SUMMARY OF U.S. GEOLOGICAL SURVEY GROUND-WATER-FLOW MODELS OF BASIN-FILL AQUIFERS IN THE SOUTHWESTERN ALLUVIAL BASINS REGION, COLORADO, NEW MEXICO, AND TEXAS

By John Michael Kernodle

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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 that, in aggregate, underlie much of the country and that 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 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.

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CONVERSION FACTORS AND VERTICAL DATUM

Multiply	By	To obtain
inch	25.4	millimeter
foot	0.3048	meter
mile	1.609	kilometer
square mile	2.590	square kilometer
cubic foot	0.02832	cubic meter
acre-foot	0.001233	cubic hectometer
	43,560	cubic feet
per foot	3.281	per meter
foot per day	0.3048	meter per day
foot squared per day	0.09290	meter squared per day
cubic foot per second	0.02832	cubic meter per second
	448.8	gallons per minute
acre-foot per year	0.0013803	cubic foot per second
	0.6184	gallon per minute
gallon per minute	0.06309	liter per second
	0.002228	cubic foot per second
gallon per minute per foot	0.2070	liter per second per meter
	0.1247	cubic foot per second per foot
gallons per day per foot	7.48	gallons
ton	907.18	kilogram

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 the United States and Canada, formerly called Sea Level Datum of 1929.

SUMMARY OF U.S. GEOLOGICAL SURVEY GROUND-WATER-FLOW MODELS OF BASIN-FILL AQUIFERS IN THE SOUTHWESTERN ALLUVIAL BASINS REGION, COLORADO, NEW MEXICO, AND TEXAS

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ABSTRACT

Four ground-water-flow models of basin-fill aquifer systems in Colorado, New Mexico, and Texas were completed in conjunction with the Southwest Alluvial Basins study that is part of the Geological Survey's Regional Aquifer-System Analysis program. The modeled areas are the San Luis Valley in Colorado, the Albuquerque-Belen Basin and the Animas Valley in New Mexico, and the Mesilla Basin in New Mexico, Texas, and Mexico. These flow models and 10 additional models of basin-fill aquifer systems in areas of Colorado, New Mexico, and Texas are described in this report. The models are summarized to identify the common simulated hydrogeologic characteristics and to isolate preferred approaches to simulating ground-water flow in the basin-fill aquifer systems.

On the basis of attributes that are common to most of these models, a set of guidelines was developed that enables the rapid construction of ground-water-flow models of specific basin-fill aquifer systems. The feasibility of this modeling approach was tested and the guidelines were refined by developing test case generalized models of the Albuquerque-Belen and La Jencia-Socorro Basins. The generalized models met the objective of being adequate representations of both the function and the response of the basin-fill aquifer systems; this demonstrates the reliability of the guidelines. The guidelines for construction of a generalized ground-water-flow model of a specific basin within the Southwest Alluvial Basins region are as follows:

(1) Perform a literature search to determine basin geometry, geologic structure, and lithology.

(2) Use a three-dimensional model to simulate the aquifer to a depth of approximately 4,000 feet or to the total depth of the basin if less than 4,000 feet. Use at least five model layers, the top layer being 200 feet or less in thickness.

(3) Simulate the basin-fill aquifer system as having a horizontal hydraulic conductivity of 20 to 45 feet per day in the open-drainage basins and 2 to 10 feet per day in the closed-drainage basins, except where field data indicate otherwise. Simulate fine-grained playa or lake deposits as having a hydraulic conductivity of 0.25 to 10 feet per day and flood-plain alluvial deposits as having a hydraulic conductivity of 50 to 70 feet per day.

(4) Do not vary horizontal hydraulic conductivity as a function of depth unless specific lithologies are being simulated. Compaction of the aquifer and increases in temperature with depth need not be simulated as affecting the apparent hydraulic conductivity, except where these specific problems are being addressed. The two factors have opposite, and potentially offsetting, effects.

(5) Use a horizontal to vertical hydraulic-conductivity ratio of from 200:1 to 1,000:1 except where geologic features such as faults, clay sequences, or steeply dipping beds exist.

(6) Simulate aquifer specific storage to be in the range of 2×10^{-6} to 5×10^{-6} per foot and specific yield in the range of 0.10 to 0.20.

(7) Include rivers and drains, if present, in the simulations as head-dependent-flux boundaries, preferably with flow routing to allow the location of the boundary to change with time.

(8) Include estimated mountain-front and tributary recharge, evapotranspiration, and net irrigation flux.

(9) Include historical ground-water withdrawals.

The general models may be rapidly assembled yet retain an accuracy that is much greater than might be expected from the small expended effort. Appropriate uses of these general models are to aid in the design of a data-collection program customized to the needs of a specific study area, to make an initial evaluation of a specific problem, or to test hypotheses regarding the hydrologic responses in a basin. A generalized flow model needs to be viewed as a preliminary effort that will be superseded by a refined and calibrated model.

INTRODUCTION

The Southwest Alluvial Basins (SWAB) study is a part of the Regional Aquifer-System Analysis (RASA) program of the U.S. Geological Survey. The SWAB study area was divided, for administrative reasons, into two parts. The western part includes the southern tip of Nevada, the eastern part of California from Hoover Dam to the Mexican border, and the southern part of Arizona. The eastern part includes parts of southern Colorado, New Mexico, and west Texas (fig. 1). This report is a product of the eastern part of the SWAB study.

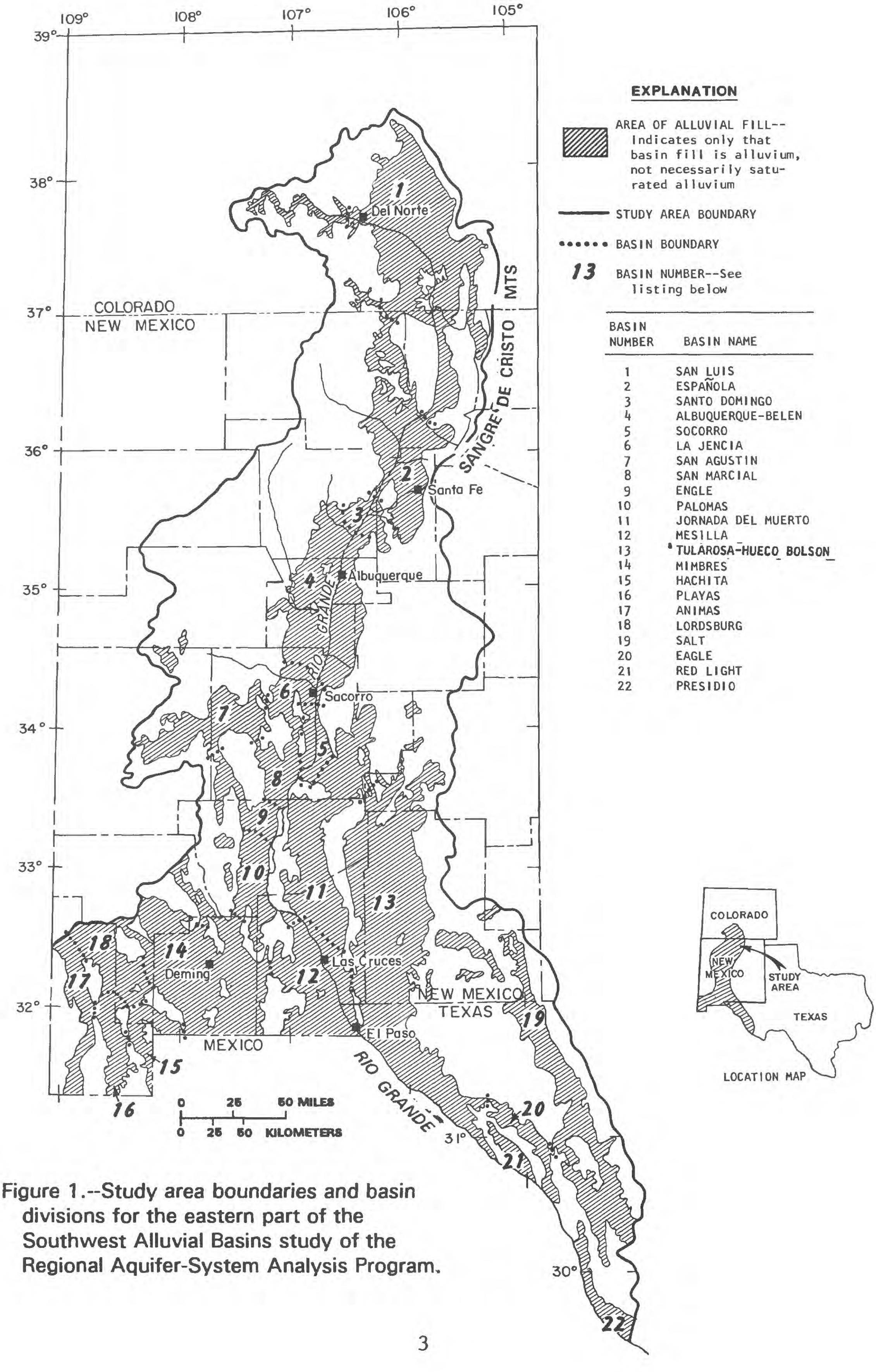


Figure 1.--Study area boundaries and basin divisions for the eastern part of the Southwest Alluvial Basins study of the Regional Aquifer-System Analysis Program.

Purpose and Scope

The main purposes of this part of the SWAB study were to enhance the understanding of the regional hydrology of the alluvial basins, in parts of Colorado, New Mexico, and Texas, and to study the hydrologic effects of stresses on the hydrologic system. Twenty-two alluvial basins were chosen for study within the area. Each study consisted of a literature review of the hydrology and geology of the basin, data compilation, data collection, data evaluation, and digital simulation of the basin hydrology where sufficient data were available. A planning report by Wilkins and others (1980) provides a more detailed description of the eastern part of the SWAB study.

The results of the SWAB study are described in Professional Paper 1407, which consists of three chapters. Chapter A is a summary of the project findings. Chapter B (this report) summarizes ground-water-flow models developed for the area and provides guidelines for developing generalized models applicable to basins within the study area. Chapter C describes the geohydrology and ground-water quality of the Mesilla Basin.

The purpose of this study is to identify specific hydrogeologic attributes that are common to alluvial-fill basins in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas. The approach taken to accomplish this objective is to review and summarize all of the documented ground-water-flow models of the U.S. Geological Survey in the study area. All models constructed in the area by the Survey are discussed and selected models constructed by the private sector are referenced.

In addition to the review and evaluation of existing models, four additional models were completed during the study. Each of the four additional models was used to test one or more approaches to modeling, and together they provided a core of knowledge about the hydrology of the alluvial-fill basins. The four flow models are the San Luis Valley (Hearne and Dewey, 1988), the Albuquerque-Belen Basin (Kernodle and Scott, 1986; and Kernodle and others, 1987), the Mesilla Basin (Frenzel and Kaehler, 1990), and the Animas Valley (O'Brien and Stone, 1983). Ten additional models, completed as part of other U.S. Geological Survey investigations, broaden the sample of modeling approaches. Therefore, the following review and summary of the 14 models are the basis of development of a unified overview of the hydrologic processes that take place in the basin-fill aquifer systems in the study area.

Location and Physical Setting of the Study Area

The eastern part of the SWAB RASA consists of part of south-central Colorado, the majority of central and southern New Mexico, and much of western Texas south of New Mexico (fig. 1). The basins in the western part of the SWAB RASA in Arizona and parts of California and Nevada were investigated by other workers (Anderson, 1980), and were not included in this study. El Paso, Tex., and Albuquerque, N. Mex., are the two largest cities in the study area, but immediately across the Rio Grande from El Paso is Ciudad Juarez, Mexico, a co-user of the water resources of the area.

A wide range of climatic conditions occurs in the study area as a result of the variability in altitude, latitude, and aspect. Altitude in the study area ranges from about 2,600 feet above sea level at Presidio, Tex., to more than 14,000 feet above sea level in the Sangre de Cristo Mountains in Colorado. Annual average precipitation ranges from less than 8 inches at low altitudes in the southern basins to more than 40 inches at high altitudes in the Sangre de Cristo Mountains in Colorado. Temperature and humidity reach comparable extremes and result in life zones ranging from Sonoran desert to alpine.

Although precipitation is scanty at low altitudes, the potential annual evapotranspiration can be as much as 6 feet of water. This imbalance is particularly significant along the Rio Grande, which progressively loses flow (primarily due to irrigation diversions and natural evapotranspiration) almost immediately after leaving the San Juan Mountains in Colorado until it eventually empties into the Gulf of Mexico. Before the construction of reservoirs, which began in the early 1900's, the Rio Grande channel frequently was dry in the Mesilla Basin upstream from El Paso, Tex.

Alluvial basins in the study area are geologically associated with the Rio Grande rift. The basins are bounded by faults and are filled with sediments derived both locally from adjacent uplifted areas and, for basins along the Rio Grande, from upstream areas. The depth of basin fill may exceed 15,000 feet (San Luis Valley and Albuquerque-Belen Basin), and total structural relief frequently exceeds 20,000 feet. Most basins are shallower, however, having fill thickness less than 10,000 feet and commonly in the range of 2,000 to 3,000 feet. The fill material, identified as the Santa Fe Group of Cenozoic age, generally is unconsolidated and consists of fine-grained playa and lacustrine deposits, conglomerates of alluvial-fan origin, fine sands of eolian origin, fluvial gravels, and lava flows, all of which may be interbedded at various scales.

There are two general types of basins: those through which surface streams flow and those with a closed surface-water drainage system. Some alluvial basins have both through-flowing streams and areas of closed drainage.

The basins along the Rio Grande support extensive irrigated agriculture, primarily in the immediate area of the river flood plains and major tributaries. Evapotranspiration by native vegetation and by agricultural crops accounts for a substantial part of the water budget of these basins. Withdrawal of ground water for municipal, industrial, or agricultural use may sufficiently lower ground-water levels in flood-plain or playa areas enough to salvage water previously lost to evapotranspiration.

Ground-water quality in basins having a through-flowing river generally is acceptable for human consumption, although there are local exceptions, especially in the southernmost basins. Saline water tends to be flushed out of the ground-water system by ground-water discharge to drains, canal leakage, and exchange of water with the through-flowing river. However, there commonly is a deep ground-water-flow system that causes an upwelling of mineralized water at the lower end of the basins. The volume of this upwelling may be small, however, depending on basin dimensions and hydraulic properties of the basin-fill material.

In basins with closed surface drainage and little, if any, ground-water outflow, dissolved minerals are concentrated in ground water near the center of the basin. Shallow ground-water levels in parts of these basins have caused and, in some basins, still cause large losses to evapotranspiration. Evapotranspiration, without a mechanism to flush the remaining salts, results in a body of brackish or saline ground water near the topographically low areas of the basin. Fresh ground water may occur only at the margin of the basin, recharged by infiltration of surface runoff from bordering mountains. However, if there is ground-water outflow from the basin, some of the dissolved salts may be flushed from the basin. Water quality may also be influenced by geothermal activity or by ground-water inflow from adjacent areas, regardless of the surface-water/ground-water relation.

DISCUSSION OF GROUND-WATER-FLOW MODELS

The term "model," even when restricted to the field of hydrology, has a confusing range of meanings. In the most restricted sense, a model is a replica or copy of an original, preserving either the appearance and function of that prototype, or both. A hydrologic example of this type of model is a device known as a sand tank, in which dyes and piezometer tubes may be used to trace the flow of water through mixtures of porous material under various conditions. This type of model has limited practical application because of the difficulty in matching the complex conditions found in most aquifer systems.

At a slightly greater level of abstraction, a model may imitate or mimic rather than duplicate the behavior of the original. A good example of this type of model is known as an electric-analog model (several of which are described in this report). An analog model uses one physical system to describe, by analogy, the behavior of another physical system. The flow of electricity is analogous to the flow of water. The flow of water through an aquifer system can be mimicked using an array of resistors, capacitors, and other electronic components in such a way that voltage correlates with potential or head, current correlates with flow of ground water, capacitance correlates with aquifer storage, and resistance correlates with the inverse of aquifer hydraulic conductance. Because of several factors, analog models have become much less common than digital models, described later. Among these factors are the inflexibility of analog models (simulated aquifer properties cannot be easily adjusted), the inability to simulate transmissivity as a function of saturated thickness, the space needed to store the model and the associated electronic equipment, and the level of electronics skill needed to assemble and maintain the model and its related equipment.

The next level of abstraction is to describe the processes taking place in an aquifer with one or more mathematical expressions. For any but the most simple problems that have direct analytical solutions, the number of computations becomes forbidding for a manual solution; therefore, digital computers are essential to complete the task. These digital models use various numerical techniques to solve or approximate a solution to the equations that describe ground-water flow. The equations of ground-water flow can be written for one to three dimensions, and for conservation of mass or momentum, or both. Historically, limitations on the complexity of the problems that can be solved have been the speed, memory, and cost of computers rather than the lack of knowledge necessary to formulate and solve the equations. As computer technology has improved, digital models are increasingly able to portray complex hydrologic conditions. However, a computer program is not itself a ground-water-flow model. The properties of a specific aquifer system need to be numerically described in a manner that accurately captures the essence of the flow system. This description or interpretation of the flow system frequently is referred to as the conceptual model of the ground-water-flow system. Confusion arises because the program is referred to as a model; there is also a conceptual model, and the union of the two is referred to as a ground-water-flow model of a specific aquifer or aquifer system.

One final type of model referenced later is known as a lumped-parameter model. These models usually describe the system by use of response functions, which may be based on theoretical development or, at the other extreme, are empirical. The lumped-parameter models are particularly useful in determining time-dependent changes in water budgets or water quality but are limited in their ability to describe processes outside the range of the response functions used in their development.

The following sections describe U.S. Geological Survey ground-water-flow models that have been completed for alluvial basins in parts of Colorado, New Mexico, and Texas. The locations of these modeled areas are shown in figure 2. References are also included for selected non-U.S. Geological Survey models. Finally, a section is included that summarizes the results of all modeling efforts in the study area and discusses the feasibility of constructing generalized flow models for specific basins.

The names of the basins used in the topic headings and text conform to common usage and the usage in the cited reports. Therefore, the area of investigation may be referred to as a valley, a basin, or a bolson. Regardless of the nomenclature, all areas are fault-controlled structural basins.

Virtually all of the models were constructed with the broad objective of providing additional knowledge of the hydrology of the alluvial basins. Most of the models also address problems very specific to the basin under investigation. As of this date (1985), there are no documented Survey models of basins in the study area that simulate solute transport or the effects of temperature or fluid density on ground-water flow.

Each basin is briefly described. Aspects of geology, hydrology, climate, evapotranspiration, water use, and water-related problems, if any, are discussed. Models are discussed and cited chronologically within each basin, and the basins are discussed in downstream order along the Rio Grande, then from west to east for the remaining basins. Virtually all of the observations presented in this report are from the references cited.

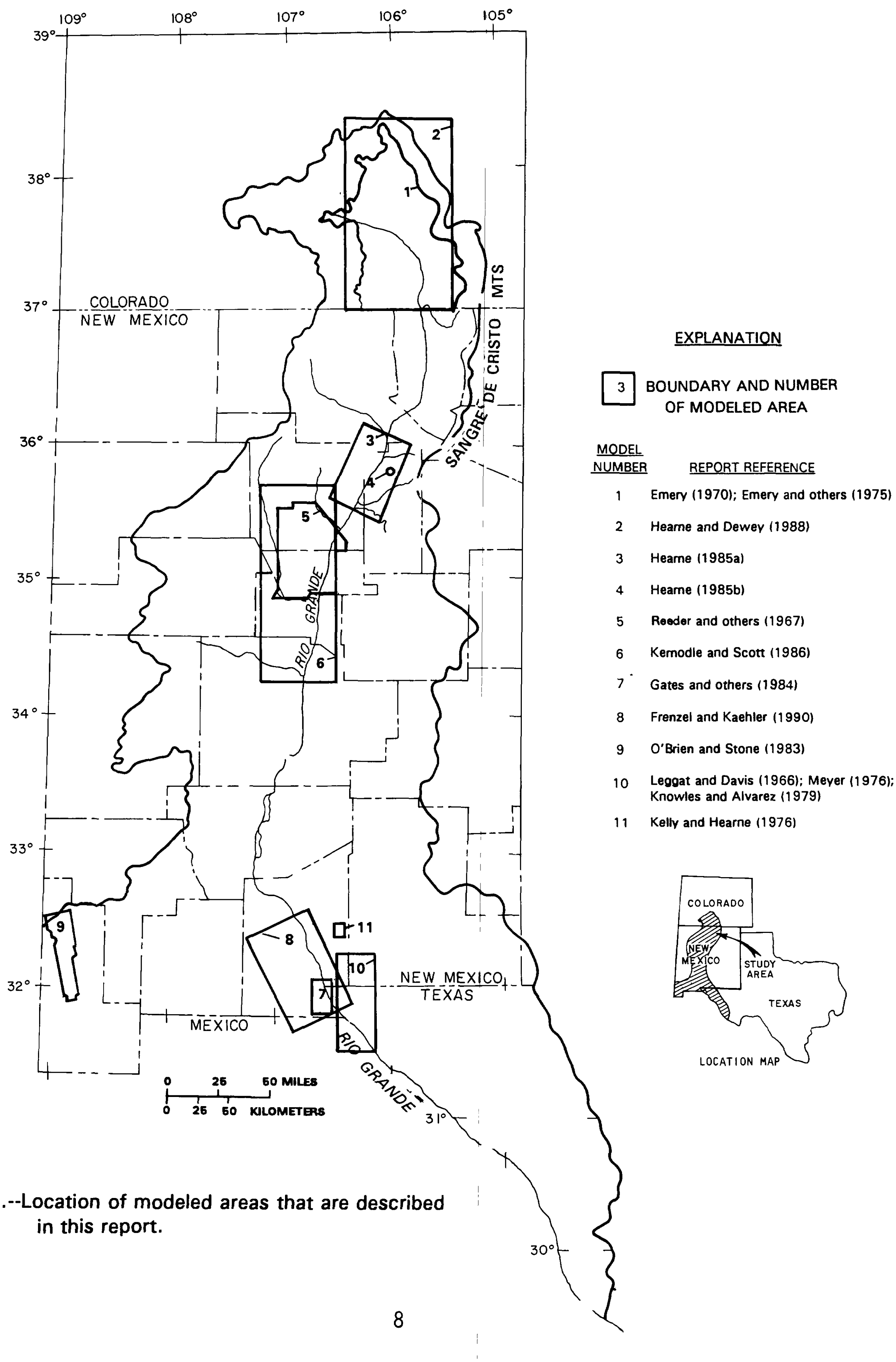


Figure 2.--Location of modeled areas that are described in this report.

San Luis Valley

The San Luis Valley is the northernmost of the large Rio Grande rift basins (fig. 1). It is unusual in that it contains the through-flowing Rio Grande, yet the northern half of the valley has closed surface-water drainage. The ground-water system is, however, continuous and unbroken between the two surface regions. Another significant hydrologic feature is a thick and persistent clay sequence that separates the aquifer system into shallow (unconfined) and deep (confined) components. A characteristic of the valley, common to many closed-drainage basins, is a shallow depth to ground water in the area of the central depression.

In spite of the high altitude and short growing season in the valley, crop productivity is high, due mainly to extensive irrigation. One-half of all water available for use in the Rio Grande drainage area is consumed by evapotranspiration in the San Luis Valley. As a direct consequence of land- and water-use patterns, soil waterlogging and high rates of evapotranspiration are persistent problems. A plan has been proposed to lower the water table by pumping ground water from wells and conveying the water to the Rio Grande through pipes or canals. The lowered water table is intended to salvage water previously lost to evapotranspiration, and the salvaged water will, among other benefits, be a credit in Colorado's required downstream delivery of water. One of the key issues in management of the water resources in the San Luis Valley is the relative amount of ground water that pumpage would salvage from loss to evapotranspiration compared to the amount that the pumpage would deplete from storage or capture from surface water. Most of the ground-water-flow models of the basin were developed to address this issue.

Two-Dimensional Analog Model

Emery (1970) designed a two-dimensional electric-analog model of the unconfined aquifer in the San Luis Valley. The extent of the modeled area coincides with the area of alluvial fill in the structural basin, except that the model terminated at the Colorado-New Mexico State boundary. The purpose of this model was to determine the probable amount of water that could be salvaged from evapotranspiration by lowering the ground-water levels.

The model simulated as much as 120 feet of the alluvial fill, in which transmissivity ranged from 10,000 to 200,000 gallons per day per foot but predominantly was 20,000 to 50,000 gallons per day per foot. Therefore, on the basis of an assumed thickness of 120 feet, most of the aquifer was simulated as having values of hydraulic conductivity of 22 to 56 feet per day. The specific yield was simulated as 0.20 (dimensionless). The surface-water systems (Rio Grande and Conejos River) were simulated as constant-head boundaries.

Emery used the model to evaluate a water-salvage plan proposed by the U.S. Bureau of Reclamation (1963). In his evaluation, Emery (1970, p. 1) found that after 50 years of pumping 84,000 acre-feet per year "the major part of the water pumped is derived from salvaged ground water that otherwise would have been lost to nonbeneficial evapotranspiration (84 percent) and the remainder is from ground-water storage (14 percent) and the Rio Grande (2 percent)." He also pointed out that because the Rio Grande was likely to have a poorer hydraulic connection with the aquifer than simulated, streamflow depletion was probably overestimated. One interesting innovation of the model was the use of an evapotranspiration versus depth-to-water function, variations of which are now a common occurrence in ground-water-flow model simulations.

Three-Dimensional Analog Model

The electric-analog model documented by Emery and others (1975) was a three-dimensional enhancement of the one previously described by Emery (1970). Their report (Emery and others, 1975, p. 2) "describes how the analog model was used to help describe the present hydrologic conditions in the valley, to predict the effects of continuing present water-use practices, and to predict the effects created by changing use."

The analog model consisted of three layers that represented the basin-fill aquifer system in the San Luis Valley. The areal extent of the modeled aquifer was the same as in the earlier two-dimensional model, as was the thickness (as much as 120 feet) of the top, unconfined layer. The other two underlying layers, representing the confined aquifer, were each 1,500 feet thick, for a total modeled thickness of 3,000 to 3,120 feet. Vertical hydraulic conductivities were used to couple the three layers and to simulate the extensive clay sequence between the confined and unconfined aquifers.

The investigators used a two-dimensional digital model in support of analytical methods to determine the vertical hydraulic conductivity of the clay "confining bed" between the upper confined layer and the unconfined layer. In this model "the configuration and altitude of the water table in the unconfined aquifer and the transmissivity of the unconfined and confined aquifers were held constant, while various values for vertical leakage were tried until a 'match' of the actual and computed potentiometric surfaces in the confined aquifer was obtained" (Emery and others, 1975, p. 9). The vertical hydraulic conductivity of the clay confining bed determined by this method was 0.059 foot per day.

The simulated hydraulic properties of the three-dimensional analog model are shown in figure 3. Although a wide range in transmissivity is shown, the horizontal hydraulic conductivity throughout most of the model was approximately 27 feet per day (not shown). The simulated vertical hydraulic conductivity of the clay sequence was 0.059 foot per day everywhere except where lava flows are intercalated with basin-fill deposits in the upper confined layer, where a vertical hydraulic conductivity of 0.00059 foot per day was simulated (not shown). Also, a vertical conductivity of 59 feet per day was simulated along a fault zone adjacent to the San Luis Hills. The vertical hydraulic conductivity between the two lower confined layers was modeled as 0.0134 foot per day, resulting in a ratio of horizontal to vertical hydraulic conductivity of 2000:1. The unconfined aquifer was modeled as having a specific yield of 0.20 and the two confined layers as having a specific storage of 5×10^{-6} per foot. Both the Rio Grande and Conejos River were simulated as specified hydraulic head having a "restricted" connection with the top, unconfined layer of the model.

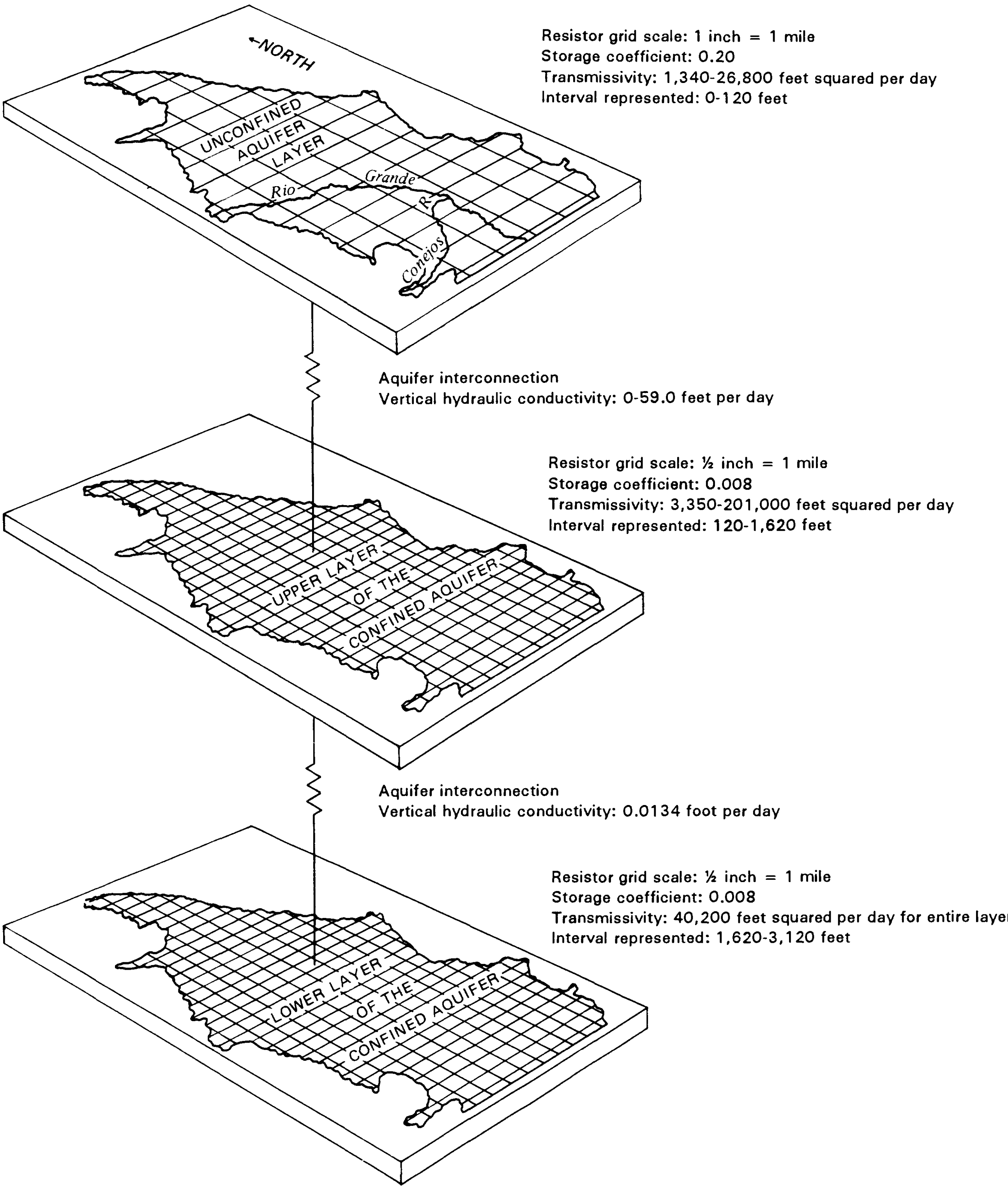


Figure 3.--Diagrammatic sketch of the three-dimensional electric-analog model of the basin-fill aquifer in the San Luis Valley. [Modified from Emery and others, 1975, fig. 4].

Emery and others (1975) used this three-dimensional model to simulate historical pumpage from the upper confined aquifer and ground-water-level changes for 1950-70 and to evaluate future ground-water-withdrawal scenarios. During these analyses the authors demonstrated that withdrawals south of the Rio Grande have a 50 times greater impact on streamflow in the Conejos River than withdrawals north of the Rio Grande. They also evaluated other scenarios but not the U.S. Bureau of Reclamation (1963) salvage plan, proposing that this could be one of the future uses of the model. The three-dimensional analog model of ground-water flow has been converted to a digital model by Leonard and Watts (1989).

Three-Dimensional Finite-Difference Model

The most recent documented model of the San Luis Valley (Hearne and Dewey, 1988) was completed as part of this RASA study. The location and finite-difference grid of the three-dimensional model are shown in figure 4. The purposes of the investigation and model were to develop an understanding of and describe the hydrologic processes in the basin-fill aquifer system in the San Luis Valley or Alamosa Basin, and to develop the ability to estimate the effects of future development on the aquifer system.

The investigators depended heavily on the results of the previous models, with two notable exceptions. First, they developed and documented a method for estimating mountain-front and tributary recharge to the aquifer system (this technique was then used in all ground-water-flow models completed in this study). Second, they used a two-dimensional cross-sectional model to test sensitivity to model cell dimensions and to the number of model layers and depth of simulation required of a three-dimensional model to obtain satisfactorily accurate results.

Other significant departures from previous models are: (1) the geometry of the basin was more closely matched, both in depth as well as in areal extent, by truncation of the model at the San Luis Hills hydrologic barrier, and (2) the investigators chose to simulate changes in stress and potentiometric-head response rather than absolute stresses and hydraulic heads.

The seven-layer, three-dimensional model simulated a total thickness of 3,200 feet of basin fill, including the upper unconfined aquifer, the clay sequence, and upper and lower parts of the confined aquifer. The two-dimensional cross-sectional model demonstrated that this simplification of the total basin depth (which exceeds 17,000 feet) caused a 2-percent error in computed hydraulic head at a deep index cell if the ratio of horizontal to vertical hydraulic conductivity was 700:1. The two-dimensional model also demonstrated that a surface dimension of 2 miles per cell side produced acceptable results.

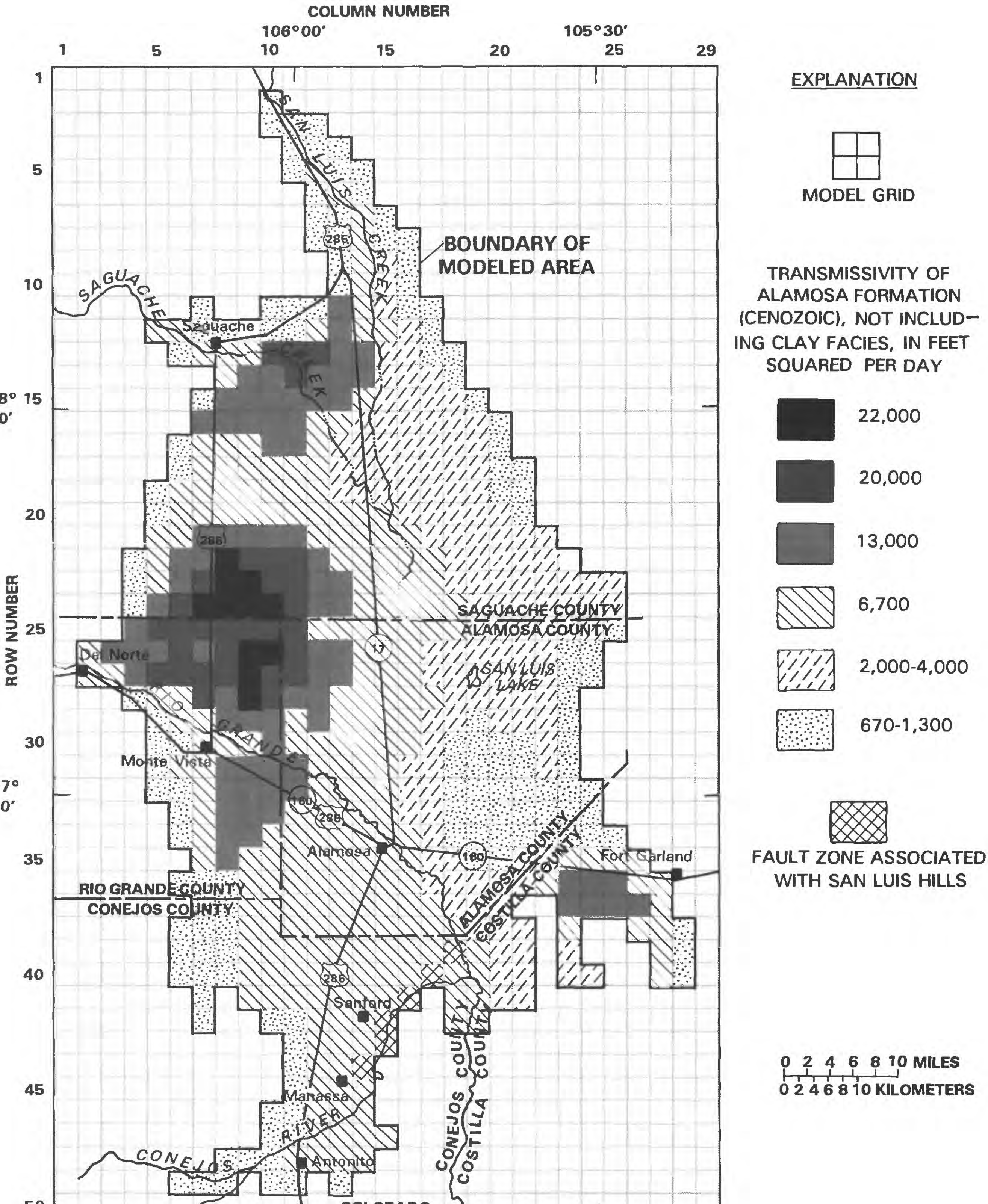


Figure 4.--Location, finite-difference grid, and transmissivity distribution of the top layer of the model of the basin-fill aquifer in the San Luis Valley. [Modified from Hearne and Dewey, 1988, fig. 16].

The distribution of simulated hydraulic conductivity along column 11 in the model is shown in figure 5. The simulated transmissivity of the top layer approximates that of Emery and others (1975, pl. 1) with calculated hydraulic conductivity ranging from 25 to 450 feet per day and a nominal layer thickness of 150 feet and saturated thickness of 100 feet. A significant part of the top layer had a modeled horizontal hydraulic conductivity of about 40 feet per day. The simulated specific yield was 0.20.

The underlying clay sequence was simulated as having a horizontal hydraulic conductivity of 10 feet per day and a vertical hydraulic conductivity of 0.06 foot per day (a horizontal to vertical ratio of about 170:1). The specific storage for this layer and the lower layers was 5×10^{-6}.

The upper part of the confined aquifer was simulated as having a horizontal hydraulic conductivity of 40 feet per day and a vertical hydraulic conductivity of 0.06 foot per day. The ratio of horizontal to vertical hydraulic conductivity was about 670:1. In the southwest part of the basin, where lava flows are intercalated with alluvium, the vertical hydraulic conductivity was modeled as 6×10^{-4}, resulting in a horizontal to vertical ratio of about 67,000:1.

The lower part of the confined aquifer was simulated as having a horizontal hydraulic conductivity of 30 feet per day, and a vertical hydraulic conductivity of 0.013 foot per day, resulting in a horizontal to vertical ratio of about 2,300:1. In the area adjacent to the San Luis Hills, the vertical hydraulic conductivity was increased to 60 feet per day to represent a highly conductive fault zone.

Flow to and from the Rio Grande and Conejos River was assumed to be a function of the difference in hydraulic head between the aquifer and the river and of the leakance and area of the streambed. The leakance times the fractional area that the streambed occupies in the model cell yielded a constant of proportionality of 0.00012 per day, which closely correlates with a value of 0.00071 per day obtained in an analog- to digital-model conversion discussed by Leonard and Watts (1989).

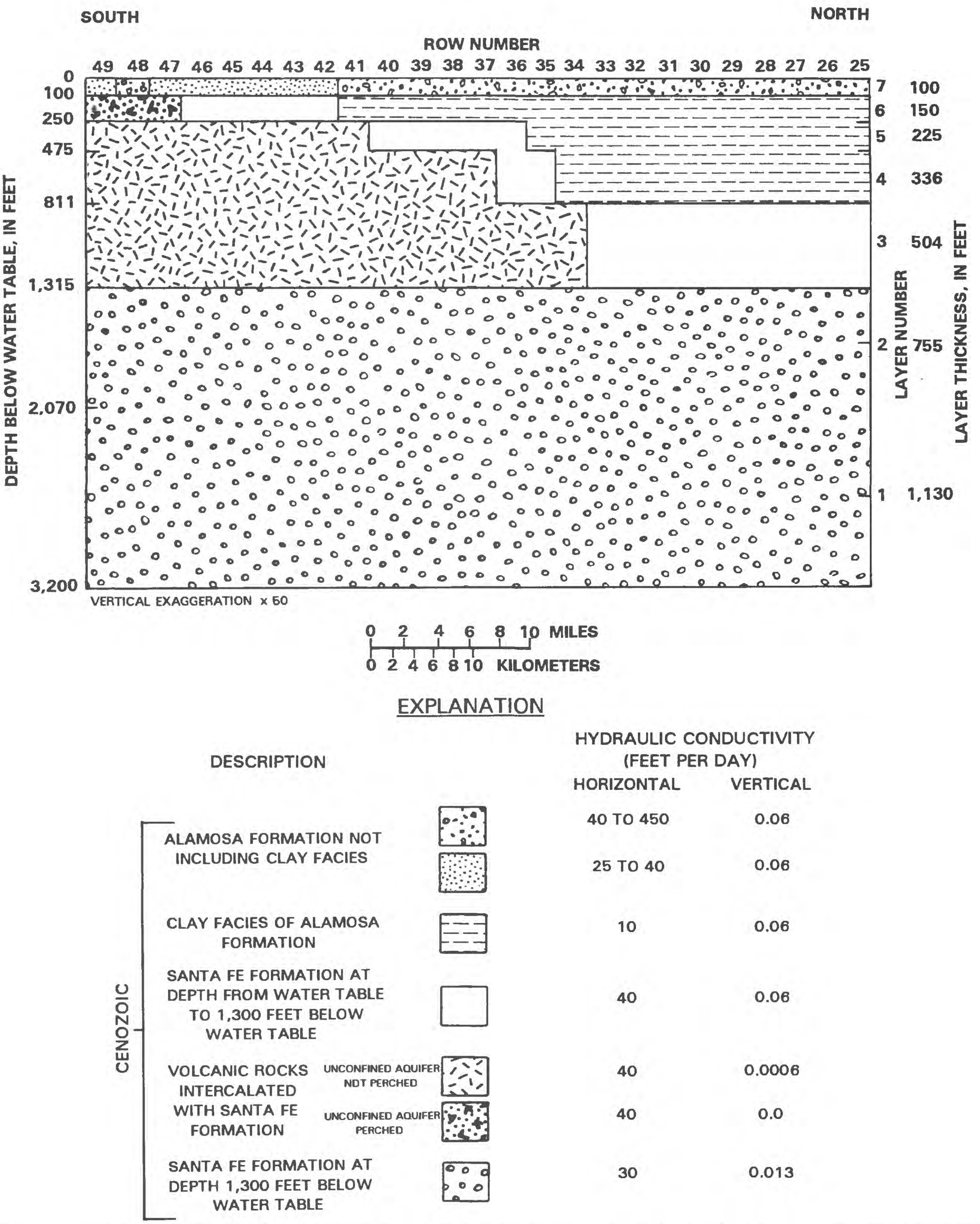


Figure 5.--Section along column 11 of the model of the basin-fill aquifer in the San Luis Valley showing layer thicknesses and distribution of simulated hydraulic conductivity. [Modified from Hearne and Dewey, 1988, fig. 18].

None of the simulated aquifer properties depart significantly from those of Emery and others (1975). The only simulated aquifer property altered during the calibration process was the constant of proportionality of the Conejos River streambed. Hearne and Dewey (1988, p. 94) described the process of calibration that they employed:

> To ensure that the assumptions made in developing the model were consistent with each other and with available data, the simulated response was compared with the measured response, and the assumptions were modified to improve the comparison--a process called "calibration." Commonly the stress is assumed to be known, and calibration results in revised estimates of aquifer and boundary characteristics. Although the results of calibration may not be unique, the calibrated model commonly is assumed to better represent the prototype. However, for the three-dimensional model of the Alamosa Basin, stresses (withdrawals for irrigation and return flows from irrigation) were assumed to be less well known than the aquifer characteristics. Therefore, calibration resulted in revised estimates of stress rather than revised estimates of aquifer characteristics.

Thus, the investigators expressed greater confidence in their understanding of the aquifer system than in the quantification of the stress applied to the aquifers in the system.

There are several significant findings from their investigation. (1) The aquifer system adjusted to a change in stress within 10 years, which the authors felt was a relatively brief time. (2) For a wide range in aquifer-system characteristics, 69 to 82 percent of all withdrawn water was derived from salvaged evapotranspiration. For 1980, salvaged evapotranspiration was computed to account for 80 percent of the pumped withdrawal, aquifer storage for 14 percent, and streamflow capture for 6 percent. (3) Truncating the simulated depth at 3,200 feet rather than simulating the entire 17,000-foot thickness produced an estimated numerical error of only 2 percent in computed hydraulic heads at selected index cells in the flow model.

Española Basin

The Española Basin lies south of the San Luis Valley and northeast of the Santo Domingo Basin (fig. 1). The through-flowing Rio Grande (fig. 6) enters the basin through the Embudo constriction and exits through White Rock Canyon as it crosses La Bajada fault. The structural basin is 25 miles north to south and 40 miles east to west even though the topographic basin is only 20 miles wide (Manley, 1978, p. 201). Manley (1978) placed the western boundary at the Nacimiento uplift and thereby included virtually all of the Jemez caldera and volcanic complex. Another investigator (Baltz, 1978, p. 212) described the basin as "a synclinal sag whose western limb is broken locally by the Pajarito fault zone and other faults." The Pajarito fault zone is north of and aligns with La Bajada fault and cuts the eastern flank of the Jemez Mountains. Displacement along these faults may be several thousand feet. The total depth of the basin may reach 7,000 feet.

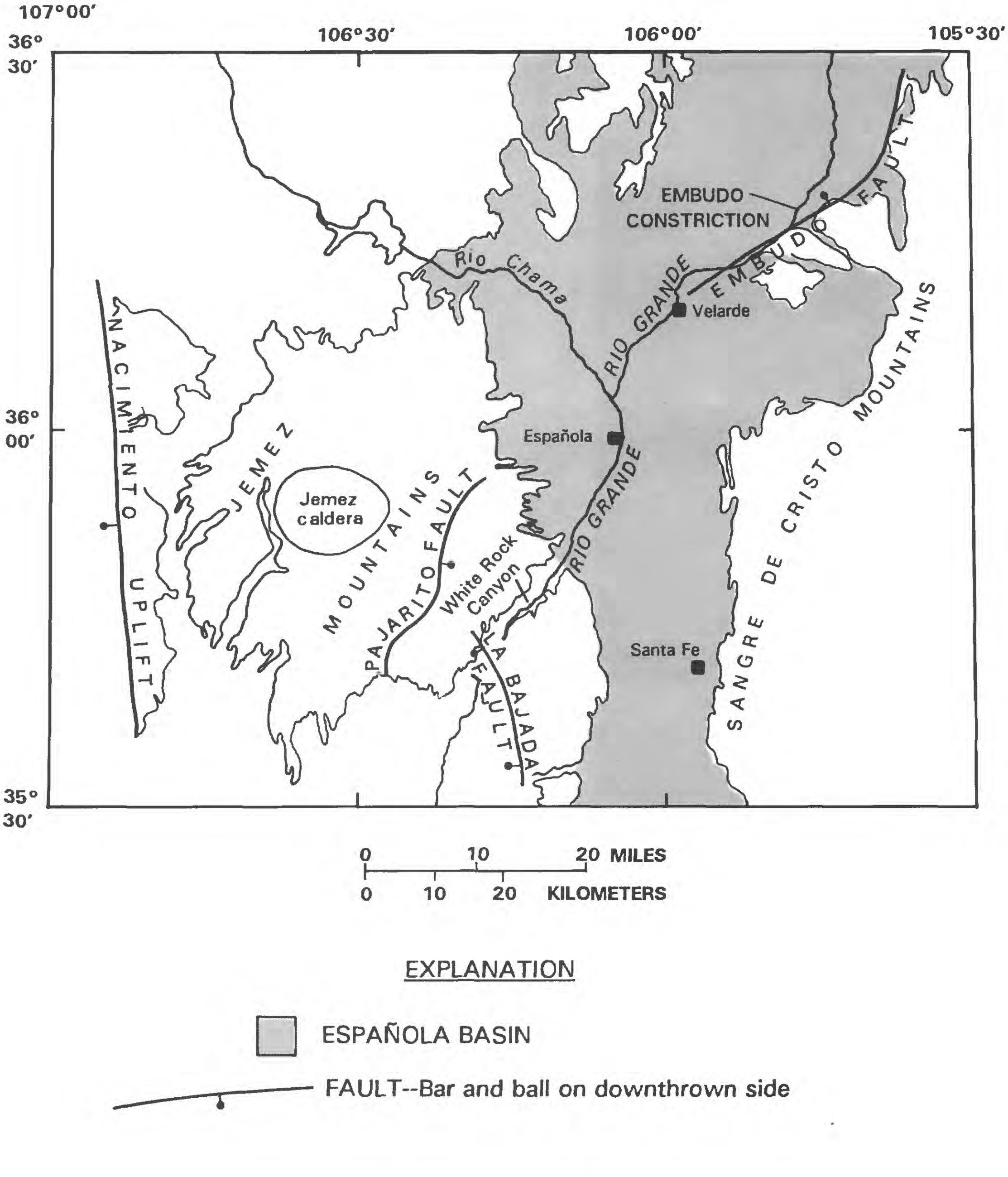


Figure 6.--Selected features in the vicinity of the Española Basin.

The mapped units of the Santa Fe Group (Miocene to Pleistocene) in the Española Basin are the Tesuque, Ancha, and Puye Formations. The Española Basin differs from others in the Rio Grande rift by having recognizable and persistent dipping beds truncated at frequent intervals by block faulting. The basin also differs in that the basin-fill deposits are generally fine grained and tuffaceous, resulting in reduced hydraulic conductivity of the aquifer system.

Withdrawal of ground water to supply the municipal and industrial requirements of the communities in the Española Basin is the major stress on the basin-fill aquifer system. The effect of these withdrawals on surface-water availability is a subject of concern in the basin. Several investigations have been undertaken to quantify these effects.

Aquifer-Test Analysis Using a Three-Dimensional Finite-Difference Model

Hearne (1985b) used a three-dimensional digital model (Posson and others, 1980) to evaluate the results of a 13-day test of the Tesuque aquifer system in the Tesuque Formation on the Tesuque Pueblo. During the test 320 gallons per minute were withdrawn from the production well and water levels were measured in that well and in 14 piezometers completed above, below, and in the producing intervals.

Prior to the test, the author used an analytical procedure to demonstrate that the ratio of horizontal to vertical hydraulic conductivity was about 250:1 under natural ambient head-gradient conditions. Geophysical logs were used to estimate the relative horizontal hydraulic conductivity and porosity of the individual beds, which were observed to dip at 7 degrees toward the northwest.

The vertical arrangement of the 15 model layers employed in the final simulation is shown in figure 7. The section is perpendicular to the strike of the beds through an assumed plane of symmetry used to reduce the size of the model by one-half. The relation between model layers, aquifer materials, screened intervals in the pumped well, and piezometer completions along a section perpendicular to the strike of the beds is shown in figure 8. The model was extended 2 miles along the strike of the beds to attain sufficient distance from the pumped well to avoid drawdown at artificial boundaries. The vertical column of cells representing the production well had surface dimensions of 1 by 2 feet. The cells in the column were coupled in the transient simulation by simulating a vertical leakance 1 million times greater than the surrounding aquifer system. In both the prestress (or initial) condition and the transient simulations, computed hydraulic heads in the model were constrained by four rows of specified-head cells (Hearne, 1985b, p. 18).

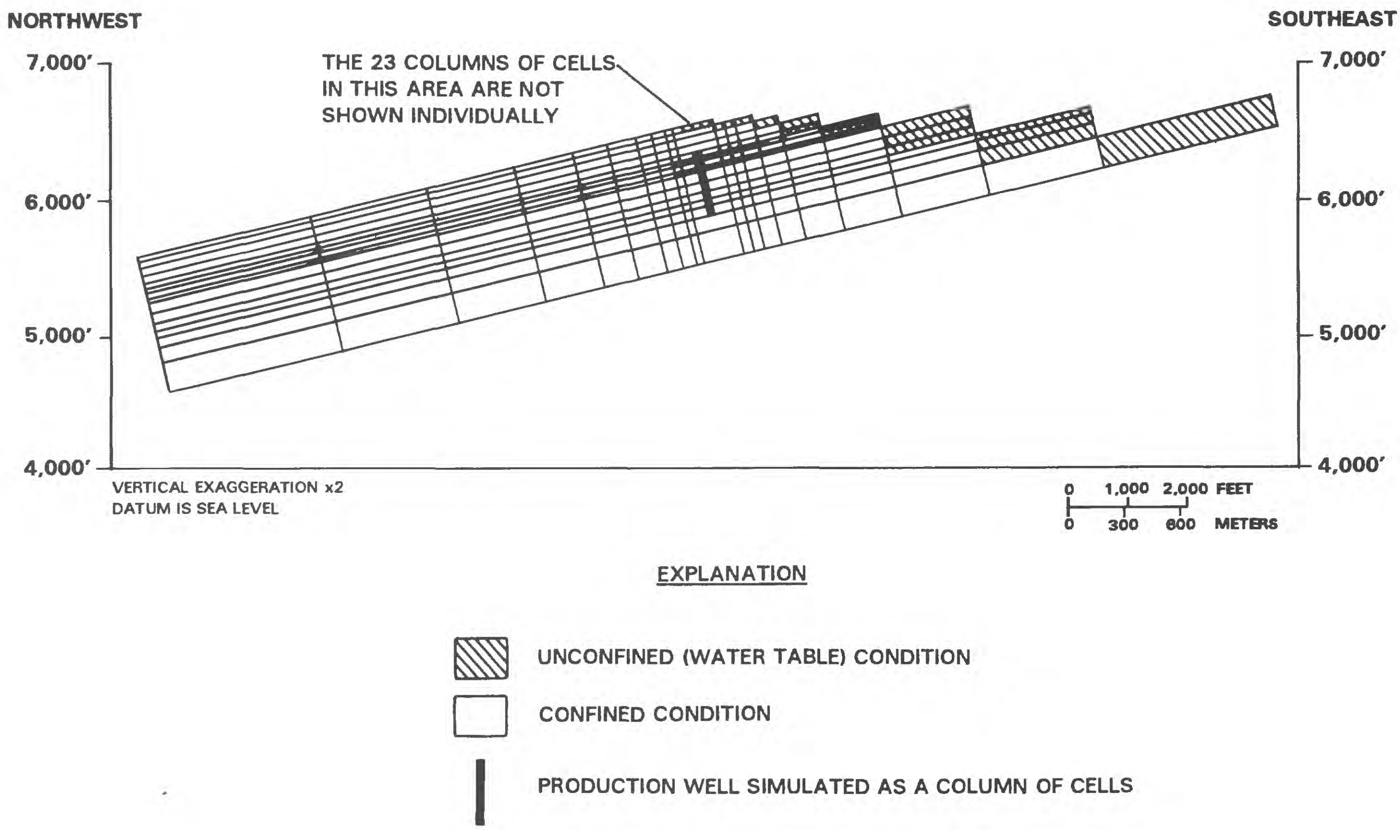


Figure 7.--Orientation of layers in the model used to evaluate the results of a test of the Tesuque aquifer system in the Española Basin. [Modified from Hearne, 1985b, fig. 12].

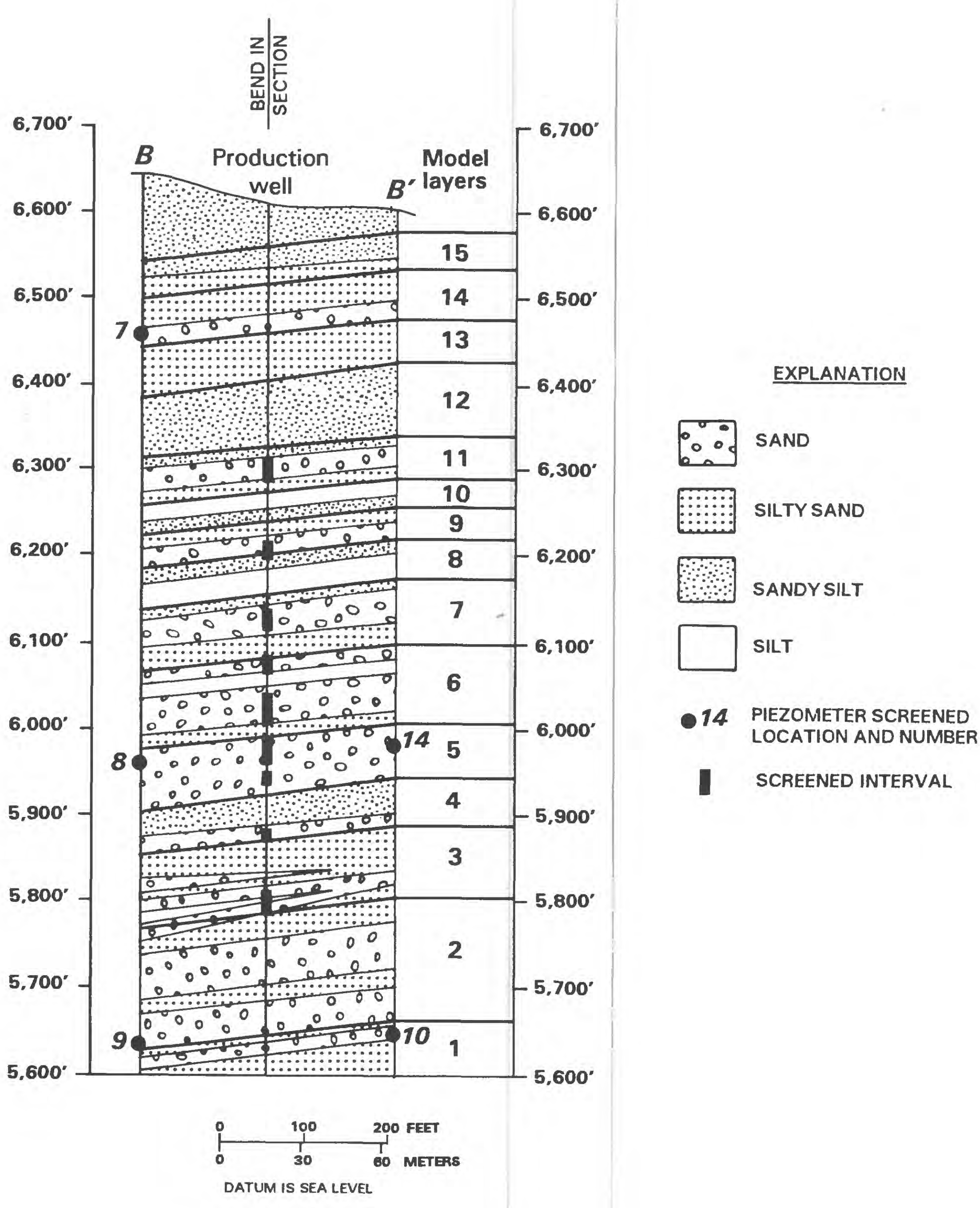


Figure 8.--Correlation between the layers in the model and the beds at the test site of the Tesuque aquifer system in the Española Basin. [Modified from Hearne, 1985b, fig. 13].

The final transient model used simulated horizontal hydraulic-conductivity values ranging from 0.30 to 2.7 feet per day. The average hydraulic conductivity was about 2 feet per day. The ratio of horizontal to vertical hydraulic conductivity ranged from 750:1 to 50,000:1. The harmonic mean of hydraulic conductances yielded a ratio of horizontal to vertical hydraulic conductivity of about 20,000:1 for the entire aquifer sequence. Specific storage was simulated as 2×10^{-6} per foot for all layers, and specific yield as 0.15 at the free-water surface. However, Hearne (1985b, p. 17) pointed out that the model was insensitive to specific yield. An acceptable match between measured and computed hydraulic heads was not obtained until two vertical impermeable barriers were simulated in the lower seven layers, one at a distance of 1,000 feet downdip and the other 2,000 feet updip from the production well.

Hearne (1985b, p. 22) attributed the discrepancy in anisotropy ratios determined for prestress and pumping conditions to "the discontinuity of less permeable beds" which allows "a tortuous path around these beds." On a long-term and regional scale that existed prior to the test, the regional ratio approached 250:1. On the short-term and local scale of the aquifer test, the local anisotropy dominated the regional discontinuities and resulted in a ratio of 20,000:1. For this and other reasons, Hearne (1985b, p. 22) warned that "Extrapolation of aquifer characteristics from a particular site to the whole basin should be done with care."

Three-Dimensional Finite-Difference Model

Hearne (1985a) documented the results of an areal three-dimensional finite-difference model of the part of the Española Basin centered about the Pojoaque River basin. The purpose of the investigation and model was to evaluate the effect of proposed ground-water withdrawals from the Tesuque aquifer system (to support irrigated agriculture) on ground-water levels and flow in streams within the Española Basin. The location of the area of investigation and modeled area are shown in figure 9.

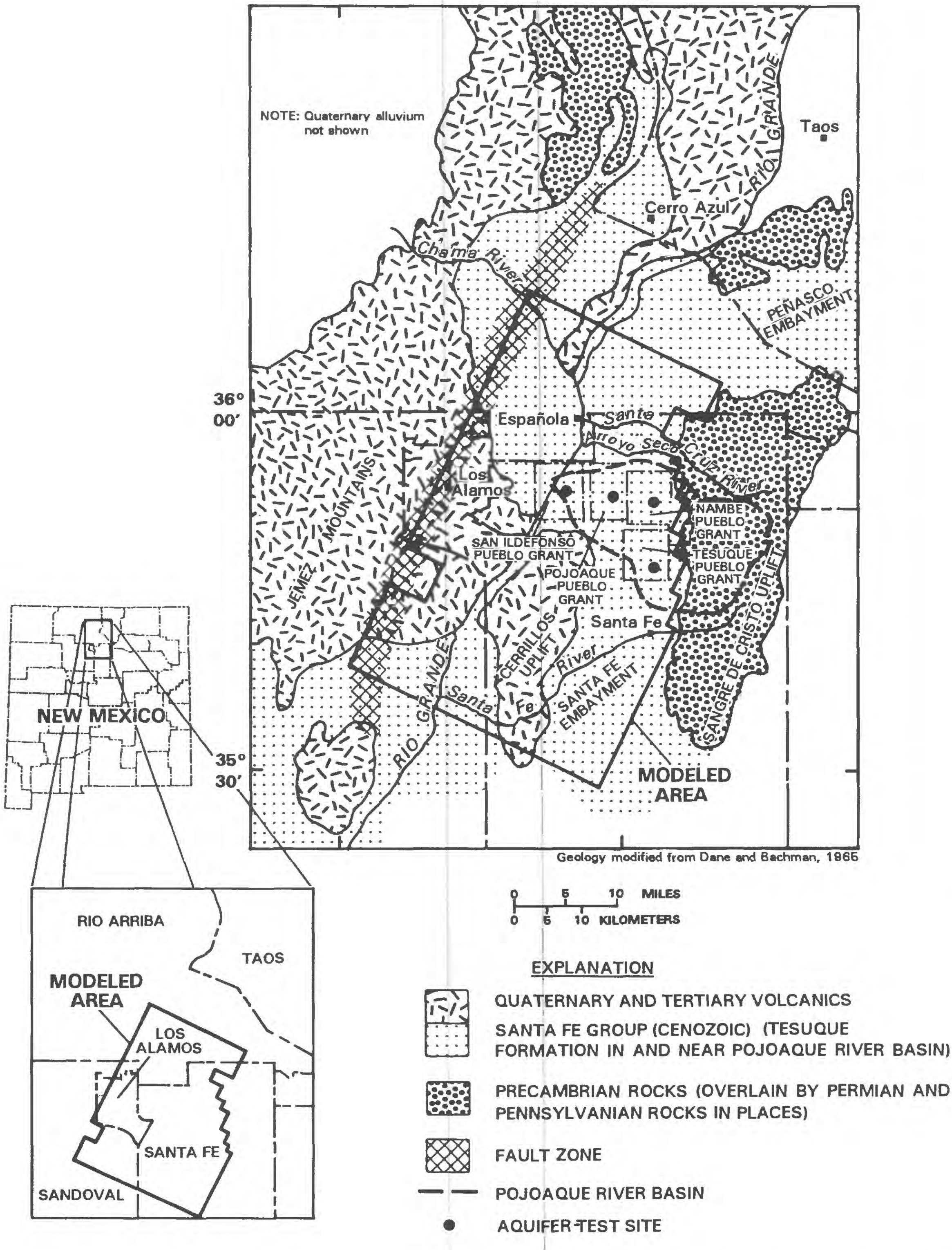


Figure 9.--Location and generalized geology of the Española Basin. [Modified from Hearne, 1985a].

The Tesuque aquifer system was simulated to a depth as great as 4,000 feet using 22 model layers. The layers were stepped downward in a northwesterly direction to conform with observed bed dips of 4 and 8 degrees. A section through the modeled area in the direction of dip is shown in figure 10. The figure is incorrect in that it portrays the bottom of the model cells as being tilted (G.A. Hearne, U.S. Geological Survey, oral commun., 1984). The cell bottoms actually were individually level with offsets at the lateral boundaries between cells within a layer. The numerical solution employed by the model (Posson and others, 1980) makes no distinction between the two portrayals. The figure from Hearne (1985a) has the advantage of being a more visually accurate representation of the actual system than a section consisting of a series of stair-stepped layers. The purpose of the stepping of the model layers was to reproduce the flow paths and distribution of hydraulic heads observed in the basin while avoiding the difficult numerical problems that arise from actually rotating the model coordinates to align with the principal conductivity tensors. The representation also is useful to approximate the indirect connection between surface-water bodies. The representation of the surface-water bodies as selected boundary types is shown in figure 11. This figure, when compared with figure 10, also illustrates the indirect connection just mentioned. For example, along row 11, column 6 is a specified hydraulic-head cell representing part of the Rio Grande (fig. 11). Just next to this cell, in column 7, is a hydraulic-head-dependent boundary representing a reach of the Pojoaque River. However, these two cells are not laterally adjacent in the same layer (fig. 10) but are connected by an indirect route consisting of vertical as well as horizontal flow paths.

In designing the flow model, Hearne attempted to maintain a simulated saturated thickness of 300 feet in the uppermost saturated cell. In the area of greatest interest, cell thickness was limited to 650 feet or less. The north and south boundaries were arbitrary. The west boundary was simulated with a specified flow across the Pajarito fault zone. The eastern no-flow boundary represented the impermeable crystalline massif of the Sangre de Cristo Mountains.

Aquifer characteristics were generalized from numerous aquifer tests in the basin, including the computer analysis just described (Hearne, 1985b). From these tests, Hearne selected and used in his simulations "most likely average values" (Hearne, 1985a, p. 10) for the aquifer properties of horizontal hydraulic conductivity (1.0 foot per day), ratio of horizontal to vertical hydraulic conductivity (about 330:1), specific storage (2×10^{-6} per foot), specific yield (0.15), and a constant of proportionality for hydraulic-head-dependent boundaries (5×10^{-10} per second or 4.32×10^{-5} per day). He also selected upper and lower limits of the plausible range of values for these properties, which he employed in the sensitivity analysis of the model. The constant of proportionality reflected an adjustment for a 7-degree intercept angle between the streambeds and the aquifer system.

NOTE: This section is through row 10 of the model

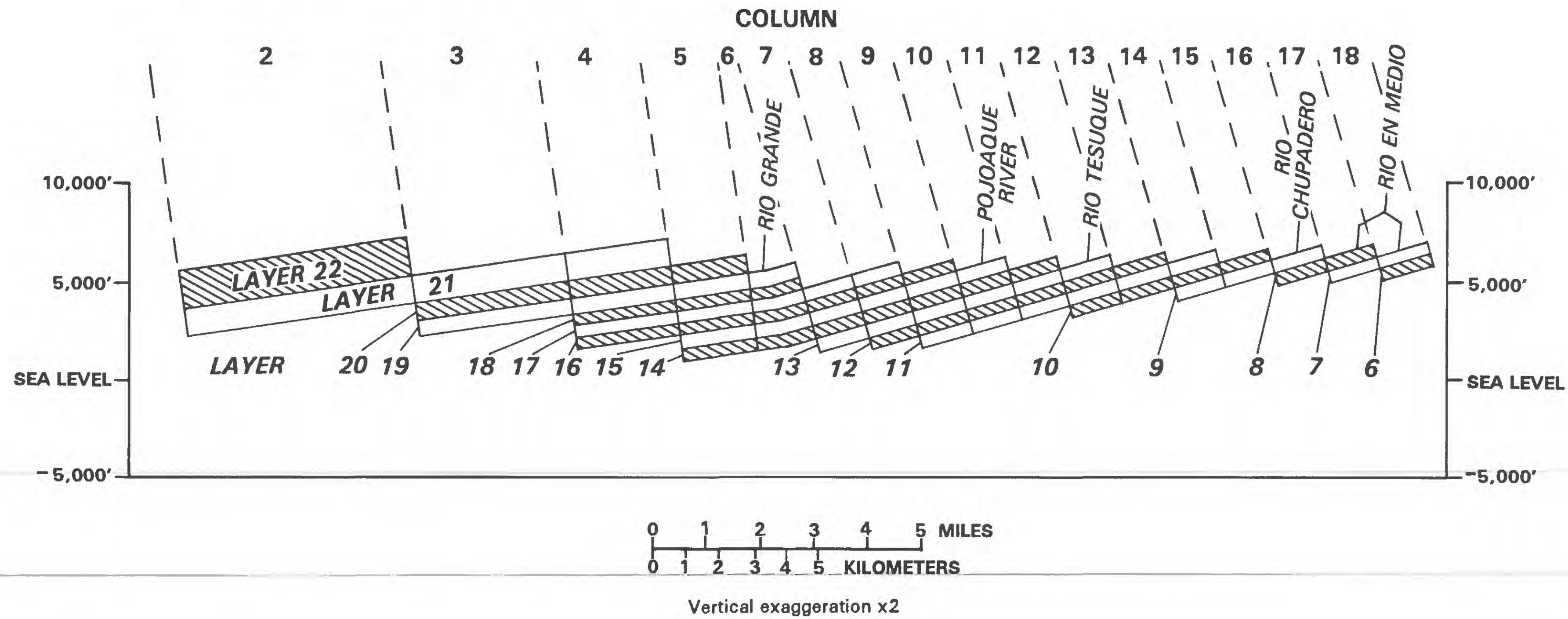


Figure 10.--Section through the modeled area of the Española Basin along the direction of dip. [Modified from Hearne, 1985a, fig. 3].

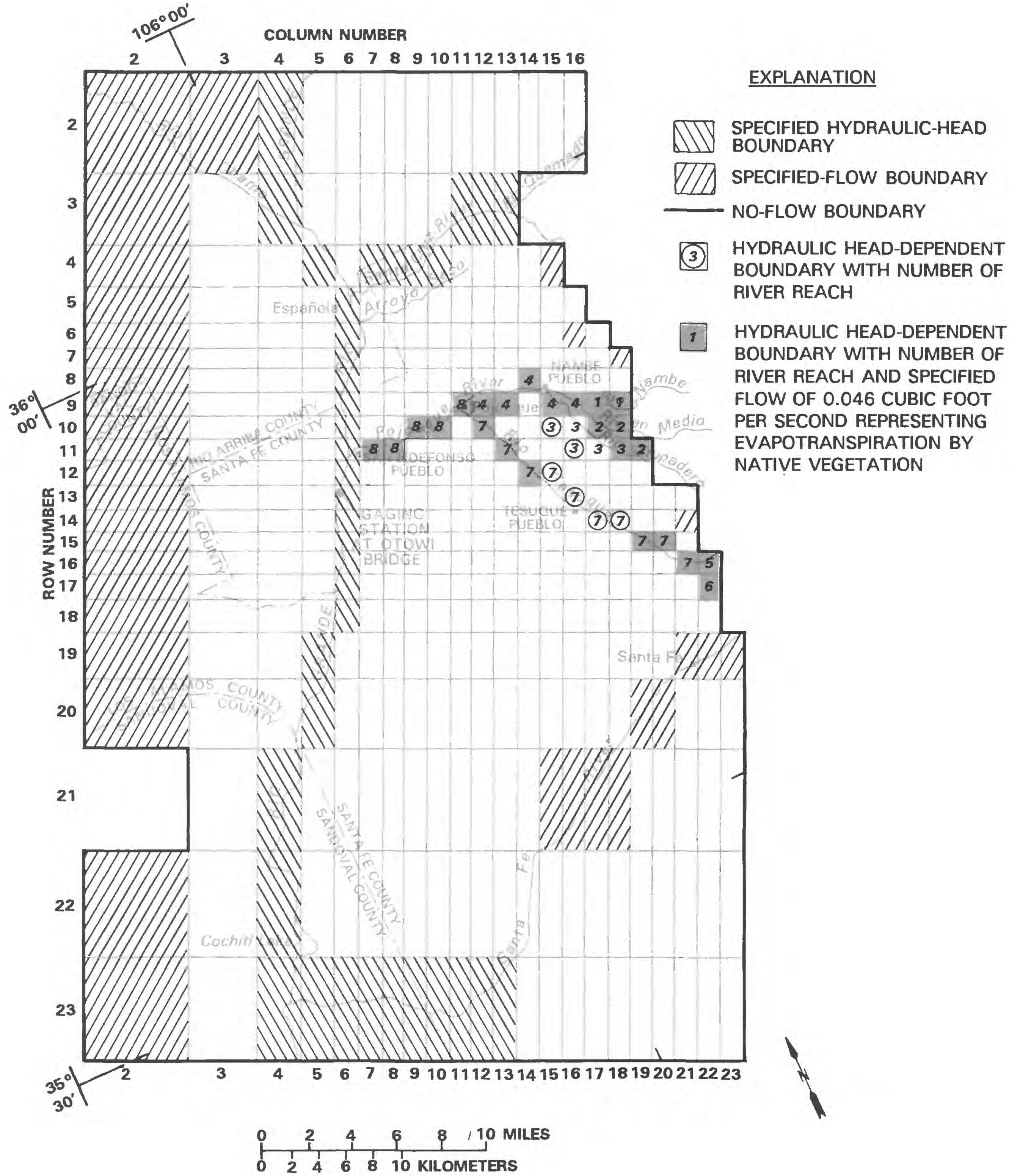


Figure 11.--Finite-difference grid and boundary conditions represented in the model of the basin-fill aquifer in the Española Basin. [Modified from Hearne, 1985a, fig. 5].

Extra attention was given to estimating surface-water budgets for both predevelopment and postdevelopment conditions. The elements of the budget then were allowed to interact with the aquifer system through the most appropriate of boundary types (fig. 11). The predevelopment simulation was considered acceptable when observed (or measured) and simulated hydraulic heads were similar and when simulated elements of the water budget were reasonable. In these simulations, Hearne reported that the Rio Grande gained a net amount of about 22.06 cubic feet per second of which 12.76 cubic feet per second came from west of the river and 9.29 came from east of the river. East of the Rio Grande, the Pojoaque River contributed 3.32 cubic feet per second, the Santa Cruz River 2.61, and the Santa Fe River 2.86.

The transient simulation for 1946-80 employed historical withdrawals as the applied stress and hydraulic-head changes as the observed and simulated response. The same criteria of reasonableness were applied to the simulation results before proceeding to project the response of the aquifer system to a 100-year program of irrigation withdrawals proposed by the U.S. Bureau of Indian Affairs, a cooperator in the investigation. The proposed irrigation program consists of a step-increase surcharge of 28.39 cubic feet per second greater than 1980 withdrawals of 11.24 cubic feet per second. The surcharge consists of a net withdrawal of 25.06 cubic feet per second for irrigation of tribal land, 2.99 cubic feet per second for irrigation of nontribal land, and 0.33 cubic foot per second for increased municipal and domestic demands.

Although the simulation period was for 1980-2080, Hearne concentrated on the simulation results 50 years after 1980 (2030). The maximum simulated drawdown in the modeled area was determined to be near the Nambe Pueblo where the drawdown was 334 feet in the uppermost confined layer and 143 feet in the top unconfined layer. These drawdowns were shown by an analytical technique to be influenced less than 5 feet by the position and type of the north, west, and south boundaries.

The computed sources of water for ground-water withdrawals to the year 2030 are summarized in table 1. The table also indicates the response of the model to simulations using the limits of plausible values for aquifer characteristics. The model was found to be most sensitive to both extremes of simulated horizontal hydraulic conductivity, less sensitive to an increase in simulated aquifer thickness, and least sensitive to the range in specific storage.

Albuquerque-Belen Basin

The Albuquerque-Belen Basin is among the largest and deepest of the Rio Grande rift basins (fig. 1). The structural basin is about 40 miles from west to east at its widest point and about 100 miles from north to south. The thickness of the basin fill locally may exceed 18,000 feet.

The southern terminus of the basin is a shallow and narrow constriction. The northern terminus has been defined differently by various investigators to either include or exclude the Santo Domingo Basin. Structurally, the Santo Domingo and Albuquerque-Belen Basins probably are parts of one rift basin. However, the ground-water systems of two basins are at least partly isolated from each other by north-northwest-trending fissure-flow volcanics.

Table 1.--Projected sources of water to pumpage and depletion of flow in the Rio Grande in the Española Basin for 2030 without and with a proposed increase in surface-water diversions and in ground-water withdrawals for irrigation of tribal and other land

[Information is from Hearne, 1985a, p. 51 and table 17]

	Total without increase (1980 pumpage)		Total with increase above 1980 base		
Source of water to pumpage	Cubic feet per second	Percentage	Cubic feet per second	Percentage	Range[1] of percentage
Withdrawn from aquifer storage	8.79	78.1	34.05	85.9	80 - 90
Capture from:					
Pojoaque River	.27	2.4	2.45	6.1	.9 - 9.2
Rio Grande	1.99	17.7	1.92	4.8	1.6 - 9.2
Santa Cruz River	.13	1.2	1.12	2.8	1.7 - 4.5
Santa Fe River	.07	.6	.14	.4	.1 - .8
Total capture from rivers	2.46		5.63		
Surface diversions	7.67		13.14		
Total depletion of flow in the Rio Grande	10.13		18.77		

[1]Range in percentage of water obtained from various sources as determined during the analysis of the sensitivity of the model to upper and lower limits of plausible values for aquifer characteristics (Hearne, 1985a, table 17).

As for the basins described earlier, the Rio Grande flows through the Albuquerque-Belen Basin. Evapotranspiration from crops and native vegetation in the flood plain is the major consumptive use of water in the basin, although most evapotranspiration loss is from surface-water, rather than ground-water, sources. Because Albuquerque, New Mexico's largest city, is in the basin, municipal, industrial, and domestic water use also is a major consumptive use of water, almost exclusively from ground water. Ground water provides a dependable source of water even though the Rio Grande may at times cease flowing in the Albuquerque area. However, the impact of ground-water withdrawals on the surface-water system has been of historical concern, and there is a more recent concern about excessive drawdown due to interference between individual wells and between well fields.

Analytical Model

The work by Reeder and others (1967) provides an exception to the general definitions of models given earlier. The purpose of their investigation and model was to quantify the impact of ground-water withdrawals on the surface-water system and on ground-water levels in the Albuquerque area. The area of investigation is shown in figure 12. Their work is an exception in that they manually completed an analytical evaluation of areally distributed withdrawals, nonuniform aquifer properties, and boundaries. Their work falls into the category of a model because they discretized the modeled area into cells and computed the interrelated and time-dependent impact of stresses on each cell. Their model essentially is a manual completion of the mathematically forbidding numerical solution referred to earlier as being relegated to digital computers.

In their analysis, Reeder and others (1967) assumed a transmissivity of 200,000 gallons per day per foot for the area east of the Rio Grande and 100,000 gallons per day per foot for the area west of the river even though they noted the presence of areas of smaller transmissivity in the Rio Puerco valley (west of the Rio Grande) and to the south and east of Albuquerque. Their estimated transmissivity was obtained by summarizing the work of earlier investigators. The investigators also assumed a specific yield of 0.20 throughout the area. The eastern and western rift boundaries were treated as gently curved no-flow boundaries. The north and south boundaries were arbitrarily located no-flow boundaries. The axial Rio Grande was treated as a constant hydraulic-head boundary.

In their analysis, the investigators developed time, distance, drawdown curves by the Theis (1935, 1952) method. Beginning with 1920, stress was applied to each appropriate cell; drawdown was determined from the Theis curves for that cell (as a function of discharge), all other stressed cells, and selected index cells. In computing cumulative drawdown, the image-well theory was used to determine the effect of the boundaries. For each step in time, computations were required for each stressed cell, each index cell, and each image cell. Computations were completed from 1920 to 1960 and 2000. Changes in water level and sources of water from 1960 to 2000 were computed as the difference between the results of the 1920-2000 and 1920-60 simulations.

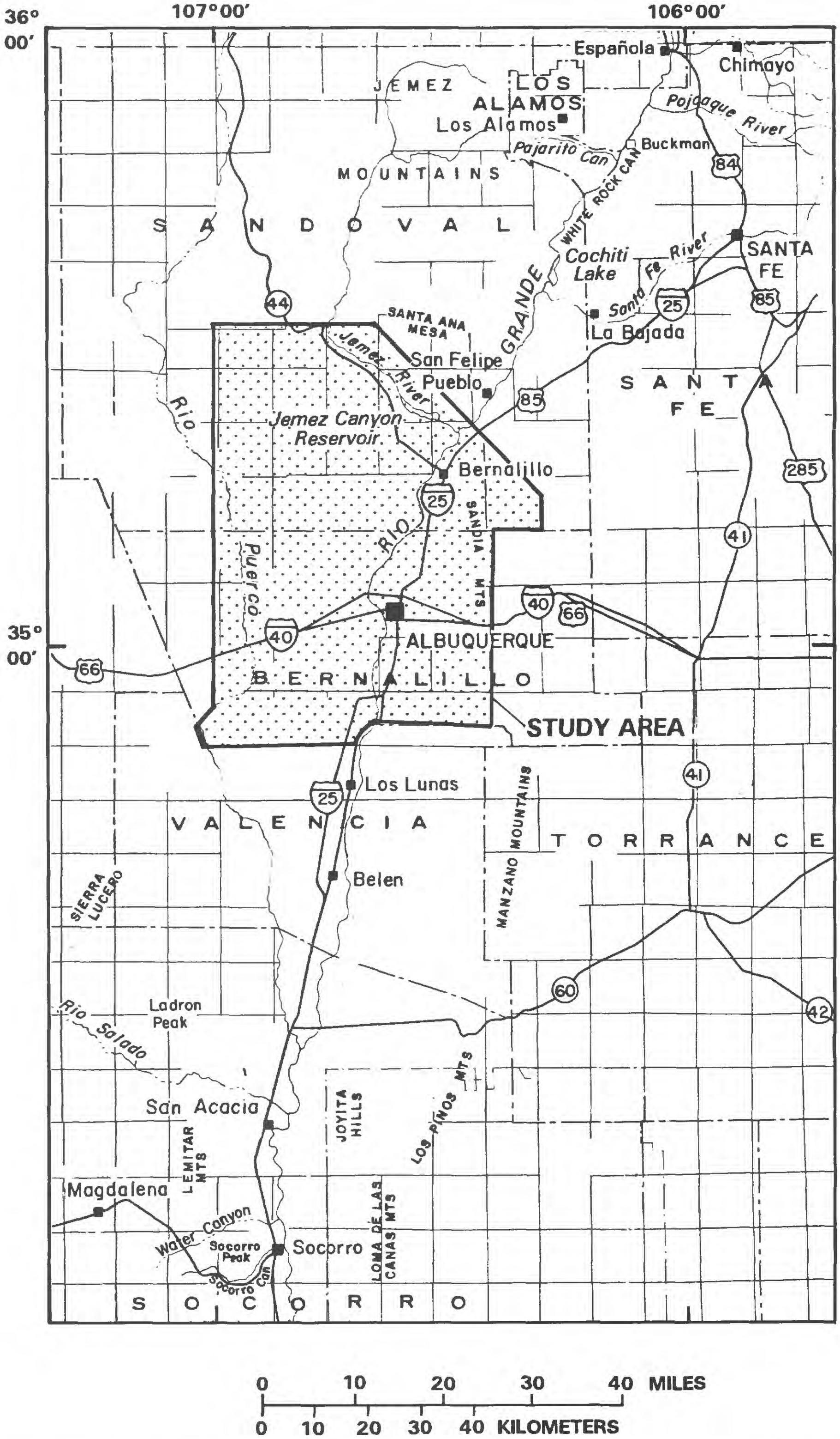


Figure 12.--Study area of the Albuquerque-Belen Basin investigated to quantify the impact of ground-water withdrawals on the surface-water system and on ground-water levels in the Albuquerque area.

Reeder and others (1967, p. 26) concluded that "About 80 percent of the water pumped by the city from 1920 to 1960 was derived from the flow of the Rio Grande, either decreasing the flow to the river or increasing the flow from the river. From 1960 to the year 2000 between 71 and 76 percent of the water pumped will be derived from the Rio Grande."

The investigators also projected maximum ground-water-level declines from 1960 to 2000 of 86 feet east of the river and 34 feet west of the river. They noted that the east and west impermeable boundaries had a large (as much as 30 percent) impact on the total drawdown.

Three-Dimensional Finite-Difference Model

The model documented in the reports by Kernodle and Scott (1986) and Kernodle and others (1987) was prepared as part of the SWAB RASA study. The modeling aspect of the investigation had several objectives: (1) To quantify the components of the available surface-water and ground-water resources; (2) to develop a better understanding of the aquifer system in the basin, in particular the significance of deeper parts of the flow system in view of vertical anisotropy and locally reduced hydraulic conductivity; (3) to test the feasibility of representing the Rio Grande flood-plain surface-water system, including irrigation drains and canals, as a constant-head boundary in a three-dimensional ground-water-flow model and to determine the degree of interaction between the surface- and ground-water systems; and (4) to quantify the dependence and impact of ground-water withdrawals on the surface-water system.

The construction of the model was almost entirely dependent on the reported work of previous investigations. From these works, a generalized concept of the basin-fill aquifer system was formulated. Among the generalizations were the following observations about aquifer properties: (1) The aquifer has larger hydraulic conductivity east than west of the Rio Grande (Reeder and others, 1967); (2) there are areas of reduced hydraulic conductivity in the northern, western, and southwestern parts of the basin (Kelley, 1977) and south of Tijeras Arroyo (Bjorklund and Maxwell, 1961); and (3) the western part of the basin has an identifiable zone of small hydraulic conductivity that can be traced into the subsurface as it dips eastward from the Rio Puerco valley (Bjorklund and Maxwell, 1961). In addition to these generalizations, other assumptions were made. One assumption was that hydraulic conductivity of the basin fill does not decrease with depth as a function of compaction and does not increase with depth as a function of increased temperature. Another was that the surface-water system in the flood plain of the Rio Grande maintains a constant hydraulic head in the immediately underlying alluvial deposits.

The modeled area includes all of the Albuquerque-Belen Basin but excludes the Santo Domingo Basin. The model-grid spacing was reduced in the Albuquerque area to achieve greater detail in the distribution of stress and observation of response. The modeled area and model grid are shown in figure 13. Six model layers were used to represent approximately the top 6,000 feet of saturated basin-fill aquifer (where present). The geometry of the basin at depth was determined from Birch (1980a), with refinements from available oil-test information and seismic profiles. The top layer represented the first 200 feet of unconfined saturated thickness.

The initial phase of the modeling process (Kernodle and Scott, 1986) employed estimated rates of mountain-front and tributary recharge (see the discussion on Hearne and Dewey, 1988) as a steady applied stress to arrive at preliminary estimates of horizontal hydraulic conductivity and vertical anisotropy. The comparison criteria were pre-1961 measured water levels and maps of contoured water levels modified from earlier investigations.

The second phase of the modeling process (Kernodle and others, 1987) included time-dependent changes in ground-water withdrawal as a transient stress, which allowed refinements of estimated hydraulic conductivity and vertical anisotropy and a determination of the specific storage and specific yield of the basin-fill aquifer. The comparison criteria were the same as the initial, steady-state phase of the modeling process.

The final model (Kernodle and others, 1987) used hydraulic conductivity of 40 feet per day for the area east of and beneath the flood-plain alluvium of the Rio Grande, 50 feet per day for the flood-plain alluvium, 30 feet per day for the area west of the Rio Grande flood plain, 2 feet per day in the northern part of the basin, 3 feet per day in an area south of Tijeras Arroyo, and 0.25 foot per day for an eastward-dipping small permeability zone in the western part of the basin. The ratio of horizontal to vertical hydraulic conductivity was simulated as 500:1 for all layers. Aquifer specific storage was simulated as 1×10^{-6} and specific yield as 0.10.

The simulation period for the final model was 1960-79; 1979 was the last year for which complete ground-water-withdrawal data were available. Because sufficient hydraulic-head data were not available for dates later than 1961, the simulation to 1979 was considered to be unverified. Two tables were presented that summarize the water budgets in the basin for 1959-61 and 1976-79 (Kernodle and others, 1987, tables 4 and 5). For the period 1959-61, 72 percent of the total ground-water withdrawals from other than flood-plain alluvium came from capture of flow to or induced recharge from the Rio Grande flood-plain system, 24 percent came from aquifer storage, and 4 percent came from induced underflow from the Santo Domingo subbasin. The total simulated withdrawal for 1959-61 was about 46,000 acre-feet per year.

Of the approximately 100,000 acre-feet per year simulated as withdrawn during 1976-79, 68 percent came from capture of water in the Rio Grande flood-plain system, 25 percent came from aquifer storage, and 7 percent came from induced underflow from the Santo Domingo subbasin. An estimated water budget for the Albuquerque-Belen Basin for 1976-79 is shown in table 2 (from Kernodle and others, 1987, table 5).

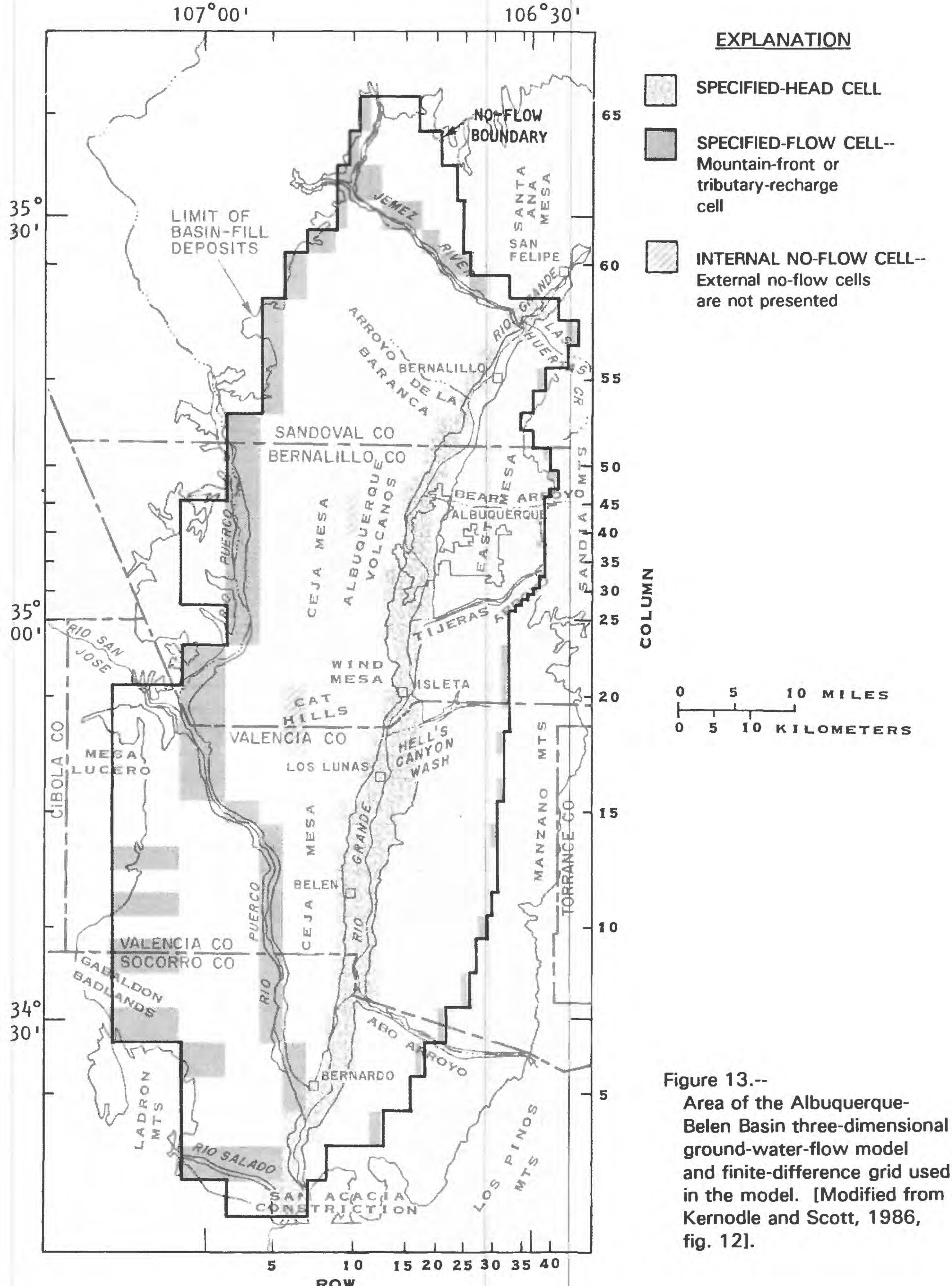


Figure 13.--Area of the Albuquerque-Belen Basin three-dimensional ground-water-flow model and finite-difference grid used in the model. [Modified from Kernodle and Scott, 1986, fig. 12].

Table 2.--Water budget for the Albuquerque-Belen Basin, 1976-79, in thousands of acre-feet per year

[Table from Kernodle and others, 1987, table 5]

Mechanism	Water inflow	Water outflow or loss
Rio Grande main stem	885	753
Rio Grande tributaries (surface flow)	80	-
Ground water	62	13
Mountain-front and tributary recharge	129	-
Evapotranspiration	-	310-390
Ground-water withdrawal:		
Intercepted ground-water discharge to flood-plain system	-	36
Induced recharge from flood-plain system	-	32
Induced inflow from Santo Domingo Basin[1]	-	[1]7
Depletion in aquifer storage[2]	-	[2]25
TOTAL	1,156	1,151-1,231

[1]Included also with ground-water inflow.
[2]Not included in totals.

Simulations were not projected beyond 1979 because of lack of pumpage data and because the treatment of the Rio Grande and other elements of the surface-water system as a constant hydraulic-head boundary was shown to be inappropriate for the later conditions of large ground-water withdrawals. The investigators suggested altering the simulated treatment of that boundary before extending simulations beyond 1979.

In the simulation of steady-state conditions, the response of the model to the removal of the lowest layer (2,250 feet thick) from the model was tested. The mean absolute error between measured and computed water levels was found to increase 1.5 feet, a significantly greater amount than any other sensitivity test involving 10-percent changes in simulated aquifer properties, but still small considering that one-third of the modeled thickness was removed.

In the simulation of transient conditions, the response of the model to paleoclimatic variations was tested. In this analysis, a 10 percent wetter climate was simulated for 400 years, followed by 800 years of current climatic conditions. Climatic changes were simulated by varying the amount of mountain-front and tributary recharge. The amount of time required for water levels to approach equilibrium was shown to be about 200 years. This analysis was performed in an effort to determine the origin of a reported ground-water trough, or linear depression, in the center of the basin and west of the Rio Grande.

Mesilla Basin

The Mesilla Basin is the southernmost Rio Grande basin in New Mexico. Most of this structural basin is in New Mexico, but the basin-fill deposits extend south into Mexico and a small part of the structural basin is in Texas. Bedrock constrictions along the Rio Grande isolate ground water in the basin from ground water in the Palomas and Jornada del Muerto Basins to the north and the Tularosa-Hueco Basin to the southeast (fig. 1). The constrictions are especially narrow and shallow. Basin fill essentially is limited to thin alluvium of the width of the flood plain of the Rio Grande. A saddle between the northeastern part of Mesilla Basin and the southern Jornada del Muerto Basin is a potential ground-water flow path through pre-Tertiary consolidated rocks. Another potential route for ground-water exchange through basin-fill deposits exists between the Mesilla and Tularosa-Hueco Basins.

The climate in the basin is arid; average annual precipitation is about 8 inches along the Rio Grande. Virtually all runoff from the Organ and Franklin Mountains that reaches the basin-fill deposits on the east side of the basin infiltrates into the aquifer, evaporates, or is transpired. Precipitation on the West Mesa evaporates or is transpired and little if any recharges the aquifer. An extensive caliche horizon on the West Mesa suggests the lack of recharge to the aquifer. Most of the surface of the West Mesa is topographically closed and drainage is internal to kryptovolcanic holes or to wind-blowout depressions.

The inner valley or flood plain of the Rio Grande is irrigated and supports intense agricultural development. This development underwent rapid growth soon after the completion, in 1916, of Elephant Butte Reservoir, which impounds the Rio Grande and supplies surface water for irrigation. As in other basins, extensive irrigation led to waterlogging of the soil and required the eventual construction of a network of drains to stabilize ground-water levels below the root zone of the crops.

Ground-water withdrawals increased substantially in response to a prolonged drought during the 1950's when there was insufficient surface water to meet the requirements of irrigated agriculture. Wells continue to supply a large percentage of the irrigation needs even when surface water is plentiful, primarily because of convenience and because the initial investment in well construction has already been made.

Withdrawal of ground water for municipal and industrial use remains a minor fraction of the total water consumption in the basin. Las Cruces, the largest city in the basin, withdrew about 10,000 acre-feet of ground water during 1975 (Frenzel and Kaehler, 1990). Well fields at Anthony and Cañutillo that supply water to El Paso, Tex., were responsible for the withdrawal of about 19,000 acre-feet of ground water, whereas the total withdrawal in the southern Mesilla Valley was about 27,000 acre-feet (Gates and others, 1984). Future municipal withdrawals in the Mesilla Basin may increase at a rate greater than the growth in population in the basin because of ground-water export from the Mesilla Basin to El Paso in the Hueco Bolson.

As in other basins, knowledge of the impact of ground-water development on the surface-water system is essential because the surface-water supply is now completely allocated to existing users. Water rights must be obtained for that quantity of water that a new well causes to be depleted from the surface-water system. A number of non-U.S. Geological Survey models have been completed to address the issue of streamflow depletion and water-quality changes in the Mesilla Basin. The majority are lumped-parameter models of the flood-plain alluvium in the inner valley of the Rio Grande, for example Lizarraga (1978), Updegraff and Gelhar (1978), or Gelhar and McLin (1979). Another approach was taken by Richardson and others (1972) and by Ball (1974), who constructed two-dimensional finite-difference ground-water-flow models of the alluvium of the inner valley.

Three-Dimensional Finite-Difference Model of the Cañutillo Area

In 1976, Gates and others (1984) completed a model of the basin-fill aquifer system in the vicinity of the Cañutillo well field in the Texas part of the Mesilla Basin. The purposes of the model were to quantify the relation between the surface- and ground-water systems, to project ground-water-level declines, and to determine the impact of a hypothetical lining of the channel of the Rio Grande on ground-water levels in the Cañutillo well-field area. A three-dimensional finite-difference flow model was constructed with three layers. The location of the modeled area is shown in figure 14. The finite-difference grid, boundary types, and simulated aquifer properties of the model are shown in figure 15.

The number and thickness of the model layers were selected to correspond to shallow, medium, and deep production zones within the aquifer system. The total simulated thickness was 1,220 feet. The Rio Grande was represented in the model by constant hydraulic-head cells in the top layer. Drains were not simulated. The eastern and southeastern model boundaries were located at the basin-bounding faults along the western flank of the Franklin Mountains. The locations of the other boundaries were arbitrarily chosen. Ground-water flow across the boundaries was computed as a function of estimated transmissivity and measured hydraulic gradient.

The calibration process consisted of matching measured and simulated predevelopment heads followed by matching water-level hydrographs in the Cañutillo well field for 1952-67 and 1968-75. Vertical leakance, seepage from irrigation, recharge, and hydraulic conductivity of the shallow aquifer under the flood plain were the values of simulated properties modified during the calibration process. In the final model, the hydraulic conductivity of the flood-plain alluvium beneath the inner valley was 117 feet per day; elsewhere the hydraulic conductivity was 13 feet per day. The ratio of horizontal to vertical hydraulic conductivity was about 20:1 outside the flood-plain area and 500:1 beneath the flood plain. Aquifer storage was estimated to be 0.15 for the flood-plain alluvium and 0.0007 for the area of the layers immediately beneath the flood plain. Outside the flood-plain area, aquifer storage was simulated to be 0.10 for each of the three model layers. Underflow into the model was simulated to be 18,000 acre-feet per year (24.8 cubic feet per second).

The final model was used to project the response of the aquifer system to a 100-percent increase in withdrawals from the Cañutillo well field for 1976-80 both with and without a hypothetical lining of the Rio Grande channel. Without the lining, the increased withdrawal resulted in a simulated "sharp decline in water levels * * * in the first few months after which the levels virtually stabilized" (Gates and others, 1984, p. 16). About 21,000 acre-feet per year was computed to leak from the shallow layer downward and about 13,000 acre-feet per year from the middle to the lowest layer. "A substantial part of the leakage from the shallow aquifer probably can be attributed to the high water table that was maintained largely by seepage of irrigation water and water in the Rio Grande" (Gates and others, 1984, p. 16). The model showed that lining the channel of the Rio Grande would cause an additional 10 feet of drawdown in the medium and deep production zones during 1976-80.

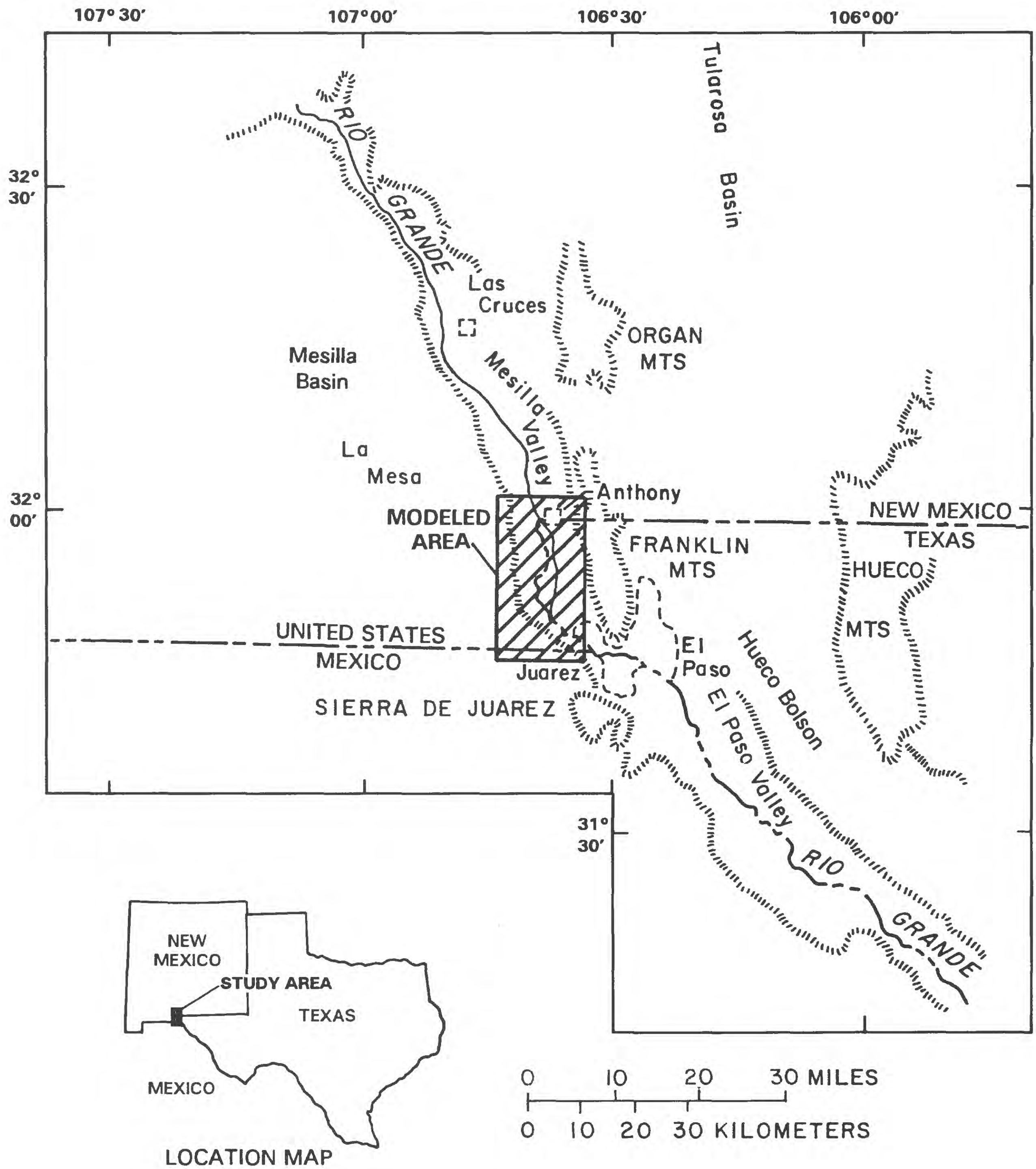


Figure 14.--Location of the lower Mesilla Valley and the modeled area near Cañutillo, Texas. [From Gates and others, 1984, fig.1].

EXPLANATION

MODELED BOUNDARY OF THE DEEP AQUIFER (LAYER 1)--Dashed line indicates subareas of different values of transmissivity or storage

MODELED BOUNDARY OF THE MEDIUM-DEPTH AQUIFER (LAYER 2)--Dashed line indicates subareas of different values of transmissivity or storage

MODELED BOUNDARY OF THE SHALLOW AQUIFER (LAYER 3)--Dashed line indicates subareas of different values of transmissivity or storage

OUTLINE OF IRRIGATED LANDS AS ASSUMED FOR THE MODEL

NODE WHERE RECHARGE IS SIMULATED TO ALL THREE LAYERS

NODE WHERE RECHARGE IS SIMULATED TO DEEP AND MEDIUM-DEPTH AQUIFERS (LAYERS 1 AND 2)

NODE WHERE RECHARGE IS SIMULATED TO MEDIUM-DEPTH AND SHALLOW AQUIFERS (LAYERS 2 AND 3)

NODE WHERE RECHARGE IS SIMULATED TO THE SHALLOW AQUIFER ONLY (LAYER 3)

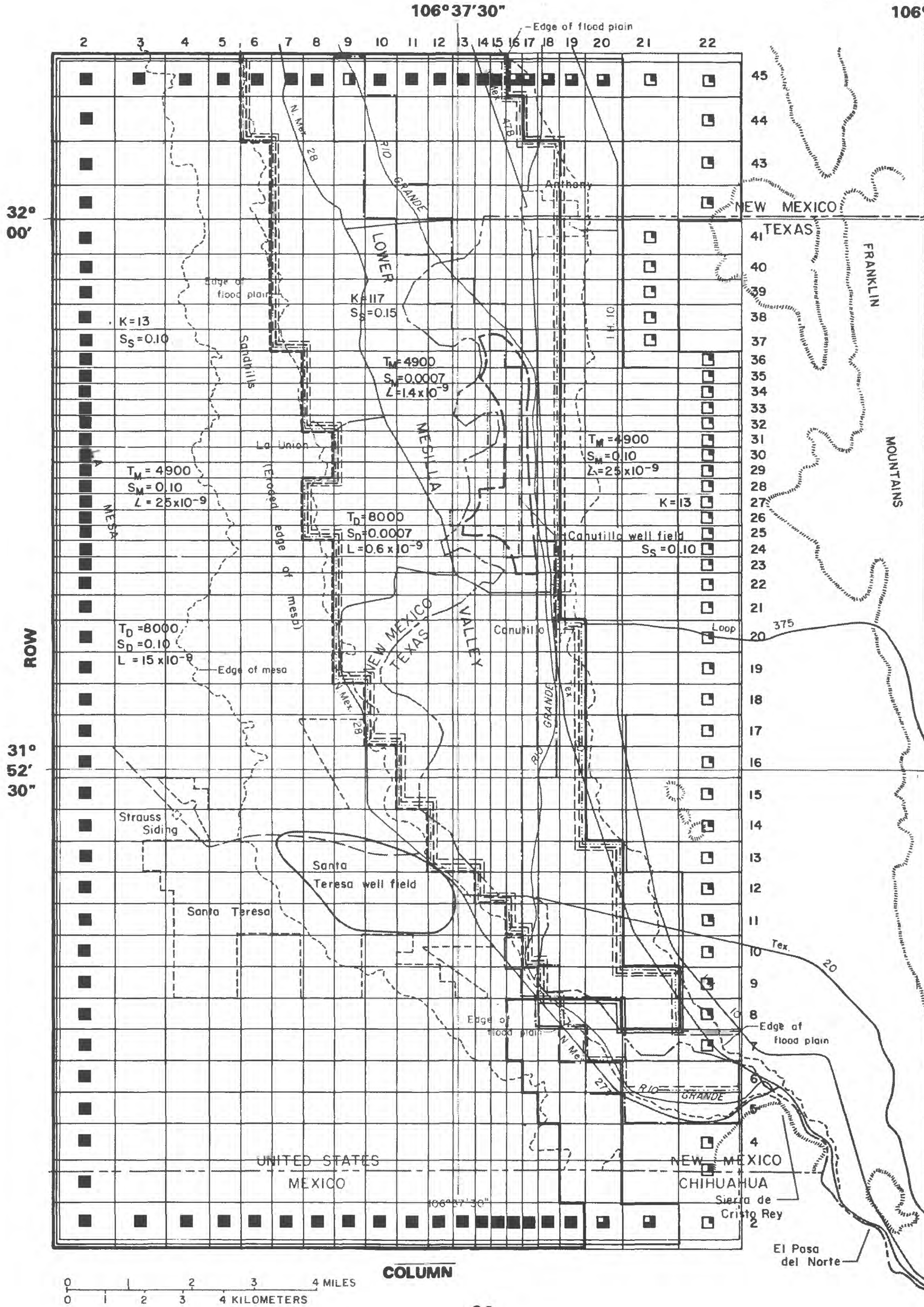
CONSTANT-HEAD NODES (RIO GRANDE) (T) (K)

AQUIFER	TRANSMISSIVITY OR HYDRAULIC CONDUCTANCE	STORAGE COEFFICIENT OR SPECIFIC YIELD	K'/b'LEAKANCE		
DEEP	T_D = FEET^2 PER DAY	S_D	L (Between deep and medium-depth aquifer)	=	in per second or feet^3 per second per feet^2 per foot
MEDIUM DEPTH	T_M = FEET^2 PER DAY	S_M	$\mathcal{L}$ (Between medium-depth and shallow aquifer)		
SHALLOW	K = FEET PER DAY	S_S			

NOTE:

T = TRANSMISSIVITY
K' K = HYDRAULIC CONDUCTANCE
b' b = THICKNESS
L $\mathcal{L}$ = LEAKANCE
S = STORAGE

Figure 15.--Model grid, boundaries, and transmissivity of the layers of the model of the lower Mesilla Valley. [Modified from Gates and others, 1984, fig. 4].



106°37'30"
106°30'
Edge of flood plain
COLUMN
ROW
32° 00'
31° 52' 30"
NEW MEXICO
TEXAS
FRANKLIN
MOUNTAINS
Anthony
LOWER
MESILLA
VALLEY
RIO GRANDE
N. Mex. 28
N. Mex. 478
I. H. 10
K = 117
S_S = 0.15
K = 13
S_S = 0.10
T_M = 4900
S_M = 0.0007
L = 1.4 x 10^{-9}
T_M = 4900
S_M = 0.10
L = 25 x 10^{-9}
T_D = 8000
S_D = 0.0007
L = 0.6 x 10^{-9}
T_D = 8000
S_D = 0.10
L = 15 x 10^{-9}
Edge of flood plain
Sandhills
La Union
(Eroded edge of mesa)
MESA
Edge of mesa
Canutillo well field
Canutillo
Loop 375
NEW MEXICO
TEXAS
Tex. 20
Strauss Siding
Santa Teresa well field
Santa Teresa
Edge of flood plain
N. Mex. 273
Edge of flood plain
UNITED STATES
MEXICO
NEW MEXICO
CHIHUAHUA
106°37'30"
Sierra de Cristo Rey
El Paso del Norte
0 1 2 3 4 MILES
0 1 2 3 4 KILOMETERS

Three-Dimensional Finite-Difference Model

As part of the SWAB RASA, Frenzel and Kaehler (1990) completed a three-dimensional finite-difference flow model of the Mesilla Basin. As with the other documented SWAB models of the Albuquerque-Belen Basin, the San Luis Valley, and the Animas Valley, the primary purpose of the model of the Mesilla Basin was to gain a better understanding of the surface- and ground-water hydrology of the basin. In addition, one of the purposes of the SWAB RASA models was to experiment with the impact of model configuration and boundary types on model accuracy and dependability. The Mesilla Basin was selected to test the feasibility of simulating drain and river flows in addition to aquifer responses. It also was anticipated that by portraying the relation between surface water and ground water as accurately as possible, the model possibly could be used as a management tool for water resources in the basin.

The geometry of the basin was determined from analysis of gravity anomalies, seismic profiles, and deep test wells. The maximum thickness of basin fill is not much in excess of 3,800 feet in each of two troughs separated by a horst in the central part of the basin. The entire thickness of basin fill was simulated using five model layers, the uppermost layer representing the top 200 feet of unconfined aquifer. The extent of the model was defined by the location of basin-bounding faults on the east and west and by shallow bedrock on the north and south. The finite-difference grid used in the simulations is shown in figure 16.

Initial estimates of hydraulic conductivity were obtained from published reports. These reports indicate a grouping of hydraulic-conductivity values for the shallow and deep deposits of the Santa Fe Group and for areas of large and moderate hydraulic conductivity within the flood-plain alluvium. The hydraulic conductivity was simulated as decreasing with depth as an assumed result of compaction of aquifer material.

Model simulations were conducted in a recursive two-step process until the difference between simulated and measured hydraulic head and surface-water flows was reduced to an acceptable level. In the first step, computed mountain-front recharge (see Hearne and Dewey, 1988) and simulated natural evapotranspiration were used as predevelopment steady stress to obtain a preliminary or initial hydraulic-head distribution. Also in this step a pre-drain and unaltered Rio Grande stream channel was simulated and flow in the braided channel was routed through the basin. Simulated hydraulic heads in the flood plain were compared with heads reported by Lee (1907, pl. X). In the next step, ground-water drains were added to the simulation, the natural braided system of the Rio Grande was replaced by the present artificially aligned channel, and transient stresses were simulated for 1915-75. The transient stresses included pumpage from municipal and industrial wells, net irrigation flux to or from ground water, evapotranspiration from nonirrigated land, and climate-induced changes in mountain-front recharge. The beginning year was selected to include the first irrigation season after completion of Elephant Butte Reservoir. The simulation was terminated at the end of 1975 because of lack of information regarding production intervals and withdrawal from (at that time) newly completed deep irrigation wells. Simulated heads were compared with measured or reported heads for 1947-48 and 1975-76. Also considered in the calibration process were drain flows, depletion of flow in the Rio Grande, and ground-water-level hydrographs.

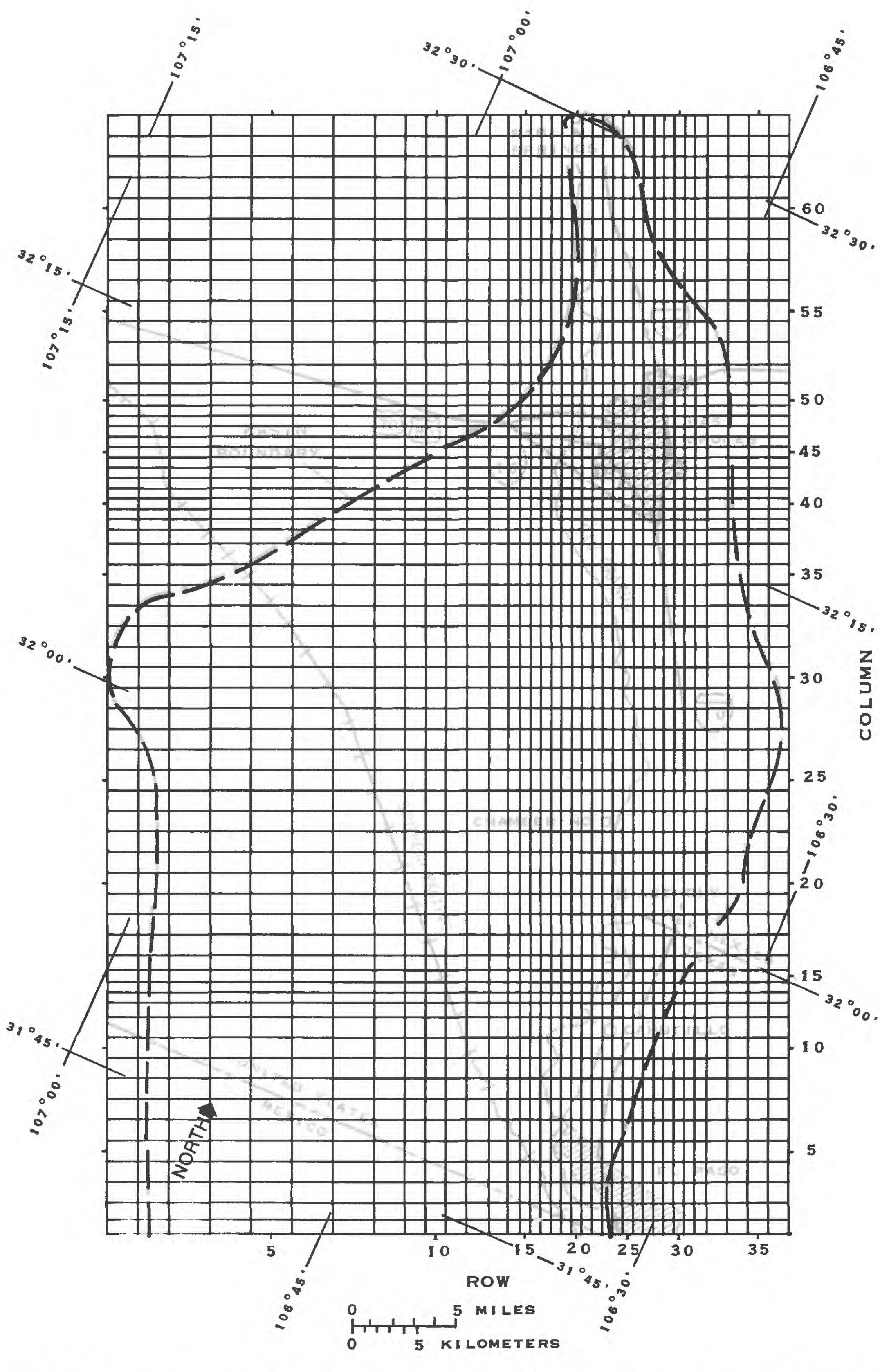


Figure 16.--Location and finite-difference grid of the ground-water-flow model of the Mesilla Basin. [From Frenzel and Kaehler, 1990, fig.19].

The horizontal hydraulic conductivity in the accepted model ranged from 3 to 140 feet per day. The hydraulic conductivity in most of the flood-plain alluvium was 70 feet per day, but there were three inclusions of 140 feet per day. The horizontal hydraulic conductivity of the Santa Fe Group ranged from a maximum of 22 feet per day in the uppermost layer to 3 feet per day in the lowest model layer. An area having a hydraulic conductivity of 13 feet per day was simulated in the vicinity of the Cañutillo well field. The simulated transmissivity of areas of the lowest two model layers was reduced to compensate for thinning of aquifer thickness. The ratio of horizontal to vertical hydraulic conductivity was simulated to be 200:1 everywhere. The specific yield was simulated to be 0.20 and the specific storage to be 1×10^{-6} per foot, with no local variations in either coefficient. The constant of proportionality (vertical leakage) was simulated to be 0.00868 per day for flow between the Rio Grande and the aquifer and 0.00341 per day for flow between drains and the aquifer.

The net irrigation flux to or from ground water was computed externally from the model and included as constant fluxes in the simulated time periods. Evapotranspiration from nonirrigated areas was computed by the model on the basis of the assumption of a maximum rate of 5.5 feet per year when the water level was at land surface and decreasing linearly to zero at a depth of 15 feet below land surface. The maximum evapotranspiration rate was adjusted in proportion to the percentage of nonirrigated acreage. Evaporation from open water was computed and included with the water budget externally from the model.

Sensitivity analyses indicated a low sensitivity of the model to aquifer storage. The probable explanation for this low sensitivity is that historical withdrawals of ground water have been located close to a source of recharge: the Rio Grande, riverside and lateral drains, and salvaged evapotranspiration. Interception of these boundaries by an expanding cone of depression is rapid; hence, the supply of water to the well from aquifer storage has historically been of short duration and small magnitude. The model was most sensitive to values of horizontal and vertical hydraulic conductivity and moderately sensitive to the removal of the lowest two model layers.

Frenzel and Kaehler (1990) observed from the simulations that approximately 80 percent of all ground water withdrawn for nonirrigation purposes was derived from the surface-water system (which includes captured ground-water discharge to the surface-water system), 10 percent from aquifer storage, and the remaining 10 percent from reduced (salvaged) evapotranspiration. These percentages remained essentially unchanged for a test of the model's response to a range of aquifer diffusivities (transmissivity divided by storage coefficient) from 0.25 to 4 times the diffusivity of the accepted model. The investigators illustrated that the importance of diffusivity increases with distance from the Mesilla Valley (the Rio Grande flood plain) by simulating an additional hypothetical withdrawal of 50 cubic feet per second for 1941-75. The additional withdrawal first was simulated along a line near and parallel with the valley and then along a line 12 to 14 miles southwest of the valley in the West Mesa area. Large (4 times the accepted value), standard (or calibrated), and small (0.25 times the accepted value) values of diffusivity were simulated. For the line near the valley, the sources of withdrawn water remained in the same proportion as for the historical simulation. For the line of wells at a distance from the valley, streamflow depletion was 53 percent of the total withdrawal for the large diffusivity, 15 percent for the standard diffusivity, and 3 percent for the small diffusivity.

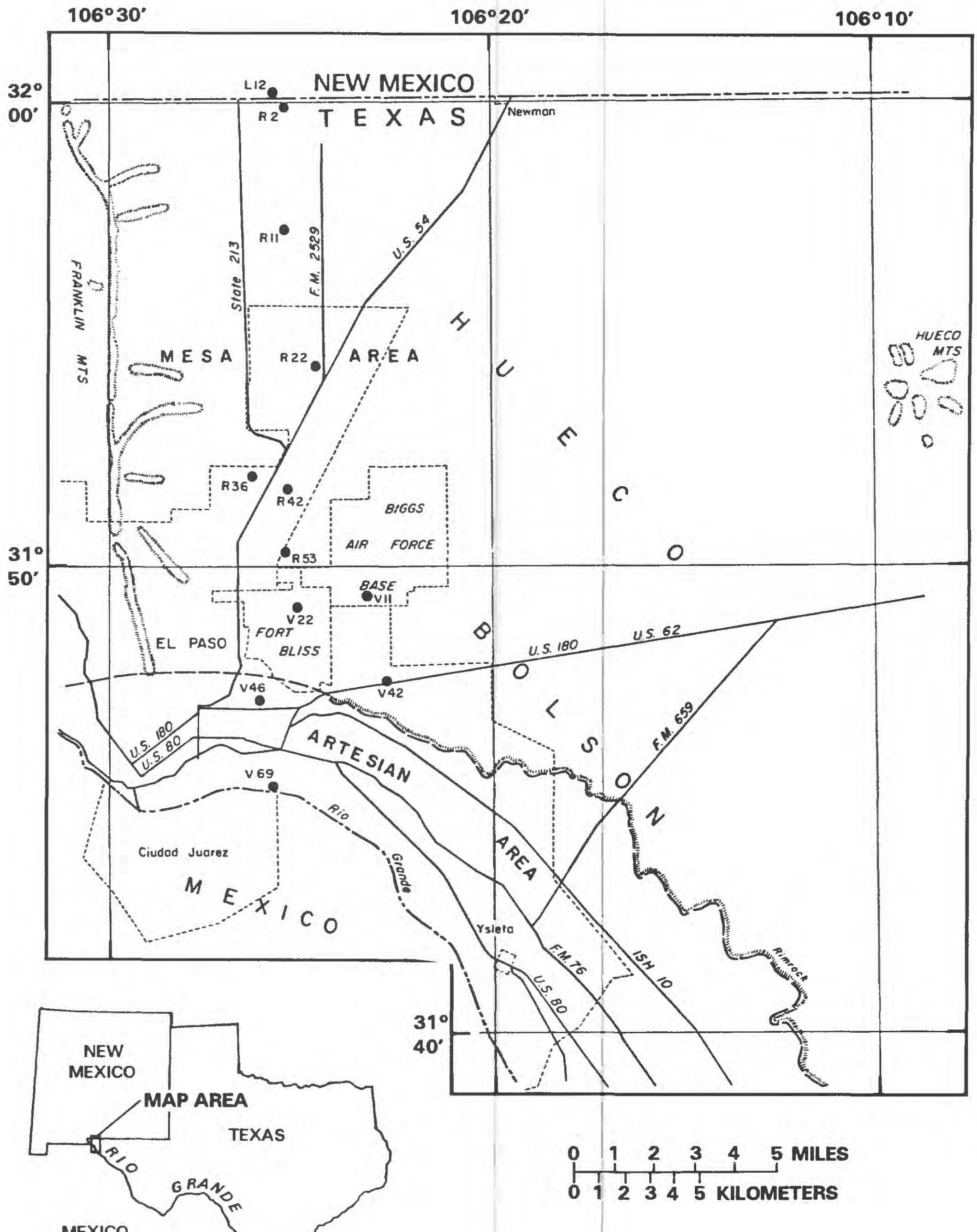
Frenzel and Kaehler (1990) indicated that refinement of the simulated aquifer properties needs to be accomplished by a program of data collection. Seasonal changes in pumpage need to be recorded. Stresses and water levels in the area of existing well fields need to be analyzed to determine storage coefficient and specific yield, and vertical conductivity. Finally, a carefully designed and executed aquifer test in the West Mesa area is needed to provide information about the aquifer in this area. The investigators also indicated that it would be useful to increase the number of model layers used to better represent the shallow part of the aquifer system, sacrificing the lowest model layer, if necessary, to minimize the computational work load.

Hueco Bolson

The Hueco Bolson (in Spanish "hueco" means hollow or empty, and "bolson" means large purse) is not actually a closed drainage basin or a basin with central drainage, as the translation implies. The Rio Grande enters the western part of the bolson through the El Paso Narrows, also known as El Paso del Norte (see fig. 15), flows between El Paso, Tex., and Ciudad Juarez, Mexico, and follows the full length of the axis of the structural basin for about 90 miles toward the southeast (fig. 17). Nevertheless, the common and published usage of the term bolson will take precedence in this report. The name lower El Paso Valley has also been applied to the Rio Grande flood-plain valley within the Hueco Bolson and in El Paso County, Tex. (The upper El Paso Valley is that part of the Mesilla Basin within Texas.)

Local naming conventions have subdivided areas that belong in one basin and have unified, in name (upper and lower El Paso Valley), basins that are distinctly separate. The Hueco Bolson and the Tularosa Basin are separately named parts of one rift basin. There is very little evidence, either hydrologic or geologic, to justify subdivision of the large structural basin into the Hueco Bolson to the south and the Tularosa Basin to the north. The two areas are delineated primarily on the presence of a low topographic divide near the Texas-New Mexico State line.

The thickness of basin fill in the Hueco Bolson has been estimated by Lovejoy and Hawley (1978, p. 57) to be about 5,000 feet at the location of the greatest structural depression at the foot of the Franklin Mountains. They also credited earlier workers for estimates of as much as 30,000 feet of structural relief between the downthrown basin and the uplifted and tilted Franklin Mountains.



EXPLANATION

– – – – – BOUNDARY BETWEEN MESA AND ARTESIAN AREAS

V69 ● OBSERVATION WELL AND NUMBER

Figure 17.--Location of the Hueco Bolson in the El Paso district. [Modified from Leggat and Davis, 1966, fig. 1].

The Hueco Bolson has been a major source of water for El Paso, Tex., and Ciudad Juarez, Mexico, which are the two largest cities in the Rio Grande rift system. Alvarez and Buckner (1980) reported that in 1973 municipal and industrial ground-water withdrawals from the bolson totaled about 14,700 acre-feet in the United States and about 30,000 acre-feet in Mexico. Irrigation withdrawal from the Rio Grande alluvial aquifer was about 24,000 acre-feet for the same year. Thus, the total withdrawal in the bolson was about 68,700 acre-feet in 1973. The city of El Paso also operates well fields in the lower Mesilla Basin, and withdrawal from these wells totaled about 22,400 acre-feet in 1969 (Meyer and Gordon, 1972).

The source and continued supply of ground water in the Hueco Bolson have been of great concern. In addition, areas of poor-quality ground water place constraints on the location, design, and withdrawal rates from wells. Thus far, depletion of aquifer storage and captured flow from the Rio Grande have supplied much of the water. However, the potential for captured flow has been diminished by lining and straightening the channel of the Rio Grande.

The primary occurrence of fresh ground water in the Hueco Bolson is in an irregularly shaped wedge of water bordering the Franklin and Organ Mountains. This wedge of freshwater overlies saline water and is kept fresh by mountain-front recharge. Sayer and Livingston (1945) estimated the amount of recharge to be 13 million gallons per day (about 14,600 acre-feet per year). The area of fresh ground water extends southward beyond the Franklin Mountains in what is commonly called the artesian area beneath El Paso and Ciudad Juarez, but is bounded on the top, bottom, and east side by saline water. The area of the Hueco Bolson outside of the inner valley (lower El Paso Valley) is often called the mesa, the bolson area, or the water-table area.

Two-Dimensional Analog Model

Leggat and Davis (1966) designed and documented a two-dimensional electric-analog model of part of the Hueco Bolson. The model was initiated because the previously employed mathematical methods "proved wholly inadequate. * * * The inadequacy was not only in the complexities of the mathematical analyses and computations, but also in evaluating the consequences of the many ways in which the reservoir might be developed" (Leggat and Davis, 1966, p. 2).

The modeled area (fig. 17) included more than the area of freshwater in the Hueco Bolson. The Franklin Mountains and Rio Grande valley were the west and southwest boundaries. The west boundary was simulated as a constant-head boundary; however, total inflow from the boundary was limited to 13 million gallons per day. The Rio Grande valley alluvium boundary was simulated as a head-dependent-flux boundary. The other boundaries (no-flow) were chosen, according to the authors, to be far enough away from pumpage stress to minimize their effect on hydraulic-head computations. The north boundary was just north of the New Mexico-Texas State line, and the east boundary was about 17 miles east of the Franklin Mountains, near the Hueco Mountains. The total modeled area was about 400 square miles.

An initial estimate of the transmissivity distribution within the aquifer was prepared in a two-step process. First, hydraulic conductivity was computed from transmissivity (determined in aquifer tests or from specific capacities of wells, if necessary) divided by the completion interval of the test well. Then, a thickness-of-freshwater map was employed to transform the hydraulic-conductivity distribution back into a map of transmissivity distribution. The range in transmissivity thus obtained was 0 to 350,000 gallons per day per foot. The hydraulic-conductivity distribution was not reported, but a quick comparison of the maps of transmissivity and thickness of freshwater indicates an approximate range in hydraulic conductivity of 125 to 1,000 gallons per day per foot squared (about 17 to 134 feet per day). The initial estimate of specific yield in the mesa area was 0.15. The initial estimate of the storage coefficient in the artesian area was 0.001. The leakance of the connection between the artesian part of the aquifer and the overlying unconfined aquifer in the valley was simulated as being 1.3×10^{-4} per day, which is equivalent to 100 feet of material having a vertical hydraulic conductivity of 0.1 gallon per day per foot squared (about 0.013 foot per day).

Leggat and Davis (1966) simulated changes in hydraulic head in response to changes in ground-water discharge for 1903-53 and 1903-63. These changes were compared with measured changes. It may be significant that the ground-water-withdrawal data reported by Leggat and Davis (1966) do not completely agree with data reported by later investigators (Alvarez and Buckner, 1980, table 1). Leggat and Davis (1966) found that their initial estimates of aquifer properties resulted in excessive simulated drawdown. Because of limitations inherent in the analog model, aquifer transmissivity and storage coefficient could only be changed as a composite term, aquifer diffusivity. The investigators modified the simulated aquifer diffusivity (transmissivity divided by storage) until measured and simulated changes in hydraulic head agreed reasonably well. The final diffusivity used in the Leggat and Davis model was about twice the original estimate.

Once they were satisfied with the analog model, Leggat and Davis (1966) projected the total change in water level for 1975 and 1990. The projections, based on a well-field design proposed by the El Paso Public Service Board and an estimated 90 million gallons per day withdrawal in 1975 and 108 million gallons per day in 1990, resulted in a simulated water-level decline of 110 feet in parts of the northern Hueco Bolson. Using the same projected withdrawals, the investigators used the model to demonstrate that an alternative well-field design would result in 10 feet less drawdown. In the El Paso-Ciudad Juarez area, Leggat and Davis (1966, fig. 11) projected a decline of 25 to 40 feet for 1903-73 and no further decline for 1974-90.

Three-Dimensional Finite-Difference Model

During 1968-69, a reach of the Rio Grande was straightened and the channel was lined, thereby making the analog model of Leggat and Davis (1966) obsolete. Rather than revise the analog model, the Texas Water Development Board requested the U.S. Geological Survey to design and construct a three-dimensional finite-difference flow model of the Hueco Bolson. Elemental in the decision was the recognized need to eventually include solute-transport capability in the simulations. A digital model would provide the definition of the flow system essential to solute-transport modeling. The area simulated and the finite-difference grid used in the model constructed and documented by Meyer (1976) are shown in figure 18. The modeled area included all of El Paso County between the Franklin and Hueco Mountains and extended about 19 miles north of the Texas-New Mexico State line.

Meyer (1976) selected a model code documented by Bredehoeft and Pinder (1970) to solve the numerical equations for a quasi three-dimensional flow system. Meyer used two model layers to simulate the unconfined and confined parts (local usage) of the aquifer system. The Rio Grande was assigned a leaky connection with the unconfined alluvial part of the aquifer.

The initial estimates of aquifer properties were made by the same method described in Leggat and Davis (1966). During the calibration process, simulated hydraulic heads were compared with measured heads for 1903 and 1936, and changes in head were compared for 1903-58 and 1903-73. During the process of matching initial heads (1903), the simulated vertical leakance between the two layers and the simulated amount of mountain-front recharge were adjusted. For simulations of transient conditions, the vertical leakance near the river and specific yield of the bolson part of the bottom layer were adjusted.

In the final model, the average transmissivity of the top layer, intended to represent 200 feet of saturated thickness of flood-plain alluvium (Knowles and Alvarez, 1979), was 30,000 gallons per day per foot (for a hydraulic conductivity of about 20 feet per day), and the specific yield was 0.20. The transmissivity of the lower layer ranged from 10,000 to 280,000 gallons per day per foot, and the specific yield in the unconfined area ranged from 0.1 to 0.3. The storage coefficient of the confined area of the lower layer ranged from 0.0001 to 0.0004. Parts of the confined area that became unconfined were assigned a specific yield of 0.14. However, transmissivity did not change.

The following observations are based on information obtained in a conversation on November 8, 1984, with Joe Gates (U.S. Geological Survey, Salt Lake City, Utah) who was involved in the investigation. The top layer of the model was simulated to be connected to the river by a leakance of 1.0 per second in only those cells over which the river passed. The top layer was connected to the underlying layer by a leakance that ranged from 0.0000001 to 9.5 per day. In the area of the model where the lower layer was simulated as being unconfined, cells in the overlying layer were made inactive.

The final model indicated that recharge along the Franklin Mountains was about 5,640 acre-feet per year. The model also showed that during 1903-73 about 50 percent of the total water withdrawn from the bolson (slightly more than 2 million acre-feet) was derived from storage in the bolson deposits (layer 2), 25 percent from downward leakage from the unconfined alluvium (layer 1), and "the rest was derived from natural recharge" (Meyer, 1976, p. 24). However, for 1968-73, only 6 percent of the withdrawn water was stated to come from recharge.

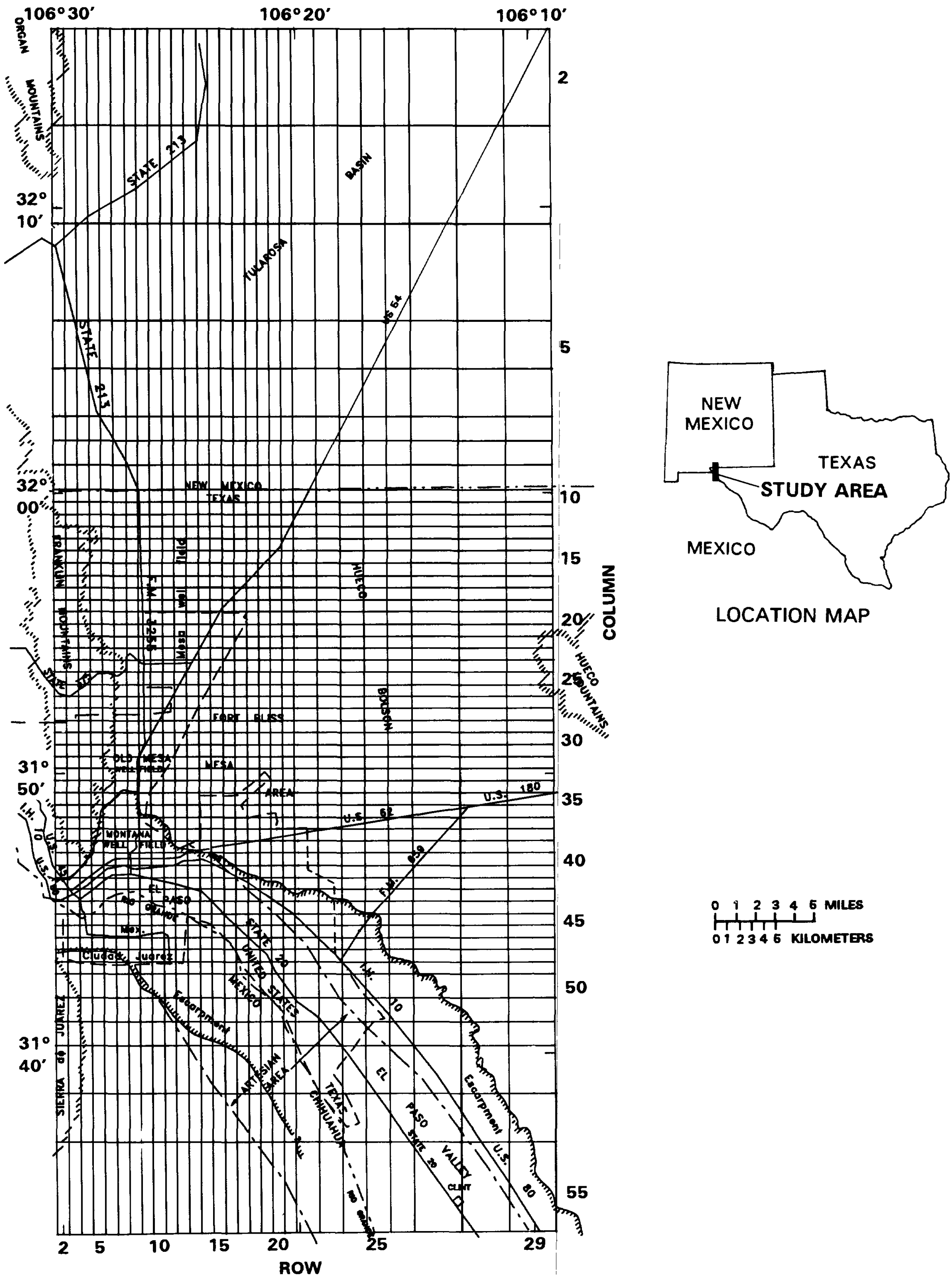


Figure 18.--Finite-difference grid for the modeled area of the Hueco Bolson. [Modified from Meyer, 1976, fig. 1].

The model was used to project water-level declines and depletion of aquifer storage for 1973-91. Because of the simulated lining of the Rio Grande channel, the declines and depletions simulated by Meyer (1976) exceeded the projections of Leggat and Davis (1966). Projected water-level declines in the El Paso-Ciudad Juarez area are as great as 140 feet for 1903-90. The model also indicated that of the total projected annual withdrawal of 115,000 acre-feet for 1974-90, 60 percent would come from aquifer storage in the bolson, 18 percent would come from induced infiltration from the Rio Grande, and 10 percent would come from depletion of storage in the unconfined alluvium. No explanation was given as to the source of the remaining 12 percent, although slightly less than 5 percent could be captured mountain-front recharge.

Because of a traditional approach to conceiving of the ground-water-flow system in the El Paso area, the full potential of a three-dimensional model devolved into a two-dimensional ground-water-flow model with an elaborate treatment of the interface between the Rio Grande and the Hueco Bolson. Because of this and the procedure used to calibrate the model, some of the simulated aquifer properties need reevaluation: Meyer (1976, p. 30) stated in his summary, "Of immediate concern is the wide range in the vertical permeabilities."

Knowles and Alvarez (1979) used the model documented by Meyer (1976) to simulate aquifer response to a revised projection of ground-water withdrawal in the Hueco Bolson. Although their work was not performed by or in cooperation with the Survey, the basis for their model was the unmodified set of simulated aquifer properties from the earlier Survey model (Meyer, 1976). "The Survey's digital model was utilized for the analysis, and aside from applying the new pumpage projections during aquifer simulation, all other parameters that had been utilized earlier by the Survey were employed in their study, including a recharge rate of 5,640 acre-feet per year" (Knowles and Alvarez, 1979, p. 4).

The revised projection of ground-water withdrawal was as much as 70 percent greater than the projection used by Meyer (1976). Knowles and Alvarez (1979) also extended their projections to 2030. The total projected withdrawal rate for 2029 was 418,000 acre-feet per year. Of this amount, the model indicated that about 5,600 acre-feet per year came from captured mountain-front recharge, 14,600 from storage in the alluvium, 83,000 from seepage from the Rio Grande, and the remaining 314,800 from storage in the bolson deposits (Knowles and Alvarez, 1979, tables 1 and 2). The investigators made four observations in their section on conclusions and recommendations. First, the simulated water-level declines in areas of the top layer exceed the thickness represented by that layer. Second, the proposed plan of future withdrawals is an efficient use of water stored in the bolson. Third, about 3.7 million acre-feet of fresh ground water will remain in aquifer storage through 2029. Finally, the authors recommended that any future model needs to include simulations of water quality.

Animas Valley

The Animas Valley (fig. 1), in extreme southwestern New Mexico, is an excellent example of an arid basin with closed surface drainage. A dune field marks the northern (downvalley) end of the topographic valley. Three large playas are south of the dune field. The Animas River, the largest stream in the valley, originates near the Mexican border and flows northward toward, but does not reach, the playas. The flow in the river ceases about midway in the basin, approximately at the point where there is a substantial increase in the thickness of the basin-fill deposits.

Although there is no surface outflow from the basin, ground water discharges northward to the Gila River. There is inflow to the valley of both surface and ground water from the Lordsburg Valley to the east. The presence of a ground-water divide between the Playas and Animas Valleys indicates that there probably is no ground-water flow between them at present. In the area of the valley where the Animas River is present there apparently is a shallow, perched ground-water system, which is adequate to supply the water needs of that area. The primary use of ground water in the basin is agricultural irrigation.

Seismic-refraction profiles indicate that there is as much as 6,000 feet of basin-fill material in the structural basin (Wilkins, 1986). A sharp contrast in thickness of fill exists between the upper (to the south) and lower Animas Valley. An abrupt thickening of fill occurs on the north end of the valley about 6 miles south of the town of Animas. As with the other rift basins, the Animas structural basin is bounded on the east and west by high-angle block faults.

Hawkins (1981) completed and documented as a master's thesis a two-dimensional finite-difference model of part of the Animas Valley. The model extended from the area of basin-fill thickening northward to about the dune area. In the preparation of this model, Hawkins (1981) first used Krieging techniques and later used flow-net analysis to determine the areal transmissivity of the aquifer. The model by Hawkins (1981) was further discussed by Hawkins and Stephens (1983).

Two-Dimensional Finite-Difference Model

O'Brien and Stone (1983) completed and documented a two-dimensional finite-difference model of the lower Animas Valley. The model was again described by the investigators in 1984 in a comparison with the techniques employed and results of an earlier model (Hawkins, 1981; Hawkins and Stephens, 1983). The work of O'Brien and Stone (1983), performed under contract with the U.S. Geological Survey as part of the SWAB RASA, explored the feasibility of simulating a closed-drainage rift basin as a two-dimensional ground-water-flow system. The model represents the culmination of a program of collection of water-level (O'Brien and Stone, 1981), water-quality (O'Brien and Stone, 1982a), and geologic data (O'Brien and Stone, 1982b) also performed under contract with the Survey.

The modeled area extends from about 6 miles south of the town of Animas to about 4 miles north of the town of Summit (fig. 19). The east and west boundaries correspond with basin-bounding faults except for the east edge of the model north of Lordsburg. The north and northeast model perimeter is simulated as a constant-head boundary. The divide between the Animas and Playas Valleys is represented by a no-flow boundary. The remaining perimeter of the model consists of constant-flux cells that represent mountain-front recharge and ground-water inflow from the upper Animas Valley.

Initial estimates of aquifer properties were taken from published reports and existing unpublished data. Where transmissivity and aquifer thickness were known, a hydraulic conductivity was calculated (but not reported). The values of hydraulic conductivity then were used in conjunction with estimated aquifer thickness to calculate a transmissivity distribution (fig. 20). The original estimate of aquifer specific yield (0.11), the average of previously reported values, was retained throughout the analysis. During the calibration process, transmissivity was altered within a 10-percent bracket and the location of simulated pumping stresses was shifted within the radius of one model cell. The final transmissivity in the calibrated model ranged from less than 50,000 to 300,000 gallons per day per foot.

In simulations of predevelopment conditions, ground-water inflow at the southern end of the basin was specified to be 4,600 acre-feet per year. Mountain-front recharge from the Pyramid Mountains on the east side of the valley was determined, by the method documented in Hearne and Dewey (1988), to be 3,000 acre-feet per year. Mountain-front recharge from the Peloncillo Mountains on the west side of the valley was determined to be 2,500 acre-feet per year. The model-computed ground-water outflow toward the Gila River was about 12,700 acre-feet per year, indicating that about 2,600 acre-feet per year of ground-water underflow enters the Animas Valley from the Lordsburg Valley. Neither evapotranspiration nor recharge from precipitation in the central part of the valley was simulated.

The transient simulations were conducted for two time spans, 1948-55 and 1955-81. For the first time span, the simulated hydraulic heads were within 10 feet of the measured or reported heads. For the second time span, the maximum error in head was 18 feet. As stated earlier, irrigation is the dominant, almost exclusive, use of ground water in the basin. O'Brien and Stone (1984) assumed that none of the water applied for irrigation returned to the ground-water system. Records of ground-water withdrawal for irrigation are not complete for the Animas Valley. In addition, the method of estimating the volume of water withdrawn changed with time. Nevertheless, after 1950 there is a consistent tendency for the irrigated area to average 12,000 to 14,000 acres and withdrawals to average about 20,000 acre-feet per year (O'Brien and Stone, 1983, table 1).

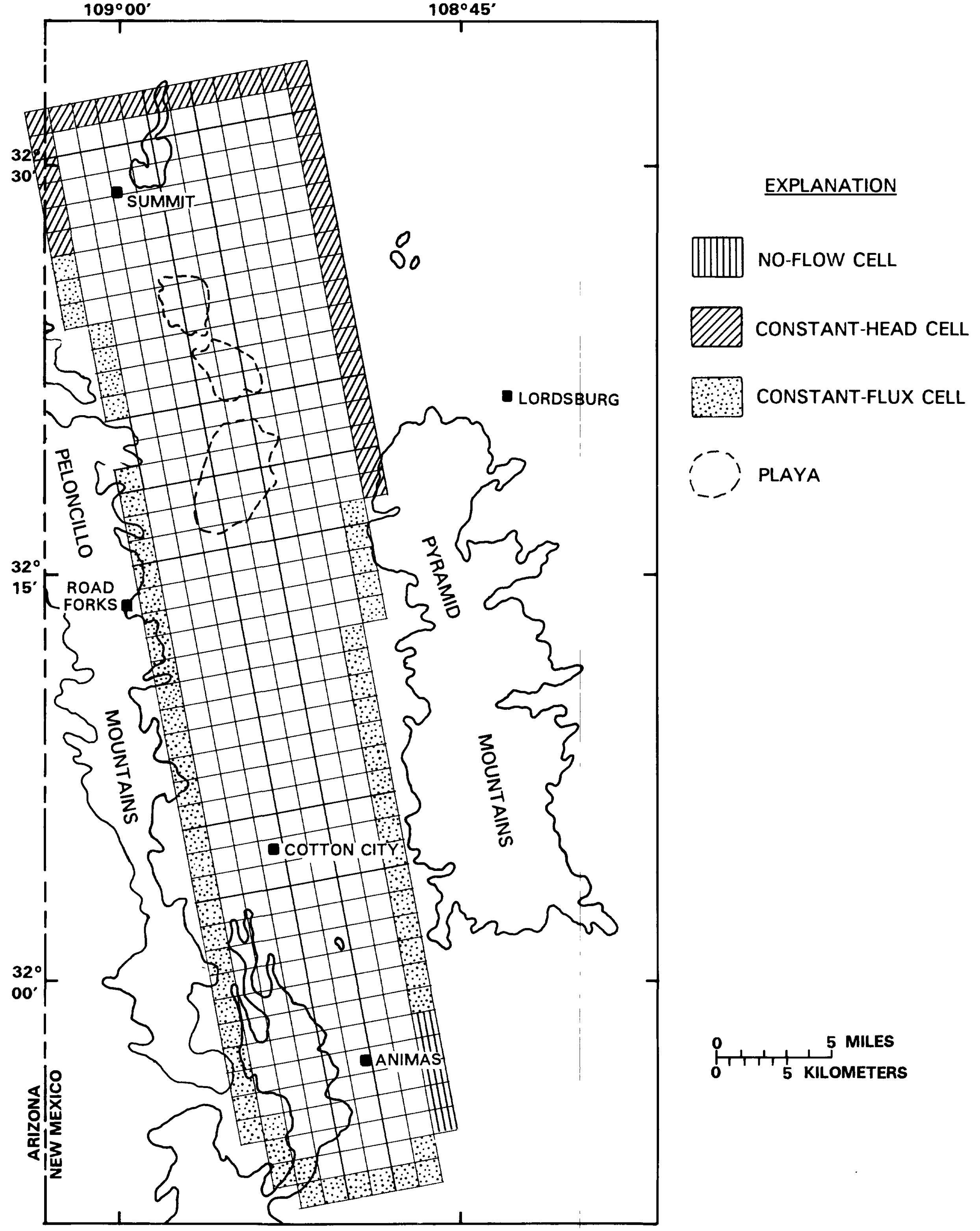


Figure 19.--Finite-difference grid and boundary conditions for the two-dimensional ground-water-flow model of the Animas Valley. [Modified from O'Brien and Stone, 1983, fig. 9].

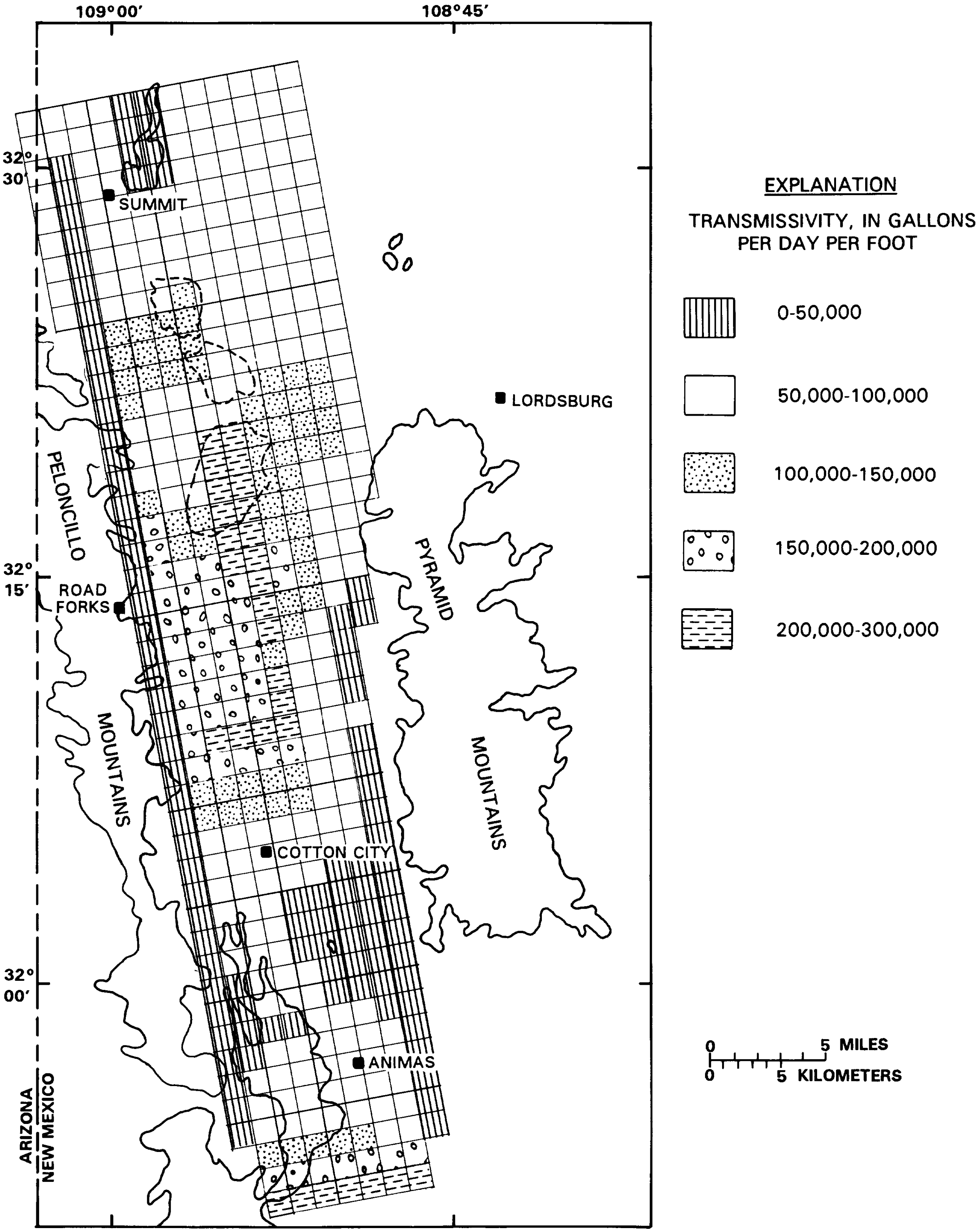


Figure 20.--Transmissivity of basin-fill aquifers in the Animas Valley. [Modified from O'Brien and Stone, 1983, fig. 12].

In their sensitivity analysis, O'Brien and Stone (1984) increased and decreased by 50 percent the modeled transmissivity, specific yield, withdrawals, and recharge. The model's sensitivity to a change in a modeled property was measured as the cumulative absolute change in hydraulic head for all active cells. Each sensitivity analysis was for a period of one calendar year, which consisted of 45 days of nonpumping, followed by 180 days of pumping and 140 days of nonpumping. The model was most sensitive to a decrease in specific yield but least sensitive to an increase in the same property. Changes in the other properties produced a relatively narrow range in model response, which was intermediate between the response to an increase and decrease in specific yield. When simulation properties were increased, changes in transmissivity and withdrawals caused the greatest error; when properties were decreased, changes in specific yield and transmissivity caused the greatest error. The model was not used to project changes in water levels beyond 1981, nor was it used to explore alternative plans of management of the ground-water resource.

Tularosa Basin

The Tularosa Basin and Hueco Bolson, as pointed out in the previous section, are the northern and southern parts of a single, large rift basin. The basin is bounded on the west by the Franklin, Organ, and San Andres Mountains and on the east by the Sacramento and Jicarilla Mountains. The northern terminus of the basin is formed by the convergence of the side boundaries. The southern end of the Tularosa Basin is defined solely by a low topographic divide near the New Mexico-Texas State line.

Surface drainage in the Tularosa Basin flows toward, but rarely reaches, the axis of the basin. Most surface runoff from the bordering mountains quickly infiltrates into alluvial-fan deposits overlying older basin-fill sediments. Lake Lucero, an alkali lake in the west-central part of the basin, is the source of windblown gypsum that forms the dunes of White Sands National Monument.

The occurrence of fresh ground water is limited to the margin of the basin, primarily in the vicinity of the mouths of the larger mountain canyons. Ground water in the center of the basin is saline and is very shallow (less than 10 feet below land surface) in a large area. Plants are sparse in the central part of the basin; therefore, transpiration is small but direct evaporation is large. Several population centers in the basin are placing increasing demands on the limited freshwater resources by expanding well fields parallel to the mountain fronts. Agricultural use of ground water is minimal.

Two-Dimensional Finite-Difference Model

Kelly and Hearne (1976) completed a water-resources investigation of the Headquarters area of the White Sands Missile Range, New Mexico (fig. 21). A two-dimensional ground-water-flow model of the immediate Post Headquarters area was included in the investigation. The investigation also included a reconnaissance of the occurrence and availability of fresh ground water in alluvial-fan deposits to the north and south of the Headquarters area, as well as water-quality data collection and a qualitative evaluation of the potential for degradation over time in the quality of withdrawn ground water.

Freshwater in the Post Headquarters area occurs in a trough-shaped volume between bedrock on the west side and saline water on the east side. The Headquarters area is located on a reentrant of basin-fill deposits into granite. Mountain-front runoff into the reentrant is the primary source of local freshwater recharge to the aquifer.

The triangle-shaped reentrant is bounded by faults that place granite (fractured and weathered near the surface) against at least 2,000 feet of basin fill. A few miles east of the Headquarters area, the thickness of fill exceeds 6,000 feet. However, a persistent and thick (as much as 100 feet) clay bed occurs at a depth of about 1,000 feet. Basin fill above the clay bed is saturated with 600 to possibly 800 feet of freshwater. Beneath the clay bed is an indurated to partly indurated conglomerate that Kelly and Hearne (1976) reported to be potentially water bearing with water of unknown quality. On the basis of 29 aquifer tests, Kelly and Hearne (1976) indicated that an area of relatively large transmissivity occurs near the center of the area of the alluvial fan. After discounting two of the tests as yielding unusually large transmissivity, the investigators reported that the values of transmissivity of the remaining tests were all less than 15,000 feet squared per day, averaged about 3,000 feet squared per day, and had a median of about 2,000 feet squared per day.

Ground-water withdrawals in the Headquarters area have remained relatively constant regardless of variations in population or staffing levels. The authors estimated that "As much as one-third of all water pumped in the Post Headquarters is used for irrigation of grass and shrubbery" (Kelly and Hearne, 1976, p. 25). These withdrawals primarily were from nine wells completed in or near the large-transmissivity area along a north-south line in the reentrant. Withdrawals of as much as 939 million gallons (about 2,890 acre-feet) in 1971 exceeded the estimated natural recharge (1,300 acre-feet per year), resulting in declining ground-water levels and encroachment of saline water. According to Kelly and Hearne (1976, p. 11), "Therefore, the Facilities Engineering Directorate, White Sands Missile Range, requested the U.S. Geological Survey to analyze the available data in order to determine if these data were adequate to predict future effects of pumping on ground-water levels, the most favorable locations and spacing of wells for future development, and the amount of potable ground water available to wells in the Post Headquarters and adjoining areas."

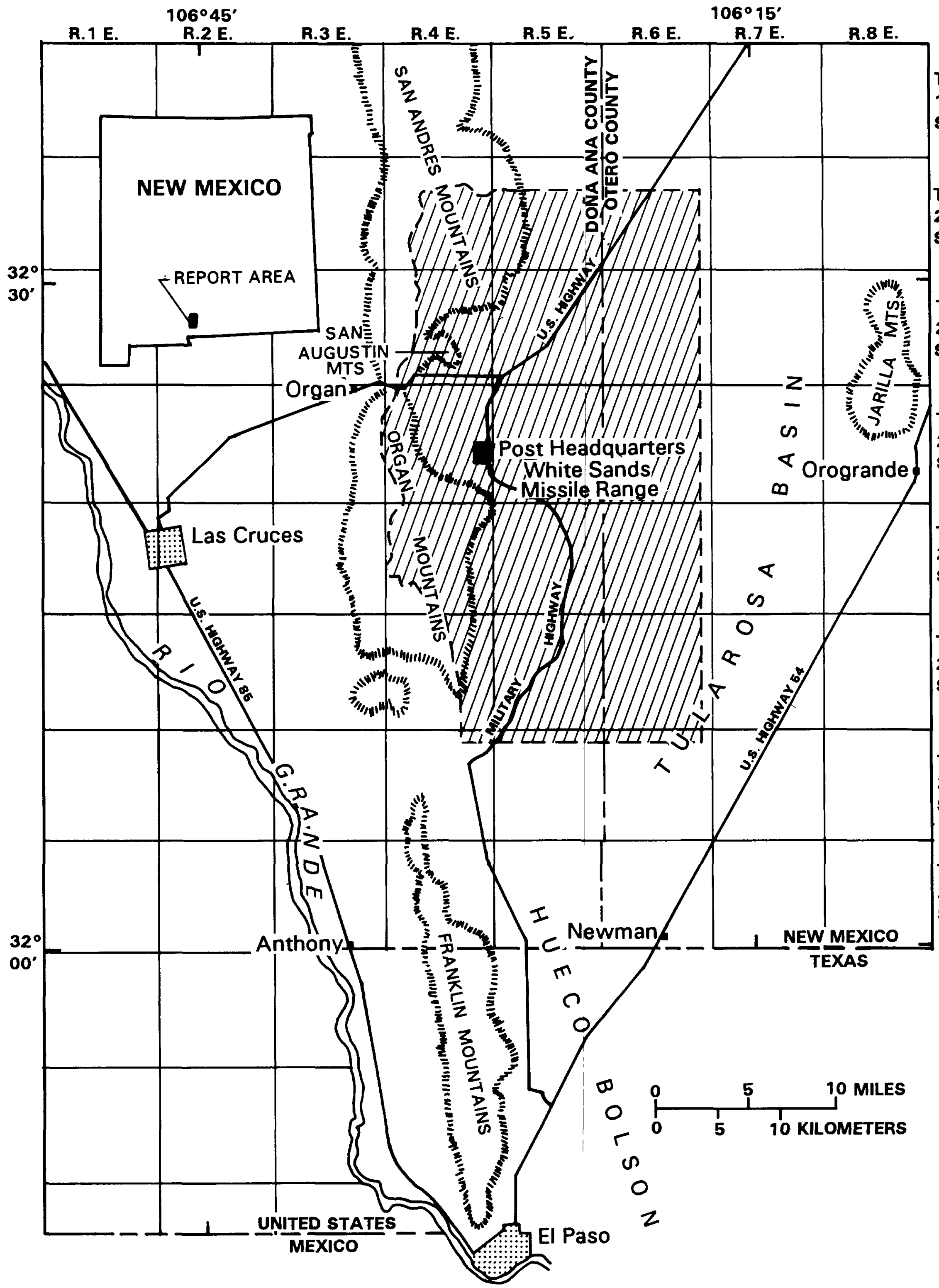


Figure 21.--Location of the White Sands Missile Range Post Headquarters and the area of investigation. [Modified from Kelly and Hearne, 1976, fig. 1].

Ground-water modeling was one aspect of the investigation. A two-dimensional finite-difference ground-water-flow model documented by Trescott (1973) was used in the simulation analysis. The modeled area, finite-difference grid, simulated aquifer thickness and hydraulic conductivity, and model boundaries are shown in figure 22. Simulated hydraulic conductivity, the parameter altered during the calibration process, ranged from 0.05 to 10 feet per day in the final model. Specific yield of the aquifer was simulated to be 0.15, and saturated aquifer thickness as either 300 or 600 feet.

Reproducing the change in ground-water levels from 1948 to 1971 was the objective of the calibration process. Once this was accomplished with acceptable accuracy, two projections of water-level decline to 1995 were made. In both projections, the 1973 rate of withdrawal was assumed to remain unchanged. The first projection, using existing well locations, resulted in a narrow cone of depression and 160 feet of drawdown at an index site. In the second projection, the total discharge was redistributed to include an additional well. The resulting cone of depression was comparatively broader but drawdown at the index site was 175 feet. However, Kelly and Hearne (1976) indicated that the second scenario was preferred because the broader cone of depression resulting from distributed withdrawals would be less likely to result in saltwater encroachment. The investigators concluded by evaluating and recommending other reentrants to the north and south of the Post Headquarters as possible future sources of freshwater.

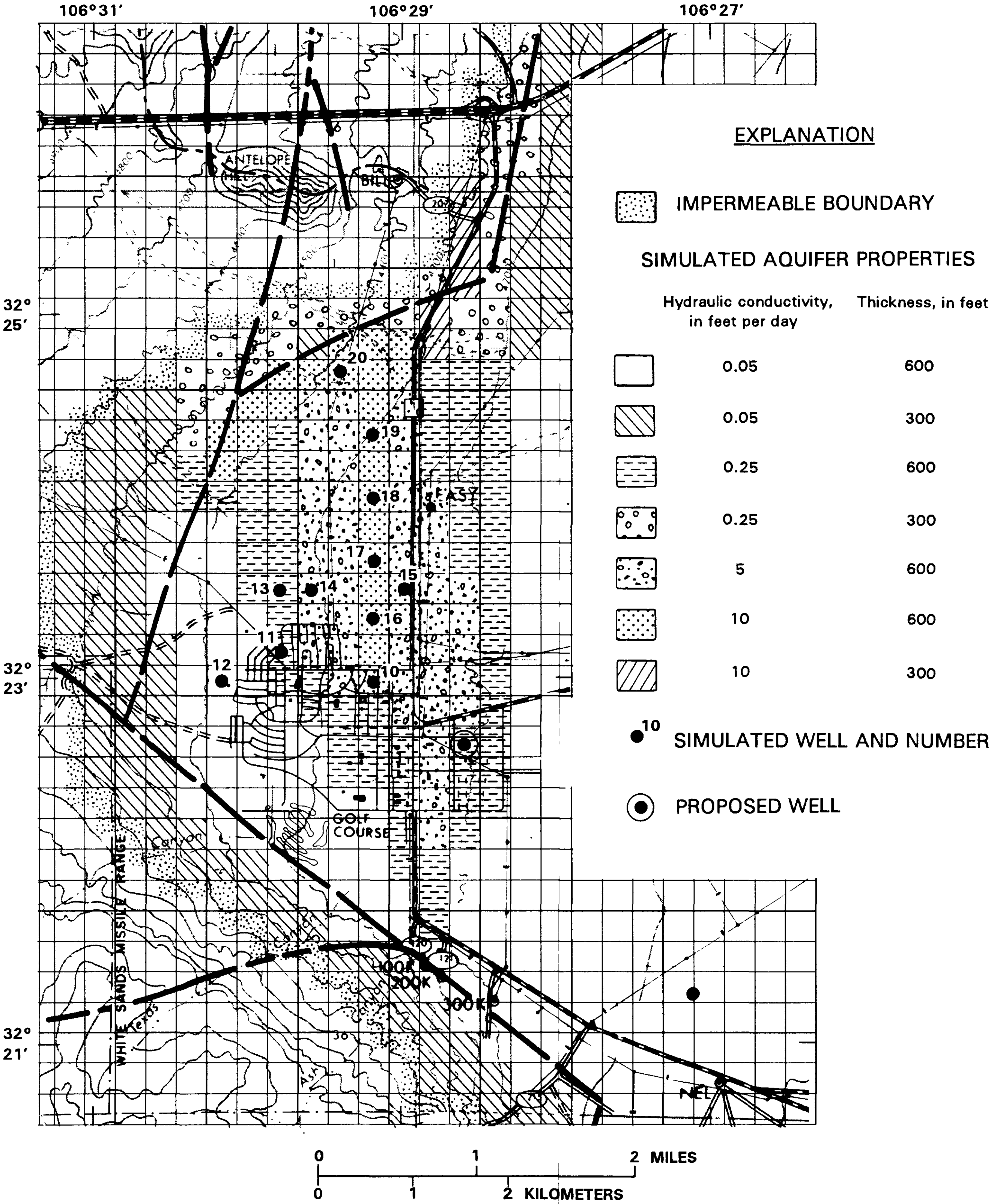


Figure 22.--Finite-difference grid, aquifer properties, and boundaries of the model of the Post Headquarters area. [Modified from Kelly and Hearne, 1976, fig. 10].

SUMMARY OF SIMULATED HYDROLOGIC CHARACTERISTICS AND MODELING APPROACHES

The varied approaches taken in preparing the models that have been described illustrate that there is no unique way to simulate ground-water flow in the Southwest Alluvial Basins. Each approach may have its own advantage, whether it be, for example, the assumed simplicity of a two-dimensional model or the assumed accuracy of a three-dimensional model. The models described generally have simulated the response of the aquifer system to the satisfaction of the investigators regardless of the initial conceptual model of the system or the choice of model code and number of dimensions simulated.

Two key factors in the development of a model are the concept of the relation between components of the flow system and the relative confidence in the accuracy of available information needed in the simulations. The first factor governs the portrayal of the system as two or three dimensional, the selection of boundary locations and types, and to some extent, the initial selection of values for simulated aquifer properties. The second factor governs the selection of those parameters that are to be altered or adjusted during the calibration process.

There is not likely to be a best approach to constructing a ground-water-flow model of an alluvial basin, but some approaches are demonstrably better than others. Because virtually any model can be made to duplicate aquifer response, the preferred approach needs to numerically duplicate the essential components of the aquifer and ground-water-flow system, within the constraints of time and cost. The preferred approaches may be determined by seeking those properties that models have in common, noting the relative success of the individual models but also remaining aware of peculiarities of the individual aquifer systems that might force departures from otherwise strong trends.

As many of the models demonstrate, two-dimensional models can successfully reproduce the response of a basin-fill aquifer system to an applied withdrawal stress. Among these models are those of Leggat and Davis (1966), Reeder and others (1967), Emery (1970), Kelly and Hearne (1976), O'Brien and Stone (1983), and finally, depending on whether the model is interpreted as being two or three dimensional, Meyer (1976). The attributes of these two-dimensional models are summarized in table 3. Although the two-dimensional models may successfully reproduce selected responses of the aquifer, they often fail to accurately mimic the function of the system. Only one (Kelly and Hearne, 1976) of the six two-dimensional models allowed simulated transmissivity to vary with time as a function of saturated thickness, and only two (Emery, 1970; Kelly and Hearne, 1976) were able to justify limiting the simulation to a shallow part of the aquifer system. For the latter two models, the presence of clays enabled the simulation of a thin and more approximately two dimensional flow system.

Table 3.—Summary of modeled aquifer properties for documented U.S. Geological Survey two-dimensional ground-water-flow models in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas

[NA, not applicable; a, estimated by this author; C, specified-head cell (constant head); L, head-dependent flux (leaky); ET, evapotranspiration (and salvaged ET); MFR, mountain-front and tributary recharge; S, aquifer storage (specific yield, specific storage, or both); T, transmissivity; Q, amount and location of ground-water withdrawals; K, horizontal hydraulic conductivity]

Senior author: Date:	Emery 1970	Reeder 1967	Leggat 1966	O'Brien 1983	Kelly 1976
Total depth (feet)	0-120	NA	NA	NA	300 and 600
Hydraulic conductivity of Santa Fe Group (feet per day)	22-56a				0.05-10
Transmissivity (1,000 gallons per day per foot)	10-200	100 and 200	0-350	50-300	
Specific yield	0.20	0.20	0.15	0.11	0.15
Storage coefficient					0.001
River boundary	C	C	L	NA	NA
Other boundaries	ET		MFR	C	
Primary properties altered during calibration			S,T	Q,T	K
Major source of water to wells	ET	C		S	S

Constant transmissivity often is assumed in two-dimensional models because aquifer thickness is great relative to the amount of desaturation of the aquifer and, therefore, the relative change in transmissivity is assumed to be negligible. This assumption often can be demonstrated to be true. However, the saturated thickness of basin fill almost always greatly exceeds the completion interval of most water wells, resulting in vertical components of flow (spherical drawdown) in a completely isotropic aquifer or confined horizontal flow in a section much thinner than the total thickness of basin fill depending on the ratio of horizontal to vertical hydraulic conductivity.

Partial penetration of the wells causes vertical components of flow, which is a violation of one of the basic assumptions upon which a two-dimensional model is founded. As a result of this violation, use of two-dimensional models tends to promote the generation of erroneous estimates of hydraulic conductivity and estimates of aquifer storage coefficient that are intermediate between values typical of confined and water-table systems. Estimated values of hydraulic conductivity may be large or small, depending on the spatial relation between the completion intervals of observation and production wells. Failure to recognize that the saturated thickness of the simulated aquifer system is great relative to the completion interval of most of the production wells is one of the greatest problems arising from the use of two-dimensional models. However, under ideal conditions (long time and small stress), the simulated values of the aquifer properties tend to converge to the actual values.

In comparison with two-dimensional ground-water-flow models, three-dimensional models may more accurately portray the flow system of the basin-fill aquifer system by simulating vertical components of flow. However, the worth of the model is still a function of the accuracy of the hydrologist's concept of the workings of the aquifer system. The models described by Emery and others (1975), Meyer (1976), Knowles and Alvarez (1979), Hearne (1985a, b), Gates and others (1984), Kernodle and Scott (1986), Kernodle and others (1987), Hearne and Dewey (1988), and Frenzel and Kaehler (1990) are the three-dimensional models discussed in this report. The attributes of these models are summarized in table 4.

The model documented by Meyer (1976) and later used by Knowles and Alvarez (1979) more closely resembles a two- rather than three-dimensional model and earlier comments about two-dimensional models apply. The model by Gates and others (1984) employed a valid but poorly explained treatment of aquifer storage that cast doubt on the validity of the simulations even though the model robustly duplicated the response of the aquifer system to historical stress.

Calibration to historical stress in no way assures that projected future responses are correct, especially if the function of the system is intentionally or inadvertently misrepresented. The closer the model is to mimicking the function of the system, the more likely it is to accurately project the response of the system. The remaining six three-dimensional models (Emery and others, 1975; Hearne, 1985a, b; Kernodle and others, 1987; Hearne and Dewey, 1988; and Frenzel and Kaehler, 1990), may be called qualified successes at reproducing both the function and response of the aquifer system.

Table 4.—Summary of modeled aquifer properties for documented U.S. Geological Survey three-dimensional ground-water-flow models in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas

[NA, not applicable; a, approximate estimate made by this author; -, indicates a range of values; /, indicates discrete values; b, storage coefficient; L, head-dependent flux (leaky); C, specified-head cell (constant head); R, head-dependent flux (with flow routing river and drains); ET, evapotranspiration (or salvaged ET); MFR, mountain-front and tributary recharge; Q, amount and location of ground-water withdrawal; K, horizontal conductivity; B, boundaries or boundary type; VK, vertical conductivity; I, irrigation-return flow; S, aquifer storage (specific yield and/or specific storage)]

Senior author: Date of publication:	Emery 1975	Hearne 1988	Hearne 1985a	Hearne 1985b	Kernodle 1987	Gates 1984 (Note 1)	Frenzel 1990	Meyer 1975
Active layers in model	3	7	22	15	6	3	5	2
Total depth (feet)	3,120	3,200	4,000±	1,166	6,075	1,200	3,450	3,000+
Thickness of top layer (feet)	0-120	0-150	300±	NA	200±	240	200±	200
Hydraulic conductivity (feet per day):								
Santa Fe Group	27a	30-45	1.0	0.3-2.7	30/40	13	3-22	17/134a
Alluvium	NA	NA	NA	NA	50	117	70/140	20
Simulated fines	NA	10	NA	NA	0.25/2/3	NA	NA	NA
Anisotropy ratio	2,000	Note 2	330	250/20,000	500	20/500	200	0.0035-33,000

Table 4.—Summary of modeled aquifer properties for documented U.S. Geological Survey three-dimensional ground-water-flow models in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas—Concluded

Senior author: Date of publication:	Emery 1975	Hearne 1988	Hearne 1985a	Hearne 1985b	Kernodle 1987	Gates 1984 (Note 1)	Frenzel 1990	Meyer 1975
Specific yield	0.20	0.20	0.15	0.15	0.10	0.15/0.10	0.20	0.1-0.3
Specific storage (x 10^{-6} per foot)	5	5	2	2	1	1.8/1.2	1	0.0001-0.0004b
River boundary	L	L	C,L	NA	C	C	R	L
Other boundaries	ET	MFR,ET	MFR	C	MFR		MFR,ET	MFR
Primary properties altered during calibration		Q	Q	K,B,VK	K,B	VK,I,K	K,VK,R	VK,S
Major source of water to wells	ET	ET	S	C	MFR	C	R	S

Note 1: (Gates, 1984) First number is the anisotropy ratio for material beneath the flood plain, specific yield for flood-plain alluvium, and specific storage for the middle layer beneath the flood plain. Second number is the anisotropy ratio for the aquifer outside of the flood plain, aquifer storage for all layers outside of the flood plain, and specific storage for the bottom layer beneath the flood plain.

Note 2: (Hearne and Dewey, 1988) 0.67/170/670/2,300/67,000. Most of the aquifer was simulated as having a horizontal to vertical anisotropy ratio of either 670:1 (shallow artesian) or 2,300:1 (deep artesian).

The following discussion is limited to the last six models, unless other models are explicitly referenced. One of the six (Kernodle and others, 1987) is a marginal candidate for further discussion because, as the investigators pointed out, the selected representation of the river and flood-plain system as a specified-head boundary is inappropriate for simulations with large nearby applied stresses. The attributes that these models have in common form the basis for developing guidelines for the construction of better (but definitely not best, optimum, or perfect) ground-water-flow models of basin-fill aquifer systems in the Southwest Alluvial Basins region.

As stated earlier, there are two key factors in the development of a ground-water-flow model: the hydrologist's perception of the mechanics of the flow system and the hydrologist's evaluation of the relative worth of the data at his or her disposal. The first factor determines the selection of the type of model employed, the simulated geometry (both external and internal), and the initial data requirements. The second factor determines which of the simulation parameters are most likely to be altered during the calibration process. The element of subjectivity at this stage of the modeling process is at least partly responsible for the variety of models described earlier.

Geometry of the Basin-Fill Aquifers

Of the six models selected for comparison, all but one (Emery and others, 1975) employed five or more model layers to portray the aquifer system. The number of layers is not necessarily a function of lithologic layering in the aquifer system although such layering is documented and explicitly represented in all of the models except Frenzel and Kaehler (1990). A common misconception is that three-dimensional data (hydraulic head, hydraulic conductivity) are required to justify a three-dimensional flow model. As a general rule, the ability of the model to mimic three-dimensional flow increases with the number of model layers. There is, however, an upper limit to the number of layers that is determined, in each instance, by the time and resources available to the investigating hydrologist.

The six models all gave definition to the shallow part of the ground-water-flow system. The top layer of most of the models was 200 feet or less in thickness and the top layer of two of the models represented a thinning to zero. A thin top layer allows separation of surface-water stresses from the bulk of the ground-water system. The need for this separation arises from the cell-centered numerical approach of most of the commonly used simulation algorithms. Simulated flow to or from a surface-water system does not actually occur across the simulated top of the aquifer but is instead simulated at the cell centers of the uppermost active model layer. If ground-water stresses also are simulated in the top layer, the simulated flow path to or from a surface-water system to the ground-water system is greatly distorted. The simplest solution is to attempt to isolate the ground-water/surface-water interaction in one layer and the remaining ground-water system with its applied stresses in one or more other layers. Care needs to be taken to transfer downward the simulated surface boundary if any of the layers are so thin that they desaturate during the simulation.

The simulation depth also influences the accuracy of the model. However, Hearne and Dewey (1988) demonstrated that the entire thickness of basin fill need not be simulated. Their work showed a very small sacrifice in accuracy resulting from simulating only the top 3,200 feet of a total thickness of more than 17,000 feet of basin fill. Likewise, Kernodle and Scott (1986) and Frenzel and Kaehler (1990) demonstrated that the models they constructed were not exceptionally sensitive to the removal of the lowest model layers. Given a ratio of horizontal to vertical hydraulic conductivity in the range of 200:1 to 700:1, a total simulated thickness of 6,000 to 3,000 feet, respectively, generally is adequate for the basin-fill aquifer systems in the study area. This guideline is, however, subject to revision should transient stresses be applied at depths approaching these limits.

Significant internal geometric features were recognized in three of the four basins that were modeled. Emery and others (1975) and Hearne and Dewey (1988) included the properties of an extensive clay series in their simulations. The 22 model layers in the model documented by Hearne (1980b) were tilted to represent a regional direction of dip that he noted in the beds of the aquifer system in the Española Basin. The model by Kernodle and others (1987) included an eastward-dipping "tight zone" as well as internal no-flow boundaries intended to represent vertical fissure-flow volcanics. This model also included areas of small horizontal hydraulic conductivity that had been noted by previous investigators. As a rule, identifiable geologic features that affect ground-water flow paths, including geologic structure and lithology of beds, need to be represented in the model.

Aquifer Properties

A remarkable uniformity in the values for simulated aquifer properties is apparent in table 4. The hydraulic conductivity of the Santa Fe Group (basin-fill deposits) has a range of 0.25 to 45 feet per day. The smaller conductivity is associated with previously recognized and mapped fine-grained deposits. The fine-grained deposits were simulated as having conductivity in the range of 0.25 to 10 feet per day. The entire basin fill in the Española Basin and most of the basin-fill deposits in the closed basins are included in the category of fine-grained deposits. The hydraulic conductivity of most of the Santa Fe Group along the Rio Grande ranges from 20 to 40 feet per day. The hydraulic conductivity of the flood-plain alluvium ranges from 50 to 140 feet per day but the majority of the alluvium was simulated with a conductivity of 50 and 70 feet per day.

Specific yield was simulated within the range of 0.10 to 0.20 and specific storage within the range of 1×10^{-6} to 5×10^{-6} per foot. Aquifer storage was the source of most water withdrawn from the aquifer in only one of the six three-dimensional models (Hearne, 1985a). Most of the six models were found to be relatively insensitive to storage coefficients for simulations of historical withdrawals. However, two-dimensional models (including Meyer, 1976) of closed basins indicated that storage depletion was a primary source of water to wells; these models were sensitive to changes in the simulated storage coefficient. The difference in sensitivity is due to the type of basin rather than the number of model dimensions.

The ratio of horizontal to vertical hydraulic conductivity was simulated to range from 0.67:1 to 67,000:1, but the range commonly was from 200:1 to 2,500:1. The higher ratios were associated with volcanics intercalated with basin-fill sediments or, in the case of Hearne (1985b), with the local response of the aquifer to a short-term, large-magnitude stress. The lower ratios were associated with large vertical conductivity along fault zones.

The outliers in the parameter values are not an artifice that the modeler evoked to bring about model calibration. Without exception, the departures have an explanation that is founded on published and well-documented phenomena and data.

Boundary Conditions

The choice of simulated boundary locations and types is a direct function of the modeler's conceptualization of the hydrologic system and, therefore, plays an especially important role in the simulation. There are five main categories of boundaries in the alluvial basins in the area of investigation: (1) internal boundaries that alter flow paths, including small-permeability beds, fissure-flow volcanics, and faults; (2) recharge boundaries, primarily around the perimeter of the basins (mountain-front recharge) and along the channels of intermittent streams, arroyos, and washes (tributary recharge); (3) recharge and discharge boundaries associated with semipermanent surface-water systems in the flood plains of major streams; (4) evapotranspiration by native vegetation and crops from shallow ground water and net irrigation flux to or from ground water; and (5) ground-water withdrawals from wells. The first category, internal boundaries, was described in the section on aquifer geometry.

Mountain-Front and Tributary Recharge

Estimates of mountain-front and tributary recharge were employed in most of the models that have been discussed. Some investigators, however, avoided the task (and a certain degree of risk) of estimating recharge by using the superposition approach to simulate only the changes in the stresses on the flow system. Recharge estimates have been prepared using several techniques, the simplest of which is the application of Darcy's law (Darcy, 1856), according to which the flux is equal to the product of hydraulic gradient, cross-sectional area, and hydraulic conductivity. Use of this method has the potential disadvantage of assuming that hydraulic conductivity is known with reasonable accuracy.

Another method of estimating recharge is to use water-budget analyses to determine what percentages of precipitation become runoff, evapotranspiration, and infiltration. Hearne and Dewey (1988) formalized this approach by using regression analyses to quantify the factors governing runoff and recharge in basins with hydraulic instrumentation, then extending the relation to drainage basins without instrumentation. As mentioned earlier, all models completed as part of this investigation (SWAB), as well as several U.S. Geological Survey models currently in progress, have used or are using the technique developed by Hearne and Dewey (1988) to estimate recharge.

When describing the hydrologic function of the system, it is more desirable to include recharge than to exclude it. Also, if time, resources, and data are available, the method developed by Hearne and Dewey (1988) or a similar method is useful for estimating recharge.

Surface-Water Boundaries

Interaction between the ground-water system and hydrologic processes taking place at or near land surface is one of the major boundary activities in the alluvial basins. The selection of the numerical representation of these boundaries is therefore of major importance to the success of the simulation in portraying the dynamics of the aquifer system. Streamflow and drain flow, and the interaction of these surface features with the ground-water system, are a major part of the total water budget of basins with through-flowing drainage. There are four common ways of representing surface-water boundaries in a ground-water-flow model: (1) as a specified-flux boundary, (2) as a specified-head boundary, (3) as a head-dependent-flux boundary, or (4) as a head-dependent-flux boundary with routing of surface flow and limiting the ground-water recharge to the available streamflow.

Using a specified-flux boundary to represent surface-water seepage to the ground-water system is a common method of simulating mountain-front and tributary recharge. This boundary type is especially useful in portraying flow of an estimated quantity of water from a perched stream across an unsaturated zone to the water table, assuming that the processes taking place in the unsaturated zone may be ignored.

Several models employed specified-head boundaries to represent all or part of the surface-water system. Of the six three-dimensional models being discussed, one (Hearne, 1985a) employed specified-head cells to represent the Rio Grande and another (Kernodle and Scott, 1986; Kernodle and others, 1987) represented the entire flood-plain alluvial system with specified-head cells. Kernodle and others (1987) pointed out that representation of the flood-plain alluvium as a constant, specified-head boundary is not appropriate when simulating large, nearby ground-water withdrawals. Otherwise, the portrayal is adequate unless the processes taking place within the flood-plain system are of concern. Two-dimensional models by Reeder and others (1967) and Emery (1970) and the three-dimensional model by Gates and others (1984) also employed specified-head boundaries to represent the Rio Grande surface-water system.

Head-dependent-flux boundaries without routing of surface-water flows were used in the three-dimensional models of Emery and others (1975), and Hearne (1985a). This representation has the advantage over a specified-head boundary in that ground-water levels can change in all parts of the simulated aquifer. Head changes may pass through the cells underlying the boundary, and changes in aquifer storage and evapotranspiration are more accurately simulated.

Head-dependent-flux boundaries with routing of surface-water flow have a number of advantages over a simple head-dependent-flux boundary when used in simulations of near-surface aquifers in arid or semiarid climates. Surface-water bodies commonly are intermittent or begin and cease to flow at varying locations. Not only does the flux to or from ground water change, but the location of the boundary also may change. The ability to simulate a moving boundary is the primary asset of this boundary type. Using this boundary type also allows gaged surface-water flows and flow depletions to be considered during the calibration process. However, comparisons of surface flow need to be given less importance than ground-water observations because none of the existing codes actually use stage-discharge relations to route flows; the simulated surface-water flows are somewhat questionable. Two of the six models routed surface flows. Hearne and Dewey (1988) used surface-water flow routing to simulate the Rio Grande in the San Luis Valley and also to represent evapotranspiration. Frenzel and Kaehler (1990) used surface-water flow routing to simulate the Rio Grande and the extensive network of drains in the Mesilla Basin. Of the four methods of representing interactions between surface water and ground water, the head-dependent-flux boundary with surface-water flow routing is the closest analogy to the function of the actual hydrologic system.

Evapotranspiration and Net Irrigation Flux

In basins having through-flowing drainage and in most basins having closed drainage, evapotranspiration by native vegetation and crops is a major part of the water budget. Evapotranspiration by native vegetation primarily is from ground water. Water evapotranspired by irrigated crops may be from either surface or ground water. Although evapotranspiration by crops is a net loss to the overall hydrologic system, the ground-water system may show a net gain or loss depending on the source of the irrigation water. In the southwest United States, any alteration of surface-water flow or any water salvage from evapotranspiration is of great importance. If the modeled area has regions where the depth to water is less than about 50 feet, including evapotranspiration in the simulations needs to be considered. If the depth to water is less than 20 feet, evapotranspiration unquestionably needs to be included in the simulation.

The possibility of irrigation water recharging ground water needs to be investigated whenever there is irrigated acreage in the area being simulated. When the source of water for irrigation is surface water, the net gain to the ground-water system may be as much as one-third of the applied water. If the source is ground water, the net loss to the ground-water system may be only two-thirds of the total withdrawn.

Simulation Parameters Altered During Calibration

Simulation parameters that are altered during the calibration process often are an indication of the hydrologist's lack of confidence in the reliability of the available data. Also, those parameters to which the models are the most sensitive are among the most likely to be altered. Finally, tradition may influence the choice of parameters to be altered during calibration; for example, there is a tendency to alter the values of simulated horizontal hydraulic conductivity instead of simulated vertical hydraulic conductivity. The variety in the simulated properties altered during the calibration process is shown in tables 3 and 4, but there is an observable pattern that could be a guide for future data-collection programs.

Electric-analog models are very difficult to modify once constructed. For this reason, the simulated hydraulic properties often are not altered, or only those that are relatively easy to change are altered. For example, Leggat and Davis (1966) were able to modify only aquifer diffusivity, the ratio of transmissivity to storage, without being able to identify which of the two or combination of the two simulated properties was being altered. Emery and others (1975) did not document any changes to improve their analog model.

Records of ground-water withdrawals for agricultural use often are the least trusted and most adjusted of the data required for models of transient ground-water flow. The location and magnitude of irrigation withdrawals commonly are changed in models of the closed-drainage basins that have extensive irrigation (O'Brien and Stone, 1983; Hearne and Dewey, 1988). Inaccurate or incomplete data for agricultural water use in basins with the through-flowing Rio Grande have been less of a problem in simulations of those basins because: (1) The majority of that water is from surface diversions rather than ground water, and (2) ground-water withdrawals for irrigation of crops in the flood plain are very near a surface-water boundary and usually have relatively little impact on ground-water levels. However, uncertainty about agricultural ground-water withdrawals caused Frenzel and Kaehler (1990) to choose 1975 as the end of the simulation period for their Mesilla Basin model.

Among the six three-dimensional models selected for close examination, the next most common simulation parameter to be altered is horizontal hydraulic conductivity (or transmissivity), and the third is boundary representation. Changes in simulated horizontal hydraulic conductivity often result in the most readily observable change in simulated water levels. The dimensions of cones of depression centered about wells or well fields are a function of hydraulic conductivity and boundary location for most of the basin models. Rate of development of a cone of depression is a function of both hydraulic conductivity and storage coefficient. However, in most basin models the simulated stress is located close enough to recharge boundaries that the importance of storage is minimized, leaving only horizontal conductivity and boundary portrayal as important simulation parameters.

The parameters least likely to be changed are aquifer storage and vertical hydraulic conductivity. In those models where the transient stress was located at a large distance from the simulated sources of recharge (the three-dimensional model of Hearne, 1985a, and the two-dimensional models of Meyer, 1976, and O'Brien and Stone, 1983), aquifer storage was the major source of water. However, Meyer (1976) altered simulated storage to attain calibration, whereas Hearne (1985a) and O'Brien and Stone (1983) did not, a reflection of the confidence these authors placed on their initial estimates of storage coefficient. Frenzel and Kaehler (1990) demonstrated the potential of their model to be sensitive to aquifer diffusivity (transmissivity divided by storage coefficient) in the event that proposed ground-water withdrawals begin away from the Rio Grande flood plain.

Vertical hydraulic conductivity can be an important simulation parameter only in those models that attempt to simulate vertical distributions of stress and aquifer response. Furthermore, the sensitivity of a three-dimensional model to changes in simulated vertical hydraulic conductivity is related to the thickness of the model layers and to the distribution of stresses within the layers. A three-dimensional model with all or most of the stress in a single, thick layer probably will be insensitive to changes in simulated vertical hydraulic conductivity.

Vertical hydraulic conductivity, as well as the ratio of horizontal to vertical hydraulic conductivity (anisotropy), is one of the least defined of the aquifer properties of the Santa Fe Group and related basin-fill deposits. Even so, for most of the three-dimensional models that have been described, the hydrologists made an initial estimate of the property that was left essentially unchanged during the calibration process. A possible explanation is that traditional analytical hydrology, specifically the field of aquifer-test analysis, has virtually ignored vertical anisotropy within an aquifer unit, usually dealing with the problem by distorting the coordinate system to account for the anisotropy. This manner of dealing with internal anisotropy seems to have carried over from analytical to numerical hydrology. Until the advent of numerical models, the effects of anisotropic conditions or of multiple boundary conditions on the analysis of an aquifer test could not be determined.

Hearne (1985b) and Kernodle and others (1987) considered the problem of vertical anisotropy within a unit as one of scale. They suggested that under the large-stress and short-duration conditions of an aquifer test, small-scale interbedding of units of contrasting hydraulic conductivity becomes important, and the apparent ratio of horizontal to vertical hydraulic conductivity approaches the ratio of the arithmetic mean of the horizontal hydraulic conductivities to the harmonic mean of the vertical hydraulic conductivities. Under the conditions of long time and small stress, the aquifer appears to be more uniform (less anisotropic) because "The discontinuity of less permeable beds may improve the crossbed communication by providing a tortuous path around * * * these beds" (Hearne, 1985b, p. 22). One conclusion to be drawn is that the degree to which the ground-water-flow system appears to be anisotropic and three-dimensional depends on the stress on the system. Another conclusion is that to simulate a stressed anisotropic system, attention needs to be given to the vertical components and internal details of that system.

GUIDELINES FOR CONSTRUCTION OF GENERALIZED MODELS

The preceding section discussed the similarity of six selected three-dimensional ground-water-flow models of the alluvial basins in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas. After accounting for those features that make each basin unique (primarily the hydrogeologic framework), the models are so nearly the same that their attributes can serve as guidelines for the rapid construction of uncalibrated, yet reasonably accurate, general models of basins that have not been extensively investigated and simulated. The most appropriate use of these general models would be to aid in the design of a data-collection program customized to the needs in the specific basin or to make an initial evaluation of a specific problem. The guidelines for construction of a generalized model of a specific basin are as follows:

(1) Perform a literature search to determine basin geometry, geologic structure, and lithology. Hawley (1978), Birch (1980b), and Wilkins (1986) are examples of suitable initial references.

(2) Use a three-dimensional model to simulate the aquifer to a depth of approximately 4,000 feet or to the total depth of the basin if less than 4,000 feet. Use at least five model layers, the top layer being 200 feet or less in thickness.

(3) Simulate the basin-fill aquifer system as having a horizontal hydraulic conductivity of 20 to 45 feet per day in the open-drainage basins and 2 to 10 feet per day in the closed-drainage basins, except where field data indicate otherwise. Simulate fine-grained playa or lake deposits as having a hydraulic conductivity of 0.25 to 10 feet per day and flood-plain alluvial deposits as having a hydraulic conductivity of 50 to 70 feet per day.

(4) Do not vary horizontal hydraulic conductivity as a function of depth unless specific lithologies are being simulated. Compaction of the aquifer and increases in temperature with depth need not be simulated as affecting the apparent hydraulic conductivity, except where these specific problems are being addressed. The two factors have opposite, and potentially offsetting, effects.

(5) Use a horizontal to vertical hydraulic-conductivity ratio of from 200:1 to 1,000:1 except where geologic features such as faults, clay sequences, or steeply dipping beds exist.

(6) Simulate aquifer specific storage to be in the range of 2×10^{-6} to 5×10^{-6} per foot and specific yield in the range of 0.10 to 0.20.

(7) Include rivers and drains, if present, in the simulations as head-dependent-flux boundaries, preferably with flow routing to allow the location of the boundary to change with time.

(8) Include estimated mountain-front and tributary recharge, evapotranspiration, and net irrigation flux.

(9) Include historical ground-water withdrawals.

Numerous publications are available from which similar guidelines may be extracted or inferred and from which estimates of aquifer properties may be obtained. Some of these are Todd (1959), Lohman (1972), Freeze and Cherry (1979), Heath (1980), Mercer and Faust (1981), Wang and Anderson (1982), Franke and others (1984), Reilly and others (1984), and McDonald and Harbaugh (1988).

Two three-dimensional models were constructed to test the feasibility of constructing general models of the basin-fill aquifer systems. These models will only be described briefly and not documented. The guidelines were refined during construction of the two models. Because the models were completed before the guidelines were fully developed, they have some attributes that are at the guideline limits. For this reason and because the models are not considered to be calibrated, the values for simulated aquifer properties are not reported.

Both models were of basins along the Rio Grande. They used the same compiled computer code that limited the dimensions of the models to 40 rows, 70 columns, and 5 layers. They both simplified the surface-water system to simulate only the Rio Grande and two drains, one drain on either edge of the flood plain. To make efficient use of the model dimensions, the model grids were aligned with the general bearing of the flood plain. Finally, the two models were completed in 6 weeks by an impartial student hydrologist.

The first model was of the Albuquerque-Belen Basin. This basin was selected because the model by Kernodle and others (1987) was readily available for comparing the results and thereby evaluating the worth of the generalized models. No data other than historical water withdrawals and estimates of mountain-front and tributary recharge were transferred directly from the model of Kernodle and others (1987) to the general model. The generalized model was as acceptable a model as the one used as a reference. In one aspect, the generalized model was superior: the flood plain was not represented as a specified-head boundary and the surface-water system was represented by a head-dependent-flux boundary with surface-water flow routing. Therefore, the simulations for later time periods produced more realistic water-level declines in the vicinity of the surface-water boundaries. Streamflow depletion and loss of water to evapotranspiration also were computed.

Kernodle and others (1987) used the mean absolute error between simulated and measured or reported hydraulic heads at 34 wells (37 values) as a measure of the degree of calibration of their model. Their reported error was 14.1 feet. In comparison, the generalized model had a mean absolute error of 19.4 feet. Most of the increase in error can be attributed to the failure to simulate in the generalized model a reported zone of small hydraulic conductivity southeast of Albuquerque. The presence of this zone was overlooked during the brief (1 week) literature search allowed the student hydrologist. The failure to simulate this zone resulted in the largest local departures between simulated and measured or reported heads.

The second generalized model was of the Socorro and La Jencia Basins. This model also reproduced measured water levels and water-level changes; however, there were far fewer data for comparison than in the Albuquerque-Belen Basin and there were no calibrated models to be used for comparison. This model was unique in simulating two basins connected by a thin ground-water flow path through faulted playa deposits. Surface-water drainage in La Jencia Basin is intermittent and northward to the Rio Salado, whereas the Socorro Basin is one of the narrowest of the basins containing the through-flowing Rio Grande. Mountain-front recharge was estimated by the technique documented in Hearne and Dewey (1988). The greatest transient stress simulated in this model was the routing of the flow of the Rio Grande into a conveyance channel that is essentially one large, straight drain.

The top layer of both generalized models exceeded the guideline thicknesses, which caused a cascade of departures from desirable values for simulated properties (but still marginally within the guidelines): a small simulated hydraulic conductivity for the flood-plain alluvium, a low ratio of horizontal to vertical conductivity, and a large simulated thickness of aquifer. These problems were recognized and the need to minimize the thickness of the top layer was emphasized in the guidelines.

Both generalized models successfully fulfilled the objectives of providing a reasonably accurate representation of both the function and the response of the basin-fill aquifer systems. The models are adequate for testing hypotheses and the subsequent designing of effective data-collection programs aimed at improving and verifying the predictive capability of either these models or their descendants. Points of greatest significance regarding the generalized models are: (1) All of the simulated aquifer properties are within the narrow range established in the guidelines; and (2) the generalized models may be rapidly assembled yet retain an accuracy that is much greater than would be expected from the small expended effort.

SUMMARY

This report describes 14 documented U.S. Geological Survey ground-water-flow models of aquifer systems in seven of the basins in the Southwest Alluvial Basins region of Colorado, New Mexico, and Texas. The models have diverse approaches to the problem of simulating the basin-fill aquifer systems. Some of the approaches are demonstrably better than others; from these preferred approaches, certain attributes common to most models can be found.

On the basis of attributes that are common to most of the ground-water-flow models, a set of guidelines was developed that enables the rapid construction of reasonably accurate generalized ground-water-flow models of specific basins. These guidelines address the type and significance of boundaries, the expected ranges in values for aquifer properties, and the hydrogeologic framework of the simulated representation of the aquifer system. The feasibility of this approach to modeling was tested using two models of three basins in the region. The first basin had been modeled previously and functioned as a benchmark. The second model was of two previously unmodeled basins hydraulically connected by a thin zone of saturation. Both generalized models met the objectives of being adequate representations of both the function and the response of the basin-fill aquifer systems.

The guidelines that were developed may be used to construct generalized models of specific basins in the region. The most appropriate uses of these generalized models are to aid in the design of a data-collection program customized to the needs of a specific study area, to make an initial evaluation of a specific problem, or to test hypotheses regarding the hydrologic responses in a basin. A generalized flow model needs to be viewed as a preliminary effort that will be superseded.

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