Numerical Ground-Water Change Model of the C Aquifer and Effects of Ground-Water Withdrawals on Stream Depletion in Selected Reaches of Clear Creek, Chevelon Creek, and the Little Colorado River, Northeastern Arizona

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Contents

Figures

Tables

Abbrevations

NAVD 88, Vertical coordinate information is referenced to the North American Vertical Datum of 1988.

NAD 27, Horizontal coordinate information is referenced to the North American Datum of 1927.

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.

acre-ft, acre foot
acre-ft/yr, acre foot per year
ft, foot
ft/day, foot per day
ft^2/d, foot squared per day
ft^3/s, cubic foot per second
ft^{-1}, per foot
gal/day, gallon per day
in./yr, inch per year
m, meter
mi, mile
mi^2, square mile

Acronyms

EDNA, Elevation Derivatives for National Applications
GIS, Geographic Information System
GWF, Ground-Water Flow (procedures of MODFLOW-2000)
KRB, riverbed conductance parameter
LCR, Little Colorado River
MODFLOW-2000, Modular Three-Dimensional Finite Difference Ground-Water Flow Model
OBS, Observation (procedures of MODFLOW-2000)
PES, Parameter Estimation (procedures of MODFLOW-2000)
RIVGRID, River Package data set program
SEN, Sensitivity (procedures of MODFLOW-2000)
USGS, U.S. Geological Survey

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Abstract

The base flow in parts of Chevelon and Clear Creeks and of the Little Colorado River near Blue Spring in northeastern Arizona is sustained by discharge from the C aquifer, and in some reaches supports threatened and endangered fish species. C aquifer water is proposed as a replacement supply to relieve pumping from the N aquifer—the current source of water for a coal slurry pipeline used to transport coal mined from Black Mesa to Laughlin, Nevada. Locations of the proposed withdrawals are in the area of Leupp, Arizona, about 25 miles from a perennial reach of lower Clear Creek. A simulation tool was needed to determine possible effects of the proposed withdrawals from the C aquifer, particularly the effects of depletion of streamflow in Clear Creek, Chevelon Creek, and the Little Colorado River in the area of Blue Spring. A numerical ground-water change model was developed for this purpose. The model uses the U.S. Geological Survey finite-difference model code MODFLOW-2000 and data sets representing key features of the C aquifer to simulate change in the system that would result from withdrawing water at proposed locations. Aquifer thickness was estimated from a hydrogeologic framework model, and values of aquifer properties such as hydraulic conductivity and specific yield were estimated from aquifer-test data. Two scenarios with differing withdrawal rates were run for a 101-year period that included 51 years of withdrawals followed by 50 years of no withdrawals. About 6 percent of the ultimate volume of depletion occurs in the 101-year period for either scenario. The maximum streamflow depletion rate for all reaches in the scenario with the greatest withdrawal rates was computed to be about 0.6 cubic foot per second. The depletion rate was highest in lower Clear Creek, the reach that is closest to the well field. A model that simulates historical conditions was used to help select the most reasonable parameter sets for a Monte Carlo analysis of computed stream depletions.

Introduction

The C aquifer underlies much of northeastern Arizona, and parts of northwestern New Mexico (*fig. 1*), and is an important source of water for many users. It is named for the primary water-bearing rock unit of the aquifer, the Coconino Sandstone. The aquifer is the most productive aquifer in the Little Colorado River Basin. Industrial users, individual homeowners, agriculture, and the municipalities of Flagstaff, Holbrook, Leupp, Show Low, St. Johns, Joseph City, Snowflake, Taylor, and Winslow, Arizona, depend upon water drawn from wells completed in the aquifer. Because the aquifer lies beneath the Navajo and Hopi Indian Reservations, it represents an important present and future source of water to support these native peoples and their economies.

Water from the C aquifer is proposed as a replacement supply for mine-related withdrawals from the N aquifer—the current source of water for a coal slurry pipeline used to transport coal mined from Black Mesa to the Mojave Generating Station, an electrical power plant in Laughlin, Nevada. Water from the N aquifer generally is of good quality. Replacement of the supply by poorer quality from the C aquifer would reduce industrial use of the N aquifer water, a current supply for tribal communities. The proposed area of withdrawals for replacement water from the C aquifer is near Leupp, Arizona. A pipline would be required to transport the water to the coal mine on Black Mesa. Along with withdrawals for the mine, withdrawals also are proposed for Hopi and Navajo water supply.

The base flow of parts of some streams, including Chevelon and Clear Creeks, and of the Little Colorado River is sustained by discharge from the C aquifer, and in some cases, these streams support threatened and endangered fish species. The C aquifer discharges into the Verde River and Salt River Basins to the south, and to underlying formations and ultimately to the lower Little Colorado River to the northwest (*fig. 1*). Blue Spring, about 13 river miles (mi) upstream from the mouth of the Little Colorado River, is the largest of many springs that discharge in the lower reach of the river.
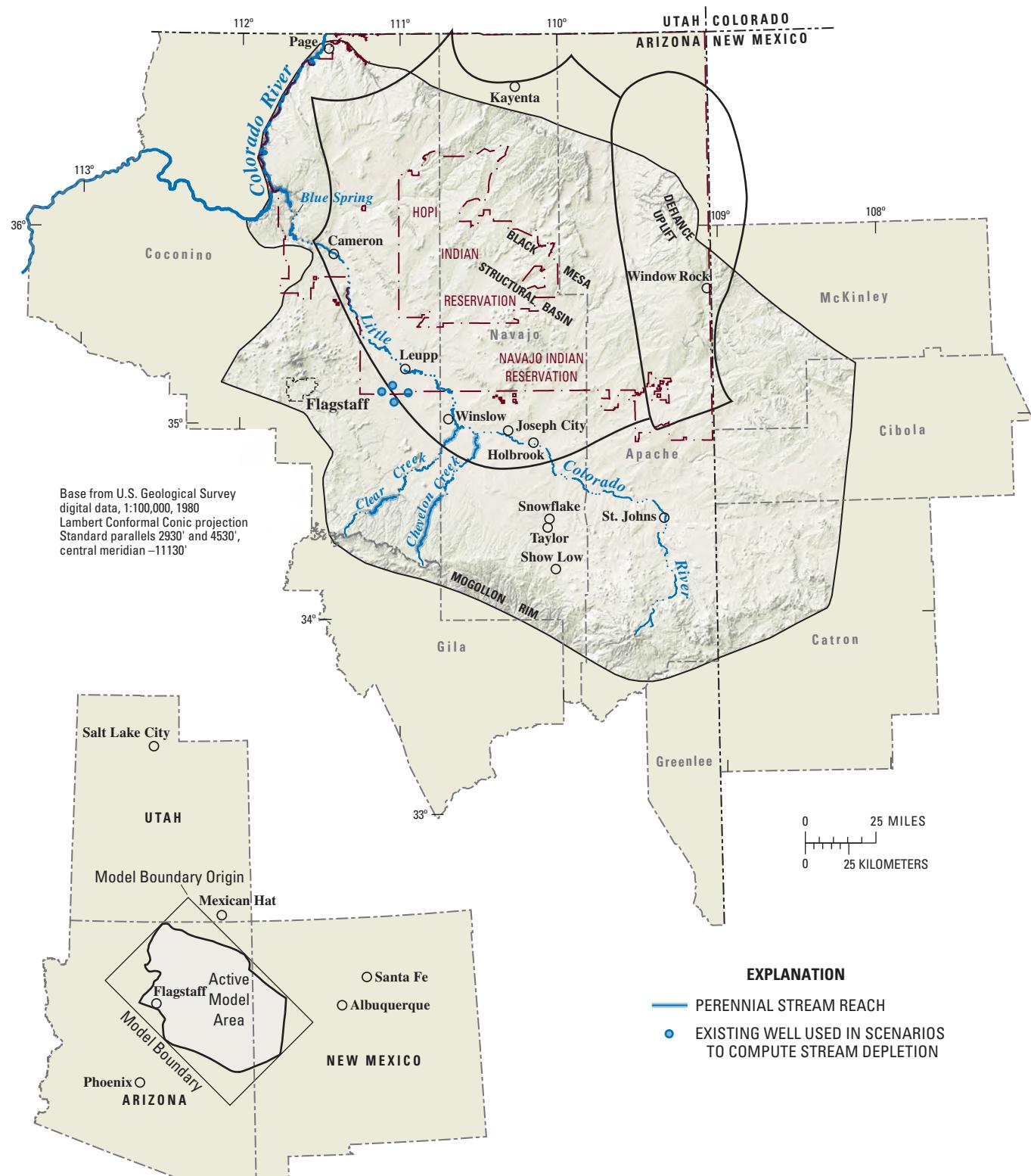
Figure 1. Location of modeled area of the C aquifer, northeastern Arizona.

Purpose and Scope

In 2004, the U.S. Geological Survey (USGS), in cooperation with the Bureau of Reclamation, was requested to participate on a Federal team to evaluate potential effects of future ground-water withdrawal scenarios from the C aquifer on base flow in the Chevelon and Clear Creek drainages, and ground-water discharge to the Little Colorado River in the area of Blue Spring. USGS participation on the Federal team involved assisting with test drilling and associated aquifer testing in the Leupp, Arizona, area (Hoffmann and others, in press) and with development of a numerical ground-water model of the C aquifer, discussed in this report.

The purpose of this report is to describe results of an analysis of pumping effects on base flow using a numerical ground-water model. The model was used to evaluate the effect of pumping on base flows of Chevelon and Clear Creeks and on ground-water discharge to the Little Colorado River near Blue Spring area for specified withdrawals and periods. Evaluations were carried out using two scenarios of withdrawals provided by a technical advisory group (Craig Sommers, Economist, ERO Resources Corporation, Denver, Colorado, written commun., 2005). Results are limited to calculated stream depletions and drawdown values.

Previous Investigations

Parts of the C aquifer have been investigated to evaluate water availability for local water supply. Peter Mock Groundwater Consulting, Inc. (2003) constructed a ground-water flow model of a 2,400-mi^2 rectangular part of the C aquifer encompassing Clear and Chevelon Creeks south and east of Leupp, Arizona. The ground-water flow model was used to estimate the changes in ground-water levels near the proposed well fields, at other existing and proposed water demand centers, and in base flow of Clear and Chevelon Creeks. Boundary conditions for this local-scale model were derived using a regional analytical model (Southwest Ground-water Consultants, Inc., 2003). The analytical model used the Theis nonequilibrium equation with the Jacob water-table correction, with a transmissivity of 25,000 (gal/d)/ft (3,300 ft^2/d) and a specific yield of 0.07.

Two recent studies have attempted to characterize the aquifer on a regional scale or over its full extent. In the area of Flagstaff, Arizona, the C aquifer was investigated by Bills and others (2000). This study included extensive structural mapping, geophysical surveying and mapping, determination of hydraulic properties, and chemical and isotopic analysis of ground water. A ground-water budget and a generalized characterization of the geohydrology of the entire C aquifer were developed by Hart and others (2002). Several other reports provide information on water resources in the study area including the C aquifer. Cooley and others (1969) describe the hydrogeology of the Navajo and Hopi Reservations and McGavock and others (1986), Mann and Nemecek (1983), and Mann (1976) describe water resources in southern parts of Coconino, Navajo, and Apache Counties, respectively.

Physiographic and Hydrogeologic Setting

The Little Colorado River Basin is in the southern part of the Colorado Plateau physiographic province (Fenneman, 1946). The Little Colorado River drains an area of about 27,000 mi^2 of northeastern Arizona and northwestern New Mexico (*fig. 1*). Elevations in the basin range from about 2,700 ft near the mouth of the Little Colorado River to 12,633 ft near Flagstaff. Average annual precipitation ranges from less than 6 in. in the lower areas to more than 30 in. in the highest elevations. Total annual average precipitation within the basin is estimated to be approximately 17,000,000 acre-ft. Potential evapotranspiration is estimated to range from less than 30 in./yr to more than 60 in./yr (Hart and others, 2002).

Boundary of the C Aquifer

The southern and southwestern boundary of the C aquifer extends south of the Little Colorado River Basin into the Verde River and Salt River Basins where rocks of the C aquifer are exposed along the Mogollon Rim (*fig. 1*). The northeastern boundary of the aquifer extends into New Mexico where the aquifer thins substantially (Hart and others, 2002). Few water-level and geologic data are available; therefore, the boundary is uncertain in these locations. The western boundary of the C aquifer is west of the Little Colorado River where the aquifer is unsaturated. Rocks of the C aquifer crop out in the southern and southwestern areas along the Mogollon Rim, in the northeastern area near the Defiance Uplift, and in the western area near the Little Colorado River (*fig. 1*).

Geologic Structure

The dominant structural feature within the C-aquifer area is the Black Mesa Basin, located in the north central part of the study area. The C aquifer generally dips toward the center of the basin from its outcrops along the margins of the study area. In the center of the Black Mesa Basin the aquifer is buried beneath nearly 5,000 ft of overlying geologic units (Southwest Groundwater Consultants, Inc., 2003; *fig. 2*) The Defiance Uplift is near the eastern edge of the C aquifer. Monoclines on either side of the uplift plunge to the west and east in the Black Mesa and San Juan Basins. Several regional-scale normal faults exist in the area and generally have two principal strike directions: north-northeast and north-northwest (Bills and others, 2000).

Geologic Units of the C Aquifer

The C aquifer comprises a sequence of sedimentary rock units between the top of the Kaibab Formation, which may underlie the Shinarump Member of the Chinle Formation or Moenkopi Formation, and the middle to upper part of the Supai Group (*fig. 2*; Cooley and others, 1969). The primary aquifer unit is the Coconino Sandstone and its lateral equivalents (the DeChelly Sandstone and Glorieta Sandstone in New Mexico); however, the overlying Kaibab and Toroweap Formations and the underlying Schnebly Hill Formation and Upper and Middle Supai Formations of the Supai Group locally can be significant water producing units. In the western part of the C aquifer near Flagstaff, units above the Supai Group are dry; the water table of the C aquifer there is in the upper part of the Supai Group (Bills and others, 2000).

The Kaibab Formation is primarily a limestone with a thickness of as much as 650 ft. It is absent in the eastern and northern part of Little Colorado River Basin. The Toroweap Formation underlies the Kaibab Formation and comprises carbonate sandstone, red beds, silty sandstone, siltstone, limestone, and thin layers of gypsum. Similar to the Kaibab Formation, the Toroweap is absent in the eastern and northern parts of the Little Colorado River Basin.

The Coconino Sandstone is a white to tan to light brown, crossbedded, eolian, fine-grained quartz sandstone. This unit generally is 300 to 1,100 ft thick (Bills and others, 2000). It can be extensively fractured where faulted. These fractured zones are likely to be areas of high permeability.

The Schnebly Hill Formation comprises a sequence of reddish-brown to reddish-orange very fine to silty sandstone, mudstone, limestone, and dolomite (Blakey, 1990). The formation ranges in thickness from a few feet to as much as 800 ft and generally underlies the Coconino Sandstone; however, the contact between the two units can be gradational and intertonguing. Where saturated, the Schnebly Hill Formation is in hydraulic connection with the Coconino Sandstone.

The Supai Group was divided into the Upper, Middle, and Lower Supai Formations by 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 a grayish-orange, calcareous, very fine grained sandstone to siltstone. The Lower Supai Formation is a red to purple sandstone and siltstone and gray limestone and dolomite. The Supai Group is typically 600 to 2,400 ft thick and is exposed in the southern and southwestern part of the study area. Where saturated, the Upper Supai Formation is the most permeable unit in the group. The Lower Supai Formation typically forms a confining unit that separates the C aquifer from underlying saturated units.
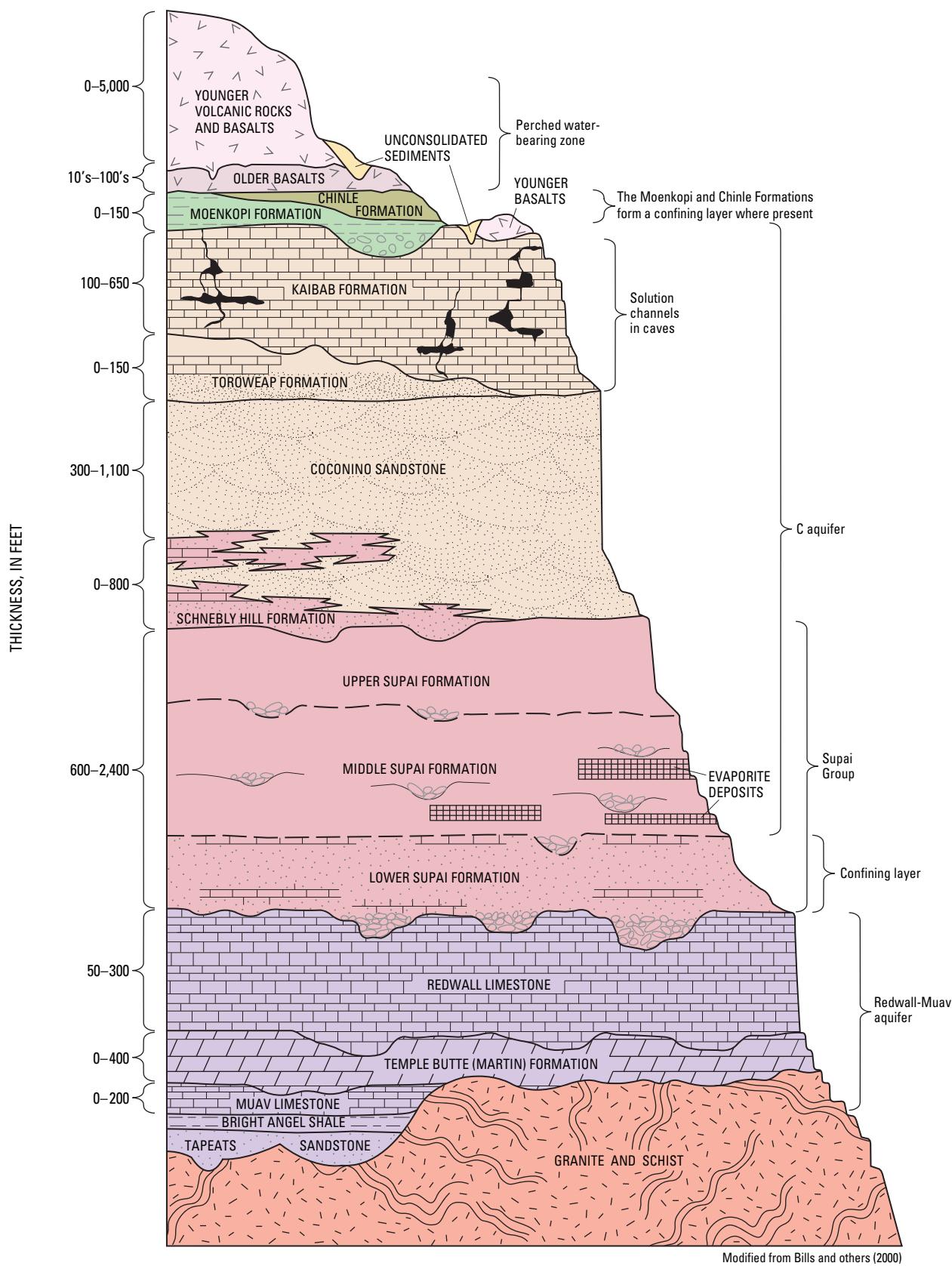
Figure 2. Generalized stratigraphic section of rock units in and adjacent to the C aquifer in the study area, northeastern Arizona.

Aquifer Properties

Aquifer tests results reported for these sedimentary units in the vicinity of Flagstaff indicate a significant variability in rock fracturing leading to a range of transmissivities, hydraulic conductivities, and storage coefficients (Bills and others, 2000; Hoffmann and others, in press). Transmissivity and hydraulic conductivity values for the Coconino Sandstone generally were higher than the values for the Kaibab Formation, the Schnebly Hill Formation, and the Supai Group. Regardless of lithology, wells with the highest transmissivity and hydraulic conductivity values were coincident with the greatest degree of fracturing and the lowest values were coincident with areas of limited or no visible fracturing. 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. In addition to the values published by Bills and others (2000), Southwest Ground-water Consultants, Inc. (2003) compiled aquifer-parameter estimates of the C aquifer from multiple sources. These sources indicate a large variation in transmissivity estimates (1 to 13,000 ft^2/d). The large variance is attributed to factors such as degree of fracturing and differences in the penetration depths of wells used for aquifer tests. Many of the wells were designed to meet small demands for water and therefore were drilled to a depth where the required yield could be obtained. In many cases these wells penetrate only a part of the total aquifer thickness, resulting in an underestimate of transmissivity. In addition to transmissivity, Southwest Ground-water Consultants, Inc. (2003) compiled estimates of hydraulic conductivity values from tests in the C aquifer (*fig. 3*).

In confined parts of the aquifer, the storage coefficient [dimensionless] is related to the compressibility of the rock matrix and water. In terms used by ground-water hydrologists, the storage coefficient, S, can be computed as

$$S = b \times S_S,$$

where b [length] is aquifer thickness, and S_s [length^{-1}] is total specific storage, which is the sum of S_{sw} and S_{sk}, the components of specific storage due to compressibility of water and the aquifer skeleton, respectively. Values of S_s can be estimated from well-designed aquifer tests and also from properties of water and rocks.

In unconfined parts of the area, the most important storage property is specific yield [dimensionless], which is a measure of the relative amount of water draining or filling pore spaces when the height of the water table changes. According to Bills and others (2000), the average of eight specific yield estimates derived from aquifer tests in the Flagstaff area was 0.077.

Several investigators (Mann, 1976; Daniel, 1981; Mann and Nemecek, 1983) have delineated areas where the C aquifer is confined and unconfined (*fig. 3*). For analyses of effects of withdrawal of water from the C aquifer, specific-yield values must be applied in areas where the aquifer is unconfined and storage coefficient values must be applied where the aquifer is confined. According to data compiled by Southwest Ground-water Consultants, Inc. (2003), the Coconino Sandstone responds to pumping as an unconfined or semiconfined aquifer, with storage properties of 0.07 for specific yield and 1×10^{-4} for the confined storage coefficient; however, there are few values of confined storage coefficient reported.

Ground-Water Flow and Discharge

Ground water flows from areas of recharge to areas of discharge. In the C aquifer, recharge occurs primarily in the southern part of the study area in the highlands north the Mogollon Rim, including the San Francisco volcanic field in the Flagstaff area, and on the Defiance Uplift in the eastern part of the area. From the recharge areas along the Mogollon Rim, some water flows to drainage basins south of the rim and some water flows northward, within the Little Colorado River drainage, ultimately toward the Blue Spring area. Ground water generally is unconfined in the southern and western parts of the aquifer (*fig. 3*) where the water table is below the elevation of the bottom of the Moenkopi Formation (*fig. 2*). The Moenkopi Formation comprises red to dark-red to reddish-brown siltstone, silty sandstone, fine- to very fine-grained sandstone, mudstone, and gypsum and is a confining layer in the north and east parts of the aquifer.

The primary discharge for the C aquifer in the Little Colorado River Basin is through a series of springs in the lower reaches of the Little Colorado River (Hart and others, 2002). The main spring is Blue Spring, which issues from solution channels in the Redwall-Muav aquifer in the lower reach of the Little Colorado River, about 13 mi upstream of the mouth of the river. The springs sometimes are collectively referred to as "Blue Springs." Discharge from Blue Spring and other springs in the area is estimated to be approximately 164,000 acre-ft/yr (Hart and others, 2002). All this discharge is thought to be downward leakage from the C aquifer because the Redwall-Muav aquifer is deeply buried and does not crop out within the Little Colorado River Basin. An interpretation of the flow system is that water moves downward through the Supai Group through a series of normal faults (Cooley and others, 1969; Hart and others, 2002).

Figure 3. Estimated hydraulic conductivity values for the C aquifer, northeastern Arizona, compiled by Southwest Ground-water Consultants (2003), hydraulic conductivity zones, and contours of drawdown to conversion to unconfined conditions used in the change model.

EXPLANATION

C AQUIFER CONFINED
C AQUIFER DRY
2 — HYDRAULIC CONDUCTIVITY ZONE USED IN THE CHANGE MODEL

ESTIMATED HYDRAULIC CONDUCTIVITY, IN FEET PER DAY:
Less than 1
1–5
5–10
Greater than 10

−100− DRAWDOWN TO CONVERSION CONTOUR, IN FEET
Modified from Mann, 1976
Modified from Mann and Nemecek, 1983
Modified from Daniel, 1981
Estimated

Base from U.S. Geological Survey digital data, 1:100,000, 1980
Lambert Conformal Conic projection
Standard parallels 2930' and 4530',
central meridian −11130'

The Redwall-Muav aquifer comprises the Redwall and Muav Limestones, the Martin Formation, and other Devonian rock units, where present. The Redwall Limestone is a light-gray to gray aphanitic to coarsely crystalline bedded limestone. Some beds within the formation are fractured, and the limestone typically contains solution cavities and caverns, some of which have collapsed. The Muav Limestone is a dolomitic limestone. Thickness of the Redwall-Muav aquifer ranges from about 200 to more than 500 ft in the study area.

Ground water from the C aquifer also has historically discharged along or adjacent to various reaches of the Little Colorado River and its tributaries. The two tributary reaches of greatest interest for this study are the lower reaches of Clear and Chevelon Creeks. USGS streamflow-gaging stations were operated from 1906 to 1982 at Clear Creek near Winslow and from 1906 to 1972 at Chevelon Creek near Winslow. Few recent base-flow measurements exist along these creeks; however, measurements made in the summer of 2005 indicate base-flow values of about 5.6 and 2.7 ft^3/s (a combined total of about 6,000 acre-ft/yr) in the lower reaches of Clear and Chevelon Creeks, respectively.

Simulation Approach and Methods

Ground-water models can be valuable tools for understanding the possible responses of aquifers to stresses such as withdrawal of water by wells. One approach to simulating the effects of withdrawal is to construct a numerical ground-water flow model with a program such as MODFLOW-2000 (Harbaugh and others, 2000). In this approach, a ground-water flow equation with head as the dependent variable is solved. Flow into and out of the aquifer system is simulated, and the computed distribution of head can be compared with observations of head in wells. A simpler approach is to simulate changes in head in an aquifer. This approach is most common with analytical models, such as the Theis (1935) equation, in which change in head or drawdown is the dependent variable. If the aquifer-system response is linear, then the principle of superposition applies and simulation of changes can be referred to as "superposition modeling." A problem with application of analytical superposition approaches (such as the Theis equation) for large-scale systems is that complex aquifer geometry and heterogeneous aquifer thickness and hydraulic properties cannot be simulated. For these situations, however, a superposition approach can be used with a numerical model program such as MODFLOW-2000. For details on the principle of superposition as used in ground-water hydrology, see Reilly and others (1987). For systems with nonlinearities such as transmissivity that changes as a function of head, the principle of superposition does not strictly apply.

Using a code such as MODFLOW-2000, a model can be set up to simulate changes in head in the aquifer and changes in flow to or from boundaries, while accounting for nonlinearities such as variable transmissivity. For the approach, referred to as "change modeling" here, the initial water surface (head) is set to a uniform elevation, referenced to an arbitrary datum. The head at all head-dependent boundary features is set to the same elevation and no recharge or discharge is specified so that no flow into or out of the aquifer or within the aquifer occurs in the initial state. Imposing a stress such as withdrawal of water by a well has the effect of propagating head changes into the surrounding model domain. In time, the effects can cause head changes in model cells containing surface-water boundaries. In response to that change in head, change in flow to or from the boundary is calculated.

In theory, any uniform initial head value could be used; however, numerical precision of calculations made with the model will be greatest when an initial elevation of 0.0 ft is used. This value also should be used for head in surface-water boundaries. To correctly simulate the aquifer thickness, the bottom of the aquifer must be set to an elevation equal to the initial head minus aquifer thickness. With an initial head of 0.0 ft, an estimated aquifer thickness of 520 ft would result in a bottom elevation of −520 ft. If at sometime during the simulation a head decline of 40 ft was computed in the change model for that location, the elevation of head in the model would be −40 ft and the saturated thickness would be updated automatically in MODFLOW to 480 ft and used in calculations of conductance between model cells.

The purpose for developing the change model described here was to estimate likely rates of capture of flow in selected streams as a result of proposed withdrawals of ground water in the area of Leupp, Arizona. Because of the short time frame provided for development of a simulation tool, a generalized simulation approach was used. Some key aspects of the approach are as follows:

1. The horizontal domain of the model is represented with a grid of cells that covers most of the C aquifer, using distant lateral boundaries so that computed effects in the area of interest will not be affected by uncertainties in the assignment of types and locations of boundaries.

2. The vertical domain of the model includes a single layer to represent the saturated part of the C aquifer, primarily the Coconino Sandstone. In the western part of the modeled area, the C aquifer is dry and the single layer represents the Redwall-Muav aquifer that receives and transmits water flowing out of the C aquifer. No attempt is made to simulate the continuum of flow from the C aquifer to the underlying rock units with a configuration other than a single model layer.

3. The system is treated as a regional, nearly homogeneous porous media. No attempt is made to account for the effects of fractures and small-scale heterogeneities in aquifer properties.

4. Key perennial stream reaches are assumed to be hydraulically connected to the C aquifer.

5. Attempts to calibrate the change model are limited. Instead, reasonable average aquifer properties are used.

6. The model considers the effects of the proposed withdrawals near Leupp. Effects from other pumping in the regional aquifer are not considered. Scenarios for withdrawal provided by a technical team include 51 years of withdrawals at various rates and locations within a well field, followed by 50 years of no withdrawals.

Further details of the simulations are included in the following sections.

Summary of Numerical Ground-Water Change Model

The numerical code used to simulate changes in water levels in the C aquifer and ground water flow at the head-dependant boundaries is the Modular Three-Dimensional Finite Difference Ground-Water Flow Model (MODFLOW-2000) developed by the USGS (McDonald and Harbaugh, 1988; McDonald and Harbaugh, 1996; Harbaugh and others, 2000). Some of the Redwall-Muav aquifer in the western part of the domain also is included to form a single continuous transmissive system that extends to the major discharge area at and near Blue Spring.

Model Discretization and Boundary

The active model domain encompasses an area of about 32,150 mi^2, which includes much of the C aquifer beneath the Colorado Plateau and parts of the Redwall-Muav aquifer near Blue Spring (*fig. 3*). The model uses a uniform grid spacing of 0.5 mi with 358 rows and 538 columns of finite difference cells. The model uses one layer; therefore, the model has a total of 192,604 cells. There are 128,609 active cells, which represent areas where the C aquifer is saturated and part of the Redwall-Muav aquifer near Blue Spring. The corner of the grid at row 1, column 1 is at Universal Transverse Mercator (UTM) zone 12 easting 518,610.25 meters, northing 4,155,142.5 meters.

The grid is rotated 45 degrees clockwise so that columns increment in a southeasterly direction and rows increment in a southwesterly direction. This rotation angle corresponds approximately with direction of regional geologic structure.

The lateral boundary of the active area of the model shown on *figure 4* is represented as a no-flow boundary. Individual segments of the boundary generally coincide with features as follows:

Boundary	Segment Approximate location of feature
A to B	Physical extent of the C aquifer along the Mogollon Rim
B to C	Transition along the Mogollon Rim to a ground-water divide
C to D	Ground-water divide
D to E	Mesa Butte and Gray Mountain Faults
E to F	Colorado River
F to G	Line from the Colorado River to ground-water divide near the Defiance Uplift
G to A	Boundary where ground-water conditions are poorly defined

The boundary segment E to F along the Colorado River is likely to have some hydraulic connection with the ground-water system yet is represented as a no-flow boundary. This representation of a boundary type, along with treatment of divides as immoveable no-flow boundaries, must be done with care in a model. In this case, the boundaries are at such great distances from the proposed withdrawals near Leupp that incorrect locations or boundary types will not affect computed results of capture in stream reaches of interest within the time frames used for the simulations.

The vertical domain of the model was simulated with one model layer. The system simulated includes the C aquifer and underlying aquifers in the western part of the area where the C aquifer is dry. For parts of the study area where the Coconino Sandstone has a substantial saturated thickness, most of the transmissivity of the C aquifer can be attributed to the Coconino Sandstone. Aquifer tests done in the C aquifer in the vicinity of Leupp, Arizona (Hoffmann and others, in press), indicate that the hydraulic conductivity of the Coconino Sandstone can be three orders of magnitude greater than that of the most permeable formation of the Supai Group, the Upper Supai Formation.
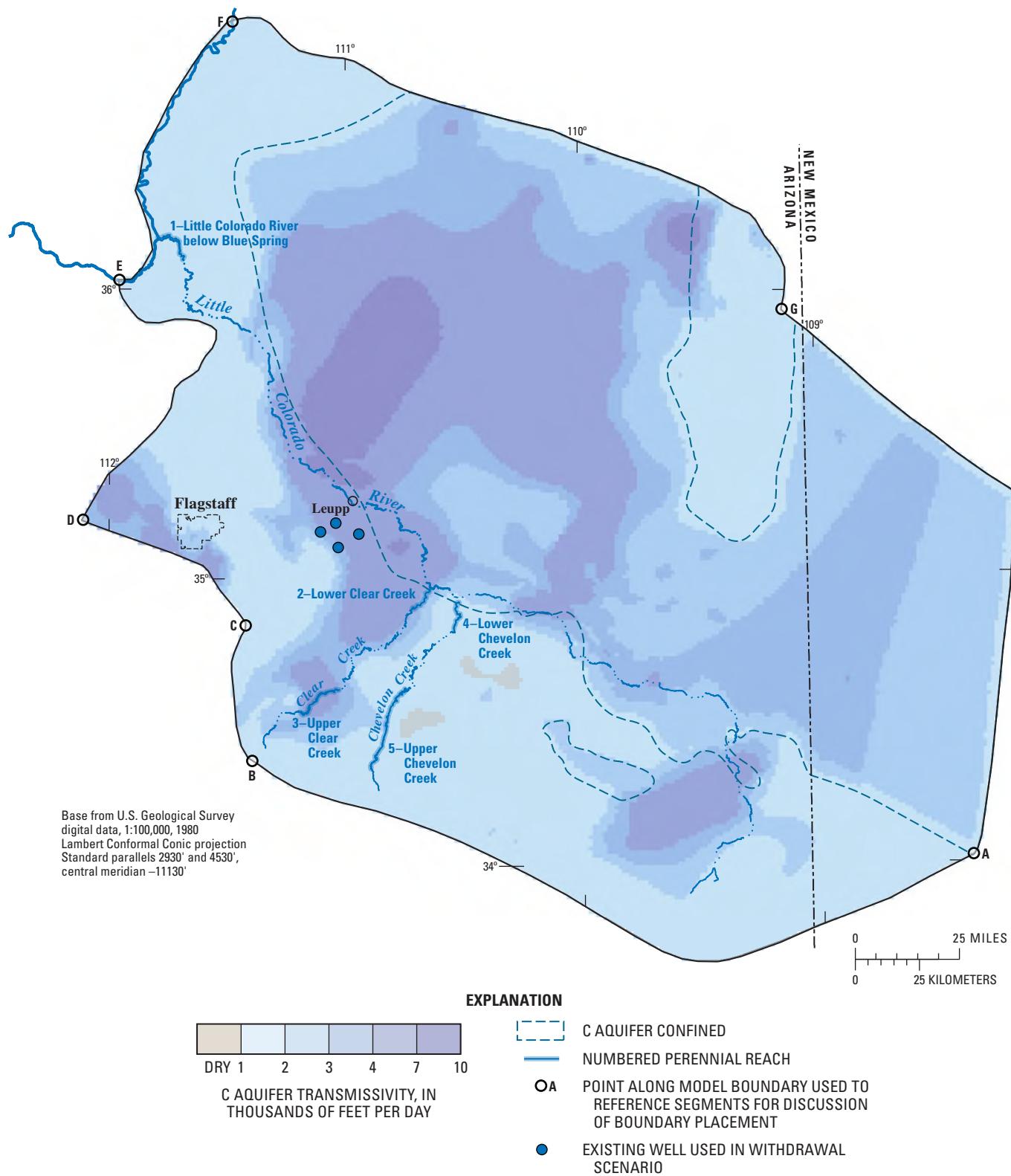
Figure 4. Lateral model boundaries, perennial reaches, confined areas, and initial transmissivity used in simulations of effects of withdrawals in the Leupp, Arizona, area.

Aquifer Thickness

A geologic framework model was used to estimate aquifer thickness throughout the area. The framework model represents three-dimensional relationships between the C aquifer, the surrounding confining units, and the water table. In the confined area, aquifer thickness is defined as the difference in elevation between the top of the Kaibab Formation or Coconino Sandstone and the top of the Supai Group. In the unconfined area, aquifer thickness is defined as the difference in elevation between the water table and the top of the Supai Group. The framework model was constructed using Dynamic Graphics, Inc. EarthVision® software. The grid consists of 358 rows and 538 columns and is rotated at 45 degrees from the horizontal, which is the same discretization and orientation of the numerical ground-water model. Thickness is calculated for coordinates that coincide with the cell centers of the numerical model.

The framework model was constrained on the basis of well logs from 795 wells throughout the study area and from elevations of stratigraphic units exposed on the Mogollon Rim and on the south rim of the Grand Canyon (*fig. 2*). Of these wells, 559 have well logs that identify elevations for the top of the C aquifer; 386 well logs have elevations for the top of the Supai Group. Geologic sections (Southwest Ground-water Consultants, Inc., 2003) and structural contours of the top of the Coconino Sandstone (Mann, 1976; Mann and Nemecek, 1983) also were used. Because cross-section and contour data were mostly interpolated or estimated on the basis of prior geologic judgment, these data were given less weight in the calculations of the top and bottom of the C aquifer. The piezometric surface was digitized from maps in Hart and others (2002). Geologic faults were not included in the construction of the framework model because of a lack of data for these features. This simplification results in a continuous layer in which steep variations in thickness can occur where faults might exist. The framework model calculated some areas with thicknesses of less than 300 ft and pinch outs of the C aquifer. The thin areas and pinch outs occurred because the framework model fits smooth planes using minimum-tension gridding to represent the top and bottom of the unit—calculations for areas having sparse data could result in thinning because one or both of the surfaces was either too low or too high in elevation. A thickness of 300 ft was assumed for these areas to assure horizontal continuity of the C aquifer. The framework model was assumed to be correct in areas having large thicknesses. The thickness of the C aquifer is generally unknown in the New Mexico part of the study area, but the productive thickness of the C aquifer is reported to be about 300 ft (Cooley and others, 1969).

The thickness of the C aquifer in the framework model varies between 300 and 2,089 ft (*fig. 5*); the average thickness of the aquifer is 390 ft. The C aquifer is dry in areas where faulting or folding has caused the Coconino Sandstone to be above the water table. Unless dry, for the purpose of this investigation, the minimum thickness applied to the aquifer is 300 ft. To the west, where the C aquifer is dry owing to drainage into the Redwall-Muav aquifer, aquifer thickness is generally held constant at about 300 ft (*fig. 5*). The thickness in the extreme western part of the model is typically greater than 700 ft, whereas much of the eastern part generally has a thickness less than 500 ft. The two thickest areas include 1,600 ft or more of saturated sediments in the western and northern parts of the study area. A thickness of 300 ft was used for the middle of the basin, northwest of Holbrook.

Surface-Water Features

For simulations of effects of possible streamflow capture by proposed ground-water withdrawals near Leupp, the following five stream reaches were represented in the model: (1) upper Clear Creek, (2) lower Clear Creek, (3) upper Chevelon Creek, (4) lower Chevelon Creek, and (5) the Little Colorado River below Blue Spring (*fig. 4*). Upper and lower limits of perennial reaches 1 and 3 were taken from Brown and others (1981). Upper and lower limits of perennial reaches 2 and 4 were taken from USGS field investigations made during the summer of 2005 (D.J. Bills, U.S. Geological Survey, written commun., 2005). Reach 5 was taken to be the trace of the Little Colorado River from Blue Spring to the mouth, where much of the discharge of water moving through the C aquifer ultimately occurs. Coordinates for traces of all segments were derived from the USGS Elevation Derivatives for National Applications (EDNA) database (U.S. Geological Survey, 2005). All five reaches were represented in the model using the River Package of MODFLOW-2000 and a river head elevation of 0.0 ft, the same elevation as starting head in the aquifer. With this approach, flow between a surface-water feature and an aquifer is proportional to the head difference between the surface water and ground water in the underlying model cell. The coefficient of proportionality is the riverbed conductance. In the case of the change model constructed for this study, the river package computes change in flow in response to change in head in model cells underlying stream segments represented with the River Package. The change model does not address whether the change is a decrease in existing ground-water discharge to a stream, or an increase in existing leakage from a stream to the aquifer. In either case, however, a computed change in response to ground-water withdrawal would result in a net decrease in stream base flow. For this study, the change model was not configured to limit the amount of capture that would be available from a given stream reach.
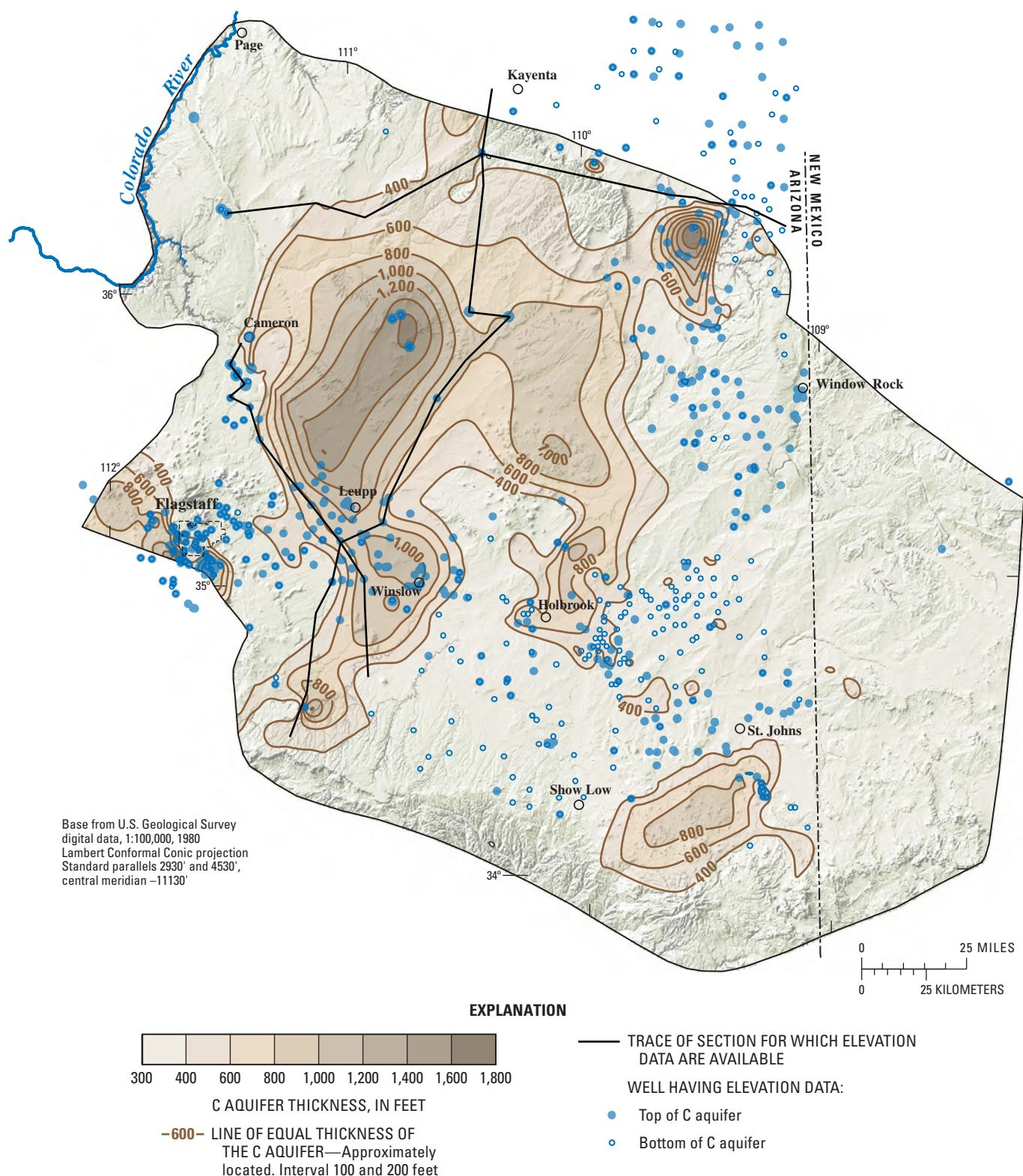
Figure 5. Thickness of the C aquifer, northeastern Arizona, as derived from a hydrogeologic framework model.

The riverbed conductance parameter required for the River Package data set is not known and cannot be estimated by calibration of the change model because observations of changes in ground-water flow to or from the streams are not available. For simulations of a 51-year withdrawal scenario, reducing the value of riverbed conductance has the effect of reducing the maximum rate of computed captured streamflow and extending the period in time over which capture occurs. Increasing riverbed conductance has the converse effect. A point exists, however, for which further increases in riverbed conductance will result in no further increases in maximum capture rate. The approach taken here is conservative with respect to the potential for capturing streamflow. The conservative approach was accomplished by using relatively large riverbed conductance values so that simulations would yield possible maximum rates in maximum capture over time.

The River Package data set for all five reaches was constructed with program RIVGRID (Leake and Claar, 1999). This program intersects a stream trace with individual model grids and computes the length of the stream in each model cell traversed. For all stream reaches in the C aquifer change model, initial riverbed conductance was computed by using a vertical hydraulic conductivity of 1 ft/d and a riverbed thickness of 20 ft. The assumed widths used for calculating riverbed-conductance values were 10 ft for reaches 1 through 4 and 20 ft for reach 5. The elevation for converting to steady river leakage was set to −30 ft so that capture would not be limited by drawdown values that might be computed under streams in this model. With these values, RIVGRID computed initial conductance values that averaged about 1,000 ft^2/d for reaches 1 through 4 and about 2,000 ft^2/d for reach 5. For final simulations, a riverbed conductance parameter (KRB) was specified in the Sensitivity Process of MODFLOW-2000 as a multiplier for all riverbed-conductance values in all five reaches. Final simulations used a value of KRB of 100, resulting in average conductance values of about 1×10^5 ft^2/d for reaches 1 through 4 and about 2×10^5 ft^2/d for reach 5. The higher conductance values lead to computation of a maximum capture rate.

Storage Properties

The change model was constructed using the Layer-Property Flow Package of MODFLOW-2000 with the option of "convertible" layer type. With this option, changes in head can cause parts of the model to switch between confined and unconfined conditions. This option requires input of specific-yield and specific-storage values. MODFLOW-2000 uses specific yield as the storage property if head in a cell is lower than the top of an aquifer, and the product of specific storage and aquifer thickness if head is higher than the top of the aquifer. For the change model in which only drawdown (downward head changes) will be simulated, the top elevation in the unconfined area can be arbitrarily specified above the initial head of 0.0 ft. In an area that is initially confined, the top elevation must be the elevation at which the system would become unconfined when drawdown occurs. The strategy for computing top and bottom elevation arrays is outlined in *figure 6*. Mann (1976) and Mann and Nemecek (1983) provide contours of the pressure head above the top of confined parts of the C aquifer in southern Navajo and Apache Counties, Arizona, respectively. These pressure heads were used to compute the "drawdown to conversion" for southern Navajo and Apache Counties. The maximum drawdown to conversion, S$_{max}$ on *figure 6*, was set at 300 ft. The shape of the surface of drawdown to conversion was extended outside of southern Apache and Navajo Counties to approximate the surface within the entire confined area (*fig. 3*).

For final simulations, a value of 0.06 was used for specific yield. This was selected on the basis of aquifer tests in the Leupp area (Hoffmann and others, in press.) A value of 2×10^{-6} ft^{-1} was used for specific storage on the basis of properties of water and estimated skeletal specific storage for sandstone.

Hydraulic Conductivity

Hydraulic conductivity in the C aquifer is highly variable. Values from individual wells are subject to local influences such as well construction and small-scale geologic features that cannot be represented in a regional flow model. On the basis of hydraulic conductivity estimates compiled by Southwest Ground-water Consultants, Inc. (2003), the four zones for hydraulic conductivity shown in *figure 3* were initially selected. Preliminary sensitivity tests using the change model showed that predicted stream depletion was sensitive to the value of hydraulic conductivity in zone 1 but was much less sensitive to the values in other zones. This result is to be expected because the proposed withdrawals near Leupp as well as all but one of the stream reaches simulated are within zone 1. The value of hydraulic conductivity for zone 1 was set at 6 ft/d on the basis of specific capacity data compiled by Southwest Ground-water Consultants, Inc. (2003) and hydraulic conductivity values determined for the Coconino Sandstone and the Schnebly Hill Formation by Hoffmann and others (in press). The value of hydraulic conductivity for zones 2, 3, and 4 was set at 5 ft/d, an estimate of the average value of the property for the C aquifer. For the Monte Carlo analysis, discussed in a later section, values of hydraulic conductivity were allowed to vary in all four zones.

Transmissivity

MODFLOW-2000 computes transmissivity as the product of hydraulic conductivity and saturated thickness for each grid cell in the active flow domain. For purposes of displaying this fundamental aquifer property, the distribution of initial transmissivity values was computed from initial thickness and hydraulic conductivity for each model cell location and mapped for the model area (*fig. 4*). This distribution is largely a reflection of the variations in the saturated thickness. The minimum value is 1,500 ft^2/d, the product of the minimum thickness (300 ft) and the minimum hydraulic conductivity (5 ft/d).

A. Actual geometry

NOTE: Initial aquifer thickness at all points, represented here with values at selected points b_1, b_2,....b_{10}, is preserved in the transformed system. Values of drawdown to conversion to water table conditions, represented here with values at selected points S_1, S_2,....S_{max}, also is preserved in the transformed system up to the value of S_{max}. Use of the transformed geometry in a change model will be valid as long as computed drawdown does not exceed S_{max}.
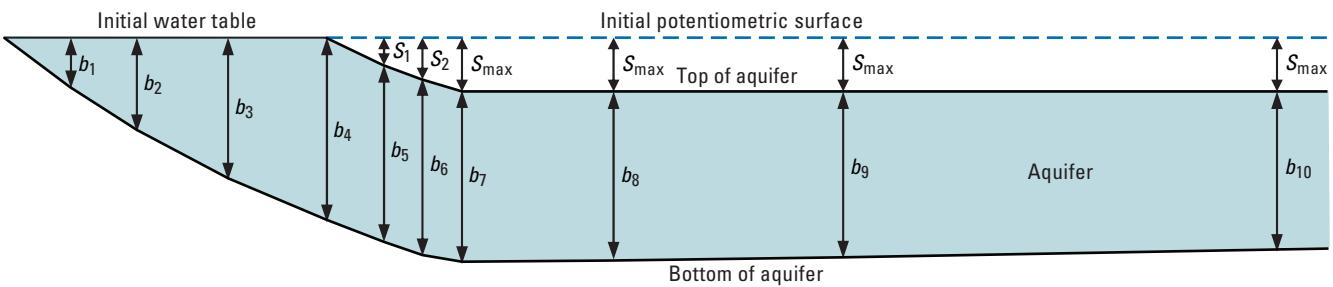
B. Transformed geometry

Figure 6. Vertical section of a hypothetical partially confined aquifer. A, Actual geometry. B, Transformed geometry for use in a change model.

Withdrawal Scenarios

Two withdrawal scenarios were simulated with the numerical change model. Both scenarios include withdrawals for the 51-year period 2010–2060 and no withdrawals for the 50-year period 2061–2110. For the period 2010–2060, multiple MODFLOW-2000 stress periods with 1-year time steps were used to represent variations in rates of withdrawals over time. A single MODFLOW-2000 stress period with fifty 1-year time steps was used to represent the recovery from 2061 to 2110. The locations of four existing wells and as many as 14 "conceptual" wells were used for both scenarios (*table 1, fig. 7*). Total withdrawals were subdivided into two use categories: (1) project withdrawals that include mine slurry and mine domestic use, and (2) withdrawals for Hopi and Navajo Tribal use. For simulations of scenarios A and B, project withdrawals for mine slurry and mine domestic use were assumed to take place at three test wells near Leupp, Arizona, drilled in 2005, and

withdrawals for tribal use were distributed among the existing Sunshine well and the conceptual well locations (*table 1, fig. 7*). Scenario A simulates a maximum withdrawal rate of about 6,500 acre-ft/yr between 2010 and 2060 (*fig. 8A*). Scenario B simulates as much as 11,500 acre-ft/yr (*fig. 8B*). Scenario A represents the minimum projected demands for the area, whereas scenario B represents the maximum projected demands for the area. For scenario A, 6,000 acre-ft/yr is withdrawn for industrial use between 2010 and 2026; the remainder (500 acre-ft/yr) is withdrawn for tribal use. Between 2026 and 2028, industrial use is reduced to a total of 505 acre-ft/yr while the tribal use is ramped up toward 6,500 acre-ft/yr (*fig. 8A*). After 2029, all pumpage is used for tribal purposes. Industrial pumpage is extracted from three wells (PW-1, PW-2, and PW-3) that are simulated in three model cells (*table 1*). Tribal pumpage is extracted from seven wells (NPW-1 through NPW-7) that are simulated in seven model cells (*table 1*).

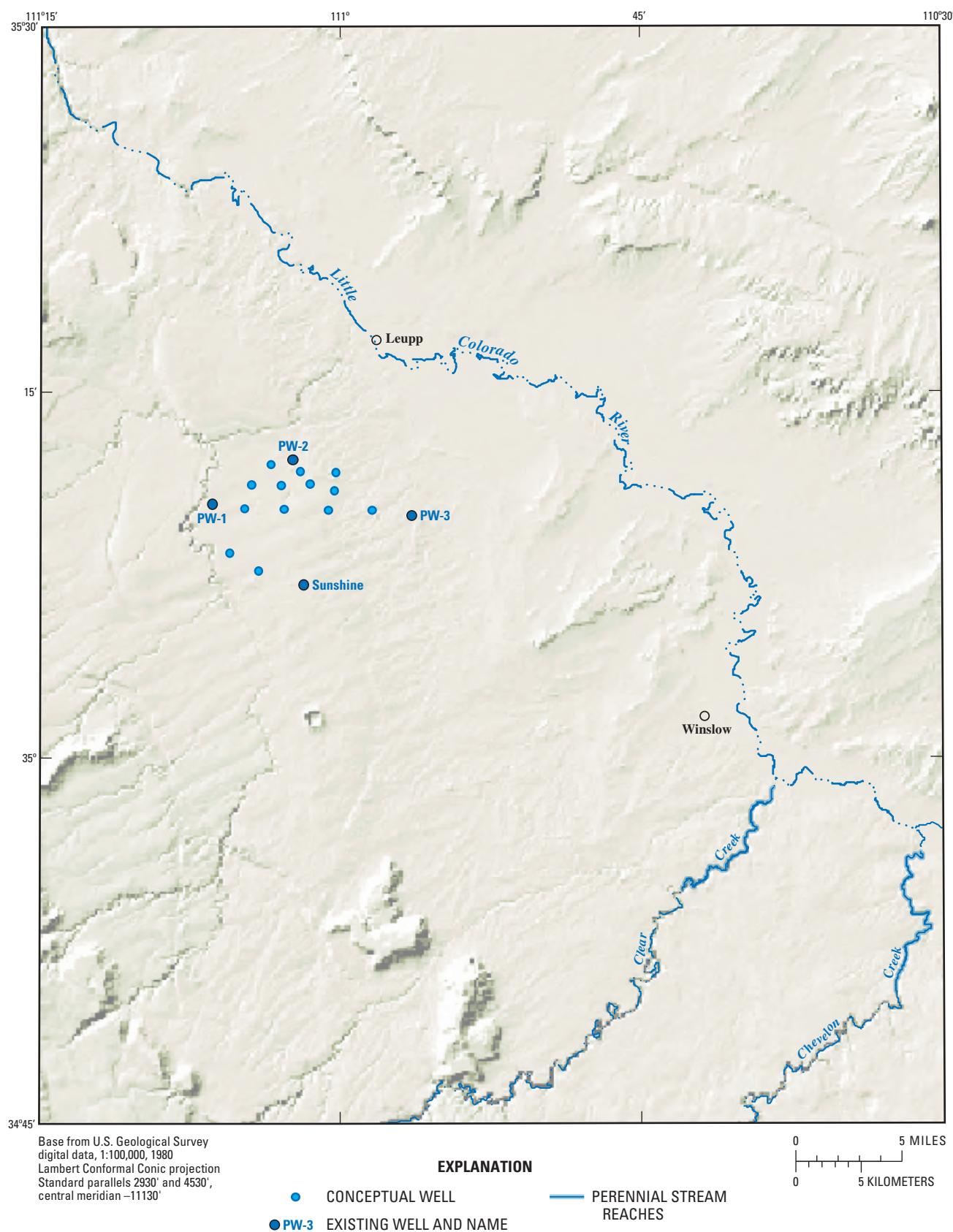**Figure 7.** Locations of existing and conceptual wells used in model simulation scenarios A and B for the C aquifer, northeastern Arizona.
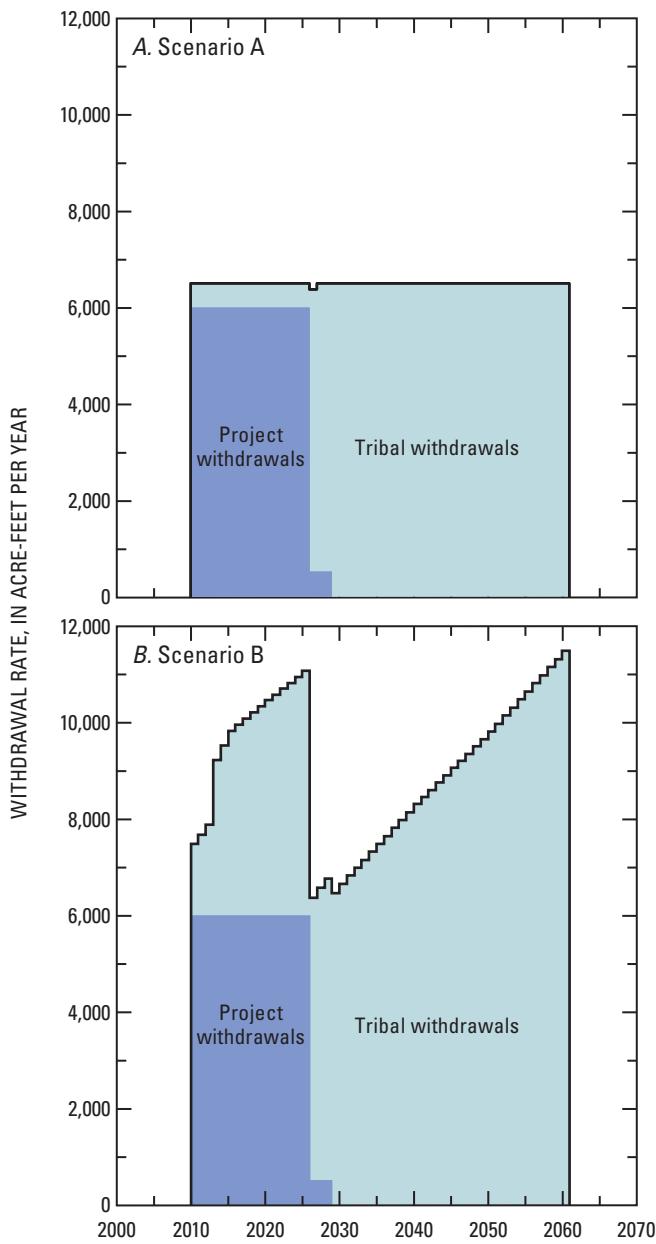
Figure 8. Withdrawal scenarios for model simulations of changes in the C aquifer, northeastern Arizona.

Table 1. Locations of withdrawals used in model simulation scenarios A and B for the C aquifer, northeastern Arizona.

Well name	UTM easting (meters)	UTM northing (meters)	Model row	Model column	Active in scenario A?	Active in scenario B?
Existing wells						
PW-1	490360	3892073	257	207	yes	yes
PW-2	496465	3895430	248	209	yes	yes
PW-3	505487	3891199	244	221	yes	yes
Sunshine	497291	3885968	256	218	no	yes
Conceptual wells						
NPW-1	492833	3891689	255	209	yes	yes
NPW-3	495797	3891654	252	212	yes	yes
NPW-6	499153	3891581	249	215	yes	yes
NPW-7	502460	3891581	246	218	yes	yes
NPW-2	493310	3893513	253	208	yes	yes
NPW-5	499732	3894482	246	213	yes	yes
NPW-4	497778	3893592	249	212	yes	yes
NPW-8	494834	3895033	250	208	no	yes
NPW-9	495593	3893475	251	210	no	yes
NPW-10	497032	3894553	248	211	no	yes
NPW-11	497471	3891556	251	214	no	yes
NPW-12	499629	3893075	247	214	no	yes
HPW-1	491677	3888360	259	211	no	yes
HPW-2	493875	3887001	258	214	no	yes

For scenario B, industrial withdrawal rates are the same as those in scenario A. The increased withdrawal rates in scenario B, relative to scenario A, result from increased withdrawals for tribal purposes. For scenario B, tribal withdrawals are assumed to occur in as many as 15 wells represented in 15 model cells (*table 1*).

The total volume withdrawn in the 51-year period is about 331,000 acre-ft for scenario A and about 464,000 acre-ft for scenario B. Because withdrawal locations are far from perennial stream reaches, depletion can occur long after withdrawals cease. This lack of immediate response occurs because recovery from shutting off withdrawals takes time to reach distant parts of the outward propagating cone of depression from the 51-year period of withdrawals. To simulate the continued depletion, each scenario therefore includes a 50-year period of no withdrawals.

Model Results

The model can make calculations of stream depletion and drawdown and display numbers with seven to eight significant digits of precision. For this discussion and the following section on the Monte Carlo analysis, most values of computed depletion of streamflow are less than 1 ft³/s. Numbers for computed depletion are presented to the nearest 0.01 ft³/s; however, this level of precision does not imply predictive accuracy to that level. Also, changes in base flow in the range of 0.01–0.1 ft³/s would be hard to detect in small streams without special monitoring equipment such as precision flumes. Furthermore, natural variations in base flow likely would be this amount or larger. This level, however, allows for comparison of relative amounts of computed depletion among different reaches, some of which have computed depletions in the range of hundredths of a cubic foot per second. Graphs constructed from model-calculated depletion rates use the full precision of numbers recorded in model runs. Drawdown values are presented here for 1- and 10-ft contour intervals.

The change model simulates stream depletions resulting from withdrawal scenarios A and B. For this analysis, the only withdrawals simulated were at rates and locations described in the previous section. For the 101-year simulation period, the original lengths of the five perennial reaches were assumed to remain unchanged by the simulated withdrawals or by any other stresses. This has the effect of representing a conservative or "worst-case" scenario of total stream depletion because actual stream depletion from the proposed withdrawals would be reduced if perennial reaches were shortened or disappeared during the simulation period. The model was configured to accumulate computed depletions separately for each of the five reaches at 5-year intervals from the start of the simulations and at additional times of 1, 51, and 101 years from the start of the simulations.

The total stream depletion from all five reaches at the end of the 101-year simulation period is about 20,000 acre-ft for scenario A and about 26,000 acre-ft for scenario B, about 6 percent of the total volume withdrawn for each scenario. About 94 percent of the depletion therefore would occur after the 101-year period, assuming that the major discharge mechanisms are represented in the model and that the withdrawals do not induce additional recharge to the aquifer.

The time distribution of the rate of total depletion from all five stream reaches simulated indicates a gradual increase in stream capture to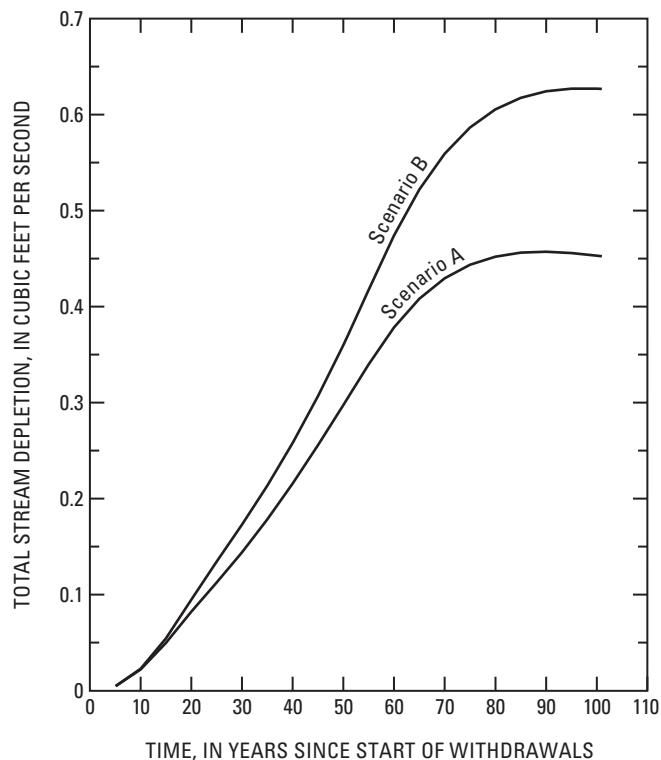 a maximum rate of depletion that occurs near the end of the 101-year simulation period (*fig. 9*). Computed depletion for scenario A was 0.31 ft^3/s at the end of the 51-year withdrawal period and increased to a maximum of about 0.45 ft^3/s around 95 years after the start of the simulation. Computed depletion for scenario B was 0.37 ft^3/s at the end of the 51-year withdrawal period and increased to a maximum of 0.63 ft^3/s around 95 years after the start of the simulation.

Of the total depletion shown in *figure 9* for scenarios A and B during the 101-year period, the relative amount occurring in each stream reach varies only slightly. The combined rates of depletion from upper Clear Creek, upper Chevelon Creek, and the Little Colorado River below Blue Spring increase from zero to about 3 percent of the total rate of depletion over the 101-year simulation period. The maximum computed depletion rate for any of these three reaches for scenario B is 0.015 ft^3/s. Because of the small values of computed depletion in these three reaches, only depletion in lower Clear Creek and lower Chevelon Creek will be discussed in detail in this report.

Computed rates of depletion for lower Clear Creek (*fig. 10* are about 83 to 84 percent of the total depletion rates for both scenarios. Most of the depletion occurs in lower Clear Creek because the creek is the closest stream reach to the withdrawal locations. For scenario A, computed depletion in lower Clear Creek was 0.26 ft^3/s at the end of the 51-year withdrawal period and increased to a maximum of about 0.39 ft^3/s around 85 years after the start of the simulation. For scenario B, computed

Figure 9. Total depletion for all stream reaches for model simulation scenarios A and B for the C aquifer, northeastern Arizona.
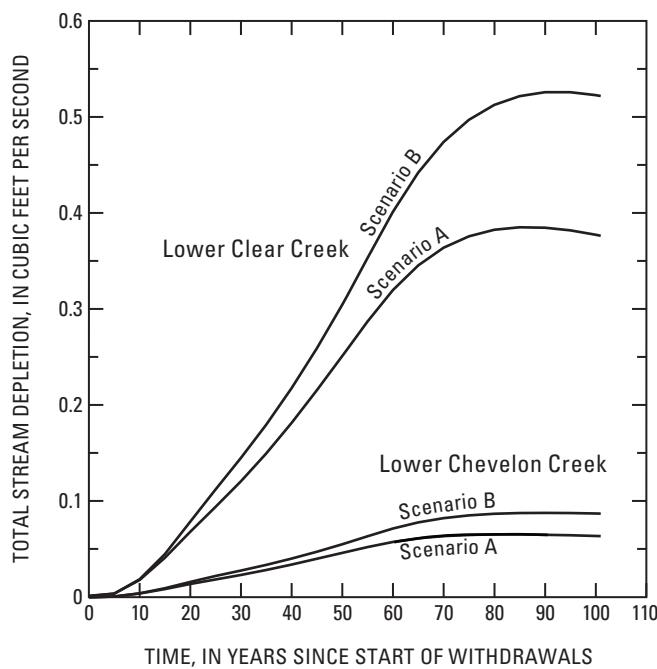

Figure 10. Depletion for lower Clear and Chevelon Creeks for model simulation scenarios A and B for the C aquifer, northeastern Arizona.

depletion was 0.31 ft^3/s at the end of the 51-year withdrawal period and increased to a maximum of about 0.53 ft^3/s around 90 to 95 years after the start of the simulation.

Computed rates of depletion for lower Chevelon Creek (*fig. 10*) are 14–17 percent of the total rate. Computed depletion for scenario A in lower Chevelon Creek reaches 0.05 ft^3/s at the end of the 51-year withdrawal period and continues to increase to a maximum of about 0.06 ft^3/s near the end of the simulation. Computed depletion for scenario B reaches 0.06 ft^3/s at the end of the 51-year withdrawal period and continues to increase to a maximum of about 0.09 ft^3/s near the end of the simulation.

Another use of the change model is to compute drawdown (negative change in head) in the aquifer. From the model results, time series of drawdown can be displayed for any model cell or the distribution of drawdown in the active model area can be displayed for different times in the 101-year simulation period. The model was configured to save head changes at 5-year intervals starting with 1 year and ending at 101 years. The model includes increases and decreases in withdrawals, and head change therefore could be negative (decreasing head) or positive (increasing head). For purposes of displaying drawdown in this report, all head changes were analyzed to determine the maximum drawdown at each active model cell for the 101-year period. Contours of 1 ft and 10 ft of maximum drawdown (*fig. 11*) reflect some of the large-scale features represented in the model. The contours bulge noticeably into the confined area, reflecting the lower storage factor in that area. The 1-ft contour is noticeably affected by the presence of lower Clear Creek, a source of water that limits the propagation of the computed cone of depression in this area.

Monte Carlo Analysis of Model Results

An indication of possible ranges of computed future stream depletions can be determined with a Monte Carlo analysis of scenarios A and B. The general procedure for these analyses is to (1) generate N sets of model parameters by perturbing each parameter with random numbers, (2) run the model with each parameter set, resulting in N sets of predictions, and (3) compute the statistical properties of the predictions. Ideally, the N sets of parameters should be generated by using standard deviations of parameters and the covariances between parameters determined by using an automatic calibration procedure.

Such an effort was attempted for the C aquifer change model using the Ground-Water Flow (GWF), Parameter Estimation (PES), Sensitivity (SEN), and Observation (OBS) Processes of MODFLOW-2000 (Harbaugh and others, 2000; Hill and others, 2000). For this purpose, a model was constructed to simulate changes in ground-water levels for the period 1961–2001, using five stress periods covering time periods 1961–1970, 1971–1980, 1981–1990, 1991–2000, and 2001, respectively. Withdrawal rates and locations for the first four periods for

various withdrawal centers (*fig. 12*, *table 2*) were taken from the Western Hopi Navajo analytical model of effects of withdrawals in the C aquifer (Bill Greenslade, Southwest Ground-water Consultants, Inc., written commun., 2005.) The final stress period, year 2001, was added because of the many water levels that were measured during that year and that were potentially useful for calibration of the model. Withdrawal locations and rates were assumed to be the same as those for the previous period, 1991–2000. For the Western Hopi Navajo withdrawal data set used here, exact locations of wells withdrawing water were not known. The Western Hopi Navajo Study assumed locations for two or more wells in each center that were generally within a 2- to 3-mi radius of the locations of the withdrawal centers shown on *figure 12*.

The OBS Process in the MODFLOW-2000 provides a means to enter observations for use in constructing an objective function. The PES Process uses the objective function, the sum of the squares of weighted differences between computed values and observations (weighted residuals), to select an optimum set of parameters. For a change model, potential observations could include changes in head and changes in some flow quantity (such as ground-water discharge to a stream) that occur in response to stresses such as withdrawals by wells. Flow data for the C aquifer were insufficient for use in the change model; however, ground-water databases of the USGS and Arizona Department of Water Resources contain more than 3,000 water-level measurements in wells in the C aquifer. Some of these form time series measurements for individual wells that could be used to compute head change for parts of the period of historical withdrawals from 1961–2001. The distribution of potentially useful time series of water-level change includes well locations adjacent to the Little Colorado River east of Winslow, locations in the Flagstaff area, and some scattered locations in the unconfined area south of the Little Colorado River.

The change model was configured to simulate historical changes using the same parameter zones and aquifer-thickness distribution described earlier for the change model used to compute stream depletion. The model for simulating historical change, however, includes additional stream reaches, shown on *figure 12*, that may have been connected to the C aquifer during the period 1961–2001. The added streams were simulated with the River Package in the same manner described earlier for the five streams in the model used to calculate future stream depletions. In addition to the streams, two ground-water discharge areas, McDonald and Obed Meadows (*fig. 12*) were added to the model. Areas were delineated from USGS topographic maps and registered with the model grid to compute meadow area in each of 12 model cells intersected. For each of those cells, conductance was calculated as the product of that area, an assumed vertical hydraulic conductivity of 1 ft/d, divided by an assumed bed thickness of 20 ft. The resulting bed conductance ranged from $1.4{\times}10^3$ to $9.7{\times}10^4$ ft^2/d with an average of $1.2{\times}10^4$ ft^2/d.
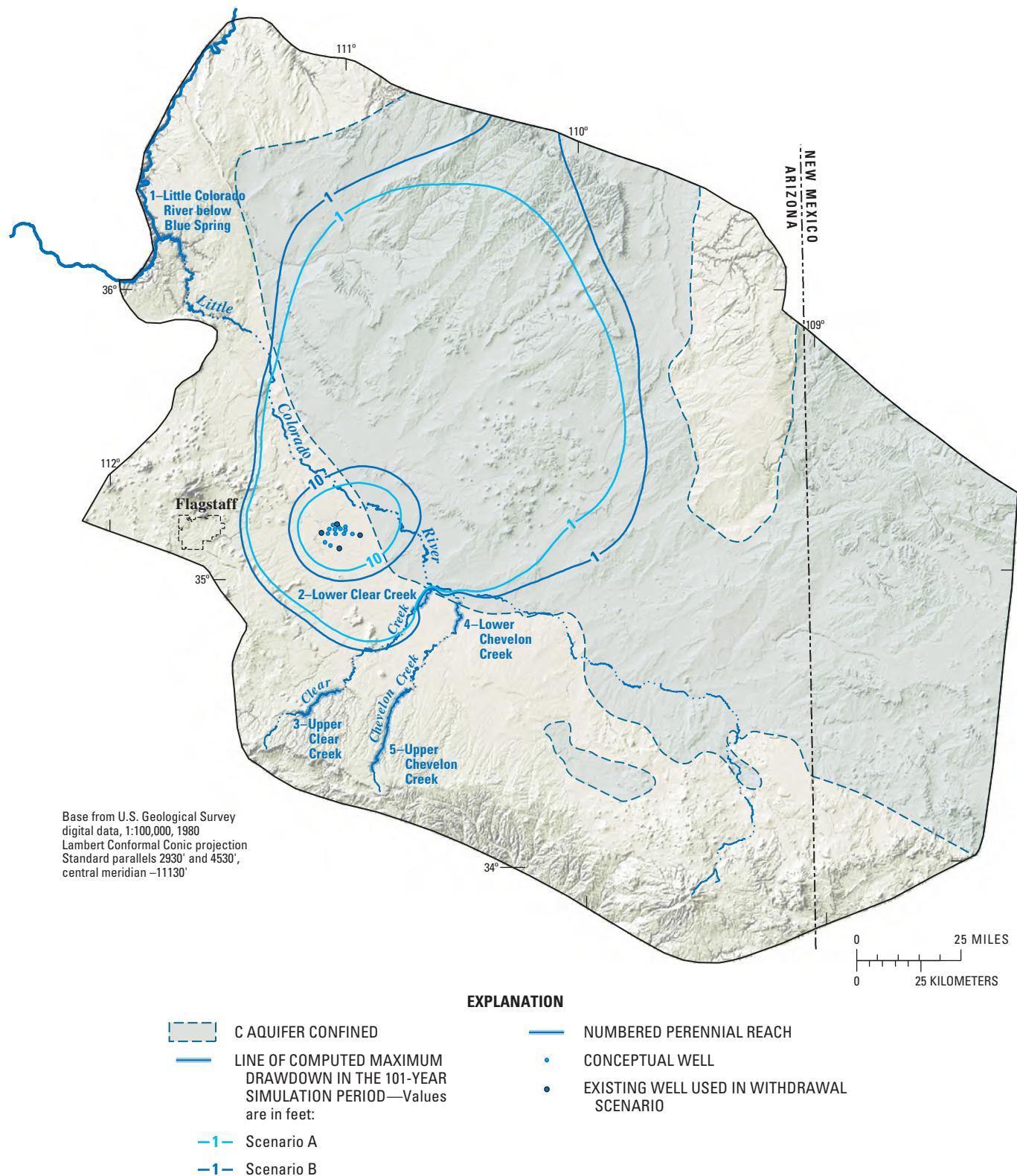
Figure 11. Computed maximum drawdown in the 101-year model simulation period for scenarios A and B for the C aquifer, northeastern Arizona.
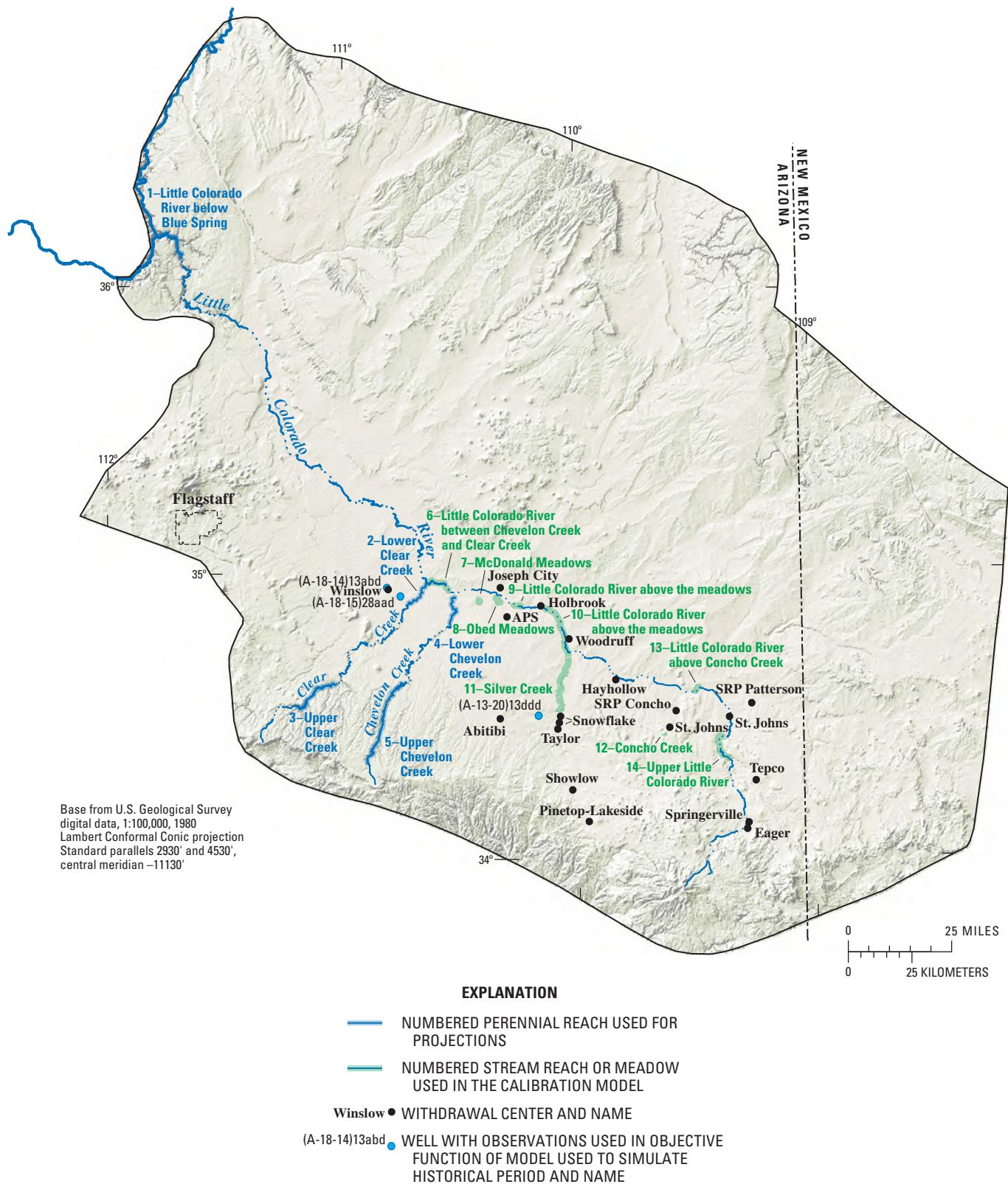
Figure 12. Locations of withdrawal centers, stream reaches, and wells with observations in a change model for the C aquifer, northeastern Arizona, that simulates historical conditions.

Table 2. Locations and rates of withdrawals from the C aquifer used in the Western Navajo Hopi Study change model to simulate historical conditions.

[Data source: Bill Greenslade, Southwest Ground-water Consultants, Inc. written commun., 2005]

Withdrawal center name	Water use	UTM easting (meters)	UTM northing (meters)	Withdrawal rate, in acre-feet per year			
				1961–1970	1971–1980	1981–1990	1991–2000
MANY FARMS	Municipal and industrial	624036	4023766	333	333	466	746
CHINLE	Municipal and industrial	630197	4001806	1133	1133	1587	2538
GANADO	Municipal and industrial	632017	3942729	701	701	982	1571
EAGER	Municipal and industrial	657570	3775603	650	650	990	1415
FLAGSTAFF	Municipal and industrial	435464	3892934	480	1800	4934	7235
HOLBROOK	Municipal and industrial	576866	3862326	1469	1469	1504	1539
PINETOP-LAKESIDE	Municipal and industrial	595824	3778254	380	380	453	674
SHOW LOW	Municipal and industrial	589344	3790577	1287	1287	1961	2801
SNOWFLAKE	Municipal and industrial	584571	3819265	474	474	564	839
SPRINGERVILLE	Municipal and industrial	658082	3778074	322	322	490	700
ST JOHNS	Municipal and industrial	650498	3819268	100	100	210	281
TAYLOR	Municipal and industrial	583426	3813899	350	350	417	621
WINSLOW	Municipal and industrial	517320	3868734	1725	1725	1878	2494
TEPCO	Industrial	660918	3794561	0	0	2484	8854
APS	Industrial	563685	3858068	3500	3182	7754	13335
SRP CONCHO	Industrial	629633	3821469	0	439	4705	5216
SRP PATTERSON	Industrial	659160	3824565	0	370	3174	3461
ABITIBI	Industrial	561051	3818273	8446	11430	15687	17415
JOSEPH CITY	Agriculture	561013	3869422	3163	7762	7815	8950
WOODRUFF	Agriculture	587813	3849446	253	621	625	716
HAY HOLLOW	Agriculture	606106	3833548	506	1242	1250	1432
SNOWFLAKE	Agriculture	584069	3816804	5568	13660	13754	15752
ST JOHNS	Agriculture	627191	3815033	3163	7762	7815	8950

Examples of computed and observed water-level time series for three locations south of the Little Colorado River [A-18-15)28aad, (A-18-14)13abd, and (A-13-20)13ddd] (*fig. 12*) show varying degrees of goodness of fit (*fig. 13*) using the parameters described earlier for the streamflow-depletion analysis. The first of these time series is for well (A-18-15)28aad, about 3 to 4 mi west of the start of the perennial part of lower Clear Creek. The time series starts in 1969, and the water level declines slightly more than 1 ft by the end of 2001. Observed head changes from 1982 to 1986 were not used because they were above zero and no process in the model could account for positive head changes. In the early 1980s, climate conditions were wetter than normal, and the positive values of head change in the well during and after this period are evidence that the aquifer responded to high runoff in lower Clear Creek and (or) other recharge. The model calculated about 5.7 ft of head decline at this location, significantly more than observed. The second time series, 1968–2001, is for well (A-18-14)13abd near the Winslow withdrawal center. Some of the individual observed values may have been influenced by recent and (or) nearby withdrawals, but the model reasonably matches the overall trend. The third time series, 1963–1985, is for well (A-13-20)13ddd near Snowflake. Observed declines are more than 40 ft over this period; the model underestimates the total decline at this well.

The entire procedure for the Constrained Monte Carlo analysis is as follows:

1. Generate N sets of parameters by perturbing values of horizontal-hydraulic conductivity for four zones, specific yield in the unconfined area, specific storage in the confined area, and riverbed conductance for all reaches.

2. For each of the N sets of parameters, run the model of historical conditions and compute the value of an objective function on the basis of the observed and computed head-change values shown in *figure 13*. Rank the N parameter sets from lowest to highest values of the objective function. Select the top M parameter sets on the basis of the lower objective function values.

3. Run the streamflow-depletion model simulating scenarios A and B with each of the top M parameter sets and compute the mean and standard deviation of the results of computed streamflow depletion in lower Clear and Chevelon Creeks.

The names of the seven model parameters allowed to vary in the Monte Carlo analysis are given in *table 3*. To generate the new parameters outlined in step 1 above, all but specific yield were log-transformed. For the log-transformed parameters, the equation for computing a perturbed value of the nth parameter, $p_n{'}$, from the base value of the parameter, p_n, is

$$p_n{'} = 10^{\left(\log_{10} p_n + \alpha_i \bar{\sigma}_n\right)},$$

where α_i is the ith number in a series of normally distributed random numbers with a mean of 0.0 and a standard deviation of 1.0 and $\bar{\sigma}_n$ is the standard deviation of the nth parameter, in $\log_{10}$ space. The equation used for generating perturbed values of the untransformed parameter, specific yield (S_y), is

$$S_y{'} = S_y + \alpha_i \sigma_{Sy},$$

where $S_y{'}$ is the perturbed specific yield, and σ_{Sy} is the standard deviation of specific yield. In generating the new parameters by using the above equations, values outside the upper and lower bounds given in *table 3* were rejected and a new value was computed by using the next random number in the series. Values of and the upper and lower bounds in *table 3* were selected on the basis of hydrologic judgment.

For the first step, 1,829 parameter sets were generated and used in the model that simulates the historical period of withdrawals as outlined in step 2. The objective function for each of these sets was computed using the head-change observations from the three time series in *figure 13*. In computing the objective function, the first point in each series is the head value from which the head changes were calculated, and thus this point was not included in the objective function. The standard deviation of measurement error (input value STATdd in Hill and others, 2000) was set to 1.0 for the head-change observations for wells (A-18-14)13abd and (A-13-20)13ddd and was set to 0.5 for the observations for well (A-18-15)28aad. This has the effect of increasing the weights for observations in the latter series, which is needed because the head-change values there are small relative to the changes in the two other series and would otherwise not contribute much to the objective function. The resulting values of the objective function ranged from 1.063×10^3 to 7.35×10^4. For comparison, the objective function for a simulation using the base parameter values is 2.405×10^3. Many parameter sets resulted in lower objective-function values for these observations than the value from the base run using the parameter sets selected for the simulation of scenarios A and B. The use of the model to simulate historical changes at three sites, however, is not considered to be a rigorous calibration for selecting parameters. The process is instead a way of selecting the most reasonable parameter sets for use in the Monte Carlo analysis and of eliminating the worst sets in matching change observations at the three sites selected.

For step 3, the sets of parameters with objective functions less than 1.5 times the objective function from the base run were used. This resulted in 900 parameter sets with objective functions ranging from 1.063×10^3 to 3.604×10^3. For each run of the streamflow-depletion model, calculated depletions in lower Clear and lower Chevelon Creeks were saved at 5-year intervals over the 101-year simulation period with additional values saved at the end of the first time step, 1 year; the end of the period of withdrawals, 51 years; and at the end of the simulation, 101 years. The saved values provide a data set for calculating basic statistics from the results for the entire simulation period. For each time interval saved, the mean and standard deviation of the 900 results were computed.
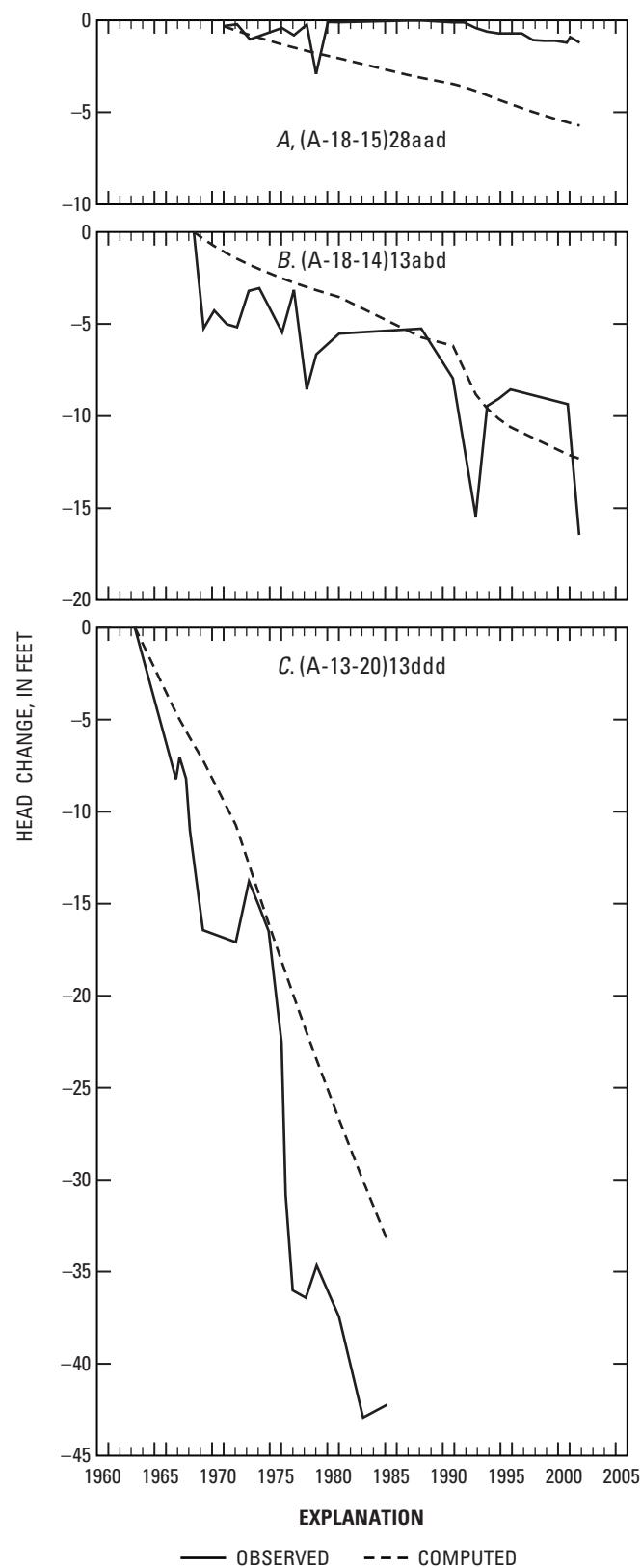
Figure 13. Observed and computed head change for three wells completed in the C aquifer, northeastern Arizona.

Table 3. Parameters used in the Monte Carlo analysis of computed stream depletions in the study area, northeastern Arizona

[Standard deviations are in transformed space for HK1, HK2, HK3, HK4, SS1, and KRB and are in native space for SY1. Upper and lower bounds are in native space. — , unitless}

Parameter name	Parameter description	Untransformed base value	Units	Transformation type	Standard deviation, σ	Upper bound	Lower bound
HK1	Horizontal hydraulic conductivity in zone 1	6	Per foot	$\log_{10}$	0.5	60	0.6
HK2	Horizontal hydraulic conductivity in zone 2	5	Per foot	$\log_{10}$	0.5	50	0.5
HK3	Horizontal hydraulic conductivity in zone 3	5	Per foot	$\log_{10}$	0.5	50	0.5
HK4	Horizontal hydraulic conductivity in zone 4	5	Per foot	$\log_{10}$	0.5	50	0.5
SY1	Specific yield in unconfined area	0.06	—	none	0.02	0.12	0.01
SS1	Specific storage in confined area	2×10^{-6}	Per foot	$\log_{10}$	1	2×10^{-5}	1×10^{-7}
KRB	Multiplier for riverbed conductance values for stream reaches used for projections	100	—	$\log_{10}$	2	1000	0.1

The results for scenarios A and B in *figure 14* show the upper and lower curves computed by adding and subtracting the standard deviation from the mean for each time interval saved. In all cases, the mean depletion from the 900 Monte Carlo runs is higher than the results for the model using the parameters selected for the depletion calculations. The first reason for the higher mean is that the model does not include a way for negative streamflow depletion to occur, and therefore the lower bound for streamflow depletion is zero. There is no corresponding bound for positive streamflow depletion, however, and the mean values from the Monte Carlo results therefore are skewed higher than the results using the standard parameters. The second reason is that the elimination of parameter sets with higher objective functions resulted in parameter distributions with median values that differ from the values of the corresponding parameters in the forward run. For example, the median of the 900 HK1 values used in the Monte Carlo analysis is 9.3 ft/d, and the value of HK1 in the forward run is 6 ft/d. The higher median HK1 value leads to a higher mean depletion rate. In computing mean minus one standard deviation, the values less than zero in the lower curves on *figure 14* were truncated at zero. The upper curves (mean plus one standard deviation) are substantially larger than the depletions simulated in the forward run, but the formulation of the Monte Carlo analyses does set upper bounds for reasonable parameter values. These curves can be viewed as a general measure of uncertainty in results using these limits and assumed standard deviations of parameters. Analyses of uncertainty could be improved with additional work in estimating parameters and determining parameter correlations, as well as in data collection to improve the model. The curves of estimated depletion in *figure 10* should be used as the best estimates of computed depletion using the change model documented in this report.

Limitations of the Model

The ground-water change model of the C aquifer described in this report was designed specifically to compute the possible effects of ground-water withdrawals in an unconfined part of the aquifer near Leupp, Arizona, about 25 mi from the nearest connected surface-water feature, lower Clear Creek. Treatment of the aquifer as a porous medium with generalized aquifer properties is reasonable for this scale of simulation. This model should not be used for purposes such as evaluation of possible drawdown in and around well fields because local conditions such as flow in fractures and heterogeneities not represented in the model may be important at that scale.

The model also was not designed to evaluate the effects of existing withdrawals throughout the C aquifer on streams of interest including lower Clear and Chevelon Creeks. That purpose would require a calibrated flow model, rather than a change model. A related caution is that the model should not be used to evaluate the effects of withdrawals in areas other than near Leupp. The perimeter boundaries are distant from this area so that possible errors in placement or types of these boundaries will not affect the calculation of depletion in stream reaches of interest. This is not true, however, for an area such as Flagstaff, which is near a model boundary. Also, in the attempt to calibrate the change model, many observations near the confined–unconfined boundary could not be matched. More work would need to be done before the model could be used with confidence to evaluate the effects of withdrawals in those areas.

The two scenarios included simulations of 51 years of withdrawals followed by 50 years of no withdrawals. A key assumption was that the lengths of the perennial reaches remain the same over this period. This assumption is conservative in that the change model calculates the maximum total depletions for lower Clear and Chevelon Creeks. If combinations of the proposed withdrawals, other C-aquifer withdrawals, and extended drought were to cause the surface-water reaches to shrink or dry up, then depletion in lower Clear and Chevelon Creeks would be less than was indicated by the model. By the principle of conservation of mass, however, the total volume of outflow captured by the withdrawal scenarios would be the same—only the timing and locations of the captured outflow would vary.
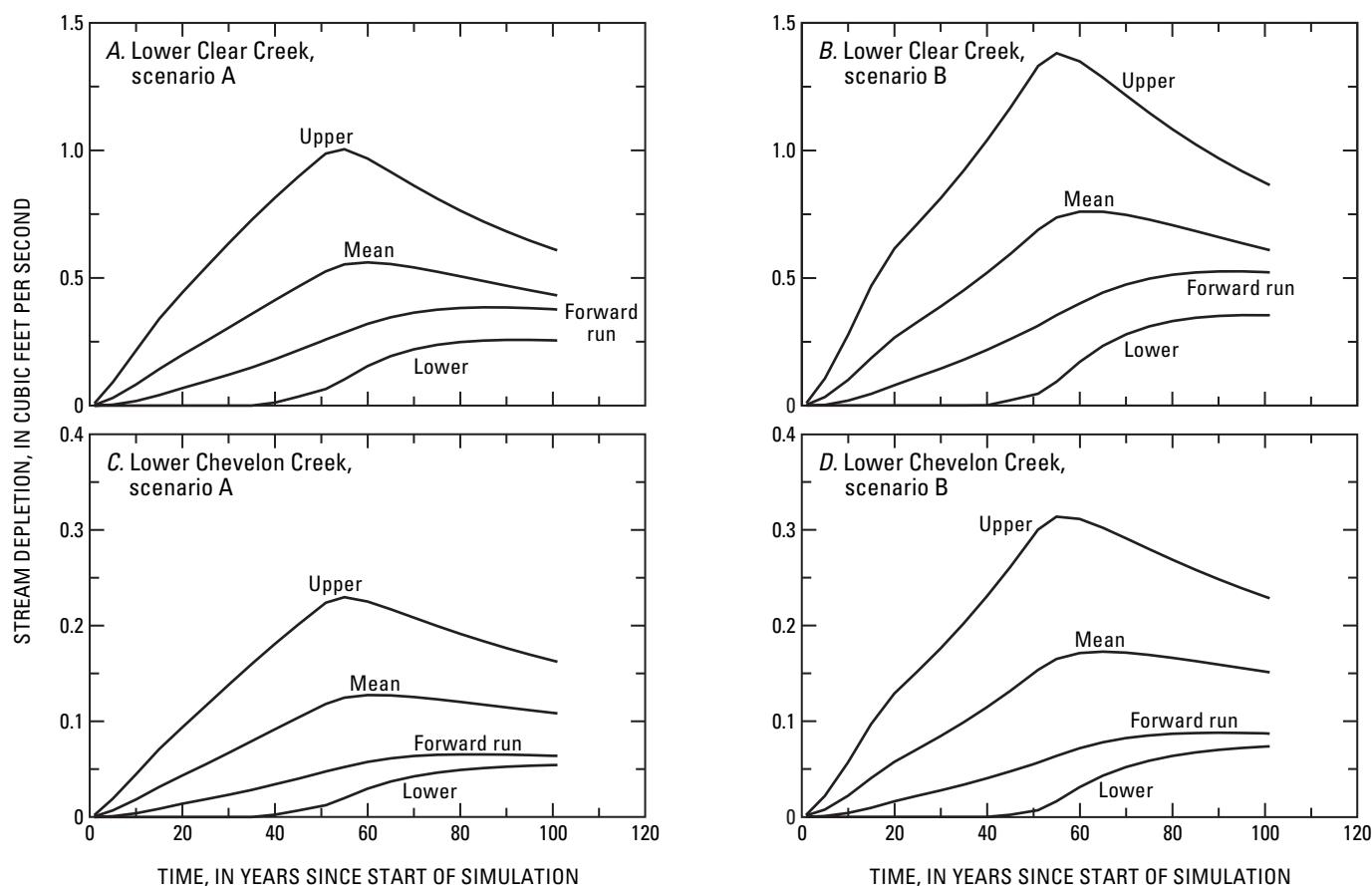

Figure 14. Results of Monte Carlo analyses of computed stream depletions in the study area, northeastern Arizona. Mean, upper and lower curves are mean results, mean plus one standard deviation, and mean minus one standard deviation, respectively. Forward-run curves are results from standard set of parameters, also shown on figure 10. *A*, Lower Clear Creek, scenario A. *B*, Lower Clear Creek, scenario B. *C*, Lower Chevelon Creek, scenario A. *D*, Lower Chevelon Creek, scenario B.

Summary and Conclusions

A ground-water change model was constructed for the purpose of evaluating possible effects of withdrawals from the C aquifer in the area of Leupp, Arizona, on connected surface-water features including Clear Creek, Chevelon Creek, and the Little Colorado River below Blue Spring. The model was constructed using the USGS model code MODFLOW-2000. The approximate aquifer thickness was mapped and four zones were used for the distribution of hydraulic conductivity. The model includes confined and unconfined areas where storage properties are represented with values of the product of specific storage and thickness, and specific yield, respectively. A limited calibration was attempted using historical withdrawals and observed head changes. The final model, however, used values of hydraulic conductivity and other properties that were deemed representative of large zones in the aquifer.

Evaluations of possible stream depletion included evaluations of scenarios A and B, with maximum withdrawal rates of about 6,500 and 11,500 acre-ft/yr, respectively. The scenarios included a 51-year period of withdrawals followed by a 50-year period of no withdrawals. The computed volume of stream depletion for the 101-year period was about 6 percent of the ultimate volume that would occur as a result of the scenarios. For both scenarios, computed maximum depletion rates in upper Clear Creek, upper Chevelon Creek, and the Little Colorado River below Blue Spring were on the order of hundredths of a cubic foot per second; detailed time series of these results were not presented. Most of the computed depletion was in lower Clear Creek; computed rates reached about 0.4 ft^3/s for scenario A and about 0.5 ft^3/s for scenario B near the end of the simulation. Computed depletion in lower Chevelon Creek is less than that in lower Clear Creek because Chevelon Creek is more distant from the withdrawal locations, and drawdown reaches Clear Creek first.

A Monte Carlo analysis was carried out to compute statistical properties of streamflow depletion using a range of parameter sets. The analysis was constrained by using a change model that simulated a period of historical withdrawals from 1961 to 2001. An objective function that used observations from 3 relevant well locations was computed for each of 1,829 sets of 7 model parameters, and the 900 sets with the best objective function were used to calculate a set of streamflow depletions by using the model that simulated the proposed withdrawals for scenarios A and B. Statistical results show that the mean and mean plus and minus standard deviations are skewed to the high side because the streamflow is bounded on the low side by zero.

The change-model approach seems to be reasonable for calculating the possible effects of withdrawals near Leupp. The model is not designed or calibrated for other purposes. Evaluation of the cumulative effects of the proposed withdrawals and existing withdrawals would best be done with a calibrated ground-water flow model.

References

Bills, D.J., Truini, Margot, Flynn, M.E., Pierce, H.A., Cathings, R.D., and Rymer, M.J., 2000, Hydrogeology of the regional aquifer near Flagstaff, Arizona, 1994–97: U.S. Geological Survey Water-Resources Investigations Report 00-4122, 142 p.

Blakey, R.C., 1990, Stratigraphy and geologic history of Pennsylvanian and Permian rocks, Mogollon Rim region, central Arizona and vicinity: Geological Society of America Bulletin, v. 102, p. 1189–1217.

Brown, D.E., Carmony, N.B., and Turner, R.M., compilers, 1981, Drainage map of Arizona showing perennial streams and some important wetlands: Phoenix, Arizona Game and Fish Department, scale 1:100,000.

Cooley, M.E., Harshbarger, J.W., Akers, J.P., and Hardt, W.F., 1969, Regional Hydrogeology of the Navajo and Hopi Indian Reservations, Arizona, New Mexico, and Utah: U.S. Geological Survey Professional Paper 521-A, 61 p.

Daniel, Debra, 1981, Total dissolved solids content of the aquifers of the Basin and Range and the C-aquifer, Arizona Department of Water Resources, 1 pl.

Fenneman, N.M., 1946, Physical divisions of the United States: U.S. Geological Survey map, scale 1:700,000.

Harbaugh, A.W., Banta, E.R., Hill, M.C., and McDonald, M.G., 2000, MODFLOW-2000, the U.S. Geological Survey modular ground-water model—User guide to modularization concepts and the Ground-Water Flow Process: U.S. Geological Survey Open-File Report 00–92, 121 p.

Hart, R.J., Ward, J.J., Bills, D.J., and Flynn, M.E., 2002, Generalized hydrogeology and ground-water budget for the C Aquifer, Little Colorado River basin and parts of the Verde and Salt River basins, Arizona and New Mexico: U.S. Geological Survey, Water-Resources Investigations Report 02–4026, 54 p.

Hill, M.C., Banta, E.R., Harbaugh, A.W., and Anderman, E.R., 2000, MODFLOW-2000, the U.S. Geological Survey modular ground-water model—User guide to the Observation, Sensitivity, and Parameter Estimation Processes and three post-processing programs: U.S. Geological Survey Open-File Report 00–184, 209 p.

Hoffmann, J.P., Phillips, J.V., Bills, D.J., and Halford, K.J., in press, Geologic, hydrologic, and chemical data from the C aquifer, near Leupp, Arizona: U.S. Geological Survey Scientific Investigations Report 2005–5280.

Leake, S.A. and Claar, D.V., 1999, Procedures and computer programs for telescopic mesh refinement using MODFLOW: U.S. Geological Survey Open-File Report 99–238, 53 p.

Mann, L.J., 1976, Ground-water resources and water use in southern Navajo County, Arizona: Phoenix, Arizona Water Commission Bulletin 10, 106 p.

Mann, L.J., and Nemecek, E.A. 1983, Geohydrology and water use in southern Apache County, Arizona: Phoenix, Arizona Department of Water Resources Bulletin 1, 86 p.

McDonald, M.G. and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, Book 6 Modeling Techniques, Chap. A1, 586 p.

McDonald, M.G. and Harbaugh, A.W., 1996, User's documentation for MODFLOW-96, an update to the U.S. Geological Survey modular finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 96–485, 56 p.

McGavock, E.H., Anderson, T.W., Moosburner, Otto, and Mann, L.J., 1986, Water resources of southern Coconino County, Arizona: Phoenix, Arizona Department of Water Resources Bulletin 4, 53 p.

Peter Mock Groundwater Consulting, Inc., 2003, in HDR Engineering, Inc. Report of Findings for the Assessment of Western Navajo-Hopi Water Supply Needs, Alternatives, and Impacts, Volume 3, Appendix B-3, variously paged.

Reilly, T.E., Franke, O. Lehn, and Bennett, G.D., 1987, The principle of superposition and its application in ground-water hydraulics: U.S. Geological Survey Techniques of Water-Resources Investigations Book 3, Chap. B6, 28 p.

Southwest Ground-water Consultants, Inc., 2003, in Report of findings for the assessment of Western Navajo-Hopi water supply needs, alternatives, and impacts, v. 3, Appendix B-1: HDR Engineering, Inc for U.S. Bureau of Reclamation, 2003, 10 p.

Theis, C.V., 1935, The relation between lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage: Transactions of the American Geophysical Union, v. 16, p. 519–524.

U.S. Geological Survey, 2005, Elevation Derivatives for National Applications, accessed June, 15, 2004, at URL *http://edna.usgs.gov/*

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