5 (upper)	1995–2016	77	<0.001	0.57	2.0	21	Up

[1]Well ER-EC-6 includes water-level data from USGS wells ER-EC-6 (1581–3820 ft) and ER-EC-6 shallow.

Pahute Mesa

The water-level trend in well *U-19bk* is representative of the recharge response in the Pahute Mesa area. Most Pahute Mesa wells have trends similar to well *U-19bk*, which has a long-term rising trend between 1995 and 2016 (fig. 25*A*). A WLM was generated using the Pahute Mesa precipitation-index record to determine whether the rising trend can be sustained by recharge from wet winters. Measured and synthetic water levels compare well when only episodic recharge and aquifer discharge are simulated (fig. 25*B*). WLM results indicate that the water-level trend in well *U-19bk*, and similar steady-state trends in the Pahute Mesa area, can be explained by recharge from the 1995, 1997, 1998, 2000, 2005, and 2011 winters. The long-term rising trend is an attenuated response to recharge that likely is attributed to a thick unsaturated zone in Pahute Mesa, which ranges between 900 and 2,400 ft.

Rainier Mesa

The water-level trend in well *ER-12-1* is representative of the recharge response in the Rainier Mesa area. Most Rainier Mesa wells have trends similar to well *ER-12-1*, which has observable recharge responses to the 1995, 2005, and 2011 winters. Recharge for the 1997–98 winters did not produce a strong response, but likely sustained the long rising trend that began in 1995 (fig. 26*A*). A WLM was generated using the Rainier Mesa precipitation-index record. Measured and synthetic water levels compare well when only episodic recharge and aquifer discharge are simulated (fig. 26*A*). This indicates that the water-level trend in well *ER-12-1*, and similar steady-state trends in the Rainier Mesa area, can be explained entirely by recharge from wet winters.

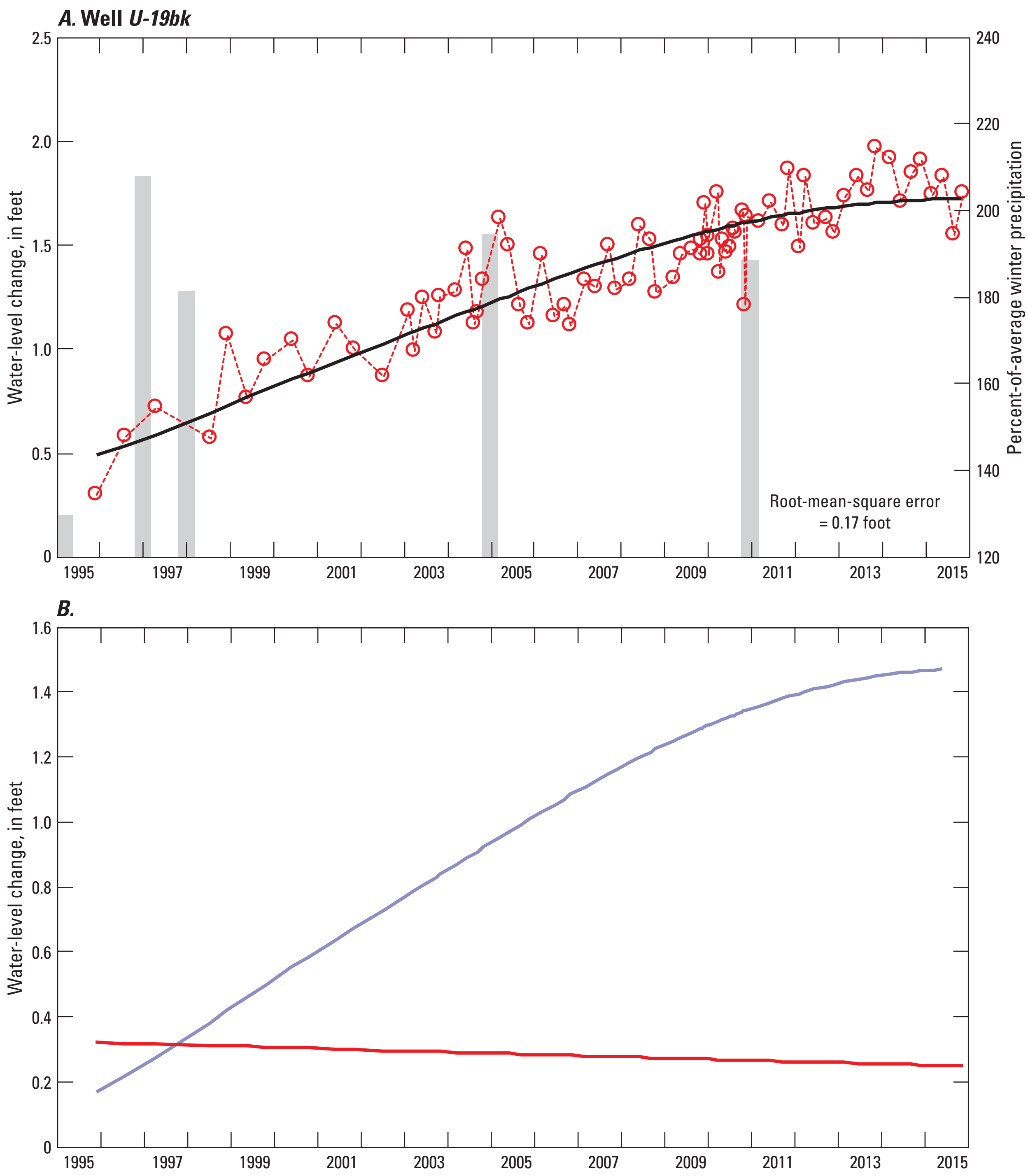


Figure 25. Water-level model results for well *U-19bk*, which has a water-level trend representative of the Pahute Mesa area, Nevada. (*A*) Comparison of steady-state and synthetic water levels and (*B*) components of synthetic curve.

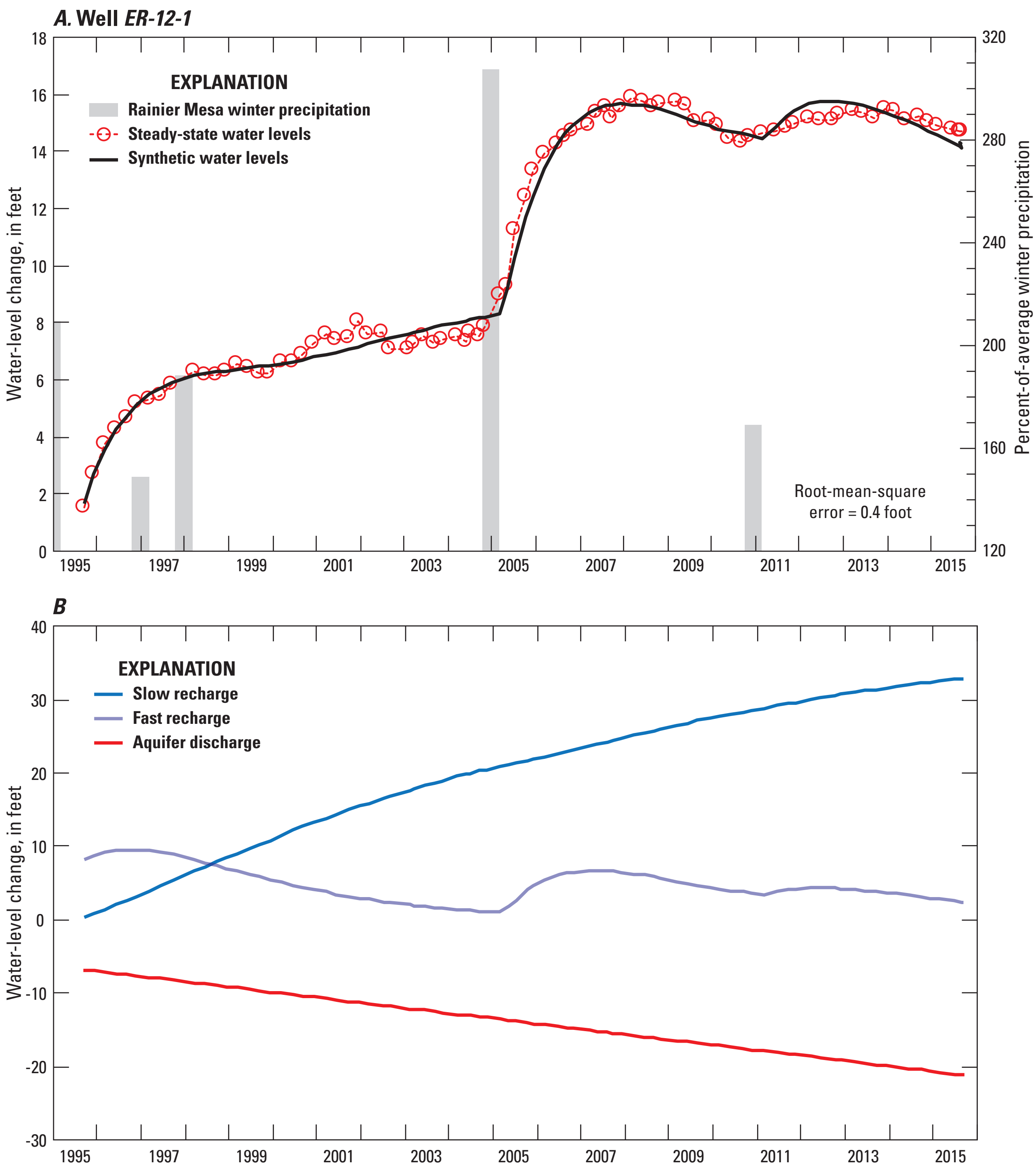


Figure 26. Water-level model results for well *ER-12-1*, which has a water-level trend representative of the Rainier Mesa area, Nevada. (*A*) Comparison of steady-state and synthetic water levels and (*B*) components of synthetic curve.

Two Gamma transforms were used to simulate water-level responses to episodic recharge in well *ER-12-1* (fig. 26*B*). The two transforms represent fast and slow recharge pathways through a thick (1,524 ft) unsaturated zone consisting of carbonate rock and siltstone. Preferential flow paths through hydraulically connected fractures in the unsaturated zone likely cause fast responses to recharge events. Slower flow through disconnected fracture networks likely cause a prolonged response to recharge that lasts many years after each wet winter.

Timber Mountain

The water-level trend in well *ER-EC-5* is representative of recharge near Timber Mountain. The well has an observable recharge response to the 2005 winter; however, a recharge response to the 1998 winter was uncertain because of a lack of data (fig. 27). WLMs were generated using recharge estimated from the Beatty precipitation index that either included or excluded recharge during the 1998 winter. A better fit was obtained between measured and synthetic water levels when the 1998 winter is included. WLM results show that the trend in well *ER-EC-5*, where the unsaturated zone is about 1,000 ft thick, can be attributed to simple recharge and discharge components. Likewise, similar trends in other wells near Timber Mountain, such as well *ER-EC-7*, also can be attributed to fluctuations in recharge.

Southern Oasis Valley

The water-level trend in well *Springdale Upper* is representative of recharge in southern Oasis Valley, where the unsaturated zone is less than 100 ft thick. The well has observable recharge responses to the 1998 and 2005 winters (fig. 28*A*). A WLM was generated using recharge estimated from the Beatty precipitation index. Measured and synthetic water levels compare well when episodic recharge and evapotranspiration are simulated (fig. 28*B*). WLM results show that the trend in well *Springdale Upper*, and other wells with similar trends, such as wells *ER-OV-04a* and *ER-OV-05*, can be explained by recharge and evapotranspiration.

Trends by Geographic Area

Wells with steady-state water-level trends were categorized into eleven geographic areas (fig. 29) based on water-level responses to episodic recharge:

1. Rainier Mesa;
2. Cactus Range;
3. Eastern Pahute Mesa and Buckboard Mesa;
4. Rocket Wash and Thirsty Canyon;
5. Gold Flat and central PMOV basin;
6. Timber Mountain and Jackass Flats;
7. Northern Oasis Valley;
8. Southern Oasis Valley;
9. Western Oasis Valley and Sarcobatus Flat;
10. Fortymile Wash and Amargosa Narrows; and
11. Yucca Mountain.

Water-level trends in geographic areas were categorized according to years that water levels responded to recharge (winters of 1995, 1997, 1998, 2000, 2001, 2005, 2010, and 2011). For each well, a water-level response to each wet winter was qualitatively flagged as either “strong,” “strong?,” “weak,” “weak?,” “no,” “no data,” or “rise.” A “strong” response is defined as a sudden abrupt change in water levels following a wet winter. A “weak” response is defined as a slow gradual change in water levels following a wet winter. A “no” response indicates that a wet winter had no observable effect on water levels. A “no data” flag indicates that no steady-state data are present in the water-level record to determine whether a recharge response occurred. A “rise” flag is used to indicate that a wet winter may have contributed to all or part of a long-term water-level rise in the water-level record. The queried “strong?” and “weak?” flags indicate uncertain water-level responses. These flags are subjective, where the year prior to and after the winter season are compared visually to determine if water levels responded to recharge. A qualitative flagging procedure was used instead of rigorous statistical approaches because sparse quarterly water-level data preclude the use of statistics and the superposition of recharge on seasonal evapotranspiration is a confounding factor for some wells. Recharge responses to winters for each well hydrograph are provided in table 6.

In each geographic area, water levels respond similarly to episodic recharge. Two different geographic areas can have similar recharge responses, but different overall water-level trends. For example, water levels in southern Oasis Valley and the Timber Mountain area respond to the 1998 and 2005 winters (table 6), but the recharge response in southern Oasis Valley is superimposed on a groundwater evapotranspiration response.

Factors that can influence water-level responses to recharge were compiled and used to explain water-level trends within geographic areas. Potential factors include unsaturated zone depth, hydrogeologic unit(s) screened in the open interval of a well, transmissivity, and distance to potential recharge areas. Average unsaturated zone thicknesses and primary water-bearing hydrogeologic units for the 62 wells are provided in table 7. Transmissivity was estimated from analytical or numerical solutions for 31 of the 62 wells (table 7). For wells with no field-measured transmissivity estimate, a qualitative estimate of transmissivity was assigned based on rock type at the well screen, well purpose, and water-level responses to various stresses. Transmissivity estimates were supplemented with a simulated transmissivity map from Fenelon and others (2016) to visualize the spatial distribution of low and high transmissivity (fig. 24).

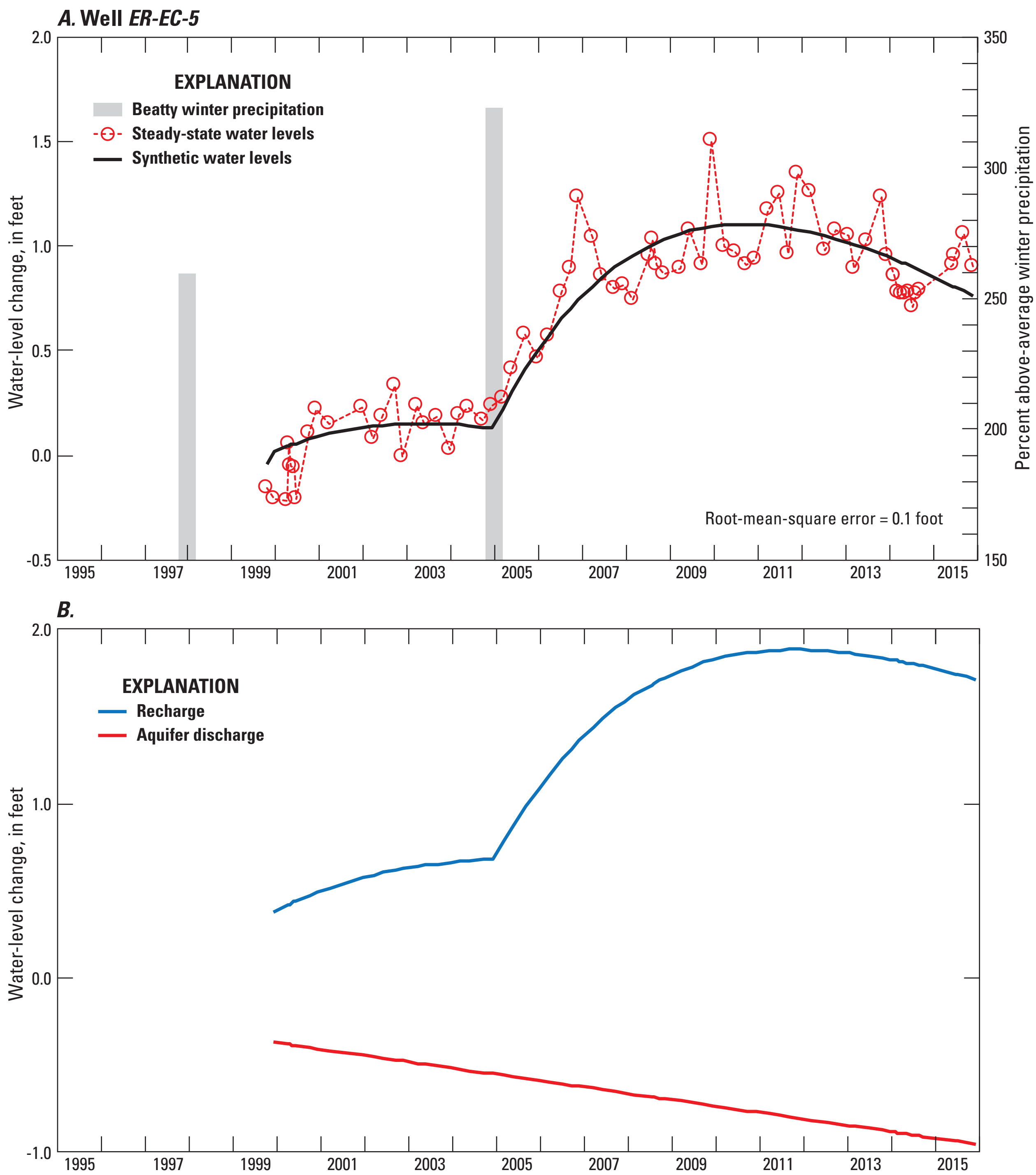


Figure 27. Water-level model results for well *ER-EC-5*, which has a water-level trend representative of the Timber Mountain area, Nevada. (*A*) Comparison of steady-state and synthetic water levels and (*B*) components of synthetic curve.

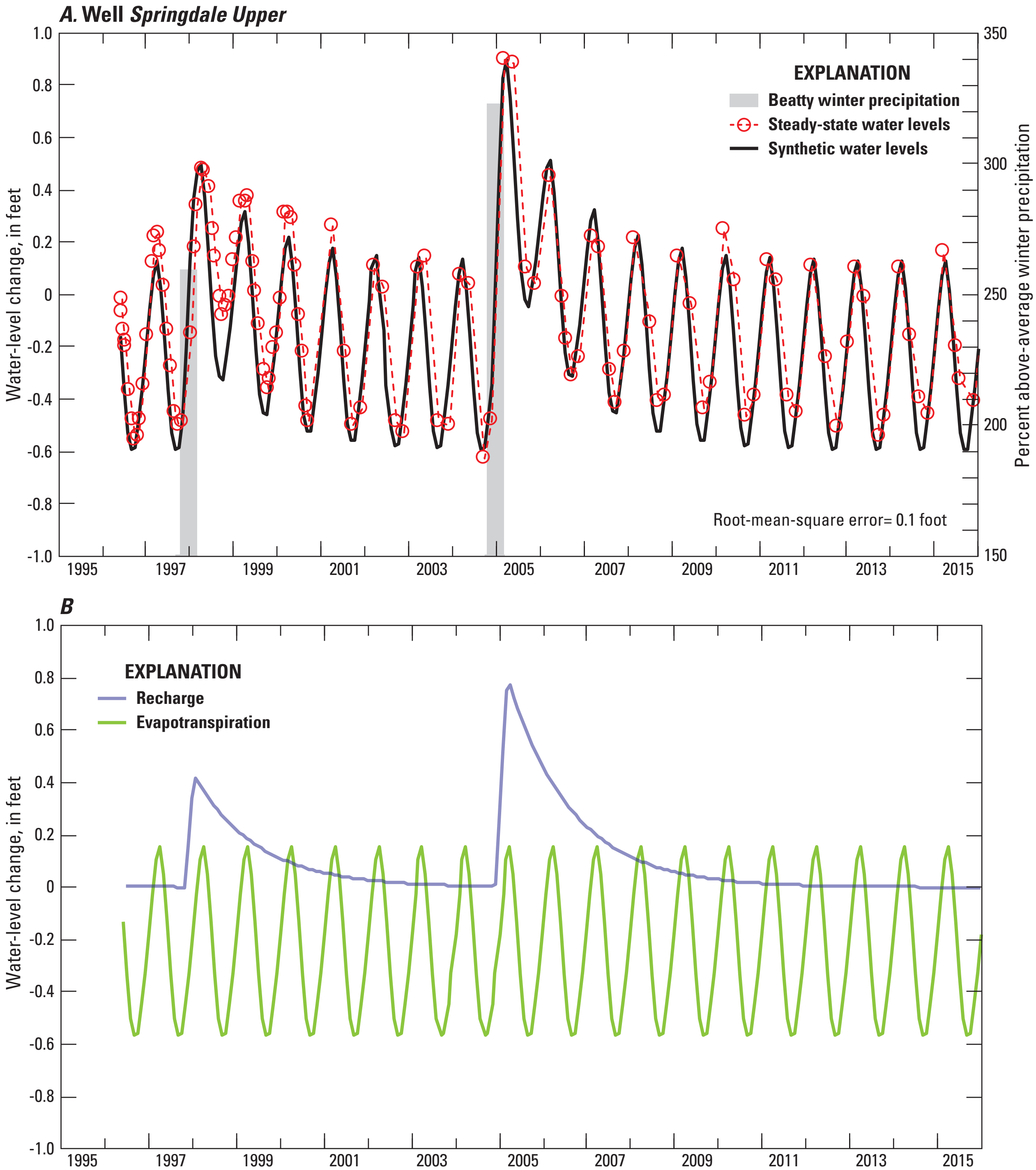


Figure 28. Water-level model results for well *Springdale Upper*, which has a water-level trend representative of southern Oasis Valley area, Nevada. (*A*) Comparison of steady-state and synthetic water levels and (*B*) components of synthetic curve.

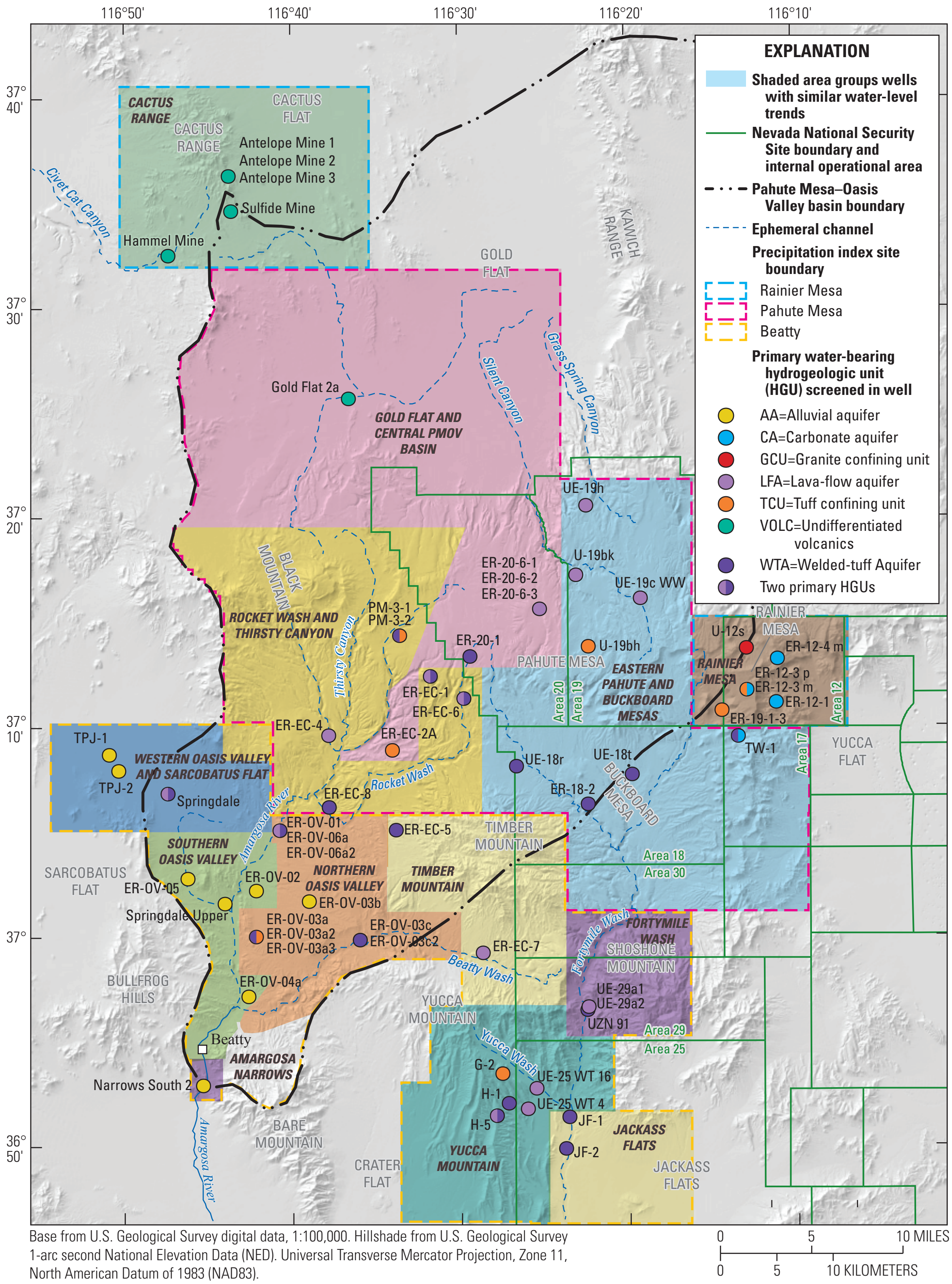


Figure 29. Geographic areas and extents used to group wells with similar steady-state trends, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

Table 6. Summary of qualitative analysis for wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**Geographic area:** Geographic areas used in the categorization of wells. Wells were categorized according to years that steady-state water levels responded to recharge, and to factors causing the shape of the water-level trend. PMOV, Pahute Mesa-Oasis Valley.
Precipitation index: Precipitation index that best characterizes recharge patterns in the geographic area, and was based on water-level responses to recharge from wetter-than-average winters.
Threshold (percent): Precipitation threshold, expressed in percent of the long-term mean that was used to determine wetter-than-average winters that contributed to recharge in a geographic area. A "High" threshold was used for wells where the unsaturated zone is thick or the well is distant from recharge areas. A "Low" threshold was used for wells with shallow unsaturated zone depths that are in close proximity to recharge areas or ephemeral channels.
Threshold explanation: Explanation of factors causing water-level trends and was used to determine the threshold applied to a precipitation index.
Short name: Reduced form of well name used in report text and figures for brevity. Well names provided in table 2. Well locations are shown in figure 3.
Years water levels responded to recharge: Water-level responses to winters of years shown. Each winter was flagged as either: "strong," "strong?," "weak," "weak?," "no," "no data." or "rise." A "strong" response is a sudden abrupt change in water levels following a wet winter. A "weak" response is a slow gradual change in water levels following a wet winter. A "no" response indicates that a wet winter had no observable effect on water levels. A "rise" flag is used to indicate that a wet winter may have contributed to all or part of a long-term water-level rise in the water-level record. The "strong?" and "weak?" flags indicate uncertain water-level responses. A "***No***" or "***No data***" flag indicates water-level data do or do not exist, respectively, for a winter that was not wet for a specified precipitation index threshold.
Water-level trend description: Description of water-level trends grouped within a geographic area. Some geographic areas have more than one type of water-level trend because of the behavior of the aquifer system to recharge]

Geographic area	Precipitation index	Threshold (percent)	Threshold explanation	Short name	Years water levels responded to recharge								Water-level trend description
					1995	1997	1998	2000	2001	2005	2010	2011	
Rainier Mesa	Rainier Mesa area	High (130)	Thick unsaturated zone	ER-12-1	Strong	Rise	Rise	***No***	***No***	Strong	***No***	Weak	Strong response to 2005 and weak response to 2011 winter
				ER-12-3 m	No data	No data	No data	***No***	***No***	Strong	***No***	Weak	
				ER-12-4 m	No data	No data	No data	***No***	***No***	Strong	***No***	Weak	
				ER-12-3 p	No data	No data	No Data	***No***	***No***	Rise	***No***	Rise	Continuous rising trend
				ER-19-1-3	No data	No data	Rise	***No***	***No***	Weak	***No***	Rise	
				U-12s	Strong	Rise	Rise	***No***	***No***	Strong	***No***	No	No response to 2011
Cactus Range[1]	Rainier Mesa area	Low (115)	Shallow water table and close proximity to recharge source	Hammel Mine	No data	No[2]	Rise	***No***	Rise	Rise	Rise	No[2]	Continuous rising trend
				Antelope Mine 1	Strong	No[2]	Strong	***No***	Strong	Strong	Strong	No[2]	Strong responses to most wetter-than-average winters
				Antelope Mine 2	Strong	No[2]	Strong	***No***	Strong	Strong	Strong	No[2]	
				Antelope Mine 3	Strong	No[2]	Strong	***No***	Strong	Strong	Strong	No[2]	
				Sulfide Mine	Strong	No[2]	Strong	***No***	Strong	Strong	Weak	No[2]	
Eastern Pahute Mesa and Buckboard Mesa	Pahute Mesa area	High (120)	Thick unsaturated zone	ER-18-2	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	Continuous rising trends
				TW-1	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				U-19bh	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				U-19bk	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				UE-18r	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				UE-18t	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				UE-19c WW	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
				UE-19h	Rise	Rise	Rise	Rise	***No***	Rise	***No***	Rise	
Rocket Wash and Thirsty Canyon	Pahute Mesa area	High (120)	Thick unsaturated zone	ER-EC-1	No data	No data	No data	No data	***No***	Weak	***No***	No	Mostly weak and attenuated recharge responses; likely attributed to proximity to ephemeral channels
				ER-EC-4	No data	No data	No data	No data	***No***	Weak	***No***	No	
				ER-EC-6	No data	No data	No data	No data	***No***	Weak	***No***	No	
				ER-EC-8	No data	No data	No data	Weak?	***No***	Weak	***No***	No	
				PM-3-1	Strong	Rise	Rise	Rise	***No***	Weak	***No***	No	
				PM-3-2	Strong	Rise	Rise	Rise	***No***	Weak	***No***	No	

Table 6. Summary of qualitative analysis for wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.—Continued

Geographic area	Precipitation index	Threshold (percent)	Threshold explanation	Short name	Years water levels responded to recharge: 1995	1997	1998	2000	2001	2005	2010	2011	Water-level trend description
Gold Flat and Central PMOV basin	Pahute Mesa area	High (120)	Thick unsaturated zone	ER-20-1	No	No	No	No	***No***	No	***No***	No	Water levels have anomalous responses not attributed to recharge
				ER-20-6-1	No data	No data	No data	No	***No***	No	***No***	No	
				ER-20-6-2	No data	No data	No data	No	***No***	No	***No***	No	
				ER-20-6-3	No data	No data	No data	No	***No***	No	***No***	No	
				ER-EC-2A	No data	No data	No data	No	***No***	No	***No***	No	
				Gold Flat 2a	No data	No data	No data	No	***No***	No	***No***	No	
Timber Mountain and Jackass Flats	Beatty, Nevada area	High (160)	Thick unsaturated zone	ER-EC-5	***No data***	***No data***	No data	***No***	***No***	Strong	***No***	***No***	Strong responses to wettest winters
				ER-EC-7	***No data***	***No data***	No data	***No***	***No***	Strong	***No***	***No***	
				JF-1	***No***	***No***	Strong	***No***	***No***	Weak	***No***	***No***	
				JF-2	***No***	***No***	Strong	***No***	***No***	Strong	***No***	***No***	
Northern Oasis Valley	Beatty, Nevada area	Low (125)	Shallow water table and close proximity to ephemeral channels	ER-OV-01	No data	No data	No	Weak	Rise	Weak	Weak	***No***	Weak responses to most wetter-than-average winters
				ER-OV-03b	No data	No data	Weak?	Weak	Rise	Weak	Weak	***No***	
				ER-OV-03c	No data	No data	Weak?	Weak	Rise	Weak	Weak	***No***	
				ER-OV-03c2	No data	No data	Weak?	Weak	Rise	Weak	Weak	***No***	
				ER-OV-06a	No data	No data	No	Weak	Rise	Weak	Weak	***No***	
				ER-OV-06a2	No data	No data	No	Weak	Rise	Weak	Weak	***No***	
				ER-OV-03a2	No data	No data	No data	No	Weak?	Strong	Weak	***No***	Strong barometric pressure fluctuations
				ER-OV-03a	No data	No data	No data	No	Weak?	Weak	Weak	***No***	Anomalous declining trend
				ER-OV-03a3	No data	No data	No data	No	Weak?	Weak	Weak	***No***	
Southern Oasis Valley	Beatty, Nevada area	High (160)	Recharge only occurs from large floods that reach groundwater discharge area	ER-OV-02	***No data***	***No data***	No data	***No***	***No***	Strong	***No***	***No***	Strong responses to wettest winters superimposed on evapotranspiration
				ER-OV-04a	***No data***	***No data***	No data	***No***	***No***	Strong	***No***	***No***	
				ER-OV-05	***No data***	***No data***	Strong?	***No***	***No***	Strong	***No***	***No***	
				Springdale Upper	***No data***	***No***	Strong	***No***	***No***	Strong	***No***	***No***	
Western Oasis Valley and Sarcobatus Flat	Beatty, Nevada area	High (160)	Lack of proximity to recharge sources	Springdale	***No data***	***No data***	Rise	***No***	***No***	Rise	***No***	***No***	Weak responses to wettest winters
				TPJ-1	***No***	***No***	Weak	***No***	***No***	Weak	***No***	***No***	
				TPJ-2	***No***	***No***	Weak	***No***	***No***	Weak	***No***	***No***	

Table 6. Summary of qualitative analysis for wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.—Continued

Geographic area	Precipitation index	Threshold (percent)	Threshold explanation	Short name	Years water levels responded to recharge								Water-level trend description
					1995	1997	1998	2000	2001	2005	2010	2011	
Fortymile Wash and Amargosa Narrows	Beatty, Nevada area	Low (125)	Shallow water table and close proximity to ephemeral channels	UE-29a1	Strong	No[3]	Strong	Strong	Weak	Strong	Strong?	***No***	Strong recharge responses to most wetter-than average winters
				UE-29a2	Strong	No[3]	Strong	Strong	Weak	Strong	Strong?	***No***	
				UZN 91	Strong	No[3]	Strong	Strong	Weak	Strong	Strong?	***No***	
				Narrows South 2	No data	No data	No data	Strong	Strong	Strong	No data	***No***	
Yucca Mountain	Beatty, Nevada area	Low (125)	Beatty precipitation index is not a perfect proxy for recharge. Low threshold needed because water levels respond to 2000 and 2001 winters	G-2	No data	No data	No data	Strong	Rise	Strong	Rise	***No***	Discernible recharge response to 2000 and (or) 2001 winter
				UE-25 WT 16	No data	No data	Strong?	Strong	Strong	Strong	Strong?	***No***	
				UE-25 WT 4	No data	No data	No data	Rise	Rise	Rise	Rise	***No***	Continuous rising trends from 2000 to 2015 with no discernible responses to a particular wet winter
				H-1 (Tube 3)	Strong	No[3]	Strong	Rise	Rise	Rise	Rise	***No***	
				H-1 (Tube 4)	Strong	No[3]	Strong	Rise	Rise	Rise	Rise	***No***	
				H-5 (lower)	Strong	No[3]	Weak	Rise	Rise	Rise	Rise	***No***	
				H-5 (upper)	Strong	No[3]	Weak	Rise	Rise	Rise	Rise	***No***	

[1]Rainier Mesa precipitation-index record best characterizes precipitation patterns and recharge responses in Cactus Range because these localities have similar altitudes of more than 7,000 ft.

[2]No water-level record showed a recharge response to either 1997 or 2011 winters, probably because the Rainier Mesa precipitation-index record is not a perfect proxy for precipitation patterns in the Cactus Range.

[3]Fortymile Wash and Yucca Mountain wells did not respond to the winter of 1997, probably because Beatty precipitation is not a perfect proxy for precipitation patterns in these areas. Precipitation data near either Shoshone Mountain or Yucca Mountain may provide a better approximation of precipitation patterns; however, no active long-term precipitation stations exist in these areas.

A generalized recharge map from Fenelon and others (2010, pl. 3) was used to discern areas of likely groundwater recharge in the study area (fig. 3).

Rainier Mesa

At Rainier Mesa, water-level responses to recharge are influenced by the types of aquifers screened in wells (fig. 30; tables 6 and 7). Well *U-12s*, screened in a granite confining unit (GCU), has a strong response to 1995 and 2005 recharge, but does not appear to respond to recharge from the 2011 winter. The response to the 2011 winter is ambiguous for wells *ER-12-3 p* and *ER-19-1-3*, screened in low-transmissivity tuff confining units (TCUs) (table 7); however, *ER-19-1-3* shows a weak response to the 2005 winter. Wells *ER-12-1*, *ER-12-3 m*, and *ER-12-4 m*, screened in a carbonate-rock aquifer (CA), have strong responses to recharge from the 2005 winter and weak responses to recharge from the 2011 winter. Well *ER-12-1* also has a strong response to recharge from the 1995 winter, whereas the 1997–98 winters probably are contributing to the continuous rise between 1997 and 2005. Most recharge responses in Rainier Mesa wells occurred in 1995 and 2005 because these years had the most total winter precipitation between 1995 and 2016 (fig. 5*A*).

Water-level responses to recharge are attenuated in wells located farther downgradient from the recharge source. Water levels in well *ER-12-1* have the fastest and largest responses to recharge because the well is drilled in the carbonate-rock outcrop, where most recharge is conceptualized to occur. Water levels in well *ER-12-4 m* have the most attenuated response to recharge in the carbonate-rock aquifer because the well is farthest downgradient of the recharge source and the well is open to low-transmissivity (9 ft^2/d) carbonate rocks.

Cactus Range

Five wells near the Cactus Range (fig. 29) have water-level responses to episodic recharge from five wetter-than-average winters (table 6). Recharge responses in wells grade from strong near the recharge area in the Cactus Range to attenuated farther downgradient (fig. 3; fig. 31). Wells *Antelope Mine 1*, *Antelope Mine 2*, *Antelope Mine 3*, and *Sulfide Mine* responded strongly to recharge from the 1995, 1998, 2001, 2005, and 2010 winters, although the responses in 2010 were less amplified in well *Sulfide Mine*. Water levels in the *Hammel Mine* well show a small continuous rise that can be attributed to recharge from one or more of the wet winters.

Water-level responses to recharge are attributed to distance from the potential recharge area in the Cactus Range and unsaturated zone depth. Water levels in the *Antelope Mine* wells and *Sulfide Mine* well responded strongly to recharge events because these wells are shallow, with depths to water of less than 50 ft, and are closer to the potential recharge area in the Cactus Range compared to *Hammel Mine* well. The small water-level rise in *Hammel Mine* well may be partly attributed to the thicker, 119-ft, unsaturated zone at the well, but distance from the recharge area likely is the primary reason for the attenuated water-level response.

Eastern Pahute Mesa and Buckboard Mesa

Water levels in wells in eastern Pahute Mesa and Buckboard Mesa (fig. 29) have attenuated responses to episodic recharge (fig. 32; table 6). Water levels in wells *TW-1, U-19bh*, *U-19bk*, *UE-19c WW*, *UE-19h*, *ER-18-2*, *UE-18r*, and *UE-18t* have long-term rising trends between 1995 and 2016. These rising trends are attributed to recharge from the 1995, 1997, 1998, 2000, 2005, and 2011 winters, in addition to recharge from winters prior to 1995. Attenuated responses to recharge likely are attributed to a thick unsaturated zone, which ranges between 900 and 2,400 ft (table 7).

Rocket Wash and Thirsty Canyon

Water levels in wells adjacent to the Rocket Wash and Thirsty Canyon drainages have consistent responses to episodic recharge (fig. 29; table 6). Wells *ER-EC-1*, *ER-EC-4*, *ER-EC-6*, *ER-EC-8*, *PM-3-1*, and *PM-3-2* have weak responses to recharge during the 2005 winter (fig. 33). Wells in the PM-3 well cluster have a strong response to recharge from the 1995 winter, and a continuous water-level rise between 1998 and 2005 that likely is attributed to recharge from the 1997, 1998, and 2000 winters. Water-level responses to recharge likely are controlled by infiltration of precipitation into ephemeral channels in the Rocket Wash and Thirsty Canyon drainages (fig. 29).

Gold Flat and Central Pahute Mesa–Oasis Valley Basin

Water levels in wells in Gold Flat and the central part of the PMOV basin have no discernible responses to recharge (figs. 29 and 34). Water-level fluctuations in wells *ER-20-1*, *ER-EC-2A*, and *Gold Flat 2A* are anomalous and the cause is unknown. Water levels in well *Gold Flat 2A* may not respond to wet winters because the well is far from recharge areas. Wells *ER-20-1* and *ER-EC-2A* are near Rocket Wash, but water levels may not respond to recharge because the wells are near low-transmissivity features (fig. 24) that limit recharge in these areas.

Table 7. Hydrogeologic and hydraulic characteristics of wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.

[**Geographic area:** Geographic areas used in the categorization of wells. Wells were categorized according to years that steady-state water levels responded to recharge, and to factors causing the shape of the water-level trend.
Short name: Reduced form of well name used in report text and figures for brevity. USGS well names provided in table 2. Locations of wells are shown in figure 3.
Average unsaturated zone thickness: Computed from the difference in land surface altitude and mean water-level altitude from 1995 to 2016, referenced to feet above National Geodetic Vertical Datum of 1929.
Primary hydrogeologic unit(s): Primary water-bearing hydrogeologic unit(s) within the open interval that likely are contributing water to the well: AA, alluvial aquifer; CA, carbonate aquifer; GCU, granite confining unit; LFA, lava-flow aquifer; TCU, tuff confining unit; VOLC, volcanic undifferentiated; WTA, welded-tuff aquifer.
Transmissivity: Estimated transmissivity, in feet squared per day (ft^2/d), within the open interval of each well as determined from either an analytical or numerical solution. Qualitative estimates of permeability at the open interval were recorded for wells with no transmissivity estimates based on rock type at the well screen, and water-level responses to recharge, evapotranspiration, pumping, and well development. Qualitative estimates: Low, low transmissivity (less than 1 ft^2/d); Low?, transmissivity assumed to be low, but uncertain; Mod, moderate transmissivity (between 1 and 1,000 ft^2/d); Mod?, transmissivity assumed to be moderate, but uncertain; High, high transmissivity (greater than 1,000 ft^2/d); High?, transmissivity assumed to be high, but uncertain; <, less than.
Testing method: Field method used to collect water-level data for estimating transmissivity. Method: CRT, constant-rate test; MWAT, multiple-well aquifer test; SC, specific capacity; Slug, slug and recovery test.
Analysis method: Analytical or numerical method used to estimate transmissivity. Method: BF, borehole flowmeter; BR, Bouwer and Rice (1976); CJ, Cooper and Jacob (1946); Halford, Halford and others (2012); MOD, numerical model; Moench, Moench (1984); Neuman, Neuman (1974); SC, specific capacity; Theis, Theis (1935); Thomasson, Thomasson and others (1960).
Reference(s)/comments: If a transmissivity (T) estimate is reported, then reference to published report, memo, or written communication where transmissivity estimate was obtained. If a qualitative transmissivity estimate, then a comment explaining qualitative transmissivity estimate; "responds well" indicates water levels in the well have a good response to recharge events; gal/min, gallon per minute]

Geographic area	Short name	Average unsaturated zone thickness	Primary hydrogeologic unit(s)	Transmissivity	Testing method	Analysis method	Reference(s)/comments
Rainier Mesa	ER-12-1	1,524	CA	22	CRT	CJ	K.J. Halford, U.S. Geological Survey (2016)
	ER-12-3 m	3,109	CA	504	CRT	Halford	K.J. Halford, U.S. Geological Survey (2016)
	ER-12-4 m	2,565	CA	9	CRT	Halford	K.J. Halford, U.S. Geological Survey (2016)
	ER-12-3 p	1,244	TCU	<1	CRT	BR	K.J. Halford, U.S. Geological Survey (2016)
	ER-19-1-3	1,005	TCU	<1	Slug	BR	K.J. Halford, U.S. Geological Survey (2016)
	U-12s	931	GCU	4	Slug	BR	K.J. Halford, U.S. Geological Survey (2016)
Cactus Range	Hammel Mine	119	VOLC	Low?	–	–	Mine shaft; likely low T rocks mined
	Antelope Mine 1	19	VOLC	Low?	–	–	Mine shaft; likely low T rocks mined
	Antelope Mine 2	23	VOLC	Low?	–	–	Mine shaft; likely low T rocks mined
	Antelope Mine 3	30	VOLC	Low?	–	–	Mine shaft; likely low T rocks mined
	Sulfide Mine	51	VOLC	Low?	–	–	Mine shaft; likely low T rocks mined
Eastern Pahute Mesa and Buckboard Mesa	ER-18-2	1,211	WTA	3	CRT	CJ; Moench	IT Corporation (2002a)
	TW-1	1,462	WTA, CA	580	SC	Thomasson	Halford, written commun. (2016)
	U-19bh	2,085	TCU	<0.004	Slug	BR	Halford, written commun. (2016)
	U-19bk	1,984	LFA	Low?	–	–	Water levels elevated
	UE-18r	1,364	WTA	1,300	CRT	CJ	Warren and others (1998)
	UE-18t	914	WTA	High?	–	–	Water level response similar to nearby UE-18r
	UE-19c WW	2,339	LFA	1,900	CRT	CJ	Warren and others (1998)
	UE-19h	2,111	LFA	19,000	CRT	CJ	Warren and others (1998)

Table 7. Hydrogeologic and hydraulic characteristics of wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.—Continued

Geographic area	Short name	Average unsaturated zone thickness	Primary hydrogeologic unit(s)	Transmissivity	Testing method	Analysis method	Reference(s)/comments
Rocket Wash and Thirsty Canyon	ER-EC-1	1,856	LFA, WTA	6,200	CRT	CJ	Garcia and others (2010)
	ER-EC-4	749	LFA	50,000	CRT	CJ	Garcia and others (2010)
	ER-EC-6	1,425	LFA, WTA	800	CRT	CJ	IT Corporation (2002b)
	ER-EC-8	323	WTA	2,800	CRT	Moench	IT Corporation (2002c)
	PM-3-1	1,458	WTA	50	CRT	Neuman	Warren and others (1998)
	PM-3-2	1,455	TCU	50	CRT	Neuman	Warren and others (1998)
Gold Flat and Central PMOV basin	ER-20-1	1,989	WTA	5,300	MWAT	MOD	Halford and others (2012)
	ER-20-6-1	2,023	LFA	3,300	MWAT	Theis	IT Corporation (1998)
	ER-20-6-2	2,024	LFA	2,000	MWAT	Theis	IT Corporation (1998)
	ER-20-6-3	2,015	LFA	1,400	MWAT	MOD	Garcia and others (2011)
	ER-EC-2A	755	TCU	230	CRT	BF	Oberlander and others (2002)
	Gold Flat 2a	233	VOLC	Mod?	–	–	Well had a pump in the past.
Timber Mountain and Jackass Flats	ER-EC-5	1,016	WTA	28,000	CRT	BF	Oberlander and others (2002)
	ER-EC-7	746	LFA	2,200	CRT	CJ	Garcia and others (2017)
	JF-1	1,161	WTA	Mod?	–	–	Responds well
	JF-2	995	WTA	Mod?	–	–	Responds well
Northern Oasis Valley	ER-OV-01	18	WTA	Mod?	–	–	Responds well
	ER-OV-03b	346	AA	Mod?	–	–	5 gal/min during well development; responds well
	ER-OV-03c	214	WTA	Mod?	–	–	7 gal/min during well development; responds well
	ER-OV-03c2	215	WTA	Mod?	–	–	7 gal/min during well development; responds well
	ER-OV-06a	16	WTA	Mod?	–	–	20 gal/min during well development
	ER-OV-06a2	18	LFA	Low?	–	–	Less than 1 gal/min during well development
	ER-OV-03a2	160	TCU	Low?	–	–	Less than 1 gal/min during well development
	ER-OV-03a	58	WTA	Mod?	–	–	20 gal/min during well development
	ER-OV-03a3	59	WTA	Mod?	–	–	20 gal/min during well development
Southern Oasis Valley	ER-OV-02	28	AA	Mod?	–	–	20 gal/min during well development
	ER-OV-04a	24	AA	Mod?	–	–	20 gal/min during well development
	ER-OV-05	32	AA	Mod?	–	–	20 gal/min during well development
	Springdale Upper	24	AA	Mod?	–	–	Water-levels respond well to recharge

Table 7. Hydrogeologic and hydraulic characteristics of wells with steady-state water-level trends in the Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada.—Continued

Geographic area	Short name	Average unsaturated zone thickness	Primary hydrogeologic unit(s)	Transmissivity	Testing method	Analysis method	Reference(s)/comments
Western Oasis Valley and Sarcobatus Flat	Springdale	94	WTA, LFA	Mod	–	–	Well equipped with windmill until 1985
	TPJ-1	43	AA	Mod?	–	–	Used as a stock well between 1960 and 1985
	TPJ-2	58	AA	Mod?	–	–	Water levels not elevated and respond to recharge
Fortymile Wash and Amargosa Narrows	UE-29a1	84	LFA	Mod?	–	–	Responds well
	UE-29a2	92	LFA	Mod?	–	–	Responds well
	UZN 91	54	WTA	Mod?	–	–	Responds well
	Narrows South 2	19	AA	720	SC	Thomasson	K.J. Halford, U.S. Geological Survey (2016)
Yucca Mountain	G-2	1,751	TCU	100	CRT	CJ	O'Brien (1998, table 4)
	UE-25 WT 16	1,549	LFA	Mod?	–	–	Responds well
	UE-25 WT 4	1,438	LFA	Mod?	–	–	Responds well
	H-1 (Tube 3)	1,878	WTA	1,600	CRT	Theis	Rush and others (1984)
	H-1 (Tube 4)	1,877	WTA	1,600	CRT	Theis	Rush and others (1984)
	H-5 (lower)	2,308	LFA	390	CRT	CJ	Robison and Craig (1991)
	H-5 (upper)	2,308	WTA	390	CRT	CJ	Robison and Craig (1991)

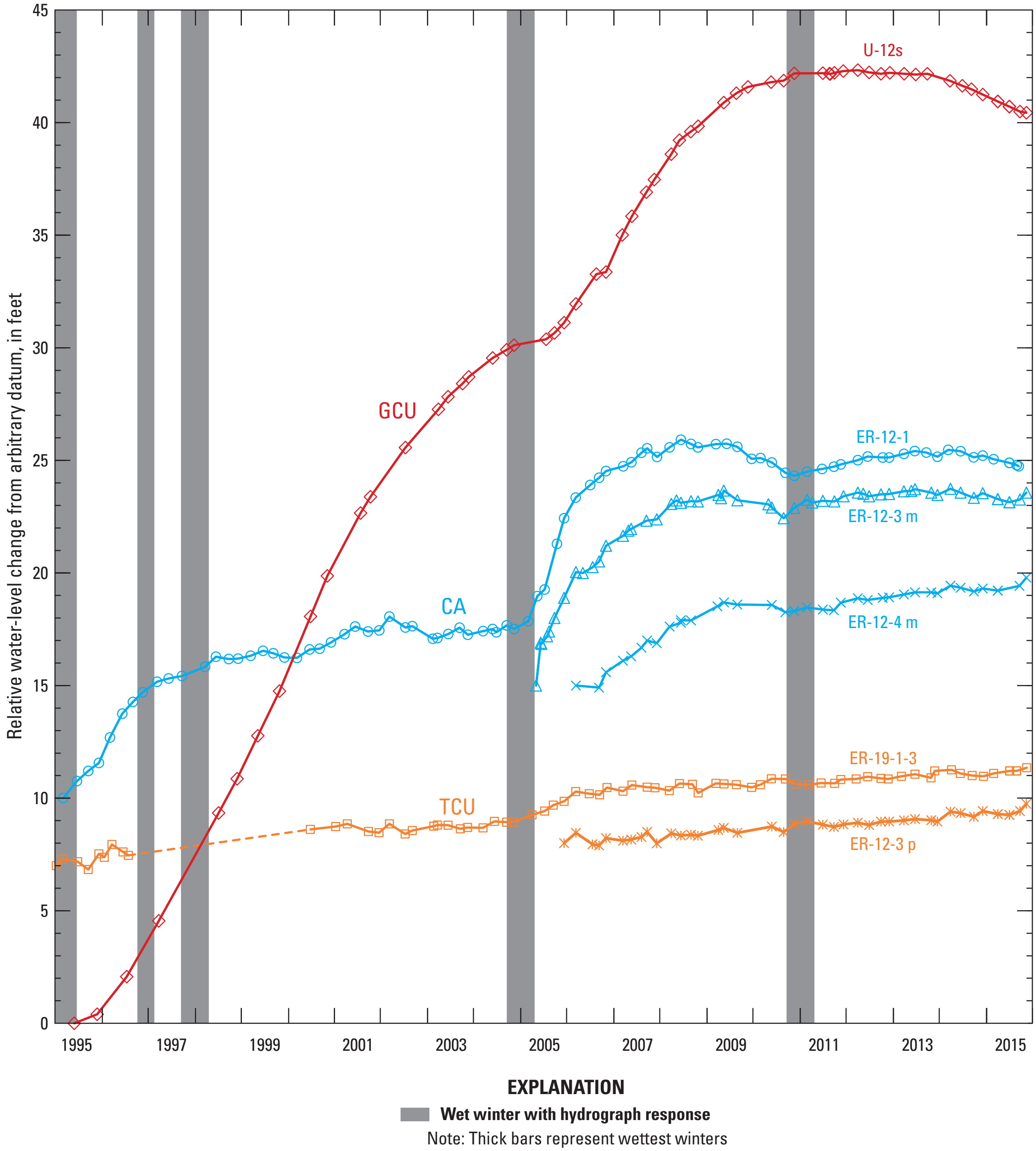


Figure 30. Relative water-level change in six wells in the Rainier Mesa area, where similar trends are grouped by primary hydrogeologic unit screened in well, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: GCU, granite confining unit; CA, carbonate aquifer; and TCU, tuff confining unit. A wet winter in this graph is defined as exceeding 130 percent of average winter precipitation for the Rainier Mesa precipitation index.

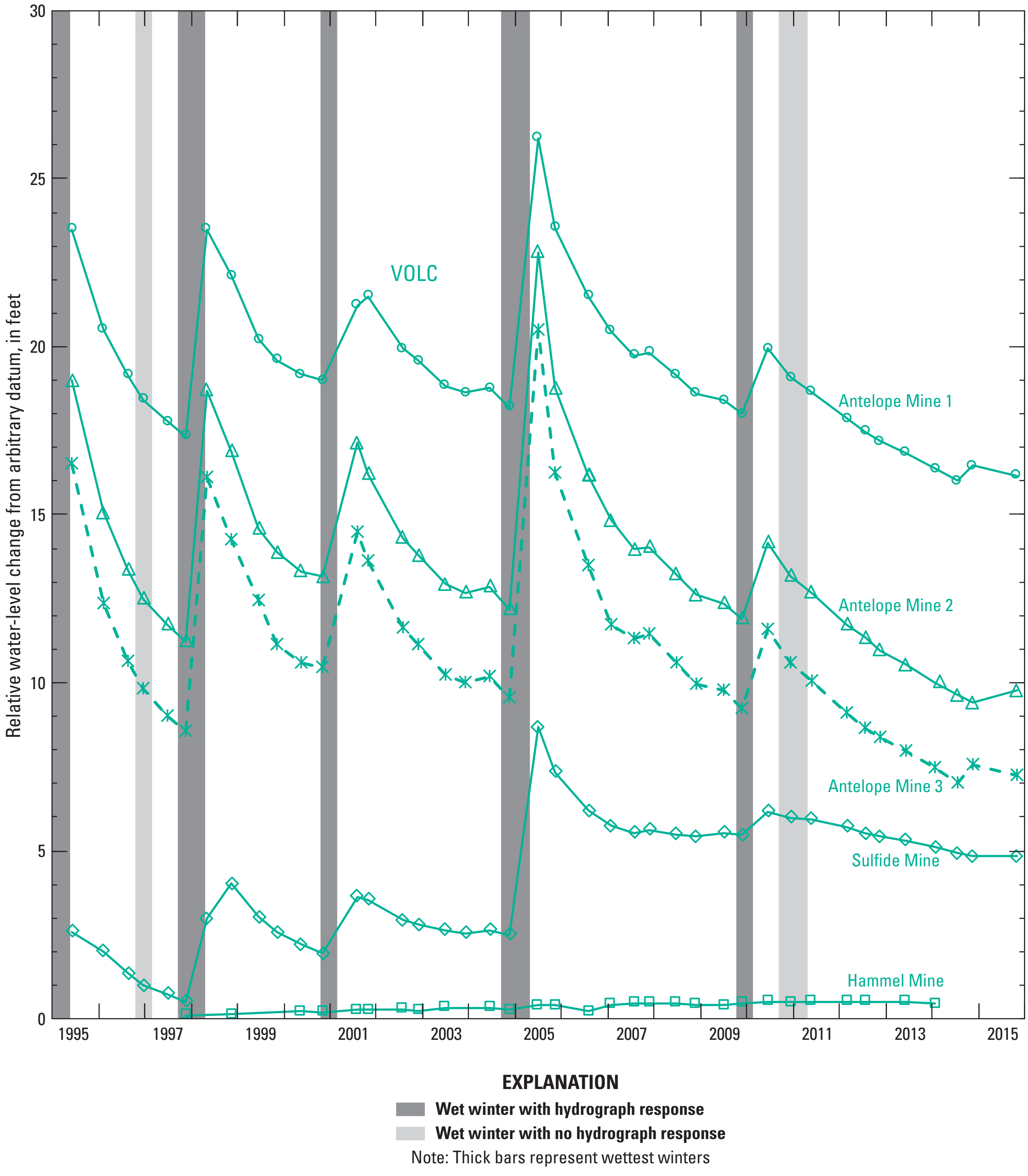


Figure 31. Relative water-level change in five wells in the Cactus Range area, which are screened in undifferentiated volcanic rocks (VOLC), Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. A wet winter in this graph is defined as exceeding 115 percent of average winter precipitation for the Rainier Mesa precipitation index.

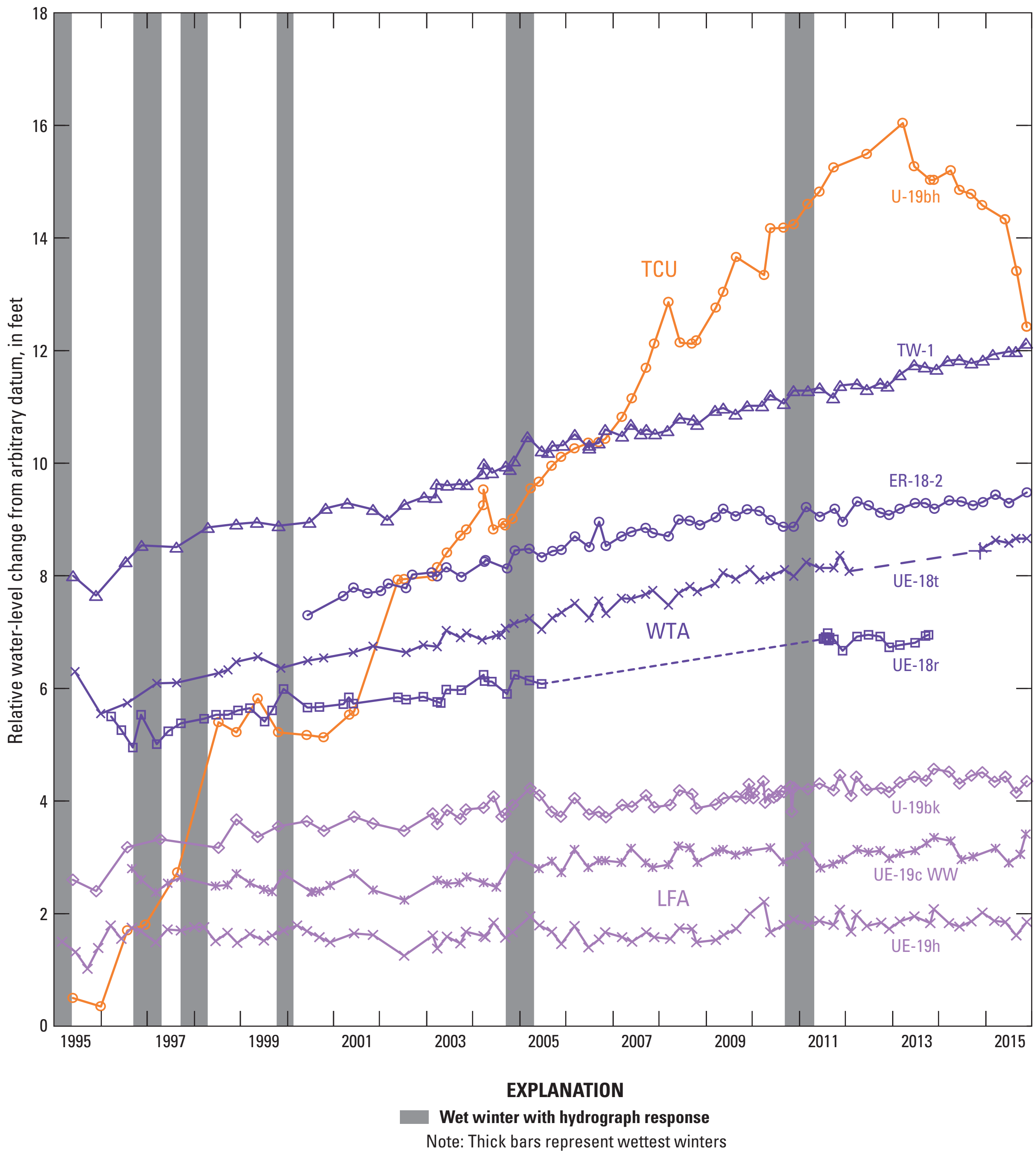


Figure 32. Relative water-level change in eight wells in eastern Pahute Mesa and Buckboard Mesa, which have long-term water-level rises in response to recharge, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: TCU, tuff confining unit, LFA, lava-flow aquifer; and WTA, welded-tuff aquifer. A wet winter in this graph is defined as exceeding 120 percent of average winter precipitation for the Pahute Mesa precipitation index.

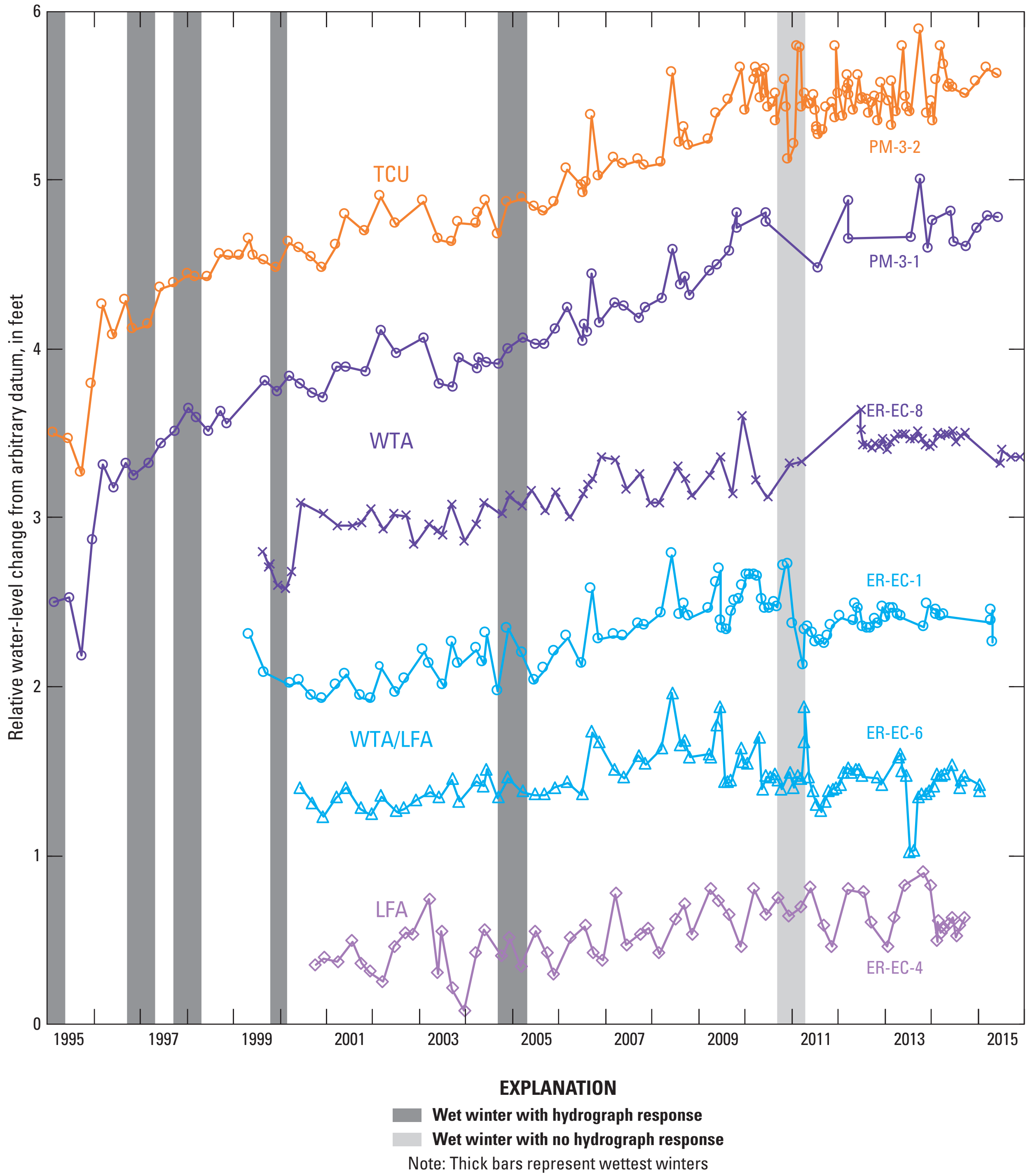


Figure 33. Relative water-level change in six wells near Thirsty Canyon and Rocket Wash areas, which have weak water-level responses to the 2005 winter, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer; WTA, welded-tuff aquifer; WTA/LFA, composite unit of welded-tuff and lava-flow aquifers; and TCU, tuff confining unit. A wet winter in this graph is defined as exceeding 120 percent of average winter precipitation for the Pahute Mesa precipitation index.

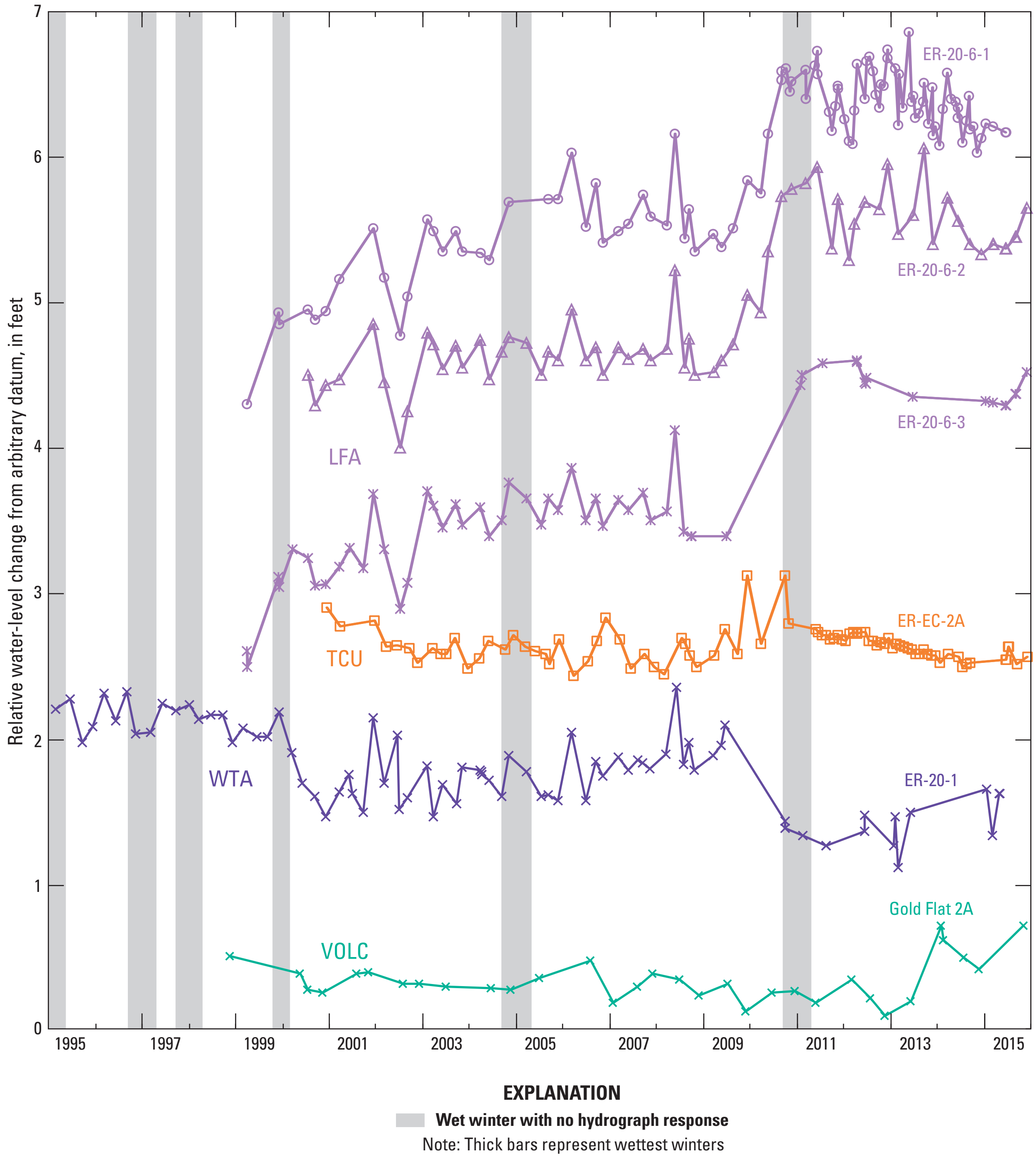


Figure 34. Relative water-level change in Gold Flat and central Pahute Mesa-Oasis Valley (PMOV) wells with ambiguous water-level responses not attributed to recharge, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer; WTA, welded-tuff aquifer; TCU, tuff confining unit; and VOLC, undifferentiated volcanic rocks. A wet winter in this graph is defined as exceeding 120 percent of average winter precipitation for the Pahute Mesa precipitation index.

Water levels in the ER-20-6 well cluster (wells *ER-20-6-1*, *ER-20-6-2*, and *ER-20-6-3*) show a sharp rise from October 2009 to August 2010, which does not correlate with a wet winter (fig. 34). This is the only large fluctuation in the hydrographs and the lack of correlation with a wet winter suggests the sharp rise resulted from some other stress. The consistency of the water-level rises, as observed in all three wells, indicates that these rises occur over a relatively large area and likely are the result of natural causes.

Timber Mountain and Jackass Flats

Wells in Jackass Flats and on the flanks of Timber Mountain have similar responses to episodic recharge (fig. 35; table 6). Water levels in these wells only respond to recharge from the wettest winters (1998 and 2005) because the unsaturated zone is more than 700 ft thick at these wells (table 7). Wells *JF-1* and *JF-2* are within 0.2 mi of Fortymile Wash, and most surface-water infiltration from precipitation during slightly above-average winters is lost to evaporation. Therefore, only streamflow derived from precipitation occurring in the wettest winters is sufficient to recharge the groundwater-flow system near wells *JF-1* and *JF-2*. Similarly, wells *ER-EC-5* and *ER-EC-7* are adjacent to Timber Mountain, and water levels only respond to recharge from surface-water infiltration along ephemeral channels at the base of the mountain.

Northern Oasis Valley

In northern Oasis Valley (fig. 29), most recharge occurs by surface-water infiltration along ephemeral washes during wet winters. In winters with average or below-average precipitation, recharge is not observed, indicating that infiltration is lost to evapotranspiration. Wells *ER-OV-01*, *ER-OV-06a*, and *ER-OV-06a2*, referred herein as the ER-OV-06 well cluster, are along an ephemeral upper reach of the Amargosa River; wells *ER-OV-03c* and *ER-OV-03c2*, referred herein as the ER-OV-03c well cluster, are along an ephemeral upper reach of Beatty Wash; and well *ER-OV-03b* is between the ephemeral channels of the Amargosa River and Beatty Wash (fig. 3). These wells have weak water-level responses to recharge from the 2000, 2005, and 2010 winters, and attenuated responses to the 2001 winter (fig. 36; table 6).

Water-level responses in the ER-OV-03a well cluster (wells *ER-OV-03a*, *ER-OV-03a2*, and *ER-OV-03a3*) are unique and differ from other water-level trends in the area. Water levels in the ER-OV-03a well cluster respond to recharge from the 2001, 2005, and 2010 winters (fig. 37; table 6); however, shallow and deep wells have differing trends. Well *ER-OV-03a2* is screened in a deep interval of a tuff confining unit, where the screened depth is 560–655 ft below land surface. Water levels in well *ER-OV-03a2* have a rising trend in response to recharge, and have seasonal water-level fluctuations likely caused by barometric pressure (Elliott and Fenelon, 2010). Wells *ER-OV-03a* and *ER-OV-03a3* are screened in shallow intervals of a welded-tuff aquifer (WTA), where screened depths are less than 250 ft below land surface. These wells have a long-term declining trend, where recharge from the 2001 winter likely caused the flattening of the trend from 2001 to 2004, and recharge from the 2005 and 2010 winters caused rising trends lasting no more than 1 year (fig. 37). The reason for the declining trend is uncertain. One explanation is that nearby spring outlets may have been altered, causing increased spring discharge and subsequent water-level declines in these wells (Elliott and Fenelon, 2010). Additionally, the wells are within 1,000 ft of an earthen dam and a pond. It is not known if these features may have affected variations in local recharge to or drainage from the shallow aquifer.

Southern Oasis Valley

Wells in southern Oasis Valley (fig. 29) are distant from high-altitude recharge areas; however, recharge occurs because the wells are shallow (depths less than 40 ft) and most are near surface-water channels. Wells *ER-OV-05* and *Springdale Upper* are near an unnamed tributary to the Amargosa River, well *ER-OV-02* is near the Amargosa River, and well *ER-OV-04a* is near the confluence of Beatty Wash with the Amargosa River (fig. 3). These wells are within a groundwater discharge area and show seasonal responses to evapotranspiration (Reiner and others, 2002). Responses to recharge during the 1998 and 2005 winters are superimposed on the evapotranspiration responses (fig. 38; table 6). Recharge occurs as surface water infiltrates into alluvial sediments underlying ephemeral channels that receive runoff from winter storms (Reiner and others, 2002).

Western Oasis Valley and Sarcobatus Flat

Only the wettest winters provide recharge to the groundwater system in western Oasis Valley and Sarcobatus Flat (fig. 29). Wells *Springdale*, *TPJ-1*, and *TPJ-2* have long-term rising trends with weak recharge responses to the wettest winters (fig. 39; table 6). Weak recharge responses likely occur because these wells are distant from high-altitude recharge areas and localized recharge in ephemeral channels. Small amounts of recharge likely are sourced from the Black and Grapevine Mountains (fig. 1).

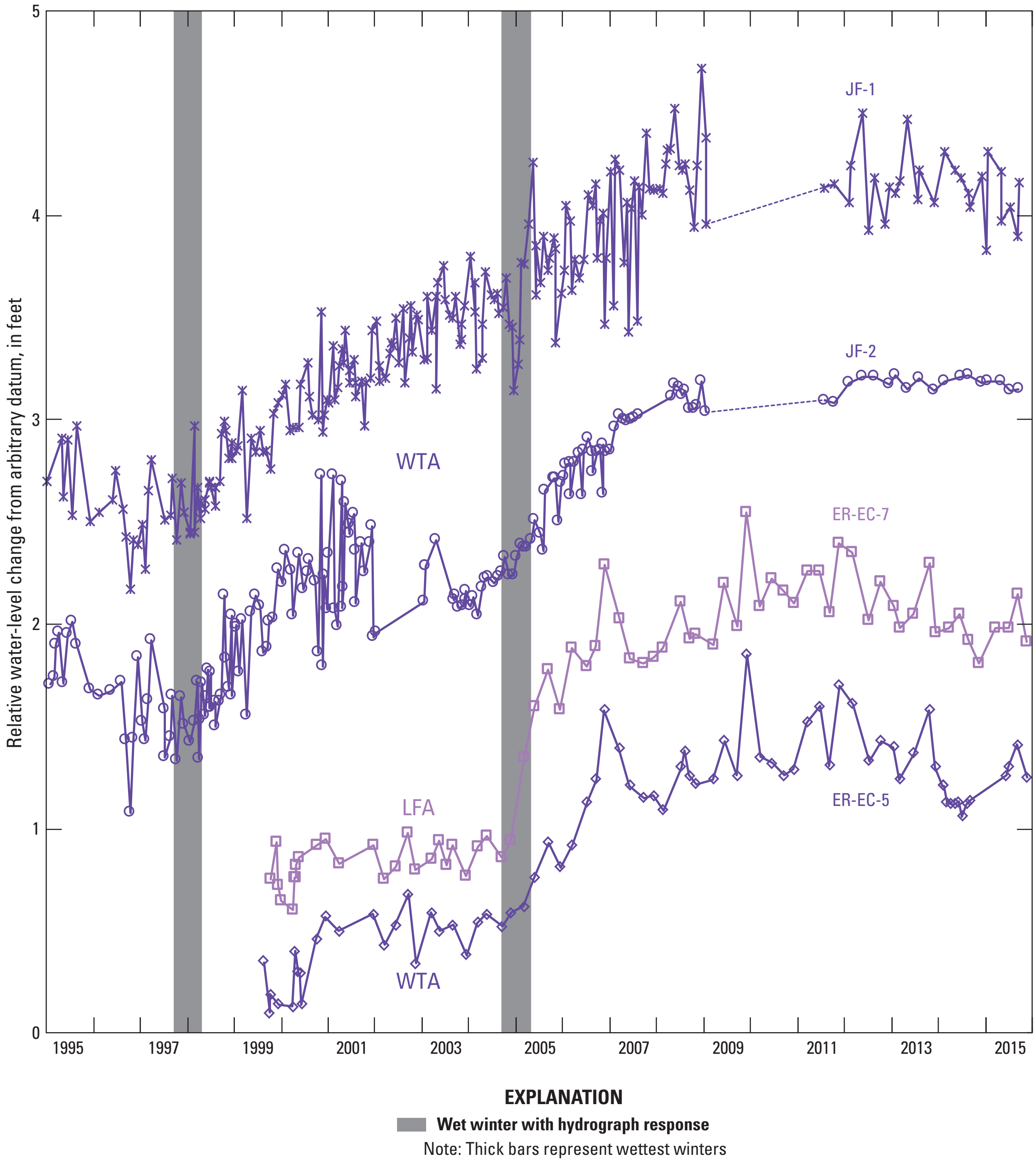


Figure 35. Relative water-level change in Jackass Flats and Timber Mountain wells, which have strong water-level responses to the wettest winters, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer; and WTA, welded-tuff aquifer. A wet winter in this graph is defined as exceeding 160 percent of average winter precipitation for the Beatty precipitation index.

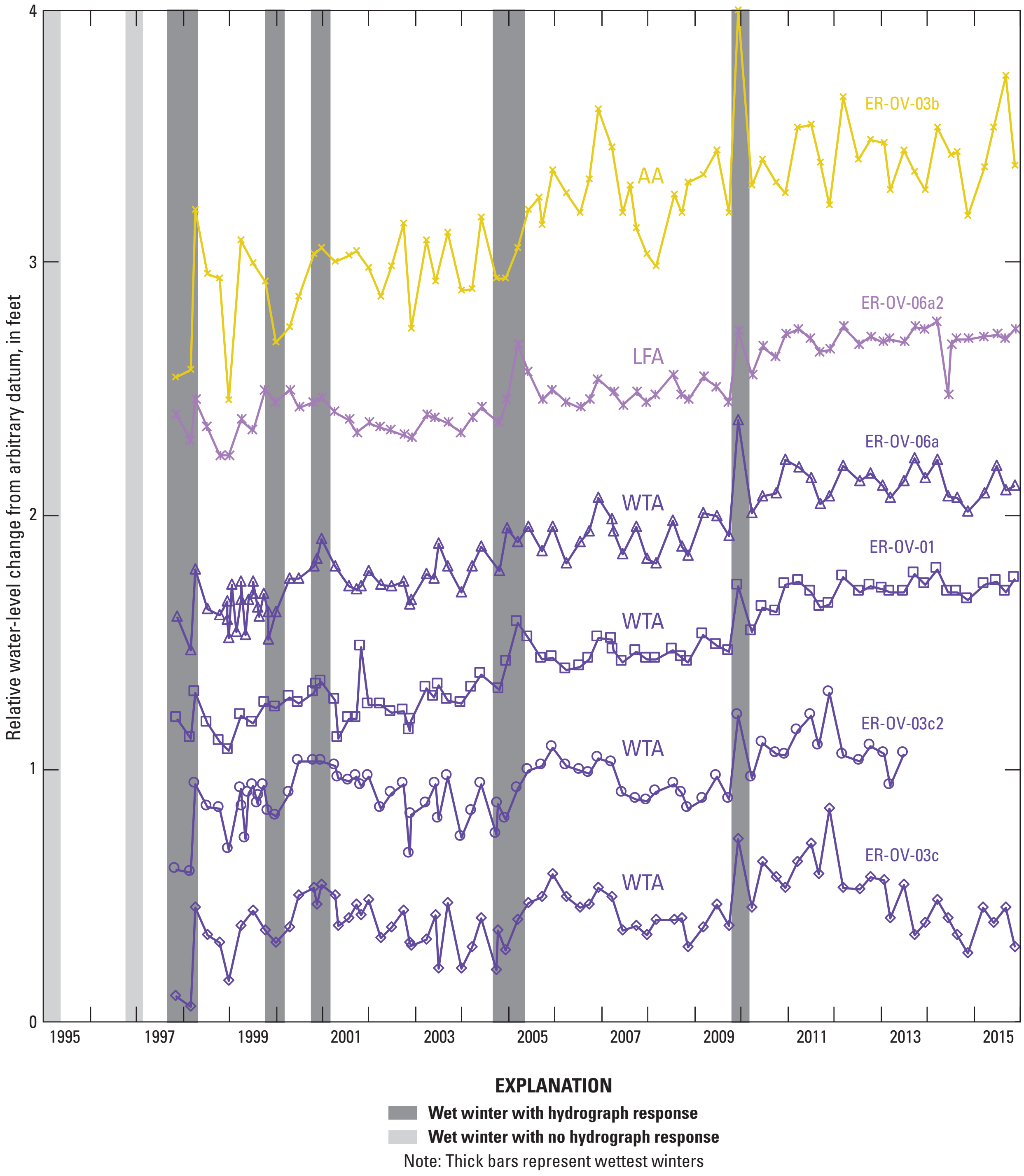


Figure 36. Relative water-level change in northern Oasis Valley wells near Amargosa River and Beatty Wash, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: AA, alluvial aquifer; LFA, lava-flow aquifer; and WTA, welded-tuff aquifer. A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

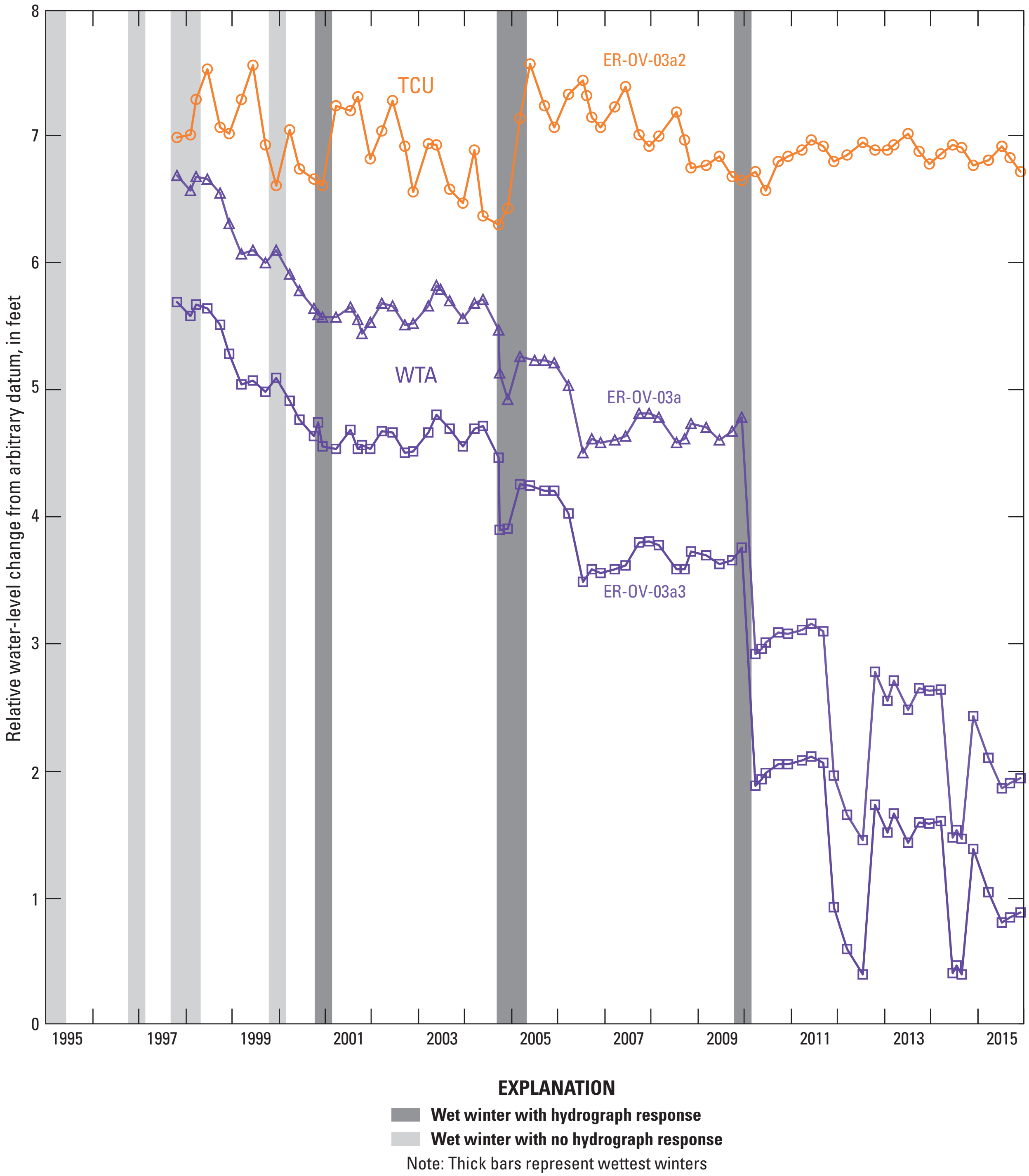


Figure 37. Relative water-level change in northern Oasis Valley wells with declining trends (*ER-OV-03a* and *ER-OV-03a3*) or trends strongly affected by barometric pressure (*ER-OV-03a2*) , Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: WTA, welded-tuff aquifer; and TCU, tuff confining unit. A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

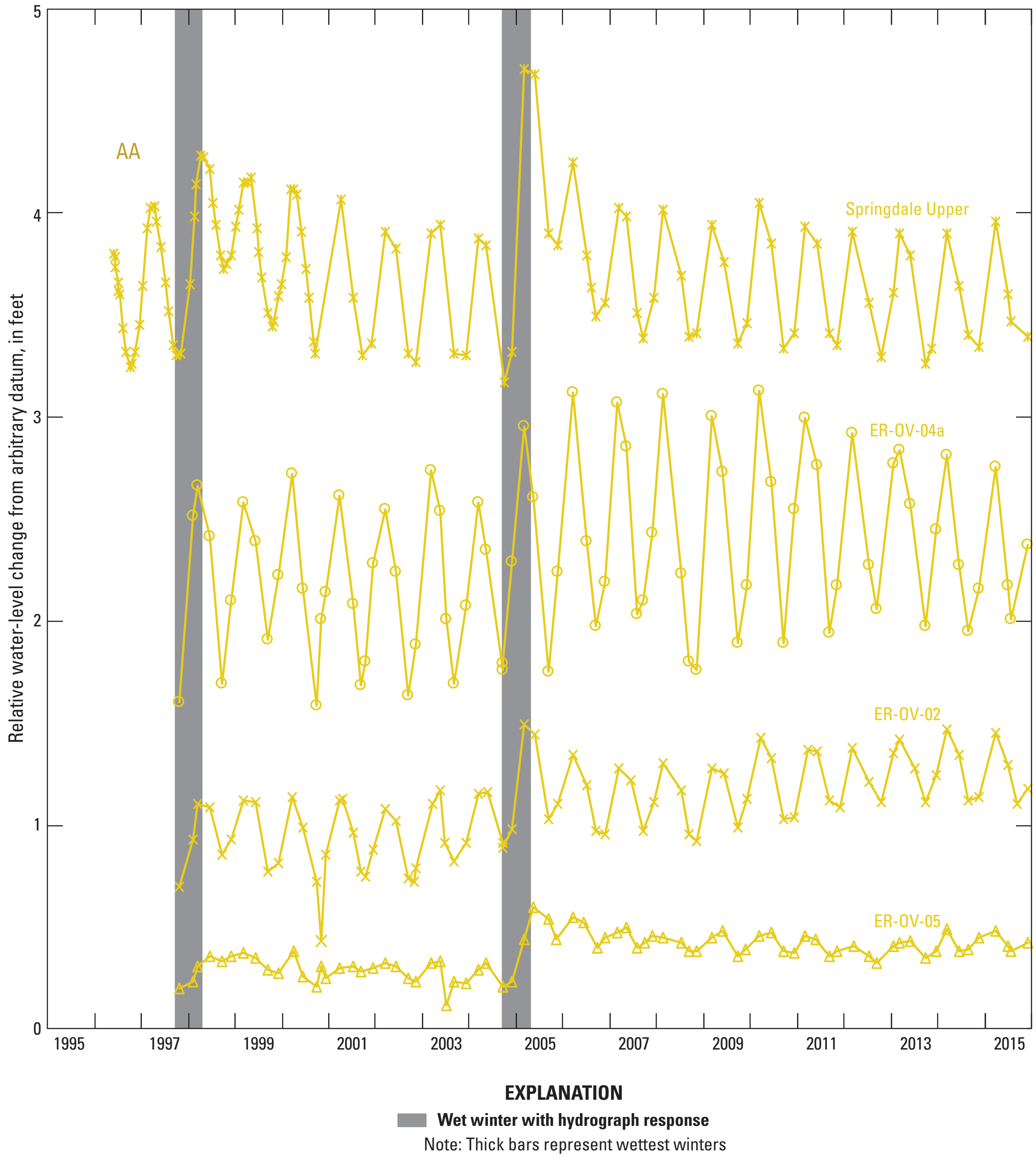


Figure 38. Relative water-level change in southern Oasis Valley wells, which have strong water-level responses to recharge superimposed on evapotranspiration, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Wells are screened in an alluvial aquifer (AA). A wet winter in this graph is defined as exceeding 160 percent of average winter precipitation for the Beatty precipitation index.

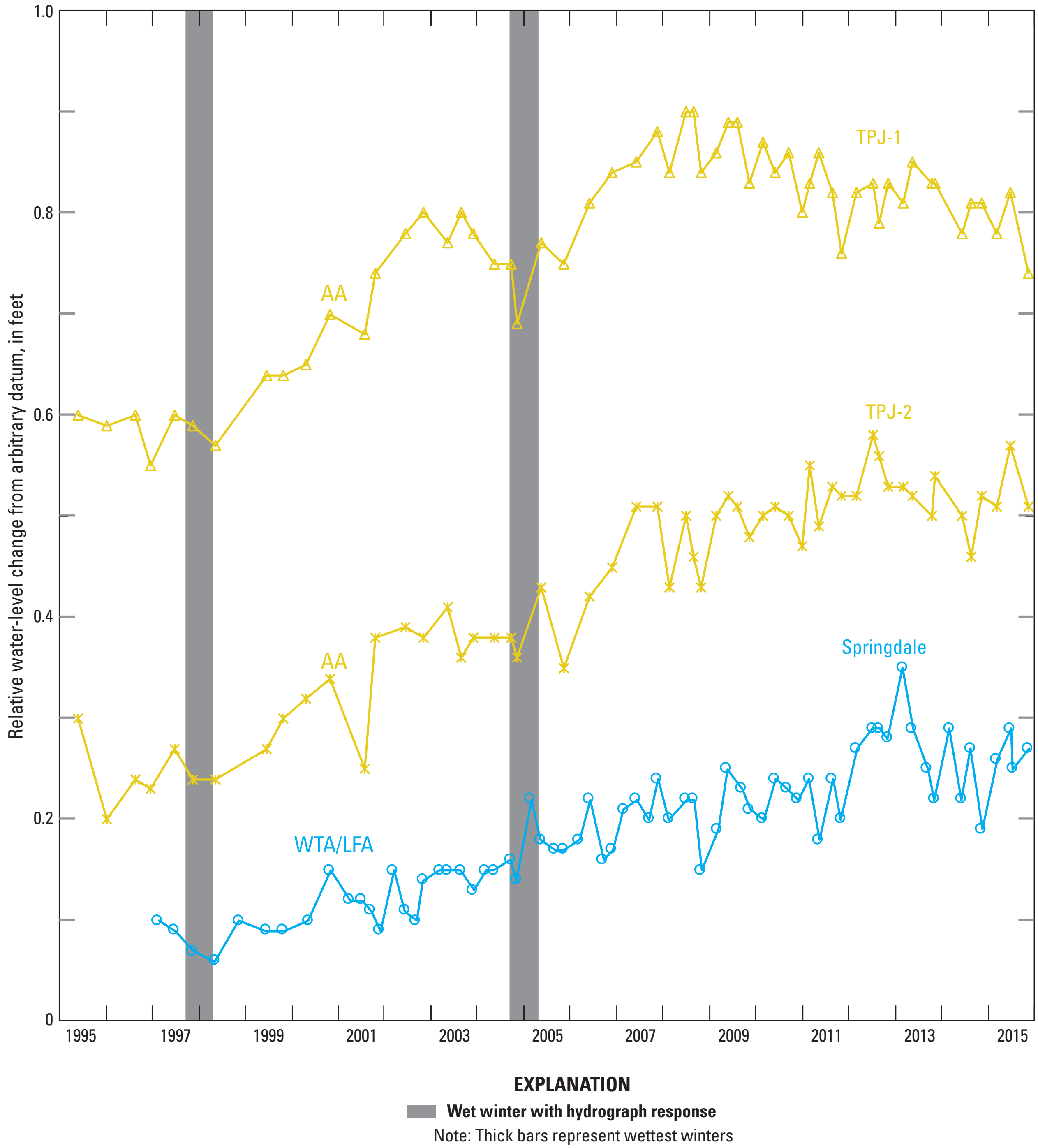


Figure 39. Relative water-level change in western Oasis Valley and Sarcobatus Flat wells, which have weak responses to recharge, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: AA, alluvial aquifer; and WTA/LFA, composite unit of welded-tuff and lava-flow aquifers. A wet winter in this graph is defined as exceeding 160 percent of average winter precipitation for the Beatty precipitation index.

Fortymile Wash and Amargosa Narrows

Along upgradient reaches of Fortymile Wash (fig. 29), water levels in wells respond to recharge from six wetter-than-average winters. Wells *UE-29a1*, *UE-29a2*, and *UZN 91*, referred herein as the UE-29 well cluster, have strong responses to recharge that occurred during the winters of 1995, 1998, 2000, 2005, and 2010, and weak responses to recharge from the winter of 2001 (fig. 40; table 6).

Water levels in the UE-29 well cluster respond to most wetter-than-average winters because the wells receive focused recharge from the infiltration of streamflow in Fortymile Wash (Savard, 1998). The recharge response is amplified at the UE-29 well cluster because the wells are located in a constrained reach of Fortymile Wash, where the narrow, steep-walled valley limits the floodplain and focuses recharge through coarse-grained streambed sediments.

Water levels in well *Narrows South 2* respond to most of the wetter-than-average winters between 1999 and 2010 (fig. 41; table 6). The well is screened in shallow coarse-grained alluvial deposits within the Amargosa Narrows, which is near the terminus of the PMOV basin (fig. 29). In the Amargosa Narrows, the ephemeral Amargosa River channel is constrained between steep bedrock highlands (Elliott and Fenelon, 2010), which focuses recharge during high-intensity streamflow events along the channel (Reiner and others, 2002). Water levels in well *Narrows South 2* show strong recharge responses from the 2000, 2001, and 2005 winters superimposed on evapotranspiration responses.

Yucca Mountain

In the Yucca Mountain area (fig. 29), water levels in wells have responded to recharge from the winters of 1995, 1998, 2000, 2001, and 2005 (table 6). Water-level responses to recharge from the winter of 2010 are unknown because of a water-level data gap from mid-2007 to early 2012; however, water levels in all wells show a long-term rise from 2005 to 2014. The rise is a response to recharge from either the winter of 2005 or the winters of 2005 and 2010 (fig. 42).

Water-level responses to recharge during the winters of 1995 and 1998 only were discernible in wells distant from previous aquifer testing conducted between 1996 and 1998. These wells include wells *H-1 (Tube3)* and *H-1 (Tube4)*, referred herein as the H-1 well cluster, and wells *H-5 (lower)* and *H-5 (upper)*, referred herein as the H-5 well cluster. Water levels in these wells have strong observable responses to the 1995 and 1998 winters (fig. 42). In other Yucca Mountain wells, water-level responses to recharge from the 1995 and 1998 winters are indeterminate because recharge, if any, was masked by drawdown from aquifer testing between 1996 and 1998. Small drawdowns of less than 1 ft were observed in wells *UE-25 WT 4* and *UE-25 WT 16* from nearby aquifer testing in borehole *UE-25c 3* between 1996 and 1998 (Geldon and others, 2002). Water levels in well *UE-25 WT 16* show a rising trend following the 1998 winter that may be attributed to recharge from this winter (fig. 43). Large drawdowns of more than 100 ft were observed in well *G-2* because of single-well aquifer testing at the well in 1996 (O’Brien, 1998), and the long recovery period following aquifer testing masked potential recharge from 1995 and 1998 winters. Drawdowns in these three wells are provided in Jackson (2018), but are not shown in the steady-state hydrographs in figures 42 and 43.

Water levels in all Yucca Mountain wells responded to recharge from the 2000, 2001, 2005, and 2010 winters. Water levels in the H-1 and H-5 well clusters show rising trends from 2000 to 2015 in response to recharge from wet winters (fig. 42), which may be attributed to local, high-altitude recharge derived from Yucca Mountain that is attenuated through an unsaturated zone between 1,800 and 2,300 ft thick (table 7). Water levels in well *UE-25 WT 4*, at the base of Yucca Mountain, show a similar rising trend from 2000 to 2015 in response to recharge from wet winters (fig. 42). Well *G-2* is a high-altitude Yucca Mountain well and water levels show strong recharge responses to the 2000 and 2005 winters (fig. 43). Water levels in well *UE-25 WT 16* show abrupt rises of less than 1 ft following the 2000 and 2001 winters, and a 6 ft rise from 2005 to 2016 that likely is a response to recharge from the 2005 and 2010 winters (fig. 43). Well *UE-25 WT 16* is about 0.3 mi from Yucca Wash, a major tributary to Fortymile Wash, and water-level responses likely are augmented by focused recharge from surface-water infiltration of ephemeral flows.

Spatial Distribution of Water-Level Responses to Recharge

Recharge is temporally and spatially variable in the study area. Spatially distributed recharge was assessed by reviewing water-level responses in wells to recharge from seven wet winters (1995, 1998, 2000, 2001, 2005, 2010, and 2011; table 6).

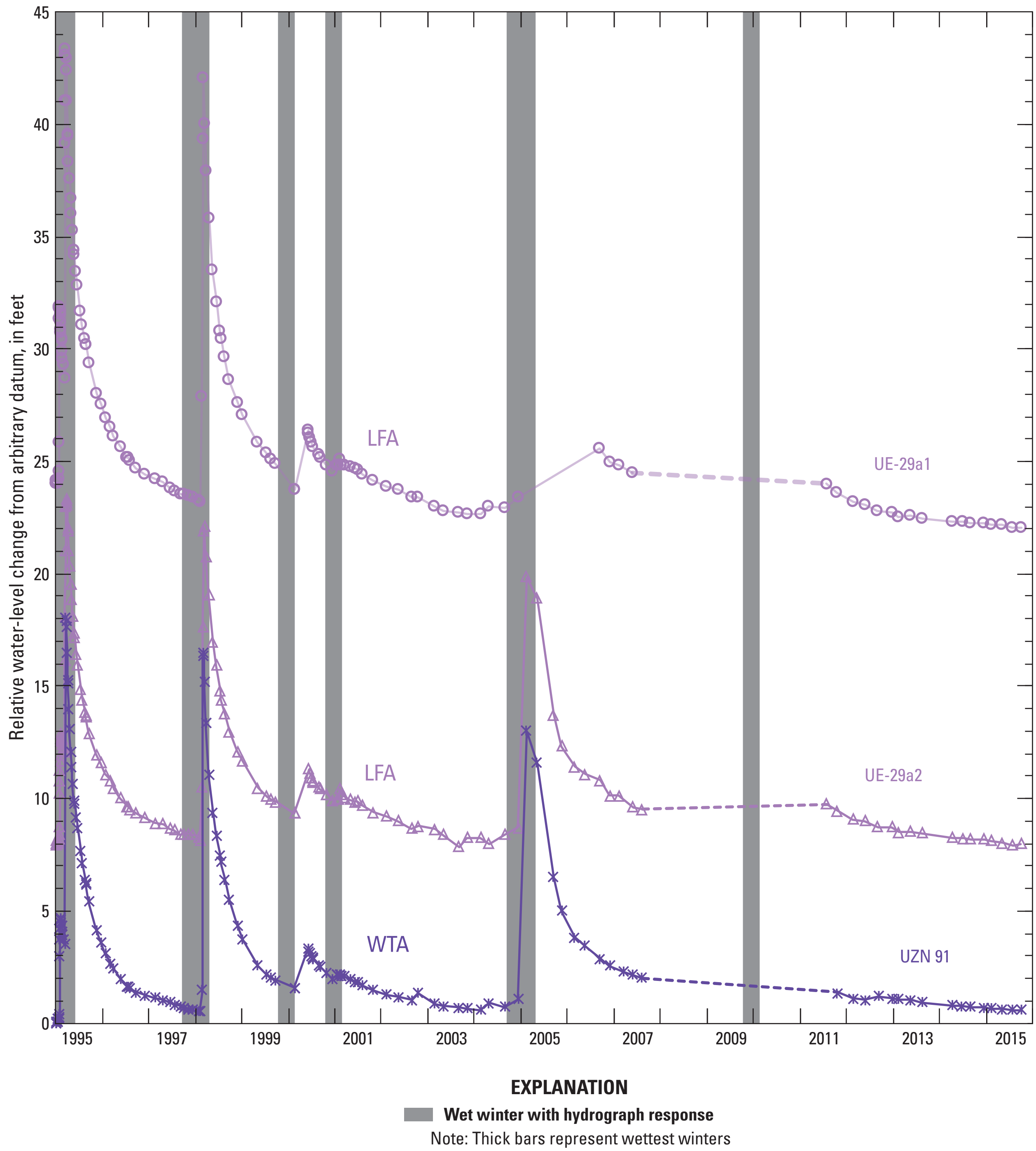


Figure 40. Relative water-level change in three Fortymile Wash wells, which have water-level responses to six wetter-than-average winters, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer; and WTA, welded-tuff aquifer. A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

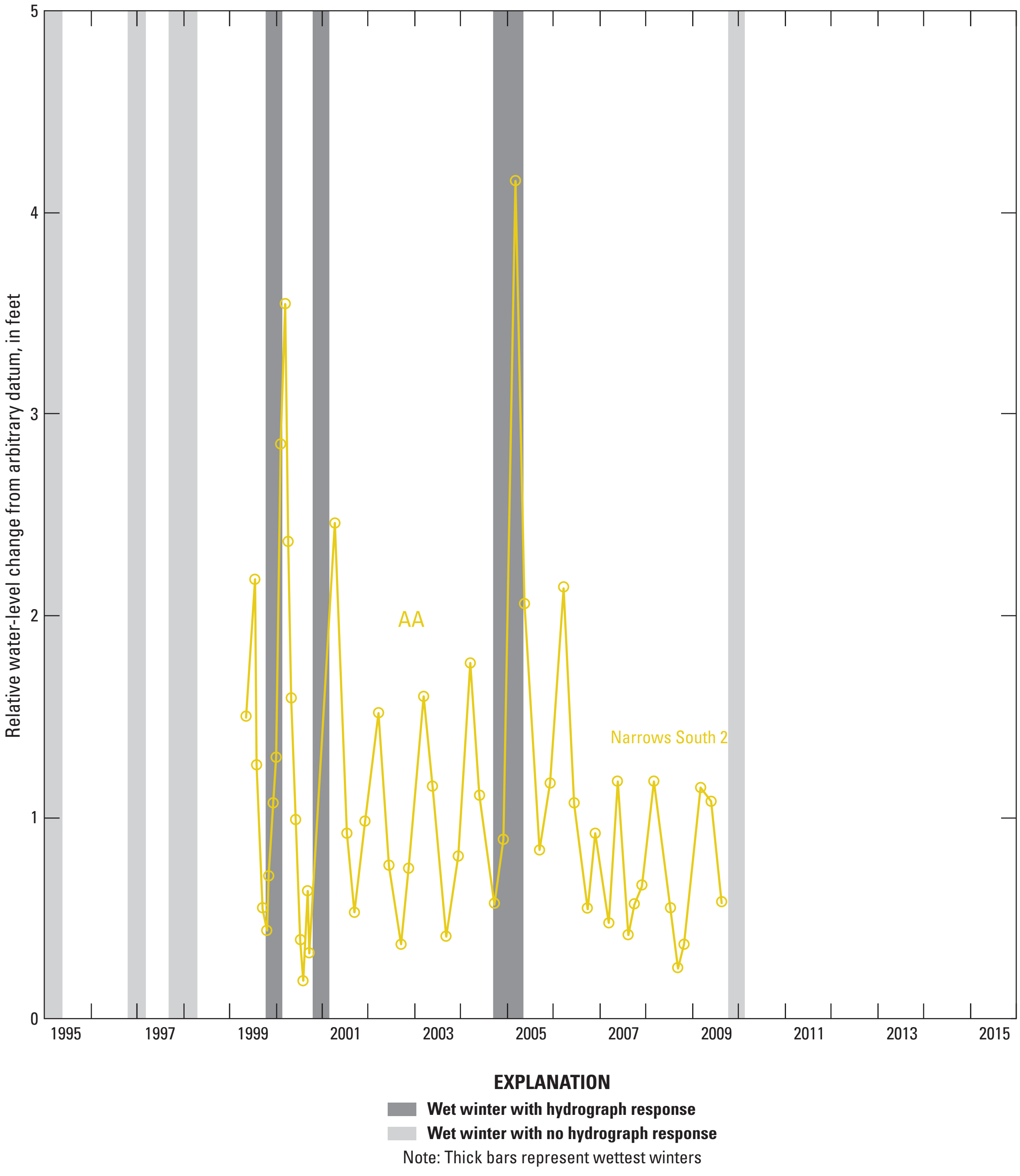


Figure 41. Relative water-level change in well Narrows South 2, which receives focused recharge from the Amargosa River, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Well is screened in an alluvial aquifer (AA). A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

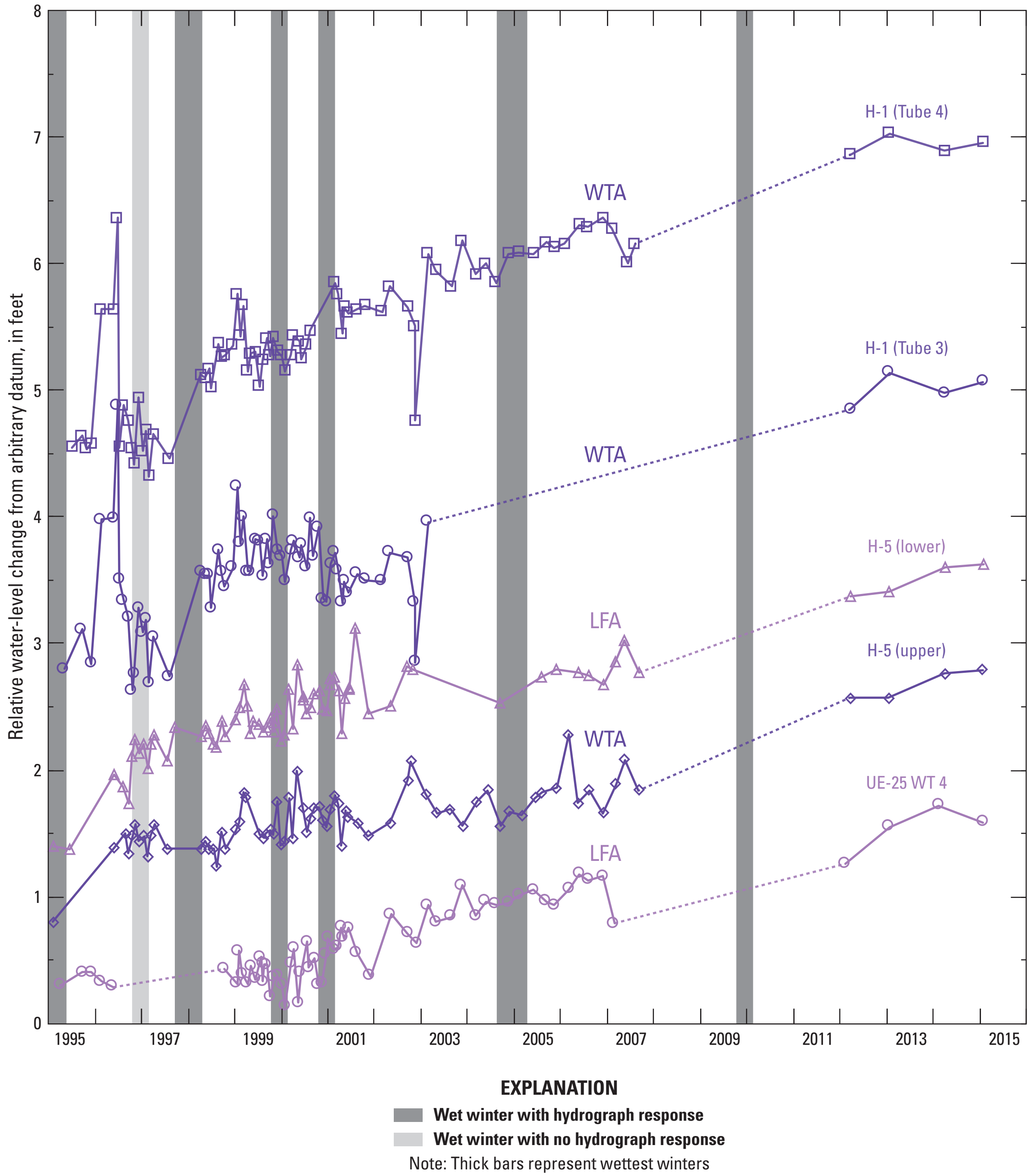


Figure 42. Relative water-level change in five Yucca Mountain wells, which have continuous rising trends from 2000 to 2015 with no discernible responses to wet winters, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer and WTA, welded-tuff aquifer. A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

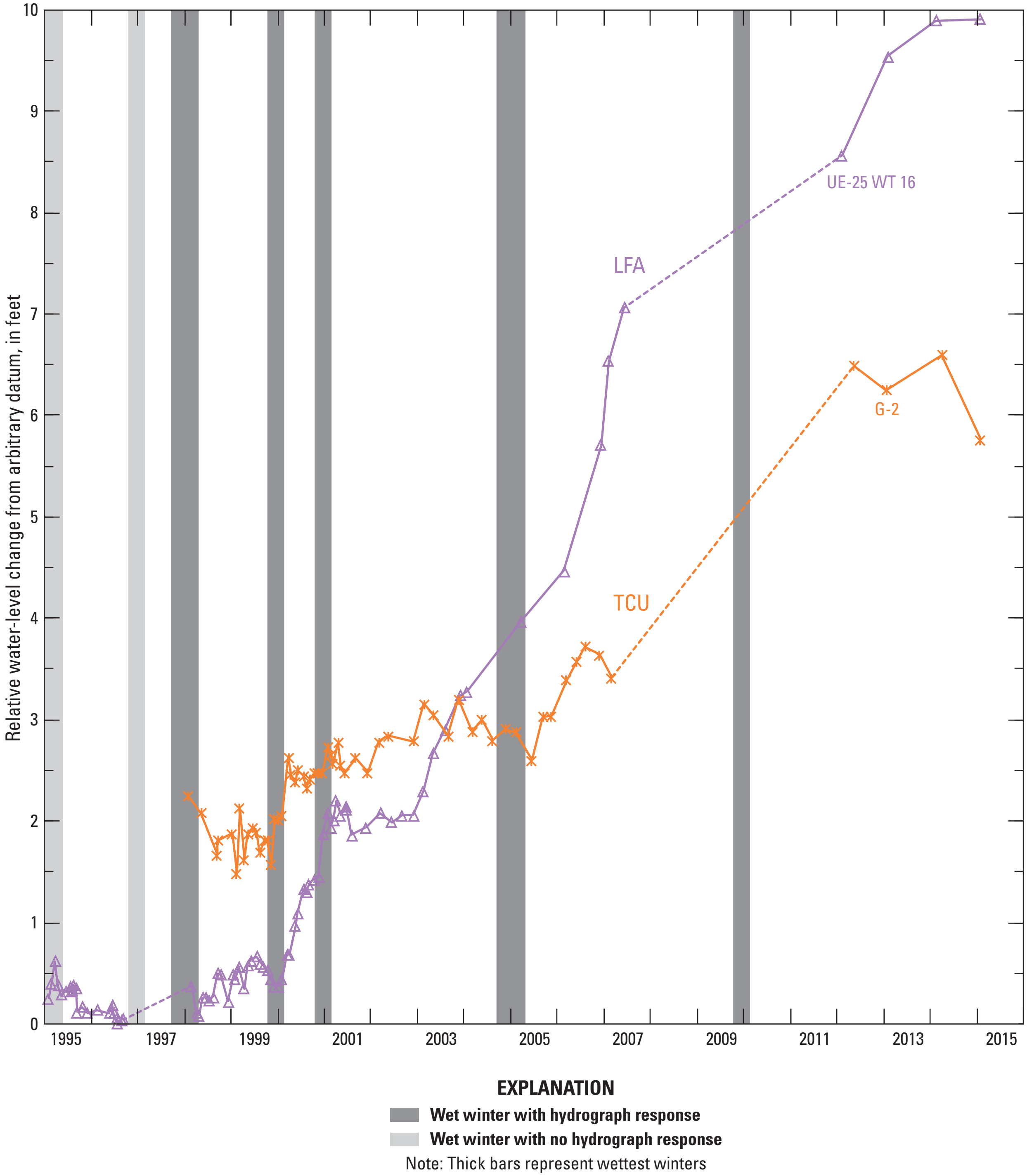
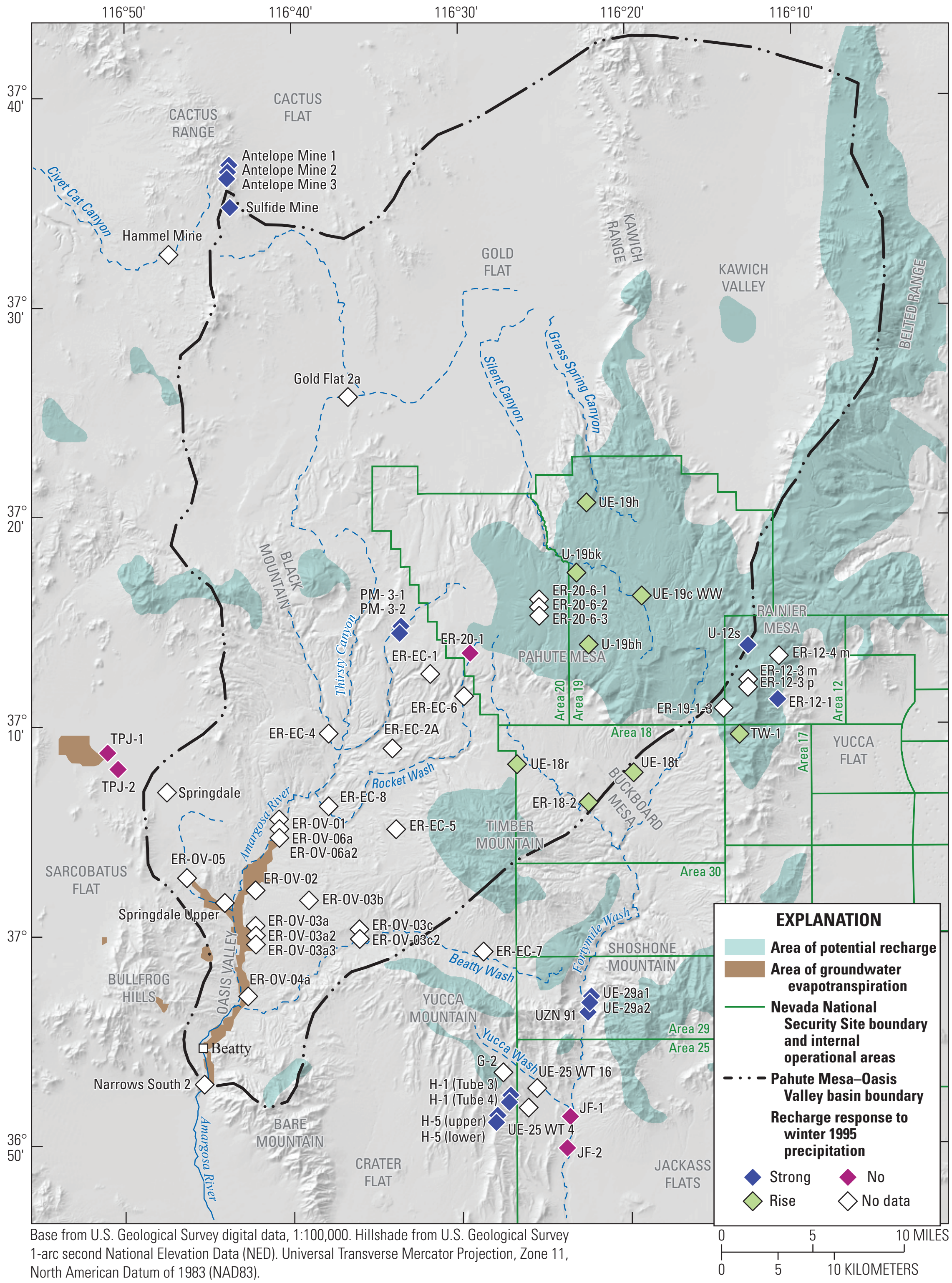


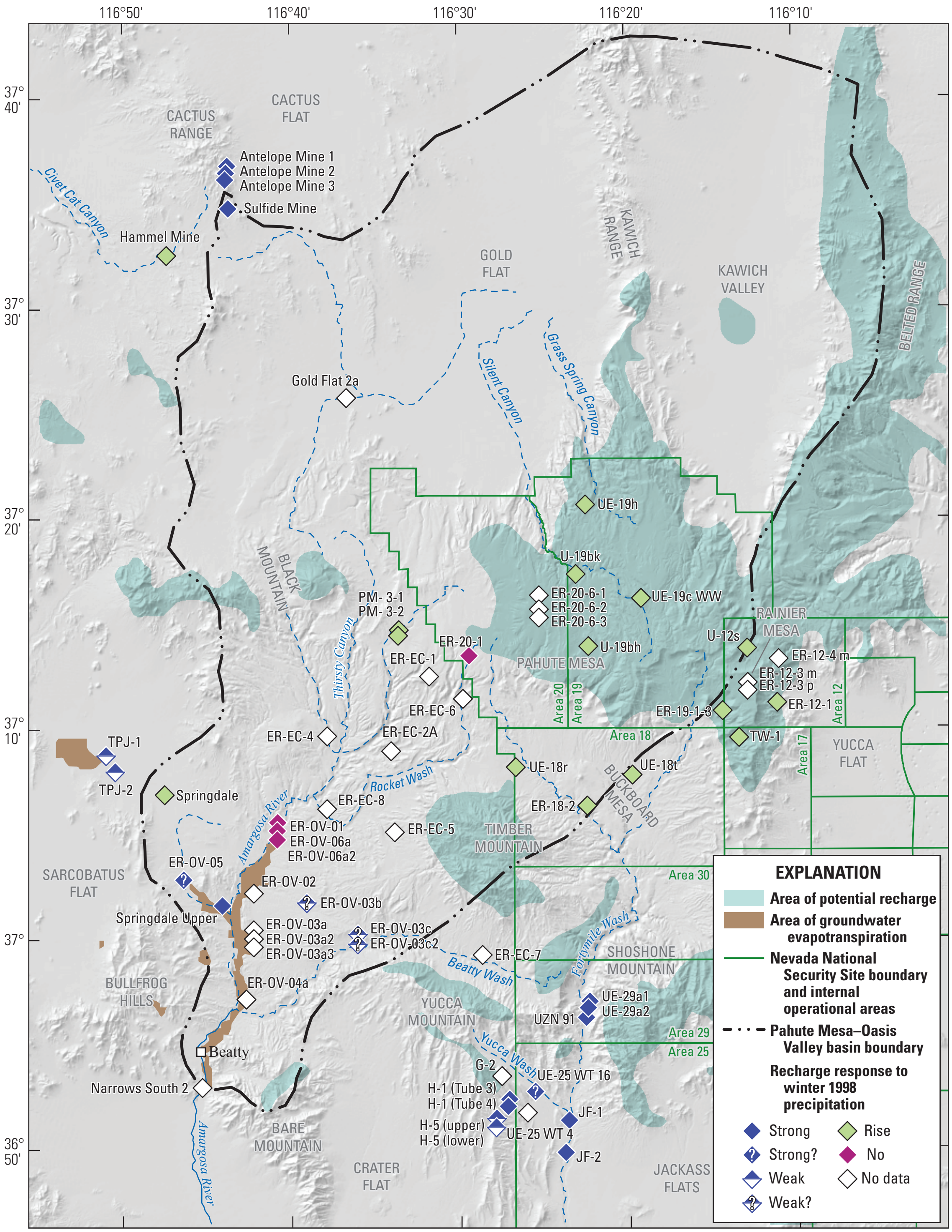
Figure 43. Relative water-level change in two Yucca Mountain wells, which have discernible recharge responses to five wetter-than-average winters, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada. Data are color coded by primary hydrogeologic unit screened in well: LFA, lava-flow aquifer and TCU, tuff confining unit. A wet winter in this graph is defined as exceeding 125 percent of average winter precipitation for the Beatty precipitation index.

Recharge responses to the 1995 and 1998 winters occur throughout the study area (figs. 44 and 45). About 45 percent of wells with steady-state trends were completed after the 1998 winter; therefore, recharge responses in these wells to the 1995 and 1998 winters could not be determined. In wells with water-level data, however, recharge responses to the 1995 and 1998 winters were determined for Rainier, Pahute, and Buckboard Mesas, the Cactus Range, Oasis Valley, Fortymile Wash, and Yucca Mountain.



Base from U.S. Geological Survey digital data, 1:100,000. Hillshade from U.S. Geological Survey 1-arc second National Elevation Data (NED). Universal Transverse Mercator Projection, Zone 11, North American Datum of 1983 (NAD83).

Figure 44. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 1995.



Base from U.S. Geological Survey digital data, 1:100,000. Hillshade from U.S. Geological Survey 1-arc second National Elevation Data (NED). Universal Transverse Mercator Projection, Zone 11, North American Datum of 1983 (NAD83).

Figure 45. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 1998.

Recharge responses to the 2000 winter were determined for wells with water-level altitudes of less than 4,800 ft (fig. 46). The strongest recharge responses occurred along Fortymile and Yucca washes, whereas weak responses were observed in wells near Beatty Wash and the Amargosa River. Recharge from the 2000 winter may have contributed to the continuous water-level rise in wells at eastern Pahute and Buckboard Mesas; the attenuated response is attributed to an unsaturated zone of more than 1,000 ft (table 7).

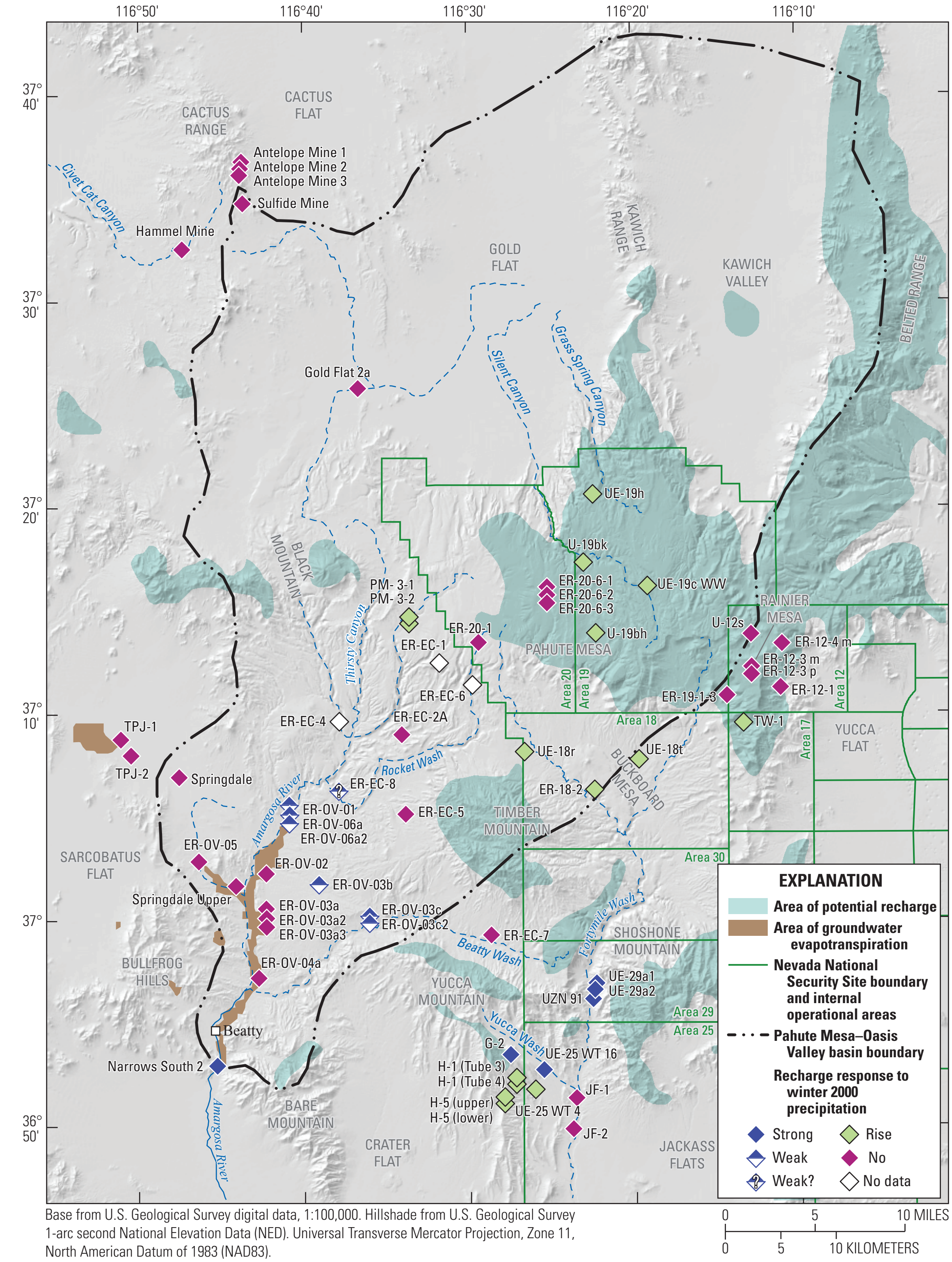


Figure 46. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 2000.

Recharge from the 2001, 2010, and 2011 winters is limited to localized regions in the study area. Recharge responses to the 2001 and 2010 winters occurred in the Cactus Range, northern Oasis Valley, the Amargosa Narrows, Fortymile Wash, and Yucca Mountain (figs. 47 and 48). Recharge responses to the 2011 winter occurred in Rainier, Pahute, and Buckboard Mesas (fig. 49).

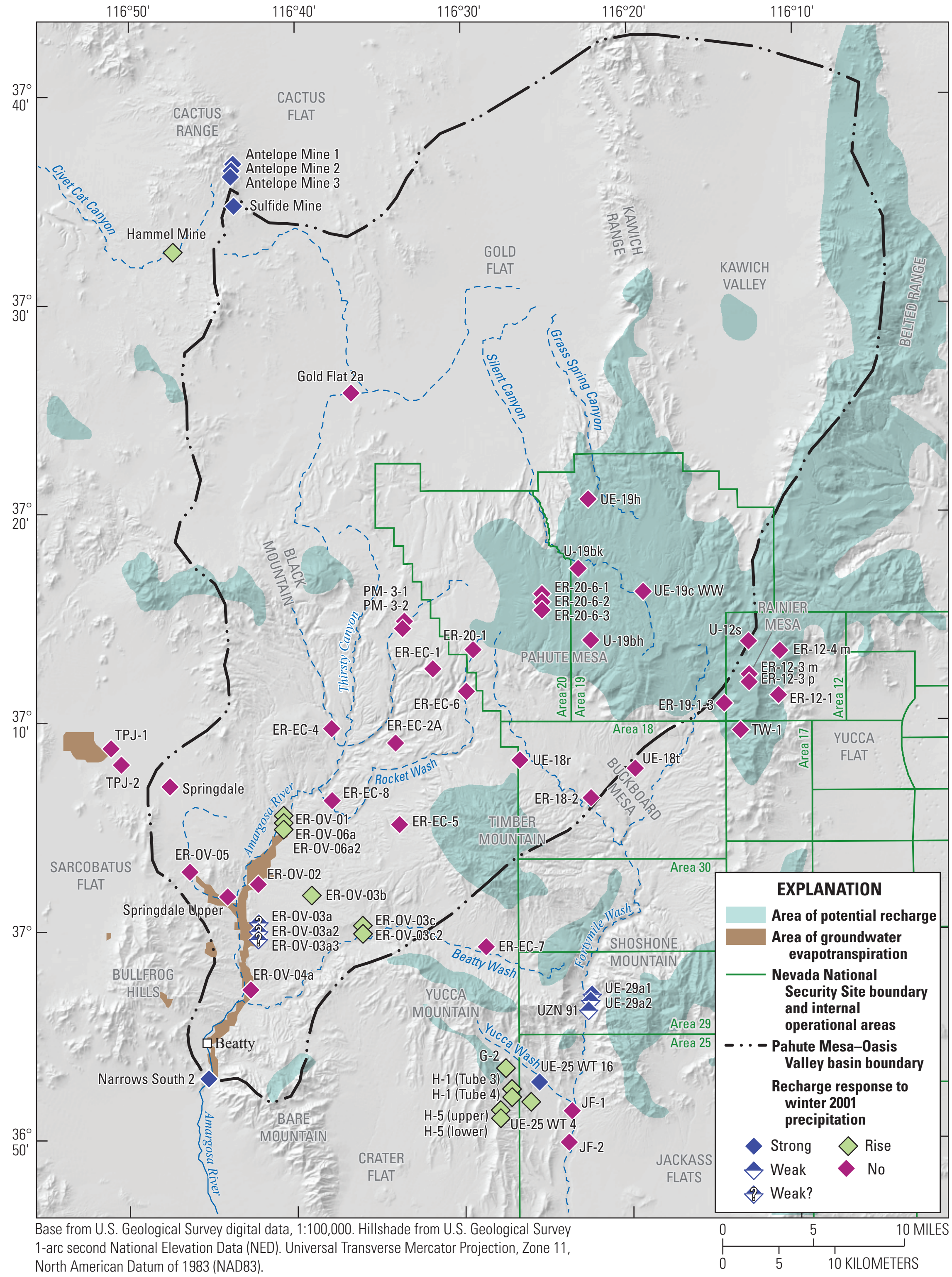


Figure 47. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 2001.

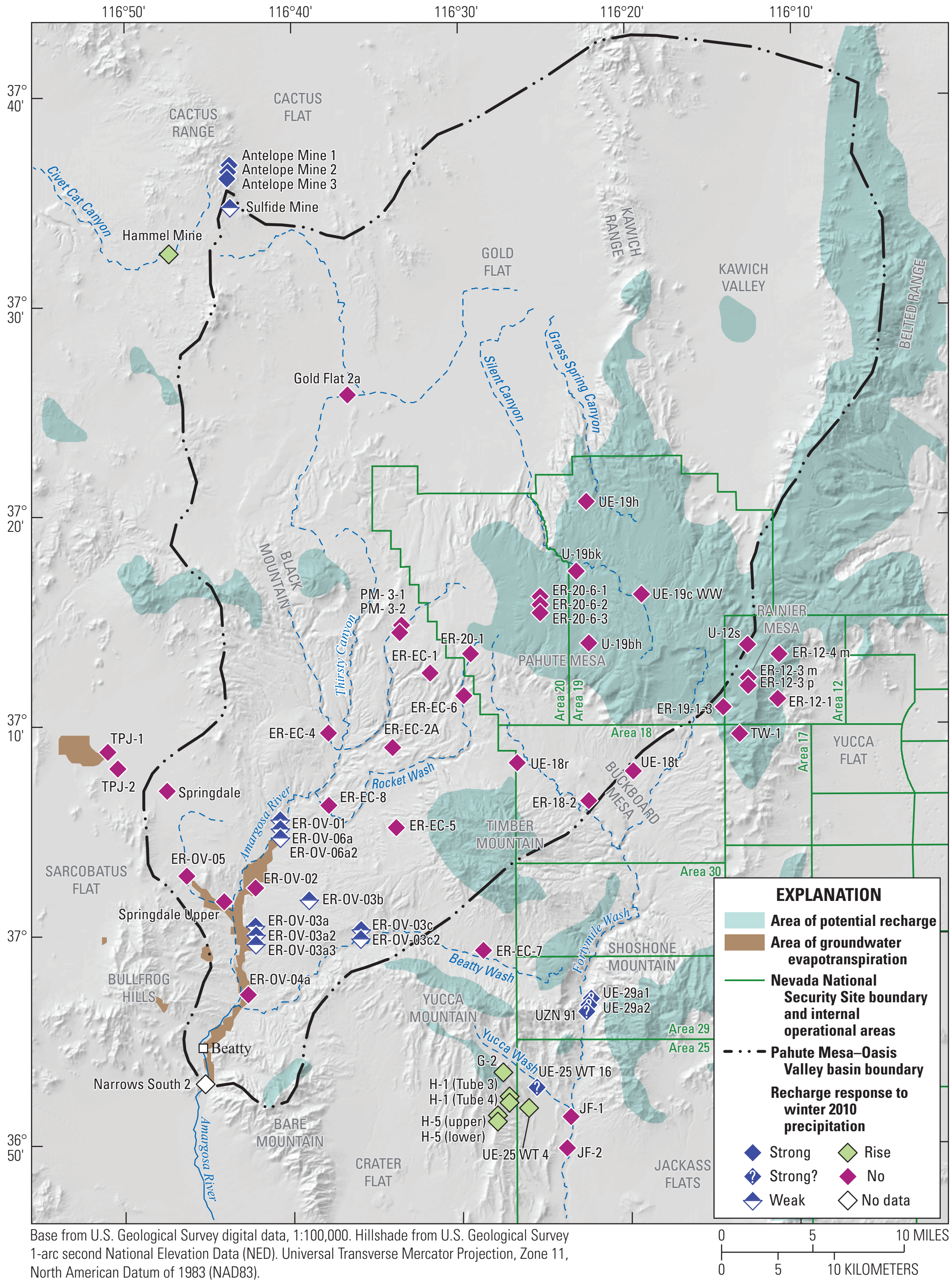


Figure 48. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 2010.

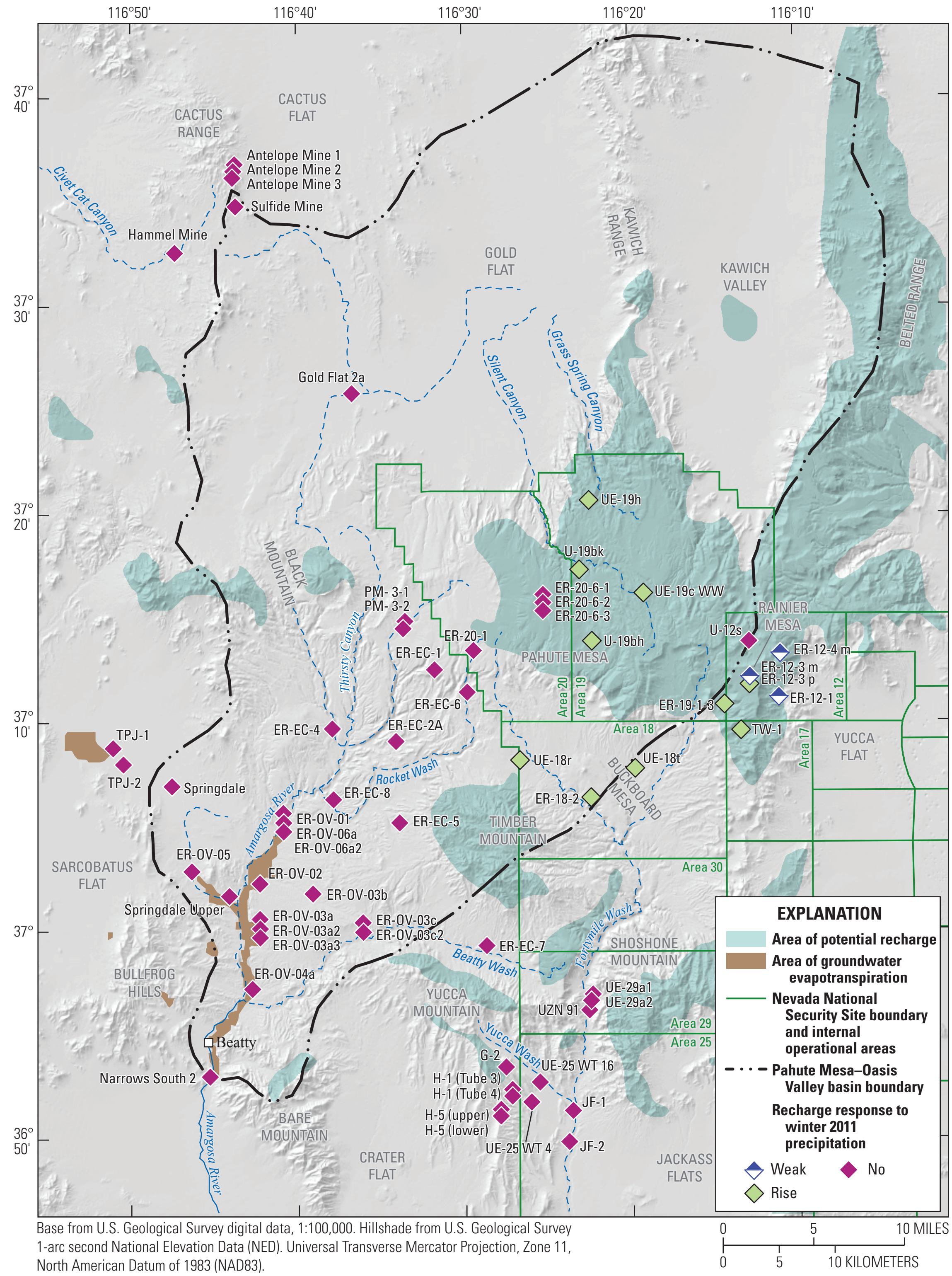
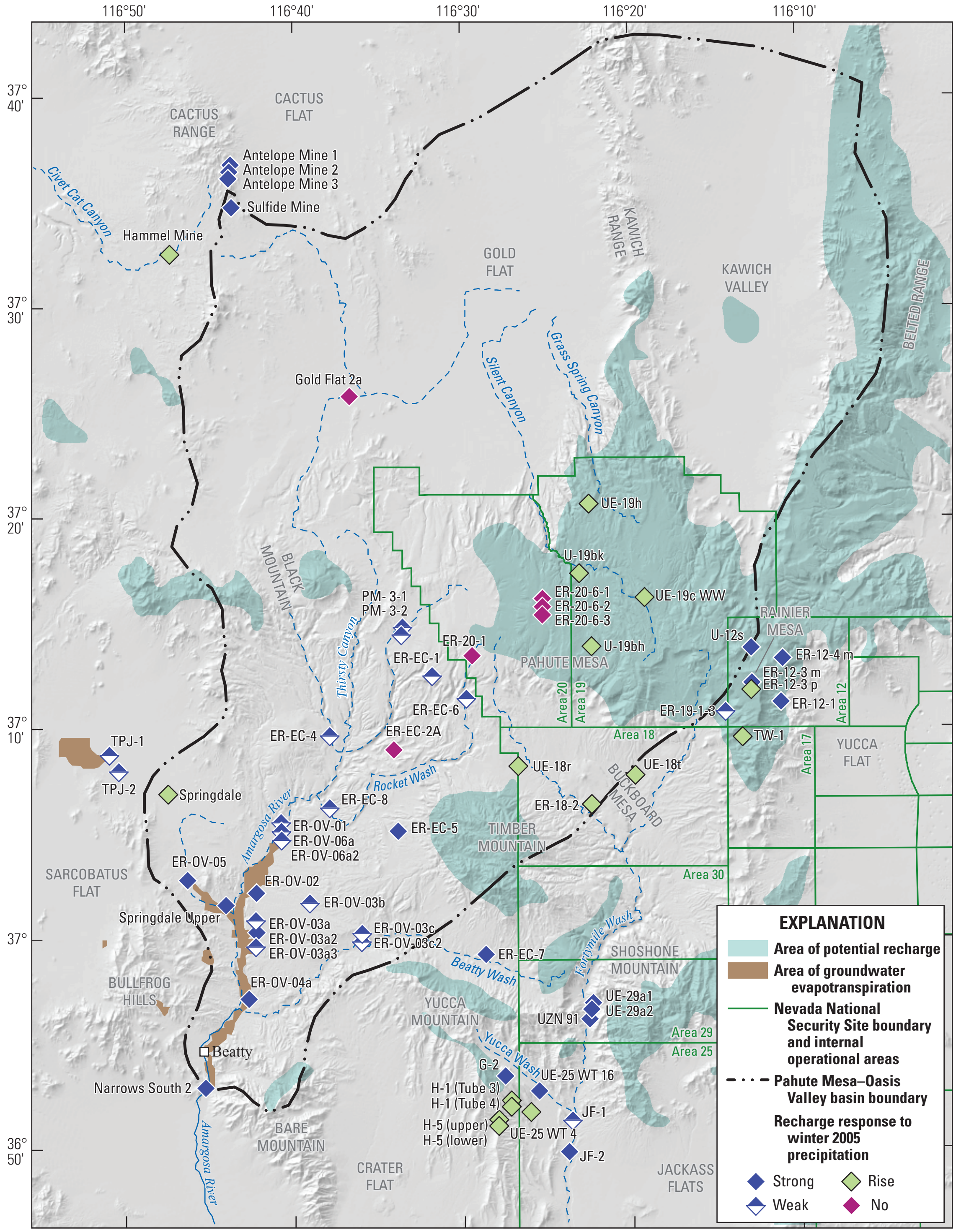


Figure 49. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 2011.

Recharge responses to the 2005 winter are ubiquitous (fig. 50). Recharge responses were determined for all geographic areas within the study area, except for wells within the Gold Flat and central PMOV basin geographic area. Water levels have either strong or weak discernible recharge responses throughout the study area. Only water levels in eastern Pahute and Buckboard Mesas and Yucca Mountain have attenuated rises, which are attributed to a thick unsaturated zone.



Base from U.S. Geological Survey digital data, 1:100,000. Hillshade from U.S. Geological Survey 1-arc second National Elevation Data (NED). Universal Transverse Mercator Projection, Zone 11, North American Datum of 1983 (NAD83).

Figure 50. Spatial distribution of recharge as determined from steady-state water-level responses in wells, Pahute Mesa–Oasis Valley groundwater basin and vicinity, Nevada, winter 2005.

Summary and Conclusions

Trends in groundwater levels were analyzed within and near the Pahute Mesa–Oasis Valley (PMOV) groundwater basin in Nye County, southern Nevada. The study objectives were to (1) identify trends in water levels in wells and (2) provide a conceptual framework that explains the hydrologic stresses and factors (or potential factors) causing the trends. Water levels in 79 wells were analyzed for trends between 1966 and 2016.

The framework consists of multiple stress-specific conceptual models to explain the water-level response to a hydrologic stress. A water-level trend reflects the summation of all hydrologic stresses affecting the aquifer at a well location. Hydrologic stresses affecting water levels include precipitation-derived recharge, evapotranspiration, pumping, nuclear testing, and water-level equilibration following localized disturbances in the wellbore. Dominant hydrologic stresses affecting water-level trends were used to categorize trends as either nonstatic, transient, or steady state.

Nonstatic water levels do not represent hydrologic conditions in the aquifer system. Instead, nonstatic levels represent equilibration of water in the borehole to water in the formation open to the well. The period of equilibration, referred to as wellbore equilibration, may take months, years, or decades because these wells are open to low-permeability units. Wellbore equilibration is observed as a steep exponential rise or decline in water levels. Equilibration may occur under non-pumping conditions after a localized disturbance to the wellbore, such as following drilling, well development, hydraulic testing, slug injection, dewatering the borehole, or bailing the well for sampling. Nonstatic water levels also are observed as large well losses (drawdowns) in a well during pumping.

Waters levels in four wells in the Pahute Mesa area are dominated by wellbore equilibration. Volcanic tuffs open to these wells have low hydraulic conductivities ranging from 2×10^{-4} to 0.01 feet per day, and the period of water-level recovery spanned from less than 1 to more than 20 years.

Transient trends are dominated by anthropogenic stresses, such as nuclear testing and groundwater pumping. The magnitude and duration of water-level responses to nuclear tests are dependent on rock hydraulic properties, distance from nuclear tests, nuclear-test yield, and the magnitude of earthquakes produced by the test. Water-level responses can occur near the nuclear-test cavity or miles from the point of detonation because of a breach scenario. In a breach scenario, a nuclear detonation hydraulically connects aquifers separated by a confining unit or hydraulic barrier. Following the breach, water levels in the two connected aquifers equilibrate to post-test conditions.

Water-level responses to nuclear testing have been observed in a few wells on Pahute Mesa. Most of the responses were relatively isolated and were observed in low-permeability rocks. Water levels in well *U-19v PS 1D* rose 860 feet (ft) over 36 years as groundwater slowly filled the cavity created by the ALMENDRO nuclear test. Water levels in wells *UE-20f* and *U-20ao* sharply rose about 50 and 39 ft, respectively, above pre-test water levels following detonations of nuclear tests within 3 miles of these wells. Six years following the nuclear test near *UE-20f*, water levels were nearly equilibrated to pre-test conditions. Water-level declines in wells *U-19bj* and *PM-2* potentially were affected by nuclear testing; however, the exact cause of water-level declines in these wells is uncertain. Water levels declined 12 ft in well *PM-2* following a nearby shallow crater test. Water levels declined for more than 20 years in well *U-19bj*, possibly because of a permanent lowering of the water table from a breach caused by a nuclear test.

Well *Beatty Wash Terrace* is the only study area well affected by long-term pumping in Beatty, Nevada. A water-level decline of about 1 ft over a 20-year period was estimated in the well from pumping in *Beatty Well 1*, about 2 mi away.

Water levels in six Pahute Mesa wells were affected by groundwater pumping from water-supply well *U-20 WW* and potentially were affected by nearby nuclear testing. Wells *U-20 WW*, *UE-20bh 1*, *UE-20n 1*, *U-20n PS 1DD-H*, *U-20bg*, and *U-20bf* have similar transient trends from about 1985 to 2016. These wells potentially were affected by nearby nuclear testing because water-level measurements in these wells from 1985 to 2000 show a declining trend of more than 10 ft that cannot be attributed to pumping from well *U-20 WW*. The declining trend potentially was caused by a hydraulic-barrier breach from a nearby nuclear test, which permanently lowered water levels in the aquifer system.

Steady-state water levels represent natural hydrologic conditions in the groundwater-flow system. This means that steady-state water levels are affected only by naturally occurring hydrologic stresses, such as recharge and evapotranspiration. Natural stresses cause steady-state water-level fluctuations over years to decades in the study area. However, by definition, steady state means that water levels do not change with time.

A conceptual model reconciles the definition of “steady state” as an unchanging condition with steady-state water levels that fluctuate with time. The conceptual model for steady-state trends in the PMOV basin assumes that water levels remain steady over a defined steady-state timescale, but may fluctuate over shorter timescales because of varying natural hydrologic stresses. In a steady-state groundwater system, long-term water levels are in a state of dynamic equilibrium, where long-term cumulative recharge is balanced by long-term cumulative discharge and the net change in long-term cumulative storage is zero.

Water-level data indicate that the steady-state timescale is more than 25 years, because water levels have been rising in the study area and throughout southern Nevada during this period. Therefore, the steady-state timescale, where water levels remain constant, must be more than 25 years. The conceptual model assumes a timescale of about a century. The timescale is tested using a precipitation data set from 1900 to 2016 to determine if recent rising water-level trends can be explained within the context of long-term steady-state conditions.

A hypothetical water-level record was constructed to explain observed rising water-level trends using the assumed century-scale period of steady state. Hypothetical recharge was determined by applying a threshold to winter (October–March) precipitation data, where winter precipitation above the threshold was assumed to recharge the groundwater system. The hypothetical record shows a declining trend from 1900 to 1968 and a rising trend from 1968 to 2016. Assuming that steady state occurs on a century timescale, measured water-level trends in the study area from 1995 to 2016 should be upward because the study area has been in a relatively wet period since 1968. The magnitude and exact pattern of water-level trends is expected to differ between the hypothetical water-level record and water-level records in study area wells because of differing rock porosities, recharge rates, and aquifer discharge rates.

The conceptual model of steady-state trends assumes that the groundwater system receives modern recharge. Thick unsaturated zones in Pahute Mesa, Rainier Mesa, and Yucca Mountain do not preclude recharge. Chemical and isotopic analyses from previous studies show that, even though the *primary source* of regional groundwater in the study area is derived from a colder climatic period (end of the Pleistocene), mixing of young groundwater with regional (older) groundwater occurs throughout the study area, including areas where the water table is deep.

Graphical, statistical, and numerical methods were used to identify and analyze steady-state trends from 1995 to 2016, a period where most wells have a consistent record. Maximum magnitudes of water-level change ranged between 0.2 and 44 ft. Graphical and statistical analyses indicate that 43 of the 62 wells with steady-state trends have significant upward trends, 6 have downward trends, and 13 have no trend. Steady-state trends in four wells were simulated using water-level models to demonstrate that the trends can be explained entirely by episodic recharge.

Wells with steady-state trends were grouped into 11 geographic areas in the study area, based on water-level responses to episodic recharge and other factors affecting the trends. Factors that influenced water-level responses to recharge include unsaturated zone depth, hydrogeologic unit screened in the open interval of a well, transmissivity, and distance to potential recharge areas. Water levels respond similarly to recharge in each geographic area.

Groundwater recharge is temporally and spatially variable in the study area. Water levels responded to recharge from the 1995, 1998, 2000, 2001, 2005, 2010, and (or) 2011 winters. Recharge responses to the 1995, 1998, and 2005 winters were ubiquitous. Recharge responses to the 2000 winter were observed in wells with water-level altitudes of less than 4,800 ft. Recharge responses from the 2001, 2010, and 2011 winters are limited to localized regions in the study area.

The trend analysis links water-level fluctuations in wells to hydrologic stresses and potential factors causing the trends to better understand and conceptualize the groundwater-flow system. Nonstatic water levels are important to recognize to avoid misinterpretation of water-level trends as representing regional groundwater conditions. Transient and steady-state trend categorizations can be used to guide future groundwater studies on the use of specific water-level data in the development of potentiometric maps, or on the appropriate water-level data to use in steady-state and transient groundwater-flow models. The sub-categorization of transient trends into trends affected by nuclear testing, pumping, or both is useful because nuclear testing effects on water levels is a unique phenomenon important to groundwater studies on the Nevada National Security Site. Steady-state trends were grouped by geographic areas to show that trends vary spatially in the study area and the variability in trends is attributed to the temporal and spatial variability in recharge. The conceptual framework of water-level responses to hydrologic stresses and trend analyses provide a comprehensive understanding of how water levels respond to natural and anthropogenic stresses in the PMOV groundwater basin.

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Appendix 1. Supplemental Notes for Selected Wells

ER-20-1

Well *ER-20-1* has a declining trend that is not attributed to pumping. Theis transforms were generated using the pumping schedule from nearby (4 mi away) water-supply well *U-20 WW*. Water-level declines were not correlated with years of pumping from well *U-20 WW*.

ER-20-6-3

A multiple-well aquifer test was done at water-supply well *U-20 WW* from October 1, 2008, to October 1, 2009 (Garcia and others, 2011). About 14 acre-ft of groundwater were withdrawn from the well during aquifer testing. Water levels were continuously monitored in well *ER-20-6-3*; however, no drawdowns were detected (Garcia and others, 2011). A water-level model was used to determine whether observation well *ER-20-6-3* has been affected by prior pumping in production well *U-20 WW*.

Measured water levels in well *ER-20-6-3* do not respond to pumping from water-supply well *U-20 WW*. Well *ER-20-6-3* is about 0.7 mi northeast of the water-supply well. A synthetic water-level curve was generated from Theis transforms of the *U-20 WW* pumping schedule. No drawdown responses were observed from pumping in well *U-20 WW*. Well *ER-20-6-3* occurs in the same structural block as water-supply well *U-20 WW*, and both wells are screened in transmissive lava-flow aquifers. Garcia and others (2011) concluded that wells *ER-20-6-3* and *U-20 WW* likely are separated by a low permeability volcanic tuff confining unit. No discernible drawdown responses were observed from nearby pumping, and *U-20 WW* pumping likely did not affect water levels in well *ER-20-6-3*.

PM-2

Water levels were flagged as nonstatic from 1966 to 1968 because of bailing the well dry. Water levels were equilibrating to the formation open to the well. Water-level recovery occurs over many years because the well is open to low transmissivity volcanic rocks (Elliott and Fenelon, 2010).

Water levels were flagged as nonstatic from 1983 to 2006 because precipitation, in the form of rain and (or) snow, was leaking down the well annulus. Leakage of water down the well occurred because the nearby shallow crater nuclear test, SCHOONER, buried the well beneath 10 ft of sediment ejected by the nuclear explosion (Russell and Locke, 1997). When the ejected sediment was removed from the vicinity of well *PM-2*, a depression was formed around the well and the top of the well casing was at land surface. Water levels fluctuated as much as 2 ft between 1983 and 1992 because of the leakage of small amounts of water from rain or snow. The 1992 winter was wet and leakage of precipitation down the well annulus caused water levels to rise about 16 ft between May 1992 and May 1993. In 1993, the depression was filled with 7 ft of earth material and a casing extension was added onto well *PM-2* to prevent leakage down the well annulus (Elliott and Fenelon, 2010). Water levels declined from 1994 to 2006 and equilibrated to steady state by November 2006.

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Director, Nevada Water Science Center
U.S. Geological Survey
2730 N. Deer Run Rd.
Carson City, Nevada 89701
https://www.usgs.gov/centers/nv-water

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