SIMULATION OF GROUND-WATER FLOW IN ALLUVIAL BASINS IN SOUTH-CENTRAL ARIZONA AND PARTS OF ADJACENT STATES

REGIONAL AQUIFER-SYSTEM ANALYSIS

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Simulation of Ground-Water Flow in Alluvial Basins in South-Central Arizona and Parts of Adjacent States

By T.W. ANDERSON *and* GEOFFREY W. FREETHEY

REGIONAL AQUIFER-SYSTEM ANALYSIS—
SOUTHWEST ALLUVIAL BASINS, ARIZONA AND ADJACENT STATES

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1406-D

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) program was started in 1978 after a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA program represents a systematic effort to study a number of the Nation's most important aquifer systems which, in aggregate, underlie much of the country and which represent important components of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system, and accordingly, transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies, both to develop an understanding of the natural, undisturbed hydrologic system, and of any changes brought about by human activities, as well as to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA program is assigned a single Professional Paper number and where the volume of interpretive material warrants, separate topical chapters that consider the principal elements of the investigation may be published. The series of RASA interpretive reports begins with Professional Paper 1400 and thereafter will continue in numerical sequence as the interpretive products of subsequent studies become available.

Gordon P. Eaton

Gordon P. Eaton
Director

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
acre	0.004047	square kilometer
square mile (mi^2)	2.590	square kilometer
gallon per minute (gal/min)	0.06309	liter per second
foot per second (ft/s)	0.3048	meter per second
foot squared per day (ft^2/s)	0.0929	meter squared per day
foot per mile (ft/mi)	0.1894	meter per kilometer
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
cubic foot per second per square mile [$(ft^3/s)/mi^2$]	0.01093	cubic meter per second per square kilometer
acre-foot (acre-ft)	0.001233	cubic hectometer

Sea Level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

SIMULATION OF GROUND-WATER FLOW IN ALLUVIAL BASINS IN SOUTH-CENTRAL ARIZONA AND PARTS OF ADJACENT STATES

By T.W. Anderson and Geoffrey W. Freethey

ABSTRACT

Numerical modeling was used to examine the characteristics of ground-water flow systems in the alluvial basins of south-central Arizona and parts of adjacent States and to evaluate similarities and differences among basins. The 72 alluvial basins of the study area were categorized into five groups that have similar geohydrologic characteristics and similar requirements for model development. Basin categorization is based on the relative importance of various components of the geologic and hydrologic systems. The principal geologic differences among groups are the presence and extent of fine-grained sediments in the basin center and the presence of saturated stream alluvium. These units represent the least and most productive parts of the aquifer, respectively. The principal hydrologic differences among groups are magnitude of the component of mountain-front recharge and presence of a perennial stream in the basin. These groups were termed southeast, central, west, Colorado River, and highland.

Ground-water flow models of 12 specific basins were developed to evaluate the utility of transferring hydrologic concepts and property values within and among groups of basins. In addition, two models representing generalized basin systems were developed that incorporated average physical and hydrologic characteristics of the southeast basin group in one model and of the central and west basin groups in the other. These basin-group models were used, along with the specific-basin models, to examine the degree of model sensitivity to changes in basin shape and the values and areal distributions of hydrologic properties.

Each model of a specific basin was based on available field data and previous conceptual or numerical models. Quantitative determinations of model-data requirements were used where field data were available. Transferability of information among basins of a particular group was tested and generally was successful. The properties that were most important to the successful development of basin models were summarized in groups including hydrologic boundaries, flow components, and hydrologic properties.

Systematic model-sensitivity analyses were done on all models to qualitatively evaluate how uncertainties in hydrologic property values and distributions might affect results of the simulations. Sensitivity analyses indicated that simulations of basins in the same group are affected most by the same changes in model input. The degree of sensitivity of a model to a specific property depends on the geologic setting of a basin, the magnitudes of predevelopment downvalley flow, and the degree of development that has taken place.

Changes in hydraulic conductivity, natural recharge quantities, and extent and integrity of confining layers affected computed water levels mainly in models of southeast basins. Changes in specific yield and location and quantity of pumping affected computed water levels the most in models of central and west basins in which storage depletion is common. Changes in quantity of water moving between streams and the aquifer, evapotranspiration, and the return quantity of excess applied irrigation water affected computed water levels the most in models of the Colorado River and probably of the highland basins in which the hydrologic system is dominated by a perennial stream.

INTRODUCTION

The Southwest Alluvial Basins, Regional Aquifer-Systems Analysis (Swab/RASA), is one of a series of projects designed to systematically evaluate the Nation's major ground-water systems (Bennett, 1979). The Southwest Alluvial Basins study began in October 1978 and was one of 29 systems included in the RASA Program (see Foreword).

The Swab/RASA study area (fig. 1) includes about 82,000 mi^2 and consists of 72 basins, most of which are virtually independent hydrologic systems. The physical character of the basins is diverse, although similarities exist in geohydrologic character. Tectonic activity, which formed the Basin and Range physiographic province (Fenneman, 1931), was areally variable. Rocks that bound the individual basins differ areally and are the source of detrital material that subsequently filled the basins. Deposition of the alluvial material that constitutes the aquifer(s) probably was unique in each basin.

The basins of the study area are considered a regional system for several reasons:

- Several basins are linked by natural hydraulic connection in which outflow from one basin represents inflow to a downgradient basin.
- The hydrologic response to development of the water resources and water-use characteristics generally have been similar.
- Ground-water resources of the basins represent a common element in the economy of the entire region.

This study integrated and explored some of the concepts used to evaluate individual ground-water systems and to compare the individual basin characteristics and hydrologic properties.

The overall purpose of the Swab/RASA study was to describe and define the hydrologic systems in the alluvial basins. The study included quantitative estimates and qualitative descriptions of the flow systems before and after development. Numerical models were developed to simulate ground-water flow in specific individual basins and in basin systems that represent individual groups of basins. The study emphasized ground-water flow simulation and was designed to—

1. Simulate ground-water flow in a few representative basins of the 72 basins in the study area.
2. Analyze findings related to geologic and hydrologic patterns that exist throughout the study area.
3. Document similarities and differences in basin and regional geohydrology to support the use of information transfer from basins that were modeled to those that were not modeled.

PURPOSE AND SCOPE

The findings of the Swab/RASA study are described in U.S. Geological Survey Professional Papers 1406–A through 1406–D, including this report, Professional Paper 1406–D. This report is closely related to Professional Paper 1406–B (Anderson and others, 1992), which describes the geohydrology and water resources of the Swab/RASA study area.

This report describes the modeling approach used in the study, the design and calibration of the ground-water models, and the results from specific-basin and basin-group models. The models were developed by using geologic and hydrologic knowledge of the systems as presented by Anderson and others (1992) and in other previous studies. The emphasis of the report is on documentation of the knowledge gained from the study of geologic and hydrologic similarities and differences among the basins of the study area. The models provide a means of testing the general understanding of the hydrologic-system function of the basins. The report also includes a general discussion of the modeling techniques used to simulate the response of a system or part of a system to a specific stress.

The transferability of information and the limitations of interpretations are discussed, and all facts and assumptions on which the interpretations depend are presented. Possible use of the models for predictive and management tools is also discussed, although the models developed as part of this study generally were not used to predict possible future conditions.

APPROACH

The general approach to this study was to analyze available data and develop numerical models of selected basins for which data were adequate for history matching. The study rationale assumes that some system functions are common and that selected information is transferable among basins or groups of basins. Information transfer was used as a means of quantitatively defining properties of the hydrologic systems and was evaluated through the use of numerical models. The importance of various hydrologic properties and flow components that define the ground-water systems also was evaluated for predevelopment and development conditions. The information-transfer approach was used to reduce the need for extensive data collection and to reduce disparities in hydrologic-system definition that result from different political jurisdictions and economies.

For studying individual basins as hydrologic units, the study area was divided into 72 basins. The boundaries were selected to represent surface-water divides, zones of minimal interbasin connection, or in some instances, arbitrary boundaries through areas of minimum ground-water development. As defined, the basins serve as the smallest practical unit for analysis and numerical modeling. Each basin can be modeled individually with minimal boundary effects because contiguous areas were not included; basins can be readily combined to model or analyze larger areas.

The 72 basins were categorized on the basis of physiographic, geologic, and hydrologic properties and water chemistry to develop and evaluate information transfer that could be used to define and describe both the individual basin systems and the regional system. The effects of poorly known system properties were explored by using a range of values in the model and evaluating the effect of each range of values on model performance. The results indicated basin conditions for which detailed and accurate values of hydrologic properties are needed and, conversely, conditions for which general approximations of these properties are acceptable.

The basins of the study area were grouped initially into categories on the basis of discernible surface features—climate, runoff, geology, altitude, shape, and general physiography—that were thought to have an influence on the ground-water hydrology of the basin (Anderson and others, 1992). A typical basin that had sufficient data available for model construction was then selected from each group, and a numerical model of the ground-water flow system was developed. Subsequently, the information gained from this model was transferred to one or several other basins for which fewer data were available. Models were then developed for those basins. If the information remained acceptable for use in the model when transferred to the basins with fewer data, then greater confidence could be placed in its applicability to basins that are less developed or undeveloped and for which virtually no data are available. Tests of the adequacy of data to define hydrologic-system properties and sensitivity of model response to variations in property values were made on all models to relate those results to the types and magnitude of data-collection programs. Two basin-group models were developed by using average values of physical and hydrologic properties of the southeast basins and of the central and west basins. These basin-group models were used to test sensitivity of hydrologic response to variations and generalizations of properties used in the models.

Data used in the numerical models consisted of information that is available for most developed basins in the Swab/RASA study area. Recharge rates were estimated from average annual precipitation. Values and areal distributions of transmissivity, hydraulic conductivity, and specific yield were estimated from aquifer-test results, specific-capacity values, or examination of descriptions of samples in drillers' logs. Flow-net analyses were used to estimate distributions of hydraulic conductivity, transmissivity, and mountain-front recharge for initial use in the models. Alternative values and areal distributions of the hydrologic properties were tested in the ground-water flow models through sensitivity analyses.

DESCRIPTION OF THE STUDY AREA

The study area encompasses about 82,000 mi^2 in southern and central Arizona, western New Mexico, and small parts of southeastern California and Nevada (fig. 1). The study area includes the surface-water drainage of the Colorado River between Hoover Dam and the international boundary near Yuma. The San Bernardino and Douglas basins in southeastern Arizona and parts of several basins in southwestern Arizona that are cut by the international boundary also are included. These basins drain into Mexico but are included in the study because of similarities in physical, climatic, and land-use characteristics to those of basins within the Colorado River drainage.

The area includes approximately equal parts of highly productive alluvial aquifers in the basins and nearly impermeable consolidated rocks in the surrounding mountains that function as boundaries to the aquifer systems. The aquifers are composed mostly of alluvial deposits and occur in the basins formed by down-faulted blocks between mountain masses, which represent the up-faulted blocks. An estimated 900 million acre-ft of recoverable ground water is stored in the upper 1,200 ft of alluvial deposits that fill the basins (Freethey and Anderson, 1986). The volume of ground water in storage, occurrence and rate of movement of ground water, and productivity of wells have a wide range from basin to basin.

Physiography, climate, and geohydrology vary widely throughout the study area. Land-surface altitude ranges from about 150 ft above sea level near Yuma, Arizona, to more than 10,000 ft in the mountainous regions of eastern Arizona and western New Mexico. Average annual precipitation ranges from less than 3 in. to more than 30 in. and generally decreases from the higher eastern basins to the lower western basins. Many streams in the northeastern part of the study area are perennial and typically recharge the alluvial aquifers over which they flow. Only the Colorado River presently contains sufficient flow to reach the dry southwesternmost part of the area. The Gila River was perennial through its entire reach in the early 1900's but has become ephemeral in places because of upstream storage and diversion of surface water and increased losses of streamflow resulting from ground-water development.

GEOHYDROLOGIC SETTING

The basins of the study area were formed as a result of the Basin and Range structural disturbance, which occurred 15 to 10 million years ago (Scarborough and Peirce, 1978). Movement along high-angle normal faults near the present basin edges resulted in the formation of a series of generally north- to northwest-trending basins and mountain ranges. Basin subsidence occurred at varying rates throughout the study area and was accompanied by deposition of locally derived sediment within the internally drained basins. Deposition continued after the establishment of external drainage in most basins. The mountains consist of igneous, metamorphic, and sedimentary rocks. The bedrock of the mountains is virtually impermeable and forms the side and bottom boundaries of the alluvial aquifers.

The basins are filled with alluvial deposits that range from less than 2,000 ft to more than 10,000 ft in thickness (Oppenheimer and Sumner, 1981). These deposits form the major aquifers of the area and store large quantities of ground water. Basin subsidence and sediment deposition occurred at different rates throughout the area, and as a result, the thickness, areal extent, and grain size of the alluvial deposits are highly variable. Sediments range in

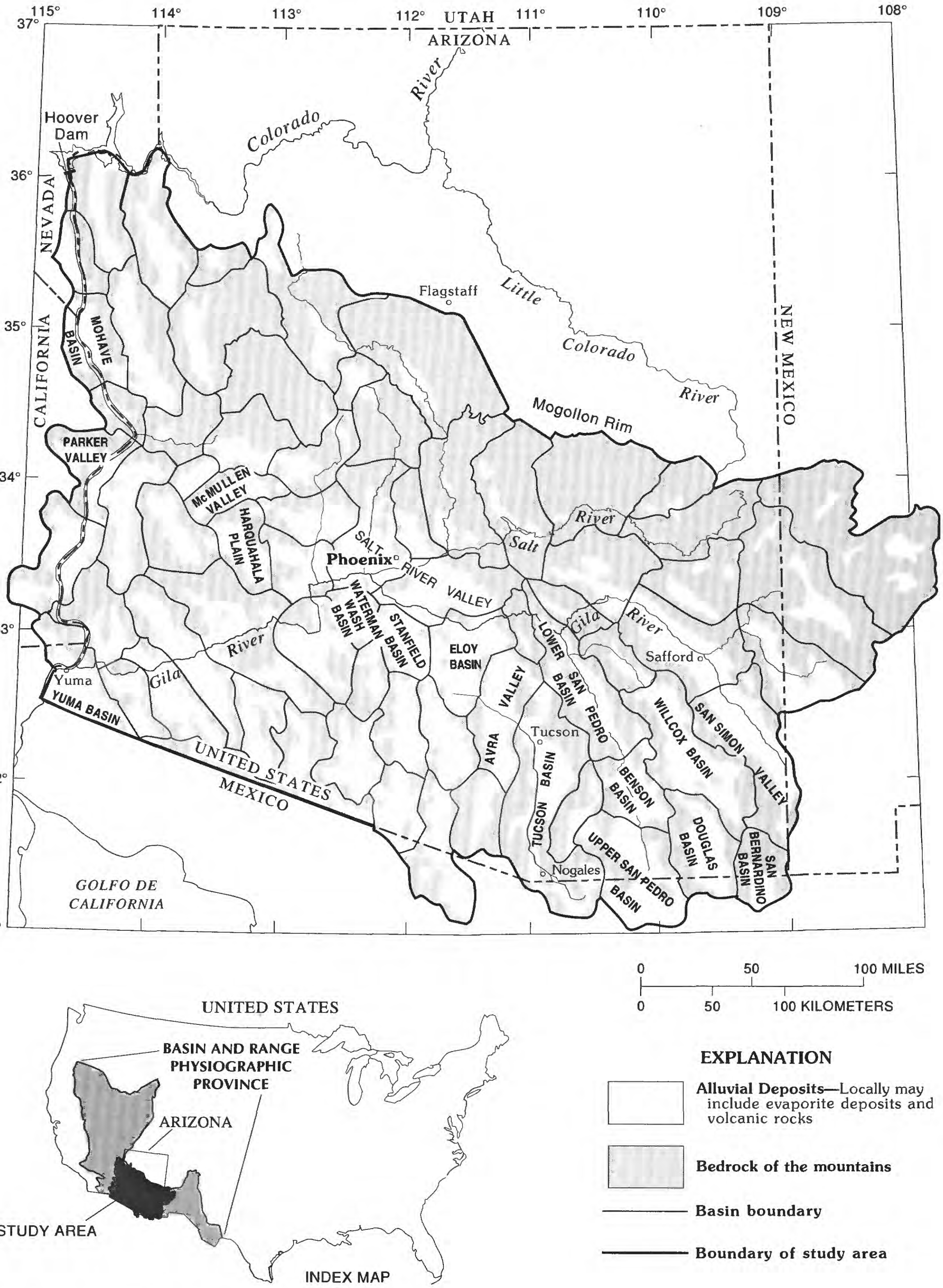


FIGURE 1.—Southwest Alluvial Basins study area.

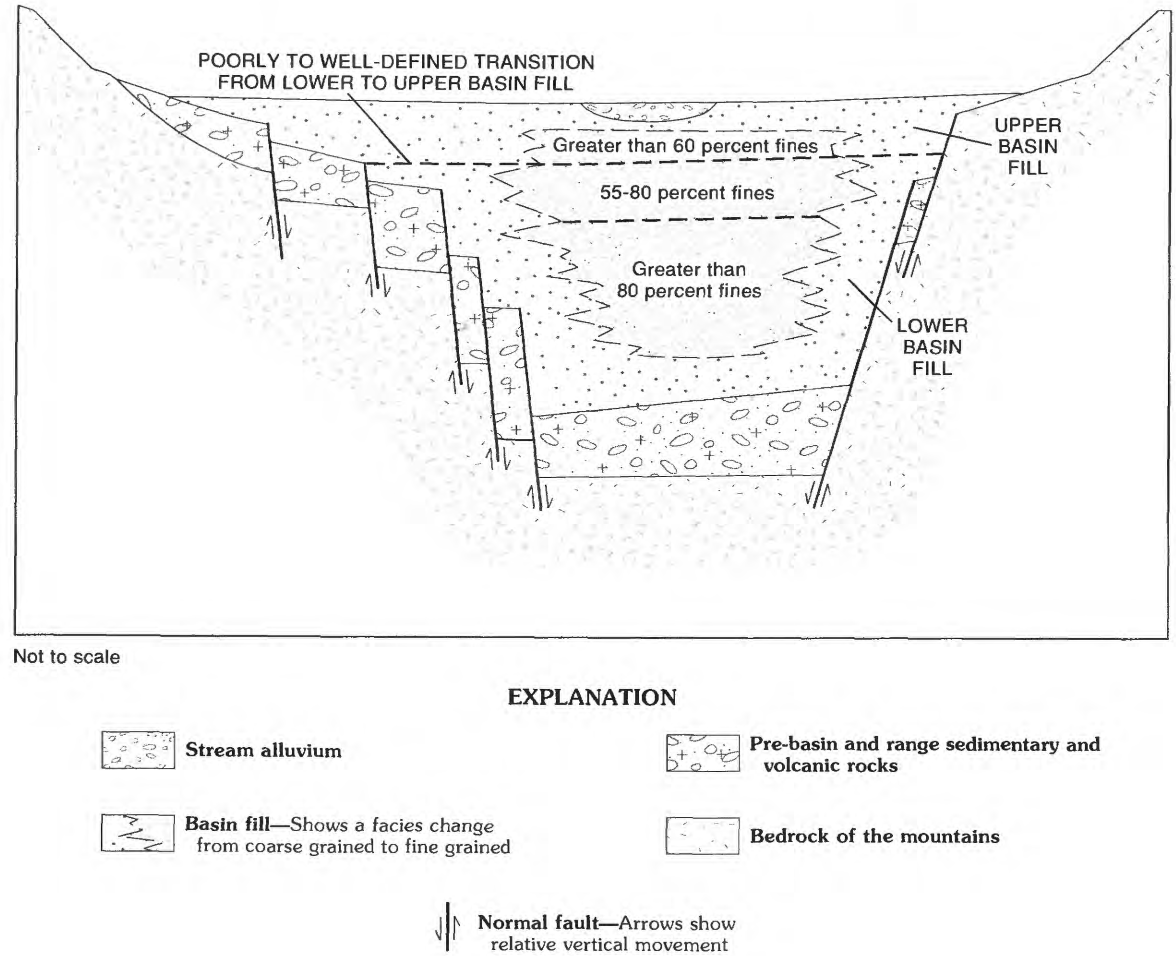


FIGURE 2.—Generalized vertical sequence of sediments that fill most basins.

grain size from clay to gravel and can be unconsolidated to highly consolidated in a single basin.

Basins are interconnected in a dendritic pattern, much like the surface drainage, and form an integrated regional-flow system. Individual basin-aquifer systems serve mainly as ground-water reservoirs with only small areas of interconnection and a small quantity of flow from basins of higher altitude to those of lower altitude.

DESCRIPTION OF SEDIMENTARY UNITS

A similar vertical sequence of sedimentary units is present in most basins and consists of pre-Basin and Range rocks, lower and upper basin fill, and stream alluvium (fig. 2). Each unit represents a different depositional environment, which is reflected in the lithology. The units are distinguishable on the basis of structural relation, degree of consolidation and deformation, source area of clasts, geologic age, and water-bearing characteristics. Although each basin is unique, general spatial patterns of textural composition, thickness, and extent of the units have been documented among groups of basins (Anderson, 1986; Freethey and others, 1986).

Pre-Basin and Range rocks consist of moderately to highly consolidated continental sedimentary deposits that range in grain size from silt to gravel. The rocks were deposited before the Basin and Range structural disturbance and were faulted and tilted at the same time as the underlying bedrock. Basin fill was deposited in the basins formed by the Basin and Range disturbance. Lower basin fill is moderately to highly consolidated, includes extensive fine-grained material near the center of basins, and contains evaporites that range from disseminated gypsum to massive halite deposits. The unit locally contains interbedded volcanic material consisting of basalt flows and tuff. Lower basin fill was deposited in topographically closed basins, and upper basin fill was deposited during a transition period from a closed to an integrated drainage. Upper basin fill is less consolidated, generally is thinner,

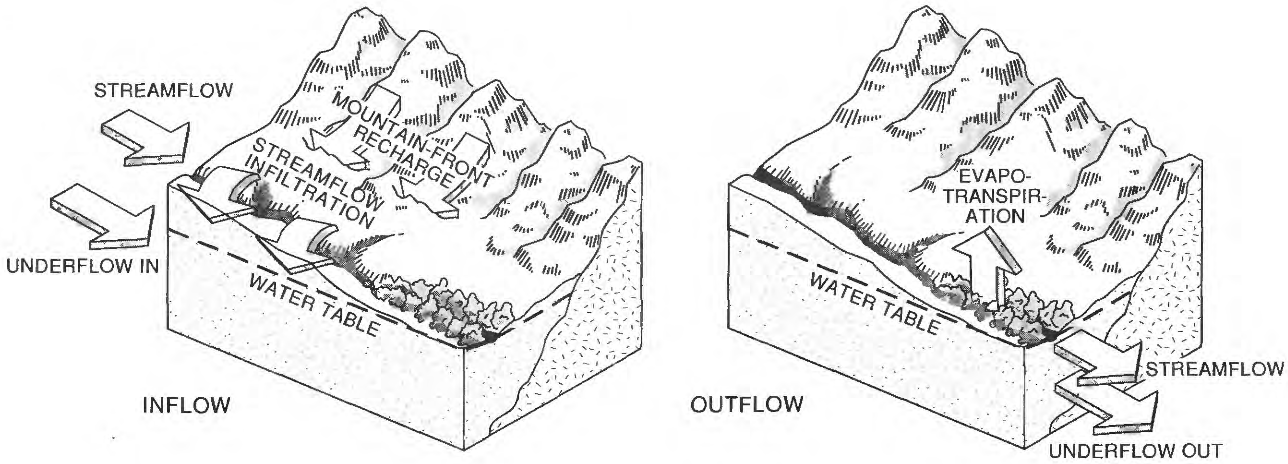


FIGURE 3.—Typical predevelopment inflow and outflow mechanisms in basins of the study area.

and contains fewer fine-grained sediments than the lower basin fill. Stream alluvium was deposited contemporaneously with the establishment of the present-day surface-drainage system. Stream alluvium ranges in texture from clay to boulders and is unconsolidated. Sediments that fill the basins generally are interconnected hydraulically and form a single aquifer system within each basin.

Upper and lower basin fill exhibit gradational patterns in grain-size distribution. The deposits become coarser grained in the direction of the source areas, which are principally the mountains that bound the basins. The change in percentage of fine-grained material from the source area to the basin center can be highly variable from basin to basin within equivalent units (Anderson and others, 1992). The sediment is continental in origin except the marine Bouse Formation in the Colorado River region (Metzger, 1968).

PREDEVELOPMENT HYDROLOGIC CONDITIONS

Ground water generally occurs under unconfined conditions in the aquifers of the study area. In several basins where extensive fine-grained deposits overlie the principal water-bearing unit, ground water occurs under confined conditions. Hydraulic heads in the basin fill range from at or slightly above land surface near perennial streams to more than 600 ft below land surface near the mountain fronts in several basins. Within the study area, an estimated 900 million acre-ft of ground water was stored in the upper 1,200 ft of alluvial deposits before development (Freethey and Anderson, 1986). The quantity of water in storage is hundreds to thousands of times greater than the quantity of water that moves into and out of the aquifers annually.

Ground-water inflow to the aquifers occurs mainly through three mechanisms: (1) infiltration of streamflow along the major streams, (2) infiltration along the mountain fronts, and (3) underflow from adjacent basins (fig. 3). Direct infiltration of precipitation that falls on the basin floor is negligible. The total average annual recharge to all the basins in the study area is estimated to be 2.5 million acre-ft for predevelopment conditions (Anderson and others, 1992). Recharge varies from year to year and from place to place within a basin. Annual recharge probably ranges from near zero in dry years to several times the long-term annual average in abnormally wet years. In general, water enters the aquifer at the upstream end and along the mountain fronts and flows toward the basin center and the outlet at the downstream end. Specific flow characteristics in a basin or for a group of basins depend on the hydrologic properties of the aquifer and on the total annual downvalley flow through the aquifer.

Ground-water outflow from the aquifers occurs through three basic mechanisms: (1) discharge to springs and streams, (2) evaporation and transpiration from the water-table zone, and (3) underflow to adjacent basins (fig. 3). The quantity of natural discharge from each basin probably was nearly constant from year to year before development. The influence of wet and dry years was dampened by the large quantity of ground water stored in the basin. Average inflow to and outflow from the aquifers before development were assumed to be equal at about 2.5 million acre-ft/yr, and no long-term change in storage occurred.

Predevelopment water-level data indicate the general direction of ground-water flow through the basins (Freethey and Anderson, 1986). The data indicate that the basins are linked to various degrees and form an integrated regional-flow system. The shapes of the water-level

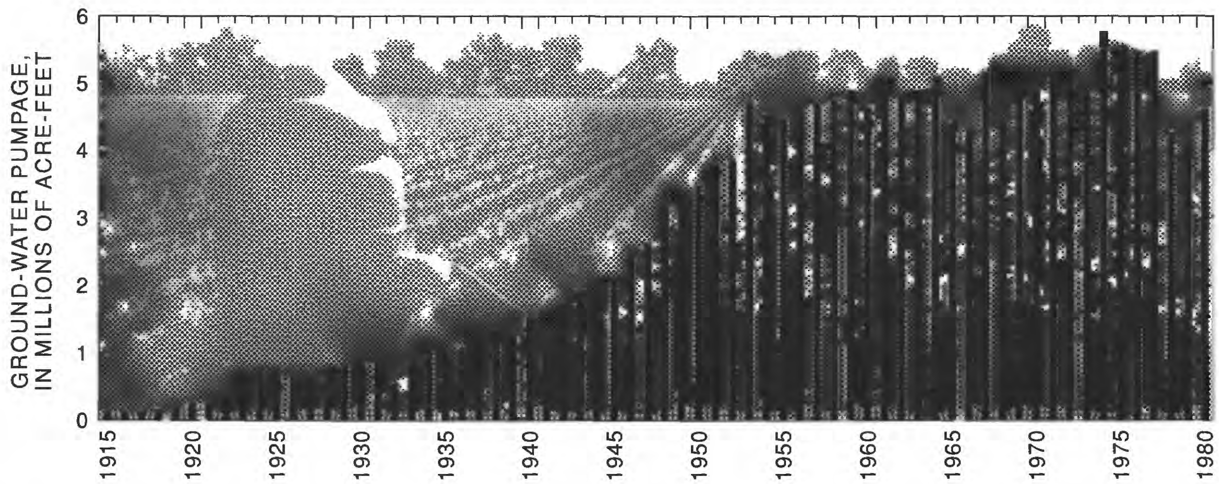


FIGURE 4.—Annual ground-water pumpage, 1915–80.

contours generally indicate areas of inflow and outflow (Anderson and others, 1992). Basins that receive a large quantity of mountain-front recharge have contours that are U-shaped and nearly parallel to the mountain fronts (Anderson and others, 1992, fig. 15*A*). Basins that receive little or no mountain-front recharge have contours that generally are straight and nearly perpendicular to the mountain fronts (Anderson and others, 1992, fig. 15*B*). Both extremes of contour shape, as well as intermediate shapes, exist within the study area and indicate the extreme areal variability in recharge magnitude. Significant variations in water-level gradients also occur in the basins and generally indicate variable physical and hydrologic properties of the aquifer material.

DEVELOPMENT OF WATER RESOURCES

Development of water resources in the study area has been principally for agriculture. Use of the water resources has resulted in depletion of base flow of formerly perennial streams in several basins and overdraft of ground water in most developed basins. Ground-water pumping began in the late 1800's to augment the highly variable surface-water supplies. Development began at different times in various basins. As early as 1920, development was mostly limited to east and west Salt River Valley, Eloy basin, and Stanfield basin (fig. 1). In 1940, nearly 1 million acre-ft of ground water was withdrawn in east and west Salt River Valley; another 0.5 million acre-ft was withdrawn in Eloy, Stanfield, Tucson, and Yuma basins. The other basins were either undeveloped or slightly developed until after 1940. Ground-water pumpage was 1.7 million acre-ft in 1942 and increased to 3.8 million acre-ft in 1952. During 1950–80, an average of 4.8 million acre-ft/yr was pumped (fig. 4), most of which was pumped from 16 basins. From the beginning of development through 1980, about 184 million acre-ft of ground water was pumped from the basins (U.S. Geological Survey, 1982). This pumpage does not represent a total loss from storage because some water recharges the systems through leakage from irrigated land and from canals. Added recharge was induced by increased streamflow infiltration, and some water that previously discharged through evapotranspiration by natural vegetation has been captured. On the basis of the volume of sediments dewatered, about 50 percent of the total pumpage is estimated to be depletion of ground water in storage.

From predevelopment time through 1980, ground-water levels declined from less than 50 ft in areas of minor development to more than 200 ft in the major developed areas (fig. 5). A maximum decline of more than 450 ft occurred in the Stanfield basin between 1940 and 1980. Average annual rates of decline have ranged from a few tenths of a foot per year to more than 10 ft/yr (fig. 6). The magnitude of ground-water depletion varies from basin to basin and reflects the influence of the geohydrologic environment, as well as the magnitude and duration of ground-water withdrawals and stream diversions. In the basins along the Colorado River in which irrigated agriculture is extensive and surface water is a major source of supply, a few feet of water-level rise has resulted from the infiltration of excess applied irrigation water.

BASIN CATEGORIES

The alluvial basins of Arizona can be grouped into five categories on the basis of hydrologic and geologic similarities (Anderson, 1984; Anderson and others, 1992). The most important hydrologic factor used in categorizing the basins is the total downvalley flow, which represents the

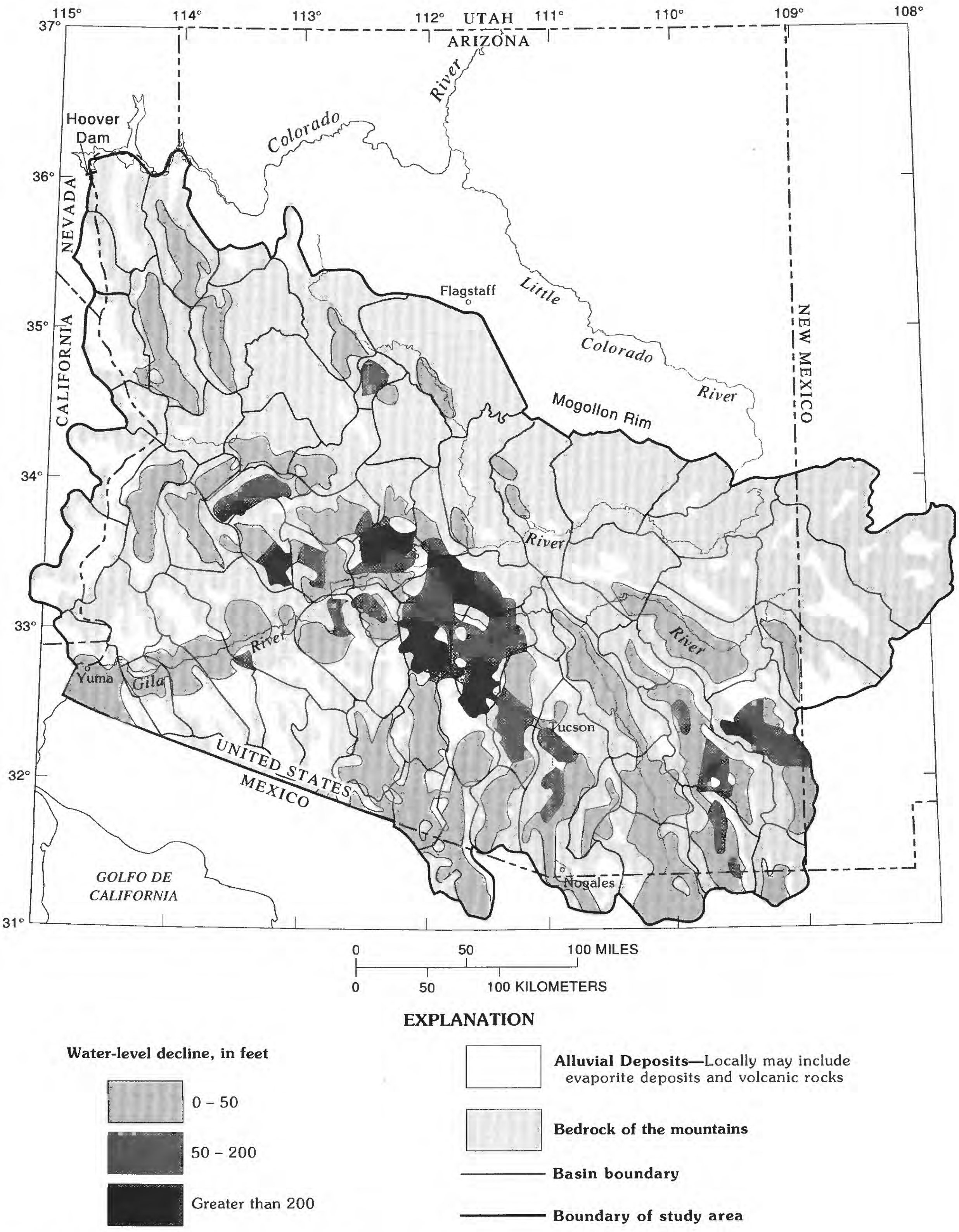


FIGURE 5.—Areas of water-level decline from the beginning of development through about 1980.

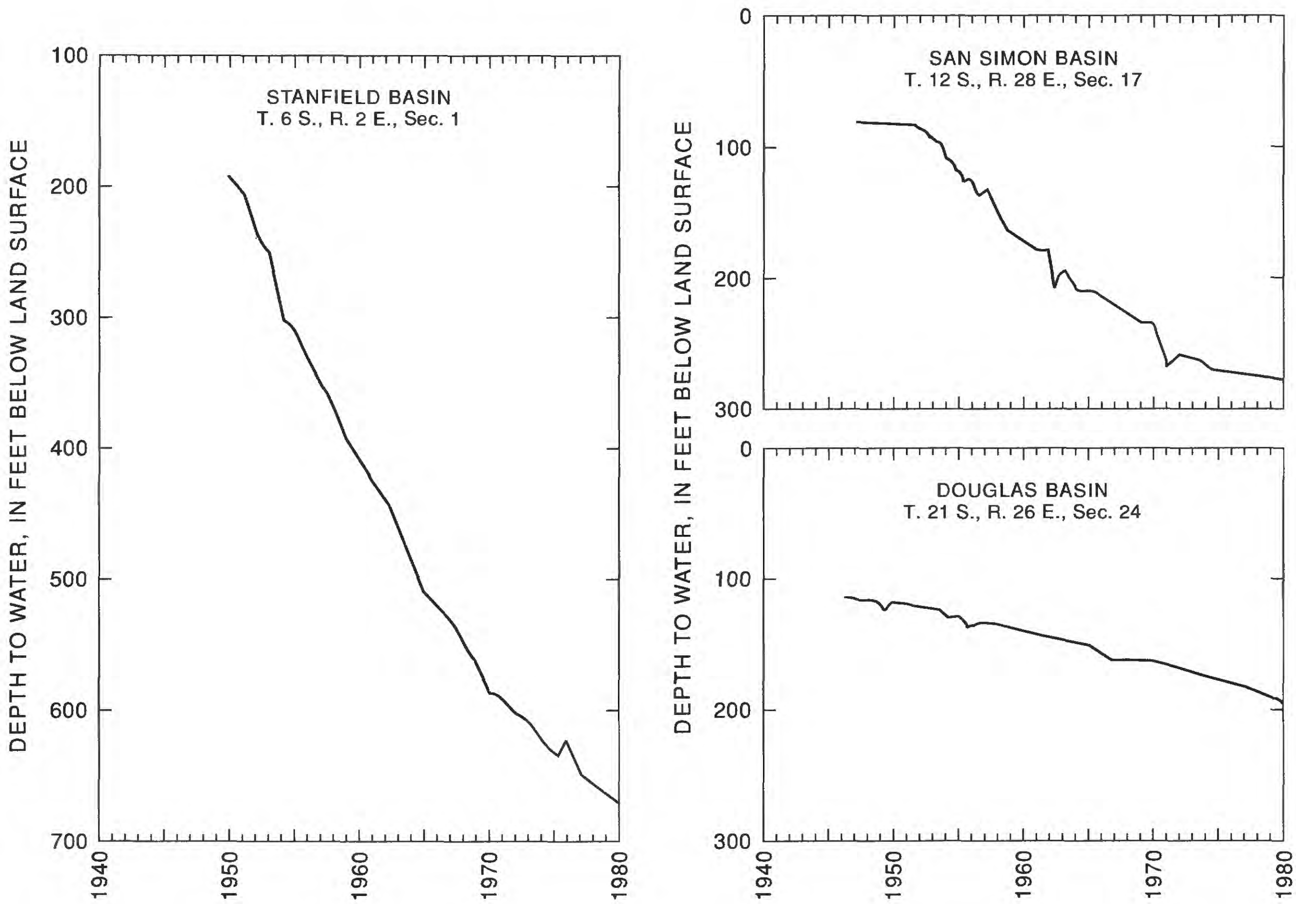
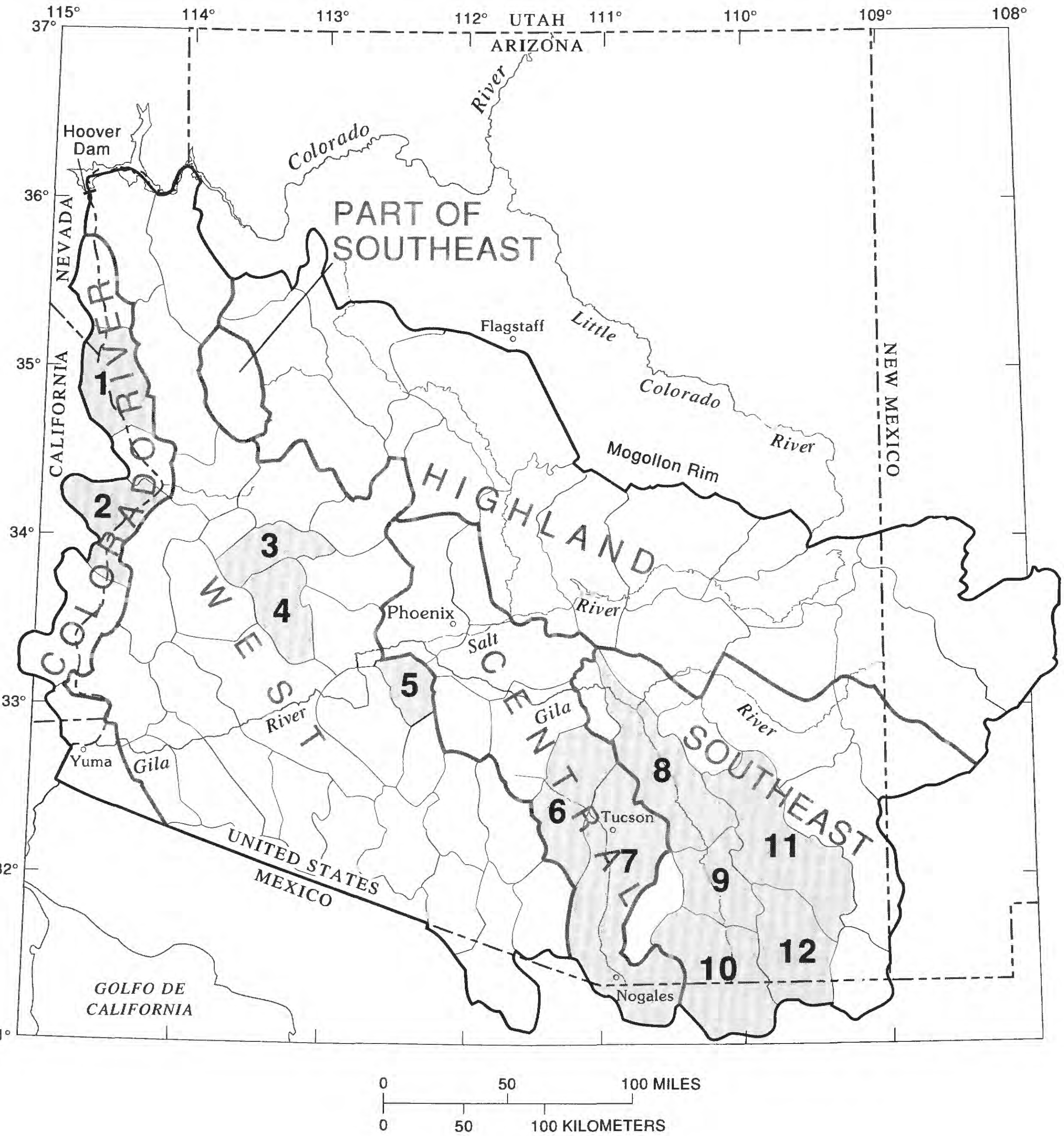


FIGURE 6.—Water-level declines in developed basins.

total renewable water resource available on an annual basis before development. The principal geologic property used in categorizing the basins is the character of the sediments that make up the aquifer system (Pool, 1984). Sediment character controls the movement of ground water. These two factors—downvalley flow and basin-sediment characteristic—control the manner in which the aquifer system will respond to a given stress. On the basis of these two factors, the basins of the study area were grouped into five categories—southeast, central, west, Colorado River, and highland (fig. 7). Generally, basins of the west category have the smallest downvalley flow, and basins of the Colorado River category have the largest downvalley flow. The principal aquifer system in each category generally consists of the following: (1) stream alluvium along the basin axis and adjacent areas in the southeast basins, (2) upper basin fill and stream alluvium in central basins, (3) lower basin fill near the basin center in west basins, (4) stream alluvium in Colorado River basins, and (5) stream alluvium that overlies pre-Cenozoic sedimentary rocks in highland basins. The general geographic grouping of basins of a category reflects the interrelation of hydrology, lithology, and physiographic factors. The response of a ground-water system to the stresses of development can be related to the basin category; therefore, the response of a ground-water system to a proposed development is qualitatively predictable.

GROUND-WATER MODELING

Two types of numerical ground-water flow models were used to analyze the hydrologic systems within the basins. One type included models of selected basins (fig. 7). The goal in developing this type was to simulate the hydrologic system of specific basins and, in doing so, establish reasonable ranges for hydrologic properties and water-budget components in each modeled basin. The other type comprised two basin-group models that simulated basin hydrology for three different geohydrologic environments of the study area. The goal in developing these basin-group models was to analyze how flow models for these three environments were affected by changes in values of individual properties, providing the



EXPLANATION

Modeled basin

1	Mohave Basin	5	Waterman Wash Basin	9	Benson Basin
2	Parker Valley	6	Avra Valley	10	Upper San Pedro Basin
3	McMullen Valley	7	Tucson Basin	11	Willcox Basin
4	Harquahala Plain	8	Lower San Pedro Basin	12	Douglas Basin

Boundary of study area — **Basin boundary** — **Boundary of general categories of basins**

FIGURE 7.—Categories of basins and basins for which ground-water models were developed.

changes were within previously established or reasonable ranges. The basin-group models eliminated site-specific problems commonly encountered in the specific-basin models, and an appraisal of the relative importance of various properties in simulating basin hydrology of different categories could be obtained.

Twelve specific-basin models were developed, and history matched to a predevelopment or another appropriate steady-state condition. Ten of the basin models were evaluated further by using available transient or development conditions. Transient conditions were not simulated in the Parker Valley and Mohave basin models because of the rapid change to a new steady-state condition in response to development in the Colorado River basins. For these two basins, two or more steady-state conditions were simulated in which the magnitude of flow components and head distributions differed from one steady-state condition to another.

Basin-group models used average values of physical and hydrologic properties that were based on actual values in the basins of the particular group being represented. These models were used to evaluate the significance of particular physical and hydrologic properties and flow components on modeling results and to systematically explore model sensitivity to variations in data values used in the models.

MODELING APPROACH

Basins selected for simulation encompassed the ranges in conditions and hydrologic-property values that exist in the study area. All modeling was accomplished by using the three-dimensional finite-difference model described by Trescott (1975) and Trescott and Larson (1976). A quasi-three-dimensional approach was used in which the aquifer system is simulated as two layers. The two layers represent upper and lower segments of the aquifer that are connected hydraulically. To allow vertical flow between simulated layers, this vertical hydraulic connection is represented by a leakance coefficient, which was either calculated by the model or assigned to each cell. Data entry to the models included the geographic extent and thickness of the two layers of the aquifer; water-bearing properties of the layers, such as horizontal and vertical hydraulic conductivity; specific yield and storage coefficient; ground-water recharge and discharge; and pumpage. The upper layer of each modeled basin was simulated by using hydraulic-conductivity values to allow transmissivity to change as water levels changed because of stress imposed on the system. The lower layer was simulated by using transmissivity values, which did not change during transient simulations. The decrease in transmissivity caused by lowering water levels below the top of the lower layer was not adjusted because the magnitude of change was not considered significant, although this was not tested as part of a model analysis. In a few simulations, ground-water withdrawals resulted in sufficient water-level declines that a secondary specific-yield value had to be used for the simulated lower layer rather than the initial storage-coefficient value. Results of the simulations consist of hydraulic head at specified locations, depths, and times; a mass balance that represents the summation of flow into and out of the simulated system; and simulated change in storage.

Each basin model was developed by using available data and known geologic and hydrologic concepts developed by previous investigators. Information transfer also was used in basins where the geohydrologic system was poorly defined. Measured and reported water levels in wells were used to develop potentiometric-surface configurations. Aquifer-test results yielded values for transmissivity and hydraulic conductivity, and analyses of various types of well logs were used to define aquifer lithology, saturated thickness, and aquifer extent in some areas. Flow-net analyses were used to estimate the quantity and distribution of recharge along aquifer boundaries where reliable water-level configurations were available.

For development of models for those basins in which data were sparse, values of hydrologic properties and mountain-front recharge were obtained by transferring values from geohydrologically similar basins and testing these values during model calibration. Empirical methods of estimating transmissivity from specific capacity (Theis and others, 1963) and of estimating transmissivity and specific yield from drillers' logs (Davis and others, 1959; Kisser and Haimson, 1981; Evans and Haimson, 1982) were used to supplement quantitative data where possible. For basins in which water-level data near mountain fronts are unavailable, mountain-front recharge was estimated by using the equation: $\log Q_{rech} = -1.40 + 0.98 \log P$, where P is volume of annual precipitation in excess of 8 in./yr. Use and limitation of the equation are explained in Anderson and others (1992).

Final development of steady-state models was accomplished through a trial-and-error process of adjusting initial values used in the models within an acceptable range. The model was assumed to be an adequate, although not unique, representation of the actual system when the standard error between field-measured and model-calculated water levels was decreased to an acceptably small value and the model-calculated ground-water budget components agreed within an acceptable range with estimates from previous studies. The acceptability of the steady-state model depended on the adequacy of field estimates and areal distribution of water-level data.

Transient models were developed by using the steady-state model and were used to simulate water-level and flow-component changes for various simulated pumping periods. During the history-matching process, values for poorly known properties, such as specific yield, were varied within previously established limits of acceptability; values for the better known properties, such as pumpage, generally were not varied. This process provided a means of obtaining representative transient models and also demonstrated the effect of possible errors in estimating a poorly known property on simulation results.

Examination of model sensitivity to variations in property values used as input data was carried out in specific-basin models and basin-group models for steady-state and transient conditions. Properties were tested for sensitivity by using the simulations developed for basins in the category in which that property was thought to be most significant. A few properties, including transmissivity, specific yield, and recharge, were tested by using simulations developed for all categories of basins. Basin-group models were used mainly to investigate the effect of property generalization on model results.

SELECTED BASIN MODELS

UPPER SAN PEDRO BASIN

Geohydrologic Conditions

Upper San Pedro basin includes 1,650 mi^2, of which about 950 mi^2 is in the United States. The basin is in a north-trending structural trough in southeast Arizona (fig. 7). The sediments that compose the aquifer system include pre-Basin and Range rocks, lower and upper basin fill, and stream alluvium. The pre-Basin and Range rocks occur at depth in most of the basin and are exposed only near the mountain fronts; they are not considered an important part of the aquifer (Freethey, 1982, p. 7). The lower basin fill is the principal water-bearing unit and is hydraulically connected to the upper basin fill, which is saturated only in places. Stream alluvium is limited in areal extent and occurs along the flood plain of the San Pedro River and major tributaries.

Average annual precipitation ranges from 12 in. in the valley to more than 25 in. in the mountains that surround the basin (University of Arizona, 1965; Sellers and Hill, 1974). Recharge to the aquifer occurs as infiltration along the mountain fronts and along the major stream channels. Ground water also enters the area from the south as underflow from Mexico. Ground water generally moves from the basin margin toward the central axis of the basin where most ground water is discharged along gaining reaches of the stream or lost to evapotranspiration by riparian vegetation. A small quantity of water leaves the basin as underflow. Estimates for inflow and outflow are given in figure 8.

The hydrologic system of the upper San Pedro basin is typical of basins of the southeast group. The upper San Pedro basin was one of the initial basins used for ground-water simulation. The model was used to evaluate sensitivity to generalizations of hydrologic properties. Knowledge about model sensitivity gained in this effort was transferred to models of other basins in the southeast group. Inflow to the ground-water system is principally through mountain-front recharge; however, a significant quantity of water is recharged through infiltration of surface flow. Under predevelopment steady-state conditions, outflow occurred through evapotranspiration and discharge to surface flow of the San Pedro River. Basin fill and stream alluvium are stratigraphically complex. Water levels in wells drilled in the basin-fill sediments indicate that ground water in the aquifer is confined in places; however, ground water generally occurs under unconfined conditions. Only the upper more permeable part of the aquifer has been drilled and is used for ground-water withdrawal (Freethey, 1982).

Ground-water resources were developed initially for agricultural purposes. The first substantial pumping occurred in the 1940's, and in 1977 about 13,200 acre-ft of water was withdrawn (fig. 8). About 70 percent of the pumpage was for agricultural use; 14 percent was for industrial use; and 16 percent was for public-supply, domestic, and livestock uses (Konieczki, 1980). As a result of increased ground-water withdrawal, surface flow has decreased, underflow entering the area at the international boundary has increased, and discharge through evapotranspiration has decreased (Freethey, 1982).

Model Characteristics

The conceptual model of the hydrologic system of the upper San Pedro basin was based on available data and interpretations from previous studies (Bryan and others, 1934; Heindl, 1952b; Brown and others, 1966; Roeske and Werrell, 1973). The numerical model was developed to simulate basinwide ground-water flow and to test the conceptual model. The numerical model developed and documented by Freethey (1982) used two layers to simulate a single aquifer system. The upper layer represents the upper 1,000 ft of the aquifer for which field data are available; the lower layer represents the aquifer deeper than 1,000 ft for which few data are available. Hydrologic properties representing the upper and lower model layers are given in figure 8. Quantities of recharge and values of hydrologic properties were adjusted within acceptable limits during the modeling process to improve the comparisons between field data and model results (Freethey, 1982, p. 49).

The following model characteristics were adopted to simulate the hydrology of the upper San Pedro basin (Freethey, 1982, p. 15–16).

- A finite-difference grid with variable cell size was used to provide high resolution in areas where data density was high or where large variations in aquifer properties or stresses occurred.
- Two layers represented (1) the basin fill less than 1,000 ft below land surface for which data were available and (2) the basin fill deeper than 1,000 ft below land surface for which few data were available. This arbitrary separation is the approximate maximum depth to which wells were drilled. Although the upper 1,000 ft of basin fill is of primary interest, the possibility of ground-water movement to or from the lower layer cannot be ignored. Hydrologic properties for this lower layer are unknown; however, values were selected in an assumed reasonable range to reflect what is known about the geology of the lower layer and to simulate the structural shape of the basin.
- The upper layer was simulated as an unconfined aquifer; the lower layer was simulated as a confined aquifer.
- Vertical connection between layers was calculated by the model from the assigned hydrologic properties of each layer. The connection is represented quantitatively by the vertical leakance, which is the vertical hydraulic conductivity (K') of the fine-grained sediment divided by its thickness (b).
- Most recharge simulated by the model occurred in the upper layer. Specified heads in the lower layer allowed a small quantity of recharge directly into the lower layer.
- Interaction takes place between perennial streams and the upper layer. Stream leakance, which represents the hydraulic property that allows or restricts vertical movement of water between the stream and the aquifer, is constant.
- Evapotranspiration discharges water from the upper layer and was simulated by a linear relation between a maximum evapotranspiration rate and a depth to water of 10 ft where evapotranspiration was assumed to cease.

The upper San Pedro basin model was used to evaluate the effect of using uniform representative values for hydrologic properties on model results, hereinafter referred to as generalizing. Freethey (1982, p. 16) lists the alternatives that were explored to evaluate model sensitivity to values and areal distributions of boundary recharge, aquifer hydraulic conductivity, and aquifer storage.

- Boundary recharge was evaluated by using two alternative modes: (1) distributed uniformly along mountain fronts to represent a situation of minimum data availability and (2) distributed on the basis of site-specific data and a flow-net analysis.
- Aquifer hydraulic conductivity was analyzed by comparing model results using (1) uniform values in three geohydrologically similar subareas of the basin and (2) an areal distribution on the basis of meager aquifer-test and specific-capacity data and a flow-net analysis.
- Aquifer storage was examined by comparing three model simulations using (1) a variable distribution of specific yield estimated from drillers' logs, (2) a uniform value for specific yield based on few data in the basin, and (3) the uniform specific-yield value used in the previous simulation and uniform values for boundary recharge and aquifer hydraulic conductivity.
- Generalized boundary recharge, hydraulic conductivity, and specific yield were used together to represent a crude approximation of the hydrologic system that might be developed where few data are available. The results were compared with the final calibrated model that used all the available information.

Results of Simulations

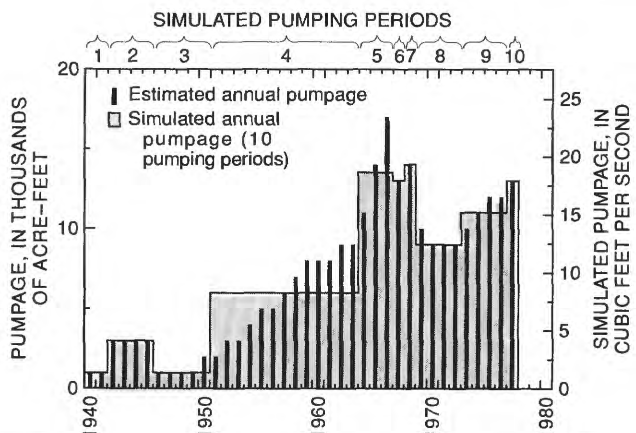
After steady-state model history matching, several model runs were made by using an area-weighted arithmetic average value to replace a single data array used in the model. Values for mountain-front recharge, hydraulic conductivity, and specific yield were independently varied in this manner to determine the effects on model results. These results were then compared with results of the steady-state simulation that most accurately matched measured water levels. Evapotranspiration rate and extinction depth, stream leakance, and ratios for vertical anisotropy of hydraulic conductivity remained unchanged during this process.

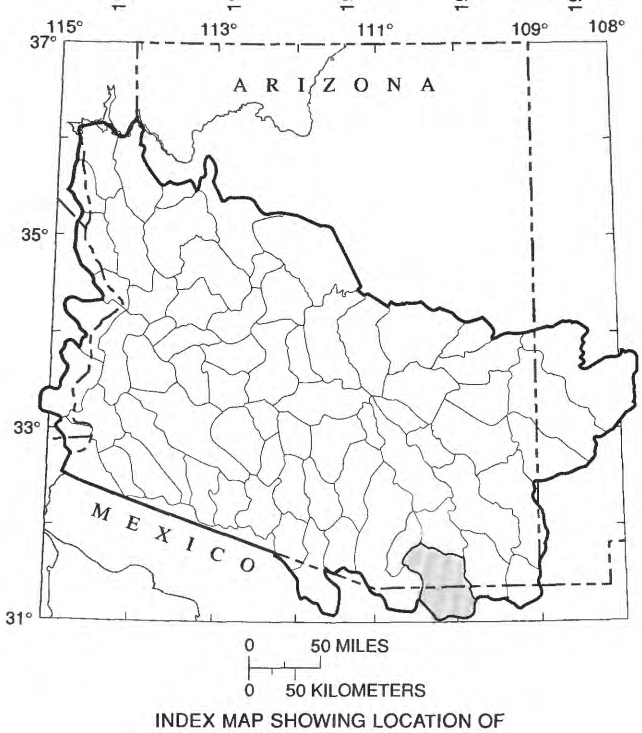
Initially, mountain-front recharge was distributed around the model perimeter on the basis of a flow-net analysis, and recharge varied from cell to cell (nongeneralized). Subsequently, a uniform average value was specified at each recharge cell (generalized). Generalizing mountain-front recharge resulted in simulated water levels that were more than 25 ft lower than water levels in the history-matched steady-state model over about 17 percent of the modeled area. The areas of greatest difference coincided with those areas of steep hydraulic gradient near the mountain fronts. The greatest water-level differences correspond with the greatest differences between the quantities of nongeneralized and generalized recharge. Water levels that are higher than those of the history-matched steady-state simulation are caused by the generalized recharge quantity that is greater than the nongeneralized recharge quantity. Water-level increases typically were less than 20 ft; however, the area influenced is

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS

[Values are in acre-feet per year; dashes indicate no data. Estimated values modified from Freethey (1982, table 2)]

Components	Steady state		Transient state: 1968		Transient state: 1977	
	Estimated	Simulated	Estimated	Simulated	Estimated	Simulated
Inflow:						
Mountain-front recharge	10,000 - 11,000	8,300	10,000 - 11,000	8,300	10,000 - 11,000	8,300
River losses	800 - 4,300	800	------	1,200	------	1,400
Underflow	700 - 3,500	7,400	------	7,500	------	7,700
Storage	0	0	------	6,000	------	5,600
Total	11,500 - 18,800	16,500	10,000 - 11,000	23,000	10,000 - 11,000	23,000
Outflow:						
River gains	1,900 - 14,300	8,300	------	5,400	------	5,900
Underflow to Benson basin	0	400	------	300	------	300
Evapotranspiration	3,600 - 12,300	7,800	------	6,500	------	6,200
Pumpage	0	0	13,500	10,800	13,200	10,600
Total	5,500 - 26,600	16,500	13,500	23,000	13,200	23,000





0 50 MILES

0 50 KILOMETERS

INDEX MAP SHOWING LOCATION OF UPPER SAN PEDRO BASIN (SHADED)

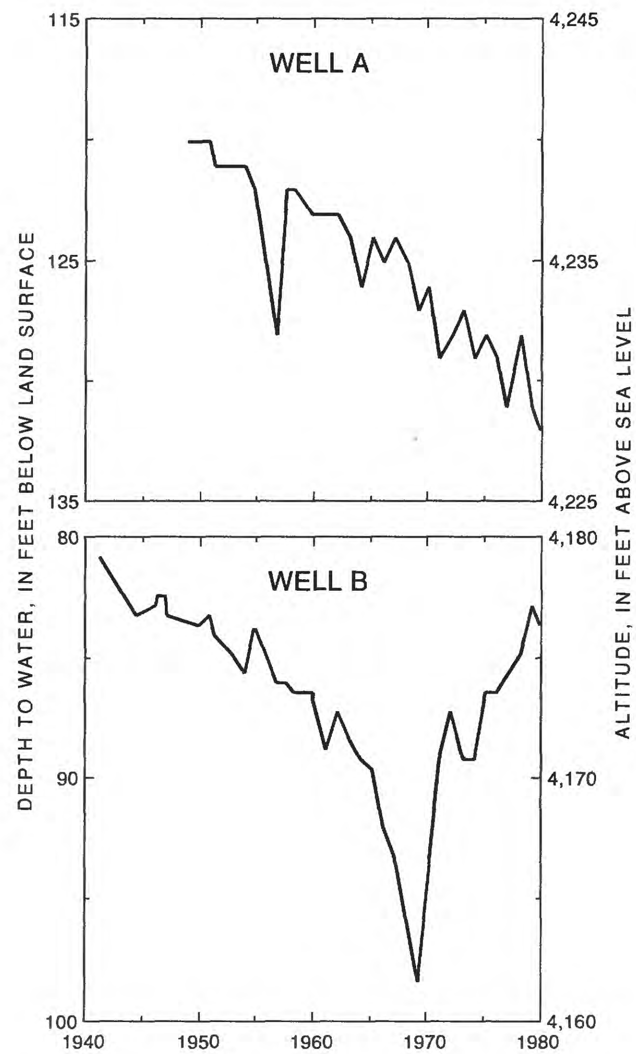
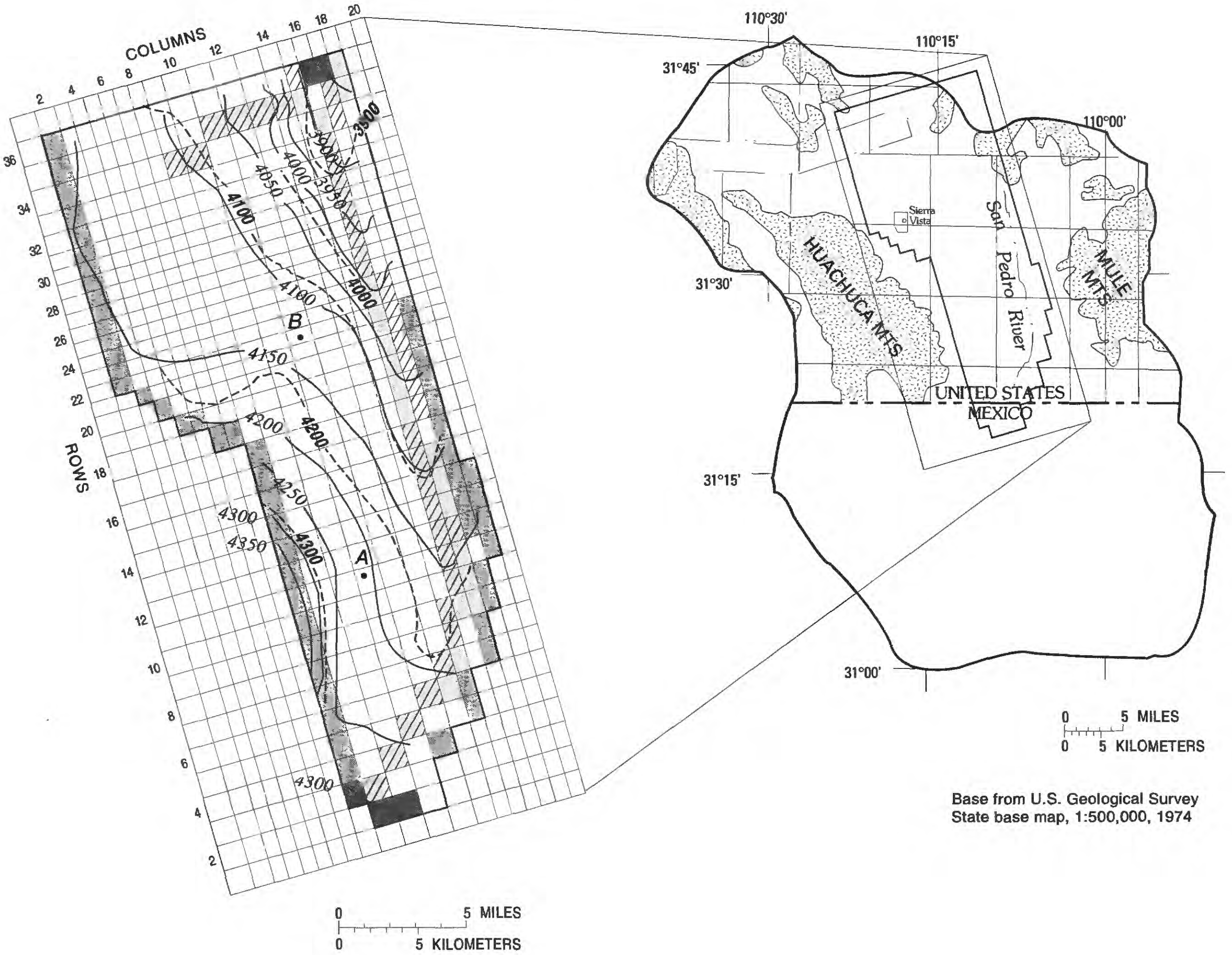


FIGURE 8.—Location of basin, estimated and simulated water-budget components and pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, upper San Pedro basin.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

Constant head

Evapotranspiration

Mountain-front recharge

River-aquifer connection

Geologic contact

Boundary of ground-water area

Boundary of modeled area—Lower and upper layers

Water-level contour, 1977–78—Shows altitude of water level. Contour interval 100 feet. Datum is sea level

-- *4200* --- Observed (Konieczki, 1980)

—*4250*— Simulated (Freethey, 1982)

• *B* **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	0.1 - 18.3
Average	4.1
Transmissivity of lower layer, in feet squared per day	
Range	9 - 2,307
Average	684
Specific yield of upper layer	0.08
Storage coefficient of lower layer	10^{-5}
Vertical leakance between layers, in feet per second per foot	
Range	$5x10^{-7}$ - $2x10^{-9}$

FIGURE 8.—Continued.

limited to a narrow zone along the basin edge where recharge occurs, gradients are large, and close matching of field- and model-head values are difficult to attain. These results indicate that generalizing recharge along a basin boundary does not greatly affect overall modeling results but can cause nonrepresentative water levels to be simulated in site-specific areas near the mountain fronts.

Specifying the same value of hydraulic conductivity throughout the basin resulted in water levels that were more than 25 ft lower than simulated water levels in about 53 percent of the modeled area. The model has a greater response to the generalization of hydraulic conductivity than it does to generalization of recharge quantity because the generalized hydraulic-conductivity value represents the arithmetic average of a data set that has a large range in values. The average value in the model can differ markedly from the field value, and large head differences will result, particularly in the basin center where field values of hydraulic conductivity are low. A conclusion of this analysis is that use of generalized hydraulic-conductivity values in a basin model is transferable but is not considered a desirable approach. The effectiveness in simulating an actual hydrologic system by using a generalized hydraulic-conductivity value depends on the variability of the hydraulic conductivity and the accuracy of the field information on which the generalized value is based.

A transient analysis of the model was used to evaluate model sensitivity to a uniform specific-yield value. A uniform specific-yield value (0.08) resulted in from 4 ft less to 1 ft more drawdown in the final pumping period of the simulation and in virtually the same volume of storage depletion in the aquifer. Generalizing recharge, hydraulic conductivity, and specific yield resulted in drawdown in the final pumping period that ranged from 8 ft less to 7 ft more than that in the simulation using nongeneralized values. The pumpage that represented a removal of water from storage increased from 53 to 58 percent; other flow components remained within previously estimated ranges.

The numerical model developed for the upper San Pedro basin simulated the hydrologic system to a degree of accuracy acceptable for most purposes related to water-resources appraisals. The accuracy, or how closely the simulation represents the actual hydrologic system, is a function of the dynamics and variability of local hydrologic conditions and the magnitude of pumpage relative to other water-budget components. By use of current knowledge and definition of the system, the numerical model can simulate the key hydrologic processes presumed to be taking place. The model can be used to simulate areal water levels that approximate the field data and to simulate ground-water-budget values that approximate values estimated in previous investigations (fig. 8). These results demonstrate that the model is a reasonable, though not unique, simulation of the real system. Predictive capabilities are limited to a general assessment of changes in inflow, outflow, and storage values and changes in regional-flow directions as a result of pumping. The purpose of the model was not to analyze site-specific ground-water conditions or to predict water-level changes in individual wells. Reliability of the predictive capability of the model also depends on future changes in the hydrologic system and how well the changes can be incorporated into numerical representations.

BENSON BASIN

GEOHYDROLOGIC CONDITIONS

Benson basin includes about 900 mi^2 in southeastern Arizona and is in the same north-trending structural trough as the upper San Pedro basin (fig. 9). Benson basin is downstream from the upper San Pedro basin; outflow from the upper San Pedro basin is inflow to Benson basin. About half the surface area is bedrock of the mountains that forms the east and west sides of the basin. The basin is about 32 mi long, and sedimentary deposits that make up the valley floor average 15 mi wide. The aquifer system is composed of the same units as in the upper San Pedro basin, and basin fill and stream alluvium are stratigraphically complex. In the Benson basin, however, the lower basin fill contains a gypsiferous, fine-grained facies as much as 800 ft thick near the basin center. This facies forms the confining unit for ground water in underlying lenses and layers of coarse-grained material. Water from wells that penetrate the coarse-grained material may flow at the land surface. On the perimeter of the fine-grained facies and in the overlying sediments, ground water occurs under unconfined conditions. Vertical leakage upward probably occurs through the confining unit and is a source of inflow to the upper, unconfined part of the aquifer.

Average annual precipitation ranges from less than 12 in. in the valley to more than 25 in. in the mountains that bound the basin (University of Arizona, 1965). Recharge to the ground-water system takes place at the mountain fronts and along the San Pedro River; ground-water movement is from the mountain fronts toward the central axis of the basin. Most discharge occurs as evapotranspiration in the zone of riparian vegetation along the San Pedro River. A small quantity of discharge occurs as underflow at the north end of the basin. Estimated and simulated inflow and outflow are given in figure 9.

Ground-water development began in the early 1900's, and pumpage is estimated to have been less than 3,000 acre-ft in 1940. In 1967, ground-water withdrawal was 12,400 acre-ft; withdrawal increased to nearly 40,000 acre-ft in 1976 and was 35,000 acre-ft in 1977 (fig. 9).

Pumping of ground water in Benson basin resulted in a decrease in the number of flowing artesian wells and a decrease in the discharge from wells that continue to flow. Reaches of the San Pedro River that were perennial now flow only intermittently.

MODEL CHARACTERISTICS

A conceptual model of the ground-water system of Benson basin was formulated on the basis of results of previous studies (Bryan and others, 1934; Roeske and Werrell, 1973; Konieczki, 1980) and through analysis of available geologic and hydrologic data. Pertinent data for Benson basin are sparse, and the approach and many of the data used in the model were transferred from the results and modeling approach used in the upper San Pedro basin (Freethey, 1982).

The numerical model of the Benson basin had the following characteristics:

- A variable grid size for higher resolution along the central axis of the basin (fig. 9).
- Two layers representing (1) the stream alluvium that extends along the axis of the basin but does not extend to the mountain boundaries and in which ground water is unconfined and (2) the lower layer that extends throughout the basin and in which ground water is confined in an area that underlies the basin-center fine-grained facies and unconfined outside this area extending to the mountain fronts.
- A simulated "quasi" third layer connects the two aquifers and simulates vertical flow through the fine-grained unit. This layer is represented quantitatively by a vertical-leakance value, K'/b, which is the vertical hydraulic conductivity divided by the thickness of the fine-grained unit.
- Recharge along the mountain fronts is to the lower layer, which is the only active layer adjacent to the mountains.
- Interaction takes place between the San Pedro River and the stream alluvium. Stream leakance, which represents the hydraulic property that allows or restricts vertical movement of water between the stream and the aquifer, is constant.
- Ground water discharges by evapotranspiration from the upper layer.
- Pumping was simulated from both layers. Irrigation pumpage from the upper layer was reduced by 30 percent to account for return of excess applied irrigation. Thirty percent of the irrigation pumpage from the lower layer was recharged to the upper layer.

The finite-difference grid for the Benson basin model consisted of 19 columns and 30 rows making 570 rectangular cells in each layer that were from 0.36 to 1.0 mi^2 in area (fig. 9). The upper layer was represented by 91 active cells, and the lower layer had 367 active cells. Mountain-front recharge was applied to the lower layer at a uniform rate per mile of mountain front. This rate differed for each mountain range on the basis of altitude of the mountains, which is related to the volume of annual precipitation in the mountains. Underflow into the basin at Fairbank and out of the basin at the Narrows was simulated by constant-head cells in both model layers. Upward leakage from lower to upper layers was controlled by the vertical-leakance value.

RESULTS OF SIMULATIONS

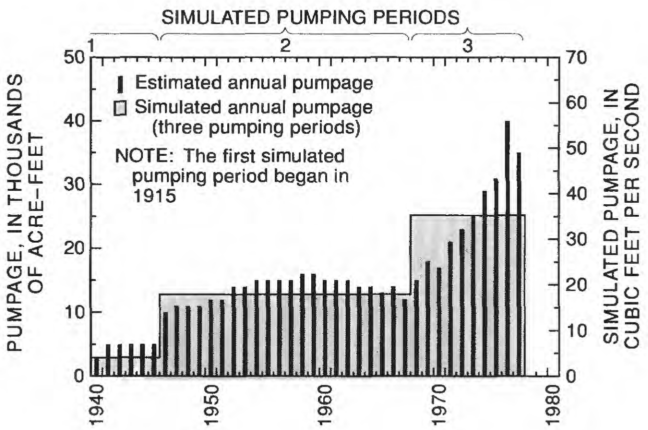
To evaluate the numerical-model simulation, steady-state and transient model runs were made. The steady-state model was used to determine if the water-budget components and hydrologic properties of the aquifer that were estimated on the basis of existing information functioned together as a reasonable representation of the hydrologic system. The transient model was used to evaluate the reliability and uniqueness of the steady-state model. The steady-state model for Benson basin verified that all the proposed hydrologic concepts were compatible in the simulation and that values of simulated flow components approximated the values of flow components of the conceptual model (fig. 9). Outflow from the upper San Pedro basin was used as inflow to the Benson basin. Simulated water levels for the predevelopment period were the same general configuration as measured water levels. The results of simulated transient conditions indicated that evapotranspiration, stream-aquifer exchange, and upward leakage from the lower layer decreased as a result of pumping and that as much as 10,000 acre-ft of water was removed from storage in the upper aquifer through 1977. On the basis of the approximate areal extent, saturated thickness, and specific yield of this unconfined aquifer, 10,000 acre-ft represents only about 1 to 2 percent of the available ground water in storage. Water-level declines are greater than 20 ft only locally near pumping wells.

The transient model simulated historical pumping conditions for three time periods—from predevelopment (arbitrarily selected as 1915) through 1945, 1946 through 1967, and 1968 through 1977 (fig. 9). Water-level declines have occurred primarily in the area underlying the flood plain of the San Pedro River where most pumping has occurred. Simulated head declines at the end of 1945 typically were less than 5 ft in the unconfined aquifer and less than 10 ft in the confined aquifer. Field evidence to corroborate these head declines consists only of early observations about the decreasing discharge of flowing artesian wells (Roeske and Werrell, 1973). Simulated head

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS

[Values are in acre-feet per year. Estimated values modified from Heindl (1952b) and unpublished records of the U.S. Geological Survey]

Components	Steady state Estimated	Steady state Simulated
Inflow:		
Mountain-front recharge	8,100 - 10,500	9,400
River losses	200 - 9,700	1,400
Underflow from upper basin	400	600
Total	8,700 - 20,600	11,400
Outflow:		
River gains	900 - 5,900	4,000
Underflow to lower basin	100 - 200	300
Evapotranspiration	5,100 - 16,700	7,100
Total	6,100 - 22,800	11,400



INDEX MAP SHOWING LOCATION OF BENSON BASIN (SHADED)

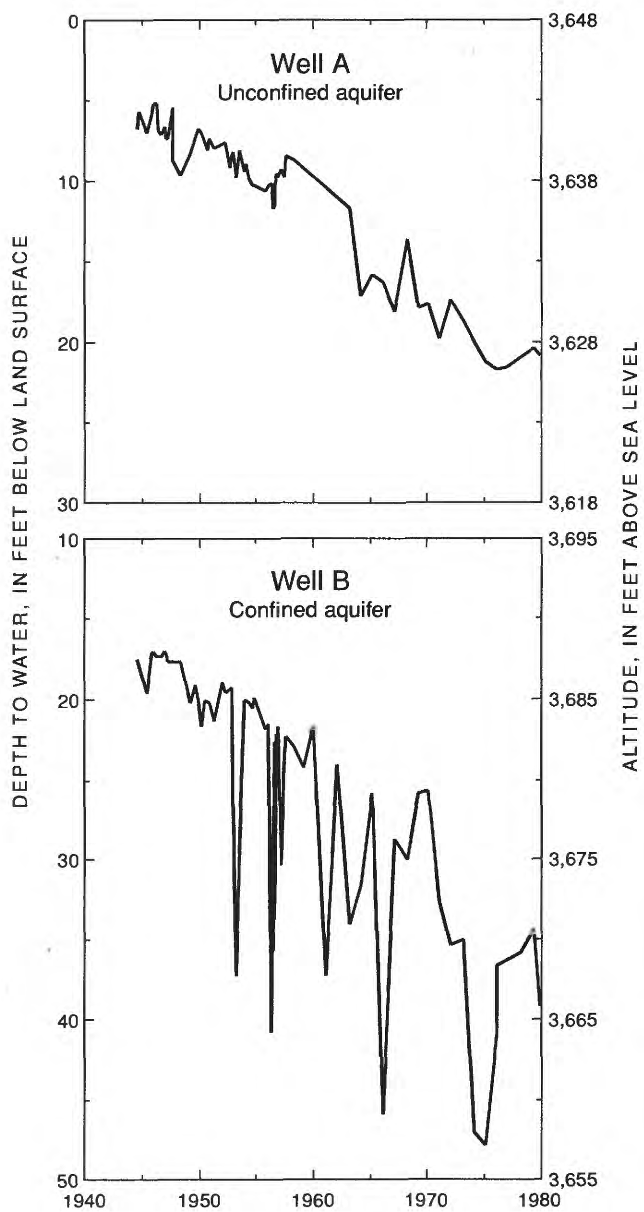
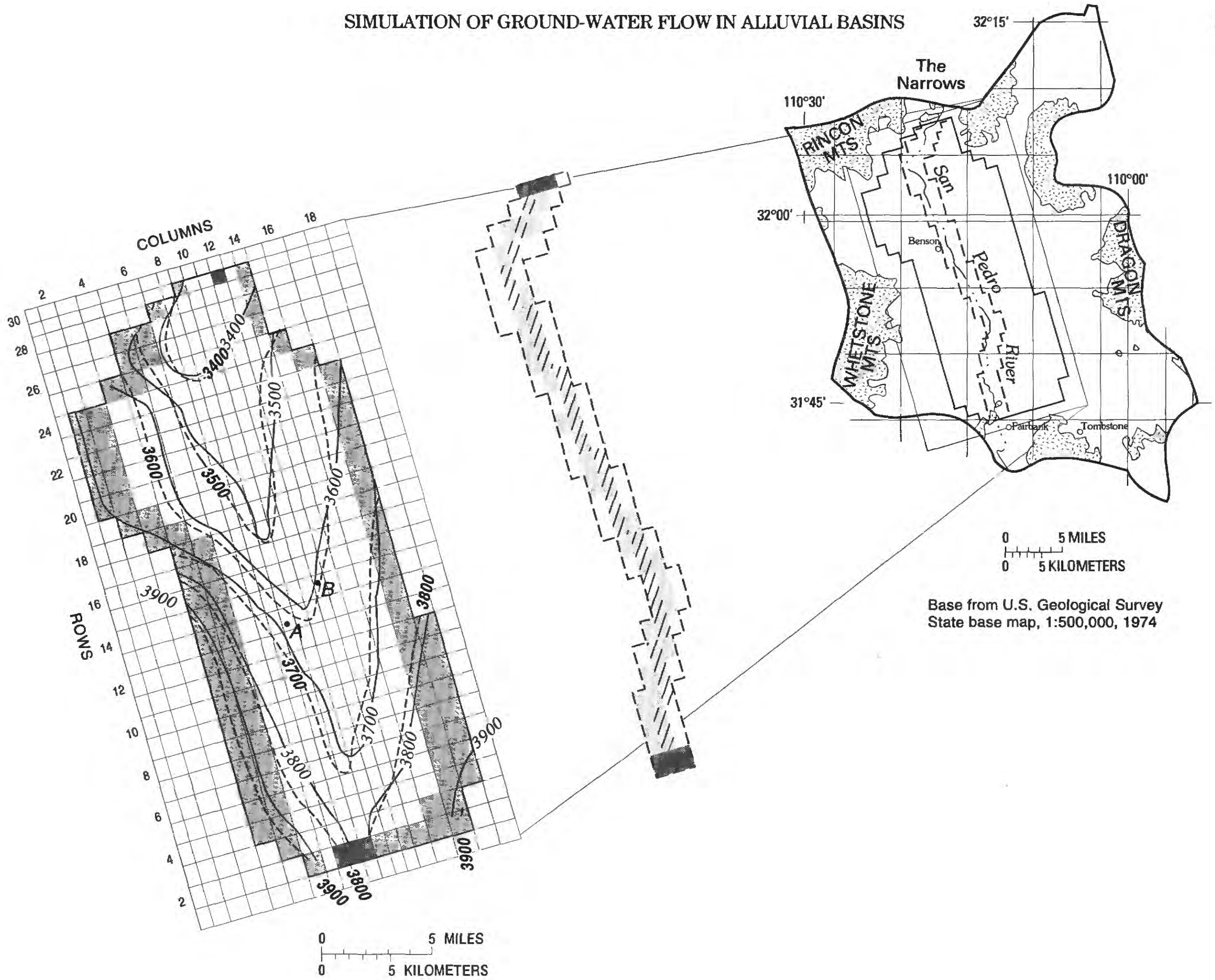


FIGURE 9.—Location of basin, estimated and simulated water-budget components and pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, Benson basin.



EXPLANATION

- Alluvial deposits
- Bedrock of the mountains
- Grid showing model cells used in simulating ground-water flow
- Cell representing
  - Constant head
  - Evapotranspiration
  - Mountain-front recharge
  - River-aquifer connection
- Geologic contact
- Boundary of ground-water area
- Boundary of modeled area—
  - Lower layer
  - Upper layer
- Water-level contour, 1977–78—Shows altitude of water level. Contour interval 100 feet. Datum is sea level
  - -- *3500* -·- Observed (Konieczki, 1980)
  - —*3500*— Simulated
- •*B* Well for which hydrograph is shown

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	2 - 45
Average	17.1
Transmissivity of lower layer, in feet squared per day	
Range	11 - 4,445
Average	832
Specific yield	
Lower layer	0.114
Upper layer	0.125
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-7}- 10^{-11}

FIGURE 9.—Continued.

decline at the end of 1967 typically ranged from 5 to 15 ft in the unconfined aquifer (hydrograph for well A, fig. 9) and from 10 to 25 ft in the confined aquifer (hydrograph for well B, fig. 9). At the end of 1977, head declines in most of the area ranged from 10 to 40 ft in the unconfined aquifer and 15 to 50 ft in the confined aquifer. A local area near the center of the modeled area had simulated head declines of 50 to 100 ft for both aquifers. This head decline was not documented by field data. The discrepancy could be caused entirely or in part by the method used to distribute simulated pumping, by no-flow boundaries in the upper layer of the model, by incorrect hydrologic properties assigned to the area, or by a combination of these factors.

LOWER SAN PEDRO BASIN

Geohydrologic Conditions

Lower San Pedro basin is the northernmost extension of the narrow north- to northwest-trending structural trough that forms the upper San Pedro and Benson basins. The lower San Pedro basin includes about 1,700 mi^2 in southeastern Arizona (fig. 10). About 40 percent of the surface area is basin fill, and 60 percent is bedrock of the mountains. The alluvial basin is about 75 mi long and averages less than 10 mi wide. Aquifer lithology in the lower San Pedro basin is similar to that of the upper San Pedro and Benson basins. The ground-water system consists of an upper unconfined aquifer and a lower confined aquifer separated by an intermediate fine-grained unit. The upper aquifer consists mainly of stream alluvium underlying the flood plain of the San Pedro River and is of limited areal extent. The lower aquifer consists of thin sand layers underlying a fine-grained unit.

Average annual precipitation ranges from less than 12 in. on the valley floor to more than 30 in. in the mountains that bound the basin (University of Arizona, 1965). Ground water moves from areas of recharge along the mountain fronts and along stream channels to areas of discharge along the central axis of the basin. Discharge as underflow from the upstream Benson basin enters the lower San Pedro basin at the south end. Some ground water moves vertically from the deep confined aquifer to the upper unconfined aquifer as leakage through the fine-grained unit. Most discharge occurs as evapotranspiration in the zone of riparian vegetation in the flood plain of the San Pedro River or as base flow to the river. Heindl (1952a) estimated that 35,000 acre-ft of water discharged from the basin through evapotranspiration annually.

Ground water is the primary source of water used in the lower San Pedro basin. Many ditches diverted streamflow to irrigate small acreages prior to 1940 and probably still do when streamflow is available. After World War II, however, ground-water withdrawal increased markedly to about 20,000 acre-ft in 1951 (Heindl, 1952a) and increased gradually to about 23,000 acre-ft in 1960 (fig. 10). From 1965 through 1976, pumpage increased to more than 50,000 acre-ft/yr (U.S. Geological Survey, 1985) and has remained near 50,000 acre-ft/yr since 1976. Most pumpage for irrigation is from the stream deposits along the San Pedro River in which ground water is unconfined. Most pumpage for industrial and public-supply uses is from the deep aquifer system in which ground water is confined. Before 1953, all ground-water withdrawal was for irrigation; however, by 1976, 53 percent was for uses other than irrigation. A summary of the ground-water pumpage history in the lower San Pedro basin is included in figure 10.

Model Characteristics

The conceptual model of the hydrologic system of the lower San Pedro basin is similar to those of the upper San Pedro and Benson basins. The results of previous studies provided the basis for developing the initial model; these studies include Bryan and others (1934), Heindl (1952a), Page (1963), Roeske and Werrell (1973), and Jones (1980). The numerical model was used principally to test the transferability of information on mountain-front recharge, river-aquifer interaction, evapotranspiration, transmissivity, hydraulic conductivity, and vertical interlayer leakance. The characteristics of the developed models of the upper San Pedro and Benson basins were used as initial property values for the model of the lower San Pedro basin.

The numerical model developed for the lower San Pedro basin had the following characteristics and assumptions:

- A variable column width is specified to enhance resolution along the axis of the basin where discharge to the stream and evapotranspiration occurs. Row width is a uniform 1 mi (see fig. 10).
- Two layers representing (1) the stream alluvium along the flood plain of the San Pedro River in which ground water is unconfined and (2) the basin-fill sediment throughout the basin extending to the mountain fronts. Ground water is confined in the basin-fill layer underlying the basin-center fine-grained facies and unconfined outside this area extending to the mountain fronts.
- The fine-grained facies separating the stream alluvium from the underlying confined aquifer is represented quantitatively by the vertical-leakance value, which is

the vertical hydraulic conductivity of the fine-grained facies divided by its thickness.

- Mountain-front recharge was applied to the lower layer only.
- Stream-aquifer interchange takes place and is controlled by the stream elevation, stream-bottom elevation, stream-leakance value, and head in the upper layer.
- Evapotranspiration was simulated as a discharge of ground water from the upper aquifer and controlled by a linear relation between the maximum evapotranspiration rate and a depth to water of 10 ft, where evapotranspiration was assumed to cease.
- Transient simulation used two pumping periods in which withdrawals from both layers were represented. Irrigation pumpage, which was simulated as withdrawal from the upper layer, was reduced by 30 percent to account for return of excess applied irrigation water.

The finite-difference grid for the lower San Pedro basin model (fig. 10) consisted of 18 columns and 69 rows making 1,242 rectangular cells in each layer from 0.4 to 1 mi^2 in area. The upper layer included 248 active cells, and the lower layer included 865 active cells. Mountain-front recharge was introduced into the lower layer as a constant flow at a rate that varied depending on the altitude of the adjacent mountain range and average annual precipitation in the mountains. Underflow into the basin at the Narrows (south end) and out of the basin at Winkelman (north end) was simulated by constant-head cells; underflow into the basin at the mouth of Aravaipa Canyon was simulated as a constant flow. Vertical leakage through the fine-grained confining unit was controlled by an array representing leakance values (K'/b) and ranged from 10^{-9} to 10^{-6} (ft/s)/ft, depending on the thickness and character of the fine-grained facies.

Results of Simulations

The steady-state model was adjusted through a trial-and-error process to simulate the general shape of the estimated water-level configuration and to obtain flow-component values that were reasonable approximations of estimated water-budget values. Because correlative field data were sparse, however, a detailed definition of the hydrologic system could not be established. Matching measured water levels to within appropriate accuracy criteria was not done because actual predevelopment potentiometric levels for the confined aquifer were not available. After an acceptable steady-state simulation was obtained, the model was stressed by simulated pumping. The response of the transient model was compared with available field evidence to determine qualitatively the representation of the actual ground-water system by the model.

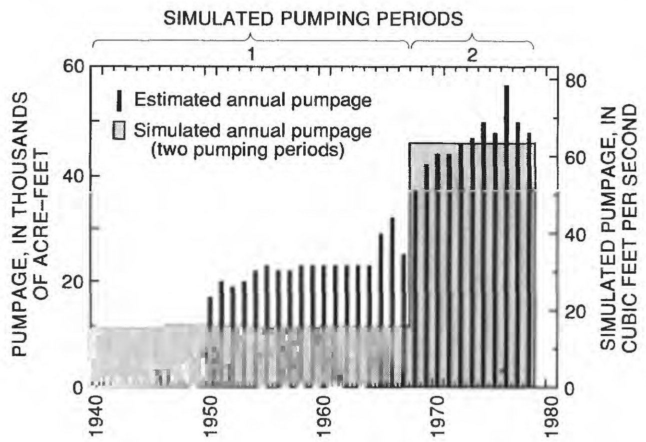
Steady-state simulation produced a hydraulic-head configuration similar to the estimated predevelopment potentiometric surface, although differences of as much as +50 ft were present in places near the mountain fronts. Water-budget outflow components generally were smaller than estimated outflow components in previous investigations; however, inflow components generally matched estimated inflow components (fig. 10). The difference is in the balance of inflow with outflow. Previous estimated inflow-outflow values were not balanced, whereas the model inflow-outflow values were balanced. The model can be adjusted to simulate larger or smaller inflow and outflow. For example, inflow and outflow components can be increased by a factor of two by increasing mountain-front recharge, transmissivity, vertical leakance, and hydraulic conductivity, with little or no resulting change in water-level configuration. The model inflow, however, would then be twice the inflow estimated on the basis of field data. The model is not sufficiently sensitive to variations in water-budget components to define flow components or aquifer characteristics with more precision.

The transient simulation represented two pumping periods. The termination of these periods represents the ends of 1967 and of 1978, which are times for which water-level maps are available (Roeske and Werrell, 1973; Jones, 1980). Differences in the water-level configuration shown on the two maps are small; however, water-level measurements in a deep well near Mammoth indicated a water-level decline of about 20 ft between 1967 and 1978. Model results indicate an average water-level decline of about 4 ft in the lower layer for the same period and localized declines of as much as 150 ft. The reason for this large difference between field and simulated conditions is unknown but could be because data are insufficient to define local geohydrologic conditions and pumping distributions. Between 1967 and 1978, net pumpage from the aquifer approximately doubled. Model results indicated that storage depletion increased from about 1,700 acre-ft in 1967 to 8,250 acre-ft in 1978 (fig. 10). The model results also indicate that natural discharge by evapotranspiration decreased from nearly 30,000 acre-ft/yr in the early 1940's to less than 23,000 acre-ft in 1967 and to about 16,000 acre-ft in 1978 as a result of water-level declines. Model results also indicate that base flow supplied by aquifer discharge decreased from more than 22,000 acre-ft/yr in the early 1940's to about 10,000 acre-ft in 1967 and to less than 6,000 acre-ft in 1978. The head decreases in both parts of the aquifer system were accompanied by a decrease in the net upward flow from the lower aquifer to the upper aquifer.

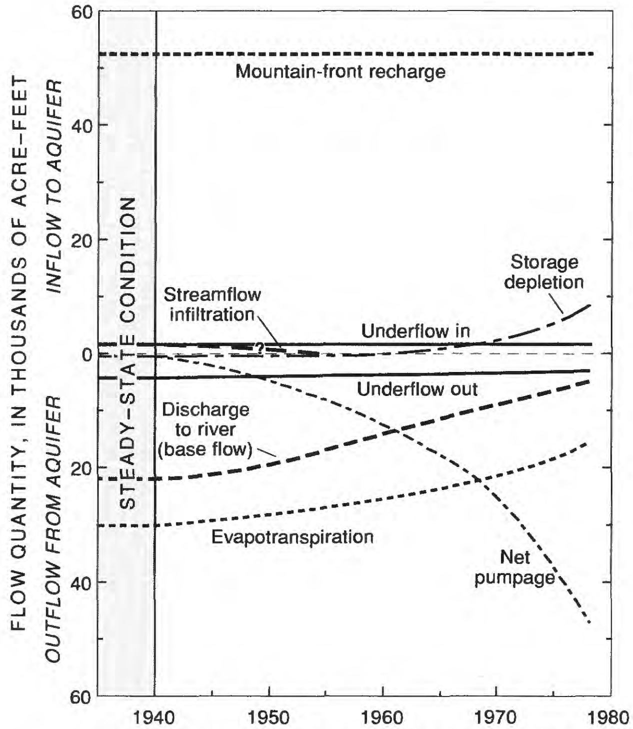
ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS

[Values are in acre-feet per year: dashes indicate no data. Estimated values modified from Heindl (1952a) and unpublished records of the U.S. Geological Survey]

Components	Steady state: Estimated	Steady state: Simulated
Inflow:		
Mountain-front recharge	25,000 - 27,500	26,400
River losses	2,200	900
Underflow from Benson basin	1,200	1,700
Total	28,400 - 30,900	29,000
Outflow:		
River gains	21,700	11,400
Underflow to Gila River	2,700	2,100
Evapotranspiration	12,700 - 38,000	15,600
Springs	3,200	------
Total	40,300 - 65,600	29,100



INDEX MAP SHOWING LOCATION OF LOWER SAN PEDRO BASIN (SHADED)



SIMULATED CHANGES IN WATER-BUDGET COMPONENTS AS A RESULT OF INCREASED GROUND-WATER WITHDRAWAL IN LOWER SAN PEDRO BASIN, 1940–78

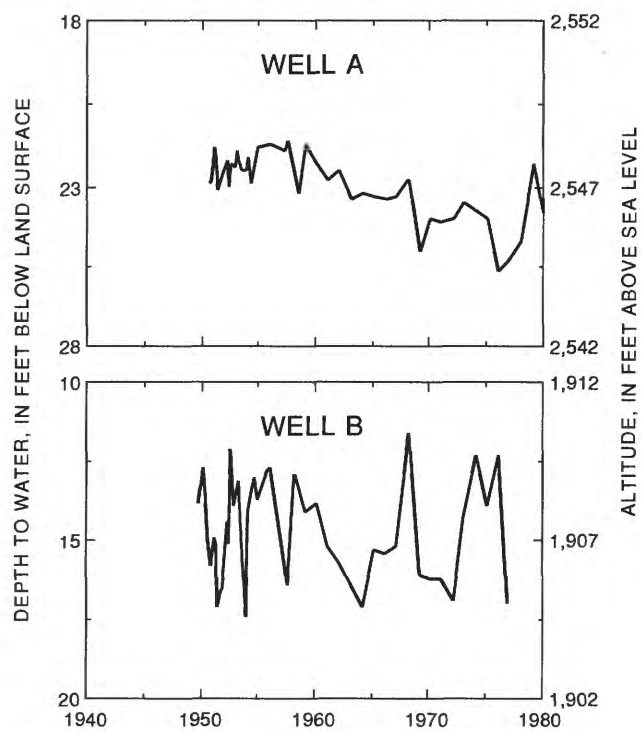
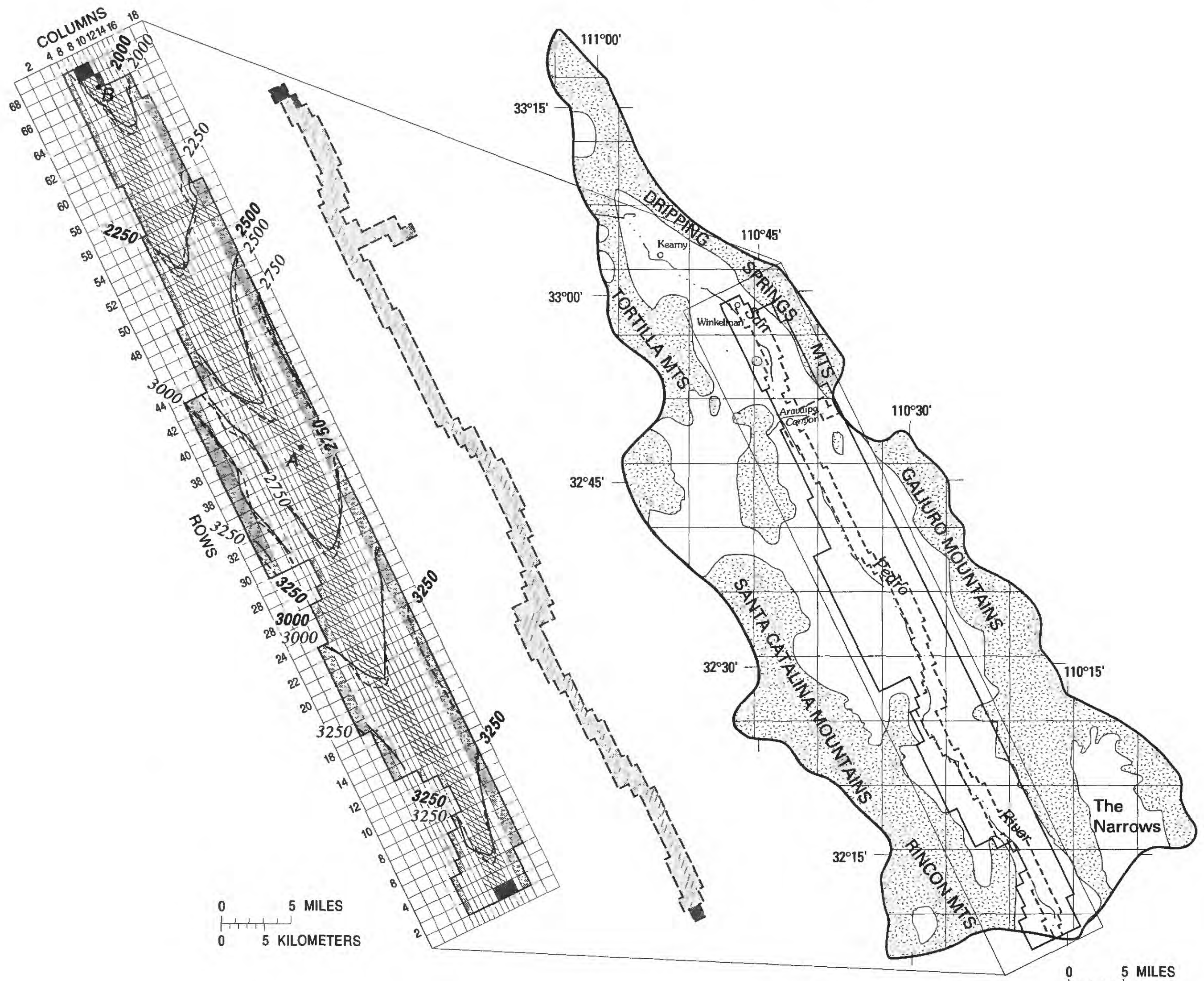


FIGURE 10.—Location of basin, estimated and simulated water-budget components and pumpage, ground-water levels, model grid and boundary, simulated changes in water-budget components, and observed and simulated water levels, lower San Pedro basin.



Base from U.S. Geological Survey State base map, 1:500,000, 1974

EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

- Constant head
- Evapotranspiration
- Mountain-front recharge
- River-aquifer connection
- Vertical leakance

Geologic contact

Boundary of ground-water area

Boundary of modeled area—

- Lower layer
- Upper layer

Water-level contour, 1978–79— Shows altitude of water level. Contour interval 250 feet. Datum is sea level

- - - *2000* - - - Observed (Jones, 1980)

—*2000*— Simulated

•*B* **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	16 - 32
Average	31.6
Transmissivity of lower layer, in feet squared per day	
Range	67 - 5,346
Average	947
Specific yield	
Lower layer	0.086
Upper layer	0.146
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-6} - 10^{-9}

FIGURE 10.—Continued.

WILLCOX BASIN

GEOHYDROLOGIC CONDITIONS

Willcox basin is a topographically closed basin in the southeastern part of the study area (fig. 7). Although small quantities of ground water may have discharged to adjacent basins before development, the basin is internally drained, and the basin center is a large, barren playa. The bedrock of the mountains crops out in about one-third of the 1,660-mi^2 basin, bounds the basin, and forms its physical and hydrologic boundaries. The remaining two-thirds of the basin area is underlain by alluvial sediments that form the aquifer system.

The sediments that fill Willcox basin consist of pre-Basin and Range rocks, lower and upper basin fill, and stream alluvium. Pre-Basin and Range rocks occur at depth in the basin and are not penetrated by wells. Lower and upper basin fill form the aquifer system. Coarse-grained facies of these units occur near the mountain fronts, and very fine grained facies, composed primarily of silt and clay, are present at the basin center underlying and forming the playa. Near the playa, the sediments are heterogeneous and consist of interfingering fine and coarse facies. Stream alluvium consists primarily of sands and gravels that underlie the many washes that enter the basin from the mountains.

Average precipitation ranges from less than 12 in./yr on part of the valley floor to more than 35 in./yr in the surrounding mountains (University of Arizona, 1965). Ground water originates as precipitation falling in the mountains. Recharge is a result of infiltration of surface runoff into the coarse sediments along the mountain fronts. Before development, ground water moved from the mountain fronts toward the playa where discharge occurred through spring flow, evaporation, and transpiration (fig. 11).

Development of ground-water resources in Willcox basin paralleled that of other basins that have agriculturally oriented uses. Ground-water withdrawal generally was less than 5,000 acre-ft/yr until the end of World War II (U.S. Geological Survey, 1982). Withdrawal increased gradually to about 40,000 acre-ft in 1952; however, in 1953, ground-water withdrawal more than doubled to about 94,000 acre-ft. Pumping increased steadily after 1953 to about 300,000 acre-ft in 1967 (fig. 11) (U.S. Geological Survey, 1982). Since 1967, rates of ground-water withdrawal generally have remained stable. Although exact quantities are not well documented, the quantity of ground water used for domestic, livestock, public supply, and industrial purposes is small compared to that used for irrigation.

MODEL CHARACTERISTICS

A numerical model of Willcox basin was developed to examine the hydrologic similarities of this basin and others of the southeast group. Only a small number of data were available to provide the initial values of hydrologic properties for use in the model. Previous studies by Meinzer and Kelton (1913), Coates (1952), Brown and others (1963), Brown and others (1966), and Brown and Schumann (1969), provided the basis for model development. One objective of developing this model was to determine if the relation between precipitation and mountain-front recharge that was used in the model for the upper San Pedro basin remained valid when used in the model for Willcox basin.

The numerical model was developed to simulate steady-state and transient conditions and had the following characteristics:

- A uniform row and column width of 1 mi.
- Two layers representing (1) the aquifer system in the upper 1,000 ft of basin fill from which most withdrawal occurs and (2) the aquifer system below 1,000 ft for which few field data were available (fig. 11).
- The two aquifer layers were connected hydraulically by a leakance value to allow vertical flow. Confined ground-water conditions, prevalent at the edge of the playa on the east side, could not be adequately represented because of the model scale. Representation of these areas as unconfined probably caused actual drawdown to exceed simulated drawdown in those areas.
- Perimeter recharge applied to the upper model layer generally was uniform within the influence of each individual range of mountains. Simulated recharge from the Chiricahua Mountains on the southwestern perimeter of the basin was distributed over a large area of the model on the basis of chemical characteristics of the ground water (F.N. Robertson, hydrologist, U.S. Geological Survey, oral commun., 1981).
- Drain cells were used to represent spring discharge at four points at the margin of the playa. Simulated spring discharge was dependent on the difference between head in the aquifer and altitude of land surface.
- In areas where depth to water was less than 25 ft (Meinzer and Kelton, 1913), evaporation from the playa and transpiration by mesquite and grass immediately adjacent to the playa were simulated with a single maximum evapotranspiration rate and extinction depth of 25 ft.
- Four periods of pumping were simulated. Irrigation return flow was not incorporated, although it probably does occur and could have been simulated to improve the match of measured and simulated water levels. The simulated pumping periods are a compromise between

periods of generally uniform pumping and periods for which pumpage and water-level records are available.

The finite-difference grid used for simulating the hydrologic system of Willcox basin consisted of 55 rows and 24 columns (fig. 11). A total of 711 cells were active in each of the two model layers. Pumpage data for the transient runs were distributed on the basis of well location and date of drilling because pumpage data for individual wells were not available.

Results of Simulations

The steady-state (predevelopment) model was matched to the water-level configuration of 1910 (Meinzer and Kelton, 1913). The transmissivity distribution obtained for the upper layer by using transmissivity estimated by Brown and Schumann (1969) worked well in the model and, for the most part, was not altered. A recharge rate of 47,000 acre-ft/yr along the mountain fronts provided the most reasonable match of hydraulic properties and water-level configurations. Estimates of mountain-front recharge from earlier investigations ranged from 20,000 to 75,000 acre-ft/yr.

The transient model was calibrated for 1910–74 with intermediate pumping periods ending in 1946, 1952, and 1962. Simulated water levels for the first 50 years of the transient period were higher than measured water levels, but in the last 10 to 15 years of the transient period, simulated water levels were lower than measured water levels. This difference may indicate that a return of excess applied irrigation water to the aquifer may have occurred after a significant time lag or that transmissivity and specific yield used in the simulation change significantly with depth. Measured water levels in the Kansas Settlement area (fig. 11) were consistently lower than simulated water levels. The difference may be caused by the extensive interfingering of coarse material extending from the basin perimeter toward the playa and fine-grained lacustrine deposits of the playa. The interfingering, in turn, causes water-level response to pumping to be more characteristic of a confined aquifer system rather than an unconfined aquifer system as was simulated. The deep, steep-sided cone of depression also suggests a confined-aquifer system response. In addition, much of the pumping in the Kansas Settlement area occurs close to the playa. The very fine grained lacustrine material that underlies the playa could effectively create a boundary condition that results in increased drawdown as the cone of depression intersects this poorly permeable unit.

The regional geohydrologic character of the Willcox basin is similar to that of other basins in southeastern Arizona. Recharge and discharge mechanisms are similar except that no underflow occurs into or out of adjacent basins. The distribution of pumping in the basin, however, is not similar because of a clay deposit at the basin center. The clay deposit extends from land surface to below the present depths of pumping wells. Boundary conditions created by this clay deposit may increase the rate of storage depletion, and delayed drainage may be a long-term consideration.

The quantity of mountain-front recharge into the basin was evaluated by using the steady-state model. The initial value simulated, about 40,000 acre-ft/yr, was based on the same relation between total rainfall greater than 8 in. and recharge used for the upper San Pedro basin and other basins in the study area (Anderson and others, 1992). The transmissivity of the upper model layer, which was simulated as the product of saturated thickness and hydraulic conductivity, was deemed to be one of the most reliable property values and was not altered during simulation. To accommodate different recharge to the system, transmissivity of the lower model layer was altered. Values of simulated mountain-front recharge tested in the steady-state model were about 24,000, 47,000, and 74,000 acre-ft/yr. After an initial model run, the largest value was not tested further because the transmissivity of the lower unit required for that annual rate of recharge was considered unreasonably large. The two smaller recharge values were tested further in the transient-model analysis. The most reasonable match between water-level-decline rates measured in observation wells and water-level-decline rates simulated was obtained by using the recharge rate of 47,000 acre-ft/yr.

DOUGLAS BASIN

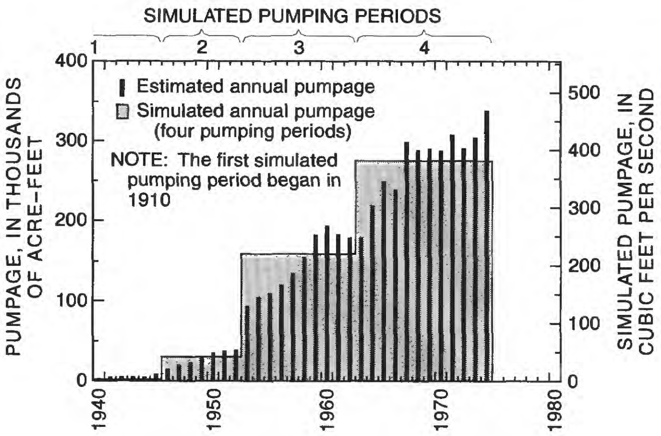
Geohydrologic Conditions

Douglas basin in southeastern Arizona includes about 1,440 mi^2 in the United States and Mexico (fig. 7). About 840 mi^2 is underlain by basin fill; the remaining area is bedrock of the mountains. Douglas basin is geohydrologically similar to the upper San Pedro basin. The principal aquifer is composed largely of fine-grained sediments in both the upper and lower basin fill, and lithologic differences between the units are difficult to distinguish. The sediments grade from coarse near the mountain fronts to as much as 80-percent silt- and clay-size particles near the basin center. Ground water generally is unconfined in the basin; however, ground water is confined in places, particularly at depth in the southern part of the basin. The model of Douglas basin served as a means of testing and verifying the applicability of information transfer and testing the empirical relation between precipitation and mountain-front recharge. Previous studies by Meinzer and Kelton (1913), Coates and Cushman (1955), White and

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS

[Values are in acre-feet per year; dashes indicate no data. Estimated values derived from information in Meinzer and Kelton (1913), Coates (1952), Brown and others (1963), and Mann and others (1978)]

Components	Steady-state Estimated	Steady-state Simulated
Inflow:		
Mountain-front recharge	20,800 - 75,000	47,000
Outflow:		
Evaporation	75,000	46,300
Springs	------	700
Total	75,000	47,000



0 50 MILES

0 50 KILOMETERS

INDEX MAP SHOWING LOCATION OF WILLCOX BASIN (SHADED)

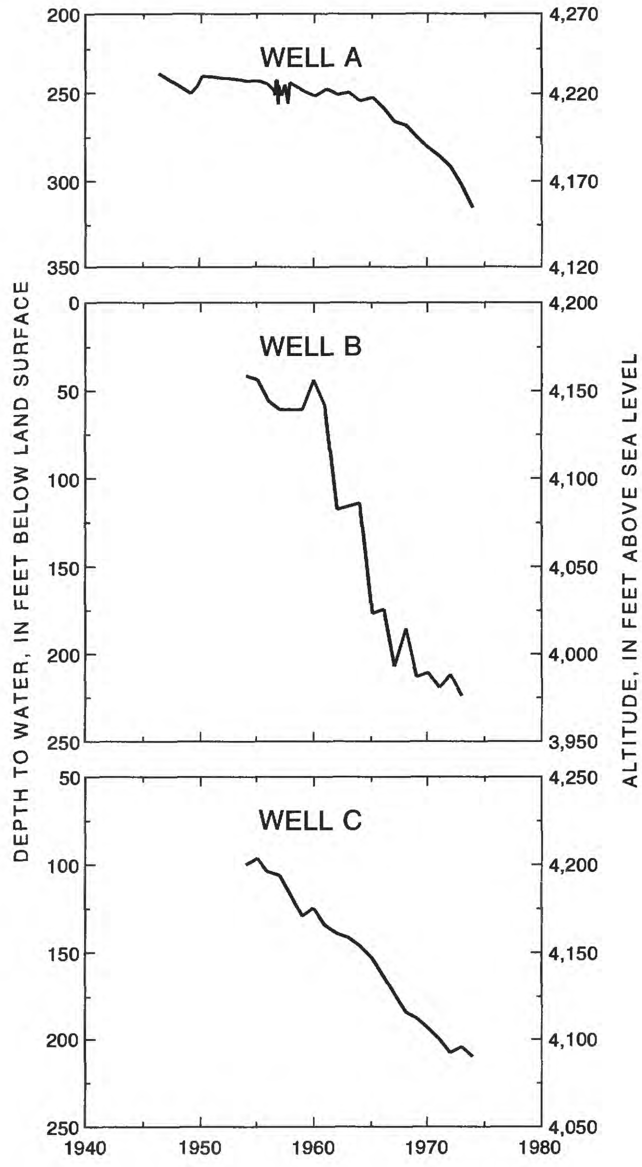
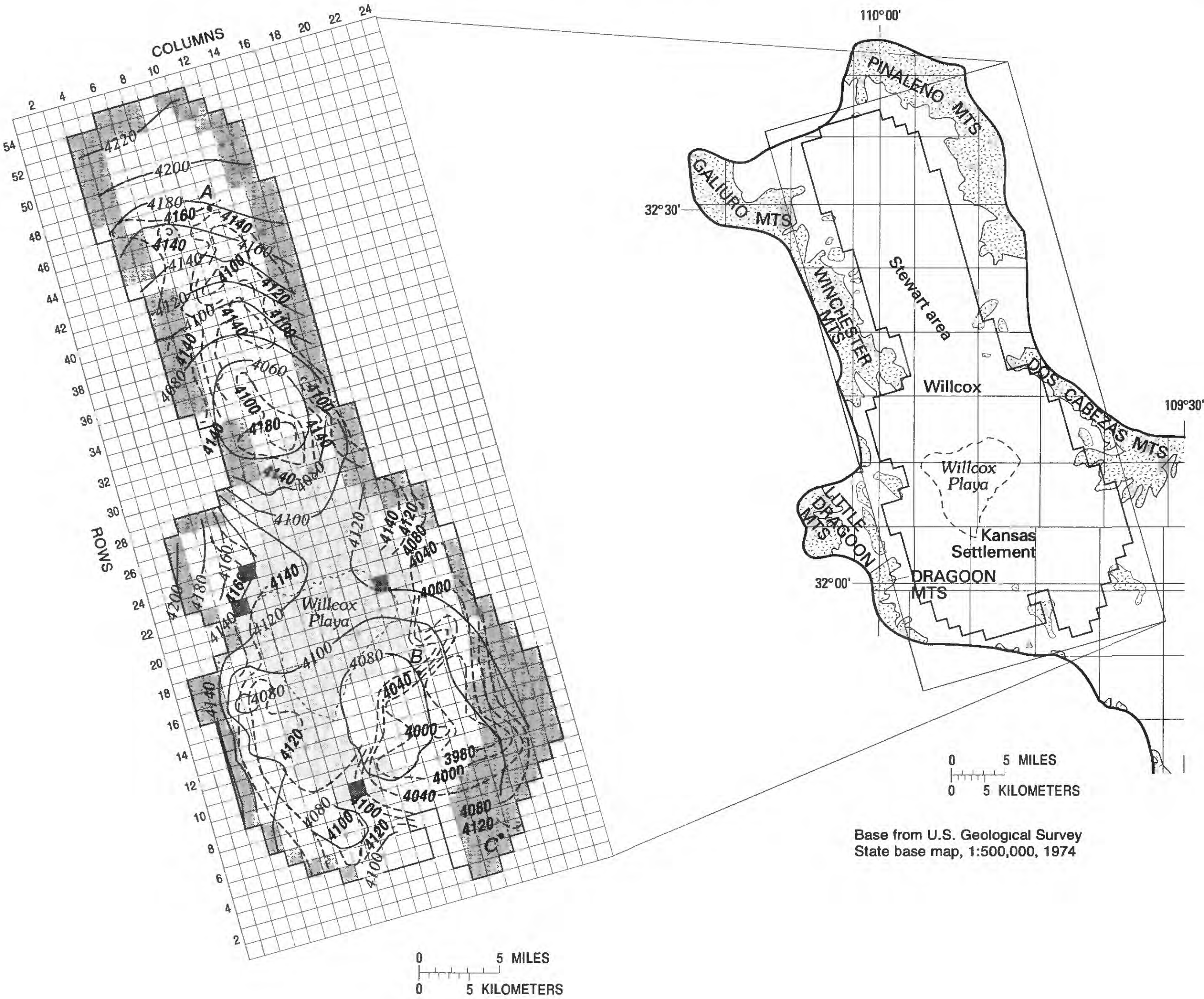


FIGURE 11.—Location of basin, estimated and simulated water-budget components and pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, Willcox basin.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

Mountain-front recharge

Evapotranspiration

Spring

Geologic contact

Boundary of ground-water area

Boundary of modeled area— Upper and lower layers

Water-level contour, spring 1975— Shows altitude of water level. Contour interval 20 and 40 feet. Datum is sea level

--- *4140* --- Observed (from Mann and others, 1978)

—*4140*— Simulated

•*B* **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	0.4 - 47
Average	8.0
Transmissivity of lower layer, in feet squared per day	
Range	400 - 54,980
Average	9,109
Specific yield of upper layer	0.098
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-7} - 10^{-9}

FIGURE 11.—Continued.

Childers (1967), and Mann and English (1980) provided the basis for model development.

Precipitation on the valley floor is less than 12 in./yr; however, the surrounding mountains receive more than 20 in./yr, and parts of the Chiricahua Mountains to the east receive as much as 35 in./yr (University of Arizona, 1965). Ground water moves from the basin margins to the basin axis and then toward the south end of the basin. The origin of ground water at the basin margins is surface runoff from the mountains that infiltrates the coarse sediments of the mountain fronts. Ground water discharges from the basin as base flow in Whitewater Draw and as underflow into Mexico. Before development, ground water also was discharged by evapotranspiration.

Ground water is withdrawn for irrigation, industrial, and public-supply uses in the Douglas basin. Withdrawal began about 1910 but was less than 5,000 acre-ft/yr until the close of World War II. At that time, about half the pumpage was for other than agricultural purposes. Beginning about 1950, irrigation became the major use of ground water. In 1974, pumpage was 138,000 acre-ft (fig. 12), and about 96 percent was used for irrigation (U.S. Geological Survey, 1982). Pumpage subsequently decreased to 80,000 acre-ft in 1978 (U.S. Geological Survey, 1982); however, the system continues to be overdrafted—that is, more water is withdrawn than recharged back into the system.

Model Characteristics

The numerical model developed for the Douglas basin had the following characteristics:

- Uniform row and column widths of 1 mi.
- Two layers representing (1) the upper part of the aquifer that was penetrated by wells and (2) the lower part of the system as yet unpenetrated by wells (fig. 12).
- A vertical leakance connecting the two layers represents the vertical anisotropy in hydraulic conductivity within the aquifer system, which results in head differential with depth.
- Perimeter recharge was applied to the upper layer of the model to simulate infiltration of runoff from surrounding mountains and underflow in the northeastern part of the basin, north of the Swisshelm Mountains.
- River cells simulate the discharge of ground water from the aquifer along the lower reach of Whitewater Draw.
- Discharge of ground water by evapotranspiration from the upper part of the aquifer in the area occurs where depth to water was less than 15 ft in 1910 (Meinzer and Kelton, 1913). The model uses a linear relation between a maximum evapotranspiration rate and the depth at which evapotranspiration ceases.
- Four periods of simulated pumping represent a generalized pattern of the actual pumping history; simulated pumping used an average rate applied over several years. Simulated irrigation pumpage was reduced by 5 to 15 percent to represent a small quantity of recharge from excess applied irrigation.

Results of Simulations

Development of the steady-state model involved matching the model-calculated water-level configuration for the upper model layer with the measured water-level configuration for 1910 (Meinzer and Kelton, 1913). Initial values of horizontal and vertical conductivity were altered until the average absolute difference between model-calculated and measured heads reached an optimum low value of about 3 ft. The largest differences were near the basin margin where only a few water-level measurements were available. Less than 2 percent of the active cells of the upper layer indicated a head difference of more than 10 ft.

The validity of the steady-state solution was tested further by transient simulation involving four pumping periods: (1) predevelopment (1910) through 1946, (2) 1947 through 1951, (3) 1952 through 1965, and (4) 1966 through 1978. Information on water-level configurations was available for the winter periods of 1951, 1965, and 1978, and some water-budget information was available for 1951 and 1965 for use in the history-matching process (fig. 12). The simulated water level for the 1951 pumping period matched well with observed water levels for 1951. However, because water-level declines at the south end of the basin may be affected by the occurrence of confined conditions locally in the aquifer system, simulated water levels for pumping periods ending in 1965 and 1978 (fig. 12) show more decline than was recorded in the northern and northwestern parts of the basin. This difference indicates that either hydrologic properties such as transmissivity, specific yield, and recharge values were too small; that actual pumpage was less than the quantity simulated; or that a combination of some of these factors occurred. More hydrologic data are needed to confirm these possibilities.

Initial values used in the Douglas basin model for rates of evapotranspiration, river leakance, and interlayer leakance were transferred from the model of upper San Pedro basin. Subsequently, only minor changes were needed in these values to reflect different basin geometry. The general relation between precipitation greater than 8 in. and mountain-front recharge developed as part of this regional study (Anderson and others, 1992) was used to determine the initial values for mountain-front recharge. Use of this relation provided adequate values, although no refinement or further definition could be accomplished because of the lack of detailed data about hydrologic properties and the exact shape of the water-level contours in the recharge areas.

TUCSON BASIN

Geohydrologic Conditions

Tucson basin includes about 2,870 mi^2 in southeastern Arizona and northern Mexico. The basin consists of about equal proportions of alluvial material and bedrock of the mountains. The length of the basin from the international boundary on the south to the outflow, about 20 mi northeast of Tucson, is about 85 mi, but only the northernmost 65 mi was considered in this study (fig. 7). The numerical model included an area about 65 mi long and from 5 to 20 mi wide (fig. 13). Average precipitation is about 12 in./yr on the basin floor and may exceed 30 in./yr in the highest parts of the surrounding mountains.

The basin occupies a structural trough that trends north to northwest. The lithology of the sediments that fill the trough are well defined (Pashley, 1966; Davidson, 1973; Anderson, 1987; Anderson, 1988). Three units have been defined—Pantano Formation, Tinaja beds, and Fort Lowell Formation—that are nearly coincident with the pre-Basin and Range, lower basin fill, and upper basin fill, respectively. The Pantano Formation (Finnell, 1970) is at depth in the basin except where it outcrops on the basin perimeter. More than 8,000 ft of basin-fill sediments, which include the Tinaja beds, Fort Lowell Formation, and stream alluvium, are known to overlie the pre-Basin and Range rocks along the central axis of the basin (Anderson, 1987). The sedimentary units are hydraulically connected and form a single aquifer system in the basin.

Ground water generally is under unconfined conditions although head variations with depth, which reflect local ground-water confinement, are associated with the presence of fine-grained facies. Recharge occurs by infiltration of streamflow along the main stream channels and along mountain fronts. Before development, ground water moved from the areas of recharge toward the center and the downstream end of the basin, where it discharged by evapotranspiration or as surface flow in the Santa Cruz River. Small quantities of water also recharged and discharged from the aquifer system as underflow. After development of the ground-water resources, capture of discharge occurred gradually until only minor quantities of natural discharge remained. The principal discharge mechanism that has continued to occur is underflow at the downgradient end of the basin. As development increased, more ground water was withdrawn from storage within the aquifer. The regional ground-water flow system has been disrupted by pumping, and most of the ground-water movement in the basin is now toward pumping wells and their associated cones of depression.

On the basis of archeological evidence, use of the available surface-water resources in the Tucson basin for irrigation of cropland predates the arrival of the Spanish in the 1500's (Harshbarger and others, 1966). Ground-water withdrawal began prior to 1900, but the quantity of ground-water withdrawal did not become significant until after 1920. Pumpage in 1940 was 62,000 acre-ft (Anderson, 1972) and averaged about 280,000 acre-ft/yr in the mid-1970's (U.S. Geological Survey, 1985).

Water levels declined as much as 150 ft between 1940 and 1978 near the center of two cones of depression in the basin (fig. 13). Rates of water-level decline were as great as 7 ft/yr in the mid-1970's (Murphy and Hedley, 1984) in the area of greatest withdrawals. Because of large-scale withdrawals and associated water-level decline, reaches of the Santa Cruz River that formerly contained perennial flow now contain flow only in direct response to runoff events.

Model Characteristics

The numerical model of the Tucson basin simulated steady-state (predevelopment) and transient (pumping) conditions. The quasi-three-dimensional model was based on an earlier two-dimensional electrical-analog model (Anderson, 1972). The steady-state model was used to refine values and areal distribution of hydrologic properties and to evaluate the rates of ground-water movement within and between the two simulated layers of the aquifer. The model also was used to explore alternative conceptual ground-water models. The transient model was used to determine if the hydrologic properties used in the steady-state model realistically represented and simulated system response to pumping stress.

The steady-state and transient models developed for the Tucson basin have the following characteristics:

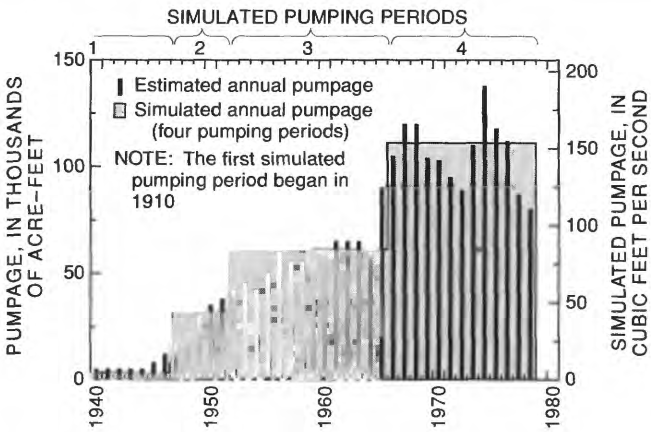
- A finite-difference grid with uniform cell size of 1 mi^2 except at the south end where an expanding row width was used because of sparse data (fig. 13).
- Two layers representing (1) the Fort Lowell Formation, which is equivalent to an upper basin-fill unit, and stream alluvium and (2) the Tinaja beds, which are a lower basin-fill equivalent, and part of the Pantano Formation, which are pre-Basin and Range rocks (Davidson, 1973).
- A vertical-leakance value was used to allow vertical ground-water flow between layers. Vertical anisotropy of hydraulic conductivity in each principal layer is used

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS

[Values are in acre-feet per year; dashes indicate no data. Estimated values modified from Coates and Cushman (1955), White and Childers (1967), and unpublished records of the U.S. Geological Survey]

Components	1951 Estimated	1951 Simulated	1965 Estimated	1965 Simulated
Inflow:				
Mountain-front recharge	20,000	15,500	20,000	15,500
Precipitation	2000	------	------	------
Storage	[1]30,000	21,600	[2]70,000	44,300
Total	52,000	37,100	90,000	59,800
Outflow:				
Stream gains	1,400	0	0	0
Underflow to Mexico	1,400	1,700	0	1,100
Evapotranspiration	8,000 - 13,000	7,100	0	2,000
Springs	200	------	------	200
Pumpage	41,000	28,500	90,000	57,300
Total	52,000 - 57,000	37,300	90,000	60,600

[1]Calculated by using specific yield of 8 percent.
[2]Calculated by using specific yield of 20 percent.



INDEX MAP SHOWING LOCATION OF DOUGLAS BASIN (SHADED)

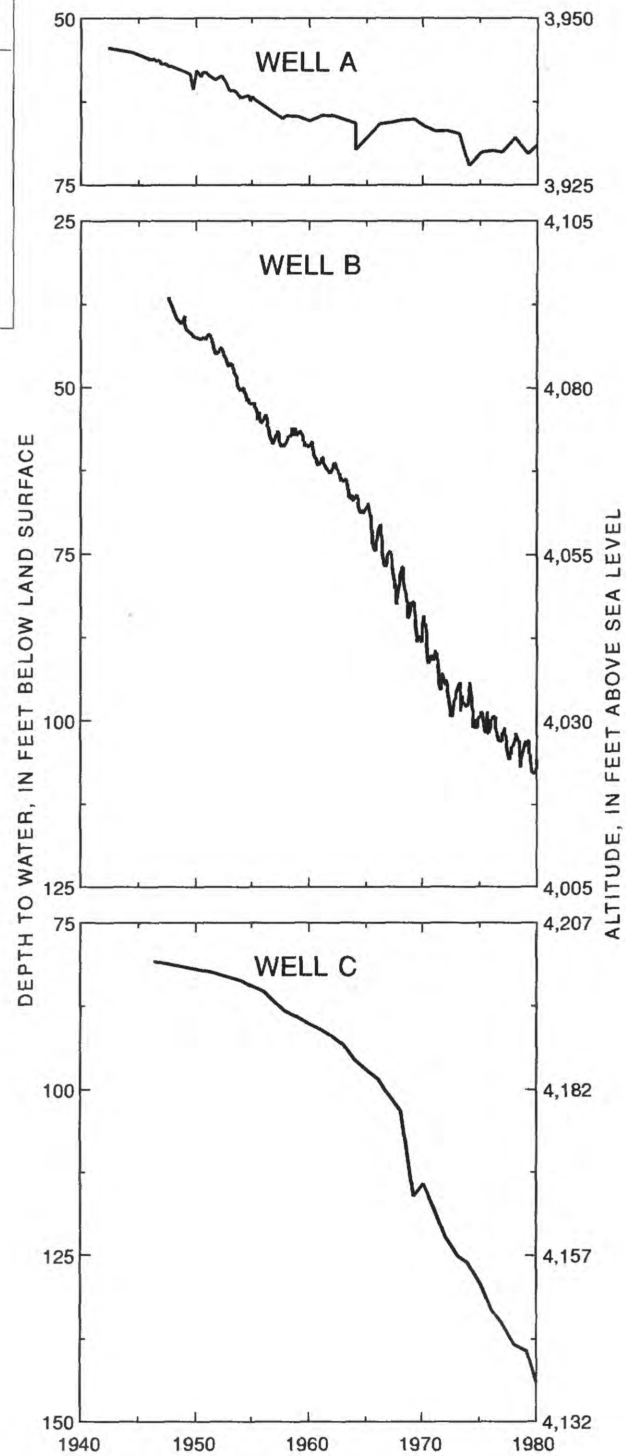
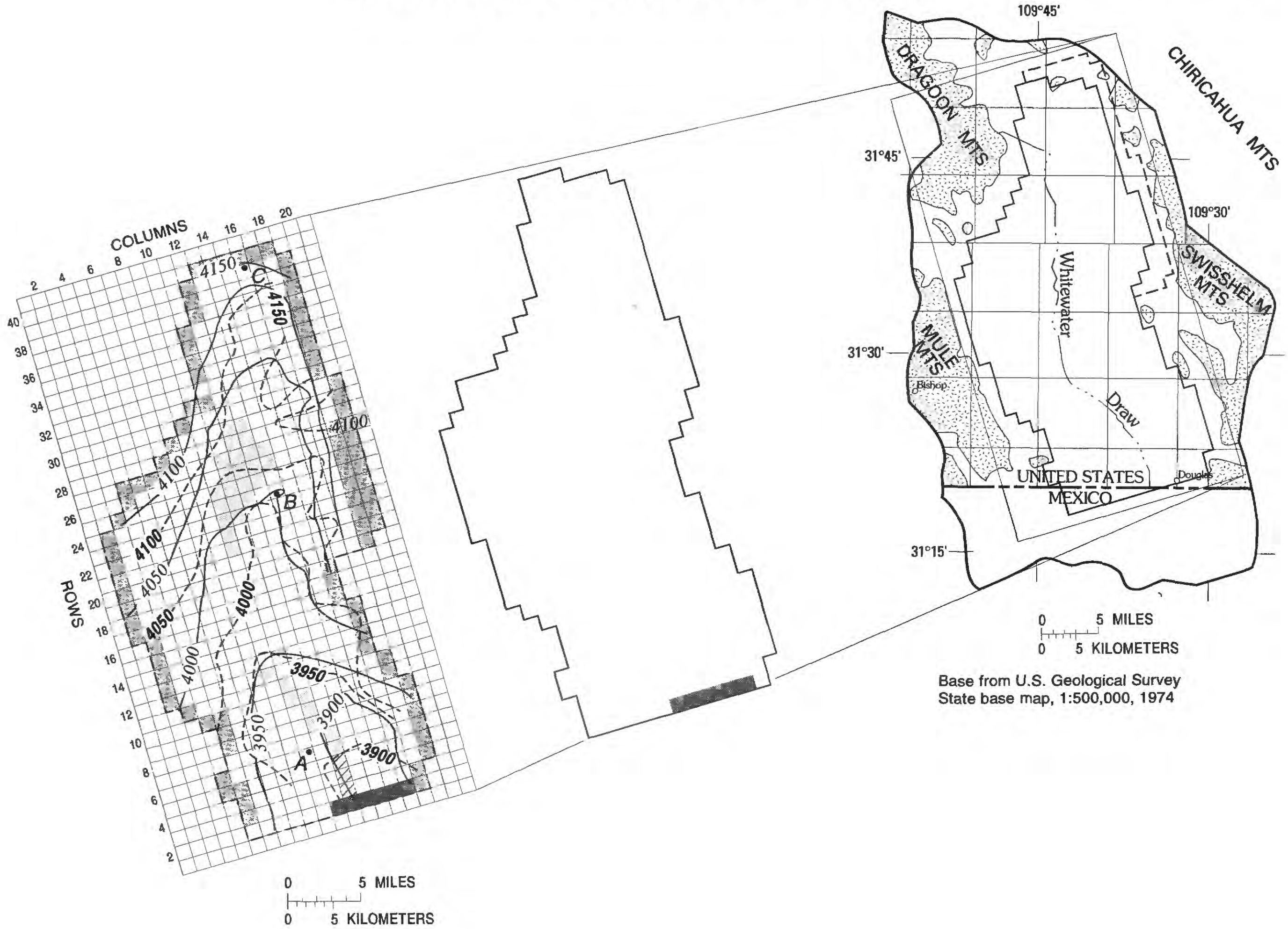


FIGURE 12.—Location of basin, estimated and simulated water-budget components and pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, Douglas basin.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

Constant head

Evapotranspiration

Mountain-front recharge

River-aquifer connection

Geologic contact

Boundary of ground-water area

Boundary of modeled area—

Lower layer

Upper layer

Water-level contour, 1978— Shows altitude of water level. Contour interval 50 feet. Datum is sea level

4050 Observed (Mann and English, 1980)

4050 Simulated

•B **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	1 - 78
Average	7.8
Transmissivity of lower layer, in feet squared per day	
Range	86 - 1,996
Average	1,008
Specific yield of upper layer	0.14
Storage coefficient of lower layer	10^{-5}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-7} - 10^{-10}

FIGURE 12.—Continued.

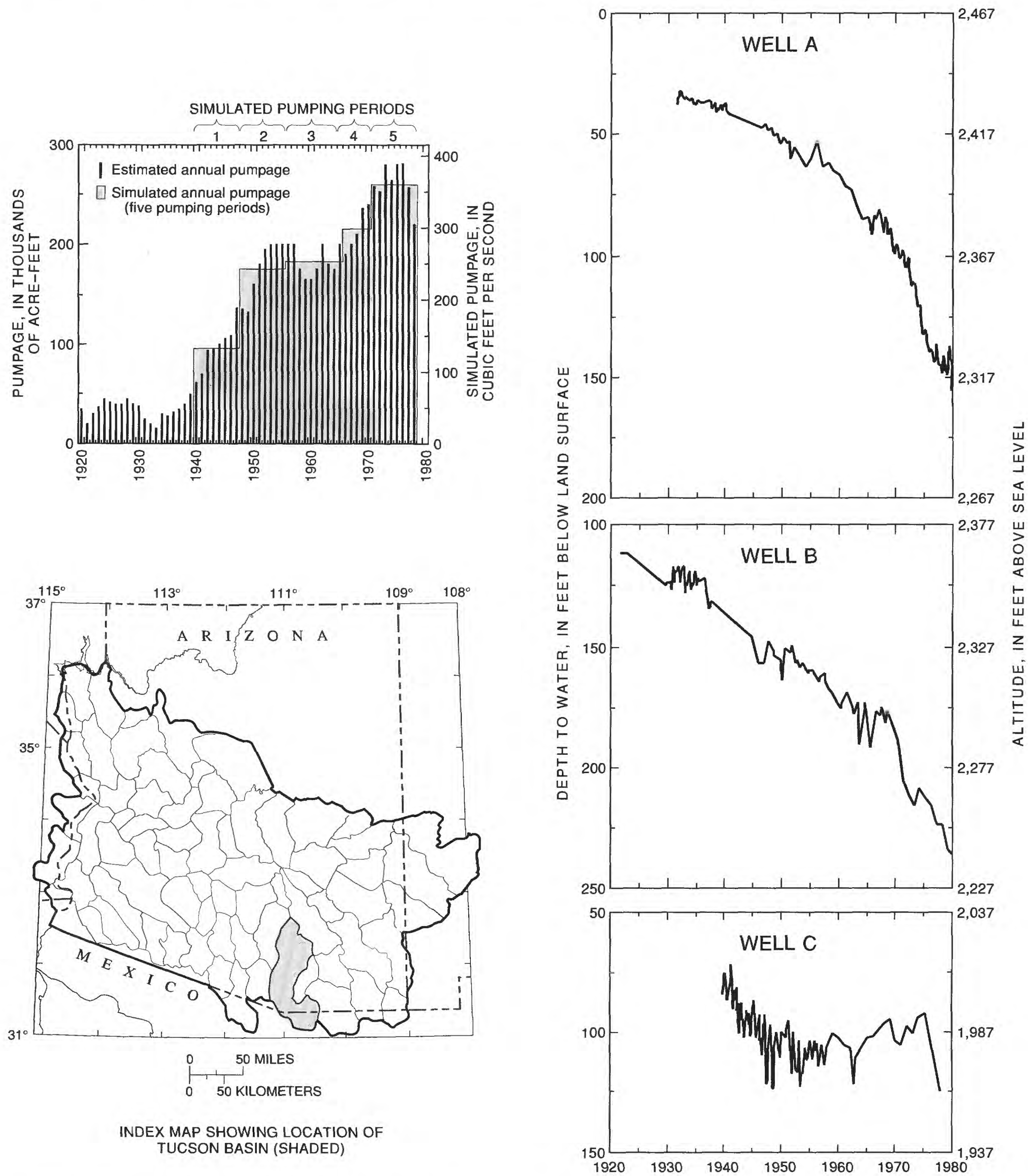
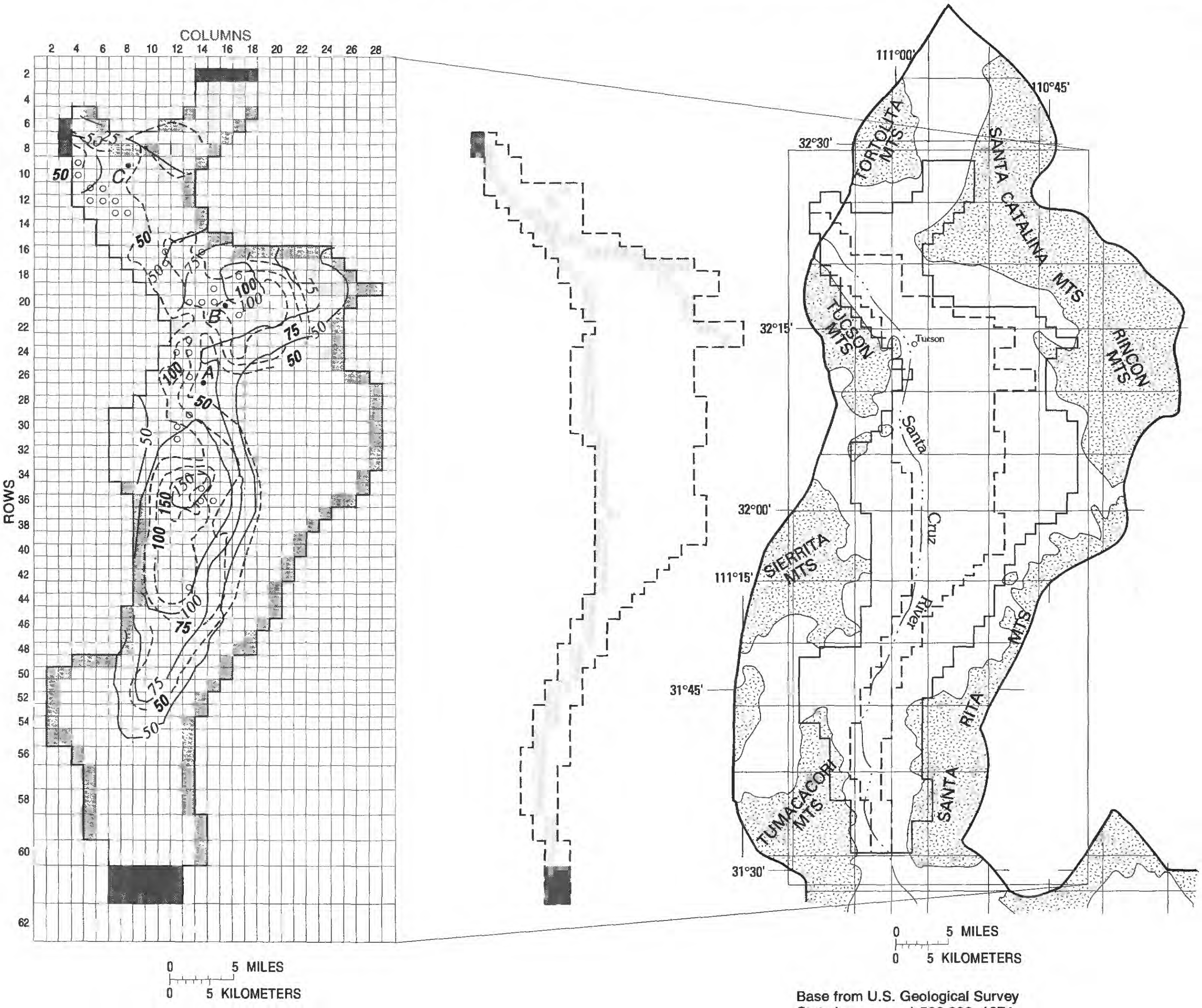


FIGURE 13.—Location of basin, estimated and simulated pumpage, ground-water levels, model grid and boundary, and observed and simulated water-level declines, Tucson basin.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid of model cells used in simulating ground-water flow

Cell representing steady-state boundary conditions

Constant head

Ground-water pumpage

Mountain-front recharge

Streamflow recharge

Geologic contact

Boundary of ground-water area

Boundary of modeled area—

Lower layer

Upper layer

Line of equal water-level decline, 1940–78—Interval 25 and 50 feet

---50--- Observed

—50— Simulated

•*B* **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	3.4 - 716
Average	86.6
Transmissivity of lower layer, in feet squared per day	
Range	18 - 40,300
Average	5,444
Specific yield of upper layer	0.133
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	5×10^{-4} - 4×10^{-6}

FIGURE 13.—Continued.

to calculate the harmonic mean of the vertical leakance at the start of the simulation. The leakance value does not change during the simulation but converts to zero if the upper layer is dewatered.

The steady-state model has the following additional characteristics:

- Recharge to the lower layer along the mountain fronts and along stream channels traversing sediments of the Tinaja beds or Pantano Formation, except in the area adjacent to the Tucson Mountains where no recharge is simulated.
- Recharge to the upper layer along the stream alluvium of the main stream channels.
- Simulated discharge from the lower layer that represents the quantity and areal distribution of pumping in 1940.
- Underflow simulated by constant heads in both layers as inflow at the south end and northeast end of the basin.
- Underflow simulated by constant heads in both layers as outflow near Rillito at the northwest end of the basin.

The transient model had these additional characteristics:

- A superposition approach was used so that only changes in the ground-water system from steady-state conditions were simulated.
- A simulated surface of uniform altitude representing the head in each of the two layers at the beginning of the transient period.
- Discharge from simulated pumping wells representing only the increase in quantity of pumpage since 1940.
- Recharge in areas where the quantity of pumpage had been reduced since 1940 and where natural discharge was captured after 1940.
- Pumping only from the lower layer of the model.

Conditions in 1940 were assumed to represent steady-state conditions of inflow equal to outflow with no long-term change in storage (Anderson, 1972). Although some development had occurred in the northwestern part of the basin before 1940, conditions were assumed to have reached a new equilibrium condition by this time. The transient phase was simulated by five pumping periods—1940–47, 1948–55, 1956–65, 1966–70, and 1971–78. The total water-level decline during 1940–78 (fig. 13) was used for matching water-level declines simulated by the model. For the superposition approach used in the transient analysis, flow quantities used in the model consisted of changes in flow components that occurred after 1940, and model output consisted of changes in water levels since 1940.

Results of Simulations

The electrical-analog model developed by Anderson (1972) was used in development of the quasi-three-dimensional numerical model of the Tucson basin. Because of differences in computational procedures, more complex configuration, and time constraints, the simulated 1940 water levels did not match measured water levels as closely as water levels generated by the electrical-analog model. The greatest difference between the observed and simulated water levels occurred near the basin margins where few water-level data are available (fig. 13). Improved matches between measured and simulated head values could have been achieved by selectively increasing recharge or decreasing transmissivity, depending on the range of values considered reasonable for these properties. Total recharge used in the numerical model was 5,000 acre-ft/yr less than that used in the electrical-analog model, and increasing this flow component appears to be a viable option in improving the match of water levels. Values of hydrologic properties used in the numerical model were about the same as those used in the electrical-analog model.

Water-level-decline maps and hydrographs of specific wells were used in the history-matching process. The distribution and magnitude of simulated head decline were consistently similar to actual measured declines. Simulated declines in the later pumping periods were less than measured declines, which indicated that simulated pumpage was too small, specific yield was too large, additional discharge was not being accounted for in the model, or a combination of these factors. Model-generated declines are much more uniform than those based on field data because of the averaging and simplification of a highly variable, complex hydrologic system that are necessary in the model. A major limitation of the superposition approach used in this basin model was the nonlinearity of the saturated thickness-transmissivity relation. Although water-level declines represented only a few to no more than 10-percent change in saturated thickness, changes in transmissivity of about 45 percent occurred in a few areas between 1940 and 1978. This large change is of special concern in areas where highly transmissive surficial deposits have been dewatered. The superposition approach is not recommended for future use because of this limitation in accuracy of simulation. An alternative approach might be to include more model layers and for each layer to represent a reasonably homogeneous segment of the aquifer.

AVRA VALLEY

Geohydrologic Conditions

Avra Valley occupies a north-trending structural basin in south-central Arizona (fig. 7). The basin includes about 520 mi^2, of which about 75 percent is basin-fill sediments, and 25 percent is bedrock of the mountains. Avra Valley is a broad, gently sloping plain that drains to the north and northwest. The valley is bounded on the east and west by mountains composed of nearly impermeable consolidated rocks. Average precipitation is about 10 in./yr on the valley floor and about 12 in./yr in the mountains (University of Arizona, 1965).

The sediments that fill Avra Valley are hydraulically connected and form a single aquifer system. Upper basin fill is as much as 700 ft thick in the northern part of the valley. Only about 400 ft of these sediments were saturated before development. Upper basin fill thins to less than 200 ft thick in the southern part of the basin. This part of the unit is above the water table and therefore is not part of the aquifer system. The lower basin fill includes extensive fine-grained material in the northern part of the valley and coarse-grained material in the south.

Before development, the ground-water system of Avra Valley consisted of a large storage reservoir having small quantities of recharge and discharge. Recharge occurs as underflow from the south; a small quantity of mountain-front recharge may enter the system from the east and west. Additional underflow enters the basin from the adjacent Tucson basin to the east. Before development, the movement of ground water generally was from south to north paralleling the direction of surface drainage and discharge of ground water occurred as underflow to the northwest. Development has resulted in local cones of depression that have altered the original ground-water flow pattern.

The hydrologic system of Avra Valley is typical of those of the central group of basins (fig. 7) but does not have perennial streams or a large recharge from streamflow infiltration. Significant ground-water development began in the early 1940's for agricultural use. Withdrawals were estimated to be 12,000 acre-ft in 1940 (fig. 14) and 159,000 acre-ft in 1975 (U.S. Geological Survey, 1985). Annual withdrawals decreased as agricultural lands were purchased and left unplanted by the City of Tucson, which purchased the land to obtain the rights to pump ground water underlying the property. Because withdrawal exceeded recharge, water levels declined as much as 150 ft between 1940 and 1985 (Cuff and Anderson, 1987). Maximum rates of water-level decline were about 10 ft/yr in the early 1960's (fig. 14).

Model Characteristics

The numerical model for Avra Valley was developed for steady-state and transient analysis and was based on previous work by Moosburner (1972). The model had the following characteristics:

- A finite-difference grid with uniform cell size of 1 mi^2. The model grid consisted of 21 columns and 44 rows (fig. 14). The upper layer had 420 active cells, and the lower layer had 368 active cells.
- Two layers connected by a vertical-leakance value were used to simulate the aquifer system. In the northern part of the valley, the upper layer represents the upper basin fill, and the lower layer represents the lower basin fill. In the south and central parts of the valley, the break between model layers is arbitrary, and both layers represent lower basin fill. The vertical-leakance value was calculated by the model as a function of the ratio of vertical to horizontal hydraulic conductivity for each layer.
- Perimeter recharge was considered negligible in the steady-state and transient simulations, especially in contrast to the large quantity of pumpage and extensive depletion of water in storage during 1940–77.
- The steady-state model used constant-head boundaries for both layers in areas of inflow and outflow. The transient model used constant-flow boundaries at the same locations. The quantity of flow in the transient model was equal to the flow that was simulated in the steady-state (predevelopment) model.
- The pumping history was simulated in six pumping periods—1940–50, 1951–55, 1956–60, 1961–64, 1965–72, and 1973–77 (fig. 14).

Results of Simulations

The steady-state model represents a conversion of the two-dimensional electrical-analog model of Moosburner (1972) to a quasi-three-dimensional numerical model. Water-level altitudes from the steady-state simulation generally matched measured water-level altitudes within 5 ft (fig 14). Larger differences were noted in the southernmost part of the area where data are sparse.

Measured water levels for 1964 and 1977 were used for the history-matching process. Simulated water levels at the end of 1964 compared closely with water levels of Moosburner (1972) and measured water levels. Simulated water levels for the end of 1977 resulted in a poor match with measured water levels, and significantly greater magnitude and rate of drawdown were simulated by the model in the north-central part of the area. These results indicated that the model could be improved by an increase in recharge, an increase in specific yield, a decrease in pumping, or some combination of these

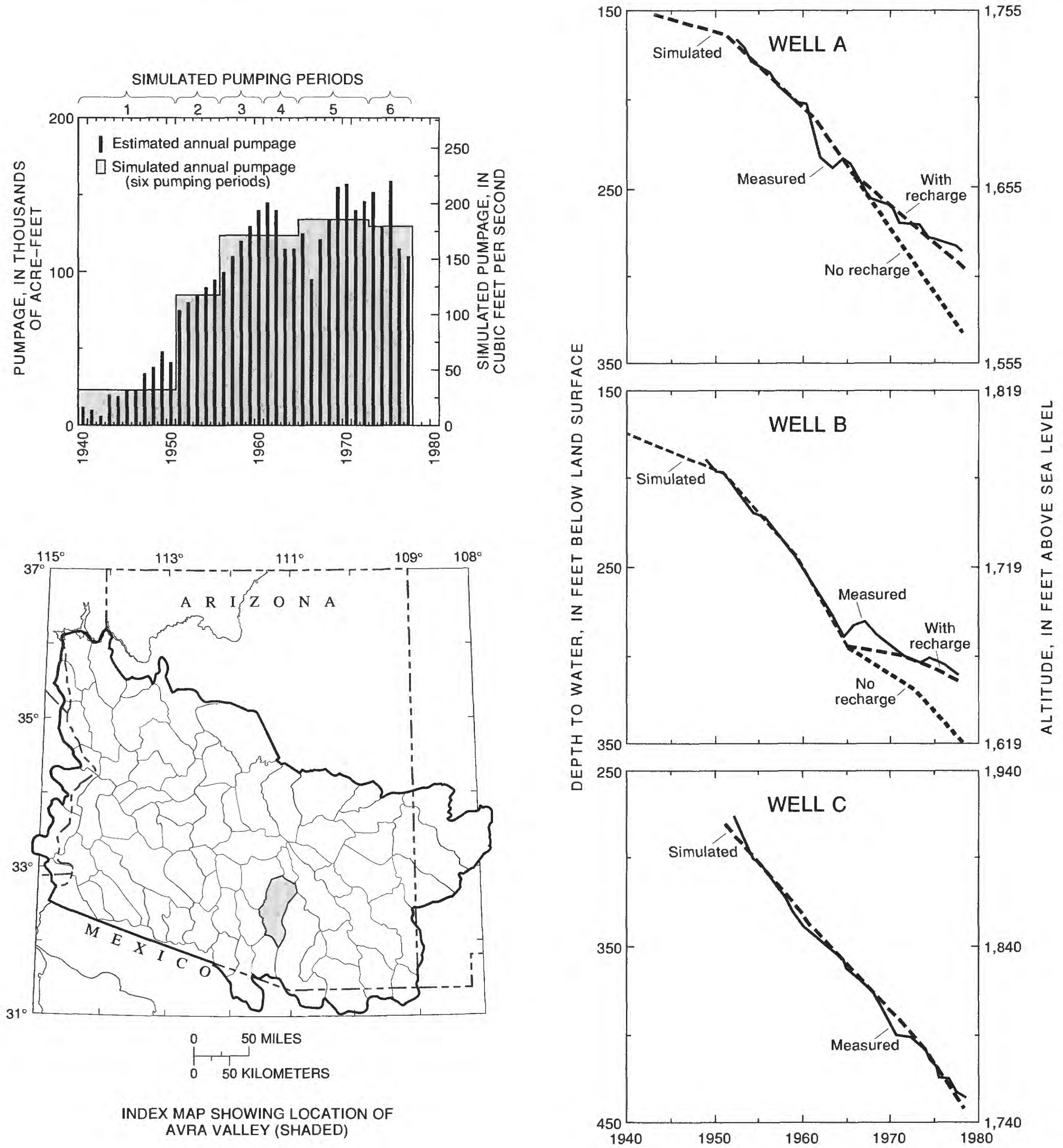
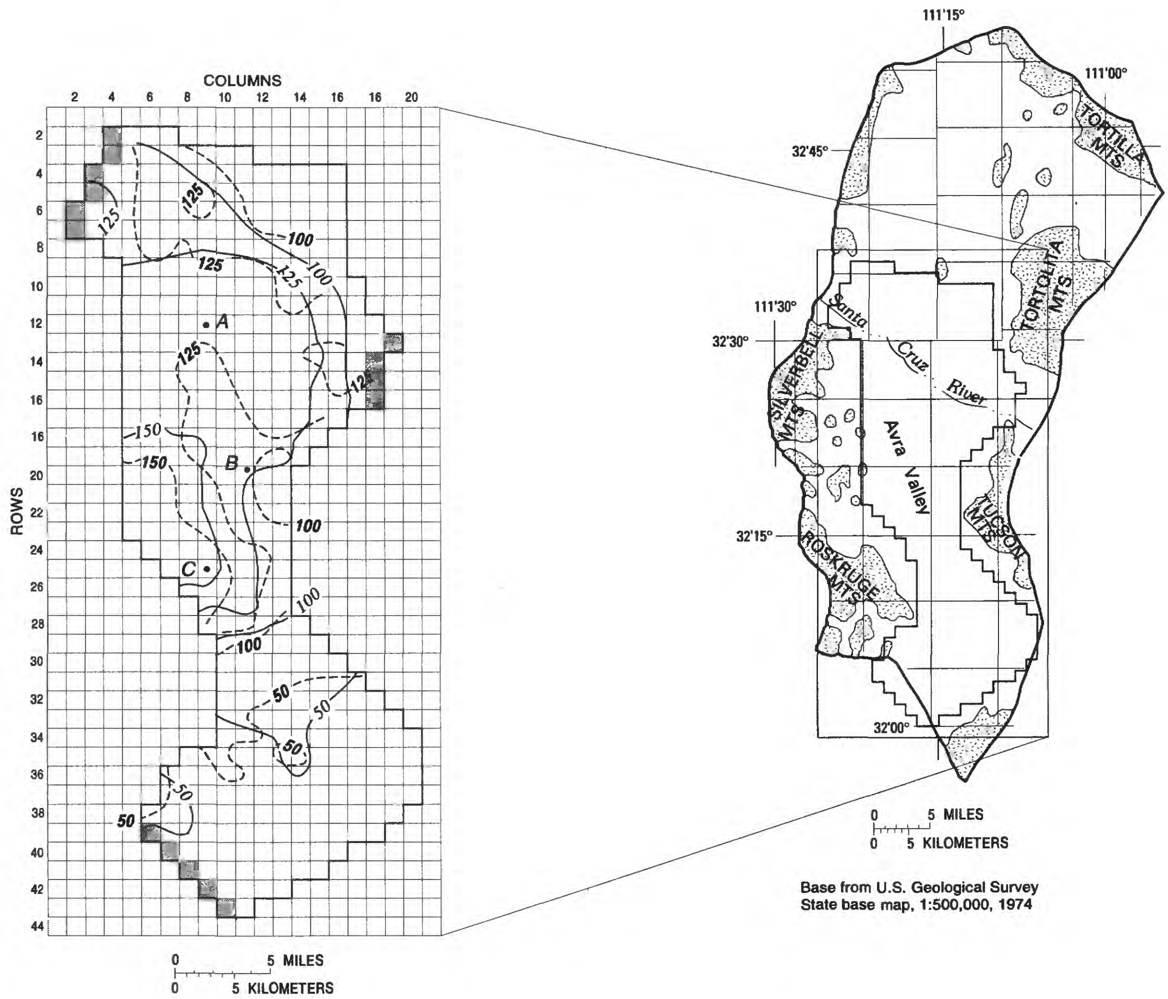


FIGURE 14.—Location of basin, estimated and simulated pumpage, measured and simulated ground-water levels, model grid and boundary, and observed and simulated water-level declines, Avra Valley.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

Inflow

Outflow

Geologic contact

Boundary of ground-water area

Boundary of modeled area—Upper and lower layers

Line of equal water-level decline, 1940-78—Interval 25 and 50 feet

---50--- Observed (based on unpublished data from City of Tucson, University of Arizona, and U.S. Geological Survey)

—50— Simulated

•*B* **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	0.6 - 110
Average	26.6
Transmissivity of lower layer, in feet squared per day	
Range	66 - 4,360
Average	1,530
Specific yield of upper layer	0.131
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, feet per second per foot	
Range	$9x10^{-4}$ - $4x10^{-6}$

FIGURE 14.—Continued.

factors. The extent of the area in which matches between simulated and measured water levels were poor was similar to the known extent of perched ground water. The correspondence in the two areas probably indicates that recharge is entering the regional ground-water system as a result of drainage from the perched ground water. The perched ground water probably originates, at least in part, from excess applied irrigation water.

The model was further used to develop quantitative information on the magnitude and rate of the possible recharge. In the area in which recharge was indicated, depths to water were 200 to 300 ft in the early to mid-1960's, which is the period when simulated and measured water levels began to diverge (hydrographs for wells A and B, fig. 14). In the area to the south where depths to water were greater than 350 ft, the measured and simulated water levels did not diverge (hydrograph for well C, fig. 14). Where recharge from deep percolation of excess applied irrigation water apparently occurred, the volumetric difference caused by the divergence of measured and simulated water levels for 1940–77 was about 500,000 acre-ft. A transit time of 15 to 20 years was indicated for the excess applied irrigation water to move through the 250-ft-thick unsaturated zone to reach the saturated zone (Anderson, 1983, p. 38).

WATERMAN WASH BASIN

Geohydrologic Conditions

Waterman Wash basin is in south-central Arizona about 25 mi southwest of Phoenix (fig. 7) and is typical of the basins of the west group. The basin is drained by Waterman Wash, an ephemeral stream tributary to the Gila River. The basin includes about 470 mi^2, of which about 75 percent is alluvial sediments, and 25 percent is nearly impermeable bedrock of the mountains. The alluvial basin is about 25 mi long and 15 mi wide. Average precipitation ranges from less than 7 in./yr on the valley floor to more than 10 in./yr in the mountains (University of Arizona, 1965).

The sediments that fill Waterman Wash basin include both upper and lower basin-fill units. Pre-Basin and Range rocks may occur at depth in the basin; however, data are not available to verify this speculation. The lithologic units are interconnected hydraulically and form a single aquifer system. Ground water occurs under unconfined conditions in the basin.

Before development, ground-water movement in Waterman Wash basin was minimal because recharge to the ground-water system was small. Ground water moved from the southeast to the northwest, where it discharged as underflow and possibly by evapotranspiration.

Natural recharge to the aquifer of Waterman Wash basin has been estimated to be 1,500 to 2,500 acre-ft/yr (Halpenny and others, 1952, p. 155). Withdrawal of ground water in Waterman Wash basin began as early as 1940 but probably did not exceed natural recharge until about 1950. During the early 1950's, the amount of cultivated acreage increased from virtually nothing to about 15,000 acres, and ground-water withdrawal increased from less than 5,000 acre-ft in 1950 to 60,000 acre-ft in 1960 (fig. 15). Since 1960, the quantity of pumpage has fluctuated between 45,000 and 72,000 acre-ft/yr (U.S. Geological Survey, 1982). Water levels have declined more than 170 ft in the area of greatest withdrawal.

Model Characteristics

The numerical model developed to simulate the hydrologic system of the Waterman Wash basin was based on a previous model by Matlock (1981). Modification of Matlock's single-layer model to a multilayered model enabled analysis of vertical components of ground-water flow. A steady-state simulation was developed to assess the previous concepts of the natural hydrologic conditions, the significance of the quantity and distribution of natural recharge, and the significance of interlayer vertical flow under steady-state conditions. A transient simulation using two periods of pumping—1951–60 and 1961–80—was used to analyze the significance of hydrologic properties during the development period (fig. 15).

The characteristics of the model were as follows:

- A finite-difference grid with uniform cell spacing of 1 mi in both horizontal directions. The model grid consisted of 14 columns and 27 rows, of which 209 cells were active.
- Two layers representing (1) the upper basin fill, which consists of poorly sorted gravel, sand, silt, and clay, and (2) the lower basin fill, which consists of moderately indurated sand and gravel and a small percentage of fine-grained material.
- A vertical-leakance value connected the two layers to represent the vertical anisotropy within the aquifer system.
- Discharge as underflow was simulated in the area where the channel of Waterman Wash exits the basin. Recharge as underflow was simulated at two cells at the south end of the model array.
- Simulated pumpage was from the upper layer only.

Results of Simulations

Although the quasi-three-dimensional steady-state model developed as part of this study was nearly identical with the two-dimensional model developed by Matlock

(1981), one significant difference was the simulation of natural recharge. In Matlock's model, natural recharge of 2,000 acre-ft/yr was distributed around the model perimeter; in this model, 2,000 acre-ft/yr was simulated as entering through two cells at the south end of the area. The differences in water-level configuration are minor and indicate that the effect of natural recharge on the system is minor and that the spatial distribution of recharge is not an important model consideration. This situation probably reflects the combination of a small quantity of recharge entering part of the aquifer having a large hydraulic conductivity. Also, few wells are available in which to measure water levels in the mountain-front area. Thus, a small change in hydraulic gradient, if it occurs, cannot be well defined. These factors result in water-level contours being interpreted as generally straight lines across the basin and nearly perpendicular to the mountain fronts.

The transient model in this study used 1980 as the final stress period, whereas Matlock's model used 1975 for the final stress period. Transient results generally were satisfactory, although the results of the first pumping period do not simulate actual water-level conditions as well as those of the second period. The conclusion was that additional time increments of pumping should have been used to better simulate the temporal and spatial changes in pumping stress.

HARQUAHALA PLAIN

Geohydrologic Conditions

Harquahala Plain is a northwest-trending structural basin in west-central Arizona about 70 mi west of Phoenix and is typical of the west group of basins (fig. 7). The basin includes about 760 mi^2, of which about 30 percent is consolidated rocks, and 70 percent is alluvial material. The area is arid, and average precipitation is about 7.5 in./yr (University of Arizona, 1965).

Basin fill in Harquahala Plain consists of upper and lower units. Pre-Basin and Range rocks are presumed to occur at depth in the basin but have not been penetrated by wells. The upper basin-fill unit includes a heterogenous mixture of coarse- and fine-grained sediments in the southeastern part of the basin and an areally extensive fine-grained facies that increases in thickness from the southeast toward the northwest. On the basis of well logs, the fine-grained facies is more than 1,300 ft thick near the center of the basin. The lower basin-fill unit includes fine-grained material near the basin center and unconsolidated silt, sand, and gravel and conglomerate at the southeast end of the basin. This material, plus the coarser material of the upper basin-fill unit, composes the principal water-bearing unit in the basin. All the sediments are interconnected hydraulically and form a single aquifer system.

Before development, ground-water movement in the aquifer was from the northwest to the southeast (fig. 16); the water-level gradient throughout much of the central and southeastern parts of the basin was about 2 ft/mi. Natural recharge to and discharge from the aquifer system before development were small. Metzger (1957, p. 32) postulated that 90 percent of natural recharge was from infiltrating streamflow and that the total was less than "a few thousand acre-feet annually." Some water enters at the northwest end of the basin as underflow from McMullen Valley. Before development, discharge occurred as either underflow or as a combination of underflow and evapotranspiration at the southeast end of the basin.

Irrigation wells were first drilled in Harquahala Plain in the late 1930's, but pumpage was small until the early 1950's when the first large-capacity wells were developed. Water-level declines of 3 to 4 ft/yr occurred during the mid-1950's. Withdrawals increased markedly from 30,000 acre-ft in 1955, to 125,000 acre-ft in 1960, and to 200,000 acre-ft/yr during 1962–65 (fig. 16). Since 1965, pumpage has diminished steadily to about 100,000 acre-ft/yr during 1978–80 (fig. 16). During 1957–63, water-level declines ranged from 60 to 160 ft in areas of ground-water withdrawals. By 1980, total water-level decline exceeded 300 ft in much of the southeastern part of the basin (Graf, 1980).

Model Characteristics

The numerical model for Harquahala Plain included both steady-state and transient analyses and had the following characteristics:

- A variable-sized finite-difference grid was used. The model grid consists of 38 rows and 17 columns and includes 298 active cells in each of two layers.
- Two layers were used to represent (1) upper basin fill, which consists of a mixture of fine- and coarse-grained sediments at the southeast end and fine-grained sediments at the center of the basin, and (2) lower basin fill, which includes conglomerate and silt, sand, and gravel at the southeast end and fine-grained sediments in the basin center.
- Interconnection between the two layers was simulated by a model-calculated vertical-leakance value, which was calculated on the basis of the harmonic mean of the vertical hydraulic conductivities of the units divided by one-half the thickness of the units.
- No mountain-front recharge was simulated. For the steady-state simulation, only underflow into the basin from the northwest and underflow out of the basin at

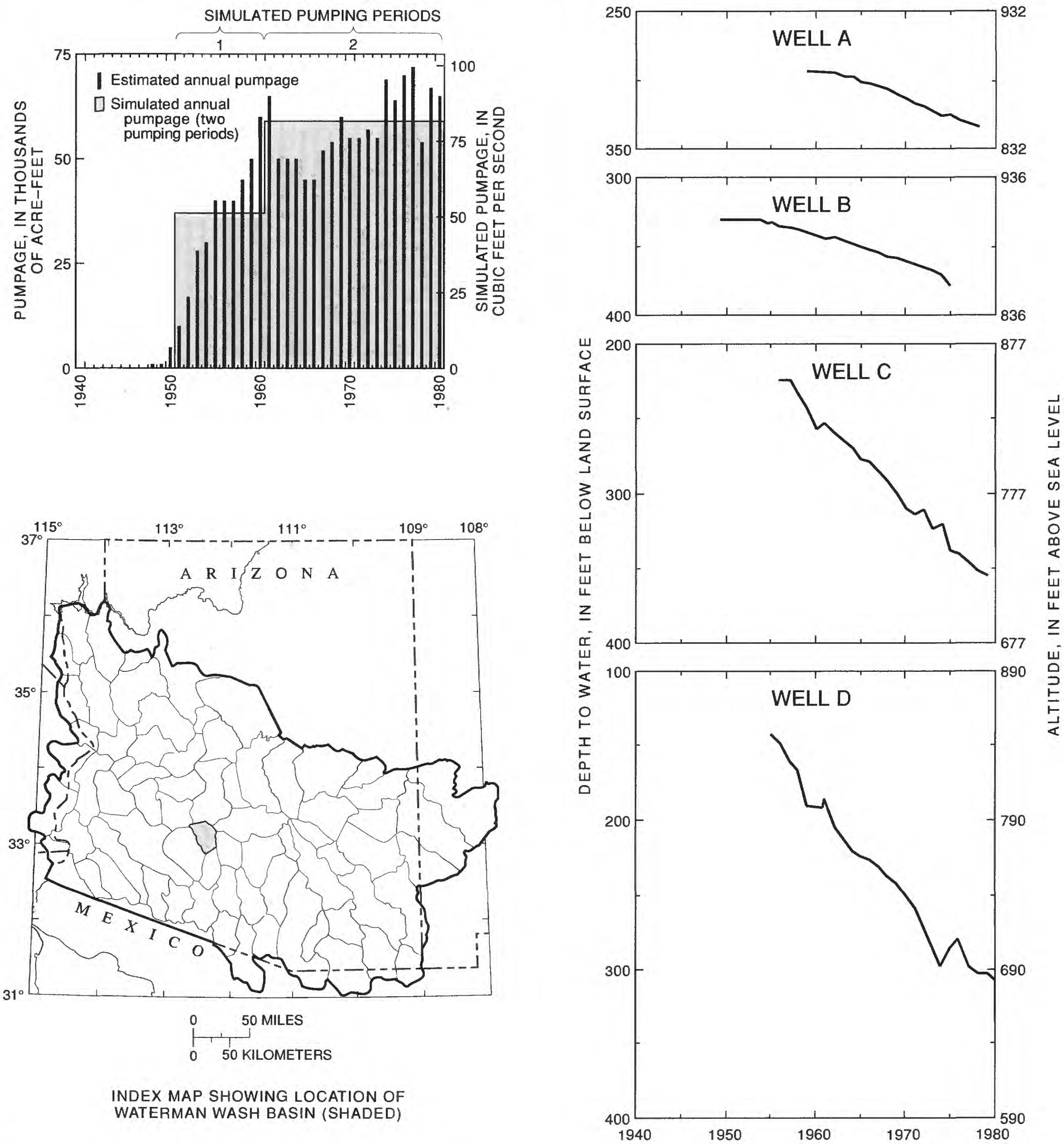
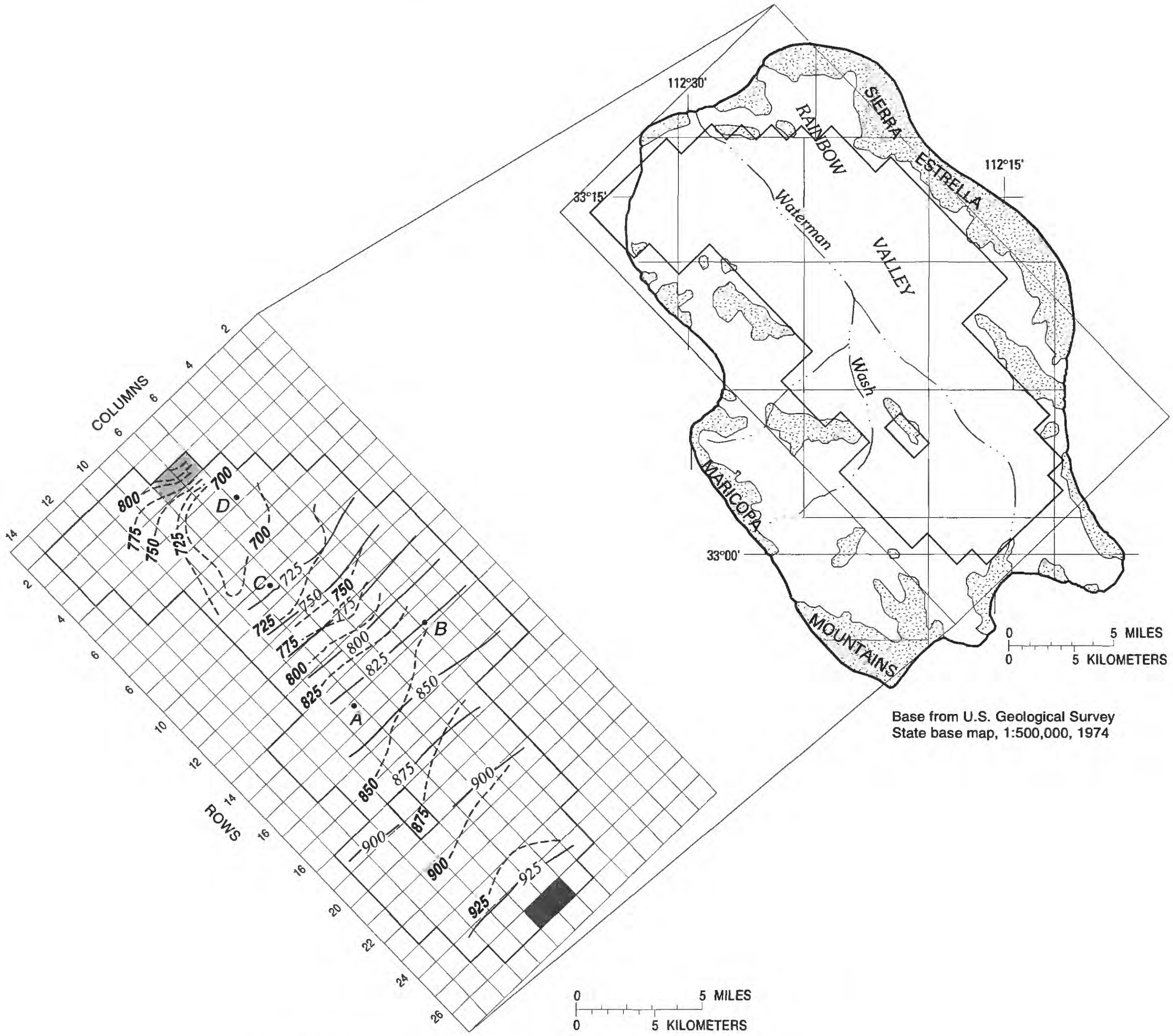


FIGURE 15.—Location of basin, estimated and simulated pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, Waterman Wash basin.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing
- Inflow
- Outflow

Geologic contact

Boundary of ground-water area

Boundary of modeled area—Upper and lower layers

Water-level contour, 1980-81—Shows altitude of water level. Contour interval 25 feet. Datum is sea level

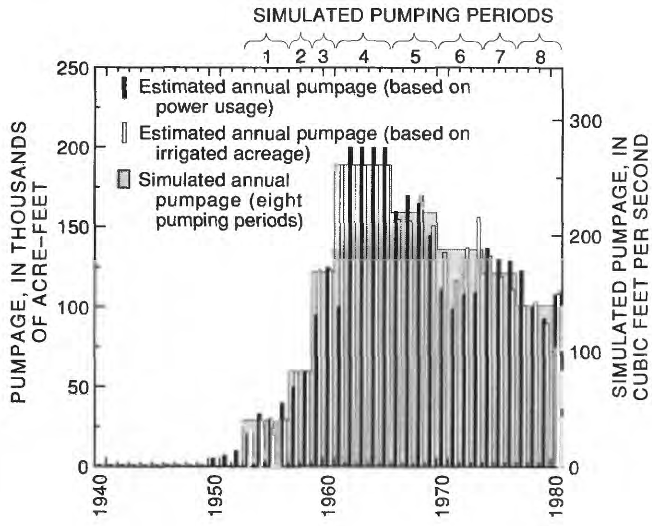
- - -750 - - - Observed (R.S. Stulik, Arizona Department of Water Resources, written commun., 1981)

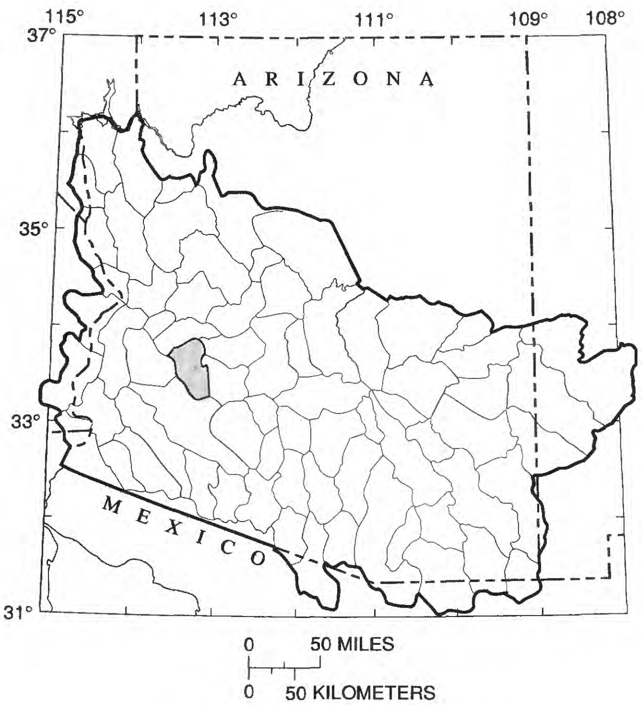
—750— Simulated

•B Well for which hydrograph is shown

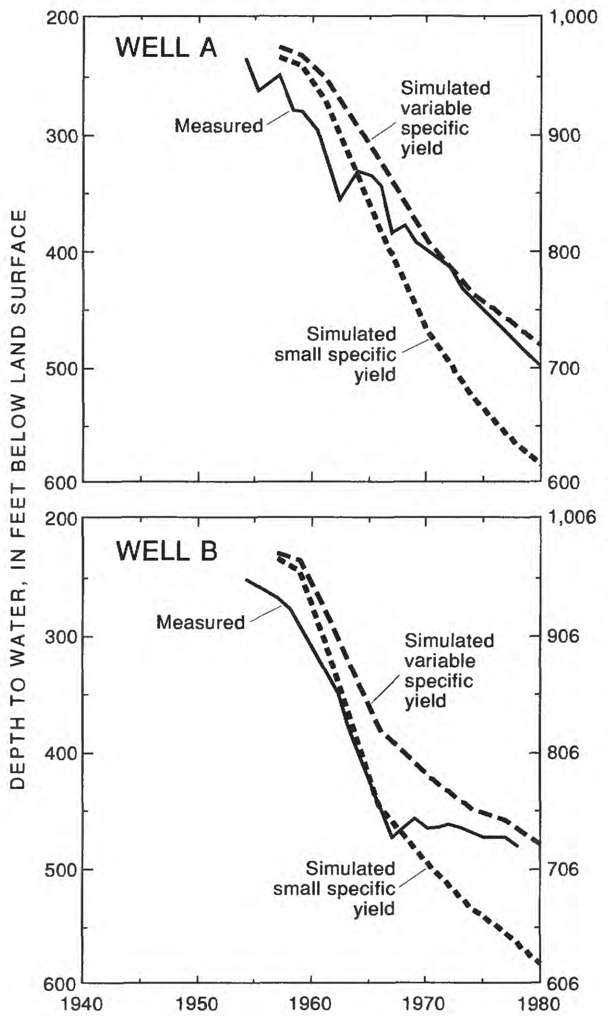
HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	8 - 20
Average	14.6
Transmissivity of lower layer, in feet squared per day	
Single uniform value	7,000
Specific yield of upper layer	0.117
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-4} - $4x10^{-5}$

FIGURE 15.—Continued.



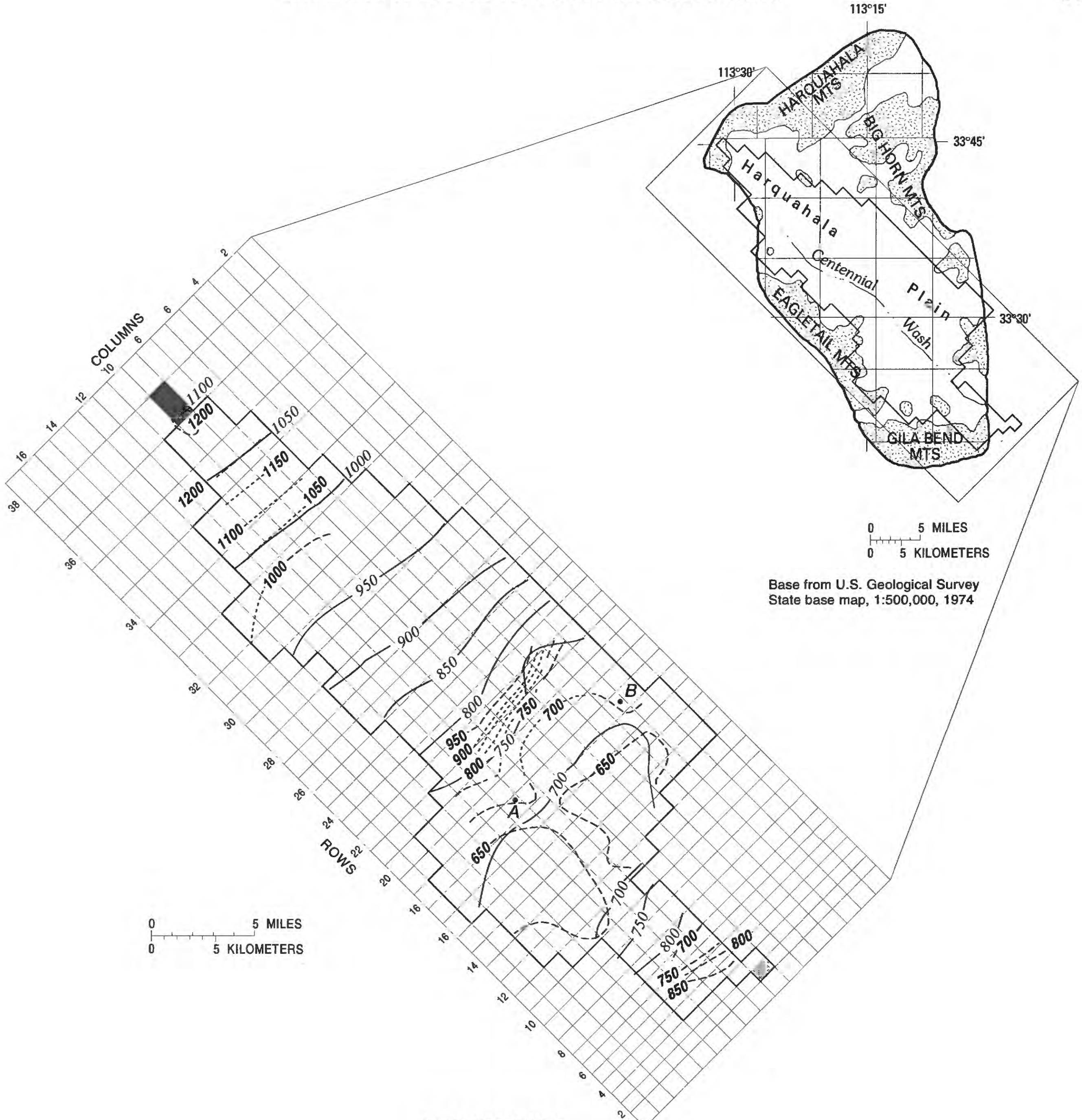


INDEX MAP SHOWING LOCATION OF HARQUAHALA PLAIN (SHADED)



HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	0.1 - 43
Average	2.2
Transmissivity of lower layer, in feet squared per day	
Range	1,000 - 41,000
Average	13,900
Specific yield of upper layer	0.05 - 0.15
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-6} - 10^{-7}

FIGURE 16.—Location of basin, estimated and simulated pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, Harquahala Plain.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing

Inflow

Outflow

Geologic contact

Boundary of ground-water area

Boundary of modeled area—Upper and lower layers

Water-level contour, 1980—Shows altitude of water level. Contour interval 20 and 50 feet. Datum is sea level

- - -750- - - - Observed (Graf, 1980)

——750—— Simulated

• *B* **Well for which hydrograph is shown**

FIGURE 16.—Continued.

the southeast end were included. These flows were simulated by use of constant-head cells.

- All simulated pumping was from the lower layer. The pumping history was simulated in eight pumping periods—1953–56, 1957–58, 1959–60, 1961–65, 1966–69, 1970–73, 1974–76, and 1977–80.

Results of Simulations

Few data are available on the magnitude and distribution of hydrologic properties of the sediments and predevelopment water-level conditions; thus, the developed steady-state model is of uncertain reliability. The steady-state simulation generally was insensitive to change in recharge location and quantity within a range thought to be practical for the basin. On the basis of the steady-state model results, natural recharge to and discharge from the basin were estimated to be about 1,200 acre-ft/yr.

Transient-model results indicate that the basin is a typical storage-depletion system in which pumpage is balanced almost entirely by removal of water from storage. The small rate of natural recharge had no discernible effect on the rate of water-level decline and the mining of ground water. Under these conditions, specific yield is a major controlling property in the simulation but is the poorest known. The effect of poorly known specific yield on model results was evaluated by making two transient-model runs using different distributions and values of specific yield. The values were assigned on the basis of extent of fine-grained sediments and were based on values determined by indirect means for other basins in the study area. The first model run used relatively small specific-yield values of 0.05 and 0.10, which corresponded to areas of fine-grained facies and to areas of coarser, heterogeneous sediments, respectively. The model response at two locations is shown as "small specific yield" on hydrographs for wells A and B in figure 16. A second model run used specific-yield values of 0.05, 0.10, 0.13, and 0.15 to represent various areas of decreased content of fine-grained sediments. The model response for this scenario is shown as "variable specific yield" on hydrographs for wells A and B in figure 16. The model-generated response indicates that use of the variable specific-yield values provided good comparisons of water-level-decline rates during 1950–65. The divergence of field and model data starting about 1965 cannot be explained by altering the specific yield and is thought to indicate recharge from excess applied irrigation water, although other factors may also be involved.

MCMULLEN VALLEY

Geohydrologic Conditions

McMullen Valley is a roughly rectangular, northeast-trending basin in west-central Arizona (fig. 17). The basin includes about 720 mi^2, of which about 25 percent is nearly impermeable consolidated rocks of the mountains, and the remaining 75 percent is permeable alluvial material. The climate is arid, and average annual precipitation is less than 10 in. (Sellers and Hill, 1974).

Basin fill, which is the major aquifer in the valley, consists of upper and lower units. The upper basin fill is unsaturated in most of the basin. The small part of upper basin fill that was saturated before development subsequently has been dewatered. In the central part of the basin, lower basin fill includes coarse-grained sediments overlain by as much as 1,000 ft of fine-grained sediments. The lithology of the basin was documented by Pool (1987, p. 14–18).

Ground-water movement through the basin was from northeast to southwest before development. Natural recharge, although small, occurs in the northern part of the basin and possibly at the base of the peripheral mountains as mountain-front recharge. Discharge occurred as underflow into Harquahala Plain and, in part, possibly as evapotranspiration.

Ground-water development altered the natural ground-water flow system after 1952. During 1954–59, 50 irrigation wells were completed in the valley (Kam, 1964). Irrigated area increased from 11,000 acres in 1958 to 16,600 acres in 1965 (Briggs, 1969). By 1980, irrigated acreage had increased to 34,200 acres (Remick, 1981). As shown in figure 17, pumpage increased from 5,000 acre-ft in 1952 to 139,000 acre-ft in 1975 and has decreased slightly since 1975 (U.S. Geological Survey, 1985). Maximum water-level declines of more than 250 and 150 ft occurred in the southwest and northeast parts, respectively, of the basin. Rates of decline ranged from 5 to as much as 15 ft/yr during the early 1970's (hydrograph for well A, fig. 17) in the principal cones of depression.

Model Characteristics

The numerical model for McMullen Valley included steady-state and transient analyses and were documented by Pool (1987). The steady-state analysis was used to evaluate areal distribution and magnitude of aquifer properties and predevelopment recharge and discharge. The transient simulation was used to analyze the relation of pumpage and water-level decline. The numerical model had the following characteristics:

- A finite-difference grid of variable cell size was designed to allow more detailed evaluation in areas of steep water-level gradients and known geohydrologic boundaries. The model grid consisted of 37 rows and 29 columns.

- Two layers were used to represent (1) upper and lower basin fill of which the lower basin fill is the principal aquifer in most of the basin and (2) pre-Basin and Range rocks. The layers simulate different areal extents because of the physical and hydrologic boundaries of the aquifer.
- Hydraulic connection between the two layers was simulated by a vertical-leakance value.
- Natural inflow and outflow were simulated as constant-head cells during steady-state calibration. The flow quantity from the constant-head cells simulated in the steady-state model was transferred to the transient simulation and held constant.
- Pumping was simulated in four periods—1951–57, 1958–65, 1966–72, and 1973–80.

Results of Simulations

Water levels in the steady-state model were matched with the pre-1952 water-level configuration (Kam, 1964), although definition of the predevelopment system was poor because of the lack of data. The transient analysis included the period 1952–80. The history-matching procedure resulted in a good comparison of simulated and field-measured water-level contours, with little change in the general distribution of hydrologic properties. Transient simulation required adjustments to vertical leakance, specific yield, and the location of an impermeable boundary as documented by Pool (1987), all of which were undefined or poorly defined by available field data. The northeastern part of the basin was simulated as a typical storage-depletion system. Stratigraphic complexities and the presence of buried consolidated rocks along the south boundary made simulation of the southwestern part of the system difficult. A fine-grained facies that inhibits vertical movement of ground water also affected the size and shape of the simulated cone of depression. Simulations indicate that development in McMullen Valley has affected water levels throughout the valley. In some areas, water-level declines have been accelerated because the cones of depression have intersected impermeable boundaries. Model results also indicated that vertical leakance increases in the northwestern part of the valley; the fine-grained facies may be absent in this part.

PARKER VALLEY

Geohydrologic Conditions

Parker Valley includes about 230 mi^2 of the Colorado River flood plain in western Arizona and is typical of the basins in the Colorado River group (fig. 7). The climate is arid, and average annual precipitation on the flood plain is less than 4 in. (Hely and Peck, 1964, pl. 3).

In Parker Valley, the sediments include the Miocene(?) fanglomerate (Metzger and others, 1973); the overlying Bouse Formation, which is equivalent to upper basin fill; older alluvial deposits, which are almost completely eroded; and stream alluvium, which makes up the flood plain of the Colorado River and fills channels cut into the Bouse Formation and older alluvium (Metzger and others, 1973; Tucci, 1982). The stream alluvium and at least the upper part of the Bouse Formation are connected hydraulically and constitute the principal aquifer in Parker Valley. Ground water occurs under unconfined conditions in the aquifer.

The hydrologic system of Parker Valley is dominated by the Colorado River, which controls the quantity of inflow to and outflow from the aquifer adjacent to the river. In response to extensive agricultural development, diversion and application of surface water to irrigated cropland have resulted in shallower depths to ground water because of recharge from excess applied irrigation water. Discharge from the ground-water system occurs as evapotranspiration from crops and riparian vegetation, as leakage to drains used to prevent waterlogging, and as discharge to the Colorado River.

Model Characteristics

The numerical model for Parker Valley was used to evaluate present knowledge and concepts of the ground-water system and to simulate the areal system (Tucci, 1982). Model results were initially assumed to be transferable to other similar basins along the lower Colorado River; this assumption was proved to be only partly valid on the basis of a model of the Mohave basin, which is about 30 mi north of Parker Valley. The numerical model for Parker Valley had the following characteristics:

- Variable spacing was used in the finite-difference grid to allow increased resolution in simulation of the extent and alignment of the Colorado River.
- Two layers were used to represent (1) stream alluvium and (2) the upper zone of the Bouse Formation (Metzger, 1968).
- Flow between layers was controlled by a leakance value to provide a vertical connection and was calculated by the model on the basis of the vertical hydraulic conductivity and thickness of the model layers.

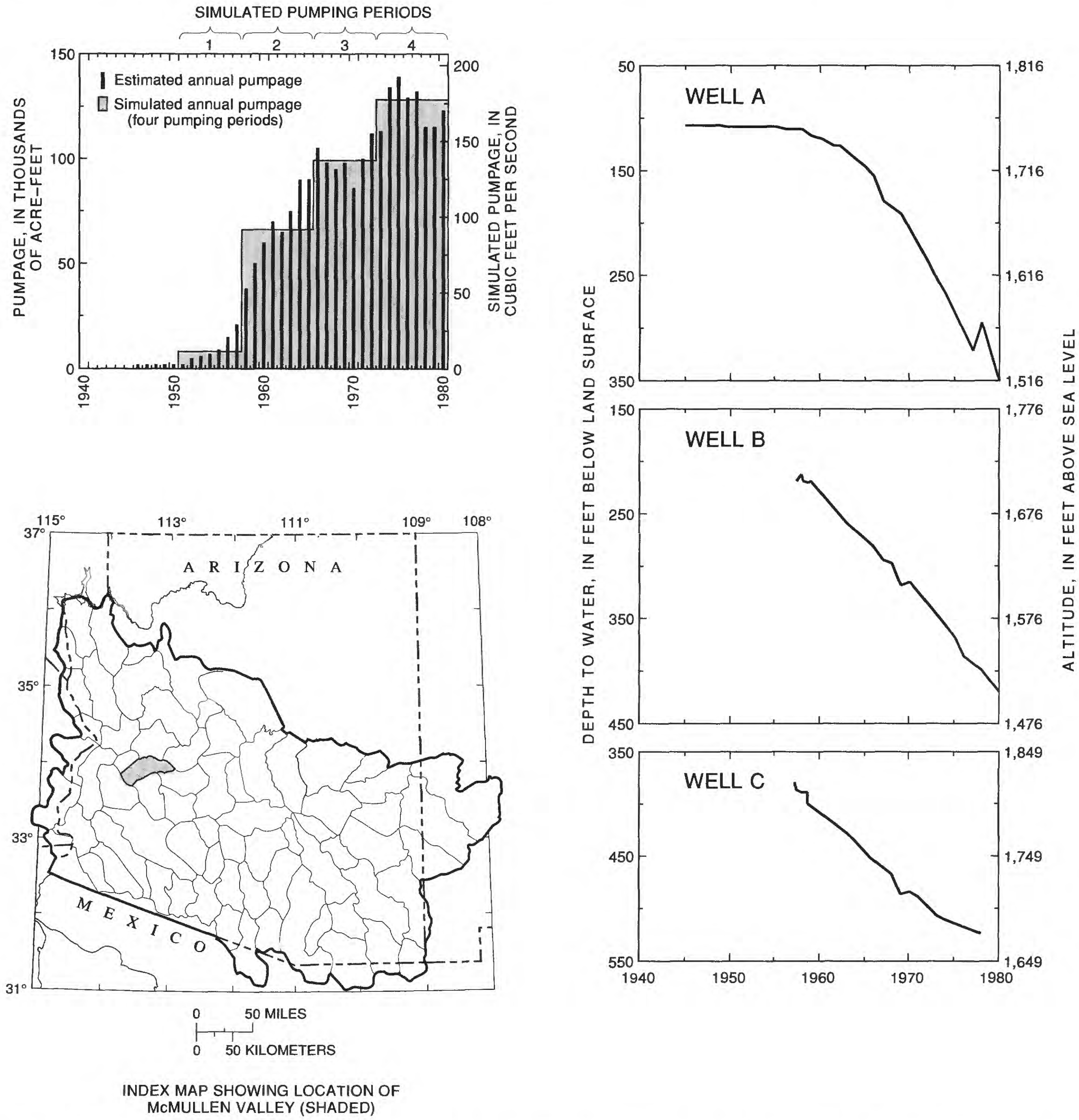
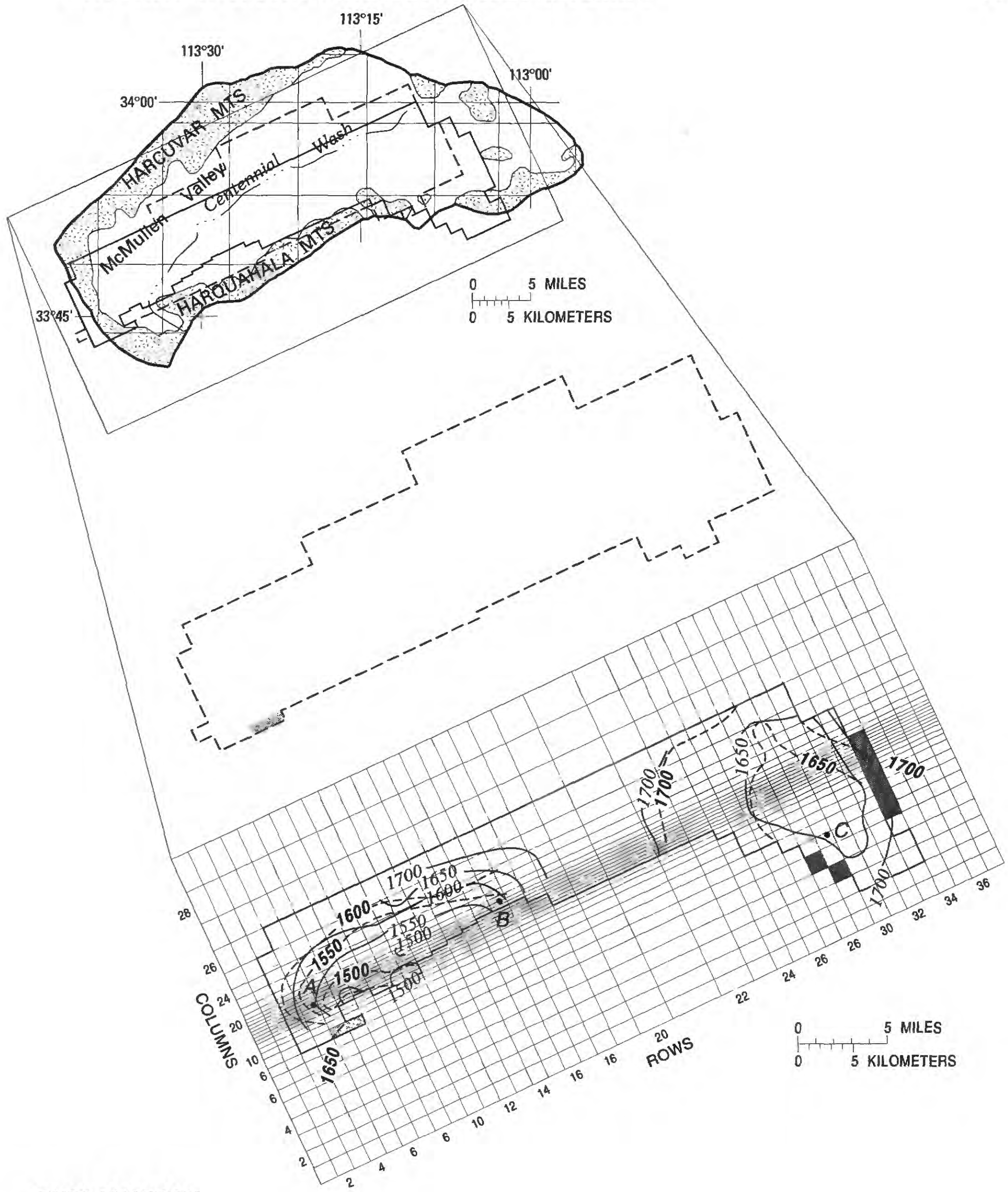


FIGURE 17.—Location of basin, estimated and simulated pumpage, ground-water levels, model grid and boundary, and observed and simulated water levels, McMullen Valley.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

2 4
2 Grid showing model cells used in simulating ground-water flow

Cell representing

Inflow

Outflow

Geologic contact

Boundary of ground-water area

Boundary of modeled area—

Lower layer

Upper layer

Water-level contour, December 1980 (Pool, 1987)—Shows altitude of water level. Contour interval 50 feet. Datum is sea level

--1500-- Observed

—1500— Simulated

•B **Well for which hydrograph is shown**

HYDRAULIC CHARACTERISTICS USED IN MODEL	
CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	1 - 20
Average	3.5
Transmissivity of lower layer, in feet squared per day	
Range	100 - 30,000
Average	6,450
Specific yield of upper layer	0.10
Storage coefficient of lower layer	10^{-4}
Vertical leakance between layers, in feet per second per foot	
Range	10^{-6} - 10^{-11}

FIGURE 17.—Continued.

- Uniform hydraulic conductivity was used throughout the upper layer.
- Uniform transmissivity was used throughout the lower layer.
- River-aquifer connection was simulated in the upper layer as leakage through a confining layer. Leakage to drains was similarly simulated (Tucci, 1982).
- Evapotranspiration by native vegetation was simulated in the upper layer, and the maximum effective depth of evapotranspiration was simulated as 20 ft.

The finite-difference grid consisted of 30 rows and 52 columns (fig. 18). A row of constant-head cells was used at the downgradient end of the model, and constant-flow cells were used at areas of tributary underflow into the basin. River and drain cells were treated similarly by using an altitude for the water level in the river or drain and a leakance value that controls the interconnection of the surface-water and ground-water systems.

Results of Simulations

Steady-state model simulations were made for ground-water conditions in 1940–41, the mid-1960's, and 1980 (Tucci, 1982, p. 38). In 1940–41, the Colorado River was a losing stream throughout Parker Valley and represented the major source of recharge to the aquifer; evapotranspiration by native vegetation on the flood plain was the dominant discharge mechanism. Model results for conditions in the mid-1960's indicate that the Colorado River was a gaining stream in the northern part of the valley as a result of ground-water-level rises caused by infiltration of applied irrigation water. Recharge to the aquifer from infiltration of surface water was decreased as a result of (1) shallower ground-water levels and (2) smaller gradients away from the river along reaches where it was still a losing stream. Model results for conditions in 1980 indicated a further decrease in surface-water infiltration associated with continued ground-water level rises.

Model results generally compared favorably with available field data. The model was most sensitive to changes in the average evapotranspiration rate. The model was less sensitive to changes in river- and drain-leakance values. The hydrologic system in Parker Valley can be adequately simulated by a simple two-dimensional approach because of the magnitude and dominance of the near-surface hydrologic interactions.

MOHAVE BASIN

Geohydrologic Conditions

Mohave basin is along the flood plain of the Colorado River in western Arizona (fig. 7). The modeled area includes 103 mi^2 near the center of the basin (fig. 19). The climate is arid, and average annual precipitation is less than 6 in. (Hely and Peck, 1964, pl. 3).

The geohydrologic setting of Mohave basin is similar to that of Parker Valley and other basins transected by the lower Colorado River. The aquifer is composed of stream alluvium of the Colorado River and upper basin fill, which includes older alluvium and the underlying upper zone of the Bouse Formation (Metzger, 1968; Metzger and Loeltz, 1973). As in Parker Valley, the stream alluvium occupies channels cut into the older alluvium and the Bouse Formation. The older alluvium occurs only laterally adjacent to the stream alluvium at the edge of the present flood plain of the Colorado River. The ground-water system of Mohave basin is dominated by infiltration of surface water from the Colorado River, which is the principal source of recharge. Evapotranspiration by crops and riparian vegetation is the principal discharge.

Model Characteristics

The numerical model for Mohave basin was developed to test the transferability of modeling concepts and hydrologic properties from the model of Parker Valley. The model results were matched to steady-state conditions in the 1960's and had the following characteristics:

- Variable spacing was used in the finite-difference grid to provide additional resolution in simulation of the extent and alignment of the Colorado River.
- Two layers were used to represent (1) stream alluvium and the adjacent older alluvium and (2) the upper zone of the Bouse Formation (Metzger, 1968).
- Flow between layers was controlled by a leakance value to provide a vertical connection and was calculated by the model on the basis of vertical hydraulic conductivity and thickness of the model layers.
- Uniform hydraulic conductivity was used throughout the upper layer.
- Uniform transmissivity was used throughout the lower layer.
- The river-aquifer connection was simulated in the upper layer as leakage through a confining layer as in the model for Parker Valley.
- Constant-head cells were used to simulate ground-water flow along the southeast edge of the model and to simulate the open-water surface of Topock Marsh adjacent to the Colorado River in the southern part of the area (fig. 19).

- Evapotranspiration by native vegetation was simulated in the upper layer. The maximum effective depth of evapotranspiration was simulated as 20 ft.

The finite-difference grid consisted of 52 rows and 23 columns (fig. 19). The outer boundary of the model consisted of no-flow cells except in the southeast corner of the model, where tributary ground-water inflow along Sacramento Wash was simulated by constant-head cells. As an initial approach to model development, hydraulic conductivity of the upper layer, transmissivity of the lower layer, river leakance, and the evapotranspiration and depth-to-water relation were transferred directly from the calibrated model for Parker Valley.

Results of Simulations

Transfer of hydrologic-property values from the model of Parker Valley to the model of Mohave basin was only partially successful. Although the geohydrologic settings are similar in both basins, sufficient differences exist that make direct transfer of some property values inappropriate. The most significant difference was the need to decrease hydraulic conductivity of the upper unit where the older alluvium occurred laterally adjacent to the stream alluvium. The older alluvium is considered to be much finer grained than the stream alluvium, on the basis of field observations and analysis of drillers' logs. The hydraulic conductivity was decreased from 311 to 6 ft/d in areas underlain by the older alluvium outside the flood plain of the Colorado River; this decrease resulted in much improved simulated water-level gradients. Hydraulic conductivity of the stream alluvium was maintained at 311 ft/d as used in the Parker Valley model. Also, transmissivity of the lower model unit—equivalent to the upper part of the Bouse Formation—was decreased by about one-third to adjust for the unit being thinner and having a higher silt and clay content than in Parker Valley. The only other modification to model properties was in the relation between evapotranspiration and depth to water. On the basis of known differences in type and density of riparian vegetation, the maximum evapotranspiration rate was decreased about 30 percent.

A water budget and a map of head distribution were used as means of evaluating the degree to which the model simulated the actual hydrologic system. Individual water-budget components based on field estimates were compared with model-calculated components (fig. 19). The small imbalance in the budget is because of intrinsic inaccuracies in estimating some of the individual budget values. Within the flood plain, where water levels were best known, model-calculated water levels were within 10 ft of measured levels. Information transfer from a model of a hydrologically similar basin provided initial estimates of aquifer properties and thus decreased the time necessary to calibrate the numerical model. The entire process served to prove the adaptability of information transfer—developing the conceptual model and the numerical model—to basins of this group.

BASIN-GROUP MODELS

Two basin-group models—one representing the southeast basins and the other representing the central and west basins—were used to explore model sensitivity to use of generalized values of geohydrologic properties (table 1). Models that used average values were compared with models that used various alternative combinations of boundary, hydrologic, and flow properties.

Because of the ready transferability of property values from the Parker Valley model to the Mohave basin model, no further basin-group modeling of the Colorado River group was undertaken. In addition, owing to the paucity of data, no models were developed for the basins of the highland group. These basins are thought to have distinct similarities to the Colorado River group in which coarse, unconsolidated sediments make up the shallow alluvial aquifer, and the hydrology is dominated by the presence of perennial surface flow. Significant differences do exist, however, in size of the basins and climatic characteristics.

SOUTHEAST BASINS

Geohydrologic Conditions Represented and Model Characteristics

The basin-group model of the southeast basins consists of two layers. The upper layer represents an unconfined aquifer 5 mi wide and extending the length of the basin. The lower layer represents the underlying basin-fill aquifer and extends the entire length and width of the 15- by 50-mi basin (fig. 20). Water in this layer is simulated as occurring under confined conditions in the part underlying the upper layer and under unconfined conditions in the rest of the aquifer. The model was developed by using a uniform finite-difference grid that has a cell size of 1 mi^2, includes 17 rows and 52 columns, and represents the average physical size of the basins in the southeast group. The upper layer was represented by 250 active cells, and the lower layer by 690 active cells. Transmissivity values for the lower layer ranged from 300 ft^2/d at the outer edge of the simulated area to 7,000 ft^2/d near the center. Transmissivity of the upper layer was initially 4,000 ft^2/d. Flow components that are typical of the southeast basins were included; initial model data and water-level configuration are shown in figure 21.

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS
[Values are in acre-feet per year. Modified from Tucci, 1982, table 1]

Components	1940–41 Estimated	1940–41 Simulated	Mid-1960's Estimated	Mid-1960's Simulated	1980 Estimated	1980 Simulated
Inflow:						
Recharge	12,000	12,000	225,000	223,000	236,000	235,000
River losses	213,000	253,000	92,000	97,000	50,000	82,000
Tributary inflow	12,000	12,000	12,000	12,000	12,000	12,000
Total	237,000	277,000	329,000	332,000	298,000	329,000
Outflow:						
Evapotranspiration	235,000	271,000	163,000	161,000	71,000	120,000
Outflow to Palo Verde Valley	2,000	1,000	3,000	3,000	3,000	2,000
Drain base flow	----------	----------	172,000	168,000	224,000	206,000
Total	237,000	272,000	338,000	332,000	298,000	328,000

HYDRAULIC CHARACTERISTICS USED IN MODEL

CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day Single uniform value	311
Transmissivity of lower layer, in feet squared per day Single uniform value	8,980
Vertical leakance between layers, in feet per second per foot Range	$5x10^{-6}$ - $7x10^{-6}$

INDEX MAP SHOWING LOCATION OF PARKER VALLEY (SHADED)

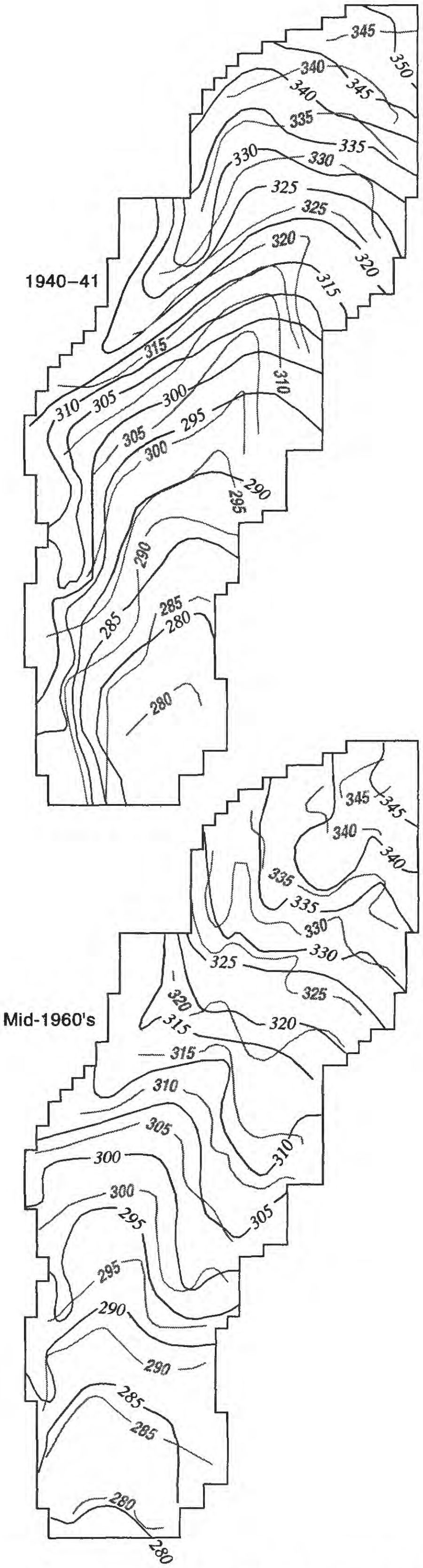
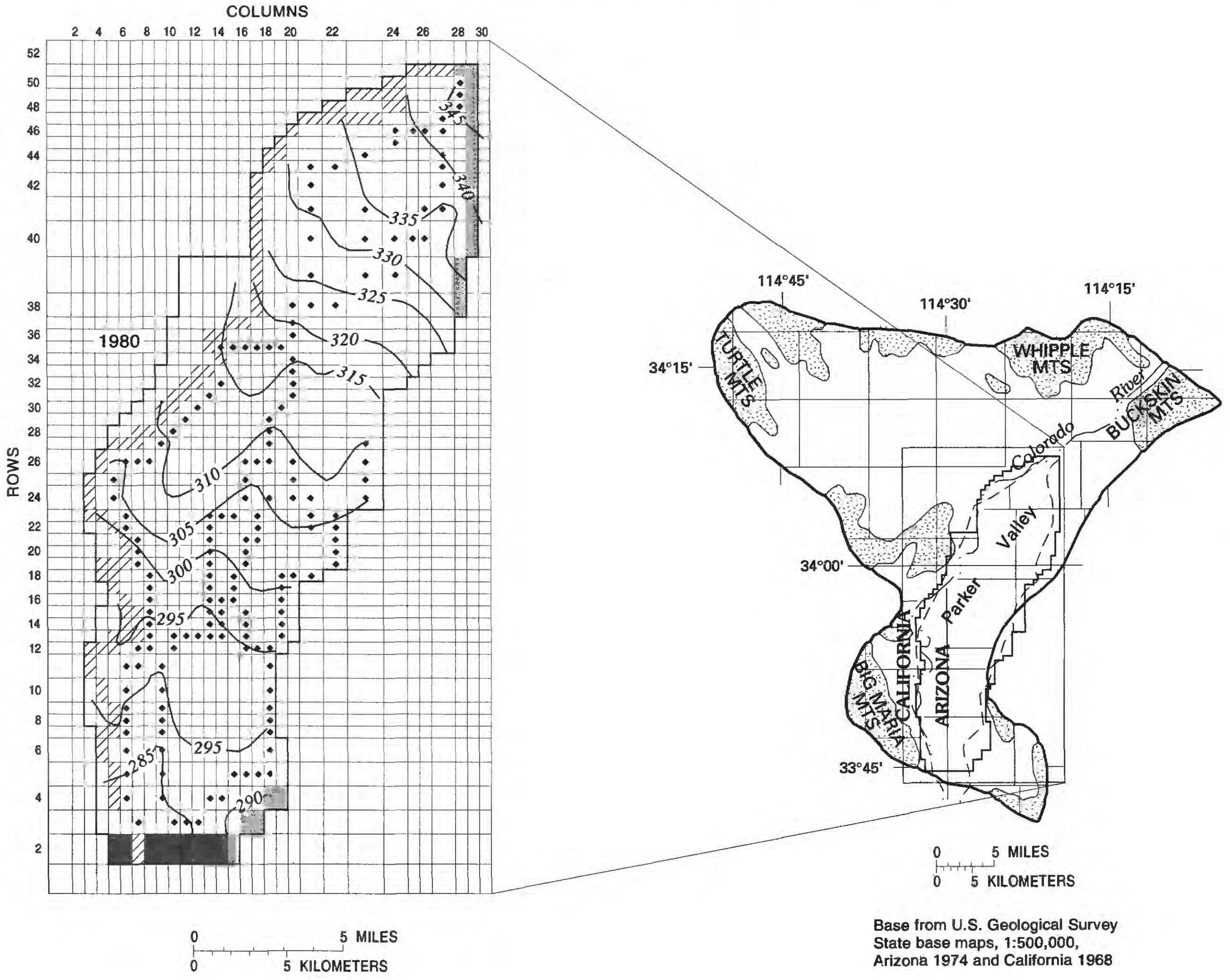


FIGURE 18.—Location of basin, estimated and simulated water-budget components, model grid and boundary, and observed and simulated water levels, Parker Valley.



EXPLANATION

Alluvial deposits

Bedrock of the mountains

Grid showing model cells used in simulating ground-water flow

Cell representing
- Constant head—inflow
- Constant head—outflow
- Drain (mid-1960's and1980 simulation)
- River-aquifer connection

Geologic contact

Boundary of flood plain

Boundary of ground-water area

Boundary of modeled area— Upper and lower layers

Water-level contour, 1940–41, mid-1960's, and 1980 (Tucci, 1982)— Shows altitude of water level. Contour interval 5 and 10 feet. Datum is sea level

295 Observed

295 Simulated

FIGURE 18.—Continued.

ESTIMATED AND SIMULATED WATER-BUDGET COMPONENTS
[Values are in acre-feet per year. Estimated values modified from Metzger and Loeltz (1973)].

Components	1960's Estimated	1960's Simulated
Inflow:		
River losses	150,000	155,000
Tributary inflow and unmeasured runoff	5,000	8,000
Total	155,000	163,000
Outflow:		
Evapotranspiration	147,000	150,000
Pumpage	11,000	11,000
Total	158,000	161,000

INDEX MAP SHOWING LOCATION OF MOHAVE BASIN (SHADED)

HYDRAULIC CHARACTERISTICS USED IN MODEL

CHARACTERISTIC	VALUE
Hydraulic conductivity of upper layer, in feet per day	
Range	6 - 311
Average	120
Transmissivity of lower layer, in feet squared per day	
Single uniform value	2,590
Vertical leakance between layers, in feet per second per foot	
Single uniform value	$6.8x10^{-6}$

FIGURE 19.—Location of basin, estimated and simulated water-budget components, model grid and boundary, and observed and simulated water levels, Mohave basin.

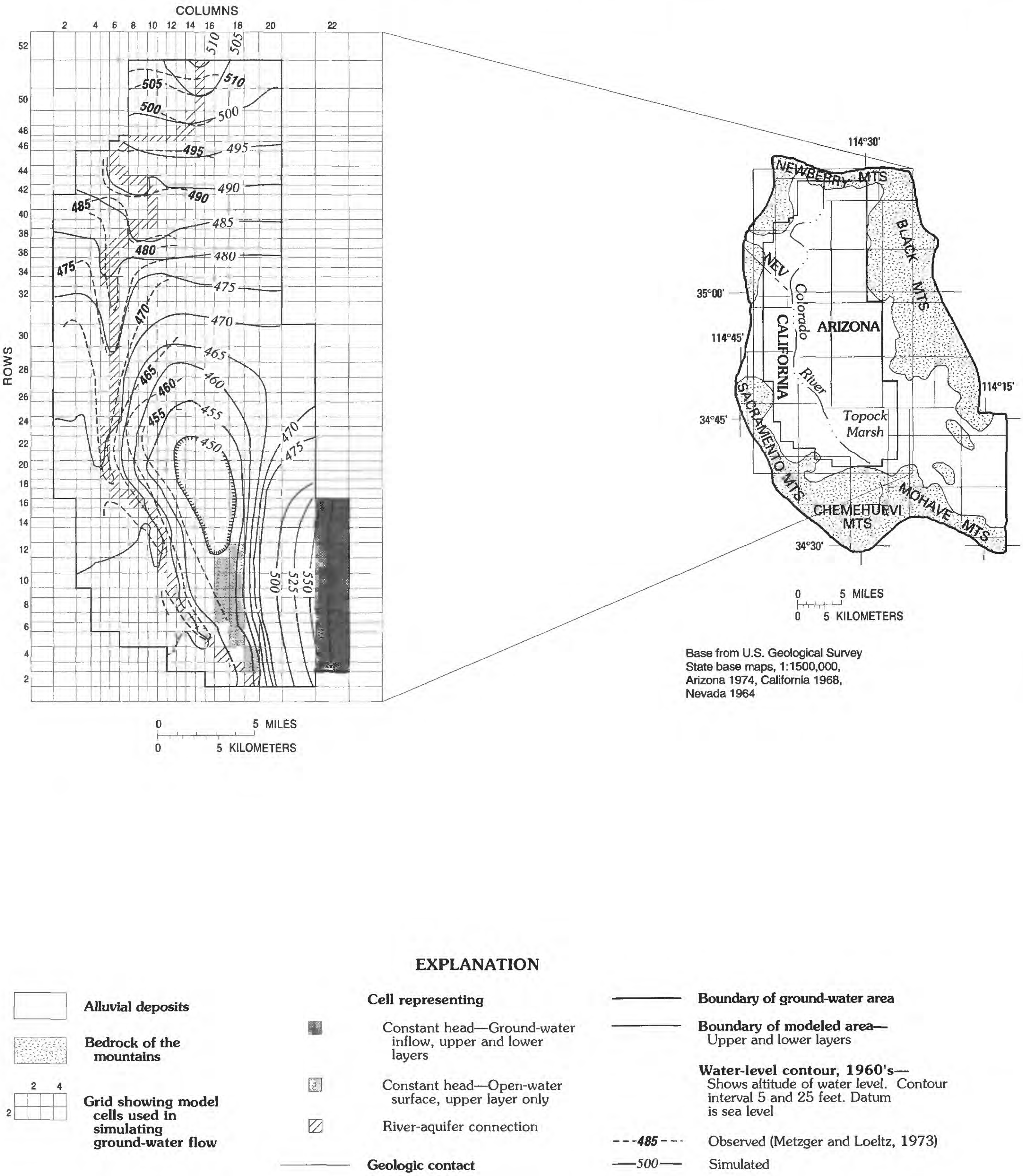


FIGURE 19.—Continued.

TABLE 1.—*Range in values of geohydrologic properties in southeast basins and central and west basins and values used in basin-group models*

[— indicates no data]

Basin characteristics	Southeast basins		Central and west basins	
	Maximum and minimum values for hydrologic property or basin characteristics	Basin-group model	Maximum and minimum values for hydrologic property or basin characteristics	Basin-group model
Basin-fill width, in miles	7–21	15	7–27	12
Basin-fill length, in miles	30–80	50	17–68	30
Length-to-width ratio	1.5–8.0	3.33	1.0–8.5	2.5
Axial water-level gradient, in feet per mile	3–45	15	1–18	3
Mountain-front recharge, in acre-feet per year	5,000–50,000	22,600	50–5,000	1,000–2,000
Saturated thickness, in feet	100–1,000	200–1,000	100–1,000	200–700
Hydraulic conductivity (*K*), in feet per day	1–50	20	3–40	9.5
Transmissivity (*T*), in feet squared per day	100–50,000	300–7,000	2,000–15,000	2,000–6,600
Specific yield(s)	.03–.25	.20	.05–.18	.12
Vertical-leakance rate (*TK*), in feet per second per foot	10^{-4}–10^{-11}	10^{-5}–10^{-11}	10^{-4}–10^{-11}	10^{-8}–10^{-11}
Riverbed-leakance rate (*RC*), in feet per second per foot	$\frac{1}{100} \times K$	.2	—	—
Maximum evapotranspiration rate (*QET*), in feet per year at 100-percent density	3–10	6	—	—
Effective depth at which evapotranspiration ceases, in feet	0–50	10	—	—
Vegetation density, in percent	20–80	50	—	—
Cell coverage by phreatophytes, in percent	10–90	50	—	—
Average annual pumpage, 1960–80, in acre-feet per year	4,000–250,000	50,000–260,000	4,000–138,000	63,700

RESULTS OF SIMULATIONS

The basin-group model of the southeast basins was used to investigate the relation of water-level contour shape to the controlling properties. Water-level contours in the southeast basins typically are U-shaped, are concave opposite the direction of ground-water flow, and qualitatively indicate the quantity of recharge entering a basin at the mountain fronts. The concave shape ranges from a broad U-shape in Douglas and Willcox basins to a V-shape in the lower San Pedro basin. Initial property values used in the basin-group model of the southeast basins were modified systematically to determine which properties exerted the greatest influence on water-level configuration. Comparison of water-level profiles at a particular section across the model (fig. 22) illustrates the effect of changing basin shape, transmissivity distribution, vertical leakance between layers, or increased recharge.

Changing the transmissivity distribution had the greatest effect on the shape of the simulated water-level profile, and altering the vertical-leakance value had the greatest effect on the computed head value. The model results indicate that areas within a basin where water-level gradients are steep—generally near mountain fronts—have either hydraulic conductivities or saturated thicknesses that are on the order of 10 times smaller than areas where water-level gradients are gentle. Shallow depth to a pedimentlike surface is the most likely cause for small saturated thickness; hydraulic conductivity may be small because of the presence of consolidated rocks indigenous to the nearby mountains or indurated pre-Basin and Range rocks.

CENTRAL AND WEST BASINS

GEOHYDROLOGIC CONDITIONS REPRESENTED AND MODEL CHARACTERISTICS

The central and west basins, which are hydrologically similar, were analyzed by use of a single basin-group model. The principal differences in basins of the two groups are size, depth, and vertical extent of the fine-grained facies. Also, because of larger and more significant inflow and outflow components in the central basins than in the west basins, the central-basin system response to development is delayed slightly as a result of capture of natural discharge. Subsequent to total capture of this flow, the response of the systems generally is the same.

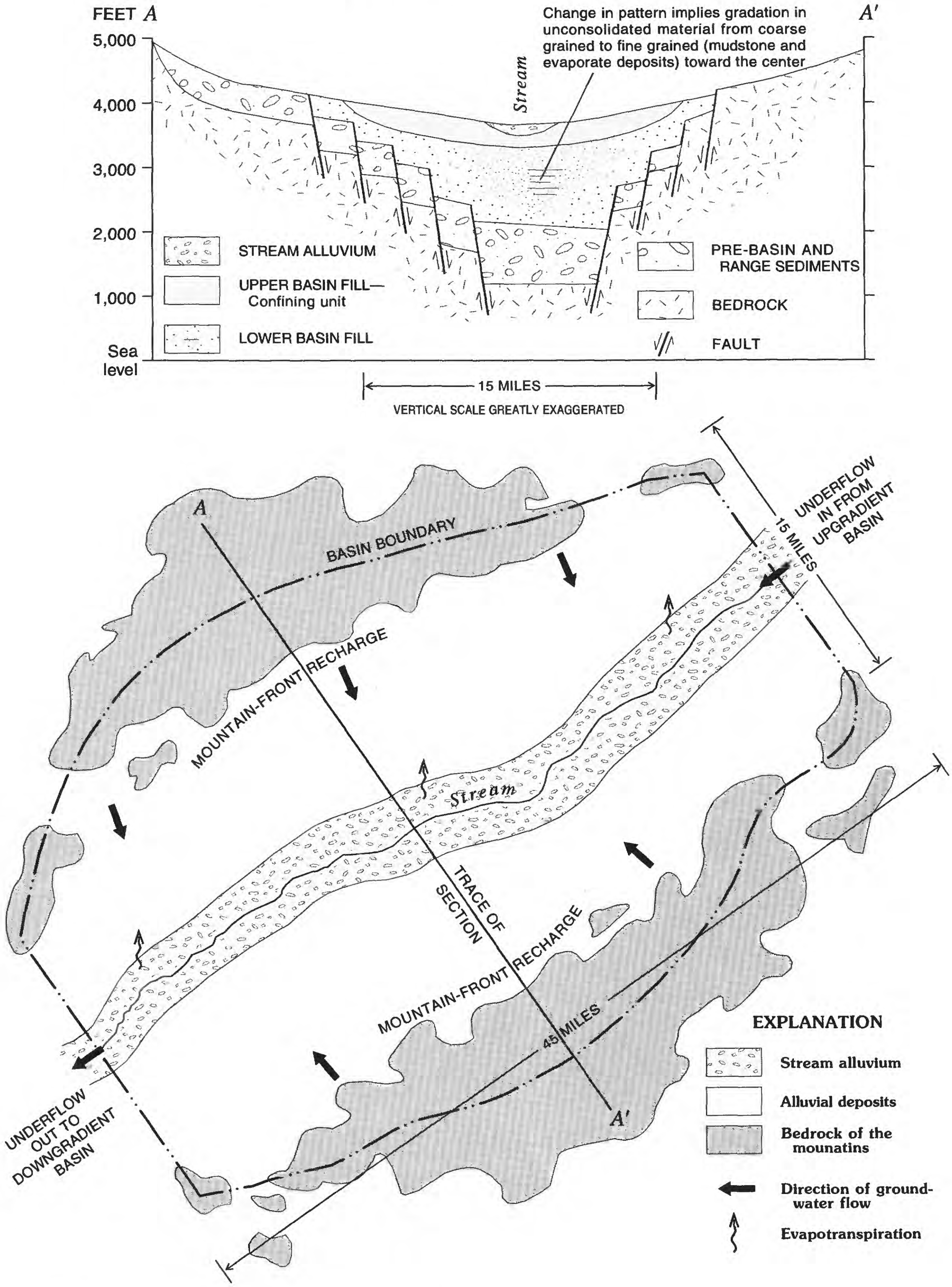
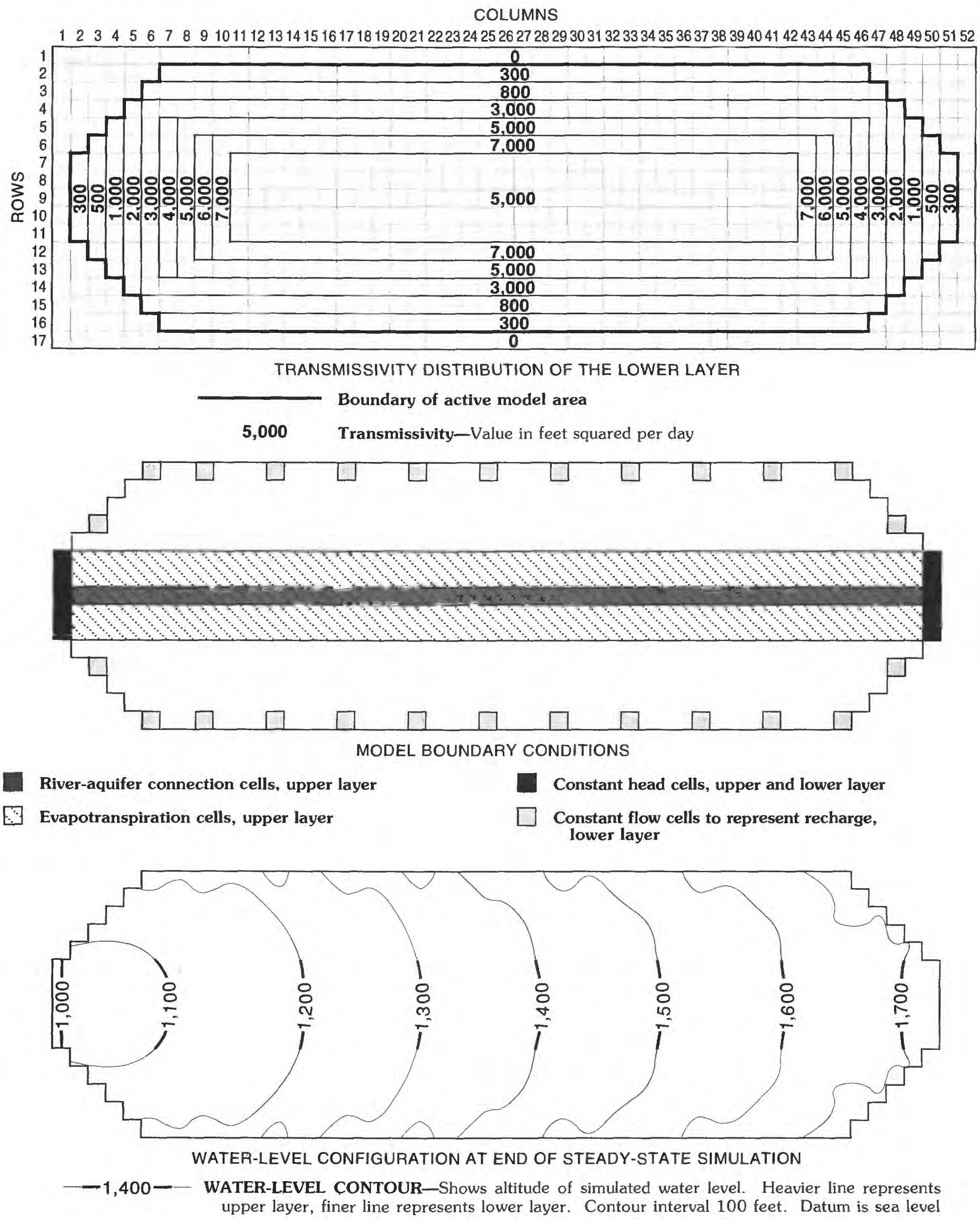


FIGURE 20.—Typical geographic and geohydrologic features in the southeast basins.



MODEL-INPUT CHARACTERISTICS	
Maximum evapotranspiration rate *(QET)*	1.5 x 10^{-8} feet per second
Depth at which evapotranspiration effectively ceases *(ETDIST)*	10 feet
Interlayer leakance *(TK)*	10^{-7} feet per second per foot
Riverbed leakance *(RC)*	10^{-8} feet per second per foot
Hydraulic conductivity of the upper layer *(K)*	20 feet per day
Saturated thickness of the upper layer *(STRT-BOTTOM)*	200 feet
Grid cell dimensions *(ΔX and ΔY)*	5,280 feet x 5,280 feet

FIGURE 21.—Model grid, boundary conditions, and data input for steady-state basin-group model of the southeast basins.

The basin-group model that represents the central and west basins includes one layer for those analyses that involve basin geometry and two layers for analyses that involve pumping stress-related reactions. The model, which was developed to represent a 12- by 30-mi area, had a uniform finite-difference grid having 360 active cells and included 14 rows and 32 columns (fig. 23). The single-layer simulation was developed to analyze the character of saturated lower-basin fill under unconfined conditions. A two-layer simulation represented saturated upper and lower basin fill separated by fine-grained sediments near the center of the basin. Evapotranspiration and river-aquifer exchange were not simulated because these components are negligible in the central basins and are not typical of the west basins. Flow components for the central and west basins can include mountain-front recharge, underflow into and out of the basin, and streamflow infiltration along the major basin drainage. The basin-group model representing these two groups was developed by using only the flow components of mountain-front recharge and underflow out of the basin because that configuration is most typical (fig. 24).

Results of Simulations

The basin-group model that represents the central and west basins was used to analyze the significance of basin shape and recharge location on water-level gradients. Ranges and location of recharge are the principal hydrologic differences between the central and west basins. In the west basins, recharge is small—on the order of a few hundred to about 5,000 acre-ft/yr. In the central basins, recharge is large—generally ranging from about 5,000 to about 50,000 acre-ft/yr—and occurs as a combination of underflow and mountain-front recharge.

Four basin shapes were considered—(1) a straight-sided channel with open ends, (2) a slope-sided trough with open ends, (3) a scoop with one end constricted and one end open, and (4) a long bowl with both ends constricted (fig. 25). Initial flow through each basin shape was established by constant-head cells at each end of the model grid and an initial water-level gradient of 3 ft/mi. Model-calculated flow through each basin was then applied as a uniformly distributed mountain-front recharge and as an intermittently spaced recharge around the basin perimeter. Results indicate that the water levels are practically insensitive to basin shape. Recharge location causes a significant change in water-level configuration. Where recharge is distributed around the margin of the basin rather than as underflow at the upgradient end, water-level gradients change from uniform to variable throughout the basin. Differences in water levels resulting either from uniformly distributed recharge or intermittently spaced recharge were negligible. Intermittently spaced recharge depicted in the model probably is more realistic and simulates recharge at intermittently spaced points where streams emerge from the mountains.

RESULTS FROM SPECIFIC-BASIN AND BASIN-GROUP SENSITIVITY ANALYSIS

Numerical models were developed for 12 of the 72 alluvial basins in south-central Arizona and for 2 idealized alluvial basins representing a typical southeast basin and a typical central-west basin. The models were analyzed to determine (1) the relation between the basin category and the importance of various model-input parameters and (2) the adequacy and transferability of information used to develop basin models.

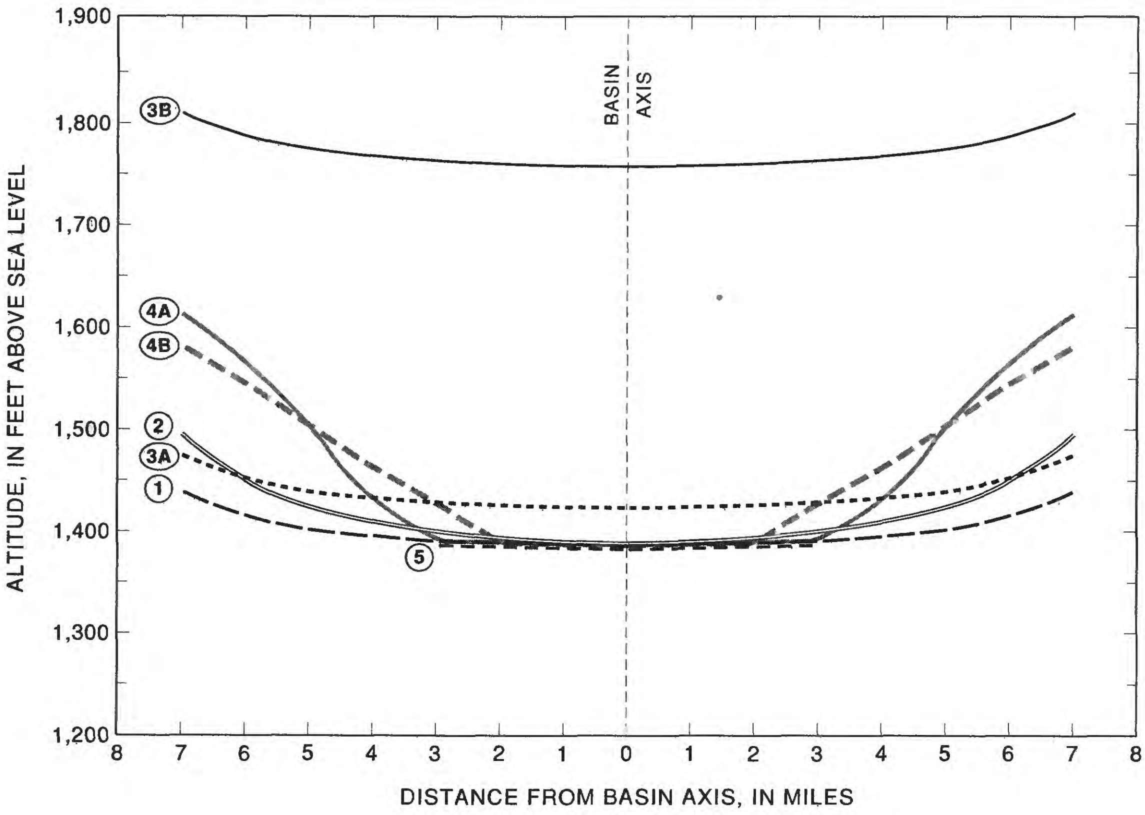
RELATION OF BASIN CATEGORY AND MODEL INPUT

A controlled examination of model response to changes in property values can be used to show the significance of each property and the data that define that property. The significance of a particular property depends on the hydrologic and geologic characteristics of a basin. The basins were grouped on the basis of these characteristics.

If large changes in simulation results take place with only minor changes in a tested property, the model reliability is highly dependent on the correct value for that property. In contrast, if large changes in property values result in insignificant changes in simulation results, the model is insensitive to that property, and an estimate of the actual value will suffice for use in the model. Knowledge of the model sensitivity to changes in each property may allow streamlining and redirection of data-collection activities. Also, the validity of transferring hydrologic boundaries, flow components, and hydrologic factor values from one basin to another can be qualitatively judged where alternative model results can be compared.

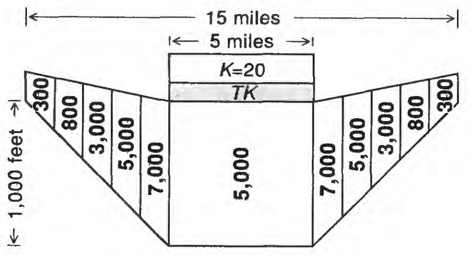
The properties that are significant in development of models of the basins are divided into three groups—hydrologic boundaries, flow components, and hydrologic factors. Components of each property are listed in table 2 with a subjective evaluation of their importance in steady-state and transient model development. The hydrologic-boundaries property represents the physical constraints on aquifer extent. The flow-components property includes recharge to and discharge from the aquifer. The hydrologic-factors property controls the movement and storage of water within the aquifer.

Selected properties are more important in steady-state simulations than in transient simulations. For steady-state conditions, water-level configuration and hydraulic conductivity are essential for model development. Flow



EXPLANATION

CASE 1



Input characteristics for steady-state model (used as standard for comparison)

Hydraulic conductivity (K=20) of upper layer, in feet per day

Transmissivity (**5,000**) of lower layer, in feet squared per day

Interlayer leakance (TK) $=10^{-7}$ feet per second per foot

Mountain-front recharge (Qm) = 26 cubic feet per second

(1) — — — **Simulated water-level profile**

CASE 2

Increased recharge

Mountain-front recharge = 52 cubic feet per second (All other parameters same as in CASE 1)

(2) ═══ **Simulated water-level profile**

FIGURE 22.—Effects of modifications of mountain-front recharge, interlayer-leakance values, lower-layer transmissivity, and basin width on water-level profiles of the steady-state basin-group model of the southeast basins.

CASE 3

Modified interlayer leakance

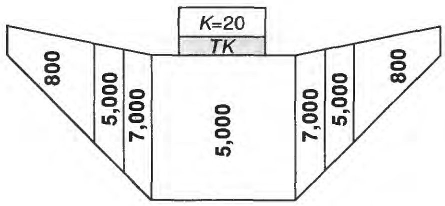
A. Interlayer leakance $=10^{-11}$ feet per second per foot (All other parameters same as in CASE 1)

(3A) **Simulated water-level profile**

B. Interlayer leakance $=10^{-12}$ feet per second per foot (All other parameters same as in CASE 1)

(3B) **Simulated water-level profile**

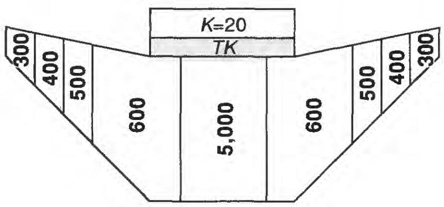
CASE 4



Modified lower-layer transmissivity

A. Transmissivity as shown at left. (All other parameters same as in CASE 1)

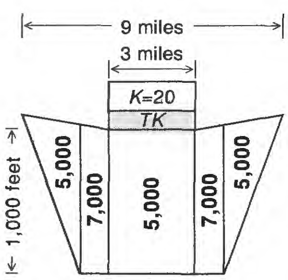
(4A) **Simulated water-level profile**



B. Transmissivity as shown at left (All other parameters same as in CASE 1)

(4B) **Simulated water-level profile**

CASE 5



Decreased basin width

Basin width as shown at left (All other parameters same as in CASE 1)

(5) **Simulated water-level profile**

FIGURE 22.—Continued.

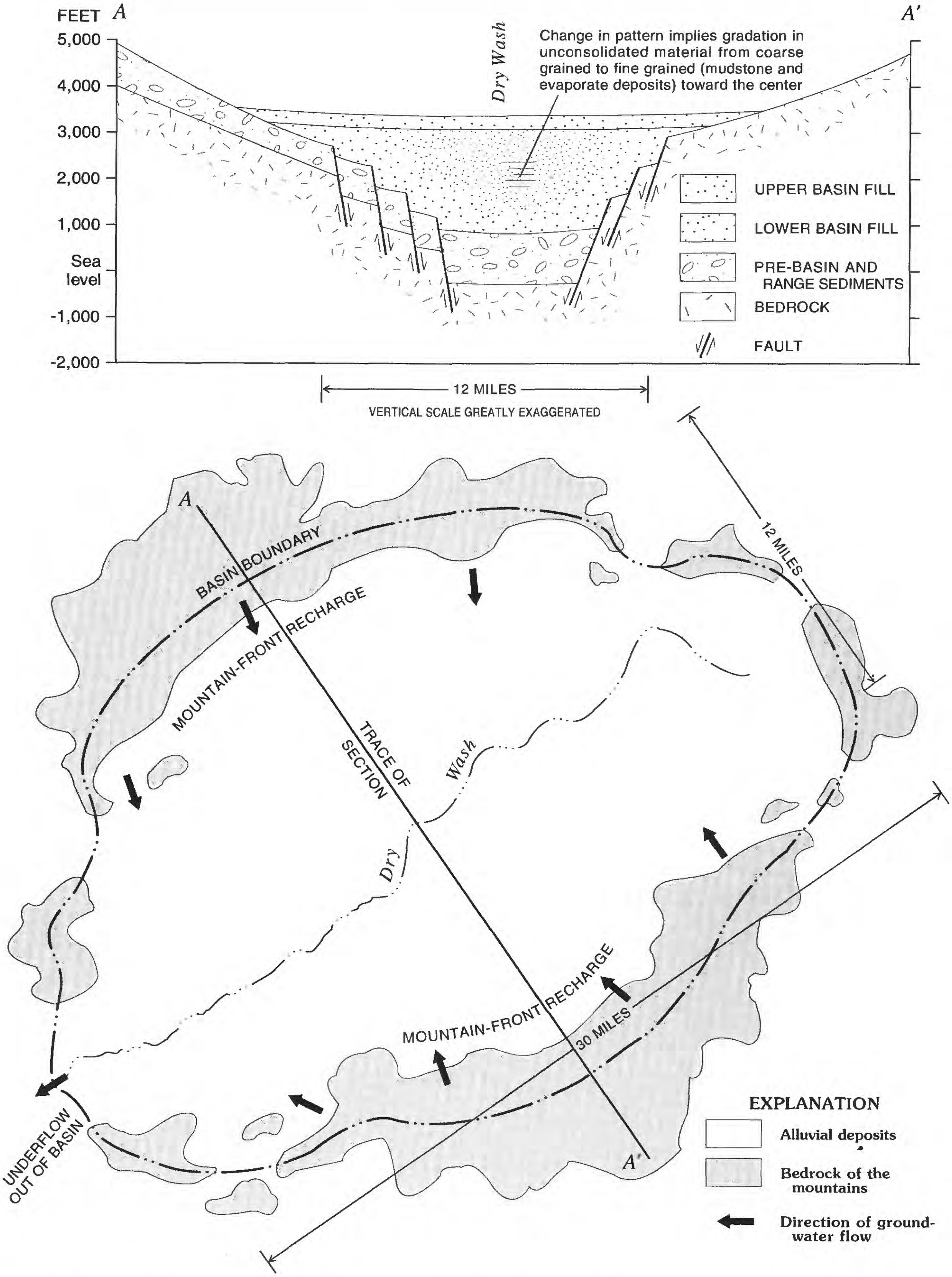
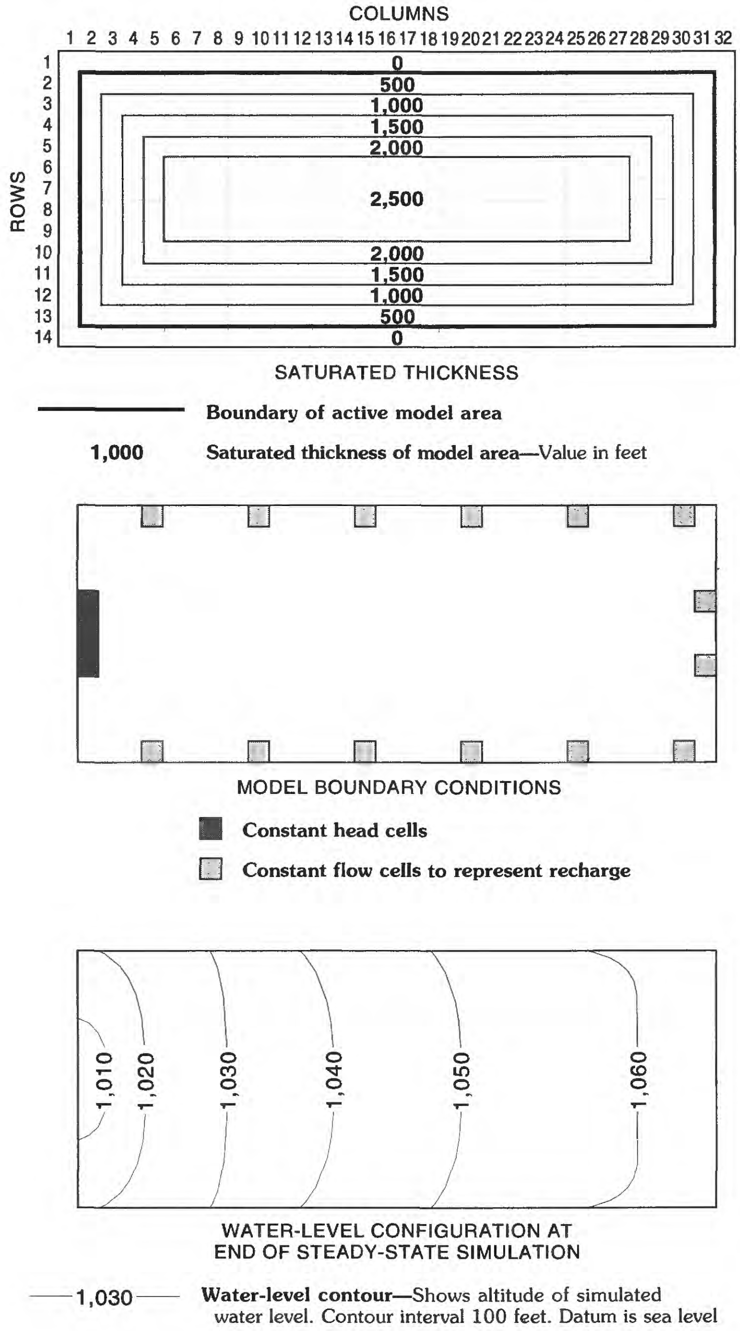
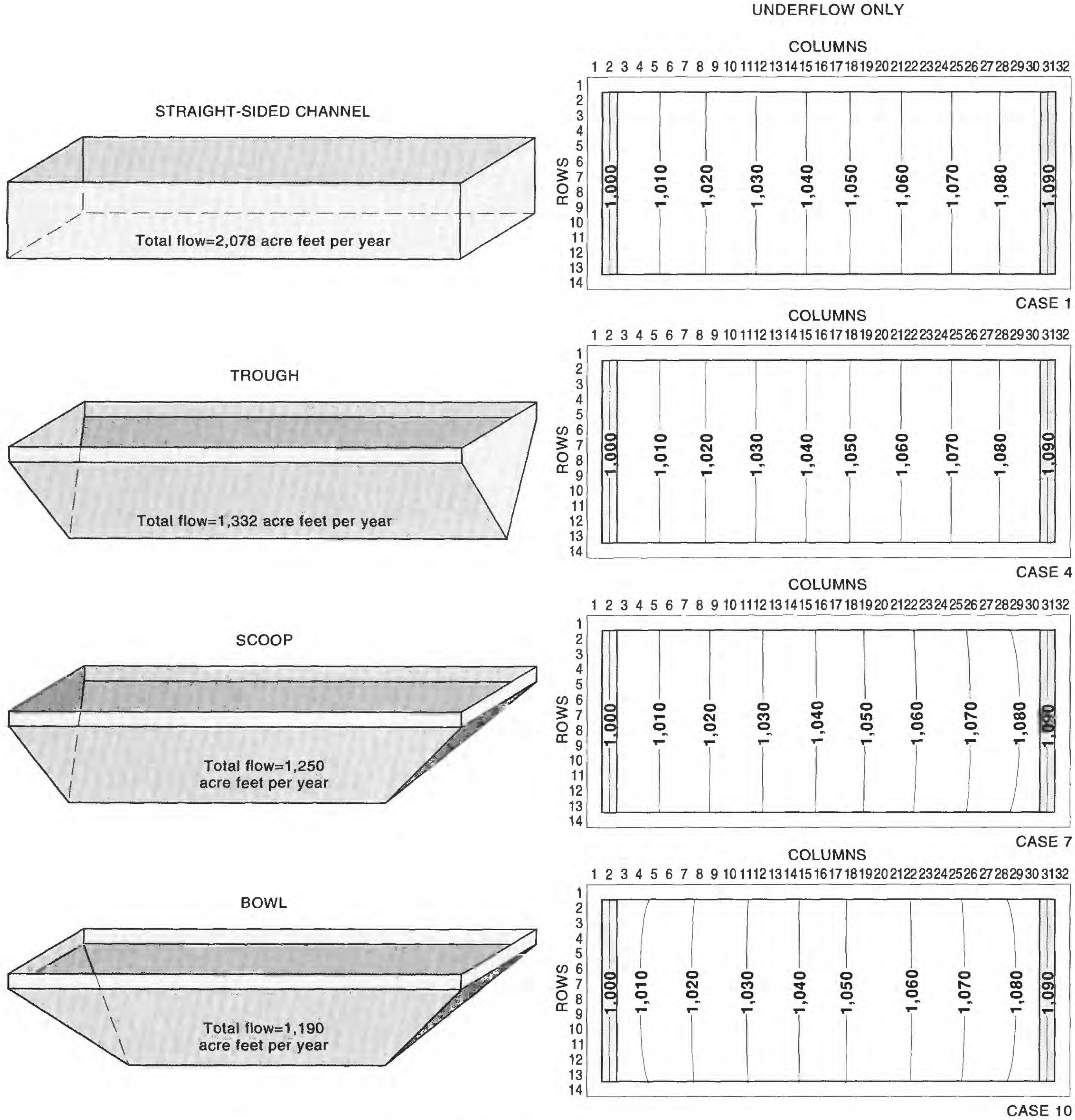


FIGURE 23.—Typical geographic and geohydrologic features in the central and west basins.



MODEL-INPUT CHARACTERISTICS	
Hydraulic conductivity of the upper layer *(K)*	20 feet per day
Grid cell dimensions *(ΔX and ΔY)*	5,280 feet x 5,280 feet

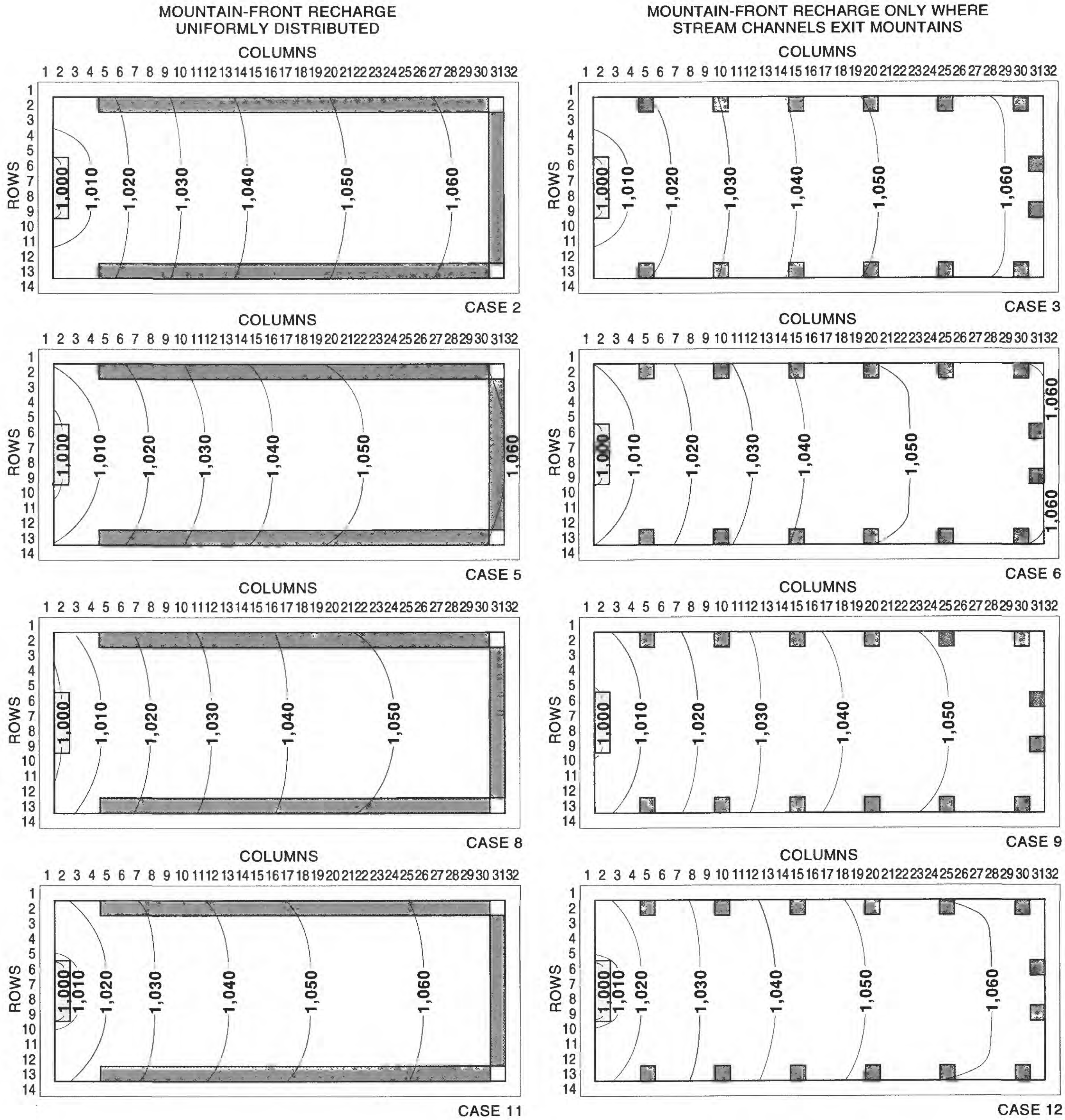
FIGURE 24.—Model grid, boundary conditions, and data input for steady-state basin-group model of the central and west basins.



WATER-LEVEL CHANGE STATISTICS, IN FEET
NOTE: Case 1 is the standard for comparison

Case:	1	2	3	4	5	6	7	8	9	10	11	12
Maximum difference	–	27.9	28.6	0.9	31.3	31.8	5.3	33.2	34.0	4.6	26.3	27.2
Minimum difference	–	–13.2	–13.5	–2.0	–9.5	–9.9	0.07	–9.0	–9.3	–4.3	–18.8	–19.1
Average difference	–	3.3	4.1	–0.6	6.3	7.1	1.7	8.3	9.3	–0.4	0.1	1.1
Average absolute difference	–	9.2	9.4	0.9	9.2	9.7	1.7	10.4	11.0	2.4	11.0	11.0

FIGURE 25.—Variations in data used and results of analysis of basin shape and recharge location and distribution in basin-group model of the central and west basins.



BASIN CHARACTERISTICS	
Basin length	30 miles
Basin width	12 miles
Maximum saturated thickness	700 feet
Hydraulic conductivity	9.5 feet per day
Initial water-level gradient	3 feet per mile
Minimum saturated thickness	200 feet

EXPLANATION

— 1,030 — **Water-level contour**—Shows altitude of water level calculated by the model. Contour interval 10 feet. Datum is sea level

Constant-head cell

Constant-flow cell

FIGURE 25.—Continued.

TABLE 2.—*Relative importance of hydrologic boundaries, flow components, and geohydrologic factors by basin group for modeling steady-state and transient conditions*

[1, important; 2, moderately important; 3, not important; — indicates property does not exist in this category]

Basin group	Steady-state conditions												
	Hydrologic boundaries			Flow components							Hydrologic factors		
	Water-level configuration	Aquifer bottom	Lateral boundaries	Areal recharge	Mountain-front recharge	Underflow in	Underflow out	Stream infiltration	Base-flow discharge	Evapotranspiration	Hydraulic conductivity	Vertical anisotropy	River/drain leakance
Southeast	1	1	3	2	1	2	2	2	2	1	1	1	2
Central	1	1	3	2	2	3	3	2	2	1	1	1	2
West	1	2	3	—	3	3	3	2	2	2	2	2	2
Colorado River	2	2	3	3	3	3	3	1	1	1	2	3	1

Basin group	Transient conditions															
	Hydrologic boundaries			Flow components									Hydrologic factors			
	Water-level configuration	Aquifer bottom	Lateral boundaries	Areal recharge[1]	Mountain-front recharge	Underflow in	Underflow out	Stream infiltration[2]	Base-flow discharge	Drain discharge	Evapotranspiration	Pumping	Hydraulic conductivity	Vertical anisotropy	River/drain leakance	Storage coefficient
Southeast	1	3	3	2	2	3	3	2	2	2	2	1	2	1	2	1
Central	1	2	2	2	3	3	3	2	3	3	3	1	2	2	3	1
West	1	2	2	3	3	3	3	3	3	—	3	1	3	2	3	1
Colorado River	2	3	3	2	3	3	3	1	1	1	1	2	3	3	1	3

[1]Includes recharge from excess applied irrigation water in some basins during recent periods.
[2]Includes recharge from canals and laterals in a few basins.

component values require varying degrees of reliability owing to the different significance of the components as related to the hydrologic categories. For example, although mountain-front recharge is important in the southeast basins because it represents the largest single predevelopment recharge source, in the west basins mountain-front recharge is practically nonexistent and therefore is not important.

For transient simulations, reliable water-level-change information, pumpage, and specific yield or storage coefficient are the most important data in the history-matching process, especially for those basins in which significant storage depletion has occurred. The relative significance of other boundary conditions and hydrologic factors is related to the individual setting of the geohydrologic system. Flow components also have differing importance depending on the individual basin setting; however, the magnitude and distribution of pumping commonly are critical items. This is particularly true in the central and west basins where large-scale pumping has resulted in a large-scale overdraft of ground-water resources.

INFORMATION ADEQUACY AND TRANSFERABILITY

Model-sensitivity analyses provided a qualitative assessment of the adequacy of values and areal distributions of properties used in the models. In general, the more sensitive a model is to a particular property, the greater the data requirements will be. Similarly, transferability of property values and distributions is related to

TABLE 3.— *Geohydrologic factors evaluated as part of the modeling process*

[• indicates this factor was evaluated in the model]

Factors	Southeast					Central		West			Colorado River		Basin group	
	Upper San Pedro basin	Benson basin	Lower San Pedro basin	Willcox basin	Douglas basin	Tucson basin	Avra Valley	Waterman Wash basin	Harquahala Plain	McMullen Valley	Mohave basin	Parker basin	Southeast basins	Central and west basins
Hydrologic boundaries														
Water-level configuration	•	•	•	•	•	•	•	•	•	•	•	•	•	•
Aquifer bottom	•						•		•	•		•	•	•
Lateral boundaries									•	•			•	•
Flow components														
Areal recharge	•	•	•	•	•	•	•	•	•	•	•	•		
Mountain-front recharge	•	•	•	•	•	•	•	•		•		•	•	•
Underflow in	•					•	•	•	•	•				
Underflow out	•					•	•		•	•		•		
Stream infiltration	•	•	•			•					•	•		
Base-flow discharge	•	•	•		•						•	•		
Drain discharge											•	•		
Evapotranspiration	•	•	•								•	•		
Pumping	•	•	•	•	•	•	•	•	•	•			•	•
Hydrologic factors														
Hydraulic conductivity	•					•	•			•		•	•	•
Vertical anisotropy	•					•	•	•	•	•		•	•	•
River/drain leakance	•	•	•								•	•	•	
Storage coefficient				•	•	•	•	•	•	•			•	•

model sensitivity. In the situation where a model is insensitive to a property, it is directly transferable to all the basins of a group. The models developed as part of this study were used to explore the effects of various geohydrologic properties on model results. The properties that were specifically addressed in each model are indicated in table 3.

HYDROLOGIC BOUNDARIES

The most significant hydrologic boundary was the water-level configuration, which represents the upper boundary of the ground-water flow system. Accurate areal and vertical head-distribution data are vital in understanding the flow system. In general, because of similarities in hydrologic properties of sediments that make up the aquifers and the existence of regional similarities or trends in magnitude of some components of inflow to the basins, hydraulic gradients are similar in basins of a particular group. Moreover, the general shape of the water-level contours can be reliably predicted on the basis of basin categorization (see section "Results of Simulation" for the basin-group models).

Explicit definition of the physical limits of the aquifer in the typical geohydrologic setting of the study area, including the bottom and sides of the aquifer system, is least important for steady-state modeling. For transient modeling, lateral boundaries of the aquifer in close proximity—within several miles—of pumping centers need to be well defined because of boundary effects on the size and shape of a cone of depression. The model for McMullen Valley was used to evaluate the sensitivity of

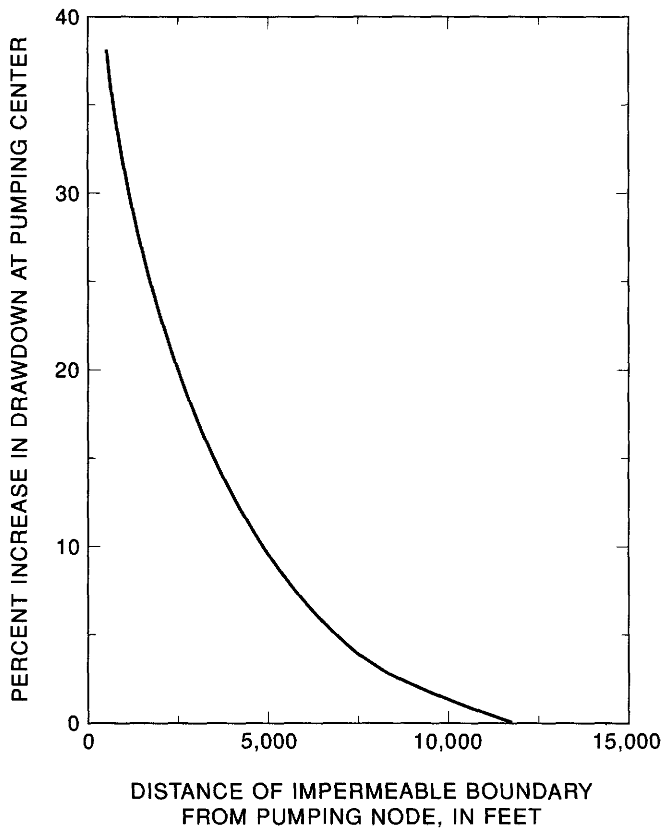


FIGURE 26.—Effect of lower-layer boundary location on simulated drawdown in McMullen Valley (from Pool, 1987).

simulated drawdown near an impermeable bedrock boundary (Pool, 1987). The location of the bedrock boundary was varied from one-quarter mile to several miles from the pumping cell. In this storage-depletion system, part of the well field is within a few miles of the lateral boundary of the basin. The size and shape of the cone of depression are influenced greatly by boundary effects. Drawdown increases significantly if the pumping center is within 7,000 ft of the aquifer boundary (fig. 26). The distance is unique to the specific situation and will change if aquifer character, geometry, or magnitude and duration of pumping are changed.

FLOW COMPONENTS

Pumpage estimates are available in most areas or can be made on the basis of irrigated area. Although pumpage is the most well-documented flow component in most basins, it still is poorly known. Data documenting other flow quantities are, for practical purposes, nonexistent, and values must be estimated by indirect means. The other flow components, however, generally are much smaller than pumpage, so that errors in these other values have less effect on results of simulations than do errors in pumpage estimates. Modeling is useful in determining the reasonableness of flow-quantity estimates. Large errors in estimated quantities are possible because the typical water-budget method of estimation allows errors to accumulate in the unknown quantities.

In the southeast basins, mountain-front recharge represents the principal inflow; however, interbasin underflow and stream-aquifer exchange can be of nearly equal importance. Evapotranspiration is the main discharge component, and model results are sensitive to the evapotranspiration rates; however, underflow and stream base flow are important in several basins. Varying the evapotranspiration rate within a reasonable range of values in the model of the upper San Pedro basin (Freethey, 1982) changed the net discharge by about 40 percent and resulted in an imbalance of the other flow components.

Results from the upper San Pedro basin model indicated that generalizing mountain-front recharge was feasible (Freethey, 1982) but not recommended if site-specific water-level matches are a goal. In models for five other southeast and central basins, the use of generalized mountain-front-recharge values presented no undue difficulty in calibration.

Other significant flow components were investigated during model development for specific basins. Results of simulations of Avra Valley indicate that return of excess applied irrigation water may be about 30,000 acre-ft/yr since the early to mid-1960's (Anderson, 1983). Results from nine other models also indicate that return of excess irrigation water to the principal basin aquifer probably is a significant flow component. Few field data exist with which to verify irrigation return flow, however, and only indirect evidence is available with which to quantify the flow component.

The models of the Colorado River basins were sensitive to the rate of river leakage to the principal aquifer and the rate of consumptive use by riparian vegetation. The magnitude of these two components overwhelms other much smaller components. The value of riverbed leakance controls the quantity of infiltration from surface flow. Drains have been installed in a few basins, and the drain leakance is not as great a control on the water budget as is the riverbed leakance (fig. 27), probably because of the much smaller areal extent of drains. The ground-water systems of the Colorado River basins are limited in areal extent, and modeling on a regional scale as in this study is useful mainly in examining the magnitude of the water-budget components. Balancing the water budget of a basin so that all flow components are within a reasonable range of values was the best method of calibration rather than relying on head values and configurations.

Models of the Colorado River basins were most sensitive to changes in the evapotranspiration rate (Tucci,

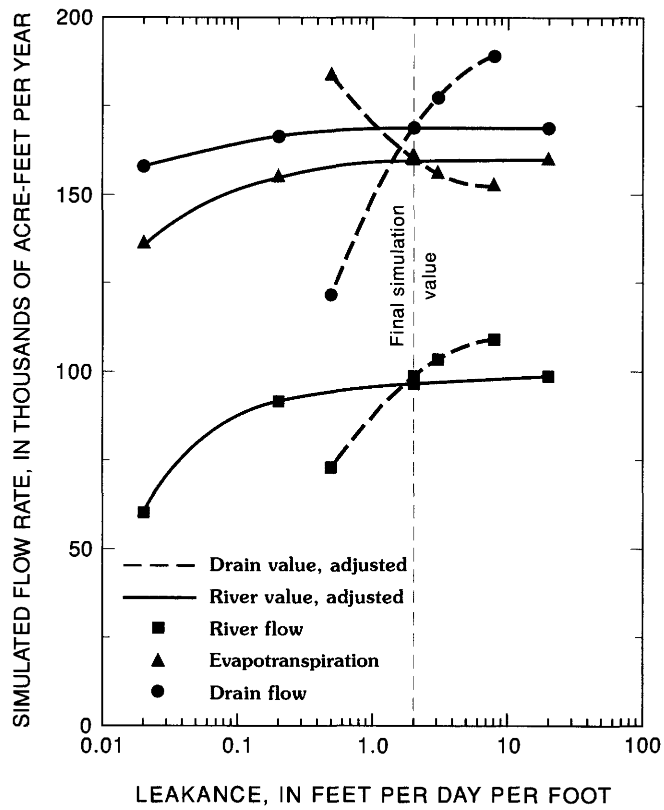


FIGURE 27.—Variation in components of loss from river flow, evapotranspiration, and drain flow resulting from changes in the leakance factor for drains and rivers in the Parker Valley model (from Tucci, 1982).

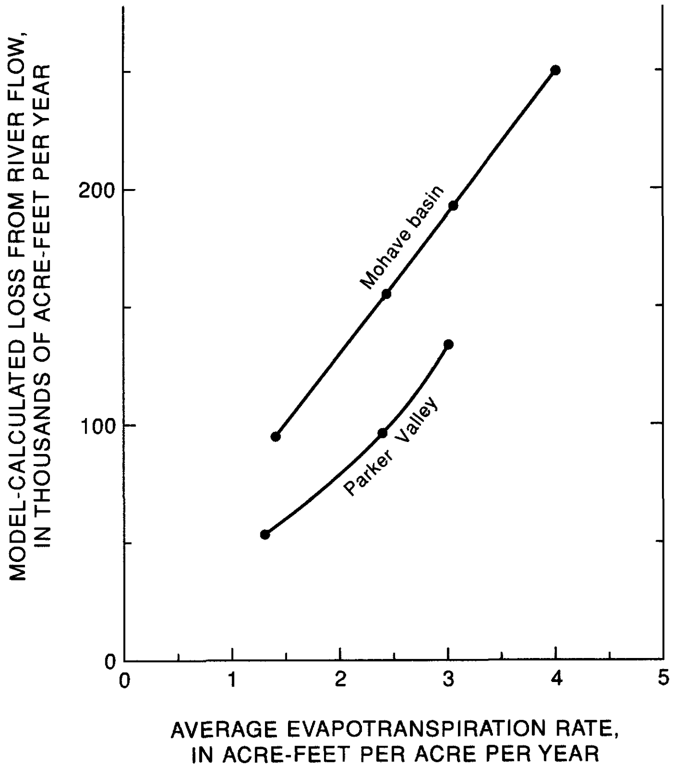


FIGURE 28.—Variation in net loss from river flow in Mohave basin and Parker Valley related to change in average evapotranspiration rate.

1982). Increasing the average evapotranspiration rate by 1 (acre-ft/acre)/yr caused about the same increase in river infiltration in Mohave basin as in Parker Valley—about 60,000 acre-ft/yr (fig. 28). Because of model sensitivity to changes in evapotranspiration, changes in vegetation type and density are important considerations for similar basins in the Colorado River area. Sensitivity to evapotranspiration also is expected for models of other alluvial aquifers traversed by a perennial stream, such as some basins of the highland and southeast groups.

HYDROLOGIC FACTORS

For a steady-state model, the most significant hydrologic factor of a basin is the hydraulic conductivity; for a transient model, specific yield or storage coefficient is the most significant property. Vertical anisotropy is important in basins containing an extensive fine-grained unit that functions as a leaky confining unit between two layers of the aquifer system.

Hydraulic conductivity and transmissivity arrays were developed for most basin models by using specific-capacity data and flow-net analyses. Hydrologic factors were generalized to analyze the adequacy of using a constant rather than a variable distribution of values for areas where no field data existed. Results from this analysis in the upper San Pedro basin indicated that the model was sensitive to generalized aquifer hydraulic conductivity (Freethey, 1982). Use of a single average value for hydraulic conductivity in the model of the Benson basin proved unsatisfactory because of the large variability in the potentiometric-surface gradient created by variations in the hydraulic conductivity. The greatest disparity occurred at the basin margins where water levels were more than 150 ft lower than those in the nongeneralized simulation; water levels in the flood-plain area were about 30 ft higher than those in the nongeneralized simulation. In the same test, simulated water-budget values also fell outside the acceptable range established from field data and previous studies.

The basin-group model of the southeast basins was used in a sensitivity analysis of system response to variations in values of hydraulic conductivity of the upper layer, transmissivity of the lower layer, interlayer leakance, and riverbed leakance. The results of varying these properties in terms of average head change and percentage change in flow quantities allow a qualitative comparison of the general significance of each property (table 4).

TABLE 4.—*Sensitivity of model-calculated heads and flow quantities to changes in values of flow components and hydrologic factors in southeast basin model*

[* indicates that reasonable range is exceeded; — indicates that multiplication factor was not tested]

	Multiplication factor	Flow components									Hydrologic factors								
		Maximum evapotranspiration rate			Riverbed leakance			Mountain-front recharge			Hydraulic conductivity, upper layer			Transmissivity, lower layer			Interlayer leakance		
		Average absolute head change, in feet		Change in quantity of flow, in percent	Average absolute head change, in feet		Change in quantity of flow, in percent	Average absolute head change, in feet		Change in quantity of flow, in percent	Average absolute head change, in feet		Change in quantity of flow, in percent	Average absolute head change, in feet		Change in quantity of flow, in percent	Average absolute head change, in fee		Change in quantity of flow, in percent
		Lower layer	Upper layer		Lower layer	Upper layer		Lower layer	Upper layer		Lower layer	Upper layer		Lower layer	Upper layer		Lower layer	Upper layer	
Decrease	0.0001	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	385.3	2.82	0.66
	.001	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	42.14	1.90	−2.06
	.01	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	4.51	.64	.78
	.1	*	*	*	2.61	2.91	−22.25	*	*	*	*	*	*	*	*	*	.46	0	−.13
	.2	*	*	*	—	—	—	*	*	*	3.13	3.07	−8.40	*	*	*	—	—	—
	.25	*	*	*	—	—	—	*	*	*	—	—	—	62.32	3.07	1.13	—	—	—
	.3	4.98	4.72	−19.93	—	—	—	*	*	*	—	—	—	—	—	—	—	—	—
	.5	—	—	—	—	—	—	14.33	3.83	−28.44	1.39	1.33	−5.11	21.64	2.17	−.35	—	—	—
	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
	2	—	—	—	—	—	—	26.60	5.51	52.42	1.35	1.23	9.69	16.24	7.80	5.78	—	—	—
	3	4.07	4.02	5.76	—	—	—	*	*	*	—	—	—	—	—	—	—	—	—
	4	—	—	—	—	—	—	*	*	*	—	—	—	41.93	30.20	19.05	—	—	—
	4.5	4.85	4.81	5.89	—	—	—	*	*	*	—	—	—	*	*	*	—	—	—
	5	*	*	*	—	—	—	*	*	*	2.57	2.25	37.25	*	*	*	—	—	—
	10	*	*	*	3.09	3.40	41.88	*	*	*	*	*	*	*	*	*	0	0	.01
Increase	100	*	*	*	*	*	*	*	*	*	*	*	*	*	*	*	0	0	.01

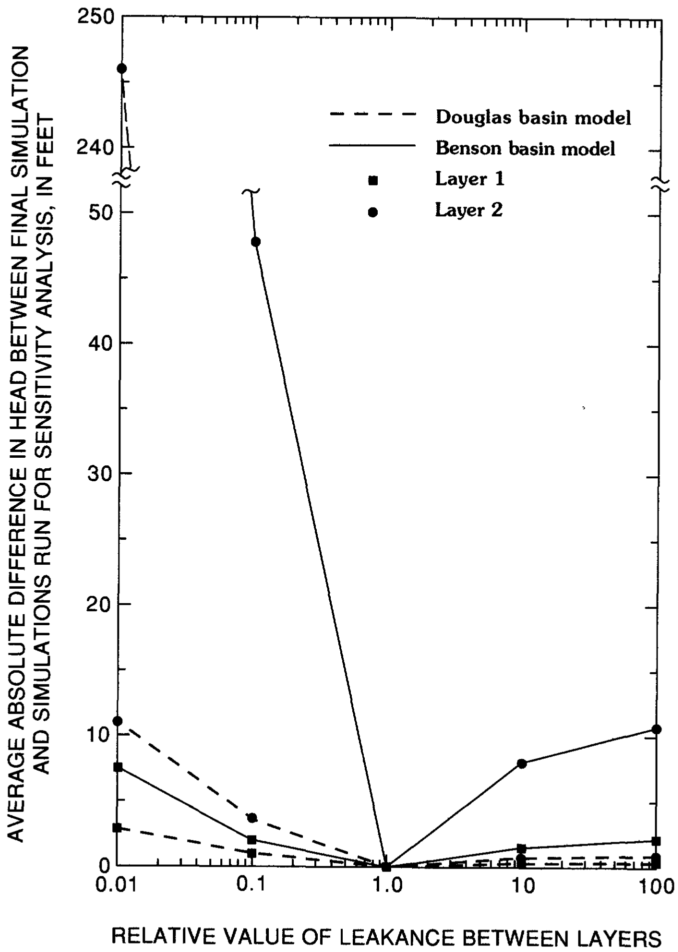


FIGURE 29.—Sensitivity of model-calculated heads in the Benson and Douglas basin models to variations in interlayer-leakance value.

The value used for vertical interlayer leakance was found to have almost no effect on simulation results until the value was less than 10^{-8} (ft/s)/ft. Head differential between layers increases significantly when smaller values of vertical leakance are used. Interlayer-leakance values were varied through several orders of magnitude in the models for Benson and Douglas basins. On the basis of well-log data, the basin-center fine-grained facies is more extensive and thicker and serves more as a confining bed in Benson basin than in Douglas basin. Accordingly, model sensitivity to variations in interlayer leakance is greater in the model for Benson basin than in the model for Douglas basin (fig. 29). Both model responses indicate an increasing sensitivity as the value of interlayer leakance is decreased. The conclusion is that a model is most sensitive to interlayer leakance where the fine-grained sediments form a massive confining bed and the vertical hydraulic conductivity is small.

In transient models of the upper San Pedro and Douglas basins, a uniform average specific yield was used in several sensitivity runs. Results indicate that using a uniform average value of specific yield does not change model results significantly compared with simulations using a moderately variable array of specific yield. As the degree of development increases, however, the difference resulting from the two arrays increases. If the deposits within the basin area affected by pumping have a large areal difference in grain size, an average specific-yield value can result in significant differences in simulated water levels.

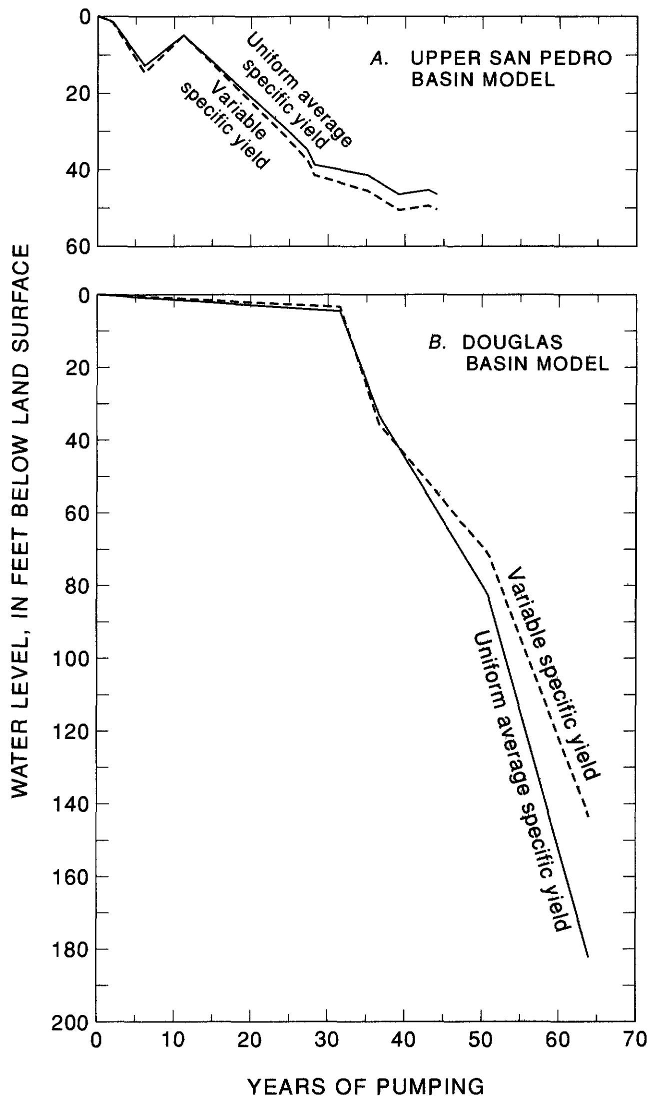


FIGURE 30.—Maximum simulated drawdown using a variable areal distribution of specific yield and a uniform specific yield representing the average value for the models of upper San Pedro basin and Douglas basin.

Comparison of maximum drawdowns using average and variable specific yield in the upper San Pedro model (fig. 30*A*), in which all other values used in the model are fixed, indicates that using a uniform average value results in a slightly smaller maximum drawdown because the average value is slightly larger than the variable value in the cone of depression. In the model of the Douglas basin, a similar analysis using an average specific yield resulted in a larger maximum drawdown (fig. 30*B*). In this area, the actual specific yield is smaller than the basinwide average. Use of a basinwide uniform value of specific yield provides satisfactory results for a general model calibration, but for site-specific analyses, especially where the aquifer material is heterogeneous, areally variable specific-yield values are needed.

The effect of varying hydrologic properties also was examined by imposing pumping stress on the basin-group model of the central and west basins in conjunction with changing specific yield and by altering the location of the area where vertical leakance is small because of the presence of a confining layer. First, the basin-group model was used to examine variations in specific yield and pumping locations and quantity. The second approach introduced a fine-grained layer in the basin-group model to examine the resulting effects on drawdown. Location and distribution of pumping were varied to approximate conditions that exist in three of the modeled basins of the west group—McMullen Valley, Harquahala Plain, and Waterman Wash basin.

In the single-layer model, a simulated pumping rate of 1 $(ft^3/s)/mi^2$ was used over an area of 88 mi^2 in the central part of the modeled area for a 20-year duration. The resulting water-level declines using six different specific-yield distributions indicate the model sensitivity to the absolute value and areal distribution of specific yield (fig. 31). Using a constant value of 0.12 for specific yield results in virtually the same water-level decline as using a random spatial distribution of specific-yield values of 0.06, 0.12, and 0.18. Water-level decline is proportional to the average value of specific yield in these storage-depletion systems. Actual ground-water withdrawal in several west basins is concentrated near the outflow end of the basin or in more than one pumping center. The effect of concentrating the pumping is shown by water-level declines on an axial profile through the basin-group two-layer model (fig. 32). The increase in decline because of the presence of a fine-grained facies is also shown. The magnitude of the increase in water-level decline depends on the quantity of pumpage in areas where the fine-grained unit is present. Redistribution of pumpage to within 2 mi of the basin boundary increased water-level declines less than 10 percent.

HYDROLOGIC INSIGHTS GAINED FROM MODELING SOUTHWEST ALLUVIAL BASINS

The degree of ground-water development represented in the 72 alluvial-basin aquifers in south-central Arizona ranges from virtually pristine to being dewatered beyond the point that results in land subsidence. Development in many basins began prior to comprehensive data-collection efforts, thus precluding conceptualization of predevelopment hydrologic conditions on the basis of fact. Results of modeling 12 alluvial basins in various stages of development aided in more clearly understanding hydrologic processes occurring before pumping began. Modeling more than one basin in four of the groups has shown that basins within each of these four groups of basins have similar hydrologic systems and demonstrate a similar response to pumping stress. Although no basins in the highland group were modeled because of a lack of information, their geographic and hydrographic environments suggest that their hydrologic systems are similar to those of the Colorado River basins; thus, insights about Colorado River basins gained from modeling results are also considered to apply to the highland basins.

Results of simulations of hydrologic systems of the southeast group indicated that (1) mountain-front recharge probably occurs nonuniformly along the margins of the basins; (2) hydrologic factors such as transmissivity, hydraulic conductivity, specific yield, and confined storage coefficient are laterally variable; (3) the distribution of storage coefficient or specific yield is not as critical as the average value of these properties in determining the amount of water-level decline caused by pumping; (4) accurate areal distribution and values of transmissivity or hydraulic conductivity are critical in determining the configuration and gradient of the potentiometric surface; (5) variably confined aquifer conditions near basin margins probably are a result of interlayering of coarse sediments from the mountains and fine sediments along the axis of basins originating from lake deposits; and (6) pumping causes discharge by evapotranspiration and stream gains to decrease, even if pumping is from substantial depth in the basin fill.

The model of Willcox basin (closed drainage) indicated that extensive lacustrine deposits in the center of a basin can create a boundary condition that will accelerate water-level declines in nearby pumping wells. The model of the Tucson basin indicated that postdevelopment water-level declines can result in a decrease in transmissivity of the principal deep aquifer and that these decreases are not proportional to the decrease in saturated thickness because the upper basin fill is coarser grained and more permeable than the lower basin fill. As the upper basin fill is dewatered, that part of the aquifer having larger hydraulic conductivity is eliminated, leaving only the underlying finer-grained basin fill to transmit water to wells.

Models of the hydrologic systems of the central and southeast basin groups indicated that (1) excess applied irrigation water can eventually percolate down to

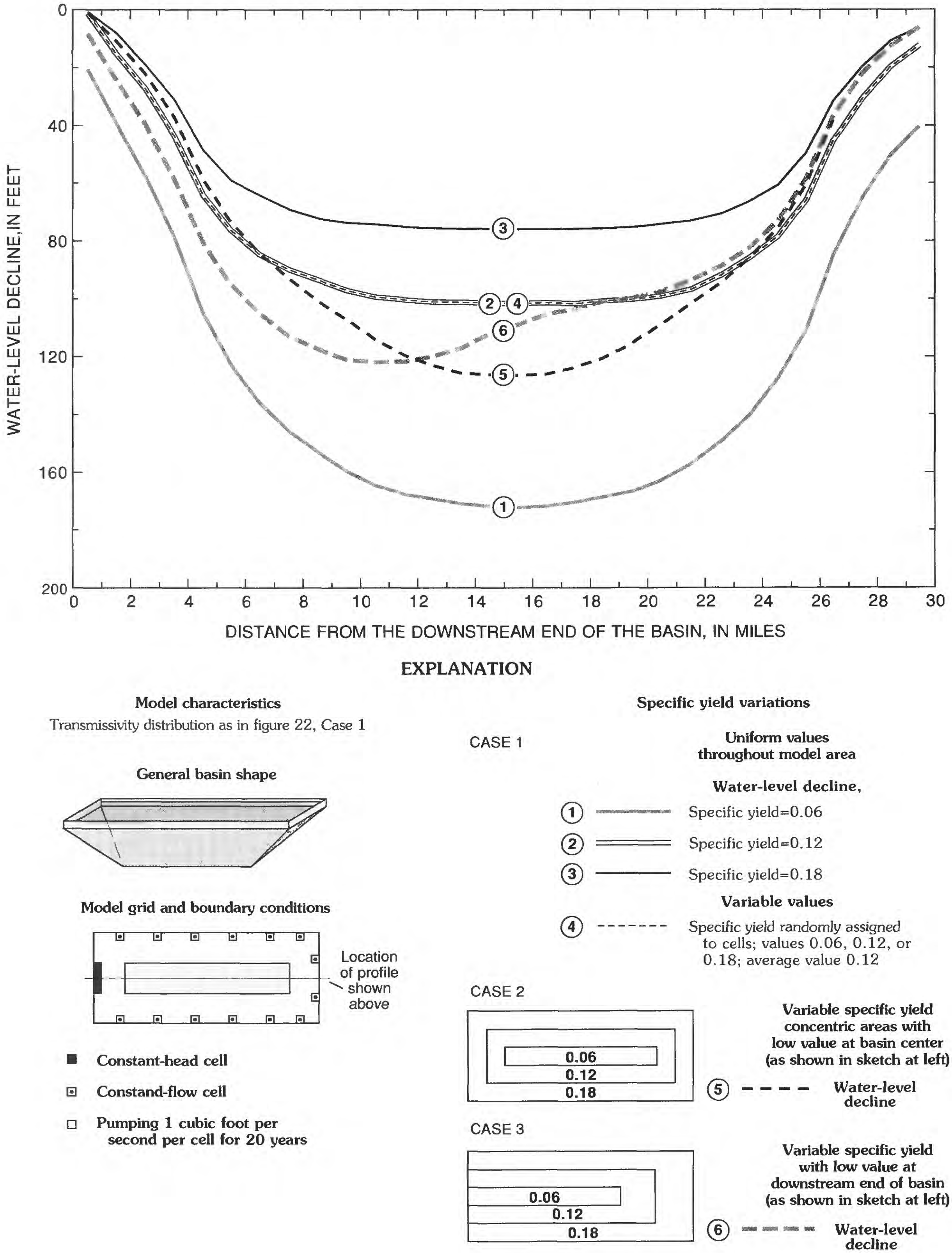
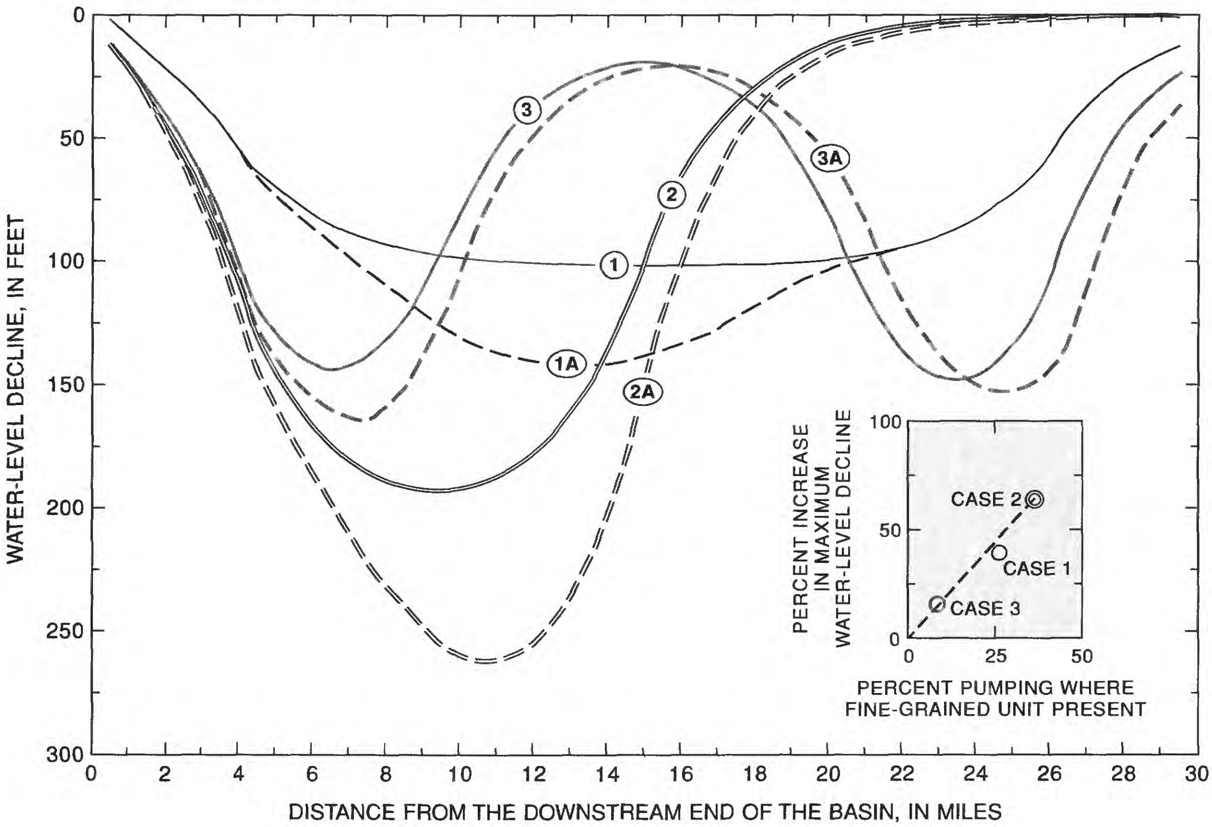


FIGURE 31.—Response of the basin-group model of the central and west basins to variations in areal distributions and values of specific yield.



EXPLANATION

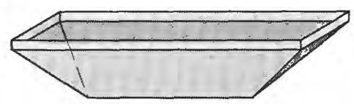
Pumping distribution model scenarios

Model characteristics

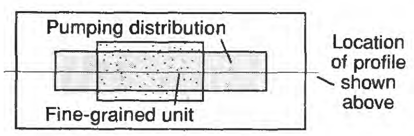
Transmissivity distribution as in figure 22, Case 1
Specific yield of upper layer= 0.12

Interlayer leakance (TK)=10^{-7} feet per second per foot

General basin shape



CASE 1

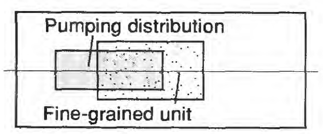


Pumping from lower layer uniformly distribution over 88 square miles at rate of 1 cubic foot per second per square mile for 20 years

(1) ———— **Water-level decline along basin axis, no fine-grained unit**

(1A) — — — **Water-level decline along basin axis with fine-grained unit at basin center, interlayer leakance=10^{-11} feet per second per foot**

CASE 2

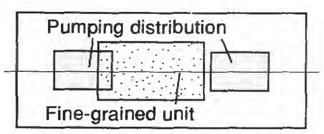


Pumping from lower layer uniformly distributed over 44 square miles at rate of 2 cubic feet per second per square mile near downstream end of basin for 20 years

(2) ═════ **Water-level decline along basin axis, no fine-grained unit**

(2A) ═ ═ ═ **Water-level decline along basin axis with fine-grained unit at basin center, interlayer leakance=10^{-11} feet per second per foot**

CASE 3



Pumping from lower layer uniformly distributed over 44 square miles in two pumping centers at rate of 2 cubic feet per second per square mile for 20 years

(3) ———— **Water-level decline along basin axis, no fine-grained unit**

(3A) — — — **Water-level decline along basin axis with fine-grained unite at basin center, interlayer leakance=10^{-11} feet per second per foot**

FIGURE 32.—Water-level declines resulting from variable areal distributions of pumping from the lower layer and presence or absence of a fine-grained facies in the basin-group model of the central and west basins.

recharge the aquifer even in basins where the aquifer is several hundred feet below land surface and (2) the time required for this water to move through the unsaturated zone may be several decades.

Models of hydrologic systems in the central and west basin groups indicated that (1) the direction and rate of ground-water movement prior to development are defined primarily by the location of recharge sources, even though the magnitude of this recharge is small; (2) after development in the basin, the direction and rate of ground-water movement are governed primarily by the location of pumped wells and the amount of withdrawal occurring at each well or well field; and (3) because recharge is small and storage depletion is large, water-level declines near pumped wells can be accelerated if the cone of depression intersects buried bedrock along the margins of the basin.

Models of hydrologic systems in the Colorado River basin group indicated that near-surface hydrologic processes, such as (1) stream-aquifer water exchange, (2) evapotranspiration prior to development, (3) drain-aquifer water exchange, and (4) recharge from irrigation after development, are the primary controlling components of ground-water movement in these basins.

GENERAL MODELING CONSIDERATIONS

The four principal items that are required in the development of a numerical model are as follows:

- Water-level maps that represent the predevelopment period and subsequent development periods, so that field-determined potential distribution can be compared to the simulation results;
- Definition of the aquifer boundaries, including information to define the size and shape of the basin;
- Areal distribution of aquifer properties, including hydraulic conductivity or transmissivity and storage characteristics; and
- Definition of quantities and areal distributions of the various flow components of the water budget of a basin.

This study developed information and understanding of these four essential items throughout the study area. The steady-state water-level map for a particular basin can be of fair to poor quality depending on the rapidity of the initial ground-water development. A representation of the steady-state flow system throughout the study area was developed as part of this study (Freethey and Anderson, 1986). The process of developing this map included the use of all available data and information transfer to extrapolate water-level shapes and gradients to areas of minimum data.

The steady-state water-level map was used in conjunction with knowledge of the physical boundaries of the basins to develop an areawide synthesis of boundary-flow conditions. Flow-net analyses of individual basins were used to develop and refine estimates of location and quantities of boundary flows. Through a process of balancing and adjusting the flow components of individual basins and the entire study area, a general synthesis of inflow and outflow quantities was developed (Freethey and Anderson, 1986).

Aquifer properties are a function of the sediments that make up the aquifers, which are geologically complex in the basins of the study area. In developing a model, the complexities of the aquifer are generalized through the use of average values of hydrologic properties to simulate a small block of the system. In this study, that concept has been extended to look at generalizations within groups of basins. The models represent another step in the development of a complete understanding of the geohydrologic systems of the basins. The systems were simulated in quasi-three-dimensional form. Previous models had simulated the basins as two-dimensional systems, although evidence documents the existence of head variation with depth. In most instances, however, the head variations are small and spatially variable. Head variations are thought to be local and associated with the presence of fine-grained facies and faults (Davidson, 1973). In the southeast basins, however, areally extensive confined ground-water conditions have been documented. Major depositional or structural discontinuities in a basin must be considered in development of a three-dimensional model because of differences in hydrologic properties. Location and extent of the fine- and coarse-grained sediments that make up the aquifers have been described in a general manner by Freethey and others (1986). These data should allow the development of first-approximation three-dimensional models of all basins in the study area.

The values of hydrologic factors for use in a model can be significantly refined if more field data are available. Two issues need to be considered, however—(1) the accuracy of existing estimates of property values and (2) the sensitivity of the model to variations in those values. A particular property may not be of equal importance in all ground-water flow models. Ranking of the properties by importance is dependent on the geohydrologic environment that defines each aquifer system. Properties that exhibit the greatest influence on steady-state ground-water flow—hydraulic conductivity, natural recharge rates, and extent and integrity of confining units—are most significant in the southeast basins. Properties that exhibit the greatest influence on transient ground-water flow—specific yield, pumping rates, and well location—are most significant in the central and west basins. Properties that exhibit the greatest influence on shallow water-table conditions—evaporation and transpiration, stream-aquifer exchange, and irrigation-return flow—are most important in those basins in which streamflow is perennial.

Information needed to adequately simulate a hydrologic system that has been stressed by pumping or land-use changes caused by development will differ between basin groups. When ground-water withdrawals in a basin become greater than the natural downvalley flow and withdrawal is sufficiently widespread, duplication of stress–strain relations, in terms of matching pumpage and water-level decline, requires the documentation of actual withdrawal rates and locations of pumped wells. In a ground-water overdraft environment such as exists in the more intensively developed basins, a model may have a generally coarse grid with the smallest cells being 1 mi^2 because the data to simulate the dominant stress, which is pumping, do not exist in greater detail. Grid sizes of less than 1-mi width were used in the models of Parker Valley, Mohave basin, upper and lower San Pedro basins, and Benson basin, in which the small spacing was required to simulate the river alignment. In the models for Harquahala Plain and McMullen Valley, smaller grid spacing was used where a pumping center was close to a poorly permeable boundary and image effects greatly influenced the size and shape of the cone of depression. In most basins, available data did not warrant the use of small grid spacing.

Models of the southeast basins need to be multilayered to allow for vertical flow through leaky confining layers. Significant head differences may exist where fine- and coarse-grained deposits are interlayered, and the value of interlayer leakance may be smaller than 10^{-10} (ft/s)/ft. History matching in steady-state and transient simulations is greatly affected by the steep hydraulic gradient at the basin perimeter; this effect may indicate a vertical component of flow that is not being simulated or the presence of relatively thin, steeply inclined, permeable deposits that transmit mountain-front recharge into the basin-fill aquifer. Because evapotranspiration is a principal discharge mechanism in the southeast basins, the mathematical relation that simulates the evapotranspiration function is important and will have a great effect on the history-matching process.

Models of the central and west basins need to be multilayered to allow adequate simulation of vertical-flow components. Although one-layer models have been developed and calibrated, their limitations must be recognized. History matching for transient simulations is most sensitive to pumping rates and specific yield. Variable recharge, delayed drainage, and depth-dependent variations in hydrologic properties are of major importance in transient simulations because of their variation with time.

Simple one-layer models of the Colorado River basins can provide satisfactory results for purposes such as those of this study. The process of matching simulated and measured water levels is sensitive to the evapotranspiration rates and riverbed-leakance value. For these basins, water-budget estimates are of equal or greater value than a water-table map in providing the data necessary for final model development. The alluvial aquifers in the highland basins are considered to be scaled-down versions of the Colorado River group; therefore, properties critical to model development should be the same. Specific basin models of this group were not developed because of insufficient data. Evapotranspiration and riverbed leakance are presumed to be the principal factors to which the model history-matching process will be sensitive.

Simulation of the stream alluvium is important in some basins, especially in the southeast basins and the Colorado River basins. This upper layer needs to be simulated in models of some basins because it is integral to the hydrologic processes of evapotranspiration, stream-aquifer exchange, and vertical movement of ground water to or from the deeper aquifer being pumped. Transient history matching for a simulation where storage depletion is important, such as those of the west basin group, depends on accurate values for pumpage and specific yield. Without reliable values for pumpage and specific yield, the range of possible matches between simulated and measured water levels is large, and the resulting simulation may provide questionable results. Other factors that result in time-dependent changes in flow quantities will be important in the simulation; these factors include recharge from excess applied irrigation water, delayed drainage from fine-grained sediments, and the quantity of water derived from compaction of fine-grained material.

SUMMARY

Numerical modeling was used as a tool to examine and evaluate regional similarities and differences in ground-water flow systems within the alluvial basins that occupy 82,000 mi^2 of south-central Arizona and parts of adjacent States. The 72 alluvial basins of the study area represent virtually independent hydrologic units but can be categorized into five groups that have similar geohydrologic characteristics and similar requirements for model development. The groups—southeast, central, west, Colorado River, and highland—are composed mainly of adjacent basins that have similar climatic characteristics and altitude ranges. Models of 12 specific basins and 2 basin groups were developed to evaluate the current understanding of the basin flow systems and to explore the utility and applicability of general information-transfer techniques.

The basins are filled with variable thicknesses of alluvial deposits that constitute the aquifer system of the basin. A general sequence of sediments is present in the

basins and includes pre-Basin and Range rocks, lower and upper basin-fill units, and stream alluvium. The sediments that fill the basins generally are interconnected hydraulically. The part of the aquifer system that includes the principal water-bearing unit in a basin depends on the geohydrologic setting, which also influences basin categorization.

The basins are interconnected in a dendritic pattern much like the surface-drainage system and form an integrated regional-flow system. Individual basin-aquifer systems serve mainly as reservoirs for the storage of ground water with only small areas of interconnection and a small quantity of flow that occurs from basins of higher altitude to those of lower altitude.

Basin categorization is based on the relative importance of various components of the geologic and hydrologic systems. The principal geologic differences between groups are the presence and extent of fine-grained sediments in the basin center and the presence of saturated stream alluvium. These units represent the least and most productive parts of the aquifer, respectively. The principal hydrologic differences between groups are (1) magnitude of the component of mountain-front recharge and (2) presence of a perennial stream in the basin.

Ground-water modeling provides a means of improving the understanding of the geohydrology by allowing evaluation of alternative property values and distributions. Representative basins from each category were analyzed by using numerical models to test the concepts of categorization. Each model of a specific basin was developed on the basis of available data and previous conceptual or numerical models. Initial property values used in the models were based on available field information or, where no field data were available, were transferred from a basin or basins within the same geohydrologic category for which data were available.

Of the 12 basin models developed, all 12 were history matched to predevelopment steady-state conditions, and 10 were history matched to transient development conditions. The history-matching procedure for steady-state models depended on the most reliable data available. Models of basins having plentiful predevelopment water-level information were history matched by reducing the standard error between observed and model-calculated water levels to an acceptably small value. Model-calculated ground-water budget values were compared with corresponding budget values from previous field investigations where the measured values were available.

Hydrologic-property values derived from steady-state simulations were used subsequently in the transient models. Transient models were used to simulate water-level and flow-component changes for various periods of pumping stress; they also served as a means of evaluating the consistency of property values determined from steady-state history matching. The least-known values, such as specific yield, were varied within acceptable limits; the best-known values, such as pumpage, were not varied.

Among the specific topics investigated by ground-water simulation were (1) magnitude and distribution of mountain-front recharge, (2) spatial distribution of hydraulic conductivity and specific yield of the aquifer material, (3) relation between pumping and capture of natural discharge, and (4) recharge from deep percolation of excess applied irrigation water. Both specific-basin and basin-group models were used to analyze the significance of these topics to the local hydrologic setting and their effect on system response to development.

Because mountain-front recharge is a major water-budget component in the southeast basins, these models were used to test the general relation between precipitation and recharge that resulted from the regional water-budget analysis of all basins. The specific-basin models and basin-group model for the southeast basins were further used to analyze the effect of using a uniform spatial distribution of mountain-front recharge. The empirical relation between precipitation and recharge proved adequate for estimating the magnitude of mountain-front recharge, but because recharge does not occur uniformly along the basin margin, this approach to spatially distributing the recharge is not recommended. The quantity of mountain-front recharge in the west basins is small; therefore, models are insensitive to the exact magnitude and spatial distribution used in the simulation.

All basin models can be used to evaluate the effect of poorly defined spatial distributions of hydraulic conductivity and specific yield. Simulation results indicate that aquifer properties are highly variable laterally and that limited generalization is acceptable for modeling purposes. The relatively large magnitude of water-budget components in the southeast and Colorado River basins results in significant long-term capture of natural discharge by pumping. The components of evapotranspiration and discharge to springs and streams are most affected by capture.

The need to include recharge from deep percolation of excess applied irrigation water in simulations is indicated in 10 of the 12 specific-basin models. Such recharge becomes important after a lag time of a few years to several decades following initiation of irrigation. The overall result has been a change in the stress-strain relation over time; a decreased rate of decline has been documented in recent time in most basins, although the pumping rate has remained nearly uniform.

Basins of the southeast group for which ground-water flow models were developed are upper San Pedro, Benson, lower San Pedro, Willcox, and Douglas basins. These basin aquifers were simulated as two layers connected through a vertical-leakance value that represented the effects of anisotropy of the basin-center fine-grained facies. The most important properties used in the models of the southeast group include specific yield, interlayer leakance, the relation between evapotranspiration and depth to water, and pumpage quantity and location.

Basins of the central group for which ground-water flow models were developed are the Tucson basin and Avra Valley. Models of the west basins group were developed for Waterman Wash basin, Harquahala Plain, and McMullen Valley. These aquifer systems are most effectively simulated by using multiple layers. Recharge and discharge mechanisms are more important in the central group than in the west group. Storage depletion is common in both basin groups as a result of development, although some capture of discharge occurs in the central basins during the early phase of development.

Basins of the Colorado River group for which ground-water flow models were developed are the Parker Valley and Mohave basin. Because the hydrologic system of these basins is dominated by streamflow in the Colorado River, hydrologic properties related to shallow water-table conditions are most important. These include river-bed leakance, the relation between evapotranspiration and depth to the water table, and return of excess applied irrigation water. Adequate ground-water flow models of these basins can be developed by using a two-dimensional approach. Because of insufficient data, no specific-basin models were developed for the highland group, although these basins are thought to be scaled-down versions of the Colorado River basins.

Systematic model-sensitivity analyses were done on all models to qualitatively evaluate uncertainty in the definition of values of hydrologic properties and the effect on model results. A direct relation exists between basin category and sensitivity of the developed model to selected hydrologic properties. The specific properties depend on the degree of development, the geologic setting, and the magnitude of predevelopment downvalley flow. The transfer of information among basins of a particular group proved generally successful. The properties that are important to adequate modeling of the basins can be summarized into three groups—hydrologic boundaries, flow components, and hydrologic factors. Properties in the hydrologic-boundaries group represent the physical limits of the aquifer system, those in the flow-components group include recharge to and discharge from the aquifer, and those in the hydrologic-factors group control the movement and storage of water in the aquifer.

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