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Regional Groundwater-Flow Model of the Redwall-Muav, Coconino, and Alluvial Basin Aquifer Systems of Northern and Central Arizona

Scientific Investigations Report 2010-5180, v. 1.1

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

Front Cover

Cathedral Rock, Sedona, Arizona. View southeast from Oak Creek near Red Rock Crossing.
(Photo courtesy of Ken Thomas.)

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U.S. Department of the Interior
KEN SALAZAR, Secretary

U.S. Geological Survey
Marcia K. McNutt, Director

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Conversion Factors

Inch/Pound to SI

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)
mile, nautical (nmi)	1.852	kilometer (km)
yard (yd)	0.9144	meter (m)
Area		
acre	4,047	square meter (m^2)
acre	0.4047	hectare (ha)
acre	0.4047	square hectometer (hm^2)
acre	0.004047	square kilometer (km^2)
square foot (ft^2)	929.0	square centimeter (cm^2)
square foot (ft^2)	0.09290	square meter (m^2)
section (640 acres or 1 square mile)	259.0	square hectometer (hm^2)
square mile (mi^2)	259.0	hectare (ha)
square mile (mi^2)	2.590	square kilometer (km^2)
Volume		
gallon (gal)	3.785	liter (L)
gallon (gal)	0.003785	cubic meter (m^3)
gallon (gal)	3.785	cubic decimeter (dm^3)
million gallons (Mgal)	3,785	cubic meter (m^3)
cubic inch (in^3)	16.39	cubic centimeter (cm^3)
cubic inch (in^3)	0.01639	cubic decimeter (dm^3)
cubic inch (in^3)	0.01639	liter (L)
cubic foot (ft^3)	28.32	cubic decimeter (dm^3)
cubic foot (ft^3)	0.02832	cubic meter (m^3)
cubic yard (yd^3)	0.7646	cubic meter (m^3)
cubic mile (mi^3)	4.168	cubic kilometer (km^3)
acre-foot (acre-ft)	1,233	cubic meter (m^3)
acre-foot (acre-ft)	0.001233	cubic hectometer (hm^3)
Flow rate		
acre-foot per day (acre-ft/d)	0.01427	cubic meter per second (m^3/s)
acre-foot per year (ac-ft/yr)	1,233	cubic meter per year (m^3/yr)

acre-foot per year (ac-ft/yr)	0.001233	cubic hectometer per year (hm^3/yr)
foot per second (ft/s)	0.3048	meter per second (m/s)
foot per minute (ft/min)	0.3048	meter per minute (m/min)
foot per hour (ft/hr)	0.3048	meter per hour (m/hr)
foot per day (ft/d)	0.3048	meter per day (m/d)
foot per year (ft/yr)	0.3048	meter per year (m/yr)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
cubic foot per second per square mile [$(ft^3/s)/mi^2$]	0.01093	cubic meter per second per square kilometer [$(m^3/s)/km^2$]
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
gallon per day per square mile [$(gal/d)/mi^2$]	0.001461	cubic meter per day per square kilometer [$(m^3/d)/km^2$]
million gallons per day (Mgal/d)	0.04381	cubic meter per second (m^3/s)
million gallons per day per square mile [$(Mgal/d)/mi^2$]	1,461	cubic meter per day per square kilometer [$(m^3/d)/km^2$]
inch per hour (in/h)	0 .0254	meter per hour (m/h)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
	Specific capacity	
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter [(L/s)/m]
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Hydraulic gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer (m/km)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)
	Leakance	
foot per day per foot [(ft/d)/ft]	1	meter per day per meter
inch per year per foot [(in/yr)/ft]	83.33	millimeter per year per meter [(mm/yr)/m]

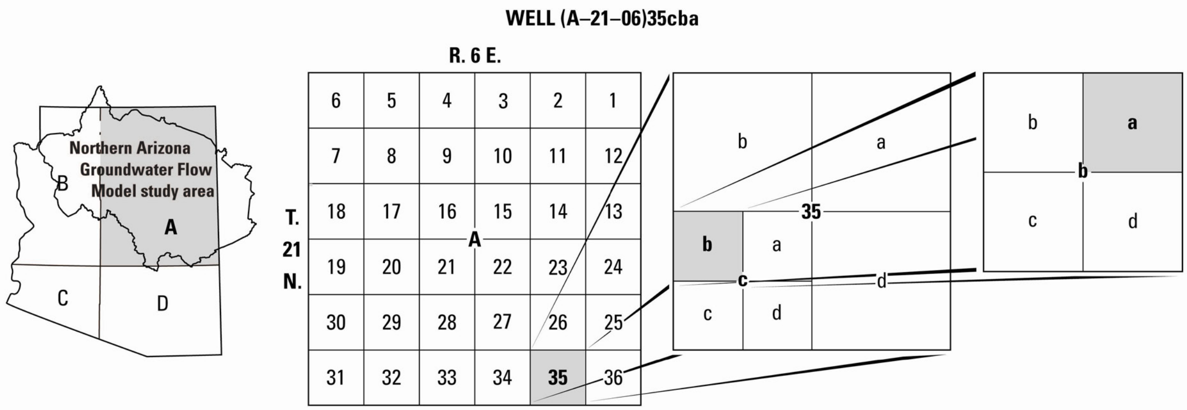
SI to Inch/Pound

Multiply	By	To obtain
	Length	
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
	Area	
square meter (m^2)	0.0002471	acre
hectare (ha)	2.471	acre
square hectometer (hm^2)	2.471	acre
square kilometer (km^2)	247.1	acre
square meter (m^2)	10.76	square foot (ft^2)
hectare (ha)	0.003861	square mile (mi^2)
square kilometer (km^2)	0.3861	square mile (mi^2)
	Volume	
cubic decimeter (dm^3)	0.03531	cubic foot (ft^3)
cubic meter (m^3)	35.31	cubic foot (ft^3)
cubic meter (m^3)	0.0008107	acre-foot (acre-ft)
cubic hectometer (hm^3)	810.7	acre-foot (acre-ft)
	Flow rate	
cubic meter per second (m^3/s)	70.07	acre-foot per day (acre-ft/d)
cubic meter per year (m^3/yr)	0.000811	acre-foot per year (ac-ft/yr)
cubic hectometer per year (hm^3/yr)	811.03	acre-foot per year (ac-ft/yr)
meter per second (m/s)	3.281	foot per second (ft/s)
meter per day (m/d)	3.281	foot per day (ft/d)
cubic meter per second (m^3/s)	35.31	cubic foot per second (ft^3/s)
cubic meter per day (m^3/d)	35.31	cubic foot per day (ft^3/d)
liter per second (L/s)	15.85	gallon per minute (gal/min)
	Specific capacity	
liter per second per meter [(L/s)/m]	4.831	gallon per minute per foot [(gal/min)/ft]
	Hydraulic conductivity	
meter per day (m/d)	3.281	foot per day (ft/d)
	Hydraulic gradient	
meter per kilometer (m/km)	5.27983	foot per mile (ft/mi)
	Transmissivity*	
meter squared per day (m^2/d)	10.76	foot squared per day (ft^2/d)
	Leakance	
meter per day per meter [(m/d)/m]	1	foot per day per foot [(ft/d)/ft]

Notes on Conversion Factors

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:
°F=(1.8×°C)+32
Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:
°C=(°F-32)/1.8
Vertical coordinate information is referenced to the insert datum name (and abbreviation) here, for instance, "North American Vertical Datum of 1988 (NAVD 88)"
Horizontal coordinate information is referenced to the insert datum name (and abbreviation) here, for instance, "North American Datum of 1983 (NAD 83)"
Altitude, as used in this report, refers to distance above the vertical datum.
*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.
Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25°C).
Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).
NOTE TO USGS USERS: Use of hectare (ha) as an alternative name for square hectometer (hm^2) is restricted to the measurement of small land or water areas. Use of liter (L) as a special name for cubic decimeter (dm^3) is restricted to the measurement of liquids and gases. No prefix other than milli should be used with liter. Metric ton (t) as a name for megagram (Mg) should be restricted to commercial usage, and no prefixes should be used with it.

ARIZONA WELL-NUMBERING SYSTEM



Quadrant A, Township 21 North, Range 6 East, section 35,
quarter section c, quarter section b, quarter section a

The well numbers used by the U.S. Geological Survey in Arizona are in accordance with the Bureau of Land Management's system of land subdivision. The land survey in Arizona is based on the Gila and Salt River meridian and base line, which divide the State into four quadrants and are designated by capital letters A, B, C, and D in a counterclockwise direction beginning in the northeast quarter. The first digit of a well number indicates the township, the second the range, and the third the section in which the well is situated. The lowercase letters a, b, c, and d after the section number indicate the well location within the section. The first letter denotes a particular 160 -acre tract, the second the 40-acre tract and the third the 10-acre tract. These letters also are assigned in a counterclockwise direction beginning in the northeast quarter. If the location is known within the 10-acre tract, three lowercase letters are shown in the well number. Where more than one well is within a 10-acre tract, consecutive numbers beginning with 1 are added as suffixes. In the example shown, well number (A–21–06)35cba designates the well as being in the NE1/4, NW1/4, SW1/4, section 35, Township 21 North, and Range 6 East.

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Abstract

A numerical flow model (MODFLOW) of the groundwater-flow system in the primary aquifers in northern Arizona was developed to simulate interactions between the aquifers, perennial streams, and springs for predevelopment and transient conditions during 1910 through 2005. Simulated aquifers include the Redwall-Muav, Coconino, and basin-fill aquifers. Perennial stream reaches and springs that derive base flow from the aquifers were simulated, including the Colorado River, Little Colorado River, Salt River, Verde River, and perennial reaches of tributary streams. Simulated major springs include Blue Spring, Del Rio Springs, Havasu Springs, Verde River headwater springs, several springs that discharge adjacent to major Verde River tributaries, and many springs that discharge to the Colorado River. Estimates of aquifer hydraulic properties and groundwater budgets were developed from published reports and groundwater-flow models. Spatial extents of aquifers and confining units were developed from geologic data, geophysical models, a groundwater-flow model for the Prescott Active Management Area, drill logs, geologic logs, and geophysical logs. Spatial and temporal distributions of natural recharge were developed by using a water-balance model that estimates recharge from direct infiltration. Additional natural recharge from ephemeral channel infiltration was simulated in alluvial basins. Recharge at wastewater treatment facilities and incidental recharge at agricultural fields and golf courses were also simulated. Estimates of predevelopment rates of groundwater discharge to streams, springs, and evapotranspiration by phreatophytes were derived from previous reports and on the basis of streamflow records at gages. Annual estimates of groundwater withdrawals for agriculture, municipal, industrial, and domestic uses were developed from several sources, including reported withdrawals for nonexempt wells, estimated crop requirements for agricultural wells, and estimated per-capita water use for exempt wells. Accuracy of the simulated groundwater-flow system was evaluated by using observational control from water levels in wells, estimates of base flow from streamflow records, and estimates of spring discharge.

Major results from the simulations include the importance of variations in recharge rates throughout the study area and recharge along ephemeral and losing stream reaches in alluvial basins. Insights about the groundwater-flow systems in individual basins include the hydrologic influence of geologic structures in some areas and that stream-aquifer interactions along the lower part of the Little Colorado River are an effective control on water level distributions throughout the Little Colorado River Plateau basin.

Better information on several aspects of the groundwater-flow system are needed to reduce uncertainty of the simulated system. Many areas lack documentation of the response of the groundwater system to changes in withdrawals and recharge. Data needed to define groundwater flow between vertically adjacent water-bearing units is lacking in many areas. Distributions of recharge along losing stream reaches are poorly defined. Extents of aquifers and alluvial lithologies are poorly defined in parts of the Big Chino and Verde Valley sub-basins. Aquifer storage properties are poorly defined throughout most of the study area. Little data exist to define the hydrologic importance of geologic structures such as faults and fractures. Discharge of regional groundwater flow to the Verde River is difficult to identify in the Verde Valley sub-basin because of unknown contributions from deep percolation of excess surface water irrigation.

Introduction

In 1999, the Arizona Department of Water Resources (ADWR) started the Rural Watershed Initiative (RWI), a program that addresses water-supply issues in increasingly populated rural areas, with an emphasis on regional watershed studies. The program encourages the development of partnerships between local stakeholders and resource agencies, such as the U.S. Geological Survey (USGS), to develop information needed to support resource planning and management decisions. The Arizona Water Science Center (AZWSC) of the USGS, in cooperation with ADWR, has completed three initial RWI studies focusing on the hydrogeologic framework and conceptual understanding of groundwater resources in northern and central Arizona (fig. 1). The three completed RWI studies include the Coconino Plateau (Bills and others, 2007), the upper and middle Verde River watersheds (Blasch and others, 2006), and the Mogollon Highlands (Parker and others, 2005). These three study areas have had, or likely will have, rapid population growth and increased use of groundwater supplies. A numerical groundwater-flow model of the region that includes the area of the RWI studies was deemed necessary so that future investigators can assess the effect of anticipated increased use of groundwater.

The Northern Arizona Regional Groundwater-Flow Model (NARGFM), described and documented in this report, was developed to help assess the adequacy of the regional groundwater supply and potential for the effects of increased groundwater use on water levels, streamflow, and riparian vegetation. Hydrologic information and understanding gained during initial RWI studies was used to develop the groundwater-flow model. The model can be used by resource managers to examine the hydrologic consequences of various groundwater development and climate change scenarios. A regional groundwater-flow model was developed rather than individual models of administratively defined groundwater basins or sub-basins because groundwater flow is continuous through aquifers that cross boundaries of the groundwater basins and increases in groundwater withdrawals in one basin can potentially capture groundwater flow from adjacent basins. Only a regional model can simulate the effect of changes in any basin or sub-basin on another. Simulation of regional groundwater flow does not diminish the ability to simulate groundwater flow in individual basins or sub-basins. Accurate simulation of groundwater flow in any sub-area of the regional model is dependent on the availability of information defining the local hydrogeologic system and changes in the system.

The regional numerical model simulates groundwater flow in the primary aquifers of the region, including the Redwall-Muav (R-aquifer) and Coconino (C-aquifer) aquifers on the Colorado Plateau and alluvial and volcanic aquifers in several basins that lie adjacent to the southern extent of the Colorado Plateau. Groundwater flow also is simulated in rocks that underlie and lie adjacent to the primary aquifers of the alluvial basins. The Redwall-Muav aquifer has been referred to as the R-aquifer. The term “Redwall-Muav aquifer” is used in this report. The Coconino aquifer has been referred to as the C-aquifer, C multiple-aquifer system (Cooley and others, 1969), the regional aquifer (Levings, 1980; Owen-Joyce and Bell, 1983; and Bills and others, 2000), and the Coconino aquifer (Mann, 1976; McGavock, 1968; McGavock and others, 1986; and Hart and others, 2002). The term “Coconino aquifer” is used in this report. Aquifers in alluvial basins have been referred to as stream alluvium and upper and lower basin fill (Anderson and others, 1992), alluvium and basin fill (Parker and others, 2005), upper alluvial unit and lower volcanic unit (Corkill and Mason, 1995), and Verde Formation (Twenter and Metzger, 1963). The terms “upper and lower basin fill” are used in this report.

The numerical groundwater model has two primary uses: (1) evaluation of the hydrologic effects of groundwater use on the groundwater-flow system and (2) identification of major hydrogeologic parameters that need improved definition. The model can be used to estimate changes in the water levels and discharge to streams, springs, and riparian evapotranspiration that may result from anticipated future groundwater use and management practices. However, the certainty of projected change is dependent on future validation of the hydrologic assumptions that are inherent in the model. The model can be used by resource managers to examine the hydrologic consequences of various groundwater development and climate-change scenarios for regions that are sub-basin or larger in area. Use of the model for site-scale investigations may require additional data to better define the local hydrogeology. The model can also provide information that should help identify data needs and refine future studies for improved simulation of the hydrologic effects of groundwater use.

For the purpose of estimating magnitude and timing of change in water levels and discharge resulting from an imposed stress using this or any groundwater-flow model, the only hydrologic parameters that are of importance are the aquifer properties of transmissivity and storage (Leake, 2011). Those properties influence the rate of propagation of changes in groundwater flow through the aquifer and release of groundwater from storage. Variations in recharge rates—natural, artificial, or incidental—can cause change in water levels and discharge to streams, but are not fundamental variables that affect the calculation of human-induced change in the model. Rather, the effects of recharge variations are independent of the effects of groundwater use and management practices, and are superimposed on these other effects. The superimposed effects of variations in recharge rates can enhance or counter the effects of groundwater use and management practices. Other hydrologic parameters that have no influence on changes in water levels and discharge to streams include directions of groundwater flow and sources of water that contribute groundwater discharge to streams.

The groundwater flow model itself, however, must consider properties and conditions that are not essential in computing effects of human activities on an aquifer system. In constructing the flow model, an attempt was made to reasonably represent (1) spatial and temporal recharge distributions, (2) transmissivity distributions, (3) distributions of hydraulic properties that control interactions between vertically adjacent aquifers, (4) spatial distributions of withdrawals and incidental recharge rates, (5) aquifer lateral and vertical extents, and (6) hydrologic barriers and conduits. It should be noted that estimation of item 2, transmissivity distributions, using the model requires independent knowledge of item 1, recharge distributions. Although this study used the most comprehensive information on these items currently available, future studies may result in an improved understanding of the groundwater-flow system that could substantially alter the fundamental conceptual hydrogeologic model in some areas. The simulated hydrologic system should not be considered a replication of the true system, but a simulation of the system as currently understood by the modeler and simplified for use in a numerical model. For additional discussion of model assumptions and limitations, please refer to the section titled “Model Applicability, Limitations, and Suggestions for Future Work.”

Purpose and Scope

The purpose of this report is to document the development of a groundwater-flow model that simulates the groundwater-flow system that existed in northern and central Arizona before development of groundwater supplies, pre-1938, and the flow system that occurred during development of the resource, 1938–2006. The groundwater-flow system of northern and central Arizona includes the basins of the Verde, Salt, and Little Colorado Rivers and adjacent areas of the western Coconino Plateau (fig. 1). Within the regional extent of the model there are several areas where management of groundwater resources are of interest. Use of the model for assisting management decisions is of particular interest for several areas. These areas include the Little Chino, Big Chino, and Verde Valleys, where groundwater discharge to the Verde River and potential capture of groundwater flow from adjacent areas including the Coconino Plateau and Little Colorado River Plateau basins is a primary interest. Included in the report are discussions of pertinent background information on the hydrology of the study area, modeling procedure, and results of steady-state and transient simulations.

Objectives

The primary objectives of the study were to (1) develop a groundwater-flow model that represents the regional groundwater-flow system and changes in the system, (2) evaluate the current concepts of groundwater flow in the region, (3) identify deficiencies in data available to define the hydrogeologic system, (4) provide boundary conditions for local-scale models, and (5) provide a tool that can be used to help assess changes in the groundwater system that may result from groundwater use strategies. Another major objective was to compile a hydrologic database that could be accessed for future hydrologic studies and hydrologic model development. Because of the interest in simulation of the groundwater-flow system in the Verde River basin and adjacent areas, particular care was taken in simulating details of the system in that area. A more generalized simulation was considered adequate for the groundwater-flow system near the northern, eastern, and western boundaries of the study area because of a paucity of data.

Approach

A groundwater-flow model of multiple interconnected regional and sub-regional aquifers in the central and northern Arizona region was developed by first constructing a hydrogeologic framework and a conceptual groundwater-flow model. Available data were used to define the extent of the aquifers, predevelopment conditions, and changes that have occurred with development of groundwater supplies. The hydrogeologic framework and a conceptual groundwater-flow model were developed by compiling three-dimensional information on the geologic units, water levels, hydraulic properties of the primary aquifers, and groundwater flow to streams and springs. Changes in the groundwater system were defined by historical groundwater and surface-water use, variations in water levels in wells, and variations in groundwater discharge to streams and springs. Several simplifications and approximations of the complex hydrogeologic system were needed to handle deficiencies in data and minimal understanding of the groundwater-flow system in some areas and to produce a representation of the groundwater-flow system that could be simulated numerically. The accuracy of the simulated groundwater-flow system in any sub-region of the modeled area is dependent on the quantity and quality of available data and hydrologic understanding of the flow system.

The groundwater-flow system was simulated using the USGS finite-difference groundwater model program MODFLOW. Construction of the model was facilitated using ArcGIS ESRI® Geographic Information System and Aquaveo GMS® graphical interface. The model runs on standard versions of MODFLOW-2000 and MODFLOW-2005 available from the USGS. Several MODFLOW Packages were used including, Recharge (RCH), Layer-Property Flow (LPF), Streams (STR), Drains (DRN), Evapotranspiration (EVT), Wells (WEL), and the solver Package Preconditioned Conjugate Gradient (PCG2). The RCH Package simulates distributions of recharge rates across the model domain. For details, see the “Inflows and Outflows” section. The LPF Package simulates the flow of groundwater through porous media from recharge areas to discharge at streams, drains, evapotranspiration through plants, or wells. For details, see the “Hydraulic and Storage Properties” section. Outflow from the groundwater-flow system is simulated using the STR, DRN, EVT, and WEL Packages. For details, see the “Inflows and Outflows” section.

The groundwater-flow model is discretized spatially and through time. The hydrogeologic framework was discretized in layered grids with a uniform grid-spacing of 0.62 mi (1km). The grid was rotated 60° in the counterclockwise direction so that columns of the model grid are approximately parallel and rows are approximately normal to regional geologic structure. The vertical discretization was limited to three layers, representing hydrogeologic units including alluvial deposits, sedimentary rocks, carbonate rocks, and crystalline rocks. Units represented in each layer vary by location within the model domain. For details, see the “Model Framework” section. The model simulates predevelopment conditions that are assumed to be steady state, followed by nine transient stress periods that encompass the period 1910–2005. Although the simulated grid spacing and time discretization includes enough detail for regional and subregional analyses, refinements may be required to improve the simulation for some subregional studies.

Several groundwater-flow model parameters were adjusted in hydrologically reasonable limits to calibrate the simulated groundwater-flow system against observations of water levels in wells and discharge to streams and springs. Parameters that were allowed to adjust during the calibration process included distributions of hydraulic conductivity, vertical anisotropy, specific storage, specific yield, streambed conductivity, and

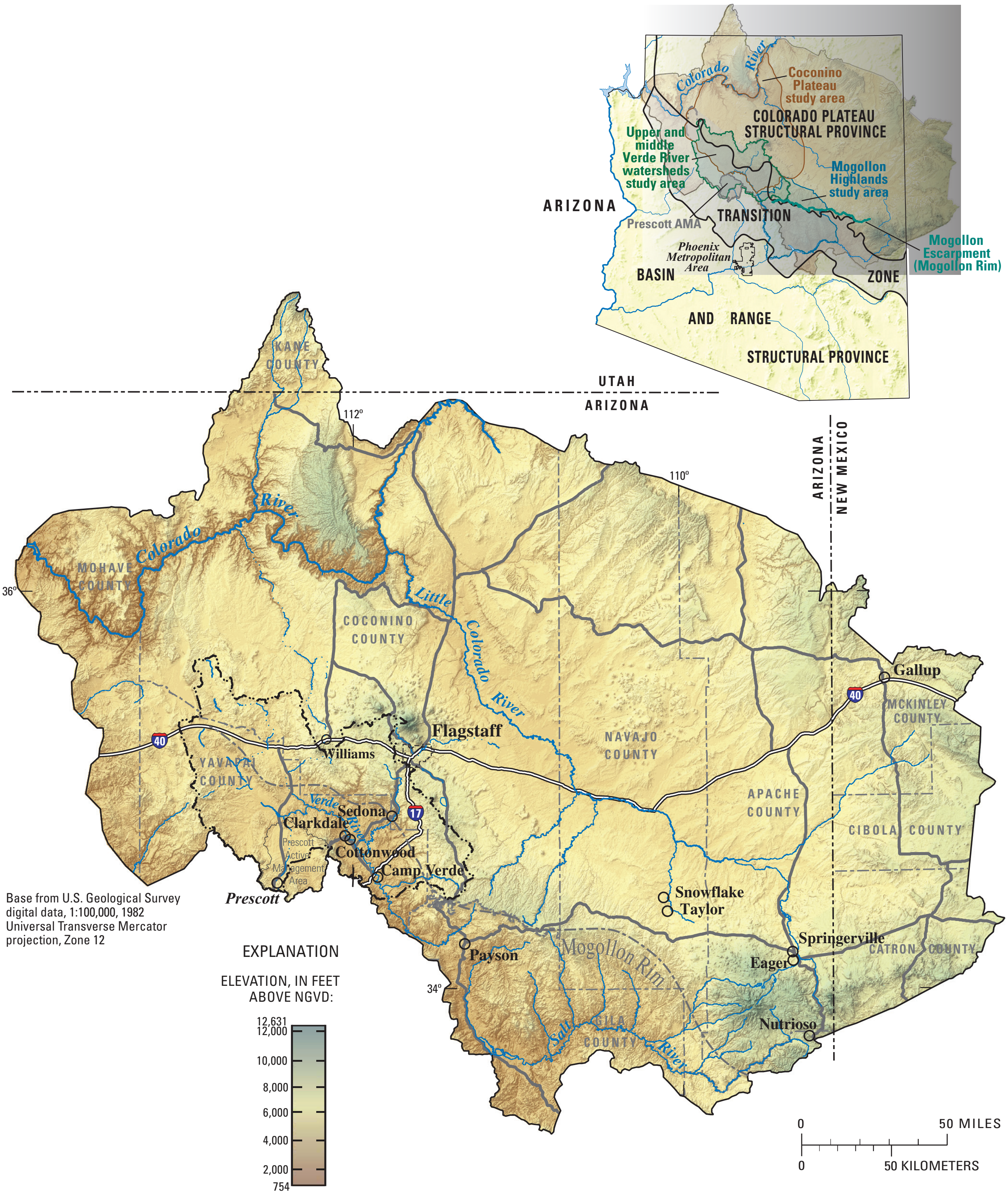


Figure 1. The Northern Arizona Regional Groundwater-Flow Model study area.

evapotranspiration rates. Recharge rates were not allowed to vary during the calibration process.

Overall goals in the model calibration process included representing regional patterns of water movement from areas of recharge to areas of discharge; representing approximate lateral extent of saturated portions of various hydrogeologic units; representing existing predevelopment groundwater levels and changes in groundwater levels in various parts of the model domain; representing vertical differences in groundwater levels, or differences between groundwater levels and levels of connected surface water, where such information is available; and representing estimated or measured groundwater discharge to surface features such as streams and springs. Sparsity of observation data in some parts of the model domain did not permit adjustment of some parameters in those areas. The product of this approach is a model that represents important elements in the regional flow system and approximates regional flow patterns and responses to stresses such as groundwater pumping and variations in recharge.

Description of Study Area

Physiography

The study area includes the watersheds of the Verde, Salt, and Little Colorado Rivers and adjacent areas of the Coconino Plateau (fig. 1). The region is predominantly in northern and central Arizona, but also includes adjacent parts of western New Mexico and southern Utah. The study area is largely in the physiographic provinces of the Colorado Plateau Structural Province and the Transition Zone, but includes parts of the Basin and Range Structural Province (Wilson and Moore, 1959; Fenneman, 1931) (fig. 1). The Colorado Plateau is an area of uplifted layered sedimentary rock and piles of layered volcanic rocks dissected by deep canyons. The Basin and Range Structural Province comprises mountainous regions of crystalline and consolidated sedimentary rocks separated by basins filled predominantly with unconsolidated alluvium derived from the surrounding mountains. The Transition Zone has physiographic characteristics of both provinces, having undergone episodes of extension and compression. The Transition Zone is deformed by faulting and uplift and contains rocks and alluvial sediments similar to the rocks and sediments in the Colorado Plateau (Anderson and others, 1992).

Land surface altitudes in the study area range from more than 7,000 feet on the boundaries of the study area on the Kaibab Plateau, Defiance Uplift, and in the central part of the study area along the Mogollon Rim, which is bounded by the White Mountains on the southeast and San Francisco Peaks on the northwest, to less than 3,000 ft at the lower reaches of the Verde and Salt Rivers, and less than 1,000 ft at the lower reaches of the Colorado River. North of the central highlands, the land surface gently slopes toward the Little Colorado and Colorado Rivers. To the south of the central divide, land surface drops steeply to the Verde and Salt Rivers (fig. 2).

The major rivers in the study area (fig. 2) are the Colorado River and Little Colorado River to the north of the Mogollon Rim and the Verde River and Salt River to the south of the Rim. Multiple major perennial streams that discharge to the Little Colorado River include Silver Creek, Clear Creek, and Chevelon Creek. Multiple perennial streams discharge to the Colorado River, including Nutrioso Creek, Kanab Creek, Havasu Creek, Nankoweap Creek, Bright Angel Creek, Hermit Creek, Monument Creek, Crystal Creek, Shinomo Creek, Tapeats Creek, Deer Creek, Diamond Creek, and Spencer Creek. Perennial streams discharging into the Salt River include the Black River, White River, Carrizo Creek, Cibecue Creek, Cherry Creek, Canyon Creek, and Pinal Creek. Perennial streams contributing flow directly to the Verde River include Sycamore Creek, Oak Creek, Wet Beaver Creek, West Clear Creek, Fossil Creek, and East Verde River (fig. 2). Additional discharge to each of these streams is supported by numerous intermittent and ephemeral streams.

Climate

The climate of the study area is primarily arid to semiarid with large spatial and temporal variations of temperature and precipitation (fig. 3). Climate conditions are strongly correlated with altitude; moderate summers and severe winters are at higher altitudes, and extreme summer heat and mild winters are at lower altitudes. Microclimates are common in the study area because of local controls on the amount of solar radiation and precipitation reaching the land surface in mountainous terrain and in the deep canyons.

The spatial distribution of precipitation is primarily influenced by the direction of approaching winds and orographic uplift of air masses. Average annual precipitation varies from about 7–15 in. in the basins to about 20–37 in. in the mountains and higher altitudes of the Colorado Plateau.

Precipitation in the study area is dominated by extended below-average periods of precipitation interspersed with occasional periods of above-average precipitation (Blasch and others, 2006). Annual precipitation is bimodally distributed between a summer monsoon period and a winter frontal storms period. The summer monsoon (also known as the North American Monsoon, or the Southwestern, Arizona, or Mexican Monsoon), generally begins in early July and extends through September. During these months, moisture-laden air from the Gulf of Mexico and the Gulf of California migrates northward to the study area. Convective uplift caused by surface heating is combined with orographic uplift to create unstable atmospheric conditions. Intense rainfall, lightning, hail, and high winds are typically associated with the instability. Convective monsoon rainstorms are characteristically short lived (less than a few hours), can be intense (greater than 1 in/hr), and localized (10s of square miles). The winter frontal season is from December through March. During this season, winds typically are from the west, bringing moisture-laden air masses from the Pacific Ocean. Frontal storms, caused by cyclonic flow systems in the area are characteristically longer (12–48 hr), less intense (less

Figure 2. Common names of physiographic features in the Northern Arizona Regional Groundwater-Flow Model study area.

than 0.25 in/hr), and more regional in extent (100s of square miles) than summer convective storms.

Precipitation also can fall during October and November as a result of tropical disturbances from the Pacific Ocean. Although precipitation during this period can be a substantial part of the annual total, the atmospheric conditions that result in precipitation do not happen regularly.

Average annual mean temperatures range from about 68°F in the basins to 43°F at higher altitudes and are inversely correlated with altitude and latitude. Average annual minimum temperatures were recorded as low as 36°F on the north rim of the Grand Canyon, and average annual maximum temperatures have been measured as warm as 81°F in the Salt River Basin. Large differences between the minimum and maximum daily temperatures are characteristic of the study area.

The average annual rainfall rate is greater than the average annual snowfall rate for all climate stations in the study area. The greater rainfall rate is attributed to warm annual mean temperatures. This greater rate is true in the basins and the higher altitudes where the ratio of rainfall to snowfall is about 3 to 1.

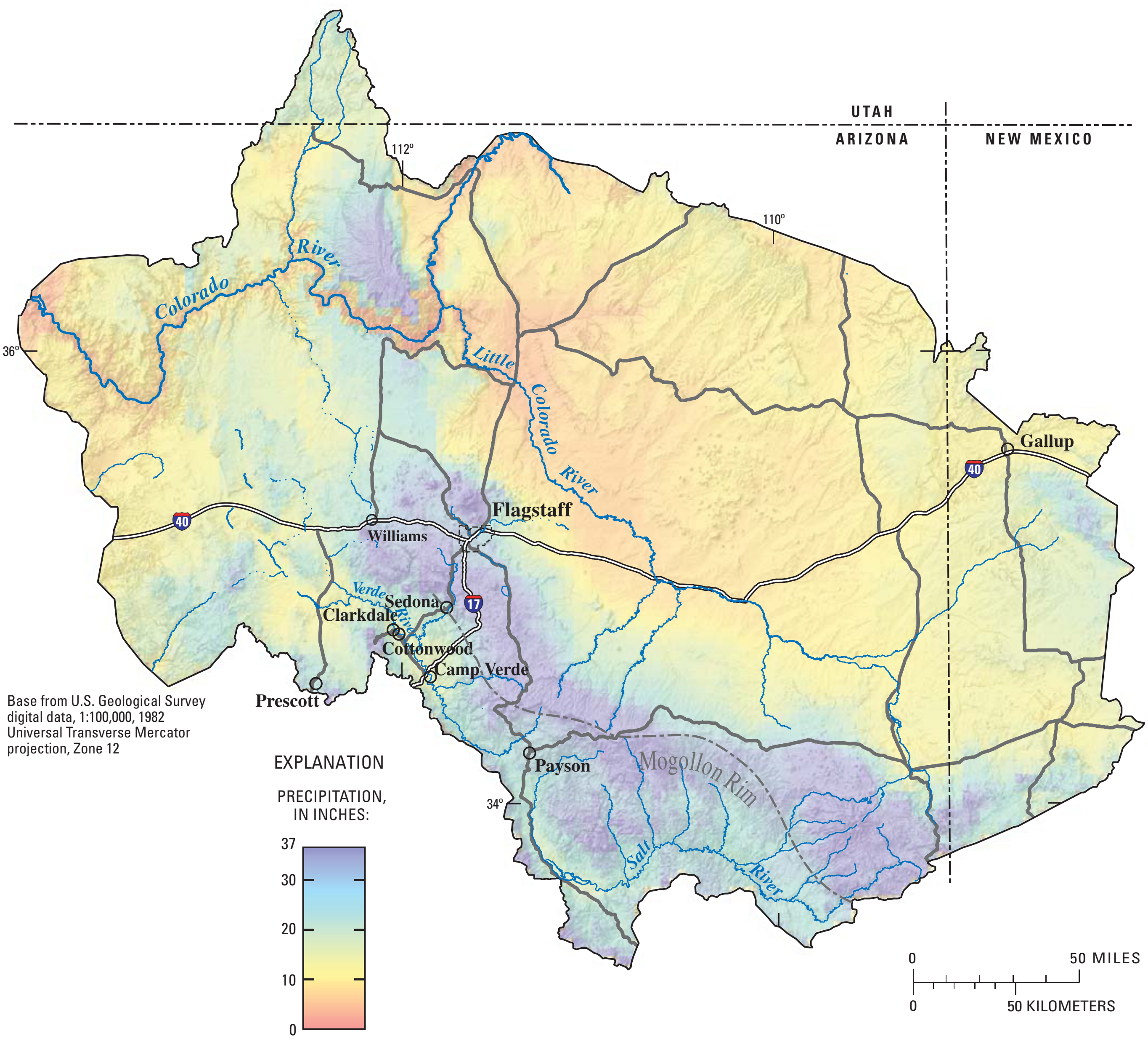


Figure 3. Average annual precipitation in the Northern Arizona Regional Groundwater-Flow Model study area 1971–2006.

Vegetation

The distribution of vegetation in the study area is influenced by temperature and water availability. Thus, plant communities are distributed on the basis of differences in latitudes, altitudes, and topography (fig. 4). Basins in the southern part of the study area are inhabited by desert scrub characteristic of the Sonoran desert. Basins in the northern part of the study are inhabited by sparse vegetation, desert scrub, and grasslands typical of the Great Basin desert. Piñon-juniper woodlands and chaparral are primarily present in the middle altitudes (about 3,900 – 5,600 ft). The predominant type of vegetation in the study area at altitudes above 5,600 ft is montane coniferous forest. Montane coniferous forest is present on the high altitude areas and mountains associated with the Kaibab Plateau, Defiance Uplift, and the Mogollon Rim including the White Mountains and a large region near San Francisco Mountain, Flagstaff, and Williams. About 85 percent of the study area includes about equal parts desert vegetation, piñon-juniper woodlands, and chaparral. The remainder is primarily montane coniferous forest.

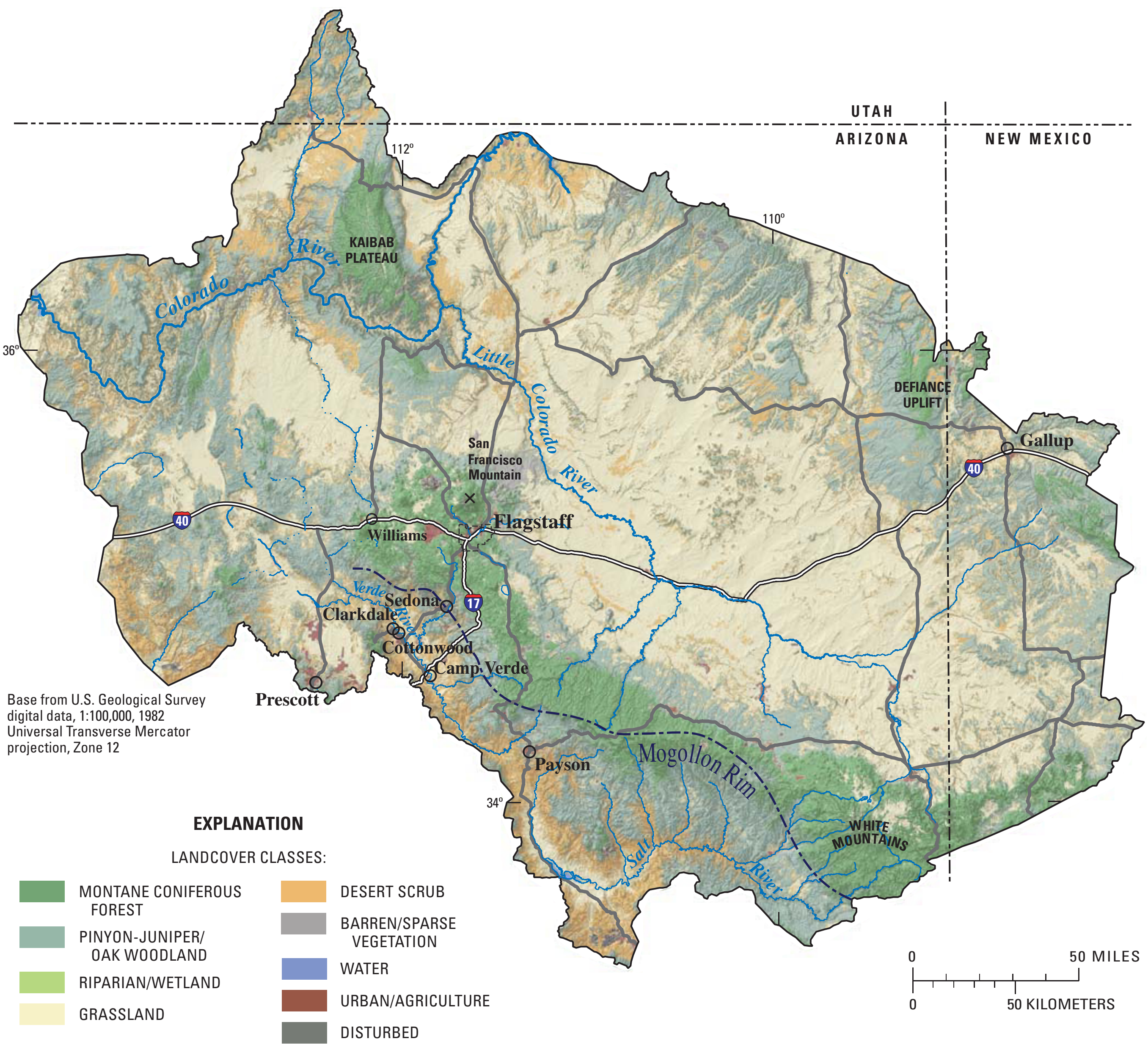


Figure 4. Landcover classes in the Northern Arizona Regional Groundwater-Flow Model study area.

Vegetation that taps groundwater supplies, phreatophytes, occur along perennial and intermittent streams. These plants are a very small percentage of overall vegetation type, but can locally use hydrologically important amounts of groundwater. Important phreatophytes of the area include Cottonwood, Willow, Sycamore, Tamarisk, and Mesquite.

Land and Water Use

Population

The 2005 estimated total population of the study area was 552,005. About 50 percent of the population lived in the incorporated cities and towns (U.S. Census Bureau, 2006b). The estimated population of the study area in Apache, Navajo, and Coconino Counties was 69,343, 108,432, and 123,866, respectively (U.S. Census Bureau, 2006a). Yavapai County had the largest population with 198,701 residents. The population of Coconino, Navajo, Apache, and Gila Counties were 123,866, 108,432, 69,343, and 51,663, respectively.

Land Use

About 47 percent of the land in the study area is publicly owned (fig. 5); 23 percent is managed by the U.S. Department of Agriculture (USDA) Forest Service, 4 percent is managed by the National Park Service, 11 percent is managed by the

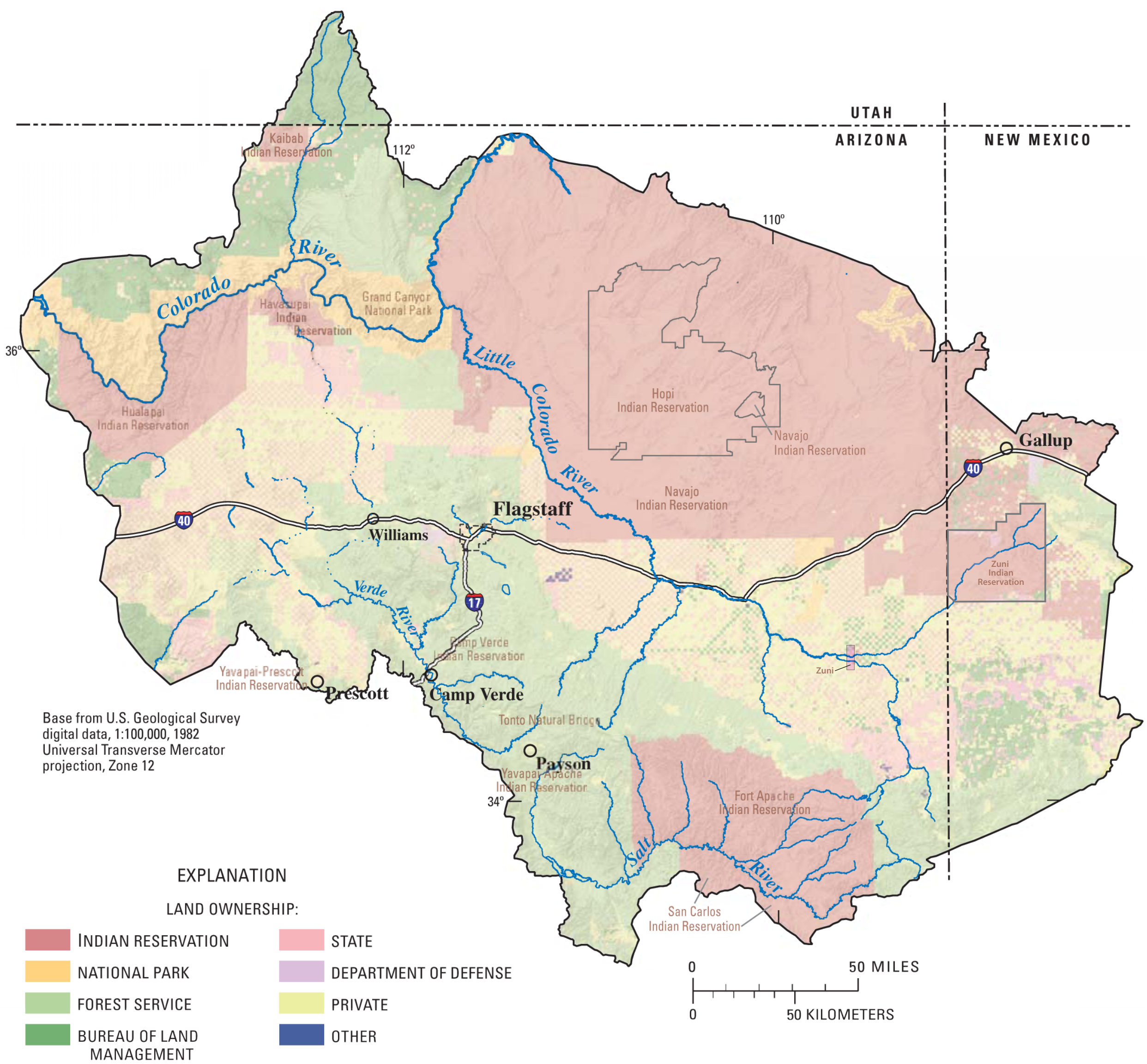


Figure 5. Land ownership in the Northern Arizona Regional Groundwater-Flow Model study area.

Bureau of Land Management, 9 percent is managed by the States, and less than 1 percent is managed by other public agencies. Native American reservations account for 36 percent of the study area while private holdings account for 17 percent of the land ownership. Recreation, agriculture, cattle ranching, mining, and urban development are the largest land uses in the region.

Water Use

Groundwater (including spring water) is the predominant source of water for domestic, industrial, and agricultural water uses. Current primary groundwater withdrawals in the study area are for industrial use (39 percent), municipal/domestic use (36 percent), and agricultural use (16 percent) (fig. 6). Groundwater near population centers is supplied primarily by private and municipal water companies. Wells are used in the rural areas to obtain groundwater for domestic and stock use. During the 1940s, the average annual groundwater withdrawal for the study area was about 15,000 acre-ft per year. The average annual groundwater withdrawal increased to about 140,000 acre-ft per year during 2000 to 2005. A more complete description of water use is described in the "Groundwater Budget Methods" and "Groundwater-Flow Model" sections of this report.

Previous Studies

Geologic and hydrologic data are available from prior studies for most of the area. Limited geologic or hydrologic data are available for parts of the study area such as western Coconino County and southern Navajo and Apache County. Subtantially more data are available for other parts of the study area such as the Grand Canyon National Park and Yavapai County. The RWI reports (Parker and others, 2005; Blasch and others, 2006; and Bills and others, 2007) provide a synopsis of previous studies for each of the RWI areas. Only the previous studies of the Upper and Middle Verde River Watersheds are discussed in detail in this report. Parker (2005) provides detailed discussion of previous studies in the Mogollon Highlands, and Bills and others (2007) has a detailed discussion of previous studies in the Coconino Plateau. Additional studies are described that were not previously included in the RWI.

Hydrogeology

Upper and Middle Verde River Watersheds

Blasch and others (2006) described the hydrogeologic framework, surface-water flow systems, and groundwater-flow

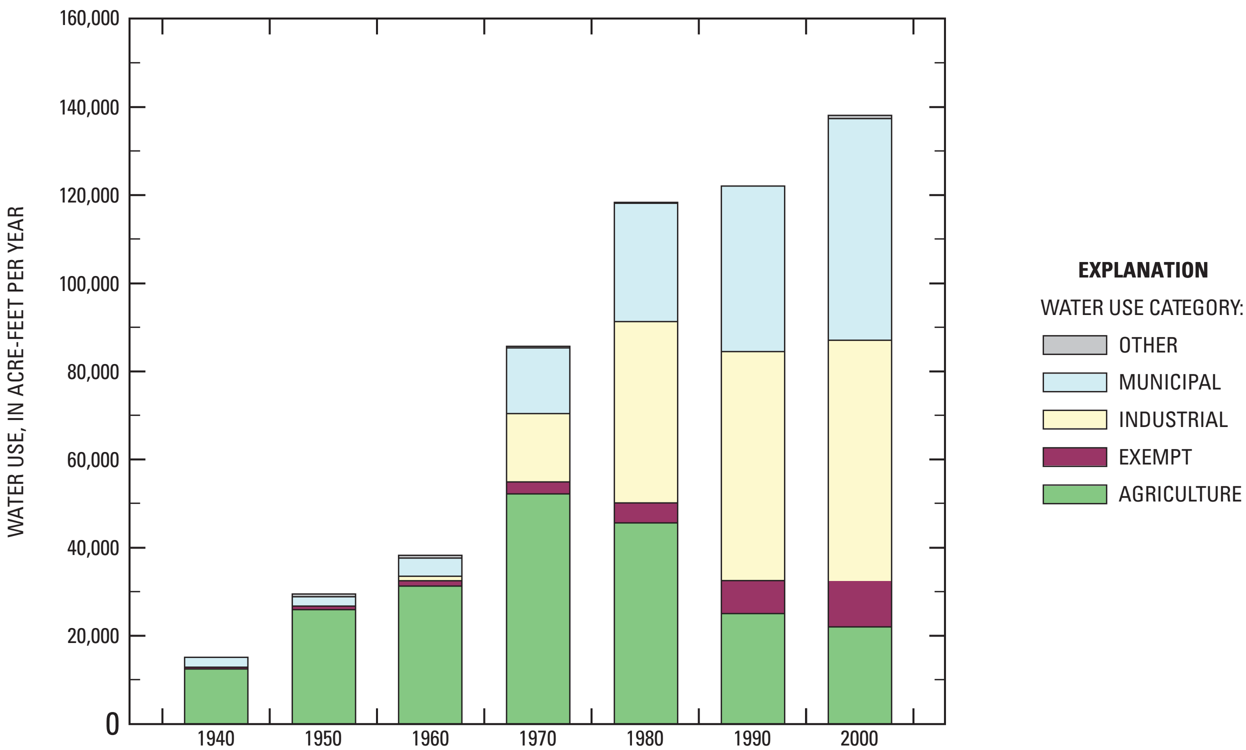


Figure 6. Water use trends from 1940 through 2005 for the Northern Arizona Regional Groundwater-Flow Model study area.

systems of the upper and middle Verde River watersheds and provide a summation of previous studies. Krieger (1965) provides a detailed discussion of the stratigraphy and structure, physiography, and mineral resources of the Prescott and Paulden areas. Anderson and Creasey (1958) described the geology of Mingus Mountain (fig. 2); Lehner (1958) described the geology of the Clarkdale quadrangle; Twenter and Metzger (1963) summarized the geology of the Mogollon Rim region surrounding the Verde Valley; and Anderson and Blacet (1972) described the rocks in the northern part of the Bradshaw Mountains. Hydrologic studies were primarily done for subregions of the study area. Schwalen (1967) described groundwater in the artesian area of Chino Valley and presented data for the period from 1940 to 1965. Wallace and Laney (1976) presented hydrographs and documented hydrologic conditions from 1975 to 1976. Matlock and others (1973) updated the work of Schwalen to include data from 1966 to 1972. Levings (1980) described groundwater availability and water chemistry in the Sedona area. Owen-Joyce and Bell (1983) presented findings of a water-resource assessment in the Verde Valley near Camp Verde, Clarkdale, and Sedona. In 1980, the Groundwater Management Act in Arizona resulted in the establishment of Active Management Areas, including the Prescott Active Management Area (PrAMA). As a result, numerous studies were made of the PrAMA that resulted in a map of groundwater conditions (Remick, 1983); a groundwater-flow model that simulates steady-state conditions (1940) and transient conditions (1940–93) (Corkhill and Mason, 1995); an updated groundwater-flow model that simulated conditions through 1998 and forecasted predictions to 2025 (Nelson, 2002); and an updated groundwater-flow model that simulated conditions through 2005, increased the model areal extent, and included new understanding of the geologic structure (Timmons and Springer, 2006). Schwab (1995) constructed a synoptic water-level map for the study area including areas outside of the PrAMA boundary. The Bureau of Reclamation (Ostenaa and others, 1993) conducted a hydrogeologic study of Big Chino Valley to identify potential sources of water for the city of Prescott. Knauth and Greenbie (1997) and Wirt and Hjalmarson (2000) used chemistry data to estimate groundwater-flow paths and source areas to the Verde River headwaters area. The Arizona Department of Water Resources (2000) compiled a summary of available water-resource data in the upper and middle Verde River watersheds. Langenheim and others (2005) calculated the depth of Tertiary alluvial sediments and volcanic deposits in the Big Chino, Little Chino, Williamson, and Verde Valleys, and identified several new faults by using aeromagnetic and gravity surveys. Wirt and others (2005) describe in a detailed study the geology, hydrogeology, and geochemistry of the headwaters region of the Verde River. In addition to these studies, digital geologic models have been constructed of the Verde River headwaters (Fry, 2006) and the Fossil Springs, Strawberry, and Pine areas (Green, 2008). Tadayon (2005) inventoried groundwater withdrawals outside of Active Management Areas. Blasch and others (2006) compiled water budget data for the upper and middle Verde River watersheds.

Colorado Plateau

The more notable reports that describe the Colorado Plateau are mentioned here to provide a regional and historical perspective. Bills and others (2007) provide an account of the previous work on the Coconino Plateau. Bills and Flynn (2002) compiled a database for the study area on geology, hydrology, climate, and other water-resources information and reports available through September 2001. The geology of various parts of the study area was investigated in some detail through the middle of the 20th century beginning with studies by Dutton (1882) and Darton (1910) who focused on the geology and structure of northern Arizona in their reconnaissance studies of the region. Robinson (1913) provided the first detailed study of the San Francisco Volcanic Field. The hydrogeologic study of the Navajo and Hopi Reservations by Cooley and others (1969) provided a geologic framework for northeastern Arizona that is the basis of all modern work in the region. The geology of the Grand Canyon has been investigated by hundreds of natural and physical scientists. The Grand Canyon Natural History Association has compiled these studies into a bibliography of the Grand Canyon and lower Colorado River as a ready reference to Grand Canyon geology (Spamer, 1990).

The surface geology and geologic structure of the Coconino Plateau south of Grand Canyon have remained largely overlooked until recent times (Huntoon, 1974; Huntoon and others, 1981, 1982, and 1986). Mineral exploration and, more recently, groundwater exploration and development are the topics of recent geologic studies on the Coconino Plateau. Billingsley (1987) and Billingsley and others (2000) provided detailed geologic descriptions of the western part of the Coconino Plateau as part of regional mineral-resource studies. Wenrich and others (1994) completed a hydrogeochemical survey to identify mineralized breccia pipes in the region. Recently, there has been renewed interest in developing more detailed geology and structural information for the Coconino Plateau because of concerns about the effects of continuing groundwater development on the sustainability of spring resources in Grand Canyon National Park. Recent geologic studies have focused on updating the Grand Canyon 30´ x 60´ quadrangle (Billingsley, 2000) and developing surface geology and structural information for the Valle and Cameron quadrangles (Billingsley and others, 2006; Billingsley and others, 2007). Geology and structural detail for the central part of the study area was described by Ulrich and others (1984) and Weir and others (1989). The regional structural framework also was revisited recently by Gettings and Bultman (2005) for identification of major hydrologically important geologic structures based on regional geophysical surveys.

The first detailed evaluations of groundwater in the Coconino Plateau part of the study area were not completed until the 1960s (Akers, 1962; Metzger, 1961; Cosner, 1962; and Twenter and Metzger, 1963). Feth (1954) and Feth and Hem (1963) focused their studies on the water-resources potential of national parks and monuments on and adjacent to the Coconino Plateau. Although Dutton, Darton, and Robertson all commented on springs and groundwater potential in their geologic reconnaissance studies, and Gregory (1916) completed hydrographic reconnaissance as part of his study of the Navajo Country, Johnson and Sanderson (1968) completed the first systemic inventory of springs along the Colorado River from Lees Ferry to Lake Mead, including a discussion of Havasu Springs and Havasu Creek. Cooley and others (1969) provided detailed discussions of the hydrogeology of the Navajo and Hopi Indian Reservations. Cooley (1976) also described, in detail, the hydrogeology of the lower Little Colorado River area, one of the primary groundwater discharge areas of the Coconino Plateau study area. Reports on the hydrology and hydrogeology for the more populated parts of the Coconino Plateau study area include studies by the city of Flagstaff (Harshbarger and Associates and John Carollo Engineers, 1972; Harshbarger and Associates and John Carollo Engineers, 1973; Montgomery, 1981; Harshbarger and Associates, 1976; Harshbarger and Associates, 1977; Duren Engineering, 1983; Errol L. Montgomery and Associates, 1992; Errol L. Montgomery and Associates, 1993; Errol L. Montgomery and Associates, 2003) the State of Arizona (McGavock, 1968; McGavock and others, 1986; Levings, 1980; Arizona Department of Water Resources, 2000), and the USGS (Feth, 1953; Cosner, 1962; Appel and Bills, 1980; Owen-Joyce and Bell, 1983; Bills and others, 2000; Bills and Flynn, 2002; Hart and others, 2002; Monroe and others, 2005). Academic studies that have contributed to the understanding of groundwater systems in parts of the Coconino Plateau include those studies of Goings (1985), Zukosky (1995), Fitzgerald (1996), Wilson (2000), and Kessler (2002).

Water-supply and water-sustainability issues on the Coconino Plateau part of the study area have changed the focus of water-resource studies away from localized studies to a more regional view of the hydrogeology and the capacity of regional systems to sustain water demands for people and ecosystems. The first of these regional studies was the Tusayan Growth Environmental Impact Statement and supplement released by the USDA Forest Service June 20, 1997, and July 17, 1998, that identified the protection of springs and seeps and associated riparian habitat in the greater Grand Canyon area as a major issue to be addressed by any new development in the area (Kaibab National Forest, 1999). A groundwater-flow model developed by Montgomery and Associates for the environmental impact statement (Errol L. Montgomery and Associates, 1999) was used to predict the effects of groundwater withdrawals for proposed development on spring flows in the area. Although the model was based on a small dataset, one of the predictions of the model was that groundwater withdrawals would have a direct, but small, effect on spring flows. Phase I of the north-central Arizona regional water study, completed in September 1998, included the compilation of current and future demands for water, currently available water supplies and costs, and possible future water supplies available to the area (Arizona Department of Water Resources, 2000). More detailed water-demand and growth studies followed (Heffernon and others, 2001; Rocky Mountain Institute and Planning and Management Consultants, Ltd., 2002). One of the current and future water supplies identified for the area is groundwater. However, a concern is the lack of information about groundwater-flow systems on the Coconino Plateau, the connectivity to surface-water resources in the area, and the variability and sustainability of these resources with time. Recent studies by Northern Arizona University indicate that spring-flow systems along the south rim of Grand Canyon may be more sensitive to changes in groundwater flow from natural and anthropogenic causes than previously known (Wilson, 2000; Kessler, 2002). The Bureau of Reclamation (BOR) North Central Regional Water Supply Appraisal Study (Kevin Black, hydrologist, BOR, written commun., 2006) report indicates that there will be an unmet water demand by the year 2025.

Several additional studies have contributed to understanding the hydrogeology of various parts of the Colorado Plateau. Harrell and Eckel (1939) studied the groundwater resources of Holbrook region and Akers (1964) completed a similar study for central Apache County. Wilson and others (1960) constructed a geologic map of Navajo and Apache Counties. Mann and Nemecek (1983) and Mann (1976) describe water resources in parts of Coconino, Navajo, and Apache Counties. Montgomery (2003) evaluated the hydrogeologic conditions in parts of Navajo and Apache Counties primarily by using geochemical analyses. Hoffmann and others (2005) completed aquifer tests near Leupp, Arizona. Leake and others (2005) constructed a groundwater model of the Coconino aquifer focusing on water development scenarios and the effect on Clear Creek, Chevelon Creek, and the Little Colorado River. E.A. Adams (unpub. data, 2005) and Adams and others (2006) investigated spring flow discharge and water quality along the south rim of the Grand Canyon.

Models of Groundwater Flow in the Study Area

Prescott Active Management Area

After the formation of the PrAMA several groundwater-flow models were developed for the region. Corkhill and Mason (1995) developed a model that simulates steady-state conditions that were assumed to occur in 1940 and transient conditions from 1940 to 1993. Nelson (2002) updated the Corkhill and Mason model and extended simulation periods to predict changes to the groundwater system through 2025. A groundwater model of the PrAMA also was developed by Southwest Groundwater Consultants, Inc. (1998). Woessner (1998) compared the models of Corkhill and Mason (1995) and Southwest Groundwater Consultants, Inc. (1998). An update of the Nelson (2002) groundwater model (Timmons and Springer, 2006) extended the model boundaries to encompass the Mint Wash area and incorporated updated geologic, hydrologic, and aquifer property data, and a greater number of calibration targets. Timmons and Springer (2006) used MODFLOW-2000 (Harbaugh and others, 2000a; Harbaugh and others, 2000b) and employed nonlinear regression to calibrate hydraulic properties of the model. Each of the PrAMA models included 2 model layers. Layer 1, the upper alluvial unit, is an unconfined aquifer consisting of alluvial and volcanic deposits, which extend throughout the Little Chino and the Upper Agua Fria sub-basins. Layer 2, the lower volcanic unit, contains volcanic flows interbedded with cinders, tuff, and alluvial materials and fractured and decomposed granite northeast of Granite Mountain near Mint Wash. The lower volcanic unit is modeled as a convertible confined/unconfined aquifer throughout the northern half of the model area. Hydraulic conductivities for the unconfined upper alluvial unit ranged from 0.3 to 25 ft/d and the specific yield values ranged from 0.05 to 0.12. Hydraulic conductivity values for the confined lower volcanic unit ranged from 0.1 to 175 ft/d and the specific storage values varied from $1x10^{-6}$ to $2x10^{-5}$.

Big Chino Valley

The Big Chino Valley is an alluvial basin west of the Verde River Headwaters. The basin has long been an attractive resource for groundwater development on the basis of shallow depth to water, large volume of saturated basin sediments, suitability for agriculture, and proximity to growing communities. Southwest Groundwater Consultants, Inc. (2005, 2007) constructed a 6-layer model for a part of the Big Chino Valley by using MODFLOW-2000 to evaluate the effects of groundwater withdrawals in the basin. The simulated area was west of the fine-grained playa deposit in the middle of the Big Chino Valley. The 6 layers were used to represent the upper shallow alluvium, volcanic rocks (basalt), lower alluvial unit below the volcanic rocks, and Redwall-Muav aquifer at depth. An aquifer test (Southwest Groundwater Consultants, Inc., 2007) performed to estimate transmissivity values for the upper alluvial units and the upper 50 ft of basalt yielded values that range from 12,000 to 25,000 ft^2/d. Estimated storage coefficient values ranged from 0.0004 to 0.002. The test also indicated that the upper alluvial layer is semiconfined by an upper silt and clay zone.

Navarro (2002) constructed a groundwater-flow model for Williamson Valley, a tributary to the Big Chino Valley, to characterize the water resources in the Williamson Valley and effects of withdrawal on riparian habitat and perennial springs. Navarro (2002) used MODFLOW-1996 to represent the groundwater system by using two layers and three distinct aquifers, including the Mint Wash aquifer, Granite Basin aquifer, and the Las Vegas aquifer. Hydraulic conductivity values were obtained through trial and error calibration. Simulated hydraulic conductivity values were 2 ft/d for basalt, 8.9 to 54 ft/d for conglomerate, 6.3 ft/day for buried conglomerate, and 2.7 ft/d for granite and gneiss. Specific yield values ranged from 0.1 to 0.2.

Colorado Plateau

A proposal to supply water withdrawn from the Coconino aquifer near Leupp, Arizona, to the Mojave Generating Station, an electrical power plant in Laughlin, Nevada (not in study area), resulted in three groundwater models for the Colorado Plateau (Peter Mock Groundwater Consulting, Inc., 2003; S.S. Papadopulos & Associates, Inc., 2005; Leake and others, 2005). The objective of these models was to estimate the effect of pumping water from the Coconino aquifer on water-level altitudes, discharge of groundwater to the Little Colorado River near Blue Spring, Clear Creek, and Chevelon Creek. Peter Mock Groundwater Consulting, Inc. (2003) constructed a 2,400-mi^2 groundwater-flow model that uses MODFLOW software to estimate the changes in groundwater levels near the proposed well fields in the study area. The authors used Southwest Groundwater Consultants, Inc., (2005) regional analytical model to enumerate boundary conditions.

S.S. Papadopulos & Associates, Inc. (2005) developed a two-layer numerical groundwater model of the Coconino aquifer and parts of the Redwall-Muav aquifers in northeastern Arizona and northwestern New Mexico. The objective of the groundwater model was to estimate effects of groundwater withdrawals near Leupp, Arizona, on groundwater discharge to Blue Spring, Chevelon Creek, and Clear Creek. The USGS finite-difference model code MODFLOW-2000 software was used to represent the groundwater system. The model domain was bounded by no-flow boundaries and recharge was assigned based on PRISM data (1961–1990) and National Climatic Center data. A maximum recharge rate of 5 in/yr was estimated for the higher altitudes near Flagstaff. The model was calibrated by using a nonlinear parameter estimation program, PEST (Watermark Numerical Computing, 2004). The calibration incorporated measured base flows, spring discharge, water levels, and water-level changes. The primary parameters adjusted during calibration were transmissivity, storage coefficient, and vertical leakance between layers. A pilot point method described by Doherty (2003) was used instead of the more common zonation approach to characterize parameters. Specific capacity for the Coconino aquifer layer ranged from 0 to 10,000 gpm/ft and transmissivity ranged from a lower bound of 0 to 1,000 ft^2/d to an upper bound between 9,000 and 10,000 ft^2/d. Storage coefficients ranged from 0.001 to 0.15. The vertical conductance between the Coconino aquifer and the Redwall-Muav aquifer was estimated to be 0.0004 d^{-1}.

Leake and others (2005) constructed a superposition model with a similar goal of estimating the potential effect of pumping from the Coconino aquifer near Leupp, Arizona, on groundwater discharge to Blue Spring, Chevelon Creek, and Clear Creek. Leake and others (2005) constructed a single layer model that represented the Coconino aquifer across most of the model domain and the Redwall-Muav aquifer in the western most part of the simulated area. The layer was simulated as convertible between confined and unconfined conditions depending on the calculated water level with respect to the overlying base of the Moenkopi Formation confining bed. Depths to the base of the Coconino aquifer ranged from 300 to 2,090 ft, however, the minimum depth was set to a constant 300 ft to assure horizontal continuity. The model grid was oriented 45 degrees west of north. Hydraulic parameters were assigned to the model on the basis of aquifer tests described in Hoffmann and others (2005). Hydraulic conductivity values used in the single layer varied from 5 to 6 ft/d, a value of 0.06 was used for specific yield, and a value of 2×10^{-6} ft^{-1} was used for specific storage, indicative of skeletal specific storage for sandstone.

Errol L Montgomery & Associates, Inc. (1999) constructed a groundwater-flow model for the Coconino Plateau groundwater sub-basin including the south rim of the Grand Canyon. The objective of the groundwater model was to quantify the effect of groundwater withdrawals in the sub-basin on Indian Garden Spring, Hermit Creek, and Havasu Springs. MODFLOW-1996 (Harbaugh and McDonald, 1996) was used to represent the groundwater-flow system as a single layer aquifer with zones of high permeability along fracture and fault zones where hydraulic conductivity is interpreted to have been enhanced by the dissolution of carbonate rock to form cavernous conduits that readily transmit groundwater in the Redwall-Muav aquifer. The grid was oriented 35 degrees west of north. Hydraulic conductivity values for the fracture zones ranged from 94 ft/d to 1,069 ft/d. Matrix rocks where solution- enhanced permeability faults were not identified were assigned storage coefficients of 0.001, whereas solution-enhanced features were assigned a storage coefficient of 0.005. Discharge from the model was represented at Hermit Creek Spring, Havasu Springs, Indian Garden Spring, and withdrawals from Tusayan and Valle. The simulation assumed discharge from Warm, Diamond, and Diamond Creek Springs were separated from the Coconino Plateau groundwater basin by the Aubrey and Toroweap fault systems. Simulations included steady-state conditions, assumed to occur before 1989, and transient responses to groundwater withdrawals.

Wilson (2000) constructed a groundwater model that utilized MODFLOW-1996 (Harbaugh and McDonald, 1996) numerical code to delineate the capture zones for the primary springs along the south rim of the Grand Canyon along the Coconino Plateau including Hermit Spring, Havasu Springs, and Indian Garden Spring. Wilson (2000) began the study by constructing a three-dimensional geologic framework model that utilized the stratigraphic geocellular modeling package Stratamodel (V4.000) (Landmark Graphics Corporation, 1998). The geologic framework model consisted of eight layers extending from the low permeability Proterozoic crystalline and Grand Canyon supergroup rocks at the bottom of the stratigraphic sequence to the Kaibab Limestone at the top. The framework model incorporated major structural features in the area and estimated unit depths by interpolating data from well logs, digital elevation models, and published maps and cross sections. The geologic layer distributions that resulted from the framework model were used to create two layers for the groundwater-flow model representing the Redwall-Muav aquifer and the Supai Formation. The model grid was oriented 60 degrees west of north. The steady-state groundwater model was calibrated to observed conditions by using water levels at eight wells, estimated discharge and hydraulic head at three springs, and estimates of recharge and hydraulic conductivity. Steady-state conditions that were assumed to occur before 1989 were simulated. No-flow boundaries were assumed for all but the spring discharge cells, which were represented as head-dependent flux boundaries. The horizontal and vertical hydraulic conductivity values of the Redwall-Muav aquifer were 1.3 ft/d and 0.3 ft/d, respectively. The horizontal and vertical hydraulic conductivity values of the Supai Formation were 0.02 ft/d and 0.0003 ft/d, respectively. Hydraulic conductivity values for the faults and fractures were 330 ft/d for the Markham Dam fracture zone, Havasu Downwarp, and 33 ft/d for other faults. Model results indicate that capture zones estimated in the model by using MODPATH (Pollock, 1994) for Hermit Spring and Indian Garden Spring were about 3 miles from the south rim of Grand Canyon. The report speculates based on the model results and discharge measurements by Goings (1985) that Hermit Spring is highly dependent on local precipitation and only partially dependent on the regional aquifer. Discharge from Havasu Springs was derived from a capture zone spanning about 99 percent of the modeled area.

A refined groundwater-flow model of the south rim of the Grand Canyon was constructed by Kessler (2002) to include additional small springs not included in the model of Wilson (2000). Particle tracking was used in the more finely gridded numerical model to delineate zones that contribute flow to each spring. The only known model of the groundwater-flow system in the region of the Kaibab Plateau was completed by Ross (2005).

Mogollon Rim

Wang and others (2001) developed a numerical fracture network model for a 2.7 by 2.7 mile area about 8 miles east of Payson. The numerical model was used to determine the effect of pumping from the fractured Payson Granite on the local hydrogeological system. The study area was divided into two regions based on fracture information obtained from boreholes. Aquifer tests were completed to provide hydraulic parameters for the model and to evaluate the 2-D fracture network. The authors concluded that fluid flow is dependent on the fracture environment and the aquifer is heterogeneous and anisotropic.

Acknowledgments

The authors acknowledge the cooperation, insight, and advice provided by many individuals, government agencies, private companies, and organizations. Particularly helpful were members of the NARGFM Model Development Panel, which included representatives from several cities, agencies, and organizations including representatives of the Arizona Department of Water Resources and the Yavapai County and Coconino County Technical Advisory Committees. Several hydrologic consulting firms also provided information and useful discussions including Southwest Ground-water Consultants, Inc.; Clear Creek and Associates, L.L.C; Errol L. Montgomery and Associates, Inc.; Leonard Rice Engineers, Inc.; and ERO Resources Corporation.

Many private water users provided helpful information about their water use practices including golf courses, industries, and power plants.

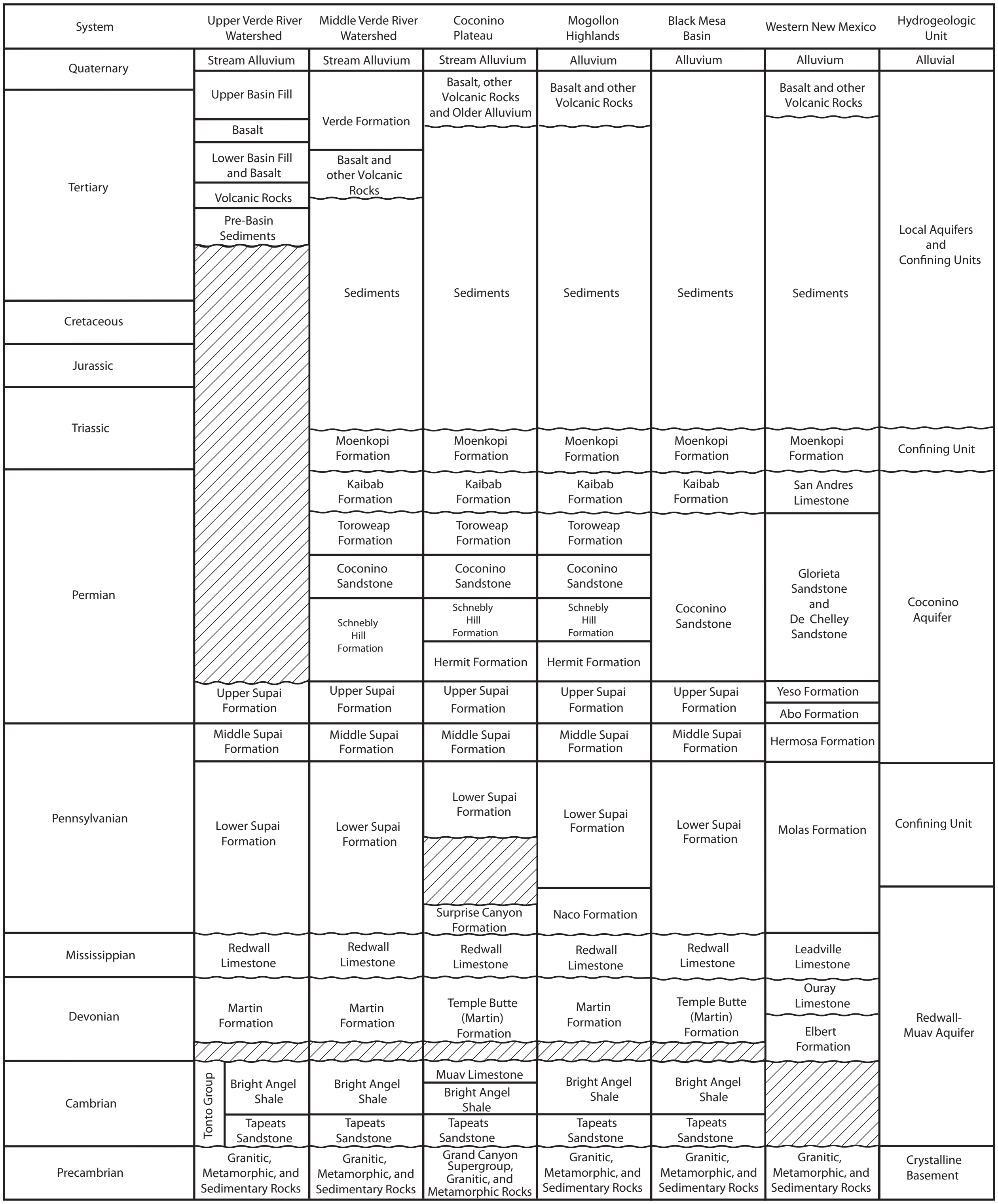


Figure 7. Generalized stratigraphic correlation of rock units across the Northern Arizona Regional Groundwater-Flow Model study area (Modified from Hart and others, 2002).

Regional Hydrogeology

Hydrogeologic Units

The stratigraphic sequence in the study area includes Proterozoic metamorphic and igneous rocks that are overlain by a sequence of Cambrian to Permian sedimentary rocks that are overlain in places by Tertiary volcanic rocks and alluvial basin deposits (fig. 7). A brief description of the stratigraphic sequences provided here was compiled from RWI reports; however, a more complete description of these units can obtained through Hart and others (2002), Leake and others (2005), and Parker and others (2005), Wirt and others (2005), Blasch and others (2006), and Bills and others (2007).

Proterozoic Basement

The oldest rocks in the study area are Proterozoic in age, underlie younger rocks throughout the study area, and are exposed north of the study area in the Grand Canyon, in the Big and Little Chino Valleys, Verde Valley, Mazatzal Mountains, and Sierra Ancha Mountains (fig. 2 and fig. 8). Proterozoic rocks consist of granite, metamorphic rocks and Grand Canyon Super Group. The Proterozoic metamorphic rocks in Verde River watersheds include regionally metamorphosed volcanic and sedimentary rocks. Proterozoic granite is exposed in the Bradshaw, and Juniper Mountains. The Grand Canyon Super Group is composed of sedimentary rocks that have been heavily eroded, faulted, and tilted (Elston, 1989). The Grand Canyon Super Group and metamorphic

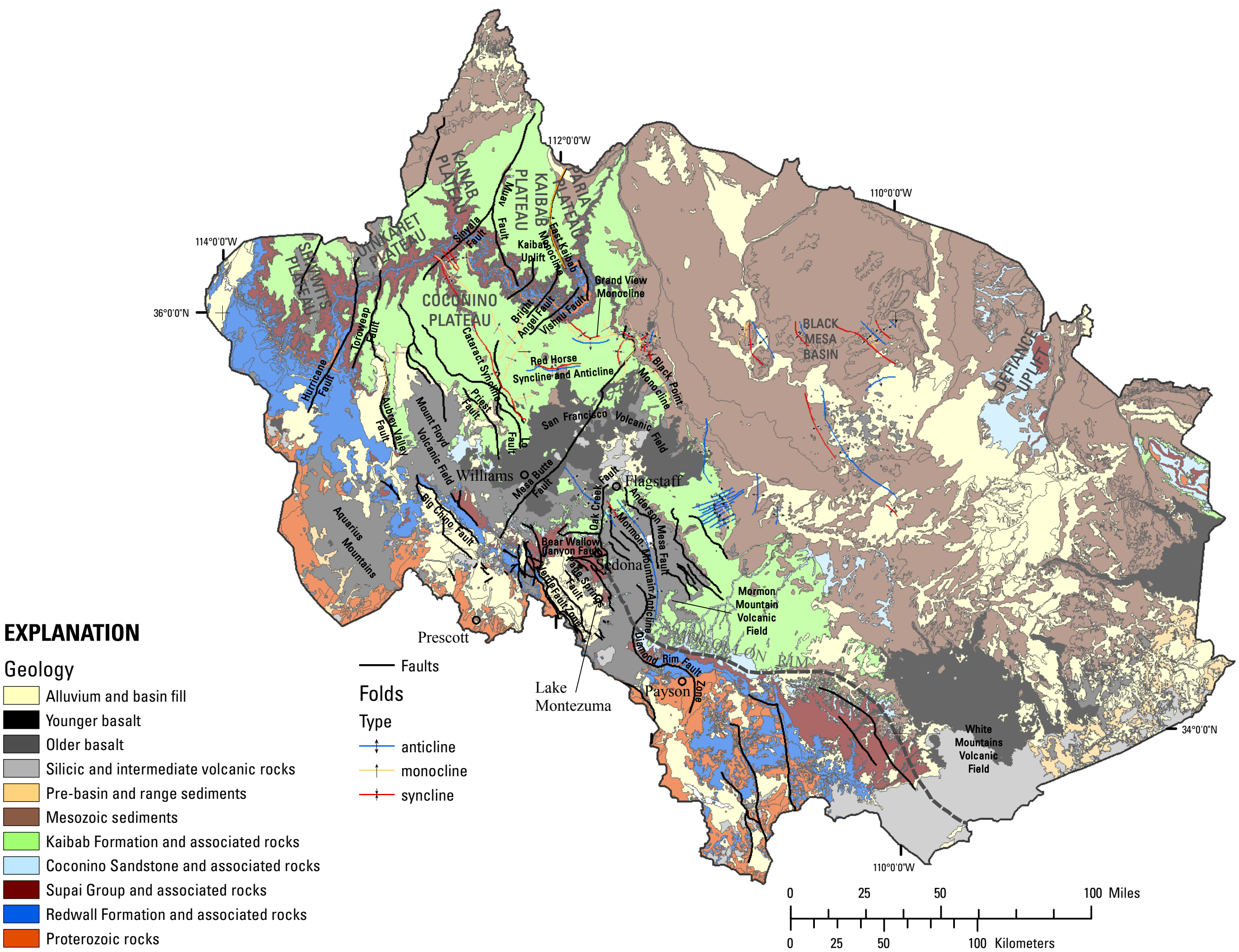


Figure 8. Surficial geology in the Northern Arizona Regional Groundwater-Flow Model study area. Geology modified from Utah State University (2004). Faults and folds modified from Ewing and others (1994), Reynolds (1988), DeWitt and others (2005), Blasch and others (2006), and Bills and others (2007).

rocks are exposed in the Grand Canyon and parts of the Verde Valley but do not appear in well cuttings from deep wells between Grand Canyon and Verde Valley (Bills and others, 2007). East of the Verde River Basin, metamorphic rocks are exposed primarily in the areas south of the Mogollon Rim in the Mazatzal Mountains and Sierra Ancha Mountains (Parker and others, 2005). The Proterozoic rocks form the base of the groundwater-flow system throughout the study area, except in areas where the rocks are highly fractured or weathered.

Paleozoic Rocks

The Paleozoic rocks are Cambrian to Permian age sandstone, conglomerate, siltstone, mudstone, shale, limestone, and dolomite. Paleozoic rocks include the Tapeats Sandstone, Bright Angel Shale, Muav Limestone, Temple Butte/Martin Formation, Redwall Limestone, Naco Formation, Supai Group, Hermit Formation, Schnebly Hill Formation, Coconino Sandstone, Toroweap Formation, and Kaibab Formation. These sedimentary rock units are characterized by east-to-west facies changes, a general westward thickening of most units of about 1,000 ft, and a regional southwest dip of about 2 degrees (Billingsley, 2000; Billingsley and others, 2006; Bills and others, 2007). Paleozic rocks underlie almost the entire study area and are exposed in deep canyons along the Colorado River (Beus and Morales, 1990) and the Mogollon Rim (Twenter and Metzger, 1963) (fig. 8).

The Tapeats Sandstone and Bright Angel Shale are discontinuous throughout much of the area, and the Tapeats Sandstone is associated with old channels and valleys on the Proterozoic surface (Bills and others, 2007). The Tapeats Sandstone is a medium-grained to very coarse-grained crossbedded feldspathic and quartz sandstone. The formation typically is cemented with silica and is similar to quartzite. The Tapeats Sandstone ranges in thickness from 0 to 400 ft and is exposed in the Grand Canyon, parts of the Chino and Verde Valleys and near Payson. The Bright Angel Shale contains particle sizes from clay to silt and is 450 ft thick near the western end of Grand Canyon and thins southward and eastward, a few feet thick in the Chino Valley area, and absent south and east of the Black Hills (Middleton and Elliott, 1990; Bills and others, 2007). Little is known about the water-bearing properties of the Tapeats Sandstone in the study area because of the lack of deep-well data; however, several springs issue from the Tapeats Sandstone in Grand Cayon and the unit is water-bearing in recently drilled wells in Williams area (Bills and others, 2007). The Bright Angel Shale is not a major water-bearing unit; however, several springs issue from the unit in Grand Canyon (Monroe and others, 2005; Bills and others, 2007).

Paleozoic rocks include a thick sequence of several limestone units including the Muav Limestone, Temple Butte and Martin Formations, Redwall Limestone, and Naco Formation. Each of the limestone units is exposed in different places in the study area, including the Grand Canyon, upper Verde River Watershed, and along the Mogollon Rim (fig. 8).

The Cambrian Muav Limestone is composed of fossiliferous limestone, silty limestone, dolomite, and mudstone with interbeds of sandstone, siltstone, mudstone, and shale (Middleton and Elliott, 1990; Billingsley, 2000; Billingsley and others, 2006). McKee and Resser (1945) described several horizons and members in the Muav Limestone, including the Rampart Cave Member that includes areas of secondary porosity that form conduits for discharge of groundwater to many springs along the south rim of the Grand Canyon. The Muav Limestone ranges from about 200 ft thick in the eastern part of Grand Canyon to more than 600 ft thick in the western part of the canyon (Middleton and Elliott, 1990; Billingsley, 2000) and is found north and west of the study area. In a few wells in the study area, the Muav Limestone lies unconformably on granite. The southward depositional extent of the Muav Limestone probably lies south of Grand Canyon (Middleton, 1989); however, outside this region the Martin Formation is lithologically similar and the two formations may be stratigraphically equivalent.

The Devonian Temple Butte and Martin Formations include beds of dolomite, limestone, shale, mudstone, sandstone, and conglomerate (Beus and Morales, 1990). The Temple Butte Formation is present in the Grand Canyon, but is absent in wells in the study area and is not exposed at the southern and western boundaries of the Plateau (Bills and others, 2007). The formation ranges from 50 to 275 ft thick where exposed (Billingsley, 2000). In the central and southern parts of the study area and along the Mogollon Rim, equivalent Devonian rocks are known as the Martin Formation (Tiechert, 1965). The Martin Formation is present in wells and crops out along the western part of the Colorado Plateau in the study area, the Mogollon Rim in the Big Chino and Verde Valleys, and the Mogollon Rim in the Payson area. The formation is about 0 to 480 ft thick and can lie unconformably on lower Cambrian units or Precambrian rocks. The Martin Formation yields water to wells in the Sedona area, near the town of Drake, and in the Black Hills. The Martin Formation is exposed along parts of the upper Verde River, in the Big Chino Valley and is extensively exposed on Big Black Mesa, in the Upper Verde River area (DeWitt and others, 2005), where groundwater discharges from the formation to the river, and in the Payson area (Middleton, 1989; Bills and others, 2007). The Temple Butte and Martin Formations are equivalent to the Ouray Limestone and Ebert Formation in western New Mexico (fig. 7).

The Redwall Limestone is a light-gray to gray, fine to coarse crystalline bedded limestone with dolomite and thin beds and lenses of chert (Beus, 1989). Some beds in the formation are fractured, and the limestone typically contains solution cavities and caverns, some of which have collapsed or filled with sediment. Those beds that have collapsed are commonly filled with conglomerate that is cemented with red, claylike sediment (Lehner, 1958, p. 530), which has been classified as the Surprise Canyon Formation (Billingsley and Beus, 1985). The Redwall Limestone ranges from about 250 to 750 ft in thickness and yields water to wells. In the Mogollon

Rim area of Payson the Redwall Limestone is primarily a layer of rubble that resulted from extensive weathering of the Redwall surface during a period of surface exposure before deposition of the overlying Naco Formation (Brew, 1965). The Redwall Limestone is equivalent to the Leadville Limestone in western New Mexico (fig. 7).

The Naco Formation consists mainly of discontinuous, ledge-forming, nodular limestones in tabular beds on the order of 1 to 10 ft thick with interbedded siltstone and shale. The unit is primarily present along the Mogollon Rim east of the Verde River and ranges in thickness from 0 to 1,000 ft. The contact generally is hidden beneath slopes of colluvium and unconformably lies on the Redwall Limestone, a basal red shale composed of eroded and reworked Redwall Formation, that fills the underlying, uneven karst surface (Brew, 1965).

The Supai Group is divided into three formations—the Upper, Middle, and Lower (Blakey, 1990). The Upper Supai Formation is a complex series of horizontally bedded reddish to brown sedimentary units that are mostly fine-grained sandstone, siltstone, and mudstone. The Middle Supai Formation is grayish-orange, calcareous, very fine-grained sandstone to siltstone. The Lower Supai Formation is red to purple sandstone and siltstone, and gray limestone and dolomite. In some locations, the base of the formation contains conglomerate or breccia. The Supai Group in the Grand Canyon area is about 1,120 ft thick and generally thins eastward and south toward the Mogollon Rim (McKee, 1982; Billingsley, 2000). The Supai Group ranges from about 300 ft thick near Sedona to about 700 ft thick in other locations along the Mogollon Rim (Blakey, 1990). The Upper Supai Formation is permeable and is included as part of the Coconino aquifer. The Middle Supai and Lower Supai Formations generally act as confining units to groundwater in the underlying Redwall-Muav aquifer, although sandstone intervals may be locally important water-bearing zones. In western New Mexico, the Supai Group is equivalent to the Yeso, Abo, Hermosa, and Molas Formations (fig. 7).

The Permian Hermit Formation is a red to dark-red, thin-bedded siltstone and fine-grained sandstone that lies unconformably on the Supai Group in Grand Canyon (Billingsley, 2000). The formation is seldom identified in well cuttings as a distinct unit except in the western part of the study area (Bills and others, 2007); however, outcrops of the Hermit Formation south of Flagstaff were described by Blakey (1990). The thickness of the Hermit Formation in Grand Canyon ranges from about 850 ft near the western end of the canyon to less than 260 ft near the eastern end. The Hermit Formation is difficult to differentiate from the underlying Upper Supai Formation in the central and eastern parts of the Mogollon Rim. Outcrops of the Hermit are 100 ft thick or less in the Sedona area, and as much as about 300 ft in the upper Verde Valley area (Blakey, 1990). The Hermit Formation is not a major water-bearing unit and may be a local confining unit.

The Schnebly Hill Formation unconformably overlies the Hermit Formation or the Supai Group and is the product of deposition in a rapidly subsiding closed basin. The formation comprises a sequence of reddish-brown to reddish-orange very fine-grained to silty sandstone, mudstone, limestone, and dolomite (Blakey, 1990). The formation underlies, and in places interfingers with, the Coconino Sandstone. In the Sedona area, the Schnebly Hill Formation is about 750 ft thick, and thickens east and west along the Mogollon Rim to more than 1,500 ft (Blakey, 1990). The formation also thins sharply to the north and is not present in well cuttings north and northwest of Flagstaff (Bills and others, 2007). The formation includes a prominent limestone bed, Fort Apache Limestone, near the base (Blakey, 1990). The Schnebly Hill Formation is locally a major water-bearing unit and is part of the Coconino aquifer.

The Permian Coconino Sandstone is a white to tan or light brown, crossbedded, aeolian, fine-grained sandstone. The Coconino Sandstone is about 700 ft thick along the Mogollon Rim but can be as thick as 1,100 ft near Flagstaff and 1,200 ft at the headwaters of the East Verde River. Exposures in the Grand Canyon indicates the formation ranges from 150 ft in the west to 500 ft in the eastern parts (Billingsley, 2000). Extensively fractured zones along faults are likely permeable and have the potential to yield large quantities of water. The Coconino Sandstone forms the primary part of the Coconino aquifer in the study area; however, the formation generally is above the water table west of Mesa Butte Fault. The Coconino Sandstone is equivalent to the Glorieta Sandstone and DeChelley Sandstone in western New Mexico (fig. 7).

The following description of the Permian Toroweap Formation by Sorauf and Billingsley (1991) is based on outcrops west of a line running from about Sycamore Canyon to Marble Canyon (fig. 2). The formation consists of red carbonate sandstone, red silty sandstone and siltstone, limestone, and thin layers of gypsum (Sorauf and Billingsley, 1991). The formation ranges in thickness from 0 to 380 ft and thickens westward. In the Flagstaff area, the transition from the upper part of the Coconino Sandstone to the Toroweap Formation is abrupt. East of Flagstaff, well data indicate that the formation is absent. The transition between the two formations is gradational northwest of Flagstaff where the two formations can be difficult to distinguish (Sorauf and Billingsley, 1991).

The Late Permian Kaibab Formation (Sorauf and Billingsley, 1991) unconformably overlies the Toroweap Formation and the Coconino Sandstone and is composed of two members: the lower Fossil Mountain Member and the upper Harrisburg Member. The Fossil Mountain Member is a light-grey, cherty, thick-bedded limestone to sandy limestone (Bills and others, 2007). The Harrisburg Member is an interbedded sequence of light-red to grey limestone, dolomite, siltstone, sandstone, and gypsum (Sorauf and Billingsley, 1991). The formation thickens from east to west from about 100 to 650 ft (Sorauf and Billingsley, 1991). The Toroweap and Kaibab Formations are exposed across a large region from the Mogollon Rim to Coconino and Kaibab Plateaus. Where exposed, the Kaibab Formation has well-developed fractures, many of which are widened by dissolution, which forms sinkholes in places. The Kaibab Formation is equivalent to the San Andres Limestone in western New Mexico (fig. 7). The

Toroweap and Kaibab Formations are part of the Coconino aquifer where the units are saturated.

Mesozoic Rocks

The Triassic Moenkopi Formation unconformably overlies the Kaibab Formation and is a sequence of fine-grained sediments including mudstone, siltstone, sandstone, and gypsum. The formation generally ranges in thickness from 0 to about 1,000 ft (Billingsley and others, 2006). The Formation has low permeability and creates confined conditions in the underlying Coconino aquifer across a broad region north of the Little Colorado River and in western New Mexico. Other Mesozoic sedimentary rocks overlie the Moenkopi Formation and include aquifer systems that are not hydraulically connected to the aquifers that are the subject of this study.

Cenozoic Rocks and Alluvial Basins

Cenozoic sediments and volcanic rocks can be divided into two groups: those sediments and rocks deposited before the basin and range structural disturbance (10–15 million years before present) and those sediments and rocks deposited during and after the disturbance (Anderson and others, 1992). Tertiary sediments and volcanic rocks that are older than the basin and range disturbance are structurally deformed and regionally discontinuous. Thick accumulations of basin-fill sediments that are associated with formation of the basin and range structural province include interbedded Tertiary basalt flows in a few areas. The basin-fill sediments tend to include coarse-grained facies of sand and gravel along the basin boundaries and fine-grained facies of silt, clay, and possible evaporate deposits in the basin center. The fine-grained facies is generally a confining unit that locally separates basin-fill aquifer into upper and lower parts. The lower basin fill generally contains more coarse-grained alluvial deposits and is more consolidated than upper basin fill. The lower basin fill is often described as conglomerate in drill logs.

Cenozoic sediments and volcanic rocks that are important regional aquifers occur primarily in the Transition Zone (fig. 1) where the basin and range structural disturbance resulted in basins filled with thick alluvial deposits. Groundwater in the Cenozoic regional aquifers may be hydraulically connected to underlying and laterally adjacent groundwater-flow systems in the Coconino and Redwall-Muav aquifers. Cenozoic alluvial deposits and volcanic rocks may be locally important aquifers in areas outside of the thick alluvial basins. These local aquifers tend to be shallow and may not be hydraulically connected to the underlying regional aquifers because of intervening confining units, such as the Moenkopi Formation, or local confining beds within the Cenozoic rocks. Cenozoic volcanic rocks may form locally important aquifers in areas outside of the thick alluvial basins along the Mogollon Rim and in the White Mountains, San Francisco Volcanic Field, and Mount Floyd Volcanic Field. Cenozoic alluvial deposits in areas outside of the thick alluvial basins that may form locally important aquifers include the rim gravels (Pierce and others, 1979; Pierce, 1984) in the region of the Mogollon Rim and Bidahochi Formation on the Little Colorado River Plateau.

The Tertiary pre-basin-and-range rocks comprise volcanic rocks and alluvial deposits that are generally poor aquifers, but may form locally important aquifers especially where the rocks are highly fractured. Sediments and volcanic rocks of primarily andesite and lati-andesite were deposited prior to the basin and range structural disturbance and have been subjected to extensional tectonics characterized by low-angle faulting and rotation. These rocks are exposed at the north end of Little Chino Valley and in the Walnut Creek area of the Big Chino Basin. An extensive region of pre-basin-and-range sediments is overlain by late Tertiary and Quaternary basalt in the Aquarius Mountains.

Late Tertiary basin-fill sediments constitute the primary aquifers in the alluvial basins of the Transition Zone and were deposited during and after the basin and range structural disturbance (Anderson and others, 1992). The basin-fill sediments include coarse-grained facies of sand and gravel along the basin margins and fine-grained facies of silt, clay, and gypsum in the basin centers that are playa and lacustrine in origin (DeWitt and others, 2005). The thickness of basin-fill sediments typically is less than several hundred feet but is a few thousand feet thick in some locations (Ostenaa and others, 1993 Langenheim and others, 2005). The playa deposit in the Big Chino Valley has an estimated surface area of about 50 mi^2 and a maximum recorded thickness of about 1,800 ft (DeWitt and others, 2005). Playa deposits of limited extent occur in the northern part of the Little Chino Basin. Basin fill in the Verde Valley includes extensive fluvio-lacustrine deposits of the Verde Formation (Jenkins, 1923) and interbedded sand, gravel, and basalt flows that also occur along the boundaries of the basin and at depth beneath the fluvio-lacustrine deposits.

The late Tertiary Verde Formation is a complex assemblage of six facies (Twenter and Metzger, 1963) that are highly variable in permeability: upper, middle, and lower limestone facies; an undifferentiated limestone facies; and sandstone and mudstone facies. The Verde Formation ranges in thickness from a few feet to more than 3,100 ft. The upper, middle, and lower limestone facies lie laterally adjacent to the thick undifferentiated limestone facies in the central part of the valley. The upper limestone facies generally extends to the basin boundaries or to where the upper limestone facies interfingers with an alluvial facies of sand and gravel. The upper limestone facies is separated from the middle limestone facies by the sandstone facies. The sandstone facies consists of sandstone and siltstone and thin interbeds of limestone. The middle limestone facies extends laterally from the thick limestone facies to the basin margins or to where the middle limestone facies interfingers with an alluvial facies of sand and gravel. The mudstone facies predominantly includes mudstone, claystone, and evaporites that interfingers with the lower limestone facies. The limestone facies are the most permeable parts of the Verde Formation and are often confined beneath the mudstone facies. The most extensive region of sand and gravel facies occurs in the northwest part of the Verde Valley, where thick intervals of sand and gravel overlie and underlie the fluvio-lacustrine deposits of the Verde Formation.

The late Tertiary to Quaternary volcanic rocks, grouped into six assemblages on the basis of age and geographic location, are (1) the Mormon Mountain Basalts, (2) the Mount Floyd Volcanic Field, (3) the San Francisco Volcanic Field, (4) White Mountains, (5) Aquarius Mountains, and (6) scattered late Tertiary intrusive rocks, pyroclastic deposits, and basalts. The volcanic rocks are exposed at land surface in nearly the entire length of the Mogollon Rim. The Mount Floyd Volcanic Field, near Seligman, is composed of olivine basalt and red cinders. The unit has an estimated age of 14.4 to 6.4 million years (McKee and McKee, 1972; Billingsley and others, 2006) and ranges in thickness from 0 to 2,000 ft (Bills and others, 2007). The volcanic rocks exposed from north of Flagstaff to west of Williams are part of the San Francisco Volcanic Field and comprise andesite, dacite, and basalt flows, and pyroclastic flows (Ulrich and others, 1984). Volcanic rocks associated with the volcanic field range from nearly 0 to 5,000 ft in thickness, and are 6.0 to 0.05 million years old (Nealy and Sheridan, 1989). Volcanic rocks of the Aquarius Mountains are late-Tertiary and Quaternary basalt (Reynolds, 1988).

Hydrologically important late Tertiary to Quaternary volcanic rocks are primarily constrained to highly permeable regions where basalt flows are interbedded with basin fill in the Little Chino and Big Chino sub-basins (Wirt and others, 2005). Late Tertiary basalt flows are common in the basin-fill sediments in the Little Chino and Big Chino Basins (DeWitt and others, 2005;Wirt and others, 2005; Owen-Joyce and Bell, 1983; Twenter and Metzger, 1963) and in the lower part of the Verde Formation. Highly permeable basalt flows are in the northern part of the Little Chino Basin where the volcanic rocks are known as the lower volcanic unit and in the Paulden area of the Big Chino Basin.The basalt flows are not known to be major water-bearing rocks in other areas.

Quaternary sediments in the study area consist of alluvial fan deposits (fanglomerate), fine-grained alluvial sediments, terrace gravels, gravel, and recent stream alluvium. These sediments are exposed at land surface in major drainages throughout the study area. Quaternary stream alluvium typically is highly permeable and locally yields water to shallow wells primarily for domestic and agricultural use. This unit is composed of unsorted, poorly bedded clay, silt, sand, pebbles, and cobbles, and typically is less than 30 ft thick. Because of the limited areal coverage and thickness, the unit is not considered a primary aquifer. In addition, the stream alluvium units are only partially saturated in many areas because the alluvium is generally above the water table except near major drainages such as the Verde River near Camp Verde and parts of the Little Colorado River.

Geologic Structure and Tectonic History

Regional tectonic stresses created the geologic structure that has shaped the landscape of the study area. Pre-Tertiary sediments were raised about 10,000 to 15,000 ft by uplift resulting from two or more tectonic compressional events that began in the Cretaceous period (Shoemaker and others, 1978). Regional folding along a general northwest trend developed with this uplift. Continued periodic tectonic compressional and extensional stresses have resulted in folds, faults, and other fractures that further modified the sediments and formed the structural depressions occupied by the current alluvial basins (Nations, 1989; Jenney and Reynolds, 1989). Northeast-, north-, and northwest-striking faults and other fractures dominate the structure of the study area (Gettings and Bultman, 2005). The extensional stresses that have weakened the regional pre-Tertiary sediments have enabled large amounts of late Tertiary and Quaternary intrusive and volcanic rocks to reach the surface (Wolfe and others, 1987a,b). In addition, zones of weakness are continuing to expand, lengthen, and deepen, in some areas into canyons, from continued interaction with water. Groundwater movement along the expanded zones of weakness enhances the preferential flow paths through dissolution of carbonate rocks and evaporites and carbonate or evaporate cement in clastic rocks.

The oldest structures in the study area are vertical to near vertical fractures that have propagated upward from Proterozoic basement rocks and strike north and northeast. These fractures are inferred to be related to reactivation of Proterozoic normal faults under tension with reversal that has further resulted in development of monoclinal structures in younger rocks (Shoemaker and others, 1978; Wolfe and others, 1987a,b). The monoclines overlying the deep-seated Proterozoic reverse faults were reactivated by late Cretaceous and early Tertiary compression commonly referred to as the Laramide Orogeny. The Laramide Orogeny uplifted and horizontally shortened the Coconino Plateau through development of folds on reactivated basement faults (Huntoon, 1974, 1990). The regional uplift, taking place in pulses, has raised the Proterozoic surface more than 10,000 ft above the Early Cretaceous level (Huntoon, 1989). The erosion that accompanied the Laramide Orogeny has stripped most of the Mesozoic rocks from the surface of the Coconino Plateau and adjacent areas and left a few well-preserved drainages that have been minimally modified by the Pliocene incision of Grand Canyon (Huntoon, 1990). The Laramide Orogeny produced a regional dip of 1–2 degrees to the east and north. Primary structural features that resulted are the Kaibab Uplift (Shoemaker and others, 1978), the Cataract Syncline (Krantz, 1989; Huntoon, 2003), the Mesa Butte Fault (Babenroth and Strahler, 1945) (fig. 8), and a large regional depression in the area of the Little Colorado River Plateau that includes the Black Mesa Basin.

The change from compressional stress to extensional stress at the onset of basin and range extension probably began by the middle Tertiary (Young, 1979) and has resulted in active faults and open fracture zones (Billingsley and others, 2006).

Recent seismic activity (Fellows, 2000) indicates ongoing extensional stresses, which have resulted in the development of local extensional basins and extensional sags and closed basins (Huntoon, 1990, 2003). Aubrey Valley is a closed topographic basin formed by extension (Billingsley and others, 2000) (fig. 8) and is similar to other extensional basins in the Transition Zone including the Big Chino and Verde Valleys (Huntoon, 2003) (fig. 8).

Ongoing erosion has produced the landscape of the modern Colorado Plateau and adjacent areas. The Colorado River was established in Grand Canyon by late Miocene, 9 to 6 Ma (Lucchitta, 1990). Downcutting through more than 1 mile of Paleozoic sediments and Proterozoic rocks progressed rapidly and resulted in dewatering of regional groundwater systems of the area. Two mature north-flowing drainages, Cataract Creek and the Little Colorado River, dominate the northern part of the study area (fig. 2). Isostatic rebound caused by the formation of Grand Canyon has resulted in the major short, steep drainages of the south rim of the canyon that are just beginning to develop southward onto the plateau (Beus and Morales, 2003). Cenozoic volcanism has modified ancestral south-flowing drainages to fairly short, steep streams that flow northward to the Grand Canyon or southward to Verde Valley (Wolfe and others, 1987a,b; Nealey and Sheridan, 1989). Faults and monoclinal structures partly define the western, southern, and eastern boundaries of the Coconino Plateau.

Coconino Plateau and Plateaus North of the Colorado River

The Coconino Plateau lies north and west of the San Francisco Mountain and includes the Cataract Creek watershed that drains north into Havasu Creek and the Colorado River (fig. 2). Along the northeastern and northern boundary of this watershed are the higher altitudes of the Coconino Plateau and along the southern boundary is the Mogollon Rim. The Mesa Butte Fault forms the eastern boundary of the Coconino Plateau. The Aubrey Cliffs are on the western edge of the Cataract Creek watershed. The Aubrey Fault and Toroweap Fault systems act as impediments to groundwater flow from western part of the Coconino Plateau basin (Errol L. Montgomery and Associates, 1999). The Aubrey Fault is a high angle, down-to-the-west normal fault with about 200 ft of offset at the north end of the fault and more than 500 ft of offset at the south end (Billingsley and others, 2000) (fig. 8). The Aubrey and Toroweap Faults represent two of the many north-striking to northeast-striking, deep-seated regional fault systems reactivated during the Laramide Orogeny and, more recently, during basin and range extension (Huntoon, 2003). Numerous other north-northeast and north-northwest trending faults occur on the Coconino Plateau (Bills and others, 2007). Several north-northeast and north-northwest trending folds occur on the Coconino Plateau (Bills and others, 2007). Some of the more prominent folds are Cataract Syncline, Red Horse Syncline, and Red Horse Anticline, Grand View Monocline, and several monoclines northeast of the Cataract Syncline.

The valleys that have developed on the interior of the Coconino Plateau have few well-defined streams and no free-flowing water. Many of these valleys have been filled to depths of 100 ft or more with gravels and other erosional materials from surrounding uplands (Billingsley and others, 2000, 2006). These drainages generally follow the regional slope to the south and southwest away from the Grand Canyon before turning into the structural trough of the Cataract Syncline (Bills and others, 2007) (fig. 8). The overall drainage pattern of the area is interrupted in places by local internal drainage areas (Bills and others, 2007). Mechanisms leading to the formation of internal drainage include (1) dissolution of gypsum in the Kaibab Formation, (2) dissolution of older limestones, (3) development of tectonically young faults and grabens that interrupt normal drainage, and (4) collapse structures associated with breccia pipe development (Billingsley and others, 2006). These internal drainage features are commonly filled with Quaternary sediments that can trap water. As a result, these areas may have a major effect on the regional groundwater movement (Bills and others, 2007).

North of the Colorado River are several broad plateaus that are bounded by monoclines. The plateaus include the Paria Plateau on the eastern edge, the Kaibab and Kanab Plateaus in the center, and the Uinkaret and Shivwits Plateaus on the western boundary of the study area (fig. 8).

Little Colorado River Plateau

The Black Mesa Basin is the dominant structural feature of the Little Colorado River Plateau (fig. 8). The primary water-bearing Paleozoic units generally dip toward the center of the basin where the Coconino aquifer is buried beneath nearly 5,000 ft of Mesozoic sediments (Southwest Groundwater Consultants, Inc., 2005). The Little Colorado River Plateau is bounded by structural uplifts in the northeast and south—the Defiance Uplift and the Mogollon Rim, respectively—and Mesa Butte Fault on the west. The Defiance uplift (fig. 8) is bounded by monoclines that plunge to the west and east in the Black Mesa Basin (Leake and others, 2005). Several regional-scale normal faults exist in the area and generally have two primary strike directions: north-northeast and north-northwest (Bills and others, 2000). Hydrologically important faults in the Little Colorado River Plateau include faults that form horst and graben structures in the Flagstaff area (fig. 8). Some of the most prominent of these faults are the Anderson Mesa, Lake Mary, and Oak Creek Faults. Numerous north-northwest trending faults in the Flagstaff area (Bills and others, 2007) are also associated with grabens that may be hydrologically important. Folds in the Little Colorado River Plateau that may have hydrologic importance include the Black Point Monocline and Mormon Mountain Anticline. The Black Point Monocline near the lower part of the Little Colorado River dips to the east-northeast and has several hundred feet of offset (Ulrich and others, 1984) (fig. 8). The Mormon Mountain Anticline, with 5 degrees of local dip, trends northwest-southeast across the southern and southeastern part of the study area (Ulrich and others, 1984; Weir and others, 1989) (fig. 2).

Mogollon Rim

The Mogollon Rim is a south-facing, mainly erosional escarpment that is retreating to the north (Elston, 1978; Pierce and others, 1979). The Mogollon Rim is the result of at least two episodes of uplift: one preceding deposition of upper Cretaceous strata and another preceding fluvial deposition of nonmarine late Tertiary rim gravels that evolved on the ancestral escarpment (Pierce and others, 1979; Pierce, 1984). Subsequent erosion of the plateau edge northward has removed most of the rim gravels and Cretaceous sediments. The rim has been subsequently segmented by more recent faulting, volcanism, and erosion into the landforms seen today (Pierce, 1984).

The Mogollon Rim east of the Verde River watershed is a combination of major Proterozoic faults and lineaments that trend northeast-southwest and more recent north-south to northwest-southeast trending structures produced by Miocene crustal extension. The earlier faults are silicified (Gæorama, Inc., 2001). The more recent faults are typically high-angle normal and reverse faults. Major extensional faults indicate displacements of as little as 50 to 100 ft locally and as much as 600 to 1,500 ft on the Diamond Rim Fault Zone north of Payson and where the Fault Zone crosses the East Verde River (Gæorama, Inc., 2001; Parker and others, 2005). The Diamond Rim Fault (fig. 8), which dips 75 to 80 degrees to the south and is upthrown on the northern side, is a major regional fault and extends 30 mi, generally east-west from east of Tonto Creek to Pine Creek (Conway 1976, Wrucke and Conway, 1987), and likely continues for some distance west of Pine Creek (Gæorama, Inc., 2001). The juxtaposition of crystalline and sedimentary formations caused by such displacements plays a major role in the development of surface drainage, determination of groundwater-flow directions, and the distribution of springs below the Mogollon Rim as faults variably enable or preclude drainage of water-bearing formations (Parker and others, 2005).

Transition Zone

The predominant structural features in the Transition Zone are the northwest- to north-trending valleys and mountains and normal faults that are typical of the basin and range structural disturbance, and pre-basin-and-range shear zones and folds. The northwest- to north-trending valleys are Big Chino, Williamson, Little Chino, Verde Valleys, and Tonto Creek Basin, which were created by extensional faulting. The extensional faulting resulted in the downdropping of Paleozoic and Tertiary deposits that sometimes resulted in the juxtaposition of Proterozoic crystalline rock against the younger deposits. The valley floors are gently sloping and consist of unconsolidated to consolidated Tertiary and Quaternary basin-fill sediments and Quaternary stream alluvium that are typically underlain by gently dipping consolidated Paleozoic sedimentary rocks or Proterozoic crystalline rocks.

The Big Chino sub-basin comprises Big Chino Valley, Williamson Valley, Big Black Mesa, and the western part of the Coconino Plateau (fig. 2). Big Chino Valley is a 28-mi-long northwest-trending valley in the northwestern most part of the upper Verde River watershed. The valley is bounded on the northeast by Big Black Mesa, and on the southwest by the Juniper Mountains. The basin underlying the valley was formed about 10 to 2 million years ago (DeWitt and others, 2005) by normal faulting in response to crustal extension during the basin and range disturbance. Normal faulting on the northeast and southwest sides of the valley created a graben, which has since filled with alluvial deposits eroded from adjacent uplands. The graben ranges in width from about 2 mi at the northwest end of the valley to about 6 mi at the southeast end near Paulden.

The Big Chino Fault (fig. 8), a normal fault which lies near the northeast boundary of Big Chino Valley and the western boundary of Big Black Mesa, is the largest documented fault in the valley. Exposure of Cambrian and Precambrian rocks on the northeast and upthrown side of the fault indicate a maximum vertical displacement of about 3,500 ft. Displacement decreases southeastward to nearly zero near the town of Paulden.

Little Chino and Upper Agua Fria Valleys, adjacent to Big Chino Valley, also trend north (fig. 8). The basin underlying these valleys was created by tensional faulting, which also produced small normal faults along the perimeter and interior of the valleys. The valleys are bounded primarily by Proterozoic crystalline and Tertiary volcanic rocks. The alluvium in Little Chino Valley is underlain predominantly by Tertiary basalt and latite volcanic rocks and only partially by Paleozoic rocks, which are present only in the northernmost part of the valley (fig. 8). The alluvium in the Upper Agua Fria Valley is underlain primarily by Proterozoic rocks.

The Verde Valley is an alluvial basin formed by a structural graben that is bounded on the north and northeast by several sub-parallel faults, the Mogollon Rim, and Mormon Mountain Anticline. The basin is bounded on the southwest by the Black Hills and other hills that are uplifted southwest of the Verde Fault Zone (fig. 8). The Verde Fault Zone is a series of northwest-striking faults and subordinate faults mostly on the east side of the Verde Fault (Anderson and Creasey, 1958). The Verde Fault is a normal fault that dips to the northeast. Estimated cumulative displacement on the southwestern side of the graben is about 3,700 ft (Ed DeWitt, research geologist, U.S. Geological Survey, written commun., 2005). The Oak Creek Fault strikes north-south and is one of the high-angle normal faults that have been reactivated by recent basin and range extension. The west side of the fault is upthrown, and displacement ranges from about 400–500 ft at the north end on the San Francisco Volcanic Field to more than about 700 ft at the south end near Sedona. Other normal faults in the area include the Page Springs, Bear Wallow Canyon, and other numerous subparallel unnamed faults associated with these major faults. The Page Springs Fault is coincident with Page Springs and may act as a highly permeable conduit for the flow of groundwater to the springs.

Aquifers and Aquifer Properties

Most rock units in the study area contain some water-bearing zones. However, structural deformation has reduced the continuity of saturated units across the study area. Groundwater systems of the study area are more complex than is indicated by the fairly simple layering of the rocks that contain the groundwater systems. The complexity is because of variations in stratigraphy, lithology, and, geologic structure (Bills and others, 2007). The Redwall-Muav aquifer, the Coconino multiple aquifer system (Coconino aquifer), and the basin-fill aquifers are the primary aquifers in the study area. The Quaternary alluvial aquifers and the Payson granite are local aquifers that are hydraulically connected to the regional groundwater system. Other local aquifers are in the alluvium, volcanic rocks, the Kaibab Formation, the Coconino Sandstone, the Supai Group, and Proterozoic rocks and often are hydraulically disconnected from each other and from the regional aquifer system. These local isolated aquifers generally are small and thus are unsuitable as long-term water supplies; however, these aquifers are used extensively to meet local water demands. These local disconnected aquifers are taken into consideration in the regional groundwater-flow system because groundwater that discharges from the local aquifers can percolate downward to the underlying regional aquifer system. The Redwall-Muav aquifer and the Coconino aquifer are the primary regional aquifers on the Colorado Plateau in the study area (Cooley and others, 1969; Cooley, 1976). The Redwall-Muav aquifer and the Coconino aquifer have southeast-northwest trending groundwater divides that are coincident with or near the Mogollon Rim and divide the regional groundwater-flow system into parts that flow northward toward the Colorado and Little Colorado Rivers and southward toward the Verde and Salt Rivers.

The Coconino aquifer comprises the Kaibab Formation, the Coconino Sandstone, the Schnebly Hill Formation, and the Upper Supai Formation (Bills and others, 2000). Although the Kaibab Formation typically is unsaturated, except for perched zones and north of the Little Colorado River, the formation provides a conduit for infiltration and percolation of precipitation and surface water (Bills and others, 2000; Wilkinson, 2000). The Coconino aquifer is the primary aquifer in the Little Colorado River watershed, but is only locally saturated west of the Mesa Butte Fault. The Redwall-Muav aquifer underlies the Coconino aquifer and comprises the Redwall, Naco, Temple Butte, and Muav Limestones. North of the Verde River and Big Chino Wash, the Martin Formation and Tapeats Sandstone underlie the Redwall Limestone and are considered part of the Redwall-Muav aquifer (Owen-Joyce and Bell, 1983). The Redwall-Muav aquifer is the primary aquifer in the Cataract Creek watershed. The basin-fill aquifers are the primary aquifers in the Chino Valleys and Verde Valley, and Tonto Creek Basin.

Knowledge of the hydrologic properties of the geologic units that constitute the regional and local aquifers in the study area is important information for conceptualizing and simulating the movement of groundwater. Aquifer properties include hydraulic conductivity, transmissivity, specific storage, and specific yield. Formation lithology, fracture development, and dissolution influence the magnitude and preferential orientation of these properties. One method that is commonly used to obtain hydraulic property estimates is aquifer tests. During an aquifer test, a well is pumped for several hours or longer while yield (volume per time) and change in water level (drawdown) in the pumped well and adjacent monitoring wells are recorded. The combined measurements of pumping and drawdown can then be used to estimate aquifer properties. The derived aquifer properties can then be used to help determine aquifer properties for a numerical groundwater-flow model.

Aquifer tests are costly and do not always provide representative results. Thus, specific-capacity values, which are more readily available, are commonly used to estimate transmissivity. Specific capacity is computed by dividing well yield by drawdown at the pumped well. Specific-capacity values are representative of aquifer properties only near the pumped well. Transmissivity can be estimated by using empirical equations that convert specific-capacity data into a transmissivity value (Theis and others, 1963; Driscoll, 1986; Razack and Huntley, 1991; and Mace, 1997). Several authors have reported specific-capacity values for the study area (Schwalen, 1967; Owen-Joyce and Bell, 1983; Navarro, 2002). A summary of specific-capacity values is presented by Blasch and others (2006) for comparison of values among water-bearing units in the upper and middle Verde River watershed.

Proterozoic Basement

In general, the Proterozoic crystalline rocks do not store or transmit substantial amounts of water and form the underlying confining bed for the Redwall-Muav, Coconino, and basin-fill aquifers. Only in a few areas with major fracturing is water found in quantities sufficient for withdrawal. One of these areas is near Payson, where the Proterozoic rock, known as the Payson granite or Payson shelf, is the primary aquifer that is exposed in more than a 128-mi^2 area. Here, Tertiary faulting has created substantial secondary permeability (Parker and others, 2005) and storage. Based on numerous aquifer tests throughout the area, transmissivity ranges from 40 to 2,270 ft^2/d and is greatest where wells are associated with major fracture systems (Southwest Groundwater Consultants, Inc., 1998). Discharge from Proterozoic rocks is low compared to the Redwall-Muav and Coconino aquifers (Flora, 2004).

Estimation of the Top of Hydrologic Basement

The lower boundary condition of the groundwater-flow model is a no-flow boundary defined by the contact between Proterozoic crystalline bedrock (sometimes called “Basement”) and overlying, more hydraulically conductive sedimentary units. The lower boundary of the groundwater-flow system was estimated as a surface defined by the altitude of the top of the Proterozoic crystalline rock determined by using outcrops, borehole data, and variations in the magnetic and gravitational fields of the Earth caused by contrasts in magnetic susceptibility and density. Gravity and magnetic data were used to estimate variations in the surface between outcrop and borehole control points. Three methods based on analysis of the horizontal and vertical gradients of the gravity and magnetic fields were used to analyze the gravity and magnetic data, including Euler deconvolution, analytic signal amplitude (ASA), and horizontal gradient magnitude (HGM). Each method calculates the altitude of susceptibility and density contrasts caused by dipping contacts, faults or other structures. For magnetic field methods, the calculated altitude is typically more sensitive to the part of the structure that is closest to the land surface. For gravity methods, the calculated altitudes are more sensitive to the center of mass of the structure. Calibration of the gravity and magnetic-field derived altitudes determined from outcrop and borehole data is necessary to adjust for altitude offsets introduced by simplifying assumptions and method sensitivities.

Gravity and magnetic-field modeling produced eight models by using the ASA, HGM, and Euler deconvolution. Analytic signal amplitude (ASA) provides depth estimates representing minimum depths with the assumption of thick sources. T is the total gravity or magnetic field.

$$\left|ASA(x, y)\right| = \sqrt{\left(\frac{\partial T}{\partial x}\right)^2 + \left(\frac{\partial T}{\partial y}\right)^2 + \left(\frac{\partial T}{\partial z}\right)^2} \qquad (1)$$

Horizontal gradient magnitude (HGM) provides depth estimates representing minimum depths with the assumption of thick sources.

$$HGM(x, y) = \sqrt{\left(\frac{\partial T}{\partial x}\right)^2 + \left(\frac{\partial T}{\partial y}\right)^2} \qquad (2)$$

In Euler deconvolution, the apparent depth to the gravity or magnetic source is derived from Euler’s homogeneity equation:

$$N(B-T) = (x-x_0)\frac{\partial T}{\partial x} + (y-y_0)\frac{\partial T}{\partial y} + (z-z_0)\frac{\partial T}{\partial z} \qquad (3)$$

This process relates the magnetic field and magnetic field gradient components to the location of the source of an anomaly (x_0,y_0, z_0), with the degree of homogeneity expressed as a structural index (*N*), a measure of the rate of change of the field with distance. *B* is the magnitude of the regional field. Structural indices of 0 and 0.5 were used for the magnetic field and 0.5 and 1 for gravity. Larger structural indices for gravity reflect the fact that the gravitational force falls off as the inverse of the second power of distance compared with the third power for the magnetic force.

Calibration and adjustment of each surface was needed to account for simplifying assumptions in each method. To calibrate, a linear regression was completed for gravity- and magnetic-derived altitudes against field-measured outcrop and borehole altitudes. The regression equation was then used to adjust gravity- and magnetic-derived altitudes. A minimum-curvature algorithm was used to interpolate among the adjusted altitudes to create a preliminary basement surface modeled by using one algorithm (ASA, HGM, or Euler deconvolution), one structural index (if required by the algorithm), and one data type (gravity or magnetics). Values from the best four of eight surfaces generated were averaged to create the final basement surface for use as the underlying no-flow boundary in the groundwater-flow model. Only surfaces computed from the ASA algorithm and Euler deconvolution with a structural index of 0.5 were used for averaging. The HGM surfaces were left out, because these surfaces were similar to the ASA surfaces but added no new information at the scale of the study area. The Euler deconvolution structural-index 0 surface had too few calculated altitudes to make a good calibration. The surface calculated by using Euler deconvolution and a structural index of 1 highlighted structures with too great a dimensionality (pipe shaped structures) to be of interest. To calibrate the gravity and magnetic estimates of the basement surface, shifts in altitude of each surface were required, typically 3,000 meters. The offset was smaller for magnetics, probably because gravity is sensitive to the center mass of structures, which typically lie at greater depths. Finally, a correction surface was constructed from the difference of conrol data and the averaged basement surface and applied to the original basement surface so that the final surface conformed locally with outcrop and borehole data. The final basement surface was used to define the base of the simulated groundwater-flow system and is discussed in the “Model Framework” section.

Redwall-Muav Aquifer

The Redwall and Muav Limestones are the primary water-bearing rock units in the region of the Coconino Plateau and underlying the Coconino aquifer in the remainder of the Colorado Plateau. Because of the regional extent of these formations in northern Arizona, Cooley (1976) defined the Redwall and Muav Limestone multiple-aquifer system as the saturated to partly saturated and hydraulically connected

Redwall, Temple Butte, and Muav Limestones (fig. 7). Hydraulically connected water-bearing zones in the Redwall and Muav Limestones underlying the Coconino aquifer have been referred to as the Redwall-Muav Limestone in the Little Colorado River Basin (Hart and others, 2002). McGavock and others (1986) characterized the limestone aquifer in the area as consisting of several hydraulically connected limestone, sandstone, and shale units including the Tapeats Sandstone, Bright Angel Shale, Muav Limestone, Temple Butte Limestone, Martin Formation, and the Redwall Limestone.

The Naco Formation, because of similar physical properties, may be considered as part of the Redwall-Muav aquifer (Parker and others, 2005). For this study, the Naco Formation, which lies between the Redwall and Lower Supai Formations in eastern Arizona, is also grouped as part of the Redwall-Muav aquifer. The Naco Formation is present near upper Fossil and Tonto Creeks and extends toward the eastern part of the study area. Springs near Fossil Creek discharge from rocks at the base of the Naco Formation or the top of the Redwall Limestone (Parker and others, 2005).

The Redwall Limestone occurs throughout the study area and is the upper rock unit of the Redwall-Muav aquifer where the Naco Formation is absent (fig. 7). The formation crops out in steep canyons and escarpments in the northern, southern, and western parts of the study area at or near locations of groundwater discharge. The Redwall Limestone is variably saturated in the study area. In a few places along the north and south rims of the Grand Canyon, the Redwall Limestone is unsaturated to partly saturated where groundwater migrates into lower units of the aquifer.

The Temple Butte Formation, Martin Limestone, and the Muav Limestone underlie the Redwall Limestone near the the Grand Canyon in the area of the Coconino Plateau (fig. 7). The Temple Butte Formation thins southward from the Grand Canyon and overlies the Martin Formation or is absent. The Temple Butte Formation and other Devonian limestone rock units are exposed along the north and south rim of Grand Canyon and locally may be partly saturated in areas where groundwater discharges from the Redwall-Muav aquifer (Huntoon, 1977). Well records indicate, however, that these rock units do not extend for substantial distances south of Grand Canyon, and, therefore, are not considered to be a major part of the Redwall-Muav aquifer on the Coconino Plateau. The Martin Formation is mainly in the central and southern part of the plateau and thickens to the south.

The Bright Angel Shale and Tapeats Sandstone underlie the Martin, Temple Butte, and Muav Limestones in the central and western parts of the study area (fig. 7). The Bright Angel Shale is several hundred feet thick. The unit is not a major water-bearing unit, however, because the Bright Angel Shale is composed of very fine-grained sediments that impede the downward migration of water (Huntoon, 1977). However, dozens of springs discharge from the Bright Angel Shale in the Grand Canyon from fine grained sandstone, sandy siltstone, and bedding plane fractures (Monroe and others (2004). The Tapeats Sandstone is a major water-bearing unit in places. Tapeats Sandstone is as a continuous unit along the south and north rims of Grand Canyon. South of Grand Canyon, the unit is mainly present as isolated erosion remnants overlying Proterozoic rocks. The Tapeats Sandstone is believed to be hydraulically connected to the overlying Redwall and Muav Limestones through faults and fractures and where the Bright Angel Shale is thin or absent.

The Redwall-Muav aquifer mostly is confined by fine-grained sediments in the overlying Lower Supai Formation and by underlying Proterozoic crystalline rocks. Water recharges the Redwall-Muav aquifer where the unit crops out and through downward leakage from overlying units through faults, fractures, and breccia pipes. Most groundwater in the Redwall-Muav aquifer discharges at springs along the Mogollon Rim, in deep canyons of the Colorado River, and along the lower part of Little Colorado River. Some groundwater in the Redwall-Muav aquifer flows laterally into the basin-fill aquifers of the Verde and Chino Valleys (Blasch and others, 2006). Some groundwater discharges through the bottom of the Redwall-Muav aquifer to underlying areas of permeable basement rocks.

Water Levels and Saturated Thickness

Bills and others (2007) constructed a potentiometric surface map of the Redwall-Muav aquifer by using water-level data for wells and the altitude of springs that discharge from the aquifer (fig. 9). Groundwater flows from recharge areas at high altitudes along the Mogollon Rim and San Francisco Mountain area toward discharge areas at low altitude springs and streams in the Verde Valley and near the Colorado River. Groundwater also discharges from the Coconino aquifer to the Redwall-Muav aquifer upstream from Blue Spring on the Little Colorado River. Some groundwater also may flow from the Coconino aquifer to the Redwall-Muav aquifer near the Mesa Butte fault. The hydraulic gradient in the Redwall-Muav aquifer ranges from about 4.4 to 88 ft/mi in the study area (Bills and others, 2007). The large range in the gradient is a reflection of the varied flow conditions in the aquifer that are largely controlled by geologic structure, solution channel features, and recharge distributions. The gradient is steeper near discharge areas along the south rim of Grand Canyon and south of the Mogollon Rim, and is lower in the interior of Cataract Creek drainage basin. The general groundwater-flow system in the Redwall-Muav aquifer has varied little on the basis of water-level variations of only a few feet at individual wells.

The static water level in wells developed in the Redwall-Muav aquifer range from about 320 ft to more than about 2,890 ft below land surface and can be several hundred feet above the top of the Redwall Limestone (Bills and others, 2007). In groundwater discharge areas in the northern and southern parts of the study area, erosion has removed overlying rock units in steep canyons and escarpments, exposing the aquifer to the atmosphere in small areas and resulting in unconfined conditions. The saturated thickness of the aquifer is roughly the same as the combined thickness of the Redwall and Muav Limestones and the Tapeats Sandstone

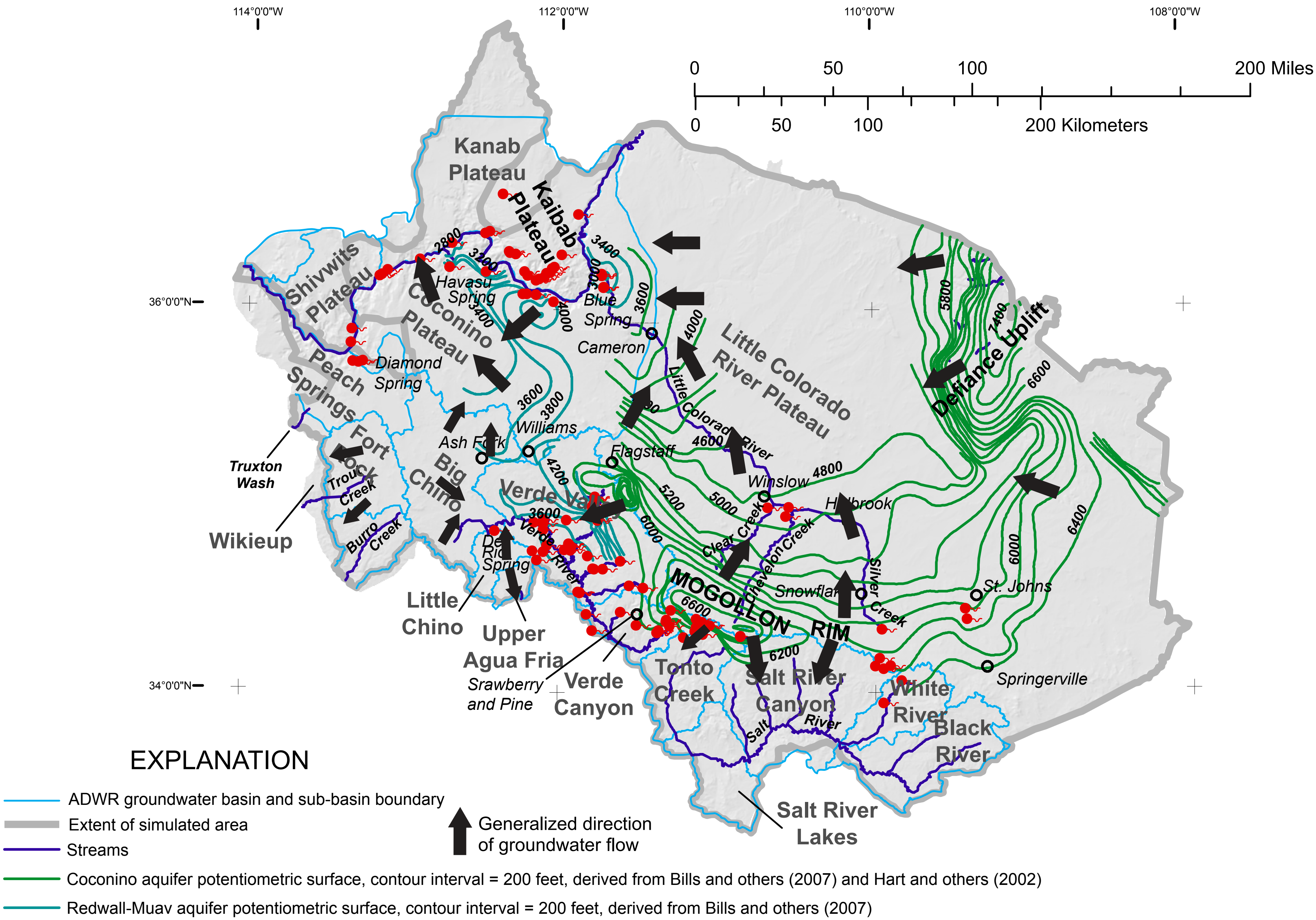


Figure 9. Generalized groundwater-flow system in the major aquifers of the Northern Arizona Regional Groundwater-Flow Model study area.

that ranges from about 640 to 2,000 ft and averages about 1,000 ft in the study area.

Recharge and Discharge

Recharge to the Redwall-Muav aquifer is primarily through downward leakage from overlying units through faults, fractures, and other geologic structures that create secondary porosity and conduits for groundwater flow. Recharge also occurs where the aquifer materials crop out, mainly in areas south of the Mogollon Rim. Normal faults and associated fractures occur throughout the study area and predominantly strike northeast to northwest. Recent analysis of surface geophysical data by Gettings and Bultman (2005) indicate that many of these structures are deep seated, penetrating the Coconino aquifer and the Redwall-Muav aquifer and bottoming in the Proterozoic crystalline rocks. Areas of substantial faulting and fracturing are (1) along the Mesa Butte Fault zone; (2) in the Cataract Creek drainage basin; (3) north of Mount Floyd; (4) along the Bright Angel and Vishnu Faults; (5) in the Cameron area coincident with several large monoclines; and (6) south of Flagstaff in association with extensional basins (Cooley, 1976; Ulrich and others, 1984; Billingsley, 2000; Bills and others, 2000; Billingsley and others, 2006).

Substantial faults and fractures probably are present in the consolidated sediments underlying the San Francisco and Mount Floyd Volcanic Fields, but if so, these faults and fractures are masked by the volcanic rocks. These volcanic fields are areas of major recharge potential because these volcanic fields are at higher altitudes in a region of substantial precipitation. On the west side of Cataract Creek drainage basin, recharge potential is enhanced by the presence of extensive deposits of unconsolidated alluvium that readily permit infiltration of precipitation and runoff. Infiltrating water percolates into the subsurface where fracture zones act as deep conduits to the aquifer. In the eastern and southeastern parts of the study area, groundwater is recharged to the aquifer by downward leakage from the overlying Coconino aquifer where very fine-grained sediments of the Lower Supai Formation have been faulted or fractured.

The Redwall-Muav aquifer also receives groundwater flow as underflow from the Little Colorado River Plateau basin; however, most of that flow is impeded by the more than 500 ft of uplifted low permeability Proterozoic crystalline rocks near the Mesa Butte Fault. As a result, most of the groundwater flow from the Little Colorado River Plateau basin likely discharges along the lower Little Colorado River. Underflow from areas adjacent to the study area is unlikely because the Redwall-Muav aquifer in the study area is topographically higher than in adjacent areas.

The Redwall-Muav aquifer discharges groundwater as (1) spring flow along the lower Little Colorado River and in tributaries of the Colorado River along the north and south rim of Grand Canyon, (2) spring flow to the Verde River and tributaries, (3) underflow into the basin-fill aquifer in Big Chino Valley, (4) underflow into basin-fill aquifer in the Verde Valley, (5) discharge from wells, and (6) evapotranspiration where the water table is at or near land surface.

Metzger (1961) noted that springs issuing from the Redwall Limestone in the lower Little Colorado River and in Havasu Creek and Warm Creek have large discharges, but that most other springs and seeps along the south rim of Grand Canyon that issue from the Redwall and Muav Limestones have small discharges. Along the North Rim of Grand Canyon, discharge of more than 1,000 ac-ft/yr (1.4 ft^3/s) at many springs is derived from recharge on the Kaibab Plateau (fig. 9). The average flow of Blue Spring (fig. 9), one of dozens of outlets from the Redwall-Muav aquifer along the lower Little Colorado River, is about 95 ft^3/s, and the combined flow from all springs in this reach of the Little Colorado River is about 237 ft^3/s. Havasu Spring in Havasu Creek (fig. 9) has a discharge of about 64 ft^3/s. Additional springs that discharge from the Redwall and Muav Limestones downstream to the mouth of Havasu Creek increase the base flow of the creek to about 71 ft^3/s. Below Havasu Creek, springs in deep canyons near the western margin of the Coconino Plateau basin (fig. 9)—including Diamond Creek, Spencer Creek, and Quartermaster Spring—discharge water from the Redwall-Muav aquifer at rates of 2.5, 5, and 5 to 30 ft^3/s, respectively. Numerous smaller springs and seeps discharge from the Redwall-Muav aquifer along the south rim of Grand Canyon (Monroe and others, 2005) (fig. 9). These springs typically are about 3,000 ft below the surface of the Coconino Plateau. Smaller springs discharge at the northwest end of the Grandview Monocline, in the Pipe Creek area on the Bright Angel and Vishnu Faults, in the Monument Creek area, in the Hermit Creek area, and in the area from Royal Arch Creek to Olo Canyon. The largest of these small spring flows from Grandview Monocline to Olo Canyon range from 0.67 to 1.11 ft^3/s and are in the Pipe Creek area and along Hermit Creek. The remaining spring flows in these areas are less than 0.22 ft^3/s. Other minor springs and seeps west of Pipe Creek typically have flow rates of a few gallons per minute or less.

Southward flowing groundwater in the Redwall-Muav aquifer discharges at springs along the upper reaches of the Verde River along lower Sycamore Creek, Fossil Creek, and the East Verde River. Geochemical results indicate that groundwater also discharges from the Redwall-Muav aquifer to springs along the lower reaches of Oak and Wet Beaver Creeks. The base flow of the Verde River increases from about 24 ft^3/s near Paulden to 78 ft^3/s near Clarkdale because of discharge from springs, including Summers Spring in lower Sycamore Creek. Fossil Creek dicharge of about 41 ft^3/s (Feth and Hem, 1963; McGavock and others, 1968) is derived from a combination of local and regional aquifer sources (Green, 2008). Spring flow of about 31 ft^3/s from the Redwall-Muav aquifer to lower Oak Creek is mainly in the Page Spring area (Owen-Joyce and Bell, 1983). This spring issues from either the Verde Formation or the Supai Group (Owen-Joyce and Bell, 1983), but the main orifice is a solution channel in limestone rubble on the west-facing slope of a limestone unit that is consistent with Redwall Limestone lithology. Groundwater in the aquifer that is not discharged as springs or withdrawn by wells flows southward into Verde Valley and is hydraulically connected to groundwater in the Verde Formation (Owen-Joyce and Bell, 1983). The amount of groundwater flow from the Redwall-Muav aquifer to the Verde Formation is unknown (Blasch and others, 2006). Two large-discharge springs flow from below the Paleozoic rocks at the southern end of the study area (Parker and others, 2005). Springs near Tonto Natural Bridge discharge from the Tapeats Sandstone (Feth and Hem, 1963), and R-C Spring east of Tonto Creek discharges about 1.8 ft^3/s from the Mazatzal Group quartzite; however, the volume of spring discharge and the nature of the rocks from which the springs discharge indicate that the main flow paths are through the Redwall-Muav aquifer.

Downward leakage from the Redwall-Muav aquifer to underlying rock units can happen throughout the Coconino Plateau study area where deep-seated faults and fractures extend through the entire sequence of Paleozoic rocks and penetrate the Proterozoic crystalline rocks. In the northern part of the study area, groundwater migrates through faults and fractures in the Bright Angel Shale and the underlying Tapeats Sandstone into the underlying crystalline rocks. Several small springs and seeps discharge less than 0.02 ft^3/s from these Proterozoic rocks several hundred feet to a thousand feet below the main discharge zone of the Redwall-Muav aquifer. Leakage from the aquifer can flow directly into the underlying rocks in the southern part of the study area because the Bright Angel Shale is absent and the Tapeats Sandstone is only locally present. Groundwater from the Redwall-Muav aquifer also can flow laterally into permeable consolidated to unconsolidated valley-fill units in the Big Chino and Verde Valleys; however, the quantity of underflow in these areas is unknown.

Evapotranspiration (ET) can cause seasonal variations in the base flow of streams where rates of evapotranspiration are a substantial percentage of base flow. In the northern part of the study area, shallow groundwater and spring flow support lush riparian habitat in the otherwise arid environment of the south rim of Grand Canyon. Evapotranspiration can account for as much as 10 percent of the base-flow component in Havasu Creek (Bills and Flynn, 2002). Variation in discharge from the Redwall-Muav aquifer because of seasonal evapotranspiration at most other springs is small.

Aquifer Properties and Well Yield

Most wells developed in the Redwall-Muav aquifer in the study area are in small communities, such as Valle, Tusayan, Ash Fork, Seligman, Drake, Tonto Village-Kohls Ranch area, and Supai. A few municipal wells have been developed in the aquifer in Verde Valley near Sedona and in the Williams area. Data on aquifer properties and well yield—transmissivity, hydraulic conductivity, storage coefficient, and specific capacity—for the Redwall-Muav aquifer are lacking because of the few wells that have been developed in this aquifer and few aquifer tests that are available for analysis. Aquifer properties are affected by formation lithology and geologic structure. Structural development (faulting and fracturing) has resulted in secondary permeability that greatly influences the movement of groundwater in the aquifer.

Available data on aquifer properties for the Redwall-Muav aquifer were compiled from previous reports. The data indicate that the aquifer is anisotropic and confined in much of the study area. Small parts of the aquifer are unconfined where the aquifer rocks crop out. Wells drilled along extension faults and fractures typically penetrate zones of increased transmissivity because of the solution-enhanced permeability (Errol L. Montgomery and Associates, 1999). Storage coefficients are not available for the Redwall-Muav aquifer; however, meager test data indicate that storage likely is influenced by structure. The storage coefficient probably is low in areas where data from wells drilled into unfractured or slightly fractured limestones indicate low transmissivity (Montgomery, 1981). Levings (1980) completed multiple aquifer tests in the Sedona area where transmissivity of the Redwall Limestone at well (A-17-05)33ada was 16,000 ft^2/d.

Yields from wells developed in the Redwall-Muav aquifer range from less than 1.0 gal/min to more than 1,000 gal/min. Factors that contribute to the large range in yields from the Redwall-Muav aquifer include lithology, degree and type of fracturing, saturated thickness penetrated by the well, and pump design and lift. Dissolution of limestone and the widening of fractures by dissolution also contribute substantially to the large range of well yields. One test well drilled into a zone of secondary fractures in the Redwall Limestone to the south of the Anderson Mesa Fault, south of Flagstaff, yielded 35 gal/min (Montgomery, 1981). Recent wells drilled into the Redwall and Muav Limestones and the Martin Formation east of Williams along the Mesa Butte Fault zone yield 7.0 to 280 gal/min (Dennis Wells, city manager, city of Williams, written commun., 2004). Wells drilled in areas not near regional faults in the Valle and Tusayan areas typically produce about 40 to 50 gal/min (Bills and Flynn, 2002). Well yields from the Redwall-Muav aquifer in Verde Valley range from less than 10 gal/min to about 1,100 gal/min (Owen-Joyce and Bell, 1983; Bills and Flynn, 2002).

Coconino Aquifer

Cooley and others (1969) defined the C multiple-aquifer system (Coconino aquifer) as the sequence of rock units between the Kaibab Formation and the Supai Group (fig. 7). This definition has been refined to include the Kaibab Formation, Toroweap Formation, the Coconino Sandstone (and lateral equivalents, the DeChelly Sandstone and Glorieta Sandstone in New Mexico), the Schnebly Hill Formation, and the Upper and Middle Supai Formations (McGavock and others, 1986; Bills and others, 2000; and Bills and Flynn, 2002; Bills and others, 2007). The Kaibab Formation underlies the Shinarump Member of the Chinle Formation or the Moenkopi Formation. The primary water producing unit is the Coconino Sandstone; however, the overlying Kaibab and Toroweap Formations and the underlying Schnebly Hill Formation and Upper and Middle Supai Formations of the Supai Group can be locally major water producing units (Leake and others, 2005). The Lower Supai Formation typically forms a confining unit that separates the Coconino aquifer from the underlying Redwall-Muav aquifer and local Proterozoic crystalline aquifers. West of the Mesa Butte Fault, the Coconino aquifer is locally present, but the primary water-bearing zones are unsaturated across broad regions because of large depths to water. The Coconino aquifer is locally perched in places on the Coconino Plateau (Bills and others, 2007). The Coconino aquifer thins toward the east. The exact eastern boundary is uncertain, however, because water level and geologic data are meager. Groundwater in the Coconino aquifer is unconfined except where the base of the Moenkopi Formation falls below the potentiometric surface across much of the region north of Little Colorado River. Many wells drilled into the confined part of the aquifer flowed at land surface before significant development of the groundwater supplies (Mann and Nemecek, 1983; Mann, 1976).

The Coconino Sandstone, the Schnebly Hill Formation, and the Upper and Middle Supai Formations, crop out in steep canyons or escarpments in the northern, southern, and western parts of the study area near groundwater discharge zones of the Verde River, Salt River, and Little Colorado River watersheds. The Kaibab Formation crops out over large parts of the Colorado Plateau, north of the San Francisco and Mount Floyd Volcanic Fields and southeast of the San Francisco Volcanic Field (fig. 9). The Kaibab Formation is dry except for perched groundwater in areas north of Williams and north of the Little Colorado River.

Water Levels and Saturated Thickness

Bills and others (2007) constructed maps of the potentiometric surface of the Coconino aquifer on the basis of earlier studies by Hart and others (2002), Bills and others (2000), and Owen-Joyce and Bell (1983) (fig. 9). A groundwater mound in the Coconino aquifer south of Flagstaff near the Mogollon Rim indicates greater recharge in that area. The groundwater mound forms a groundwater divide that is not fixed spatially nor temporally and can be affected by groundwater withdrawals and variations in recharge distributions. Groundwater flows from the divide northward toward the Little Colorado River and from the divide southward to the Salt River and Verde River basins. A groundwater mound also is coincident with the Defiance Uplift (fig. 9). This groundwater mound also forms a groundwater divide that defines the northeast extent of the regional groundwater-flow system and study area. Groundwater from the Defiance Uplift flows westward beneath Black Mesa and discharges at springs on the lower Little Colorado River and in Marble Canyon (Cooley and others, 1969; Cooley, 1972). Estimated hydraulic gradients in the aquifer range from about 40 to 100 ft/mi (Bills and others, 2000). The varied hydraulic gradient is a reflection of the varied flow conditions in the aquifer that are largely controlled by geologic structure. As groundwater flows from recharge areas toward discharge areas, groundwater also tends to migrate deeper into the subsurface along fractures and faults.The saturated thickness of the Coconino aquifer ranges as much as about 2,000 ft between the Defiance Uplift and Black Mesa, north of Leupp, and near Flagstaff. North and west of Flagstaff, the Kaibab Formation, the Coconino Sandstone, and the Schnebly Hill Formation are unsaturated, and the underlying sandstone units of the Upper and Middle Supai Formations are the primary water-bearing units in the aquifer. The Kaibab Formation and the Coconino Sandstone also are largely unsaturated south of the Mogollon Rim where the Schnebly Hill Formation and the Upper and Middle Supai Formations are the primary water-bearing units. The Coconino aquifer is locally saturated west of the Mesa Butte Fault where it may be locally perched.

Water-level trends in the Coconino aquifer at wells that have repeated observations are varied in the study area. Variations since 1983 range from a few feet of decline or rise in areas with little groundwater withdrawal to more than 100 ft of decline near pumping centers for municipal and industrial supply including three power plants. Several Coconino aquifer wells in areas of scant groundwater withdrawal indicate water-level rises that appear to correspond with variations in precipitation and recharge including a few wells near the Mogollon Rim and a few wells below the Mogollon Rim near Oak Creek and Dry Beaver Creek in the Verde Valley. As groundwater withdrawals progressed, most of the flowing wells in the confined part of the aquifer ceased flowing and some historical spring areas stopped flowing (Mann and Nemecek, 1983; Mann, 1976).

Recharge and Discharge

Recharge to the Coconino aquifer is primarily in the southern part of the study area in the highlands of the Mogollon Rim and on the Defiance Uplift in the northeastern part of the study area. Recharge mechanisms include direct infiltration of precipitation and infiltration of runoff. A major part of the recharge process is the interception of runoff by open fractures and solution channels developed on the Kaibab Formation surface (Bills and others, 2000; Wilkinson, 2000). The aquifer also is recharged by downward leakage of groundwater from overlying perched zones and through the overlying volcanic rocks. Small quantities of water are recharged to the aquifer from infiltration of treated municipal effluent along drainages in the Kaibab Formation near Flagstaff (Bills and others, 2000). No substantial quantities of groundwater likely enter the Coconino aquifer from areas adjacent to the study area. Groundwater in the aquifer generally is unconfined except north of the Little Colorado River where the base of the Moenkopi Formation lies below the aquifer water level (Mann, 1976; Daniel, 1981; Mann and Nemecek, 1983; Leake and others, 2005). The transition between confined and unconfined conditions generally is near the Little Colorado River.

Groundwater flows from recharge areas along the Mogollon Rim and Defiance Uplift toward discharge to other aquifers and at springs and perennial streams in the Little Colorado River and Verde River drainage systems. Groundwater in the Coconino aquifer in the Little Colorado River Plateau basin discharges to the Redwall-Muav aquifer before primarily discharging to a series of springs, collectively referred to as Blue Springs, directly into the Little Colorado River about 13 mi upstream from the mouth of the Colorado River (Hart and others, 2002). Discharge from the predominant spring, Blue Spring, and other springs in the area is estimated to be about 237 ft^3/s (Hart and others, 2002). Some groundwater discharge also likely occurs directly to the Colorado River, but at an uncertain rate (Cooley and others, 1972: Bills and others, 2007). Groundwater discharge occurs to Little Colorado River tributary streams Clear, Chevelon, and Silver Creeks. Most of this discharge, however, reinfiltrates the stream channels and recharges the aquifer in downstream reaches along the Little Colorado River. Groundwater from the Coconino aquifer also discharges to springs on the Defiance Uplift in a few the drainages such as Canyon de Chelly (Cooley and Others, 1969; Hart and others, 2002). Groundwater from the Coconino aquifer also discharges as spring flow into tributaries of the Verde and Salt Rivers in a region between Sycamore Creek and the Black River. Groundwater from the aquifer that does not discharge to springs or withdrawn by wells flows south toward the Verde Valley and discharges as groundwater flow to the Verde Formation (Owen-Joyce and Bell, 1983) and to the basin-fill aquifer in the Tonto Creek Basin. Small amounts of groundwater also discharge from the aquifer as evapotranspiration in a few riparian areas south of Flagstaff where the roots of phreatophytes can access shallow water tables. Municipal and public supply wells discharge water from the Coconino aquifer, sometimes in large amounts, near Flagstaff and in Verde Valley near Sedona. Large amounts of withdrawals for industrial use have occurred at power plants near Springerville, St. Johns, Holbrook, and at a paper mill near Snowflake. Agriculture in the Little Colorado River Plateau basin is partially supported by several high-yield irrigation wells developed in the Coconino aquifer in a large region from St. Johns to Winslow. A few wells drilled north and west of Williams penetrated small perched zones in the Coconino Sandstone or in sandstone beds in the Upper and Middle Supai Formations at depths of about 1,200 ft to more than 2,000 ft below land surface.

Aquifer Properties and Well Yield

Aquifer properties for the Coconino aquifer have been estimated by many previous studies (Cooley and others, 1969; Owen-Joyce and Bell, 1983; McGavock and others, 1986; Bills and others, 2000; Bills and Flynn, 2002; Southwest Groundwater Consultants, Inc., 2003; Hoffmann and others, 2005). Data indicate that the Coconino aquifer on the Colorado Plateau is anisotropic (Cooley and others, 1969; McGavock and others, 1986; Bills and others, 2007). Bills and others (2000) noted that hydraulic conductivity values generally are greater near wells developed in the Coconino Sandstone or sandstone beds of the Upper and Middle Supai Formations and are lower near wells developed in the Kaibab or Schnebly Hill Formations. Hydraulic conductivity values generally are greater where extensive fracturing is present. Transmissivity values range from 10 to 4,700 ft^2/d and hydraulic conductivity values range from 0.019 to 6.88 ft/d for the Flagstaff area (Bills and others, 2000). Mann and Nemecek (1983) report a transmissivity range of 940 to 9,100 ft^2/d for the aquifer in southern Apache County. Transmissivity values in Chinle Valley and on the Defiance Plateau range from 31 to 35,000 ft^2/d (Cooley and others, 1969). Data from aquifer tests near Leupp, Arizona (Hoffmann and others, 2005), indicate that hydraulic conductivity of the Coconino Sandstone ranges from 11 to 28 ft/d; values range from 0.9 to 8 ft/d where the Coconino Sandstone interfingers with the Schnebly Hill Formation; the Upper Supai Formation had the smallest values (0.1 and 0.2 ft/d). Transmissivity of the Supai Group at well (A-17-05)19aaa near Sedona was 10,000 ft^2/d (Levings, 1980). The large variance in the transmissivity of the Coconino aquifer is attributed to degree of fracturing and differences in the penetration depths of wells used for aquifer tests (Leake and others, 2005). In addition, Bills and others (2000) report that higher well yields correlate to north-to northwest-trending faults and fractures that are largely extensional and that wells near-northeast trending structures, which are dominantly compressional, had lower yields. Many wells were designed to meet small demands for water and, therefore, were drilled to a depth where the required yield could be obtained. Some of these wells penetrate only a part of the total aquifer thickness, resulting in an underestimate of total transmissivity for the aquifer.

The storage coefficient and specific yield of the Coconino aquifer are related to the lithology and geologic structure of the rock units. Estimated storage properties for the Coconino Sandstone are about 0.07 for specific yield and 1×10^{-4} for storage coefficient based on the few aquifer tests available for the study area (Cooley and others, 1969; McGavock and others, 1986; Mann, 1976; Mann and Nemecek, 1983; and Southwest Groundwater Consultants, Inc., 2003).

Several factors contribute to a large range in well yields from the Coconino aquifer: (1) formation lithology, (2) degree and type of fracturing, (3) degree of secondary mineralization of the aquifer, (4) saturated thickness penetrated by the well, (5) well efficiency, and (6) pump design. Bills and Flynn (2002) report yields that range from about 1 to 1,700 gal/min. Bills and others (2000) indicated that the degree and type of fracturing has the greatest effect on yields for wells developed in the Coconino aquifer near Flagstaff. Wells that yield less than 100 gal/min generally are not completed in or near faults or other fractures, and wells that yield greater than 100 gal/min are completed in or near faults and fractures. Large well yields from the Coconino aquifer in the central part of the Little Colorado River Plateau basin occur where at least 75 percent of rock units that comprise the aquifer are saturated (Nemecek and Mann, 1983). Kaczmarek (2003) observed pumping rates of 20–80 gal/min in the Schnebly Hill Formation near Strawberry and only 10–30 gal/min in the Supai Formation near Pine. Kaczmarek (2003) hypothesizes that even though proximity to fractures in the host rock is consequential, the greater primary porosity of the Schnebly Hill Formation than the Supai Formation should be considered as contributing to the greater withdrawal rate.

Alluvial Basin Aquifers

Five primary alluvial basin aquifers occur in the study area, including Big Chino Valley and Williamson Valley, Little Chino Valley and Lonesome Valley, Upper Agua Fria basin, Verde Valley, and the Tonto Creek Basin. Aquifers in the alluvial basins include lower and upper parts of the basin fill, two primary facies of coarse- and fine-grained alluvial deposits, and intercalated basalt flows. The coarse-grained facies of the basin fill generally lies on the basin boundaries. The fine-grained facies generally lies in the deepest part of the structural basin. Both facies of the upper basin fill are generally more permeable than the equivalent facies in the lower basin fill. Basalt flows are intercalated with alluvial deposits in the upper and lower basin fill. Basalt flows are associated with highly permeable zones in the upper basin fill in the Paulden area of the Big Chino Basin and in the lower basin fill in the Little Chino basin where the lower basin fill is known as the lower volcanic unit. Elsewhere, intercalated basalt flows are poor aquifers. The basin-fill aquifers are generally unconfined on the basin boundaries and confined in the basin centers where overlain by partially saturated confining beds of basalt or the fine-grained facies.

Thin stringers of Quaternary flood-plain alluvial and terrace deposits occur along major streams, but are not major aquifers because the materials are rarely saturated. An exception is along the Verde River in the Camp Verde area where depths to water are shallow and the unit receives discharge from permeable zones in the Verde Formation. Where unsaturated, these highly permeable deposits are effective conduits for the transmission of infiltrated streamflow to the underlying aquifer. Perched aquifers may form locally where the Quaternary alluvium overlies low permeability rocks or the fine-grained facies of basin fill. Perched aquifers form in the Paulden area where the alluvial deposits are shallowly unerlain by the fine-grain facies of basin fill and along the Verde River upstream from the confluence with Oak Creek where the alluvial deposits overlie confining beds of the Verde Formation. Thickness of the alluvium is 60 to 100 ft in the Verde Valley (Owen-Joyce, 1984) but is generally less elsewhere. Only about 30 ft or less of the alluvium is commonly saturated.

Climate and stable-isotope data on precipitation and groundwater indicate that recharge in the alluvial basins is derived from precipitation that falls on the surrounding mountains. Blasch and others (2006) estimated that less than 2 percent of precipitation recharges the aquifer in the Big Chino sub-basin. Recharge happens through direct infiltration of precipitation that exceeds runoff and evapotranspiration at high altitudes and as infiltration of runoff in low-lying areas and ephemeral stream channels where local infiltration of accumulated runoff in highly permeable alluvium exceeds evapotranspiration rates. Much of the recharge preferentially infiltrates permeable rocks and solution channels in Paleozoic limestone outside of the alluvial basin. Most of the remaining recharge infiltrates through the ephemeral stream channels in the alluvial basin. The relative distributions of the recharge source are poorly understood. However, the greatest percentage of recharge is likely in the mountains and near the base of the mountains through a process that has been commonly known as “mountain front” recharge. Substantial recharge may occasionally happen during large runoff events along major ephemeral channels that cross alluvial basins.

Big Chino Sub-Basin

The lower and upper basin fill form an alluvial aquifer system in the Big Chino sub-basin, including Williamson Valley. The extent of the basin-fill aquifer is less than the sub-basin extent, and it occurs only in the alluvial basin that lies southwest of Big Black Mesa. Groundwater in two areas of the Big Chino sub-basin does not flow into the basin-fill aquifer in the Big Chino alluvial basin including an extensive region north of Big Black Mesa and a small region northeast of Granite Mountain near Mint Wash. The basin-fill aquifer is hydraulically connected to local granite and Redwall-Muav aquifers in the surrounding mountains, but the connection between the aquifers is poorly understood. The basin-fill aquifer is hydraulically connected to the underlying Redwall-Muav aquifer in the lower part of the Big Chino sub-basin. The two aquifers are likely hydraulically connected in the upper part of the sub-basin, although supporting data are limited. Lower basin fill is the most extensive water-bearing unit in the basin. The fine-grained facies of the basin fill, which is in the southeast and central parts of the alluvial basin, creates confined conditions in underlying aquifers including coarse-grained portions of the lower basin fill and Redwall-Muav aquifer. Basalt flows are intercalated with the basin fill and are associated with a region of highly permeable deposits in the Paulden area. Basalt flows also occur in the upper part of the Big Chino alluvial basin; however, these volcanic rocks are not known to be productive aquifers. A thin layer of Quaternary alluvium overlies the fine-grained facies of upper basin fill near the Big Chino Wash and forms a local perched aquifer that has hydraulic heads that are as much as 100 ft above the hydraulic heads of the lower basin fill. A thin layer of Quaternary alluvium also is a local aquifer near Williamson Valley Wash. Average thickness of the lower and upper basin fill is about 1,000 ft; however, the maximum thickness is more than 3,000 ft (Langenheim and others, 2005). Depths to water are about 100 ft or less in the central parts of the Big Chino basin-fill aquifer, but as much as 300 ft near the boundaries of the basin-fill aquifer.

Water-level declines of less than 15 ft have been measured in the basin-fill aquifer since development of groundwater supplies. In the Williamson Valley area, water-level declines of about 5 ft or less were measured before the mid-1960s, but have not varied substantially since that time. Declines of less than 15 ft were measured before the mid-1970s in the agricultural area above Walnut Creek. Less than 5 ft of decline were measured before the mid-1960s in the Paulden area. Water levels in the Big Chino rose a few feet after the mid-1960s in the agricultural area upstream from Walnut Creek and in the Paulden area before declining slightly beginning in the late 1990s. Causes of water-level variations after the initial declines are uncertain, but likely include variations in irrigation practices and variations in rates of recharge and withdrawals.

Recharge and Discharge

In addition to infiltration of precipitation and runoff, undetermined amounts of groundwater recharge likely enters the basin-fill aquifer in the Big Chino sub-basin as groundwater flow from the Redwall-Muav aquifer near the Juniper Mountains, Big Black Mesa, and from the Little Chino sub-basin downstream from Del Rio Springs (Nelson, 2002). Water Resources Associates (1990) estimated that groundwater underflow also enters the basin-fill aquifer along the northwestern boundary of the Big Chino sub-basin. As a result of groundwater use, a part of the withdrawn water also recharges the aquifer through infiltration beneath septic systems and as excess applied irrigation water.

Groundwater in the Big Chino basin-fill aquifer discharges primarily to the Redwall-Muav aquifer in the Paulden area, which then discharges at the Verde River headwater springs and through well withdrawals. Groundwater also discharges from the Big Chino sub-basin to the Little Chino sub-basin from local alluvial or granitic aquifers in the Mint Wash area northeast of Granite Mountain. Groundwater discharges from the Big Chino basin-fill aquifer through evapotranspiration in two areas prior to development of groundwater supplies. Phreatophytes access shallow groundwater that has persisted near Williamson Valley Wash since before development of groundwater supplies. Wirt and others (2005) report that phreatophytes likely accessed shallow groundwater in the region between the confluence of Big Chino Wash with Pine and Walnut Creeks prior to local well withdrawals. Some of this groundwater use by phreatophytes may have continued through the period of groundwater development.

Water-level variations in the Big Chino sub-basin closely correspond with variations in precipitation and runoff. Some increased recharge in the Big Chino and Williamson Valley Wash areas may result from slight lowering of shallow water levels near streams, which allows greater storage volume for the infiltration of runoff. Records of relatively stable water levels in the Williamson Valley area while irrigation withdrawals were occurring may have resulted from several possible changes in the local groundwater-budget including increased recharge, reduced net groundwater withdrawals, or reduced evapotranspiration rates resulting from greater depths to water. Each of these mechanisms may have contributed to locally balanced groundwater budgets and stable water levels after the early 1960s near the Big Chino and Williamson Valley Washes.

Aquifer Properties

Aquifer and specific capacity tests for the Big Chino basin-fill aquifer that have been done by several investigators resulted in large range in hydraulic properties. Aquifer tests done in the Tertiary alluvial and basalt layers of Big Chino Valley by Water Resources Associates (1990) resulted in a range in transmissivity values of 19,100 to 334,000 ft^2/d. Ewing and others (1994) reanalyzed the test data and calculated a range in transmissivity values of 21,500 to 246,000 ft^2/d. Clear Creek and Associates (2008) did specific capacity and aquifer tests at 24 wells in the upper part of the alluvial basin. Transmissivity estimates derived from the specific capacity tests ranged from about 100 to 32,000 ft^2/d. Transmissivity estimates derived from four aquifer tests ranged from about 15,000 to 18,000 ft^2/d. Hydraulic conductivity estimates ranged from 1 to 65 ft/d.

Specific-capacity values, in Big Chino Valley, are largest for the water-bearing unit(s) near the junction of the valley with Williamson Valley and west of Paulden, where values are as great as 52 gal/min/ft. Near the eastern extent of the basin-fill aquifer, specific-capacity values have greater variability. Specific-capacity values associated with the basin fill in Williamson Valley are greatest in the north-central part of the valley, where values range from 5 to 42 gal/min/ft, and values are less than 1.5 gal/min/ft near the boundaries of the basin-fill aquifer. Specific capacity estimates for the upper part of the alluvial basin (Clear Creek and Associates, 2008) ranged from 0.5 to 160 gal/min/ft.

Navarro (2002) estimated that channel permeability along Mint Wash ranged from 0.4 ft/d to more than 10 ft/d on the basis of permeability measurements and soil classification information. Navarro (2002) relied on groundwater-flow model calibration for estimates of hydraulic properties because of mixed results from multiple aquifer tests. Hydraulic conductivity for basalt was 2 ft/d, values for granite and gneiss were 2.7 ft/d, conglomerate values averaged 6.3 ft/d and ranged from 8.9 to 53.9 ft/d. Estimates of specific yield values ranged from 0.1 to 0.2.

Little Chino and Upper Agua Fria Sub-Basins

The basin-fill aquifer in Little Chino Valley has been described by Matlock and others (1973), Corkhill and Mason (1995), Nelson (2002), and DeWitt and others (2005). Volcanic and basin-fill deposits of Quaternary and Tertiary age are described by Krieger (1965) and Wilson (1988). The lower basin fill and intercalated basalt flows in the Little Chino sub-basin (lower volcanic unit) and upper basin fill (upper alluvial unit) form the basin-fill aquifer system in the Little Chino and Upper Agua Fria sub-basins. The basin-fill aquifer may be hydraulically connected to local granite aquifers in the surrounding mountains. The aquifer also may be hydraulically connected to the underlying Redwall-Muav aquifer in the northern part of the Little Chino sub-basin, but the degree of connection is poorly understood. Lower basin fill and intercalated basalt flows form the primary aquifer in the basin. Unconfined conditions generally exist except in the lower part of the Little Chino sub-basin where confined areas (Schwalen, 1967) in the lower basin fill result from interbedded low permeability zones and an overlying fine-grained facies of the upper basin fill. Thin stringers of Quaternary alluvium overlie the upper basin fill along Granite Creek and the Agua Fria River. However, the Quaternary alluvium along Granite Creek is unsaturated. The Quaternary alluvium along the Agua Fria River is saturated near Humbolt and serves as a highly permeable conduit that drains the basin-fill aquifer. Thickness of the basin-fill sediments generally ranges from about 100 to 800 ft (Ed DeWitt, research geologist, U.S. Geological Survey, written commun., 2005). Depths to water are about 100 ft or less in the lower parts of both sub-basins, but as much as 300 ft near the margins of the basin-fill aquifer and at the groundwater divide between the Little Chino and Upper Agua Fria sub-basins.

Water levels declined as much as 100 ft in the lower basin-fill aquifer since development of groundwater supplies began in the 1930s with the drilling of several flowing wells (Schwalen, 1967). Similar to the Big Chino sub-basin, the greatest rates of water-level decline occurred in the agricultural area of the Little Chino Valley upstream from Del Rio Springs before the mid-1960s and water levels later stabilized or declined at a reduced rate. Causes of water-level variations after the initial declines likely include variations in withdrawals and irrigation practices and variations in recharge rates. Water levels in the area of the upper basin-fill aquifer that lie above the confined lower basin fill in the Little Chino Basin have not declined substantially because the aquifer is not as heavily used and because of recharge from excess applied irrigation water and ephemeral channel infiltration along Granite Creek (Nelson, 2002). Water-level declines in the Upper Agua Fria River basin have generally been less than about 40 ft, except for an area of greater decline near agricultural withdrawals near the Agua Fria River.

Recharge and Discharge

The primary mechanism of recharge to the basin-fill aquifer system in the Little Chino and Upper Agua Fria sub-basins is likely infiltration and deep percolation of runoff into highly permeable alluvial deposits along ephemeral streams. Variable recharge rates along ephemeral channels are likely and have been simulated in groundwater-flow models (Timmons and Springer, 2006; Nelson, 2002; Corkhill and Mason, 1995). In addition to infiltration of precipitation and runoff, the basin-fill aquifer in the Little Chino and Upper Agua Fria sub-basins also is recharged by groundwater flow from local granite aquifers in the Bradshaw Mountains, local alluvial aquifers near the mountains that may be perched above low permeability granitic rocks, and potentially could receive flow from local disconnected Redwall-Muav aquifers in the Black Hills and Mingus Mountain. The basin-fill aquifer receives some groundwater flow from the part of Mint Wash that lies in the Big Chino sub-basin near Granite Mountain. As a result of development of surface-water and groundwater supplies, groundwater also recharges along irrigation distribution systems and a part of the withdrawn groundwater recharges the aquifer through septic systems and as excess applied irrigation water. The groundwater divide between the Little Chino and Upper Agua Fria sub-basins is well defined by recent water-level data, but the predevelopment divide is poorly defined by sparse data.

Groundwater discharge from the Little Chino and Upper Agua Fria basin-fill aquifer primarily discharges to Del Rio Springs and the Agua Fria River upstream from Humbolt, and to the Redwall-Muav and Big Chino alluvial aquifers near Del Rio Springs. Some groundwater likely flows through alluvial and volcanic deposits near Del Rio Springs. Evapotranspiration from the groundwater system by phreatophytes occurs near Del Rio Springs and along the Agua Fria River near Humbolt.

Longer-term water-level declines in the Little Chino basin are related to groundwater withdrawals in excess of recharge rates. Short-term, decadal or shorter, trends in water levels closely correspond with variations in precipitation and runoff. The correspondence of water-level trends and precipitation likely results from variations in irrigation practices and recharge rates.

Aquifer Properties

Aquifer property estimates for the basin-fill aquifer in the Little Chino and Upper Agua Fria sub-basins are available from aquifer tests and numerical simulations of groundwater flow (Timmons and Springer, 2006). Published results are available for aquifer tests of the alluvial and volcanic units in Little Chino Valley near Del Rio Springs (Allen, Stephenson, & Associates, 2001). Reported transmissivity values range from 51,000 to 73,500 ft^2/d and average 59,000 ft^2/d. Storativity values range from $1.2x10^{-8}$ to $7.17x10^{-5}$ ft^{-1} and average $2.9x10^{-5} ft^{-1}$. Simulated values of hydraulic conductivity for the upper basin fill range from 0.3 to 25 ft/d. Simulated values of hydraulic conductivity for the lower basin fill range from 0.1 to 175 ft/d and specific storage values range from $1x10^{-6}$ to $2x10^{-5}$ ft^{-1}. Specific yield values for the lower and upper basin fill range from 0.05 to 0.12.

Specific-capacity values for Tertiary rock units in Little Chino and Lonesome Valleys are in general smaller than those values in Big Chino Valley. The median value for wells near the town of Chino Valley is 5 gal/min/ft. Values tend to increase southward in part because of fewer confining units in the subsurface. Unconfined conditions tend to produce larger values than confined conditions. Near Prescott, values range from <1 to 520 gal/min/ft; median values are about 1 gal/min/ft. In general, specific-capacity values in Lonesome Valley are less than those values in Little Chino Valley; however, large ranges of values are in both areas.

Verde Valley Sub-Basin

The basin-fill aquifer in the Verde Valley includes lower and upper parts. The upper basin fill in the Verde Valley is predominately a fine-grained facies that is a fluvio-lacustrine deposit known as the Verde Formation, which also is the primary aquifer in the basin (Twenter and Metzger, 1963). Groundwater is primarily stored in the limestone, sandstone, and conglomerate beds that are intermixed with less permeable mudstone, claystone, and basalt flows (Owen-Joyce, 1984). Groundwater flow in the Verde Formation is enhanced by solution channels and collapse structures. A coarse-grained facies of basin fill surrounds and is intercalated with the fine-grained facies of the Verde Formation (Twenter and Metzger, 1963), but is much less extensive. The greatest accumulation of the coarse-grained facies of basin fill is north and west of Cottonwood. Lower basin fill is not extensively developed in the Verde Valley because it is commonly found at great depth. Groundwater is commonly confined in the upper basin-fill except where the water table lies in the coarse-grained facies on the basin margins. Narrow and thin stringers of Quaternary alluvium along the major stream channels form an effective water-bearing unit that functions as a receptacle for stream channel recharge along tributary streams, is a conduit for drainage of water from the Verde Formation in the Camp Verde area, and includes perched aquifers especially along the Verde River upstream from Oak Creek. The maximum estimated thickness of the basin fill sediments is 4,200 ft from gravity models (Langenheim and others, 2005). The deepest well known in the basin was drilled to a depth of 2,078 ft in the south-central part of the valley and did not fully penetrate basin-fill sediments. Depths to water in the basin-fill aquifer are generally a few tens of feet or less, but as much as 400 ft near the margins of the basin.

Water levels have declined as much as 20 ft in many parts of the Verde Valley basin-fill aquifer. As much as 100 ft of decline have been detected in parts of the Verde Formation near Cottonwood and near Lake Montezuma. Water-level records in several areas, however, have not indicated any consistent downward or upward trends. Early water-level changes near the Verde River and other tributaries likely resulted from diversion of water from streams for agricultural use, but these changes are not documented. Local perched aquifers likely developed in agricultural areas as a result of recharge in excess of irrigation needs.

Recharge and Discharge

The primary mechanism of recharge to the basin-fill aquifer system in the Verde Valley is groundwater flow from the Redwall-Muav and Coconino aquifers, which received recharge in the higher altitude near the Mogollon Rim through infiltration of precipitation that is in excess of evapotranspiration rates and runoff. Groundwater also recharges along tributary streams where the stream alluvium is in contact with underlying coarse-grained facies of basin fill such as along the lower reaches of West Clear Creek and Dry Beaver Creek. As a result of development of surface-water supplies, groundwater recharges along irrigation distribution systems. Some recharge also occurs through septic systems and in a few areas as deep percolation of excess irrigation water derived from groundwater supplies.

Groundwater discharges from the basin-fill aquifer in the Verde Valley primarily as discharge to the Quaternary alluvium and Verde River and through well withdrawals. Some groundwater may discharge from the basin through the Verde Fault zone, but the rate is uncertain. Evapotranspiration from the basin-fill aquifer by phreatophytes occurs near perennial springs, and along tributary streams. Evapotranspiration does not happen directly from the basin-fill aquifer near the Verde River but from hydraulically connected Quaternary alluvium along the Verde River and perched zones in the Quaternary alluvium along tributaries.

Aquifer Properties

Levings (1980) completed multiple aquifer tests in the Sedona area. The aquifer test at well (A-16-04)27dcc indicated a transmissivity of 20 ft^2/d for the Verde Formation, and the test at well (A-15-04)12abd indicated a transmissivity of 50 ft^2/d for the combined Verde Formation and parts of the Supai Group. Wells in Verde Valley generally have specific-capacity values that are intermediate between wells in Little Chino and Big Chino Valleys. Most of these wells are completed in the Verde Formation or the unconsolidated Quaternary alluvium in areas adjacent to the Verde River and major tributaries. Values range from <1 to 1,500 gal/min/ft; most values are in the range of 6 to 110 gal/min/ft. The large range is attributed to varying degrees of fracturing, faulting, and (or) solution-channel development.

Tonto Sub-Basin

Upper and lower layers of the basin-fill aquifer in the Tonto Creek sub-basin have been detected (Schumann and Thomsen, 1972). The upper layer includes fine-grained facies of poorly consolidated clay and silt and a coarse-grained facies that includes sand and gravel near the basin fill boundaries. The lower layer is composed of interbedded silt, sand, sandstone, and gravel and interbedded basaltic flows and unconformably overlies Proterozoic schist. The lower layer is confined where it underlies the fine-grained facies of the upper layer. Permeability in the basin-fill aquifer is low because of the silt content and cementation of the two layers. Infiltration of streamflow runoff is the primary recharge mechanism (Schumann and Thomsen, 1972). Well yields are typically less than 10 gal/min; however, isolated parts of the basin fill have produced wells yielding a few hundred gallons per minute (Schumann and Thomsen, 1972). Groundwater discharged from the basin to the Salt River prior to construction of Roosevelt Reservoir and thereafter discharged to the reservoir.

Groundwater Budget Methods

Methods for estimating components of the groundwater budgets are discussed in this section. Estimated rates for each major component are mentioned here but described more thoroughly for each basin and sub-basin later in the discussions that evaluate the simulation of groundwater flow in each major groundwater basin.

Inflows

One objective for this study is to estimate rates and distributions of recharge to the aquifers in the study area. The primary methods that were used to estimate natural recharge included the Basin Characterization Model (BCM) developed by Flint and Flint (2008) and isotopic analyses developed by Blasch and Bryson (2007). Methods for estimating incidental and artificial recharge because of development of the surface and groundwater supplies are also discussed.

Natural Recharge

Natural recharge to the groundwater system is primarily through mountain block recharge and focused mountain front recharge in stream channels. The spatial and temporal distribution of recharge through direct infiltration was estimated by using a BCM that estimates monthly runoff, evapotranspiration rates, and direct infiltration for about 300-m-square grid cells across the western United States (Flint and Flint, 2008). The BCM considers multiple parameters that influence recharge such as rainfall, snowfall, solar radiation, wind speed, vegetation, soils, aspect, slope, temperature, humidity, and rock type. Many of these parameters were measured directly at specific locations and indirectly computed for other areas. These parameters were used to partition precipitation at the soil surface into runoff, soil storage, evapotranspiration rates, and infiltration (eventual recharge). The model is calibrated through adjustment of parameters, mainly adjustment of hydraulic conductivity values for each soil and rock type, to match simulated rates of runoff and base flow with observed and estimated rates. This method for estimating recharge is based on numerous characteristics of the watershed and has an advantage over other methods in that the estimated values can be calibrated to runoff and base flow. Streamflow gages across many watersheds across the western United States were used as control for the calibration, including several gages in the NARGFM area. The parameters used for control included multiple assemblages of soil, vegetation, slope, altitude, and rock type that are common in the study area. Additionally, stable isotope results for precipitation samples at a range of altitudes (Blasch and Bryson, 2007) were used to further constrain the altitude and spatial distribution of BCM estimated recharge for the Big Chino, Little Chino, and Verde Valley sub-basins.

The resulting BCM estimated recharge distribution (fig. 10*A*) has a strong relation to altitude because of greater precipitation rates, lower temperatures, and lower rates of transpiration by vegetation at high altitude than at lower altitude. BCM estimated recharge rates tended to underestimate overall recharge rates in alluvial basins on the basis of base flow estimates in four alluvial basins. BCM estimates of recharge rates in other areas, however, compared well with previously estimated rates. The underestimate in alluvial basins likely results from the inability of the BCM to estimate recharge that happens through streamflow infiltration. Greater total recharge is through infiltration of stream flow in ephemeral channels in the alluvial basins in comparison to other areas. The BCM estimates were modified by adding recharge to four alluvial basins, including Big Chino, Little Chino, upper Agua Fria, and Verde Valley, to approximate the estimated predevelopment water budget. The additional recharge was distributed as a uniform rate throughout the surficial permeable sediments including the Supai Formation, Coconino Sandstone, and Tertiary and Quaternary alluvial deposits except for the extents of fine-grained facies of alluvial deposits in the Big Chino and Verde Valley (Verde Formation) sub-basins. These modified BCM estimates of recharge distributions were not modified during the model calibration process as a way to better simulate observed water levels and discharge to streams and springs. The modified BCM recharge rates were similar to estimated recharge rates derived from groundwater-budget analyses for several basins and considered to adequately represent true rates across the NARGFM. Consideration of the approximate recharge distribution as a known variable allowed for adjustment of only simulated hydraulic parameters to better simulate observed water levels and discharge to streams and springs.

The monthly BCM recharge estimates were used to estimate average-annual and decadal variations in recharge distributions. Average-annual BCM estimated recharge rates from 1971 to 2005 were used to estimate average-annual recharge rates for the predevelopment period. Scaled decadal variations in recharge were estimated by using an extended BCM dataset for the period 1940–2005 (fig. 10*B*). Scaled variations in decadal recharge are similar to variations in precipitation and ranged from 1.0 for predevelopment, to 0.60 for the 1950s, 1.6 for the 1970s, and 0.7 for 2000–2005 (fig. 10*B*). Decadal recharge rates for the NARGFM were calculated by multiplying the average-annual and scaled decadal BCM estimates.

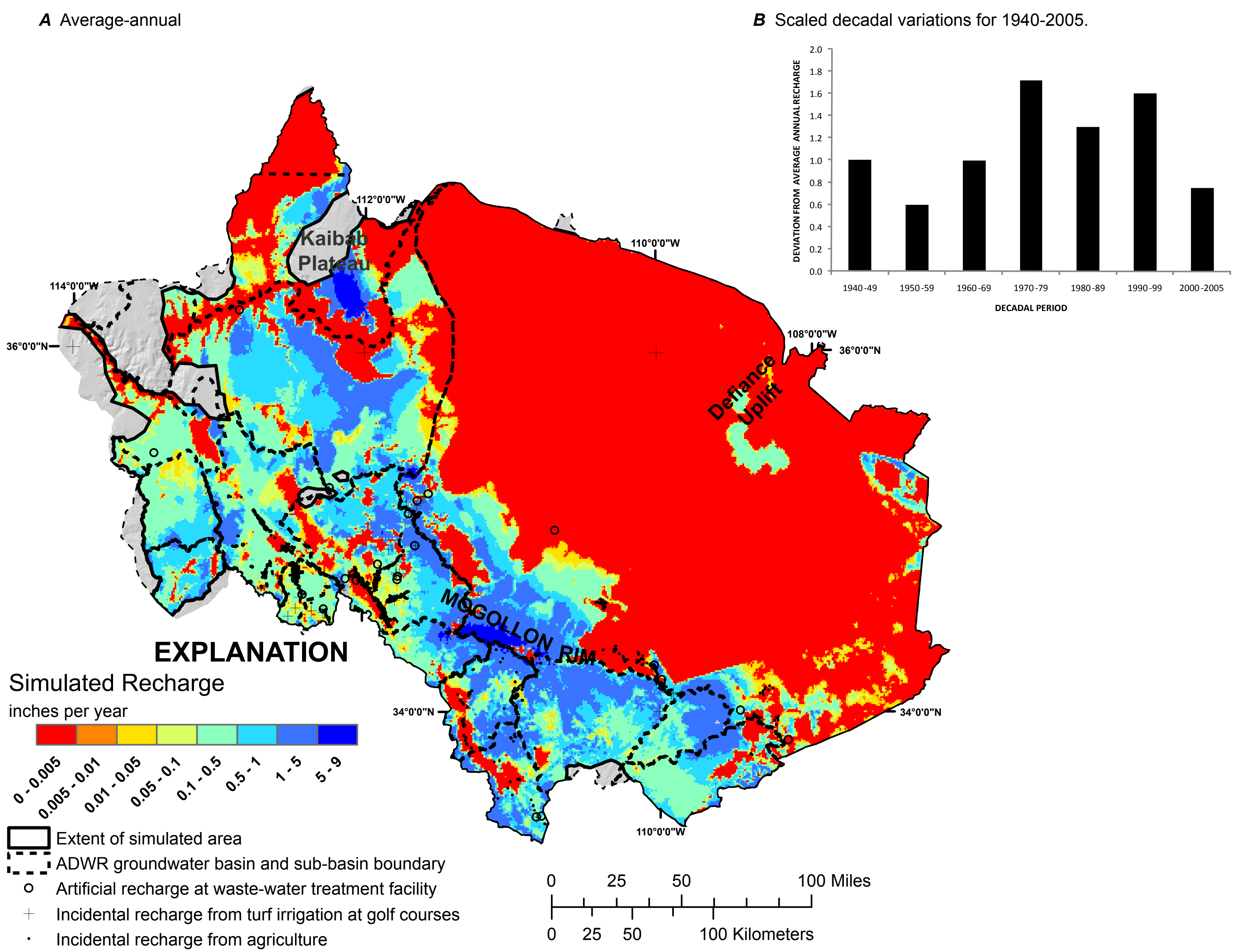


Figure 10. Sites of incidental and artificial recharge and simulated rates of average-annual natural recharge in the Northern Arizona Regional Groundwater-Flow Model by using a Basin Characterization Model for (*A*) average-annual and (*B*) scaled decadal variations for 1940–2005.

Incidental and Artificial Recharge

After development of surface-water and groundwater supplies, additional sources of recharge developed either incidentally as a result of water-use practices or intentionally as artificial recharge. Incidental recharge results from deep percolation of water that is in excess of evaporative demands at septic systems, in stream channels that receive waste-water effluent, and for irrigation of agriculture, golf courses, and greenways. Recharge rates for most of these activities in the study area have not been estimated. Previous protocols for estimating these recharge rates in other areas that have been used by ADWR, Bureau of Reclamation, and Yavapai County also were used as a basis for estimating incidental recharge for this study. Recharge rates from agricultural irrigation and industrial water use were based on coefficients calculated as a percentage of withdrawals that are estimated to be in excess of evaporative demands. These coefficients were 0.5 (50 percent of water applied) for agriculture irrigation, 0.2 for industrial use, and 0.1 for golf courses. These coefficients are not based on actual measurements and likely are not uniform in the study area. Incidental agricultural recharge was assumed to be coincidental with the groundwater withdrawal. Some crops are subirrigated, that is, the crop is not intentionally irrigated by using surface or groundwater supplies, and rely on precipitation and shallow groundwater as water sources. Therefore, no incidental recharge is assumed to occur in areas of subirrigated crops.

Incidental Recharge from Agricultural Irrigation

Incidental recharge from agricultural irrigation (fig. 10*A*) was estimated for agricultural areas in the Big Chino, Little Chino, and Verde Valley sub-basins and for the Little Colorado River Plateau and sub-basins along the Salt River on the basis of estimated irrigation requirements and sources of irrigation water, that is, surface or groundwater supplies. Estimates for the Little Chino sub-basin were made for both surface- water and groundwater irrigation on the basis of a groundwater-flow model developed for the PrAMA (Timmons and Springer, 2006). Estimates of incidental agricultural recharge in other basins are directly related to estimates of agricultural groundwater withdrawals, commonly estimated on the basis of crop requirements, and crop use for surface water irrigated agriculture. Discussions of estimates of agricultural groundwater withdrawals for each basin are in the “Groundwater Withdrawal” section. A general estimate of 50 percent of agricultural groundwater withdrawals was used to estimate incidental recharge from groundwater irrigated agriculture on the basis of estimates for irrigation practices in the PrAMA (Corkhill and Mason, 1995). The incidental agricultural recharge from groundwater irrigated crops generally was assumed to be at the centroid of the agricultural field.

Agricultural incidental recharge in the Big Chino sub-basin is concentrated at about 8,000 acres of agricultural land in five agricultural regions, including the Paulden area, the upper Big Chino alluvial basin area, Williamson Valley, Walnut and Pine Creeks, and near Big Chino Wash immediately upstream from the Big Chino alluvial basin. About 40 percent of the irrigated acreage lies in the upper part of the Big Chino alluvial basin. Estimated incidental agricultural recharge in the Big Chino sub-basin varied from about 2,100 ac-ft/yr before 1950 to about 4,000 ac-ft/yr during the 1960s and about 2,100 ac-ft/yr after 1980.

Agricultural incidental recharge from surface-water supplies in the Little Chino sub-basin is along the Chino Valley irrigation canal and ditches that deliver water from Watson Lake near Prescott to agricultural fields south of Del Rio Springs. Estimated incidental agricultural recharge from surface-water supplies in the Little Chino sub-basin was about 2,000 ac-ft/yr before 1940 and decreased to about 1,600-1,400 ac-ft/yr after 1940 as groundwater withdrawals augmented the surface-water supplies.

GIS datasets that describe agricultural irrigation in the Verde Valley sub-basin, including field extents, crop type, and crop water use (Arizona Department of Water Resources, 2000), were used to estimate crop water use and incidental recharge from agricultural irrigation by using surface-water supplies. Incidental agricultural recharge derived from groundwater supplies was considered negligible, because agricultural water supplies in the Verde Valley sub-basin are primarily derived from surface-water supplies (Arizona Department of Water Resources, 2000). Based on scant historical data, the agricultural fields were assumed to operate continuously since the ditches were constructed. However, about 3 percent of the total irrigated acreage was assumed to be fallow during any year, according to Arizona Department of Water Resources, (2000). Crop type and water use rates included alfalfa (3.57 ac-ft/yr), corn (1.86 ac-ft/yr), orchard (3.36 ac-ft/yr), pasture (3.11 ac-ft/yr), turf (3.35 ac-ft/yr), vegetables (1.13 ac-ft/yr), and landscaping (3.35 ac-ft/yr). A total of about 5,800 agricultural acres are in the Verde Valley sub-basin with a crop use of about 17,000 ac-ft/yr. Most of the fields lie along the Verde River and major tributaries including Oak Creek, Wet Beaver Creek, and the lower part of West Clear Creek. Incidental recharge rates were assumed to be equal to the consumptive crop use and were assumed to be at the centroid of mapped fields.

Withdrawals for agriculture in the Little Colorado River Plateau basin and sub-basins in the Salt River drainage were estimated based on groundwater irrigation data developed by USGS (Tadayon, 2005) and recent aerial and satellite photography for effluent irrigated areas. The datasets include only the locations of agricultural fields, but no information for crop type, applied irrigation, or period of irrigation. Consequently, the fields were assigned a weighted irrigation factor of 3.15 ac-ft/yr that was developed from agricultural irrigation data from the Verde Valley (Arizona Department of Water Resources, 2000). Agricultural fields overlying the Moenkopi Formation confining unit were considered disconnected from the regional aquifer and were removed from the withdrawal and recharge accounting. Centroid points for the remaining fields were used for recharge accounting (fig. 10*A*). Effluent irrigated areas are near Spingerville, Eager, and Nutrioso and near a paper mill west of the Snowflake-Talylor area. Irrigation by using effluent from the paper mill began in 1992. During 1961–1992, paper mill effluent was discharged into the nearby Dry Lake. Recently, 3,100 acres adjacent to Dry Lake have been irrigated (Arizona Department of Water Resources, 2010.) Data on irrigation of agricultural fields near Springerville, Eager, and Nutrioso were obtained from personal communications with city administrators. Agricultural fields near perennial streams in the Salt River sub-basins were assumed to be irrigated by surface-water diversions.

Incidental Recharge from Golf Course Irrigation

Incidental recharge from irrigation of golf courses (fig. 10*A*) was estimated by using data from USGS annual records, ADWR annual reports, Arizona Corporation Commission (ACC) annual reports, and direct contact with golf course managers. (The “Groundwater Withdrawal” section of this report contains a more detailed discussion of sources of groundwater withdrawal data.) Scant data exist for irrigation at individual golf courses; therefore, an average of golf course usage was computed for golf courses that lacked irrigation data. The beginning of irrigation was assumed to be 1990 for a few golf courses that lacked construction dates. Incidental recharge rates at golf courses are assumed to be the non-consumptive use part of the applied irrigation water, or 10 percent of the estimated applied water. Two reports have estimated zero recharge at golf courses in semi-arid environments (Foster and others, 1999; New Mexico State Engineers Office, verbal communication, 2008; http://www.ose.state.nm.us/publications/waterlines/wl-winter-2001/golf.html, accessed May 10, 2008). However, based on consultation with golf course operators and water managers in the study area, flushing of salts from soils is a primary reason for applying more water than needed for turf consumptive use requirements.

Incidental Recharge from Septic Systems

A thorough estimate of recharge from septic systems was beyond the scope of this study; therefore, estimates were based on assumptions. Incidental septic recharge is based on several considerations for households that are not connected to sewer systems, including the estimated number of individuals per household, estimated per capita water use for each household, and all water entering a home is assumed to discharge through the septic system. As a simplification, 35 percent of the water that was withdrawn for an exempt well was estimated to recharge the aquifer at each exempt well. This value was based on the knowledge that some leach fields are near the surface and above the depths of plant roots and some water is applied for irrigation.

Artificial Recharge

Artificial recharge in the study area is through the deep percolation of water beneath infiltration ponds and lagoons at municipal and community wastewater treatment facilities. Estimates of artificial recharge rates used in the model of the Prescott Active Management Area (Timmons and Springer, 2006) were used for that area. Historical records for effluent discharge were sparse in other areas; therefore, artificial recharge for the areas outside the PrAMA boundaries were obtained from previous reports, the Arizona Water Atlas (Arizona Department of Water Resources, 2010), and from communications with water treatment plant operators. With the exception of a few facilities, recharge studies do not exist and recharge rates must be inferred. Discharge of effluent from waste-water treatment plants in the study area has many destinations and uses that result in different recharge rates including irrigation (golf courses, agriculture, landscaping), evaporation ponds, wildlife lagoons, recharge ponds, and discharge into watercourses. Many facilities discharge water to several uses, but the distribution is poorly documented. Oral communications with many operators have indicated a high demand for effluent, and most irrigators must supplement their allotment of effluent with other sources. Little of the applied effluent water is likely in excess of use demands and available for recharge where effluent deliveries were insufficient for demands. Additionally, many evaporation ponds and lagoons were designed to minimize recharge rates as mandated by permit authorities. Consequently, only facilities that potentially recharge more than 50 ac-ft/yr in 2000 were considered.

Artificial recharge rates from effluent were estimated based on the effluent destination. Recharge factors for effluent applied as irrigation for agriculture or greenways was considered the same as incidental recharge, or 50 percent of applied effluent. Recharge of water in natural watercourses or recharge ponds was estimated to be 80 percent of discharge. Effluent recharge rates at non-groundwater irrigated golf courses are assumed to be equivalent to the recharge rates at golf courses that use groundwater, or 10 percent of the estimated applied water. Only 5 percent of effluent distributed to lagoons or evaporation ponds was estimated to recharge the aquifer.

Outflows

Discharge from the aquifers is primarily as spring flow and base flow to the major river systems in the study area, groundwater pumpage, and to a lesser extent evapotranspiration.

Spring Discharge

Numerous springs are present in the study area associated with mountain ranges, river systems, and exposed canyon walls. Spring discharge is observed where changes in permeability between geologic layers exits, high permeability faults and fractures are present, and/or where water-level altitudes are above ground-surface altitudes (Cooley, 1976; Wirt and Hjalmarson, 2000; Hart and others, 2002; Monroe and others, 2005; Blasch and others, 2006; Bills and others, 2007). Total average-annual and average-seasonal discharge is dependent on the relation of the spring to recharge zones, hydrogeologic units, weather patterns, and anthropogenic development (Hart and others, 2002; Flora, 2004; Monroe and others, 2005; Blasch and others, 2006; Bills and others, 2007).

Estimates of spring discharge were derived from several souces. The U.S. Geological Survey National Water Information System-Groundwater Site Inventory System (NWIS-GWSI) database contains records for springs for the study area. Additional springs are identified on USGS topographic maps for which no hydrologic or water-quality information was available in the database. Additional discharge and water-quality measurements have been collected for springs in the study area (Feth, 1954; Feth and Hem, 1963; Twenter and Metzger, 1963; Owen-Joyce and Bell, 1983; Wirt and Hjalmarson, 2000; Ingraham and others, 2001; Hart and others, 2002; Flora, 2004; Monroe and others, 2005; Blasch and others, 2006; Bills and others, 2007; Rice, 2007). Additional information resides in U.S. Department of Agriculture Forest Service unpublished datasets (Springs database for Sycamore Creek, Oak Creek, Wet Beaver Creek, West Clear Creek, Fossil Creek, and East Verde River watershed, 2000; Spring sampling and flow measurements along the Mogollon Rim between Canyon Creek and Sycamore Creek, 2001, prepared by Hydro Geo Chem, Inc., Tucson, Ariz.). Given the size of the study area and remoteness of many springs, only a few springs have repeat measurements, and fewer still have long-term discharge records. For this study, only springs with discharge more than 50 ac-ft/yr were incorporated into the numerical groundwater-flow model because springs discharging less than 50 ac-ft/yr account for less than 1 percent of the total discharge from all springs and a much smaller part of overall groundwater budget.

Base Flow

Most groundwater flowing through the regional aquifers of the study area discharges to streams as stream base flow. Estimates of stream base flow are, therefore, needed for comparison with simulated rates of discharge to streams. Streamflow includes two components—surface runoff and base flow. The surface runoff component must be removed from streamflow records at gaging stations to isolate the base flow component of the record. The base flow component of streamflow records was estimated at many gage sites along the major drainage systems in the study area. Streamflow records were available for estimation of base flow at gage sites that include most of the largest and major sources of groundwater discharge, including several streamflow gages on the the Verde River and tributaries, downstream from Blue Springs on the Little Colorado River, downstream from Havasupai Springs, on the lower reaches of the Salt River, and at the lower reaches of the streams that drain the western basins groundwater systems. Base flow records are difficult to remove from the streamflow records on the Colorado River because the streamflow is dominated by reservoir releases. Base flow estimates were derived from previous investigators and reported discharge estimates at a few other major streams, including Sycamore and Fossil Creeks in the Verde River drainage system, Clear and Chevelon Creek in the Little Colorado River drainage system, and many springs that discharge to the Colorado River.

Annual base flow for streamflow records was estimated as the minimum of average flow for any 7-day period during winter (January-February). Winter low-flow estimates were considered representative of annual rates of groundwater discharge because some streamflow-gage records include seasonal variations related to seasonal abstractions through evapotranspiration and agricultural diversions. Steady state base flow was estimated as the average of annual base-flow estimates for the period of record at streamflow gages that had long-term records. Long-term records include at least one full wet-dry decadal variation in precipitation and runoff. Many streamflow-gage records include dry periods before and after the wet period during the late 1970 to mid-1990s.

Evapotranspiration

Evapotranspiration of groundwater is through phreatophytes and subirrigated agriculture where depths to water are very shallow near stream channels. Minor rates of groundwater discharge happens through evapotranspiration by phreatophytes. The large depths to groundwater in the study area limit the accessibility of groundwater by phreatophytes to narrow areas near perennial streams and springs. Rare areas of subirrigated agriculture that can access groundwater are locally substantial areas of groundwater discharge, but are of minor importance on a regional scale.

Natural Vegetation

Estimated rates of groundwater evapotranspiration by natural vegetation are derived from existing reports for the Verde River and Coconino Plateau areas. Many areas have no published estimates of evapotranspiration rates by phreatophytes, such as along the Little Colorado River and Salt River drainage systems. Streamflow records in these areas indicate little or no seasonal variations in base flow, which suggests that groundwater evapotranspiration rates are not substantial in comparison to groundwater discharge to the stream.

Requirements for simulating groundwater evapotranspiration rates in the numerical groundwater-flow model include spatial extent of phreatophytes, depths to water, type of phreatophyte, maximum rates of water use for each phreatophyte type, and maximum depths of groundwater withdrawal for each phreatophyte type. Maps or GIS coverages describing phreatophyte type are generally unavailable for the area. The primary vegetation type was, therefore, assumed to be deciduous trees that may include various species of cottonwood, willow, sycamore, and saltcedar. Commonly used maximum water use rates, 4.92×10^{-4} to 6.54×10^{-4} ft/d, and depths to water for deciduous trees, 16.4 ft (Leenhouts and others, 2005) also were assumed to apply to groundwater use by phreatophytes in the study area. Extents of phreatophytes were approximated for each area of potential groundwater use by phreatophytes. Depths to groundwater were allowed to be calculated by the groundwater-flow model based on the difference of land surface and calculated water-level altitude in unconfined aquifers. Simulated evapotranspiration only occurs where calculated depths to water are less than 16.4 ft in the extent of potential phreatophytes.

Subirrigated Agriculture

The only known area of subirrigated crops that may access shallow groundwater supplies is in the Williamson Valley and Big Chino Valley areas. Yavapai County surveyed 1,325 acres of subirrigated crops in these areas consisting entirely of pasture grasses (John Munderloh, Yavapai Water Coordinator, written commun., 2004). Depths to groundwater may be sufficiently shallow across a part of this subirrigated area to be accessible by the pasture crop.

Possible evapotranspiration rates from subirrigated crops was estimated by using two methods. One method results in an estimate of maximum evapotranspiration rates by using average monthly potential evapotranspiration estimates, crop water demands, and effective precipitation. The second method uses crop water demands and effective precipitation (Arizona Department of Water Resources, 2000). Effective precipitation is the amount of moisture retained by the soil after precipitation and is influenced by factors such as slope of the agricultural field, soil properties, rainfall intensity, and rainfall frequency (U.S. Department of Agriculture Soil Conservation Service, 1970). Average monthly potential evapotranspiration rates (ETo) were derived by using SOLPET V1.0 (Flint and others, 2004). Monthly potential evapotranspiration rates were multiplied by the monthly water demand for pasture grasses (Shuttleworth, 1993) to obtain monthly evapotranspiration rate. Effective monthly precipitation was subtracted from monthly evapotranspiration rates to obtain the maximum amount of water that can be removed from the aquifer through evapotranspiration, about 4,300 ac-ft/yr. Use rates of this magnitude assume the pasture crop is extensive and that groundwater is sufficiently shallow to be accessed by crop roots across the 1,325 acres. These conditions likely occur, however, across only a part of the pasture extent, and actual evapotranspiration of groundwater is likely much less than estimated by the method.

The crop water demands and effective precipitation method used monthly water demand estimates for pasture crops corrected for monthly effective precipitation estimates to estimate evapotranspiration rate of groundwater. Results indicate that about 2,600 ac-ft/yr is transpired from the subirrigated crops. This method also assumes the pasture crop is extensive and that groundwater is sufficiently shallow to be accessed by crop roots across the 1,325 acres. These conditions are likely to occur, however, across only a part of the pasture extent and actual evapotranspiration of groundwater is likely much less than estimated by the method.

Groundwater Withdrawals

Groundwater, including spring water, is the predominant source of water for domestic, industrial, and agricultural water uses in the study area. The recent primary uses of groundwater in the study area are for industrial (39 percent), municipal/domestic (44 percent), and agriculture (16 percent). Groundwater is supplied primarily by private and municipal water companies near population centers. Private wells are the primary source of groundwater in the rural areas. Average-annual groundwater withdrawal in the study area has increased from about 15,000 ac-ft/yr during 1940 to about 140,000 acre-ft per year after 2000 (fig. 6).

Groundwater withdrawal datasets include many users whose regulation authority ranges from the State level to tribal lands that have no State-imposed regulations. Data were compiled for the period of 1938 through 2005 from three primary data sources: USGS annual records, ADWR annual reports, and Arizona Corporation Commission (ACC) annual reports. Additional groundwater withdrawal data were obtained from reports on specific groundwater subbasins or regions, ADWR well drilling records, historical publications, and personal communications. The USGS collected groundwater withdrawal records from about 1975 to 1995. The USGS collected the data through voluntary correspondence with water users and published annual results in groundwater use and water-level reports (for example, Babcock, 1977). The early publications were superseded with decadal reports (for example, Tadayon, 2005). The ADWR annual records for active management areas and irrigation nonexpansion areas were obtained through the Arizona Water Commission beginning in 1978; however, electronic records are not available until about 1983. ACC records prior to 1996 are held in State archives. This study accessed ACC annual reports that date back to 1980 and hardcopy and electronic files through 2005. Records prior to 1980 are unavailable.

Groundwater withdrawals were assigned to 12 water-use categories: municipal-large, municipal-small, agriculture, industrial-turf-golf courses, industrial-turf-school, industrial-sand and gravel, industrial-power plant, industrial-dairy, industrial-mining, industrial-other, exempt, and other. Withdrawals in regulated areas, including PrAMA, Joseph City Irrigation Non-expansion Area, and the ACC Certificate of Convenience and Necessity boundaries, had been assigned water use categories through ADWR management plans.

Outside of the regulated areas, water users were not required to meter and report annual water withdrawals. However, many private and public entities, including Native American Tribes, have voluntarily shared data invaluable for use in the groundwater model. For various areas and data availability, assumptions had to be made. Some assumptions were made across the model area; while others are basin or user specific. Listed are some of the broad assumptions:

- Throughout the model area, records of water withdrawals were incomplete and it was necessary to: (1) extrapolate records back in time, (2) interpolate for missing records, or (3) extrapolate the records forward in time. In the first and third situations, the last reported value was used to extrapolate pumping. When withdrawal data were missing in a record, the last known value was repeated until the next known value in the record. ADWR well records were critical for determining extent of extrapolation for a particular well based on the drill date, ownership, and use. Additionally, county assessor records provided more information regarding ownership and lot splits.
- Occasionally, multiple uses served by one entity were not separated. If volumes could not be separated, the entire volume was assigned to the dominate use.
- PrAMA (Little Chino Valley and Upper Agua Fria sub-basins)—ADWR PrAMA modeling datasets were incorporated and water use codes were assigned based on ownership shown in annual reports. In the PrAMA pumping files, a single point in a cell could represent an agricultural user and the exempt wells in that cell.

Exempt wells, as determined by the State of Arizona, pump 35 gal/min or less and are considered domestic wells for this analysis. Information for withdrawals by exempt wells is meager, with the exception of the Little Chino sub-basin and the Upper Agua Fria sub-basin in the PrAMA. Exempt well withdrawal rates applied to the PrAMA model (Timmons and Springer, 2006) also were applied to the Little Chino sub-basin and the Upper Agua Fria sub-basin in the NARGFM model. However, the PrAMA model does not include the spatial extent of exempt wells that lie outside of the alluvial basins. For the NARGFM, exempt well withdrawals outside of the alluvial basin were included and two rules were implemented. The first rule is that once an exempt well is inserted in the flow model that well is used to the end of the simulation. The second rule is that exempt wells in the basin-fill aquifers are estimated to withdrawal 0.5 ac-ft/yr and exempt wells in all other units are estimated to withdraw 0.33 ac-ft/year.

Agricultural irrigation has been the predominant groundwater use in alluvial basin areas suitable for agriculture, including Big Chino sub-basin, Little Chino sub-basin, and Upper Agua Fria sub-basin. Withdrawals for agriculture in the Big Chino sub-basin were estimated by using current pumping records and estimated irrigation data provided by Yavapai County (John Munderloh, Yavapai Water Coordinator, written commun., 2004). The pumping data provided by Yavapai County contains estimated irrigation and crop information for the years particular fields were active. Crop use was recorded by using historical aerial photography. The photographic record was discontinuous through time, so periods missing records were interpolated based on known data. Withdrawals were estimated to be twice the consumptive use value because 50 percent of withdraw are estimated to return to the aquifer as incidental recharge. The pumpage dataset, which contains actual well locations and pumping quantities, was used as the preferred dataset.

Pumping withdrawals for agriculture in the Verde Valley sub-basin were considered negligible based on agricultural data from ADWR (2000). The almost exclusive reliance on surface water diversions for irrigation and lack of reporting of groundwater withdrawals prohibits any reasonably accurate record of pumping.

Withdrawals for agriculture in the Little Colorado River Plateau basin and sub-basins along the Salt and Verde Rivers were estimated on the basis of irrigation data provided by ADWR that includes only locations of agricultural fields and no information for crop type, irrigation applied, or period of irrigation. Consequently, the fields were assigned a weighted irrigation factor of 3.15 ac-ft/yr that was derived for the Verde Valley (ADWR, 2000). Assuming that fields that do not lie adjacent to perennial streams use groundwater, about 2,200 ac-ft/yr of water was estimated withdrawn for agricultural irrigation in the Salt River sub-basin.

About 57 percent of groundwater use in the model area during 2000 to 2005 was extracted from the Little Colorado River Plateau basin. Groundwater withdrawal for industrial use is the primary use of groundwater in the Little Colorado River Plateau basin. Groundwater withdrawals for industrial uses accounted for about 30 percent of groundwater use in the basin during the 1970s but increased to about 64 percent of groundwater for the period 2000 to 2005. Municipal/domestic/exempt use accounted for about 28 percent of groundwater use during the 2000 to 2005 period. Agricultural fields in the Little Colorado River Plateau basin that do not use the Coconino aquifer as a source were removed from the withdrawal and recharge accounting. Agricultural fields near Springerville, Eager, and Nutrioso were obtained from Google Earth imagery and personal communications with city administrators.

Withdrawals in most remaining sub-basin areas, western basins, and Coconino Plateau are dominated by municipal/domestic/exempt groundwater use, with the exception of the Burro Creek sub-basin that is dominated by industrial (mining) use. Municipal/domestic/exempt groundwater use accounts for more than 90 percent of water use in the sub-basins. Industrial water use (including golf course irrigation) was the predominant groundwater use (about 25 percent) during the past 5 years in the Verde Valley. About 33,900 acre-ft of annual surface water diversions for primarily agricultural use in the Verde Valley sub-basin between 1990 and 2003 is not included in this groundwater withdrawal analysis.

Groundwater-Flow Model

A numerical groundwater-flow model was constructed to simulate the conceptualized pre-development and developed groundwater-flow systems for the period 1910 through 2005. The three-dimensional finite-difference modular groundwater-flow model code MODFLOW-2005 (Harbaugh, 2005) was selected for simulation of the regional groundwater-flow system. MODFLOW-2005 has been rigorously evaluated, is supported with many optional packages, freely available, and widely used by governmental and nongovernmental agencies in the State of Arizona. The numerical code is specific for simulation of saturated flow in porous media. The aquifer systems in the northern Arizona model domain are primarily porous sedimentary rocks; however, fractures and solution channels likely provide locally effective flow paths, especially in areas of crystalline rock and limestone. Approximation of groundwater flow in these areas as flow through porous media is likely adequate for simulation of regional groundwater flow. Detailed simulation of groundwater flow may require appropriate numerical code for simulation of fracture and solution channel flow.

Construction of the numerical model required several tasks. The first task was to define the model framework with the assignment of boundaries, including streams, springs, and lateral and vertical extents of aquifers. The model domain was then discretized in space into a 3-dimensional grid and discretized in time into several stress periods representing major periods of changes to the system. Rates of natural, artificial, and incidental recharge were estimated and distributed across the model domain and through time. Distributions and rates of evapotranspiration were assigned. Initial estimates of aquifer properties, including horizontal and vertical hydraulic conductivity, specific storage, and specific yield, were assigned to the model domain. Hydrologic control data for the simulation of predevelopment and transient conditions were selected, including water levels in wells and base flow at streams and springs. Groundwater withdrawal datasets were constructed. Predevelopment conditions were then simulated and the initially estimated aquifer hydraulic properties were adjusted in an acceptable range to match hydrologic control data. Simulated predevelopment conditions formed the initial conditions for simulation of changes to the groundwater-flow system during transient conditions. Aquifer storage properties were adjusted to simulate observed changes in the groundwater system during the transient simulation period. Further adjustments also were made to hydraulic conductivity distributions to help match simulated with observed transient conditions.

The model simulates several discrete groundwater-flow systems that are hydraulically connected. The focus of the model, however, is on the systems in the Verde River Basin and the interaction of those systems with adjacent systems. Therefore the simulation of each major groundwater-flow system is discussed separately and in varying detail commensurate with the purpose of the model and availability of hydrologic control data in each area. More detailed discussions of hydraulic property distributions and water budgets are provided for the Big Chino, Little Chino, and Verde Valley groundwater systems than for other systems. Simulation of adjacent systems that include the Coconino aquifer and Redwall-Muav aquifer south of the Colorado and Little Colorado Rivers are discussed in lesser detail. Other systems such as areas north of the Colorado and Little Colorado Rivers, Salt River Basin, and western basins are discussed at a generalized level. Greater accuracy in the simulation of observed water levels and estimated water budgets was generally required for the systems in the Verde River Basin. Less accuracy was accepted for other systems because of fewer control data such as water levels in wells, and greater uncertainty in water budgets.

Model Framework

Boundaries

The spatial extent of the model was determined so that major groundwater-flow divides that define the groundwater systems of the Verde River and adjacent basins could be simulated rather than set at predetermined locations. This determination required that model boundaries include the extent of the aquifers that are adjacent to the Verde River Basin, where possible, including the aquifers in the Colorado and Little Colorado River Basins, Salt River Basin, and several small basins adjacent to the west and southwest boundaries of the Verde River Basin (fig. 11). The boundaries for the numerical model were, therefore, set at watershed boundaries that are coincident with low-permeability crystalline rocks along the southern boundary of the Verde River Basin and adjacent basins along the southern boundary of the model, including Truxton Wash, Trout Creek, Burro Creek, Tonto Creek, and part of the Upper Agua Fria sub-basin. Other boundaries are coincident with assumed groundwater divides at watershed boundaries of the Colorado, Little Colorado, and Salt Rivers that include likely sedimentary rock aquifers. An exception to the watershed boundaries is along the northeastern and eastern boundaries that are identical to the boundary of the Coconino aquifer model of Leake and others (2005), which almost follows the watershed boundary of the Little Colorado River. The Colorado River watershed north of the Grand Canyon was included in the model to allow for the simulation of groundwater flow beneath the river. The entire boundary perimeter is represented as no flow with the exception of groundwater outflow at discrete locations along streams that are represented in the model by using the MODFLOW Stream (STR) (Prudic, 1989) or Drain (DRN) Packages. Groundwater outflow occurs at the intersection of the upper model boundary with the multiple springs and streams, including the Verde River, Little Colorado River, Colorado River, Agua Fria River, Salt River, Burro Creek, Trout Creek, Truxton Wash, and several streams in the northeast extent of the

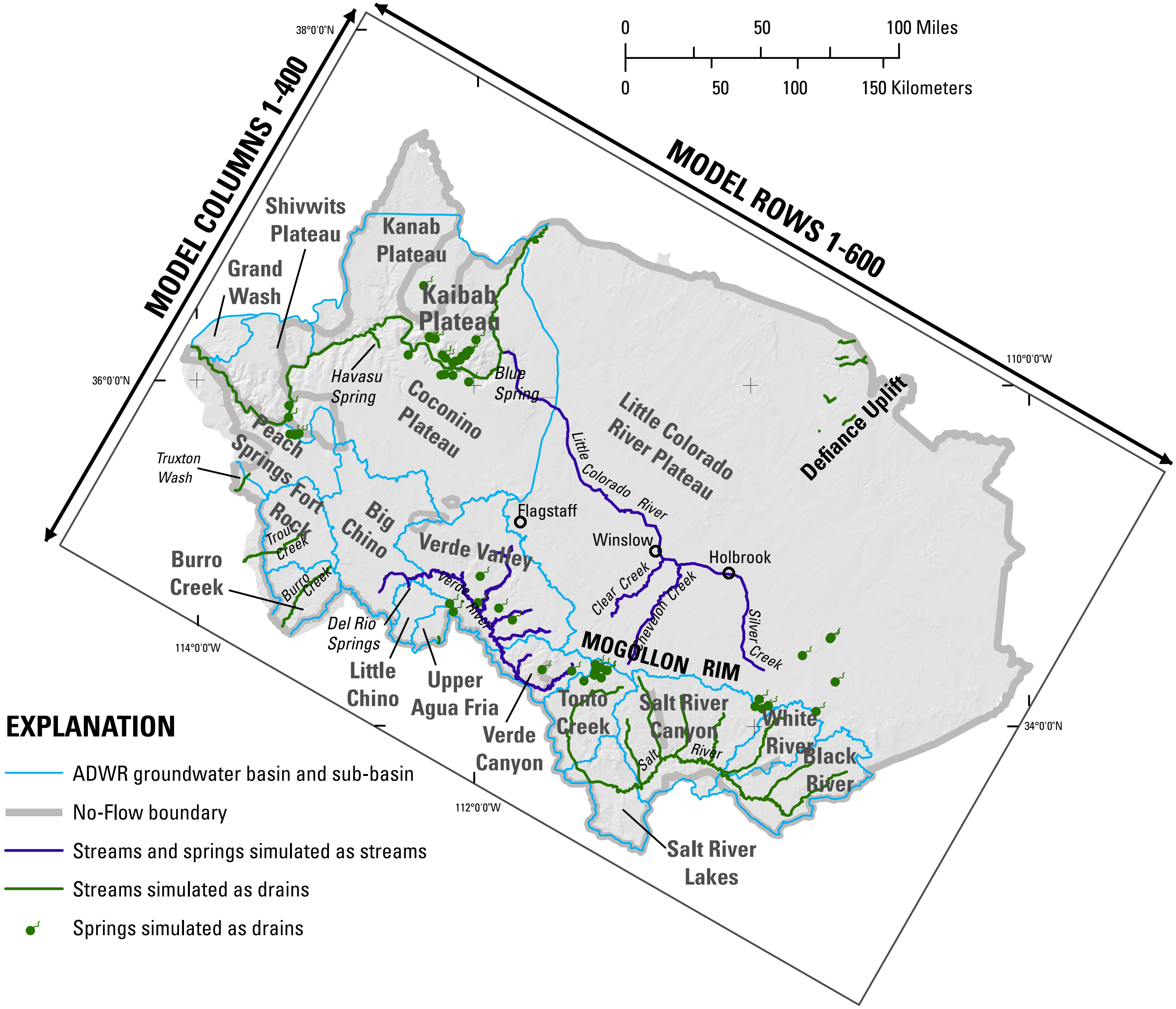


Figure 11. Boundary conditions for the Northern Arizona Regional Groundwater-Flow Model.

model. No groundwater inflow was assumed to be anywhere along the model boundary. Surface-water inflow occurs along the Colorado River at Page, Arizona, but was not simulated.

Spatial Discretization

A model grid of 600 rows, 400 columns, and three layers is used to represent the northern Arizona regional groundwater-flow system. Cell sizes are 0.62 by 0.62 mi (1 km by 1 km) (fig. 11). The model grid is rotated counterclockwise 60 degrees west of north. Three layers are used to represent the primary aquifers (figs. 12 and 13). Layer 3 is the lowest of the layers, extends across the entire model domain, and represents the Redwall-Muav aquifer and crystalline rocks that are exposed at the land surface in the southern and eastern parts of the model domain where the Redwall-Muav aquifer is absent (figs. 13*C* and 13*D*). Layer 2 extends only partially over the model domain and represents the Supai Formation on the Colorado Plateau, sand and gravel in the Verde and Big Chino Valleys, and the lower volcanic unit in the Little Chino Valley and Upper Agua Fria sub-basin (fig. 13*B*). Layer 1 is the uppermost and least extensive model layer and represents the Coconino aquifer on the Colorado Plateau, the thick silt and clay and adjacent interbedded alluvial deposits in the Big Chino Valley, the fine-grained part of the Verde Formation in the Verde Valley, and the upper alluvial layer in the Little Chino Valley and Upper Agua Fria sub-basin (fig. 13*A*).

Temporal Discretization

Groundwater flow is simulated for steady-state conditions that were assumed to exist in 1910 and transient conditions during 1910 through 2005. The simulation period is divided into nine multi-year stress periods. No seasonal or annual variations were simulated. The groundwater flow system in 1910 was dominated by natural conditions across most of the study area except in the Little Chino and Verde Valley sub-basins where the natural groundwater-flow system was altered by surface-water diversions for agricultural use. Natural predevelopment conditions prior to surface water diversions were not simulated because data to define that system are sparse. Groundwater conditions during 1910–1938 were the same as conditions in 1910 across the model extent, with the exception of some groundwater development in the Little Chino sub-basin after the mid-1920’s. The initial period of substantial groundwater development—primarily in the Little Chino sub-basin—is only two years, 1938 and 1939; thereafter, each stress period except for the final stress period is decadal in length. The final stress period is 6 years in length including 2000 through 2005. Each stress period was simulated using 5 time steps with each subsequent time step increasing in length by a factor of 1.2 over the previous time-step length.

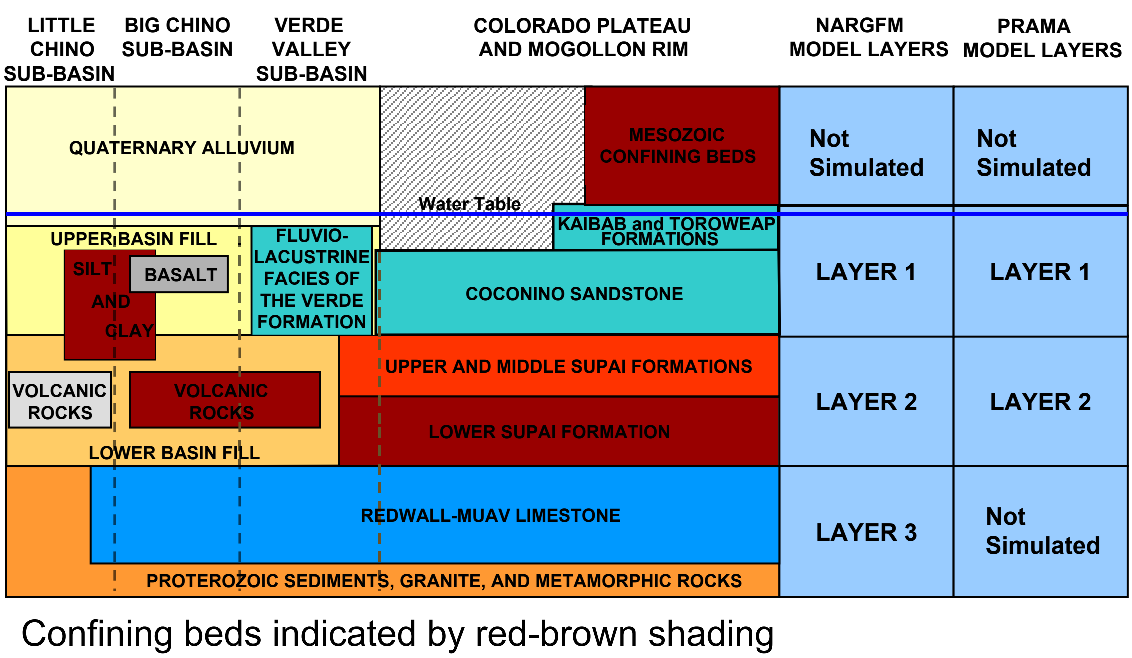


Figure 12. Conceptualized relations among major hydrogeologic units and Northern Arizona Regional Groundwater-Flow Model layers.

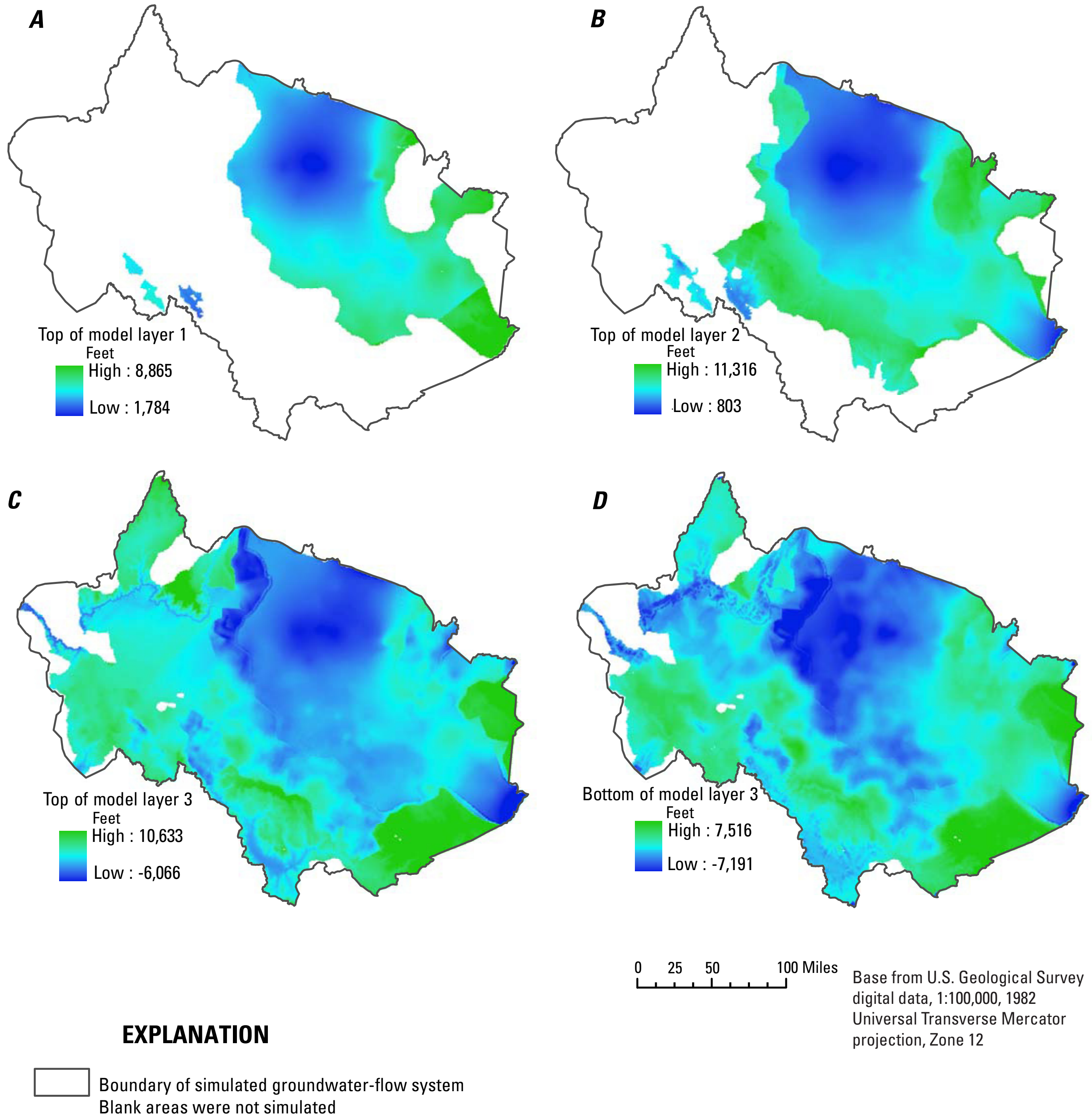


Figure 13. Altitude and upper and lower extents of Northern Arizona Regional Groundwater-Flow Model layers. *A*, Top of model layer 1, *B*, Top of model layer 2, *C*, Top of model layer 3, and *D*, Bottom of model layer 3.

Inflows and Outflows

Natural Recharge

Natural recharge, based on the modified BCM estimates, was applied to the uppermost model cell in the spatial extent of the model domain by using the MODFLOW Recharge Package. Modifications to the original BCM estimates of direct infiltration included additions to recharge resulting from ephemeral channel recharge in the Verde Valley basins.

Artificial and Incidental Recharge

Artificial recharge of sewage effluent and incidental recharge from excess applied irrigation water and golf courses was assigned to the centroid of the area of water application. This simplification was considered adequate for the 0.62 mi x 0.62 mi grid-cell sizes and the small size of many of the areas. The well package was used to apply artificial and incidental recharge at each cell.

Evapotranspiration

Evapotranspiration by phreatophytes was simulated along perennial stream reaches along the Agua Fria River and in the Verde River and Little Colorado River drainage systems by using the Evapotranspiration Package. Extents and rates of water use by pheatophyte type is poorly defined in the study area; however, rates of evapotranspiration from the groundwater system are a minor water-budget component at basin and regional scales. As a result, simulation of evapotranspiration was necessarily generalized and restricted to floodplain areas near the perennial reaches along the Agua Fria River and the main stem and tributaries of the Verde River, including Williamson Valley Wash and Little Chino Wash, and the Little Colorado River and tributary streams Chevelon Creek and Clear Creek. The Evapotranspiration Package calculates rates of evapotranspiration on the basis of a linear depth and rate relation. Rates in a model cell are a maximum where the water table is at or above the evapotranspiration surface and decrease linearly with depth to a rate of zero at the maximum depth below the evapotranspiration surface. Evapotranspiration altitudes, maximum evapotranspiration depths, and maximum rates of evapotranspiration were assigned to each potentially active evapotranspiration region. The evapotranspiration surface was estimated as 3.28 ft (1 m) below the minimum altitude of the land surface in each model cell. Maximum depths of evapotranspiration were assumed to be about the maximum depth of cottonwood roots, 16.4 ft. Maximum rates of evapotranspiration were assumed to range from 4.92×10^{-4} ft/d in most of the simulated evapotranspiration areas to 6.56×10^{-4} ft/d in areas near the Verde River and lower parts of tributary streams.

Streams and Springs

Perennial and intermittent streams and springs were simulated by using the STR and DRN Packages. STR allows discharge of groundwater to simulated streams and springs in gaining reaches and downstream infiltration in losing reaches. DRN allows discharge of groundwater where the water table is above the specified drain altitude at the simulated stream and spring, but does not allow infiltration of water when the water table is below that altitude. STR was used to simulate stream-aquifer interactions along the perennial reaches of the Verde River drainage system, including the Verde River, Sycamore Creek, Oak Creek, Wet Beaver Creek, West Clear Creek, and East Verde River (fig. 11), and several major springs. The Verde River stream network also included intermittent stream reaches of two major tributaries to the Verde River, Williamson Valley Wash and Little Chino Wash downstream from Del Rio Springs. STR also was used to simulate stream-aquifer interactions along major perennial and ephemeral reaches of the Little Colorado River drainage system, including Silver Creek, Chevelon Creek, Clear Creek, and the Little Colorado River downstream from Silver Creek. Several major springs, such as Page Spring near Sedona and Blue Spring near Cameron, were included in the simulation by using STR along the Verde and Little Colorado Rivers. DRN was used to simulate groundwater discharge to other streams, including the Colorado River, Salt River, Burro Creek, Trout Creek, Truxton Wash, and the Agua Fria River. DRN was also used to simulate groundwater discharge to springs that discharge 50 ac-ft/yr or more from the simulated aquifers and are not a part of the stream network, including many springs near the Colorado River—several in the Grand Canyon area, Havasu Spring, several springs in the Peach Springs area—and many springs in the Verde Valley and Salt River Basin that are not part of the stream and drain networks.

The STR Package also was used to simulate average annual agricultural diversions for use near 10 ditches along the Verde River the Verde Valley sub-basin. A total consumptive use of 10,200 ac-ft/yr was diverted from the stream system and not allowed to reinfiltrate along downstream reaches. The same rate was applied to steady-state and transient conditions as varied stream diversions for agricultural use have not been documented. Stream diversions and agricultural use were, therefore, assumed to have not varied since before 1910.

Input to the STR Package requires stage, altitude of the bottom of the streambed, altitude of the top of the streambed, streambed hydraulic conductivity, width, slope, sinuosity, and Manning's roughness coefficient. STR routes the streamflow through a network of channels and uses Manning's equation to calculate stream stage assuming a hypothetical rectangular channel cross section. The application of STR in this model does not allow for simulation of surface runoff from individual or seasonal runoff events; therefore, no stream inflow from upland channels was simulated.

Simulated streams in the Verde River and Little Colorado River networks were divided into 1,620 segments corresponding with each model cell in 377 stream segments. Each segment was assigned top and bottom altitudes of the streambed at the upper and lower extent of the segment, streambed hydraulic conductivity, stream width, sinuosity, and Manning's roughness coefficient. Slope was calculated by STR for each reach in each segment. Altitudes of the top of the streambed were estimated by using the minimum 10-m DEM altitude in a grid cell traversed by a stream reach. The bottom of the streambed was set at 29.5 ft below the streambed altitude for most stream segments. Stage initially was set to 3.28 ft (1 m) above the streambed and allowed to vary with head in the adjacent model cell. Streambed conductance values varied from 0.00328 to 328 ft/d (0.001 to 100 m/d). Low values of 0.0328 ft/d or less were primarily assigned to reaches of the Little Colorado River and tributaries where the streams overlie the Moenkopi Formation. Low values of streambed conductance in the Verde River stream system, 0.0328 ft/d , were assigned to areas where the Verde River and tributaries overlie the lower part of the Supai Formation and where the Verde Formation is known from seepage runs to be poorly hydraulically connected to the aquifer. High values of hydraulic conductivity, 0.328 to 164 ft/d (0.1 to 50 m/d), were assigned where perennial reaches of the Little Colorado River and Verde River stream networks overlie the primary aquifers and where the Verde Formation is known from seepage runs to be hydraulically connected to the Verde River. Streambed hydraulic conductivity values of 328 ft/d were assigned only at Del Rio Springs. Ditch diversions from the Verde River system were assigned streambed hydraulic conductivity values of zero to prevent the ditch from gaining from or losing to the aquifer. Stream widths varied from narrow widths of 3.28 ft along ditches, 6.56 ft in the upper stream reaches of tributaries, to 147 ft along the lower reaches of the Verde River. Stream sinuosity was set at 1.0, which indicates a straight channel in a model cell. Manning's roughness coefficient was 0.22 for all segments, which is an average value for intermittent streams in Arizona.

Simulation of groundwater discharge to streams and springs by using the DRN Package requires altitude of the drain and estimated vertical hydraulic conductivity for cells that include the stream or spring. Vertical hydraulic conductance values ranged from 0.328 to 32.8 ft/d for drains. Altitudes were estimated as the lowest 10-m DEM altitude in the model cell containing the drain.

Groundwater Withdrawals

Groundwater withdrawals were simulated by using the MODFLOW Well Package. Input from the Well Package includes well location, withdrawal rate, and model layer from which withdrawals are taken. Knowledge of the vertical distribution of withdrawals in a multiple aquifer system is important for proper simulation of transient groundwater-flow response. Withdrawals were assigned to model layers based on supporting data from USGS and ADWR groundwater databases where the well construction data were available. Many wells, however, lacked depth information. Withdrawals at these wells were assigned to the uppermost layer or to the primary layer used by other nearby wells. Wells were assigned decadal withdrawal rates from 1940 to 2005. Decadal withdrawal rates were computed as the average of estimated annual rates. Further temporal refinement is not warranted for early decades because withdrawal data are insufficiently documented. Annual withdrawal data after about 1990 are much better documented than earlier data and may be sufficient for further temporal refinement.

Hydraulic and Storage Properties

The ability of aquifers and confining units to transmit and store groundwater through the model from areas of recharge to discharge is dependent on the spatial distribution of transmissivity (the product of hydraulic conductivity and saturated layer thickness), horizontal anisotropy, vertical anisotropy, specific storage, and specific yield in the model domain. Hydraulic conductivity has units of length per unit time. Hydraulic conductivity is assigned to the model cells as an array of values for flow in the direction along rows. Hydraulic conductivity along model columns is calculated as the product of hydraulic conductivity along rows and a horizontal anisotropy ratio that also is assigned to each model cell as an array of values. The model grid was rotated 60 degrees counterclockwise toward west to match the primary structural trends of the region that also are assumed to strongly influence anisotropy of groundwater flow. Estimates of hydraulic conductivity were distributed across the model layers based on aquifer and well pumping tests and simulated values used in previous groundwater models; initial values ranged from very low values for low-permeability crystalline rocks and confining units to high values for permeable sedimentary rocks of limestone, sandstone, and alluvium. Initially, the aquifer systems across the entire model domain were assumed to be isotropic and the horizontal anisotropy was assigned a unit value for the entire model domain. Vertical anisotropy values (Kh/Kv) initially ranged from 10 for aquifers to 500 for model layers that included numerous confining beds. Initial hydraulic and storage property distributions were adjusted during the process of constraining the model to improve representation of hydraulic head and discharge to streams and springs. Aquifer-storage properties of specific storage and specific yield define the ability of the aquifers and confining units to store and yield groundwater. Specific storage has units of inverse length and specific yield is dimensionless. Specific storage values determine the ability of the aquifer to release water from storage through deformation of saturated pore spaces and expansion of water with a decline in hydraulic head throughout the aquifer thickness. The lowest specific-storage values were initially applied to areas with poorly compressible rocks such as crystalline rock and limestone. The highest specific-storage values were initially applied to areas with compressible alluvial deposits in alluvial basins. Specific yield is the ratio of the volume of water which the porous medium, after being saturated, will yield by gravity to the volume of the porous medium (Lohman and others, 1972). Specific yield values that were initially applied to the model layers range from very low values, 0.01 or less, for hydrogeologic units that have little primary effective porosity, such as granite, Redwall-Muav limestone, fine-grained sediments of the lower part of the Supai Formation, and fine-grained facies of basin fill, to high values for hydrogeologic units that have high effective porosity, such as coarse-grained alluvial deposits.

The combination of storage and transmissive properties influence the rates of change in water levels and discharge to streams and springs that result from changes in aquifer recharge and withdrawals. Aquifer storage properties are poorly defined throughout most of the model domain because of sparse data. Values are scarce because the properties are derived from expensive and labor-intensive aquifer tests that are infrequently performed. Initial storage and transmissive property values assigned to model layers were derived from available aquifer tests and previous groundwater-flow models, including models of similar aquifers. Initial storage property distributions were adjusted during the process of model calibration to improve representation of changes in hydraulic head and discharge to streams and springs.

Several groundwater-flow model parameters were adjusted within hydrologically reasonable limits to calibrate the simulated groundwater-flow system against observations of water levels in wells and discharge to streams and springs. Parameters that were allowed to adjust during the calibration process included distributions of hydraulic conductivity, vertical anisotropy, specific storage, specific yield, streambed conductivity, and evapotranspiration rates. Recharge rates were not allowed to vary during the calibration process.

Overall goals in the model calibration process included representing regional patterns of water movement from areas of recharge to areas of discharge; representing approximate lateral extent of saturated portions of various hydrogeologic units; representing existing predevelopment groundwater levels and changes in groundwater levels in various parts of the model domain; representing vertical differences in groundwater levels, or differences between groundwater levels and levels of connected surface water, where such information is available; and representing estimated or measured groundwater discharge to surface features such as streams and springs. Sparsity of observation data in some parts of the model domain did not permit adjustment of some parameters in those areas. The product of this approach is a model that represents important elements in the regional flow system and approximates regional flow patterns and responses to stresses such as groundwater pumping and variations in recharge.

Simulated Transmissive Properties

Simulated hydraulic properties are discussed by major hydrogeologic unit in each model layer. First, hydraulic conductivity, horizontal anisotropy, and vertical anisotropy distributions are discussed, followed by a discussion of transmissivity distributions. Hydraulic properties are distributed across each model layer by assigning values to multiple polygons that represent variations in lithology or hydraulic gradients in the primary hydrogeologic units (fig. 14). The individual hydraulic property polygons generally encompass larger regions on the Colorado Plateau where the primary hydrogeologic units are lithologically similar across broad regions than in the alluvial basins where lithology is variable across small regions because of structural deformation and alluvial facies variations. In general, simulated hydraulic properties for each hydrogeologic unit vary across about three orders of magnitude. However, lower hydraulic conductivity values were required for each layer in the northeast part of the simulated area than elsewhere. The low hydraulic conductivity values in the northeast area may indicate a poor conceptual model of the hydrogeologic system, including inaccurate recharge rates and a poorly understood groundwater-flow system. Uncertainties in simulation of the northeast area, however, should not result in greater uncertainty in simulation of the focus area of the model, that is the Verde River groundwater-flow system and adjacent areas.

Hydraulic Conductivity

Simulated distributions of hydraulic conductivity along rows are described below and shown for each layer in fig. 14. However, the full range of simulated hydraulic conductivity values, both along rows and along columns, includes the effects of simulated anisotropy. As a result of simulated anisotropy, hydraulic conductivity along columns may be greater or less than values shown on figure 14. Several areas include hydraulic conductivity values that are greater along columns by a factor of as much as 5, including areas near the Mogollon Rim and parts of the Big Chino and Verde Valley sub-basins, and by a factor of 10 in the Little Chino sub-basin. Also, a few areas include hydraulic conductivity values that are less along columns than along rows by a factor of as great as 10.

Layer 1

Hydraulic conductivity along rows for the Coconino aquifer in layer 1 (fig. 14*A*) ranges from low values (2.0×10^{-4} to 0.3 ft/d), in the northeast area to higher values (6.6×10^{-3} to 33 ft/d) across the remainder of model layer 1. The highest values of hydraulic conductivity in layer 1 in the region of the Coconino aquifer is in a broad region east of Flagstaff. Hydraulic conductivity for the part of layer 1 in the region of the Verde Formation was more uniform than many other hydrogeologic units, varying across a small range, 3.3×10^{-1} to 13 ft/d. Hydraulic conductivity for interbedded intervals of basin fill in the Big Chino Basin also varied across a small range, 6.6 to 66 ft/d, with the greatest values associated with intervals that include basalt flows in the Paulden area. Hydraulic conductivity for silt and clay intervals of basin fill in the Big Chino Basin was 6.6×10^{-1} ft/d, but greater values, 66 ft/d, were applied to intervals that include basalt flows in the Paulden area. Hydraulic conductivity for basin fill in the Little Chino and Upper Agua Fria basins varied from about 3.3×10^{-1} to 32.8 ft/d.

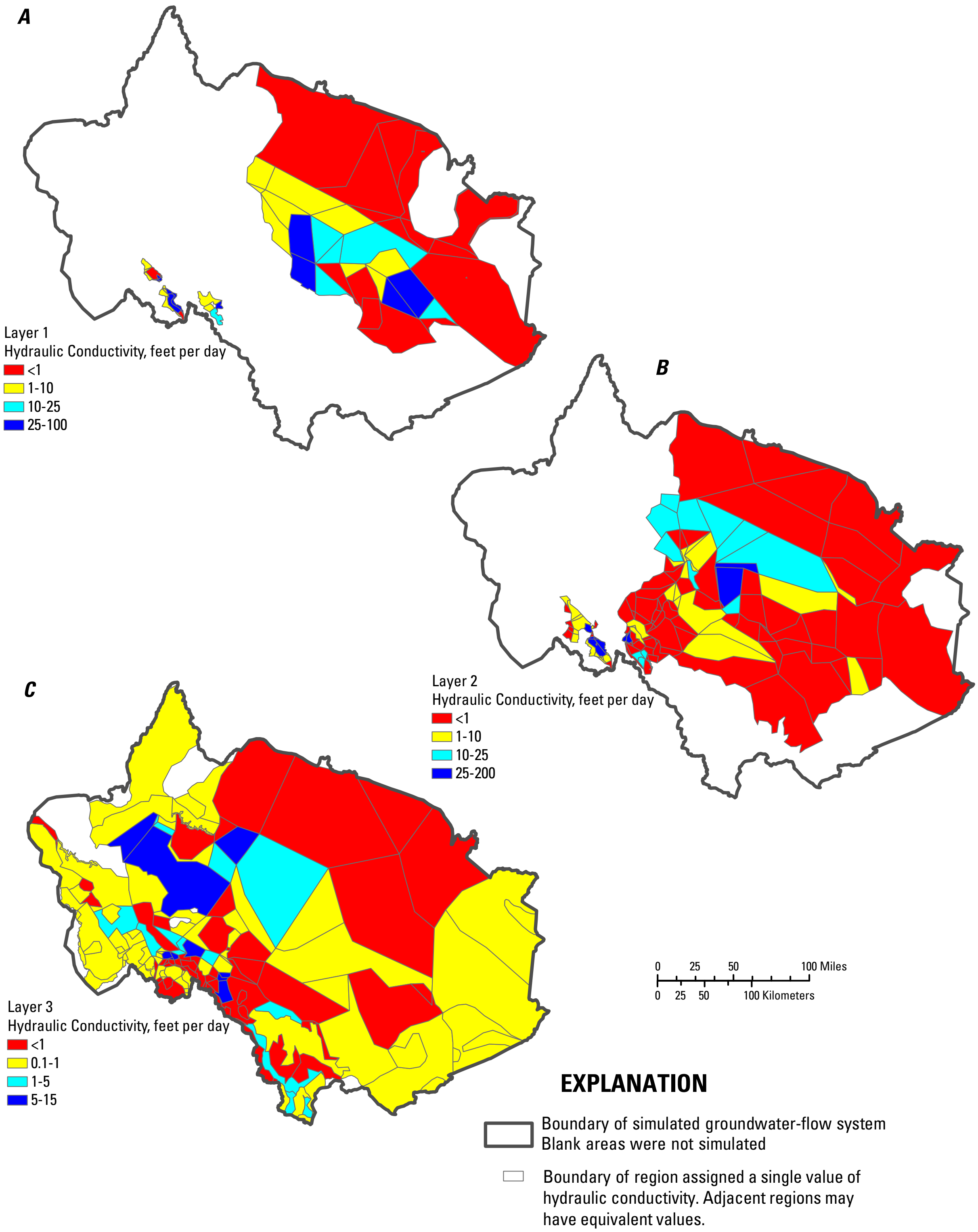


Figure 14. Distributions of hydraulic conductivity along rows for the Northern Arizona Regional Groundwater-Flow Model layers. *A*, Layer 1. *B*, Layer 2. *C*, Layer 3.

Layer 2

Simulated hydraulic conductivity values along rows for the Supai Formation in layer 2 (fig. 14*B*) vary from low values in the northeast area, 6.6×10^{-5} to 3.3×10^{-1} ft/d, to 1.3×10^{-2} to 39 ft/d across the remainder of layer 2. Similar to hydraulic conductivity distributions in layer 1, the highest values of hydraulic conductivity for the Supai Formation in this layer are across a broad region east of Flagstaff. Hydraulic conductivity for sand and gravel and conglomerate intervals of basin fill in the Big Chino and Verde Valleys varied across a broad range, 1.3×10^{-2} to 66 ft/d, with the greatest values in the Paulden area. Hydraulic conductivity for the lower volcanic unit in the Little Chino sub-basin ranged from 9.8 to 66 ft/d with lowest values near Del Rio Springs and the highest values in the central part of the basin. Hydraulic conductivity for basin fill conglomerate in the upper Agua Fria basin ranged from about 3.2×10^{-1} to 66 ft/d, with the lowest values in the southernmost part of the basin and greatest values in the north-central part of the basin.

Layer 3

Simulated values of hydraulic conductivity for the Redwall-Muav aquifer are less than 0.1 ft/d for much of the layer 3 extent (fig. 14*C*). Areas of higher hydraulic conductivity for the Redwall-Muav aquifer, more than 1 ft/d, are along the lower reaches of the Little Colorado River, Big Chino Basin, Verde Valley, Coconino Plateau, and parts of the Mogollon Rim. The highest hydraulic conductivity values for the Redwall-Muav aquifer, 13 ft/d, are on the Coconino Plateau. Hydraulic conductivity values for crystalline rocks of layer 3 are less than 1 ft/d, with the lowest values of 3×10^{-3} ft/d or less at the northern extent of the Little Chino sub-basin. Areas of basalt and basin fill in the western basins were simulated by using low hydraulic conductivity values of less than 1 ft/d.

Anisotropy

Variations in the original distributions of horizontal and vertical anisotropy were required in some areas of the model to reproduce observed hydraulic head distributions and distributions of stream discharge. Anisotropic conditions are where groundwater flows more readily in one direction than another.

Horizontal

In the NARGFM, anisotropy of the horizontal hydraulic conductivity distribution is approximated by specifying a ratio of hydraulic conductivity in the direction along rows to hydraulic conductivity along columns. An anisotropy value that is greater than 1 indicates that hydraulic conductivity along columns is greater than along rows by the specified ratio. An anisotropy value that is less than 1 indicates that hydraulic conductivity along columns is less than along rows by the specified ratio. Anisotropy of the vertical hydraulic conductivity distribution is approximated by specifying a ratio of horizontal hydraulic conductivity along rows to vertical hydraulic conductivity between model layers. Horizontal hydraulic conductivity along rows is normally greater than vertical hydraulic conductivity between model layers and most vertical anisotropy ratios are greater than 1 throughout the NARGFM, except for one small area. Areas that required the greatest horizontal anisotropy included areas where groundwater flow may be influenced by geologic structures.

Horizontal anisotropy was applied to some areas of the model, on the basis of alluvial depositional conditions and likely orientation of secondary porosity related to structural deformation, to better match simulated and observed hydraulic head and simulated and observed groundwater discharge rates. Anisotropy was assigned to parts of the northeast area of the model and regions near the Mogollon Rim where anisotropy is related to secondary porosity caused by structural deformation. Anisotropy in parts of the alluvial basins is primarily related to greater hydraulic conductivity in the orientation of the primary orientation of deposition, which is normally parallel to the primary northwest to southeast orientation of the basins. Hydraulic conductivity along columns is calculated as the product of hydraulic conductivity along rows and the anisotropy value. Groundwater is more readily transmitted along columns where anisotropy values are >1. Layer 1 generally was simulated by using isotropic conditions, except where a value of 5 was assigned throughout the layer extent in the Big Chino sub-basin and along the boundaries of the Little Chino and Upper Agua Fria sub-basins. Layer 2 generally was simulated by using isotropic conditions, except where a value of 5 was assigned in the Big Chino sub-basin outside of Williamson Valley, along parts of the boundaries of the Little Chino sub-basin, for the Supai Formation in the region of the Mogollon Rim, and parts of the northeast model extent where values of 3 to 5 were assigned. One region east of Flagstaff was assigned an anisotropy value of 0.3. Layer 3 generally was simulated by using isotropic conditions, except areas in the northeast model extent, Mogollon Rim, Coconino Plateau, and Big and Little Chino sub-basins. Anisotropy values of 0.1 and 0.5 were assigned to granite and limestone areas near Mingus Mountain and to a region of granite in the hills on the west side of the Little Chino sub-basin, respectively. Anisotropy values of 0.2 were assigned to part of the northeast area, northern Coconino Plateau, and near Oak Creek. Anisotropy values of 3 to 5 were assigned to areas along the Mogollon Rim, areas of Redwall-Muav Limestone in the Big Chino sub-basin, and some granite areas on the boundaries of the Little Chino and Upper Agua Fria sub-basins. The highest horizontal anisotropy values of 10 were assigned to the lower volcanic unit in the Little Chino sub-basin.

Vertical

Anisotropy of the vertical hydraulic conductivity distribution is approximated by specifying a ratio of horizontal hydraulic conductivity along rows to vertical hydraulic conductivity between model layers. Horizontal hydraulic conductivity along rows is normally greater than vertical hydraulic conductivity between model, layers and most vertical anisotropy ratios are greater than 1 throughout the NARGFM except for one small area. The greatest vertical anisotropy was required in areas of thick silt and clay in alluvial basins where water-level data indicated major vertical differences in hydraulic head. Initial vertical anisotropy values were altered during the model development process to better match simulated and observed hydraulic head distributions and to better match simulated and observed groundwater discharge to streams and drains in some areas. Simulated values of vertical anisotropy in the Coconino aquifer part of layer 1 ranged from 1 at the lower reaches of the Little Colorado River to 10 across much of the remaining extent of the Coconino aquifer and 200 in the northeast area. Simulated values of vertical anisotropy in the interbedded parts of alluvial basins in layer 1 ranged from 10 to 100 in the Big Chino sub-basin and along the margins of the Little Chino and Upper Agua Fria sub-basins to as great as 5,000 in the central parts of the Little Chino and Upper Agua Fria sub-basins. Simulated values for fine-grained basin fill of layer 1 included values of 5,000 in the region of silt and clay in Little Chino sub-basin, values of 2,000 in the region of silt and clay in Big Chino sub-basin, and values of 500 to 5,000 for the Verde Formation. Simulated values of vertical anisotropy in the Supai Formation part of layer 2 ranged from isolated regional values of 10 in areas where good hydraulic connection was required among all model layers to values of 500, 2,000, and 5,000 in the central part of the Coconino aquifer where poor vertical hydraulic connection between layers was required. Simulated values of vertical anisotropy in the alluvial basin part of layer 2 included values of 10 across the Big Chino sub-basin and along boundaries of the Verde Valley, values of 100 to 500 along the margins of the Little Chino sub-basin and across most of the Verde Valley, and values of 1,000 to 5,000 in the central parts of the Little Chino and Upper Agua Fria sub-basins. Vertical anisotropy values of 5,000 were required for the lower volcanic unit in the Little Chino and Upper Agua Fria sub-basins to produce large differences in hydraulic head between layers that have been observed by water levels measured in wells that tap the lower volcanic unit and upper alluvial unit. Simulated values of vertical anisotropy in layer 3 remained at 10, except in a region along the lower reaches of the Little Colorado River where values of 2 were assigned.

Simulated Storage Properties

Specific Storage

Specific storage values for the Coconino and Kaibab Formations of layer 1 ranged from 3 x 10^{-6} to 3 x 10^{-3} ft^{-1}. Specific storage values for the interbedded basin fill of layer 1 ranged from 3 x 10^{-6} to 3 x 10^{-5} ft^{-1}. Specific storage for the fine-grained intervals of basin fill in layer 1 in the Big Chino and Little Chino sub-basins and the Verde Formation was uniformly assigned a value of 3 x 10^{-4} ft^{-1}. Specific storage values for the Supai Formation of layer 2 ranged from 1.0 x 10^{-5} to 1.0 x 10^{-4} ft^{-1}. Specific storage for the coarse-grained basin fill of layer 2 in the Big Chino sub-basin, upper Agua Fria sub-basin, and Verde Valley was assigned a uniform value of 1.0 x 10^{-4} ft^{-1}. Specific storage values for the lower volcanic unit of layer 2 in the Little Chino sub-basin ranged from 3 x 10^{-7} to 3 x 10^{-5} ft^{-1}. Simulated specific storage for crystalline rock and Redwall-Muav aquifer in layer 3 was a uniform value of 3.0 x 10^{-7} ft^{-1}. Alluvial basins simulated as layer 3 in the groundwater sub-basins in the Salt River Basin were assigned a uniform value of 3.0 x 10^{-6} ft^{-1}.

Specific Yield

Distributions of specific yield are shown for each model layer in figure 15. Specific yield values for the Coconino and Kaibab Formations of layer 1 ranged from 0.06 to 0.25. Values of specific yield for the interbedded basin fill of layer 1 ranged from 0.05 to 0.15. Fine-grained intervals of basin fill in layer 1 in the Big Chino sub-basin were assigned a specific yield value of 0.01. Fine-grained intervals of basin fill in layer 1 in the Verde Formation were assigned a specific yield value of 0.10. Fine-grained intervals of basin fill in layer 1 in the Little Chino sub-basin were assigned large values of 0.25 to allow for variations in storage in the shallow aquifer. Specific yield values for the Supai Formation of layer 2 ranged from 0.06 to 0.25. Specific yield values for the coarse-grained basin fill of layer 2 in the Big Chino, upper Agua Fria, and Verde Valley sub-basins and for the lower volcanic unit of layer 2 in the Little Chino sub-basin ranged from 0.03 to 0.20. Specific yield for crystalline rock, Redwall-Muav aquifer, western alluvial basins, and White Mountains area of layer 3 was simulated by using a uniform value of 0.01. Alluvial basin portions of layer 3 in the Little Chino and upper Agua Fria sub-basins were assigned specific yield values of 0.05 to 0.10. Alluvial basins simulated as layer 3 in the Salt River sub-basins were assigned a uniform specific yield value of 0.10.

Evaluation of the Simulation of Groundwater Flow

Simulation of the calibrated groundwater-flow system is evaluated by comparing simulated and observed hydraulic head and groundwater discharge to streams and springs for predevelopment steady-state and transient conditions. Evaluations are discussed by major groundwater basins and sub-basins defined by ADWR including Big Chino, Little Chino, upper Agua Fria, Verde Valley, and Verde Canyon sub-basins, the Little Colorado River Plateau and Coconino Plateau basins, the west basins as a group, and the sub-basins along the Salt River. The Kaibab Plateau also is simulated, but few hydrologic data are available to evaluate the simulated groundwater-flow system in this area. Simulated groundwater-flow divides were locally different from the divides between ADWR groundwater basins and sub-basins. The influence of simulated groundwater divides that are different from those divides defined by ADWR is discussed for groundwater basin where the difference is important.

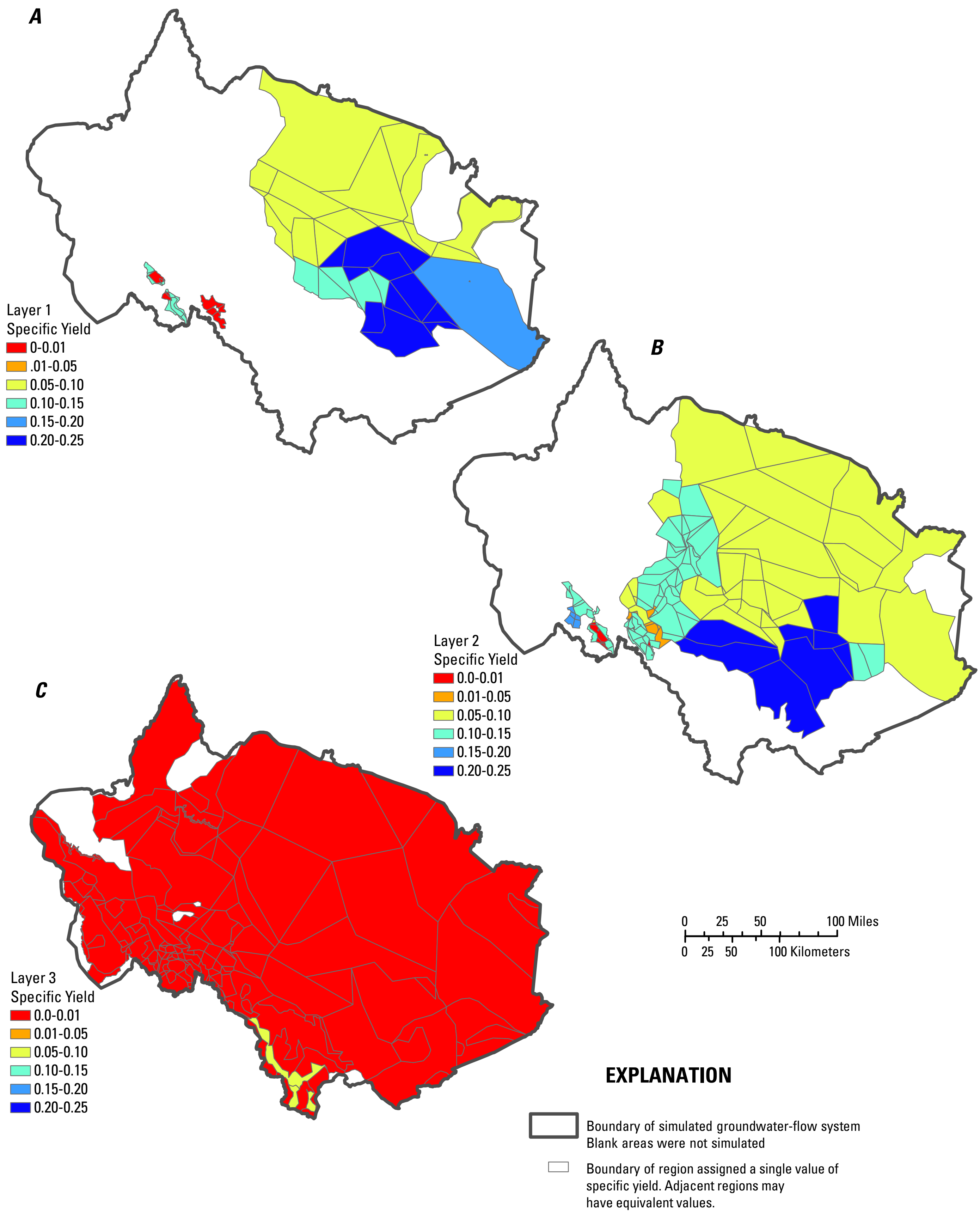


Figure 15. Distributions of specific yield for the Northern Arizona Regional Groundwater-Flow Model layers (*A*) layer 1, (*B*) layer 2, and (*C*) layer 3.

The simulated groundwater budgets for each basin and sub-basin are discussed for predevelopment and transient conditions. Predevelopment groundwater budgets are presented in a table for each basin or sub-basin. Evaluation of the transient groundwater budgets is discussed and presented using graphs showing changes in major groundwater budget components, graphs showing simulated and estimated groundwater discharge at several streamflow-gaging stations, and a table of the groundwater budget for the final year of simulation, 2005.

Observation Data

Data from hydrologic observations that define the groundwater-flow system prior to development for supply and during development (transient conditions) include the altitude of water levels in wells and estimates of groundwater discharge to streams based on measured streamflow at gaging stations. Early water levels in wells were assumed to represent the hydraulic head in the groundwater-flow system prior to development of the groundwater supply (appendix 1). Multiple water-level measurements have been made at many wells in the study area; however, reliable decadal or longer trends in water levels require more than two water-level measurements made during periods of two or more decades. Wells with more than 10 water-level measurements during multiple decades were used to define transient changes to the hydraulic head distribution that were simulated by the model (appendix 2). Base flow, the part of streamflow derived from groundwater discharge, was estimated at streamflow-gaging stations by using methods appropriate for each gage. For comparison with model simulation of base flow, average annual estimates of base flow were required because annual estimates include seasons where flow may be diminished by evapotranspiration. Steady-state base flow was estimated for streamflow-gaging stations with a long period of record, more than 10 years, as the average of monthly minimum daily flows for the steady-state period, which varied across the study area. Many streamflow gages lacked sufficient continuous records prior to development of groundwater supplies to reliably estimate steady-state base flow, so steady-state base flow was estimated on the basis of published measurements or estimates. Changes in base flow at the streamflow-gaging stations were used to define variations in groundwater discharge during the transient simulation.

Predevelopment Conditions

The predevelopment simulation is evaluated by comparison of simulated and observed hydraulic head at control points, mostly wells, and comparison of simulated groundwater discharge and groundwater discharge estimated from streamflow records, mostly at streamflow-gaging stations. In addition, estimated rates of predevelopment evapotranspiration were available in some areas for evaluation of simulated evapotranspiration rates. Most values of simulated predevelopment hydraulic head at control points should fall within the estimated error of the observations, which includes error in the measurement of depth to water and error in the estimated altitude of the measurement reference, usually land surface. Error in the measured depth to water at wells was estimated for each basin or sub-basin on the basis of the standard deviation of the earliest depths-to-water measurements that were repeated at individual wells. For most basins and sub-basins, the repeated measurements used to estimate depth-to-water error occurred before 1961. Error in the estimated measurement reference altitude was derived from the ADWR Groundwater Site Inventory well records, which include estimated land-surface error for each well. Total hydraulic head error was estimated for each well with repeated predevelopment depth-to-water measurements as the square root of the sum of the squared depth-to-water standard deviation and squared altitude error. Estimated hydraulic head error was calculated for each basin or sub-basin as the average of the estimated hydraulic head error at each well that had records of repeated predevelopment depth-to-water meaurements. Simulated predevelopment groundwater discharge was evaluated for several basins or sub-basins on the basis of the estimated groundwater discharge component of early streamflow records at one or more streamflow-gaging stations. Many basins and sub-basins lacked predevelopment streamflow records, however, and later records were used as a guide to evaluate the simulation in those areas.

Water-level data used for comparison with simulated steady-state hydraulic heads during periods before development of the groundwater system for supply include the earliest pre-1961 data available at wells (figs. 16 and 17). Extensive development of groundwater supplies happened after 1960 in many areas. The earliest pre-1961 water-level data were used to define steady-state hydraulic head distributions in most of those areas. Pre-1961 water-level measurements in wells that tap water-yielding zones in crystalline rocks, alluvial aquifers, and the Coconino aquifer include 38 pre-1940 water levels, 95 measured during 1940–49, and 318 measured during 1950–59. Much of the earliest water-level data is concentrated in the Little Chino sub-basin because the earliest groundwater development was in that area. Only six water levels measured before 1961 are available for wells that tap the Redwall-Muav aquifer. However, most wells that tap the Redwall-Muav aquifer are in areas where the aquifer is poorly developed or was developed much later than other areas. The earliest water-level data available for the Redwall-Muav aquifer in these poorly developed areas were collected at 21 wells after 1960 and as late as 2003.

The variability of the steady-state water-level data was evaluated by calculating the standard deviation of the depths to water at wells that had multiple observations before the groundwater system was heavily developed for supply. Predevelopment observations at single wells were repeated once to as many as 56 times. Estimates of water-level variability were made for each major aquifer. Predevelopment depths to water in the basin fill and volcanic aquifers of the Big Chino, Little Chino, and Upper Agua Fria sub-basins had an

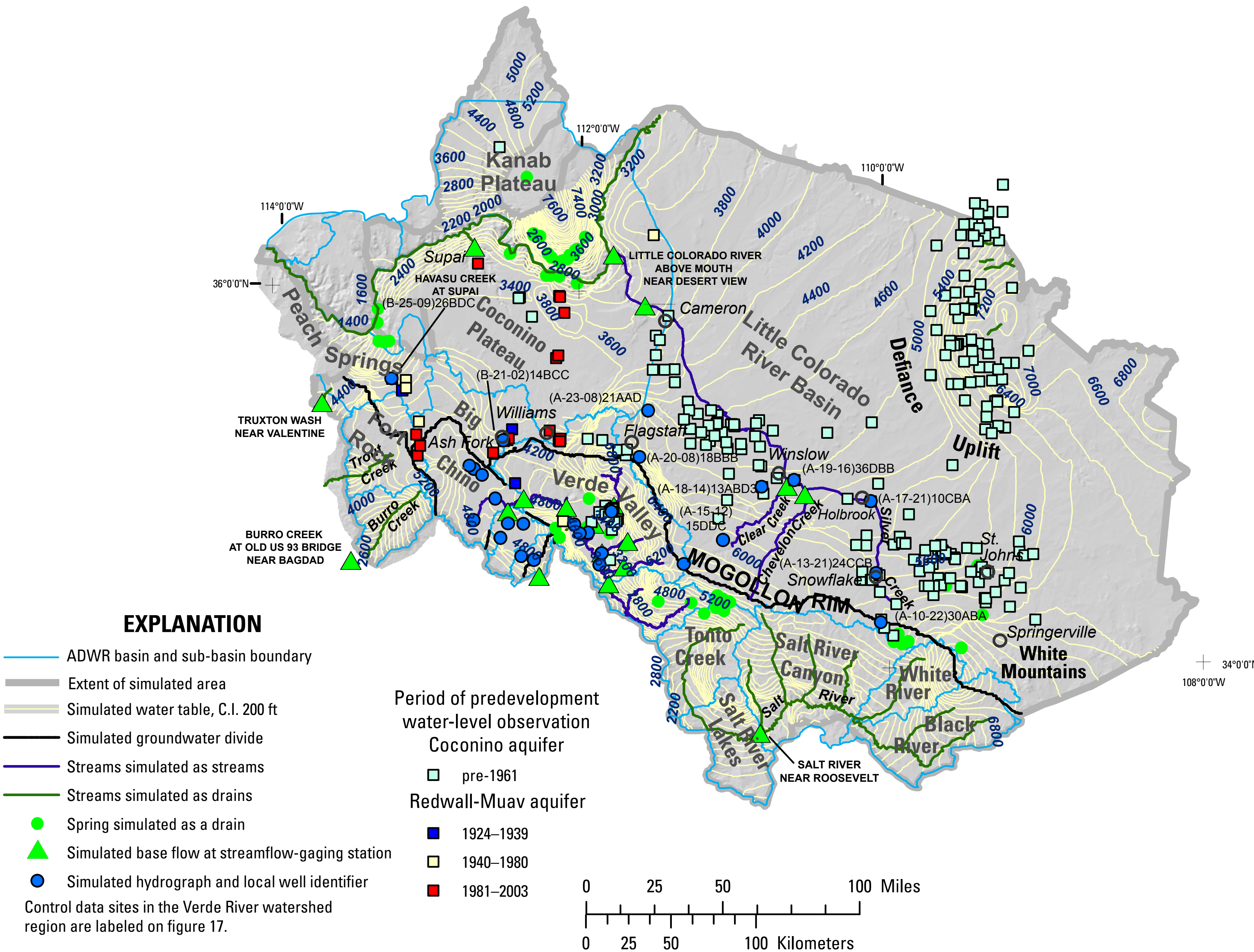


Figure 16. Simulated predevelopment groundwater-flow system, Arizona Department of Water Resources (ADWR) basins and sub-basins, and predevelopment control data sites including wells with water levels and streamflow-gaging stations in the area of the Northern Arizona Regional Groundwater-Flow Model.

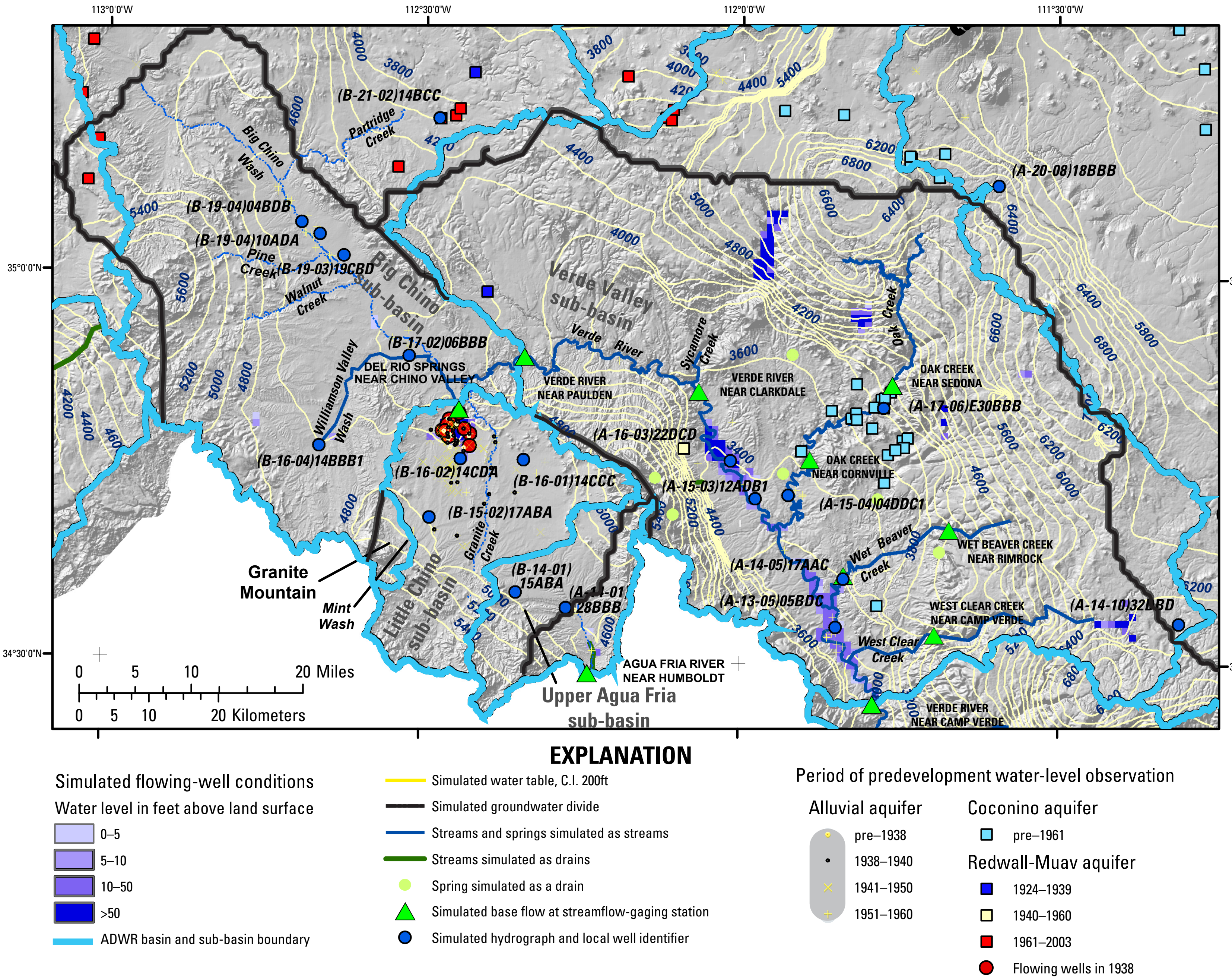


Figure 17. Simulated predevelopment groundwater-flow system in the Arizona Department of Water Resources (ADWR) sub-basins in the Verde River Watershed region of the Northern Arizona Regional Groundwater-Flow Model and predevelopment control data sites including wells with water levels and streamflow-gaging stations.

average standard deviation of about 9 ft. Predevelopment water levels in the Verde Formation of the Verde Valley sub-basin had an average standard deviation of 2.7 ft. The standard deviation of pre-1961 water levels in the Coconino aquifer was 4.8 ft. Predevelopment water levels in poorly developed areas of the Redwall-Muav aquifer had a standard deviation of 7.2 ft.

Predevelopment steady-state base flow was estimated from pre-1961 streamflow records and estimates from previous studies. Five streamflow-gaging stations have sufficient early records to estimate steady-state base flow at streams that receive discharge from the major simulated aquifers, including Verde River near Clarkdale, Verde River near Camp Verde, Oak Creek near Cornville, Salt River near Roosevelt, and Little Colorado River near Cameron. Records at the Verde River near Camp Verde are greatly influenced by seasonal streamflow diversions for agricultural use and are not useful for estimating groundwater discharge without more information about the rate and timing of the diversions. Average-annual base flow at the Verde River near Camp Verde may provide, however, a minimum estimate of average annual groundwater discharge upstream from the streamflow-gaging station. Other estimates of steady-state base flow that are available from published reports include the Agua Fria River near Humbolt and Del Rio Springs (Nelson, 2002) and the lower reaches of Chevelon and Clear Creek (Leake and others, 2005). One-time estimates of base flow also are available for many springs across the study area; however, these estimates generally were not used to evaluate the simulation of the groundwater-flow system because variability of spring flow is poorly known. Early streamflow records at a few streamflow-gaging stations may have been influenced by development of the groundwater system. Base flow was estimated by using early records at these stations and can be considered an estimate of minimum steady-state base flow at several streamflow gages, including Verde River near Paulden (beginning 1963), Little Colorado River upstream from the mouth (beginning 1990), Wet Beaver Creek near Rimrock (beginning 1963), West Clear Creek near Camp Verde (beginning 1966), and East Verde River near Childs (beginning 1962).

Transient Conditions

Water-level records at 83 wells with 10 or more observations across multiple decades were selected to compare observed and simulated changes in hydraulic head during the transient simulation period; however, only a subset were selected as representative of water-level changes in each of the basins and sub-basins (figs. 16 and 17). Thirty hydrograph wells are in the Little Chino and Upper Agua Fria sub-basins, including eight wells that lie outside of the extent of the alluvial and volcanic aquifers and tap water in crystalline rocks or thin alluvial deposits that are simulated as part of model layer 3. Six hydrograph wells are in the Big Chino sub-basin, all of which tap the basin-fill aquifer. Five hydrograph wells in the Verde Valley sub-basin tap the Verde Formation and adjacent alluvial deposits, layers 1 and 2, respectively. Thirty-nine hydrograph wells tap the Coconino aquifer. Two hydrograph wells tap the Redwall-Muav aquifer on the Coconino Plateau. One hydrograph well taps water in the crystalline rocks west of the town of Payson.

Base-flow estimates from records at 15 streamflow-gaging stations were available to compare 10 years or more of continuous observed and simulated changes in groundwater discharge from the major simulated aquifers from 1940 through 2005 (figs. 16 and 17). Ten of the streamflow-gaging stations are along the Verde River and major tributaries. Several of the streamflow-gaging station records of multi-decadal length had substantial variations in estimated base flow. The variations at any particular streamflow-gaging station may be related to groundwater withdrawals, variations in recharge, or variations in runoff.

Evaluation of Simulated Predevelopment Conditions

Simulation of the groundwater-flow system in the NARGFM region prior to extensive development of the groundwater supplies for human use is evaluated for each major groundwater basin and sub-basin defined by the ADWR. Simulated and estimated predevelopment water levels and groundwater budgets are discussed for each area. Areas of primary focus of the NARGFM are discussed first, including the Big Chino, Little Chino, Upper Agua Fria, and Verde Valley sub-basins. Other basins that are adjacent to the primary focus areas are then discussed, including the Little Colorado River Plateau and Coconino Plateau basins. The western basins that include the Burro Creek and Fort Rock (Trout Creek) sub-basins and a part of Peach Springs basin (Truxton Wash watershed) are discussed. Finally, the Verde Canyon sub-basin, Tonto Creek sub-basin, and several sub-basins in the Salt River drainage are discussed. Groundwater budgets also were developed for several other regions that conform with simulated groundwater-flow systems that differ substantially from the groundwater basins defined by the ADWR and for sub-regions where base flow derived from groundwater is defined by observations. These simulated groundwater systems include the part of the Big Chino sub-basin that contributes groundwater flow to the Verde River, the combined areas of the Big Chino and Little Chino sub-basins that contribute groundwater flow to the Verde River, Verde Valley sub-basin upstream from the streamflow-gaging station near Clarkdale, Verde Valley sub-basin between the streamflow-gaging stations near Clarkdale and near Camp Verde, and the groundwater-flow systems for the Little Colorado and Coconino Plateau basins. The predevelopment groundwater-flow systems are not discussed, but simulated groundwater budgets are provided for a few areas of the NARGFM that are lacking observations, including much of the Peach Springs basin and several basins north of the Colorado River, including the Grand Wash, Shivwits Plateau, and Kanab Plateau basins.

Little Chino and Upper Agua Fria Sub-Basins

Previous studies of the groundwater-flow system in the Little Chino and Upper Agua Fria sub-basins have assumed that steady-state conditions existed in 1940 following a few decades of agricultural irrigation derived from diverted surface water. However, an expansion of groundwater use followed the drilling of the first flowing well near Del Rio Springs in 1930 (Schwalen, 1967). The first study of the groundwater system documented some useful early groundwater information beginning in 1938, when 13 flowing wells and one pumped well were reportedly used for irrigation and the Santa Fe Railroad reportedly had exported water from two wells adjacent to Del Rio Springs since 1926 (Schwalen, 1967). An evaluation of predevelopment groundwater conditions in the area suffers from a lack of pre-1938 data including few water levels in wells. Only seven pre-1938 water levels and 16 water levels measured in 1938 are available (appendix 1). No streamflow or spring-flow records are available for the period before 1940. In addition, many irrigation system improvements were made prior to 1938, such as diversions from impounded runoff for irrigation that likely resulted in incidental recharge (Schwalen, 1967) and uncertain transient hydrologic effects. Simulated conditions prior to known groundwater withdrawals and after surface-water irrigation system improvements are evaluated as steady-state for the Little Chino and Upper Agua Fria sub-basins, with the understanding that limitations in hydrologic data define the groundwater system during this time and that changes in the groundwater-flow system may have happened before collection of the earliest water-level, streamflow, and spring-flow measurements.

Evaluation of the simulated steady-state groundwater-flow system in the Little Chino and Upper Agua Fria sub-basins was made through the comparison of simulated and observed hydraulic head measured at 103 wells before 1951, including seven pre-1937 wells (appendix 1), and estimated groundwater discharge from the basins through evapotranspiration and flow to Del Rio Springs and the Upper Agua Fria River (fig. 17). Water levels at two wells—(B-16-01)20CBD1 and CBD2—measured in 1940 near Granite Creek were eliminated from the comparison because water levels were abnormally high, more than 100 ft above other nearby wells, and likely represent recharge along the ephemeral stream channel, a process that was not explicitly simulated in this numerical model.

Efforts were made to accurately simulate pre-1938 water-level altitudes as steady-state water level altitudes. Water levels collected during early transient conditions of 1938–1950 were considered to underestimate steady-state hydraulic head and represent a minimum target head for simulation. Average simulated error at seven wells where depths to water were measured before 1938 was 5.1 ft with an average absolute error of 17.1 ft . In comparison, the standard deviation of depth to water measured at the seven wells before 1938 was about 8.9 ft. The standard deviation of observed water-level altitudes, which include errors in the estimated land-surface altitude at the well, at the same seven wells was about 17.1 ft (fig. 18). Comparison of simulated steady-state water-level altitudes with observed post-1937 water-level altitudes indicate an increasing deviation of simulated from estimated steady-state water-level altitude, with the simulated steady-state water levels increasingly higher than estimated water-level altitudes with time. Simulated steady-state water levels at 38 wells measured during 1938–40 were above measured values by an average of 7.5 ft with an average absolute error of 20.7 ft for estimated water-level altitudes that have a standard deviation of about 15.6 ft. Simulated steady-state water levels at 58 wells measured during 1941–1950 were above measured values by an average of 10.1 ft with an average absolute error of 26.3 ft for estimated water-level altitudes that have a standard deviation of about 15.8 ft. Simulated hydraulic heads in the lower volcanic unit were as much as 100 ft above the land surface in 1938 near Del Rio Springs. This region of simulated flowing wells, where hydraulic heads of more than 5 to 75 ft above land surface were observed, includes all of the 11 flowing wells that were documented at the time (Schwalen, 1967) (fig. 17). Considering uncertainty in early pumping and recharge distributions and inability to accurately define water-level altitudes that existed before the initial development of the groundwater supplies, the simulated steady-state hydraulic heads in the Little Chino sub-basin reasonably represent the observed heads.

Inflow to the steady-state groundwater-flow systems within the Little Chino and Upper Agua Fria sub-basins includes natural recharge, incidental recharge associated with agricultural irrigation, and groundwater flow from adjacent basins. Rates of simulated natural recharge through direct infiltration of precipitation in the Little Chino sub-basin under steady-state conditions was about 2,400 ac-ft/yr (table 1) based on BCM results that were modified to include about 8.4×10^{-3} ft/yr of ephemeral channel recharge applied to the alluvial surface. Additional incidental recharge along canals and at irrigated fields of about 2,000 ac-ft/yr (table 1) was simulated as deep percolation of excess irrigation water in the Little Chino sub-basin by using the MODFLOW Well Package. Recharge in the Little Chino sub-basin is augmented by groundwater inflow from adjacent areas of about 2,200 ac-ft/yr (table 1), including 800 ac-ft/yr from the Big Chino sub-basin in an area near Granite Mountain, 900 ac-ft/yr from the Upper Agua Fria sub-basin, and 500 ac-ft/yr from the Black Hills in the Verde Valley sub-basin. Total inflow to the Little Chino sub-basin groundwater system was about 6,600 ac-ft/yr (table 1). Simulated inflow to the groundwater-flow system in the Upper Agua Fria sub-basin included steady-state natural recharge of about 1,300 ac-ft/yr and groundwater inflow of about 200 ac-ft/yr from the area of the Black Hills (table 1). The total simulated steady-state inflow to the Little Chino and Upper Agua Fria groundwater sub-basins—the PrAMA—of about 7,700 ac-ft/yr includes natural recharge of 4,100 ac-ft/yr, incidental recharge of 2,000 ac-ft/yr, and a net groundwater inflow from adjacent areas of about 1,500 ac-ft/yr. Inflow to the groundwater-flow system in the PrAMA area was less than the estimated steady-state recharge of about 10,000 ac-ft/yr simulated by Nelson (2002) and about 9,900 ac-ft/yr simulated by Timmons and Springer (2006). The NARGFM

Table 1. Predevelopment groundwater flow budgets for Arizona Department of Water Resources groundwater basins and sub-basins and selected regions of the Northern Arizona Regional-Flow Model (NARGFM).

[Values are in acre-feet per year except for cumulative values, which are in acre feet.]

		Verde River Basin groundwater-flow system								Colorado River Basin groundwater-flow system				Western basins groundwater-flow system				Salt River sub-basins groundwater-flow system		
Groundwater Basin	Upper Agua Fria	Little Chino	Little Chino and Upper Agua Fria sub-basins	Big Chino	Big and Little Chino	Verde Valley sub-basin above the streamflow-gaging station near Clarkdale	Verde Valley sub-basin between the streamflow-gaging stations near Clarkdale and near Camp Verde	Verde Valley	Verde Canyon	Little Colorado	Coconino Plateau	Peach Springs	Kanab Plateau and adjacent areas	Burro Creek	Fort Rock (Trout Creek)	Truxton Wash	Western basins	Tonto Creek	Salt River Lakes	Salt River above the streamflow-gaging station at Roosevelt
Arizona Department of Water Resources basin type	sub-basin	sub-basin	Prescott Active Manage-ment Area	sub-basin	sub-basins	part of sub-basin	part of sub-basin	sub-basin	sub-basin	basin	basin	Part of basin that lies outside of Truxton Wash watershed	several basins north of the Colorado River	sub-basin	sub-basin	Parts of Peach Springs basin and Wikieup sub-basin	simulated ground-water basin[1l]	basin	basin	Salt River Canyon, Black River, and White River sub-basins
Groundwater-budget component Inflow																				
Natural recharge	1,300	2,400	4,100	41,600	43,900	30,000	61,400	91,400	35,200	206,600	152,000	6,700	100,300	17,900	19,500	4,500	41,900	52,800	37,800	177,800
Recharge from infiltration of streamflow derived from base flow[1a]	N/A	0	0	6,500	6,500	13,400	22,000	35,400	2,000	12,600	300	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Incidental recharge[1b]	0	2,000	2,000	0	2,000	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Groundwater inflow from adjacent areas[1c]	200	2,200	1,500	2,700	4,900	8,800	1,700	8,300	31,200	2,600	192,300	5,400	36,400	5,400	1,200	1,700	2,700	24,400	2,500	10,900
Total Inflow[1d]	1,500	6,600	7,700	50,800	57,400	52,200	85,100	135,100	68,300	221,800	344,600	12,100	136,700	23,300	20,700	6,100	44,600	77,200	40,300	188,600
Outflow																				
Groundwater discharge to streams (base flow)[1e]	N/A	5,300	5,300	22,800	28,200	34,900	61,700	96,500	56,600	12,600	169,200	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Discharge to streams and springs simulated as drains (base flow)[1f]	1,000	0	1,000	0	0	0	100	100	300	400	151,400	7,400	128,200	23,300	11,600	2,800	37,800	74,300	40,300	188,300
Evapotranspiration by phreatophytes[1g]	200	100	300	2,200	2,300	2,600	9,500	12,100	N/A	0	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Groundwater withdrawals	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
Groundwater outflow to adjacent areas[1h]	800	1,000	1,100	25,800	26,900	14,800	13,800	26,400	11,400	208,900	24,000	4,800	8,500	0	9,000	3,300	6,800	2,900	0	300
Total Outflow[1i]	2,000	6,600	7,700	50,800	57,400	52,400	85,100	135,100	68,300	222,000	344,600	12,100	136,700	23,300	20,700	6,100	44,600	77,200	40,300	188,600
Net groundwater flow to (-) and from (+) adjacent basins[1j]	-600	1,000[2]	400	-23,100[3]	-22,000[3]	-5,900	-12,100	-18,000	19,800	-206,300	168,300	600	27,800	5,400	-7,900	-1,600	-4,100	21,500	2,500	10,600
Net streamflow[1k]	N/A	5,300	5,300	16,400	21,700	21,600	39,700[4]	61,200[4]	54,600	0	168,900	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A

1a- Includes the total of simulated streamflow infiltration; all of which is derived from discharge of groundwater to streams (simulated using the MODFLOW Streams Package (STR)) within the groundwater basin and in upgradient groundwater basins.

1b- Includes recharge resulting from excess applied irrigation water derived from surface water and groundwater supplies, discharge from waste-water treatment facilites, and golf courses.

1c- Includes components of groundwater inflow from multiple adjacent basins and sub-basins.

1d- Represents the sum of inflow components including groundwater flow from adjacent basins and reinfiltration and recharge of groundwater discharge to streams in the groundwater basin.

1e- Includes only discharge to streams simulated using the MODFLOW Streams Package (STR).

1f- Streams and springs simulated using the MODFLOW Drain Package (DRN).

1g- Evapotranspiration was simulated in only the Verde River Basin above the streamflow-gaging station near Camp Verde and in the Little Colorado River Basin.

1h- Includes components of groundwater outflow to multiple adjacent basins and sub-basins.

1i- Calculated as Total Inlow minus Total Outflow. Multiple areas of groundwater inflow and outflow may occur for any basin.

1j- Calculated as Groundwater Infow minus Groundwater Outflow.

1k- Net streamflow is equivalent to stream base flow simulated as exiting the basin. Calculated as Recharge from infiltration of streamflow derived from base flow minus Groundwater discharge to streams. Both components are simulated using the Streams Package (STR).

1l- Includes the sum of groundwater budget components for the Burro Creek and Fort Rock sub-basins and the parts of the Peach Springs Basin and Wikieup sub-basin within the Truxton Wash watershed.

2- Includes groundwater outflow of about 2,100 acre-feet per yr to the Big Chino sub-basin and inflow from portions of the Big Chino (about 1,000 acre-feet per yr) and Upper Agua Fria sub-basins.

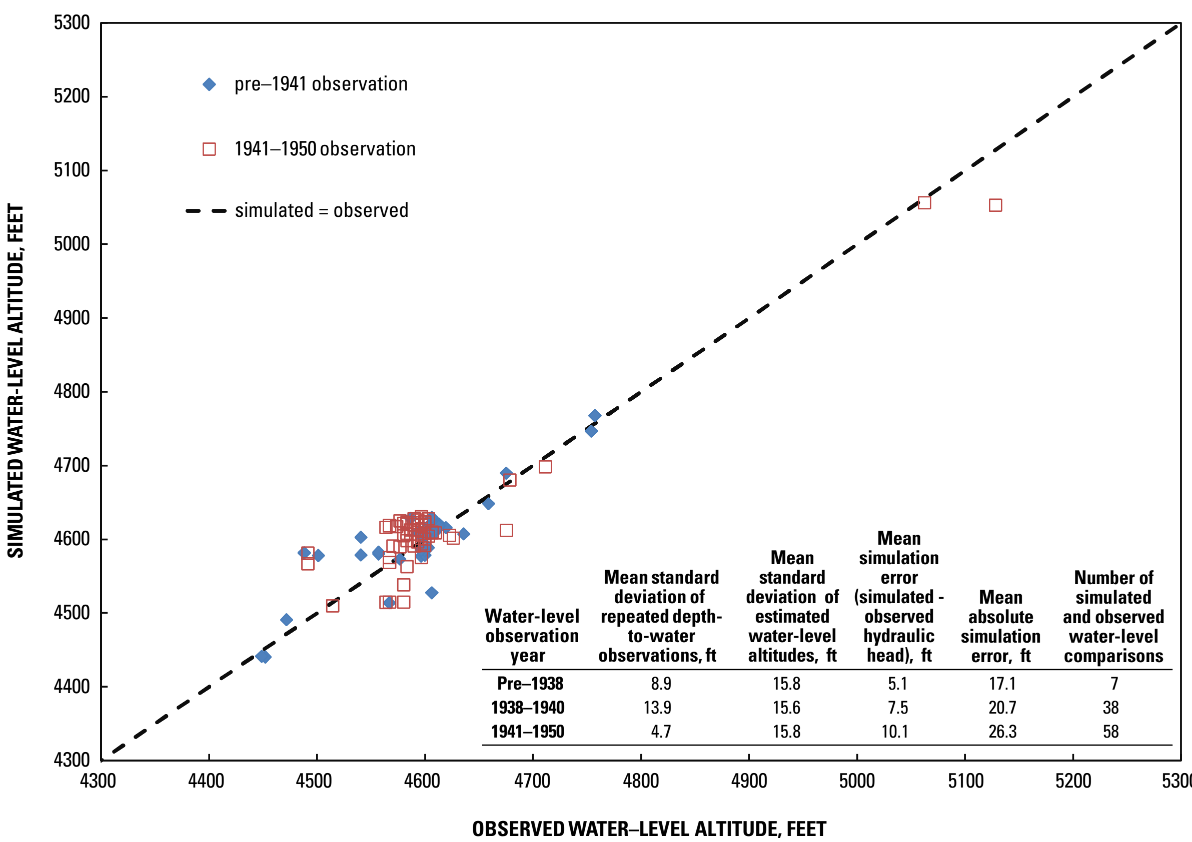
3- Includes the balance of outflow of groundwater to adjacent basins and inflow from adjacent basins, primarily from the Little Chino sub-basin.

4- Does not include diversion of 10,200 acre-feet per year of base flow that is transpired by crops.

simulated less natural recharge than previous models because the average annual amount of water available for recharge—the sum of average annual recharge and runoff that was calculated by the BCM—was less than rates of natural recharge that were used in previous models. Neither the previous nor the BCM estimates of recharge may be accurate, but the range of the recharge estimates indicate that natural recharge rates are not well defined.

Groundwater outflow from the Little Chino and Upper Agua Fria sub-basins occurred through discharge to springs, streams, ET, and groundwater flow before initial well withdrawals. Estimates of groundwater discharge to Del Rio Springs of about 2,300–3,400 ac-ft/yr were measured during 1940–45 (Schwalen, 1967). Estimates of base flow for the Upper Agua Fria River near Humbolt during 1940 are reported as about 2,000 ac-ft/yr by the ADWR (Corkhill and Mason, 1995). Additional steady-state discharge through ET is estimated for areas near Del Rio Springs, 100–200 ac-ft/yr, and the perennial reach of the Upper Agua Fria River, 300 ac-ft/yr (Nelson, 2002). Groundwater discharge is reported as groundwater underflow from the Little Chino sub-basin at estimated rates of 2,000–5,600 ac-ft/yr (Corkhill and Mason, 1995; Nelson, 2002; Salt River Project (SRP), 2000, written commun. to ADWR).

Simulated rates of steady-state groundwater outflow from the Little Chino and Upper Agua Fria sub-basins were similar to estimated rates. Steady-state base flow at Del Rio Springs was simulated as net streamflow of about 5,300 ac-ft/yr (table 1). Steady-state base flow to the Upper Agua Fria River was simulated as about 1,000 ac-ft/yr of discharge to drains (table 1). Groundwater discharge to ET was simulated as about 100 ac-ft/yr near Del Rio Springs in the Little Chino sub-basin and 200 ac-ft/yr along the perennial reach of the Upper Agua Fria River near Humbolt (table 1). Simulated groundwater discharge from the Upper Agua Fria sub-basin occurred as about 800 ac-ft/yr of underflow to the Little Chino sub-basin. Simulated groundwater discharge from the Little Chino sub-basin occurred near Del Rio Springs as about 1,100 ac-ft/yr of groundwater underflow through the limestone of layer 3 and lower volcanic unit of layer 2. The difference in simulated and estimated base flow at Del Rio Springs may be partly caused by groundwater withdrawals near Del Rio Springs at about the time of the observations. Variations in springflow with withdrawals from



Water-level observation year	Mean standard deviation of repeated depth-to-water observations, ft	Mean standard deviation of estimated water-level altitudes, ft	Mean simulation error (simulated - observed hydraulic head), ft	Mean absolute simulation error, ft	Number of simulated and observed water-level comparisons
Pre–1938	8.9	15.8	5.1	17.1	7
1938–1940	13.9	15.6	7.5	20.7	38
1941–1950	4.7	15.8	10.1	26.3	58

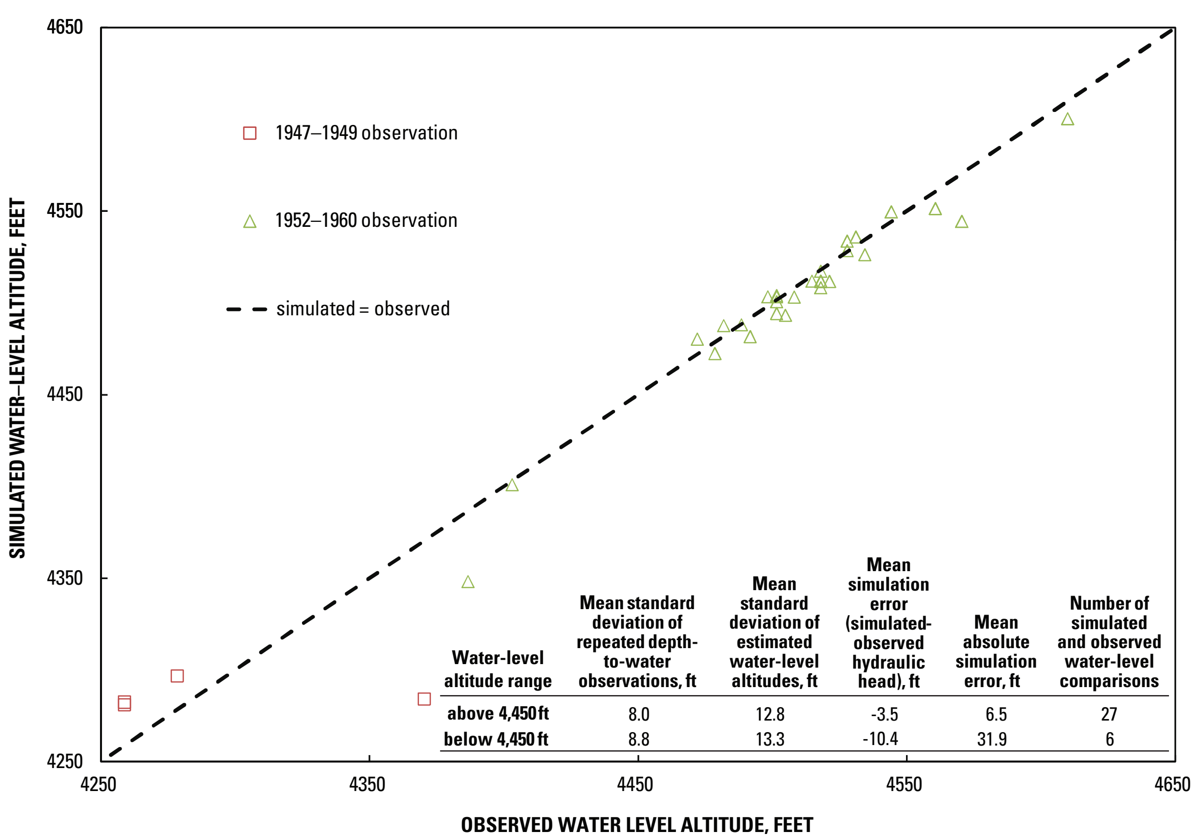
Figure 18. Simulated and observed predevelopment hydraulic head at wells in the Little Chino and Upper Agua Fria sub-basins.

the Sante Fe Railroad wells near the springs were documented by Schwalen (1967). Local groundwater withdrawals from the Santa Fe Railroad wells and other wells may have contributed to the variability of spring discharge that was observed during 1940–44. Simulated steady-state groundwater budgets deviate from estimated water budgets, but not substantially more than is likely because of uncertainties in the estimated values. Simulation of discharge to Del Rio Springs that is near the greatest range of measured rates is not surprising because nearby groundwater withdrawals may have reduced spring discharge and resulted in low rates of observed discharge.

Some differences in ADWR groundwater sub-basins and simulated groundwater flow divides occur in the region of the Little Chino and Upper Agua Fria sub-basins. The simulated groundwater divide between the Little Chino and Upper Agua Fria sub-basins lies a few miles southeast of the groundwater basin divide and slightly southeast of the divide simulated by previous models (fig. 17). The simulated western flow divide of the Little Chino sub-basin includes much of the Granite Mountain and upper part of Mint Wash areas. The simulated flow divide in the Black Hills includes a slightly greater area for the Little Chino sub-basin than defined by the ADWR groundwater sub-basin boundary. The net effect of the difference in administratively defined and simulated groundwater basins for the Little Chino and Upper Agua Fria sub-basins is that simulated recharge areas for the sub-basins include a larger area than defined by the administrative groundwater sub-basins, especially for the Little Chino sub-basin. The simulated recharge area contributing to the Upper Agua Fria sub-basin is less than defined by the administrative groundwater sub-basin. Few early water-level data are available to verify the location of the steady-state groundwater divide between the two sub-basins.

Big Chino Sub-Basin

Few data are available to define the steady-state groundwater-flow system in the Big Chino sub-basin. Early water-level records are available beginning in the late 1940s after initiation of agricultural withdrawals. The first streamflow records at Paulden began in 1963, long after the beginning of substantial withdrawals. Steady-state conditions of hydraulic head and groundwater discharge must, therefore, be estimated on the basis of the earliest



Water-level altitude range	Mean standard deviation of repeated depth-to-water observations, ft	Mean standard deviation of estimated water-level altitudes, ft	Mean simulation error (simulated-observed hydraulic head), ft	Mean absolute simulation error, ft	Number of simulated and observed water-level comparisons
above 4,450 ft	8.0	12.8	-3.5	6.5	27
below 4,450 ft	8.8	13.3	-10.4	31.9	6

Figure 19. Simulated and observed predevelopment hydraulic head at wells in the Big Chino sub-basin.

available data. Steady-state water-level altitudes can only be estimated as greater than or equivalent to the earliest measured water-level altitudes. Water levels at four wells are available during the late-1940s (fig. 17). Water levels collected during 1952–60 are available at 29 wells. Other useful information that helps define steady-state hydraulic head includes perennial streamflow and riparian conditions along part of Williamson Valley Wash and Big Chino Wash near the confluence of Pine Creek (Wirt and others, 2005). Steady-state groundwater discharge through streamflow was likely greater than or equivalent to the earliest estimated base flow at the streamflow-gaging station near Paulden because the record began following the worst drought on record and decades after the earliest nearby agricultural withdrawals in the Paulden and Little Chino sub-basin areas.

Simulated water-level altitudes compare favorably with 27 observed values above an altitude of about 4,450 ft—the altitude of water levels in Williamson Valley and the upper part of Big Chino sub-basin—and less favorably with the 6 observed water levels below an altitude of 4,450 ft in the lower part of the Big Chino sub-basin (fig. 19). The average error of simulated water levels in the Williamson Valley and the upper part of the Big Chino sub-basin was -3.5 ft, with an average absolute error of 6.5 ft. The standard deviation of estimated pre-1961 water-level altitudes was about 12.8 ft. Observed saturated conditions were simulated at the land surface near Williamson Valley Wash and along Big Chino Wash near the confluence with Pine Creek during predevelopment. The average error of pre-1961 simulated water levels in the lower part of the Big Chino sub-basin was -10.4 ft, with an average absolute error of about 31.9 ft. Standard deviation of the estimated water-level altitude for the pre-1961 observed water-level altitudes in the lower part of the Big Chino sub-basin was about 13.3 ft. Simulated water levels at three wells in the lower Big Chino sub-basin were 18 to 24 ft above observed water levels. Part of the simulation error at the three wells can be attributed to pre-observation water-level decline. Two other pre-1961 estimated water-level altitudes in the lower part of the sub-basin were simulated as lower than observed altitudes. The simulation error at the two wells can be attributed to observed water levels in the wells that may represent elevated water levels in a shallow aquifer that overlies the thick silt and clay layer in the area and was not simulated. Water-level altitudes in the shallow system are as great as 100 ft above deeper water-level altitudes. In addition, at least three aquifers are present in the region and the aquifer that is represented by water levels at each well is uncertain. Simulated vertical head-differences among layers varied from differences of about 5 ft that indicate downward hydraulic gradients to differences of about 30 ft that indicate upward hydraulic gradients. Accurate simulation of steady-state conditions in the lower part of the Big Chino sub-basin also is compromised by a lack of early water-level data in the area immediately upstream from the headwater springs of the Verde River. The lack of early water-level data in the region results in uncertain definition of the steady-state hydraulic gradients and transmissivity. Considering the complicated hydrogeology defined by sparse water-level data, this simulation of steady-state groundwater conditions in the lower Big Chino sub-basin represents a generalized approximation of the groundwater-flow system that compares well with available water-level data.

Rates of simulated steady-state natural recharge in the ADWR Big Chino sub-basin were about 41,600 ac-ft/yr (table 1). Additional inflow includes streamflow infiltration of discharge to Del Rio Springs, about 5,300 ac-ft/yr of the 6,500 ac-ft/yr of streamflow infiltration in the sub-basin (table 1), and net groundwater inflow of about 2,700 ac-ft/yr from adjacent basins (table 1) including the balance of groundwater flow to and from the Little Chino sub-basin, about 200 ac-ft/yr, and groundwater flow of about 2,500 ac-ft/yr from adjacent areas northwest of the sub-basin. Total simulated steady-state inflow to the Big Chino sub-basin was about 50,800 ac-ft/yr (table 1). However, the part of streamflow infiltration that was simulated along Williamson Valley Wash, about 1,200 ac-ft/yr, was derived from discharge of groundwater that recharged within the Big Chino sub-basin and does not represent additional water to the groundwater system. Therefore, the total simulated inflow that does not include water that has recirculated through the surface water system was about 48,600 ac-ft/yr.

Groundwater that originates as recharge within the Big Chino sub-basin exits the sub-basin as groundwater flow across the sub-basin boundaries or discharge to streams and ET within the sub-basin. Total groundwater outflow to adjacent areas before development, about 25,800 ac-ft/yr, includes outflow to three different regions. A large region of the sub-basin contributes about 16,200 ac-ft/yr to the groundwater-flow system in the Coconino Plateau basin. Smaller regions contribute about 2,600 ac-ft/yr of groundwater flow to the Burro Creek and Fort Rock sub-basins. About 7,000 ac-ft/yr exits the sub-basin as groundwater flow across the common boundary with the Verde Valley sub-basin. Groundwater underflow from the Big Chino sub-basin also occurs to the Little Chino sub-basin in the Granite Mountain area at a rate of about 1,000 ac-ft/yr. However, nearly all of groundwater discharge to the Little Chino sub-basin re-enters the Big Chino sub-basin as surface flow below Del Rio Springs or as groundwater flow near Del Rio Springs; a small part is lost to ET near the springs. Simulated total steady-state discharge to streams, including Verde River headwater springs, Verde River upstream from the streamflow-gaging station near Paulden, and Williamson Valley Wash, was about 22,800 ac-ft/yr (table 1). Groundwater discharge to Williamson Valley Wash contributed about 400 ac-ft/yr, which reinfiltrated along downstream parts of the wash. Simulated rates of groundwater losses to ET were about 2,200 ac-ft/yr within the Big Chino sub-basin.

Simulated steady-state groundwater discharge to the Verde River above the streamflow-gaging station near Paulden is represented by the net streamflow for the simulated groundwater-flow systems in the Little Chino and Big Chino sub-basins, about 21,700 ac-ft/yr (table 1). ET rates of about 2,300 ac-ft/yr in the two sub-basins were simulated primarily near Del Rio Springs, along Williamson Valley Wash, and along the Verde River.

Substantial differences occur between the Big Chino groundwater sub-basin boundary defined by the ADWR and the simulated groundwater flow divide (fig. 17). The greatest difference has been previously observed in the region north of Big Black Mesa, where groundwater flows northward across the ADWR boundary of the Coconino Plateau basin and toward discharge areas near the Colorado River. Other substantial differences are along the north boundary, west boundary, and the southeast boundary near Granite Mountain and Mint Wash. The net effect of the difference between simulated and previously defined boundaries is a smaller region that contributes recharge and groundwater flow to the Big Chino sub-basin. Few water-level data are available to verify the location of the steady-state groundwater divide along the north and west boundaries of the Big Chino sub-basin.

Verde Valley Sub-Basin

Few data are available to define the steady-state groundwater-flow system in the Verde Valley sub-basin prior to surface-water irrigation of agriculture along the Verde River and tributary streams. Therefore, steady-state conditions are assumed to have occurred after surface-water irrigation had been ongoing for several decades but before the initiation of extensive groundwater withdrawals. The earliest available information that describes the steady-state groundwater system includes water levels in wells before 1961 and estimates of base-flow discharge at several streamflow-gaging stations for a large range of time periods. Water-level data are available at 62 wells before 1961, including 21 water levels in wells in the Coconino aquifer and two water levels in wells in the Redwall-Muav aquifer. The earliest available streamflow data include records for the Verde River near Clarkdale beginning in 1915, Verde River near Camp Verde beginning in 1934, Oak Creek near Cornville beginning in 1941, Wet Beaver Creek near Rimrock beginning in 1963, West Clear Creek near Camp Verde beginning in 1966, and Oak Creek near Sedona beginning in 1982. Estimated groundwater discharge for the earliest period at these streamflow gages may represent steady-state conditions for the period of surface-water diversions and irrigation because groundwater use upstream from each streamflow gage was minimal. However, groundwater discharge upstream from these streamflow gages is highly variable and likely sensitive to variations in recharge rates (Blasch and others, 2006). Therefore, the period of record and potential variations in recharge must be considered in the evaluation of groundwater discharge at these streamflow-gaging stations as estimates of steady-state conditions.

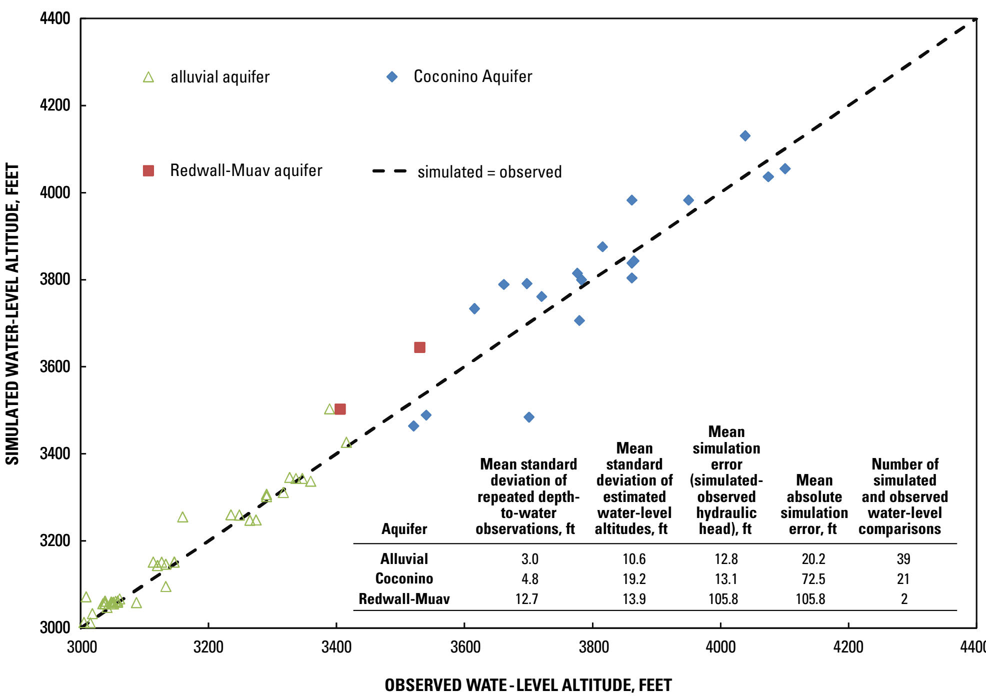
The most effective streamflow gage for estimating base flow in the Verde Valley sub-basin is arguably the Verde River near Camp Verde. Steady-state base flow estimates at this streamflow gage are also an estimate of total groundwater budgets and recharge rates for the groundwater-flow system upstream from the gage. Unfortunately, estimates of steady-state base flow for the Verde River near Camp Verde are uncertain because of the influence of seasonal surface-water diversions for the irrigation of crops that transpire water at the rate of about 10,200 ac-ft/yr near the Verde River. Natural base flow would typically change seasonally with variations in riparian ET rates from the groundwater system. After the advent of surface-water diversions for agricultural irrigation, base flow also began to vary with the diversions and with irrigation-caused variations in discharge from the shallow groundwater-flow system to the stream. The part of diversions that are applied to crops in excess of crop requirements infiltrates beneath the canals and fields, where the water is temporarily stored in the unsaturated zone and shallow aquifers before slowly returning to the stream as groundwater discharge. The process of infiltration, recharge, and groundwater discharge to the stream takes an unknown amount of time, but likely enhances the seasonal base flow variations at the streamflow-gaging station near Camp Verde. Base flow at the streamflow gage varies seasonally from about 134,000 ac-ft/yr during winter to about 45,000 ac-ft/yr during summer on the basis of records during 1936–45. Winter base flow is augmented by discharge of excess irrigation water that was temporarily stored in the subsurface before returning to the stream as groundwater discharge. Winter base-flow estimates are, therefore, an overestimate of pre-agriculture groundwater discharge. Summer base-flow estimates, however, are an underestimate of base flow because of ET losses through crops and riparian vegetation and the temporary subsurface storage of excess irrigation water. The best estimate of groundwater discharge to the Verde River upstream from the streamflow gage near Camp Verde is between the winter and summer estimates.

A better estimate of groundwater discharge to the Verde River upstream from the streamflow-gaging station near Camp Verde could be obtained through a detailed analysis of irrigation system losses, temporary subsurface storage of excess irrigation water, and return of the temporarily stored water through delayed discharge to the stream. The analysis would require irrigation records, seepage runs along the stream (a few of which exist), and seasonal records of water levels in the shallow aquifer in several agricultural areas. These data generally do not exist and would require a concerted monitoring effort. Such an effort also would allow improvements to the simulation of groundwater flow in the Verde Valley through simulation of the seasonal irrigation and shallow groundwater processes. In lieu of a detailed analysis of irrigation system losses and return of excess irrigation water, steady-state base flow upstream from the streamflow-gaging station near Camp Verde was estimated as about 98,200 ac-ft/yr. Average annual minimum monthly streamflow during the early record, 1936–45, which likely includes some surface water runoff, probably overestimates rates of groundwater discharge by a small amount.

A simplified method was employed in the numerical groundwater-flow model to simulate the poorly defined and complex process of diversion, ET, infiltration, recharge, and returns to the stream through ditches and groundwater discharge. The simplified method included the simulation of net losses through agricultural ET at a rate of about 10,200 ac-ft/yr as diversions from the streams at the head of each of nine ditchs along the Verde River and one ditch near the Verde River along West Clear Creek.

Simulated steady-state water-level altitudes in the Verde Valley area generally compare favorably to observed water levels; however, simulated errors are substantially greater than estimated measurement errors at many wells (fig. 20). Much of the error is attributed to substantial and poorly defined vertical hydraulic gradients in the major aquifers throughout the groundwater-flow system in the Verde Valley. Simulated water-level altitudes at 39 wells in the alluvial aquifer, which were near the major streams, averaged 12.8 ft above the observed water-level altitude with an average absolute error of about 20.2 ft in comparison to an estimated altitude observation error of about 10.6 ft. Steady-state water-level data are sparse between major streams. Simulated water-level altitudes at 21 wells in the Coconino aquifer averaged about 13.1 ft above the observed water-level altitude with an average absolute error of about 72.5 ft in comparison to a estimated water-level altitude error of about 19.2 ft. Most of the Coconino aquifer wells are concentrated in the Oak Creek area where large vertical hydraulic gradients are common and the greatest simulation error occurred. Simulated water-level altitudes at two wells in the Redwall-Muav aquifer were 97 and 114 ft above the observed water-level altitude, which had an estimated observation error of about 13.9 ft. Both wells in the Redwall-Muav aquifer are in a region of steep simulated hydraulic gradients where slightly inaccurate well locations could result in large differences in simulated hydraulic head.

Potential flowing well conditions were simulated under steady-state conditions in several areas of the Verde Valley (fig. 17). Hydraulic heads in layer 2 were simulated as above land surface in several areas near the Verde River, Oak Creek, and Wet Beaver Creek. Hydraulic heads in layer 2 were simulated as more than 50 ft above land surface north of Cottonwood and near Page Springs. Hydraulic heads in layer 2 were simulated as 0–50 ft above land surface in several areas near the Verde River between Cottonwood and the confluence with Oak Creek, in the Camp Verde area, and near the lower parts of Wet Beaver Creek. Other potential flowing well conditions were simulated under steady-state conditions in the upper reaches of Sycamore Creek, Oak Creek, Dry Beaver Creek, and West Clear Creek.



Aquifer	Mean standard deviation of repeated depth-to-water observations, ft	Mean standard deviation of estimated water-level altitudes, ft	Mean simulation error (simulated-observed hydraulic head), ft	Mean absolute simulation error, ft	Number of simulated and observed water-level comparisons
Alluvial	3.0	10.6	12.8	20.2	39
Coconino	4.8	19.2	13.1	72.5	21
Redwall-Muav	12.7	13.9	105.8	105.8	2

Figure 20. Simulated and observed predevelopment hydraulic head at wells in the Verde Valley sub-basin.

Rates of simulated inflow to the steady-state groundwater-flow system in the Verde Valley sub-basin include natural recharge of about 91,400 ac-ft/yr, stream base flow of about 21,700 ac-ft/yr contributed from the region above the gaging station near Paulden, and about 8,300 ac-ft/yr of groundwater flow from adjacent areas (table 1) including about 6,000 ac-ft/yr contributed from the Little Chino and Big Chino sub-basins. The BCM estimated recharge rate includes ephemeral channel recharge of about 1.8 x 10^{-2} ft/yr, a rate similar to ephemeral channel recharge rates applied to the alluvial surface in the Big and Little Chino sub-basins, in areas of non-Verde Formation alluvium and sedimentary rock, primarily where layer 2 of the model is at the land surface. The total additional ephemeral channel recharge was 30,100 ac-ft/yr.

Simulated steady-state groundwater discharge from the Verde Valley sub-basin was through discharge to several streams, springs, ET, a small amount of groundwater withdrawal (2 ac-ft/yr), and groundwater underflow to adjacent basins. Net steady-state groundwater discharge to the Verde River downstream from the streamflow-gaging station near Paulden and along tributary streams including Sycamore Creek, Oak Creek, Wet Beaver Creek, and West Clear Creek was simulated as about 61,200 ac-ft/yr using the STR Package (table 1). However, only 26,500 ac-ft/yr net groundwater discharge to the stream in the Verde Valley sub-basin was simulated as flowing out of the sub-basin because about 10,200 ac-ft/yr of diverted stream base flow was simulated as lost through crop use. Major springs adjacent to the tributary streams were implicitly simulated as part of the stream network, including Page Spring near Oak Creek and Montezuma Well near Wet Beaver Creek. Groundwater discharge to Haskell Spring near Cottonwood was simulated as about 100 ac-ft/yr by using the DRN Package. Total steady-state discharge through ET in the Verde Valley was simulated at a rate of about 12,100 ac-ft/yr along the major perennial stream reaches (table 1), which is greater than the estimated rate of 10,800 ac-ft/yr (Blasch and others, 2006). Simulated steady-state groundwater discharge to adjacent basins was about 12,400 ac-ft/yr, most of which discharged to the Verde Canyon sub-basin, about 11,700 ac-ft/yr.

Simulated steady-state groundwater discharge to streams within the Verde Valley sub-basin was about 96,500 ac-ft/yr (table 1). However, about 35,400 ac-ft/yr of the groundwater discharge reinfiltrated to the aquifer within the sub-basin and about 10,200 ac-ft/yr of diverted stream base flow was simulated as lost through crop use (table 1). The net steady-state groundwater discharge to streams in the Verde Valley sub-basin was about 61,200 ac-ft/yr. Simulated steady-state groundwater discharge to the Verde River upstream from the streamflow-gaging station near Clarkdale was about 43,100 ac-ft/yr, including about 21,700 ac-ft/yr of groundwater inflow from the Little Chino and Big Chino sub-basins (table 1) and about 21,400 ac-ft/yr of groundwater discharge between the streamflow-gaging stations near Paulden and Clarkdale (table 1). In comparison, the annual average of minimum monthly flows at the Clarkdale gage that do not include substantial runoff from precipitation was about 56,600 ac-ft/yr during 1915–1921. Simulated steady-state groundwater discharge to the Verde River upstream from the streamflow-gaging station near Camp Verde was about 82,800 ac-ft/yr, including about 21,700 ac-ft/yr contributed from the Little Chino and Big Chino sub-basins, 21,400 ac-ft/yr contributed between the Paulden and Clarkdale gages, and 39,700 ac-ft/yr contributed between the Clarkdale and Camp Verde gages (table 1). The simulated steady-state base flow at the streamflow-gaging station near Camp Verde is 48,400 ac-ft/yr after accounting for crop use of diverted base flow. In comparison, the annual average of minimum monthly non-flood flows was about 98,600 ac-ft/yr for available record during 1934–1945. The difference in simulated and estimated base flow at the streamflow-gaging station near Camp Verde indicates that either the ephemeral channel infiltration-enhanced BCM recharge rates are underestimated for the Verde Valley or that the surface-water diversion and agricultural irrigation system has enhanced infiltration of streamflow at a rate that is comparable to rates of crop transpiration. A more thorough simulation of the diversion, irrigation, and recharge to the shallow aquifer may help understand changes in the groundwater budget caused by irrigation practices.

Steady-state net groundwater discharge upstream from the streamflow-gaging stations along major streams that are tributary to the Verde River between the streamflow-gaging stations at Paulden and Camp Verde was simulated as about 42,700 ac-ft/yr in comparison with estimated base flow of about 45,600 ac-ft/yr upstream from the gages. Net simulated groundwater discharge is the difference of simulated discharge from the groundwater system to streams and streamflow losses to the groundwater system of about 53,700 and 11,000 ac-ft/yr, respectively. Simulated groundwater discharge was similar to estimated values at the tributary streamflow-gaging stations. Simulated net groundwater discharge upstream from the streamflow-gaging station on Oak Creek near Sedona was about 14,300 ac-ft/yr in comparison with an estimated base flow of about 13,800 ac-ft/yr. Simulated net groundwater discharge upstream from the streamflow-gaging station on Oak Creek near Cornville, including Page Spring, was about 8,500 ac-ft/yr in comparison with an estimated base flow of about 20,500 ac-ft/yr. Simulated net groundwater discharge upstream from the streamflow-gaging station on Wet Beaver Creek near Rimrock was about 1,500 ac-ft/yr in comparison with estimated base flow of about 5,400 ac-ft/yr. A better representation of simulated groundwater discharge at the Cornville and Rimrock streamflow-gaging stations likely requires greater aquifer anisotropy including barriers and conduits that were not used in this simulation. Simulated net groundwater discharge upstream from the streamflow-gaging station on West Clear Creek near Camp Verde was about 10,900 ac-ft/yr in comparison with estimated base flow of about 11,800 ac-ft/yr. Simulated net groundwater discharge to Sycamore Creek was about 17,800 ac-ft/yr in comparison with an estimated discharge rate of about 8,000 ac-ft/yr along a part of the stream discharging from Sycamore, Lower Parson, and Summer Springs (Bills, 2006). In addition to streams, steady-state groundwater discharge to springs that were simulated as drains included about 100 ac-ft/yr at Haskell Spring (table 1).

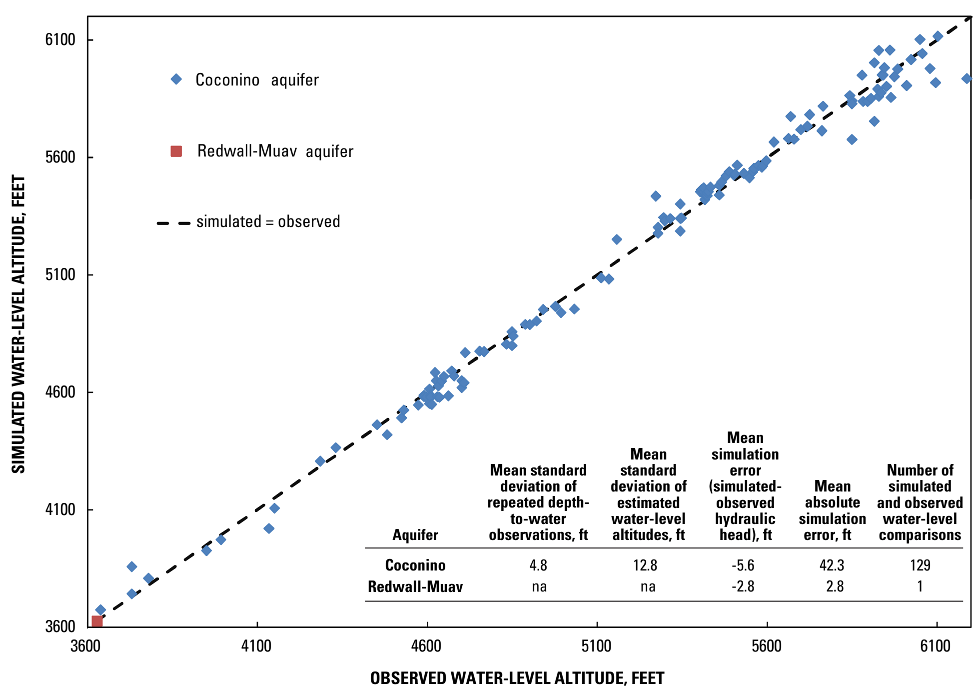
Major differences occur between the Verde Valley groundwater sub-basin boundary defined by the ADWR and the simulated groundwater flow divide (fig. 17). The greatest difference was in the upper part of the Oak Creek and Sycamore Creek drainage basins where recharge contributes to groundwater that flows toward the north and away from the Verde Valley. Available water-level data are consistent with the simulated groundwater divide in this area. A region in the upper parts of the Wet Beaver and West Clear Creek drainage basins also contributes to groundwater flow to the Little Colorado River Plateau basin and away from the Verde Valley. A small region of the Little Colorado River Plateau basin northwest of Mormon Mountain contributes flow to the Verde Valley groundwater system**.** Few water-level data are available to verify the location of the steady-state groundwater divide in the region of the upper parts of the Little Colorado River Plateau basin. The net effect of the difference between simulated and previously defined boundaries is a smaller simulated region that contributes recharge and groundwater flow to the Verde Valley.

Little Colorado River Plateau Basin

The steady-state groundwater altitude distributions in the Little Colorado River Plateau basin are defined by water levels measured before 1961 at 129 wells that tap the Coconino aquifer and one well that taps the Redwall-Muav aquifer (figs. 16, 21). Some areas of the basin were lacking early water-level records, and later water-level observations were used to help define the flow system in those areas. Estimates of steady-state groundwater discharge from the basin are based on Little Colorado River records of streamflow at the gaging station near Cameron that begin in 1948 and the streamflow records at the gaging station near Desert View beginning in 1990. Estimates of steady-state groundwater discharge are also available at Clear Creek and Chevelon Creek (Leake and others, 2005).

Evaluation of the simulated groundwater-flow system is focused on the region that lies between the Mogollon Rim, White Mountains, and areas near the Little Colorado River. The region north of the Little Colorado River has little control data except in the region of the Defiance Uplift. The flow system north of the Little Colorado River was simulated with the goal of reproducing the general distribution of hydraulic head and groundwater flow. Extremely low values of hydraulic conductivity were required in each of the layers in the region of the Defiance Uplift to reproduce locally steep hydraulic gradients and may indicate that the simulated flow system may not accurately represent the groundwater-flow system in this region.

Simulated steady-state water-level altitudes in the Little Colorado River Plateau basin were comparable to observed water levels. Simulated errors are substantially greater than estimated measurement errors at many wells, however, especially at water-level altitudes greater than 6,000 ft that are dominated by a large number of observations in the area of



Aquifer	Mean standard deviation of repeated depth-to-water observations, ft	Mean standard deviation of estimated water-level altitudes, ft	Mean simulation error (simulated-observed hydraulic head), ft	Mean absolute simulation error, ft	Number of simulated and observed water-level comparisons
Coconino	4.8	12.8	-5.6	42.3	129
Redwall-Muav	na	na	-2.8	2.8	1

Figure 21. Simulated and observed predevelopment hydraulic head at wells in the Little Colorado River Plateau basin.

the Defiance Uplift (fig. 21). Much of the error is attributed to large anisotropy and locally steep vertical hydraulic gradients in the Coconino aquifer that could not be simulated accurately without greater anisotropy, by using a finer grid resolution, or barriers and conduits for groundwater flow. Much anisotropy is simulated, however, without the use of barriers and conduits. Simulated water-level altitudes at the 129 wells in the Coconino aquifer averaged 4.8 ft below the observed water-level altitude with an average absolute error of about 42.3 ft in comparison to an average observation altitude error of about 12.8 ft (average repeated depth to water measurement error of 4.8 ft and altitude error of 10–20 ft) . The simulated water level at the single well that taps the Redwall-Muav aquifer northeast of the confluence of the Little Colorado and Colorado Rivers was within about 3 ft of the observed value.

Rates of simulated inflow to the steady-state groundwater-flow system in the Little Colorado River Plateau basin included recharge of about 206,600 ac-ft/yr and groundwater underflow from adjacent basins of about 2,600 ac-ft/yr (table 1). Inflow also occurs through infiltration of streamflow—about 12,600 ac-ft/yr; however, the infiltrated streamflow is water that discharged from the groundwater-flow system to the stream in upstream areas of the basin and is not additional input to the system. Steady-state discharge from the basin was simulated as groundwater discharge to streams using the STR Package, ET (less than 100 ac-ft/yr), and groundwater flow to adjacent basins. Steady-state discharge to streams simulated by using STR of about 12,600 ac-ft/yr was primarily along the Little Colorado River and Clear, Chevelon, and Silver Creeks (table 1). The entire simulated discharge to streams was simulated as reinfiltration along stream channels in the basin. Discharge to streams and springs simulated by using the DRN Package in the areas of the Defiance Uplift and White Mountains was about 400 ac-ft/yr and was lost from the flow system. The remainder of discharge from the steady-state groundwater system, about 208,900 ac-ft/yr, was primarily simulated as groundwater flow to the Coconino Plateau basin (table 1), about 164,600 ac-ft/yr, and groundwater flow to Verde and Salt River groundwater-flow systems—about 21,200 ac-ft/yr and about 23,100 ac-ft/yr, respectively.

Simulated steady-state groundwater discharge from the Little Colorado River Plateau basin to the Coconino Plateau basin is similar to the annual average of monthly minimum flows at the streamflow-gaging station at the Little Colorado River near Desert View, about 163,400 ac-ft/yr, which likely represents most of the discharge from the the Little Colorado River Plateau basin. Simulated steady-state discharge to Clear and Chevelon Creeks upstream from the confluence of each stream with the Little Colorado River was about 3,700 and 3,200 ac-ft/yr, respectively, in comparison with estimated summer base flow of about 4,100 and 2,000 ac-ft/yr, respectively (Leake and others, 2005). Simulated steady-state discharge to Silver Creek upstream from the confluence with the Little Colorado River was about 800 ac-ft/yr. No streamflow records or base-flow estimates are available for Silver Creek.

Differences in the boundaries of the simulated steady-state groundwater-flow system and the defined ADWR groundwater basin occur along the northwest basin boundary and in the Mogollon Rim region. The northwest basin boundary does not conform to a simulated groundwater-flow divide except near Flagstaff. Nearly all of the groundwater in the Little Colorado River Plateau basin discharges as groundwater flow across the northwest basin boundary near Cameron and between Cameron and the north model boundary. Near the Mogollon Rim the simulated and defined basin boundaries differ by less than 10 mi at three locations—near the eastern extent of the Verde Valley sub-basin, south of Snowflake, and south of Springerville (fig. 16). This region includes the greatest rate of recharge in the basin, and slight variations in the basin extent can result in substantial variations in recharge rates to the basin.

Coconino Plateau Basin

The Coconino Plateau basin defined by the ADWR includes the primary groundwater discharge area for the Little Colorado River Plateau basin near Blue Spring (fig. 16). The steady-state groundwater-flow system in the Coconino Plateau basin and nearby areas is defined by a minimal amount of data including post-1961 water levels in 19 wells that tap the Redwall-Muav aquifer or Tapeats sandstone (fig. 22) and estimates of groundwater discharge from estimates of flow at many springs (NWIS-GWSI; Johnson and Sanderson, 1968). A few Coconino aquifer water levels also are available; however, that system is not simulated because the aquifer is thin and saturated only locally in the region. Groundwater discharge from the springs in the Grand Canyon region include the largest spring, Havasu Spring (about 45,800 ac-ft/yr), and several smaller springs including Diamond (about 1,500 ac-ft/yr), Hermit (about 600 ac-ft/yr), Pumphouse (about 700 ac-ft/yr), Warm (about 500 ac-ft/yr), and several smaller springs that total about 2,000 ac-ft/yr. Total estimated spring discharge of about 51,200 ac-ft/yr represents only a part of the total discharge. Groundwater likely discharges to the Colorado River where the river is hydraulically connected to the Redwall-Muav aquifer. Groundwater discharge to the Colorado River, however, has not been measured or estimated. Total spring discharge is, therefore, considered a minimum estimate of groundwater discharge from the Coconino Plateau basin aquifer system.

Simulated steady-state water-level altitudes in the Coconino Plateau area generally compare favorably with observed water-levels; however, errors in the simulation are subtantially greater than estimated measurement errors at most wells (fig. 22). Much of the error is attributed to poorly defined but substantial vertical and horizontal anisotropy in the aquifer system that results in locally steep hydraulic gradients and large differences in hydraulic head between wells that are finished at different depths. An example of large vertical hydraulic gradients may occur in the Ash Fork area where the water-level altitude in deep well (B-22-01)30ADA (2,624 ft deep) is about 600 ft lower than at three wells—(B-22-02)12DAB, (B-22-02)14BCC, and (B-22-02)15BAD—that are less deep (1,300 to 1,700 ft). Simulation of this steep gradient is not possible with the single layer that represents the aquifer system in this numerical model. Simulated water-level altitudes at the 19 wells averaged 17.9 ft below the observed water-level altitude with an average absolute error of about 70 ft in comparison to a estimated observation altitude error of about 23 ft (repeated depth-to-water standard deviation of 16 ft and altitude error of 5 to 50 ft).

Simulated inflow to the steady-state groundwater-flow system in the Coconino Plateau basin included recharge, streamflow infiltration, and groundwater underflow from adjacent basins (table 1). Rates of simulated steady-state natural recharge to the Coconino Plateau basin were about 152,000 ac-ft/yr. Simulated streamflow infiltration of about 300 ac-ft/yr occurred along the lower part of the Little Colorado River. Simulated groundwater inflow of about 192,300 ac-ft/yr was primarily derived from the Little Colorado River Plateau basin, about 162,300 ac-ft/yr, and the remainder was primarily derived from the Big Chino and Verde Valley sub-basins—about 14,900 and 14,000 ac-ft/yr, respectively.

Simulated outflow from the steady-state groundwater-flow system in the Coconino Plateau basin included discharge to streams and springs, ET, and groundwater underflow to adjacent basins (table 1). Simulated steady-state groundwater discharge from the Coconino Plateau basin to the lower part of the Little Colorado River, which was simulated as a stream by using the STR Package, was about 173,000 ac-ft/yr. Net simulated groundwater discharge to the lower part of the Little Colorado River was about 172,700 ac-ft/yr. Simulated steady-state groundwater discharge from the Coconino Plateau basin to springs and the Colorado River, which were simulated as drains by using the DRN Package, was about 160,700 ac-ft/yr. Simulated groundwater discharge through ET was minor. Simulated groundwater flow from the Coconino Plateau basin to adjacent basins was about 24,000 ac-ft/yr of discharge to basins north of the Colorado River.

The net discharge to the lower part of the Little Colorado River simulated using the STR Package, 172,700 ac-ft/yr (table 1), is similar to the annual average of monthly minimum flows at the streamflow-gaging station at the Little Colorado River near Desert View, about 163,400 ac-ft/yr. Groundwater discharge from the Coconino Plateau basin to other springs and the Colorado River simulated using the DRN Package, 160,700 ac-ft/yr (table 1), is much greater than the sum of observed

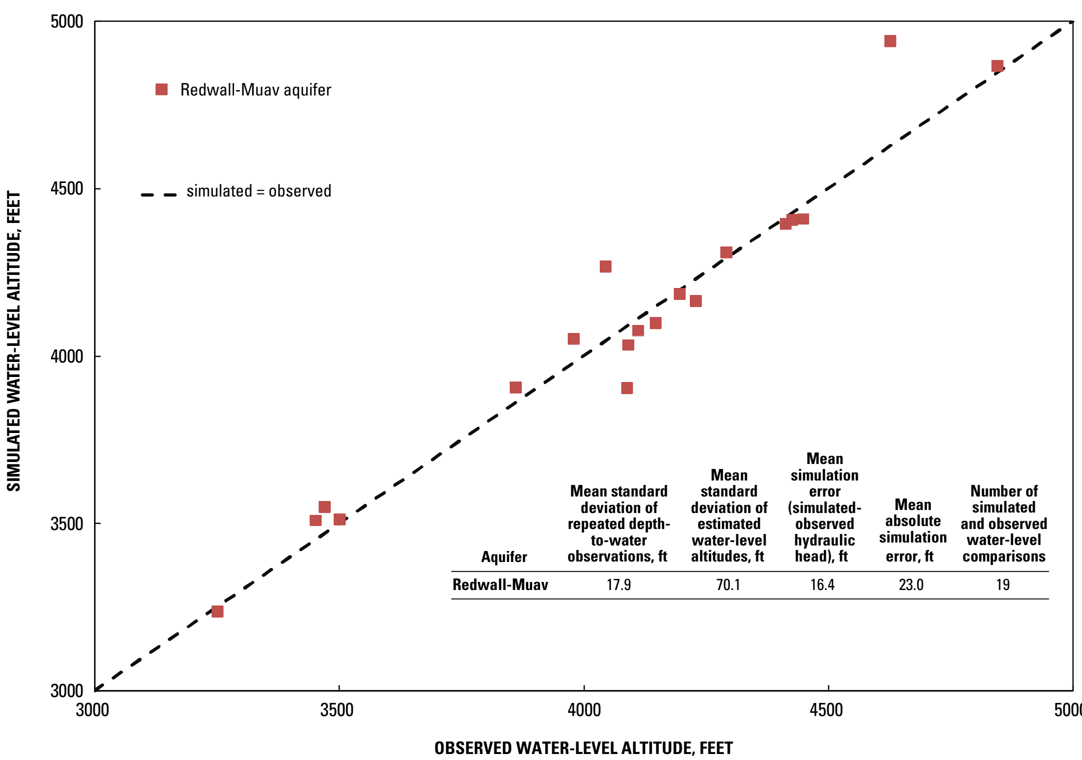


Figure 22. Simulated and observed predevelopment hydraulic head at wells in the Coconino Plateau basin and adjacent area.

estimates of groundwater discharge to springs in the Coconino Plateau, about 51,200 ac-ft/yr. This difference indicates that either recharge rates simulated by using the BCM estimates are too great by a factor of 2 or more, or estimates of discharge to springs represent only a part of total groundwater discharge. The BCM estimated recharge to the Coconino Plateau basin and adjacent areas could be inaccurate because of potentially different recharge processes than other regions and because of little data to calibrate the BCM in the area. Discharge of groundwater to the Colorado River may occur where the river flows over the Redwall-Muav aquifer, which is not estimated but could be large in comparison to estimated discharge to springs.

Major differences occur in the boundaries of the simulated and defined Coconino Plateau groundwater basin (fig. 16). Large parts of the groundwater system in the upper parts of the Big Chino and Verde Valley sub-basins were simulated as flowing into the Coconino Plateau basin. In addition, the simulated groundwater divide between the Peach Springs and Fort Rock ADWR sub-basins is substantially different from the defined groundwater divide. The groundwater divides that define these groundwater sub-basins have been uncertain because of a lack of water-level data.

Western Basins

The western basins region includes the drainage areas of Truxton Wash, Trout Creek, and Burro Creek. The Trout Creek and Burro Creek drainage areas generally correspond to the Fort Rock and Burro Creek groundwater sub-basins, respectively (fig. 9). The Truxton Wash drainage area within the region of the NARGFM includes the southeastern part of the Peach Springs groundwater basin and a small part of the adjacent groundwater basin to the west, the Wikieup sub-basin. Each of the three basins discharges groundwater to areas outside of the numerical model domain. The distribution of hydraulic head for the steady-state groundwater-flow system is defined by water-level data only near the upper reaches of the western basins and along the perennial reaches of each stream (fig. 16). The groundwater budget for the steady-state flow system is defined by estimated base flow, based on non-flood streamflow for winter months, for Truxton Wash at the streamflow-gaging station near Valentine (continuous record available 1993–2005) and for Burro Creek at the streamflow-gaging station near Bagdad (record available 1980–2005). No streamflow records are available for Trout Creek. Streams in the western basins were simulated as drains by using the DRN Package. ET from the groundwater-flow system in the western basins was not simulated.

Simulated inflow to the steady-state groundwater-flow system in the western basins included recharge and groundwater underflow from adjacent basins (table 1). Steady-state recharge to the western basins was simulated as about 41,900 ac-ft/yr, including about 4,500 ac-ft/yr in the Truxton Wash watershed, about 19,500 ac-ft/yr in the Fort Rock sub-basin, and about 17,900 ac-ft/yr in the Burro Creek sub-basin. Simulated groundwater inflow to the western basins included about 1,700, 1,200, and 5,400 ac-ft/yr to the Truxton Wash watershed and the Fort Rock and Burro Creek sub-basins, respectively. Some of the groundwater inflow to each basin includes flow from an adjacent western basin. As a result, the net simulated groundwater inflow to the western basins is about 2,700 ac-ft/yr, all of which was simulated as flowing from the Juniper Mountains in the Big Chino sub-basin.

Simulated outflow from the steady-state groundwater-flow system in the western basins included discharge to streams and groundwater underflow to adjacent basins (table 1). Simulated steady-state discharge to Truxton Wash was about 2,800 ac-ft/yr compared with estimated base flow of about 500 ac-ft/yr for available record. Simulated steady-state groundwater discharge to Trout Creek was about 11,600 ac-ft/yr. Simulated steady-state discharge to Burro Creek was about 23,300 ac-ft/yr compared with estimated base flow of about 13,900 ac-ft/yr for 16 years of available record for 1980–2005. The estimated groundwater discharge at the Burro Creek streamflow-gaging station is likely a low estimate of discharge from the groundwater basin because of substantial streamflow losses upstream from the streamflow gage to industrial and municipal water use near Bagdad and infiltration of streamflow between the streamflow gage and the groundwater basin boundary that lies about 5 mi upstream from the streamflow gage. Total steady-state groundwater discharge to the three streams was simulated as about 37,800 ac-ft/yr. Groundwater outflow was simulated from the Truxton Wash watershed and Fort Rock sub-basin at rates of about 3,300 and 9,000 ac-ft/yr, respectively. Groundwater outflow from the steady-state groundwater-flow system in the western basins was simulated as about 6,800 ac-ft/yr, of which about 2,500 ac-ft/yr discharged to the Big Chino sub-basin and 4,100 ac-ft/yr discharged to the Peach Springs basin.

Potential changes to the steady-state simulation of groundwater flow in the western basins that would improve simulation of groundwater discharge to Truxton Wash and Burro Creek include reductions in natural recharge and modification of hydraulic properties. Rates of recharge simulated by the BCM could be in error because of a lack of data to calibrate recharge through the basalt outcrops of the Aquarius Mountains and the assemblage of alluvium and Paleozoic carbonate rocks in the Truxton Wash area. Because there are few wells to define groundwater flow divides with adjacent basins, hydraulic property distributions could be modified to simulate a further westward extent of the groundwater divide with the Big Chino sub-basin, which would reduce recharge and groundwater flow rates in the western basins while increasing rates in the Big Chino sub-basin.

Differences occur in the boundaries of the simulated and ADWR-defined Fort Rock and Burro Creek groundwater sub-basins. The Truxton Wash groundwater basin is not recognized as an ADWR groundwater sub-basin and is included in the Peach Springs basin and Wikieup sub-basin. The simulated extent of the recharge area that contributes to groundwater flow in the Fort Rock (Trout Creek) sub-basin is much less than defined by the administrative groundwater sub-basin. The simulated extent of the Burro Creek sub-basin is slightly greater than defined by the ADWR groundwater sub-basin. These differences in simulated and defined groundwater basins are expected because the boundaries of the Truxton Wash, Trout Creek, and Burro Creek groundwater basins are poorly defined by available water-level data. The simulated groundwater divides also are uncertain, however, because the divide locations can vary with a range of reasonable hydraulic property and recharge distributions.

Verde Canyon Sub-Basin

The steady-state groundwater-flow system for the Verde Canyon sub-basin, a minimally developed area, is defined by stream altitudes along the Verde River, Fossil Creek, and the East Verde River and groundwater discharge estimates along Fossil Creek upstream from the streamflow-gaging station near Childs and the East Verde River upstream from the streamflow-gaging station near Childs. Few water levels in wells are available in the area for comparison with simulated steady-state water levels. Streams and major springs in the Verde Canyon sub-basin were simulated by using the STR Package. ET from the groundwater-flow system in the Verde Canyon sub-basin was not simulated.

Simulated inflow to the steady-state groundwater-flow system in the Verde Canyon sub-basin included recharge, infiltration along streams, and groundwater underflow from adjacent basins (table 1). Steady-state recharge to the Verde Canyon sub-basin was simulated as about 35,200 ac-ft/yr. Steady-state discharge to streams in the Verde Canyon sub-basin, including Verde River, Fossil Creek, and the East Verde River, was simulated as about 56,600 ac-ft/yr. About 2,000 ac-ft/yr of simulated discharge to streams reinfiltrates along the streams in the sub-basin for a net groundwater discharge of about 54,600 ac-ft/yr. Groundwater inflow from adjacent basins was simulated as about 31,200 ac-ft/yr (table 1). Most of the groundwater inflow was derived from the Little Colorado River Plateau basin, about 19,500 ac-ft/yr, and the Verde Valley sub-basin, about 11,700 ac-ft/yr.

Simulated outflow from the steady-state groundwater-flow system in the Verde Canyon sub-basin included discharge to streams and groundwater underflow to adjacent basins (table 1). Simulated steady-state discharge to streams in the sub-basin was about 56,600 ac-ft/yr. Simulated groundwater flow to adjacent basins was about 11,400 ac-ft/yr, all of which flowed into the Tonto Creek sub-basin.

Simulated steady-state discharge to Fossil Creek and the East Verde River upstream from each of the the streamflow-gaging stations near Childs was about 36,000 ac-ft/yr compared with estimated groundwater discharge of about 34,200 ac-ft/yr. The simulated distribution of groundwater discharge to each of the two streams is, however, nearly the reverse of estimated rates for the two streams. Estimated groundwater discharge to Fossil Creek is about 30,000 ac-ft/yr (Hart, 2002) in comparison to simulated discharge of about 2,900 ac-ft/yr. Estimated groundwater discharge to the East Verde River upstream from the streamflow-gaging station is about 4,200 ac-ft/yr for winter records for 1969–2004 that were not dominated by flood flows in comparison to simulated discharge of about 32,400 ac-ft/yr. Simulation of the correct distribution of base flow between the two streams requires some variations in the simulated anisotropy or aquifer system structure that would direct simulated groundwater flow toward Fossil Creek instead of simulated flow toward the East Verde River.

The simulated groundwater divide between the Verde Canyon sub-basin and Little Colorado River Plateau basin is about 5 miles north of the defined ADWR basin boundary and includes a greater area and greater simulated recharge than would be included in the defined basin (fig. 16). As a result, about 19,500 ac-ft/yr of groundwater flow simulated as entering the Verde Canyon sub-basin was derived from the Little Colorado River Plateau basin. However, much of the groundwater flow from Little Colorado River Plateau basin exited from upper altitudes of the Verde Canyon sub-basin and flowed into the Tonto Creek sub-basin (fig. 16).

Salt River Sub-Basins

Multiple groundwater basins and sub-basins are defined in the Salt River drainage. For this evaluation of the simulation of steady-state groundwater flow, only the Tonto Creek sub-basin and the combined basins upstream from the streamflow-gaging station on the Salt River near Roosevelt are discussed. The steady-state groundwater-flow system for the Salt River sub-basins is primarily defined by the altitudes of the Salt River and several tributary streams that were simulated as drains by using the DRN Package.

Simulated inflow to the steady-state groundwater-flow system in the Tonto Creek sub-basin included recharge and groundwater underflow from adjacent basins (table 1). The simulated steady-state groundwater-flow system for the Tonto Creek sub-basin included about 52,800 ac-ft/yr of recharge and 24,400 ac-ft/yr of groundwater underflow from the Little Colorado River Plateau basin, some of which flowed through the Verde Canyon sub-basin before flowing into the Tonto Creek sub-basin.

Simulated outflow from the steady-state groundwater-flow system in the Tonto Creek sub-basin included discharge to streams and groundwater underflow to adjacent basins (table 1). Simulated steady-state groundwater discharge to Tonto Creek was about 74,300 ac-ft/yr. Simulated groundwater flow to adjacent basins was about 2,900 ac-ft/yr, part of which flowed to the Salt River Lakes sub-basin, about 2,200 ac-ft/yr, and part of which flowed to the Salt River Canyon sub-basin, about 700 ac-ft/yr.

The steady-state groundwater-flow system for the Salt River Basin upstream from the streamflow-gaging station near Roosevelt includes the Salt River Canyon, White River, and Black River sub-basins. Simulated inflow to the steady-state groundwater-flow system in the three sub-basins included recharge and groundwater underflow from adjacent basins (table 1). Simulated inflow to the three sub-basins was about 177,800 ac-ft/yr of recharge and 10,900 ac-ft/yr of groundwater underflow, most of which flowed from the Little Colorado River Plateau basin, about 10,100 ac-ft/yr. Simulated outflow from the steady-state groundwater-flow system in the three sub-basins included discharge to streams and groundwater underflow to adjacent basins (table 1). Simulated steady-state discharge upstream from the streamflow gage near Roosevelt was about 188,400 ac-ft/yr, compared with estimated groundwater discharge from streamflow records at the gage of about 169,000 ac-ft/yr for 1914–2005. Simulated groundwater flow to adjacent basins was only about 300 ac-ft/yr.

The simulated groundwater divide between the Salt River sub-basins and Little Colorado River Plateau basin is generally several miles north of the defined basin boundary, except in the White Mountains where the divide is a few miles south of the defined boundary. As a result, the three sub-basins received a net simulated steady-state groundwater flow of about 10,100 ac-ft/yr from the Little Colorado River Plateau basin.

Evaluation of Simulated Transient Conditions

Simulation of the groundwater-flow system in the NARGFM region during development of the groundwater supplies for human use is evaluated for each major groundwater basin and sub-basin defined by the ADWR. Simulated and observed changes in water levels and groundwater budgets are discussed for each area for the period from before predevelopment through 2005 and illustrated by using graphs and a table showing the groundwater budget at the end of 2005 for selected basins, sub-basins, and other regions. Two statistics were calculated to describe how well the model simulates changes in water levels through time for individual sub-basins that had five or more water-level observations during a period of two decades or longer. The first statistic is the average difference between simulated and observed water-level change from the initial measurement for each subsequent observation. This statistic is a measure of model bias in estimating water-level change. A negative value indicates that the model, on average, simulates less rise and (or) more decline than observed; a positive value indicates that the model, on average, simulates more rise and (or) less decline than observed; a zero value indicates no overall bias in simulated water-level change. The second statistic is the average absolute difference between simulated and observed water-level change from the initial measurement for each subsequent observation. This statistic is a measure of model overall goodness of fit between simulated and observed water-level changes.

Areas of the primary focus of the NARGFM are discussed first, including the Little Chino, Upper Agua Fria, Big Chino, and Verde Valley sub-basins. Major basins that are adjacent to the primary focus areas are then discussed, including the Little Colorado River Plateau and Coconino Plateau basins. The western basins that include the Burro Creek and Fort Rock (Trout Creek) sub-basins and a part of Peach Springs basin (Truxton Wash watershed) are discussed. Finally, the Verde Canyon sub-basin, Tonto Creek sub-basin, and several basins in the Salt River drainage are discussed. Changes in groundwater-flow systems are not discussed for a few areas that lack observations—the Peach Springs basin and several basins north of the Colorado River including the Grand Wash, Shivwits Plateau, and Kanab Plateau basins—but simulated groundwater budgets are provided for these areas.

Little Chino and Upper Agua Fria Sub-Basins

The simulation of the transient groundwater-flow system in the Little Chino and Upper Agua Fria sub-basins from 1938 to 2005 is evaluated through comparison of simulated and observed changes in water-level altitude at wells with long-term records and changes in groundwater discharge at Del Rio Springs and the Upper Agua Fria River near Humbolt. Long-term hydrographs are available at many wells for comparison of simulated and observed hydraulic head in the sub-basins; however, records at only a few representative wells are discussed.

Groundwater budgets for the Little Chino and Upper Agua Fria sub-basins indicate that groundwater withdrawals exceeded recharge during much of the simulated transient period (table 2, figs. 23*A*1,*A*2). Simulated natural recharge rates in the Little Chino sub-basin varied from about 2,400 ac-ft/yr or less before 1950 to as much as about 4,100 ac-ft/yr during the 1970s and 1990s and about 1,800 ac-ft/yr after 1999 (fig. 23*A*1). Simulated natural recharge rates in the Upper Agua Fria sub-basin varied from about 1,700 ac-ft/yr or less before 1950 to as much as about 3,000 ac-ft/yr during the 1970s and 1990s and about 1,300 ac-ft/yr after 1999 (fig. 23*A*2). Incidental recharge that was simulated in both sub-basins was about 2,000–9,200 ac-ft/yr (fig. 23*A*3) and included several components—losses from surface-water distribution systems, excess irrigation water that was derived from surface water and groundwater sources and applied to crops and golf courses, and infiltration of discharge from sewage-treatment facilities. Recharge from losses of surface water diverted for agricultural irrigation varied from about 2,000 ac-ft/yr before the first irrigation wells were drilled to about 1,400 ac-ft/yr after 1999. Recharge of excess applied irrigation water that was derived from irrigation well withdrawals was simulated as 50 percent of the withdrawals and varied from about 700 ac-ft/yr in 1938 to an average of more than 7,000 ac-ft/yr during the 1970s and an average of about 2,300 ac-ft/yr after 1999. Recharge of excess applied irrigation water at five golf courses in the Little Chino and Upper Agua Fria sub-basins was simulated as less than 200 ac-ft/yr beginning with the final 10-year stress period in 1990. Simulated recharge from deep percolation of effluent from wastewater treatment facilities for the cities of Prescott and Prescott Valley ranged from about 100 ac-ft/yr beginning with the stress period that starts in 1980 to about 4,400 ac-ft/yr during 2000 through 2005.

The excess of withdrawals over recharge in the Little Chino and Upper Agua Fria sub-basins is supplied by removal

Table 2. Groundwater-flow budgets for Arizona Department of Water Resources groundwater basins and sub-basins and selected regions of the Northern Arizona Regional Groundwater-Flow Model during 2005.

[Values are in acre-feet per year except for cumulative values, which are in acre feet.]

		Verde River Basin groundwater-flow system							Colorado River Basin groundwater-flow system					Western basins groundwater-flow system				Salt River sub-basins groundwater-flow system		
Groundwater Basin	Upper Agua Fria	Little Chino	Little Chino and Upper Agua Fria sub-basins	Big Chino	Big and Little Chino	Verde Valley sub-basin above the streamflow-gaging station near Clarkdale	Verde Valley sub-basin between the streamflow-gaging stations near Clarkdale and near Camp Verde	Verde Valley	Verde Canyon	Little Colorado	Coconino Plateau	Peach Springs	Kanab Plateau and adjacent areas	Burro Creek	Fort Rock (Trout Creek)	Truxton Wash	Western basins[1g]	Tonto Creek	Salt River Lakes	Salt River above the streamflow-gaging station at Roosevelt
Arizona Department of Water Resources basin type	sub-basin	sub-basin	Prescott Active Manage-ment Area	sub-basin	sub-basin	part of sub-basin	part of sub-basin	sub-basin	sub-basin	basin	basin	Part of basin that lies outside of Truxton Wash watershed	several basins north of the Colorado River	sub-basin	sub-basin	Parts of Peach Springs basin and Wikieup sub-basin	simulated groundwater basin[1l]	basin	basin	Salt River Canyon, Black River, and White River sub-basins
Groundwater-budget component Inflow																				
Natural recharge	1,300	1,800	3,100	31,200	33,000	22,500	46,100	68,600	26,400	154,900	113,900	5,100	75,200	13,400	14,600	3,300	31,400	39,700	28,300	133,300
Recharge from infiltration of streamflow derived from base flow[1a]	N/A	700	700	1,900	2,600	13,800	24,700	38,500	1,800	13,100	300	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Incidental recharge[1b]	2,300	7,000	9,200	4,200	11,200	0	1,600	1,600	0	49,800	500	0	0	0	0	100	100	1,900	1,300	1,200
Groundwater inflow from adjacent areas[1c]	1,000	1,500	1,800	3,500	5,100	8,800	1,700	8,000	33,100	2,600	190,400	5,400	38,100	6,200	1,500	1,700	9,400	26,400	1,900	10,800
Total Inflow[1d]	4,600	11,000	14,800	40,800	51,800	45,100	74,000	116,700	61,300	220,300	305,100	10,500	113,300	19,700	16,100	5,100	40,900	67,900	31,500	145,200
Outflow																				
Groundwater discharge to streams (base flow)[1e]	N/A	1,800	1,800	18,100	19,900	36,100	62,200	98,400	62,800	13,100	173,000	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Discharge to streams and springs simulated as drains (base flow)[1f]	1,000	0	1,000	0	0	0	0	0	500	300	160,700	7,600	135,900	25,800	13,300	2,900	42,000	84,100	43,200	203,200
Evapotranspiration by phreatophytes[1g]	100	100	200	1,900	2,000	2,600	9,300	11,900	N/A	0	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Groundwater withdrawals	6,400	17,800	24,200	11,200	28,900	200	19,300	19,500	1,600	81,700	800	0	0	400	0	100	500	1,100	1,300	600
Groundwater outflow to adjacent areas[1h]	0	2,600	1,800	26,900	29,500	14,900	13,400	25,800	11,400	210,400	25,100	4,900	9,000	0	9,700	3,300	13,000	2,700	200	0
Total Outflow[1i]	7,500	22,200	28,900	58,100	80,400	53,900	104,200	155,600	76,400	305,600	359,500	12,400	145,000	26,100	23,100	6,300	55,500	87,900	44,700	203,700
Net groundwater flow to (-) and from (+) adjacent basins[1j]	1,000	-1,100[2]	0	-23,400[3]	-24,400[3]	-6,000	-11,800	-17,800	21,700	-207,900	165,300	600	29,100	6,200	-8,200	-1,600	-3,600	23,600	1,600	10,800
Net streamflow[1k]	N/A	1,100	1,100	16,200	17,300	22,300	37,600[4]	59,900[4]	61,000	0	172,700	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A	N/A
Net rate of groundwater storage change	-2,900	-11,200	-14,100	-17,400	-28,600	-8,800	-30,200	-39,000	-15,200	-85,000	-54,000	-1,900	-31,800	-6,500	-6,900	-1,200	-14,600	-20,100	-13,200	-58,400
Cumulative groundwater storage change since predevelopment	-117,100	-586,500	-703,600	116,700	-469,800	242,200	295,500	537,700	201,200	669,500	1,275,100	70,800	1,021,000	130,600	156,700	37,900	325,200	322,400	326,600	1,205,600

1a-Includes the total of simulated streamflow infiltration; all of which is derived from discharge of groundwater to streams (simulated using the MODFLOW Streams Package (STR)) within the groundwater basin and in upgradient groundwater basins.

1b-Includes recharge resulting from excess applied irrigation water derived from surface water and groundwater supplies, discharge from waste-water treatment facilites, and golf courses.

1c-Includes components of groundwater inflow from multiple adjacent basins and sub-basins.

1d-Represents the sum of inflow components including groundwater flow from adjacent basins and reinfiltration and recharge of groundwater discharge to streams in the groundwater basin.

1e-Includes only discharge to streams simulated using the MODFLOW Streams Package (STR).

1f-Streams and springs simulated using the MODFLOW Drain Package (DRN).

1g-Evapotranspiration was simulated in only the Verde River Basin above the streamflow-gaging station near Camp Verde and in the Little Colorado River Basin.

1h-Includes components of groundwater outflow to multiple adjacent basins and sub-basins.

1i- Calculated as Total Inlow minus Total Outflow. Multiple areas of groundwater inflow and outflow may occur for any basin.

1j- Calculated as Groundwater Infow minus Groundwater Outflow.

1k- Net streamflow is equivalent to stream base flow simulated as exiting the basin.
Calculated as Recharge from infiltration of streamflow derived from base flow minus Groundwater discharge to streams. Both components are simulated using the Streams Package (STR).

1l- Includes the sum of groundwater budget components for the Burro Creek and Fort Rock sub-basins and the parts of the Peach Springs Basin and Wikieup sub-basin within the Truxton Wash watershed.

2- Includes groundwater outflow of about 2,100 acre-feet per yr to the Big Chino sub-basin and inflow from portions of the Big Chino (about 1,000 acre-feet per yr) and Upper Agua Fria sub-basins.

3- Includes the balance of outflow of groundwater to adjacent basins and inflow from adjacent basins, primarily from the Little Chino sub-basin.

4- Does not include diversion of 10,200 acre-feet per year of base flow that is transpired by crops.

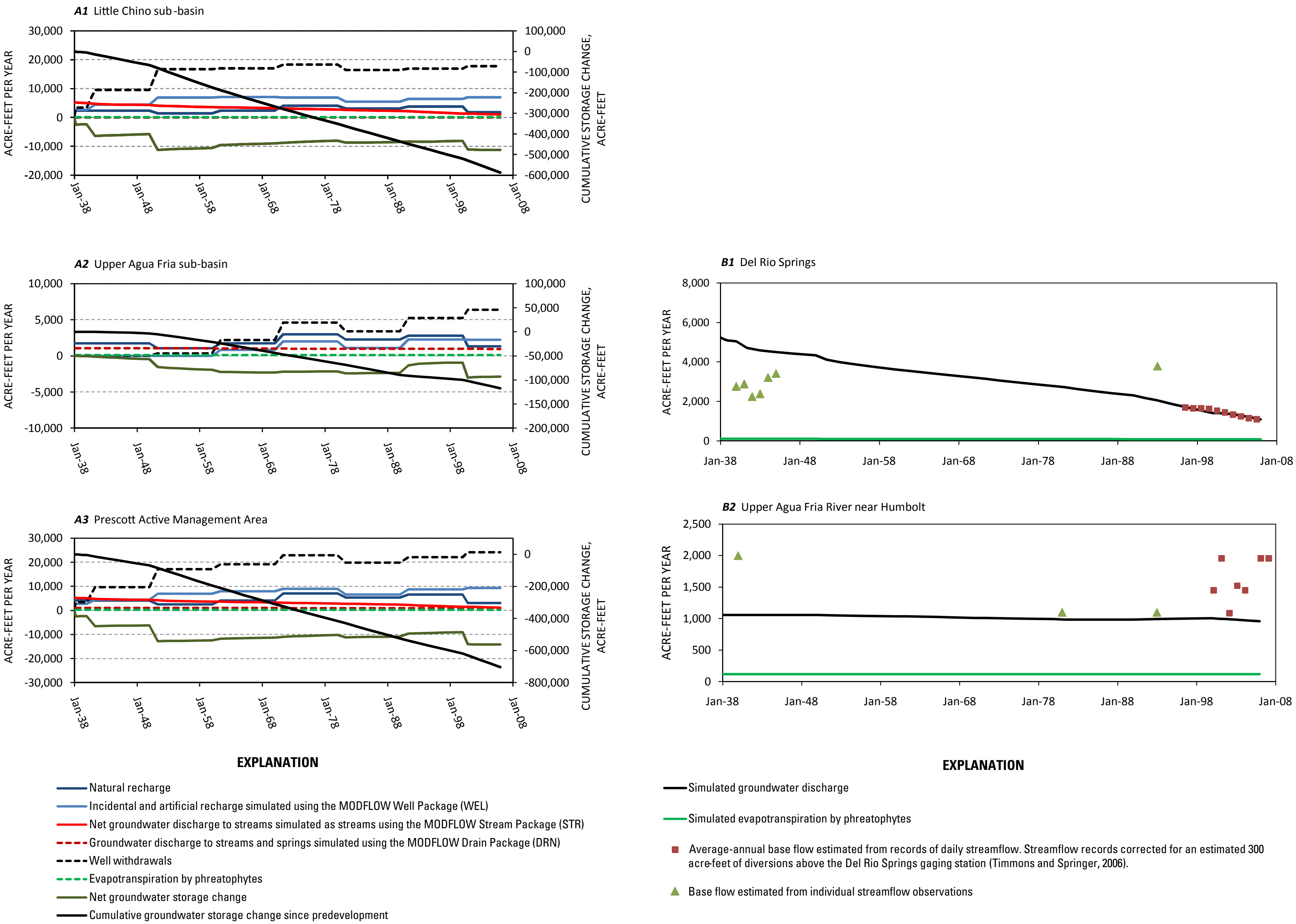


Figure 23. Simulated groundwater budgets for the (*A1*) Little Chino sub-basin, (*A2*) Upper Agua Fria sub-basin, and (*A3*) Prescott Active Management Area and simulated and estimated base flow discharge at (*B1*) Del Rio Springs and (*B2*) Agua Fria River near Humbolt.

of water from storage, which resulted in water-level declines and diminishment of groundwater discharge to adjacent basins and to streams and springs. Storage losses and capture of groundwater discharge to adjacent basins were substantial during the transient simulation in both the Little Chino and Upper Agua Fria sub-basins. Rates of storage loss were greater than 5,000 ac-ft/yr after 1950 in the Little Chino sub-basin and about 1,000 ac-ft/yr or greater in the Upper Agua Fria sub-basin (figs. 23*A*1,*A*2). Total simulated storage loss through 2005 was 586,500 ac-ft in the Little Chino sub-basin and 117,100 ac-ft in the Upper Agua Fria sub-basin for a total of 703,700 ac-ft for the two sub-basins (table 2, figs. 23 *A*1,*A*2,*A*3). Net groundwater flow from adjacent areas of about 500 ac-ft/yr before development (table 1) diminished to zero by the end of 2005 (table 2). Groundwater withdrawals resulted in reduced simulated discharge at Del Rio Springs from about 5,300 ac-ft/yr before development (table 1) to about 1,700 ac-ft/yr in 2005 (table 2, fig. 23*B*1). Simulated discharge at Del Rio Springs is greater than estimated discharge rates in early times, but similar to measured rates during 1995 through 2005 (fig. 23*B*1). Poor simulation of early discharge rates may result from poorly defined early rates of groundwater withdrawal, especially rates before 1938, and reduced springflow observations resulting from nearby groundwater withdrawals. Simulated discharge at the Upper Agua Fria River near Humbolt did not vary greatly during 1938 to 2005, and simulated rates during 2005, about 1,000 ac-ft/yr, were slightly less than simulated pre-development rates of about 1,100 ac-ft/yr (fig. 23*B*2). Simulated rates of groundwater discharge above the streamflow-gaging station at the Upper Agua Fria River near Humbolt are similar to minimum estimates of base flow discharge after 1980, but about 50 percent of a single estimate made in 1940. Annual estimates of base flow discharge based on streamflow records during 2000–06 were variable, ranging from about 1,100 to 2,000 ac-ft/yr. The variability in base flow estimates may result from several causes, including variations in annual recharge from streamflow infiltration, inclusion of a component of runoff during some wet years, or variable agricultural irrigation practices. The lower range of base flow estimates is likely a better target for this regional simulation of groundwater discharge because streamflow infiltration and stream runoff were not simulated.

Water-level records at three wells—(B-15-02)17ABA, (B-16-02)14CDA, and (B-16-01)14CCC—are selected as representative of the groundwater system response to groundwater withdrawals in areas of differing hydrogeologic conditions in the Little Chino basin (figs. 17, 24). Well (B-15-02)17ABA is in layer 3 on the west side of the Little Chino basin and west of the region of confined conditions in layer 2. Observed water levels varied little during 1940–2005, except that a slight water-level recovery happened after about 1980. Simulated water levels at well (B-15-02)17ABA included a slight downward trend during 1950–2000, followed by a slightly accelerated downward trend. Differences in simulated and observed trends at well (B-15-02)17ABA are not great, but likely indicate that variations in actual recharge rates were greater than simulated. Observed water-level declines of 70–100 ft at wells (B-16-02)14CDA and (B-16-01)14CCC are representative of trends in the lower volcanic unit (layer 2). Simulated declines in layer 2 at the wells are similar to observed; however, a period of a smaller decline in observed water levels during 1970–95 is not simulated. The lack of simulated variation in water-level trends at wells (B-16-02)14CDA and (B-16-01)14CCC is likely because of misrepresentation of variations in recharge rates or withdrawal locations and rates. Better simulation of observed water-level trends might be improved by including variations in ephemeral channel recharge and improved information on specific yield in areas of unconfined groundwater conditions.

Water-level records at 2 wells—(B-14-01)15ABA and (A-14-01)28BBB—are representative of the groundwater system response to groundwater withdrawals in the Upper Agua Fria sub-basin. Observed water levels at well (B-14-01)15ABA are variable partly because of water that occasionally is cascading into the well and that makes measurement difficult; however, an overall water-level decline of more than 300 ft was determined during about 1970–2002 (fig. 24). Simulated declines were about 270 ft during the same time period in layer 2 at the well (fig. 24). Greater amounts of decline are difficult to simulate by using the provided groundwater-withdrawal rates and locations because the simulated aquifer properties of hydraulic conductivity and specific storage in the confined conditions of the lower volcanic unit are extremely low and lower values that would result in greater simulated water-level decline are unlikely. Greater rates of withdrawal or a withdrawal location that is closer to well (B-14-01)15ABA would result in greater simulated water-level decline at the well. Another condition that could result in greater rates of simulated water-level decline at the well include a local impermeable aquifer boundary that is closer to the pumped well than was simulated. Observed water levels at well (A-14-01)28BBB vary across a large range during multi-year periods, as much as 50 ft, and indicate very little trend before about 1985, after which a slight downward trend of about 5 ft/yr occurred. Simulated water levels at well (A-14-01)28BBB began a slight downward trend about 1960 before declining at a constant rate of about 5 ft/yr after 1980. Simulated early water levels at well (A-14-01)28BBB are about 35 ft less than observed; however, two of three simulated steady-state water levels in the area matched early water levels very well and one water level was about 35 ft less than observed. The substantially different hydraulic heads among nearby wells in the area indicates that multiple poorly connected aquifers may exist, which may contribute to the poor simulation of overall hydraulic head at well (A-14-01)28BBB.

Simulated water-level trends in these sub-basins were compared with observed trends from 247 water-level measurements at 5 wells. Observed changes averaged 15.3 ft of decline and ranged from about 107 ft of decline to about 36 ft of water-level rise. The average and average absolute difference between simulated water-level changes and corresponding observed changes were about -8.6 ft and 14.7 ft, respectively (table 3).

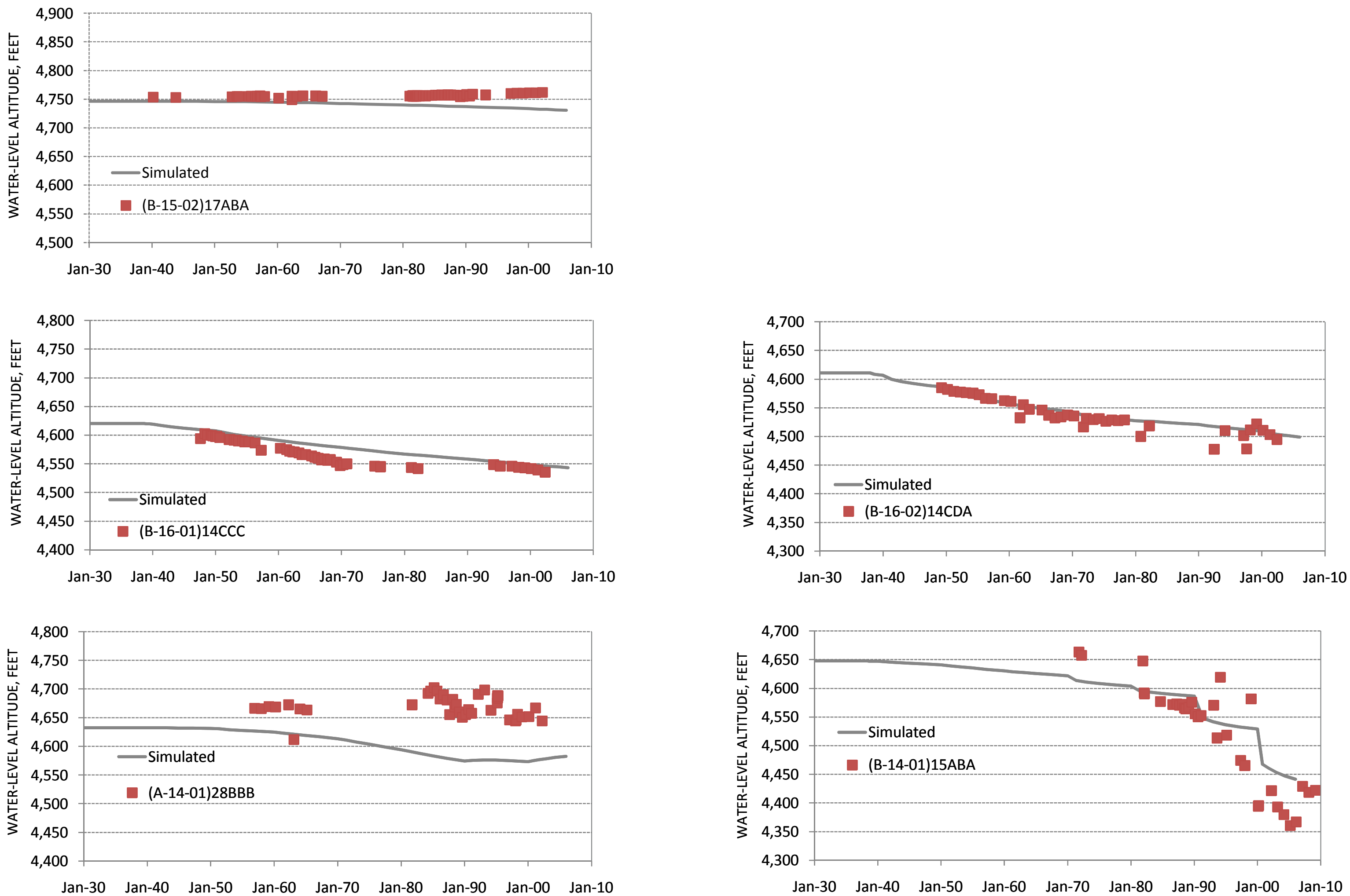


Figure 24. Hydrographs of simulated and observed water levels at selected wells in the Little Chino and Upper Agua Fria sub-basins.

Table 3. Statistical comparison of simulated and observed water-level trends at selected wells with long-term water-level records for selected regions of the Northern Arizona Regional Groundwater-Flow Model (NARGFM). [Values are in feet.]

Groundwater Basin[1]	Basin category	Average simulated minus observed change[2]	Average absolute simulated minus observed change	Number of wells	Number of water-level observations
Little Chino and Upper Agua Fria	Prescott Active Management Area	-8.6	14.7	5	247
Big Chino	ADWR groundwater sub-basin	-7.9	8.5	5	189
Verde Valley	ADWR groundwater sub-basin	0.0	8.2	7	265
Little Colorado River Plateau	ADWR groundwater basin	-11.0	18.0	8	412
NARGFM	Simulation extent	-9.0	21.4	88	3,490

[1]Statistics were calculated for only basins and sub-basins that had 5 or more wells with 10 or more water-level observations during a period of 2 decades or longer.

[2]Calculated as simulated change from the simulated hydraulic head on the initial measurement date minus observed change from the initial observed hydraulic head for each repeated observation at each well.

Big Chino Sub-Basin

Simulation of the transient groundwater-flow system in the Big Chino sub-basin during 1938–2005 is evaluated through comparison of simulated and observed changes in groundwater discharge at the streamflow-gaging station at the Verde River near Paulden and changes in water-level altitude at a few wells with long-term records. The evaluation suffers from a scarcity of streamflow and water-level records before about the mid-1960s that could document response of the groundwater system to the applied stresses of groundwater withdrawal and variations in recharge. As a result, simulation of the response of the groundwater system to these stresses is uncertain.

Simulated groundwater budgets for the Big Chino sub-basin indicate that groundwater withdrawals exceeded BCM estimated natural recharge rates before 1960 and after 1990, but increased recharge rates during 1970–2000 resulted in storage recovery and an exceedance of predevelopment volumes of water in storage (tables 1,2; fig. 25*A*). Simulated natural recharge rates in the Big Chino sub-basin varied from about 41,600 ac-ft/yr before 1950 to as little as about 24,900 ac-ft/yr during the 1950s, to the greatest rate of about 71,500 ac-ft/yr during the 1970s (fig. 25*A*). Additional incidental recharge of excess irrigation water was simulated as a maximum rate of about 6,200 ac-ft/yr during the 1960s to less than about 4,400 ac-ft/yr or less before 1950 and after 1980 (fig. 25*A*). Groundwater withdrawals peaked in the 1960s at about 14,300 ac-ft/yr before decreasing to about 8,700 ac-ft/yr during the 1990s and about 11,200 ac-ft/yr during 2000–05. Simulated rates of ET declined from about 2,200 ac-ft/yr before 1940 to about 1,900 ac-ft/yr during 2005 (table 1, fig. 25*A*). Decreasing simulated groundwater discharge to the Verde River upstream from the Paulden streamflow gage, from about 21,700 ac-ft/yr before development (table 1, Net streamflow for the Big and Little Chino sub-basins) to a minimum of about 17,600 ac-ft/yr in 1970 (fig. 25*B*), resulted from multiple causes. Groundwater withdrawals in the Big and Little Chino sub-basins and below average recharge rates during the 1950s contributed to the pre-1970 decrease in simulated groundwater discharge above the Paulden gage. The greatest rates of simulated recharge beginning in the 1970s resulted in a slight recovery of groundwater discharge above the gage to about 18,900 ac-ft/yr in 1980. Lower, but above average, simulated rates of natural recharge during the 1980s and 1990s resulted in decreasing groundwater discharge above the Paulden gage. Below average simulated rates of natural recharge during 2000–05 resulted in further declines in groundwater discharge above the Paulden gage to about 17,300 ac-ft/yr in 2005 (table 2, Net streamflow for the Big and Little Chino sub-basins; fig. 25*B*). The simulated decrease in groundwater discharge above the Paulden gage during 1980–2000 opposes a slight declining trend in simulated groundwater withdrawals in the Big and Little Chino sub-basins over the same period. The lack of simulated increases in groundwater discharge during this period of decreasing withdrawals and recharge emphasizes the importance of variations in recharge rates as important source of variations in base flow at the Paulden gage. Simulated storage change from before development to 2000 in the Big Chino sub-basin was an increase of more than 200,000 ac-ft (fig. 25*A*). Simulated storage decreased with below average recharge rates after 2000 to about 116,700 ac-ft in 2005. A net loss of storage was simulated during 2005 at a rate of about 17,400 ac-ft/yr. Simulated groundwater discharge above the Paulden gage does not increase with the increased groundwater in storage during the 1980s and 1990s because of the relative spatial distributions of the overall increased storage and localized groundwater withdrawals. Most groundwater withdrawals and associated local storage losses are nearer to the discharge area than most of the storage increase, especially withdrawal areas in the lower part of the Little Chino sub-basin and near Paulden. As a result, the withdrawals near the discharge area have a more rapid influence on groundwater discharge than storage increases that primarily lie at greater distance. Net groundwater flow to adjacent areas of about 23,100 ac-ft/yr before development (table 1) changes little during the transient simulation to about 23,400 ac-ft/yr by the end of 2005 (table 2).

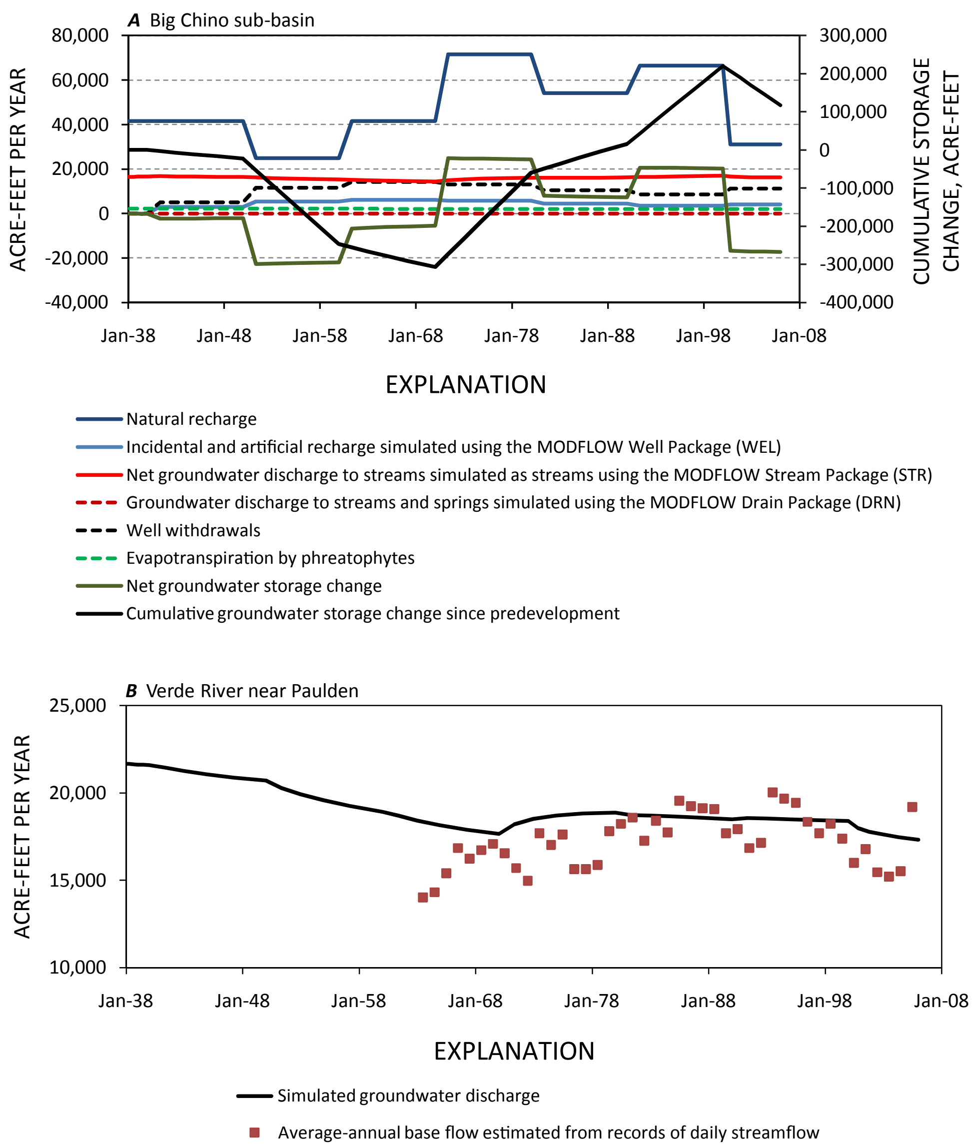


Figure 25. Simulated groundwater budgets for the (*A*) Big Chino sub-basin and (*B*) simulated and estimated base flow discharge at the Verde River near Paulden, which includes contributions from both the Big and the Little Chino sub-basins.

Simulated groundwater discharge at the streamflow-gaging station near Paulden roughly matches the discharge rates of estimated annual base flow during 1964–2005 (fig. 25*B*), whereas estimated base flow during multi-year wet periods is greater than simulated. Low estimates of annual base flow during the 1960s and 1970s, about 15,000 ac-ft/yr or less, may be related to locally greater rates of short-term groundwater withdrawals in the Paulden area. The variability of estimated annual base flow, as much as 5,000 ac-ft/yr of inter-annual variation, may also occur because of local and temporary variations in recharge rates, possibly along the Big Chino and Williamson Valley Washes, that are much shorter than the decadal length of simulated variations in recharge rates.

Water-level records at five wells were selected as representative of the groundwater system response to groundwater withdrawals in the Big Chino alluvial basin (figs. 17, 26). Three of the wells are in the upper part of the Big Chino alluvial basin—(B-19-04)04BDB, (B-19-04)10ADA, (B-19-03)19CBD—and represent water-level variations in the unconfined aquifer of model layer 1 and the deeper, but hydraulically connected layer 2. One well—(B-16-04)14BBB1—was selected to represent water-level variations in the unconfined alluvial aquifer in the Williamson Valley area, which is simulated as layer 2. And one well—(B-17-02)06BBB—was selected to represent water-level variations in the confined aquifer underlying the thick silt and clay layer, simulated as layer 2, in the lower part of the Big Chino sub-basin. No long-term water-level records are available for the Redwall-Muav aquifer in the area, layer 3. Some long-term water-level records are available for the shallow aquifer overlying the thick silt and clay layer; however, the shallow aquifer is not explicitly represented in the numerical flow model.

Long-term water-level records in the upper part of the Big Chino alluvial basin are few; however, available records indicate downward trends of water levels before about 1970 followed by recovery. Water levels at well (B-19-04)04BDB indicate a slight downward trend during 1953–67 (fig. 26). Water levels at wells (B-19-04)10ADA and (B-19-03)19CBD indicate a trend of recovery during 1972–2002 (fig. 26). Simulated water levels in the upper part of the Big Chino alluvial basin indicate trends that are similar to observed, but with recovery that begins with the stress period beginning in 1980 (fig. 26). Greater water-level recovery was observed during the early part of the record at well (B-19-04)10ADA than was simulated. Only a decrease in rate of decline was simulated at well (B-19-03)19CBD. Differences in the magnitude and timing of simulated and observed water levels may be related to several causes, including inaccurate simulation of the spatial and temporal distributions of recharge, inaccurate estimates of the spatial and temporal distribution of groundwater withdrawals, and spatial discretization of the model domain. Simulated water-level recovery about a decade after observed recovery may be related to the inaccurate simulation of the spatial and temporal distributions of recharge. Greater variations in simulated recharge near the observed water-level records would result in less delay and greater variations in simulated water levels. Both the timing and magnitude of simulated water-level variations could be improved with the addition to the model of periodic recharge near Big Chino Wash and more frequent variations in recharge rates. Inaccurate estimates of the spatial and temporal distribution of groundwater withdrawals also could, however, result in inaccurate simulation of the magnitude and timing of observed water-level variations. Spatial discretization of the model domain may also be an important consideration in the simulation of water-level variations near recharge and discharge features such as streams and withdrawal wells. Spatial discretization does not allow the precise simulation of the locations of the features. Therefore, the precise simulation of water-level observations that are located within a few grid cells, 0.62 mi for the NARGFM, of recharge or discharge features can not be expected, but simulation of general trends in observations should be possible. Some water-level observations in the upper part of the Big Chino alluvial basin are within a few grid cells of recharge sources—Big Chino Wash—and withdrawal wells, and only approximate simulation of observed trends is possible.

Water-level variations in the Williamson Valley area indicate minimal long-term variation. No major water-level trends have occurred at well (B-16-04)14BBB1 since records began in 1949 (fig. 26). In contrast, the simulated water levels at well (B-16-04)14BBB1 indicate a downward trend of about 15 ft during 1950–80 followed by stable water levels (fig. 26). The lack of observed water-level decline in this area of agricultural groundwater withdrawal indicates that withdrawal rates are compensated by almost equivalent rates of natural and incidental recharge. Some additional recharge may have been induced by slight water-level decline near Williamson Valley Wash. Induced recharge may happen in ephemeral stream channels where water levels are naturally near the streambed, such as near the part of Williamson Valley Wash immediately upstream from the streamflow-gaging station. Water-level decline in such areas may allow local infiltration of greater amounts of streamflow than would otherwise exist at the expense of reduced downstream channel flow and possible reduced recharge in downstream areas. Variations in stream-channel infiltration and recharge is difficult to explicitly simulate along this part of Williamson Valley Wash, because no streamflow records are available for upstream reaches that contribute flow to the stream in the agricultural area.

Water-level trends in the primary aquifer in the lower part of the Big Chino alluvial basin, downstream from the confluence of Walnut Creek and Big Chino Wash are similar to trends in the upper part of the basin. The best long-term record of water levels in the primary aquifer is at well (B-17-02)06BBB, which includes records beginning in 1952 (fig. 26). Water levels declined a few feet before about 1970 followed by recovery of a few feet until about 1995, after which levels declined a few feet. Simulated trends at well (B-17-02)06BBB are similar to observed (fig. 26). However, total declines of about 12 ft were simulated before 1970 followed by a couple of feet of recovery and declines of a couple of feet during the final simulated stress period after 2000.

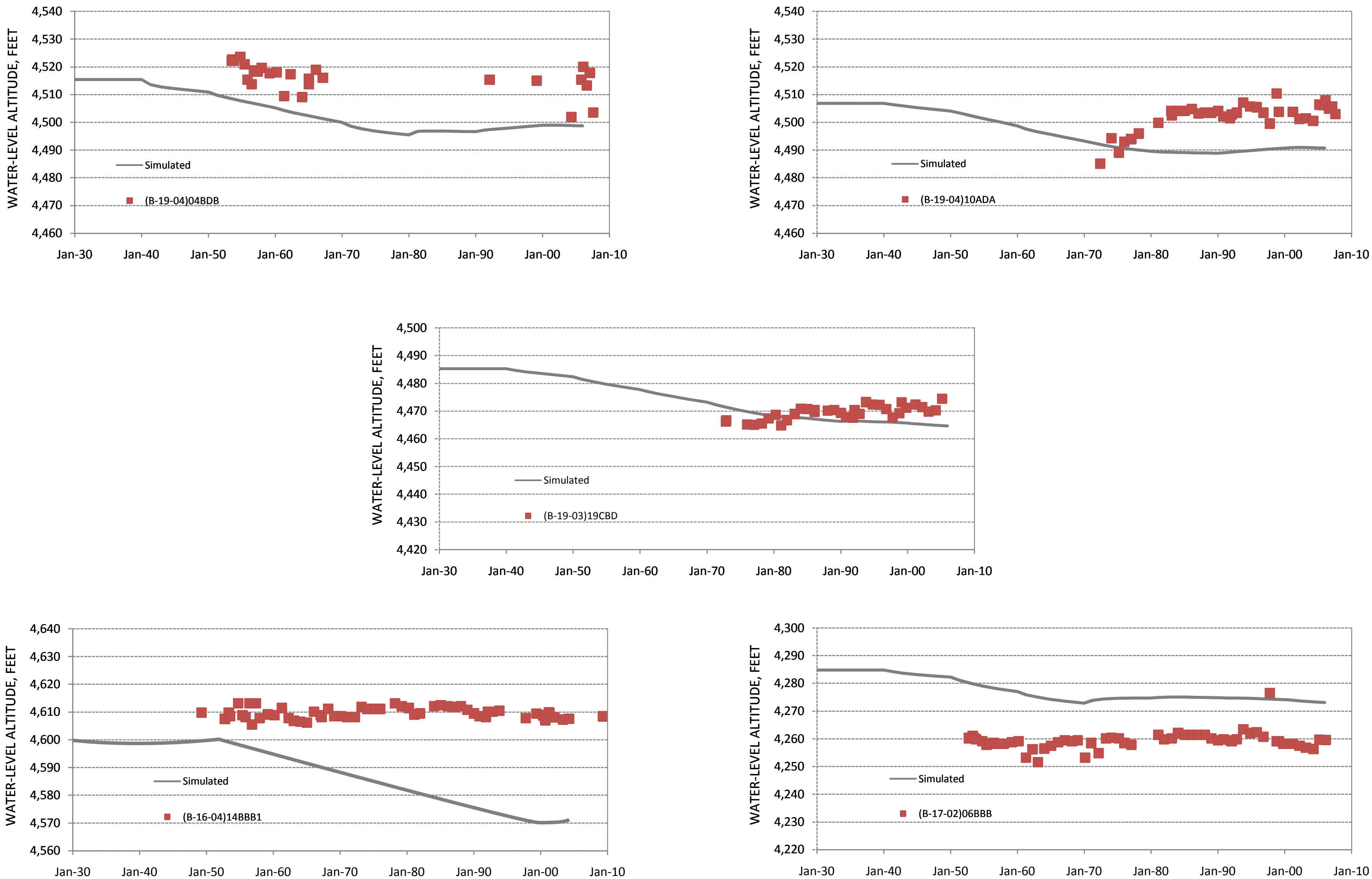


Figure 26. Hydrographs of simulated and observed water levels at selected wells in the Big Chino sub-basin.

Simulated water-level trends in this sub-basin were compared with observed trends from 189 water-level measurements at five wells including one well in Williamson Valley. Observed changes averaged 2.4 ft of rise and ranged from about 13 ft of decline to about 7 ft of water-level rise. The average and average absolute differences between simulated water-level changes and corresponding observed changes were about -7.9 ft and 8.5 ft, respectively (table 3).

Verde Valley Sub-Basin

Simulation of the transient groundwater-flow system in the Verde Valley sub-basin for 1938–2005 is evaluated through comparison of simulated and observed changes in groundwater discharge at streamflow-gaging stations and changes in water-level altitude at a few wells with long-term records. The Verde Valley sub-basin groundwater-flow system is separated into two parts for discussion—the upper part of the flow system upstream from the streamflow-gaging station near Clarkdale and the lower part of the flow system between the streamflow-gaging stations near Clarkdale and Camp Verde. The part upstream from the streamflow gage near Clarkdale is minimally developed with few pumping wells and little agriculture. The part downstream from the streamflow gage near Clarkdale is developed with surface-water diversions for irrigation and many wells that pump water for agricultural, municipal, and industrial use. Few early water-level and streamflow records are available that could document the response of the groundwater-flow system in the Verde Valley sub-basin to the applied stresses of groundwater withdrawal and variations in recharge. As a result, simulation of the response of the groundwater-flow system to these stresses is highly uncertain.

Changes to the upper part of the Verde Valley sub-basin groundwater-flow system between the Paulden and Clarkdale streamflow-gaging stations were dominated by variations in recharge rates during 1940–2005 (fig. 27*A*1). Groundwater storage and discharge to streams decreased with below average recharge rates during 1950–1960 before increasing with greater than average recharge rates during 1970–2000 and decreasing with below average recharge rates during 2000–2005 (fig. 27*A*1). Simulated rates of groundwater discharge at the Clarkdale streamflow gage are about 10,000–20,000 ac-ft/yr less than estimated annual-average base flow during 1915–1921 and 1965–2005 (fig. 27*B*1). However, simulated variations in groundwater discharge to the Verde River upstream from the streamflow-gaging station near Clarkdale during 1915–2005 are similar to variations in groundwater discharge estimated from the available record. Simulated capture of discharge by well withdrawals upstream from the streamflow-gaging station near Paulden may cause a part of about 4,000 ac-ft/yr of decline in simulated streamflow at the Clarkdale streamflow gage before about 1965. Simulated net groundwater flow to and from adjacent basins changed by only about 100 ac-ft/yr during the simulation (tables 1, 2).

Simulated groundwater budgets for the Verde Valley area downstream from the Clarkdale streamflow-gaging station indicate that changes in the groundwater system are dominated by variations in recharge rates (table 2, fig. 27*A*2). Simulated natural recharge rates were average or below average before 1970 and after 2000. Greater recharge rates during 1970–2000 resulted in storage recovery, with the greatest storage in the groundwater system happening in 2000 (fig. 27*A*2). Simulated natural recharge rates in the Verde Valley sub-basin varied from about 91,400 ac-ft/yr—30,000 and 61,400 ac-ft/yr in the upper and lower parts of the sub-basin, respectively—before 1950 to as little as about 54,700 ac-ft/yr during the 1950s and as great as about 157,300 ac-ft/yr during the 1970s. Additional incidental recharge at golf courses and near sewage treatment facilities, primarily in the lower part of the basin, peaked at about 1,600 ac-ft/yr after about 1999. Groundwater withdrawals in the Verde Valley sub-basin peaked at about 19,500 ac-ft/yr after 1999, almost all of which was in the lower part of the sub-basin (table 2). Simulated ET rates in the Verde Valley sub-basin varied little, from about 12,100 ac-ft/yr before development to about 11,900 ac-ft/yr after 2000 (table 1, fig. 27*B*2). Most of the ET was simulated as occurring in the lower part of the sub-basin, about 9,300 ac-ft/yr after 1999. Agriculture in the lower part of the sub-basin consumed surface water diverted from the Verde River at a constant rate of about 10,200 ac-ft/yr throughout the transient simulation and had no influence on variations in the simulated groundwater budget for the sub-basin. Simulated groundwater storage in the lower part of the Verde Valley sub-basin increased by about 295,500 ac-ft/yr since before development through 2005 (table 2) as a result of greater than average estimated recharge rates from 1970 to 2000. Simulated storage during 2005 was decreasing, however, at a rate of about 30,200 ac-ft/yr as a result of the combination of groundwater withdrawals and below average recharge rates (table 2). Little net change in groundwater flow between adjacent sub-basins resulted from the transient simulation.

Water-level records at seven wells were selected as representative of the groundwater system response to groundwater withdrawals in the Verde Valley sub-basin (figs. 17, 28). Five of the wells are in the alluvial basin—(A-13-05)05BDC, (A-14-05)E17AAC, (A-15-03)12ADB1, (A-15-04)04DDC1, and (A-16-03)22DCD—and primarily represent water-level variations in the confined parts of the Verde Formation and underlying conglomerate. Observed water levels at these wells are compared with the simulated heads in model layers 1 and 2 because of the large vertical hydraulic gradients in the area. Two wells—(A-17-06)E30BBB and (A-14-10)32DBD—were selected to represent water-level variations in the unconfined part of the Coconino aquifer, which is simulated as layer 2. No long-term water-level records are available for the Redwall-Muav aquifer, layer 3. Some long-term water-level records are available for the shallow aquifer overlying the thick silt and clay layer; however, the shallow aquifer is not explicitly represented in the numerical flow model.

Long-term water-level records in the alluvial basin part of the Verde Valley are few; however, available records

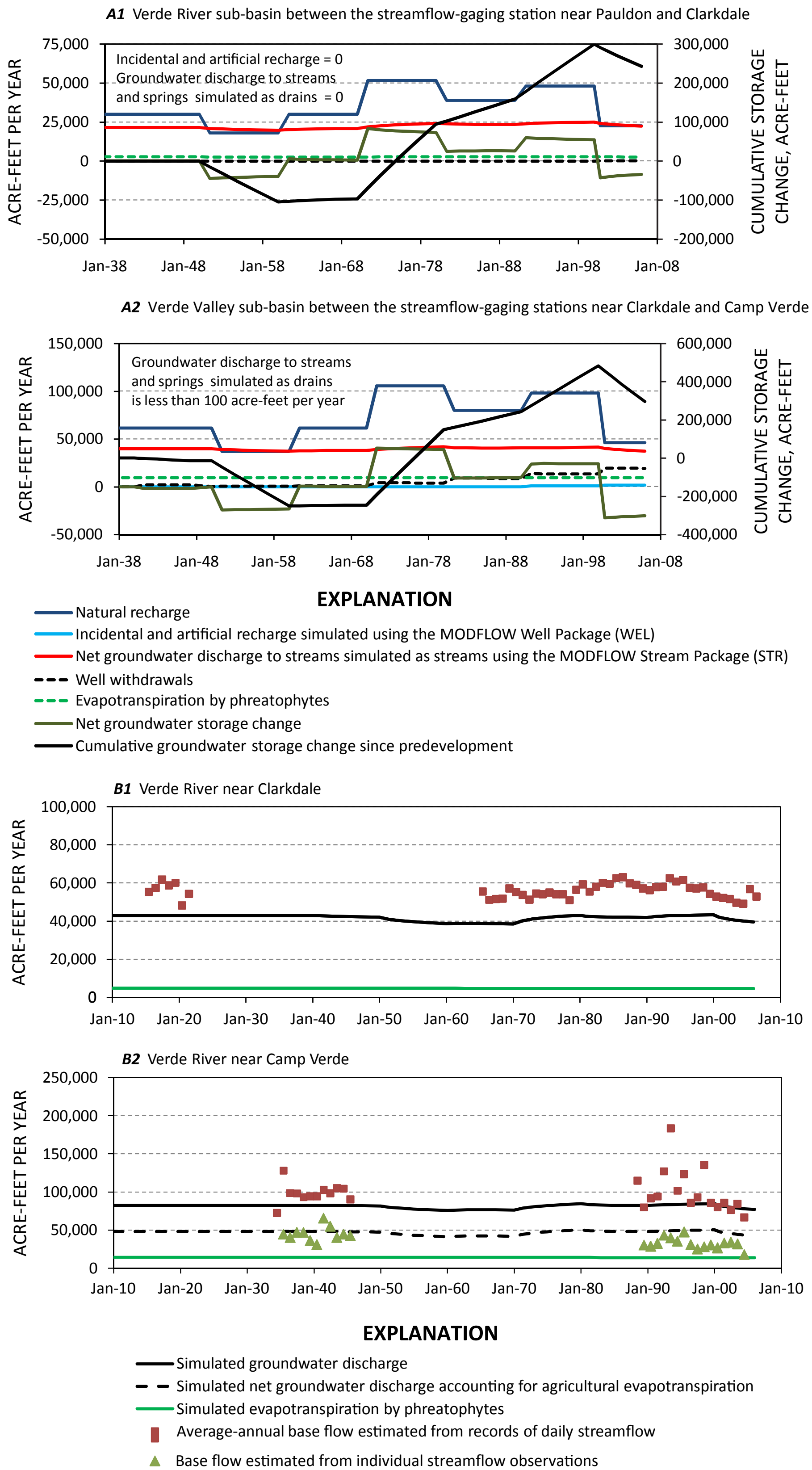


Figure 27. Simulated groundwater budgets for the Verde Valley sub-basin between the streamflow-gaging stations near (*A1*) Paulden and Clarkdale, (*A2*) Clarkdale and Camp Verde, and simulated and estimated base flow at (*B1*) Verde River near Clarkdale and (*B2*) Verde River near Camp Verde.

Figure 28. Hydrographs of simulated and observed water levels at selected wells in the Verde Valley sub-basin.

indicate moderate water-level declines of 0–30 ft in several areas and declines of tens of feet in the Lake Montezuma area. No major water-level change or large variations in water levels have happened in some areas. Little overall trend in water levels is evident in the southern part of the alluvial basin based on water-level records at wells (A-13-05)05BDC and (A-14-05)E17AAC (fig. 28). Water levels at well (A-14-05)E17AAC have varied more than 40 ft during the period of record, 1958–2006; however, no clear trend is observed. Much of the observed water-level variation may be due to variations in nearby withdrawals. Simulated water levels at the two wells also did not indicate any major trends except for a downward trend of a few feet after 1980. Simulated hydraulic head differences of about 30 and 5 ft between model layers at wells (A-13-05)05BDC and (A-14-05)E17AAC, respectively, indicate upward hydraulic gradients and groundwater flow from layer 2 to layer 1 at both wells. Water levels at both wells are representative of the Verde Formation that was simulated as model layer 1; however, vertical differences in water levels of several feet are likely at different depths in the Verde Formation. Simulated water levels in the unconfined layer 1 best represented the water levels at both wells.

Water-level records near the confluence of the Verde River and Oak Creek indicate water-level decline of 20–25 ft. Slightly different trends were indicated, however, at wells (A-15-03)12ADB1 and (A-15-04)04DDC1 (fig. 28), which tap the Verde Formation at depths of 400 and 250 ft, respectively. Water levels at well (A-15-03)12ADB1 declined about 20 ft during 1958–90 followed by a period of variable water levels and little overall decline. In contrast, water levels at well (A-15-04)04DDC1declined about 20 ft during 1958–75 followed by an extended period of little decline. Simulated water-level declines at the two wells do not match the magnitude or timing of variations in the observed water-level records. Simulated decline at well (A-15-03)12ADB1 was more than 100 ft greater than observed in model layers 1 and 2. At well (A-15-04)04DDC1, however, very little decline was simulated before 1990. Simulated hydraulic head in layer 2 was substantially greater than layer 1, 45–75 ft at both wells, which indicates the potential for upward groundwater flow from layer 2 to layer 1 and the need for knowledge of the vertical distribution of groundwater withdrawals. The poor simulation of the timing and magnitude of water-level changes at both wells may be due to uncertain withdrawal rates and locations, including vertical distributions of withdrawals. In addition, the simulated hydraulic connection between layers in this area possibly may be too restricted. However, improved information on vertical variations in water levels with depth in the Verde Formation could better constrain the simulated vertical distribution of hydraulic head.

Water-level records in the northern part of the alluvial basin near Cottonwood at well (A-16-03)22DCD indicate water-level decline of about 30 ft during 1960–2006. Water levels at the 1,400-ft-deep well represent model layer 2 (fig. 28). Simulated water-level declines in layer 2 at the well (layer 1 is not present at the well) are substantially less, about 5 ft. Observed and simulated declines at the well are likely related to the development of a cone of depression several miles to the southeast. The poor simulation of the magnitude of water-level decline at the well may be due to simulated hydraulic conductivity values that are too low in the region. Higher simulated hydraulic conductivity values would have resulted in a more broad and shallow cone of depression than was simulated and greater water-level decline at well (A-16-03)22DCD.

Water-level records in the Coconino aquifer near Sedona at well (A-17-06)E30BBB indicate very little water-level variation before about 1995 followed by about 15 ft of decline (fig. 28). Simulated water-level declines in model layer 2 of about 30 ft at the well are greater than observed and begin much earlier, about 1970 (layer 1 is not present at the well). The simulated water-level altitude at the well is much greater than observed, about 4,010 ft in 1960 compared to the observed altitude of about 3,860 ft. The poor simulation of hydraulic head reflects uncertainty in the local hydrogeologic system. Reasons for uncertain hydraulic head in the area include local hydrogeologic complexity and variable rates of streamflow losses in nearby Oak Creek, which result in variable water levels at wells. Multiple permeable zones may exist in the Coconino aquifer in the area, including faults and fractures that result in barriers or conduits for groundwater flow. In addition, multiple permeable layers may exist in the Supai Formation in the area. A conceptual model of the local hydrogeologic system that includes multiple permeable layers is supported by the presence of perched water tables in the area including at well (A-17-06)E30BBB, where cascading water has been observed (ADWR GWSI records) at about the altitude of the simulated water level.

Water-level variations in the Coconino aquifer in the far eastern part of the Verde Valley generally correspond with variations in recharge rates (fig. 28). Water levels at well (A-14-10)32DBD indicate a water-level rise of about 20 ft during about 1970–95. Because there has been little development in this area, the trend is probably natural. This natural water-level trend in model layer 2 at well (A-14-10)32DBD was simulated as a result of variations in simulated recharge rates (layer 1 is not present at the well).

Simulated water-level trends in this sub-basin were compared with observed trends from 265 water-level measurements at seven wells. Observed changes averaged about 2.1 ft of decline and ranged from about 35 ft of decline to about 34 ft of water-level rise. The average and average absolute differences between simulated water-level changes and corresponding observed changes were about 0 ft and 8.2 ft, respectively (table 3).

Little Colorado River Plateau Basin

Simulation of the transient groundwater-flow system for the Little Colorado River Plateau basin for 1938–2005 is evaluated through comparison of simulated and observed changes in water-level altitude at wells with long-term records.

Simulated variations in groundwater discharge to streams are discussed but can not be evaluated against observations because no long-term records are available at streamflow-gaging stations. Uncertainty in the vertical distribution of well withdrawal rates and screened intervals of wells has a major influence on the evaluation of simulated changes in some areas.

Groundwater budgets for the simulated Little Colorado River Plateau basin indicate that changes in the groundwater system have been dominated by variations in recharge rates and to a lesser amount by variations in groundwater withdrawals (table 2, fig. 29*A*). Rates of simulated incidental and artificial recharge varied from about 7,000 ac-ft/yr during the 1960s to about 49,800 ac-ft/yr during 2000–2005 (table 2, fig. 29*A*). Simulated natural recharge rates were at or below average, about 206,600 ac-ft/yr, before 1970 and after 2000. Above average natural recharge rates during 1970–2000, 268,600 to 355,400 ac-ft/yr, resulted in storage recovery with the greatest groundwater storage happening in 2000. Below average rates of natural recharge after 1999 resulted in storage loss at the rate of about 85,000 ac-ft/yr. Total change in groundwater storage from before development to 2005 was an increase of 669,500 ac-ft (fig. 29*A*). Simulated groundwater withdrawal rates increased

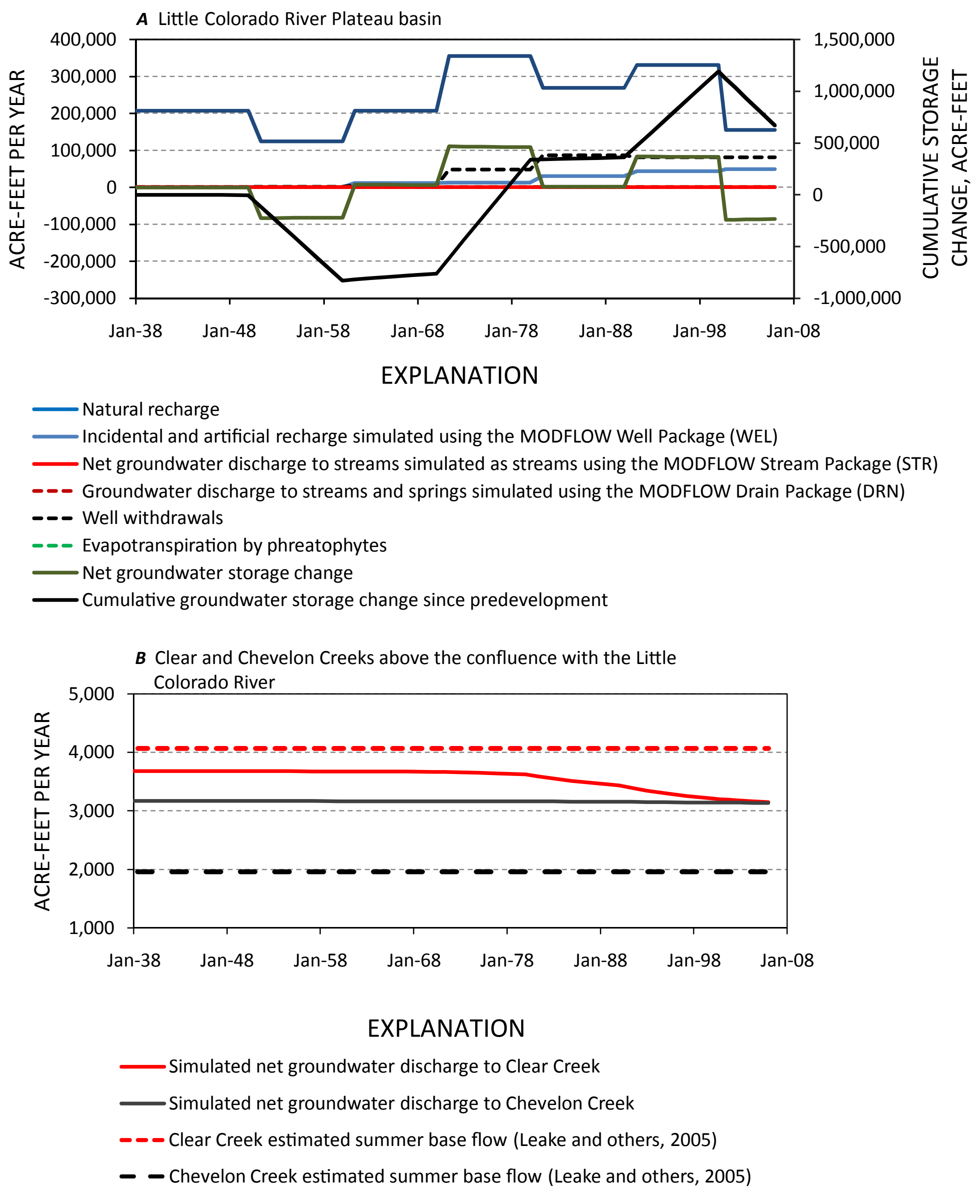


Figure 29. Simulated (*A*) groundwater budget for the Little Colorado River Plateau basin, and (*B*) estimated base flow at Clear and Chevelon Creeks.

from less than 7,000 ac-ft/yr before 1970 to more than 80,000 ac-ft/yr after 1980. Simulated net groundwater discharge to the Little Colorado River and tributaries (fig. 29*A*) was zero because all of the streamflow that discharged to streams reinfiltrated within the basin. Local withdrawals and storage losses, however, resulted in diminishment of simulated groundwater discharge to Clear Creek from about 3,700 ac-ft/yr before about 1970 to about 3,100 ac-ft/yr in 2005 (fig. 29*B*). A slight dimishment of groundwater discharge to Chevelon Creek was also simulated (fig. 29*B*).

Water-level records at eight wells were selected as representative of the groundwater system response to groundwater withdrawals in the Little Colorado River Plateau basin (figs. 16, 30). Most records indicate some groundwater level decline after about 1970. Slight water-level rises at wells (A-23-08)21AAD and (A-15-12)15DDC after 1970 were reproduced by the simulation as a result of the simulated variations in recharge rates including above average recharge rates from 1970 to 1990. Water-level declines of 0-40 ft at five wells—(A-19-16)36DBB, (A-18-14)13ABD3, (A-17-21)10CBA, (A-13-21)34DCC2, and (A-10-22)30ABA—were approximately simulated. Simulated water-level decline of about 200 ft at well (A-20-08)18BBB is much greater than observed declines of less than 100 ft. Most of the poor simulation of water-level decline at the well is because of short periods of water-level recovery of about 50 ft from 1978 to 1984 and from 1991 to 1996 that were not simulated. The short-term trends may result from brief periods of recharge shorter than a decade, inaccurate simulation of the withdrawal location because of spatial discretization of the model domain, or local variations in withdrawal rates of less than a decade. Well (A-20-08)18BBB is near the Lake Mary Well Field, a major water supply for Flagstaff, which receives local recharge from infiltration at Lake Mary. This local recharge source was not simulated, which is a likely cause of the greater than observed rates of water-level decline. Major local variations in recharge rates that were not simulated also occur in other parts of the Little Colorado River Plateau basin, especially near major ephemeral streams including along reaches of the Little Colorado River. Large vertical hydraulic head gradients (downward) developed in one area of the Little Colorado River Plateau basin near well (A-13-21)34DCC2 during the transient simulation. The accuracy of the simulated vertical gradients can not be confirmed because of inadequate information on the vertical distribution of well withdrawals and water levels in wells.

Simulated water-level trends in this basin were compared with observed trends from 412 water-level measurements at eight wells (fig. 30). Observed changes averaged about 5 ft of decline and ranged from about 59 ft of decline to about 35 ft of water-level rise. The average and average absolute differences between simulated water-level changes and corresponding observed changes were about -11 ft and 18 ft, respectively (table 3).

Coconino Plateau Basin

Simulation of the transient groundwater-flow system for the Coconino Plateau for 1938–2005 is evaluated through comparison of simulated and observed changes in water-level altitude at wells with long-term records. Simulated variations in groundwater discharge are discussed but can not be evaluated against observations because no long-term records are available at streamflow-gaging stations.

Groundwater budgets for the simulated Coconino Plateau basin indicate that changes in the groundwater system are dominated by variations in recharge rates (fig. 31). Simulated recharge rates were at or below average, about 152,000 ac-ft/yr, before 1970 and after 1990. Above average recharge rates during 1970–2000, 197,600 to 261,500 ac-ft/yr, resulted in storage recovery with the greatest groundwater storage in 2000. Groundwater storage had increased since before development by 1,275,100 ac-ft by the end of 2005 despite losing storage at a rate of about 54,000 ac-ft/yr during 2005 (table 2). In comparison, withdrawal rates are less than 1,300 ac-ft/yr throughout the transient simulation period, more than two orders of magitude less than simulated recharge rates. Simulated groundwater discharge to springs and streams simulated as drains varied slightly with variations in recharge rates from 151,400 ac-ft/yr before 1940 to 147,900 ac-ft/yr during the 1960s and 164,200 ac-ft/yr in 2000.

No long-term water-level records are available to evaluate changes to the groundwater-flow system in the Coconino Plateau basin. However, long-term records are available at two wells near the groundwater flow divide in the Big Chino sub-basin, well (B-21-02)14BCC, and Peach Springs basin, well (B-21-02)14BCC (figs. 16, 32). Data from well (B-21-02)14BCC, a public-supply well, indicates no water-level trend during 1974–2002 in comparison to simulated decline of more than 80 ft. The stable observed water-level record at the well is not possible to simulate using the estimated withdrawal distribution and rates or using the simulated conceptualization of groundwater-flow system in the area of the well. Data from the other well, (B-25-09)26DBC, an unused well, indicates a water-level trend that generally mimics variations in recharge rates for 1960–2004. An observed water-level recovery of about 45 ft during 1960–80 was simulated as a recovery of about 30 ft during 1970–2000.

Western Basins

Simulation of the transient groundwater-flow system for the western basins—Truxton Wash watershed and Fort Rock and Burro Creek sub-basins—for 1938–2005 is difficult to evaluate because of a lack of observed long-term changes in stream base flow and water-level records at wells. Streamflow records are available for streamflow-gaging stations at Burro Creek near Bagdad and Truxton Wash near Valentine. Continuous records are short, however, for each streamflow gage. At Burro Creek, records are available for 1980–93 and after 2003. At Truxton Wash, records are

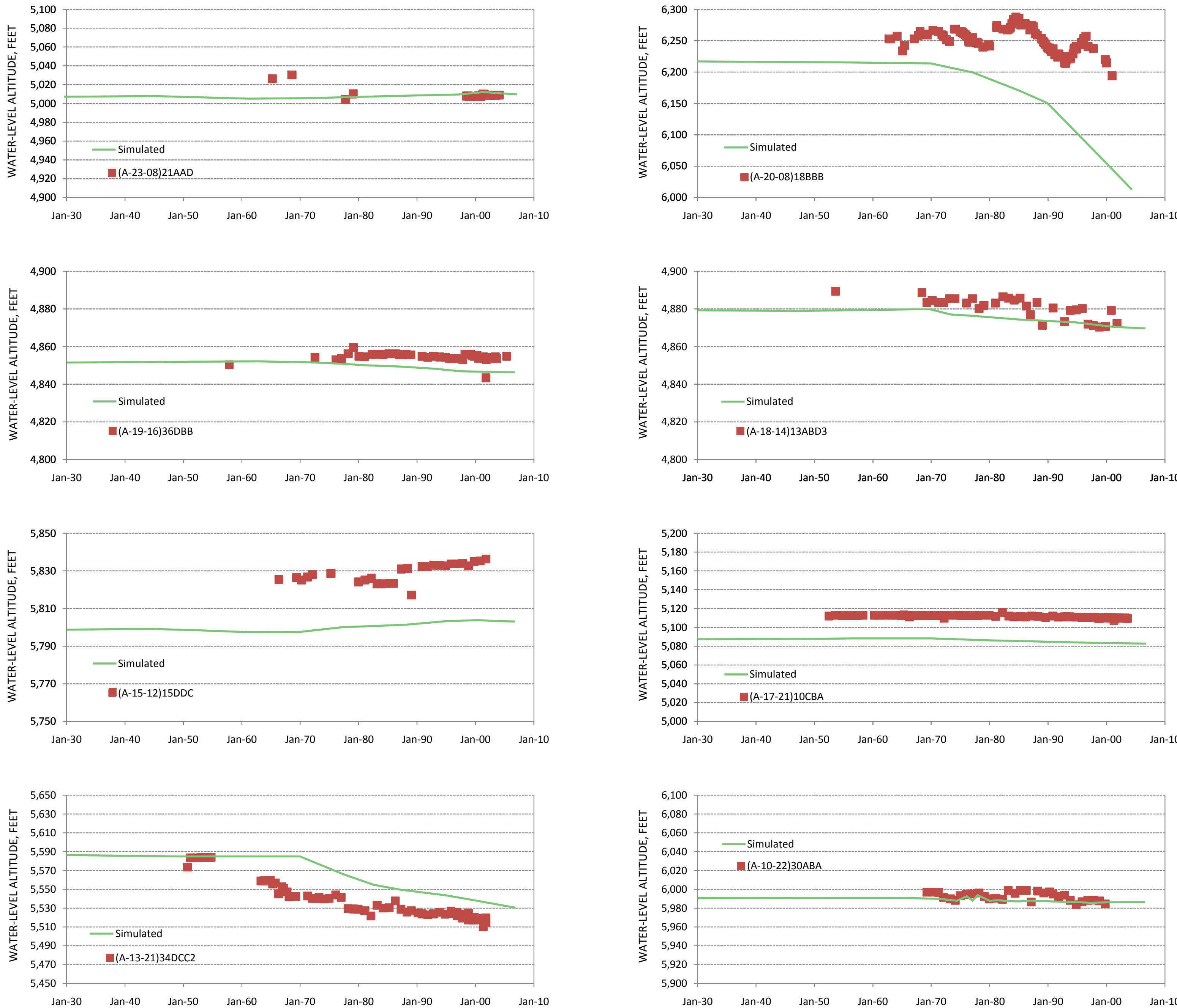


Figure 30. Hydrographs of simulated and observed water levels at selected wells in the Little Colorado River Plateau basin.

available for 1993–2001 and after 2003. Simulated variations in groundwater discharge are, therefore, discussed but can not be evaluated.

Similar to the simulated transient groundwater systems in many other regions included in the numerical groundwater-flow model, the simulated groundwater budgets for the western basins indicate that changes in the groundwater system are dominated by variations in recharge rates (fig. 33*A*). Simulated recharge rates were at or below average, about 41,900 ac-ft/yr, before 1970 and after 2000. Above average recharge rates during 1970–2000, 54,400 to 72,000 ac-ft/yr, resulted in storage recovery with the greatest groundwater storage happening in 2000. Groundwater storage was simulated as decreasing during 2005 at a rate of about 14,600 ac-ft/yr as a result of below average recharge rates after 2000. The amount of groundwater in storage was about 325,200 ac-ft by the end of 2005, more storage than before development (table 2, fig. 33*A*). In comparison, groundwater withdrawal rates in the western basins were less than 600 ac-ft/yr throughout the transient simulation period. Simulated groundwater discharge to springs and streams, simulated as groundwater discharge to drains, varied slightly with variations in recharge rates, from 37,800 ac-ft/yr before 1940 to a maximum of 45,100 ac-ft/yr during 2000.

Verde Canyon Sub-Basin

The transient groundwater-flow system for the Verde Canyon sub-basin during 1938–2005 can not be evaluated because of a lack of long-term water-level and long-term streamflow records for the largest discharge source, Fossil Springs. Simulated variations in groundwater discharge, however, are discussed.

Simulated groundwater budgets for the Verde Canyon sub-basin indicate that changes in the groundwater system are dominated by variations in recharge rates (fig. 33*B*). Simulated recharge rates were at or below average, about 35,200 ac-ft/yr, before 1970 and after 2000. Above average recharge rates during 1970–2000, 45,700–60,500 ac-ft/yr, resulted in storage recovery with the greatest groundwater storage occurring in 2000. Groundwater storage was simulated as decreasing during 2005 at a rate of about 15,200 ac-ft/yr (table 2) as a result of below average recharge rates after 2000. By the end of 2005, the amount of groundwater in storage was about 201,200 ac-ft more than before development (table 2, fig. 33*B*). Groundwater withdrawal rates are small in comparison to rates of recharge and storage change, about 1,700 ac-ft/yr or less, throughout the transient simulation period. Simulated groundwater discharge to springs and streams (net streamflow) varied slightly with variations in recharge rates from 54,600 ac-ft/yr before 1940, to a minimum of less than 50,000 ac-ft/yr during the 1960s, and a maximum of about 67,800 ac-ft/yr during the 1990s.

Salt River Sub-Basins

Simulation of the transient groundwater-flow system for the three Salt River sub-basins upstream from the streamflow-gaging station near Roosevelt—Salt River Canyon, Black River, and White River sub-basins—can be evaluated through comparison of simulated groundwater discharge and groundwater discharge that is estimated from streamflow records at the gaging station near Roosevelt for 1913–2005. Long-term water-level records at wells in

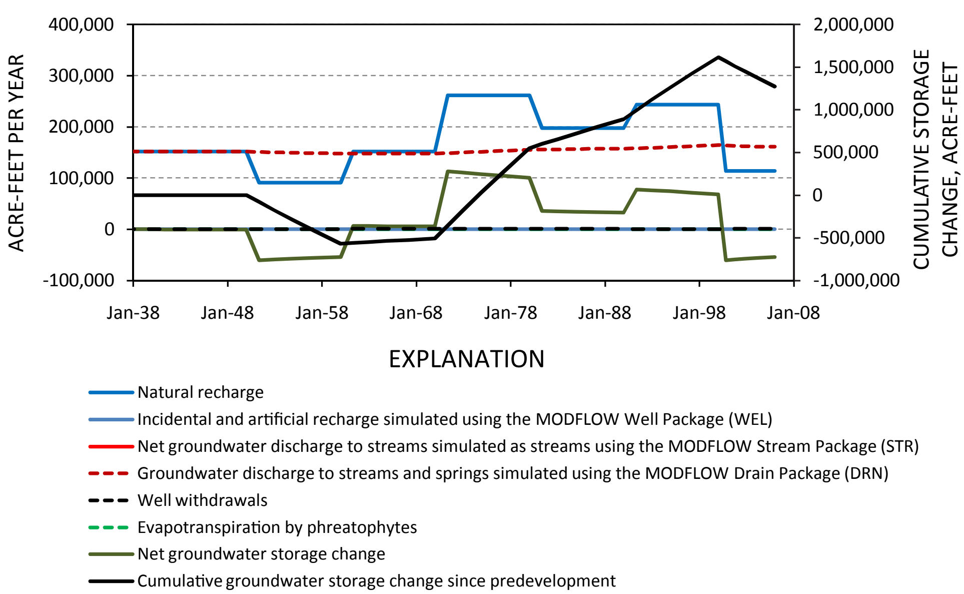
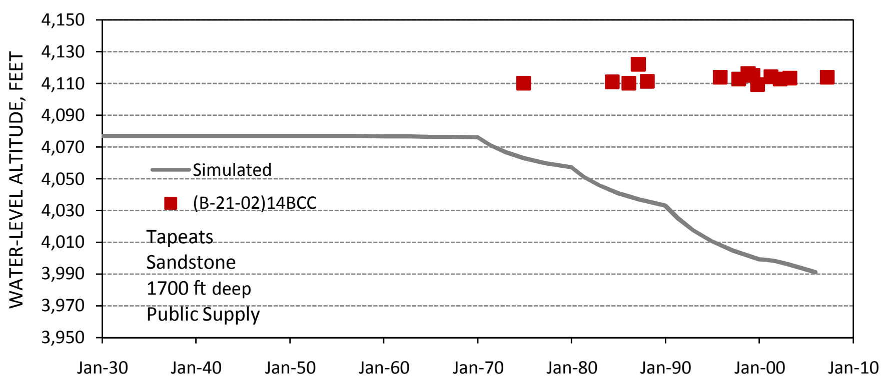


Figure 31. Simulated groundwater budget for the Coconino Plateau basin.

the area are unavailable, however. Similar to the simulated transient groundwater systems in many other regions included in the numerical groundwater-flow model, the simulated groundwater budgets for the Salt River sub-basins indicate that changes in the groundwater system are dominated by variations in recharge rates (fig. 33*C*1). Simulated recharge rates were at or below average, about 177,800 ac-ft/yr, before 1970 and after 1990. Above average recharge rates during 1970–2000, about 231,100 to 305,800 ac-ft/yr, resulted in storage recovery with the greatest groundwater storage happening in 2000. Groundwater storage was simulated as decreasing during 2005 at a rate of about 58,400 ac-ft/yr (table 2) as a result of below average recharge rates after 2000. At the end of 2005, the amount of groundwater in storage was about 1,205,600 ac-ft more than before development (table 2, fig. 33*C*1). Simulated groundwater discharge to springs and streams simulated as groundwater discharge to drains changed with variations in recharge rates from 188,400 ac-ft/yr before 1940, to a minimum of 170,100 ac-ft/yr in 1960, and a maximum of 230,700 ac-ft/yr in 2000 (fig. 33*C*1). Groundwater withdrawal rates were about 500 ac-ft/yr or less throughout the transient simulation period, much less than rates of recharge or groundwater discharge to springs and streams. Groundwater withdrawals in the basin were therefore a minimal influence on simulated and observed hydrologic records.

Winter low flow at the streamflow gage at the Salt River near Roosevelt indicates a trend in groundwater discharge to streams that is similar to simulated rates but varies more greatly and more frequently than the simulated stress periods of 10-year duration (fig. 33*C*2). Trends in groundwater discharge before 1950, which declined from more than 200,000 ac-ft/yr around 1920 to about 158,000 to 170,000 ac-ft/yr during 1925–48, are not simulated because no variations in recharge rates were simulated during that period. An approximation of the pre-1940 variations in recharge is needed to better simulate changes in the groundwater-flow system during that time period. Simulated decadal and longer trends in groundwater discharge after 1940 were similar to winter low-flow records including a decline in flow that occurred with below-average recharge during the 1950s and increasing flow during the 1970s. Periods of maximum winter low flow during the 1980s were not simulated, but the low-flow records likely include significant amounts of runoff that contribute to the extended period of elevated low flows. Trends in winter low flow of less than a decade in length are common in the record, but were not simulated in the numerical model. A simulation that varies recharge rates at



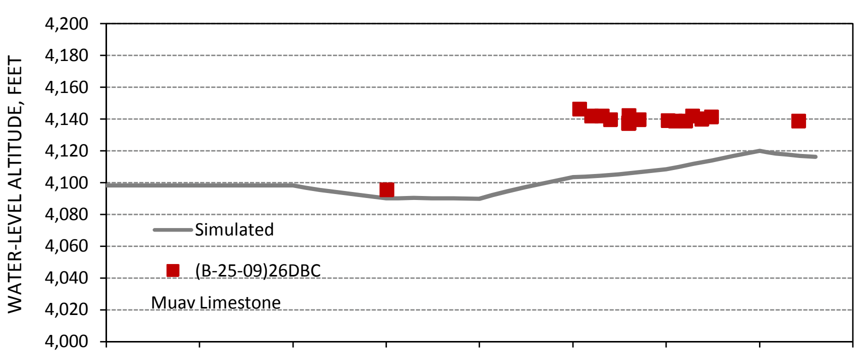


Figure 32. Hydrographs of simulated and observed water levels at selected wells in the Coconino Plateau basin.

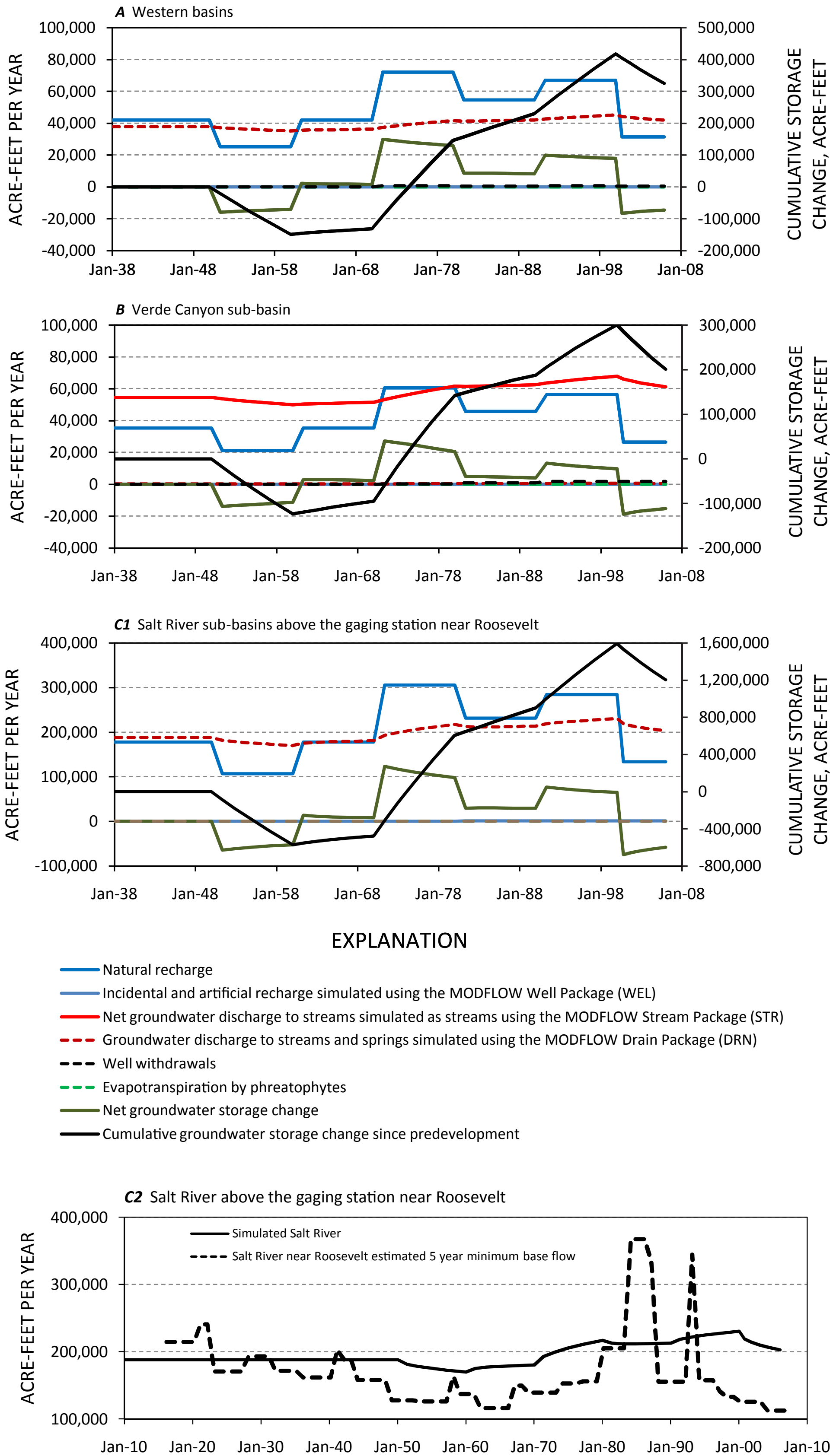


Figure 33. Simulated groundwater budgets for the (*A*) Western basins, (*B*) Verde Canyon sub-basin, (*C1*) Salt River basin above the streamflow-gaging station near Roosevelt, and (*C2*) estimated base flow discharge at the Salt River near Roosevelt.

smaller time intervals is needed to reproduce the frequency and magnitude of the estimated variations in base flow.

Model Applicability, Limitations, and Suggestions for Future Work

The numerical groundwater model has two primary uses: (1) evaluation of the hydrologic effects of groundwater use on the groundwater-flow system and (2) identification of major hydrogeologic parameters that need improved definition. The model can provide information that should help identify data needs and refine future studies for improved simulation of the hydrologic effects of groundwater use. The model also can be used to estimate changes in the water levels and discharge to streams, springs, and riparian evapotranspiration that may result from anticipated future groundwater use and management practices. However, the certainty of projected change is dependent on future validation of the hydrologic assumptions that are inherent in the model.

For the purpose of estimating magnitude and timing of change in water levels and discharge resulting from an imposed stress using this or any groundwater-flow model, the only hydrologic parameters that are of importance are the aquifer properties of transmissivity and storage (Leake, 2011). Those properties influence the rate of propagation of changes in groundwater flow through the aquifer and release of groundwater from storage. Variations in recharge rates—natural, artificial, or incidental—can cause change in water levels and discharge to streams, but are not fundamental variables that affect the calculation of human-induced change in the model. Rather, the effects of recharge variations are independent of the effects of groundwater use and management practices and are superimposed on these other effects. The superimposed effects of variations in recharge rates can enhance or counter the effects of groundwater use and management practices. Other hydrologic parameters that have no influence on changes in water levels and discharge to streams include directions of groundwater flow and sources of water that contribute groundwater discharge to streams.

The groundwater-flow model itself, however, must consider properties and conditions that are not essential in computing effects of human activities on an aquifer system. In constructing the flow model, an attempt was made to reasonably represent (1) spatial and temporal recharge distributions, (2) transmissivity distributions, (3) distributions of hydraulic properties that control interactions between vertically adjacent aquifers, (4) spatial distributions of withdrawals and incidental recharge rates, (5) aquifer lateral and vertical extents, and (6) hydrologic barriers and conduits. It should be noted that estimation of item 2, transmissivity distributions, using the model requires independent knowledge of item 1, recharge distributions. Although this study used the most comprehensive information on these items currently available, future studies may result in an improved understanding of the groundwater-flow system that could substantially alter the fundamental conceptual hydrogeologic model in some areas.

Appropriate uses of the model are limited to sub-basin-scale or larger applications and to regions of the model where geologic and hydrologic data are available, including long-term records of transient changes in water levels, base flow, and withdrawals. Basin-scale applications refer to the simulation of overall water budgets and groundwater flow in a groundwater basin or large parts of basins. Few hydrologic records that document withdrawals and major changes in the hydrologic system are available across broad regions of the study area, resulting in uncertain simulated changes in those areas. Differences between the simulated and observed conditions at individual wells, springs, and stream reaches were included as measures of simulation accuracy. However, accurate simulation of observed water levels or stream base flow should not be interpreted as an indicator of accurate simulation of the local flow system. Many modifications to local hydrologic parameters or even major modification to the hydrogeologic conceptual flow system could reproduce similarly accurate simulations of local observations that are different from the true system and could result in inaccurate projected changes. The simulated hydrologic system should not be considered a replication of the true system, but a simulation of the system as currently understood by the modeler and simplified for simulation by using a numerical model.

Several observations were made during the construction of the groundwater model that, if incorporated into a future version, would improve simulations. The accuracy of simulations in many groundwater basins could be improved by acquiring more hydrologic data, including additional streamflow discharge information and water-level records at more wells, improved definition of vertical groundwater flow between vertically adjacent aquifers, improved definition of recharge distributions, improved definition of storage properties, improved information on geologic structures that may be hydrologic-flow barriers or conduits, and improved water-use information. Additional streamflow monitoring would better define ephemeral channel recharge and base-flow distributions. Additional water levels at wells would better define hydraulic gradients and allow improved simulation of transmissivity. Groundwater likely flows vertically between aquifers in many areas, but this flow component is poorly defined throughout most of the study area. Vertical groundwater flow could be better defined by the development of water-level records at colocated deep and shallow wells that monitor different aquifers or permeable zones. Broad assumptions were used to distribute recharge throughout the model. Better information defining recharge distributions, especially ephemeral channel recharge, would result in improved transmissivity distributions. Reasonably accurate storage and transmissive properties are needed for the proper simulation of the effects of withdrawals on water levels and groundwater discharge to streams and springs. Storage properties are a difficult parameter to obtain; however, estimates that are only as accurate as a half-order of magnitude would be an improvement over existing information for some areas. Better definition of the extents of confined and unconfined conditions of groundwater flow are needed for improved storage coefficient distributions in many areas. Extents of individual aquifers and confining beds, including extents of silt and clay facies of basin fill, are not well defined in several areas. Accurate understanding of the aquifer boundaries is most important in the alluvial basin areas, but is locally important on the Colorado Plateau where the primary aquifers are much more extensive. Many geologic structures, such as faults and fractures, are hydrologically influential features that may act as barriers to groundwater flow that hydraulically separate regions of groundwater flow or may act as conduits that readily transmit groundwater across large distances and between aquifers. However, the hydraulic information that could verify the influence of most of these structures is largely unavailable. Accurate water-use information including well withdrawals and incidental and artificial recharge rates are needed along with water-level and streamflow records to accurately define the response of the aquifer system to withdrawals.

Each basin and sub-basin within the groundwater-flow model has hydrologic data deficiencies that could be minimized by long-term monitoring and future investigations. The primary data deficiencies for each major basin and sub-basin are discussed below.

Hydrologic data deficiencies in the Big Chino sub-basin include a lack of long-term hydrologic records, poorly defined interactions between multiple aquifers, a lack of defined recharge distributions, poorly defined storage property distributions, and poorly defined aquifer and confining unit extents in some areas. Long-term records are needed that document water levels at wells that tap the major water-yielding zones. Unfortunately, water levels have not been collected continuously at many wells since early development of the groundwater resources. Interactions among the major aquifers are poorly defined in some influential areas, including the Paulden area where the major water-bearing zones include the Redwall-Muav aquifer, lower basin fill, and upper basin fill including interbedded basalt flows, and the upper part of the basin where interactions between the Redwall-Muav aquifer and basin-fill aquifers are poorly understood. The cause of steep hydraulic gradients near Walnut Creek is not well understood, but may be related to extents of aquifers or the fine-grained facies of basin fill. A reasonably accurate understanding of the recharge distributions influence the simulated distributions of transmissivity and, thereby, also influence simulated rates of changes in water level and stream base flow. The source of water that recharges the aquifer system in the Big Chino sub-basin has been partly defined by using geochemical data (Blasch and others, 2006); however, the location of the recharge, especially ephemeral stream channel recharge, has not been defined. Aquifer-storage properties are poorly defined throughout the basin and have an influence on simulated rates of change in water level and stream base flow that is equivalent to the influence of transmissivity distributions.

The primary hydrologic data deficiency in the Big Chino sub-basin is a lack of long-term records that could document the response of the aquifer system to changes in withdrawals and recharge. Without these records, numerical simulations of the system response, including changes in water levels and stream base flow, are not well constrained. A good dataset to define response of each aquifer to changes in rates of withdrawal and recharge and interactions among the primary aquifers is water-level records at multiple colocated wells that monitor the major permeable zones—Redwall-Muav, lower basin fill, upper basin fill, and the basalt zone near Paulden. Streamflow records began at the gaging station near Paulden in 1964. Earlier records would have captured any stream base- flow response that may have accompanied early development and water-level change and climate variability. Aquifer-system monitoring in the Big Chino sub-basin has substantially improved recently with the addition of water-level monitoring at many wells by ADWR and USGS in cooperation with Yavapai County and the City of Prescott. Streamflow records are being collected by USGS at Williamson Valley Wash and Verde River near Paulden and by private agencies on the Verde River near Paulden. These data-collection sites should be maintained with annual or more frequent measurements for a sufficient period of time to document changes in the system that may result from variations in withdrawal or recharge. Loss of these future long-term records would result in little improvement in understanding of the hydrologic system.

Hydrologic data deficiencies in the Little Chino and Upper Agua Fria sub-basins, PrAMA, include poorly defined interactions between the lower and upper aquifers, poorly defined ephemeral channel recharge distributions, and poorly defined storage property distributions. The ADWR is monitoring water levels at an impressive array of well sites in the PrAMA that have yielded a better understanding of the hydrologic system. Extents and degree of the hydraulic connection between upper and lower aquifers could be better defined in the PrAMA by monitoring colocated wells that tap both aquifers. Substantial ephemeral channel recharge has been identified in the area and estimated on the basis of scant data (Corkhill and Mason, 1995; Nelson, 2002). Ephemeral channel recharge rates and variations in rates are, however, poorly defined because of insufficient data. Likely ephemeral channel recharge areas have been identified recently through the ADWR monitoring network by large annual or more frequent water-level variations near major channels. Unfortunately, quantification of recharge rates from available data is highly uncertain because extents of the recharge mounds and specific yield values are poorly defined. Recharge rates at each ephemeral channel could be better quantified by using more frequent or continuous water-level monitoring along with colocated observations of water-mass change by using gravity methods. Aquifer storage properties, including extents of confined and unconfined groundwater in each aquifer and perched aquifers, are poorly defined. Additional monitoring of water-mass change by using colocated gravity monitoring at monitor wells and other areas could improve estimates of specific yield and extents of confined and unconfined groundwater (Pool, 2008).

Hydrologic data deficiencies in the Verde Valley sub-basin include a lack of long-term hydrologic records including records of surface-water diversion, poorly defined stream-aquifer interactions including groundwater discharge to the Verde River and tributary streams, poorly defined interactions between multiple aquifers including perched aquifers and multiple permeable zones in the Verde Formation, poorly defined extents of lithologic facies and permeable zones in the alluvial basin aquifer, poorly defined recharge distributions, poorly defined storage property distributions, and uncertain hydrologic influence of geologic structures. The use of surface-water diversions for agricultural use has resulted in complex hydrologic interactions between shallow groundwater in the Quaternary alluvium and streams. As a result, separation of the base flow and runoff components of surface-water flow in the Verde River and tributaries is difficult to estimate. Understanding of this shallow flow system and stream-aquifer interactions could be improved with detailed monitoring of diversions, agricultural water use, water levels in the shallow aquifer, and streamflow. The alluvial aquifer system in the Verde Valley sub-basin includes poorly defined extents of multiple highly permeable zones that are interbedded with multiple confining beds. This highly heterogeneous aquifer system has resulted in a complex groundwater-flow system that includes large head differences among wells that tap multiple zones in local areas. Detailed and frequent monitoring of withdrawals and water levels or multiple aquifer tests would be required to better understand the groundwater-flow system in regions of the aquifer. A more general understanding of the groundwater-flow system and distributions of storage properties could be developed through improved mapping of extents of confined and unconfined conditions that coincide with intersection of the piezometric surface and the interface of the sand and gravel and fluvio-lacustrine facies of the Verde Formation (Twenter and Metzger, 1963). Although ephemeral channel recharge in the Verde Valley may be limited by large extents of confining beds in the Verde Formation, substantial recharge may exist along losing reaches of streams that flow over permeable Supai Group rocks, sand, and gravel facies of the Verde Formation, and permeable fractures, faults, and collapse structures. Recharge through stream-channel infiltration occurs along Oak Creek downstream from Sedona. Recharge likely occurs along losing reaches of other streams, but is not quantified. The hydrologic importance of geologic structures such as fractures, faults, and collapse structures in each of the primary aquifers in the Verde Valley is not well understood. A better understanding requires detailed monitoring of groundwater systems near these structures, including discharge from streams and springs and water-level monitoring in wells.

Hydrologic data deficiencies in the Little Colorado River Plateau and Coconino Plateau basins include a lack of water-level data in some areas, a lack of long-term hydrologic records, poorly defined interactions between the Redwall-Muav and Coconino aquifers, poorly defined storage property distributions, and poor understanding of the hydrologic influence of geologic structures. Water-level data that define hydraulic gradients are lacking in areas between the Mogollon Rim and Little Colorado River, north of the Little Colorado River, and across broad areas of the Coconino Plateau. Long-term water-level records are available in a few areas near centers of groundwater withdrawal, but many other areas are lacking records, especially across the Coconino Plateau. Long-term streamflow records that document variations in groundwater discharge are lacking at perennial streams in the area. Monitoring should be continued at many areas of groundwater discharge that are currently monitored, such as the Little Colorado River upstream from the outflow to the Colorado River and Havasu Creek. The hydrologic influence of geologic structures in the Little Colorado River Plateau and Coconino Plateau basins is evident from variations in water-level gradients adjacent to major structures. However, the importance of many structures is poorly defined because hydraulic data are lacking.

Hydrologic data deficiencies in other basins, including the sub-basins along the Salt River, Tonto Creek sub-basin, and the western basins, generally include all the same deficiencies as the other basins. Long-term records and often basic records such as water levels and stream base flow are lacking for these areas. Locally, an understanding of aquifer extents, transmissivity, storage properties, and the hydrologic influence of geologic structures could be improved. The groundwater supplies in these areas are not extensively developed, and improved understanding of the hydrologic systems may not be locally influential. However, these hydrologic systems interact with adjacent aquifers such as in the Big Chino sub-basin and Little Colorado River Plateau basin. Therefore, the extents of groundwater-flow systems in the poorly developed basins can be essential to understanding the relations with adjacent groundwater-flow systems. In addition, discharge features in the poorly developed basins are available for capture by withdrawals in the adjacent and hydraulically connected developed basins. Capture of discharge in adjacent poorly developed basins may be at greater rates than capture of discharge features in the developed basins.

Athough varying availability of data describing aquifer characteristics and historical hydrologic changes limits the absolute accuracy of simulation results, the model was developed by using the best available information compiled through an extensive search of existing literature and databases. Accordingly, simulation results, particulary with respect to trends and differences, can serve a significant use in informing the decisions of resource managers so long as model limitations are considered. The benefits also include definition of data needs that can beneficially direct future studies.

Summary

A numerical flow model (MODFLOW) of the groundwater-flow system in the primary aquifers in northern Arizona was developed. Interactions between the aquifers, primary perennial streams, and springs were simulated for the predevelopment flow system and for the transient flow system for the period 1910 through 2005. Primary aquifers that were simulated included the Redwall-Muav, Coconino, and alluvial basin aquifers. In addition, regions of the study area that lacked the primary aquifers but may contribute or receive groundwater flow to and from the primary aquifers and perennial streams, including areas of crystalline rock, volcanic rocks, and pre-basin sediments, also were simulated, primarily as poorly permeable aquifers. Major simulated streams and springs that derive base flow from groundwater discharge from the primary aquifers included the Colorado River and tributary streams and springs; the Little Colorado River and perennial tributaries Chevelon, Clear, and Silver Creeks; the Salt River and tributaries; the Verde River and major tributaries Williamson Valley Wash, Sycamore Creek, Oak Creek, Wet Beaver Creek, West Clear Creek, Fossil Creek, and the East Verde River; and other small streams including Burro Creek, Trout Creek, Truxton Wash, Upper Agua Fria River, and several streams in northeastern Arizona. Many major springs that discharge groundwater from the primary aquifers also were simulated, including Del Rio Springs, Blue Spring, Havasu Springs, Verde River headwater springs, several springs adjacent to major Verde River tributaries such as Page Spring and Fossil Springs, and many springs that discharge to the Colorado River. Most of the major springs lie near major tributary streams and are often the primary source of streamflow. These springs were not simulated discretely, but as parts of the streams.

Several published reports that describe the hydrogeologic systems, groundwater-flow models, or other topics relevant to development of a numerical groundwater-flow model were reviewed. These documents provided basic information for the conceptualization of the hydrogeologic system-including hydraulic properties and groundwater budgets. Most of the basic information from a groundwater-flow model that was developed for the Prescott Active Management Area also was incorporated into the NARGFM, including tops and bottoms of aquifers and groundwater withdrawals.

A major task in the development of the numerical groundwater-flow model was the creation of a hydrogeologic framework model that defines the spatial extents of the primary aquifers and confining units. The creation of the framework model required the compilation of geologic data from GIS coverages of surficial geology, geophysical models, and well drilling information including drill logs, geologic logs, and geophysical logs. Surfaces defining the top and bottom altitudes of the major hydrogeologic units were developed from the intersection of the GIS geology coverages with land-surface altitude coverages, borehole data, and a groundwater-flow model for the PrAMA area. A hydrologic basement surface was developed from gravity and magnetic data. The surfaces that define the extents of the simulated hydrogeologic units, including hydrologic basement, Redwall-Muav aquifer, Coconino aquifer, and lower and upper basin fill, were transformed to the numerical model grid.

GIS datasets that describe the locations and altitudes of perennial and intermittent streams and springs were developed from existing GIS datasets including stream coverages, land-surface altitude, and GWSI datasets. The GIS stream and spring datasets were used to develop stream and drain datasets for the groundwater-flow model. Perennial and intermittent streams in the Verde River and Little Colorado River drainage systems were simulated as streams by using the STR Package.

Another task for model construction was the development of a natural recharge distribution. A Basin Characterization Method (BCM) was used to distribute recharge as a result of direct infiltration of precipitation that is in excess of ET demands and runoff. The method distributes monthly precipitation from 1971 to 2006 into multiple components of ET rates, soil storage, runoff, and recharge based on multiple attributes for 984-ft (300-m) grid cells including plant type, soils, slope, aspect, rock type, and solar energy. Monthly BCM estimates of runoff were calibrated against runoff observed at numerous streamflow gages. The BCM average annual recharge for 1971–2006 was found to underestimate predevelopment recharge in alluvial basins. This underestimate was not unexpected because direct infiltration of precipitation is not as great a contribution to overall recharge as in high altitude areas such as the Mogollon Rim. The BCM estimates of recharge in the Big Chino, Little Chino, Upper Agua Fria, and Verde Valley areas were augmented with recharge rates in areas of permeable sediments that balance the total estimated recharge with estimated predevelopment groundwater discharge as stream base flow and ET. Decadal variations in recharge rates were developed from the decadal BCM estimates for the period 1940–2005. The decadal variations in recharge rates resulted in improved simulation of water-level variations at wells in the alluvial basins and near the Mogollon Rim. Artificial recharge at wastewater treatment facilities and incidental recharge at agricultural fields and golf courses also was estimated for the transient groundwater-flow system.

Groundwater discharge to streams, springs, ET, and as well withdrawals was estimated on the basis of existing information from hydrologic records and available publications. Predevelopment rates of groundwater discharge to streams, springs, and ET rates were estimated where possible on the basis of streamflow records at gages and from previous reports. Estimates of groundwater discharge to streams, springs, and ET rates were used as observation datasets to compare with BCM recharge estimates and the numerical groundwater-flow model results. Groundwater withdrawal estimates were important datasets for model input. Annual withdrawal estimates from individual wells for agricultural, municipal, industrial, and

domestic uses were developed from several sources including reported withdrawals for non-exempt wells, estimated crop requirements for agricultural wells, and estimated per-capita water use for exempt wells.

Simulations of the groundwater-flow system in northern Arizona include conditions that existed before development of groundwater supplies and subsequent transient conditions through 2005. The accuracy of simulations was evaluated by using observational control from water levels in wells and estimates of base flow from streamflow records and estimates from previous studies. Because groundwater data became available with development of groundwater supplies and, therefore, were influenced by development to some extent, data from periods of early groundwater development in each groundwater basin were assumed to represent predevelopment conditions. Control data for transient groundwater conditions included variations in water levels and variations in stream base flow. Many areas had little control data for evaluation of predevelopment and transient conditions, resulting in uncertain simulations. Transient control data for some areas was sufficient to adequately evaluate the simulations.

Substantial groundwater development began with the discovery of flowing well conditions in the Little Chino Valley about 1930. Development of groundwater supplies began about 1940 or later in other alluvial basin areas and much later in areas where surface-water supplies are plentiful or depths to water are great, such as on the Coconino Plateau. Groundwater data accumulated with development of the groundwater supplies in each area. As a result, the predevelopment groundwater-flow system is not defined by data collected during a specific time period, such as during 1940, but is defined by the earliest water levels in wells and rates of discharge to streams that became available at different times across the study area. The earliest data used to define the predevelopment groundwater-flow system are available for the Little Chino sub-basin during or before the 1940s, for the Big Chino and Verde Valley sub-basins before 1961, for the Little Colorado River Plateau basin largely before 1961, except for poorly developed areas, and for other areas as late as the 1990s.

Many data deficiencies are evident from the NARGFM simulations that result in uncertain response of the groundwater system to variations in withdrawal and recharge. Mostly, a lack of long-term hydrologic records exists in many areas that could document response of the groundwater systems to changes in withdrawals and recharge. Recharge along ephemeral and losing stream reaches is poorly defined, resulting in uncertain transmissivity distributions and simulations of hydrologic response to changes in recharge and discharge. Vertical groundwater flow between vertically adjacent aquifers and water-bearing units is an influence in many areas, but is poorly defined by existing water-level data. Vertical groundwater flow could be better understood by developing long-term records from colocated wells that tap the locally important water-bearing zones. Improved knowledge of the extents of aquifers and hydrologically influential alluvial facies would reduce uncertainties in the simulation of the groundwater-flow systems in the Big Chino and Verde Valley sub-basins. Storage properties are poorly defined throughout most of the regional aquifers. Storage properties and aquifer transmissivity are equally as important for the simulation of the response of the aquifer system to changes in withdrawals and recharge. Mostly, the distribution of confined and unconfined groundwater-flow conditions needs to be better defined, especially in the alluvial basins. Many geologic structures such as faults, fractures, and collapse structures are expected to be hydrologically influential. However, little hydraulic data exists that can be used to infer the effect of most individual structures or groups of structures. Use of surface water for irrigation has complicated estimates of pre-development groundwater budgets in the Little Chino and Verde Valley sub-basins. Diversion of surface water for application at agricultural fields modified the groundwater system before the development of groundwater supplies. In the Little Chino sub-basin, diversion of impounded runoff likely resulted in greater recharge rates through the incidental deep percolation of excess irrigation water applied to fields at the expense of surface-water runoff in Granite Creek. The increased recharge likely resulted in elevated water levels in the upper alluvial unit and increased discharge at Del Rio Springs. Extensive diversion of surface water from the Verde River and several tributaries during the growing season results in the temporary storage of incidental deep percolation of excess irrigation water in the local Quaternary aquifer, which is perched in some areas. The temporarily stored water slowly drains to the Verde River. This artificial shallow groundwater-flow system has resulted in modified seasonal variations in Verde River base flow and difficulty in separating base flow into contributions from recirculated agricultural water and contributions from regional groundwater flow. A poor understanding of base flow results in a poor understanding of groundwater budgets. A more detailed understanding of the shallow groundwater-flow system and streams is needed to better define base flow and groundwater budgets for the Verde Valley sub-basin.

The simulations resulted in a better general understanding of the groundwater-flow system, including the important influence of ephemeral and losing stream reaches in alluvial basins, recognition that recharge variations are essential to understanding water-level and base-flow variations, that administrative groundwater basin boundaries and simulated groundwater divides are substantially different in several areas, and that groundwater discharge from the regional aquifers through phreatophyte ET is not an influential component of the groundwater budgets. Results of the NARGFM support the findings of other investigators that recharge along ephemeral and losing stream reaches is a major component of overall recharge in alluvial basins. Estimates of direct recharge by using BCM methods in the Big Chino alluvial basin and PrAMA required augmentation through other recharge mechanisms. The most likely other recharge mechanism is recharge along ephemeral and losing stream reaches. The influence of recharge along ephemeral and losing stream reaches in other areas is uncertain but should be considered. The simulation of BCM-based decadal variations in recharge rates improved the simulation of water-level trends in the alluvial basins and near the Mogollon Rim. Administrative groundwater basin and sub-basin boundaries defined by ADWR are substantially different from simulated groundwater divides in several areas. Some differences were expected based on water levels in wells. In areas lacking water-level control, the simulated groundwater divides are uncertain because divide locations result from simulated spatial distributions of recharge and transmissivity that have limited control. The difference in administrative and simulated groundwater divides influences the simulated groundwater budgets in the affected basins. Much of the recharge to the Big Chino groundwater sub-basin north of Big Black Mesa actually contributes to groundwater flow in the Coconino Plateau basin and discharges to the Colorado River. About 15 percent of recharge in the administratively defined Little Colorado River Plateau basin was simulated as contributing to the Verde River, Coconino Plateau, and sub-basins in the Salt River drainage. Discharge of groundwater from the regional aquifers through phreatophytes is not a major part of the groundwater budgets in any of the simulated groundwater basins. Much of the phreatophyte water use in the Verde Valley is withdrawn from perched aquifers that receive incidental recharge from excess irrigation water and unsaturated zones along losing stream reaches.

The simulations resulted in some major insights about the groundwater-flow systems in individual basins. Simulated stream base flow responded quickly to, and base flow trends mimicked variations in, recharge rates in many groundwater basins, including the Big Chino sub-basin, Verde Valley sub-basin, Coconino Plateau basin, each of the western basins, Verde Canyon sub-basin, and the sub-basins along the Salt River. The simulated baseflow response to recharge was verified by available long-term records at only one streamflow gaging station, the Salt River gage near Roosevelt. The Little Colorado River Plateau basin was the only basin where simulated groundwater discharge did not vary greatly or rapidly with variations in rates of recharge. The relative lack of response of the Little Colorado River Plateau basin to variations in recharge rates is due to the large distance between areas of greatest recharge in the high elevation areas and discharge near the Colorado River and relatively large storage and low transmissive properties of the aquifer in comparison to other groundwater basins in the NARGFM. Base flow to the Verde River headwater springs and the Verde River upstream from the streamflow-gaging station near Paulden likely diminished before installation of the station in 1964. The diminished flow resulted from groundwater withdrawals in the Big and Little Chino sub-basins and below average recharge rates during the 1950s. Geologic structures associated with Big Black Mesa are efficient barriers to the northward flow of groundwater from the Big Chino alluvial basin toward the Coconino Plateau basin. As a result of the flow barrier, groundwater in the Big Chino alluvial aquifer discharges to the Verde River rather than the Colorado River. A part of groundwater discharge from the Big Chino sub-basin may occur as groundwater flow near the streamflow-gaging station near Paulden rather than stream base flow. Geologic structures in the Verde Valley sub-basin also are hydrologically important features, groundwater flow barriers or conduits, that locally influence the flow of groundwater. Stream-aquifer interactions along the lower part of the Little Colorado River are an influential control on head distributions throughout the Little Colorado River Plateau basin. However, the groundwater-flow system and distributions of hydraulic head in the region are not well defined.

Simulation of the multiple hydraulically connected aquifers across the study area allows for evaluation of the effects of groundwater withdrawals and management practices in each basin on the groundwater system in that basin and adjacent basins. This simulation allows a more rigorous estimate of the effects of development than can be attained by individually simulating groundwater systems in each basin or part of a basin. Individual simulations of groundwater flow in sub-basins that are a part of hydraulically connected regional aquifer systems require simplifying assumptions of groundwater flow across the boundaries of the sub-basin. The most common assumption is that no flow of groundwater occurs across an artificial groundwater boundary. The no flow assumption forces all of the changes that result from groundwater withdrawals, such as diminished groundwater discharge, to occur within the simulated sub-basin. In reality, a part of the change happens in adjacent hydraulically connected basins. Therefore, no-flow boundaries result in the simulation of the maximum possible change that can happen in the simulated sub-basin. Alternative assumptions about groundwater flow across the boundaries, such as constant flux, variable flux, or variable head, result in reduced rates of change in the simulated sub-basin and only general estimates of the effect on the groundwater-flow systems in adjacent basins. Simulation of multiple hydraulically connected aquifers in the NARGFM included as much hydrogeologic information as possible to simulate the hydraulic connections, including regionally consistent methods of distributing recharge, aquifer structure, hydraulic property estimates, and locations of streams and drains. The degree of hydraulic connection in some areas is, however, poorly defined and could be improved with additional information.

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