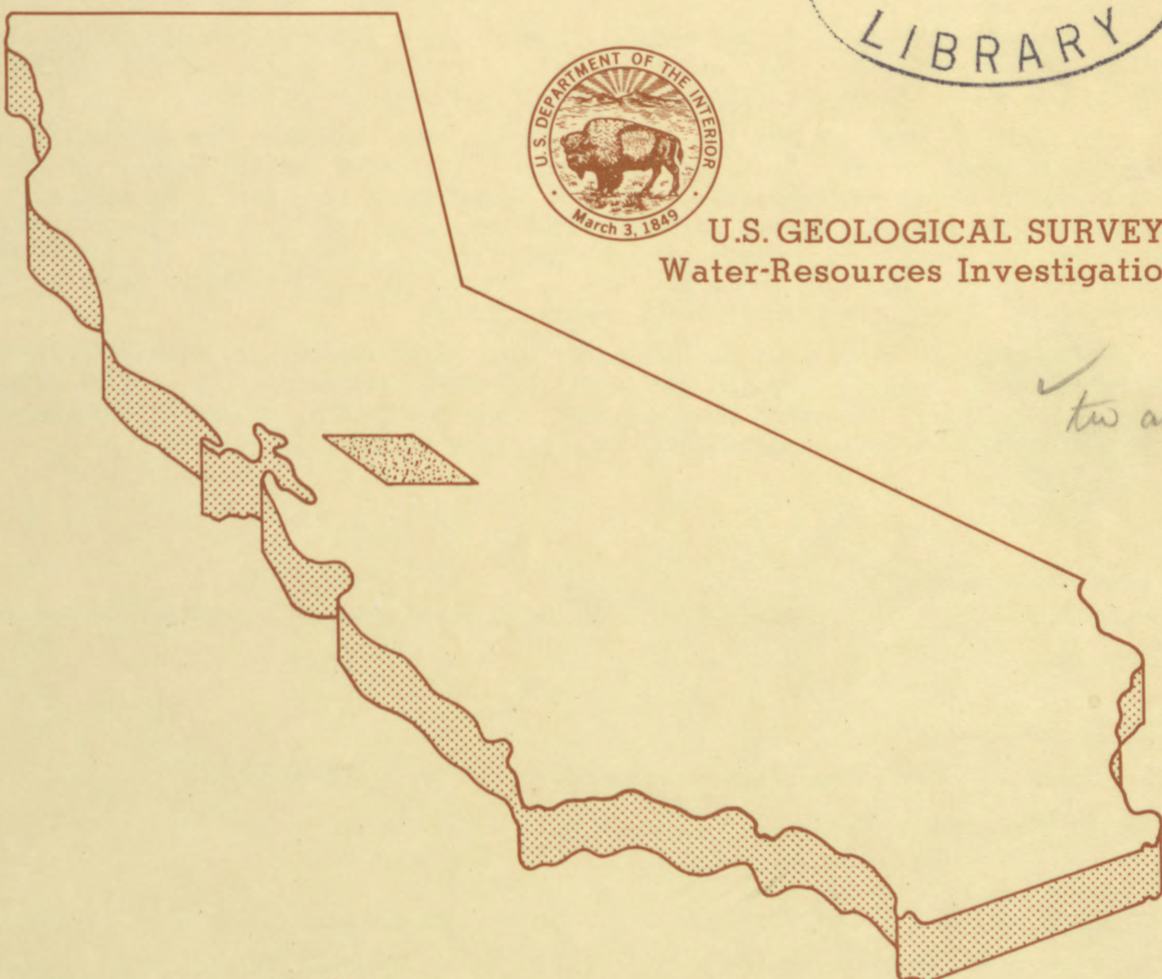
DIGITAL MODEL OF THE UNCONSOLIDATED AQUIFER SYSTEM IN THE MODESTO AREA, STANISLAUS AND SAN JOAQUIN COUNTIES, CALIFORNIA



U.S. GEOLOGICAL SURVEY
Water-Resources Investigations 81-12

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A digital mathematical model of the unconsolidated alluvial aquifer system in the Modesto area has been developed which can be used to determine the effects of increased pumping and water use on future water levels in the aquifer system.

The model is divided into two layers or units that represent the unconsolidated alluvial deposits in the area. The lower unit in the model, called unit 1, represents an aquifer that is both confined and unconfined depending on location. The aquifer is confined in the western part of the study area by a clay bed 20 to 100 feet thick. Elsewhere, the aquifer is unconfined. The upper unit, called unit 2, represents an unconfined aquifer and lies above the clay bed or its extension. Where the clay bed is absent, units 1 and 2 are considered a single aquifer.

The model, as calibrated, can evaluate, with reasonable accuracy the effects on water levels of changing stresses and stress patterns only within the area of primary interest for unit 2. In other areas of unit 2 and for unit 1, predicted changes should be looked on as, at best, representing only general trends.

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July 1981

UNITED STATES DEPARTMENT OF THE INTERIOR

JAMES G. WATT, Secretary

U.S. GEOLOGICAL SURVEY

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CONVERSION FACTORS

The inch-pound system of units is used in this report. For readers who prefer to use metric units, the conversion factors for the terms used in this report are listed below.

Multiply	By	To obtain
acre-ft (acre-foot)	0.001233	hm^3 (cubic hectometer)
ft (foot)	0.3048	m (meter)
ft/d (foot per day)	0.3048	m/d (meter per day)
ft^2/d (foot squared per day)	0.09290	m^2/d (meter squared per day)
ft^3/s (cubic foot per second)	0.02832	m^3/s (cubic meter per second)
in/yr (inch per year)	2.54	cm/yr (centimeter per year)
mi (mile, statute)	1.609	km (kilometer)
mi^2 (square mile)	2.590	km^2 (square kilometer)

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called mean sea level.

DIGITAL MODEL OF THE UNCONSOLIDATED AQUIFER SYSTEM IN THE MODESTO AREA, STANISLAUS AND SAN JOAQUIN COUNTIES, CALIFORNIA

By Clark J. Londquist

ABSTRACT

The city of Modesto is undergoing a period of rapid population growth and is faced with the problem of managing the water resources of the area in order to best meet future needs. A digital mathematical model of the unconsolidated alluvial aquifer system in the Modesto area has been developed to aid the city in meeting this problem. This model can be used to estimate the effects of increased pumping and water use on future water levels in the aquifer system.

The model is divided into two layers or units that represent the unconsolidated alluvial deposits in the area. The lower unit in the model, called unit 1, represents an aquifer that is both confined and unconfined depending on location. The aquifer is confined in the western part of the study area by a clay bed 20 to 100 feet thick. Elsewhere, the aquifer is unconfined. The upper unit, called unit 2, represents an unconfined aquifer and lies above the clay bed or its extension. Where the clay bed is absent, units 1 and 2 are considered a single aquifer.

Prior to the 1952 pumping season, the ground-water system in the area was approximately in a steady-state. Since 1952, ground-water pumpage has been increasing, water levels have been changing, and the state of the system has become transient. Unit 2 of the model was calibrated to the nearly steady-state conditions existing in the spring of 1952 and then calibrated to transient-state conditions for the 26-year period from spring 1952 to spring 1978. Computed water levels, within the area of primary interest for unit 2, had a root-mean-square deviation from the measured water levels of 2.8 feet with a maximum deviation of 9 feet. Because few data are available for unit 1, no attempt was made to calibrate this part of the model.

The model, as calibrated, can evaluate, with reasonable accuracy the effects on water levels of changing stresses and stress patterns only within the area of primary interest for unit 2. In other areas of unit 2 and for unit 1, predicted changes should be looked on as, at best, representing only general trends.

INTRODUCTION

The city of Modesto is undergoing a period of rapid population growth and is faced with the problem of managing the available water resources of the area in order to best meet future needs. In 1971, the city entered into a cooperative agreement with the U.S. Geological Survey to make a water-resources investigation of the city and the surrounding area. The purpose of this investigation was to provide information on the ground-water system and to construct a mathematical model that could be used as a predictive tool for testing alternative ground-water management plans. The primary purpose of this model is to simulate ground-water flow beneath the projected future service area of the city of Modesto and the immediate surrounding area. For purposes of this report, this area has been designated as the area of primary interest (fig. 1).

This is the fourth report prepared by the Geological Survey under this cooperative agreement. The first two reports (Page, 1972; Page and others, 1974) described the hydrology of the area and the hydrologic data that had been collected. The third report (Page, 1977) described a single-layer finite-difference digital model of the unconfined aquifer system in the area. That model could not be calibrated with the existing data, but it did indicate areas where additional data were necessary in order to successfully construct and calibrate a digital model of the ground-water-flow system. The study continued with the objectives of gathering the additional necessary data and then constructing a digital model of the Modesto area. The resulting model, described in this report, is a quasi-three-dimensional, finite-difference digital model that simulates movement of water in both the confined and unconfined parts of the unconsolidated aquifer system underlying the Modesto area.

The author wishes to thank the city of Modesto personnel who assisted in the collection and assimilation of the large quantity of data gathered during this investigation. The author also acknowledges the cooperation and assistance provided by the local water agencies and electric power companies who furnished the data necessary for estimating the ground-water pumpage and surface-water distribution.

GEOHYDROLOGIC SETTING

The model covers an area of about 600 mi^2 in the northeastern part of the San Joaquin Valley. The primary surface drainage in this area is by means of the San Joaquin River and two of its large tributaries, the Tuolumne and Stanislaus Rivers (fig. 1). In addition to these river systems, there are numerous smaller streams, canals, and sloughs that carry water throughout the area. The average annual surface-water discharge from the study area is 4,326 ft^3/s (U.S. Geological Survey, 1979, p. 333).

Precipitation in the area averages about 12 in/yr and occurs mostly from October through May (Page, 1972, p. 4).

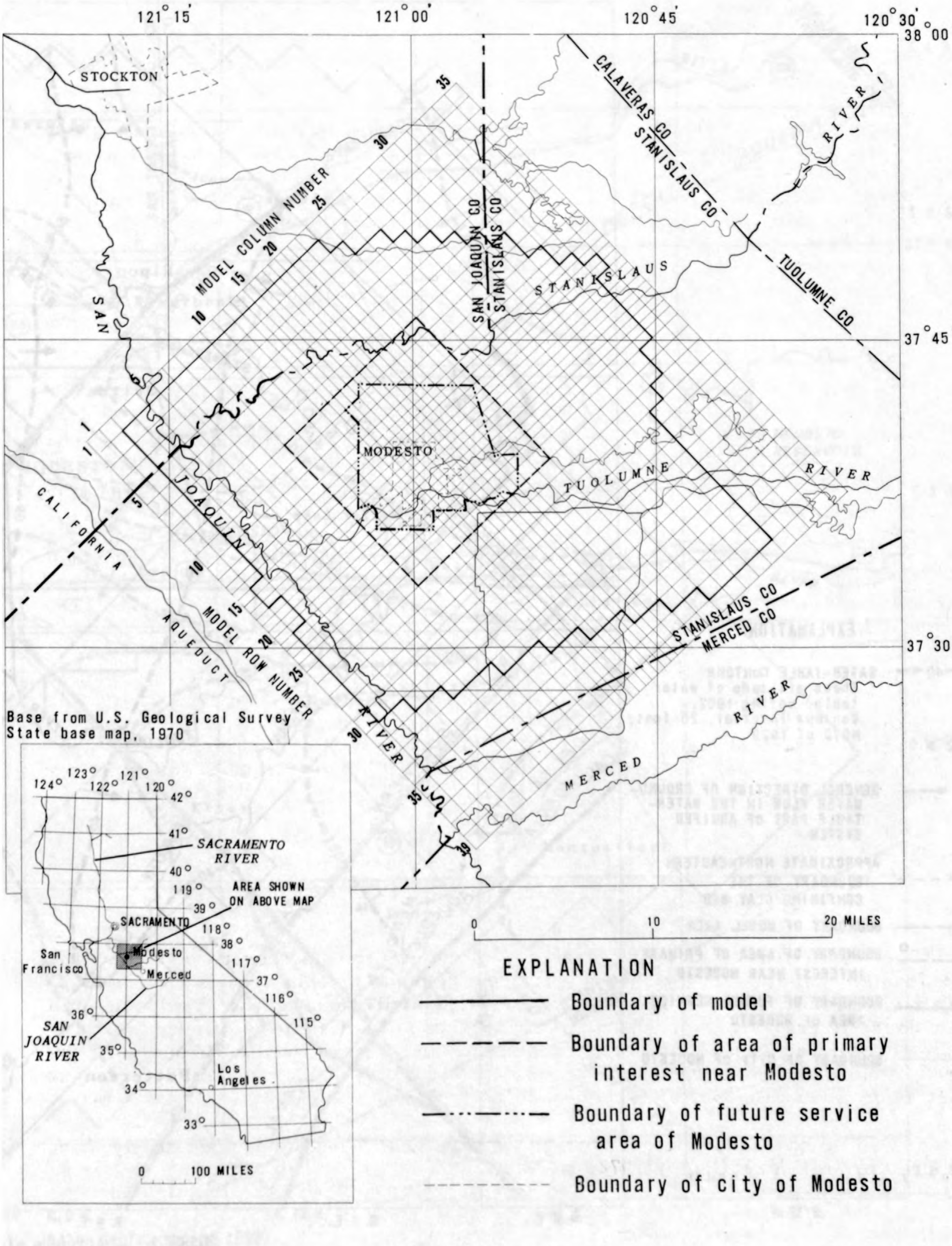


FIGURE 1. — Location of study area and model grid network.

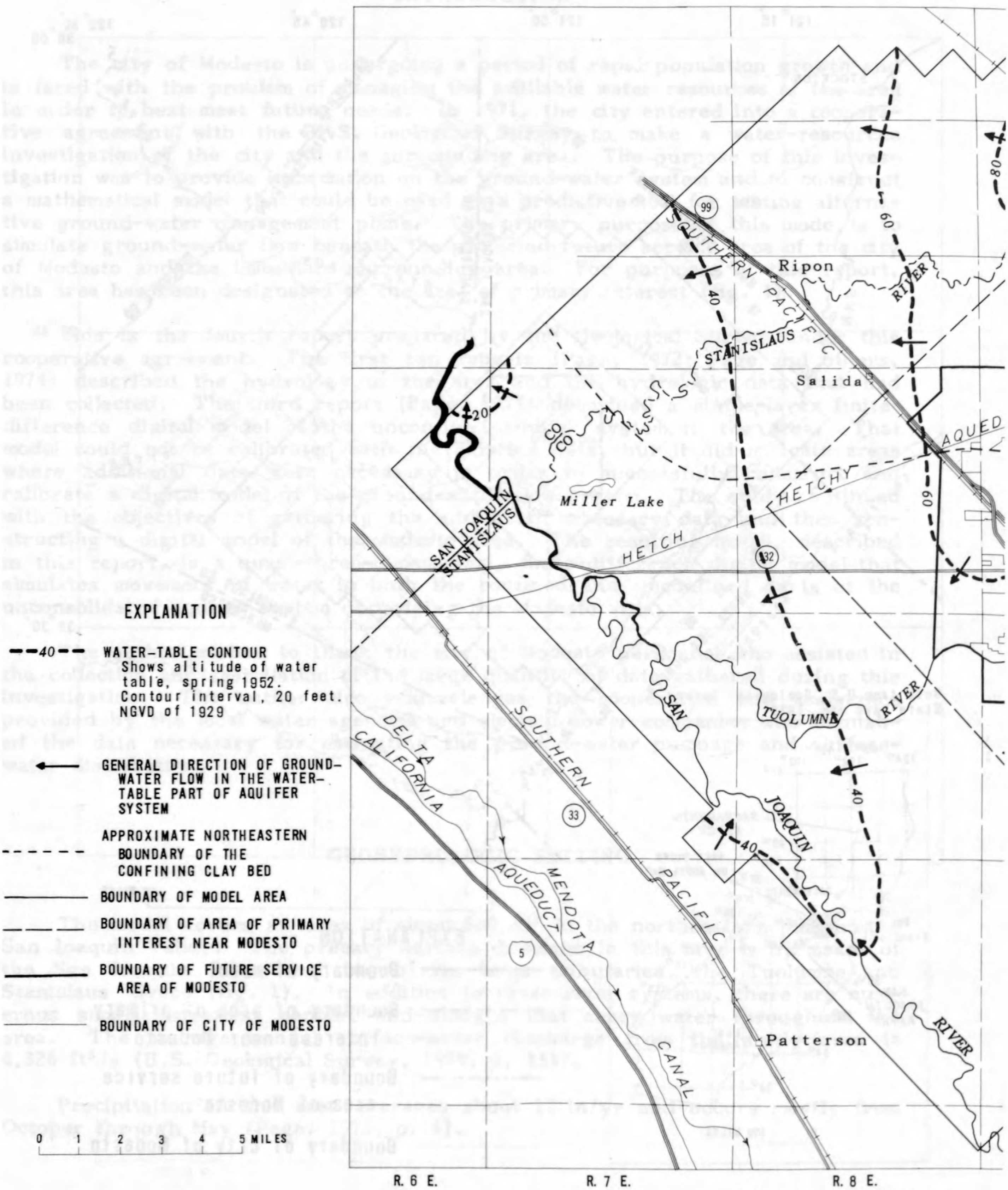
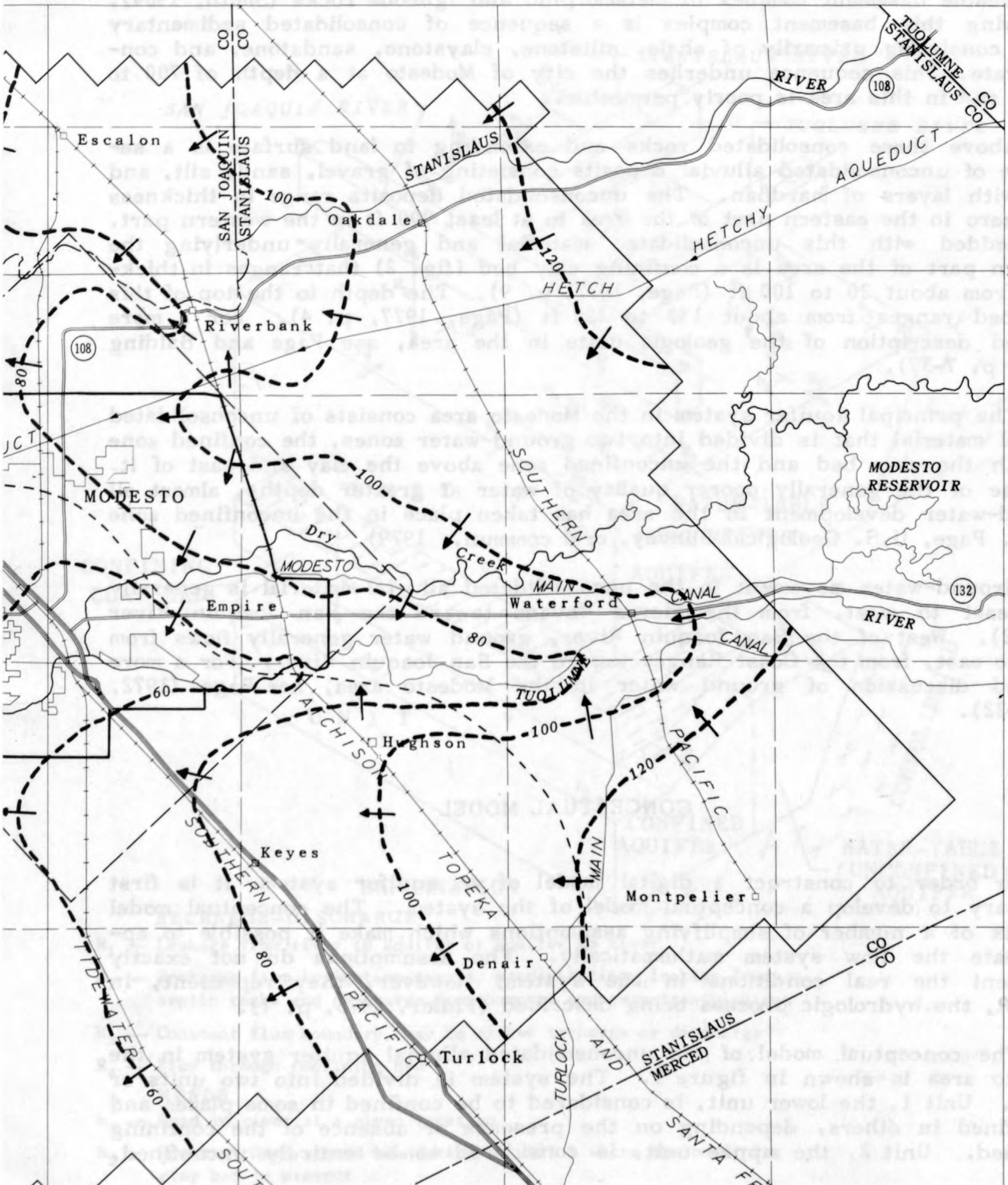


FIGURE 2. — Altitude of



water table, spring 1952.

The Modesto area is underlain at a depth of about 11,000 ft by a virtually impermeable basement complex of metamorphic and igneous rocks (Smith, 1964). Overlying this basement complex is a sequence of consolidated sedimentary rocks consisting primarily of shale, siltstone, claystone, sandstone, and conglomerate. This sequence underlies the city of Modesto at a depth of 700 to 800 ft and in this area is poorly permeable.

Above these consolidated rocks and extending to land surface is a sequence of unconsolidated alluvial deposits consisting of gravel, sand, silt, and clay with layers of hardpan. The unconsolidated deposits range in thickness from zero in the eastern part of the area to at least 900 ft in the western part. Interbedded with this unconsolidated material and generally underlying the western part of the area is a confining clay bed (fig. 2) that ranges in thickness from about 20 to 100 ft (Page, 1977, p. 9). The depth to the top of this clay bed ranges from about 130 to 220 ft (Page, 1977, p. 4). For a more detailed description of the geologic units in the area, see Page and Balding (1973, p. 7-37).

The principal aquifer system in the Modesto area consists of unconsolidated alluvial material that is divided into two ground-water zones, the confined zone beneath the clay bed and the unconfined zone above the clay and east of it. Because of the generally poorer quality of water at greater depths, almost all ground-water development in the area has taken place in the unconfined zone (R. W. Page, U.S. Geological Survey, oral commun., 1979).

Ground-water movement in the unconsolidated alluvial material is generally from east to west, from the Sierra Nevada toward the San Joaquin River (fig. 2). West of the San Joaquin River, ground water generally flows from west to east, from the Coast Ranges toward the San Joaquin River. For a more detailed discussion of ground water in the Modesto area, see Page (1972, p. 12-22).

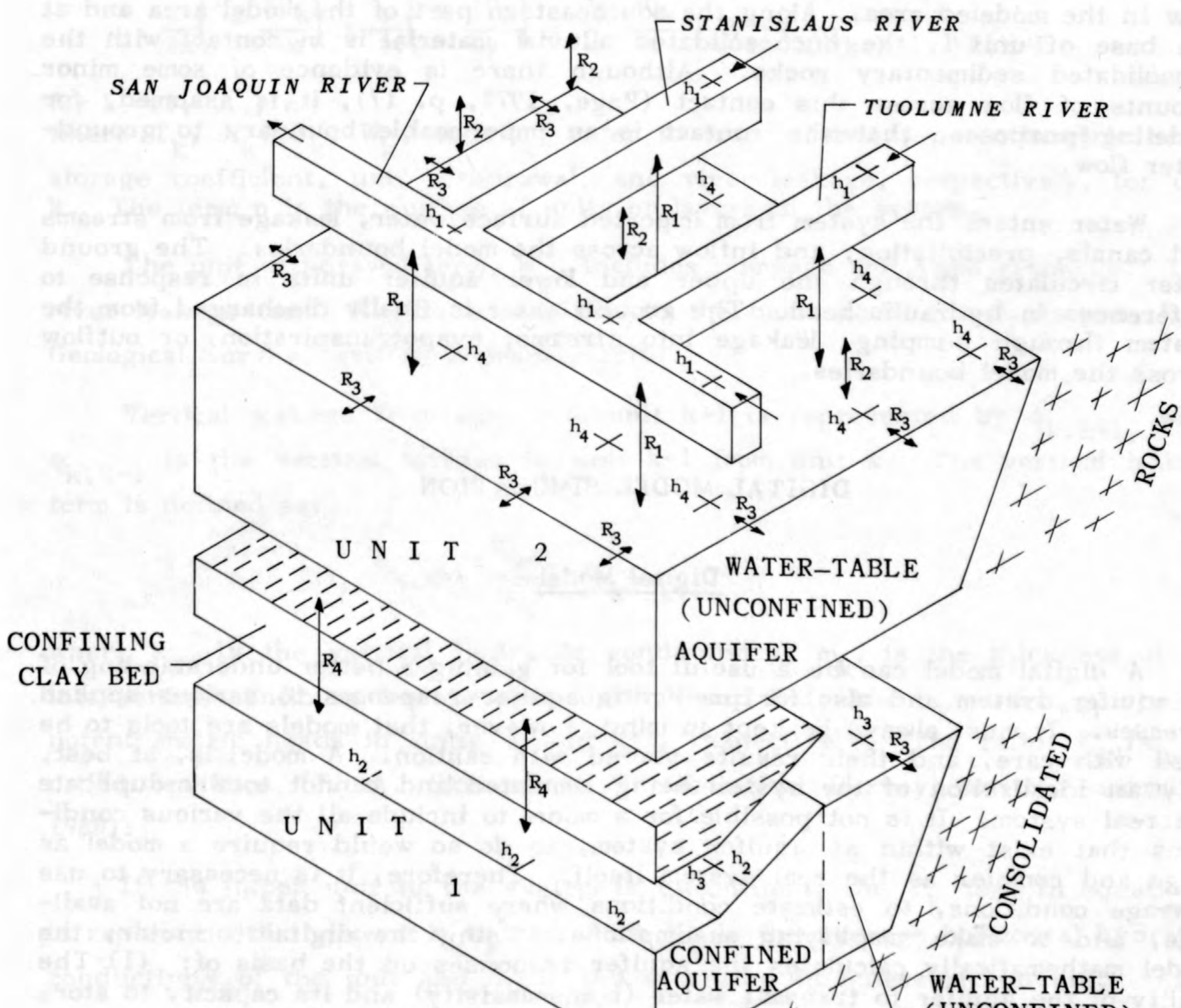
CONCEPTUAL MODEL

In order to construct a digital model of an aquifer system, it is first necessary to develop a conceptual model of the system. The conceptual model consists of a number of simplifying assumptions which make it possible to approximate the flow system mathematically. The assumptions do not exactly represent the real conditions in the system; however, they represent, in concept, the hydrologic process being described (Fidler, 1975, p. 7).

The conceptual model of the unconsolidated alluvial aquifer system in the Modesto area is shown in figure 3. The system is divided into two units or layers. Unit 1, the lower unit, is considered to be confined in some places and unconfined in others, depending on the presence or absence of the confining clay bed. Unit 2, the upper unit, is considered to be entirely unconfined.

The interbedded clay layer separates unit 1 from unit 2 and acts as a confining bed where it is present. This confining bed is assumed to permit flow in only the vertical direction and to have no storage component. For the part of unit 1 that is not overlain by the clay bed, it is assumed that flow is not restricted between units 1 and 2, and that this part of unit 1 is unconfined.

CONCEPTUAL MODEL



EXPLANATION

RECHARGE-DISCHARGE

R_1 — Leakage from river to aquifer or aquifer to river

R_2 — Recharge from irrigation return, precipitation, leakage from septic tanks and discharge from pumpage and evapotranspiration

R_3 — Constant flux boundary, may be either recharge or discharge

R_4 — Flow through confining bed

HEADS

h_1 — Head in river at a given location

h_2 — Constant head around boundaries of unit 1 where the confining clay bed is present

h_3 — Head generated by model at a given location in unit 1

h_4 — Head generated by model at a given location in unit 2

FIGURE 3. — Conceptual model of the hydrologic system in the Modesto area. (Modified from Page, 1977, fig. 6).

In most areas, the unconsolidated alluvial aquifer system extends well beyond the limits of the model and, except for part of the southeastern boundary of the model, there are no natural hydrologic boundaries to ground-water flow in the modeled area. Along the southeastern part of the model area and at the base of unit 1, the unconsolidated alluvial material is in contact with the consolidated sedimentary rocks. Although there is evidence of some minor amounts of flow across this contact (Page, 1977, p. 17), it is assumed, for modeling purposes, that the contact is an impermeable boundary to ground-water flow.

Water enters the system from imported surface water, leakage from streams and canals, precipitation, and inflow across the model boundaries. The ground water circulates through the upper and lower aquifer units in response to differences in hydraulic head. The ground water is finally discharged from the system through pumping, leakage into streams, evapotranspiration, or outflow across the model boundaries.

DIGITAL MODEL SIMULATION

Digital Model

A digital model can be a useful tool for gaining a better understanding of an aquifer system and also for predicting aquifer responses to various applied stresses. It must always be kept in mind, however, that models are tools to be used with care, and their results viewed with caution. A model is, at best, only an idealization of the system being simulated and cannot totally duplicate the real system. It is not possible for a model to include all the various conditions that exist within an aquifer system, to do so would require a model as large and complex as the real system itself. Therefore, it is necessary to use average conditions, to estimate conditions where sufficient data are not available, and to make simplifying assumptions. Using the digital computer, the model mathematically calculates the aquifer responses on the basis of: (1) The ability of the aquifer to transmit water (transmissivity) and its capacity to store and release water (storage coefficient); (2) the quantity of water entering the aquifer (recharge); and (3) the quantity of water leaving the aquifer (discharge) (Swain, 1978, p. 13).

Mathematical Description

The basic model used in this study is the quasi-three-dimensional version of the finite-difference model described by Trescott (1975). In this model the three-dimensional aquifer system is approximated by layered, horizontal units separated by confining beds. The model solves the two-dimensional ground-water flow equation for each layer and couples the layers by terms representing flow through the intervening confining beds.

The basic equations used to describe the flow within the system for confined units are:

$$\frac{\partial}{\partial x}\left(T_k \frac{\partial h_k}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_k \frac{\partial h_k}{\partial y}\right) = S_k \frac{\partial h_k}{\partial t} + W_k + R_k + q_{k,k+1} - q_{k,k-1} \quad (1)$$

for $k = 1, \ldots, n$

where T_k, h_k, S_k, W_k, and R_k, refer to transmissivity, potentiometric head, storage coefficient, unit withdrawal, and river leakage, respectively, for unit k. The term n is the number of units or layers in the system.

The unit withdrawal term, W_k, includes pumpage and areal recharge. The river leakage term, R_k, is an addition to the basic model (S. P. Larson, U.S. Geological Survey, written commun., 1976).

Vertical leakage from unit k to unit k+1 is represented by $q_{k,k+1}$, while $q_{k,k-1}$ is the vertical leakage to unit k-1 from unit k. The vertical leakage term is defined as:

$$q_{k,k+1} = \frac{K_{k'}}{m_{k'}} (h_k - h_{k+1})$$

where $K_{k'}$ is the vertical hydraulic conductivity, $m_{k'}$ is the thickness of the intervening confining bed between units k and k+1, and h_k and h_{k+1} are the potentiometric heads in units k and k+1, respectively. The ratio $K_{k'}/m_{k'}$ is called leakance (J. A. Skrivan, U.S. Geological Survey, written commun., 1980).

If the upper unit in the system is unconfined, the T_k term in equation 1 is replaced with the term $K_k b_k$, in which K_k is the average horizontal hydraulic conductivity of the unit and b_k is the saturated thickness of the unit. Also, specific yield of the unconfined unit is inserted into equation 1 in place of storage coefficient (Trescott, 1975, p. 9).

The model calculates potentiometric heads or water-table altitudes for each unit. In general, through time, heads depend on lateral flow in each aquifer unit, vertical leakage between units, change in ground-water storage, leakage from rivers, areal recharge, and pumpage. The model also produces a head map and a drawdown map for each unit at the end of the simulation period. Drawdown is computed as the difference in head between the start and end of the simulation period.

The theoretical development of the model, numerical-solution techniques, computer program, and data requirements to run the program are discussed by Trescott (1975).

Grid Network

In order to enter data into a finite-difference model, it is first necessary to establish a rectangular-grid system that includes the entire model area. Each rectangle is called a grid element or cell, and the center point of each of these elements is a node.

This grid system is used to overlay maps showing the areal distribution of the various aquifer parameters. The average value of the parameter within that grid element is entered into the model as the value of that parameter for the entire element. This process is repeated until a value for the specific parameter has been assigned for every element in the model area.

Figures 1 and 4 show the grid layout for the Modesto area model. An identical grid network is used for each layer of the model. The size of the grid elements varies, depending on their location. In the area of primary interest the elements are generally one-half mile on a side. The size of the elements increases toward the boundaries of the model to a maximum of 1 mi on a side. The smaller elements allow for a more detailed approximation of the flow system in the area of primary interest where more intensive data collection took place. The larger elements provide a more general approximation of the flow system but allow the model area to be covered with fewer elements. Using fewer elements helps to reduce computer storage and execution time.

Data Requirements

In order to use the model for approximating the flow system and predicting water-level changes, it is necessary to establish the location and nature of the lateral boundaries of the model area, to estimate the aquifer properties within the model area, and to estimate the rates and distribution of water entering and leaving the model area (Barker, 1979, p. 58). This information is supplied to the model in the form of the following parameters:

1. Specific lateral boundary conditions for each aquifer unit.
2. Transmissivity for confined units, and hydraulic conductivity and saturated thickness for unconfined units.
3. Storage coefficient of each unit.
4. Initial head distribution in each unit.
5. Leakance values between adjacent units.
6. Leakage to or from rivers.
7. Areal recharge to or discharge from upper unit.
8. Quantity and distribution of pumpage from each unit.

The accuracy of the model predictions is directly related to the accuracy of the data defining these required parameters. Data describing the parameters for the part of unit 1 below the confining clay bed are sparse and, therefore, average transmissivity and storage values were used in constructing this part of unit 1. For the part of unit 1 not overlain by the confining clay bed, aquifer parameters were modeled as though they are similar to those of model unit 2.

Lateral Boundary Conditions

Boundary conditions must be specified in order to simulate lateral flow at the edges of the modeled aquifer system. Two types of lateral hydrologic boundaries can be used in this model, constant-head boundaries and constant-flux boundaries. At a constant-head boundary, the water levels in the elements at the boundary do not change with time, but the flux, or amount of water coming into or leaving the element, varies in response to water-level changes in adjacent elements. At a constant-flux boundary, the flux remains constant for the element, but the water levels vary in response to water-level changes in adjacent elements. When the constant flux is set at zero, the boundary is referred to as a no-flow boundary.

Hydrologic boundaries may be simulated where actual boundaries exist or, if the system is so large that actual boundaries do not exist within the limits of the model area, model boundaries must be placed at a sufficient distance from the area of interest that they will have little or no effect on model predictions for the critical areas. The only natural lateral hydrologic boundary in the Modesto model area is the contact between the unconsolidated alluvial material and the less permeable consolidated rocks along the southeastern part of the model area. Although there is undoubtedly some flow across this boundary, it is probably slight and, for modeling purposes, the boundary was considered a no-flow boundary.

Constant-head boundaries were used around unit 1 for all areas where the confining clay bed is present. Constant-flux boundaries were used around the remainder of the area (fig. 4). Values for constant-head and constant-flux boundaries were generated by the model when the system was considered to be under nearly steady-state conditions (see discussion on steady state, p. 26).

For unit 2, constant-flux boundaries were simulated around the entire area. The rate of the flux for each boundary element was generated by the model when the system was considered to be under steady-state conditions. Flux was set at zero around the southeastern part of the area, where the boundary is the contact between the unconsolidated alluvial material and the less permeable consolidated rocks (fig. 4).

To determine the influence of the model boundaries on predicted water levels in the area of primary interest, two model runs were made using different boundary conditions. These runs were for the period spring 1952 to spring 1978. The predicted water levels, in the unconfined layer, for each of these runs were compared on an element-by-element basis to those obtained using the boundary conditions selected for use in the model. For the first run the constant-flux rates used in the model were doubled and for the second run all of the constant-flux boundaries, except those that were no-flow, were changed to constant-head boundaries. The heads used were those that existed when the system was considered to be under steady-state conditions. The predicted head distribution using the doubled flux rates departed from the predicted head distribution using the original flux rates within the area of primary interest by a root-mean-square (rms) deviation of 0.4 ft and the maximum departure was 1.4 ft. The rms deviation is analogous to the standard deviation of statistics (Hoxie, 1977, p. 27) and is a measure of the agreement between the two predicted head distributions. Comparison of the predicted heads within the area of primary interest using the constant-head boundaries with those using the constant-flux rates used in the model showed a rms deviation of 1.6 ft and maximum departure of 4 ft. The actual measured drawdown for the area of primary interest over this same 26-year period averaged about 18 ft per element and the maximum drawdown was 32 ft.

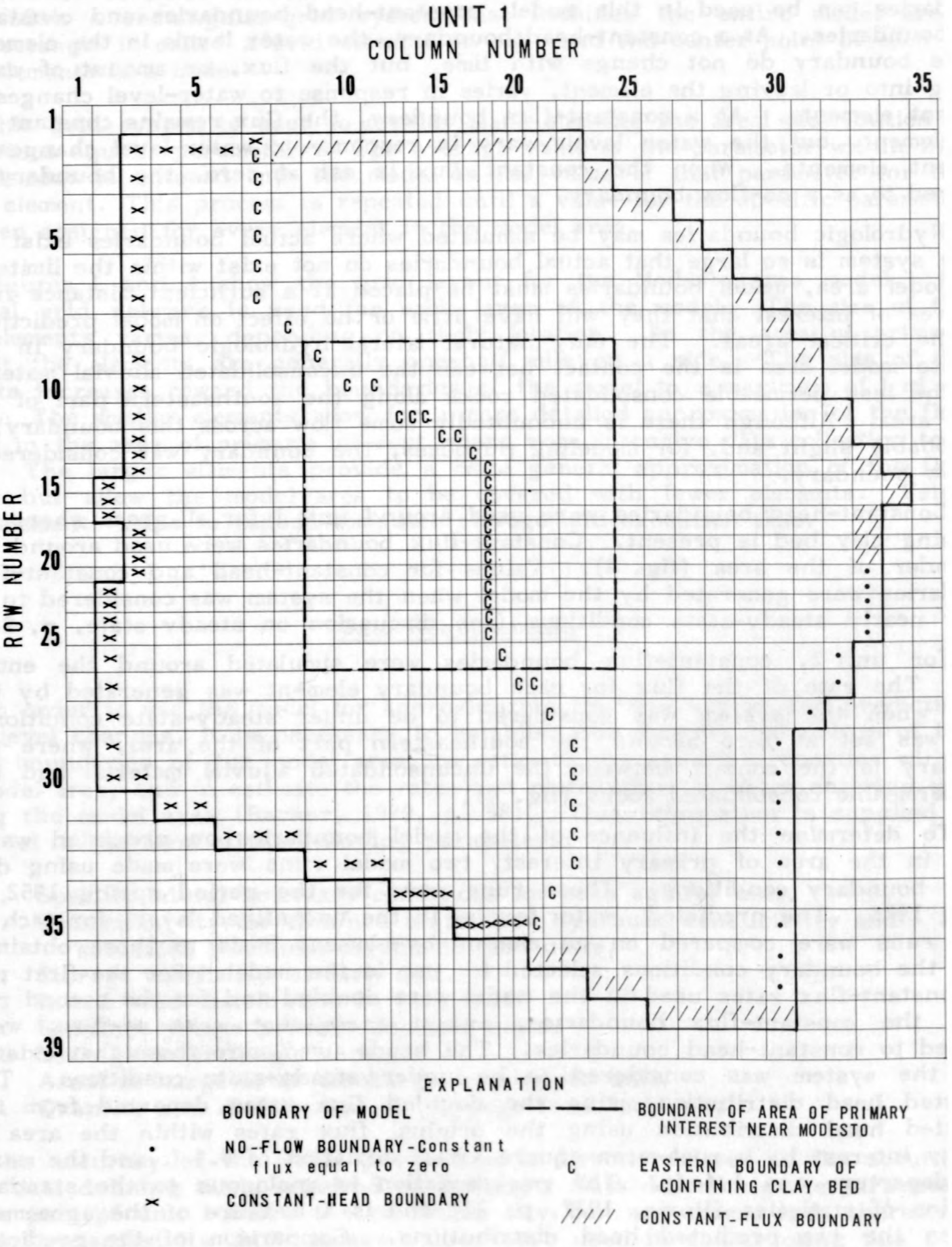


FIGURE 4.—Grid configuration with model boundaries and model location of rivers.

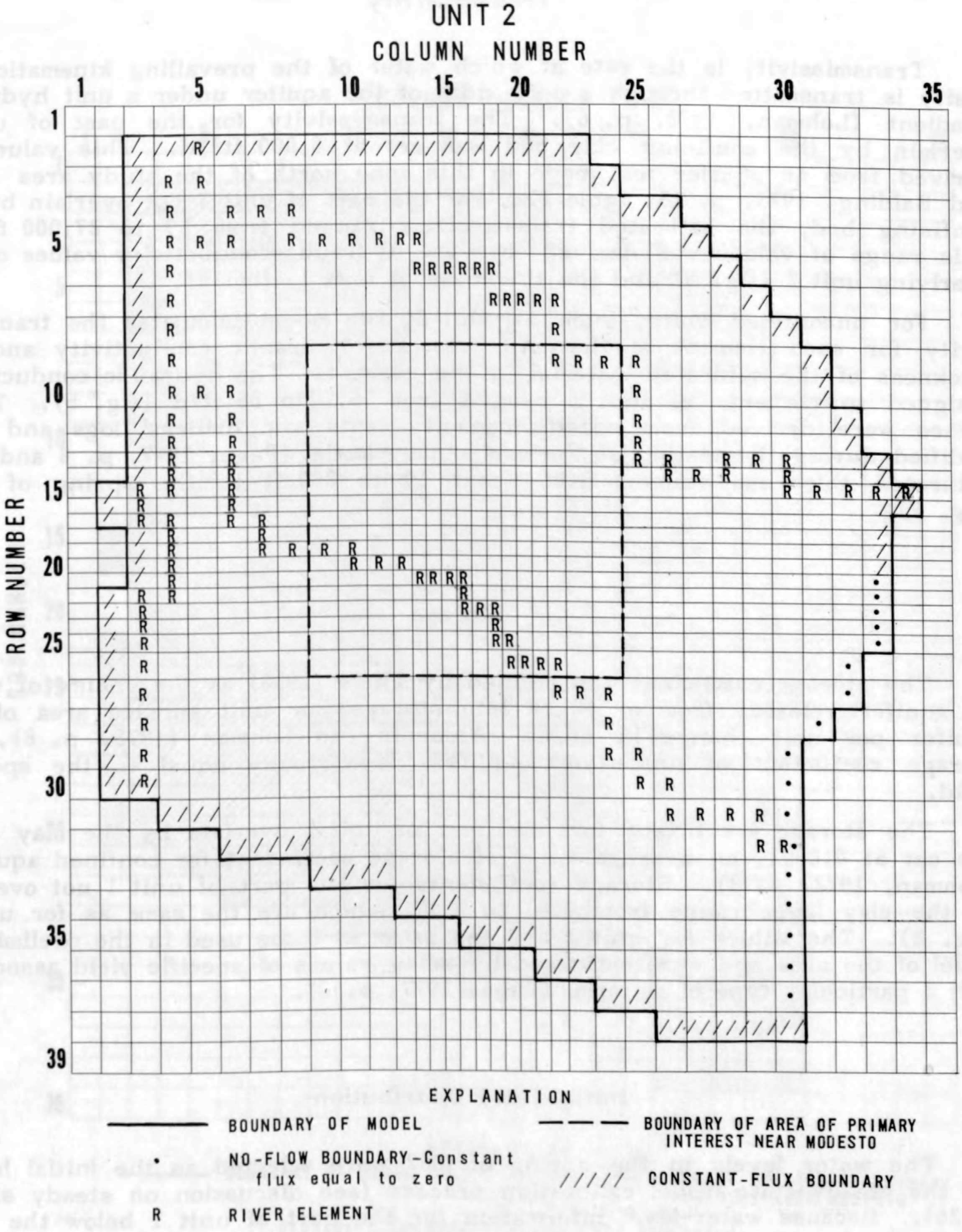


FIGURE 4.— Continued.

These comparisons indicate that the boundaries have been placed far enough from the area of primary interest so as to have only minimal effect on predicted water levels within this area. The constant-flux boundaries selected for use in the model were chosen because they produced a better match between the predicted and measured water levels around the boundary of the model.

Transmissivity

Transmissivity is the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient (Lohman, 1972, p. 6). The transmissivity for the part of unit 1 overlain by the confining clay bed was set at 8,000 ft^2/d. This value was derived from an aquifer test made in this zone north of the study area (Page and Balding, 1973, p. 13, table 1). For the part of unit 1 not overlain by the confining bed, the estimated transmissivity ranged from 69 to 27,000 ft^2/d. This range of values was derived from the hydraulic-conductivity values of the overlying unit 2 (fig. 5) and the thickness of unit 1 (fig. 6).

For unconfined units, such as unit 2, the model calculates the transmissivity for each element of the unit from the hydraulic conductivity and the thickness of the saturated material in the element. The hydraulic conductivity assigned to elements of unit 2 ranged from 6.7 to 86 ft/d (fig. 5). These values were derived from specific-capacity tests and drillers' logs and then modified during the testing of the preliminary model (Page, 1977, p. 8 and 22). Saturated thickness ranged from about 10 to 250 ft in the spring of 1952 (fig. 7).

Storage

The storage coefficient was defined by Theis (1938) as the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. According to Lohman (1972, p. 8), the storage coefficient of unconfined aquifers is virtually equal to the specific yield.

The storage coefficient for the part of unit 1 overlain by the clay layer was set at 0.0001, an average value of storage coefficient for confined aquifers (Lohman, 1972, p. 8). Storage coefficients for the part of unit 1 not overlain by the clay layer range from 0.07 to 0.17, which are the same as for unit 2 (fig. 8). The values for unit 2 are the same as those used in the preliminary model of the area and were estimated by using values of specific yield associated with a particular type of material (Page, 1977, p. 8).

Initial Head Distribution

The water levels in the spring of 1952 were selected as the initial heads for the steady-state model calibration process (see discussion on steady state, p. 26). Because water-level information for the part of unit 1 below the confining clay layer is limited, the initial head distribution for unit 1 (fig. 9) was generated by the model during the steady-state calibration process rather than derived from actual water-level measurements.

The initial head distribution for unit 2 (fig. 2) was derived from water-level measurements taken in the spring of 1952 and from projected water-level contours based on these measurements (Page, 1977, p. 9). East of the confining clay bed, the water-level altitudes are the same for both model units.

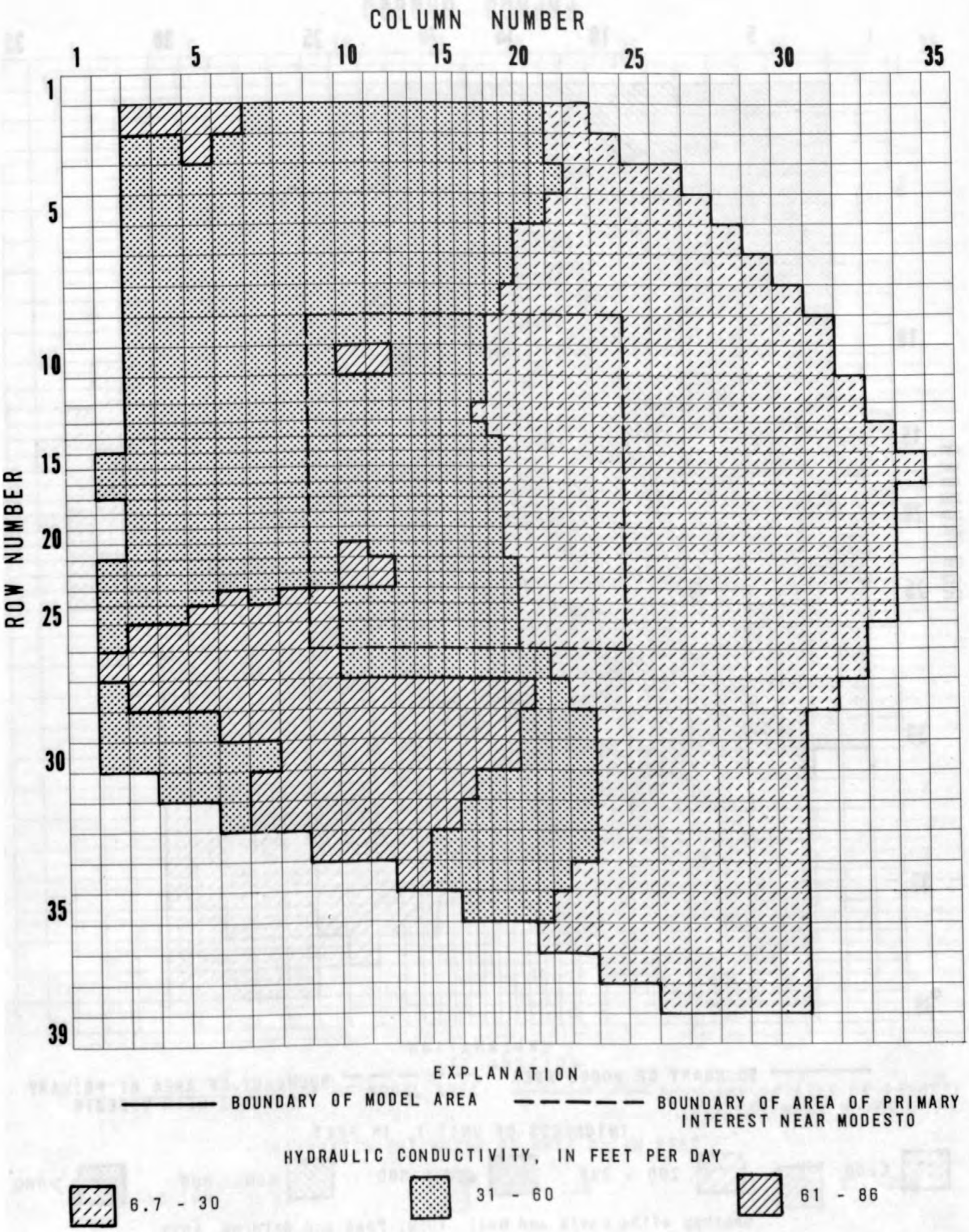


FIGURE 5.—Areal distribution of hydraulic conductivity of unit 2, as simulated in the model.

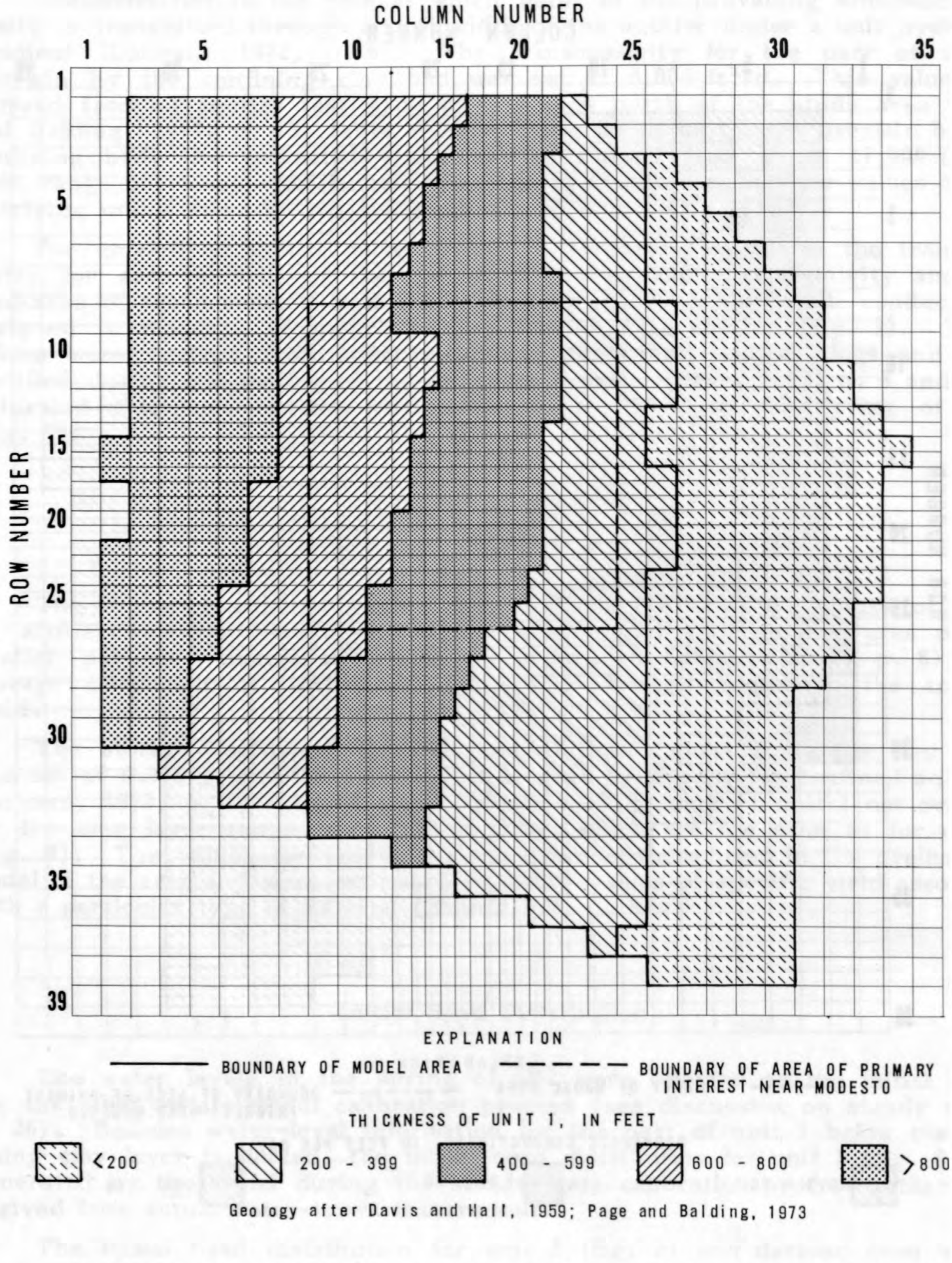


FIGURE 6.—Areal distribution of thickness of unit 1, as simulated in the model.

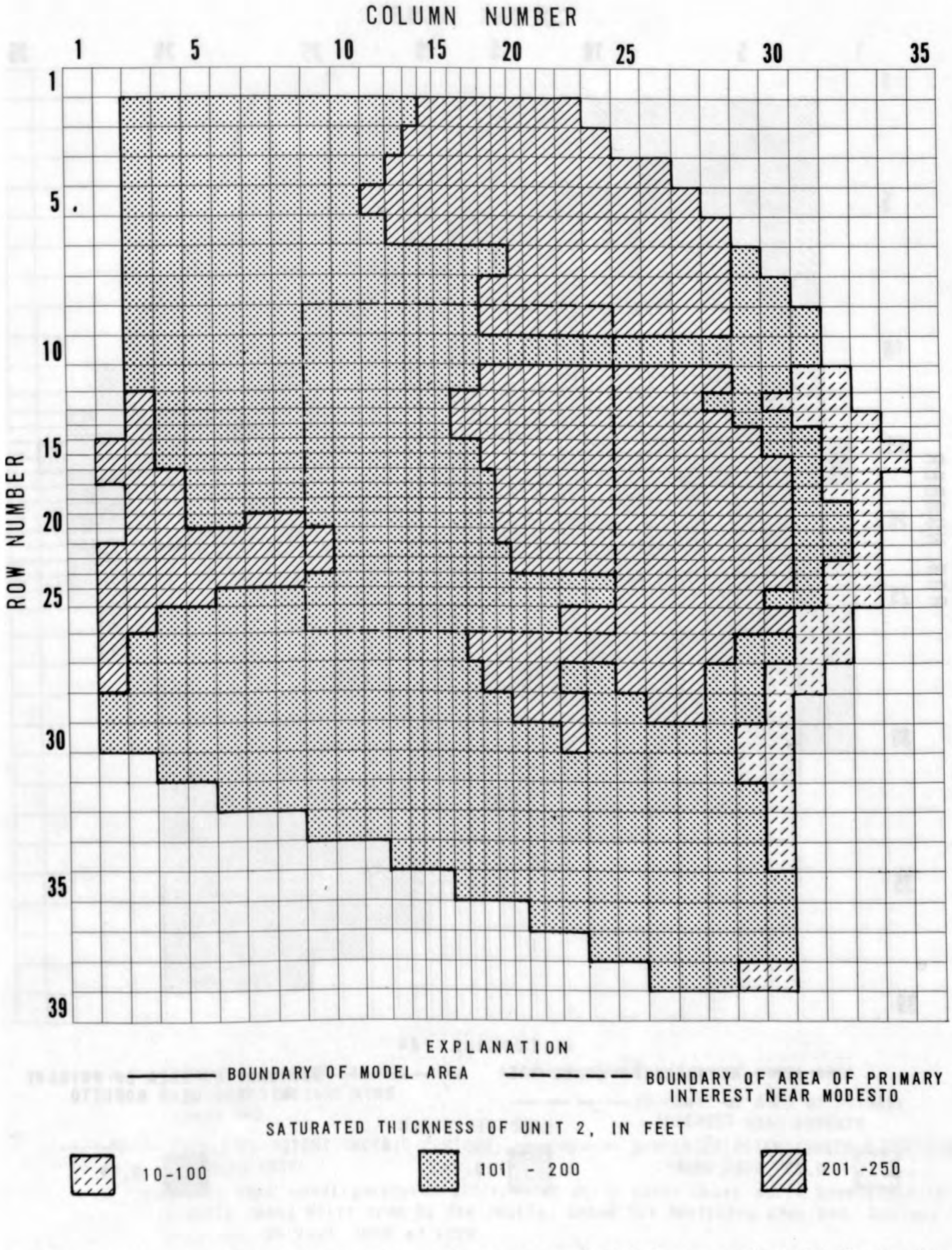


FIGURE 7.—Areal distribution of saturated thickness of unit 2, spring 1952, as simulated in the model.

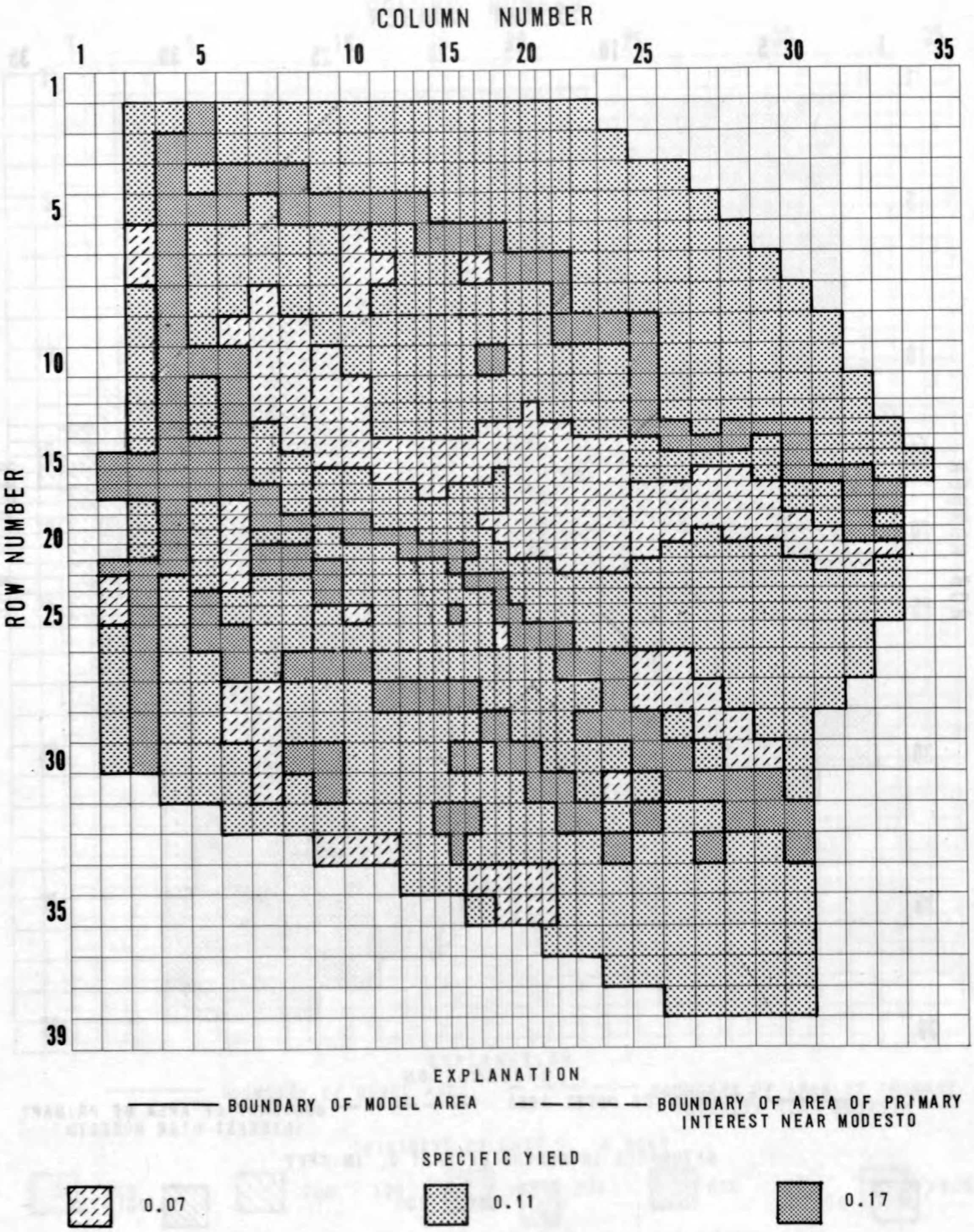


FIGURE 8.—Areal distribution of specific yield of unit 2, as simulated in the model.

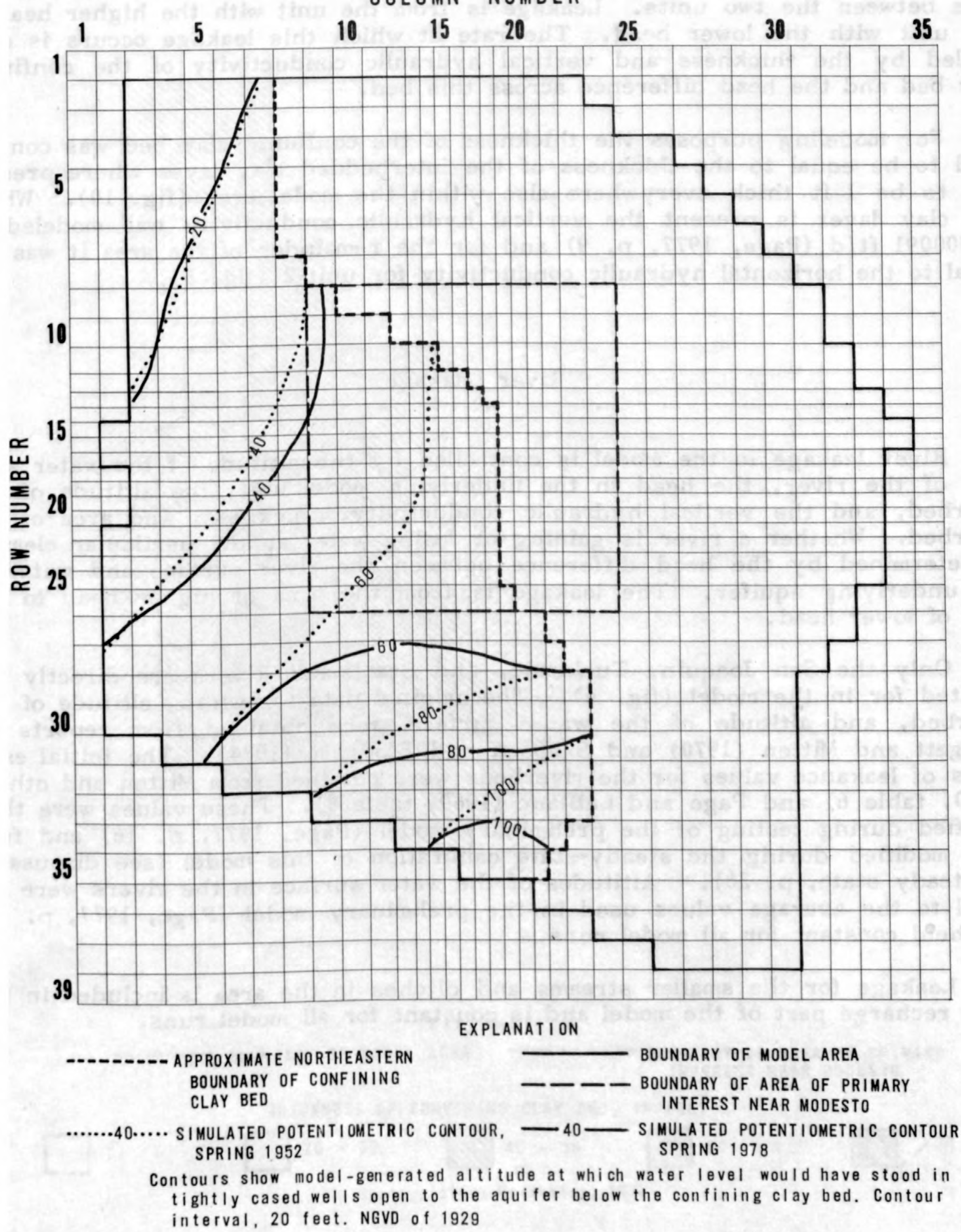


FIGURE 9.—Simulated potentiometric surface below the confining clay bed, spring 1952 and spring 1978.

Leakage Between Units

Leakage from one unit to another occurs whenever there is a head difference between the two units. Leakage is from the unit with the higher head to the unit with the lower head. The rate at which this leakage occurs is controlled by the thickness and vertical hydraulic conductivity of the confining clay bed and the head difference across this bed.

For modeling purposes the thickness of the confining clay bed was considered to be equal to the thickness of the interbedded clay layer where present and to be 1 ft thick everywhere else within the model area (fig. 10). Where the clay layer is present the vertical hydraulic conductivity was modeled as 0.0000091 ft/d (Page, 1977, p. 9) and for the remainder of the area it was set equal to the horizontal hydraulic conductivity for unit 2 (fig. 5).

River Leakage

River leakage in the model is controlled by the altitude of the water surface of the river, the head in the underlying model unit, the altitude of the riverbed, and the vertical hydraulic conductivity, thickness, and area of the riverbed. Whether a river is gaining or losing water at any particular element is determined by the head difference between the river surface and water in the underlying aquifer. The leakage is from the area of higher head to the area of lower head.

Only the San Joaquin, Tuolumne, and Stanislaus Rivers are directly accounted for in the model (fig. 4). The original data for width, altitude of the riverbed, and altitude of the water surface were obtained from reports by Blodgett and Mitten (1970) and Simpson and Blodgett (1974). The initial estimates of leakance values for the riverbeds were obtained from Mitten and others (1970, table 6) and Page and LeBlanc (1969, table 8). These values were then modified during testing of the preliminary model (Page, 1977, p. 16) and further modified during the steady-state calibration of this model (see discussion on steady state, p. 26). Altitudes of the water surface in the rivers were set equal to the average values used in the preliminary model (Page, 1977, p. 16) and held constant for all model runs.

Leakage for the smaller streams and ditches in the area is included in the areal recharge part of the model and is constant for all model runs.

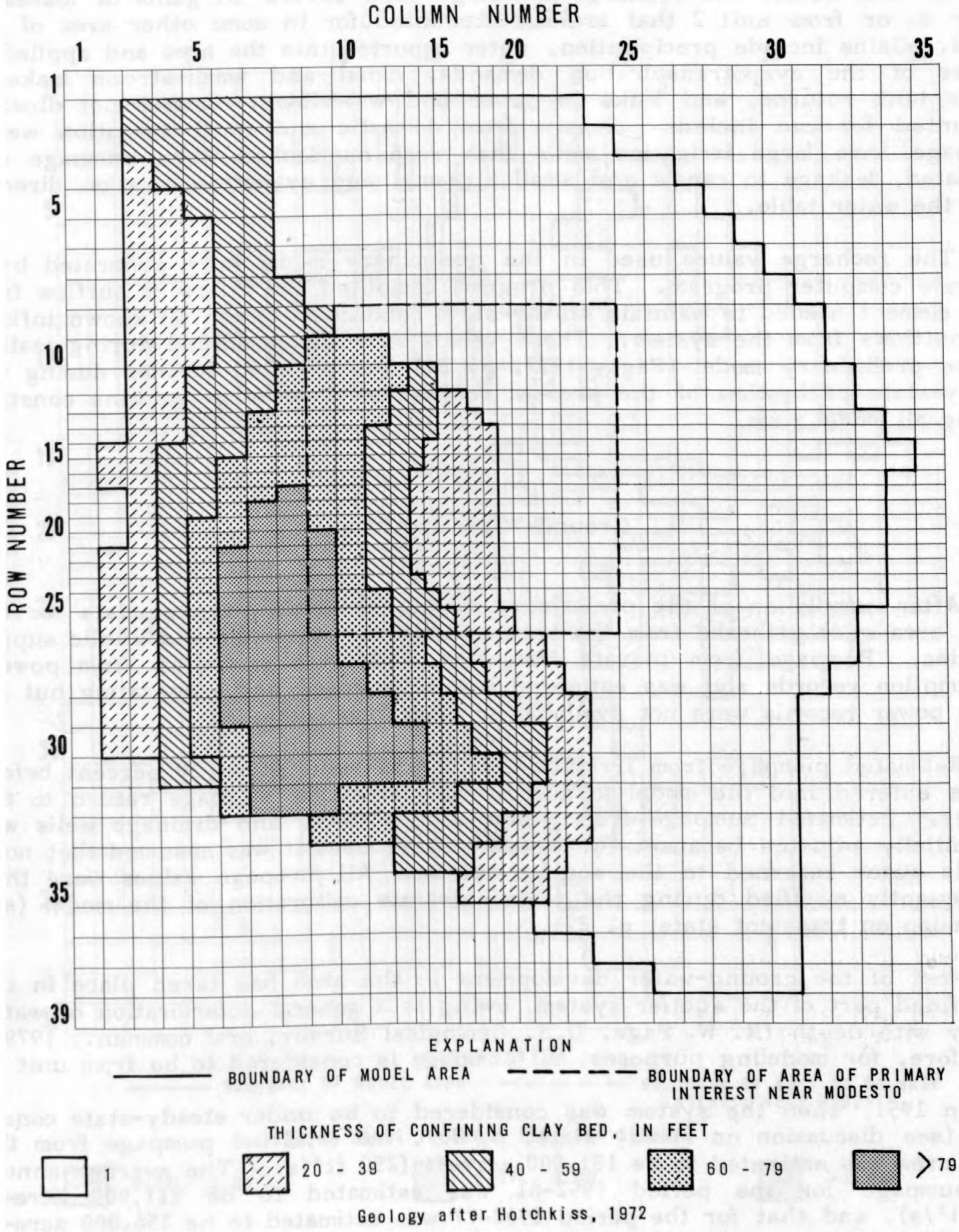


FIGURE 10.—Areal distribution of thickness of confining clay bed between units 1 and 2, as simulated in the model.

Areal Recharge-Discharge

In this model, the recharge-discharge term covers all gains or losses of water to or from unit 2 that are not accounted for in some other area of the model. Gains include precipitation, water imported into the area and applied in excess of the evapotranspiration demands, canal and small-stream leakage, septic tank returns, and leaks in sewer and waterlines. Losses not directly accounted for can include pumpage from domestic and small irrigation wells, pumpage from large irrigation wells that were overlooked when pumpage was estimated, leakage to canals and small streams, and evapotranspiration directly from the water table.

The recharge values used in the preliminary model were generated by a separate computer program. This program computed the inflow or outflow from each element needed to maintain steady-state conditions, given all known inflows and outflows from the system. These values were then modified during testing of the preliminary model (Page, 1977, p. 21) and modified further during the steady-state calibration of the present model. Areal recharge remains constant during all model runs.

Ground-Water Pumping

After completion of the preliminary model, available pumpage data for the model area were gathered from the various irrigation districts and public supply agencies. Pumpage from private irrigation wells was computed from power-consumption records and was estimated for wells known to be operating but for which power records were not available or were incomplete.

Estimated pumpage from irrigation wells was reduced by 30 percent before it was entered into the model to account for irrigation pumpage return to the aquifer. Estimated pumpage from public supply wells and drainage wells was not initially adjusted because, for modeling purposes, it was assumed that none of this water returned to the aquifer system. All pumpage values were then subsequently modified during the transient-state calibration of the model (see discussion on transient state, p. 27).

Most of the ground-water development in the area has taken place in the unconfined part of the aquifer system, owing to a general deterioration of water quality with depth (R. W. Page, U.S. Geological Survey, oral commun., 1979). Therefore, for modeling purposes, all pumpage is considered to be from unit 2.

In 1951, when the system was considered to be under steady-state conditions (see discussion on steady state, p. 26), the total net pumpage from the model area was estimated to be 181,000 acre-ft (250 ft^3/s). The average annual net pumpage for the period 1952-61 was estimated to be 211,000 acre-ft (291 ft^3/s), and that for the period 1962-77 was estimated to be 256,000 acre-ft (353 ft^3/s). The pumping distribution for these periods is shown in figures 11, 12, and 13, respectively.

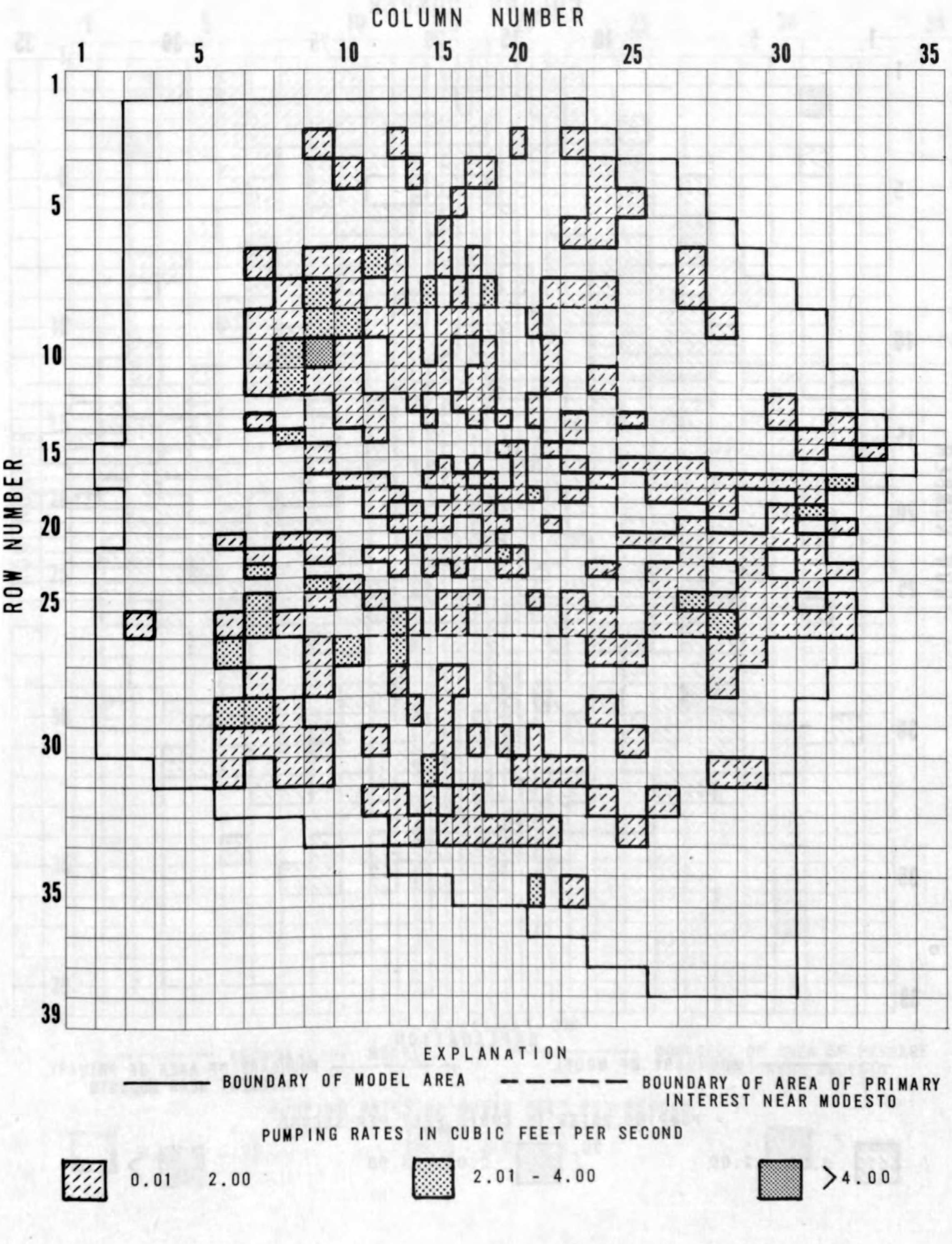


FIGURE 11.—Areal distribution of net pumping rates from model unit 2 for 1951.

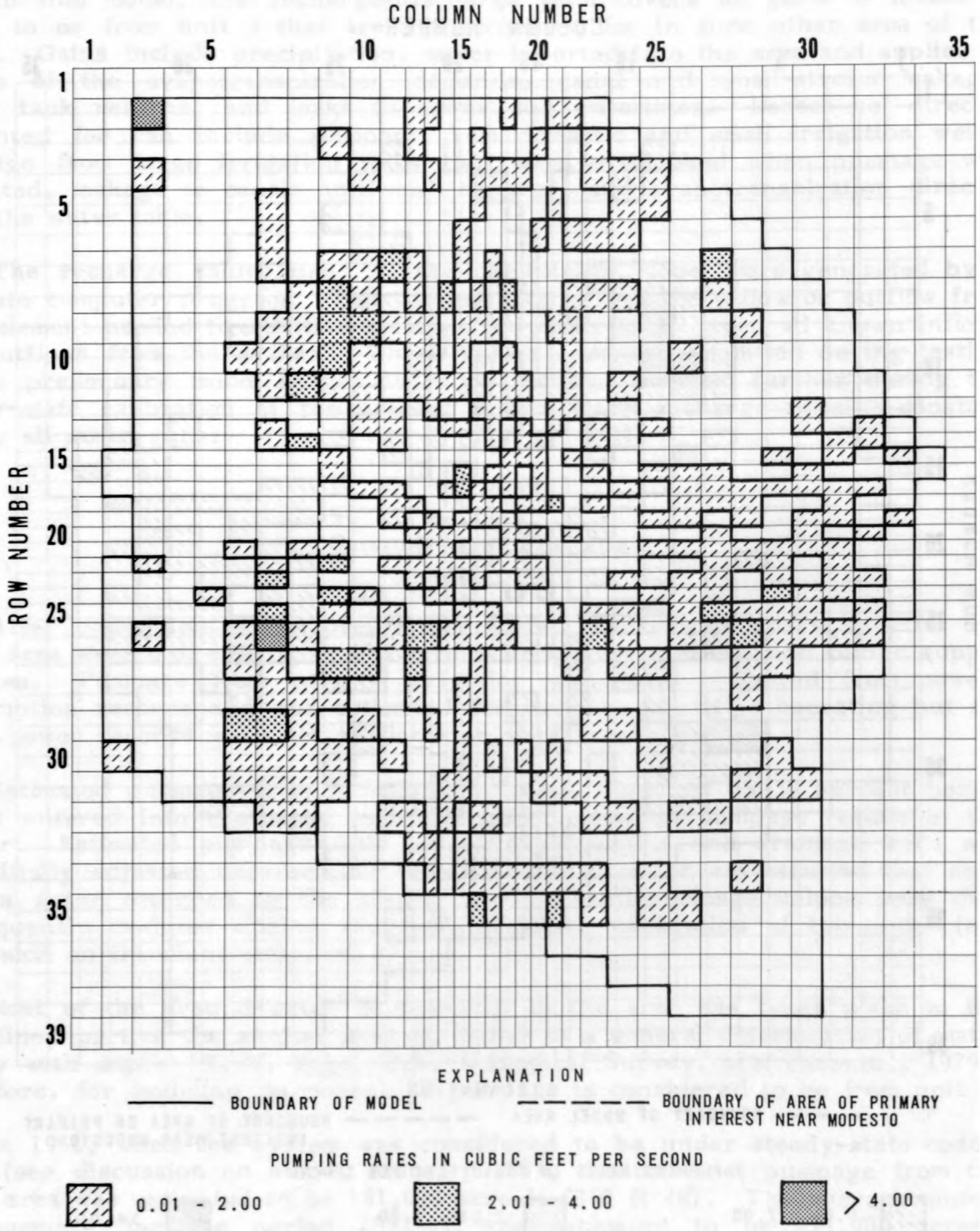


FIGURE 12.—Areal distribution of average net pumping rates from model unit 2 for 1952-61.

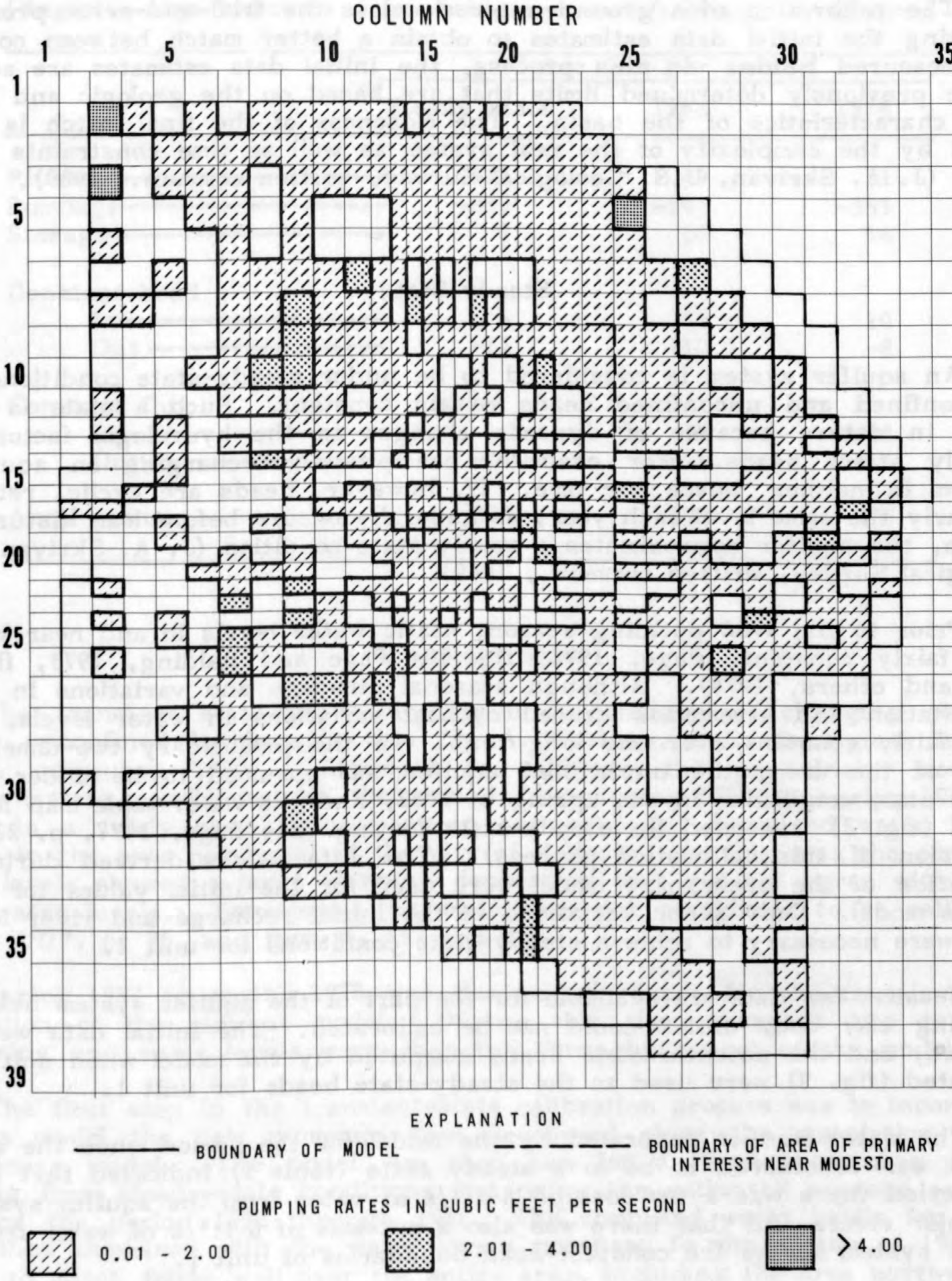


FIGURE 13.—Areal distribution of average net pumping rates from model unit 2 for 1962-77.

Model Calibration

The calibration of a ground-water model is the trial-and-error process of adjusting the initial data estimates to obtain a better match between computed and measured heads. In this process, the initial data estimates are adjusted within previously determined limits that are based on the geologic and hydrologic characteristics of the basin. The closeness of the final match is determined by the complexity of the real system as well as time constraints on the study (J. A. Skrivan, U.S. Geological Survey, written commun., 1980).

Steady State

An aquifer system is considered to be under steady-state conditions when the confined and unconfined heads remain constant. Such a state is rarely found in nature because of dynamic changes in the hydrologic factors that directly affect heads. For example, precipitation recharging an aquifer is uniform in neither space nor time. If, however, heads are cyclic, returning to nearly the same level each year, as generally occurs before man disturbs the system, the aquifer approximates a steady-state condition (J. A. Skrivan, U.S. Geological Survey, written commun., 1980).

Prior to the 1952 pumping season, static water levels in and near Modesto were fairly constant (Page, 1972, fig. 10; Page and Balding, 1973, fig. 16; Page and others, 1974). Although seasonal pumpage and variations in annual precipitation and streamflow caused cyclical variations in water levels, levels were fairly constant over the long term. In the preliminary two-dimensional model of the area, the unconfined aquifer was assumed to be under nearly steady-state conditions in the spring of 1952, and the water-table map for this period (fig. 2) was used for steady-state testing (see Page, 1977, p. 22 for a discussion of this calibration process). The data values derived during the calibration of the preliminary model were used for the initial values for unit 2 in this model. Only minor variations in the initial recharge and river leakage rates were necessary to achieve steady-state conditions for unit 2.

Because few data are available for the part of the aquifer system below the confining clay bed, unit 1 could not be calibrated. The initial data were not adjusted, and the potentiometric heads computed by the model when unit 2 was calibrated (fig. 9) were used as the steady-state heads for unit 1.

The water budget generated by the model for the period when the aquifer system was considered to be in a steady state (table 1) indicated that during this period there was a net loss of 18 ft^3/s of water from the aquifer system to the major rivers and that there was also a net loss of 6 ft^3/s of water from the aquifer system across the constant-head boundaries of unit 1.

TABLE 1. -- Water budget for the steady-state condition, spring 1962, and spring 1978, as generated by the model

[Negative sign indicates water being removed from aquifer system]

	Cubic feet per second		
	Steady state	1962	1978
Rivers	-18	-9	21
Pumpage	-250	-291	-353
Storage	0	30	56
Constant head			
In	5	6	10
Out	-11	-10	-8
Constant flux			
In	14	14	14
Out	-15	-15	-15
Areal recharge	275	275	275

Transient State

The transient state of an aquifer is a condition of changing confined and/or unconfined heads with time (J. A. Skrivan, U.S. Geological Survey, written commun., 1980). Attempts to calibrate the preliminary model to transient-state conditions for the period from spring 1952 to spring 1962 resulted in a fairly close match between computed and measured water levels, except in the area northwest of the city of Modesto. In this area the computed water levels showed a large pumping depression that did not agree with measured water levels. Therefore, the model was not considered to be calibrated (Page, 1977, p. 35, and figs. 5, 15, and 16).

Spring 1952 to spring 1978 was the period chosen for the transient-state calibration of the current model. During this time, pumpage was generally increasing and water levels were changing throughout most of the model area.

The first step in the transient-state calibration process was to incorporate in this model the new pumpage data developed since the completion of the preliminary model. The model was then run for a 10-year pumping period, starting from steady-state conditions and using the estimated average pumping rate for the period 1952 through 1961. The computed water levels for unit 2 were then compared with the water levels measured in the spring of 1962 and found to match fairly well over the entire area, including the area northwest of the city of Modesto. Several more runs were made, each time varying the pumping rates on an element-by-element basis until an acceptable match was achieved between measured and computed water levels (fig. 14). In the area of primary interest for unit 2, the rms deviation between measured and computed water levels was 2.4 ft and the maximum deviation was 8 ft. The rms of the measured drawdown for the area of primary interest over this time period was 12.5 ft. During this calibration process the total net pumpage was decreased about 8 percent.

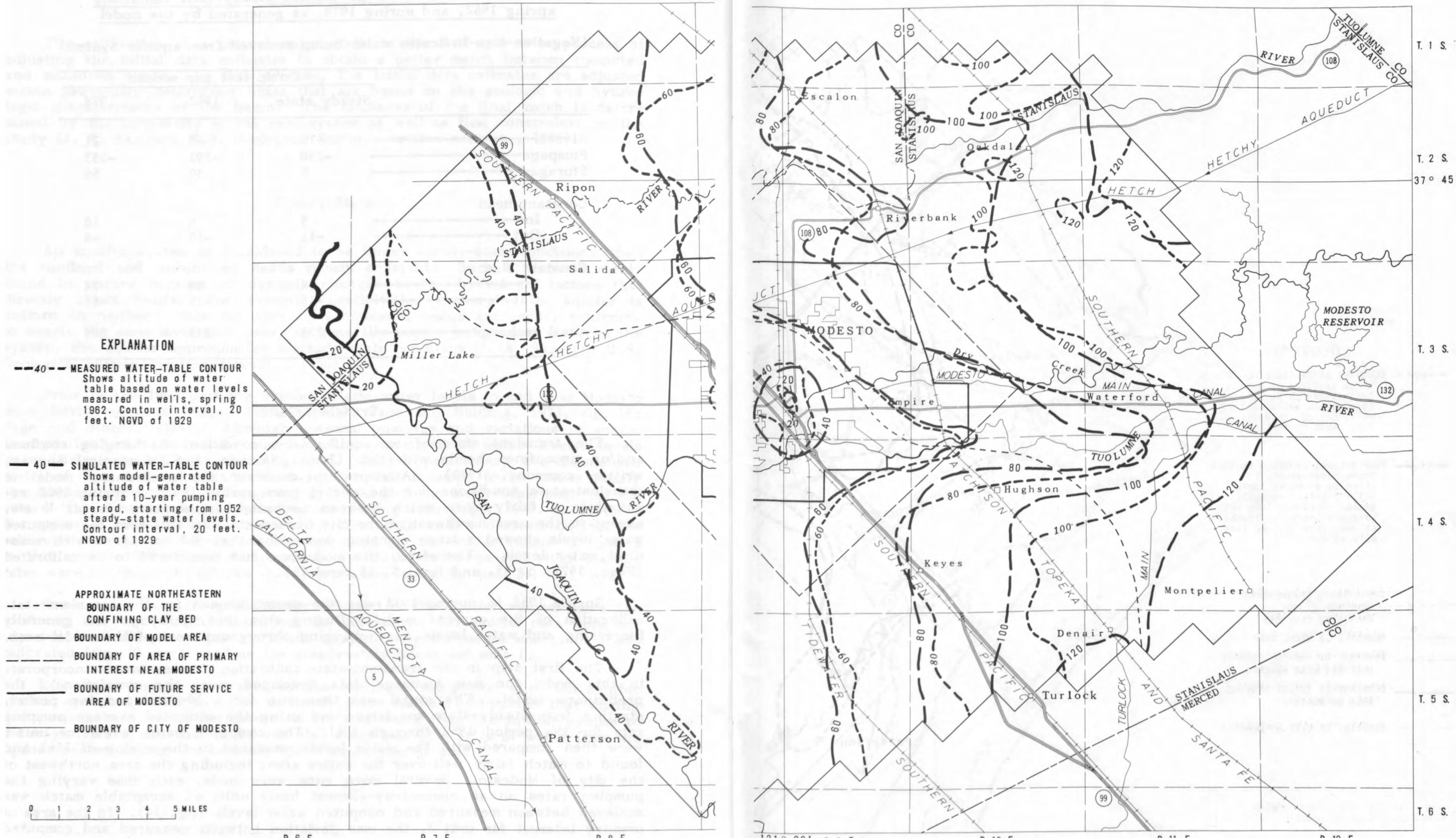


FIGURE 14. — Measured and model-generated water levels in the unconfined aquifer, spring 1962.

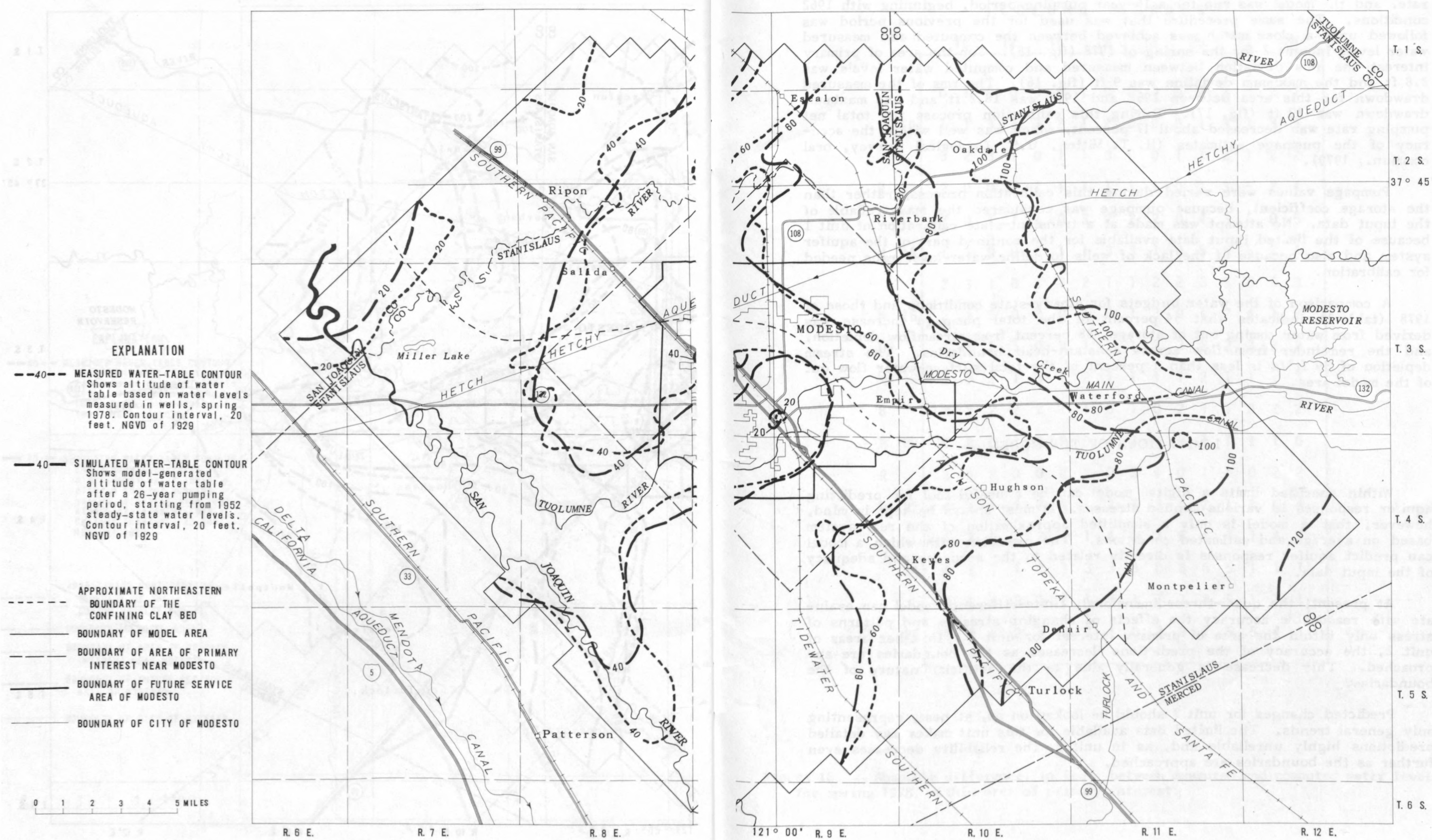


FIGURE 15. — Measured and model-generated water levels in the unconfined aquifer, spring 1978.

The 1952-61 average net pumping rate was then replaced with the 1962-77 rate, and the model was run for a 16-year pumping period, beginning with 1962 conditions. The same procedure that was used for the previous period was followed until a close match was achieved between the computed and measured water levels in unit 2 for the spring of 1978 (fig. 15). For the area of primary interest the rms deviation between measured and computed water levels was 2.8 ft and the maximum deviation was 9 ft (fig. 16). The rms of the measured drawdown for this area between 1952 and 1978 was 18.2 ft and the maximum drawdown was 32 ft (fig. 17). During this calibration process the total net pumping rate was decreased about 17 percent, which was well within the accuracy of the pumpage estimates (H. T. Mitten, U.S. Geological Survey, oral commun., 1979).

Pumpage values were varied during this calibration process, rather than the storage coefficient, because pumpage was considered the least reliable of the input data. No attempt was made at a transient-state calibration of unit 1 because of the limited input data available for the confined part of the aquifer system and also because of the lack of wells to define water-level maps needed for calibration.

A comparison of the water budgets for steady-state conditions and those of 1978 (table 1) indicates that 54 percent of the total pumpage increase was derived from water coming out of storage, 38 percent from streamflow depletion, and the remainder from flow across constant-head boundaries. The stream depletion of 39 ft^3/s is less than 1 percent of the total surface-water flow out of the model area.

LIMITATIONS OF THE MODEL

Within specified limits a digital model can be a useful tool for predicting aquifer responses to various applied stresses. It must always be kept in mind, however, that a model is only a simplified approximation of the real system based on average and estimated conditions. The precision with which a model can predict aquifer responses is directly related to the accuracy and adequacy of the input data.

At present, the quasi-three-dimensional, finite-difference model can evaluate with reasonable accuracy the effects of changing stresses and patterns of stress only within the area of primary interest for unit 2. In other areas of unit 2, the accuracy of the predictions decreases as the boundaries are approached. This decrease is generally due to the artificial nature of the boundaries.

Predicted changes for unit 1 should be looked on as, at best, representing only general trends. The limited data available for this unit makes any detailed predictions highly unreliable and, as in unit 2, the reliability decreases even further as the boundaries are approached.

COLUMN NUMBER

ROW NUMBER		10		12		14		16		18		20		22		24
	0	1	0	0	0	0	1	0	0	0	0	2	3	3	3	7
10	1	1	3	3	3	2	1	2	0	0	0	1	3	1	2	5
	3	2	3	1	0	1	1	0	0	0	1	1	2	1	2	2
12	2	0	0	2	1	1	1	1	1	1	0	1	3	5	2	4
	2	0	1	1	1	0	3	2	2	2	0	4	3	5	0	5
14	1	1	0	0	0	1	1	3	3	1	0	5	5	5	2	4
	2	3	1	0	1	1	2	1	1	2	2	3	5	6	3	3
16	2	3	3	0	0	0	1	0	1	3	0	1	2	4	1	2
	3	1	2	1	0	3	1	1	1	2	3	2	2	3	2	2
18	0	0	2	1	3	1	2	1	1	1	1	2	1	1	1	3
	3	2	2	2	2	2	1	0	1	0	2	1	1	3	2	2
20	4	2	0	1	2	0	0	4	3	0	0	1	1	1	1	0
	3	0	2	3	0	2	2	2	2	1	0	1	1	0	2	2
22	4	1	1	3	2	4	1	2	7	3	1	2	1	3	4	4
	5	3	1	2	2	1	1	4	6	7	6	4	1	1	4	4
24	4	3	1	1	3	3	3	3	4	8	8	6	5	1	1	1
	4	6	4	5	1	2	1	1	1	7	7	3	5	1	1	1
25	7	7	6	2	2	3	2	4	0	2	9	7	5	5	4	1

FIGURE 16. — **Absolute difference, in feet, between measured and computed water levels for spring 1978, within area of primary interest.**

COLUMN NUMBER

ROW NUMBER		10		12		14		16		18		20		22		24
	1	4	4	5	6	8	10	11	13	13	14	12	11	10	11	9
10	2	2	2	3	5	7	10	11	14	15	16	17	21	20	15	12
	1	3	5	9	12	14	16	17	17	18	18	18	17	17	20	17
12	4	6	10	12	15	19	22	24	22	22	21	19	17	15	22	24
	6	10	14	16	19	23	27	29	26	25	23	19	18	16	22	25
14	10	13	17	19	22	26	29	31	28	26	26	20	18	18	22	26
	10	13	18	20	23	25	28	28	27	26	27	22	19	18	21	26
16	12	15	18	22	23	25	26	26	24	23	26	25	23	20	23	27
	17	20	19	21	23	23	23	23	23	22	27	27	23	22	22	28
18	14	17	19	20	20	32	20	22	24	22	26	27	25	24	23	27
	10	16	20	18	17	17	17	19	21	19	24	26	26	28	25	23
20	11	11	15	15	12	12	11	10	12	15	20	23	25	27	23	20
	14	15	13	11	11	10	9	9	9	11	16	18	21	22	18	16
22	13	17	16	15	15	11	12	8	7	7	12	14	16	16	14	13
	11	14	17	19	19	19	17	10	6	5	8	10	14	15	13	11
24	11	14	18	22	24	24	21	15	12	7	6	9	10	14	15	14
	9	9	11	15	25	27	25	21	19	9	8	11	9	13	16	14
26	7	9	11	16	22	24	26	25	26	20	10	8	10	8	19	14

FIGURE 17. — Measured drawdown, in feet, between spring 1952 and spring 1978, within area of primary interest.

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