GROUND-WATER FLOW IN THE CENTRAL VALLEY, CALIFORNIA

REGIONAL AQUIFER-SYSTEM ANALYSIS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1401–D

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By ALEX K. WILLIAMSON, DAVID E. PRUDIC, *and* LINDSAY A. SWAIN

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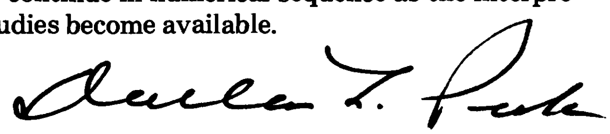
FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The Regional Aquifer-System Analysis (RASA) Program was started in 1978 following a congressional mandate to develop quantitative appraisals of the major ground-water systems of the United States. The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which in aggregate underlie much of the country and which represent an important component 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 the changes brought about in it by human activities, and 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.



Dallas L. Peck
Director

CONTENTS

ILLUSTRATIONS

* Note: These maps are also printed on transparent material and included in a pocket at the back of this report. The transparent maps may be used to easily compare any two or three maps to see cause and effect relationships as well as interrelationships between different characteristics or conditions of the aquifer system.

TABLES

CONVERSION FACTORS

For readers who may prefer to use International System of Units rather than inch-pound units, the conversion factors for the terms used in this report are listed below:

Multiply inch pound units	*By*	*To obtain SI units*
inch (in)	25.40	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
acre	0.4047	hectare
square mile (mi^2)	2.590	square kilometer
foot per day (ft/d)	0.3048	meter per day
foot per year (ft/yr)	0.3048	meter per year
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
gallon per minute (gal/min)	0.06309	liter per second
acre-foot (acre-ft)	1,233	cubic meter
square foot per second (ft^2/s)	0.09290	square meter per second

Sea level: In this report "sea level" refers to the 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 of 1929."

GROUND-WATER FLOW IN THE CENTRAL VALLEY, CALIFORNIA

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ABSTRACT

The agricultural productivity of the Central Valley is dependent on the availability of water from irrigation. About 7.3 million acres of cropland in the Central Valley receives about 22 million acre-feet of irrigation water annually. One half of this irrigation water is supplied by ground water, which amounts to about 20 percent of the Nation's ground-water pumpage. Ground water is important as a stable supply of irrigation water because of the high variability of surface-water supplies in the Central Valley. This large ground-water development during the past 100 years has had major impacts on the aquifer system, such as decline in water levels, land subsidence, depletion of the aquifer storage, and increase in recharge. The flow conditions before and during development were simulated on a regional scale using a three-dimensional finite-difference flow model.

The Central Valley is a large (20,000-square-mile) structural trough filled with poorly permeable marine sediments that are overlain by coarser continental sediments. In general, previous investigators have conceptualized the northern one-third of the valley—the Sacramento Valley—as a water-table aquifer and the southern two-thirds—the San Joaquin Valley—as a two-aquifer system separated by a regional confining clay layer. A somewhat different concept of the aquifer system was suggested during this study by analyses of water-level measurements, texture of sediments interpreted from electric logs, and flow-model simulations. Vertical hydraulic head differences are found throughout much of the Central Valley. Early in development, flowing wells and marshes were found throughout most of the central part of the valley. More than 50 percent of the thickness of the continental sediments is composed of fine-grained lenticular deposits that are discontinuous but are distributed throughout the stratigraphic section in the entire Central Valley.

The concept presented in this report considers the entire thickness of the continental deposits as one aquifer system which has varying vertical leakance that depends on several factors, including amount of fine-grained sediments. The average horizontal hydraulic conductivity is about 6 feet per day, and the average thickness of the continental deposits is about 2,400 feet.

Irrigation use, which averaged 22 million acre-feet of water per year during 1961–77, increased evapotranspiration about 9 million acre-feet per year over its predevelopment value. This is a large figure compared with the average annual surface-water inflow to the Central Valley of 31.7 million acre-feet per year. Precipitation on the valley floor is mostly lost to evapotranspiration. The overall postdevelopment recharge and discharge of the aquifer system was about 6 times greater than the predevelopment estimated values. The increases of pumpage associated with development mostly in the San Joaquin Valley have caused water-level declines that exceed 400 feet in places and have resulted in the largest known volume of land subsidence due to fluid withdrawal in the world. Water in aquifer storage declined about 60 million acre-feet from predevelopment to 1977; 40 million acre-feet were derived from the water-table zone, 17 million acre-feet from compaction of sediments, and 3 million acre-feet from elastic storage. During 1961–77, ground water withdrawn from aquifer storage averaged about 800,000 acre-feet per year.

The flow model constructed during this study was calibrated principally in accordance with the hydrologic data observed during 1961–75 because little predevelopment data were available for analysis. An explicit algorithm to simulate land subsidence was developed and calibrated. The simulated land subsidence was within 6 percent of the estimated observed volume; however, the time lag associated with this type of subsidence was not adequately simulated. Simulated water-level changes averaged 2.6 and 12 feet higher than observed water-level changes for the water table and the lower pumped zones, respectively, and the standard deviation of the simulated changes minus the observed change was 22 and 27 feet, respectively. The flow model was tested for the period of 1976–77 drought with good results. The simulations indicated that vertical leakance greatly increased from the predevelopment values as a result of water flowing through some of the more than 100,000 irrigation well casings that are open to different aquifer layers.

The simulation results are shown on maps for comparison with observed hydrologic data. A description of the computer-tape file, which contains estimates of recharge/discharge, and the aquifer properties used in the simulation are included in appendix A and B, respectively. The theoretical basis of calculating borehole hydraulic conductance of multilayer wells which cause increases in vertical leakance during the post-development period is discussed in appendix C.

INTRODUCTION

The Central Valley of California (fig. 1) has fertile soil and a long growing season, conditions that are conducive to farming. Almost 40 percent of the total U.S. production of vegetables, fruits, and nuts come from this valley (U.S. Department of Commerce, 1978). The valley floor, where agricultural production is most intense, has an average water deficiency (precipitation minus potential evapotranspiration) under natural conditions of as much as 40 in/yr (Thomas and Phoenix, 1976, p. 2). Thus, agricultural development in the valley is dependent on water from sources other than direct precipitation.

The water needed for agricultural production is obtained from two sources. The first source is streams and rivers that enter the valley from the surrounding mountain ranges, where there is a surplus of water. The surface water is diverted by canals to areas of farming. The second source is ground water, which is used primarily where surface-water supplies are not available or are not sufficient or dependable enough to support agricultural activities.

The amount of water required to support agriculture averages about 22 million acre-ft/yr. Ground-water withdrawals in the Central Valley account for about one-half of the total water used. This amount is equal to 74 percent of the total annual ground-water pumpage in California (Kahrl, 1978) and is more than 20 percent of the total annual ground-water pumpage for the entire United States (Murray and Reeves, 1977).

This large demand for ground water has placed considerable stress on the aquifer system within the valley. Ground-water pumpage has exceeded recharge in several parts of the valley and has caused water levels to decline more than 400 ft. In some areas, water levels have declined below sea level (Thomas and Phoenix, 1976; Bertoldi, 1979). The effect of excessive pumpage in the valley has been the greatest volume of land subsidence due to fluid withdrawal recorded anywhere in the world (J.F. Poland, U.S. Geological Survey, oral commun., 1982). More than 5,200 mi^2 of land surface has subsided more than 1 ft, and at one location subsidence exceeds 29 ft (Ireland and others, 1984).

PURPOSE AND SCOPE

The Central Valley aquifer system was studied as part of the nationwide Regional Aquifer-System Analysis (RASA) program of the U.S. Geological Survey. The valley was chosen for study because of (1) its long history of intensive ground-water development, (2) its dependence on ground water to maintain agricultural productivity, (3) previous studies of the aquifer system were limited to localized geographic areas or to defining only part of the system, and (4) the large size (20,000 mi^2) and complexity of the system. The scope of the overall project was to collect, interpret, and verify hydrologic information from numerous sources with the goal of quantifying the hydrologic conditions of the entire system and to develop methods of evaluating aquifer responses to changes in ground-water-management practices (Bertoldi, 1979, p. 9). The purposes of the study reported herein, which is part of the overall Central Valley RASA project, are to (1) evaluate the aquifer system on a regional basis, mainly through the use of a mathematical (computer) model, (2) simulate conditions that existed before development of the ground-water resources (prior to 1870), (3) simulate present conditions, and (4) identify changes in the ground-water system caused by development of the valley's water resources. Simulation of the aquifer system using a mathematical model was chosen as a method for analysis because it integrates large amounts of diverse types of data, testing both the conceptualization of the system and the aquifer characteristics.

Only those aspects that directly apply to the analysis of aquifer properties and to ground-water flow within the system between Red Bluff in the north and Bakersfield at the south end of the valley (fig. 1) are included in this report. Detailed descriptions of the water quality and geology of the Central Valley are discussed in separate reports, as is information that pertains to the drilling of test holes. This report presents information on recharge, evapotranspiration, and pumpage. The methods of computation of these hydrologic variables are discussed in detailed reports by Diamond and Williamson (1983) and Williamson (1982).

PREVIOUS INVESTIGATIONS

No comprehensive report on the modeling of ground-water flow of the entire Central Valley of California has been published. The Central Valley has been studied or modeled in different areas by several investigators since about the late 1880's. The earliest reliable systematic study was by W. Hammond Hall (1886), the California State engineer from 1878 to 1889. Hall's work, together with Mendenhall and others' (1916) study of ground-water resources of the San Joaquin Valley and Bryan's (1923) study of the Sacramento Valley, helped formulate the concepts of the aquifer system in the valley during a period when there was little stress on the system.

Between 1923 and the end of World War II (1945), virtually no quantitative investigative reports for the Central Valley were published; however, ground-water data were being accumulated. It was during the period 1923–45 that hundreds of exploratory gas and oil wells were drilled and logged in the valley, and these logs provided basic information on the lithologic character of the aquifer system, including the lower boundary of alluvium, the distribution of coarse- and fine-grained materials, and the distribution of minerals.

Post-World War II agricultural growth and attendant ground-water use in the valley increased so rapidly that by 1950 California pumped nearly 50 percent of all the ground water pumped in the United States. With this increased pumping, virtually tens of thousands of wells were drilled in the Central Valley, making available a greatly expanded set of data upon which to renew scientific investigation of the ground-water resources. The new data allowed Croft (1968, 1972) to map an important confining bed that extends over nearly 5,000 mi^2 of the San Joaquin Valley and four other lesser confining beds. From the data gathered from 1945 to 1960, Davis and others (1959) and Olmsted and Davis (1961) were able to define geologic features and to estimate the storage capacity of the upper 200 ft of the aquifer system in the San Joaquin and Sacramento Valleys. Eighty-six papers reporting on subsidence research were published by the U.S. Geological Survey between the years 1950 and 1983. These papers describe the mechanics of subsidence caused by compaction of both shallow deposits (hydrocompaction) and deep deposits (owing to withdrawal of ground water, oil and gas fluids) in the San Joaquin Valley. These reports contain valuable data that were used to form the initial model values of specific-storage coefficients, specific yields, and vertical and horizontal hydraulic conductivity.

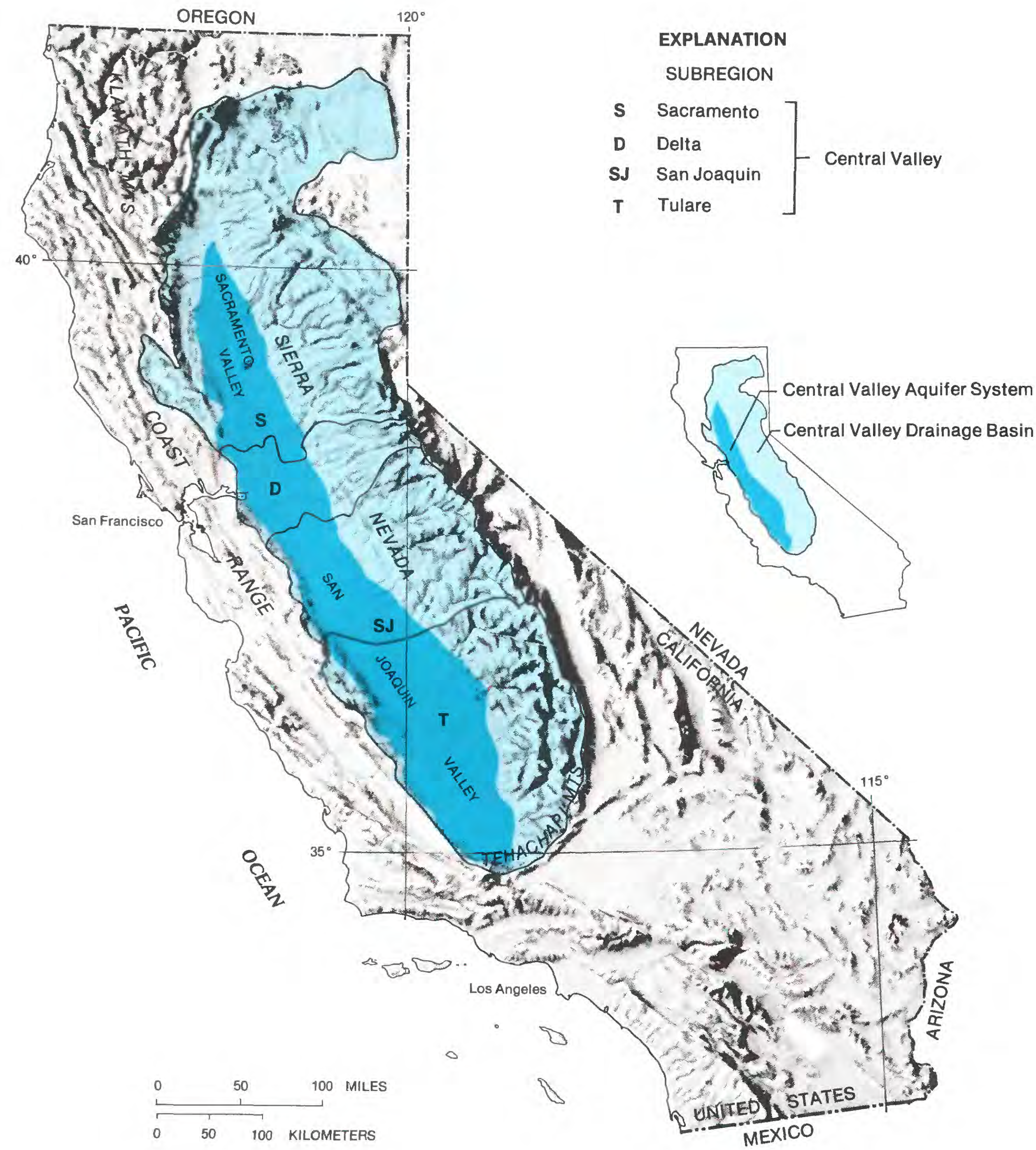

FIGURE 1.—Location of the Central Valley, Calif. (modified from Thomas and Phoenix, 1976).

The California Department of Water Resources administers two programs—one that provides ground-water-level data dating back to 1921 and another that provides comprehensive land-use data, with resurveying in most areas every 5 to 10 years. These basic data provided valuable data on head distribution, evapotranspiration, recharge, distribution of pumpage, and irrigation return flow.

Since about 1970, several investigators have developed ground-water-flow models for parts of the valley. Bloyd (1978) designed an uncalibrated, unverified flow model for natural flow conditions in the Sacramento Valley. The California Department of Water Resources (1977b), in cooperation with the Kern County Water Agency, developed a calibrated flow model for the Kern County area of the Tulare Basin in part of the southern San Joaquin Valley. Londquist (1981) and Page (1977) developed models of parts of the aquifer system in areas of San Joaquin and Stanislaus Counties; the California Department of Water Resources (1974a) designed a mathematical model to simulate man's impact on the water resources of Sacramento County. A contractor for the California Department of Water Resources (1982) has developed a calibrated three-dimensional flow model of the San Joaquin Valley for use in coordination with an economic optimization model. Mitten (1983) and C.J. Londquist (U.S. Geological Survey, written commun., 1983) are using ground-water-flow models to study the aquifer system in the Fresno and Madera areas, respectively. Corapcioglu and Brutsaert (1977) developed a model to simulate land subsidence caused by pumping in a few sites in the San Joaquin Valley. These models provided some information for estimation of initial boundary conditions and comparative values for hydraulic conductivity and storage coefficients where applicable to the regional model discussed in this paper.

Although the foregoing studies provided the bulk of the background information, it would be negligent to omit mention of other sources of information. Nearly 600 reports (Bertoldi, 1979) and numerous data obtained from 300 local agencies, farmers, and industrial managers were used in formulating and corroborating the characteristics of the regional aquifer system of the Central Valley.

WELL-NUMBERING SYSTEM

The well-numbering system commonly used in California is shown and explained in figure 2.

DESCRIPTION OF THE CENTRAL VALLEY

Surrounded by mountains and filled with alluvium and other sediments, the Central Valley extends more than 400 mi from near Red Bluff in the north to near Bakersfield in the south (fig. 1). The valley ranges in width from about 20 to 70 mi and covers an area of approximately 20,000 mi^2. Geologically, it is one of the most notable structural troughs in the world.

The Central Valley is subdivided into two distinct valleys, each drained by a major river after which that part of the valley is named. As a result, the northern one-third of the valley is called the Sacramento Valley and the southern two-thirds is called the San Joaquin Valley. The southern part of the San Joaquin Valley, sometimes called the Tulare Basin, is a basin of interior drainage where water often collects in nearly dry-lake areas known as Kern Lake, Buena Vista Lake, and Tulare Lake beds (informal usage) (fig. 3). The two valleys are separated by an area commonly called the Delta where the Sacramento and San Joaquin Rivers meet and discharge through a natural outlet at Suisun Bay and into San Francisco Bay. The Central Valley basin can be subdivided for study into four subregions: Sacramento, Delta, San Joaquin, and Tulare (fig. 1).

Topographically, the Central Valley is relatively flat and of low altitude. The only feature of prominent relief within the valley is the Sutter Buttes, which rise about 2,000 ft above the valley floor near the center of the Sacramento Valley. Altitudes in the valley are mostly less than 500 ft above sea level. Maximum altitudes of about 1,800 ft occur at the apexes of some alluvial fans along the south and northwest perimeters and on the Sutter Buttes to the north. Two areas within the valley—the Sutter Buttes and the Kettleman Hills—(fig. 3) are not part of the aquifer system.

HYDROLOGY

The climate in the valley is of Mediterranean type (dry summers). Average annual precipitation ranges from 13 to 26 inches in the Sacramento Valley and from 5 to 16 inches in the San Joaquin Valley (fig. 4). About 85 percent of the annual precipitation occurs in the 6 months from November through April (fig. 5). Summers are hot, and winters are moderate and allow a long growing season.

Streamflow, a very important factor in the water supply of the valley, is entirely dependent on precipitation in the Sierra Nevada and in parts of the Klamath Mountains in the north (fig. 1). No perennial streams of any significant size enter the valley from the west side except those in the northwest end of the valley. The mean annual streamflow entering the Central Valley around its perimeter is 31.7 million acre-ft. Mean annual precipitation in the mountains increases with altitude to as much as 90 in (Rantz, 1969). Much of the precipitation in the mountains occurs in the form of snow, especially in the higher southern Sierra Nevada.

The resulting snowpack delays runoff so that about 78 percent of the total unimpaired streamflow to the valley occurs during the 6 months from January through June (fig. 5).

Precipitation and runoff in the valley vary greatly from year to year as well as within the year (fig. 6). The standard deviation of annual flows ranges from 40 to 80 percent of the mean among the major streams. Years

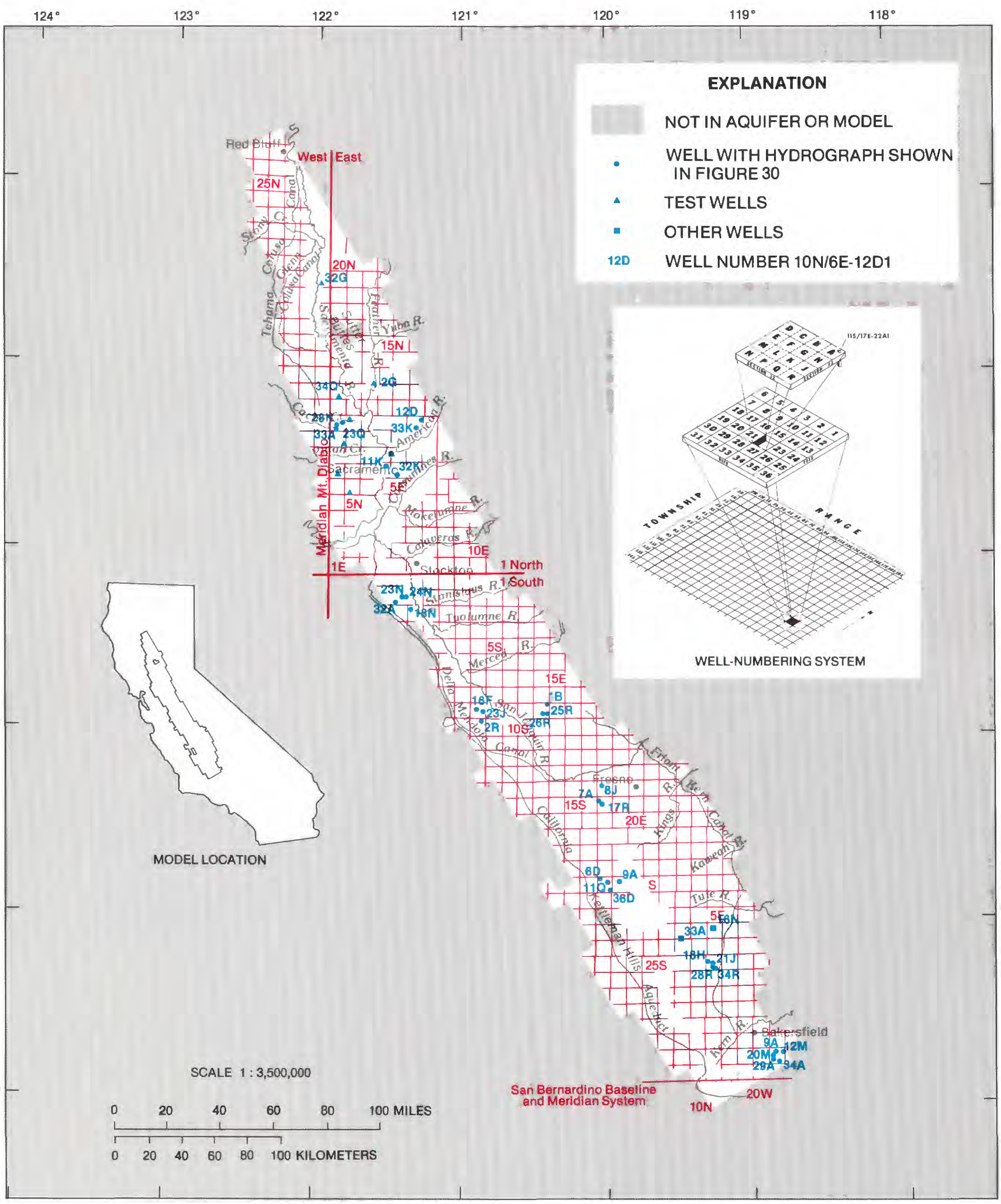

FIGURE 2.—Well-numbering system and well locations.

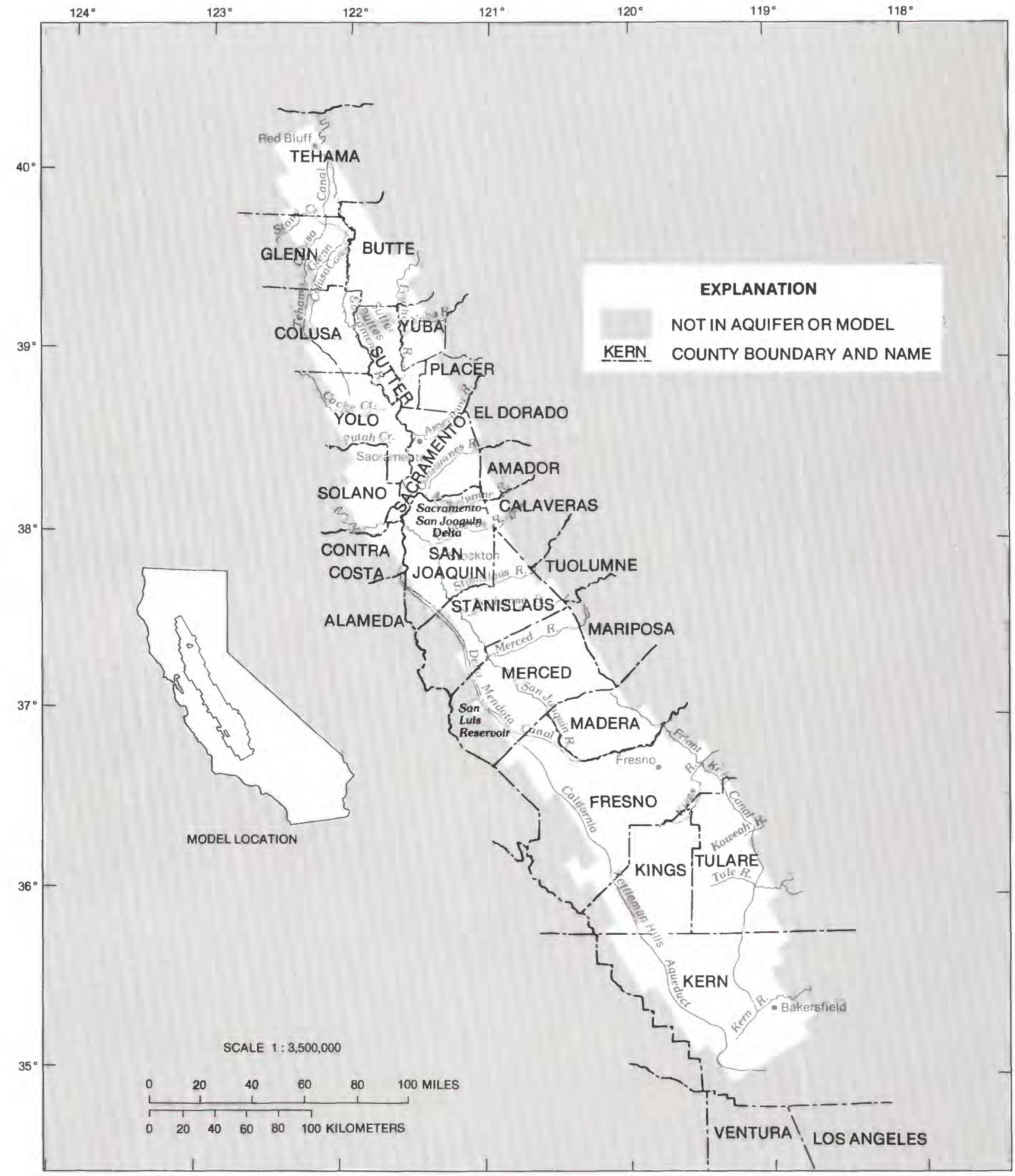

FIGURE 3.—Geographic features in the Central Valley.

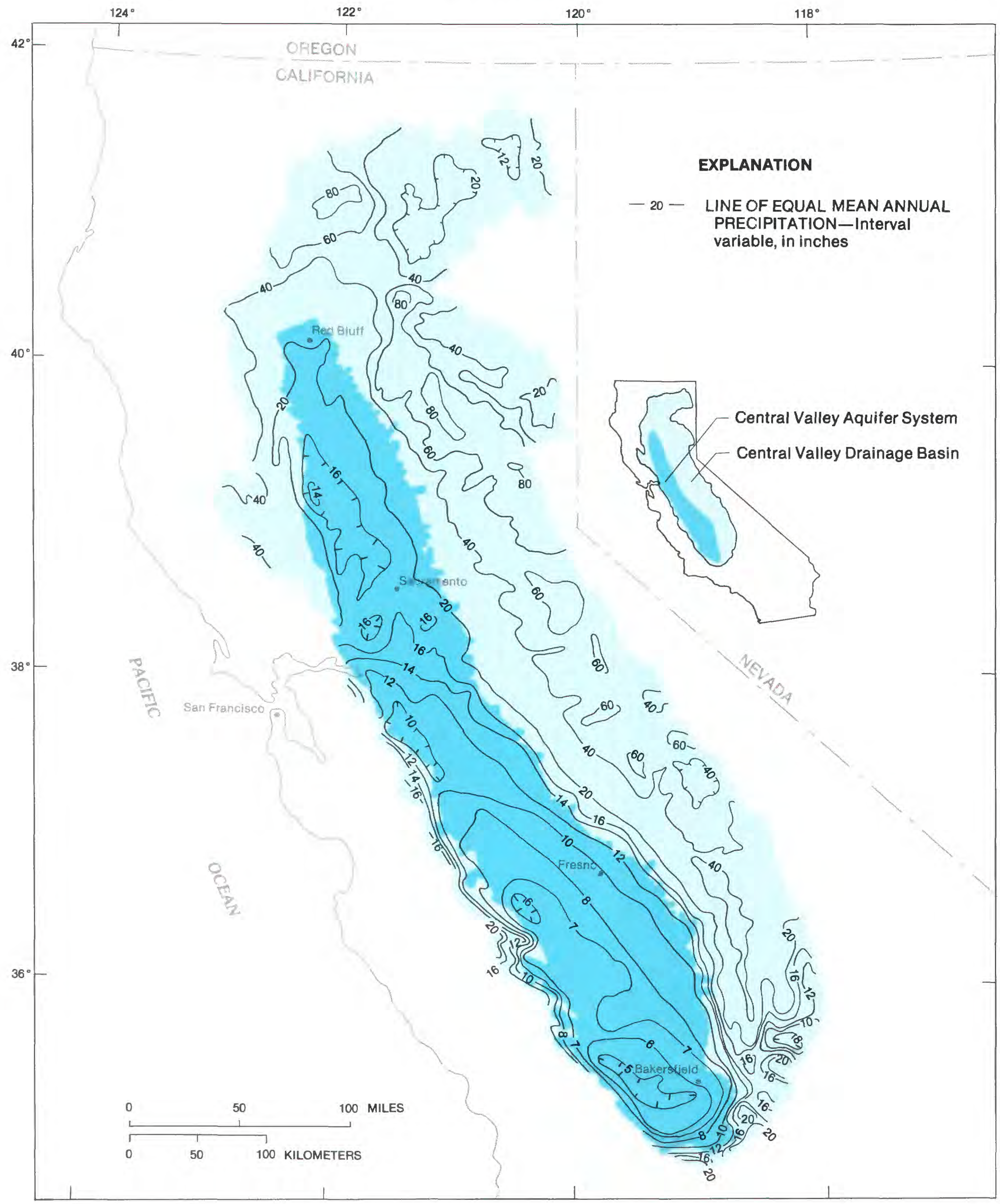

FIGURE 4.—Mean annual precipitation in the Central Valley drainage basin (from Rantz, 1969).

when precipitation is near the mean are somewhat rare. A relatively stable measure of variability in the valley would be the sum of the 15 largest streams' annual flow, because often one end of the valley is wetter or drier than the other. However, for this flow (sum of the 15 largest streams' annual flow), only 2 (1962 and 1975) of the 17 years (1961–77) and only 16 percent of 44 years of record were within 10 percent of the mean annual flow. Figure 7 shows the periods of greater (curve rises) and less (curve falls) than normal precipitation since the late 1800's.

GEOLOGY

The geology of the Central Valley is described in an accompanying report (Page, 1986); therefore, this section contains information pertinent only to an understanding of the ground-water-flow system.

In general, the Central Valley is a long, northwest-trending, asymmetric structural trough that is filled with sediments. Along the eastern part of the valley the sediments are underlain by pre-Tertiary crystalline and metamorphic rocks of the Sierra Nevada block (Davis and others, 1959, p. 40; Olmsted and Davis, 1961, p. 39). The sediments are thought to be underlain by a pre-Tertiary mafic and ultramafic complex in the west side and part of the east side of the valley (Cady, 1975, p. 17–19; Suppe, 1978, p. 7). Generally, only minor quantities of water are present in the joints and cracks of these pre-Tertiary rocks.

Rocks of the Coast Ranges on the west side of the valley consist mainly of pre-Tertiary and Tertiary semiconsolidated to consolidated clastic sediments of marine origin that have been folded and faulted. These deposits extend eastward underneath the Central Valley where, near the east edge, they become thinner (Davis and others, 1959, p. 40; Olmsted and Davis, 1961, p. 42). The marine sedimentary rocks contain saline water except in a few areas where freshwater has apparently

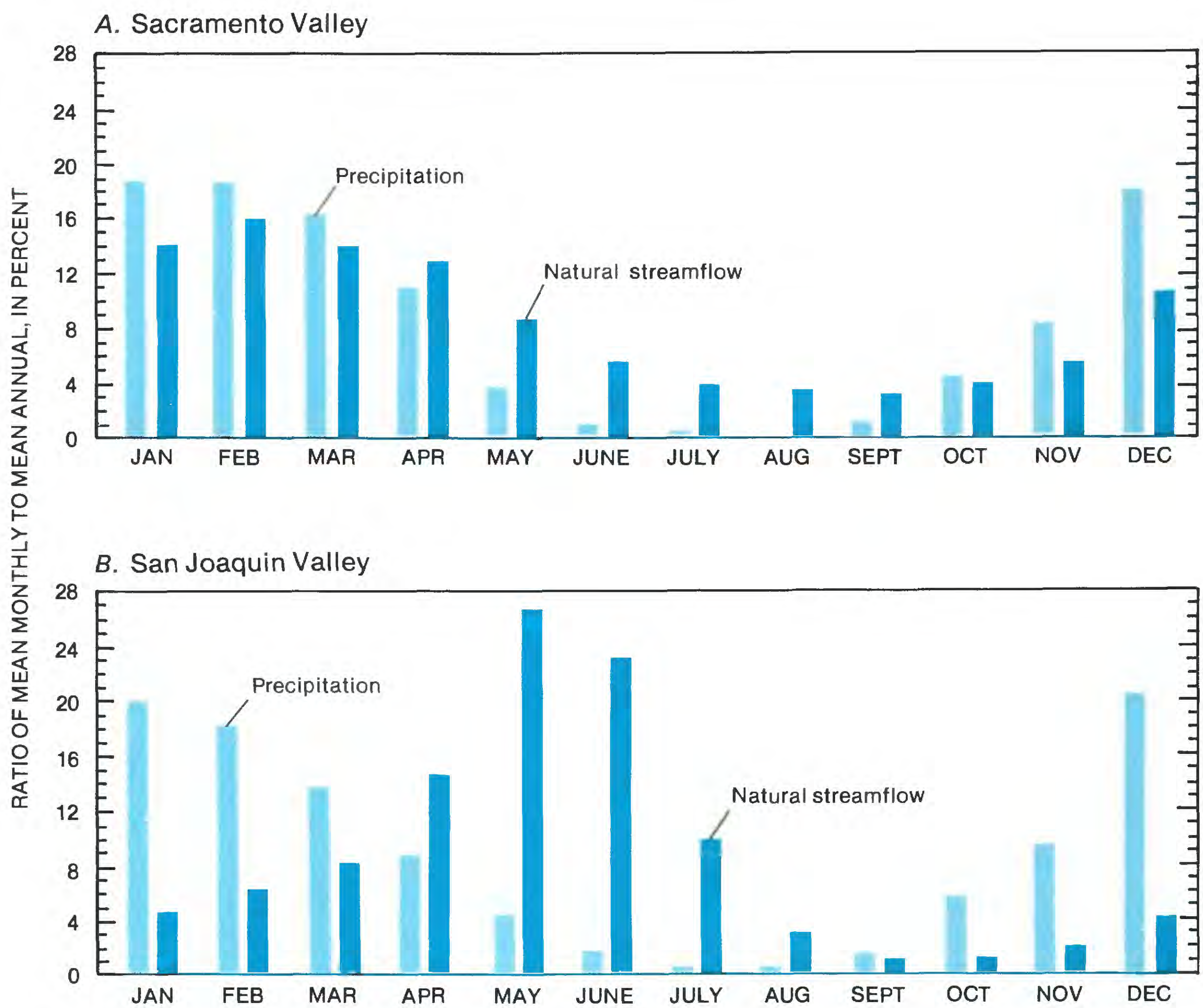

FIGURE 5.—Mean monthly precipitation and streamflow in the (*A*) Sacramento Valley and (*B*) San Joaquin Valley, as a percentage of the mean annual precipitation and streamflow, respectively.

flushed out some of the saline water (Davis and others, 1959, p. 44; Olmsted and Davis, 1961, p. 134; Page, 1986).

Continental deposits of post-Eocene to Holocene age overlie the marine sedimentary rocks (fig. 8) (R.W. Page, written commun., 1981, and Page, 1974). The continental deposits include some volcanic material but contain mostly fluvial deposits with lesser amounts of interbedded lacustrine deposits. The continental deposits consist predominantly of lenses of gravel, sand, silt, and clay. The numerous lenses of fine-grained deposits (clay, sandy clay, sandy silt, and silt) are distributed throughout the valley and constitute over half of the total thickness penetrated by wells, as determined from electric logs (Page, 1986, fig. 35). Most of these lenses are not widespread, although several major ones have been mapped in the valley—principally beneath the axis of the San Joaquin Valley. The most notable deposit is the Corcoran Clay Member (Pleistocene) of the Tulare Formation (Pliocene and Pleistocene), which is part of the E-clay of Croft (1972) in the San Joaquin Valley. This diatomaceous clay bed covers an area of approximately 5,000 mi² (Page, 1986, plate 4) and ranges in thickness from near zero to at least 160 ft beneath the present bed of Tulare Lake (Davis and others, 1959; Page, 1986, p. 16). The northern extent of the Corcoran Clay Member is not known because of the absence of data north of Stockton, particularly in the Delta area. A diatomaceous clay similar in composition to that of the Corcoran Clay Member was found in a test hole (12N/1E-34Q) drilled in the Sacramento Valley (Page and Bertoldi, 1983). The location of this hole is shown in figure 2. Laboratory tests of the clay indicate that it is highly susceptible to compaction, like the Corcoran Clay Member; however, the clay was not found in six other test holes in the area (fig. 2), and the full extent of this clay is not known.

LAND SUBSIDENCE

The many fine-grained (clayey) lenses in deposits of the Central Valley are conducive to subsidence, both

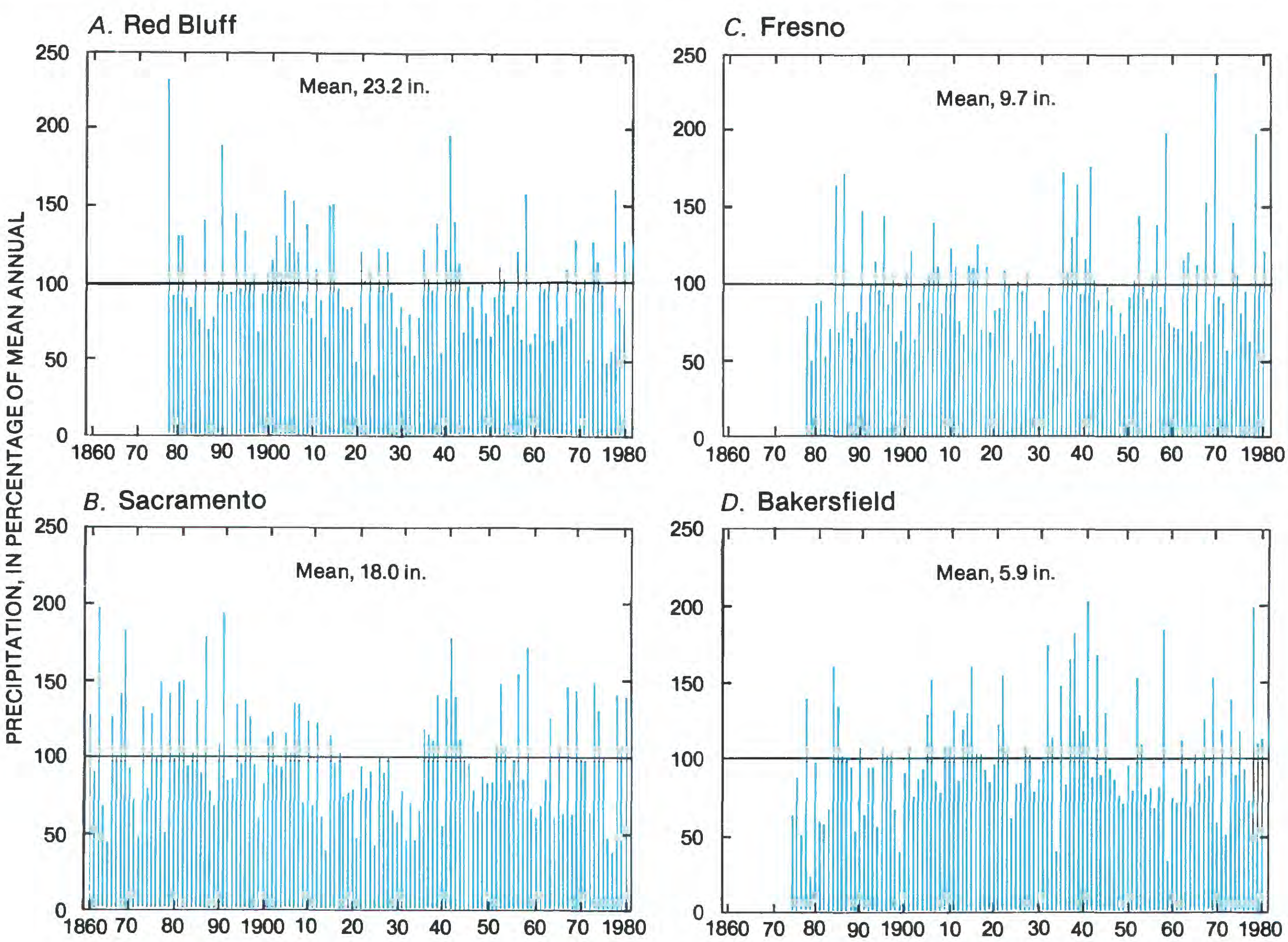

FIGURE 6.—Annual precipitation, 1860–1980, at four locations in the Central Valley: (A) Red Bluff, (B) Sacramento, (C) Fresno, and (D) Bakersfield.

naturally and by man-induced activities. The five processes that are known to cause land subsidence in the Central Valley, in order of their magnitude, are

1. Compaction of the aquifer system caused by lowering of the hydraulic head in the aquifer system;
2. Oxidation and compaction of peat soils caused by draining the lands near the confluence of the San Joaquin and Sacramento Rivers;
3. Compaction of moisture-deficient deposits above the water table (referred to as "hydrocompaction") caused by applying water at land surface to previously dry sediments;
4. Compaction of deposits below the aquifer system caused by fluid withdrawal from oil and gas fields; and
5. Deep-seated tectonic settling.

Of these five processes that cause land subsidence in the Central Valley, only the first two listed have altered the ground-water system or changed the physical properties of the aquifer materials. The other three processes have had little impact on the ground-water flow system as a whole. All five processes are briefly discussed in the following paragraphs.

Compaction of the aquifer system caused by the lowering of the hydraulic head has caused the greatest amount of subsidence over the largest area in the Central Valley (fig. 9). Most of the land subsidence has occurred in the San Joaquin Valley south of the Merced River where approximately 5,200 mi^2 had subsided at least 1 ft by 1970 and a maximum subsidence of 29.6 ft was measured at one location in 1977 (Ireland and others, 1984, p. 2). In the Sacramento Valley, the maximum amount of subsidence by 1973 was about 2 ft in at least two small areas in the southwestern part of that valley (Lofgren and Ireland, 1973, p. 6). Lofgren and Ireland (1973, p. 6) noted that other areas may have also subsided, but precise leveling data were not available for several parts of that valley. Leveling data near Zamora in the Sacramento Valley (J.C. Blodgett, U.S. Geological Survey, written commun., 1979) indicate that subsidence in that area has increased between 1973 and 1979.

Compaction of the aquifer system occurs mainly in the fine-grained sediments. When the hydraulic head in the aquifer system declines to a level below the preconsolidation stress, the fine-grained sediments compact and release water. Such compaction is a one-time source. Thus, the storage capacity of the aquifer system is reduced, even though the storage capacity of the coarse-

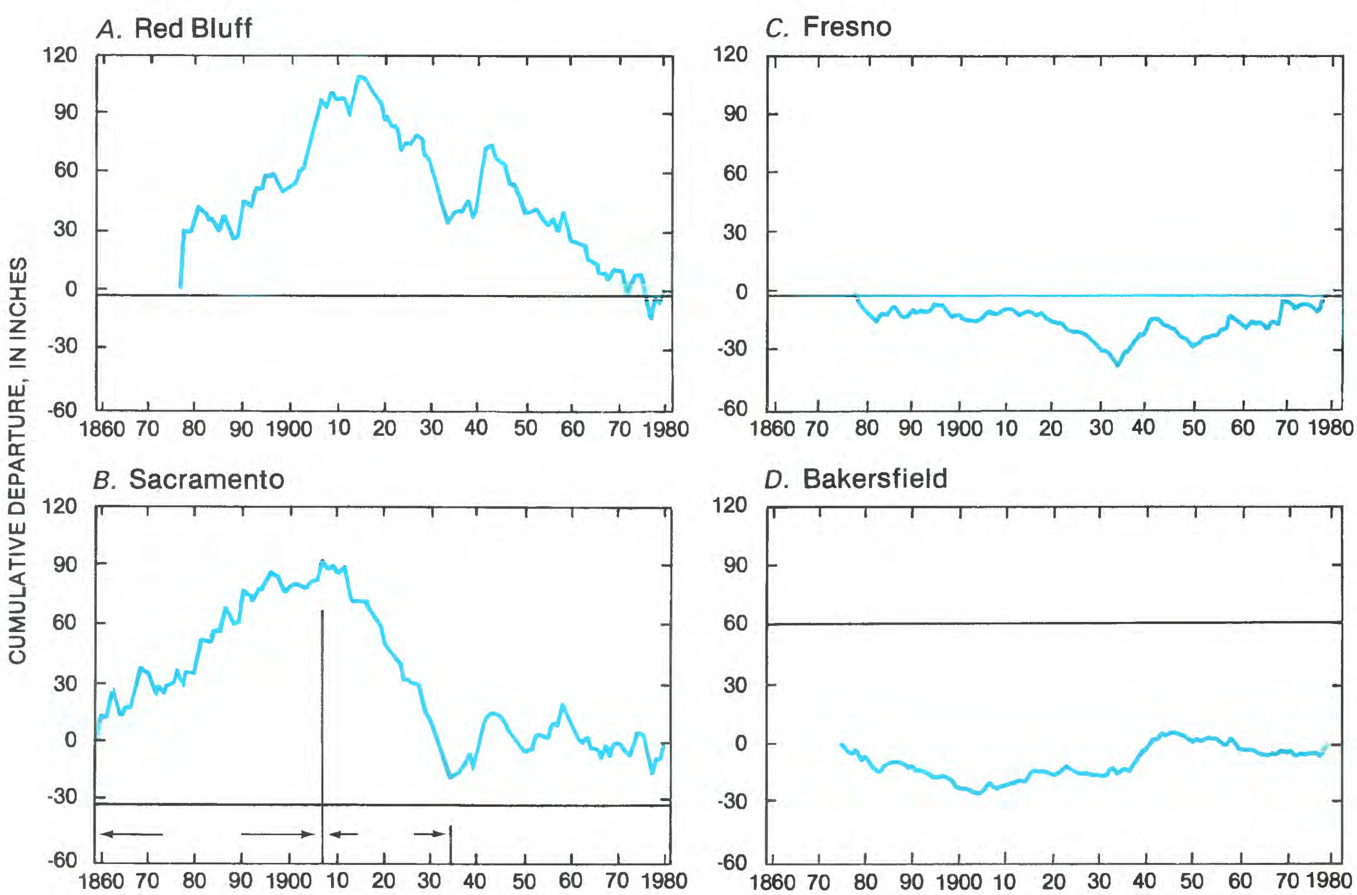

FIGURE 7.—Cumulative departure of precipitation, 1860–1980, from the mean annual at four locations in the Central Valley: (*A*) Red Bluff, (*B*) Sacramento, (*C*) Fresno, and (*D*) Bakersfield.

grained parts of the system may remain constant. During periods of water-level decline, compaction reduces the amount of drawdown by providing a source of water from the fine-grained sediments to pumping wells. On the second cycle of drawdown, after recovery of water levels due to cessation of pumping or to

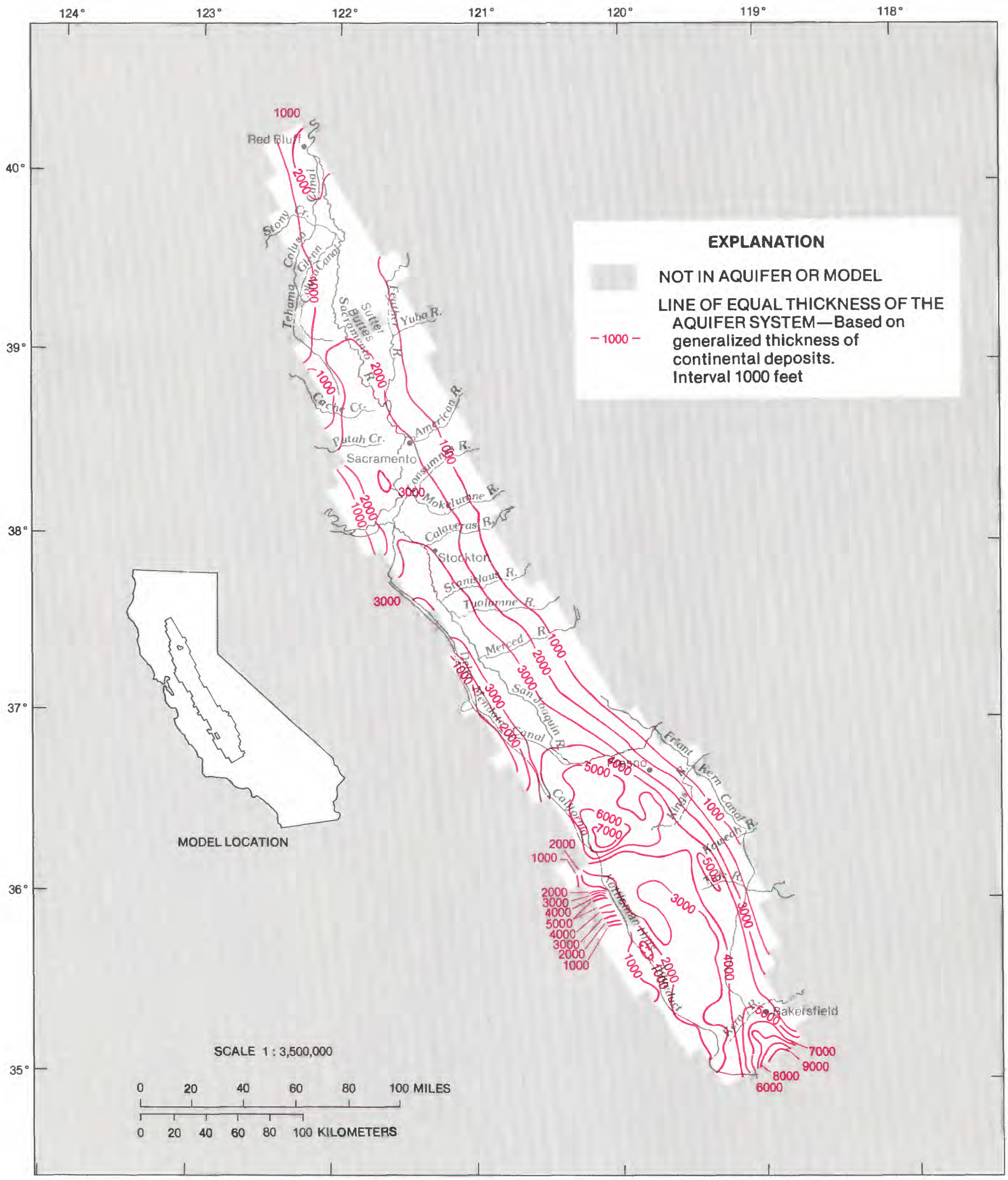

FIGURE 8.—Thickness of the aquifer system, based on the generalized thickness of continental deposits. (After R.W. Page, 1981, written commun., and Page, 1974.)

recharge, the water release from the compacted fine-grained sediments does not occur and water levels decline rapidly.

Land subsidence caused primarily by the oxidation and compaction of peat soils after marshlands were drained to grow crops has affected an area about 410

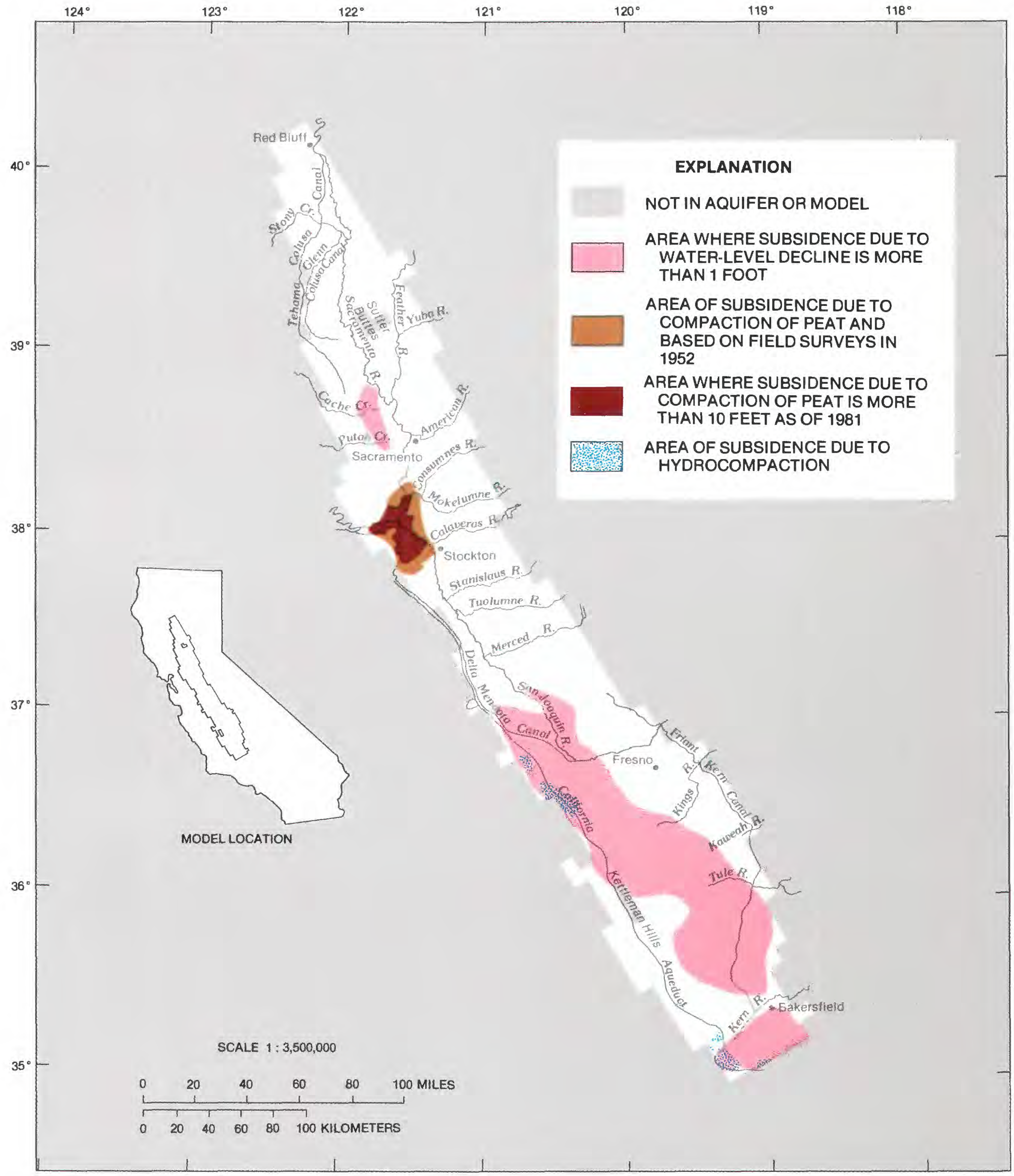

FIGURE 9.—Areas affected by land subsidence.

mi^2 near the confluence of the Sacramento and San Joaquin River systems (Poland and Evenson, 1966, and fig. 9). Based on a map by Newmarch (1981), an area of about 170 mi^2 has subsided at least 10 ft since reclamation began to 1980. Drainage for cultivation of this low-lying area began in 1850, and by at least 1922 the entire area was under cultivation (Weir, 1950). Today, the area is a complex system of manmade islands and channels. Prior to development much of the marshland was at or above sea level, but since development much of the area is below sea level and is continuing to subside about 3 in/yr (Newmarch, 1981). In some places as much as 40 ft of loose organic peat overlies the sediments. Weir (1950) estimated that subsidence in the lower Jones Tract was 4.5 ft for the period 1902 (when the tract was first drained) to 1917. Poland and Evenson (1966) reported that subsidence on one island was more than 9 ft from 1922 to 1955, and Newmarch (1981, p. 135) reported a maximum of 21 ft on one island as of 1980.

Perhaps the most critical problem in the area near the confluence of the Sacramento and San Joaquin Rivers is that the peat lands continue to subside. To allow farming, the water table in the islands has to be lowered by pumping water out of drains and discharging into the rivers, thus increasing the hydraulic gradient from the river toward the island.

Compaction of deposits above the water table after water was applied at the surface (called hydrocompaction) has been documented in a few areas on the west and south ends of the San Joaquin Valley (Bull, 1964; Lofgren, 1969; Poland and others, 1975, p. H8). The total area that was affected by hydrocompaction in the San Joaquin Valley is about 210 mi^2 (fig. 9). Subsidence of 5 to 10 ft is common in these areas and, locally, subsidence of 15 ft has been observed (Poland and Evenson, 1966, p. 244).

Compaction of deposits beneath the aquifer system caused by fluid withdrawal from oil and gas fields may cause local land subsidence. Lofgren (1975, p. D33) noted that subsidence around oil fields south and west of Bakersfield was generally less than 1 ft during the period of leveling from 1935 to 1965. However, the maximum amount of subsidence may have occurred earlier because peak production from these fields was before 1935. Lofgren and Ireland (1973) noted that some subsidence caused by fluid withdrawal from oil and gas fields in the Sacramento Valley may have also occurred, although data are sparse. Similarly, Newmarch (1981, p. 140) indicated that as much as a foot of subsidence could be attributed to the removal of fluids from a few gas fields near the Delta and noted that the subsidence was probably limited to areas close to the fields.

Little information is available for the rates of tectonic downwarping in the Central Valley. Lofgren (1975) indicated that structural downwarping has been uniform since the Pleistocene in the southwestern part of the San Joaquin Valley based on calculations of average depositional rates from carbon-14 dates and that the rate of downwarping is sufficiently slow that it has not affected the historical span of leveling. Newmarch (1981, p. 138) estimated a rate of tectonic downwarping of 0.006 in/yr for the southern end of the Sacramento Valley, assuming that downwarping began 6 million years ago, that the approximately 3,000 ft of alluvial materials were deposited at sea level, and that the base of these deposits moved downward owing to tectonic downwarping. Evidence of tectonic movement was noted by Poland and others (1975, p. H8) in the southern Coast Ranges near the southwestern end of the Central Valley and in the Tehachapi Mountains to the south, where apparent movements of as much as 0.8 ft have been measured at bench marks. During the period of development in the Central Valley (about 130 years), the overall effect of this process on the total observed land subsidence has been minimal compared with the effect of other processes.

ACKNOWLEDGMENTS

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MODEL DEVELOPMENT

A three-dimensional ground-water-flow model, developed for this study, was used to analyze the aquifer system in the valley. This section describes (1) the concepts and development of the flow model, (2) the initial estimates of recharge, discharge, and hydraulic properties of the aquifer system used in the model, and (3) the procedure used to calibrate the flow model by modifying the initial estimates of recharge, discharge, and aquifer properties.

SIMULATION OF GROUND-WATER FLOW

Ground-water flow in the Central Valley was simulated with a finite-difference model. A finite-difference model is a set of ground-water-flow equations with representative aquifer properties which can describe ground-water flow in the aquifer system. The set of ground-water-flow equations then can be solved simultaneously with the aid of a computer. A computer program written by Trescott (1975) and modified by Trescott and Larson (1976) and Torak (1982) was chosen for this study because (1) it simulates ground-water flow in three dimensions, (2) it has been successfully used to simulate ground-water flow in many aquifer systems, and (3) it has been successfully modified to incorporate the effects of inelastic compaction of fine-grained sediments in an aquifer system near Houston, Tex. (Meyer and Carr, 1979). The three-dimensional ground-water-flow equation the program solves simultaneously can be written as follows (Trescott, 1975, eq. 3):

$$Ss\frac{\partial h}{\partial t} + w(x, y, z, t) = \frac{\partial}{\partial x}\left(Kxx\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(Kyy\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(Kzz\frac{\partial h}{\partial z}\right) \quad (1)$$

where

h = hydraulic head, in feet;
Ss = specific storage, in feet^{-1};
w = volumetric flux of recharge/discharge per unit volume, in seconds^{-1};
t = time, in seconds;
Kxx,Kyy = hydraulic conductivity in the principal horizontal directions, in feet per second;
Kzz = hydraulic conductivity in the vertical direction, in feet per second, and
x,y,z = cartesian coordinates.

To solve the three-dimensional ground-water-flow equation, Trescott's program replaces the continuous derivatives in the flow equation with finite-difference approximations at a point or node. An example of a group of nodes used in the finite-difference approximation is shown in figure 10. Surrounding each node is a block with dimensions *x*, *y*, and *z* in which the hydraulic properties are assumed to be uniform. The result is *N* number of unknown head values at *N* nodes, which results in *N* number of equations, where *N* is the number of blocks that represent the aquifer system.

In Trescott's program, the time derivative $\frac{\partial h}{\partial t}$ is approximated by the backward-difference technique (Remson and others, 1971, p. 78). The approximation for each node may be given as

$$\frac{\partial h}{\partial t} = \frac{(h_i - h_o)}{\Delta t}, \quad (2)$$

where

h_o = the hydraulic head in a node at the beginning of a time step, in feet;
h_i = the hydraulic head in a node at the end of a time step (unknown), in feet; and
Δt = the time-step interval, in seconds.

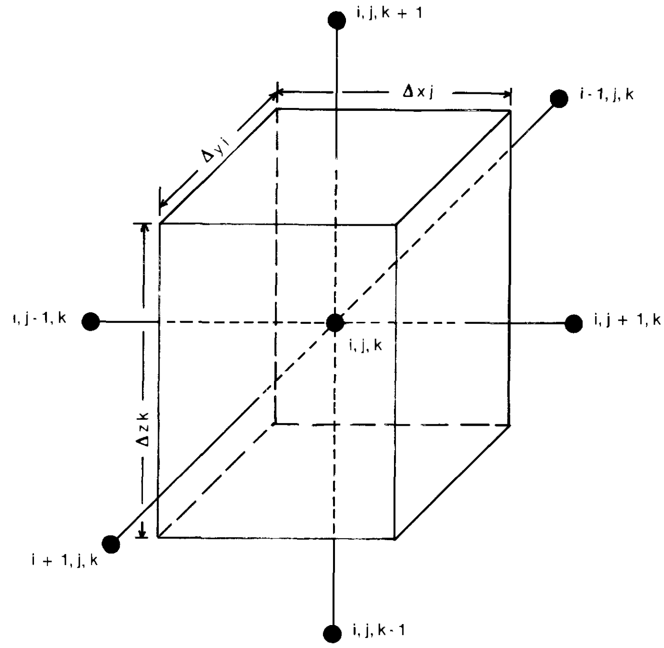
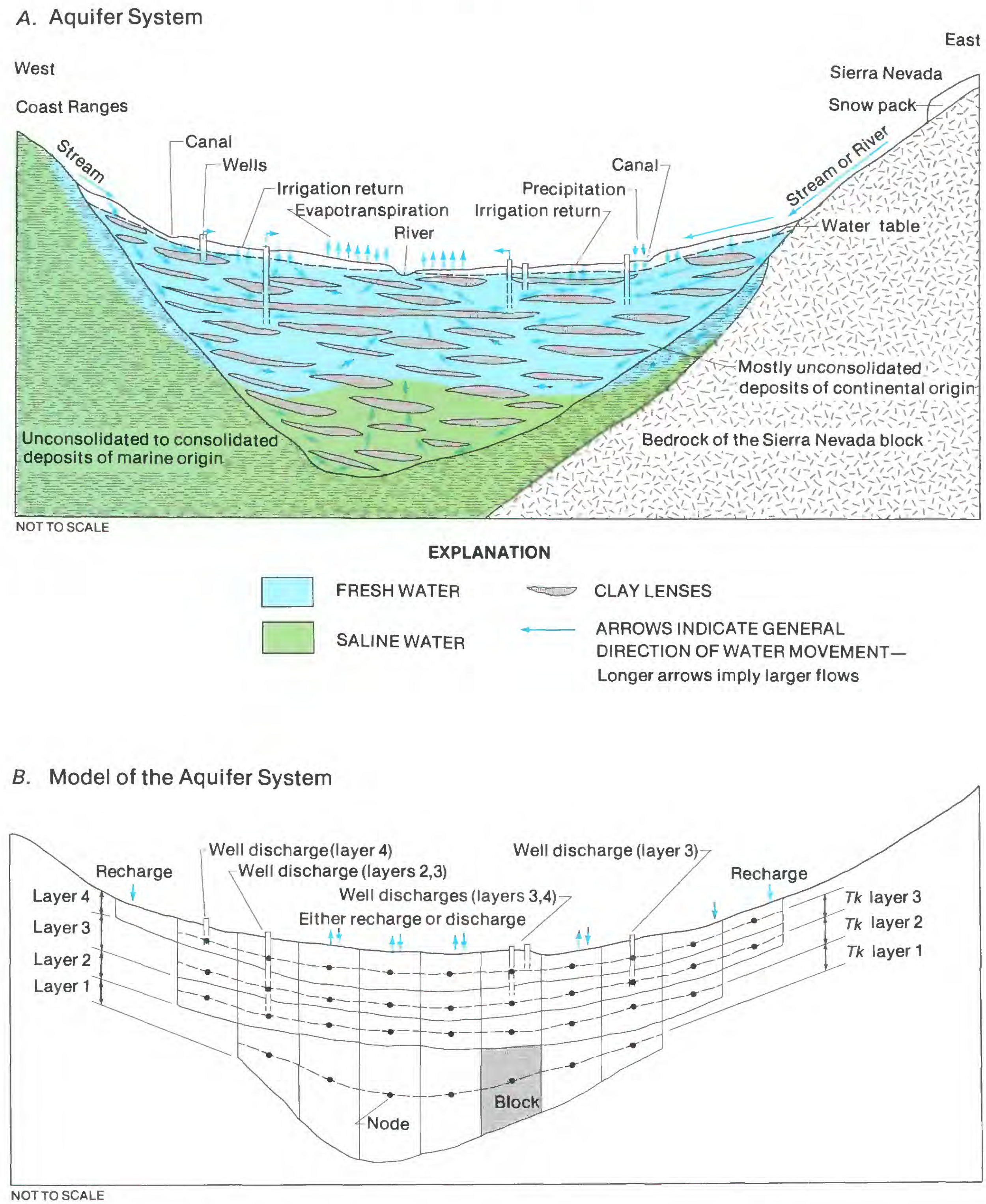


FIGURE 10.—Node array for finite-difference formulation showing model block associated with node i, j, k. (From Bennett and others, 1982.)

The program solves the unknown head for each time step using the strongly implicit procedure (Trescott, 1975, p. 11). This is done by iterating through the finite-difference equations for each node until the head change between the previous iteration and the current iteration is less than a specified amount for all nodes. Once this criterion is met, the program advances to a new time-step interval and the process of computing head values at each node is repeated. Both the ground-water-flow equation and the numerical technique are discussed in detail in Trescott (1975) and Trescott and Larson (1976). In the following paragraphs, the basic concepts and structure of the model are described.

GENERAL CONCEPTS AND FEATURES OF THE MODEL

In general, ground water moves from the margins of the valley toward the center and, since development, to major pumping centers. A simplified section (fig. 11*A*) shows the general patterns of recharge, discharge, and ground-water flow at present (1983) in the Central Valley aquifer system. The computer model can simulate many elements of the real aquifer system, as shown in figure 11*B*, including recharge from precipitation, streams, and irrigation return flow, and discharge as evapotranspiration, to streams as baseflow, and to wells as pumpage. The aquifer system is heterogeneous and consists of many discontinuous beds of clay, silt, sand, and gravel. The model simulates the heterogeneity in the aquifer system by (1) varying the aquifer proper-



Vertical leakance (*Tk*) values between layers are calculated by dividing the harmonic mean of the vertical hydraulic conductivity of the aquifer materials by the thickness between nodes. *Tk* layers 2 and 3 may be increased by wells that are screened in both layers 2 and 3 and less frequently in layers 3 and 4.
Discharge from all wells in the block simulated at the node.

FIGURE 11.—Conceptualization of (*A*) the aquifer system and (*B*) a model of the aquifer system.

ties from block to block and (2) averaging values to represent the aggregate of the heterogeneity within each block.

Dividing the Aquifer System Into Finite-Difference Blocks

The aquifer system was divided into blocks by superimposing a grid over a map of the study area and orienting it such that a minimum number of the blocks were outside the study area. A uniform planimetric grid spacing of 6 mi by 6 mi was used in the study (fig. 12). The vertical dimensions of the blocks vary and are incorporated into several terms that quantify the aquifer properties. For example, the horizontal transmissivity term for each node equals the product of the thickness of the block and the average horizontal hydraulic conductivity of the sediments. Similarly, the leakance (*Tk*) term, which affects vertical flow between layers, equals the equivalent vertical hydraulic conductivity divided by the thickness between nodes (one-half of each adjacent block thickness).

The valley was also subdivided by grouping model blocks into areas and subareas for analysis (see fig. 27). In the San Joaquin Valley, subarea boundaries approximate the ground-water-management boundaries outlined by the California Department of Water Resources (1980).

Four model layers were used to simulate the three-dimensional flow in the Central Valley aquifer system. The lowest model layer (layer 1 in fig. 11*B*) consists of the continental deposits below the depth penetrated by any production wells in the area. Most of the pumpage comes from layers 3 and 4. The division between the water table (layer 4) and the lower pumped zone (layer 3) was determined on the basis of the following criteria:

1. In areas where there was a large amount of well-construction data, the division between the shallow and the deep zones (model layers 3 and 4) was based on the vertical zonation of perforation intervals. A depth near which the majority of wells had no perforation was chosen as the boundary between the two zones.

2. In most of the area where the E-clay, which includes the Corcoran Clay Member of the Tulare Formation (Croft, 1972, p. 18), has been mapped, the division made by the criteria coincided with the depth above the E-clay. In the Westside subarea, the division based on criterion 1 was above the Corcoran Clay Member. The E-clay underlies more than half of the San Joaquin Valley (Croft, 1972, pl. 4).

3. In the remaining areas, the division was interpolated and extrapolated from adjacent areas.

Layer 2 extends to the depth of the deepest wells in the area. In model blocks where the wells are not as deep as they are in the adjacent general area, layer 3 extends to the deepest wells in the block. This layer definition reduces the effect of well leakage between nonadjacent layers (model layers 2 and 4) and allows for a simple adjustment of the *Tk* term between adjacent layers to account for well leakage during transient analyses (Bennett and others, 1982, p. 338).

Transmissivities were assumed constant in all model layers, including the uppermost layer, which incorporated the water table. Commonly, the transmissivity of the uppermost layer is allowed to vary depending on the saturated thickness in the layer, which can change during a simulation period owing to pumping or recharge. However, unless the changes in the water table are large compared with the thickness of the uppermost model layer, the change in the transmissivity is small and assigning a constant value makes little difference. In simulating the Central Valley aquifer from 1961 to 1977, the water table in a few model nodes in the uppermost layer changed about 60 ft but the initial saturated thickness was more than 500 ft. The maximum error in assuming a constant transmissivity was 12 percent, which is within the limits of this large-scale study.

Boundaries

The modeled aquifer system is surrounded by impermeable (no flow) boundaries except at Suisun Bay (fig. 12). Generally, the boundaries along the west side of the valley and beneath the aquifer system represent less permeable marine deposits; along the east side, the boundary is represented by less permeable igneous or metamorphic rocks. At the south end of the Central Valley, the boundary of the modeled aquifer system is the White Wolf fault, which acts as a barrier to flow (Wood and Dale, 1964). At the north end, the boundary is the Red Bluff arch, which is a series of low-lying hills consisting of northeast-trending anticlines and synclines. The series of hills acts as a barrier to ground-water flow (California Department of Water Resources, 1978, p. 39). In addition, both the Sutter Buttes and the Kettleman Hills within the valley restrict ground-water flow and were assumed virtually impermeable (Page, 1986, fig. 2 and p. C19).

Along the three model blocks that coincide with the discharge point (Suisun Bay) of the San Joaquin and Sacramento Rivers (fig. 12), constant hydraulic heads were specified in all model runs in the uppermost model layer (layer 4 in fig. 11*B*). During steady-state (predevelopment) simulations, the hydraulic head in the entire model layer 4 was held constant to aid in estimating recharge and discharge.

SIMULATION OF LAND SUBSIDENCE

The computer program of Trescott (1975) was modified to account for the release of water from the inelastic (irreversible) compaction of clay beds in the aquifer system. In general, the ratio of subsidence to head decline in an aquifer system, which is related to the irreversible compaction of the clayey beds, is small until after the head

FIGURE 12.—Model grid and boundaries.

declines below the preconsolidation head (a critical head); then the ratio of subsidence to head decline increases to a constant (Holzer, 1981). Water released from storage during the interval of head decline when the aquifer head is still above the preconsolidation stress comes mostly from expansion of the water and elastic compression of the aquifer materials (referred to in this report as "elastic storage"). Water released from storage in the interval of head declines when the aquifer head is below the preconsolidation stress comes mostly from the compaction of the clayey beds (referred to in this report as "inelastic storage"). Riley (1969) compared the compaction measured in an extensometer with head declines in an area in southeastern San Joaquin Valley and noted that the relation between compaction and head declines changed during large annual head fluctuations. Compaction was small and recoverable (elastic) during the initial part of the seasonal head decline. However, when heads declined below a certain altitude, which also declined each year, compaction per unit head decline increased and compaction became mostly irreversible (inelastic). Riley (1969) interpreted the head where the change in the rate of compaction to unit head decline occurred to be the new man-induced preconsolidation stress.

When pumping of ground water ceases, as in the example of seasonal pumping for irrigation in the Central Valley, head recovers and, in general, the compaction of the clayey beds ceases. The amount of water that can be stored in the aquifer system during the recovery period by elastic storage is much less than the amount released by inelastic compaction. If the head declines again, because of pumping of ground water, compaction of the clayey beds will not recur until the head in these beds again decreases below the preconsolidation stress, providing that residual compaction from the previous drawdown phase has been completed (Poland and Davis, 1969, p. 263). The amount of water released from storage during the same interval of head decline that occurred during the first drawdown period (assuming the head recovered to the initial level) is much less for the second drawdown period. This concept is illustrated in figure 13*A*. Poland (1961, p. B54) estimated that as little as 10 percent of the water released during the first drawdown period in which the clayey beds were compacted would be released by elastic compression of the clayey beds in a subsequent recovery of head to the initial level and a head decline over the same interval, again assuming that residual compaction from the previous drawdown period was largely complete.

In a real compacting system, the mechanics of subsidence are not as simple as shown in figure 13*A*. For example, at the Pixley well-field site (23S/25E-16N), about 3 mi south of Pixley, compaction was approximately 3 ft for the period 1959–71 yet the long-term head decline was negligible (fig. 13*B*). Helm (1975) related this to the continued compaction in the middle of the thicker clayey beds because of the time needed for pressure head changes to reach the middle of these beds. The cyclic nature of the compaction curve is produced by the seasonal periods of drawdown; during each seasonal drawdown period, the middle zones of the clayey beds were equilibrated to a new, lower head before the head in the more permeable zones of the aquifer recovered. Helm (1978, p. 195) estimated that the time for nonrecoverable compaction to be complete, assuming that the head was lowered instantaneously a specified amount and remained constant, was 5 years for the Pixley site. At six other sites in the San Joaquin Valley, it ranged from 40 to 1,350 years.

The modification used in the Central Valley flow model differs from the method used by Meyer and Carr (1979) in a study near Houston, Tex. In the Central Valley flow model, values of lowest critical heads (hydraulic head at which inelastic compaction of the clay beds begins) and inelastic storage are read into the computer program for each block in model layers 2 and 3. These layers were the intervals where compaction of the clayey beds was most prevalent in the aquifer system. Meyer and Carr (1979), in their analysis, assumed that the initial critical heads were 80 ft below the initial hydraulic heads (predevelopment or steady-state heads) and a single multiplier was used to change the storage value from elastic (recoverable) to inelastic (nonrecoverable). However, in this study the calibration period (1961–77) began when subsidence in the aquifer system had been occurring for many years. Therefore, the approach used in this study allowed for an inelastic storage to be simulated in the first time step when the starting head was below the critical head. The approach also allowed for varying inelastic-storage values from block to block because of differences in the percentage of fine-grained (clayey) beds.

The modification in the computer program allows for the compacting clayey beds within a model layer, in an individual block, at the start of a time step to respond with either an elastic- or an inelastic-storage value depending on whether or not the hydraulic head is below the lowest previous critical head. If the initial hydraulic head (starting water level) is above the initial critical head, the elastic-storage value is used until the hydraulic head falls below the critical head (see fig. 13*A*). When this happens, the elastic-storage value changes to an inelastic-storage value, associated with inelastic compaction, at the beginning of the next time step. The inelastic-storage value is used until the hydraulic head begins to recover; then the inelastic-storage value returns to the elastic-storage value, again at the beginning of the next time step, and the hydraulic head at which recovery started is recorded as a new critical head. When the hydraulic head falls below the new critical head, the elastic-storage value is changed to an inelastic-storage value and the cycle repeats itself. Subsidence is computed only if the head declines below the critical head. It is calculated by multiplying the drop in head below the critical head by the inelastic-storage

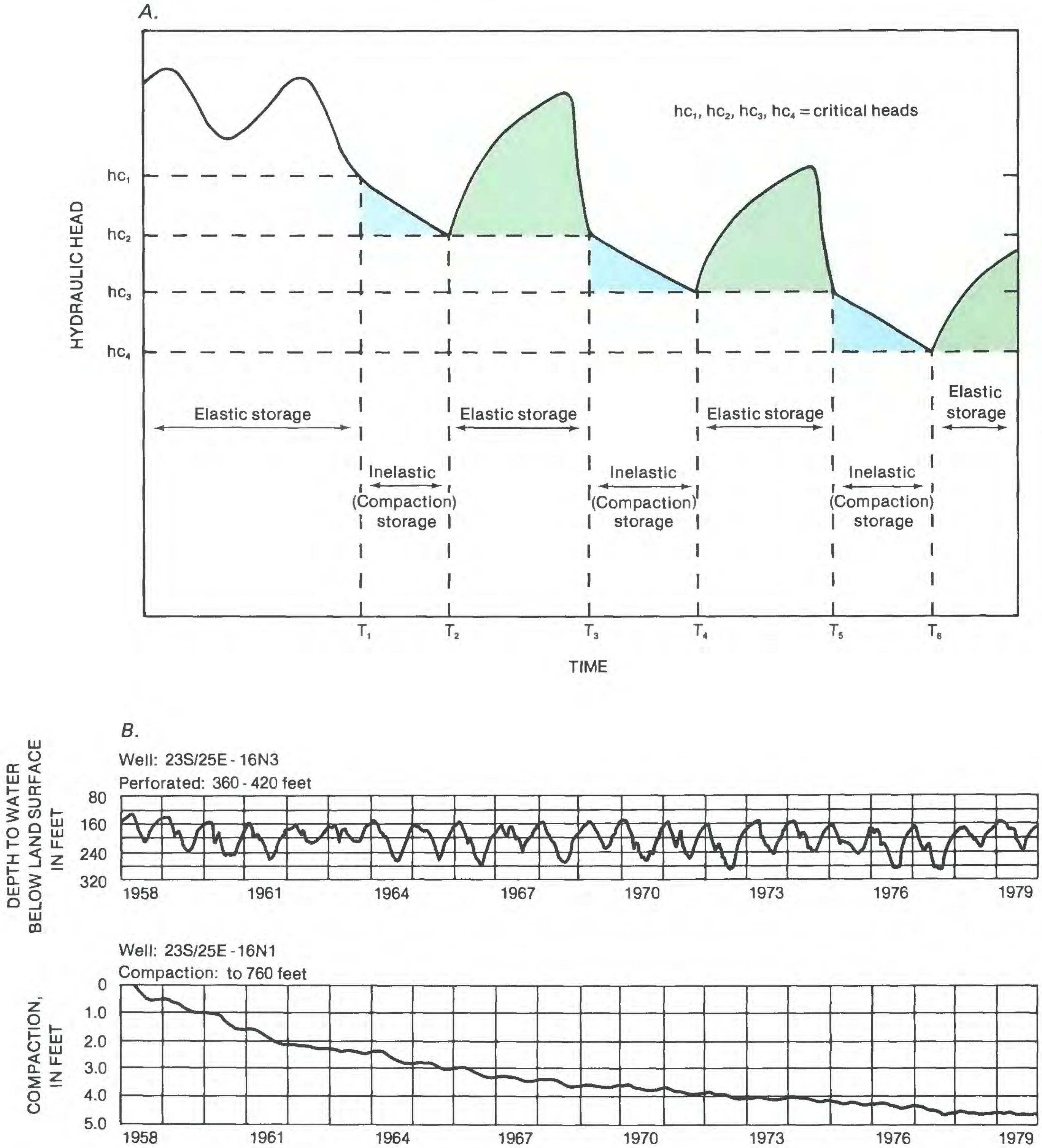


FIGURE 13.—(*A*) Relation of storage coefficient to the hydraulic head in a compacting interval of the aquifer system, and (*B*) hydrograph and compaction record for a well in the Tulare Basin. Location shown in figure 2.

coefficient. This value is calculated at the end of a time step and is accumulated throughout the simulation.

The modification has a few drawbacks. First, the change in head in an aquifer system actually propogates slowly through the included clayey beds in the vertical direction because of the low vertical hydraulic conductivity and the large inelastic specific storage of the clayey beds. This causes a gradual rather than an abrupt release of water from inelastic storage. In the simulations, however, all of the water is released from inelastic storage within the time step in which the head change occurs. Therefore, the time lag between stress change and compaction is not adequately simulated. This error is mostly canceled when looking at periods of several years or more. Second, the inelastic-storage term is assumed constant even though laboratory consolidometer tests of small clay samples indicate that the amount of water released from inelastic storage is a function of the applied stress. In addition, the vertical hydraulic conductivity of the compacting clayey beds, in theory, decreases as the beds are compacted. However, on the basis of soil consolidation theory, Helm (1977) was able to simulate the total compaction with reasonable results at seven sites in the San Joaquin Valley for periods of decades using constant values for aquifer properties.

In the computer program, the change from an elastic-storage value to an inelastic-storage value or vice versa was done at the beginning of each time step even though the change actually occurred during the previous time step. This means that unless small time steps are used in the simulation, the change from one storage value to another could lag greatly, thus causing errors in the simulation. A better technique would be to have the storage values change while iterating through the finite-difference equations within the time step. However, attempts to do this caused instability in the program and the difference in computed head values between iterations did not converge to an acceptable value.

ESTIMATES OF RECHARGE AND DISCHARGE

Methods used to estimate the initial values of recharge, discharge, and hydraulic properties of the aquifer system used in the simulations were selected on the basis of two criteria: (1) a method should be as independent as possible of the other methods being used in order to avoid situations in which an error or a wrong assumption would carry through the analysis, and (2) a method should be applicable throughout the valley so that if there is a bias error, at least the relative differences between one area and another would be apparent. These criteria eliminate some methods of estimation. However, the benefits of maintaining independence and consistency in a regional aquifer analysis were judged more important than being able to use all available methods.

Recharge and discharge can be considered at various scales of detail. The scale chosen is important because the hydraulic effect on some unit volume of the aquifer equals the difference between recharge and discharge in that unit. When a larger unit of the aquifer is considered, more canceling effects occur and, consequently, the variation of net recharge/discharge per unit area is smaller. Consideration of this principle requires that care be taken when comparing values of recharge/discharge. Because this is a regional analysis, the geographic units chosen (model blocks) were designed with a 6-mi^2-grid spacing. Equal values of recharge and discharge within the same model block are ignored because their net effect on flow to or from adjacent nodes or deeper layers is zero. The total recharge minus the total discharge into or out of a particular model block of the aquifer system is termed "net recharge/discharge."

Surface-water bodies, such as rivers and lakes, can be recharging the aquifer system or receiving discharge from the aquifer system depending on head difference between the surface body and the aquifer at a particular location and time. Precipitation can recharge the aquifer directly through the soil. Irrigated agricultural land usually recharges the aquifer system by irrigation return flow but can receive discharge from the aquifer under particular conditions. Wells usually discharge water but can be used for recharge, although this is uncommon in the Central Valley.

The only component of net recharge/discharge that can be measured directly is pumpage. Because net recharge/discharge is a sum of components, there are many ways to categorize the components by type and in time or space. The result is that there appear to be many ways to calculate the components (Wilson and others, 1980). However, most of these methods can be classified as one of, or a combination of, the following four types.

1. *Proportional.*—The proportional method assumes that a constant proportion of the inflow term becomes ground-water recharge. The inflows are measured or estimated and the proportions are compared with or taken from values calculated from the results of other methods, such as the water-budget method. In evaluating recharge from irrigation return flow, this proportion is equal to 1 minus the irrigation efficiency minus the proportion of irrigation water that becomes surface runoff.

2. *Rate-time.*—The rate-time method is also called the infiltration-duration method. It uses the equation

$$Qr = iAt, \qquad (3)$$

where

Qr = recharge volume in the specified time period, in acre-feet;

i = infiltration rate, in feet per year;

A = wetted area for infiltration, in acres; and

t = time duration of infiltration during the time period, in years.

The infiltration rate (i), if measured, is measured for a short time and in a small area, and involves a small measured water budget (such as a stream-seepage measurement or a percolation test). This rate must be extrapolated in time and space, which is difficult owing to its high variability and its poor relation to other conditions.

3. *Ground-water flow.* — The ground-water-flow method assumes that the flow across a plane, as calculated by Darcy's law, is equal to the net recharge upgradient from that plane. This calculation is made by analytical techniques or simulation-model calibrations. This also assumes that the flow system is in equilibrium (steady-state condition) and that the aquifer properties are estimated correctly. This would be a poor method to use for input to a flow model because it violates the principle of independence.

4. *Water budget.*—The water-budget method is based on the continuity equation:

$$\Sigma \text{ Inflows} - \Sigma \text{ Outflows} + \Delta \text{ Storage} = 0 \qquad (4)$$

The terms in this basic equation have been divided by many investigators in various ways. Net recharge/discharge is a component of one of the terms. It is assumed that all the significant components of each of these terms, except net recharge/discharge, can be measured or estimated. The equation is then used to solve for the dependent variable, net recharge/discharge, which is sometimes referred to as a residual quantity. In this type of equation, in which the dependent variable, net recharge/discharge, is equal to the difference of the independent terms, the random error in the dependent variable will be large if the difference between the independent terms is small relative to the size of the terms themselves.

There are also various ways to extrapolate the results of the methods described above to other locations or other time periods. These include other types of regression models that relate net recharge/discharge to flow, storage, or conveyance properties of water sources.

The water-budget method was the principal method used in this study because budgets could be designed to minimize the random error by adhering to the following guidelines:

1. Categorizing components so that recharge was relatively large compared with the other terms in the equation.

2. Choosing budget-unit boundaries at points where:
a. reasonably accurate data for flows across boundaries were available;
b. the number of significant flow components was minimal;
c. boundaries are compatible with other flow components such that water is not missed or counted twice; and
d. the geographic units for which average flow components are calculated are similar in size to the nodal spacing for the ground-water model.

Recharge and discharge values were estimated for the 17-year period 1961–77 by several types of water budgets. This period was chosen because recent data were available and because it includes a variety of dry and wet conditions as well as changes in water development. These stresses on the ground-water system aid in understanding the flow system because they require a more rigorous test of the simulation and the concepts upon which it is based. The estimates of the various components of recharge and discharge are given in appendix A.

The model does not automatically adjust certain components of recharge and discharge, as might be desired for head-dependant functions such as river leakage or evapotranspiration. By regression analysis, the authors found that the dominant factors affecting recharge and discharge rates in the aquifer system are the amount of surface-water flow, land use, and canal systems; these factors affect net recharge/discharge more than the head change in the aquifer. Therefore, the authors did not use the head-dependent function for net recharge/discharge in the model.

STREAMFLOW

Streamflow losses (ground-water recharge) and streamflow gains (ground-water discharge) were estimated by Mullen and Nady (1985) using the water-budget method. This was done for all major streams, for each reach bounded by gages, according to the following equation:

$$\text{Loss} = Q_{ups} + Q_{in} - ET - D_{iv} - Q_{dns} \qquad (5)$$

where (all in acre-feet per year):

Q_{ups} = flow at the upstream gage;

Q_{in} = inflow from tributaries or drains;

ET = evapotranspiration from the channel and riparian vegetation;

D_{iv} = diversions for irrigation; and

Q_{dns} = flow at the downstream gage.

Generally, all of these terms except ET are measured quantities, except where part of the record has been estimated. Evapotranspiration cannot be measured from riparian vegetation with any degree of accuracy.

Evapotranspiration from streams and riparian vegetation was not estimated because of uncertainty about the width of the channel and adjacent land with riparian vegetation and uncertainly about the evapotranspiration rate. Therefore, the stream-loss values estimated for the simulation model include evapotranspiration from the stream surface and from riparian vegetation. This error was considered in the calibration process, which is described later. The stream-loss values also include some unmeasured accretions (gains) from surface drains and unmeasured diversions for irrigation. In the Sacramento River, unmeasured gains from small creeks are of significant size. This causes an underestimate of stream losses and a corresponding overestimate of ungaged runoff infiltration. The Tule River also has unmeasured diversions that are significant. This causes an overestimate of stream losses and a corresponding underestimate of irrigation return flow.

The results of the stream-water budgets for 69 reaches of 20 major streams are summarized by Mullen and Nady (1985) and in table 1. The total length of the gaged reaches of major stream channels (accounting for 30.1 of the 31.7 million acre-ft/yr mean inflow) in the valley is about 1,200 mi. Average annual rates of exchange in the different reaches ranged from a gain of 13,400 to a loss of 23,800 acre-ft/yr per mile of channel. The sum of gaining reaches was 1,300, and the sum of losing reaches was 1,650 acre-ft/yr. These values were prorated and summed for each model block on the basis of the proportion of the length of the reach in the model block.

The minor streams that are gaged account for 7 percent (2.1 million acre-ft/yr) of the valley's inflow. Other minor streams that are not gaged account for less than 1 percent of the total inflow (Nady and Larragueta, 1983b). Ungaged flow was estimated by a multiple-regression analysis based on 60 gaged small streams. Most of the flow of the ungaged minor streams is applied on fields as artificial recharge.

PRECIPITATION

Ground-water recharge by precipitation occurs when precipitation is greater than the potential evapotranspiration and when the soil-moisture storage capacity is full. In general, precipitation exceeds potential evapotranspiration in the winter while the reverse is true in the summer; thus, most of the ground-water recharge from precipitation occurs during the winter and spring months. The method of estimating ground-water recharge from precipitation is described below.

Estimates of monthly soil-moisture budgets for the 50-year period 1922 through 1971 were computed for native vegetation by the California Department of Water Resources and the U.S. Bureau of Reclamation (John Renning, U.S. Bureau of Reclamation, written commun., 1979). They assumed 2-, 3-, and 4-ft rooting depths for the Sacramento Valley, Delta, and San Joaquin Valley areas, respectively, and a moisture-holding capacity of 1.5 in per foot of root depth to determine soil-moisture storage capacity. The monthly precipitation that exceeds monthly potential evapotranspiration is added to soil-moisture storage until the capacity is filled. Excess precipitation for any month is accumulated with the excess precipitation from previous months of that year and becomes a recharge value for the ground-water system. The soil-moisture storage is carried over into the summer, when it is depleted because the potential evapotranspiration exceeds precipitation. Linear regressions for the three areas were computed, relating excess precipitation to annual precipitation. The results are shown in table 2. Total precipitation on the valley floor averages about 12.4 million acre-ft/yr. Excess precipitation, which averages 1.5 million acre-ft/yr, includes ground-water recharge and surface runoff. The surface runoff is not added in any other water-budget term, so it is counted here even though it may actually become recharge downgradient in the valley. Total annual precipitation for each model block was estimated on the basis of mean annual precipitation (fig. 4) and measured ratios of annual to mean annual precipitation for each year during the period 1961–77.

IRRIGATION

Recharge and discharge resulting from irrigation is very important in understanding the aquifer system in the Central Valley because 57 percent of the total area of 20,000 mi^2 is irrigated. During 1961–77, water use for irrigation averaged about 22 million acre-ft/yr.

To determine net recharge/discharge from irrigated areas and unlined canals, a water budget was designed to examine the artificial components (such as canal losses and irrigation return flow) of the hydrologic cycle, which have greater values than the natural components because of extensive agricultural development. A major component in many areal water budgets is evapotranspiration. Estimation of evapotranspiration is difficult and subject to large errors. However, evaluation of the artificial components of the cycle allows the use of evapotranspiration values from irrigated agriculture, where the environment is much more uniform. The relatively uniform agricultural evapotranspiration contributes less variation and uncertainty to the water-budget analysis.

The spatial boundaries chosen for a water budget of irrigated lands are land surface at the top and the depth of crop roots at the bottom and, horizontally, the model block boundaries or the boundaries of geographic units of similar size whose data could be translated to model blocks by an areal proportion.

The water budget is defined as follows:

$$\overset{Inflow}{(SW + GW)} - \overset{Outflow}{(ETAW + GWR_4)} \pm \Delta SMS = 0, \quad (6)$$

where

- SW = surface inflow, measured at the diversion point to an area, minus surface outflow, if any, from that area;
- GW = pumped ground water;
- $ETAW$ = evapotranspiration of applied water;
- GWR_4 = recharge to the top layer (layer 4); and
- ΔSMS = change in soil moisture storage in time (using 1-year intervals, ΔSMS is assumed to be zero).

This calculation includes recharge from irrigated areas with recharge from unlined distribution canals; this approach has several advantages in addition to making it possible to consider one less term. A regional scale analysis does not require detailed separation of hydrologic features. Flow measurements of smaller, unlined distribution canals (such as ditchtender records) usually are approximate and may contain significant errors. This equation also makes GWR_4 as large as possible compared with the other terms, and this tends to minimize the effects of errors in the smaller terms.

Removing ΔSMS from equation 6 and solving for GWR_4,

$$GWR_4 = SW + GW - ETAW. \quad (7)$$

Separating GW into layers of origin, layer 4 (top) and layer 3 (deeper),

$$GWR_4 = SW + GW_4 + GW_3 - ETAW. \quad (8)$$

For a water-table aquifer, assuming the time lag for recharge is less than the periods of interest for modeling, the net recharge between the upper land surface and the water table is the desired result. This assumption was tested by checking response-time lags in water-table well hydrographs; it appears to be valid for simulation periods of 6 months to 1 year for much of the valley. The net recharge/discharge to the water table (net R/D_4) is then

$$\text{Net}R/D_4 = GWR_4 - GW_4. \quad (9)$$

Substituting equation 8 into equation 9 gives

$$\text{Net}R/D_4 = (SW + GW_4 + GW_3 - ETAW) - GW_4. \quad (10)$$

GW_4 cancels out, yielding

$$\text{Net}R/D_4 = SW - ETAW + GW_3. \quad (11)$$

The net recharge/discharge (net R/D_3) for the lower pumped zone (model layer 3) is

$$\text{Net}R/D_3 = -GW_3. \quad (12)$$

Equations 11 and 12 indicate that pumpage from the lower zone (layer 3) can be represented in the water budgets as a transfer of water to the water table (layer 4). Adding these two equations together shows that where the layer definition can be ignored, the composite net flow (net F) is

$$\text{Net}F = \text{Net}R/D_3 + \text{Net}R/D_4 = SW - ETAW. \quad (13)$$

Equation 13 has the advantage of having only one component that needs to be estimated because net surface inflow (SW) is measured.

Ideally, all components should be calculated for identical areas. However, the most accurate land-use and surface-water data are not collected or summarized for areas that have coincidental boundaries. Therefore, it was necessary to apportion the data values among model blocks on the basis of the area in that model block.

Surface-water-delivery data for the San Joaquin Valley and southern Delta areas were collected as irrigation district totals and prorated to the model blocks in each district. The evenness of distribution within a district varies from one district to another, but the distribution was compared in the Turlock Irrigation District against more detailed records of deliveries. In that district, which is large and has a large supply of surface water, the assumption of uniform distribution was adequate for the water years tested, 1962 and 1970.

In the Sacramento Valley, surface-water-delivery data are often misleading. Because of the abundance of water, much of the water delivered drains off one field to another field or to another irrigation district downslope. There is very little detailed data for drain flows. Therefore, it is possible to count water delivered to crops more than once. The most detailed surface-water-use data available are estimated from land use and unit applied-water values (Bloyd, 1978, p. 120). Another source of error in these data is the practice of determining from aerial photographs whether the fields are irrigated by surface water or ground water. Many fields are equipped for both types of irrigation, so it is difficult to determine which is used primarily. To make adjustments for these errors, water budgets for subareas 12 to 15 (fig. 27) were developed.

From these subareas, the ratios of net surface water used to total delivered average 77, 47, 57, and 83 percent, respectively. These ratios were used to adjust downward the total surface-water delivery presented by

TABLE 1.—*Summary of major stream losses and gains*

[Totals may not agree because of rounding]

Stream name	Reach	Upstream gage	Reach length (mi)	1961-77 mean: Inflow (1000 acre-ft/yr)	1961-77 mean: Diversion (1000 acre-ft/yr)	1961-77 mean: Losses (negative shows gains) (1000 acre-ft/yr)	1961-77 mean: Standard deviation of loss (1000 acre-ft/yr)	Unit loss (1000 (acre-ft/yr)/mi)
Kern River	1	Below Isabella Dam	19.9	646.0	0.0	12.7	14	0.6
	2	Near Democrat Springs	23.8	636.3	0.0	-24.0	28	-1.0
	3	Near Bakersfield	20.7	678.1	427	67.3	89	3.3
			64.4		427	56.1		0.9
Tule River	1	Below Success Dam	11.9	141.8	69.1	18.2	8.7	1.5
	2	Below Porterville	2.7	54.5	0.0	20.1	21	7.5
	3	At Oettle Bridge	23.0	34.4	0.0	18.4	27	0.8
	4	Porter Slough at Porterville	5.9	15.1	2.0	6.4	9.3	1.1
	5	Porter Slough near Porterville	3.7	6.9	0.0	2.1	5.4	0.6
			47.2		71.1	65.3		1.4
Kaweah River	1	Below Terminus Dam	2.8	421.3	71.7	-11.8	6.4	-4.2
	2	Below McKays Point	4.5	215.2	82.5	-2.0	5.5	-0.5
	3	Below Peoples Ditch	9.5	158.1	117	20.2	4.9	2.1
	4	St. Johns below Mckays Point	27.1	202.3	90.1	46.7	29	1.7
			43.9		362	53.0		1.2
Kings River	1	Below Pine Flat Dam	21.9	1,707	956	-53.8	34	-2.5
	2	At Reedly Narrows	13.0	805.4	184	16.5	27	1.3
	3	Below Peoples Weir	16.9	605.4	263	65.9	15	3.9
	4	Below Lemoore Weir	5.4	276.5	90.2	10.1	4.5	1.9
	5	North Fork below Island Weir	5.3	176.2	17.0	4.9	7.1	0.9
	6	Fresno Slough below Crescent Weir	9.5	154.2	6.7	11.2	14	1.2
	7	Fresno Slough at Stinson Weir	18.1	136.3	3.2	4.6	12	0.3
	8	South Fork below Army Weir	37.6	86.7	59.0	4.5	11	0.1
			127.7		1,578	63.9		0.5
San Joaquin River	1	Below Friant Dam	64.9	2,697	2,250	165	30	2.5
	2	Near Mendota	20.7	283.4	164	6.9	16	0.3
	3	Chowchilla Bypass at Head	81.0	458.1	0.0	147	92	1.8
	4	Near Dos Palos	46.8	379.6	0.0	-42.2	35	-0.9
	5	Near Stevinson	7.3	510.1	0.0	157	530	21
	6	At Fremont Ford	7.0	1,124	1.6	157	260	22
	7	Near Newman	9.9	1,006	7.1	-44.8	38	-4.5
	8	At Crows Landing Bridge	9.5	1,590	63.1	-62.1	11	-6.5
	9	At Patterson Bridge	20.7	1,610	107	-44.2	130	-2.1
	10	At Maze Road Bridge	5.1	2,570	10.7	-47.1	92	-9.2
			272.9		2,600	392		1.4
Fresno River	1	Near Daulton	14.8	107.9	54.4	3.4	14	0.2
	2	At Madera	8.2	51.8	0.0	9.7	8.9	1.2
			23.0		54.4	13.1		0.6
Chowchilla River	1	Below Buchanan Dam	13.0	164.0	0.0	4.9	17	0.4
Merced River	1	Below Merced Falls	7.3	867.0	534	-0.5	18	-0.1
	2	Below Snelling	18.7	320.0	31.7	-60.2	20	-3.2
	3	Near Cressey	23.6	362.1	17.6	-43.9	10	-1.9
			49.6		584	-104.6		-2.1
Tuolumne River	1	Below Lagrange Dam	20.7	1,488	898	-90.7	42	-4.4
	2	At Hickman Bridge	16.3	756.7	1.3	-33.9	32	-2.1
	3	At Modesto	13.0	790.7	6.5	-44.1	47	-3.4
			50.0		906	-168.7		-3.4

TABLE 1.—*Summary of major stream losses and gains*—Continued
[Totals may not agree because of rounding]

Stream name	Reach	Upstream gage	Reach length (mi)	1961-77 mean Inflow (1000 acre-ft/yr)	Diversion	Losses (negative shows gains)	Standard deviation of loss	Unit loss (1000 (acre-ft/yr)/mi)
Stanislaus River	1	At Goodwin Dam	11.0	1,054	519	-40.7	33	-3.7
	2	At Orange Blossom Bridge	13.6	575.7	1.5	-3.2	37	-0.2
	3	At Riverbank	16.7	585.9	3.6	-67.2	38	-4.0
	4	At Ripon	6.7	658.2	3.6	-2.6	38	-0.4
			48.0		528	-113.6		-2.4
Calaveras River	1	Below New Hogan Dam	6.8	139.2	0.0	-1.5	1.5	-0.2
	2	At Jenny Lind	11.1	249.0	2.7	13.7	18	1.2
	3	At Bellota	16.8	30.1	2.8	17.7	7.1	1.1
			34.7		5.5	29.9		0.9
Mokelumne River	1	Below Comanche Dam	24.3	499.8	118.1	48.0	17	2.0
Comsumnes River	1	At Michigan Bar	25.5	346.4	9.8	2.5	17	0.1
American River	1	At Fair Oaks	16.0	2614	34.6	382	140	24
Yuba River	1	Below Englebright Dam	17.8	1848	188	-49.0	71	-2.8
Feather River	1	At Oroville	15.6	4,310	582	-10.9	57	-0.7
	2	Near Gridley	21.7	3,550	42.0	-178	120	-8.2
	3	At Yuba City	5.0	5,391	0.7	-3.9	130	-0.8
	4	Below Shanghai Bend	13.8	5,738	56.6	-186	220	-13
			56.1		681	-378		-6.7
Sacramento River[1]	1	Near Red Bluff	43.2	--	--	44.0	58	1.0
	2	Near Vina Bridge	17.0	--	--	-5.3	44	-.31
	3	At Hamilton City	18.7	--	--	22.0	56	1.2
	4	At Ord Ferry	15.0	--	--	-1.6	64	-.11
	5	Butte City	26.4	--	--	1.5	54	.06
	6	At Colusa	26.5	--	--	-30.3	66	-1.1
	7	Below Wilkins Slough	28.9	--	--	-106	54	-3.7
	8	At Knights Landing	14.4	--	--	41.4	53	2.9
	9	At Verona	19.0	--	--	-16.6	74	-.87
			209.1			-51		-0.2
Stony Creek	1	Below Black Butte Dam	18.5	421.5	72.7	49.1	18	2.7
Cache Creek	1	At Rumsey	21.3	507.6	0.0	-0.2	19	0.0
	2	Near Capay	20.3	530.1	134	23.2	18	1.1
			41.6		134	23.0		0.6
Putah Creek	1	Near Winters	10.9	346.2	181	13.9	5.6	1.3
	2	Below Winters	4.3	111.3	0.1	1.3	5.6	0.3
	3	Above Davis	5.6	110.0	0.0	3.2	4.6	0.6
			20.8		181	18.3		0.9
		TOTAL --------------------	1,204.1		--	336		0.3

[1]Sacramento River flows are for the April to October (7 month) period; they are not annual figures. Inflow and diversions not listed.

Bloyd (1978, p. 130–132). Though reported by Bloyd as totals for townships (36 mi²), these data were available on a quarter-township basis (Phil Lorens, California Department of Water Resources, unpub. data, 1978). These values were available only for 1961 and 1970; therefore, they were adjusted for other years on the basis of a regression of known surface-water diversions for the other major streams (Mullen and Nady, 1985). This regression accounted for variation from wet years to dry years and for long-term trends. Evapotranspiration (ET) of applied water values was calculated on the basis of land-use data, which are summarized for 7.5-minute quadrangles of latitude and longitude, and unit ET values. Each quadrangle includes an area about 1.64 times the area of a model block. Details of estimating evapotranspiration of applied water are presented by Williamson (1982). Average unit ET of applied-water values was used, causing an overestimate in wet years and an underestimate in dry years. The variation in ET between dry and wet years, however, is small.

Pumpage data were collected for quarter township areas (0.25 times the area of the model block). Pumpage data were estimated from power consumption records and from pumping plant efficiency tests (Diamond and Williamson, 1983). Data for missing years were estimated by regression analysis. Estimates were not available for most of the Delta area. Pumpage in the Delta area was estimated for the simulations by the water-budget method assuming an irrigation efficiency of 55 percent, estimated values of crop needs (ET of applied water), and amounts of surface water diverted for irrigation.

There is some error in all the prorations. The effect of these errors is equivalent to a transfer of a volume of water from a model block to an adjacent model block. For this reason, constant additive adjustments to net recharge estimates were calibrated for each model block to account for balancing the errors in the volumes between adjacent model blocks.

The proportion of pumpage from the deeper zone in the aquifer was estimated by several methods. These methods assume that the proportion of flow from different zones into a well is proportional to the length of perforations in that zone. In the Central Valley, irrigation well casings are usually perforated throughout the lower two-thirds of the well depth. Construction data for more than 3,300 irrigation and public-supply wells were used to calculate the proportion of perforated intervals in each zone for each model block. To extend this analysis, discharge water temperature measurements for 35,000 pumping plant efficiency tests from about 13,000 wells were analyzed. Temperature data from 3,000 wells having construction information established a relation between temperature and perforated interval.

TABLE 2.—*Regression results—Excess precipitation* (PPT_{ex}) *as a function of annual precipitation* (PPT)

equation: $PPT_{ex} = m\ PPT + b$

Area	Slope(m)	Intercept(b)	R^2
Sacramento	0.64	-9.1	0.85
Delta	0.63	-7.3	0.79
San Joaquin	0.64	-6.2	0.64

This relation was used to approximate perforated intervals for each of the 13,000 wells. These approximate predicted perforated intervals were used to estimate the proportion of perforated intervals in each zone. These proportions were averaged with those previously determined using appropriate weighting factors. Where no data existed, the proportion was interpolated from adjacent areas. The effect of errors in estimating these proportions is discussed in the section "Changes in Recharge and Discharge."

ESTIMATES OF AQUIFER PROPERTIES

The methods used to estimate aquifer-system properties such as thickness, hydraulic conductivity, and storage are described in the following sections. The same principle of using consistent methods for the entire valley, as previously described, was applied. Some measures (such as the mean) of the estimates made are given in this section; others are given in the sections on predevelopment and postdevelopment ground-water flow. These estimates were adjusted during calibration of the model. The complete data set of final values after calibration is given in appendix B.

THICKNESS

Post-Eocene deposits of continental origin constitute the primary ground-water reservoir in the Central Valley. The thickness of these deposits (fig. 8) was estimated by R.W. Page (Page, 1974; U.S. Geological Survey, written commun., 1981) from interpretation of electric logs and from published reports. The thickness of these deposits averages about 2,400 ft and increases from north to south, with a maximum thickness of more than 9,000 ft near Bakersfield. However, the contact between continental deposits and the underlying marine deposits is not always certain because the two types of deposits interfinger in some places, particularly near the southern end of the valley. For example, de Laveaga (1952, p. 102) suggested that the continental deposits may be as much as 15,000 ft thick in places where 9,000 ft is shown in figure 8. Thus, the thickness of continental deposits in the Central Valley, particularly in the southern part, used in the analyses of the system may be less than what is actually present. Excluding the deeper continental deposits (which interfinger with marine deposits) probably does not greatly affect the analyses of ground-water flow in the Central Valley because the amount of flow in the deeper parts of the continental deposits is considered small.

HYDRAULIC CONDUCTIVITY

The hydraulic conductivity (K) of a saturated, porous medium is the volume of water it will transmit in a unit time, through a cross section of unit area, under a hydraulic gradient of a unit change in head through a unit length of flow (Lohman, 1972, p. 6). In this report, hydraulic conductivity is expressed in units of feet per day.

HORIZONTAL

Two sources of data were considered to estimate horizontal hydraulic conductivity (Kh) values—specific-capacity data from power company pump-efficiency tests and drillers' logs. Because pump-efficiency tests are not available for the entire valley, that source was used only to spot check the results of the other methods.

Drillers' logs contain descriptions of the formations drilled through in each depth interval. Each formation description was assigned to one of five categories of formations with similar properties described by Davis and others (1959, p. 202–206). The depth interval and the category was coded for each well log for computer tabulation. More than 10,000 well logs in the Sacramento Valley and more than 7,400 logs in the San Joaquin Valley were coded for the analysis.

Hydraulic conductivities were assigned to formation categories that were characterized by grain size using values determined by Johnson and others (1968), Morris and Johnson (1967), and the California Department of Water Resources (1966, p. 137). Although there is considerable variation in Kh values within a category, the method should still give a good indication of relative differences in Kh because of the large sample size. Table 3 shows the categories and their corresponding Kh values and specific yields which are discussed in the section on "Aquifer Storage."

An equivalent Kh value was computed for each segment of each well which corresponded to the appropriate model layer, by the following equation:

$$Kh_{eq} = \frac{\Sigma\,(b\ Kh)}{\Sigma\, b}\,, \qquad (14)$$

where

Kh_{eq} = equivalent horizontal hydraulic conductivity,

b = thickness of the interval reported on the drillers' log, and

Kh = horizontal hydraulic conductivity of the interval.

These equivalent Kh values for individual wells were averaged for each layer in each model block. Values for model blocks having no data were interpolated and extrapolated from nearby model blocks. The resulting Kh values for all of the model blocks have a mean of 25 ft/d and a standard deviation of 13 ft/d. The resulting Kh values were compared with values reported by other investigators. The comparison showed that estimates of Kh obtained in the above manner were not consistently larger or smaller than other estimates. It also showed that in 57 percent of the 244 model blocks that could be compared, the present estimates are within a ratio of 0.6 to 1.67 of the other estimates. Estimated values were also compared with values estimated from specific-

TABLE 3.—*Hydraulic conductivity and specific yield values used for aquifer materials for initial estimates*

[Hydraulic conductivities were reduced by a factor of 4 during model calibration]

Aquifer material	Hydraulic conductivity (ft/d)	Specific yield (percent)
Bedrock	0.0	0.0
Clay	.00053	3
Sandy clay	1.1	5
Fine sand	11	10
Sand and gravel	110	25

capacity data collected by utility companies in pump-efficiency tests. In two-thirds of the 251 model blocks that could be compared, the values from drillers' logs were larger than those estimated from specific capacity. Only 46 percent of the model blocks were within the ratios discussed on p. 27.

VERTICAL

The aquifer system is composed of many interbedded lenses of coarse- and fine-grained deposits in which the vertical hydraulic conductivity varies according to the type of deposit. Because it is impossible to model every lens in the aquifer system, an equivalent vertical hydraulic conductivity of the lenses in each model layer in each block was calculated by applying the principle of conductances in series as

$$Kz_{eq} = \frac{\Sigma b}{\frac{b_1}{Kz_1} + \frac{b_2}{Kz_2} + \ldots + \frac{b_n}{Kz_n}}, \tag{15}$$

where

Kz_{eq} = equivalent vertical hydraulic conductivity;
Σb = total thickness between the centers of two adjacent model layers;
b_1, b_2, b_n = thickness of individual lenses; and
Kz_1, Kz_2, Kz_n = vertical hydraulic conductivity of corresponding lenses in the aquifer system.

The lenses were categorized into coarse- and fine-grained deposits. This simplified equation 15 is as follows:

$$Kz_{eq} = \frac{\Sigma b}{\frac{\Sigma b_c}{Kz_c} + \frac{\Sigma b_f}{Kz_f}}, \tag{16}$$

where

$\Sigma b_c, \Sigma b_f$ = sum of the thicknesses of coarse and fine beds, respectively, and
Kz_c, Kz_f = vertical hydraulic conductivities of coarse and fine sediments, respectively.

In general, the vertical hydraulic conductivity of the fine-grained lenses is much less (by at least two orders of magnitude) than that of the coarse-grained lenses, and this causes the term $\Sigma b_c / Kz_c$ to be negligible. Thus, equation 16 can be simplified to

$$Kz_{eq} \cong \frac{\Sigma b \, Kz_f}{\Sigma b_f} \tag{17}$$

The ground-water-flow model used in this investigation incorporated the vertical hydraulic conductivity into the term known as leakance. Leakance (*Tk*) is defined by Lohman (1972, p. 30) as the ratio of *Kz* to the thickness of the confining beds. In an aquifer system composed of many interbedded lenses of coarse- and fine-grained deposits, an equivalent *Tk* can be computed as

$$Tk_{eq} = \frac{Kz_{eq}}{\Sigma b}. \tag{18}$$

where Tk_{eq} is equivalent leakance.

Substituting the right side of equation 17 for Kz_{eq} in equation 18 yields

$$Tk_{eq} \cong \frac{Kz_f}{\Sigma b_f}. \tag{19}$$

Thus, the flow between model layers is controlled by the vertical hydraulic conductivity of the fine-grained deposits divided by the thickness of the fine-grained deposits.

Tk values were calculated for each well using equation 19 on the basis of thicknesses of coarse- and fine-grained beds developed by Page (1986, p. 20) from 690 electric logs, selected at a density of one per quarter township (9 mi^2). The initial value of *Kz* used for fine-grained beds was 1×10^{-4} ft/d. These equivalent *Tk* values for individual wells were averaged for each model block.

Laboratory values of vertical hydraulic conductivity, given by Johnson and others (1968), were tested in the early phases of calibration, but they were not used because they represent point data rather than areal and depth integrated averages necessary in the regional model.

In some areas, many wells are perforated for long intervals across two adjacent model layers. Bennett and others (1982) discuss this problem, noting that where wells penetrate two adjacent layers, by using the Thiem equation, Tk_{eq} values for the wells can be calculated. All of the well Tk_{eq} values can be summed with the aquifer Tk_{eq} because the flows are parallel. Because of the large variation in values and the model's high sensitivity to Tk_{eq}, these values were substantially adjusted in the calibration process. This is further discussed in the section, "Changes in Ground-Water Flow."

AQUIFER STORAGE

The term "storage coefficient" is used to describe water that is released from or taken into storage. Theis (1938, p. 894) defined it as the volume of water (in cubic feet) released from storage in each column of the aquifer having a base 1 ft^2 and a height equal to the thickness of the aquifer when the water table or the piezometric surface is lowered 1 ft. The storage coefficient is equal to the specific storage times the thickness of the aquifer, where the specific storage of a saturated aquifer is the volume of water that a unit volume of aquifer releases from storage under a unit decline in hydraulic head. Jacob (1940) noted that water released from an elastic artesian aquifer was derived from three sources: (1) expansion of the water, (2) compression of the aquifer, and (3) compression of the adjacent and included clay beds. Poland (1961) assumed that the third source of water was caused by inelastic compaction of the adjacent and included clayey beds. Water is also released from the shallow part of the aquifer by gravity drainage when the water table is lowered (known as specific yield). However, the volume of water released by gravity drainage, or the aquifer's specific yield, is usually much greater than the volume released from the other sources. Thus, for the upper part of the aquifer system in the Central Valley, specific yield was used as the storage coefficient. Specific yield was estimated by the same method of weighted averages as described in the section on "Hydraulic Conductivity," except specific yield replaced horizontal hydraulic conductivity. The values used for each formation are given in table 3. The mean specific yield is 0.09 and the standard deviation is 0.03.

Jacob (1940) concluded that in an elastic artesian aquifer, the water released from compression of the adjacent and included clayey beds was the chief source of water released from storage in the aquifer. In the analyses of the Central Valley aquifer system, the system below the uppermost part was considered confined in the sense that the vertical permeabilities of the sediments are much lower than the horizontal permeabilities, a condition that restricts the vertical movement of water.

Jacob (1940), in defining the elastic-storage coefficient for an uncemented granular material, assumed that water stored in clayey beds was released instantly so as to avoid mathematical complications (although Jacob recognized there would be a time delay between the lowering of the head in the aquifer and the release of water from the clays because of their low permeability):

$$S = \gamma \Theta m \left(\frac{1}{Ew} + \frac{1}{\Theta Es} + \frac{c}{Ec} \right) , \qquad (20)$$

where

S = storage coefficient, dimensionless;

γ = specific weight of water (0.434 pound per square inch per foot);

Θ = porosity of the sediments, dimensionless;

m = thickness of the aquifer, in feet;

Ew = bulk modulus of elasticity of the water (3×10^5 pounds per square inch);

Es = bulk modulus of elasticity of the aquifer matrix, in pounds per square inch;

Ec = modulus of compression of clay beds, in pounds per square inch; and

c = a dimensionless quantity that depends largely on the thickness, configuration, and distribution of the clay beds.

Replacing the storage coefficient with specific storage (Ss), and rearranging terms, the equation can be

$$Ss = \frac{S}{m} = \frac{\gamma\Theta}{Ew} + \frac{\gamma}{Es} + \frac{\gamma\Theta c}{Ec} . \qquad (21)$$

The elastic specific storage (Ss_e) of the aquifer system is equal to

$$Ss_e = \frac{\gamma\Theta}{Ew} + \frac{\gamma}{Eas} , \qquad (22)$$

where Eas is the weighted average bulk modulus of elasticity of the aquifer system, in pounds per square inch.

Estimates of the elastic-storage term were calculated by adding the product of the thickness of coarse-grained deposits times its specific storage to the product of the thickness of the fine-grained deposits times its specific storage. Values of the elastic specific storage of the coarse- and fine-grained deposits were obtained from Poland (1961), Riley and McClelland (1972), and Helm (1978).

Poland (1961, p. B53) assumed that release of water from storage during short-term pumping tests was primarily caused by the expansion of water and the elastic compression of the coarse-grained part of the aquifer. He approximated the contribution of water derived from each of the two mechanisms for the aquifer system in the southwestern part of the San Joaquin Valley. In the calculations, he used an aquifer thickness of 700 ft and a storage coefficient of 0.001, which is the average of aquifer tests of wells for the area studied by McClelland (1962). Clayey interbeds were not included in Poland's calculations and they accounted for another 300 ft of the aquifer system. The estimated elastic specific storage value of the coarse-grained deposits in the aquifer system was 1.4×10^{-6} per foot, with about 40 percent contributed by the expansion of water and 60 percent contributed by the elastic compression of the aquifer matrix. Similarly, Riley and McClelland (1972, p. 77d) estimated the elastic specific storage of the more permeable layers (coarse-grained deposits) in the aquifer system near Fresno to be between 0.7×10^{-6} and 1×10^{-6} per foot. These results were based on several detailed aquifer tests.

In contrast, Helm (1978, p. 193) calculated an elastic specific storage value of the fine-grained (clayey) deposits at seven sites in the San Joaquin Valley. The values ranged from 2.0×10^{-6} to 7.5×10^{-6} per foot, with an average value of 4.5×10^{-6} per foot. Thus, on the basis of somewhat limited information, the range of elastic specific storage for the Central Valley aquifer system was estimated to be between 1×10^{-6} per foot for parts of the aquifer system that are all coarse-grained deposits to 4.5×10^{-6} per foot for parts of the system that are all fine-grained deposits. This results in an average elastic specific storage value of about 3×10^{-6} per foot where the deposits are one-half coarse grained and one-half fine grained.

Poland (1961) estimated that the volume of stored water released by the inelastic compaction of clayey beds in the highly compressible aquifer system was 50 times greater than the volume of water released by the elastic expansion of water and the elastic compression of the aquifer system. In the southwestern part of the San Joaquin Valley the ratio of subsidence to head decline ranged from 0.04 to 0.1. Poland concluded that land subsidence in areas of heavy ground-water pumpage was caused almost totally by "... the compaction of the clay, silty clay, and clayey silt beds, both by plastic deformation and mechanical rearrangements of grains, and to that extent is inelastic and permanent." However, water is not always released from the compaction of the clayey beds, but is dependent on the change in head in the aquifer system. The theory and mechanics of how the clayey beds in an aquifer system compact, causing land subsidence, is presented in detail by Lofgren (1968) and Poland and Davis (1969).

Estimates of inelastic (compaction) storage were calculated by (1) estimating the thickness of fine-grained beds in the aquifer system and (2) multiplying that value by the mean inelastic specific storage of 3×10^{-4} per foot. The mean inelastic specific storage value was calculated by Helm (1978, p. 193), who estimated an inelastic specific storage value at each of seven sites in the San Joaquin Valley, where the values ranged from 1.4×10^{-4} to 6.7×10^{-4} per foot. Another estimate of the inelastic specific storage was calculated from Poland (1961) to be about 2×10^{-4} per foot assuming a 300-ft-thick clayey section in the aquifer system and an inelastic storage coefficient calculated by Poland of 5×10^{-2}. This value is reasonably close to the mean value estimated by Helm (1978).

WATER-LEVEL ANALYSIS

Two major data bases of water-level measurements were accessed and analyzed to provide estimates of model-block-averaged water levels during the calibration period and also during predevelopment.

A statistical analysis of the data was chosen over the more traditional method of drawing contour maps for each time period of interest. Contour maps of water levels were available from the California Department of Water Resources but were used only for verification of the estimates because of the following limitations:

1. Water levels of the entire valley were not mapped, and only one depth zone was mapped in any area.
2. Temporal trends determined by using values interpolated from successive contour maps can be erroneous owing to the cumulative effect of variation of subjective input in compiling each map.
3. It was unclear which wells were used for the water-level mapping and what well construction information was available.
4. Confinement exists in areas where no extensive clay layers have been mapped, because numerous discontinuous clay layers collectively act as confining units. The absence or presence of clay layers was not considered in compiling the water-level maps.
5. Only a part of the data was used because of the time required to incorporate a large volume of available data.

The data base from the California Department of Water Resources was copied, edited, and analyzed; more than 460,000 ground-water-level measurements were available from more than 18,000 wells for the years between 1920 and 1979. Depth and (or) construction information was available for about 8,000 of the wells, which allowed assigning the wells to the depth zones in the model. About 32 percent were in the top (water table) zone, 6 percent were in the next two lower layers, 10 percent possibly spanned the top two layers, and 52 percent were of unknown depth. Most of the wells were measured biannually, though about 6 percent were measured at least monthly. Of the biannually measured wells, the autumn measurements were almost always taken in October. Most of the spring measurements were taken during March in the Sacramento Valley and Delta areas, during February in the upper San Joaquin Valley area, and during January in the Tulare area. These times of measurement cause a slight problem because the usual months of high and low water levels are February and August, respectively. The effect of the water level in spring is slight because the monthly change is small, but the effect in the autumn is substantial because the recovery of water levels is very rapid at the end of the pumping season. This condition occurs because water levels in the aquifer systems respond fastest immediately following a change in stress, with the rate of change decreasing with time. Therefore, a measurement taken early will more accurately reflect the seasonal maximum or minimum than one taken late, after the next change in stress occurs. Often by the time of the autumn measurement, more than half of the postseason recovery has taken place. More measurements are taken in spring (57 percent) than in autumn (43 percent). The Bureau of Reclamation has a ground-water-level data base that was used as a supplement. Many, but not all, of the 112,000 measurements in the Bureau of Reclamation file are duplicates of measurements found in the California Department of Water Resources file.

In order to use the large file of data, several steps were taken. First, depth and well-construction information was added for about 2,000 wells that had drillers' logs available. Then, the data were plotted by making computer-generated hydrographs for the period 1960–80 with all of the wells in a township plotted on the same page using different symbols. This allowed easy location of large errors and comparison of adjacent well hydrographs. Because well-construction information and depth zones were assigned to some of the wells, other wells could be seen to have similar responses and were coded to depth zones accordingly. They were assigned only if there was substantial evidence to indicate similarity.

The next step was to convert all of the records to seasonal values, whether the actual data were monthly or biannual. Means were calculated for each group of water-level measurements within the same year, season, and model block. These means were plotted on the same page with all of the depth zones of one model block. The hydrographs were compared with the California State Department of Water Resources contour map for specific times as a check for the spatial variation of water levels among blocks.

The data were also averaged by area and subarea to determine long-term trends. If a block contains rolling terrain, the average depth to water showed trends more consistently than the average altitude of water levels within the block, because some wells may be measured in one year and may not be measured in other years. The results are described in the sections, "Effects of Development" and "Change in Aquifer Storage."

SEQUENCE OF CALIBRATION OF THE MODEL

Calibration of a ground-water-flow model is achieved by adjusting the values of one or more aquifer properties or recharge/discharge such that the computer-simulated hydraulic heads match (within the limits of the investigation) the observed heads in the aquifer system. The normal sequence of calibration of most model studies is to first adjust values of aquifer properties (usually terms that incorporate vertical and horizontal hydraulic conductivity) assuming steady-state conditions (no head change with time) and to then adjust values of aquifer properties (usually the storage term) assuming transient conditions (changes in head with time). However, in the Central Valley, the system as a whole has been in a state of continual change since agricultural development began in the late 1800's. Few data are available for the natural recharge rates to and discharge rates from the ground-water system or for the distribution of hydraulic heads before agricultural development began. Thus, the computer model that numerically represents the Central Valley aquifer system was calibrated under transient conditions.

Transient simulations were run for the period spring 1961 to autumn 1977 because there were for this period (1) both natural variations in recharge and discharge to the system and changes in man's operation of the water system and (2) adequate data for the distribution of head in the aquifer system and for estimates of recharge from precipitation, streams, and applied irrigation water and discharge from evapotranspiration and pumpage. These data were compiled for water years (October 1 to September 30) and were allocated to 6-month (spring–autumn and autumn–spring) periods. All river recharge and discharge and precipitation recharge was assumed to occur in the autumn-spring period. Municipal pumpage was divided equally between the two 6-month periods. All of the agricultural pumpage was assumed to occur in the spring–autumn period. Analysis of well hydrographs indicates that irrigation return flow reaches the water table after about 6 months; therefore, recharge from irrigation was assumed to occur in the autumn–spring period. Because of a data-manipulation difficulty, this recharge was allocated to the winter season before the irrigation season instead of after.

Calibration of the model of the Central Valley aquifer system was done in three phases. In each phase, pumpage in the lower pumped zone (model layer 3) was held constant (the values were assumed correct), while one set of values (transmissivity, leakance, storage, or recharge) was adjusted at a time. Repeated adjustments were made to each of the sets of values. A discussion of each phase is presented in the following paragraphs.

In the first phase of model calibration, the simulation period 1961–76 was divided into two separate periods: spring 1961 to spring 1970 and spring 1970 to spring 1976. The rates of recharge and discharge during the 6-month period were summed and averaged for the particular period. These periods were selected because (1) in the west side of the San Joaquin Valley, hydraulic heads during the earlier period (1961–70) declined as much as 60 ft because of heavy pumpage and the land subsided as much as 8 ft, and (2) in the same area, hydraulic heads during the latter period (1970–76) recovered as much as 120 ft following deliveries of surface water from the California aqueduct. The modification of the computer program that automatically changed the storage term from elastic to inelastic depending on the head in the aquifer system was not used during the first phase of calibration. Instead, the storage term for blocks that correspond to areas actively subsiding were assigned an inelastic-storage value. The inelastic-storage value was estimated by dividing the amount of observed land subsidence in the model block by the observed head decline during the particular calibration period. An elastic-storage value was assigned to all other blocks that were outside the areas of active subsidence. The storage term was held constant throughout the first phase of calibration.

During early calibration of the model, it was obvious that the model-computed heads were more sensitive to the leakance (*Tk*) value than to any other value. Therefore, the sequence of calibration in the first phase was to uniformly adjust all vertical hydraulic conductivities (incorporated in the *Tk* values) and then, on the basis of a relation between observed and computed vertical head differences, to individually adjust the values of *Tk* for each block. The relation is expressed in the following equation:

$$Tk_{new} = Tk_{old}\, FAC\, \frac{\Delta HV_{mod}}{\Delta HV_{obs}}, \qquad (23)$$

where

ΔHV_{mod} = the computed difference between model layers 4 and 3 at the end of the pumping period;

ΔHV_{obs} = the observed vertical head difference between the water-table zone (model layer 4) and the lower pumped zone (model layer 3);

Tk_{new} = the adjusted leakance value;

Tk_{old} = the previous leakance value; and

FAC = 0.9 when the ratio of ΔHV_{mod} to ΔHV_{obs} is less than 1, and 1.1 when the ratio is greater than 1.

Second, horizontal hydraulic conductivities were adjusted uniformly throughout all layers to achieve the best fit of horizontal hydraulic gradients. At this point, it became obvious that the net recharge/discharge from streams was in error because simulated heads were either too high or too low at points that correlated with the stream values. Because no reasonable change in any other parameter could solve this problem, all net recharge/discharge values calculated from stream budgets were divided by 5. The best results in fitting horizontal head gradients were obtained when the initial estimates of horizontal hydraulic conductivity were reduced by a factor of 4.

Next, the amounts and distribution of recharge and discharge in the uppermost model layer (layer 4) were adjusted in blocks whose heads could not be matched by changing the other model values. Simple linear regression analysis showed that for a 1 ft change in head at the end of the 15-year simulation period, a 0.25 ft/yr change in net recharge/discharge in the top layer was required. The recharge and discharge adjustments were made for the two calibration periods and the differences in the adjustments between the two calibration periods were averaged at each block. The result was a reduction in the overall amount of recharge to the uppermost layer by 20 percent and, in places, a substantially different distribution of recharge and discharge. The result of the first phase of model calibration was a model that simulated the overall changes in head in the aquifer system from 1961 to 1970 and from 1970 to 1976.

In the second phase of model calibration, the two calibration periods remained the same but the computer program was modified to account for water released from compaction of the clay beds. The inelastic-(compaction-) storage term was then calibrated for the period 1961 to 1970, first by uniformly adjusting the inelastic-storage term throughout layers 2 and 3, and finally by adjusting individual values assigned to the blocks. Individual adjustments occurred mostly in the Westside subarea (see fig. 27). In addition to adjusting values of inelastic storage, minor adjustments were made for both horizontal and vertical hydraulic conductivity values, particularly where individual adjustments of inelastic storage were made to improve model results.

The third and final phase of model calibration was done while simulating 6-month periods from spring 1961 to spring 1976. The simulations included the modified version of the computer program that accounted for subsidence. These simulations were used to calibrate the elastic-storage term and to slightly readjust all other values in the model. In general, the adjusted elastic-specific-storage values were a factor of 2 times greater than the average initial estimate discussed in the section on "Aquifer Storage," except in the Westside subarea, where the adjusted values approximated the initial specific storage estimates. The results obtained from this calibration phase and the sensitivity of aquifer properties are discussed in following sections.

PREDEVELOPMENT GROUND-WATER FLOW

Water development for irrigation began in 1850 in the Central Valley. This irrigation development affected the ground-water system, which previously had been in hydrologic equilibrium (also referred to as a "steady-state condition" because there was no trend of changing aquifer storage with time). Consequently, most of the hydrologic data were collected after changes had already taken place in the system. However, some water-level measurements made by the State engineer's office before development were available, and they are a good indication of what ground-water conditions were like in those areas. Most of the water-level measurements used in the analysis of predevelopment ground-water flow were obtained for the periods 1905–07 in the San Joaquin Valley (Mendenhall and others, 1916, p. 15) and 1912–13 in the Sacramento Valley (Bryan, 1923, p. 18). Some earlier (late 1800's) information was obtained from Hall (1886). Some adjustments to the data from the early 1900's were required because effects of development were already occurring. Also, strong inferences about ground-water conditions can be made from other evidence, such as areas of marsh and swamp. Simulation of the predevelopment flow system using the available information has expanded the understanding of how the system operated.

WATER LEVELS AND FLOWS

The aquifer system in the Central Valley is a single and heterogeneous system in which flows and heads vary in all three dimensions. This type of system is difficult to understand and describe. To simplify the discussion, horizontal and vertical variations in flow and head are discussed separately, while attempting to show the relations. This is compatible with the description of the simulation because the model also considers horizontal- and vertical-flow components separately.

HORIZONTAL

Ground water moves from areas of recharge to areas of discharge, in the direction of decreasing hydraulic head. In the Central Valley, ground-water flow in the predevelopment system began as recharge in the low hills along the perimeter of the valley and in the upper reaches of streams and moved toward the topographically low areas in the center of the valley.

Under natural conditions, the water table roughly paralleled the land surface and the direction of ground-water flow was approximately coincidental with the slope of the land (fig. 14). Recharge occurred in high-altitude areas and discharge occurred in low-altitude areas where the water table was close to land surface.

The Central Valley has only one outlet for discharge of surface water and ground water from the Delta west to San Francisco Bay (fig. 12). Because this outlet is only about one-third of the way from the north end of the valley, the head gradient has to be steeper in the Sacramento Valley. Notice that the trough of lowest head in the San Joaquin Valley is to the west of the center (fig. 14*B*). This also coincides with the topography.

Much of the ground-water discharge from the southern part of the valley was to Tulare Lake and the area surrounding it (note the depression in fig. 14*B*). Because of the characteristics of the surface-water drainage system and the variability of surface runoff, the volume and therefore the level of the lake varied tremendously. From records obtained between 1853 and 1908 (Grunsky, 1898a; Mendenhall, 1908), the water level of the lake varied more than 40 ft, from an altitude of 220 ft during the wet years 1862–68 to 180 ft (altitude at bottom of lake) in 1906 when the lake was dry. This natural fluctuation would have significantly affected ground-water levels and flows. Also, it was reported that deep and very shallow ground water was fresh, while a zone of intermediate depth was alkaline. This is an additional indication that although the system was probably in equilibrium during a long period, there were short-term variations from that state.

VERTICAL

Under natural conditions, recharge and discharge occur at the water table. If the lower part of an aquifer is to contribute to the horizontal flow between recharge areas and discharge areas, there must be vertical flow downward in the recharge areas and upward in the discharge areas (figs. 15 and 16). Downward head gradients are often not discovered because they occur in recharge areas where deep wells are not commonly drilled. Upward head gradients along the trough of the valley, indicated by large areas of wells that flowed without pumps prior to development, were documented as early as the 1880's (Hall, 1889). Figure 17 shows the area of flowing wells documented by Hall and the areas outlined as artesian in the San Joaquin Valley in the early 1900's (Mendenhall and others, 1916).

Most investigators have conceptualized the ground-water system in the Sacramento Valley as a single water-table aquifer (Bloyd, 1978, p. 102) and in the San Joaquin Valley as two aquifers, a water-table aquifer and a confined aquifer below the Corcoran Clay Member of the Tulare Formation. The Corcoran Clay Member is a very notable marker bed in the valley and has been geologically correlated from well logs over much of the San Joaquin Valley (Page, 1986, plate 4). Its lateral boundary, where known, roughly coincides with the area of predevelopment flowing wells (fig. 17). In many areas, water levels in wells completed above and below the Corcoran Clay Member are substantially different. These factors are the basis for the assumption that other fine-grained beds in the valley are much less significant than the Corcoran Clay Member in their effect on confinement. However, there is substantial evidence to suggest that this assumption is not valid.

As stated earlier, there are numerous fine-grained beds throughout the Central Valley. Though they individually have small lateral extent, the aggregate thickness of these beds is as much as several thousand feet (Page, 1986, p. C15), whereas the Corcoran Clay Member thickness ranges from zero to 160 ft with a mean thickness of 55 ft. Water-level differences with depth have been measured in many areas, such as the northwestern Sacramento Valley and the southeastern San Joaquin Valley, where the Corcoran Clay Member has not been mapped. Also, in several areas on the west side of the San Joaquin Valley, the Corcoran Clay

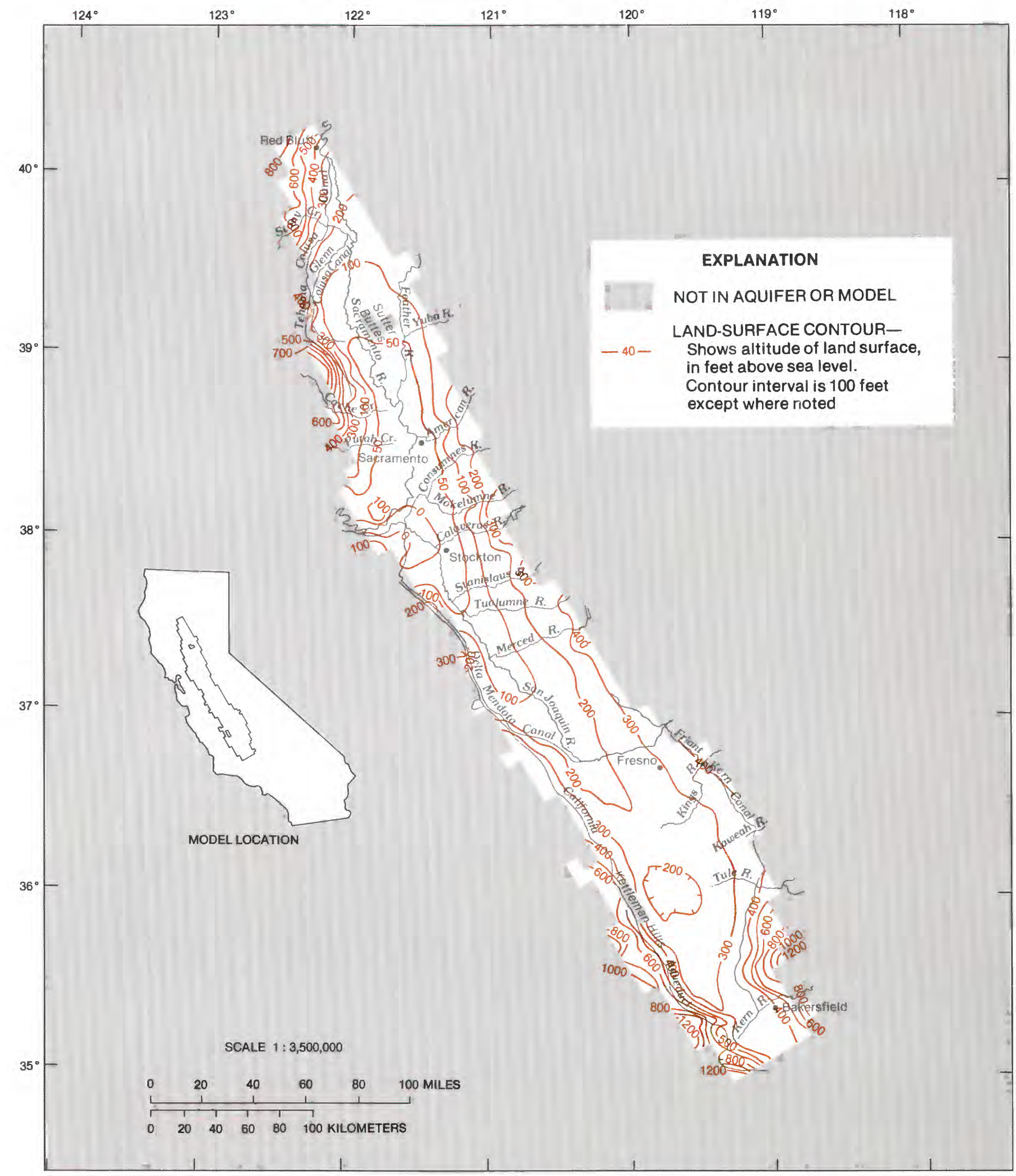


FIGURE 14A.—Land-surface altitude, circa 1957.

Member has had numerous wells drilled through it and the wells commonly are perforated immediately above and below the clay layer. This condition has allowed almost free flow through the well casings and gravel packs, with the result that the piezometric head has been equalized in the vicinity of the clay. Despite this head

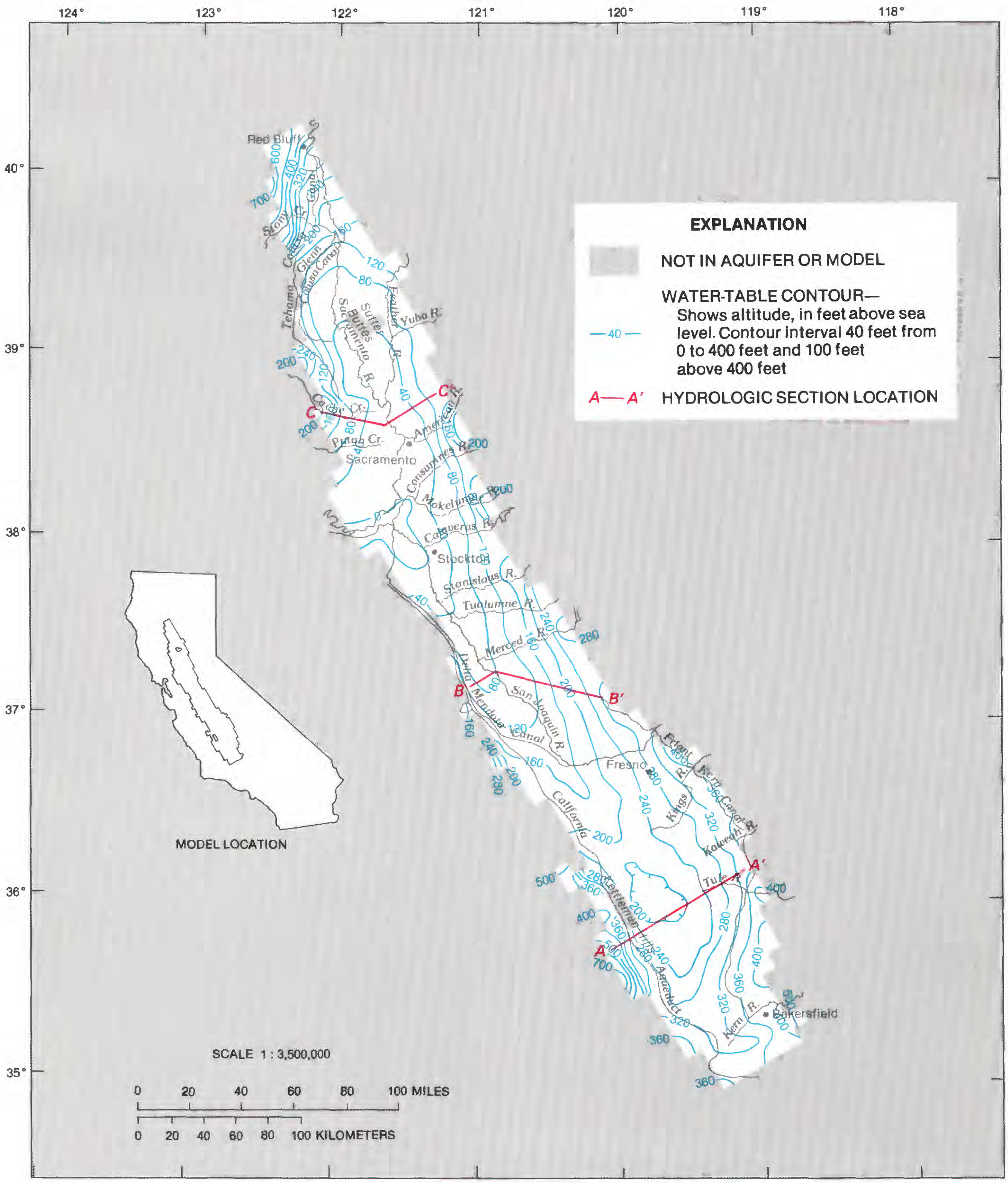


FIGURE 14*B*.—Predevelopment water-table altitude and hydrologic section locations.

equalization through wells adjacent to the Corcoran Clay Member, head differences as much as 400 ft have occurred between very shallow wells (less than 250 ft deep) and deeper wells. These head differences are the result of numerous clay stringers between the shallow wells and the deeper wells which, when combined, have a low enough vertical permeability to restrict the vertical movement of water.

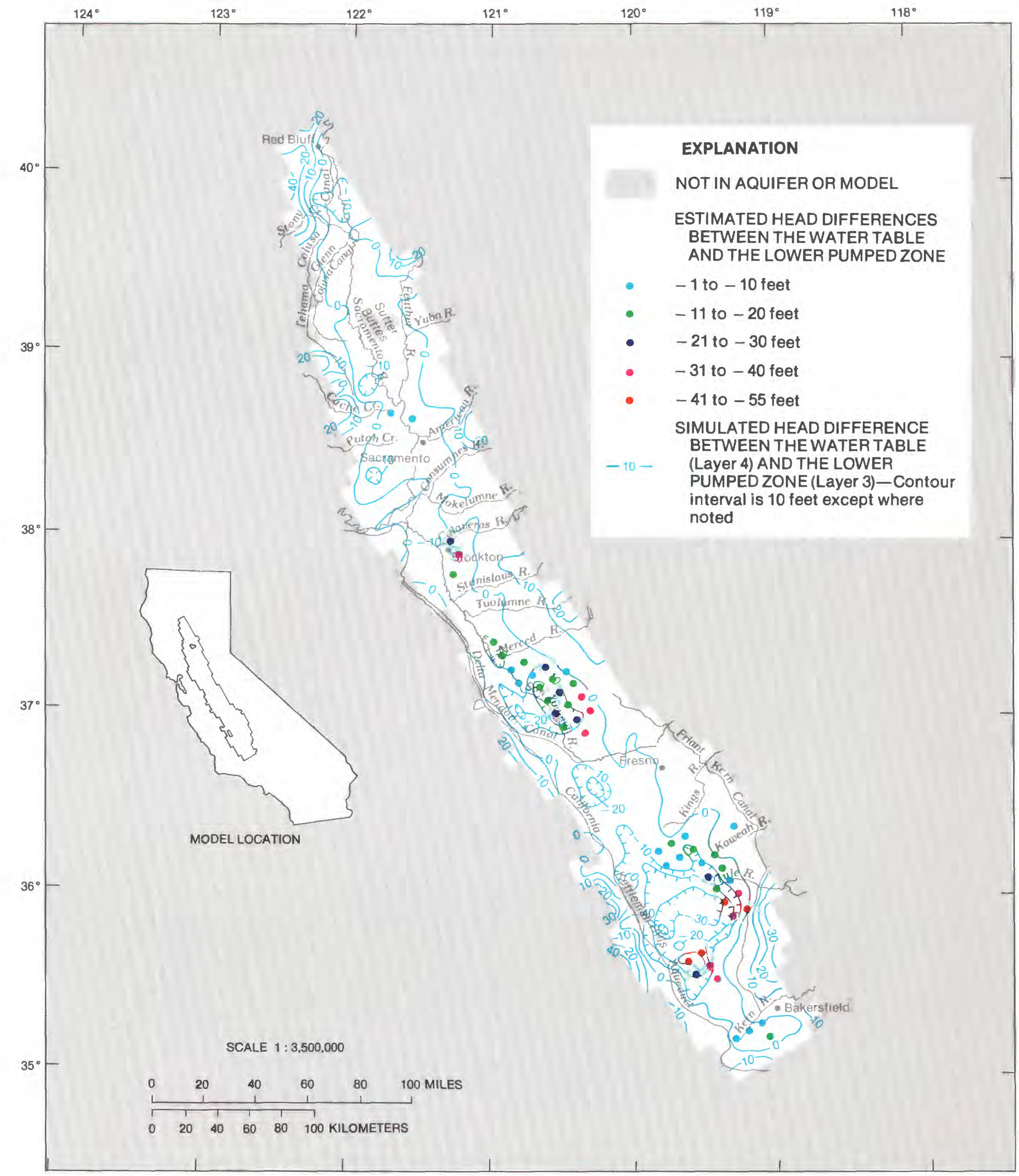


FIGURE 15.—Simulated and estimated head difference between the water table (layer 4) and the lower pumped zone (layer 3) under predevelopment conditions.

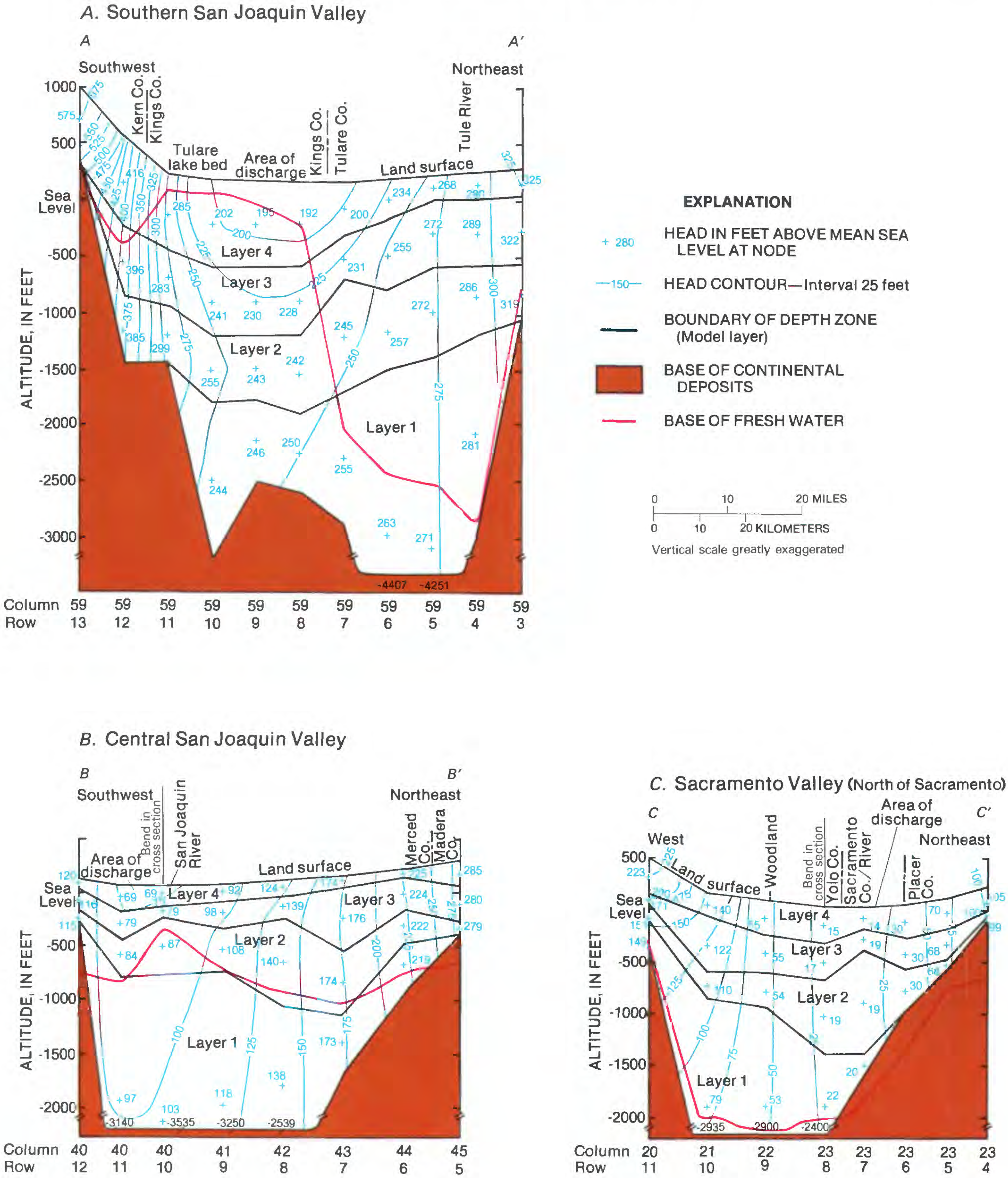


FIGURE 16.—Hydrologic sections across the Central Valley showing simulated predevelopment equipotential lines: (*A*) Southern San Joaquin Valley, (*B*) Central San Joaquin Valley, and (*C*) Sacramento Valley north of Sacramento. Locations of sections shown in figure 14*B*.

The amount of vertical flow and head gradient depends mainly on vertical hydraulic conductivity (Kz) and on the thickness of the aquifer system. The aquifer system in the Central Valley is composed of interbedded coarse- and fine-grained beds, with about 55 percent of the thickness composed of fine-grained beds (Page, 1986, fig. 35). This percentage varies little (standard deviation of about 8 percent) and is usually in the range from 40 to 70 percent. Therefore, under predevelopment conditions, significant vertical head gradients probably existed throughout the valley except where the flow was entirely horizontal or in local areas where sediments were predominantly coarse grained.

Predevelopment vertical head differences are difficult to estimate because they are very sensitive to ground-water development and there are few data for heads at depth before development occurred. Hall (1886) reported data on about 350 deep wells that had been drilled between 1858 and 1885. Most of these wells were flowing artesian wells ranging in diameter from 2 to 12 in and in depth to 1,200 ft (one was 2,160 ft). Only one had a measured static head (water level was reported as 11 ft above land surface), though most had a reported flowing head and flow rate. The flows ranged up to 1,100 gal/min. To convert the flowing head measurements to static head values, a form of the Thiem equation was used to compute drawdown:

$$\Delta h = Q \frac{ln\left(\frac{Ra}{Rw}\right)}{2\pi T}, \qquad (24)$$

where

Δh = static head minus flowing head in the well, in feet;

Q = discharge of the well, in cubic feet per second;

Ra = radius from the well where water level is static, in feet;

Rw = radius of the well, in feet; and

T = transmissivity of the aquifer penetrated by the well, in feet per second.

Several assumptions had to be made to apply the equation. The value chosen for Ra (2,100 ft) is somewhat arbitrary; however, changing it will not have a great effect on the result because the ratio of radii is in a logarithm term. The transmissivity chosen was equal to the depth of the well times the estimated hydraulic conductivity. The well radius used was 0.58 ft (7 in), an average for the reported wells. The estimated static head varied from nearly zero to more than 50 ft above the flowing-head measurement.

Vertical head differences were estimated by subtracting the static water levels in the deeper aquifers calculated from Hall's data from the estimates of the water-table altitudes reported by Bryan (1923) and by Mendenhall and others (1916). In areas with large lakes, the lake level was used for the water-table altitude. The vertical resistance to flow in the model (Tk) was adjusted where data were available so that the simulated head difference approximated the observed head difference between layers 3 and 4. Observed head differences between layers 3 and 4 ranged from zero to 40 ft; in the Tulare Lake area, the observed difference was 55 ft.

Ground-water development in the valley has caused the hydraulic head to decline at depths where water is partially confined; currently, artesian water rises above land surface in only a very few areas. This occurs in some areas of the central Sacramento Valley that have very little deep pumping; wells drilled by the U.S. Geological Survey (fig. 2) in 1979–80 near Zamora (12N/1E–34Q) (French and others, 1982) and Butte City (19N/1W–32G) (French, Page, Bertoldi, and Fogelman, 1983) with 2,500- and 1,500-ft depths, respectively, had water levels rising above land surface.

RECHARGE AND DISCHARGE

Natural recharge to the valley occurs from precipitation in excess of direct evapotranspiration and seepage through stream channels along their upper reaches. Most stream recharge occurs on the east side of the valley where large streams flow from the Sierra Nevada. The Coast Ranges on the west side are not as high and have much less precipitation and smaller drainage areas available to sustain streamflow. Mean annual inflow to the valley in stream channels is about 31.7 million acre-ft/yr. Ground-water discharge occurs mainly through evapotranspiration and discharge to streams where ground-water levels are near land surface. Ground-water outflow to Suisun Bay is negligible. Estimating predevelopment recharge and discharge is difficult because of the lack of data before the system changed substantially owing to water development. There is no evidence to indicate that the streamflow into the valley or precipitation have changed much since the 1800's, but ground-water flows have changed dramatically. Ground-water discharge also occurs to stream channels, generally in parts of their lower reaches, where the head in the aquifer is higher than the water level in the channel. Stream channels gain about 0.3 million acre-ft/yr from the aquifer system and lose about 0.5 million acre-ft/yr, according to estimates made by stream water budgets calculated during 1961–77, with adjustments made during model calibration.

Deep percolation of precipitation on the valley floor, upgradient from the swampy areas and lakes, is a significant source of recharge in the wetter areas and during the wetter years. Average annual precipitation on the valley floor is about 12.4 million acre-ft/yr (11.6 in/yr). The potential evapotranspiration (calculated as the evapotranspiration of irrigated grass) is about 49 in/yr. This value varies little across the valley or from year to year (California Department of Water Resources, 1975), but is concentrated in the summer. Precipitation

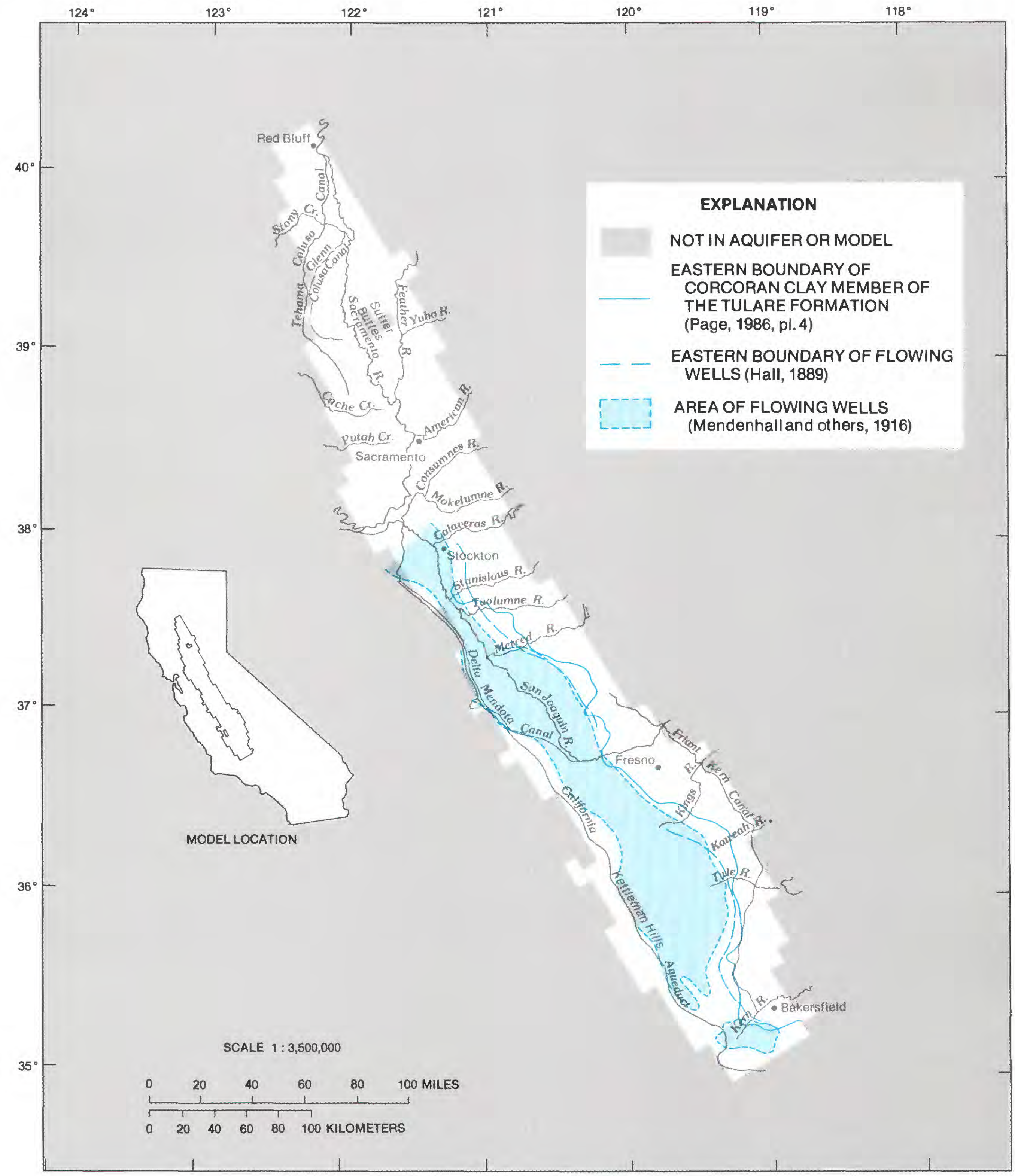


FIGURE 17.—Areas of predevelopment flowing wells and the Corcoran Clay Member of the Tulare Formation.

occurs mainly in the winter (fig. 5). Therefore, in the winter, precipitation exceeds evapotranspiration so that excess is stored in the soil until all of its storage capacity is filled. Additional precipitation will either run off or percolate into the aquifer. In the summer, evapotranspiration in excess of precipitation is withdrawn from soil storage until it is depleted. Monthly soil-moisture budgets (see section on "Precipitation") indicate that no recharge occurs until annual precipitation exceeds about 12 in. This occurs in most years on the north and east sides of the valleys but in only extremely wet years in the southwest part, where average annual precipitation is less than 6 in. The long-term average rate of precipitation in excess of direct evapotranspiration for the Central Valley is about 1.5 million acre-ft/yr.

Evapotranspiration directly from ground water can be roughly estimated assuming none occurs from ground water where the depth to water is greater than 10 ft, and also assuming a linear increase in evapotranspiration to its potential of about 4.1 ft/yr where the water table is at land surface. Using these assumptions and the estimated predevelopment depth to water, there could have been about 13 million acre-ft/yr of evapotranspiration directly from ground water in the central part of the valley where the water table was close to the land surface (about 8,000 mi^2 or 62 percent of the valley area). Only about 40 percent of that amount (5 million acre-ft/yr) could have been supplied from local direct precipitation because the central part of the valley has less precipitation than the north and east sides. Most of the remainder must have been supplied from local ground-water flow out of the stream channels to adjacent areas because that volume could not have been supplied from regional ground-water flow as will be demonstrated below.

Natural recharge and discharge to the regional ground-water flow system can be calculated by the model. These calculations were made using the aquifer properties calibrated during the 1961–77 period, with adjustments for changes because wells were not present during the predevelopment period. The head in the uppermost model layer (layer 4) was held constant at the best estimates of the predevelopment water table altitude (fig. 14*B*). Simulations with these constant heads produced an estimate of the net amount of water that recharged and discharged the deep regional aquifer system from the uppermost model layer (layer 4), as shown in figure 18. These values represent total recharge/discharge to the deep regional aquifer system in the Central Valley. Thus, the values in figure 18 representing the amount of water that recharged and discharged the deep regional aquifer system in the Central Valley are smaller than the total recharge and discharge estimates for the whole aquifer system, including water moving through the upper model layer (layer 4), described previously. Total simulated recharge and discharge to the deep regional ground-water system were slightly more than 0.2 million acre-ft/yr each. In general, more recharge than discharge occurs along the margins of the valley and more discharge than recharge occurs in the low-lying central parts. In the San Joaquin Valley, the areas of discharge generally corresponded to areas of flowing wells (compare figs. 17 and 18). A schematic summary of the predevelopment water budget for the valley is shown in figure 19, showing the relationship between recharge and discharge and ground water on a local and regional scale.

EXTENT OF FRESHWATER

The post-Eocene continental deposits constitute the primary fresh ground-water reservoir in the Central Valley. Freshwater in the Central Valley is defined as water that has a specific conductance of less than 3,000 micromhos per centimeter at 25°C (Olmsted and Davis, 1961, p. 134; Berkstresser, 1973; Page, 1973). This corresponds to about 2,000 mg/L of dissolved solids. Beneath the body of freshwater is saline water. In general, the salinity of the water beneath the base of freshwater increases gradually with depth, at least in the San Joaquin Valley; however, at certain locations it may increase rapidly (Page, 1973).

The vertical extent of freshwater varies greatly throughout the valley (fig. 20). The greatest thickness of freshwater occurs near Bakersfield, where it exceeds 4,500 ft. In the San Joaquin Valley, the occurrence of freshwater is not related to any specific formation, but rather is generally within the post-Eocene continental deposits. The base of freshwater in the San Joaquin Valley in places reflects the underlying structure of the thick Tertiary basin, particularly near Bakersfield. It also reflects the anticlinal structures of some of the oil and gas fields in that valley (Page, 1973). In the Sacramento Valley, the base of freshwater is generally coincident with the base of continental and volcanic deposits and rarely reflects deeper structures such as faults and gas reservoirs (Berkstresser, 1973). The shallow body of saline water west of Sutter Buttes (fig. 20) is found in marine deposits, while the shallow body of saline water south of Sutter Buttes may be a body of evaporation residue. Another possible cause of this shallow body of saline water was thought to be from upward migration of marine connate waters through defective, abandoned, or improperly constructed deep wells (Olmsted and Davis, 1961, p. 136). However, after investigation, G.H. Davis (oral commun., 1983) could not find evidence of more than one or two deep wells ever drilled in this area.

Within the freshwater body are zones of water that approach and exceed the specific conductance limit that defines freshwater. These zones of saline water are surrounded by freshwater and may represent evaporation

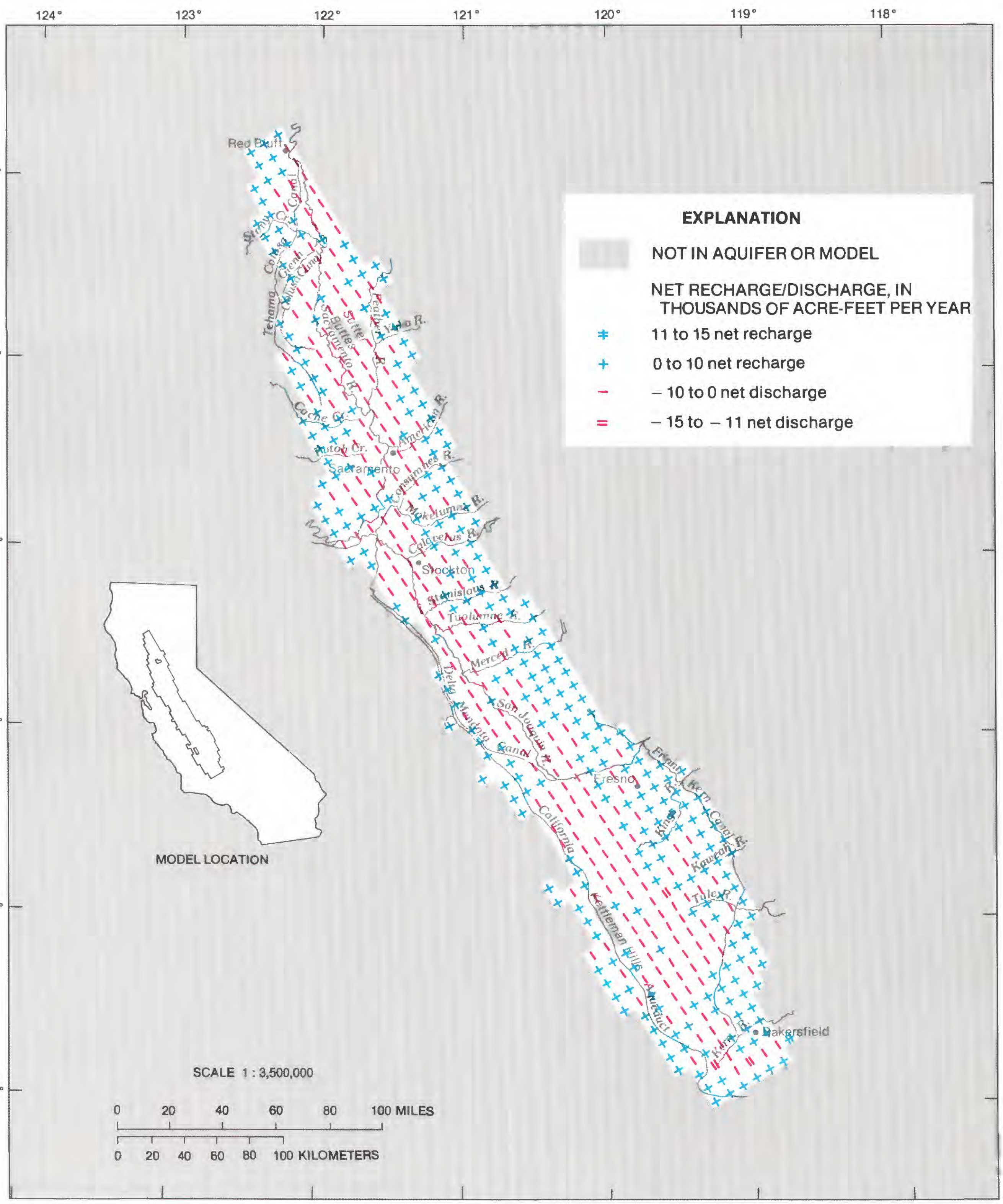


FIGURE 18.—Simulated predevelopment net recharge/discharge.

residues or bodies of estuarine marine water trapped when the sediments were deposited (Davis and others, 1959, p. 181; Olmsted and Davis, 1961, p. 136).

The initial simulation assumptions were that the interface between fresh and saline water was static and that the thickness of the aquifer system was equal to the thickness of the freshwater body. However, simulation results indicated that the assumption of a static interface between fresh and saline waters was not correct. Where the thickness of freshwater was small, the simulation required hydraulic conductivities in the aquifer system that were unrealistically large, and where the thickness of freshwater was large, the hydraulic conductivities required were unrealistically small. Davis and others (1959, p. 43) suggest that because there is little evidence of the marine sediments being flushed with freshwater (except on the southeast side of the San Joaquin Valley) and because of comparatively recent structural deformation, not enough time has elapsed for the interface between the freshwater and the saline water to reach a stable position. Thus, the thickness of the aquifer system used in

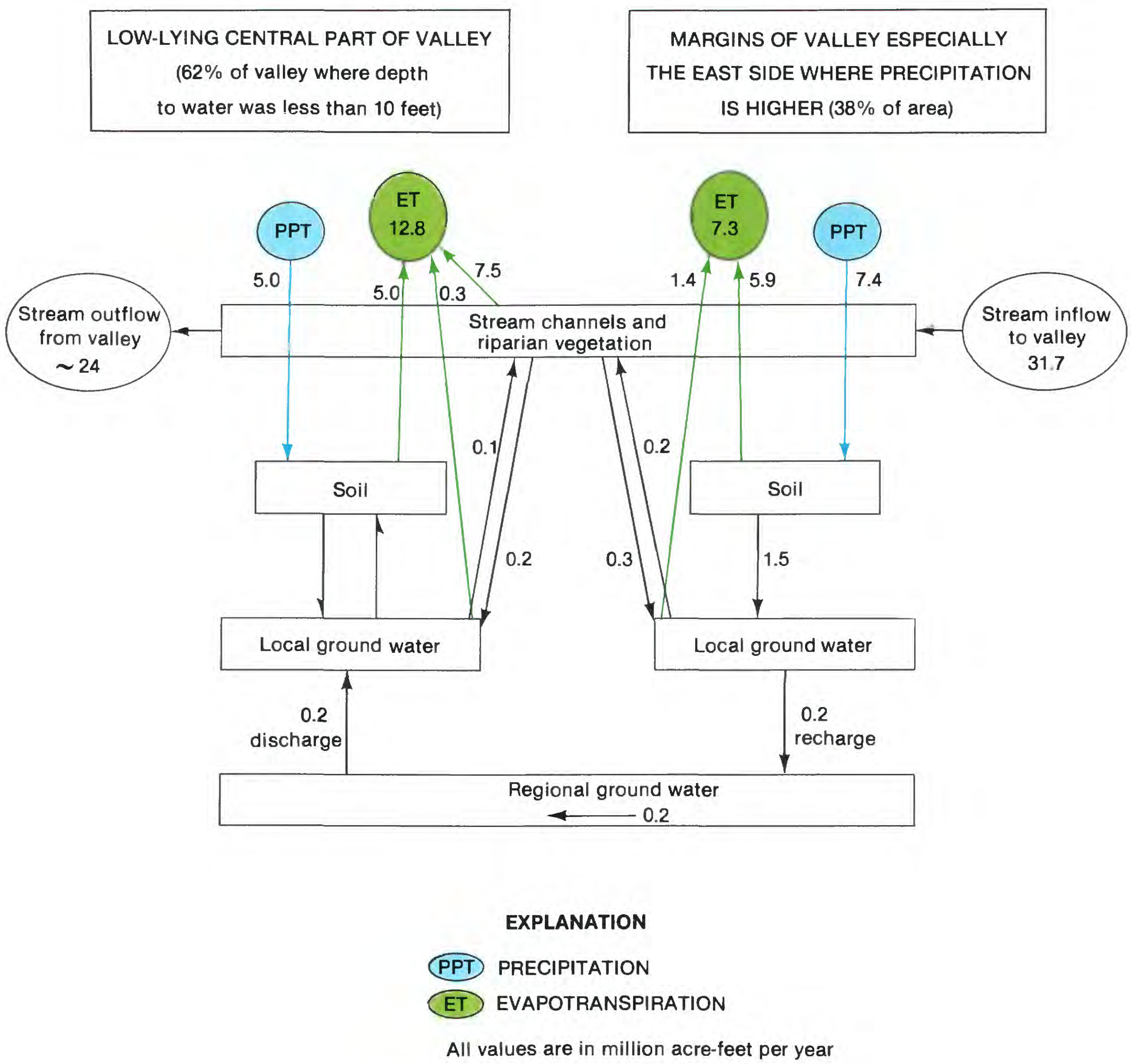


FIGURE 19.—Predevelopment water budget.

the final analysis of ground-water flow was increased to include most of the post-Eocene continèntal deposits.

Density variations between the freshwater and the saline water were not accounted for in the analysis of

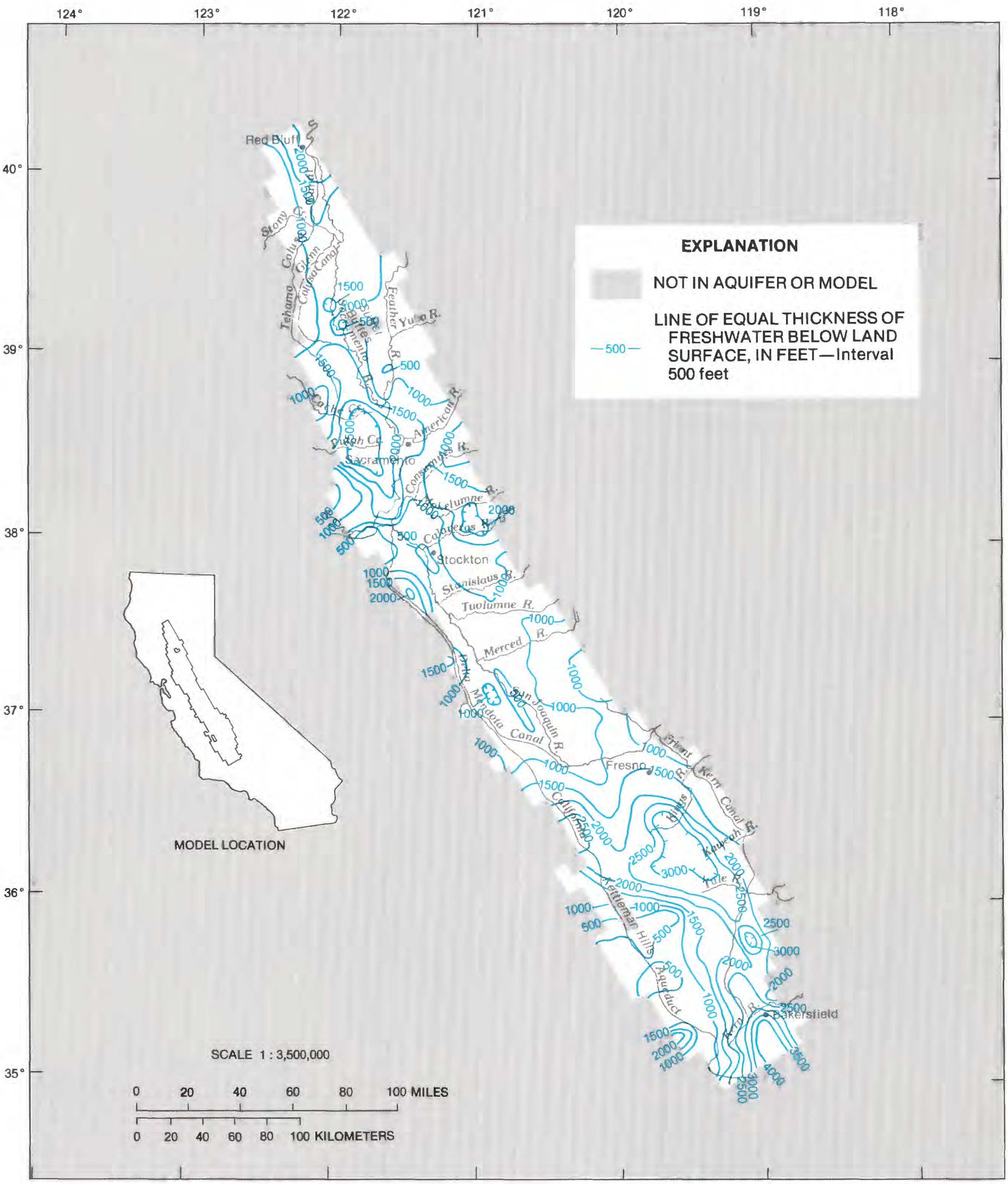


FIGURE 20.—Depth to base of freshwater.

ground-water flow, nor was any analysis done to determine the effect of pumping in the freshwater body on the movement of the saline waters. Not incorporating density differences in the analysis was thought to yield only minor errors in the overall analysis of ground-water flow because most of the flow occurs in the upper part of the aquifer system. Most of the post-Eocene continental deposits that contain saline water were incorporated in the lowest model layer where hydraulic head data are largely unknown and where essentially no ground water is pumped. Simulation results indicate that the amounts of water that move into and out of the lowest model layer are small. Under predevelopment conditions, only about 70,000 acre-ft/yr (23 percent of the layer 4 vertical flow) flows into or out of layer 1. In 1961, the total layer 1 vertical flow is only 6 percent of the layer 4 flow. These simulations assume that only hydraulic gradients cause the movement of brine waters.

POSTDEVELOPMENT GROUND-WATER FLOW

The period 1961–77 was studied intensively to understand the present flow system and to attempt to detect trends. This period was not affected by natural climatic trends (fig. 7). The period from predevelopment (before about 1860) until 1961 was not studied intensively because very little data are available and it would be difficult to extrapolate back in time because so many conditions have changed.

HISTORY OF WATER DEVELOPMENT

The favorable climate for agriculture in the Central Valley combined with water management and transferring water from areas of abundant water to areas of scarcity has resulted in one of the most productive agricultural areas in the Nation that is dependent on irrigation. This agricultural area is further expanded such that the valley is one of the Nation's largest users of ground water. Water development for irrigation has had a major effect on the hydrologic budget of the valley, in both ground water and surface water. Development of both surface- and ground-water sources for domestic and industrial needs has also expanded greatly over the years. The quantity of domestic and industrial water needed, however, has always been small compared with the quantity needed for irrigation.

IRRIGATION

Irrigation was introduced to California around 1790 by Roman Catholic priests from Mexico (Hall, 1889). From 1790 to about the late 1860's, development spread into the Central Valley in a sporadic manner. In the initial phases of irrigation development, local interests were responsible for developing and managing their own resources. In the foothill area of the Sierra Nevada and adjacent sections of the valley, development after 1849 was accelerated as a result of the Gold Rush. After mining ceased, the ditches were used to convey water for irrigation.

In 1857, an act was passed by the California State Legislature that offered patents to anyone who would drain and reclaim river-bottom lands (Manning, 1967). As a result, most of the earliest expansion in irrigation was concentrated on the valley floor, where broad plains had been subject to annual flooding from the main rivers that traversed these lowland areas. Thousands of miles of canals and laterals were constructed to drain the wetlands. Additional diversion began as a result of appropriation of sustained flows from the main rivers. By 1900, the entire flow of the Kern River and much of the flow of the Kings River had been diverted by a series of canals constructed to serve lands throughout the southern San Joaquin Valley (Nady and Larragueta, 1983a). Because no significant construction of storage facilities accompanied these earliest diversions, the amount of irrigation water was limited by the low summer flow.

When the drought around 1880 caused a great decrease in surface water in the San Joaquin Valley, ground water began to be developed to supplement the decreased supply as well as to serve lands beyond the reach of the diversion canals (Manning, 1967). In the earliest period of ground-water development, shallow ground water was plentiful and flowing wells were common, especially around the old lake basins in the central parts of the San Joaquin Valley. By 1910, almost all of the surface-water supply in the San Joaquin Valley had been diverted, causing an increased impetus to develop ground-water resources.

Even though ground-water use prior to 1900 was increasing, it was only a very minor part of the total irrigation supply. With increased production from the ground-water system, flow rates declined steadily in the once naturally flowing wells and it became necessary to install pumps for irrigation. Around 1930, the development of a greatly improved deep-well turbine pump spurred additional ground-water development for irrigation, because it allowed more efficient pumping from greater depths.

Further expansion of irrigation development was dependent on the provision of additional sources or more elaborate means for transporting existing streamflow to the land. Again, it was local efforts that conceived and completed the first reservoirs along the eastern margin of the valley.

Construction of larger storage reservoirs, major canals, and large-scale pumping plants was expensive and, therefore, beyond the means of most groups of water users. It was in response to this need that the Federal government became involved with irrigation and was responsible for construction of substantial storage, pumping, and conveyance facilities in California, beginning in the 1940's. Tables 4 and 5 summarize the development of major water facilities in the valley.

The Bureau of Reclamation's Central Valley Project (CVP) is one of these large-scale projects. The CVP, consisting of major storage and conveyance facilities, is a major conservation and reclamation project designed to be a multipurpose development to supply water for irrigation, municipal, industrial, and other uses. The project has several key features. Shasta Dam on the upper Sacramento River was built to store winter flows to be released during the summer irrigation season and the following year, if necessary. Sacramento River water is diverted from the Delta south through the Delta-Mendota Canal to meet irrigation needs in the southern San Joaquin Valley (see fig. 3). This allows diversion of San Joaquin River water from below Friant Dam, north in the Madera Canal, and south in the Friant-Kern Canal.

In the late 1950's and early 1960's, the California State Water Plan (SWP) was initiated. Because of the great cost, this project was an effort of the entire State. A major project of the SWP is the Oroville Dam on the

TABLE 4.—*Surface-water stor4age reservoirs in the Central Valley*

[Abbreviations: USBR, U.S. Bureau of Reclamation: CoE, U.S. Army Corps of Engineers; SWP, California Department of Water Resources State Water Project; Priv., private]

	Average annual flow (acre-ft/yr)	Dam/Reservoir	Storage capacity (acre-ft)	Year completed	Owner
Putah Cr.	373,000	Monticello Dam/ Lake Berryessa	1,592,000	1957	USBR
Stony Cr.	458,600	Black Butte	147,600	1963	CoE, USBR
Sacramento R.	6,223,000	Shasta	4,436,000	1949	USBR
Feather R.	4,263,000	Oroville	2,685,000	1968	SWP
Yuba R.	1,800,000	Englebright	70,000	1941	CoE
North Yuba R.	112,300	New Bullards Bar	727,400	1969	Priv.
Bear R.	326,700	Camp Far West	102,200	1963	Priv.
American R.	2,714,000	Folsom	1,010,000	1956	USBR
Mokelumne R.	577,400	Camanche	431,500	1963	Priv.
Calaveras R.	158,700	New Hogan	323,700	1963	CoE
Stanislaus R.	974,500	New Melones	2,420,000	1978	USBR
Tuolumne R.	1,826,000	New Don Pedro	2,030,000	1970	Priv.
Merced R.	969,400	New Exchequer Dam/ Lake McClure	1,024,000	1967	Priv.
Chowchilla R.	71,870	Buchanan Dam/ Eastman Lake	150,600	1975	CoE
Fresno R.	78,970	Hidden Dam/ Hensly Lake	85,300	1975	CoE
San Joaquin R.	1,721,000	Friant Dam/ Millerton Lake	503,200	1942	USBR
Kings R.	1,655,000	Pine Flat	1,001,000	1951	Priv., CoE
Kaweah R.	475,300	Terminus Dam/ Lake Kaweah	142,900	1962	Priv., CoE
Kern R.	668,000	Isabella	567,900	1954	Priv., CoE
Tule R.	134,800	Success	81,700	1961	Priv., CoE
Calif. Aqueduct[1]	N/A	San Luis	2,040,000	1967	SWP, USBR
TOTAL	25,580,000		21,572,000		

[1]Not a river, but a major water conveyance connected to large reservoir.

TABLE 5.—*Major water-conveyance facilities in the Central Valley*

[Abbreviations: USBR, U.S. Bureau of Reclamation; CoE, U.S. Army Corps of Engineers; SWP, California Department of Water Resources State Water Project; Priv., private]

Stream	Average annual flow (acre-ft/yr)	Canal	[1]Normal flow (acre-ft/yr)	Year com-pleted	Owner
Sacramento R.	9,629,000	Tehama-Colusa	[2]509,500	1971	USBR
Sacramento R.	11,510,000	Glenn-Colusa	811,200	1905	Priv.
Putah Cr.	373,100	Putah So.	222,500	1959	USBR
Delta	N/A	Delta-Mendota	2,348,000	1951	USBR
Delta	N/A	Calif. Aqueduct	1,510,000	1968	SWP, USBR
San Joaquin R.	1,721,000	Madera	226,000	1944	USBR
San Joaquin R.	1,721,000	Friant-Kern	1,002,000	1949	USBR
TOTAL			6,630,000		

[1]Based on a near-normal year, 1975.

[2]Based on 1978-81 average.

Feather River, which allows diversion of water in the Delta (fig. 3) into the California Aqueduct. From the Delta, water flows south, to San Luis Reservoir and then to the southern San Joaquin Valley, and is pumped over the Tehachapi Mountains (fig. 1) to southern California.

Figure 21 shows the increase in irrigated acreage in California from 1870 to 1975 and in the Central Valley and its subregions from 1959 to 1975. The proportion of irrigation from ground water to irrigation from surface water has changed greatly over the years, as well. Until 1900, only a small amount of the irrigation was from ground water. T.R. Simpson (Pacific Gas and Electric Co., unpub. rept., 1949) states that in the San Joaquin Valley, the combined capacity of wells south of Chowchilla was 5.3 million acre-ft/yr in 1919 and about 14.9 million acre-ft/yr by 1929. The combined gross pumpage of more than 35,000 wells in the San Joaquin Valley south of Merced in 1948 was close to 6 million acre-ft/yr. As the amount of ground water pumped increased, so did its proportion of total irrigation because surface-water use did not increase as much. Davis and others (1964) reported that in the San Joaquin Valley in 1952, gross diversion of surface water was about 8.5 million acre-ft/yr and ground-water pumpage for irrigation was about 7.5 million acre-ft/yr.

During the period 1961–77, ground-water use accounted for about 50 percent of the irrigation supply in the Central Valley. As shown in figure 22, the proportion between surface water and ground water varies substantially from dry to wet years. Many farms are equipped to use either ground water or surface water. Therefore, in wet years abundant and inexpensive surface water is used, whereas in dry years (note 1976–77) ground-water use is predominant. Most surface water is distributed from the streams or Federal and State canals or reservoirs to one of several hundred irrigation districts that distribute to individual farms. Most of the fields are irrigated by some type of flooding method (border or furrow), but in about 20 percent of the area, sprinklers are used (Stewart, 1975, p. 20). Based on the number of agricultural power accounts in the late 1960's, there were about 100,000 active irrigation wells in the valley. The distribution of ground-water pumpage, shown in figure 23, is more toward the southern and eastern parts in the valley where irrigation is most extensive. The distribution and magnitude varies, as shown by comparing the two dry years (1961 and 1977) with the near-normal years (1962 and 1975). Trends through the period are also evident. Well-construction data for about 3,000 irrigation wells show that most wells are perforated throughout the lower two-thirds of their depth. The distribution of the approximate depth to the weighted center of the pumped zone is shown in figure 24. Variation in the depth of major production zones is because of water quality and aquifer-yield considerations. A more complete treatment of the distribution of ground-water pumpage is given by Diamond and Williamson (1983).

DOMESTIC AND INDUSTRIAL

A small proportion of water used in the valley is for domestic and industrial purposes. Ground-water pumpage for domestic use increased about 3 percent per year, from about 300,000 acre-ft in 1961 to about 490,000 acre-ft in 1977 (Diamond and Williamson, 1983). Industrial water use in 1970 was 132,000 acre-ft (California Department of Water Resources, 1977c, p. 74, 75). This figure includes both surface-water and ground-water use.

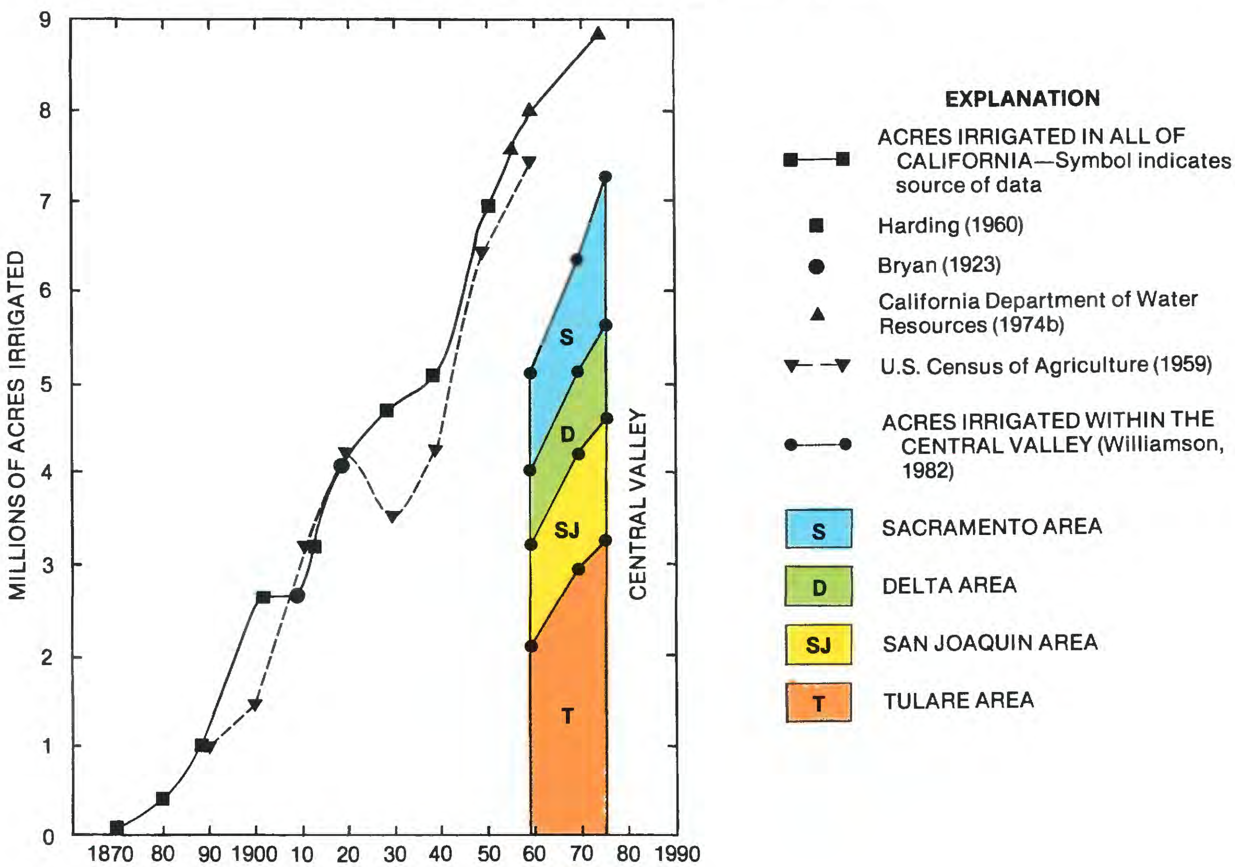


FIGURE 21.—Increase of irrigated acreage in California since 1870 and in the Central Valley since 1959 (modified from Nady and Larragueta, 1983a, sheet 2).

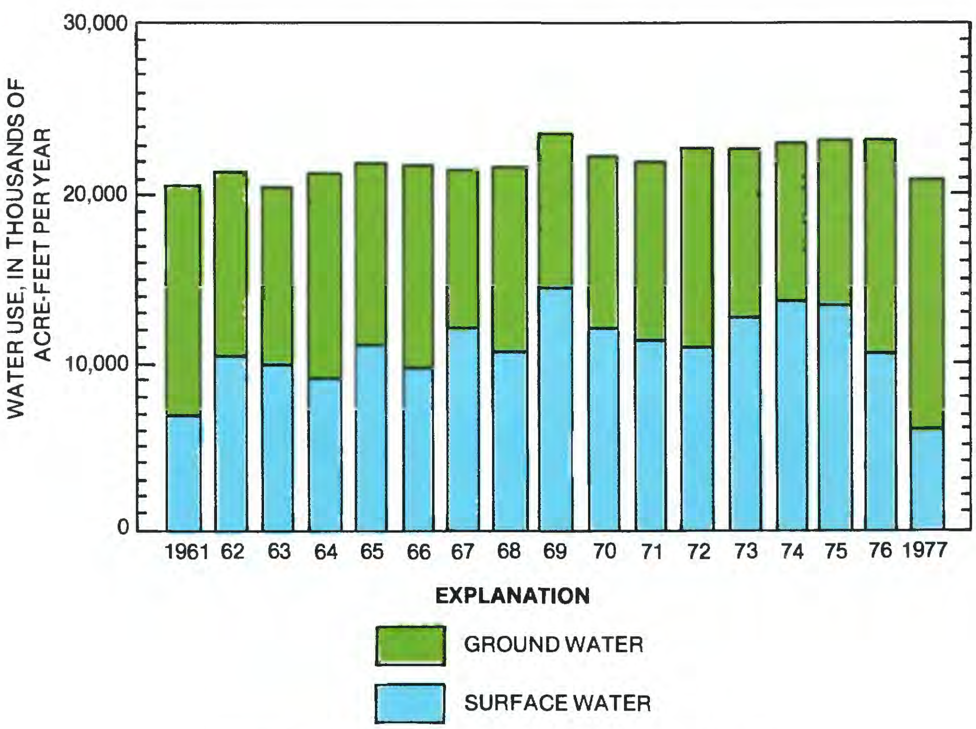


FIGURE 22.—Irrigation water use from 1961 through 1977.

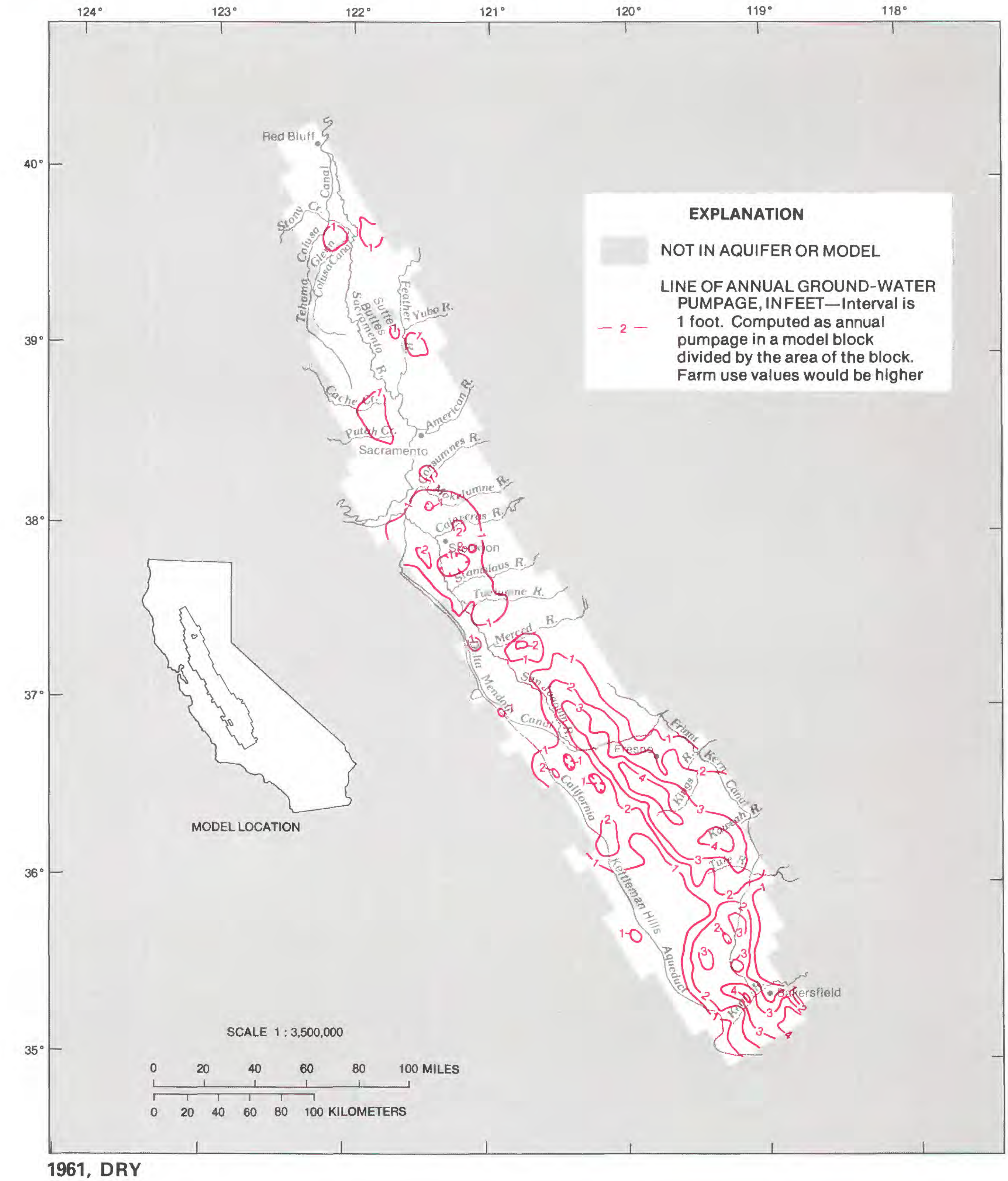


FIGURE 23A.—Ground-water pumpage for 1961.

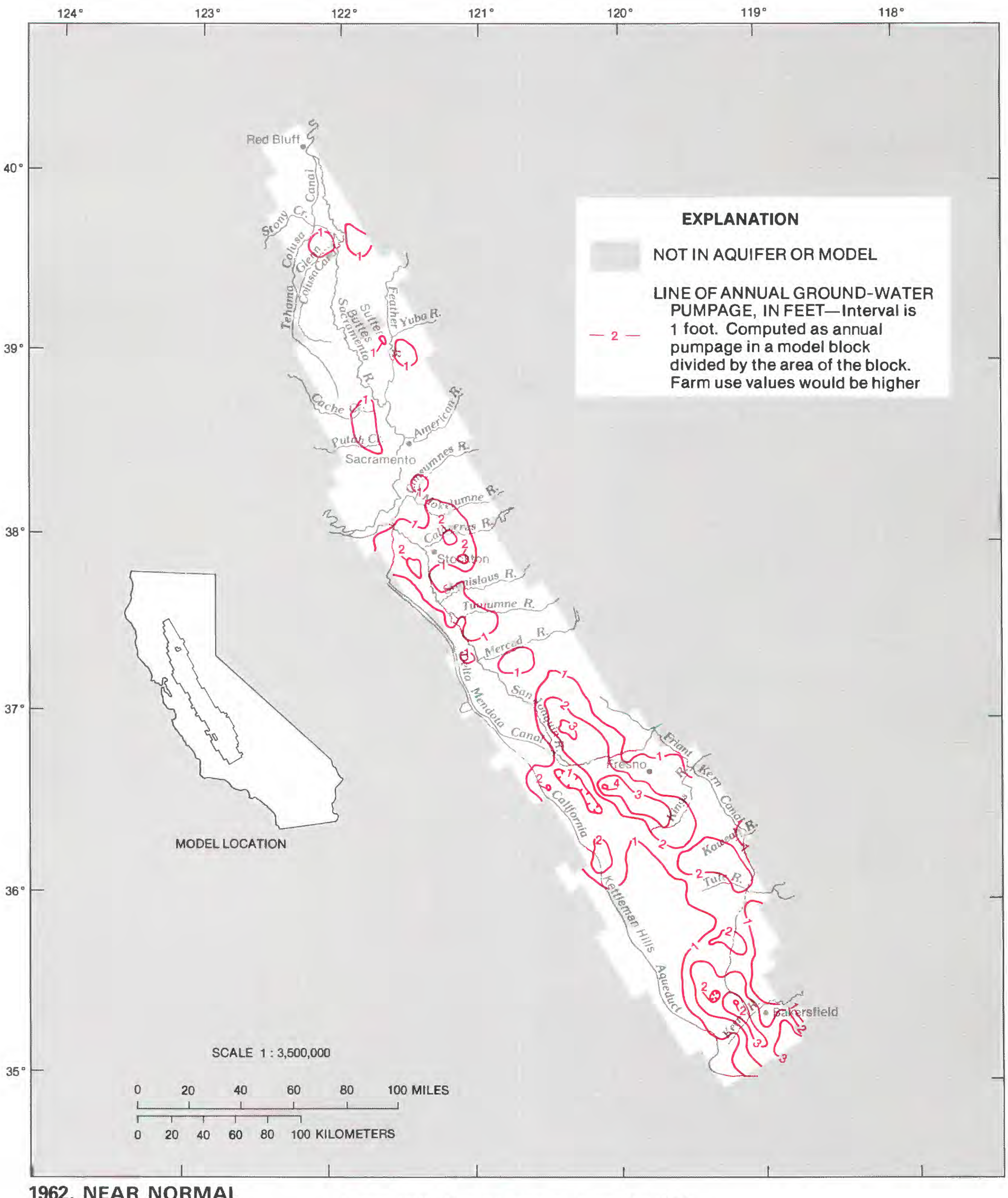


FIGURE 23B.—Ground-water pumpage for 1962.

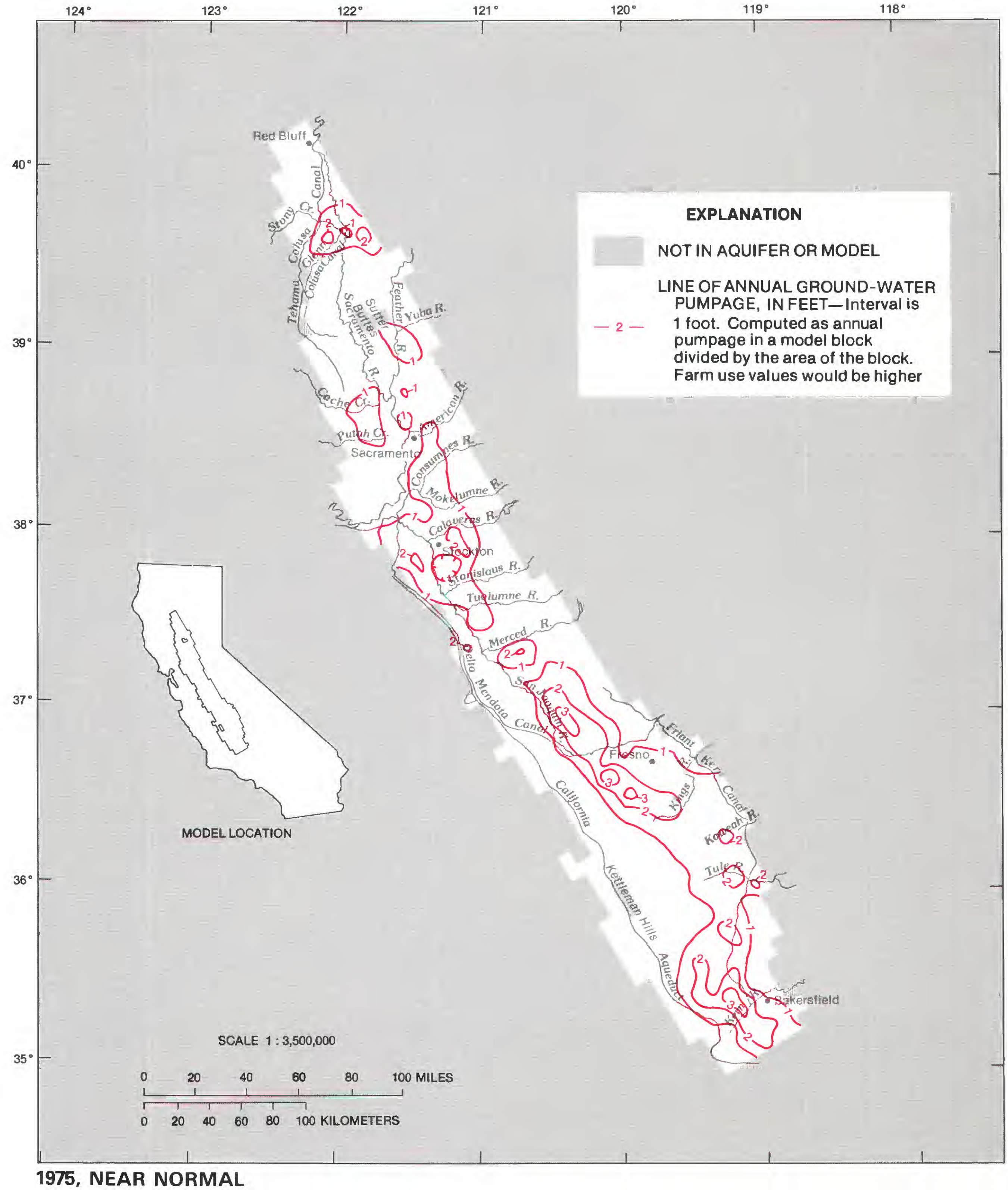


FIGURE 23C.—Ground-water pumpage for 1975.

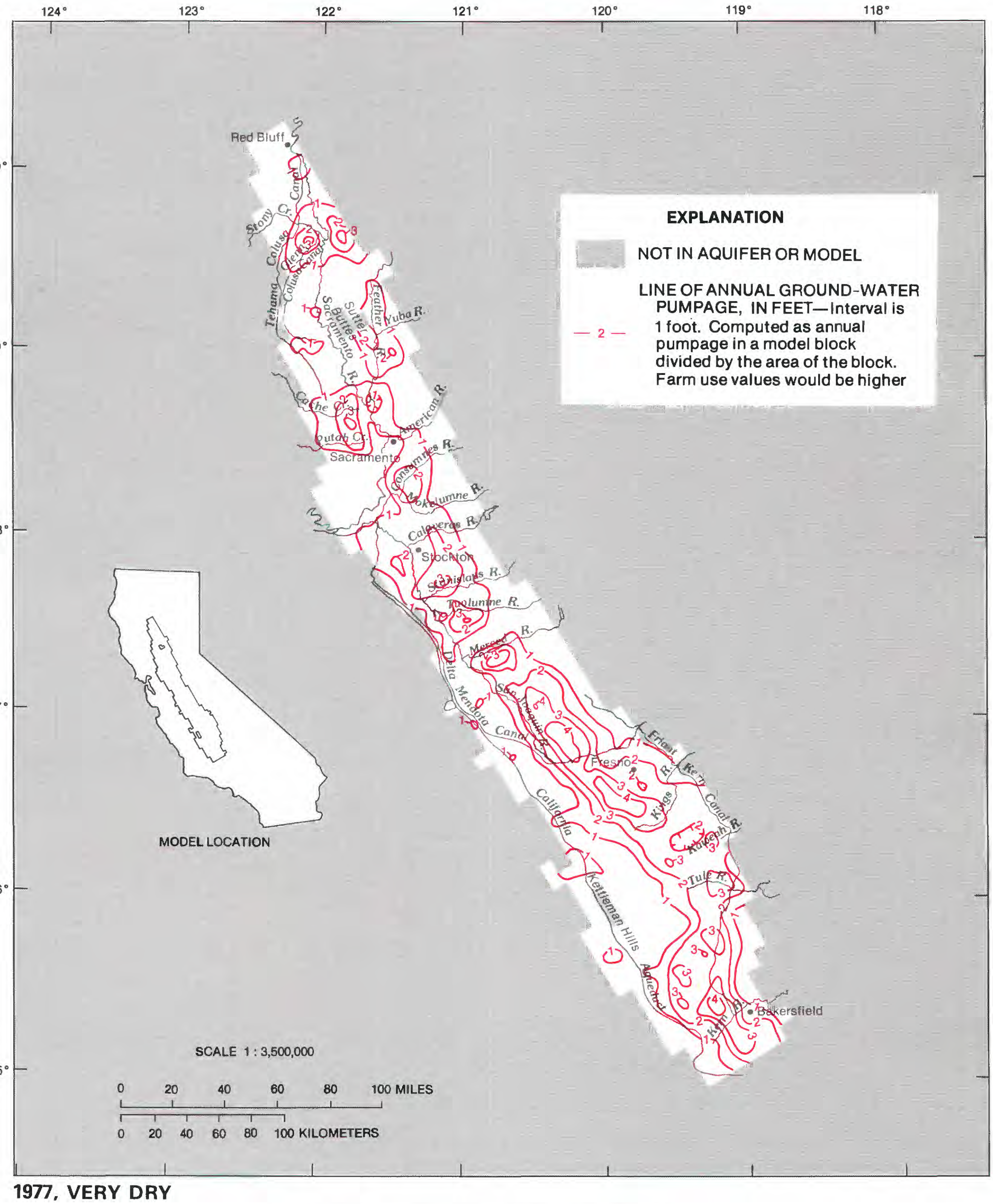


FIGURE 23*D*.—Ground-water pumpage for 1977.

EFFECTS OF DEVELOPMENT

Development of water resources has had a major effect on the aquifer system. In many areas, pumpage has lowered water levels, which has altered the direction and rates of ground-water flow (fig. 25) and, in places, has caused the land to subside. Large diversion of surface water for irrigation has altered the amount and distribution of recharge to the aquifer system, which has caused a change in the configuration of the water table. All of these causes, but principally surface-water diversions, have decreased the volume of surface water discharged into Suisun Bay. Changes in or to the aquifer system caused by development are discussed in the following paragraphs.

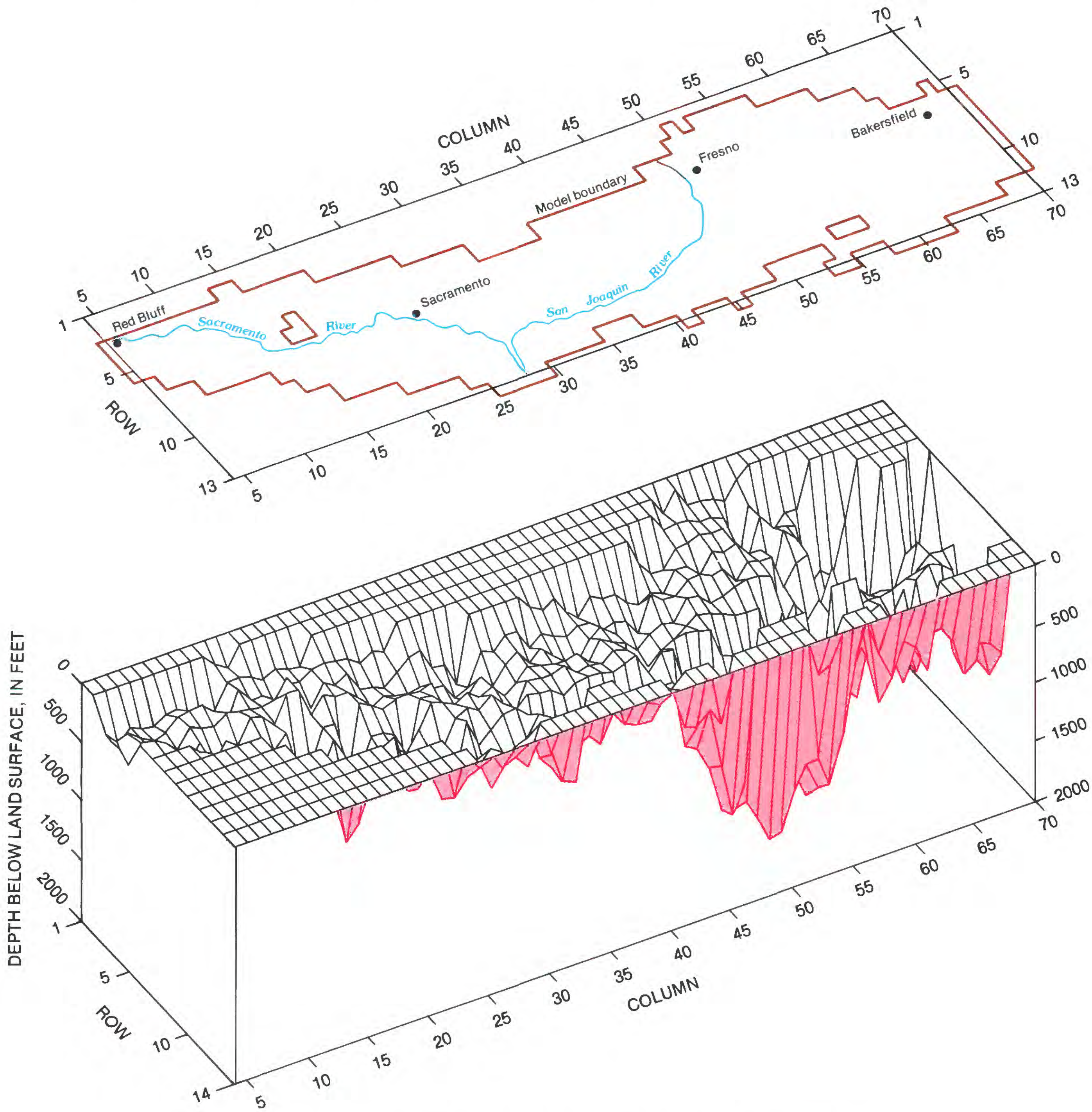


FIGURE 24.—Approximate depth to the weighted center of the pumped zone.

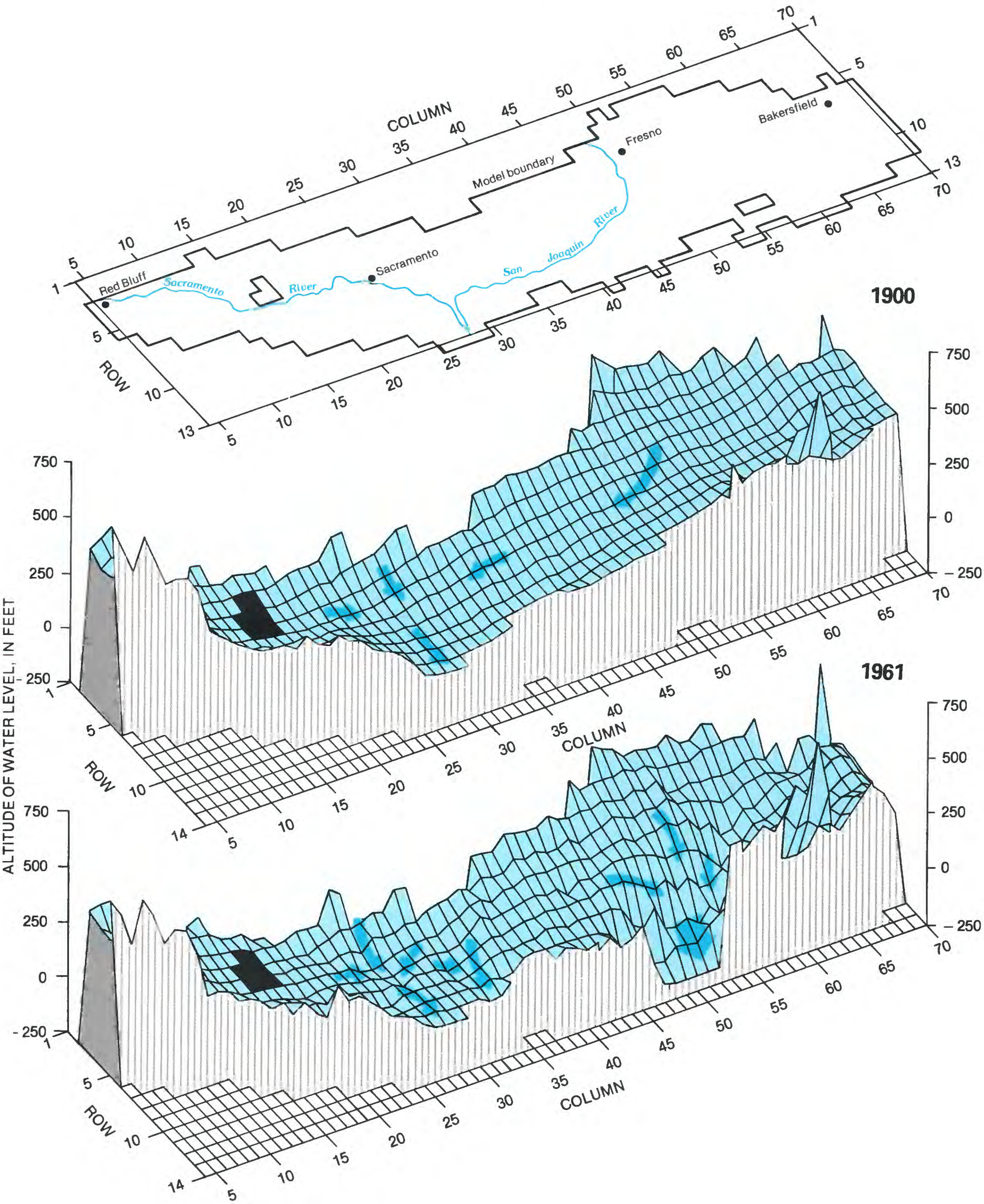


FIGURE 25.—Change in water level and direction of flow in the lower pumped zone, 1900–61, due to ground-water pumpage.

CHANGES IN RECHARGE AND DISCHARGE

Development of irrigated agriculture has had major effects on the volume and distribution of ground-water recharge and discharge in the valley. This is shown by comparing recharge and discharge values from the predevelopment and postdevelopment simulations. During predevelopment conditions, the recharge and discharge was about 2 million acre-ft/yr each (fig. 26*A*). From predevelopment times to the period 1961–77, average discharge increased to 12.2 million acre-ft/yr and average recharge increased to 11.4 million acre-ft/yr.

Agricultural development in the valley has changed the paths of most of the 31.7 million acre-ft of surface-water inflow. Figure 26 shows the magnitude and postdevelopment changes in the major components of a hydrologic budget for the valley. More detail on how the budget components were estimated can be found in the "Model Development" section. Average budget components for 1961–77 for each area and subarea (fig. 27) are given in table 6.

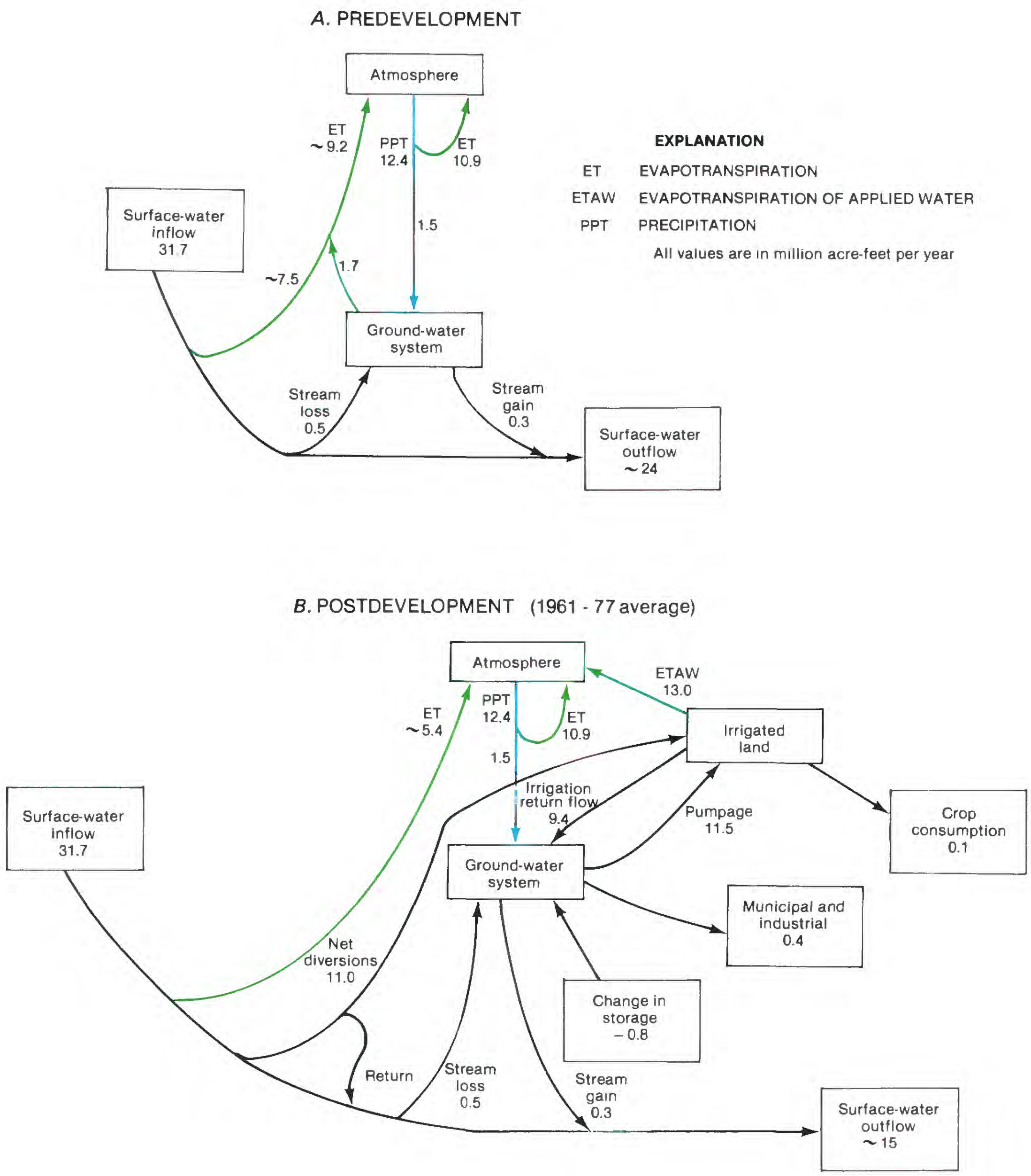


FIGURE 26.—Change in water-budget terms due to development in the Central Valley.

An index of surface-water outflow from the Delta was estimated for the period 1922–80 by summing the gaged annual flows into the Delta and adjusting for use, precipitation, and export. A linear multiple regression was used to relate Delta outflow to year and annual precipitation as a mean of four gaging stations; a

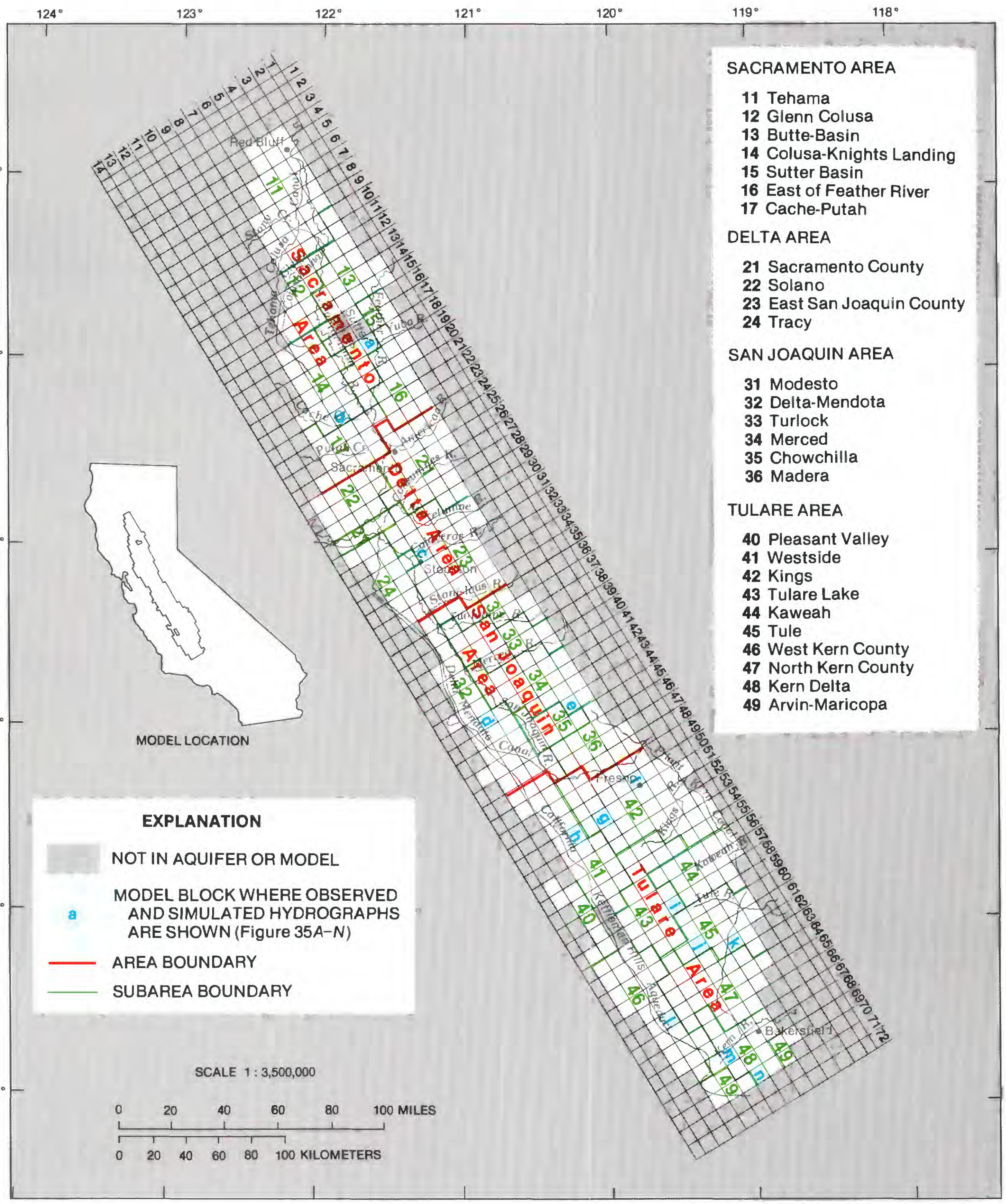


FIGURE 27.—Area and subarea boundaries.

TABLE 6.—*Summary of the components of ground-water recharge and discharge, 1961–77 average*

[All values in 1000 acre-ft/yr. Columns 8–13 are calculated as follows: Column 8 = columns 4/(5 + (6+14–7)); Columns 9 and 10 are sums of − (discharge) and + (recharge) values of = columns (1 + 2 − 3 − 4 + 5 − 7 + 14); Columns 11 and 12 are sums of +, − adjustments to layer 4 net recharge; Column 13 = columns (− 9 + 10 − 11 + 12). Totals may not agree because of rounding]

Subarea No. (See fig. 27)	Excess precipitation	River Loss	River Gain	Evapotranspiration of applied water	Surface water diverted	Pumpage, layer 4	Municipal pumpage	Irrigation efficiency	Net estimated Discharge	Net estimated Recharge	Adjustments +	Adjustments −	Net recharge/discharge, layer 4	Pumpage, layer 3
	(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)	(12)	(13)	(14)
11 -----	220	57	6	281	178	182	6	0.64	33	282	195	5	59	87
12 -----	40	7	3	373	649	44	7	0.51	3	356	337	--	16	40
13 -----	134	17	10	476	351	188	18	0.77	50	144	98	45	41	95
14 -----	83	9	11	362	258	111	--	0.76	28	113	70	53	68	108
15 -----	43	10	45	345	538	76	1	0.53	8	251	236	--	7	43
16 -----	158	42	59	389	324	215	10	0.65	74	214	116	59	83	73
17 -----	67	13	3	378	343	140	29	0.67	58	180	99	72	95	109
Sacramento --	746	155	137	2,605	2,642	956	70	0.64	254	1,539	1,150	234	369	555
21 -----	200	87	4	381	299	217	61	0.63	28	322	218	9	86	155
22 -----	80	0	--	219	297	21	1	0.67	17	183	185	12	7	9
23 -----	153	49	16	905	638	391	4	0.65	58	335	210	198	266	362
24 -----	23	3	11	396	173	170	6	0.69	51	77	13	215	227	241
Delta -------	456	139	31	1,901	1,407	799	73	0.66	153	917	626	434	572	767
31 -----	41	0	26	236	353	97	27	0.50	12	168	105	0	51	50
32 -----	29	56	38	835	1152	231	8	0.54	51	582	360	47	218	176
33 -----	44	8	42	424	465	110	9	0.65	33	161	107	11	33	85
34 -----	59	27	10	497	489	92	17	0.53	64	498	243	62	253	382
35 -----	8	0	3	296	187	110	1	0.53	17	178	--	41	202	266
36 -----	21	17	1	321	182	189	6	0.47	15	231	1	44	258	324
San Joaquin -	203	109	120	2,609	2,828	830	68	0.54	193	1,819	816	204	1,015	1,284
40 -----	5	2	--	63	18	18	3	0.75	14	25	5	77	83	51
41 -----	8	3	0	804	425	89	2	0.83	63	154	1	216	307	463
42 -----	50	46	7	1,601	1,107	1,303	90	0.56	351	419	50	460	478	562
43 -----	9	16	0	748	473	249	17	0.92	215	54	--	267	105	104
44 -----	16	12	--	627	571	335	19	0.49	60	415	32	81	405	402
45 -----	14	13	--	607	428	194	12	0.59	43	306	8	129	383	426
46 -----	5	2	--	228	221	149	2	0.50	49	133	33	81	132	86
47 -----	8	5	--	678	386	152	5	0.49	12	566	45	126	635	839
48 -----	2	19	3	390	355	234	32	0.38	23	434	112	93	392	460
49 -----	1	13	--	144	97	36	2	0.46	5	152	44	96	200	182
Tulare ------	119	131	10	5,891	4,081	2,758	183	0.58	834	2,657	330	1,626	3,119	3,576
Central Valley ----	1,524	534	299	13,005	10,958	5,342	395	0.59	1434	6,933	2,922	2,499	5,076	6,181

decrease of outflow with time was noted. Average Delta outflow declined from about 24 million acre-ft/yr to about 15 million acre-ft/yr during the period 1920–80. The adjusted *R*-square for the relation (a coefficient of determination of the regression) was about 0.67. This decrease was caused mainly by increased evapotranspiration within the valley because of irrigation. Irrigation had other substantial effects on the hydrology of the valley. A large volume of water flows through the irrigation cycle in the form of net surface-water diversions and ground-water pumpage, becoming evapotranspiration of applied water, infiltration, and crop consumption. Net surface diversions do not include volumes that are reused by other irrigators or returned to some surface-water body. In figure 26, the term evapotranspiration (ET) from streams includes ET from nonirrigated lands and was calculated as residuals in the budgets presented. The losses and gains from streams for the predevelopment conditions are poor estimates because they were derived from the postdevelopment estimates which are not necessarily the same. The values shown in figure 26*A* do not correspond to the previously mentioned sums of the simulated predevelopment recharge and discharge (0.2 million acre-ft/yr each, p. D40) because the previous values were summed from simulation output, which causes some cancellations of recharge and discharge within model blocks. Figure 26 shows a more realistic difference between overall pre- and postdevelopment recharge and discharge of a factor of about 6 to 1.

Postdevelopment average overall recharge comes mostly from irrigation return flow (83 percent), but also from precipitation (13 percent) and streams (4 percent). The actual proportion of overall recharge from streams to the aquifer system is probably larger; however, some recharge will discharge to nearby streams through local or intermediate flow systems, which are not modeled in the regional model.

Variations in the components of the water budget during the simulation period are shown in figure 28; wet years (1967, 1969, and 1973) and dry years (1961, 1976, and 1977) are easily identified. It is notable that overall irrigation efficiency improved from about 53 percent to about 64 percent during the period 1961–77. This can be inferred from the growth rate of irrigated acreage (fig. 21) because it exceeds the growth rate of irrigation water use (fig. 22). This is probably a result of economic and other conditions that encouraged irrigators to conserve water.

During early calibration of the simulation model, it was obvious that the estimates of river losses/gains and small stream recharge were too large. Water levels in aquifers in some losing sections of rivers rose hundreds of feet and in some gaining sections of rivers, water

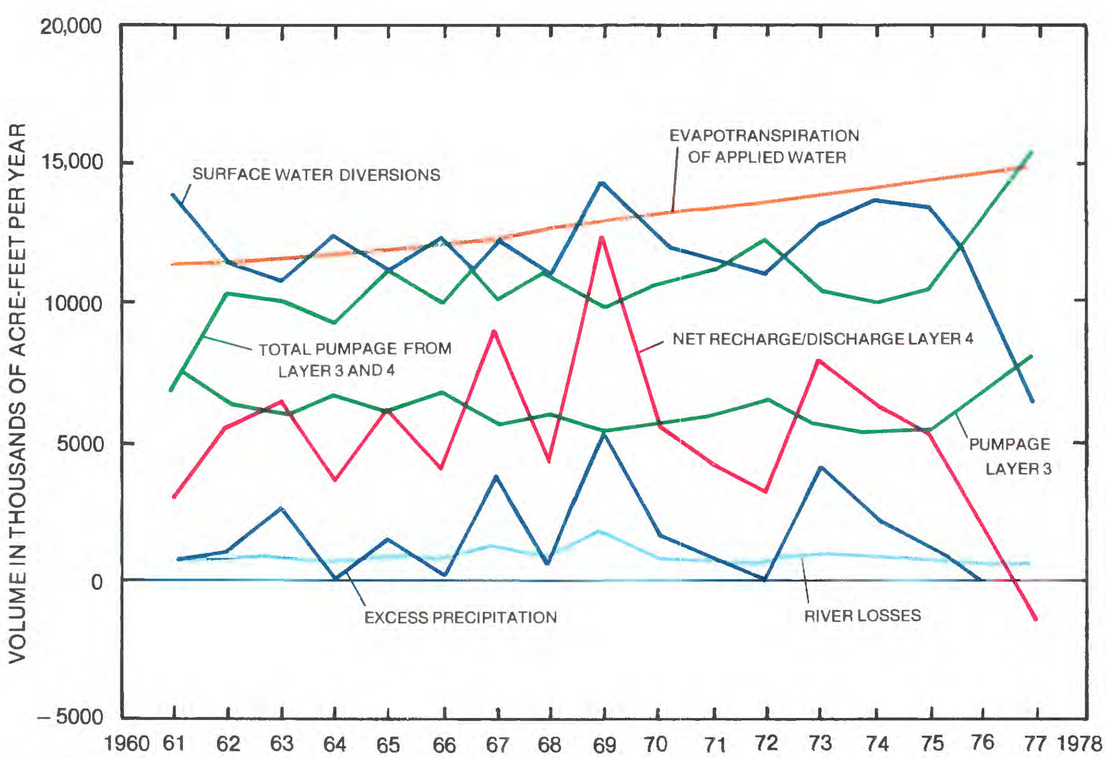


FIGURE 28.—Components of a recharge and discharge water budget, 1961–77. Components are all shown with positive signs. Net recharge/discharge for layer 4 equals surface water minus evapotranspiration of applied water (SW-ETAW) plus net precipitation (Net PPT) plus river losses/gains (Riv Loss); thus it can be positive or negative. Pumpage from layer 3 is discharge for that layer.

levels in aquifers dropped a similar magnitude. No reasonable adjustment in any other model value could correct the imbalance. Individual values of stream losses could be greatly in error owing to the increase of the measurement error in the residual analysis of the water budgets in the streams. However, long-term averages should be closer to the actual values if the errors are randomly distributed. Nevertheless, all of the estimated values of stream losses/gains were divided by five to allow the model to respond within the limits of reasonable adjustments in other values. This adjustment was necessary because of systematic errors in estimating stream losses/gains, local recharge and discharge within a model block, and inability of the model to simulate the aquifer system to match the observed water levels and water level changes.

After this calibration, the simulated water levels in the Sacramento Valley remained too high compared with observed values. To adjust for apparent overestimates of surface water diverted for irrigation, the diversion values in the Sacramento area (fig. 1) were multiplied by 0.75. This improved the simulation substantially.

In order to fit the observed water-table altitudes, additional small adjustments in the net recharge/discharge term were necessary. This was done because the process of allocating water-budget volumes to model blocks introduced errors that would result in too much water in one model block and too little in an adjacent block. The adjustment was made by relating change in simulated head to change in net recharge/discharge. The distribution of the resulting adjustments to net recharge/discharge is shown in figure 29. A spatial trend in these values of adjustment would indicate an underlying problem in the concepts or methods, such as a missing component of recharge. No such trend was detected, indicating that the net recharge/discharge errors were a result of random measurement and distribution errors.

CHANGES IN WATER LEVELS

Water-level changes resulting from water-resources development have occurred over most of the valley and have been of major proportions in many large areas. Generally, deeper pumped zones have much smaller storage coefficients than the specific yield of water-table systems because changes in head do not result in immediate dewatering of aquifer materials. Consequently, in deeper pumped zones, heads decline more rapidly and the cone of depression extends farther out than in a water-table aquifer that is stressed by similar amounts of pumpage. This is generally true in the Central Valley and the result is that water-level changes have been more pronounced in the lower pumped zone than at the water table. When water levels decline to a point that compaction of sediments begins to occur, the amount of water released from fine-grained sediments increases and tends to slow the rate of water-level decline. Figure 30 shows long-term hydrographs for wells that were chosen for the length of their record and the different stages of development that they represent (locations are shown on fig. 2). Each hydrograph (lettered *A–J*) in figure 30 shows wells that are located near each other to demonstrate patterns of hydraulic head change, both long term and seasonally, which differ primarily due to the well depth. Generally the deeper wells show more seasonal fluctuation and greater long-term declines than do shallow wells. Wells in the Sacramento Valley (fig. 30*A–C*) show a slow, but steady, decline beginning in the 1950's. Water development and water-level declines began earliest in the southern end of the Central Valley and moved north as time passed. Figure 30*D* shows that some heads in deep wells in the Delta area have been below sea level since before 1960. Somewhat farther south (figs. 30*E–F*), declines began occurring in the 1940's. Figure 30*G* shows an area in central Fresno County where the head decline in relatively shallow wells has been substantial, starting in the early 1940's. The Westside area wells shown in figure 30*H* show large declines until the late 1960's, followed by significant recovery due to decreases in pumpage because of importation of surface water, and then steep drawdown during the 1976–77 drought. Wells in the two other major subsidence areas, Tulare-Wasco (fig. 30*I*) and south of Bakersfield (fig. 30*J*), show complicated and highly variable patterns, with declines beginning before 1940.

PREDEVELOPMENT TO 1961

Water-table altitudes and lower pumped zone heads for spring 1961 are shown in figure 31. The changes in water level that have occurred since predevelopment conditions are shown in figure 32. Note that the changes shown in figure 32*B* were calculated from the observed 1961 heads and the simulated 1860 heads in the lower zone. The most substantial changes were in the western and southern parts of the San Joaquin Valley. There were smaller changes in most of the remaining areas of the Central Valley. The period between predevelopment conditions and 1961 was not simulated because of the absence of data for many critical components of recharge and discharge.

Just north of the Delta area (fig. 27), a depression in the water table to below sea level developed (fig. 31*A*). In the lower pumped zone, a depression developed north of Sacramento. These areas rely on ground-water pumpage for irrigation. Much of the lowlands of the Sacramento Valley sustained a small rise in the water table because of recharge from surface-water irrigation. Water levels for both the shallow and deep zones of eastern San Joaquin County declined substantially. The area encompassed by the zero-altitude contour grew much larger, especially in the lower pumped zone, indicating seawater intrusion which has caused difficulties for the city of Stockton (fig 2).

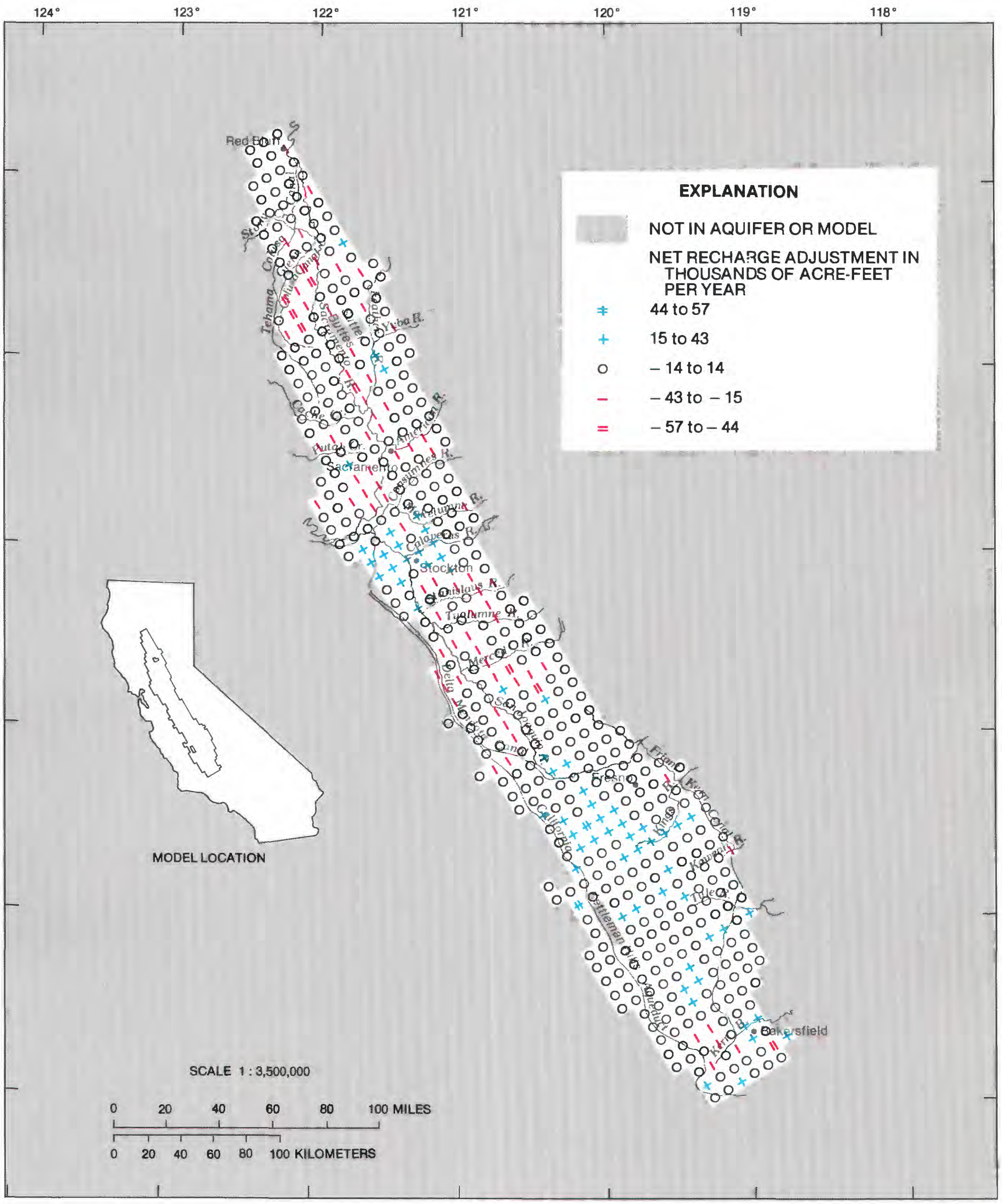


FIGURE 29.—Calibration adjustments to postdevelopment water-table net recharge/discharge estimates.

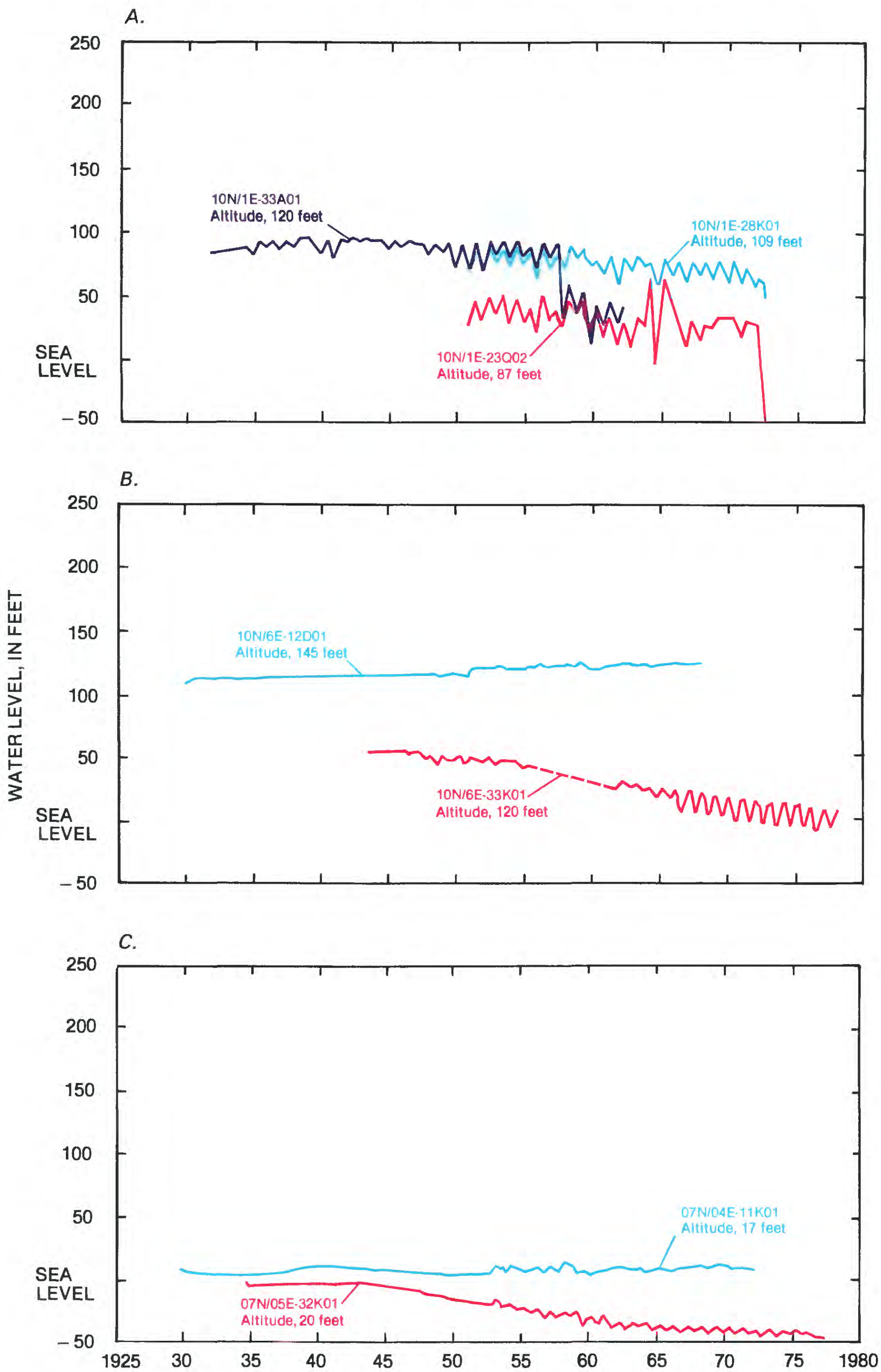


FIGURE 30*A–C*.—Measured water level in wells showing long-term water-level change, 1925–80. (Altitude shown is that of land surface at the well.)

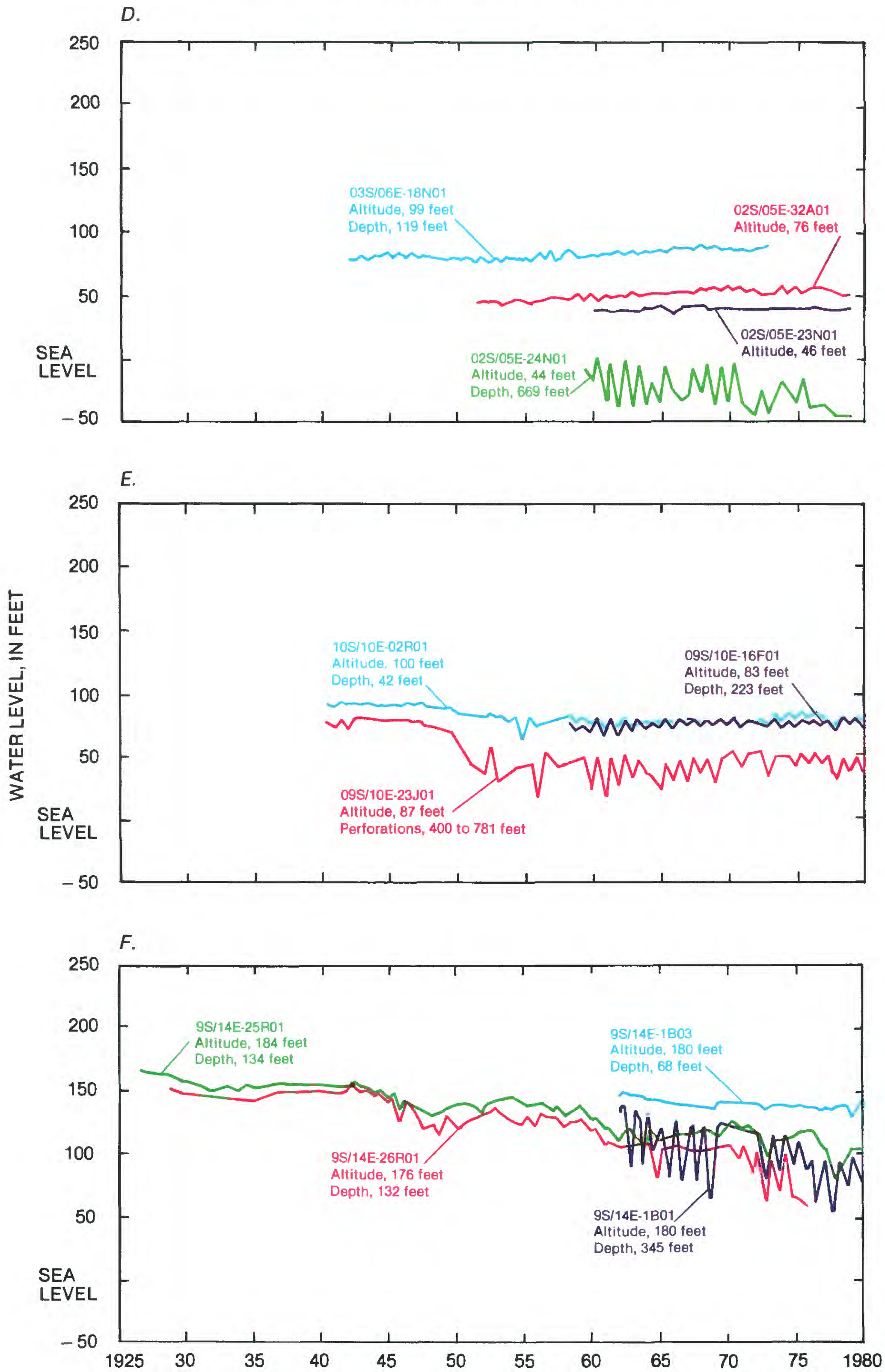


FIGURE 30*D–F*.—Measured water level in wells showing long-term water-level change, 1925–80. (Altitude shown is that of land surface at the well.)

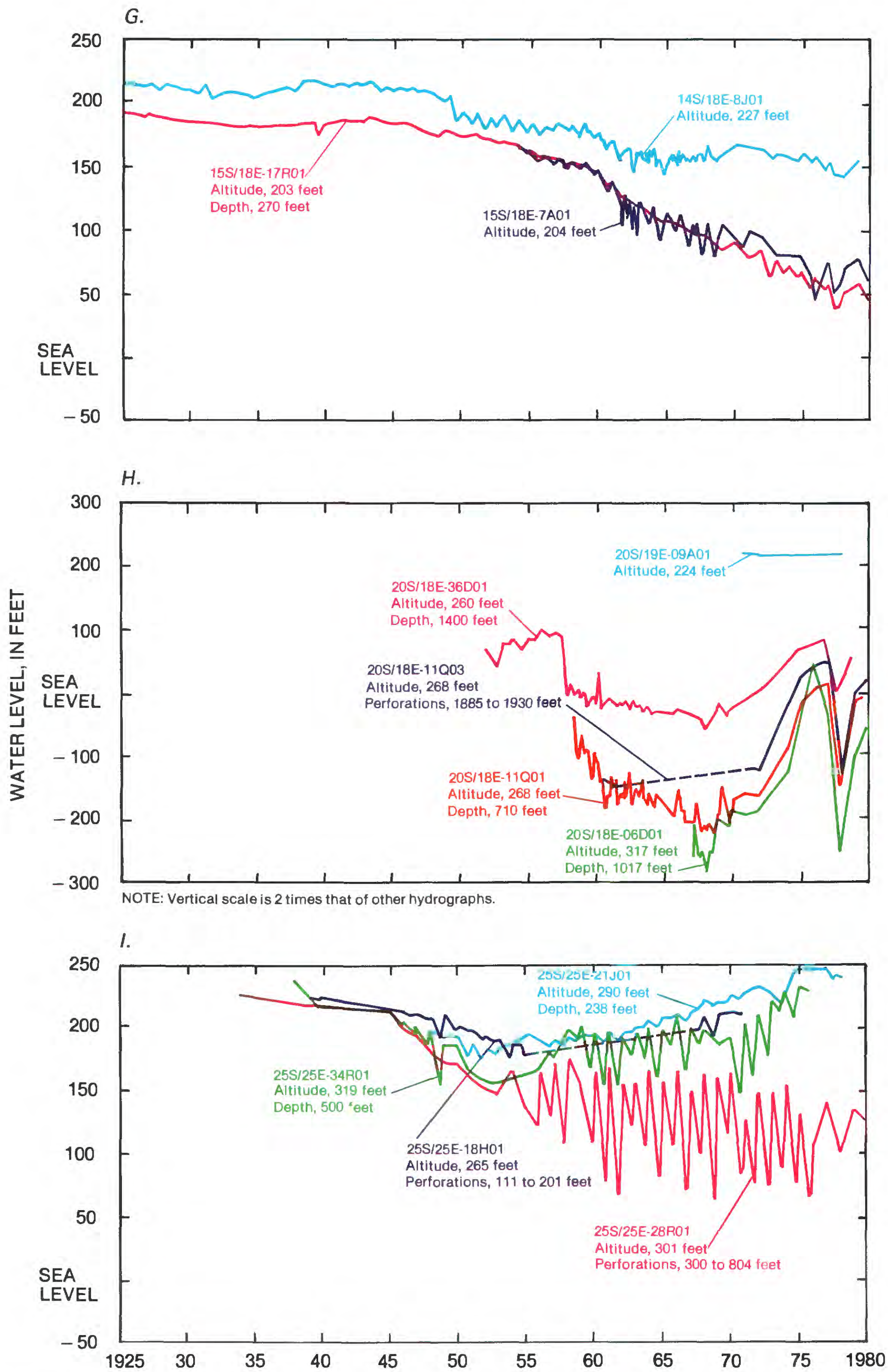


FIGURE 30G–I.—Measured water level in wells showing long-term water-level change, 1925–80. (Altitude shown is that of land surface at the well.)

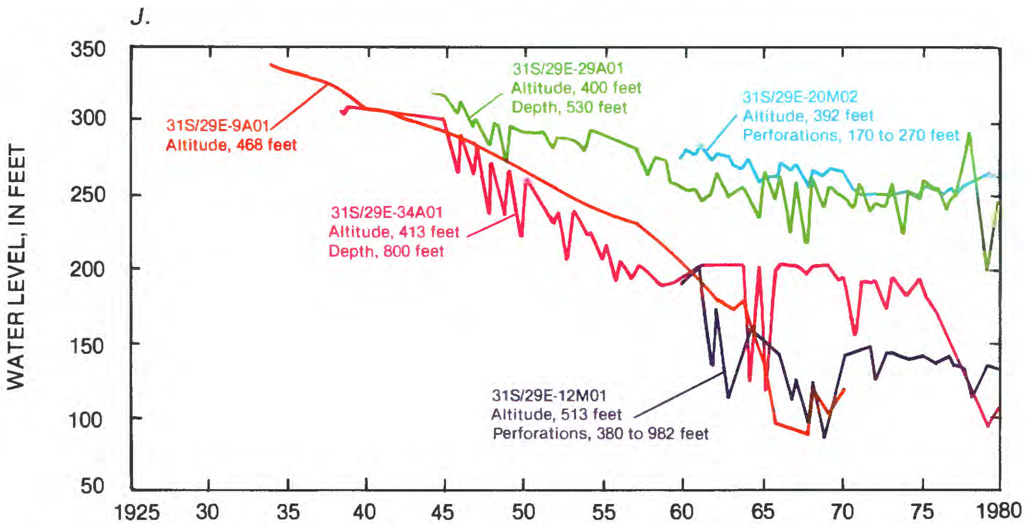


FIGURE 30*J*.—Measured water level in wells showing long-term water-level change, 1925–80. (Altitude shown is that of land surface at the well.)

The water table rose in the Delta-Mendota and the Westside areas (figs. 27 and 32*A*) because of recharge from surface-water irrigation. The water table declined substantially in the Chowchilla, Madera, western Kings, Pleasant Valley, Tule, and Kern County areas, which depend heavily on ground water for irrigation and which have many relatively shallow irrigation wells. In 1950, the Friant-Kern Canal (fig. 3) began delivering surface water along the east side of the San Joaquin Valley. In parts of the service area, water-level declines were reversed because of reduction in pumping (fig. 30*I*).

Water levels in the lower pumped zone declined as much as 400 ft in the Westside area from predevelopment to 1961 (figs. 27 and 32*B*). Until 1968, the irrigation in this area was supplied almost entirely by ground water. Around 1960, the lower pumped zone water levels were declining at a rate of about 10 ft/yr.

In the southeast and southern areas of the San Joaquin Valley, water levels in the lower pumped zone were declining, though not as dramatically as in the Westside area because there was some surface water available for irrigation.

1961 TO 1977

The observed and simulated water-table altitude for spring 1976 and the change in water table from 1961 to 1976 are shown in figure 33. In the Sacramento Valley, areas of past water-level decline showed continued and often accelerated decline. The depression of water level in some areas north of the Delta dropped to more than 40 ft below sea level. The area with water-table altitudes below sea level enlarged substantially. The water-level depression in eastern San Joaquin County developed in magnitude and areal extent.

In the San Joaquin Valley, the rate of water-table decline increased in the Chowchilla, Madera, and western Kings areas. Significant water-table declines occurred in the Kern Delta area as well. In parts of the eastern side of the Tule area, water-table rises continued as a result of recharge from the delivery of surface water begun in 1950 through the Friant-Kern Canal and of reduction of pumpage (Poland and others, 1975, p. 46).

The simulated changes in water-table altitude agree well with the observed data (fig. 33*B*), except in a few areas. The model simulates too much decline in the Chowchilla and eastern San Joaquin areas and the area just north of the Sutter Buttes in the Sacramento Valley. The boundaries of the various areas of similar change (decline or rise) are often shifted slightly from their position on the observed map. This is probably because the location of values of recharge and discharge is not precise.

The observed and simulated spring 1976 water-level altitudes in the lower pumped zone and 1961–76 changes are shown in figure 34. Water levels in the lower pumped zone in the Sacramento Valley continued to decline, especially in the areas east of the Feather River, the Cache-Putah area, and the areas just north and south of Sacramento (fig. 34). Two depressions developed in the Delta area with minimum water levels more than 40 ft below sea level (fig. 34).

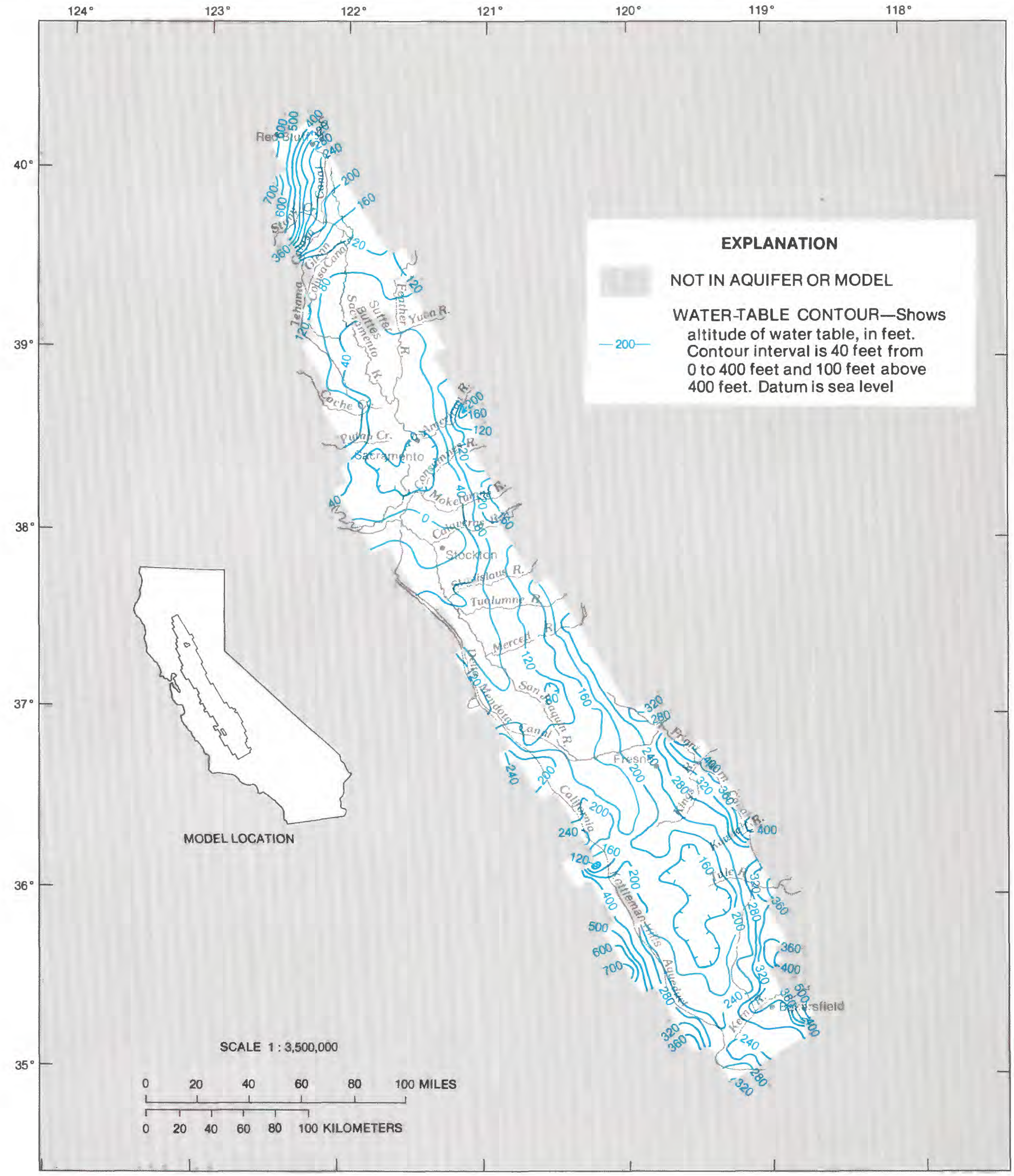


FIGURE 31A.—Spring 1961 water-table altitude.

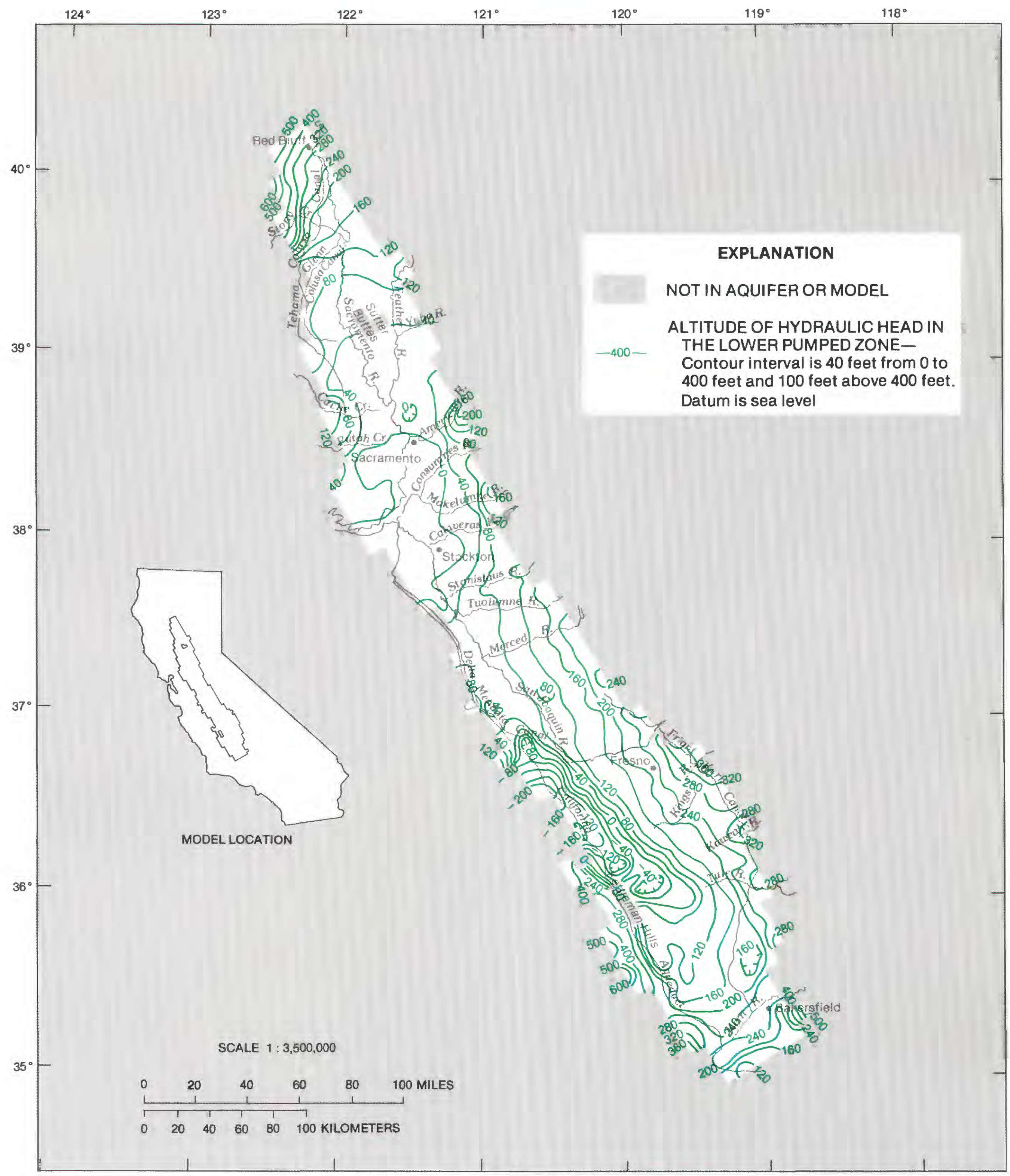


FIGURE 31*B*.—Spring 1961 hydraulic head in the lower pumped zone.

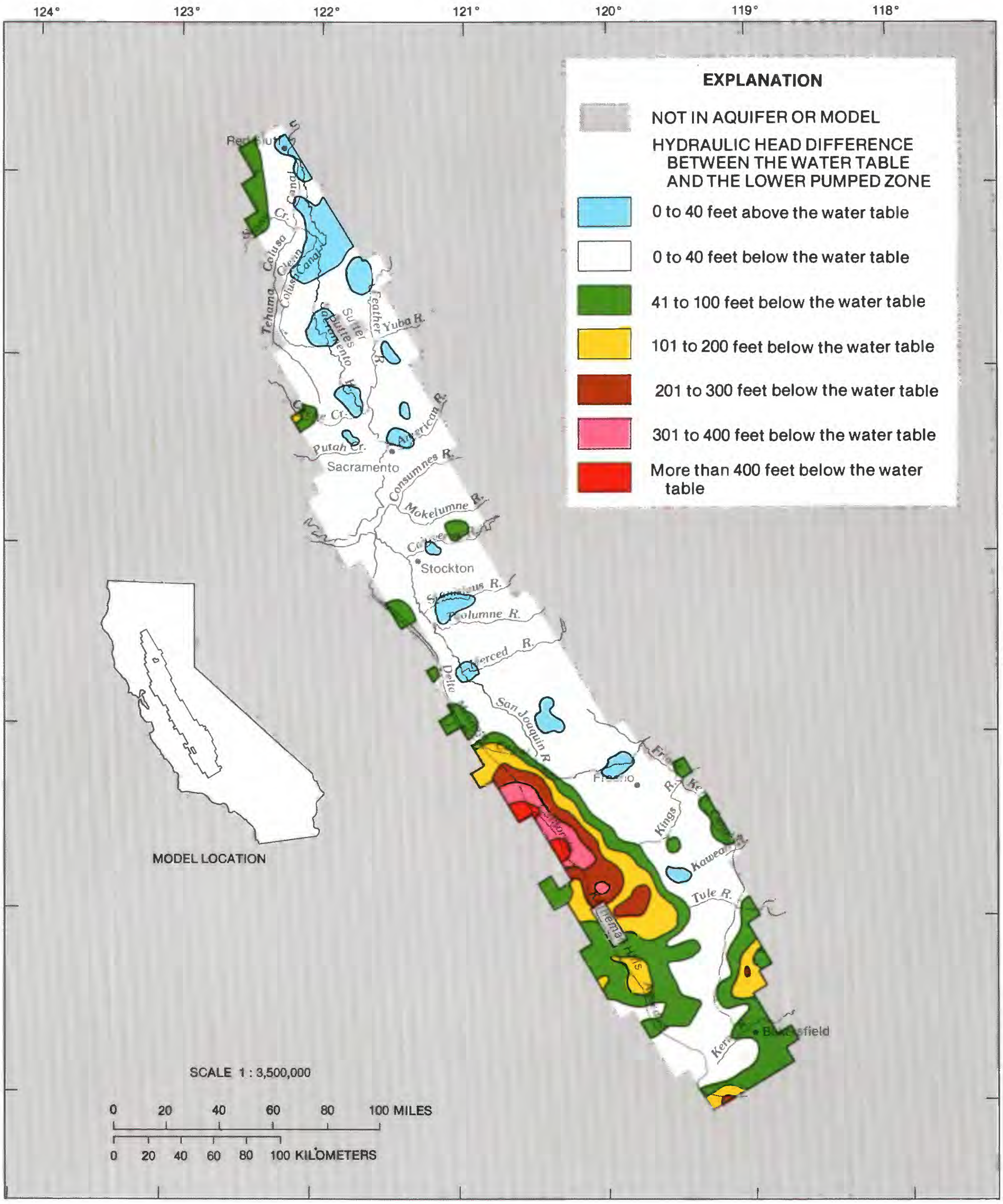


FIGURE 31C.—Spring 1961 hydraulic head difference between the water table and the lower pumped zone.

In some areas of the San Joaquin Valley, lower pumped zone water levels continued to decline whereas other areas showed a reversed trend. In 1967, the California Aqueduct began delivering surface water to farms along the west side and near the southern end of the San Joaquin Valley. Ground-water pumpage began decreasing as farms converted to surface-water irrigation, with the result that water levels in the Westside area rose as much as 200 ft by spring 1976 (Ireland and others, 1984, p. 72). In the western Kings area, just to the east, the decline continued because there was still very little surface water delivered to this area. Also, because most of the wells in this area are perforated through the water table and the lower pumped zone, the two zones react to the pumping stress as one zone. Some of the areas in the east side, where surface water is now being delivered by the Friant-Kern Canal, showed continued water-level rises in the lower pumped zone through the 1960's. Most of Kern County showed a continued or slightly increased decline.

The simulated changes in the lower pumped zone water level also agree well with the observed data (fig. 34), except in a few areas. The model simulated too little decline in the central part of Kern County and the western Kings area. It simulated too much decline in eastern San Joaquin County, apparently owing to an overestimated amount of discharge, because the water table decline was also too large. In the Westside area, the 1961–76 period included a period of moderate decline and a period of large recovery. The average simulated overall rise matched the observed average well but was quite variable, as shown on figure 34. The cause is not known but may be related to the size of the model blocks.

The first year of the 1976–77 drought produced very little surface-water runoff, yet most of the reservoirs were near capacity at the beginning of the season, so that there was little effect on the amount of surface water delivered for irrigation (fig. 22). This was especially true in the areas served by the State Water Project. The operation of the Federal Central Valley Project was more conservative, and, as a result, relatively less water was delivered in 1976 so that relatively more water was left to deliver in 1977 as the drought continued and became more severe. As a result of the drought, many farmers drilled or restored the operation of wells to compensate for anticipated surface-water shortages. The State Department of Water Resources received about 4,500 new drillers' logs for irrigation and municipal wells that were drilled in 1977 and 1978 in the San Joaquin Valley. The total number of wells drilled in the valley was probably larger. Water levels declined substantially all over the valley, as shown in the selected hydrographs of observed and simulated water levels in figure 35. The very steep decline in the lower pumped zone shown in figure 35*H* was caused by a reduction of the amount of water released from compaction during a second period of drawdown for the same head interval. The seasonal decline was much greater than during the 1960's, though the pumpage in the Westside area was only one-half as much.

These hydrographs represent average water levels for a given model block (locations shown on fig. 27) and were selected because they represented different conditions for the valley where substantial data were available. The hydrographs were prepared in the final stages of calibration, therefore prompting little additional calibration of these particular model blocks. The accuracy of the model simulations is shown during the calibration period, 1961–75, and also through the drought, during which time the capabilities of the model were tested.

Rapidly changing water levels at the beginning of a simulation period would indicate that the initial conditions were incorrectly specified. The consistent trends in water-level decline or rise shown in figure 35 suggest that initial conditions were reasonable. Hydrographs for each model block were prepared to check for this problem, and no significant problems were discovered. The hydrographs also allowed comparison of the simulated and observed seasonal water-level fluctuation. This comparison was somewhat hampered because most of the autumn observations were not representative of the lowest water level. The simulated seasonal fluctuation is probably too large (for example, see fig. 35*E*) because of the allocation of the components of recharge and discharge entirely to one season or the other.

The simulated water levels for model blocks in the southern end of the valley did not decline as much as the observed water levels did during the drought. In the Westside area (for example, fig. 35*H*, column 51, row 10), the observed decline during 1977 was very large because water levels had been substantially above the record lows and, therefore, little subsidence occurred and the water levels reacted to the small confined storage coefficient. The model simulated this occurrence, but with a smaller magnitude than the observed data. Also, the hydrograph for column 61, row 7 (fig. 35*J*) shows the observed water table rising slightly and the simulated heads dropping slightly.

CHANGES IN GROUND-WATER FLOW

Changes in ground-water flow are a secondary effect of changing water levels resulting from changes in recharge and discharge owing to development. In a heterogeneous ground-water system like that in the Central Valley, there are changes to vertical and horizontal flow which, though closely interrelated, will be discussed separately, for clarity.

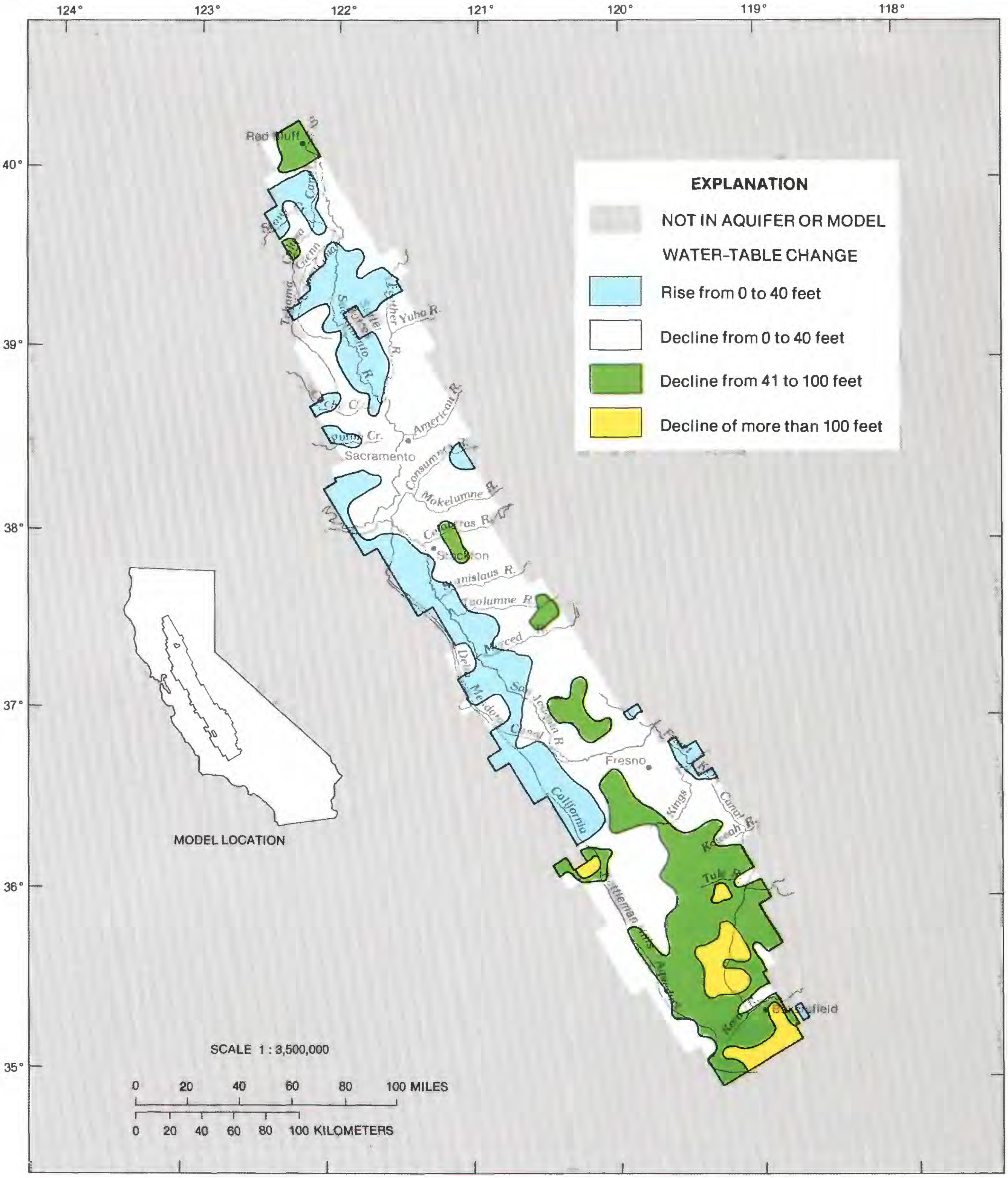


FIGURE 32A.—Change in water-table altitude from 1860 to spring 1961.

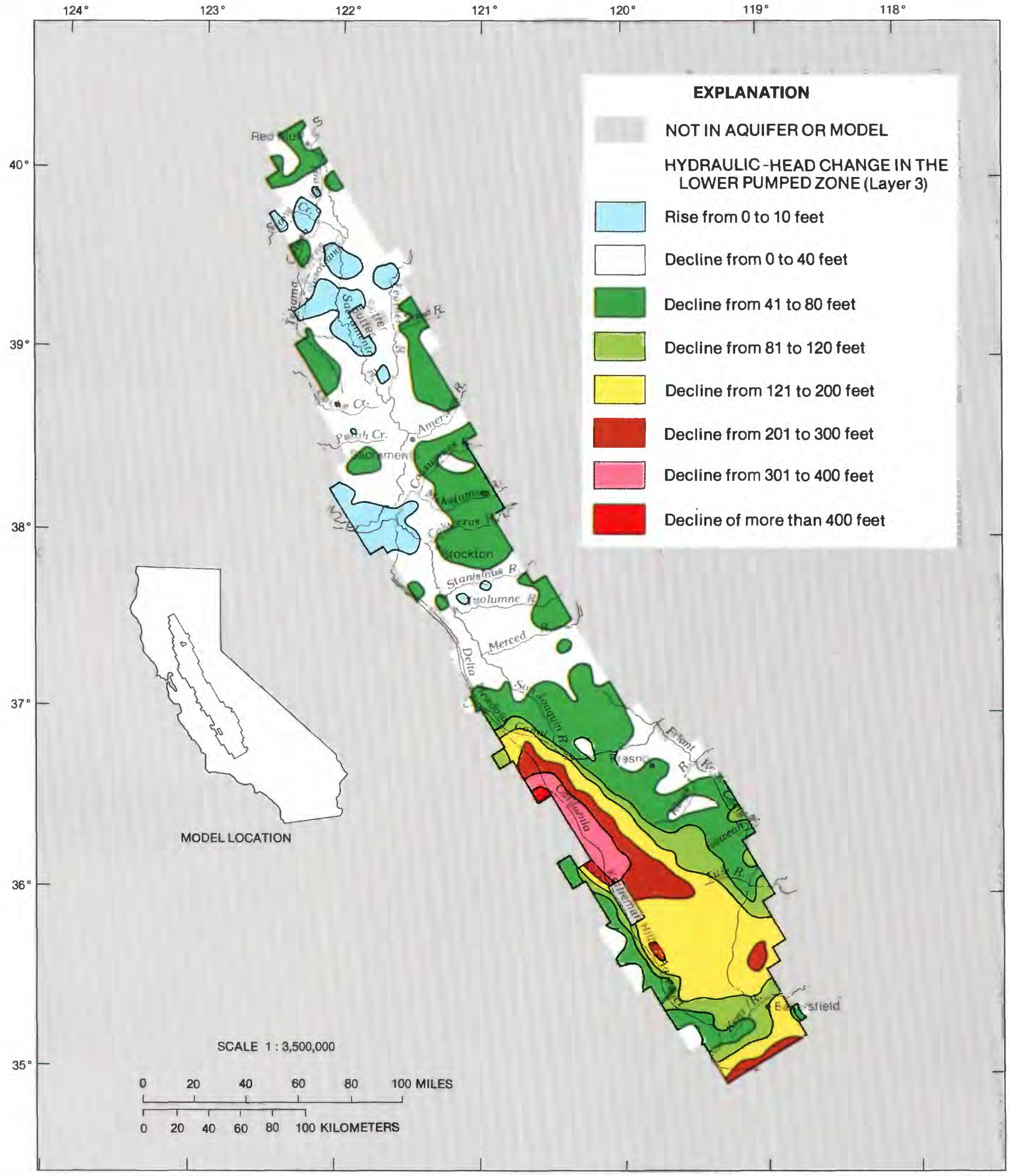


FIGURE 32*B*.—Change in hydraulic head in the lower pumped zone (layer 3) from 1860 to spring 1961 (using observed values for 1961 and simulated values for 1860).

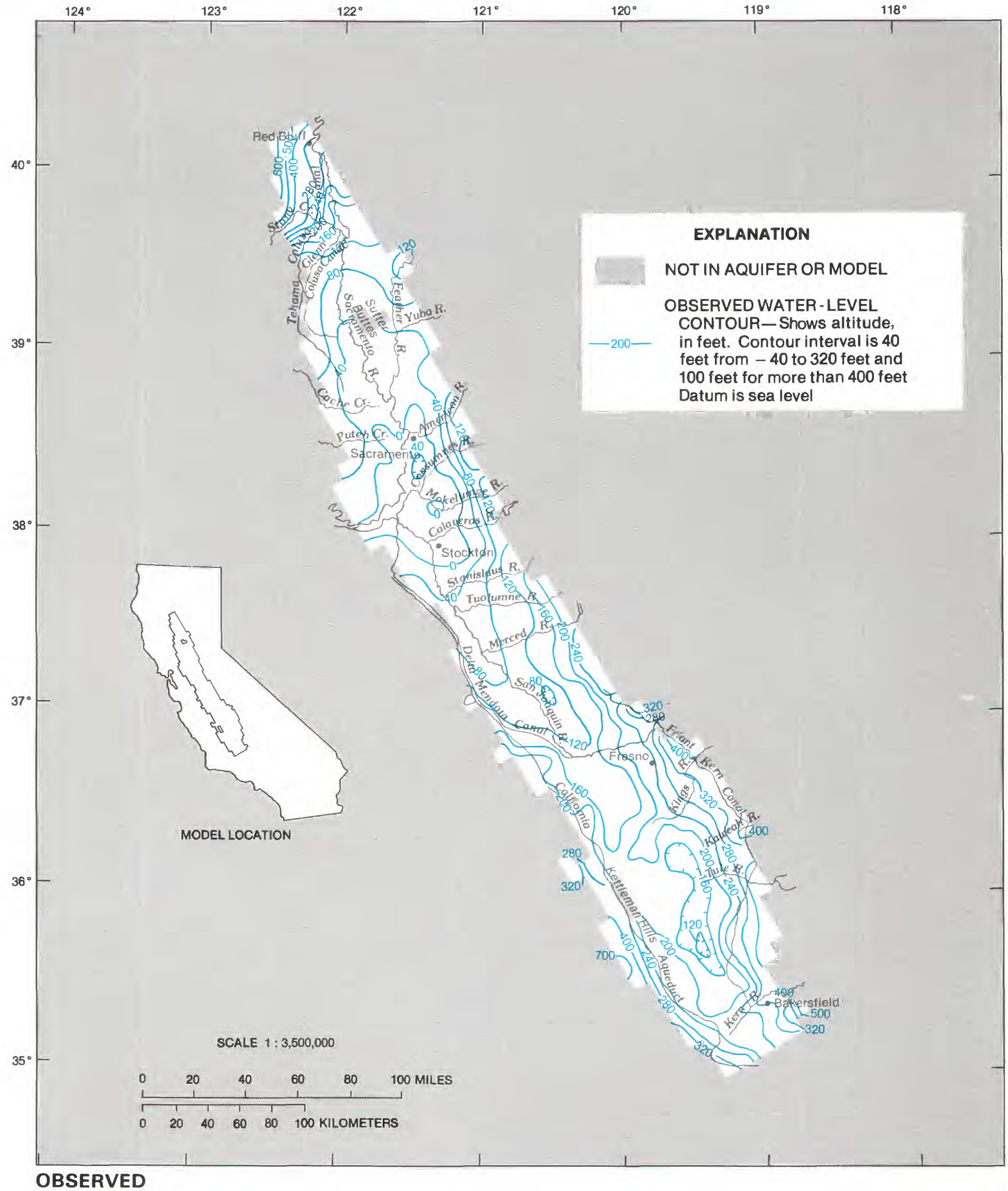
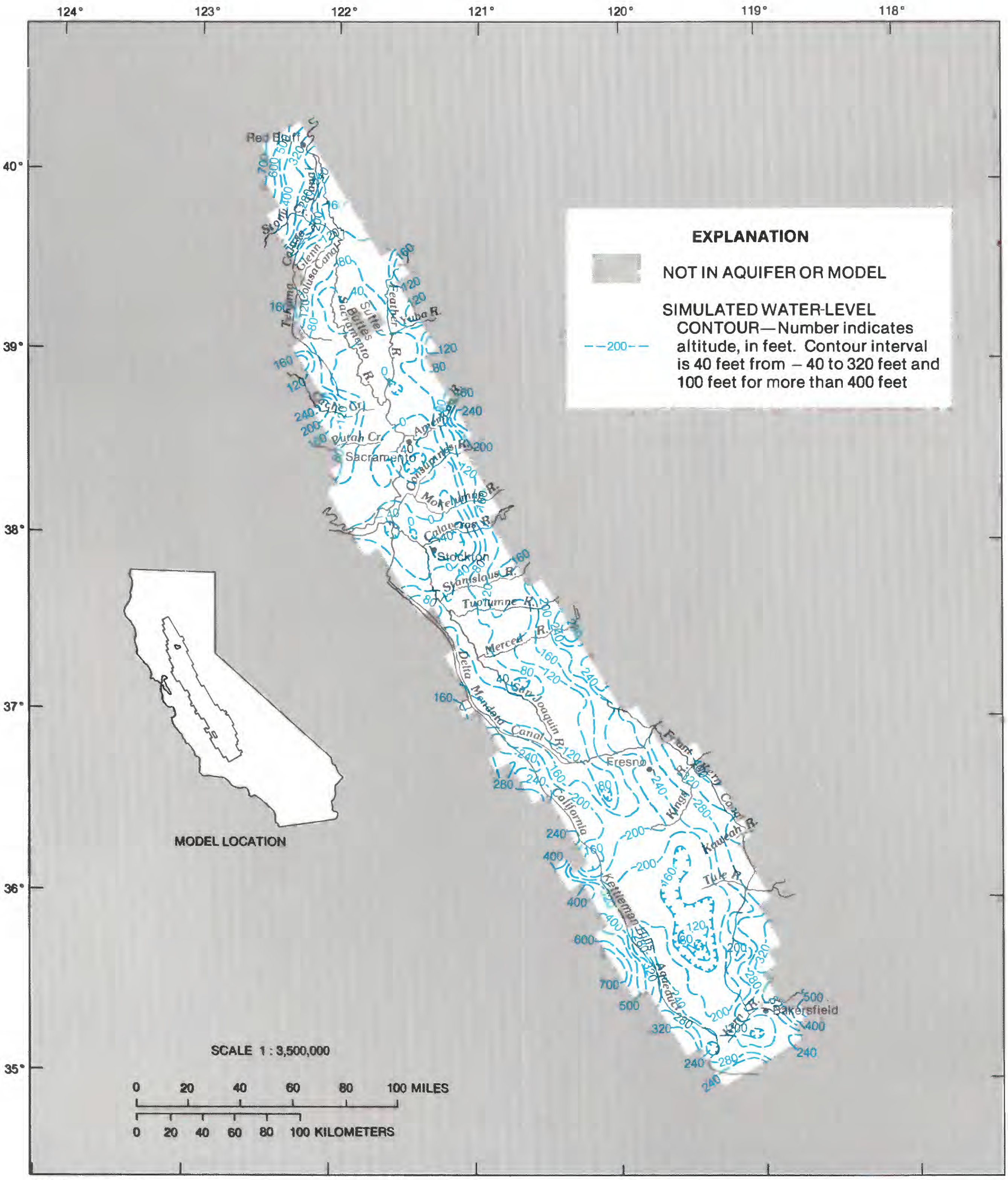


FIGURE 33A.—Observed water-table altitude, spring 1976.



SIMULATED

FIGURE 33A.—Simulated water-table altitude, spring 1976.

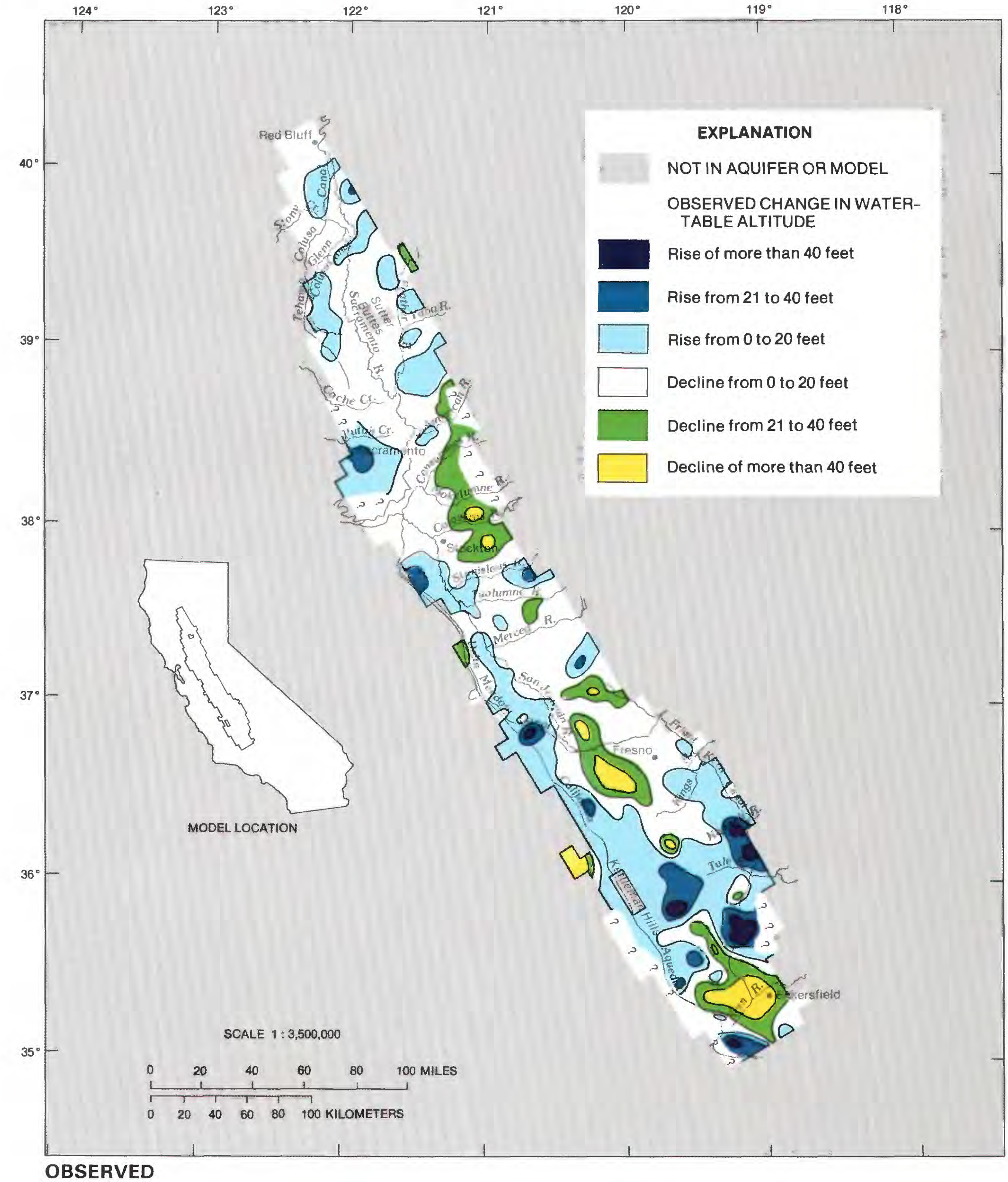


FIGURE 33*B*.—Observed change in water-table altitude, spring 1961 to spring 1976.

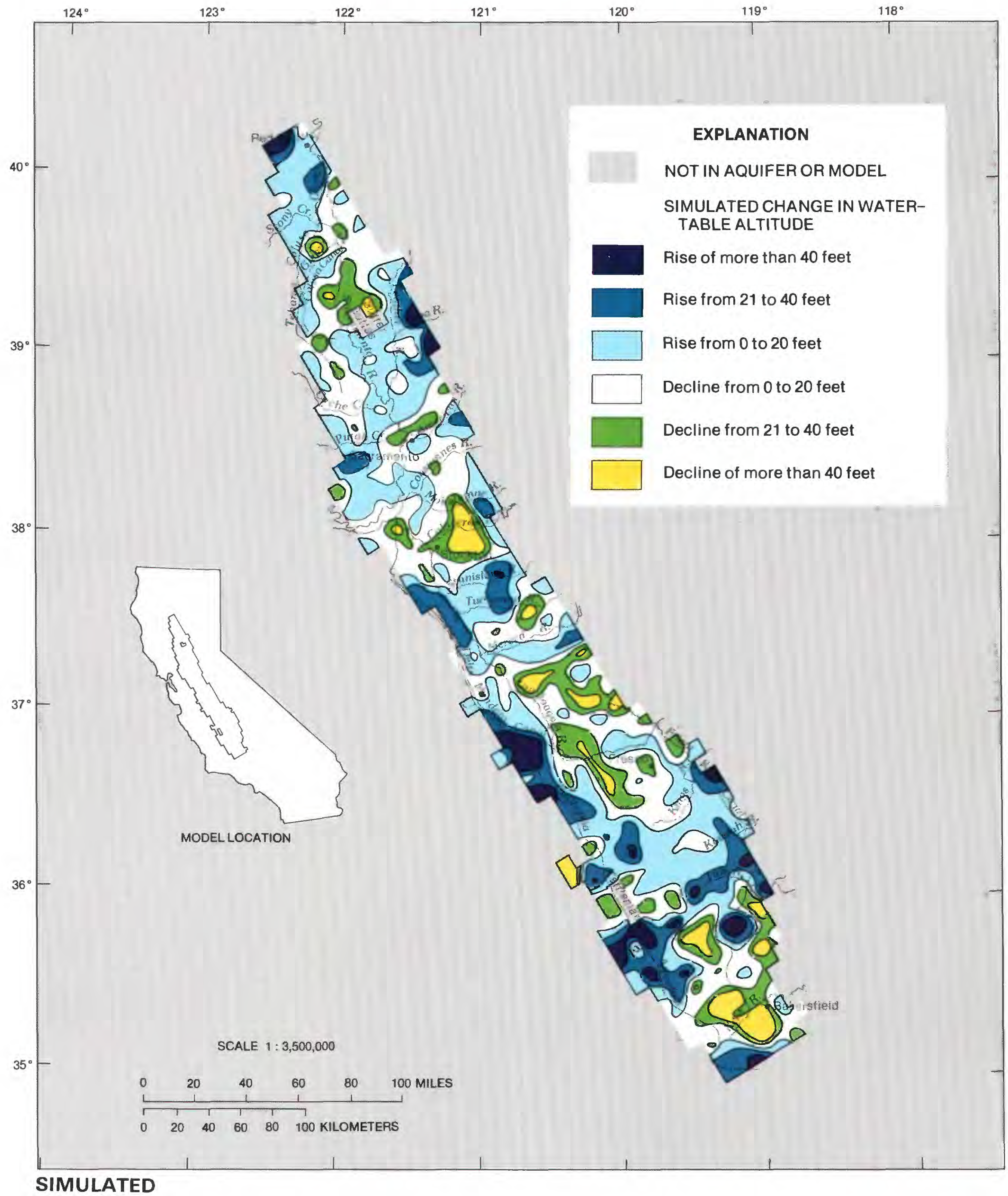
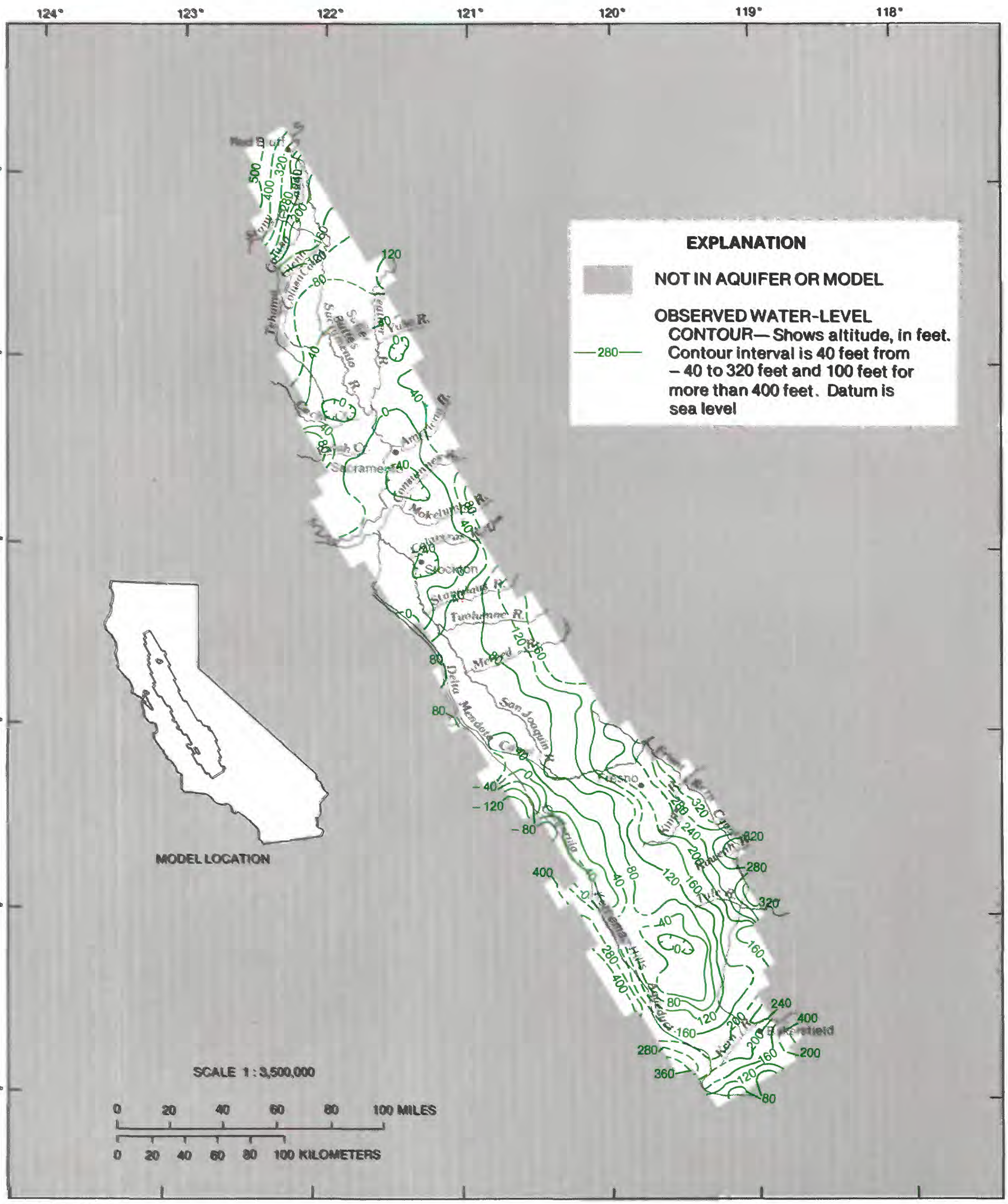


FIGURE 33B.—Simulated change in water-table altitude, spring 1961 to spring 1976.



OBSERVED

FIGURE 34A.—Observed hydraulic head in the lower pumped zone (layer 3), spring 1976.

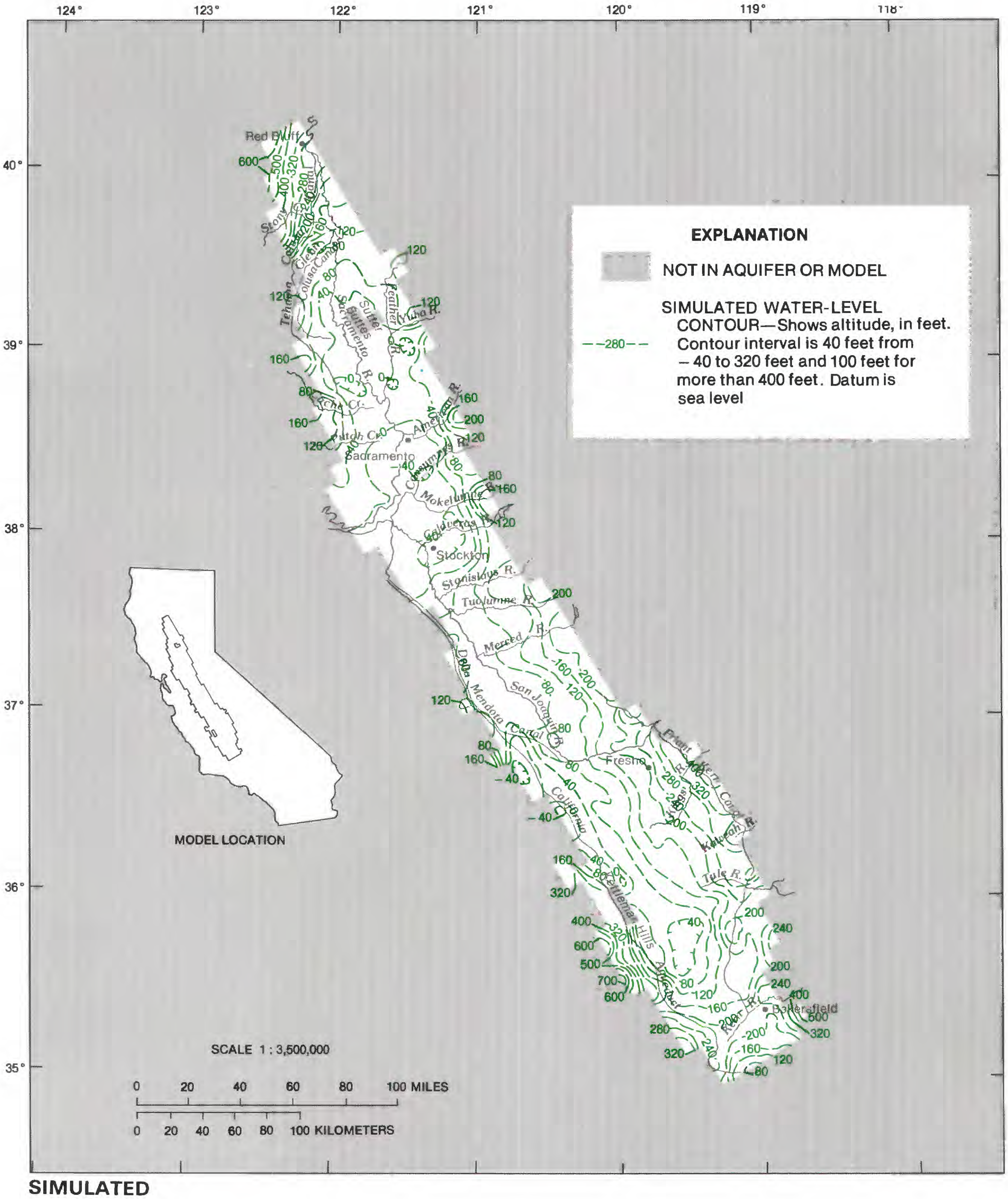
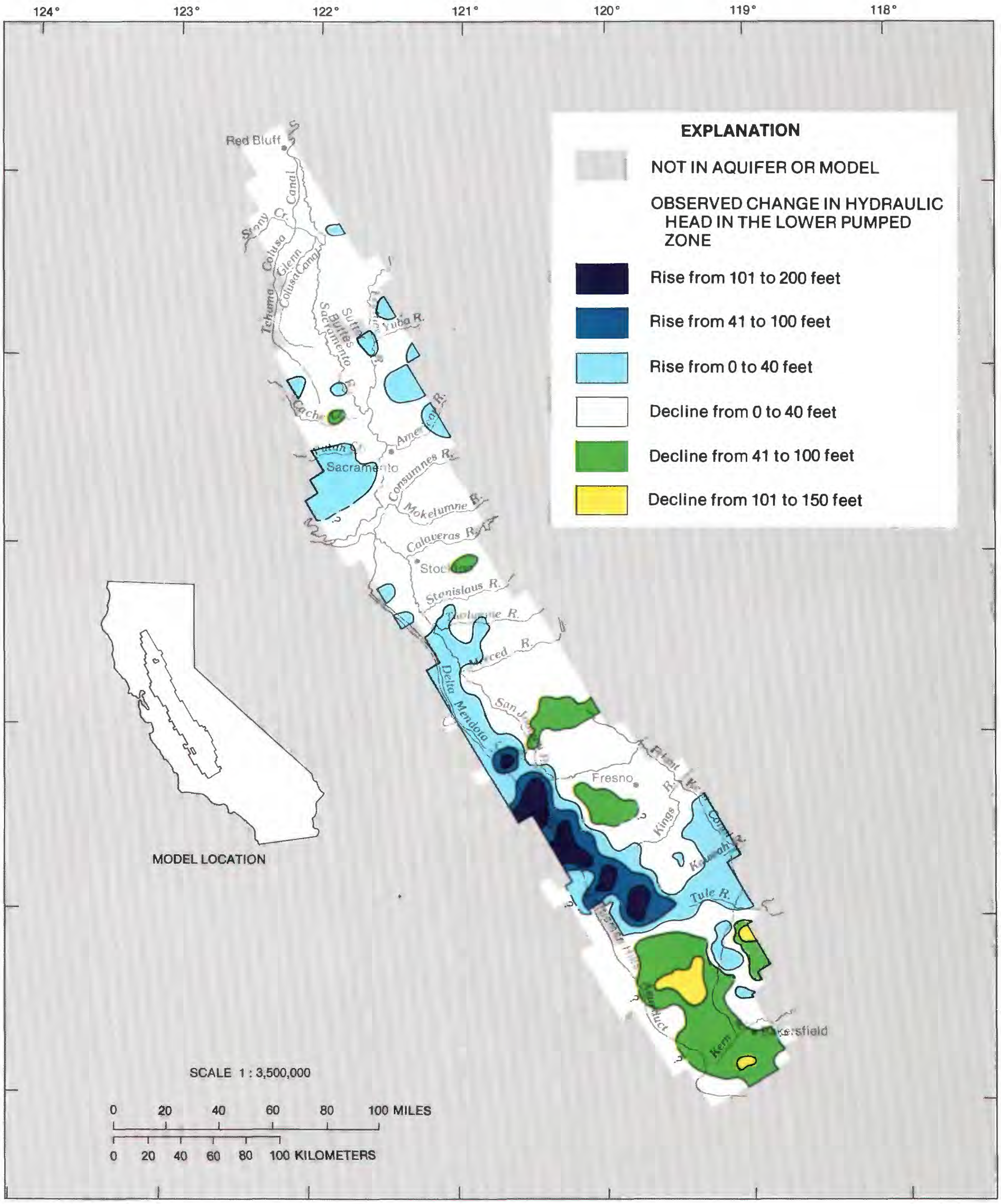
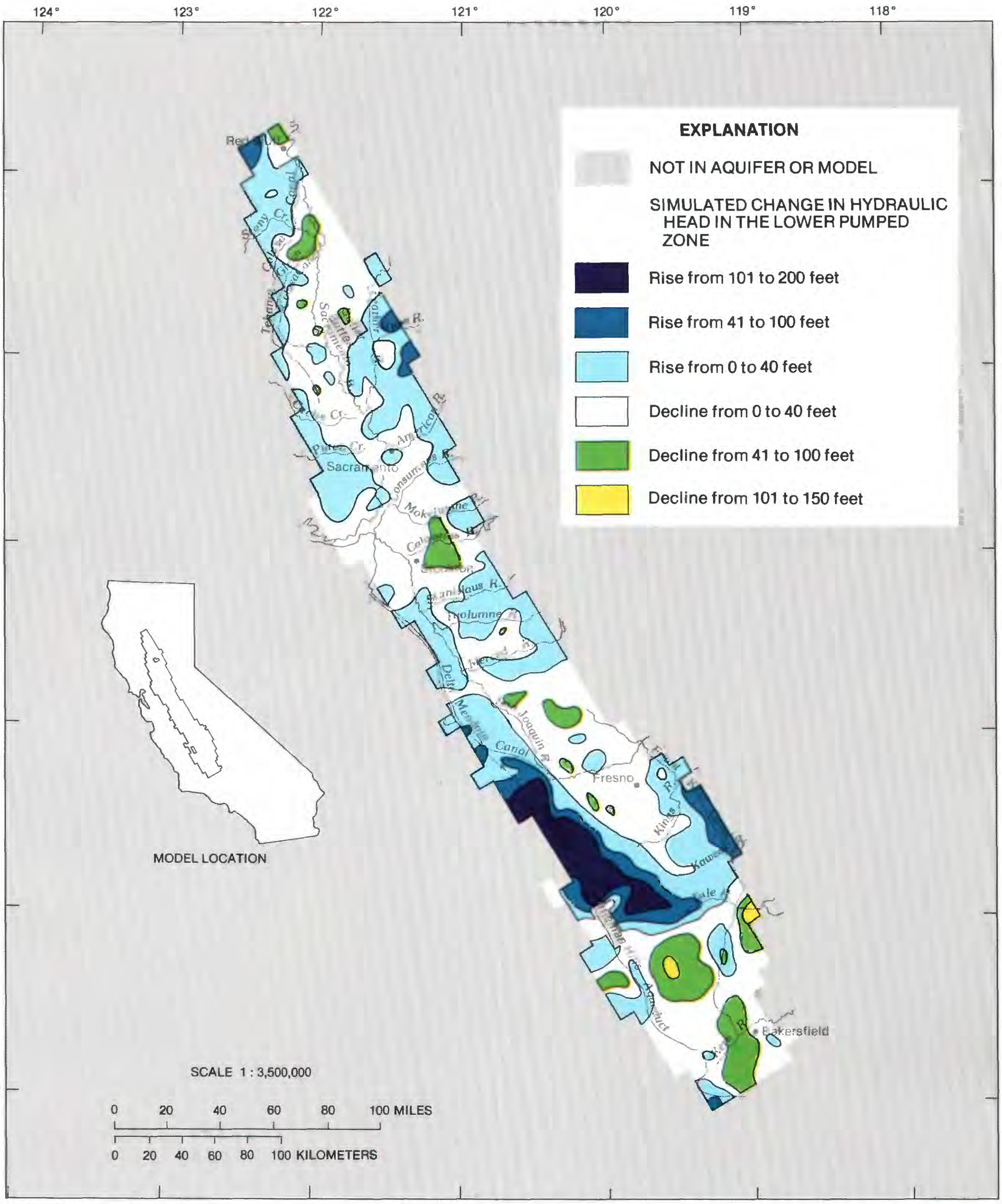


FIGURE 34*A*.—Simulated hydraulic head in the lower pumped zone (layer 3), spring 1976.



OBSERVED

FIGURE 34*B*.—Observed change in hydraulic head in the lower pumped zone (layer 3), from spring 1961 to spring 1976.



SIMULATED

FIGURE 34*B*.—Simulated change in hydraulic head in the lower pumped zone (layer 3), from spring 1961 to spring 1976.

The dramatic change since development in the pattern of flow, especially the location of major ground-water discharge, is shown in figure 25. Before development, the lower pumped zone heads were near the water-table altitudes and flow was toward the Delta because that was the location of the lowest head. By 1961, pumping in the Westside area had lowered water levels enough so that it became a major discharge area, receiving flow from much of the San Joaquin Valley. In this area, heads in the lower pumped zone were far below sea level

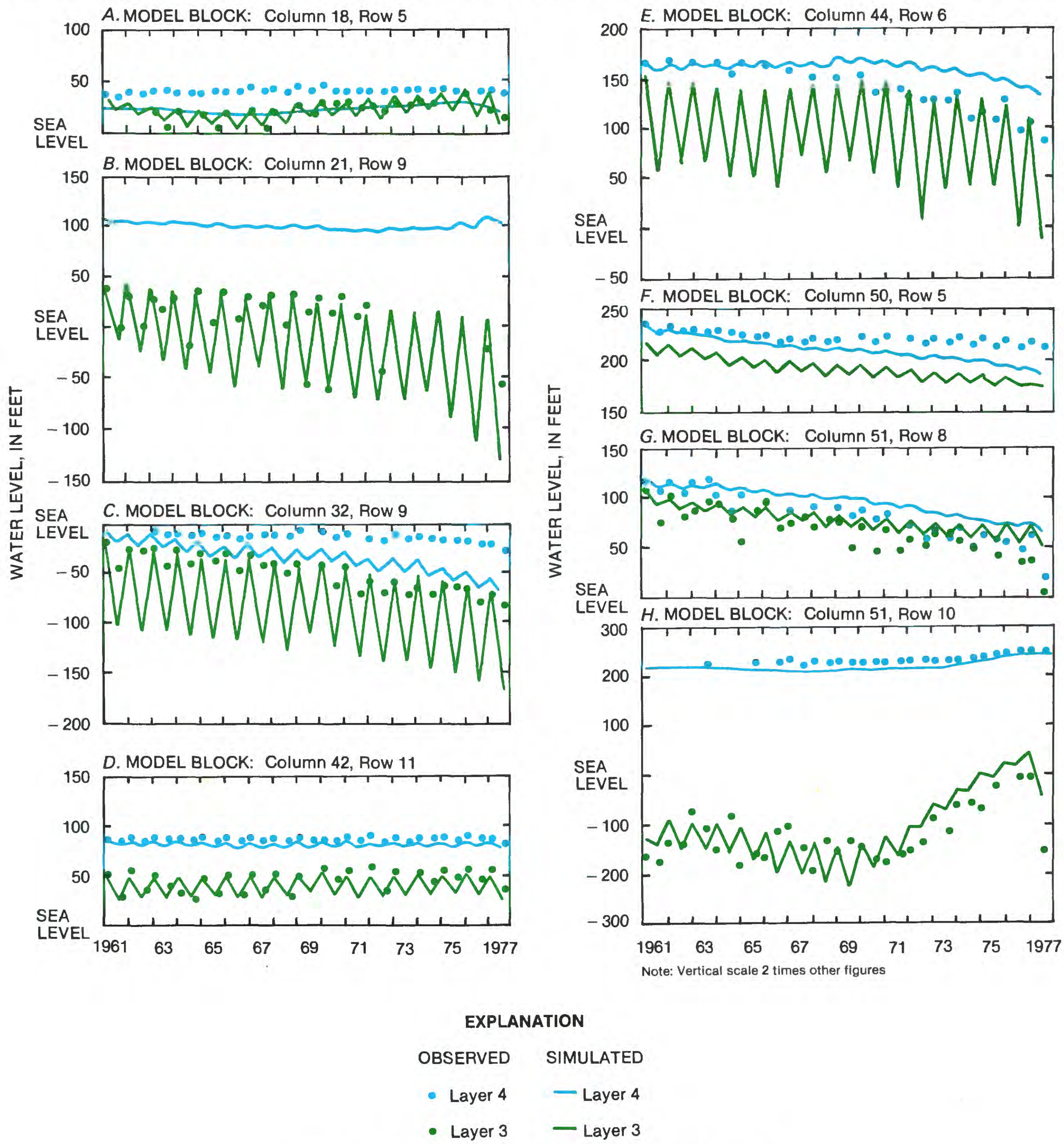


FIGURE 35A–H.—Observed and simulated water levels, 1961–77.

in the early 1960's. Notice the very steep gradient toward this area from all sides (figs. 31*B* and 25),which indicates flow, especially from the east side of the valley toward the west. This large, well-developed depression of water levels in the San Joaquin Valley simplified calibration of the transmissivities for the simulation model. Calibration of transmissivities requires detailed and accurate knowledge of the volumes of recharge and discharge. There is a greater certainty for the estimates of pumpage during 1961–77 than for values

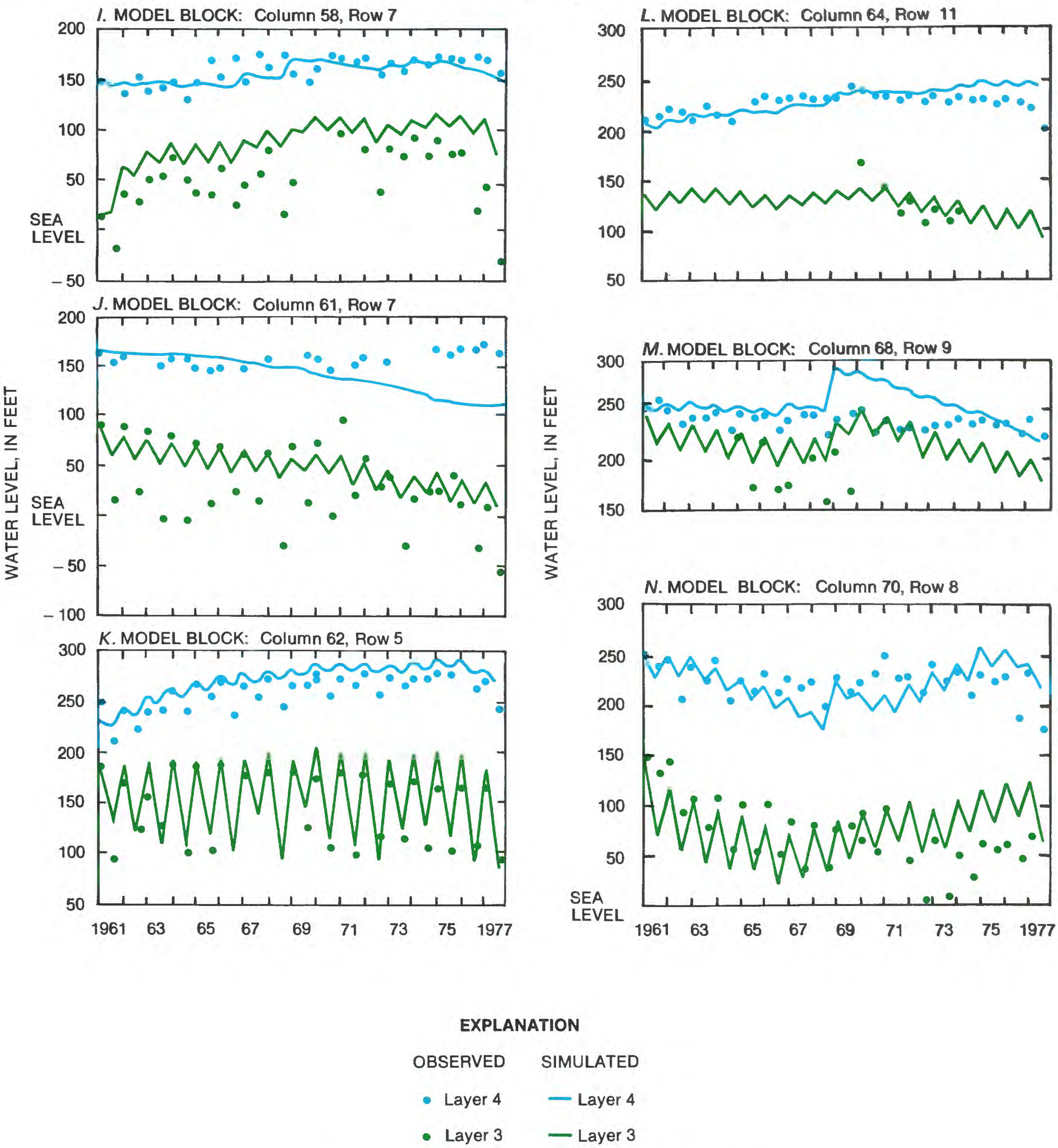


FIGURE 35*I–N*.—Observed and simulated water levels, 1961–77.

of recharge and discharge during predevelopment. In calibrating transmissivities, the relative differences in thickness and permeabilities among areas were preserved, with the factor for the whole set of values being adjusted so that the gradients and the amounts of land subsidence matched observed values. The simulated flow from adjacent areas into the Westside area during the early 1960's accounted for about 13 percent of the ground water withdrawn from the area. The remainder was supplied from inelastic compaction (about 47 percent), leakage from the water table (about 32 percent), and elastic storage and upward leakage from below the lower pumped zone (about 8 percent).

Table 7 shows thickness and hydraulic conductivity (*K*) for all four model layers, and specific yield for the water table. All *K* values shown have been reduced by a factor of 4 as a result of model calibration. Specific yield and *K* values are both related to the coarseness of sediments, which increases toward the south. The average *K* value for the San Joaquin Valley is almost double that for the Sacramento Valley in layers 1 and 2, and about 50 percent larger for layers 3 and 4. This may be a result of the higher proportion of finer grained volcanic sediments in the Sacramento Valley. The larger proportion of fine-grained sediments may also mean that there is significant potential for future land subsidence in the Sacramento Valley if enough pumpage develops at depth in some locations. The areas that have large alluvial fan deposits (especially Kings and Kern Delta) have the largest *K* values. The smallest values are found in the flood plains and along the west side of the Central Valley.

To study changes in flow conditions before and after development, the authors used simulations to calculate

TABLE 7.—*Summary of specific yield, thickness, and hydraulic conductivity values*

[Totals may not agree because of rounding]

Subarea No. (See fig. 27)	Specific Yield	Thickness (feet)				Horizontal hydraulic conductivity after model calibration (ft/d)				Volume of water in storage (1961) to depth ≤ 1000 feet (million acre-ft)
		Layer 1	Layer 2	Layer 3	Layer 4	Layer 1	Layer 2	Layer 3	Layer 4	
11 ----	0.077	759	227	245	174	3.0	3.1	4.4	4.9	43
12 ----	0.062	423	414	235	200	2.0	2.0	3.0	3.7	14
13 ----	0.074	487	301	304	246	4.1	4.1	5.0	4.7	27
14 ----	0.072	870	340	340	219	2.3	2.3	4.7	4.8	25
15 ----	0.079	670	609	313	220	2.0	2.0	3.0	5.5	14
16 ----	0.074	45	179	233	228	3.0	3.2	4.9	4.7	17
17 ----	0.081	1,029	491	421	321	3.4	3.4	5.2	6.0	26
Sacramento --	0.074	622	328	292	223	2.9	3.0	4.5	4.9	166
21 ----	0.076	580	367	267	237	1.8	1.8	4.1	5.0	40
22 ----	0.080	1,050	377	282	228	2.0	2.0	4.1	5.3	21
23 ----	0.084	965	418	257	243	3.7	3.7	5.1	5.6	51
24 ----	0.103	1,925	439	315	175	2.3	2.3	4.4	7.9	22
Delta -------	0.084	998	398	273	228	2.6	2.6	4.5	5.7	134
31 ----	0.098	507	595	238	191	4.6	4.6	5.8	6.1	14
32 ----	0.112	1,205	519	370	207	5.4	4.6	5.5	9.1	51
33 ----	0.093	1,148	546	268	199	4.9	4.9	6.3	6.8	23
34 ----	0.097	1,114	404	293	119	5.3	5.3	6.5	6.9	37
35 ----	0.090	1,562	434	498	201	3.9	3.9	4.6	6.1	15
36 ----	0.096	921	696	360	219	5.7	5.7	6.4	7.3	24
San Joaquin -	0.100	1,094	522	333	185	5.2	4.9	5.9	7.5	164
40 ----	0.099	878	356	404	213	6.7	3.7	5.5	7.1	4
41 ----	0.103	2,234	908	1,073	267	7.1	3.6	4.2	7.5	52
42 ----	0.113	1,734	984	319	281	6.9	6.9	8.1	9.4	93
43 ----	0.083	1,328	802	696	576	7.2	6.7	8.6	7.1	37
44 ----	0.109	1,147	803	507	266	6.4	6.7	7.5	8.6	34
45 ----	0.085	1,339	832	642	306	5.0	5.0	6.4	5.5	33
46 ----	0.090	163	501	461	356	3.8	4.2	3.9	5.1	15
47 ----	0.094	1,141	950	746	322	5.6	5.5	6.3	6.4	42
48 ----	0.124	3,437	1,015	688	379	6.8	6.8	9.0	10.1	42
49 ----	0.124	1,530	856	846	306	7.5	7.1	7.7	10.0	13
Tulare ------	0.101	1,488	835	614	331	6.2	5.7	6.7	7.6	365
Central Valley ----	0.092	1,121	578	424	260	4.6	4.3	5.6	6.6	828

the amount of flow across each block face. Owing to the difficulty of summarizing the changes in flow across the great number of block faces, the flows are summarized in cumulative frequency distributions to compare them. The downward flow across a block face is assigned a negative sign, and the upward flow is assigned a positive sign. Because there are four block faces in a horizontal plane, the flow direction cannot be meaningfully summarized; therefore, the authors grouped the calculated horizontal flows by magnitudes but without consideration of flow direction. The authors also calculated flow velocity in both horizontal and vertical directions by dividing the flow quantity by the product of the respective block face area and an assumed effective porosity of 30 percent. The cumulative frequency distributions of flow quantity and flow velocity are shown in figures 36*A*–*H*.

Figure 36*A* suggests that the amount of vertical flow was balanced between upward and downward flow before development. This is required under the assumption of steady-state flow conditions before development. In this situation, the long-term recharge was equal to discharge; therefore, the downward flow in recharge areas was balanced by upward flow in discharge areas. However, this balanced flow condition in the vertical direction was changed by development. Figure 36*E* shows the distribution of vertical flow during simulation of 1961 flow conditions. Most of the pumping in the Central Valley in 1961 was located in layers 3 and 4; therefore, the amount of downward flow from surface-water bodies to layer 4 (a water-table aquifer) and from layer 4 to layer 3 was increased by an order of magnitude greater than that of the predevelopment amounts. The downward flow from layer 3 to layer 2 and from layer 2 to layer 1 was reduced somewhat. The upward flow from layer 3 to layer 4 and from layer 4 to surface-water bodies was also reduced, and the upward flow from layer 1 to layer 2 and from layer 2 to layer 3 was increased (figs. 36*A* and *E*). This indicates that pumping has induced recharge and has captured natural discharge. One interesting point is that in a very small area there was more downward flow from layer 3 to layer 2 during development than during predevelopment (17 acre-ft/yr versus 5.7 acre-ft/yr, figs. 36*A* and *E*). This probably was caused by inducing more recharge from upper layers owing to pumping; thus, there was more water recharging into layer 2 from layer 3.

The amounts of horizontal flow reveal more interesting points. About one-half of the total block faces in the horizontal direction have very little flow, as indicated by figure 36*B*, because the block faces parallel to the main flow direction have little horizontal flow. The amount of horizontal flow in layer 3 was increased by pumping; however, horizontal flow in layer 4 shows very little effect of pumping even though there were wells in that layer. This probably was due to plenty of recharge to layer 4 (a water-table aquifer), and because the pumping in layer 4 was fairly evenly distributed valleywide. On a regional scale there probably was little change in the magnitude of the hydraulic gradient in layer 4 before and after development. The interesting point is that the change in horizontal flow in layer 2 was the same magnitude as the change in layer 3 (figs. 36*E* and *F*), even though there was little pumping in layer 2. The probable explanation is that after development more downward flow was induced by pumping in recharge areas from layer 3 to layer 2, as suggested by figure 36 *E*. This increased downward flow moved horizontally and flowed upward in pumping or natural discharge areas (fig. 36*E*). Because there was very little horizontal flow in layer 1, the cumulative frequency curve would not show on the scale chosen to present flow for the other layers.

Figures 36*A*, *B*, *E*, and *F* suggest that the magnitudes of flow in the vertical direction are much larger than those in the horizontal. Yet the horizontal flow velocities are larger than the vertical flow velocities (figs. 36*C* and 36*D*). This contrast in flow magnitudes and flow velocities is due to the geometry of the aquifer and its discretization for simulation. The flow area for vertical flow across horizontal planes is much greater than the area for horizontal flow across vertical planes. This length of the flow paths for vertical flow is much shorter than the length of the flow paths for horizontal flow. The magnitudes of flow are proportional to the area of flow and are inversely proportional to the length of the flow paths. Therefore, even though horizontal permeabilities are mcuh larger than vertical permeabilities, vertical flows on a regional scale can be very large. On a local scale, of course, the flow near a well is mostly horizontal.

Factors Affecting Vertical Flow

Water development has changed vertical flows due to (1) changes in the direction and magnitude of the vertical hydraulic gradient caused by changes in recharge and discharge, (2) an increase in the effective or apparent values of vertical leakance (vertical hydraulic conductivity divided by thickness of the layer) caused by wells with long lengths of perforated openings connecting adjacent layers (Bennett and others, 1982), and (3) a possible decrease in vertical leakance caused by compaction of sediments (Helm, 1976, p. 389).

The vertical hydraulic gradient changed dramatically from predevelopment to 1961, as can be seen by comparing figures 15 and 31*C*. Under predevelopment conditions, the vertical gradient was downward around the margins of the valley and upward in the center. Model

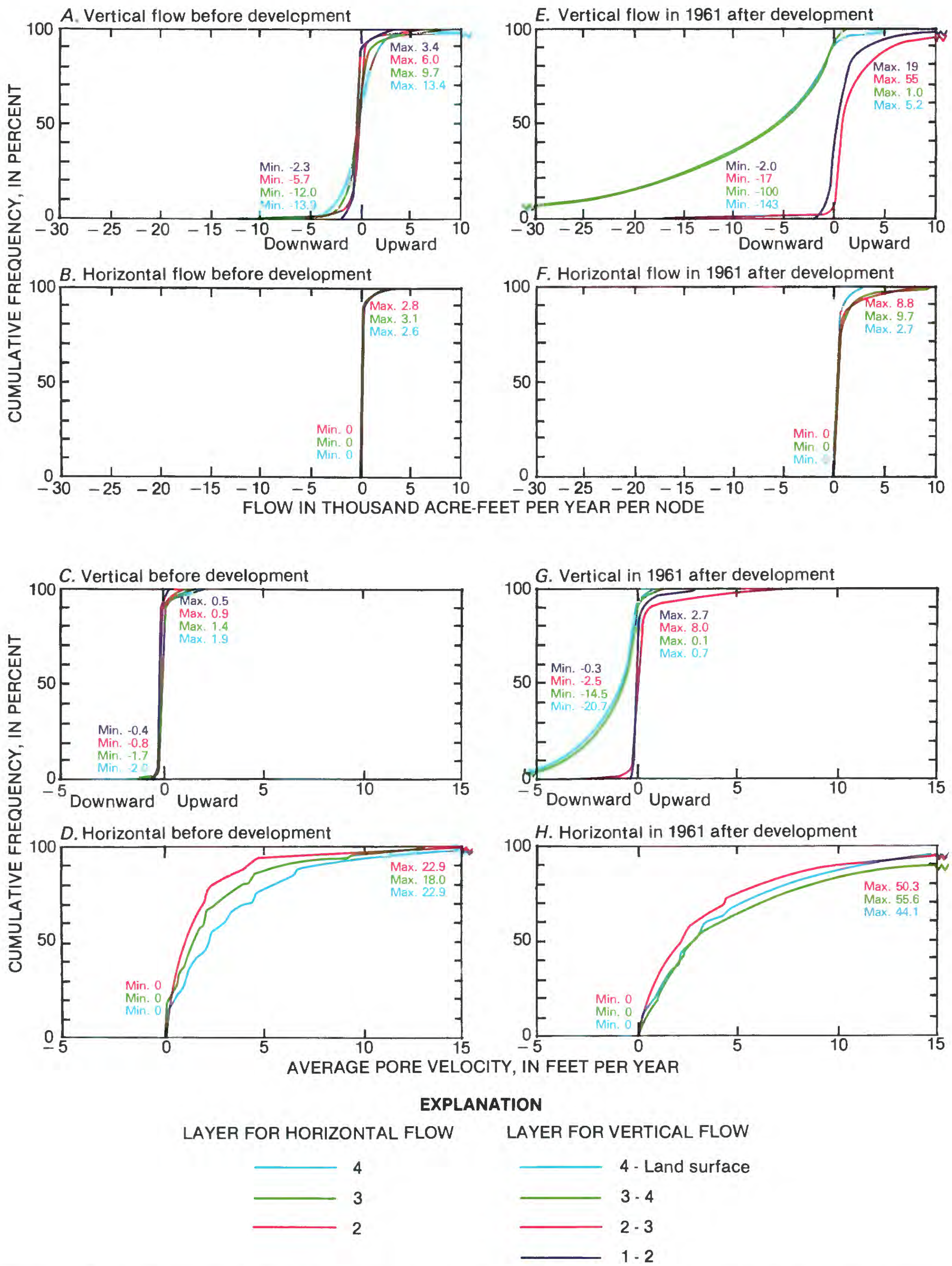


FIGURE. 36*A–H*.—Variation in horizontal and vertical flows and average pore velocities during predevelopment and 1961 flow conditions. Layer 1 horizontal flows are too small to be shown at this scale.

simulations indicate that the predevelopment head difference between water-table altitudes and water levels in the lower pumped zone was always less than 85 ft and generally less than 25 ft. Irrigation development had two effects on this head difference. First, canal losses and deep percolation of water from irrigated fields added to the recharge of the water table, which caused water-table rises in several areas. Second, ground-water pumpage, about one-half of which was withdrawn from the lower pumped zone (layer 3), increased the discharge from the deep zone. The cumulative effect of these development impacts was to reverse the head gradient in the center of the Central Valley so that the head gradient was in a downward direction almost everywhere. Some exceptions where the head gradient is still upward in test holes with multiple piezometers are in the center of the Sacramento Valley at Zamora (12N/1E-34Q, fig. 2) and Butte City (12N/3E-2G, fig.2), (French and others, 1982,1983b).

Vertical leakance values were determined largely by model calibration. The division of the aquifer system into layers was planned to minimize the complexities of model calibration of leakance which is affected by multilayer wells. Where possible, layer boundaries were chosen so that the perforated interval of most wells was entirely within one layer. Where this was not possible, boundaries were chosen so that perforated intervals of wells would span no more than two adjacent layers. This occurred between layers 3 and 4 in several areas of the valley. In the Westside area, most well perforation intervals spanned most of layers 2 and 3, but very few spanned layers 3 and 4. The vertical leakance used in the predevelopment simulations should reflect only the undisturbed characteristics of the sediments. The leakance used in postdevelopment simulations could reflect substantial alterations due to interconnection by multilayer wells and also due to compaction of sediments. Using hydraulic parameters and well densities typical in the Central Valley, trial calculations show that the effect of multilayer wells should dominate the postdevelopment values of leakance and should result in a significant increase in that value over predevelopment conditions.

Calculations of the multilayer well effect were made using the method of Bennett and others (1982) (see appendix C). Using values typical of the Central Valley, these calculations indicate that an irrigation well connecting two vertically adjacent model blocks should have a vertical hydraulic conductance on the order of 800 ft^2/day. In contrast, the natural vertical hydraulic conductance (leakance times the area of the block) between the centers of two adjacent model blocks should be about 4,000 ft^2/day. Thus, in areas where the density of wells reaches one per square mile (or 36 per block), the vertical hydraulic conductance provided by the wells could be expected to be roughly seven times the natural vertical hydraulic conductance of the clay beds between two adjacent layers. The total hydraulic conductance is equal to the sum of the two sources of conductance and therefore would be about eight times its predevelopment value. However, if the leakance of the sediments were significantly reduced by compaction, as Helm (1976) indicates is possible, the contribution of natural conductance under postdevelopment conditions could be reduced from small to negligible.

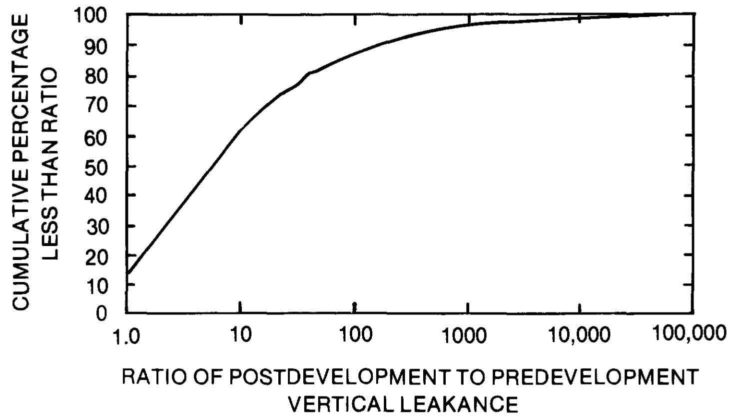


FIGURE 37.—Ratio of postdevelopment to predevelopment vertical leakance between layer 4 and layer 3 in 51 model blocks where predevelopment heads could be estimated. See figure 15 for block locations.

In general, the calibration results support the inferences developed from the trial calculations described above. Figure 37 shows the comparison of calibrated postdevelopment and predevelopment leakance values between layers 3 and 4 in 51 model blocks where predevelopment heads could be estimated. The locations of these model blocks are shown in figure 15. In 44 model blocks the postdevelopment leakance was higher, while in 7 blocks it was lower. The median ratio of the postdevelopment to predevelopment leakance was about 6. Thus the median value agrees reasonably well with the trial calculations of the effects of well interconnections.

This analysis of leakance also indicates that in the Westside area of the San Joaquin Valley in the 1960's the flow of water from the water-table zone (layer 4) down to the lower pumped zone (layers 2 and 3) as described by Davis and others (1964, pp. 81–88), must have been circulation within and between layers 2 and 3 instead. Using the estimated hydraulic conductance of multilayer wells discussed above, 1,000 active irrigation wells as estimated by Davis and others, and the vertical head difference of 400 ft between layers , which was common in the early 1960's in the Westside area, the estimated flow through the multilayer wells would be about 10 times the leakage simulated in this study. This does not count leakage that could have occurred through the 2,000 abandoned wells. This volume of leakage would have dissipated the vertical head difference between layers 4 and 3 to about one-tenth of the observed difference. In contrast, the head differences that occurred within the lower pumped zone during the pumping season, due to unequal pumping stresses, were on the order of 40 ft, which is consistent with the well conductance estimates. Furthermore, nearly all of the non-zero current meter measurements made in the 1964 study were at depths well within the lower pumped zone (Davis and others, 1964, table 13, p. 84).

LAND SUBSIDENCE

The extent and magnitude of land subsidence in the San Joaquin Valley that exceeded 1 ft from 1926 to 1970 is shown in figure 38*B*. Comparing this figure with figure 17, which shows the area of the Corcoran Clay Member and areas of flowing wells in the late 1800's, it is noted that land subsidence occurs mostly where the clay exists. Poland and others (1975, p. H8) separated the subsidence area into three areas (fig. 38*A*): (1) the Los Banos-Kettleman City area west of Fresno, where a maximum subsidence of 29.6 ft was observed in 1977 (Ireland and others, 1984); (2) the Tulare-Wasco area between Fresno and Bakersfield, which includes two areas where subsidence has exceeded 12 ft; and (3) the Arvin-Maricopa area 20 mi south of Bakersfield, where maximum subsidence exceeded 9 ft as of 1970.

Man-induced subsidence in the Central Valley probably began in the middle to late 1800's when the peat soils of the Sacramento-San Joaquin Delta were drained for cultivation. Weir (1950) noted that in 1922 the entire Delta area was in cultivation, and that farmers in the area were concerned about subsidence. Weir also estimated that subsidence in the lower Jones Tract was 4½ ft between 1902 (when the tract was first drained) and 1917. This type of subsidence is caused mainly by the oxidation and compaction of the organic peat soils since the lands were drained (Weir, 1950; Newmarch, 1981). The peat lands had to be drained in order to cultivate, which meant that the water table had to be lowered. The draining of the lands is done by a series of ditches that drain to a central location, from where the water is pumped out into the nearby surface channels. During the summer growing season, water is siphoned back into these same ditches to raise the water level in the ground to within the root zone. However, because the land continues to subside, the water table must continually be lowered. The volume of water removed from storage in this area is equal to the specific yield times the change in the water table because the removal of water is more a function of draining the sediments than of water being released from compaction.

Subsidence caused primarily by compaction of the fine-grained sediments in the aquifer system began in the San Joaquin Valley in the middle 1920's. However, the cumulative volume of subsidence and hence the volume of water released from compaction remained small until after World War II (Poland and others, 1975). Subsidence in the Sacramento Valley presumably began in the early 1950's, although data are sparse (Lofgren and Ireland, 1973). This type of subsidence caused problems, such as cracks in road and canal linings, changing slopes of water channels, and ruptured well casings. During the early 1960's, in parts of the Westside area, large and expensive irrigation wells had a useful life of about 7 years because of casing failures.

Figure 39 shows the cumulative volume of subsidence in the San Joaquin Valley. The total volume of subsidence in the San Joaquin Valley by 1970 was 15.6 million acre-ft (Poland and others, 1975, p. H9). Also included in figure 39 are cumulative volumes of subsidence for each of the three major subsiding areas. The volume of subsidence in the Los Banos-Kettleman City area (fig. 38*A*) accounted for nearly two-thirds of the total volume of subsidence as of 1970. From 1970 through 1975 there was little subsidence in this area because of surface-water imports from the California Aqueduct, which greatly reduced the amount of ground-water pumpage. However, subsidence recurred during the drought of 1976 through 1977 owing to an increase in ground-water withdrawal. In addition to the cumulative volume of subsidence, ground-water pumpage was also plotted for the Los Banos-Kettleman City area. The correlation between pumpage and the volume of subsidence is good, indicating that about one-third of the water pumped was derived from compaction of the aquifer system (Poland and others, 1975). The pumpage, however, included all pumpage in the area (both shallow and deep). Bull and Miller (1975) estimated that at least 75 to 80 percent of the water pumped came from the lower pumped zone. Assuming that compaction occurs only in the lower zone, about 43 percent of the water pumped from the lower pumped zone came from compaction of the fine-grained beds. Similar comparisons of water pumped versus volume of subsidence from 1926 to 1970 were not done in the Tulare-Wasco or Arvin-Maricopa areas, mostly because of the absence of pumpage data and partly because the relation between pumpage and subsidence is not as pronounced, as discussed in the section, "Factors that Affect the Relation of Subsidence to Pumpage."

Observed land subsidence in the San Joaquin Valley reported by Poland and others (1975) and Ireland and others (1984) was primarily dependent on periods when detailed leveling lines were made in the areas of major land subsidence. However, the level lines were not always measured during the same years for each of the major subsiding areas. The last detailed leveling for the

TABLE 8.—*Comparison of estimated and simulated volumes of land subsidence in the San Joaquin and Sacramento Valleys from 1961 to 1977*

[In million acre-feet]

Years	San Joaquin Valley		Sacramento Valley	
	Estimated[1]	Simulated	Estimated[2]	Simulated
1961-69	5.2	4.8	0.17	0.10
1970-75	1.1	.48	.12	.04
1976-77	.60	1.2	.06	.22
1961-77	6.9	6.5	0.35	0.36

[1]Estimates obtained from Poland and others (1975), Ireland and others (1984), and unpublished data filed in the U.S. Geological Survey office in Sacramento, Calif.

[2]Estimates obtained from Lofgren and Ireland (1973) and unpublished data filed in the U.S. Geological Survey office in Sacramento, Calif.

Tulare-Wasco area was done in 1969–70, for the Arvin-Maricopa area, in 1970, and for the Los Banos-Kettleman City area, in 1971–72 (Ireland and others, 1984, p. 14). Since 1972, only partial leveling of selected lines (particularly along the California Aqueduct) has been done.

Because the times of detailed leveling did not always correspond among areas of subsidence and because the principal simulation period of the aquifer system was from spring 1961 to autumn 1977, yearly estimates of land subsidence from 1961 to 1977 were made primarily on the basis of average rates of subsidence between times of leveling and were prorated to individual years according to extensometer data from wells as reported in Poland and others (1975) and Ireland and others (1984). An estimate of land subsidence was also made for the period during the drought largely on the basis of extensometer data in wells and from a few level lines. The yearly estimated rate of subsidence in the San Joaquin Valley decreased in the 1970's (fig. 39), mostly because of decreased subsidence in the Los Banos-Kettleman City area (figs. 41 *A–D*), although the yearly estimated subsidence rate increased during the drought of 1976 through 1977 when ground-water pumpage increased greatly. Estimates of pumpage from 1973 through 1977 in the Los Banos-Kettleman City area were also added to figure 39. The relation between pumpage and land subsidence changed following 1970, after which a reduced proportion of the water pumped came from compaction of fine-grained sediments. This reduction probably is due to hydraulic head recovery which accompanied the reduction in pumpage during 1968–75.

SIMULATED SUBSIDENCE, 1961 TO 1977

Overall, the simulated volume of subsidence from 1961 to 1977 in both the San Joaquin and Sacramento Valleys compared well with the estimated volumes of subsidence from leveling and extensometer data for the same period (table 8). Simulated and estimated volume of subsidence for both the Arvin-Maricopa and the Tulare-Wasco areas also compared closely (table 9 and fig. 40). In both areas, the simulated subsidence from 1961 to 1969 was slightly more than the estimated subsidence, while during the period 1970–75 it was slightly less. This is consistent with the simplified approach to land subsidence in the simulation processes because all water is assumed to be released simultaneously during a given head decline in the simulations, whereas in the actual aquifer system, water may be released slowly owing to compaction of the fine-grained (clayey) beds for some time after a given head decline. In the area between the Tulare-Wasco and the Los Banos-Kettleman City areas, simulated subsidence was slightly less than estimated subsidence.

In the Los Banos-Kettleman City area, simulated subsidence west of the Fresno Slough and San Joaquin Rivers was generally less than estimated subsidence (table 9). Simulated subsidence for the period 1961–69 should have been more than estimated subsidence because the time lag was not simulated, and presumably it should have been as much as the amount estimated for 1961–75. The period 1970–76 was a time when generally the water levels recovered and subsidence was probably caused by the time lag between the head change in the aquifer materials and the water released

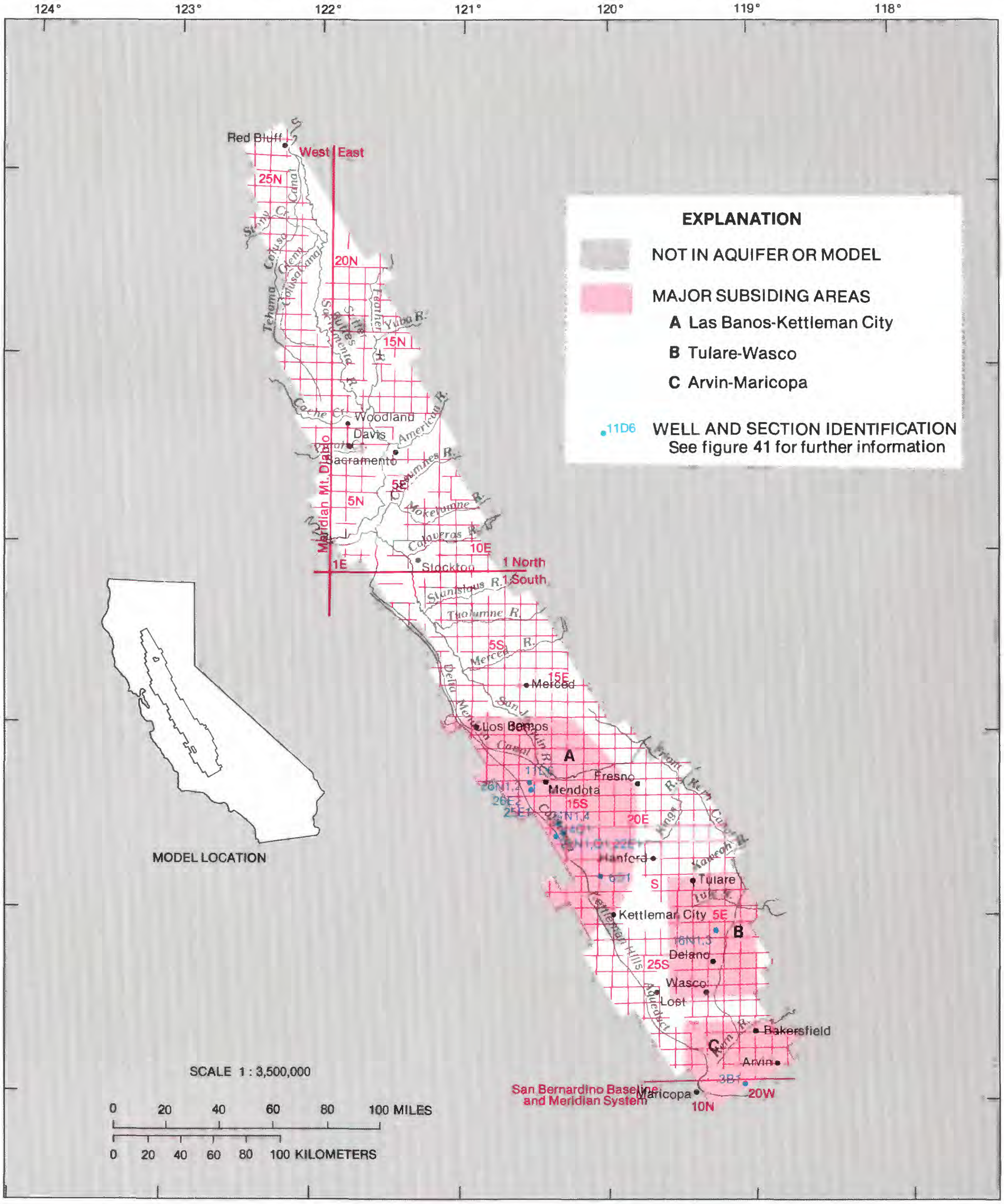


FIGURE 38A.—Major subsiding areas and locations of wells with water level and compaction data (modified from Ireland, Poland, and Riley, 1984, figs. 6 and 32).

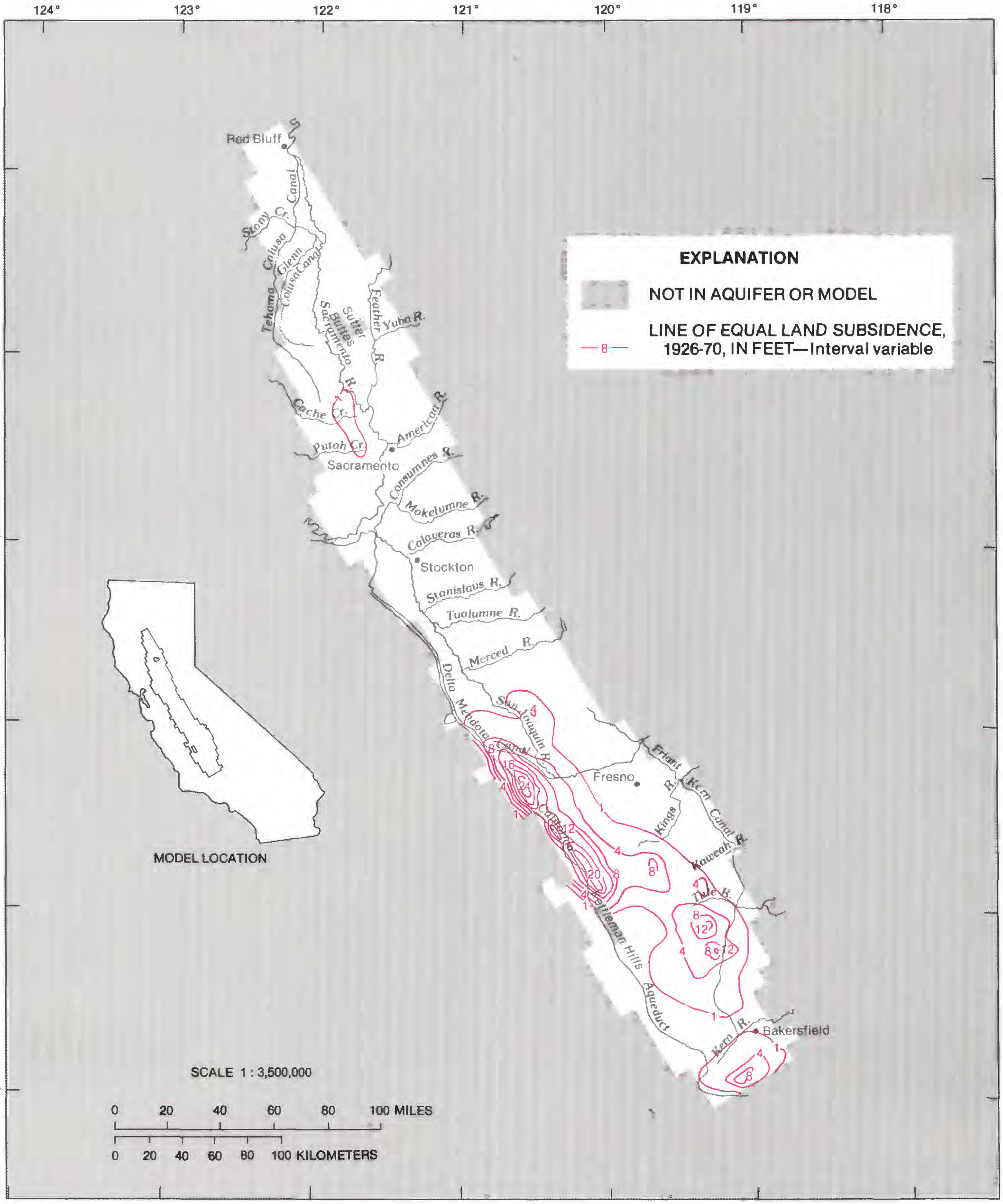


FIGURE 38B.—Land subsidence, 1926–70 (modified from Poland and others, 1975, fig. 6).

from compaction of the fine-grained (clayey) beds to the aquifer system. During the drought of 1976–77, the water levels in the lower pumped zone did not decline below the previous lows observed in the 1960's, yet subsidence was observed along the California Aqueduct and in the few wells with extensometers (Ireland and others, 1984, and fig. 41). Simulated subsidence in the same area was very small, as expected, because most of the heads in the model blocks did not decline below previous lows. Some of the observed subsidence may have been elastic, as indicated by negative compaction values following 1977 (fig. 41).

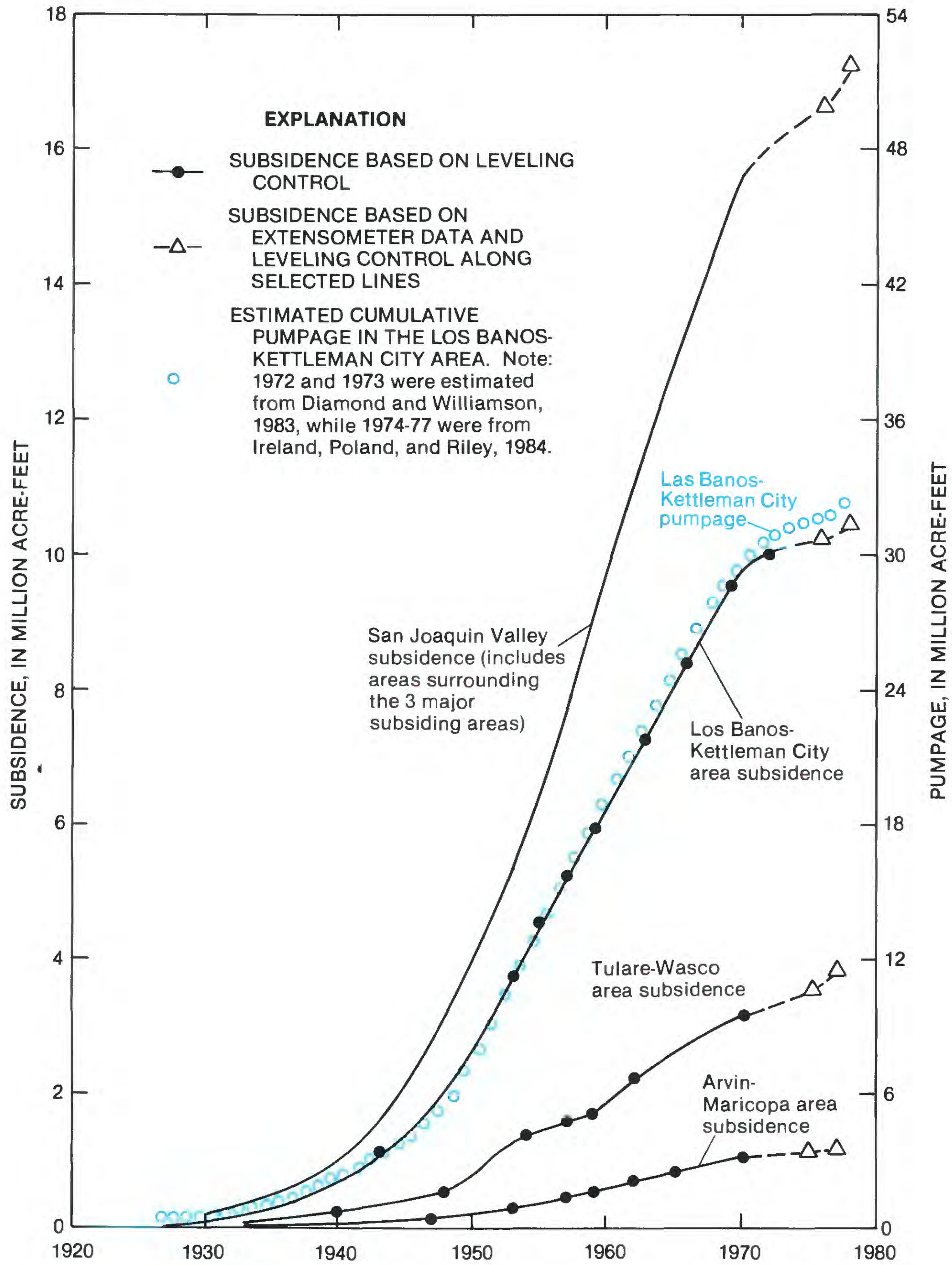


FIGURE 39.—Volumes of land subsidence in the major subsiding areas of the San Joaquin Valley, and pumpage in the Los Banos-Kettleman City area, 1925–77 (modified from Poland and others, 1975, figs. 6, 19, 29, and 38).

The distribution of estimated and simulated subsidence is shown in figure 40. The variations in simulated versus estimated subsidence may be explained in several ways:

1. In the simulation, pumpage from the lower pumped zone was the primary cause of land subsidence. The estimates of pumpage were summed by quarter townships and then transferred as the model input. The model grids, however, did not correspond to the township grid. Errors in transferring the pumpage from the township grid to the model grid can cause the amount and distribution of subsidence to be shifted in the model simulations.

TABLE 9.—*Comparison of estimated and simulated volumes of subsidence to pumpage for the major subsiding areas from 1961 to 1977*

[Pumpage and land subsidence are in million acre-feet. Pumpage for the lower pumped zone only]

Years	Total pumpage from lower pumped zone	Estimated volume of subsidence	Estimated percentage of pumpage from compaction	Simulated volume of subsidence	Simulated percentage of pumpage from compaction
			Arvin-Maricopa area		
1961-69	6.8	0.41	6	0.46	7
1970-75	6.8	.11	2	.08	1
1976-77	1.4	.04	3	.11	8
1961-77	12.6	0.56	4	0.65	5
			Tulare-Wasco area		
1961-69	7.5	1.0	13	1.2	16
1970-75	5.4	.36	7	.19	4
1976-77	2.2	.31	14	.30	14
1961-77	15.1	1.7	11	1.7	11
			Los Banos-Kettleman City area		
1961-69	8.0	3.3	42	2.6	32
1970-75	2.8	.51	18	.14	5
1976-77	1.0	.23	23	.05	5
1961-77	11.8	4.1	35	2.8	24
			Davis-Zamora area		
1961-69	2.0	0.17	9	0.03	2
1970-75	1.4	.12	9	.01	1
1976-77	.46	.06	12	.07	14
1961-77	3.9	0.35	9	0.11	3

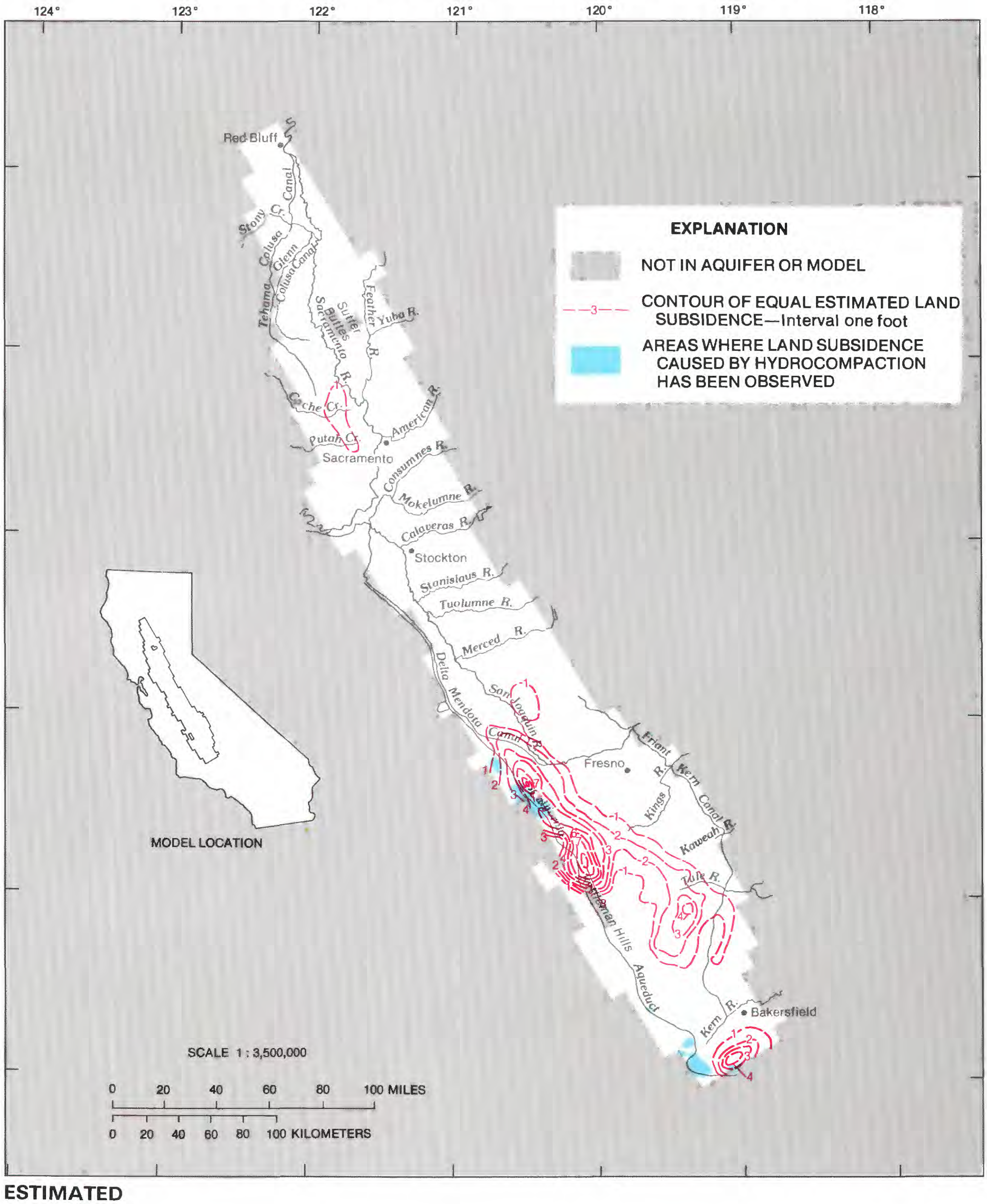


FIGURE 40A.—Estimated land subsidence, 1961–75.

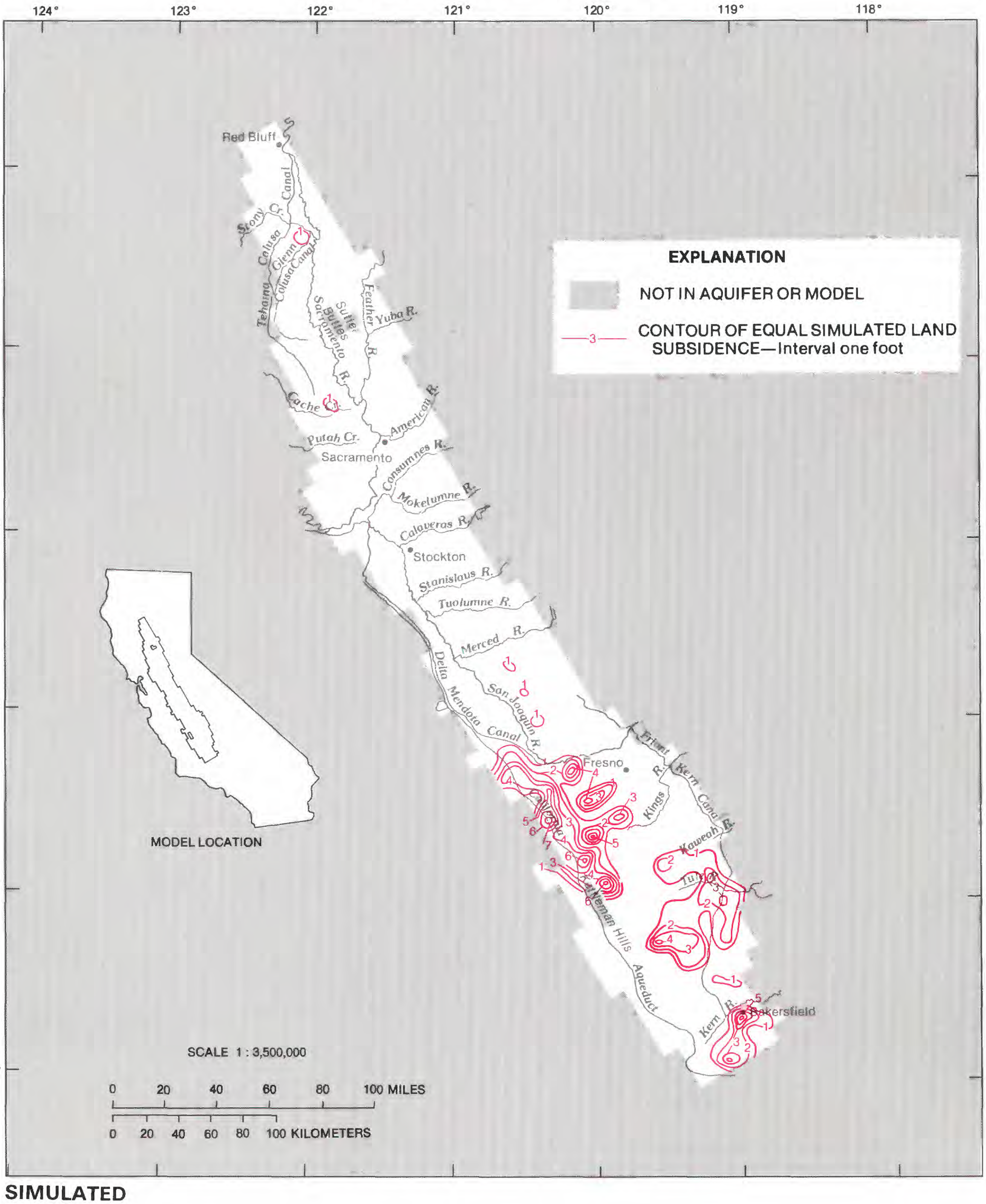


FIGURE 40*B*.—Simulated land subsidence, 1961–75.

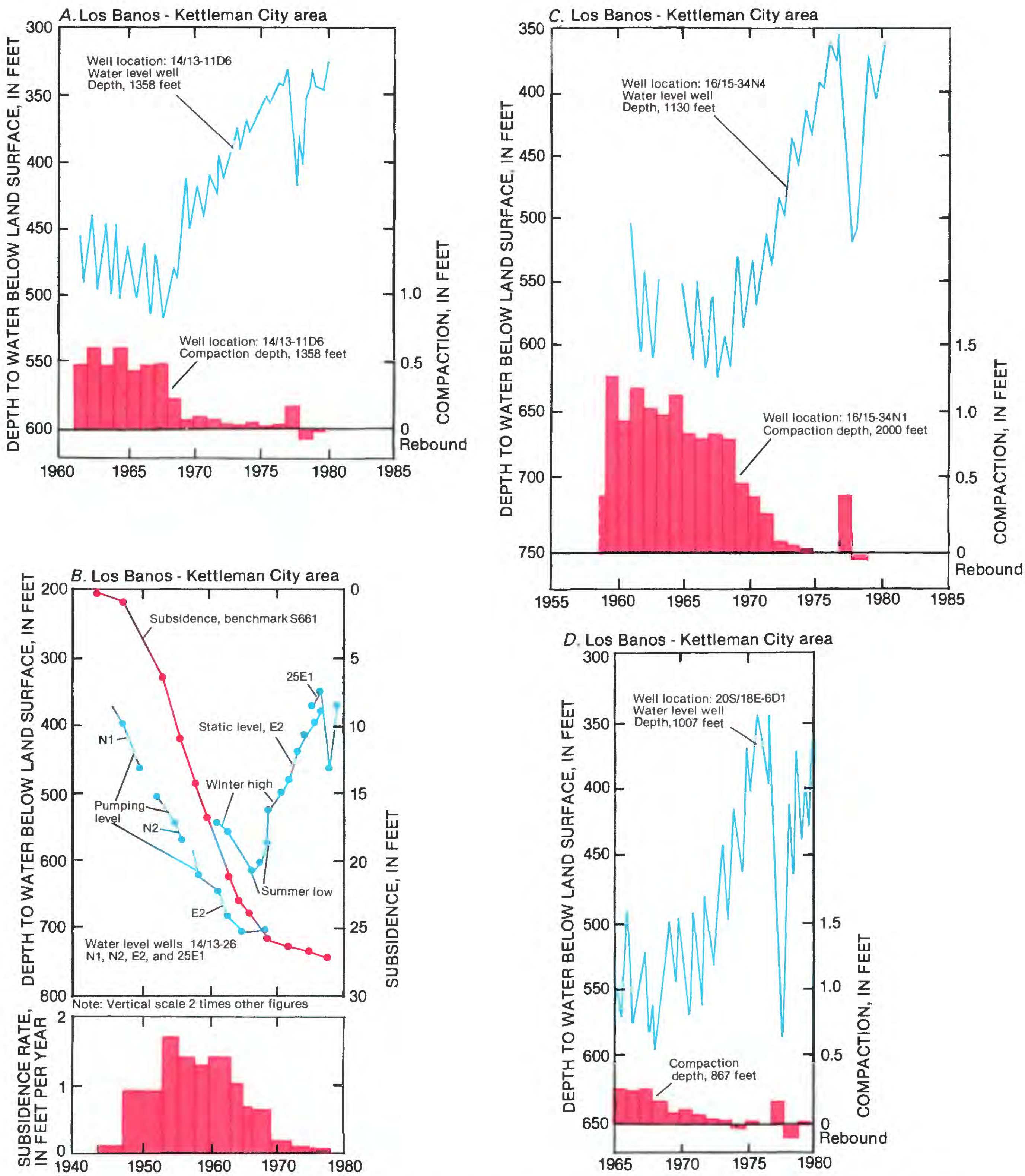


FIGURE 41*A–D*.—Measured water levels and compaction of selected wells in the major subsiding areas of the San Joaquin Valley, 1940–80. (After Ireland and others, 1984, figs. 22, 16, 21 and 24.)

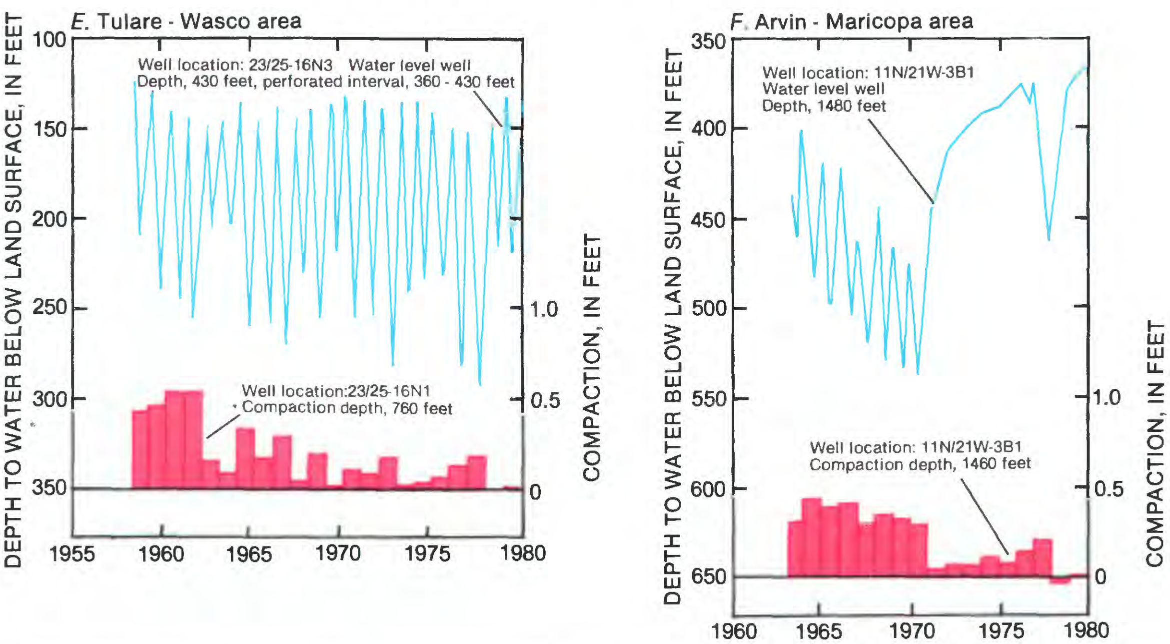


FIGURE 41E and F.—Measured water levels and compaction of selected wells in the major subsiding areas of the San Joaquin Valley, 1940–80. (After Ireland and others, 1984, figs. 29 and 31.)

2. Estimates of land subsidence, particularly after 1972, are based primarily on projections of localized data to areas without data. Because several parts of the Central Valley have not been releveled since 1970, these estimates of subsidence are subject to error.

3. The simulated amount of subsidence in any model block is dependent on the head at which inelastic compaction begins (the critical head). In the simulations, head in the clayey beds within the aquifer system was assumed to immediately equal the head in the aquifer system. Without considering the time lag, this assumption involves error because sufficient time is needed for a change in head in the aquifer to propagate through the thicker clayey beds.

4. Estimates of the critical head initially used in the simulation from 1961 through 1977 were made for areas of known subsidence by subtracting an estimated average head fluctuation in the 1960's from the heads of spring 1961. For critical heads in areas outside areas of known subsidence, a head of 80 ft less than the simulated steady-state head was used. Holzer (1981) estimated a change in head of 85 ft before significant subsidence occurred in the Tulare-Wasco area. He made this estimate on the basis of the observation that the ratio of subsidence to water-level decline increased dramatically in two wells in the area. The critical head in several of the model blocks, particularly in the active subsiding areas, were adjusted such that the simulated and estimated subsidence and drawdowns corresponded. The adjustments of head were usually small, less than 20 ft in most model blocks. These adjustments were not significant because the method used to estimate critical heads was not exact. Errors in estimating the critical head for each model block affect the simulated distribution and amount of subsidence as well as the heads in the lower pumped zone.

5. Subsidence was computed during simulations by multiplying the inelastic storage value by the amount of drawdown that was simulated when the inelastic storage value was actively used. However, if the computed head decreased below the critical head during the first time step of a pumping period, no subsidence was computed. This error was reduced by using a short initial time step.

6. In the simulations, when heads declined below the critical head values, water was released from compaction instantaneously. When the heads recovered above the lowest computed head, subsidence would not begin again until after the head was lower than the new critical head value. However, continuation of subsidence in the aquifer system (although at greatly reduced rates) has been observed for years after the time that heads recovered in the aquifer system. These observations are supported by water levels and extensometer data in the major subsiding areas (figs. 41*A*–*F*). In fact, observed subsidence in figures 41*A*, 41*C*, 41*D*, and 41*F* increased during the drought of 1976–77 even though water levels in wells did not go below the previous low water level. However, some of the observed subsidence during the drought may have been caused by elastic compression,

as indicated by the negative compaction (rebound) values following the drought. Similarly, water levels in a well near Delano in the Tulare-Wasco area did not show a continued yearly water-level decline, yet compaction (although somewhat variable) was continuous from 1958–77 (fig. 41*E*). The yearly simulated subsidence for this area was zero for the periods when the heads did not decline below the previous lowest head. Not being able to simulate subsidence during these conditions is the result of using a simplified approach to the complicated mechanics of subsidence. In particular, the assumption that the head in the coarse-grained deposits in the aquifer system is equal to the heads in the fine-grained deposits is not true (see "Model Limitations" section).

Factors that Affect the Relation of Subsidence to Pumpage

Estimates of ground-water pumpage, determined from electric power consumption and pump-efficiency tests, have been compiled yearly from 1961 through 1977 for most of the Central Valley (Diamond and Williamson, 1983). In addition, pumpage estimates were divided between the upper water-table zone and the lower pumped zone. A comparison of subsidence or the amount of compaction of the fine-grained sediments and pumpage in the lower pumped zone was done for each of the major subsidence areas (table 9).

The percentage of the total water pumped that was released from the fine-grained (clayey) sediments and caused compaction varied from area to area (table 9). The lowest overall percentage from 1961 through 1977 occurred in the Arvin-Maricopa area, where presumably only 2 to 6 percent of the water pumped from the lower pumped zone came from compaction. In contrast, as much as 42 percent of the pumpage came from compaction in the Los Banos-Kettleman City area during a period of major subsidence in 1961 through 1969.

The difference in the proportion of water released during compaction to total pumpage among the major subsidence areas is probably caused by (1) variations in amount, compressibility, and origin of the fine-grained sediments, and (2) variations in applied stress that compacts the deposits (Poland and others, 1972, p. 6). These variations are discussed in the following paragraphs.

Texture maps showing the amount of coarse-grained deposits with depth were prepared by Page (1986, figs. 6–21, 29–34). These maps indicate that the amount of coarse-grained material is consistently less to depths of 2,100 ft in the Los Banos-Kettleman City area than in the other major subsidence areas. The Arvin-Maricopa area consistently shows more coarse-grained material (Page, 1986, fig. 34). Thus, the variations in proportions of water released during compaction to total pumpage can generally be explained by differences in the percentage of fine-grained deposits.

Meade (1968, p. 4) indicates that montmorillonite is more susceptible to compaction than either illite or kaolinite. In each of the major subsidence areas in the San Joaquin Valley, montmorillonite was determined to be the major clay mineral, and was between 65 to 75 percent of the total clay minerals, as shown in the table below (from Meade, 1967, p. C18, C34, C46).

Clay mineral	Los Banos-Kettleman City (percent)	Tulare-Wasco (percent)	Arvin-Maricopa (percent)
Montmorillonite	70	60	75
Illite	10	20	10
Chlorite	10	0	10
Kaolinite-type mineral	5	10	5
Vermiculite	—	10	—
Mixed-layer montmorillonite-illite and low grade illite-montmorillonite	5	Trace	—

The results are based on 85 samples from four deep test holes in the Los Banos-Kettleman City area, 26 samples from two test holes in the Tulare-Wasco area, and 8 samples from one test hole in the Arvin-Maricopa area.

In contrast, the principal clay mineral found in soils and alluvium of the upper San Joaquin River basin was kaolinite and in many of the samples montmorillonite was absent (Meade, 1967, p. C21). Similarly, analyses of core samples from three test holes in the Sacramento Valley (one near Zamora) indicate that kaolinite is also the dominant clay mineral and that no montmorillonite was found in any of the samples to a measureable extent (French and others, 1982).

The montmorillonite in the Los Banos-Kettleman City area is in part derived from transport by streams that originate in the Diablo Range to the west (Meade, 1967, p. C18); aggregates of montmorillonite clays were found in the fan deposits. Some of the montmorillonite was also formed after the sediments were deposited. The source of montmorillonite in sediments from the Sierra Nevada is uncertain. Meade (1967, p. C18) listed possible sources as the belt of metamorphic rocks in the western foothills of the Sierra Nevada or clays from the Coast Ranges which were mixed with sediments from the Sierra Nevada; the montmorillonite clays may have formed by alteration or transformation of other minerals soon after they were deposited in the valley.

Reasons for the absence of montmorillonite in test holes in the Sacramento Valley and in analyses of soils and alluvium in the upper San Joaquin River basin are unknown, because the source areas of the sediments (Coast Ranges and the Sierra Nevada) are essentially the same. Although the major subsidence areas in the San Joaquin Valley contain principally montmorillonite, differences in the amount of compaction compared to pumpage cannot be explained by differences in the types of clay minerals. However, the absence of montmorillonite in other areas might contribute to a lesser amount of subsidence.

The origin of deposition of the sediments may also contribute to differences in the amounts of water contributed to pumpage from compacting clays in the major subsidence areas. Bull (1975) determined that in the Los Banos-Kettleman City area the highest apparent compressibility of the sediments in the lower pumped zone coincides with the area of flood-plain deposits, as opposed to areas of alluvial fan deposits, and that the bedding of the deposits is an important factor controlling the magnitude and rate of compaction. In the Arvin-Maricopa area, the proportion of flood-plain or lacustrine sediments is small (Lofgren, 1975, pl. 1), and in the Tulare-Wasco area, the proportion of flood-plain or lacustrine sediments increases to the west, where beneath the present-day Tulare Lake bed the sediments are largely lacustrine or flood plain in origin (Lofgren and Klausing, 1969, p. B9). Also, Meade (1967, p. C27) noted that the alluvial fan deposits in the Tulare-Wasco area differed from those in the Los Banos-Kettleman City area because the deposits in the Tulare-Wasco area are generally coarser grained and contain fewer fine clays. For these reasons, the variations in the amount of water contributed to pumpage from compacting clays may, in part, be explained by the depositional environment of the sediments.

Variations in the change in the effective stress among major subsidence areas may also affect the proportion of water contributed to pumpage from compacting clays. The change in effective stress in a confined aquifer system is proportional to the head difference between the hydraulic head in the confined zone and the water table (Lofgren, 1968). Thus, the greatest change in effective stress occurs when the hydraulic head in the lower confining zone is declining and the head in the water-table zone is rising or staying nearly constant. However, when water levels in both the confining zone and the water-table zone are declining, the change in effective stress would be small. Thus, variations in well construction or in the amount of water pumped that came from the water-table zone in the major subsidence areas may cause variations in the amount of water released due to compaction.

Differences in well construction in the major subsidence areas may in part explain the differences in the ratio of the amount of water released from compaction to the amount of water pumped. The amount of water pumped per unit area in the Los Banos-Kettleman City area is smaller than it is in the Arvin-Maricopa area (see fig. 23 for pumpage and fig. 38*A* for location), yet the amount of water released from compaction compared with pumpage is high (table 9). Most of the wells in the Los Banos-Kettleman City area are perforated below the shallow water-table zone because of poor quality water in the water-table zone (Davis and others, 1959, p. 184; Bull and Miller, 1975, p. E25). However, in the Tulare-Wasco and the Arvin-Maricopa areas, water is obtained from a greater interval of the aquifer system (Lofgren and Klausing, 1969, p. 43; Lofgren, 1975, p. D44) and the perforated intervals commonly extend from the water-table zone into the lower pumped zone.

The effect of this type of well construction is threefold: (1) some of the water pumped from the wells in the Tulare-Wasco and the Arvin-Maricopa areas probably came from the water-table zone, (2) the water levels in both the water-table zone and the lower pumped zone were lowered, thus reducing the vertical hydraulic gradient and consequently the rate of compaction of the fine-grained sediments, and (3) the wells with perforations open to both the water-table zone and the lower pumped zone essentially increased the vertical hydraulic conductance and hence the amount of circulation between the water-table zone and the lower pumped zone, as described in the section, "Changes in Ground-Water Flow."

In summary, the variations in the ratio of the amount of water released during compaction to the amount of water pumped can be explained by several factors. These are the amount of fine-grained sediments, the types of clay minerals, the environment of deposition of the sediments, and the change in vertical hydraulic gradient which is dependent on the perforated intervals of wells.

CHANGE IN AQUIFER STORAGE

Increase in discharge (such as pumpage) or decrease in recharge causes decline in water levels, which indicates release of water from storage in the aquifer system. There are three types of release from aquifer storage: (1) water-table release (water released from storage is a result of gravity drainage of water stored in pores of the sediments); (2) elastic release (water released from storage is a result of the expansion of the compressed water and sediments when the hydraulic pressure is reduced); and (3) release from inelastic compaction, which occurs only when applied stress exceeds preconsolidation stress so that the pores of the sediments are rearranged and pore volume is reduced (the action is irreversible, i.e., permanent).

The total estimated decrease in ground-water storage from predevelopment conditions until 1961 was about 47 million acre-ft and through 1977, 60 million acre-ft.

The decrease in aquifer storage for the period 1961 through 1977 was estimated to be about 13 million acre-ft, or about three-quarters of a million acre-ft/yr. This decrease in aquifer storage is due to discharge (mainly pumpage) in excess of recharge. The amount of water released from water-table and elastic storage were calculated as the product of water-level changes, covered area, and the appropriate storage coefficients. This calculation probably is better than the calculation of storage changes from a water-budget approach, because small errors in recharge/discharge can cause large errors in the calculations of aquifer-storage changes. It would be desirable to determine aquifer-storage changes for shorter time periods to see the status of the system before and after the major water-importation development began. However, it is not feasible to determine aquifer-storage changes accurately for any shorter period of time because of the high variability in climatic conditions which overwhelms the short-term effects of development.

The volume of aquifer-storage change is substantial; however, it is still very small compared with the total volume of water in aquifer storage (table 7). The storage values shown in table 7 were calculated from the product of the specific yield and the thickness determined from the difference between the altitude of the 1961 water table and the shallower of (1) a depth of 1,000 ft, or (2) the base of continental deposits, or (3) the base of freshwater. There was more than 800 million acre-ft of freshwater in storage in the aquifer system at depths of 1,000 ft or less in the Central Valley as of spring 1961.

Water-Table Zone

The volumetric change in storage resulting from head changes in the water-table zone was estimated by analyzing the water-level data. The model-simulation results were not used because slight differences in the balance of recharge and discharge causing a small mean difference in observed and simulated water levels would substantially affect the simulated changes in aquifer storage in the water-table zone.

Seasonal high or low water levels for each measured well (usually spring high and autumn low) were averaged for the four geographic areas of the Central Valley (see fig. 27). December to May was used as the spring season, and June to November as the autumn season. Depth to water was chosen over water-level altitudes because its variation was less dependent on the selection of wells in a given season. Variation in water-level altitude is largely related to variations in land-surface altitude, and so it is dependent on the selection of wells measured. Averages were made over large areas to minimize the effect of outliers. The change in depth to water was multiplied by the land area where the changes occurred and the average specific yield to obtain the values of changes in aquifer storage in the water-table zone. Using the average specific yield introduces some errors if the specific-yield values are not distributed evenly with respect to the distribution of depth-to-water measurements. There were more than 2,000 water-level measurements for most of the spring seasonal averages. Estimates of the change in aquifer storage in the water-table zone were 34 million acre-ft for the period from predevelopment until 1961, and about 5.5 million acre-ft for 1961–77.

Elastic Storage

Elastic storage is a result of the expansion of water and the compression of sediments because of change in fluid pressure. Change in elastic storage is computed as the product of the elastic specific storage, the thickness of the confined aquifer, the aquifer area, and the decline in head. This was calculated for each of the 484 model blocks that had head declines, using the thickness of layer 3, or the sum of the thicknesses of layers 2 and 3 in the 163 model blocks where many wells penetrated layer 2. The thickness of layer 1 was ignored because the drawdown was negligible. The change in elastic storage in layer 4 is obscured by and included with the change in water-table storage. The average estimated elastic specific storage was 3×10^{-6} per ft. The estimates of elastic specific storage were increased by a factor of two in most areas during calibration of the model with 6-month time periods. The calibrated elastic specific storage may be too large because allocating all agricultural pumpage to the autumn period and allocating all recharge to the spring period exaggerated the seasonal change in stress. The average lower pumped zone head decline was 80 ft. The amount of water released from elastic storage was about 3 million acre-ft from predevelopment to 1961.

The average head decline in the lower pumped zone from spring 1961 to spring 1976 was small because in many areas water levels declined; however, in other areas, they rose sharply. Therefore, the net change in elastic storage during that period was negligible.

Water Released From Inelastic Compaction

The process of compaction of fine-grained sediment in the aquifer system caused by head decline was discussed in the sections on land subsidence. When the fine-grained sediments in the aquifer are compacted, grains are reoriented and there is a reduction in the pore space within the compacted beds, thus releasing water. The volume of water released by compaction is approximately equal to the volume of land subsidence observed at the surface. Four other processes also cause land subsidence in the Central Valley (Poland and others, 1975): oxidation and compaction of peat soils, compaction of moisture-deficient sediments near land surface when water is first applied, compaction of deep deposits caused by the withdrawal of gas and oil, and tectonic settling. These processes cause only localized subsidence or a small rate of subsidence compared with subsidence caused by the decline of hydraulic heads within the aquifer system. Thus, the amount of water that has been released from compaction in the Central Valley was estimated by the volume of land subsidence through 1977, which is 17 million acre-ft.

The loss of pore space is a loss of storage capacity in the aquifer system. Therefore, if water levels recover to their previous highest altitude, the amount of water stored in the aquifer system is not the same as the amount stored before compaction; it is less. Inelastic compaction means permanent compaction. This type of land subsidence represents a one-time withdrawal of water from storage. However, the storage capacity of the coarse-grained sediments is unchanged.

Table 10 compares the amounts of water released from inelastic compaction to ground-water pumpage and water released from the water-table zone. From 1961 to 1978, about 7.3 million acre-ft of water was released from inelastic compaction, or about 4 percent of the total estimated pumpage of 189 million acre-ft for the entire Central Valley. Almost three-fourths of the water released from inelastic compaction occurred between 1961 and 1970, a period of major subsidence in the Los Banos-Kettleman City area (see table 9).

Most of the water released from inelastic compaction occurred in the Tulare area (see fig. 27 for location). In that area, the amount of water released from inelastic compaction during the period 1961–70 was about 8 percent of the estimated pumpage in the Tulare Basin (table 10). The amount of water released from inelastic compaction in the other areas during the same period was 3 percent or less. For the entire Central Valley, the amount of water released from the water-table zone was about 3 percent of the estimated pumpage for spring 1961 to spring 1978 (table 10). Thus, it can be concluded that most of the water pumped from 1961–78 came from increased recharge and decreased natural discharge.

MODEL LIMITATIONS

The model represents only the significant features of the aquifer system. It grossly simplifies the system, both in its temporal and spatial variability and in its processes. The following discussion is intended to alert readers not to overextend conclusions drawn from results of the simulations and to provide suggestions for further study.

CALIBRATION

Calibration of the flow model during this study is achieved by adjusting the values of one or more aquifer properties or recharge/discharge such that the computer simulated hydraulic heads match (within the limits of the investigation) the observed heads in the aquifer system. Calibration is a continuous process until a point that the head difference between the simulated and observed values reaches a preset value (a criteria set by the authors). Further improvement is still possible because of the vast number of values that can be adjusted. However, the process is constrained by the amount of data available to determine how closely the observed data can be reproduced by simulation. The differences among observed and simulated water-level changes from 1961 through 1975 are summarized in table 11. The following are discussions of these differences:

1. The errors in matching observed water-level changes in layer 4 (the water-table zone) are less than those in layer 3 (the lower pumped zone). This is not surprising because the smaller elastic-storage coefficient in layer 3 causes the hydraulic head in layer 3 to respond faster to pumpage; hence, any head change is magnified.
2. Simulated water levels in layers 3 and 4 at the end of the calibration period are too high, by a modelwide average of 2.6 ft in layer 4 and 12.0 ft in layer 3. This probably indicates that the estimates of recharge were too high, or that the estimates of discharge were too low, or both. This systematic error, which is cumulative, as indicated by the increasing average observed minus simulated head difference with time (fig. 42), could have been adjusted by multiplying recharge and discharge values by a factor. This adjustment was not made because there is no hydrologic basis for it and because it would not really add significantly to the overall fit or to the understanding of the system. This error appears to have little relation to whether or not the block was one where the observed water levels rose or declined.
3. Figure 43 indicates that 80 percent of the simulated minus observed water-level differences are within +23 to −26 ft for the water table, and +15 to −45 ft for the lower pumped zone.

Comparison of observed and simulated water levels would not have much meaning unless something is known about the errors in estimating observed average water level for a block at a time period. Because of the size of the blocks chosen and the variability of water levels in space, time, and depth, the accuracy of

TABLE 10.—*Proportion of pumpage from water table and compaction storage*

[Pumpage and water released from water table and compaction storage are in millions of acre-feet. Note that the main source of water for pumpage is not storage, but increased recharge and decreased natural discharge. Locations of areas in the Central Valley are shown in figs. 1 and 27].

		Estimated water released from or recharged into aquifer storage[1]			
	Pumpage[2]	Water table zone	Contributed to pumpage in percent	Compaction	Contributed to pumpage in percent
Sacramento Valley - area 1					
Spring 1961 to spring 1970	11.3	0.6	5	0.17	2
Spring 1970 to spring 1976	9.0	1.6	18	.12	1
Spring 1976 to autumn 1977	4.7	.6	13	.06	1
Autumn 1977 to spring 1978	([3])	-1.8	--	--	--
Spring 1961 to spring 1978	25.0	1.0	4	0.35	1
Delta Area - area 2					
Spring 1961 to spring 1970	12.3	-0.6	--	([4])	--
Spring 1970 to spring 1976	8.9	.05	1	--	--
Spring 1976 to autumn 1977	3.7	1.1	30	--	--
Autumn 1977 to spring 1978	([3])	-1.0	--	--	--
Spring 1961 to spring 1978	24.9	-0.5	--	--	--
San Joaquin Valley - area 3					
Spring 1961 to spring 1970	17.0	-0.02	--	0.48	3
Spring 1970 to spring 1976	12.3	1.3	11	.18	1
Spring 1976 to autumn 1977	5.4	3.9	72	.08	1
Autumn 1977 to spring 1978	([3])	-2.3	--	--	--
Spring 1961 to spring 1978	34.7	2.9	8	0.74	2
Tulare Basin - area 4					
Spring 1961 to spring 1970	58.9	-1.6	--	4.7	8
Spring 1970 to spring 1976	32.1	1.8	6	.89	3
Spring 1976 to autumn 1977	13.6	5.0	37	.54	4
Autumn 1977 to spring 1978	([3])	-2.3	--	--	--
Spring 1961 to spring 1978	104.6	2.9	3	6.1	6

See footnotes at end of table.

TABLE 10.—*Proportion of pumpage from water table and compaction storage*—Continued

[Pumpage and water released from water table and compaction storage are in millions of acre-feet. Note that the main source of water for pumpage is not storage, but increased recharge and decreased natural discharge. Locations of areas in the Central Valley are shown in figs. 1 and 27]

		Estimated water released from storage[1]			
	Pumpage[2]	Water table	Percentage of pumpage	Compaction	Percentage of pumpage
Entire Central Valley - Total					
Spring 1961 to spring 1970	99.5	-1.6	--	5.4	5
Spring 1970 to spring 1976	62.2	4.8	8	1.2	2
Spring 1976 to autumn 1977	27.4	10.6	39	.7	2
Autumn 1977 to spring 1978	([3])	-8.3	--	--	--
Spring 1961 to spring 1978	189.1	5.5	3	7.3	4

[1]Negative values indicate an increase in the volume of water stored in the aquifer system. Estimates of the amount of water released from elastic storage in the lower pumped zone is not shown because the values are small (less than 0.05 million acre-ft) for each of the major areas, even though head declines may be large in the lower pumped zone at several locations.

[2]Pumpage includes estimates of all pumpage from both the water-table zone and the lower pumped zone. Estimates in the Delta area are considerably more than those shown in table 2 of Diamond and Williamson (1983). In this table the estimates represent the entire Delta area.

[3]Pumpage that occurs during this period is excluded from the study period.

[4]Water released from compaction of sediments (land subsidence) in the Delta area is caused primarily by drainage of peat lands, and the amount of water released is incorporated into the specific yield of the water table.

TABLE 11.—*Summary of water-level changes, observed and simulated, 1961–75, in feet*

Layer	Number of blocks	Observed decline or rise	Observed water-level change: Mean	Observed water-level change: Standard deviation	Observed change - simulated change: Mean[1]	Observed change - simulated change: Standard deviation	Absolute value of observed change - simulated change: Mean	Absolute value of observed change - simulated change: Standard deviation
4	529	both	5.1	20.3	-2.6	21.9	16.5	14.6
	396	decline	15.0	16.2	-2.3	21.9	17.1	13.8
	133	rise	-13.0	13.5	-3.1	22.0	15.5	16.0
3	529	both	8.0	48.8	-12.0	27.4	22.0	20.2
	435	decline	30.3	28.4	-10.8	24.9	20.9	17.4
	94	rise	-41.6	48.1	-14.5	32.3	24.5	25.4

[1]Observed change-simulated change: negative sign means simulated water level above observed water level.

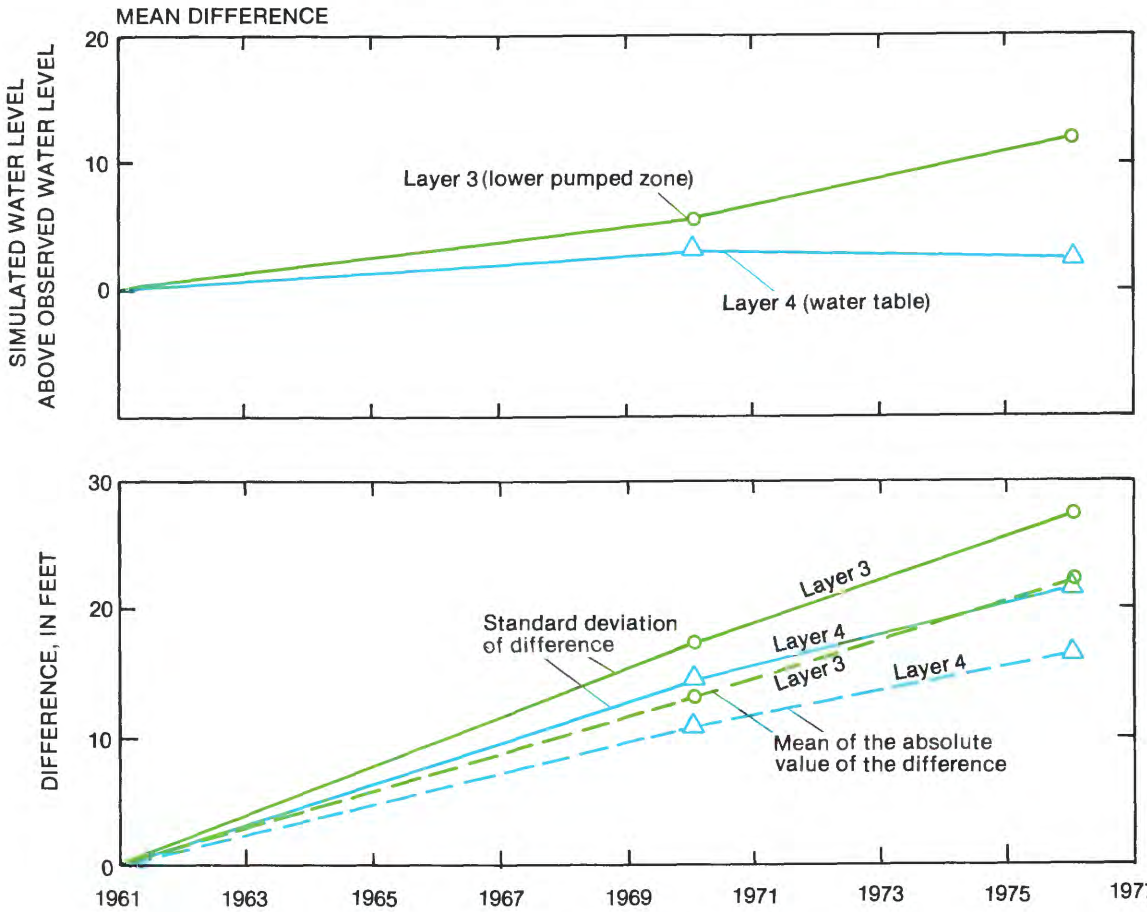


FIGURE 42.—Departure of simulated and observed water levels, 1961–76.

estimating a block's water level is approximately 20 ft. In light of this fact, the statistics about the model fit seem reasonable.

The absence of knowledge about water levels is even more pronounced at depth. In addition, two-thirds of the wells in which water levels are monitored do not have drillers' logs or other construction data available. Only three known piezometers measure water levels in the deep zone (layer 1) below the lower pumped zone, and these are all in the Sacramento Valley. There are other indications of water level at depth, such as gas-well shut-in pressures. A problem in interpreting these gas-well data is that the shut-in pressures were observed only when the wells were drilled, whereas gas pressure changes as the field is developed.

VARIABLE DENSITY

As previously described in the section, "Extent of Freshwater," saline water is found below the freshwater body throughout much, if not all, of the Central Valley. Salinity of water in these deeper zones may exceed that of seawater (Hill, 1972). Model simulations made during this study did not account for the differences in density of the waters. Because the ratio of seawater density to freshwater density is 41 to 40, a freshwater head of 41 ft would be equal to a seawater head of 40 ft. Ignoring the density difference introduces an error of about 2.5 percent in the head values from the deepest part of the aquifer system where saline water occurs. The source and movement of this saline water is not known. A preliminary analysis of shut-in pressure data shows that the simplest assumption of a static head distribution in the saline water system is invalid. The rate of movement of the interface between the fresh and saline water has not been analyzed.

RECHARGE AND DISCHARGE ESTIMATES

A significant limitation of the simulation of the aquifer system is the inability to relate variability of recharge and discharge to water-table fluctuations. Regression analyses using estimated values of recharge from, and discharge to, streams showed a poor correlation with depth to water. This poor correlation is probably due to the depth-to-water data, which were not always observed

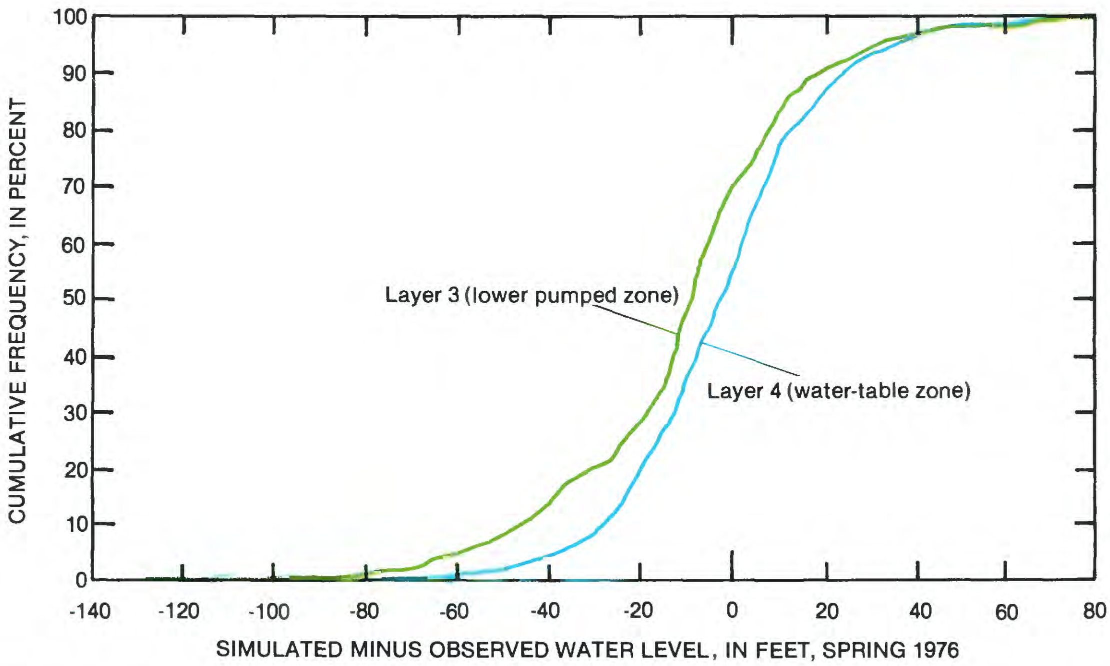


FIGURE 43.—Cumulative distribution of the deviation of simulated water levels from observed water levels for the end of the calibration period, spring 1961 to spring 1976, for the water table (layer 4) and the lower pumped zone (layer 3).

near the streams. Recharge and discharge did not need to be head-dependent in the simulation algorithm because there was no need for prediction capabilities in the simulation. The relation was assumed to be inherent in the estimated data collected for the calibration period.

As mentioned earlier, the estimates of net recharge/discharge were adjusted during calibration by adding a factor that was constant in time for each block. The relation of the final calibrated estimates to the initial estimates is shown in figure 44. These values represent 1961–77 averages of net recharge/discharge to and from the water-table zone. As shown by figure 44, there were many values that were changed by a factor several times greater than the initial estimated values. This may not be indicative of a large absolute change, because some values were very small to start with. However, there is a definite need for improvement in data, methods of estimating, and methods of distributing the values geographically.

SUBSIDENCE

The modification of the Trescott (1975) ground-water-flow computer program which was used to simulate land subsidence had two major shortcomings. First, the subsidence resulting from head declines was simulated as if it all occurred during the same time step as the head decline, whereas in the aquifer system there is a significant time lag before all of the subsidence occurs. Therefore, the short-term subsidence simulations are in error, but the magnitude of the error decreases with time. Second, the change from one storage value to another was explicit; it was done at the beginning of each time step based on whether or not the head in the previous time step dropped below the critical head. Thus, small time steps were necessary in the simulations to minimize this error, and this increased the computer time and the cost of each simulation.

The method of simulating subsidence used during this investigation also did not accurately simulate the effects of the 1976–77 drought. Simulated subsidence was less than observed subsidence because in many model blocks, the head did not decline below the previous lowest head. However, some of the observed compaction, as measured from wells with extensometers, was elastic. This is demonstrated by the negative compaction after the drought, indicating elastic rebound.

Another problem with the technique of simulating water released from compaction was the value used for the starting "critical" head—the head at which inelastic compaction begins. The simulated volume of subsidence, especially for the early years, was sensitive to the initial estimate of the critical head. Initial critical-head values were estimated to be 80 ft less than the predevelopment water levels of the early 1900's. The 80-ft difference was based on estimates by Holzer (1981) at a few locations in California. Model simulations began in 1961 during a period of major subsidence in several parts of the Central Valley, and water levels in several areas were already many feet below the initial estimate

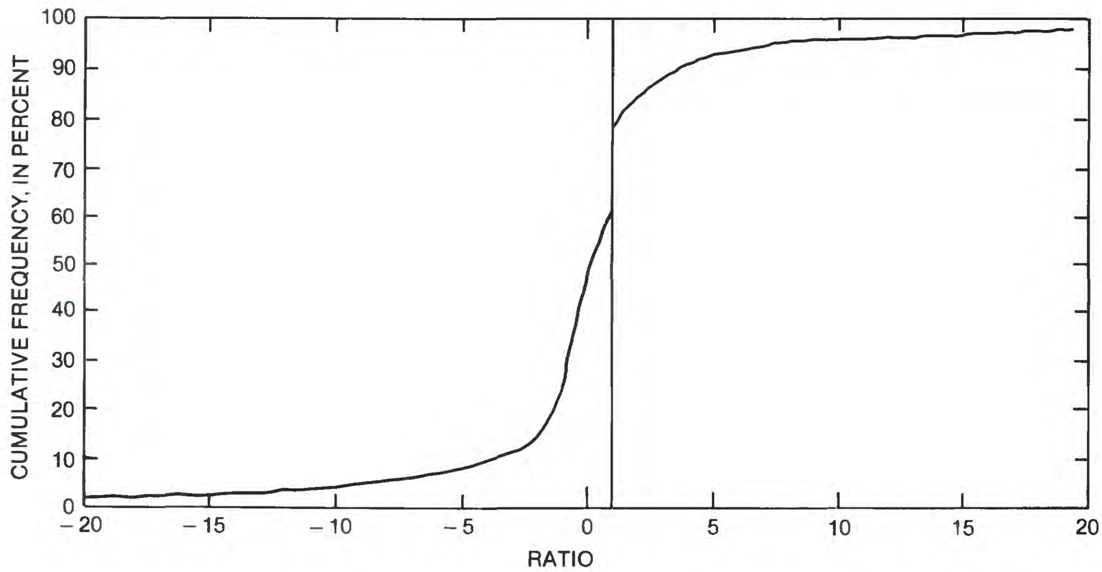


FIGURE 44.—Ratio of calibrated to estimated net recharge/discharge to and from the water-table zone. Negative values indicate that the sign changed during calibration.

of the critical heads. Thus, in areas where the water levels in 1961 were below the initial estimate of critical head, the critical head was estimated to be the previous observed low water level, which commonly had occurred during the 1960 irrigation season. Critical-head values were adjusted as much as 15 ft in several model blocks during the calibrations.

An approach suggested by Helm (oral commun., 1979), coupling a three-dimensional flow model with his one-dimensional (vertical) subsidence model (Helm, 1975), was investigated but abandoned because of the potential numerical instability of coupling the two models.

Another approach which used several layers at the bottom of the three-dimensional flow model to simulate the processes that operate within individual fine-grained beds was only preliminarily tested owing to insufficient time and the uncertainty of success associated with the application of new approaches. These lower layers in the model would have simulated only one-half (edge to center) of one fine-grained bed, so the flow from the top aquitard layer to the lower pumped zone of the aquifer system would have to be multiplied by two times the number of aquitards to simulate the combined effect of all of the aquitards on the lower pumped zone.

Though not thoroughly applied, this approach has several potential advantages: it is implicit, it allows for the time lag, it relies wholly on the numerical stability of the three-dimensional flow model, which has been extensively tested, and it allows detailed vertical discretization where necessary. A major problem with this approach is that it would not permit simulation of aquifer zones below the lower pumped zone because they would be totally confined from the lower pumped zone by the simulated aquitard. In the real system, fine-grained beds confine flow only in a very local area because their lateral extent is usually small. The approach was tested in a 3-by-3 areal grid to compare it with the results of Helm's model. A 3-by-3 areal grid will have only one vertical set of active blocks, so it essentially becomes a one-dimensional vertical system. Helm's simulation results were duplicated with only four layers representing the half aquitard. However, it would not accurately simulate the second cycle of renewed water-level declines that occurred in the Westside area during the drought. Helm's model was not tested under these conditions. This approach appears to have some promise based on a small-scale test, but it needs further refinement and testing.

SUMMARY AND CONCLUSIONS

Agricultural production of the Central Valley is dependent on the availability of water for irrigation. One-half of this irrigation water is supplied by ground water. Ground-water pumpage in the Central Valley accounts for 74 percent of California's total pumpage and about 20 percent of the Nation's irrigation pumpage. Ground-water pumpage is especially important in dry years because it supplements highly variable surface-water supplies. In 1975, about 57 percent of the total land area (12.8 million acres) in the Central Valley was irrigated. This heavy agricultural development during the past 100 years has had major impacts on the aquifer system. Ground-water flow before and during development was simulated using a three-dimensional finite-difference flow model on a regional scale.

The Central Valley is a large structural trough filled with marine sediments that are overlain by continental deposits. More than half of the thickness of the continental sediments is composed of fine-grained sediments. When development began in the 1880's, flowing wells and marshes were found throughout most of the central part of the Central Valley. Most previous investigators have conceptualized the northern one-third of the valley, the Sacramento Valley, as one water-table aquifer and the southern two-thirds, the San Joaquin Valley, as a two-aquifer system separated by a regional confining clay layer. A somewhat different conceptual model of the aquifer system is suggested during this investigation by analysis of water-level measurements, lithologic analysis, and the simulated flow conditions. Vertical hydraulic-head differences are present nearly throughout the valley. The new conceptual model assumes that the entire thickness of the continental deposits is one aquifer system that has varying vertical leakance and confinement depending on the proportion of fine-grained sediments.

The average horizontal hydraulic conductivity for the Central Valley is about 6 ft/d and the average thickness of the continental sediments is about 2,400 ft. The average horizontal hydraulic conductivity for the Sacramento Valley is about one-half of the average for the San Joaquin Valley, probably because of the greater amount of volcanic sediment found in the Sacramento Valley. These conditions could be significant in evaluating the potential for land subsidence in the future. Saline water underlies the freshwater throughout most of the Central Valley. The difference in density between fresh and saline waters was not considered in the simulations during this investigation because the aquifer system below the base of freshwater is poorly understood.

During 1961–77, an average of 22 million acre-ft/yr of water was used for irrigation; about one-half of the water was ground water. This level of development has increased evapotranspiration and decreased surface-water outflow by about 9 million acre-ft/yr from its predevelopment value (24 million acre-ft/yr). This is a large value compared with the average annual surface-water inflow to the Central Valley of 31.7 million acre-ft. Precipitation on the valley floor (12.4 million acre-ft/yr) is mostly lost to evapotranspiration. The overall irrigation efficiency (an average of 59 percent) increased during the 1961–77 period, apparently as the result of water conservation. Overall, the postdevelopment recharge and discharge values for the aquifer system were about 6 times greater than the predevelopment values. Postdevelopment average recharge came mostly from irrigation return (83 percent), but also from precipitation (13 percent) and infiltration from streams (4 percent). The actual proportion from streams is probably larger, but owing to the scale of the regional model constructed during this investigation, some stream recharge cancels with local discharge to other nearby stream reaches.

The increases in pumpage because of agricultural development, especially where little surface water was available, have caused water-level declines that exceed 400 ft in places and have contributed to the largest volume of land subsidence in the world due to ground-water withdrawal. From predevelopment until 1977, the volume of water in aquifer storage declined about 60 million acre-ft, with 40 million acre-ft from the water-table zone, 17 million acre-ft from inelastic compaction of fine-grained sediments, and 3 million acre-ft from elastic storage. During 1961–77, ground water withdrawn from storage averaged about 800,000 acre-ft/yr. As of 1977, more than 800 million acre-ft of freshwater was in aquifer storage in the upper 1,000 ft of sediments. Aquifer storage greatly exceeds surface-water storage, which is about equal to the average annual surface-water inflow (31.7 million acre-ft). This was evident during the 1976–77 drought, when surface storage was depleted and many farmers switched to ground water for irrigation.

The simulation model was calibrated principally according to the hydrologic data observed during the 1961–75 period because little predevelopment data are available. The simulated water levels were found to be most sensitive to the leakance value. Of the five types of causes that resulted in land subsidence occurring in the valley, the most significant cause is that resulting from withdrawal of ground water. Subsidence of this type was incorporated into the flow model. The computer program was modified to include both an elastic-storage and an inelastic-storage coefficient, using the inelastic-storage coefficient values only if the aquifer head for the previous time step was lower than the estimated critical head below which compaction of fine-grained sediments would begin. The simulated volume of land subsidence was within 6 percent of the total estimated volume. However, the time lag associated with this type of subsidence was not adequately simulated, nor was the subsidence during periods when the aquifer head was not lower than its previous lowest head (critical head) as occurred at times during the 1976–77 drought. At the end of the 1961–75 calibration period, simulated water-level changes averaged 2.6 and 12 ft above observed water-level changes for the water-table zone and the lower pumped zone; the standard deviation was 22 and 27 ft, respectively, which is nearly within the error of the estimated average observed water-level changes in a model block.

The simulations showed that vertical leakance greatly increased from the predevelopment values as a result of water flowing through some of the more than 100,000 irrigation well casings that are open to different aquifer layers. This may affect ground-water quality by allowing poor quality water in one of the aquifer layers to mix with good-quality water in another aquifer layer. The simulations also showed that on a regional scale the volume of vertical flow was more than horizontal flow, despite the fact that vertical velocities are much lower. This is due to the larger area of the aquifer in a horizontal plane than in a vertical plane. These factors should be considered in plans for improving and protecting ground-water quality in the valley.

During 1961–77, only 7 percent of the annual pumpage (11.9 million acre-ft) was being taken from aquifer storage. The remainder was being supplied primarily by recharge, from irrigation return flow but also from other increased recharge and decreased natural discharge. Only about 7 percent of the total freshwater in aquifer storage in the upper 1,000 ft of the aquifer system had been removed as of 1977. In addition, as water levels decline, more recharge is captured and less discharge to surface water bodies would occur. Therefore, at the present level of development, the withdrawal from aquifer storage will eventually diminish and the aquifer system will reach a new equilibrium condition. However, if ground-water development continues at an increasing rate, the aquifer system will take a longer time to reach a new equilibrium. The continuation of ground-water development is one of the reasons that a goal of the U.S. Bureau of Reclamation's Central Valley Project to eliminate depletion in aquifer storage has not been reached. Although the Bureau of Reclamation imported surface water into the Central Valley to decrease ground-water pumpage in some areas, ground-water development was allowed to continue in other areas.

There are other impacts from water-level declines that need to be considered. Land subsidence continues to be a problem in some areas of the Sacramento and San Joaquin Valleys, though the areas of greater subsidence have been controlled by importing surface water and decreasing ground-water pumpage. In these areas, the recovery of lower pumped zone water levels to nearly their predevelopment altitude may lead to an overestimate of the available ground-water resources in those areas. If pumpage increases again, water levels will drop rapidly toward the previous lows, as happened in the Westside area during the 1976–77 drought. This is because loss of aquifer storage capacity resulted from the compaction of fine-grained sediments. Water-level declines also cause increased energy consumption and associated costs. The effect (if any) on the movement of the deeper saline waters in response to water-level declines is unknown and was not evaluated during this study.

The regional aquifer-system analysis during this investigation indicates that, although there are local areas of severe aquifer depletion in the Central Valley, the ground-water resources of the entire valley are sufficient to meet the existing needs, assuming that development is carefully planned and managed. Ensuring adequate ground-water resources in the future will require a cooperative effort by local water districts and State and Federal agencies to monitor ground-water conditions in the Central Valley.

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APPENDIX A: RECHARGE AND DISCHARGE VALUES USED IN SIMULATIONS

Recharge and discharge data consisting of 10 variables for 529 nodes for a period of 17 years were stored on a machine-readable magnetic tape in a standard sequential format. The volume of data is too large to be printed here. Most of the data are not available elsewhere (at least not in machine-readable form) and may be useful to other investigators.

The tape-file format (on standard labeled tape) is as follows: File number is 1, data set name is APENDX.A.RECHARGE; tape is a high-density (6250 BPI) tape with EBCDIC coding; record format is fixed blocked; logical record length is 80; block size is 4,000, number of blocks is approximately 223; and number of records is 11,107.

Each record contains 10 data fields, each field is of length 8 in G8.0 format. The first 3 data fields are (1) year as number past 1900 (for example, "77" is 1977), (2) column in model grid, and (3) row in model grid. The other 7 data fields, all in 1,000 acre-ft/yr are (1) excess precipitation, (2) ungaged runoff from small streams, (3) river losses (+, or positive) and gains (−, or negative), (4) evapotranspiration of applied irrigation water, (5) surface water diverted to irrigation districts, (6) agricultural pumpage, and (7) municipal pumpage.

A duplicate of the tape (tape no. 112312) may be obtained from

U.S. Geological Survey, WRD
ATTN: Computer Specialist
Federal Building, Rm. W–2234
2800 Cottage Way
Sacramento, CA 95825

APPENDIX B: *Aquifer properties used in simulations*

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d)				Aquifer thickness (ft)				Percentage of fine-grained sediment				Leakance (TK) (x 10^{-6} d^{-1})					Inelastic storage coefficient	1961 critical head (ft)
			Layer				Layer				Layer				Predevelopment			Postdevelopment			
															Between layers			Between layers		Layers	Layers
			1	2	3	4	1	2	3	4	1	2	3	4	1-2	2-3	3-4	2-3	3-4	2-3	2-3
4	3	0.09	6.5	6.6	6.5	6.5	1,350	150	250	100	57	58	56	61	1.9	7.2	8.1	140	8.1	0.0370	340
4	4	0.04	1.3	1.3	1.3	1.4	1,020	288	100	87	57	58	56	61	0	0	11	0	11	0.0260	480
4	5	0.08	–	–	0.2	4.8	0	0	180	100	57	58	56	61	0	0	31	0	31	0.0000	555
5	3	0.07	1.3	1.3	1.3	3.0	1,220	100	500	200	57	58	56	61	6.7	15	13	92	13	0.0560	247
5	4	0.06	1.3	1.2	1.2	3.2	1,180	382	218	200	57	58	56	61	0.64	1.7	2.3	90	2.3	0.0410	330
5	5	0.05	–	0.3	3.1	1.7	0	162	238	200	57	58	56	61	0	3.8	3.4	90	3.4	0.0370	470
6	3	0.08	5.3	5.3	5.3	6.0	1,150	303	300	200	57	58	56	61	37	90	120	90	120	0.0620	160
6	4	0.06	2.5	2.5	2.5	2.4	1,540	255	230	200	57	58	56	61	5.2	19	21	110	21	0.0390	270
6	5	0.05	2.1	2.1	2.1	2.1	150	340	170	100	57	58	56	61	0.14	0.14	0.26	110	0.26	0.0390	305
6	6	0.04	0.3	0.3	0.3	0.3	260	50	100	62	57	58	56	61	0.3	0.63	0.57	75	0.57	0.0130	580
7	3	0.10	7.3	7.2	7.3	7.0	1,110	400	300	300	57	58	56	61	36	77	140	77	140	0.0570	158
7	4	0.07	6.1	6.1	6.1	3.8	1,600	100	500	267	57	58	56	61	4.4	13	9.6	92	9.6	0.0560	212
7	5	0.06	2.8	2.8	2.8	2.8	489	460	251	200	57	58	56	61	0.31	0.41	0.64	76	0.64	0.0540	269
7	6	0.05	1.8	1.8	1.8	1.8	37	100	225	200	57	58	56	61	17	7.3	5.4	170	5.4	0.0280	442
8	3	0.10	6.7	6.7	6.7	6.7	885	200	300	300	57	58	56	61	0.011	0.026	0.02	110	0.02	0.0440	148
8	4	0.07	3.0	3.0	3.0	3.5	1,460	300	290	200	57	58	56	61	19	56	66	92	66	0.0580	175
8	5	0.07	2.8	2.8	2.8	4.4	1,000	200	300	178	57	58	56	61	0.0022	0.0065	0.0063	110	0.0063	0.0380	198
8	6	0.06	1.3	1.3	3.4	3.4	218	300	160	52	57	58	56	61	0.39	0.44	0.95	120	0.95	0.0400	327
8	7	0.10	2.5	2.5	5.0	7.7	78	50	75	22	57	58	56	61	2.4	2.5	3.2	110	3.2	0.0110	412
9	3	0.10	2.0	2.0	7.5	7.5	1,440	125	200	200	57	58	56	61	8	38	30	170	30	0.0330	140
9	4	0.08	2.8	2.8	5.2	5.2	1,290	50	240	200	57	58	56	61	0.17	0.78	0.5	190	0.5	0.0290	137
9	5	0.09	2.5	2.5	5.0	6.5	1,290	100	200	150	57	58	56	61	0.026	0.13	0.1	180	0.1	0.0230	139
9	6	0.08	2.5	2.5	5.2	5.1	421	175	184	100	57	58	56	61	2.1	3.5	4.3	150	4.3	0.0320	250
9	7	0.09	2.5	2.5	5.0	7.9	375	50	50	25	57	58	56	61	0.94	4	5.3	150	5.3	0.0081	386
10	3	0.08	1.2	1.2	1.2	5.9	1,050	250	400	352	–	42	42	63	29	66	46	150	46	0.0520	106
10	4	0.10	1.6	1.6	7.8	7.8	965	415	300	285	–	42	42	63	0.0086	0.019	0.02	140	0.02	0.0570	104
10	5	0.10	2.1	2.1	8.6	8.6	510	245	500	265	57	58	56	61	40	41	39	73	39	0.0640	105
10	6	0.09	1.6	1.6	6.5	6.5	620	400	400	200	57	58	56	61	36	46	60	68	60	0.0560	160
10	7	0.05	0.8	0.8	1.3	1.9	42	50	100	25	57	58	56	61	27	17	20	76	20	0.0110	289

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) ($\times 10^{-6}$ d^{-1}) Predevelopment between layers 1-2	2-3	3-4	Postdevelopment between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
11	3	0.07	2.4	2.5	3.7	3.7	725	200	405	400	-	42	42	63	92	160	100	160	100	0.0510	80
11	4	0.11	8.3	8.3	8.3	9.2	830	500	250	250	-	42	42	63	12	23	28	130	28	0.0560	82
11	5	0.10	8.6	8.6	8.6	8.6	557	550	233	200	57	58	56	61	6.6	9.4	17	69	17	0.0530	72
11	6	0.09	2.2	2.2	6.1	6.1	840	400	300	300	57	58	56	61	4.9	8.8	10	77	10	0.0610	58
11	7	0.07	-	3.9	4.1	4.1	0	70	200	200	57	58	56	61	0	33	21	110	21	0.0210	64
12	3	0.07	5.3	5.3	3.2	3.2	599	400	300	341	-	42	42	63	88	140	200	140	200	0.0810	62
12	4	0.09	6.6	6.6	6.9	6.9	790	365	340	300	-	42	42	63	7.2	13	12	140	12	0.0580	46
12	5	0.09	6.6	6.6	6.6	6.6	695	760	140	100	-	42	42	63	15	27	84	110	84	0.0510	25
12	6	0.09	0.5	0.6	3.5	6.6	505	400	300	300	59	51	42	60	63	95	100	120	100	0.0510	30
12	7	0.06	2.0	2.1	2.9	2.9	25	300	300	200	59	51	42	60	160	98	110	150	110	0.0590	44
13	3	0.07	6.5	6.5	6.5	3.5	242	100	483	400	-	42	42	63	250	170	160	170	160	0.0670	53
13	4	0.06	4.0	4.0	4.0	2.9	390	400	400	200	-	42	42	63	57	61	70	120	70	0.0730	16
13	5	0.10	8.6	8.6	8.6	8.7	397	650	228	200	-	42	42	63	15	19	31	110	31	0.0460	3
13	6	0.08	5.3	5.3	5.2	5.2	775	825	100	75	59	51	42	60	9.1	17	92	89	92	0.0580	1
13	7	0.06	4.0	4.0	3.2	3.2	580	500	300	200	59	51	42	60	40	63	97	110	97	0.0920	6
13	8	0.05	0.1	0.1	2.4	2.3	45	600	200	175	59	51	42	60	13	11	23	110	23	0.0580	38
14	2	0.06	-	2.5	2.5	2.5	0	50	100	100	-	42	42	63	0	1.7	1	330	1	0.0110	80
14	3	0.06	-	5.1	4.3	3.1	0	200	250	300	-	42	42	63	0	0.026	0.016	160	0.016	0.0270	38
14	4	0.06	2.8	2.8	2.9	2.9	585	420	275	200	-	42	42	63	14	22	26	140	26	0.0310	10
14	5	0.07	1.3	1.2	5.8	5.0	670	320	300	280	-	42	42	63	18	33	28	160	28	0.0370	-6
14	6	0.09	2.5	2.5	7.0	7.0	983	400	282	200	-	42	42	63	5.2	12	14	140	14	0.0470	-12
14	7	0.02	1.3	1.2	3.7	3.4	755	400	300	200	59	51	42	60	61	120	160	120	160	0.0500	-10
14	8	0.05	-	0.8	1.2	2.6	0	400	300	200	59	51	42	60	0	78	100	120	100	0.0610	7
14	9	0.05	-	0.3	1.0	2.0	0	50	75	88	59	51	42	60	0	7.1	4.8	180	4.8	0.0110	45
15	2	0.10	-	7.0	6.9	7.0	0	50	75	100	-	42	42	63	0	20	11	330	11	0.0096	65
15	3	0.08	2.4	2.4	5.0	5.5	120	300	275	200	-	60	49	40	4.6	3.5	5.2	110	5.2	0.0370	18
15	4	0.06	3.1	3.1	3.1	3.1	785	300	290	200	-	42	42	63	29	61	61	170	61	0.0120	-3
15	5	0.06	2.5	2.5	9.6	3.3	665	300	290	200	-	42	42	63	8.9	16	16	170	16	0.0280	-22
15	6	0.09	2.5	2.5	3.1	6.9	750	300	300	200	-	42	42	63	27	54	54	160	54	0.0390	-28
15	7	0.08	2.5	2.5	5.1	5.1	840	230	300	270	-	42	42	63	26	61	46	180	46	0.0260	-26
15	8	0.06	0.1	0.1	1.3	2.6	850	300	200	300	64	60	58	56	12	30	31	100	31	0.0500	-15
15	9	0.05	-	0.1	0.1	1.5	0	200	300	288	64	60	58	56	0	2	1.7	100	1.7	0.0470	45

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK)(x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	Predevelopment Between layers 2-3	Predevelopment Between layers 3-4	Postdevelopment Between layers 2-3	Postdevelopment Between layers 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
16	3	0.07	-	4.4	3.4	3.4	0	105	275	100	-	60	49	40	0	15	17	180	17	0.0150	6
16	4	0.06	0.4	0.4	1.3	2.9	378	450	250	200	-	42	42	63	30	38	49	140	49	0.0280	-2
16	5	0.06	2.2	2.2	2.5	3.0	460	310	390	300	-	42	42	63	1.4	1.8	1.5	140	1.5	0.0470	-20
16	7	0.08	8.6	8.6	8.6	5.6	40	400	400	200	64	60	58	56	44	25	34	63	34	0.0500	-40
16	8	0.06	1.3	1.3	2.5	2.5	650	340	510	150	64	60	58	56	47	58	76	60	76	0.0790	-28
16	9	0.06	1.3	1.2	2.9	3.7	370	450	300	300	64	60	58	56	39	45	58	67	58	0.0680	33
16	10	0.04	-	0.1	0.3	1.4	0	50	150	147	64	60	58	56	0	5.4	3.7	70	3.7	0.0150	75
17	3	0.06	-	3.7	2.8	2.8	0	75	200	200	-	60	49	40	0	31	25	250	25	0.0110	-10
17	4	0.08	1.0	1.0	2.5	6.0	230	400	430	200	-	60	49	40	99	78	150	78	150	0.0290	0
17	7	0.08	2.5	2.5	5.1	5.6	1,030	500	300	200	-	60	49	40	19	35	68	79	68	0.0660	-49
17	8	0.07	2.5	2.5	4.4	4.4	1,150	500	300	190	64	60	58	56	17	38	64	63	64	0.0650	-42
17	9	0.07	1.3	1.2	4.4	4.2	980	600	300	140	64	60	58	56	9.7	18	38	56	38	0.0820	10
17	10	0.10	-	2.3	5.0	8.3	0	100	150	200	64	60	58	56	0	6.9	5.1	84	5.1	0.0220	110
18	3	0.08	-	2.5	5.4	5.1	0	75	140	100	-	-	-	-	0	27	24	200	24	0.0091	12
18	4	0.10	-	2.5	6.3	8.6	0	400	200	275	-	60	49	40	0	23	37	100	37	0.0230	-10
18	5	0.08	2.5	2.5	2.5	5.8	448	440	275	275	-	60	49	40	0.052	0.063	0.1	88	0.1	0.0200	-40
18	6	0.08	2.5	2.5	2.6	5.1	1,050	500	275	200	-	60	49	40	18	34	69	81	69	0.0570	-54
18	7	0.08	2.5	2.5	3.0	5.8	1,450	325	350	300	64	60	58	56	13	36	39	75	39	0.0740	-55
18	8	0.07	2.5	2.5	7.9	4.1	1,320	325	375	300	64	60	58	56	9.1	23	25	72	25	0.0530	-58
18	9	0.07	2.5	2.5	4.2	4.8	1,120	500	300	200	64	60	58	56	4	8.5	14	63	14	0.0750	-24
18	10	0.08	-	4.2	5.7	5.7	0	50	150	100	64	60	58	56	0	3.2	2.6	84	2.6	0.0150	45
19	3	0.11	-	2.5	5.0	8.6	0	150	150	100	-	-	-	-	0	3.8	4.6	280	4.6	0.0200	-58
19	4	0.10	-	2.0	5.0	7.7	0	100	300	350	-	-	-	-	0	5.7	3.5	170	3.5	0.0160	-38
19	5	0.08	6.6	6.6	6.0	6.0	235	500	300	400	-	60	49	40	66	62	90	79	90	0.0280	-46
19	6	0.07	3.8	3.7	3.7	4.7	496	900	222	100	-	60	49	40	6	7.2	32	54	32	0.0620	-60
19	7	0.08	2.5	2.5	3.7	5.6	1,120	700	200	300	64	60	58	56	5.3	11	21	56	21	0.0730	-26
19	8	0.07	3.8	3.7	8.6	4.9	1,420	400	500	300	64	60	58	56	9.5	21	24	56	24	0.0750	-60
19	9	0.06	4.9	4.9	4.9	3.5	745	500	450	275	64	60	58	56	8.6	12	16	53	16	0.0780	-40
19	10	0.05	0.7	0.7	1.6	1.6	125	100	175	100	59	61	60	59	0.0065	0.0065	0.0057	170	0.0057	0.0210	5
20	3	0.06	-	-	2.8	2.8	0	0	195	100	-	-	-	-	0	0	4.9	0	4.9	0.0000	-30
20	4	0.08	-	2.8	5.8	5.8	0	100	150	200	-	-	-	-	0	64	46	170	46	0.0120	-62
20	5	0.08	6.3	6.2	8.5	5.8	35	100	400	300	-	60	49.	40	54	16	13	140	13	0.0310	-44
20	6	0.07	1.3	1.3	2.1	4.5	118	800	400	300	-	60	49	40	35	28	60	52	60	0.0420	-61
20	7	0.08	1.5	1.5	1.6	5.2	715	800	600	200	-	60	49	40	42	46	140	46	140	0.0970	-29
20	8	0.08	2.5	2.5	7.2	5.3	1,370	600	400	400	64	60	58	56	3.7	7.8	10	50	10	0.0830	-24
20	9	0.06	2.5	2.5	8.2	4.0	1,090	600	450	300	64	60	58	56	2.1	3.6	5.3	48	5.3	0.1100	-32
20	10	0.10	-	2.5	3.3	7.3	0	350	400	250	59	61	60	59	0	8.1	9.5	64	9.5	0.0360	40
20	11	0.07	1.3	1.2	3.0	4.8	300	150	100	50	59	61	60	59	0.16	0.29	0.49	190	0.49	0.0190	104

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK)(x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	2-3	3-4	Postdevelopment Between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
21	4	0.06	-	3.3	3.3	3.3	0	75	100	200	-	-	-	-	0	23	13	170	13	0.0110	4
21	5	0.06	-	3.0	3.0	3.2	0	200	300	250	-	-	-	-	0	40	36	140	36	0.0390	-32
21	6	0.08	3.0	3.0	5.9	5.9	468	400	350	300	-	60	49	40	7.3	8.4	12	86	12	0.0560	-60
21	7	0.11	2.5	2.5	5.0	9.5	1,120	600	175	225	64	60	58	56	10	23	47	65	47	0.0800	-31
21	8	0.08	2.5	2.5	7.2	6.1	1,740	225	500	275	59	61	60	59	20	54	51	66	51	0.0860	-23
21	9	0.09	2.5	2.5	7.0	6.7	1,490	370	400	200	59	61	60	59	5.6	13	17	62	17	0.0480	-3
21	10	0.09	2.5	2.5	2.0	7.6	2,080	265	535	200	59	61	60	59	3.6	10	11	60	11	0.0910	12
21	11	0.05	-	2.5	4.6	2.6	0	100	400	160	59	61	60	59	0	9.2	8.2	96	8.2	0.0390	66
22	4	0.06	-	3.3	3.3	3.3	0	50	100	200	-	-	-	-	0	32	16	200	16	0.0110	8
22	5	0.06	-	5.4	5.4	3.0	0	200	250	300	-	-	-	-	0	3.1	2.5	98	2.5	0.0340	-23
22	6	0.06	2.5	2.5	4.8	3.4	120	325	375	300	-	-	-	-	130	86	89	98	89	0.0670	-51
22	7	0.11	5.0	5.0	7.5	9.5	715	600	200	200	59	61	60	59	17	27	56	59	56	0.0810	-31
22	8	0.09	5.0	5.0	9.0	7.6	1,420	400	300	300	59	61	60	59	27	68	5	68	93	0.0690	-27
22	9	0.09	5.0	5.0	10.1	7.8	1,850	350	352	300	59	61	60	59	22	68	200	68	200	0.0570	-21
22	10	0.07	2.5	2.5	4.1	4.7	1,710	160	520	300	59	61	60	59	22	60	51	70	51	0.0680	43
22	11	0.08	2.5	2.5	3.3	6.0	645	400	300	200	59	61	60	59	21	31	44	68	44	0.0550	8
23	4	0.07	-	0.3	4.3	4.3	0	50	50	100	-	-	-	-	0	1.9	1.3	230	1.3	0.0091	14
23	5	0.06	-	1.3	1.9	1.8	0	100	300	250	49	48	52	40	0	39	31	160	31	0.0190	-37
23	6	0.07	-	2.5	10.7	4.8	0	400	300	300	49	48	52	40	0	93	120	96	120	0.0590	-52
23	7	0.11	2.5	2.5	5.0	9.1	214	1,000	200	200	49	48	52	40	25	25	7.3	57	81	0.0850	-44
23	8	0.10	2.5	2.5	3.1	7.7	1,020	700	250	350	59	61	60	59	28	45	89	45	89	0.0970	-37
23	9	0.07	2.5	2.5	4.6	4.7	1,440	600	450	350	59	61	60	59	24	45	71	45	71	0.0890	-30
23	10	0.07	2.5	2.5	3.6	4.9	1,280	650	450	350	59	61	60	59	25	43	88	43	88	0.1000	-25
23	11	0.08	2.5	2.5	2.9	5.0	1,320	600	550	300	59	61	60	59	22	36	50	41	50	0.1100	-14
24	4	0.08	-	-	2.5	6.2	0	0	50	81	49	48	52	40	0	0	1.9	0	1.9	0.0000	18
24	5	0.07	1.3	1.2	3.7	3.9	40	50	260	200	49	48	52	40	68	19	14	74	14	0.0150	-26
24	6	0.08	6.0	6.0	6.0	5.1	30	500	400	400	49	48	52	40	1.6	0.93	1.1	74	1.1	0.0410	-48
24	7	0.09	2.5	2.5	5.3	6.9	212	800	500	400	59	61	60	59	13	10	15	37	15	0.0950	-63
24	8	0.11	9.8	9.7	9.9	9.9	610	800	550	450	59	61	60	59	14	14	20	35	20	0.1100	-54
24	9	0.07	2.5	2.5	4.2	4.7	1,120	800	550	450	53	64	58	52	27	36	72	36	72	0.1100	-74
24	10	0.07	1.3	1.2	2.7	4.2	1,230	500	550	550	59	61	60	59	0.011	0.017	0.017	45	0.017	0.0930	-78
24	11	0.08	2.5	2.5	4.9	5.3	1,100	500	500	350	53	64	58	52	2.8	4.1	5.3	49	5.3	0.1100	-40
24	12	0.08	-	2.8	2.8	4.7	0	200	300	200	53	64	58	52	0	33	36	98	36	0.0420	-2

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	Predevelopment 2-3	Predevelopment 3-4	Postdevelopment Between layers 2-3	Postdevelopment 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
25	4	0.07	-	1.2	2.5	3.8	0	50	50	100	49	48	52	40	0	3	2.3	92	2.3	0.0087	139
25	5	0.09	2.5	2.5	5.7	5.4	180	250	350	300	49	48	52	40	46	32	32	95	32	0.0450	10
25	6	0.08	2.5	2.5	6.4	5.9	630	350	350	300	49	48	52	40	70	95	180	95	180	0.0540	-44
25	7	0.08	2.5	2.5	5.7	5.6	968	590	260	250	49	48	52	40	7.9	14	26	79	26	0.0650	-61
25	8	0.10	2.5	2.5	6.2	7.8	1,400	550	250	200	53	64	58	52	2.7	6	12	60	12	0.0650	-55
25	9	0.12	2.5	2.5	5.0	10.5	2,010	700	150	150	53	64	58	52	0.86	2.4	7.9	56	7.9	0.0670	-70
25	10	0.07	2.5	2.5	4.0	4.4	1,910	300	350	350	53	64	58	52	7.2	22	23	75	23	0.0590	-68
25	11	0.07	2.5	2.5	2.8	4.3	1,410	400	350	350	53	64	58	52	0.22	0.49	0.58	65	0.58	0.0690	-74
25	12	0.06	-	1.7	1.7	2.8	0	50	500	300	53	64	58	52	0	33	24	71	24	0.0480	-60
25	13	0.08	-	0.3	2.5	4.9	0	50	100	175	53	64	58	52	0	22	13	59	13	0.0130	-70
26	4	0.06	-	1.2	2.2	2.2	0	50	100	100	49	48	52	40	0	1.9	1.6	190	1.6	0.0150	140
26	5	0.09	-	1.3	5.4	5.4	0	160	200	140	49	48	52	40	0	4.5	5.1	180	5.1	0.0260	10
26	6	0.08	1.3	1.2	5.4	5.4	440	600	300	200	49	48	52	40	7.6	8.6	16	75	16	0.0640	-49
26	7	0.07	1.3	1.2	7.4	4.5	483	600	400	225	49	48	52	40	6.5	6.9	11	67	11	0.0650	-90
26	8	0.09	1.3	1.2	3.3	7.0	1,090	750	264	200	49	48	52	40	7.3	13	30	67	30	0.0770	-94
26	9	0.11	1.3	1.3	3.7	9.7	1,730	800	200	173	53	64	58	52	7.2	16	50	47	50	0.0860	-76
26	10	0.06	1.3	1.2	3.0	3.9	1,830	600	370	200	53	64	58	52	9.5	21	40	50	40	0.0730	-74
26	11	0.09	1.3	1.3	5.0	6.6	1,400	900	200	100	53	64	58	52	3	5.7	23	43	23	0.0940	-74
26	12	0.06	-	7.2	7.3	3.3	0	200	550	450	53	64	58	52	0	10	8.3	67	8.3	0.0670	-70
26	13	0.06	-	0.3	1.3	3.1	0	50	100	50	53	64	58	52	0	7.1	7.6	110	7.6	0.0140	-74
27	5	0.08	-	4.3	4.3	4.3	0	100	200	100	49	48	52	40	0	8.2	8.6	220	8.6	0.0210	7
27	6	0.07	1.3	1.3	1.2	4.0	380	345	325	300	49	48	52	40	62	65	75	99	75	0.0460	-26
27	7	0.07	1.3	1.2	1.8	4.5	948	260	375	325	49	48	52	40	0.57	1.1	1	100	1	0.0440	-85
27	8	0.07	2.5	2.5	5.0	4.7	1,320	480	250	250	49	48	52	40	13	30	48	92	48	0.0590	-100
27	9	0.09	1.3	1.2	6.8	6.8	1,790	600	215	200	49	48	52	40	1.9	5.6	12	83	12	0.0690	-76
27	10	0.12	1.3	1.3	6.2	6.2	1,710	725	175	100	49	48	52	40	3.4	9.2	31	76	31	0.0690	-75
27	11	0.09	1.3	1.3	5.3	6.8	1,610	300	400	296	53	64	58	52	8.7	21	24	70	24	0.0580	-72
27	12	0.06	2.5	2.5	5.7	3.6	1,080	100	350	350	49	48	52	40	17	44	31	140	31	0.0410	-73
27	13	0.07	-	2.5	4.1	4.1	0	100	150	150	53	64	58	52	0	25	23	98	23	0.0230	-79
28	5	0.05	0.3	0.2	0.5	0.7	10	325	50	25	49	48	52	40	0.83	0.74	3.7	180	3.7	0.0260	45
28	6	0.06	1.3	1.2	1.9	2.4	395	275	350	350	49	48	52	40	48	50	49	110	49	0.0270	-48
28	7	0.07	1.3	1.3	2.6	4.5	900	300	350	350	49	48	52	40	21	38	39	100	39	0.0360	-66
28	8	0.07	1.3	1.3	4.6	4.7	1,420	160	450	375	49	48	52	40	19	48	39	110	39	0.0440	-65
28	9	0.07	1.3	1.3	5.0	4.9	1,600	800	100	100	49	48	52	40	5.8	15	73	76	73	0.0900	-74
28	10	0.07	1.3	1.2	5.0	6.3	2,000	500	200	200	100	44	37	57	4.1	31	49	160	49	0.0660	-80
28	11	0.07	1.3	1.2	5.0	7.0	1,490	375	325	300	-	64	56	62	30	71	81	73	81	0.0610	-82
28	12	0.07	1.3	1.3	5.0	7.5	800	200	300	300	49	48	52	40	54	110	96	130	96	0.0460	-84
28	13	0.07	-	2.5	4.3	4.3	0	300	200	200	-	64	56	62	0	14	18	100	18	0.0460	-95

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) ($\times 10^{-6}$ d^{-1}) Predevelopment Between layers 1-2	Predevelopment 2-3	Predevelopment 3-4	Post-development Between layers 2-3	Post-development 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
29	5	0.06	-	0.5	0.5	0.7	0	100	200	200	100	44	37	57	0	4.4	2.7	400	2.7	0.0200	32
29	6	0.07	2.5	2.5	3.0	3.6	175	375	300	325	100	44	37	57	6.1	7.5	7	170	7	0.0190	-33
29	7	0.06	1.3	1.3	3.7	3.1	1,050	250	300	350	100	44	37	57	13	69	49	210	49	0.0220	-70
29	8	0.09	1.3	1.2	4.0	7.4	1,340	370	400	225	100	44	37	57	8.3	40	45	150	45	0.0820	-80
29	9	0.08	1.3	1.2	2.5	6.0	1,780	600	223	200	100	44	37	57	2.4	14	25	130	25	0.0630	-71
29	10	0.07	1.3	1.3	2.5	4.2	1,900	900	50	50	100	44	37	57	1.6	8.8	78	110	78	0.0860	-81
29	11	0.10	2.5	2.5	5.0	8.2	1,600	700	150	150	-	64	56	62	7.3	17	51	58	51	0.0770	-82
29	12	0.14	2.5	2.5	6.2	13.6	600	375	325	300	-	64	56	62	57	73	200	73	200	0.0650	-80
29	13	0.09	-	1.2	4.5	4.5	0	50	150	150	-	64	56	62	0	73	48	130	48	0.0150	-80
30	5	0.05	-	0.2	0.2	1.3	0	200	200	200	100	44	37	57	0	7.4	6.4	160	6.4	0.0260	119
30	6	0.06	-	5.5	2.4	2.3	0	350	350	300	100	44	37	57	0	17	16	170	16	0.0120	18
30	7	0.06	1.3	1.3	2.5	3.0	1,050	300	300	300	100	44	37	57	11	52	45	190	45	0.0067	-70
30	8	0.08	2.5	2.5	6.9	5.7	1,550	300	390	300	100	44	37	57	7.9	48	42	170	42	0.0430	-70
30	9	0.08	2.5	2.5	5.0	6.1	1,610	450	275	275	100	44	37	57	26	160	1200	160	1200	0.0620	-69
30	10	0.10	2.5	2.5	5.0	7.6	1,600	600	200	200	100	44	37	57	9.7	54	96	140	96	0.0630	-80
30	11	0.08	1.3	1.3	2.5	4.7	2,500	800	100	100	-	64	56	62	12	38	180	54	180	0.0800	-81
30	12	0.08	2.5	2.5	4.1	5.5	1,200	450	300	250	-	64	56	62	26	51	72	68	72	0.0680	-100
31	5	0.05	-	0.3	0.2	0.5	0	150	200	300	100	44	37	57	0	6.8	3.9	240	3.9	0.0230	130
31	6	0.06	-	1.1	0.8	0.8	0	215	285	300	100	44	37	57	0	5.6	4	230	4	0.0056	-7
31	7	0.07	1.3	1.3	2.7	2.9	1,080	250	200	330	100	44	37	57	5.8	37	26	250	26	0.0063	-60
31	8	0.07	1.3	1.3	4.3	3.9	1,460	305	300	380	100	44	37	57	29	190	160	190	160	0.0270	-80
31	9	0.07	1.3	1.2	3.8	5.2	1,620	225	300	375	100	44	37	57	20	170	0.17	220	110	0.0340	-110
31	10	0.08	0.8	0.7	5.0	5.2	2,550	300	300	300	-	64	56	62	21	86	200	86	200	0.0370	-100
31	11	0.14	1.3	1.2	6.2	11.7	2,600	420	300	180	-	64	56	62	20	71	280	71	280	0.0340	-100
31	12	0.07	1.3	1.2	2.5	3.7	1,710	700	100	90	-	64	56	62	6.4	17	75	61	75	0.0890	-90
32	6	0.05	-	0.7	0.6	0.7	0	210	120	100	100	44	37	57	0	4.5	6.1	260	6.1	0.0050	60
32	7	0.07	3.5	3.5	3.9	3.9	245	300	300	300	100	44	37	57	15	23	20	190	20	0.0240	-60
32	8	0.07	4.8	4.8	4.8	4.4	1,050	250	250	300	100	44	37	57	10	59	45	230	45	0.0250	-100
32	9	0.06	3.8	3.7	3.8	3.7	1,780	200	300	300	100	44	37	57	5	47	0.26	240	33	0.0300	-120
32	10	0.08	5.0	5.0	5.0	6.4	2,300	320	300	280	100	44	37	57	1.7	16	15	190	15	0.0440	-100
32	11	0.13	3.8	3.7	3.7	11.8	3,200	200	400	200	-	64	56	62	16	77	78	87	78	0.0440	-110
32	12	0.08	2.5	2.5	2.3	5.5	2,600	350	300	150	-	64	56	62	3.9	15	23	79	23	0.0530	-80

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Col-umn	Row	Spe-cific yield	Hydraulic conductivity (ft/d)				Aquifer thickness (ft)				Percentage of fine-grained sediment				Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment			Leakance (TK) (x 10^{-6} d^{-1}) Post-development		Inelas-tic storage co-effi-cient	1961 crit-cal head (ft)
			Layer 1	Layer 2	Layer 3	Layer 4	Layer 1	Layer 2	Layer 3	Layer 4	Layer 1	Layer 2	Layer 3	Layer 4	Between layers 1-2	Between layers 2-3	Between layers 3-4	Between layers 2-3	Between layers 3-4	Layers 2-3	Layers 2-3
33	6	0.08	-	4.9	4.9	4.9	0	100	100	90	100	44	37	57	0	13	12	580	12	0.0059	60
33	7	0.09	-	3.8	6.0	6.9	0	700	250	250	100	44	37	57	0	79	140	120	140	0.0620	-32
33	8	0.10	6.3	6.3	6.8	7.0	582	600	300	288	100	44	37	57	55	120	180	120	180	0.0730	-80
33	9	0.09	5.0	5.0	5.0	7.1	1,390	650	300	250	100	44	37	57	3.7	15	24	120	24	0.0930	-64
33	10	0.10	5.0	5.0	7.5	8.8	2,350	800	200	173	-	64	56	62	17	46	0.97	49	130	0.0780	-57
33	11	0.12	3.8	3.7	4.9	10.1	2,500	500	493	232	-	64	56	62	14	38	54	52	54	0.1100	-80
33	12	0.12	2.5	2.5	4.8	10.4	2,000	500	485	185	-	64	56	62	4.1	9.2	14	52	14	0.0820	-70
34	6	0.06	-	2.5	2.5	2.5	0	50	275	200	100	44	37	57	0	10	6	380	6	0.0150	70
34	7	0.08	5.0	5.0	10.3	6.8	145	450	250	250	100	44	37	57	3.9	4.6	5.7	160	5.7	0.0600	12
34	8	0.20	7.5	7.5	9.2	9.1	885	275	425	300	100	44	37	57	2.1	7.6	6.4	170	6.4	0.0510	-20
34	9	0.11	7.5	7.5	8.7	9.8	1,760	550	250	200	100	44	37	57	4.1	25	40	140	40	0.0620	-23
34	10	0.11	7.5	7.5	8.7	9.5	2,380	600	250	150	100	44	37	57	13	100	200	130	200	0.0780	-44
34	11	0.11	3.8	3.7	4.5	8.5	2,410	400	500	115	100	44	37	57	4.4	31	45	130	45	0.0680	-80
34	12	0.10	2.5	2.5	5.5	7.1	1,480	400	500	45	-	64	56	62	6	11	20	58	20	0.0800	-40
35	6	0.14	13.0	13.0	13.0	13.0	5	300	100	50	100	44	37	57	4.2	3.4	8.7	280	8.7	0.0270	89
35	7	0.12	-	7.5	9.7	9.8	0	600	300	280	100	44	37	57	0	83	120	120	120	0.0650	49
35	8	0.14	6.3	6.2	12.5	12.9	295	1,000	400	300	-	-	-	-	48	44	88	49	88	0.1100	13
35	9	0.14	7.5	7.5	11.3	12.7	1,350	1,200	224	200	100	44	37	57	16	49	150	76	150	0.1200	-15
35	10	0.09	5.0	5.0	5.0	6.5	1,480	1,300	320	140	-	-	-	-	25	43	230	43	230	0.1400	-43
35	11	0.12	5.0	5.0	7.5	9.7	1,360	900	700	192	-	64	56	62	8.1	10	20	32	20	0.1900	-60
36	6	0.11	-	9.9	9.9	9.9	0	100	200	200	-	-	-	-	0	0.2	0.15	230	0.15	0.0230	80
36	7	0.11	-	7.5	9.7	9.8	0	500	450	350	-	-	-	-	0	69	82	73	82	0.0750	51
36	8	0.07	4.8	4.8	4.8	4.8	580	800	310	300	-	-	-	-	17	21	38	62	38	0.0970	10
36	9	0.11	7.5	7.5	8.7	9.5	1,320	1,100	250	161	-	-	-	-	2.9	5.2	17	51	17	0.1200	-19
36	10	0.09	5.0	5.0	6.2	7.7	1,540	1,200	350	135	-	-	-	-	5.4	9.6	31	45	31	0.1400	-31
36	11	0.12	3.8	3.7	5.0	10.5	1,540	900	700	172	-	-	56	67	25	37	170	37	170	0.1400	-29
37	5	0.20	-	1.3	1.3	1.3	0	50	100	100	-	-	-	-	0	1.4	1.1	460	1.1	0.0120	170
37	6	0.05	0.1	0.1	2.5	2.4	25	300	200	100	-	-	-	-	5.7	3.7	6.1	140	6.1	0.0450	95
37	7	0.06	1.3	1.2	3.3	3.3	320	500	200	260	-	-	-	-	7	8.2	13	98	13	0.0570	45
37	8	0.09	5.0	5.0	6.2	6.5	1,120	450	300	250	-	-	-	-	6.7	14	19	92	19	0.0660	13
37	9	0.14	7.5	7.5	11.2	13.2	1,780	700	150	149	-	-	-	-	6.2	18	52	81	52	0.0760	-8
37	10	0.08	3.8	3.7	5.0	5.0	2,300	600	250	165	-	-	-	-	13	44	91	81	91	0.0760	-18
37	11	0.09	3.8	3.7	4.8	6.7	1,590	575	480	219	-	-	56	67	29	56	130	55	130	0.0900	-19

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) ($\times 10^{-6}$ d^{-1}) Predevelopment Between layers 1-2	Predevelopment 2-3	Predevelopment 3-4	Post-development Between layers 2-3	Post-development 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
38	5	0.04	–	1.2	1.2	1.2	0	100	130	100	–	–	–	–	0	1.6	1.6	300	1.6	0.0180	160
38	6	0.06	–	1.7	2.5	2.5	0	500	250	200	–	–	–	–	0	0.32	0.54	92	0.54	0.0740	95
38	7	0.10	6.3	6.3	7.3	7.3	385	600	300	300	–	–	–	–	3.8	4.2	6.3	77	6.3	0.0730	46
38	8	0.08	5.0	5.0	5.8	5.8	1,150	700	320	200	–	–	–	–	12	22	44	68	44	0.0890	14
38	9	0.06	2.5	2.5	3.7	2.0	1,900	900	200	143	–	–	–	–	25	63	250	63	250	0.0970	7
38	10	0.10	2.5	2.5	3.0	6.6	2,140	800	200	234	–	–	–	–	2.8	8.2	1.8	69	19	0.0870	-13
38	11	0.12	5.0	5.0	6.8	9.8	2,230	500	350	178	–	–	56	67	6.7	21	29	69	29	0.0710	-13
38	12	0.06	–	2.9	2.9	2.9	0	50	100	200	–	–	56	67	0	58	35	58	35	0.0066	38
39	5	0.14	–	12.5	13.5	13.5	0	140	160	100	–	–	–	–	0	3.5	4	230	4	0.0250	160
39	6	0.08	4.5	4.5	5.4	5.4	150	200	410	300	–	–	–	–	24	14	12	110	12	0.0450	110
39	7	0.09	5.0	5.0	6.5	6.5	935	250	370	300	–	–	–	–	29	55	51	110	51	0.0490	73
39	8	0.10	6.3	6.2	7.5	7.5	1,620	450	310	200	–	–	–	–	2.7	7.5	11	91	11	0.0640	28
39	9	0.11	3.8	3.7	6.2	8.5	2,210	600	250	104	–	–	–	–	7.6	25	61	81	61	0.0730	18
39	10	0.11	5.0	5.0	7.5	8.1	2,390	550	250	202	–	–	–	–	7.7	28	1.7	86	50	0.0670	0
39	11	0.12	7.5	7.5	9.5	9.8	1,910	600	440	243	–	–	56	67	25	57	210	57	210	0.0830	-8
39	12	0.11	–	8.5	8.5	8.5	0	50	200	170	–	–	56	67	0	23	14	190	14	0.0110	41
40	5	0.08	–	2.5	4.0	4.0	0	100	200	200	–	–	–	–	0	14	10	230	10	0.0250	207
40	6	0.09	3.8	3.8	5.0	5.0	375	300	230	100	–	–	–	–	13	17	27	130	27	0.0520	124
40	7	0.09	5.0	5.0	6.3	6.4	990	250	325	200	–	–	–	–	33	71	77	120	77	0.0440	81
40	8	0.13	6.3	6.3	7.5	10.0	1,790	500	200	100	–	–	–	–	12	39	91	98	91	0.0570	47
40	9	0.16	6.3	6.2	8.7	14.3	2,580	550	100	151	–	–	–	–	8.1	39	0.42	110	100	0.0540	25
40	10	0.12	6.3	6.2	7.5	8.2	2,770	550	100	192	–	–	–	–	2.2	11	39	110	25	0.0520	9
40	11	0.15	5.0	5.0	6.2	13.6	2,360	350	250	251	–	–	56	67	2.1	9	9.2	98	9.2	0.0460	-3
40	12	0.09	–	3.8	5.8	5.8	0	100	200	80	–	–	56	67	0	17	17	190	17	0.0180	34
41	5	0.07	–	3.5	3.5	3.5	0	50	100	100	–	–	–	–	0	6.5	4.9	460	4.9	0.0130	200
41	6	0.08	–	2.5	4.0	4.5	0	450	300	100	–	–	–	–	0	13	25	92	25	0.0690	122
41	7	0.08	3.8	3.7	4.5	4.5	890	400	280	100	–	–	–	–	21	39	70	100	70	0.0510	85
41	8	0.09	3.8	3.7	5.0	5.3	1,820	400	290	95	–	–	–	–	6	19	0.69	100	35	0.0550	46
41	9	0.10	6.3	6.3	7.5	7.8	2,530	400	250	159	–	–	–	–	1.8	8.3	32	110	13	0.0520	26
41	10	0.10	5.0	5.0	6.3	7.5	2,940	300	200	185	–	–	56	67	0.097	0.6	1.1	120	0.67	0.0380	14
41	11	0.10	6.3	6.2	7.0	7.4	2,240	350	250	235	–	–	56	67	4.3	18	19	98	19	0.0440	2
41	12	0.08	3.0	3.0	5.0	5.6	487	350	200	248	–	–	56	67	1.3	2	2	110	2	0.0370	8
41	13	0.07	–	2.5	3.8	5.0	0	50	100	100	–	–	56	67	0	10	6.8	380	6.8	0.0063	98

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) ($\times 10^{-6}$ d^{-1}) Predevelopment between layers 1-2	Predevelopment between layers 2-3	Predevelopment between layers 3-4	Post-development between layers 2-3	Post-development between layers 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
42	5	0.06	-	3.2	3.2	3.2	0	50	150	100	-	-	-	-	0	5	4	170	4	0.0170	190
42	6	0.06	-	2.5	3.3	3.3	0	305	600	100	-	-	-	-	0	53	69	76	69	0.0750	124
42	7	0.07	3.8	3.7	3.3	3.3	581	500	512	46	-	-	-	-	51	55	3.2	68	99	0.0750	81
42	8	0.09	5.0	5.0	10.0	5.9	1,480	840	200	141	-	-	-	-	7.4	16	1.5	66	50	0.0820	40
42	9	0.10	5.0	5.0	6.3	7.5	2,240	500	420	181	-	-	-	-	25	75	2.7	75	120	0.0710	0
42	10	0.14	5.0	5.0	5.0	13.4	2,560	450	400	221	-	-	56	67	8	27	32	69	32	0.0630	22
42	11	0.14	6.3	6.2	7.5	12.4	1,940	450	396	255	-	-	56	67	2.3	6.2	7.1	69	7.1	0.0840	14
42	12	0.11	7.5	7.5	9.1	9.1	16	500	355	219	-	-	56	67	6.9	4	5.2	69	5.2	0.0630	5
43	5	0.16	-	15.9	15.8	15.9	0	100	250	100	-	-	-	-	0	22	22	200	22	0.0310	185
43	6	0.09	3.8	3.7	7.1	7.1	38	400	600	25	-	-	-	-	33	15	23	69	23	0.0760	74
43	7	0.10	7.3	7.2	7.5	7.5	547	600	600	102	-	-	-	-	60	58	2.5	58	130	0.0880	27
43	8	0.06	3.8	3.8	4.0	1.7	1,200	800	340	136	-	-	-	-	19	33	1.8	61	79	0.0890	3
43	9	0.10	2.5	2.5	3.7	7.2	1,890	500	600	206	-	-	-	-	24	52	1.3	63	71	0.0840	8
43	10	0.17	7.5	7.5	10.0	16.7	2,340	800	100	237	-	-	56	67	1.5	5.1	11	68	11	0.0900	20
43	11	0.13	6.3	6.3	7.5	12.2	720	600	297	291	-	-	56	67	0.035	0.05	0.065	66	0.065	0.0860	35
43	12	0.12	-	5.3	9.2	9.2	0	500	508	143	-	-	56	67	0	40	57	58	57	0.0780	60
44	5	0.14	-	12.5	12.5	12.5	0	200	100	100	-	-	-	-	0	32	48	170	48	0.0260	191
44	6	0.09	2.4	2.4	2.4	6.7	455	300	340	20	-	-	-	-	22	25	45	110	45	0.0450	12
44	7	0.08	3.0	3.0	3.7	4.7	1,070	400	400	71	-	-	62	66	38	62	0.02	62	200	0.0820	30
44	8	0.07	5.0	5.0	5.4	3.5	1,800	400	400	151	-	-	62	66	25	62	0.78	62	220	0.0950	-11
44	9	0.10	5.0	5.0	5.8	7.1	1,540	800	600	209	-	-	62	66	11	16	0.73	36	25	0.1400	4
44	10	0.18	7.5	3.7	4.1	18.9	1,690	1,000	395	239	-	-	56	67	2.7	5.1	9.6	43	9.6	0.1300	50
44	11	0.09	7.5	3.8	4.2	5.3	1,120	765	500	328	-	-	56	67	0.7	0.99	1.3	47	1.3	0.1200	5
44	12	0.05	-	6.5	6.5	2.2	0	400	440	1	-	-	56	67	0	1.7	3.1	62	3.1	0.0650	0
45	5	0.05	-	2.3	2.3	2.3	0	100	320	200	-	70	50	62	0	21	17	66	17	0.0360	109
45	6	0.07	3.0	3.0	3.5	3.5	332	450	300	300	-	-	62	66	64	61	65	68	65	0.0500	76
45	7	0.08	3.8	3.8	5.1	5.1	1,300	350	500	225	-	-	62	66	34	57	0.064	57	150	0.0660	52
45	8	0.11	6.3	6.2	7.8	9.3	2,160	250	660	189	-	-	62	66	23	52	1	52	90	0.1000	15
45	9	0.11	2.5	2.5	2.1	8.5	2,400	325	520	258	-	-	56	67	14	42	2.1	68	41	0.0800	94
45	10	0.13	3.8	1.9	2.7	12.3	2,080	500	500	306	-	-	56	67	5.3	13	14	58	14	0.1000	65
45	11	0.08	7.5	3.2	4.5	4.7	1,340	500	410	364	-	-	56	67	0.13	0.25	0.25	64	0.25	0.0480	-120
45	12	0.16	-	2.2	3.2	13.4	0	300	270	1	-	-	56	67	0	1.6	3.1	67	3.1	0.0430	15
45	13	0.10	-	2.5	2.5	6.2	0	50	100	1	-	-	56	67	0	0.51	0.72	190	0.72	0.0110	130

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Col-umn	Row	Spe-cific yield	Hydraulic conductivity (ft/d) Layer 1	2	3	4	Aquifer thickness (ft) Layer 1	2	3	4	Percentage of fine-grained sediment Layer 1	2	3	4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	2-3	3-4	Post-development Between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
46	5	0.06	-	2.5	3.0	3.0	0	300	460	200	-	70	50	62	0	21	27	79	27	0.0640	185
46	6	0.09	5.0	5.0	6.3	6.9	711	1,100	200	127	-	70	50	62	9	12	57	40	57	0.1100	110
46	7	0.10	6.3	6.2	6.5	7.6	1,580	900	460	132	-	-	62	66	22	38	0.55	38	170	0.1300	53
46	8	0.11	7.5	7.5	8.0	9.3	1,970	700	600	263	-	-	62	66	21	39	140	39	140	0.1100	36
46	9	0.10	2.5	2.5	1.9	8.1	2,170	900	265	299	-	-	56	67	2.4	6	10	130	10	0.1100	90
46	10	0.13	6.3	3.1	3.5	11.6	2,540	600	500	375	-	-	56	67	9.6	26	28	880	28	0.0540	65
46	11	0.09	6.3	2.7	4.5	6.0	1,630	600	598	295	-	-	56	67	1.5	2.5	3	53	3	0.0510	-70
46	12	0.06	-	2.8	2.7	3.8	0	775	600	1	-	-	56	67	0	0.68	1.5	86	1.5	0.1400	-155
47	4	0.08	-	6.3	6.3	6.3	0	100	127	145	-	70	50	62	0	22	19	260	19	0.0300	245
47	5	0.08	3.8	3.7	4.4	4.4	18	600	300	270	-	70	50	62	20	15	27	61	27	0.0760	182
47	6	0.10	7.5	7.5	8.3	8.3	890	1,000	350	250	-	70	50	62	29	37	99	39	99	0.1100	122
47	7	0.13	6.3	6.2	5.8	11.6	1,640	1,000	510	230	-	70	50	62	23	36	220	36	220	0.1300	86
47	8	0.12	7.5	7.5	10.0	10.2	2,190	1,100	415	325	-	70	50	62	19	35	110	35	110	0.1500	63
47	9	0.14	7.5	7.5	4.9	12.6	2,300	1,100	500	342	-	-	56	67	19	40	64	980	64	0.0870	110
47	10	0.10	3.8	1.9	2.5	7.1	2,280	1,100	659	255	-	-	56	67	2.1	3.9	6.6	37	6.6	0.1500	-55
47	11	0.10	8.8	4.4	5.9	7.7	1,620	1,050	903	245	-	-	56	67	1.3	1.7	2.7	720	2.7	0.1500	-155
47	12	0.18	-	3.5	5.3	17.9	0	920	935	187	58	59	62	65	0	0.69	1.1	460	1.1	0.0770	-195
48	4	0.10	-	7.6	7.6	7.6	0	100	200	190	-	70	50	62	0	13	9.8	200	9.8	0.0370	166
48	5	0.10	6.3	6.2	7.7	7.7	245	900	400	245	-	70	50.	62	9.1	8.3	20	42	20	0.1200	147
48	6	0.11	6.3	6.2	7.5	8.6	1,260	900	400	220	-	70	50	62	27	41	100	42	100	0.1000	149
48	7	0.11	5.0	5.0	6.2	8.9	2,400	950	300	265	-	70	50	62	16	36	94	42	94	0.0870	110
48	8	0.11	5.0	5.0	6.3	9.6	3,010	800	389	364	-	70	50	62	19	52	93	130	93	0.0660	130
48	9	0.16	8.8	4.3	4.9	15.3	3,740	800	307	433	-	-	56	67	0.33	1.3	1.6	54	1.6	0.0400	80
48	10	0.10	7.5	3.7	4.4	7.7	3,330	600	630	337	58	59	62	65	1.2	3.8	4.7	42	4.7	0.0970	-85
48	11	0.09	8.8	4.4	5.6	6.0	2,180	700	957	245	58	59	62	65	1.9	3.1	4.2	990	4.2	0.1700	-220
48	12	0.09	2.9	2.6	1.3	5.6	358	700	730	80	58	59	62	65	2	1.4	2.4	600	2.4	0.1100	-250
49	4	0.11	-	9.6	9.6	9.6	0	700	300	200	-	62	61	50	0	46	200	46	200	0.1100	223
49	5	0.13	7.5	7.5	11.0	10.9	489	1,100	125	200	-	70	50	62	34	41	190	41	190	0.0930	182
49	6	0.14	7.5	7.5	10.0	13.0	1,680	1,100	145	200	-	62	61	50	19	37	220	37	220	0.1100	150
49	7	0.11	6.3	6.2	7.5	9.1	3,060	1,180	225	300	-	62	61	50	4	11	32	33	32	0.1100	110
49	8	0.11	5.0	5.0	5.0	8.8	3,460	1,240	125	442	-	62	61	50	8.3	25	70	34	70	0.1100	125
49	9	0.10	3.8	1.9	2.1	7.1	3,550	700	920	187	-	62	61	50	0.82	1.8	2.8	36	2.8	0.1100	60
49	10	0.10	6.3	3.1	3.2	7.6	2,840	700	997	180	58	59	62	65	0.92	1.8	2.6	33	2.6	0.0770	-35
49	11	0.08	7.5	3.6	5.2	5.6	2,570	800	915	255	58	59	62	65	1.8	3.4	4.8	540	4.8	0.1800	-220

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment between layers 1-2	Predevelopment 2-3	Predevelopment 3-4	Postdevelopment between layers 2-3	Postdevelopment 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
50	3	0.06	–	–	3.0	3.0	0	0	100	100	–	62	61	50	0	0	90	0	90	0.0000	309
50	4	0.10	6.3	6.2	7.1	7.1	74	750	126	200	–	62	61	50	10	9.2	28	52	28	0.0930	213
50	5	0.14	10.0	10.0	12.1	12.1	667	1,200	233	275	–	62	61	50	23	28	89	32	89	0.1200	175
50	6	0.11	6.3	6.2	7.5	8.7	2,200	1,200	250	328	–	62	61	50	14	29	83	32	83	0.1400	150
50	7	0.09	5.0	5.0	6.2	7.5	3,320	900	585	415	–	62	61	50	9.6	23	38	31	38	0.1400	33
50	8	0.06	5.0	5.0	6.3	7.1	3,870	900	472	515	–	62	61	50	29	86	310	86	310	0.1200	20
50	9	0.09	5.0	1.8	2.6	6.0	3,130	800	898	237	–	62	61	50	1.1	2.1	3.3	33	3.3	0.0900	30
50	10	0.08	7.5	3.4	3.7	5.2	3,790	700	1,050	145	58	59	62	65	1.1	2.6	3.7	800	3.7	0.1500	-135
50	11	0.10	5.0	2.2	2.6	7.2	1,750	1,500	860	440	58	59	62	65	0.88	1.2	2	450	2	0.1200	-195
51	2	0.06	–	–	1.4	1.4	0	0	75	75	–	62	61	50	0	0	15	0	15	0.0000	370
51	3	0.09	–	6.3	6.3	6.3	0	200	100	78	–	62	61	50	0	97	180	150	180	0.0340	273
51	4	0.12	–	7.5	8.8	9.9	0	1,500	135	200	–	62	61	50	0	6.8	38	28	38	0.1600	217
51	5	0.11	7.5	7.5	8.9	8.9	1,370	800	750	275	–	62	61	50	24	30	61	30	61	0.1800	183
51	6	0.13	7.5	7.5	8.7	11.0	2,810	1,500	200	300	–	62	61	50	2	4.4	17	27	17	0.1500	120
51	7	0.12	6.3	6.2	8.7	10.0	2,810	1,500	160	378	–	62	61	50	1.9	4.4	16	28	16	0.1100	45
51	8	0.13	7.5	7.5	8.8	10.7	2,420	1,600	350	492	–	62	61	50	7.3	13	35	130	35	0.0890	98
51	9	0.15	8.8	4.3	5.0	13.5	3,140	1,400	895	205	–	62	61	50	6.7	12	25	86	25	0.0920	45
51	10	0.10	7.5	3.4	5.3	6.6	3,840	1,120	1,310	220	58	59	62	65	1.1	2.2	3.4	690	3.4	0.1100	-135
51	11	0.13	8.8	3.4	4.6	10.4	2,230	1,410	1,180	340	58	59	62	65	1.1	1.5	2.4	580	2.4	0.1500	-201
52	3	0.18	–	12.5	17.7	17.7	0	580	120	140	–	62	61	50	0	60	180	65	180	0.0700	242
52	4	0.14	11.3	11.2	12.5	12.7	398	1,500	100	170	–	62	61	50	7.1	8.1	55	29	55	0.1500	213
52	5	0.11	7.5	7.5	8.7	9.1	1,610	1,500	200	285	–	62	61	50	7.2	12	47	27	47	0.1300	186
52	6	0.13	7.5	7.5	10.0	11.3	3,270	1,500	200	335	–	62	61	50	11	27	110	27	110	0.2200	162
52	7	0.13	7.5	7.5	10.0	12.1	2,670	1,500	165	392	–	62	61	50	3.3	7.4	26	27	26	0.1700	56
52	8	0.11	7.5	7.5	8.8	9.0	3,500	1,200	335	505	–	62	61	50	2.9	7.6	16	130	16	0.1300	58
52	9	0.12	7.5	7.5	4.1	9.9	3,700	800	1,560	175	58	59	62	65	5.7	10	14	130	14	0.2300	-40
52	10	0.09	6.3	2.8	3.7	5.7	4,360	1,000	1,580	275	58	59	62	65	1.5	3	4.1	21	4.1	0.2400	-148
52	11	0.08	7.5	3.4	3.6	4.5	2,370	1,310	1,210	320	58	59	62	65	1.8	2.6	4.1	820	4.1	0.1900	-215
53	2	0.07	–	–	4.4	4.4	0	0	150	90	–	62	61	50	0	0	150	0	150	0.0000	300
53	3	0.13	–	10.0	11.3	11.6	0	800	100	250	–	62	61	50	0	4	12	27	12	0.0810	241
53	4	0.14	10.0	10.0	11.3	11.6	811	1,600	100	276	–	62	61	50	3	4	21	27	21	0.1400	209
53	5	0.16	8.8	8.7	12.5	15.8	2,110	850	850	270	–	62	61	50	11	16	26	27	26	0.1400	135
53	6	0.11	7.5	7.5	9.6	9.6	3,840	800	860	340	–	62	61	50	12	28	54	28	54	0.1300	130
53	7	0.12	8.8	8.7	11.2	10.9	2,880	1,400	200	398	–	62	61	50	12	29	130	29	130	0.1200	160
53	8	0.12	7.5	7.5	12.5	10.4	4,000	1,300	260	488	–	62	61	50	1.5	4.4	11	86	11	0.0810	74
53	9	0.13	8.8	4.4	6.1	10.5	4,870	700	1,060	240	58	59	62	65	1	3.1	4.1	33	4.1	0.1700	-25
53	10	0.11	7.5	3.4	3.9	8.7	5,200	800	1,260	210	58	59	62	65	1.4	3.9	5.3	980	5.3	0.1900	-195
53	11	0.12	7.5	2.2	3.9	8.8	3,120	900	1,100	345	58	59	62	65	3.5	6.7	8.9	1100	8.9	0.1400	-155
53	13	0.13	–	–	10.8	10.8	0	0	100	100	69	48	23	47	0	0	15	0	15	0.0000	320

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	2	3	4	Aquifer thickness (ft) Layer 1	2	3	4	Percentage of fine-grained sediment Layer 1	2	3	4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	2-3	3-4	Postdevelopment Between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
54	2	0.06	–	3.6	3.6	3.6	0	100	100	100	–	62	61	50	0	81	90	150	90	0.0170	226
54	3	0.08	–	5.0	6.3	6.3	0	700	220	330	–	62	61	50	0	8.5	16	27	16	0.0760	240
54	4	0.14	11.3	11.2	12.5	13.2	676	1,350	360	290	–	62	61	50	9.3	10	30	27	30	0.1800	203
54	5	0.13	7.5	7.5	10.0	11.4	1,800	1,520	150	325	–	62	61	50	15	27	140	27	140	0.0740	188
54	6	0.14	6.3	6.3	7.4	12.8	2,480	1,350	400	350	–	62	61	50	12	22	58	26	58	0.1300	146
54	7	0.11	5.0	5.0	6.7	8.4	3,200	1,100	600	400	–	62	61	50	6.4	14	26	27	26	0.1500	126
54	8	0.12	7.5	7.5	8.7	10.1	3,010	800	800	515	–	62	61	50	4.1	8.4	11	29	11	0.0450	70
54	9	0.08	6.3	6.3	8.3	3.8	1,200	700	912	480	58	59	62	65	2.2	2.5	2.7	760	2.7	0.1000	-60
54	10	0.09	6.3	2.8	3.3	6.1	965	700	1,060	295	58	59	62	65	3.6	3.2	4.1	1100	4.1	0.1300	-125
54	11	0.12	9.2	4.1	4.1	9.2	945	700	995	280	69	48	23	47	2.9	5	7.8	150	7.8	0.1200	-115
54	12	0.12	9.8	4.4	3.8	9.8	252	500	958	270	69	48	23	47	11	9.6	13	190	13	0.0870	45
54	13	0.10	–	–	6.9	6.7	0	0	50	50	69	48	23	47	0	0	25	0	25	0.0000	320
55	2	0.06	–	–	2.9	2.9	0	0	160	100	–	62	61	50	0	0	900	0	900	0.0000	250
55	3	0.08	–	4.6	4.7	4.7	0	922	766	305	–	62	61	50	0	26	44	27	44	0.1800	225
55	4	0.12	6.3	6.3	8.7	10.4	2,200	1,250	470	275	–	62	61	50	10	18	46	27	46	0.1700	184
55	5	0.10	6.3	6.3	6.5	7.0	2,180	1,250	460	290	–	62	61	50	2.6	4.6	11	170	11	0.1600	100
55	6	0.11	3.8	3.7	4.1	8.8	2,730	1,200	500	350	74	50	42	57	2	6.6	0.17	170	13	0.1300	130
55	7	0.12	6.3	6.2	7.5	10.0	2,890	800	775	475	74	50	42	57	4	14	0.036	57	17	0.1000	140
55	8	0.09	5.0	5.0	5.5	5.8	1,560	700	768	532	74	50	42	57	1.9	4.2	29	61	4.5	0.0940	45
55	9	0.10	5.0	5.0	5.8	8.2	522	800	1,120	250	58	59	62	65	3.2	2.1	2.9	29	2.9	0.1500	-55
55	10	0.09	6.3	2.8	3.1	4.6	910	900	1,330	225	58	59	62	65	1.8	1.4	2	1400	2	0.1700	-155
55	11	0.11	–	3.5	3.5	7.7	0	1,370	1,240	290	58	59	62	65	0	0.75	1.2	920	1.2	0.2500	-95
55	12	0.11	–	3.8	3.2	8.4	0	700	1,130	110	69	48	23	47	0	14	27	1700	27	0.1700	147
56	2	0.06	–	–	3.1	3.1	0	0	100	150	–	60	66	63	0	0	900	0	900	0.0000	165
56	3	0.08	5.4	5.4	5.4	5.4	18	400	600	200	–	62	61	50	110	46	460	46	210	0.0970	205
56	4	0.12	5.0	5.0	5.6	9.9	775	800	800	200	–	60	66	63	30	26	180	26	180	0.1700	150
56	5	0.10	7.5	7.5	10.4	6.6	1,380	1,150	450	300	–	60	66	63	20	29	58	52	58	0.1300	104
56	6	0.10	5.0	5.0	6.3	7.7	1,790	1,200	340	360	–	60	66	63	27	46	3.9	130	97	0.1100	140
56	7	0.13	6.3	6.3	6.4	11.7	1,740	700	700	510	74	50	42	57	9.4	24	5.3	130	26	0.0880	60
56	8	0.12	10.0	10.0	16.5	9.6	919	850	696	555	74	50	42	57	3.1	4.8	47	57	5.7	0.0940	-20
56	9	0.04	8.8	7.9	11.0	1.6	735	850	900	650	58	59	62	65	1.8	1.6	1.7	26	1.7	0.1100	-45
56	10	0.09	6.3	2.8	1.3	4.6	885	700	1,270	360	58	59	62	65	0.71	0.55	0.65	560	0.65	0.0860	-30
56	12	0.08	3.8	3.4	4.5	5.8	4,520	1,000	500	100	69	48	23	47	0.33	2	7.3	130	7.3	0.1400	153

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	2	3	4	Aquifer thickness (ft) Layer 1	2	3	4	Percentage of fine-grained sediment Layer 1	2	3	4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	2-3	3-4	Postdevelopment Between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
57	2	0.13	–	9.8	9.8	9.8	0	100	100	100	–	60	66	63	0	140	900	140	900	0.0170	268
57	3	0.11	7.5	7.5	8.9	8.9	40	1,000	200	100	–	60	66	63	57	48	180	86	180	0.1100	240
57	4	0.11	6.3	6.2	8.7	8.4	1,620	1,000	481	284	–	60	66	63	1.2	1.9	3.5	29	3.5	0.1100	162
57	5	0.13	7.5	7.5	8.7	11.2	2,270	900	763	337	–	60	66	63	54	86	0.75	86	260	0.1800	150
57	6	0.13	8.5	8.5	8.6	11.4	1,950	700	900	400	–	60	66	63	31	42	0.052	130	51	0.1100	110
57	7	0.15	10.0	10.0	13.4	13.7	1,080	700	758	542	74	50	42	57	36	62	180	62	180	0.1100	85
57	8	0.06	8.8	7.9	8.1	3.7	882	700	740	560	74	50	42	57	0.13	0.19	0.2	62	0.2	0.0790	-70
57	9	0.02	–	18.4	20.4	20.7	0	600	700	700	74	50	42	57	0	0.17	0.15	69	0.15	0.0880	-117
57	10	0.05	8.8	4.0	4.9	2.8	1,120	600	650	650	74	50	42	57	0.0022	0.0043	0.0036	72	0.0036	0.1100	-70
57	12	0.06	1.3	1.1	2.2	2.4	3,100	500	300	440	69	48	23	47	0.14	1.1	1.2	240	1.2	0.0800	155
58	3	0.08	5.0	5.0	5.6	5.6	30	700	400	200	–	60	66	63	69	44	77	520	77	0.0850	150
58	4	0.12	6.3	6.2	8.8	9.5	1,300	1,400	350	250	–	60	66	63	45	62	170	86	170	0.1700	180
58	5	0.13	6.3	6.2	7.5	11.7	2,990	1,400	525	295	–	60	66	63	11	22	0.55	22	60	0.1800	160
58	6	0.14	10.0	10.0	12.6	12.7	2,460	900	800	340	–	60	66	63	12	20	3.4	25	29	0.1300	120
58	7	0.17	10.0	10.0	14.7	16.8	810	900	793	557	74	50	42	57	2.9	3.8	4.6	53	4.6	0.0930	25
58	8	0.06	7.5	6.8	8.4	2.8	490	900	800	650	74	50	42	57	1.4	1.4	1.6	52	1.6	0.1000	0
58	9	0.03	7.5	6.7	7.9	2.5	425	800	800	800	74	50	42	57	0.4	0.39	0.36	56	0.36	0.1600	0
58	10	0.04	6.3	5.6	6.8	2.5	1,020	700	700	740	69	48	23	47	2.2	4.5	4.4	150	4.4	0.1100	70
58	12	0.08	–	–	2.3	2.5	0	0	50	50	69	48	23	47	0	0	3	0	3	0.0000	280
58	13	0.13	–	–	10.0	11.1	0	0	50	100	69	48	23	47	0	0	3.3	0	3.3	0.0000	320
59	3	0.07	–	3.7	4.2	4.2	0	500	600	225	–	60	66	63	0	20	26	430	26	0.1400	110
59	4	0.11	6.3	6.3	8.6	8.6	1,780	600	600	225	–	60	66	63	21	35	140	35	140	0.1100	227
59	5	0.14	7.5	7.5	10.0	12.3	3,410	800	593	217	52	42	50	62	16	55	0.29	55	100	0.1300	155
59	6	0.09	7.5	7.5	8.0	6.1	2,960	700	636	350	52	42	50	62	19	57	3.2	57	160	0.1100	109
59	7	0.11	7.5	7.5	14.6	8.7	1,190	1,000	400	460	52	42	50	62	3.3	5.6	7.1	130	7.1	0.1000	10
59	8	0.07	8.8	7.9	12.1	3.4	713	700	620	740	74	50	42	57	0.87	1.2	1.1	67	1.1	0.1500	30
59	9	0.05	5.0	4.5	9.0	2.5	730	600	600	750	74	50	42	57	0.34	0.52	0.42	75	0.42	0.0950	45
59	10	0.05	2.5	2.2	4.5	3.7	1,400	600	596	784	74	50	42	57	0.51	1.2	0.97	75	0.97	0.0520	75
59	11	0.08	–	5.6	5.6	6.3	0	500	495	697	69	48	23	47	0	0.37	0.29	210	0.29	0.0680	105
59	12	0.11	–	1.0	0.9	1.0	0	600	600	550	69	48	23	47	0	9.2	9.9	180	9.9	0.0880	221
59	13	0.10	–	–	4.6	5.0	0	0	25	50	–	–	–	48	0	0	3.6	0	3.6	0.0000	295

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	Layer 2	Layer 3	Layer 4	Aquifer thickness (ft) Layer 1	Layer 2	Layer 3	Layer 4	Percentage of fine-grained sediment Layer 1	Layer 2	Layer 3	Layer 4	Leakance (TK) ($\times 10^{-6}$ d^{-1}) Predevelopment between layers 1-2	Predevelopment 2-3	Predevelopment 3-4	Postdevelopment between layers 2-3	Postdevelopment 3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
60	3	0.06	-	2.9	2.9	2.9	0	100	650	200	52	42	50	62	0	44	36	94	36	0.0570	170
60	4	0.08	3.8	3.8	5.2	5.2	1,130	800	800	200	52	42	50	62	37	47	150	47	150	0.1500	230
60	5	0.10	6.3	6.2	8.0	7.1	3,040	900	719	217	52	42	50	62	0.8	2.1	0.006	47	3.2	0.1500	111
60	6	0.09	8.8	8.7	11.0	5.3	2,700	800	600	365	52	42	50	62	6.9	19	1.1	170	23	0.0790	80
60	7	0.09	5.0	5.0	5.4	7.2	650	600	600	600	52	42	50	62	3.7	4	3.3	130	3.3	0.0770	69
60	8	0.06	2.5	2.3	3.1	2.7	855	500	900	440	74	50	42	57	0.065	0.093	0.092	65	0.092	0.1700	45
60	9	0.07	-	2.3	2.3	3.7	0	450	753	647	74	50	42	57	0	0.78	0.62	76	0.62	0.0370	-114
60	10	0.08	0.5	0.4	0.6	3.8	830	1,000	446	804	74	50	42	57	0.035	0.056	0.06	60	0.06	0.0680	10
60	11	0.09	-	22.7	10.2	5.5	0	100	190	670	-	-	-	48	0	0.0022	300E-6	240	300E-6	0.0230	37
60	12	0.09	-	1.1	0.4	1.6	0	600	500	500	-	-	-	48	0	3.5	4	63	4	0.1100	235
60	13	0.08	-	5.4	4.8	5.4	0	25	50	100	-	-	-	48	0	2.6	1.3	350	1.3	0.0061	308
61	3	0.07	-	3.4	3.4	3.4	0	200	200	200	52	42	50	62	0	10	8.5	220	8.5	0.0330	170
61	4	0.07	4.0	4.0	4.0	4.0	217	1,000	800	200	52	42	50	62	59	38	60	42	60	0.1500	234
61	5	0.09	5.0	5.0	6.7	6.7	2,140	1,000	1,000	250	52	42	50	62	22	36	0.14	69	51	0.1800	150
61	6	0.07	2.5	2.5	3.7	3.2	2,240	1,600	500	290	52	42	50	62	17	35	1.7	86	74	0.1400	125
61	7	0.06	2.5	2.5	2.5	1.9	82	1,200	800	398	52	42	50	62	4.7	2.9	4	130	4	0.1700	69
61	8	0.09	7.5	6.0	12.2	6.1	225	1,200	700	330	52	42	50	62	1.6	1.2	1.8	170	1.8	0.0540	60
61	9	0.09	7.5	6.3	11.2	6.9	682	1,000	655	345	-	48	53	55	1.9	1.9	2.9	39	2.9	0.0510	66
61	10	0.12	1.3	1.1	0.4	9.7	1,240	700	500	398	-	-	-	48	2.9	4.8	6.5	58	6.5	0.0810	-20
61	11	0.08	-	4.5	5.2	5.7	0	300	195	475	-	-	-	48	0	0.067	0.052	140	0.052	0.0380	60
61	12	0.06	-	0.3	1.1	1.9	0	590	400	450	-	-	-	48	0	5.4	6.4	86	6.4	0.0960	228
61	13	0.06	-	2.5	2.3	2.5	0	50	50	50	-	-	-	48	0	5.2	5.4	690	5.4	0.0081	394
62	4	0.06	-	2.4	2.4	2.4	0	900	600	350	52	42	50	62	0	1.1	1.5	51	1.5	0.1000	200
62	5	0.08	5.0	5.0	5.4	5.4	1,760	1,000	700	300	52	42	50	62	14	24	34	45	34	0.1600	143
62	6	0.08	3.8	3.7	3.8	3.8	1,940	900	800	300	52	42	50	62	25	44	97	44	97	0.0820	100
62	7	0.10	7.5	7.5	8.9	8.1	588	1,000	700	320	52	42	50	62	21	20	28	45	28	0.1400	93
62	8	0.07	5.0	5.0	5.6	4.7	540	1,000	750	275	-	48	53	55	43	37	69	37	69	0.1300	65
62	9	0.10	7.5	6.7	8.9	6.9	994	1,200	495	298	-	48	53	55	3.8	4.8	0.49	960	9.5	0.0620	33
62	10	0.09	6.3	5.6	5.3	6.3	762	800	1,000	197	-	48	53	55	7.3	6.2	0.61	36	8.8	0.1100	-40
62	11	0.04	-	2.3	2.3	1.4	0	1,000	775	325	-	-	-	48	0	0.011	0.019	39	0.019	0.1600	80
62	12	0.06	-	2.1	2.1	2.3	0	100	350	350	-	-	-	48	0	3.5	2.3	150	2.3	0.0420	230
62	13	0.05	-	2.5	2.2	2.5	0	25	100	100	-	-	-	48	0	2.8	1.8	550	1.8	0.0100	419
63	4	0.08	-	4.2	4.2	4.2	0	500	500	355	52	42	50	62	0	2	1.9	75	1.9	0.1000	100
63	5	0.08	-	5.0	4.8	4.8	0	1,800	1,000	350	52	42	50	62	0	6.1	11	27	11	0.2600	139
63	6	0.09	5.0	5.0	6.4	6.4	910	1,400	1,000	300	-	48	53	55	29	27	120	27	120	0.2600	120
63	7	0.08	3.8	3.7	3.5	4.2	1,220	1,200	1,060	235	-	48	53	55	14	15	25	29	25	0.2100	130
63	8	0.10	5.0	5.0	5.4	6.1	367	1,000	725	275	-	48	53	55	48	37	59	38	59	0.1200	70
63	9	0.14	8.8	7.9	10.2	11.7	710	800	805	355	-	48	53	55	29	26	2.7	1300	34	0.1400	-60
63	10	0.10	6.3	5.6	6.5	7.1	925	700	800	348	-	48	53	55	2.5	2.6	5.3	43	3.2	0.0830	-15
63	11	0.09	-	1.6	1.6	4.2	0	1,000	670	395	-	48	53	55	0	39	60	39	60	0.2200	174
63	12	0.08	-	2.3	2.2	2.5	0	125	200	325	-	-	-	48	0	8.7	5.5	230	5.5	0.0300	170

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d) Layer 1	2	3	4	Aquifer thickness (ft) Layer 1	2	3	4	Percentage of fine-grained sediment Layer 1	2	3	4	Leakance (TK) (x 10^{-6} d^{-1}) Predevelopment Between layers 1-2	2-3	3-4	Postdevelopment Between layers 2-3	3-4	Inelastic storage coefficient Layers 2-3	1961 critical head (ft) Layers 2-3
64	4	0.09	–	5.4	5.4	5.4	0	100	100	200	–	48	53	55	0	14	8.6	320	8.6	0.0093	280
64	5	0.07	–	3.1	3.1	3.1	0	2,000	1,120	375	–	48	53	55	0	3.2	6.3	21	6.3	0.2900	135
64	6	0.06	2.5	2.5	2.4	2.5	2,040	1,000	800	350	–	48	53	55	29	47	69	86	69	0.1900	115
64	7	0.11	6.3	6.2	7.5	8.7	2,260	1,000	1,000	390	–	48	53	55	18	28	38	32	38	0.1900	129
64	8	0.10	6.3	6.2	7.5	7.5	685	1,000	800	285	–	48	53	55	24	22	34	36	34	0.1300	14
64	9	0.12	9.5	9.5	9.5	9.5	290	1,100	660	340	–	48	53	55	23	18	900E-6	1400	29	0.1100	-15
64	10	0.11	6.4	6.4	6.4	8.2	445	1,000	1,000	320	–	48	53	55	40	28	40	32	40	0.1200	85
64	11	0.10	9.9	9.9	9.9	6.3	125	1,300	1,300	357	–	–	–	48	3.9	2.1	3.4	27	3.4	0.1600	115
64	12	0.08	–	5.0	5.0	5.0	0	500	500	350	–	–	–	48	0	3.7	4.4	69	4.4	0.0900	180
65	5	0.07	1.3	1.3	1.3	2.5	100	900	250	300	–	48	53	55	1.4	1.2	2.3	58	2.3	0.0860	185
65	6	0.10	5.0	5.0	6.4	6.4	1,980	800	450	350	–	48	53	55	0.55	1.2	1.8	52	1.8	0.1100	80
65	7	0.08	5.0	5.0	5.0	5.3	3,280	700	600	300	–	48	53	55	6.5	20	27	290	27	0.1000	80
65	8	0.09	6.1	6.1	6.1	6.1	2,740	600	700	300	–	48	53	55	46	120	140	250	140	0.0800	70
65	9	0.13	8.8	8.7	11.5	11.5	687	1,000	350	350	–	48	53	55	9	11	20	49	20	0.0760	43
65	10	0.11	7.5	7.5	8.4	8.0	428	1,000	975	347	–	–	–	48	46	33	50	230	50	0.1300	92
65	11	0.12	–	0.5	0.5	8.4	0	1,300	1,280	455	–	–	–	48	0	5.7	8.6	27	8.6	0.1800	180
65	12	0.09	–	1.2	1.2	6.3	0	400	500	400	–	–	–	48	0	6.7	6.8	77	6.8	0.0790	230
66	6	0.07	1.3	1.2	1.7	1.7	1,530	500	500	400	–	48	53	55	6.7	13	14	65	14	0.0670	255
66	7	0.10	5.0	5.0	6.3	6.9	3,040	700	600	420	–	48	53	55	1.8	5.2	6.3	50	6.3	0.0940	60
66	8	0.13	8.8	8.7	10.0	11.0	3,130	400	850	400	–	48	53	55	19	51	130	51	130	0.0750	60
66	9	0.14	7.5	7.5	10.0	11.3	1,620	1,100	295	415	–	54	46	44	19	37	84	51	84	0.1000	110
66	10	0.11	5.0	5.0	6.3	7.3	1,740	750	735	500	–	54	46	44	30	51	72	51	72	0.1200	103
66	11	0.11	–	2.5	2.5	5.0	0	50	100	200	–	–	–	48	0	2.2	1.1	460	1.1	0.0100	220
66	12	0.09	–	1.2	1.2	5.0	0	50	150	200	–	–	–	48	0	2.1	1.2	350	1.2	0.0160	266
67	6	0.13	6.3	6.3	10.0	11.3	2,340	600	500	440	–	48	53	55	9.1	24	26	59	26	0.0500	236
67	7	0.13	7.5	7.5	10.0	11.6	3,440	800	500	475	–	54	46	44	5.3	17	26	190	26	0.0690	230
67	8	0.15	7.5	7.5	11.3	13.3	3,270	600	900	370	–	54	46	44	13	34	44	390	44	0.0890	210
67	9	0.14	6.3	6.2	10.0	11.8	1,440	1,300	396	400	–	54	46	44	8.9	14	35	140	35	0.1300	200
67	10	0.11	–	9.1	9.1	9.1	0	1,000	400	400	–	–	–	48	0	42	74	49	74	0.1200	157
67	11	0.13	–	10.6	10.6	10.6	0	1,000	1,220	350	–	–	–	48	0	31	94	31	94	0.1400	247
67	12	0.09	–	2.5	2.5	5.0	0	50	100	100	–	–	–	48	0	23	18	460	18	0.0100	292
68	6	0.13	7.5	7.5	10.0	10.5	1,440	1,200	1,200	200	–	54	46	44	5.3	6	11	650	11	0.0930	125
68	7	0.11	7.5	7.5	11.3	8.8	4,070	1,000	997	196	–	54	46	44	15	38	120	38	120	0.1200	240
68	8	0.13	7.5	7.5	10.0	10.9	5,010	1,200	530	205	–	54	46	44	35	120	3000	350	330	0.0710	210
68	9	0.11	6.3	6.2	7.5	8.1	2,450	1,000	518	382	–	54	46	44	14	33	340	220	63	0.0560	223
68	10	0.10	5.0	5.0	5.0	8.2	1,360	500	845	655	–	54	46	44	130	190	130	190	180	0.0450	150
68	11	0.13	7.5	7.5	10.9	10.9	300	400	660	400	–	–	–	48	1.2	0.8	0.81	65	0.81	0.0450	190

APPENDIX B: *Aquifer properties used in simulations*—Continued

[Layer 1 is deepest zone of aquifer, layer 4 is water-table zone]

Column	Row	Specific yield	Hydraulic conductivity (ft/d)				Aquifer thickness (ft)				Percentage of fine-grained sediment				Leakance (TK) (x 10^{-6} d^{-1})					Inelastic storage coefficient	1961 critical head (ft)
			Layer				Layer				Layer				Predevelopment			Postdevelopment			
															Between layers			Between layers		Layers	Layers
			1	2	3	4	1	2	3	4	1	2	3	4	1-2	2-3	3-4	2-3	3-4	2-3	2-3
69	5	0.10	–	6.3	6.3	6.4	0	800	800	200	–	56	49	47	0	76	130	170	130	0.1200	475
69	6	0.15	6.3	6.2	7.3	13.5	1,140	1,200	1,200	185	–	56	49	47	6.5	6.5	12	1100	12	0.0640	120
69	7	0.10	6.3	6.2	8.2	7.0	4,920	1,000	1,020	296	–	54	46	44	16	48	81	340	81	0.1500	180
69	8	0.12	7.5	7.5	8.8	10.0	4,930	1,200	500	332	–	54	46	44	25	90	50	320	210	0.0910	207
69	9	0.12	6.3	6.2	7.6	9.0	3,080	900	900	513	–	56	49	47	5.1	11	15	390	15	0.0600	150
69	10	0.11	6.3	6.2	2.1	7.5	1,100	800	1,200	530	–	56	49	47	4.4	4.3	5.3	170	5.3	0.0470	155
69	11	0.11	7.5	7.5	8.4	8.4	990	1,000	700	460	–	56	49	47	25	30	48	120	48	0.0620	210
70	6	0.17	7.5	7.5	10.0	15.3	3,580	1,000	1,120	148	–	56	49	47	27	57	100	1700	100	0.0440	80
70	7	0.12	5.0	5.0	5.0	10.2	7,740	1,400	610	410	–	54	46	44	9.4	42	94	4300	94	0.0340	40
70	8	0.12	6.3	6.3	7.5	8.0	7,200	1,440	941	419	–	54	46	44	8.8	32	63	2600	63	0.0450	86
70	9	0.12	7.5	6.5	10.2	10.1	4,660	1,000	1,000	478	–	54	46	44	0.76	2.2	3.2	520	3.2	0.0600	85
70	10	0.14	12.5	10.8	11.1	12.5	760	1,000	650	350	–	56	49	47	0.5	0.54	0.98	220	0.98	0.0520	40
70	11	0.09	–	6.0	6.2	6.2	0	50	100	100	–	56	49	47	0	2.4	1.9	460	1.9	0.0091	100

APPENDIX C: ESTIMATES OF BOREHOLE HYDRAULIC CONDUCTANCE OF MULTILAYER WELLS

Wells constructed with perforated intervals which span across adjacent aquifer layers, whether or not they are pumped, can have a major effect on the effective vertical hydraulic conductance and therefore flow between the layers. The well bores establish a direct hydraulic link which bypasses the vertical resistance to flow of the clay beds between the centers of the aquifer layers. Bennett and others (1982) suggest that this hydraulic effect can be evaluated approximately by adaptation of the Thiem equation.

Let C_w be borehole hydraulic conductance, which is the increase in vertical hydraulic conductance caused by a well open to aquifer layers above and below the clay beds. Then, by definition,

$$C_w = \frac{Q}{H_u - H_l} \qquad (25)$$

where Q is flow through the well casing, and H_u, H_l are head in aquifer layers above and below the clay beds, respectively, at some radial distance, R, from the well assumed to be the limit of the local cone in the potentiometric surface due to the influence of the well. R is further defined below.

For the purpose of this discussion we assume that the head in the aquifer layer above the clay beds is higher than the head in the aquifer layer below the clay beds, so that water will flow from the aquifer layer above and recharge the aquifer below through the well openings. The amount of the flow can be estimated by the Thiem equation, if the following two assumptions are valid: (1) well entrance losses and head losses within the well are negligible when compared with head losses in the aquifer, and (2) storage effects in the aquifers within the cone of influence in each aquifer also are negligible. According to the Thiem equation, flow leaving the aquifer layer above the clay beds can be described by the equation,

$$Q = \frac{2\pi b_u K_u (H_u - h_w)}{ln(R_a/R_w)} \qquad (26)$$

For flow recharging to the aquifer layer below the clay beds, the Thiem equation is

$$Q = \frac{2\pi b_l K_l (h_w - H_l)}{ln(R_a/R_w)} \qquad (27)$$

where

R_a = radial distance from center of the well to a concentric circle along which the head is assumed to be the average head in the aquifer block, H_u or H_l, respectively,

R_w = radius of the well,

K_u, K_l = hydraulic conductivity of the aquifers above and below the clay beds, respectively, and,

b_u, b_l = thickness of the upper and lower aquifer layers, respectively.

The right side of equations 26 and 27 can be equated, and the resulting equation solved for h_w, because the flow into the well out of the upper aquifer must equal the flow out of the well into the lower aquifer. Thus,

$$h_w = \frac{b_u K_u H_u + b_l K_l H_l}{b_u K_u + b_l K_l} \qquad (28)$$

Substituting equation 28 into either equation 26 or equation 27, the following expression is obtained:

$$Q = \frac{2\pi b_u K_u b_l K_l (H_u - H_l)}{ln(R_a/R_w)(b_u K_u + b_l K_l)} \qquad (29)$$

Substituting equation 29 into equation 25, C_w is given by

$$C_w = \frac{2\pi b_u K_u b_l K_l}{ln(R_a/R_w)(b_u K_u + b_l K_l)} \qquad (30)$$

If $K_u = K_l$, then equation 30 can be simplified and is given by (31).

$$C_w = \frac{2\pi K}{ln(R_a/R_w)} \frac{b_u b_l}{(b_u + b_l)} \qquad (31)$$

In the calculations for the Central Valley, R_a was assumed to be about 6,500 ft. The average irrigation well radius R_w in the Central Valley is about 0.75 ft. The thickness of aquifers above (layer 4) and below (layer 3) is about 250 ft and 1,000 ft, respectively. The hydraulic conductivity (K) of both aquifer layers is about 6 ft/d (the valley average), so the conductance per well (C_w) is estimated to be 830 ft/d.

The conductance of the clay beds (C_c) can be estimated by the Darcy equation:

$$C_c = \frac{Q}{H_u - H_l} = \frac{KA}{dL} \quad (32)$$

where

A = area of the model block, and
dL = length over which the vertical head difference is measured.

Using 4.1×10^{-6} per day (the model-calibrated average for the Westside area) and $A = 10$ ft^2, C_c is about 4,100 ft^2/d. According to these calculations, the conductance of about five wells in one model block would be equal to the conductance of the clay beds. There is a range of conductance that can be computed with reasonable variable values; however, this at least shows that wells probably have a significant contribution to conductance between layers.