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MATHEMATICAL GROUND-WATER MODEL OF INDIAN WELLS VALLEY, CALIFORNIA

By

R. M. Bloyd, Jr., and S. G. Robson

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MATHEMATICAL GROUND-WATER MODEL OF INDIAN WELLS VALLEY, CALIFORNIA

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ABSTRACT

A mathematical model of the Indian Wells Valley ground-water basin was developed and verified. The alternating-direction implicit method was used to compute the mathematical solution. It was assumed that there are only two aquifers in the valley, one being deep and the other shallow. Where the shallow aquifer occurs, the underlying deep aquifer is confined or artesian. Flow between the aquifers under steady-state conditions is assumed to be in one direction, from deep to shallow. The transmissivity of the deep aquifer ranges from about 250,000 to 22,000 gallons per day per foot and from about 25,000 to 5,000 gallons per day per foot for the shallow aquifer. The storage coefficient for the deep aquifer ranges from 1×10^{-4} to 0.20.

Steady-state recharge and discharge in each aquifer was estimated to be 9,850 acre-feet per year. Ground-water pumping, sewage-effluent recharge, and capture of ground-water discharge occurred under non-steady-state conditions. Most of the ground-water pumpage is near Ridgecrest and Inyokern and in the area between the two towns. By 1968 pumpage in the deep aquifer had caused a reversal in the ground-water gradient south of China Lake and small water-level declines over most of the aquifer. The model for the deep aquifer was verified under steady-state and non-steady-state conditions. The shallow aquifer was verified under steady-state conditions only.

The verified model was then used to generate 1983 water-level conditions in the deep aquifer.

INTRODUCTION

Purpose and Scope

The purpose of the investigation was to make a quantitative hydrologic study of the Indian Wells Valley area (fig. 1) and to make available to the cooperators a working hydrologic model for use as a management tool. The scope of the investigation consisted of:

1. Developing a digital-computer program to model the ground-water basin of Indian Wells Valley.

2. Organizing, analyzing, and evaluating hydrologic data and reports for Indian Wells Valley.

3. Estimating and verifying hydrologic parameters that are necessary inputs to the ground-water model.

4. Making an initial prediction of ground-water levels in Indian Wells Valley for 1983.

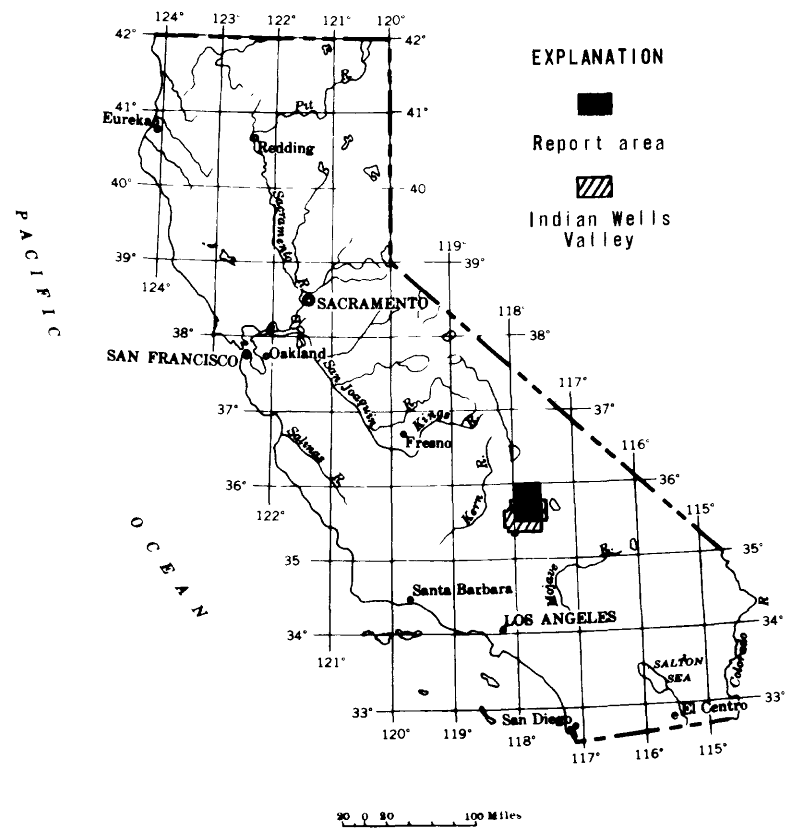


FIGURE 1.--Index map.

Acknowledgments

The investigation by the U.S. Geological Survey, in cooperation with the U.S. Department of the Navy and the Indian Wells Valley County Water District (formerly Ridgecrest County Water District), was made under the general supervision of R. Stanley Lord and N. C. Matalas, Water Resources Division. Immediate supervision was by L. C. Dutcher and J. L. Cook, chiefs of the Garden Grove subdistrict.

Location of the Area

Indian Wells Valley is in the Mojave Desert east of the Sierra Nevada in southern California (figs. 1 and 2), about 125 miles north of Los Angeles. The valley is bounded on the north by a low ridge of volcanic rocks and the Coso Range, on the east by the Argus Range, on the south by the El Paso Mountains, and on the west by the Sierra Nevada. Most of the central part of the valley is at an altitude between 2,150 and 2,400 feet above sea level. The largest and lowest playa in the valley, China Lake, is at an altitude of 2,152 feet.

The area considered in detail in this report is shown as the report area in figure 1. The flat playas and the alluvial slopes of Indian Wells Valley and most of the U.S. Naval Weapons Center, China Lake, are within the report area.

General Discussion of the Ground-Water Model

Models, or idealized representations, are integral parts of everyday life. Common examples are model airplanes, portraits, and globes. Such models can be used to abstract the essence of a subject of inquiry, showing interrelations and facilitating analysis. A mathematical model, such as the model of Indian Wells Valley, is an idealized representation of a ground-water basin and is designed to describe in mathematical language how the basin would function under various conditions.

One advantage that a mathematical model has over a verbal description of a problem is that the mathematical model describes a problem in concise quantitative terms. Such a description facilitates considering a problem in its entirety and considering all interrelations simultaneously. For example, the mathematical model of Indian Wells Valley facilitates a description of the mutual influence of the climatic, geologic, hydraulic, and manmade conditions that affect the ground-water basin.

The real world is seldom simple enough to be described exactly by any practical model. Therefore, simplifying assumptions or approximations are generally required if a model is to be feasible. Also, a model is only as accurate as the assumptions used in its construction. The assumptions used in constructing the Indian Wells Valley model are listed later in the report. These assumptions must be kept in mind when evaluating the model output.

A detailed theoretical development of the digital model was reported by Pinder and Bredehoeft (1968) and is not repeated in this report. The programing techniques used in the Indian Wells Valley model are described in a report by Thomas Maddock III (1970).

Nodal Network for the Ground-Water Model

To model Indian Wells Valley, a 60 by 40 nodal network (fig. 2) was used to specify the data points used on the model. The network intersections are spaced on one-half mile centers. The east-west lines are called rows, and the north-south lines are called columns. For ease of notation the intersection of row 2 column 5 is defined as node 2,5. The location of any data point for the model can be specified in terms of a row and a column number. For example, the approximate center of China Lake playa is at node 37,33 (fig. 2).

HYDROLOGY OF INDIAN WELLS VALLEY

Aquifer Characteristics and Ground-Water Flow

Geology, aquifer characteristics, and ground-water flow are described and documented in two previous hydrologic studies of Indian Wells Valley--Kunkel and Chase (1969) and Moyle (1963). These two reports described the geologic framework from which the hydrologic parameters used in this study were derived.

The boundary of the model area (fig. 2) approximates that of the ground-water basin (fig. 3) as described by Kunkel and Chase (1969). The aquifers in the model represent a system of multilayered three-dimensional heterogeneous alluvial deposits that fill the basin. The permeabilities of the deposits are nonhomogeneous and anisotropic. As a result perched or semiperched aquifers may occur above low permeability zones overlying the main aquifer. The model area is an intensely faulted structural depression. Many of the faults act as barriers to ground-water movement.

A deep aquifer (main water body of Kunkel and Chase (1969)) extends over most of the model area. The deep aquifer is a water-table aquifer in the southwestern part of the model area and is a confined aquifer in the northern and eastern part of the area. The deep aquifer is confined by clay zones in the eastern part of the area and by volcanic rocks in the northern part of the area.

A shallow aquifer overlies the clay zone that confines the deep aquifer (fig. 3). Kunkel and Chase (1969, fig. 6) showed water-level contours on the shallow aquifer and postulated its approximate areal extent.

A semiperched aquifer occurs northwest of the low permeability ground-water barrier or fault zone that trends southwest-northeast through sec. 3, T. 25 S., R. 38 E. (fig. 3). The water-level change across the boundary of the low permeability zone is about 5 to 10 feet.

Kunkel and Chase (1969, p. 41) also mentioned other minor water bodies that discharge across barriers or cascade over consolidated rock to the deep aquifer. These water bodies occur at the margins of the valley and are very thin marginal parts of the deep aquifer. The largest of the marginal aquifers lies beneath the extensive upland valley area southwest of the fault zone that trends northwest-southeast through sec. 1, T. 27 S., R. 38 E. (fig. 3). Because this marginal aquifer is of low transmissivity, extensive ground-water development of the aquifer is not feasible; therefore, the aquifer is not included in the ground-water model.

Under natural conditions ground water moves through the deep aquifer from the areas of recharge along the southwest, west, north, and northeast toward China Lake playa in the east-central part of the basin (fig. 4). The 1920-21 water levels are assumed to approximate natural conditions because little ground-water development had taken place in the basin before that time. Near the China Lake playa the deep aquifer discharges into the shallow aquifer. This discharge is the only significant source of natural discharge from the deep aquifer and is the only significant source of natural recharge to the shallow aquifer. Evaporation from the playa surfaces and transpiration from local phreatophyte growth are the only significant sources of natural discharge from the shallow aquifer.

Water-level contour maps (figs. 3, 4, and 5) constructed from water-level data for wells for 1920-21, 1953, and 1968 show the pattern of ground-water movement in the deep aquifer for those years. These maps are used for comparison of measured water levels with water levels generated by the model.

The results of this investigation suggest that before the establishment of the Naval Weapons Center, the shallow aquifer covered less area than at present. The main reason for the increase in size is the recharge from the Navy sewage ponds. Because of the paucity of water-level data, the extent of the shallow aquifer before the sewage ponds were constructed is not known. The boundary of the shallow aquifer shown in figure 3 was estimated from data collected after the construction of the sewage ponds.

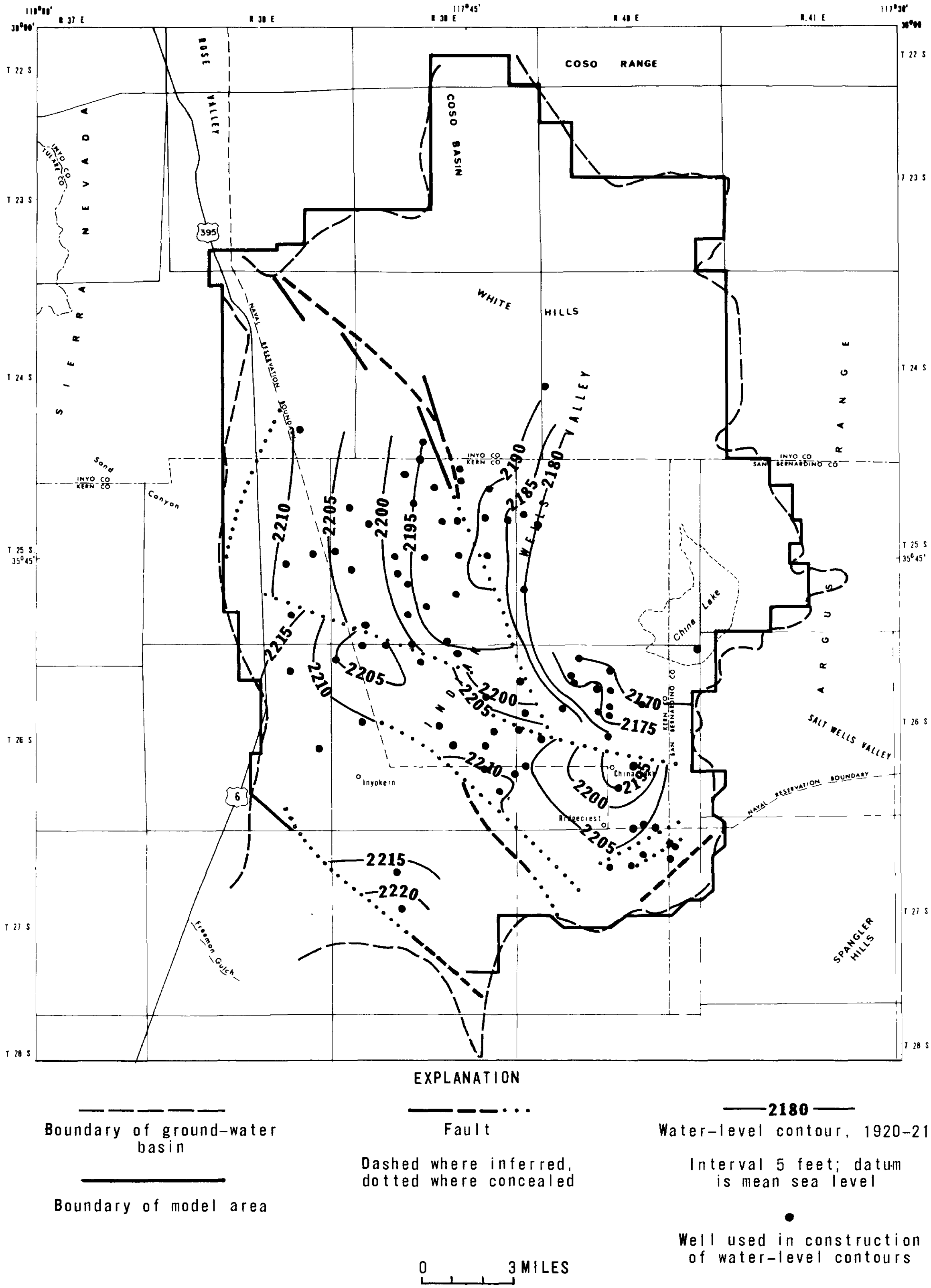


FIGURE 4.--Water-level contours for deep aquifer, 1920-21, constructed from water-level data.

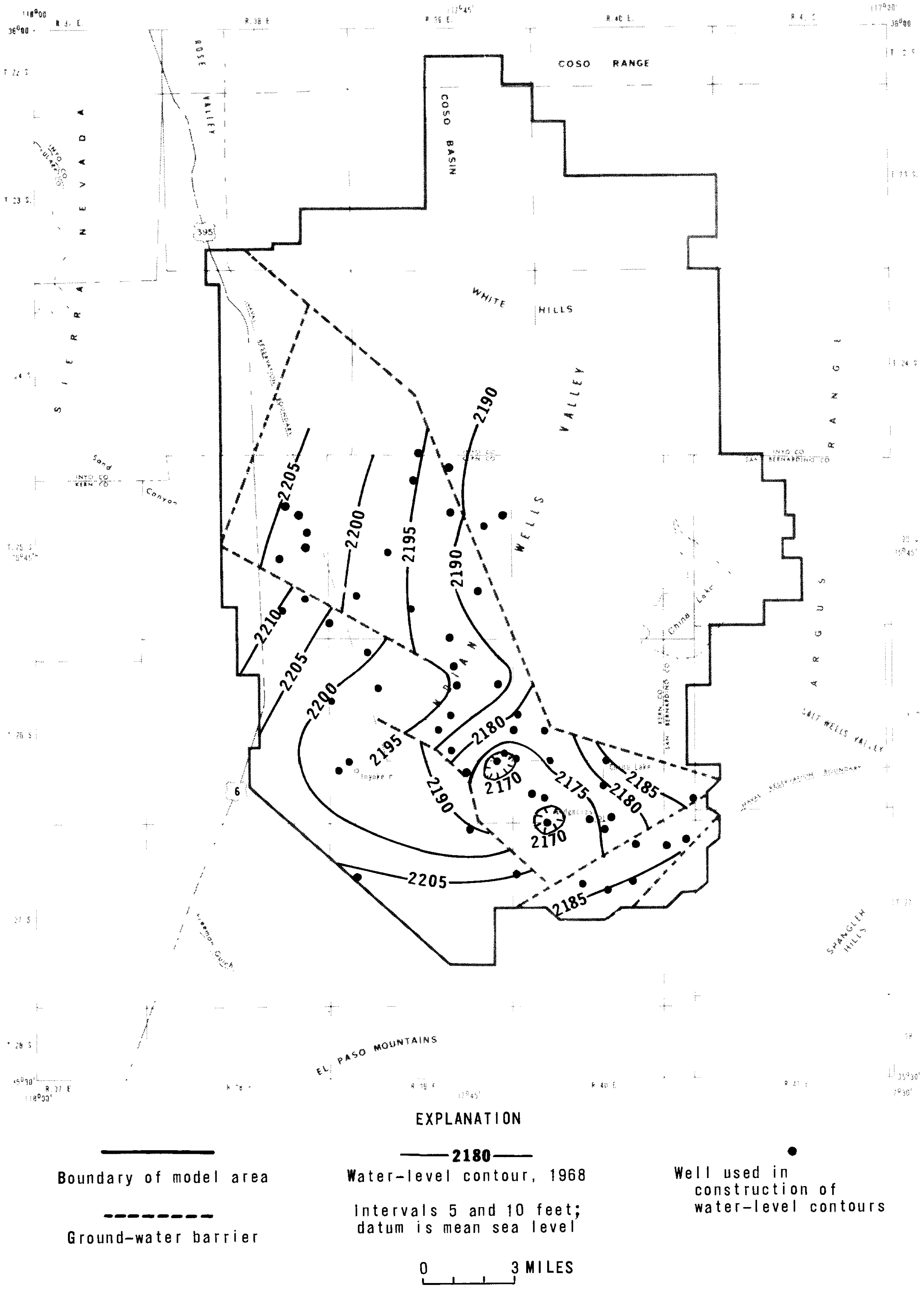


FIGURE 5.--Water-level contours for deep aquifer, 1968, constructed from water-level data.

Because much of the deep aquifer in the eastern part of the valley is confined, water must move from the confined aquifer into the overlying shallow aquifer before the water can be evaporated from the playas. Although the general movement of ground water is rather easily determined, the actual flow patterns and the quantities and the location of flow between aquifers are difficult to determine. One of the major problems of the model study was determining or defining the quantities and the location of flow between the deep aquifer and the shallow aquifer. This problem will be discussed in a following section of the report.

Aquifer Parameters

Two aquifer parameters (transmissivity and storage coefficient) describe the ability of an aquifer to transmit and to store and release water. Transmissivity (T) is the rate of flow in gallons per day, at prevailing water temperature, through a 1-foot-wide vertical strip of aquifer extending the full saturated height of the aquifer under a unit hydraulic gradient. The storage coefficient (S) is the volume of water an aquifer releases from or takes into storage per unit surface area (such as per square foot) of the aquifer per unit change in the component of head normal to that surface. These parameters are needed for each nodal point of the ground-water model.

Initial estimates of transmissivity and storage coefficient for the aquifers were made by L. C. Dutcher and W. R. Moyle, Jr. (written commun., 1970). Refinements of the estimates were made during verification of the model. The final values for these coefficients are hydrologically reasonable and are in general agreement with the estimates made by L. C. Dutcher and W. R. Moyle, Jr.

The transmissivity of the deep aquifer ranges from about 250,000 gpd/ft (gallons per day per foot) in the south-central part of the valley to less than 22,000 gpd/ft in the extreme southeastern part of the valley (fig. 6). The effect of the ground-water barriers was simulated in the model by use of a narrow zone of transmissivity ranging from 200 gpd/ft to 24,500 gpd/ft, coincident with the trace of the barriers.

Transmissivity values for the shallow aquifer have a smaller range than those for the deep aquifer and are generally of smaller magnitude (fig. 7). Because of the paucity of hydrologic data for the shallow aquifer, estimates of transmissivity were refined by trial and error with a series of computer runs. The final results are consistent with initial estimates and are considered to be hydrologically reasonable.

The storage coefficient for the deep aquifer ranges from 1×10^{-4} to 0.20 (fig. 8). Where the deep aquifer is overlain by the shallow aquifer or, as in the White Hills area, by volcanic rocks, the deep aquifer is assumed to be confined and to have a storage coefficient of 1×10^{-4}. Elsewhere the deep aquifer is assumed to be unconfined and to have a storage coefficient of 0.05 to 0.20.

Because of a paucity of data, a storage coefficient of 0.05 was assumed for the entire shallow aquifer.

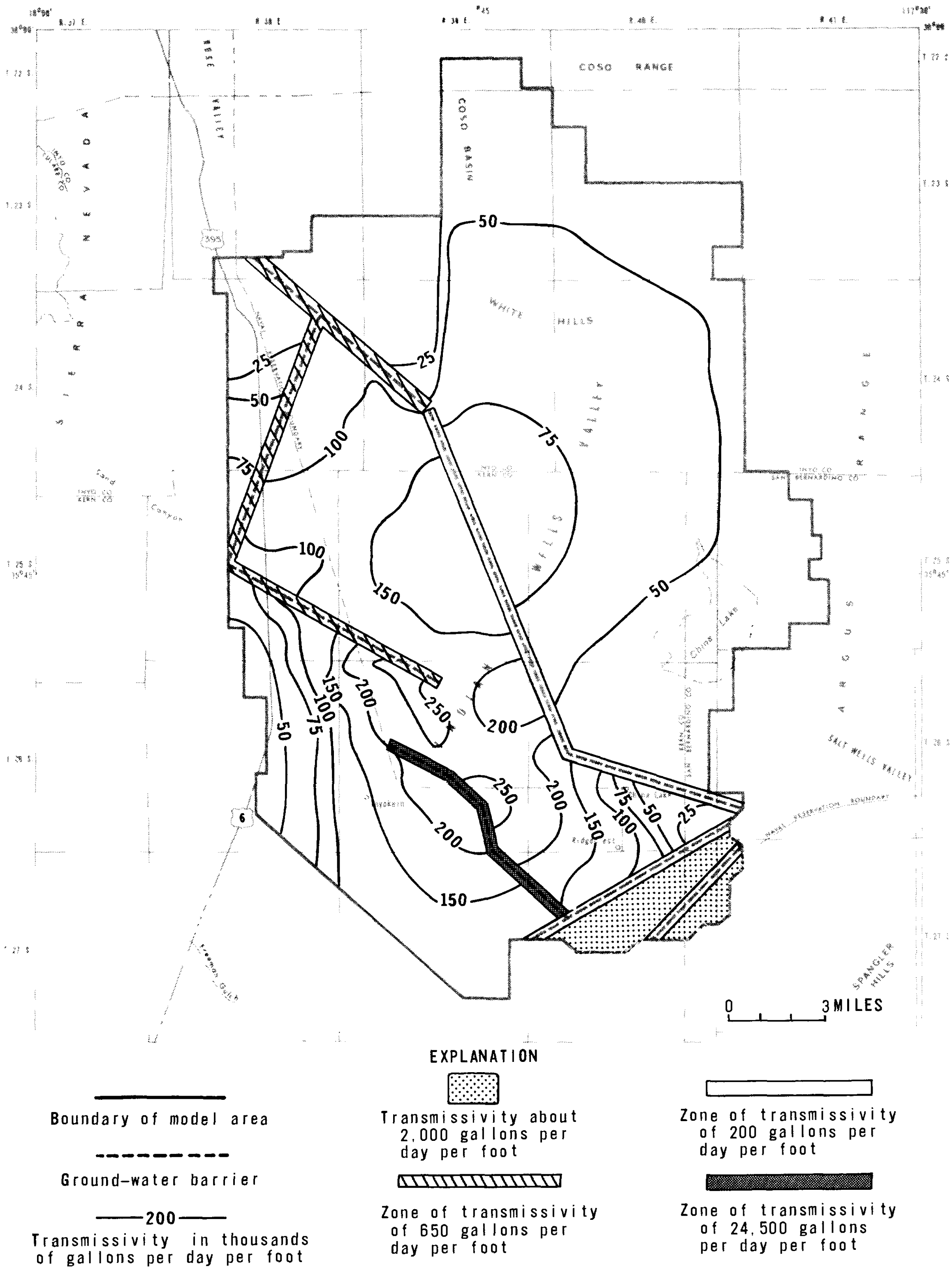


FIGURE 6.--Transmissivity of deep aquifer.

FIGURE 7.--Transmissivity of shallow aquifer.

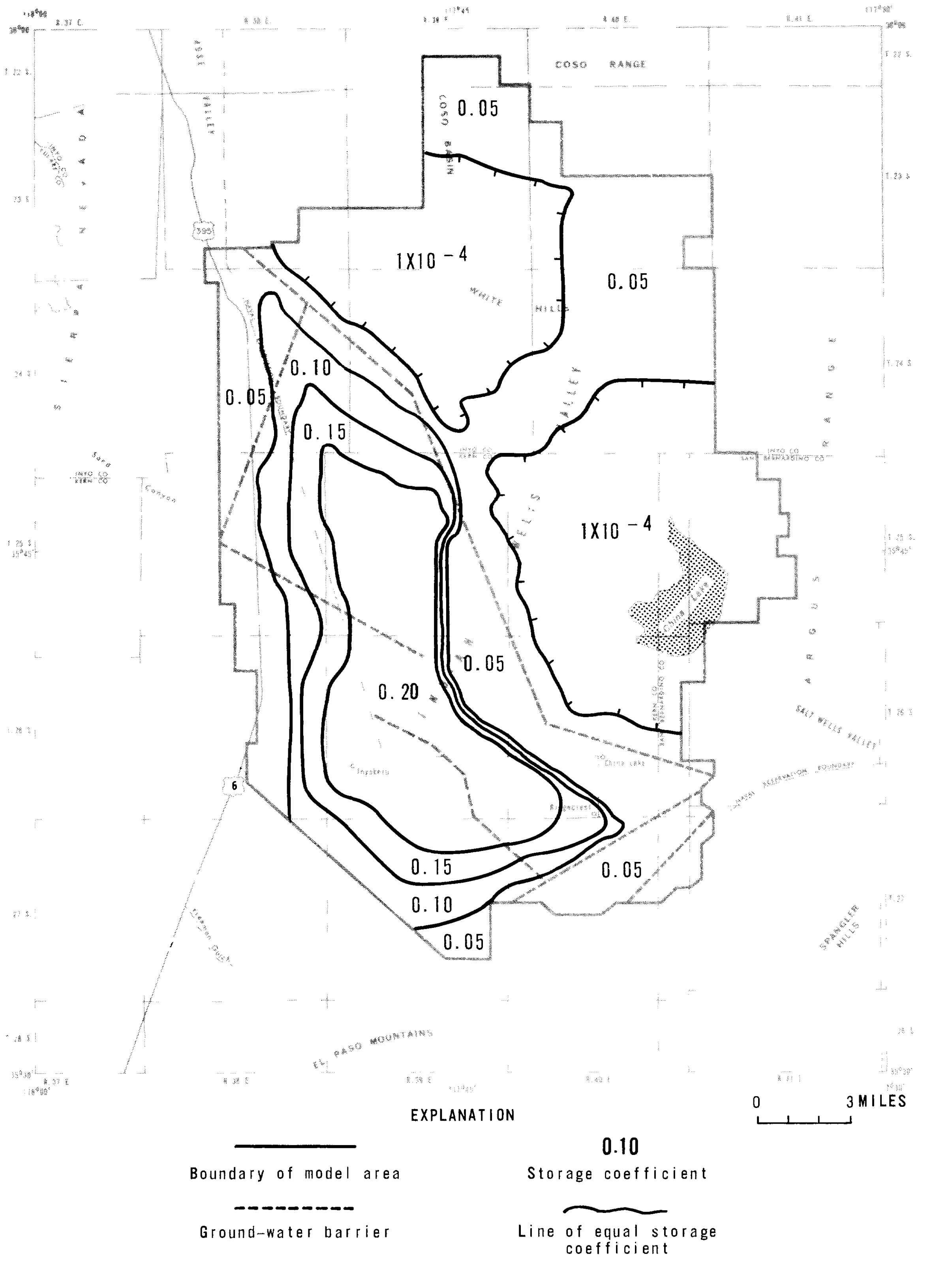


FIGURE 8.--Storage coefficient for the deep aquifer.

Steady-State Ground-Water Recharge and Discharge

The configuration of the two aquifers in Indian Wells Valley is such that under natural conditions recharge and discharge for the deep aquifer and recharge and discharge for the shallow aquifer were all approximately equal and the basin was in steady-state conditions. The digital model was used to simulate these steady-state conditions in each aquifer.

Estimates of average annual evaporation from the basin by Kunkel and Chase (1969, p. 68-69) were used to make the initial estimate of average annual discharge from the shallow aquifer; that estimate was then assumed to be equal to average annual recharge to both the deep and shallow aquifers and equal to average annual discharge from the deep aquifer.

Tables 1 and 2 list the verified steady-state recharge and discharge of 9,850 acre-feet per year for the deep aquifer. Figure 9 shows steady-state recharge values for the deep aquifer by areas. Approximately two-thirds of the total recharge to the deep aquifer originates in the mountainous area southwest of the model area.

Because areas of recharge and discharge for the shallow aquifer may coincide, a node in the shallow aquifer can have both recharge and discharge. When this occurs, the difference between the quantities of recharge and discharge is modeled at that node. These differences are shown in tables 3 and 4, and the areas of recharge and discharge are shown in figure 9.

Recharge to the deep aquifer occurs as ground-water underflow from the permeable materials in canyons of the Sierra Nevada and the Coso and Argus Ranges and as deep percolation of some of the streamflow from Rose Valley and Freeman Gulch (fig. 9). The assumption was made that there was no recharge from deep percolation of precipitation on the valley floor.

Although Kunkel and Chase (1969) estimated total steady-state recharge to the deep aquifer, additional work was required to properly distribute this recharge over the model area. Because no stream-gaging stations or precipitation gages are in the drainage area above the valley floor, a simple altitude-area relation was used to apportion recharge to the various parts of the valley.

The orographic effects in the Sierra Nevada were assumed to be greater than in the Coso and Argus Ranges because more moisture is present in the air as it passes over the Sierra Nevada. Therefore, recharge was assumed to be available from areas in the Sierra Nevada above 4,500 feet altitude and from other areas above 5,000 feet. Within the surface drainage area of Indian Wells Valley there are 88 square miles above 4,500 feet altitude in the Sierra Nevada and 102 square miles above 5,000 feet in the Coso and Argus Ranges. Recharge was apportioned to the individual streams in these categories on the basis of their drainage areas. The resulting recharge was distributed to nodes near the model boundary adjacent to the mouth of the canyons. For example, recharge from Sand Canyon in the Sierra Nevada (fig. 9, table 1) was assumed to occur at nodes 29,2 and 30,2.

TABLE 1.—*Steady-state recharge for the deep aquifer, in acre-feet per year*

Area	Node number	Recharge
Coso Wash	2,16	200
	2,17	200
	2,18	200
	2,19	200
	2,20	200
Petroglyph Canyon	3,21	148
	3,22	148
	4,21	148
	4,22	148
Renegade Canyon	6,24	42
	7,24	93
	8,24	93
Mountain Springs Canyon	9,25	93
	9,26	93
	9,27	93
	9,28	93
	9,29	93
	9,30	93
	9,31	93
	9,32	93
	9,33	94
	9,34	94
Unnamed Canyon	12,34	59
Wilson Canyon	15,34	175
	16,34	175
Burro Canyon	27,36	4
	27,37	4
El Paso drainage	59,16	100
	59,17	100
	59,18	100
	59,19	100
Little Lake	14,2	43

Area	Node number	Recharge
Freeman Gulch	50,6	187
	51,7	207
	52,8	207
	53,9	207
	54,10	207
	55,11	207
	56,12	207
	57,13	207
	57,14	207
	58,15	199
Freeman Canyon	46,4	107
	47,4	107
	48,4	107
	49,5	107
Indian Wells Canyon	44,5	201
	45,5	201
Grapevine Canyon	34,2	205
	35,2	228
	36,2	236
	37,3	200
	38,3	200
	39,3	200
	40,3	200
	41,3	150
Sand Canyon	29,2	248
	30,2	248
Ninemile and Noname Canyons	21,2	145
	22,2	145
	23,2	145
	24,2	145
	25,2	195
Fivemile and Deadfoot Canyons	16,2	95
	17,2	95
	18,2	95
	19,2	95
	20,2	95
Total recharge:		9,850

TABLE 2.--*Steady-state discharge for the deep aquifer, in acre-feet per year*

Node Number	Discharge	Node Number	Discharge	Node Number	Discharge	Node Number	Discharge
21,23	50	26,18	50	32,22	54	38,22	98
21,24	50	26,19	50	32,30	48	38,23	98
21,25	50	26,22	18	32,31	64	38,24	50
21,26	50	26,23	24	32,32	40	38,27	48
21,27	50	26,24	24	32,33	22	38,28	48
21,28	50	26,25	30	32,34	8	38,29	48
21,29	50	26,26	30	33,20	70	38,30	28
21,30	50	26,27	46	33,21	54	38,31	16
21,31	25	26,28	16	33,22	30	38,32	16
21,32	25	26,29	16	33,30	48	38,33	16
22,21	50	26,30	16	33,31	58	38,34	12
22,22	50	26,31	18	33,32	38	39,23	50
22,23	50	26,32	16	33,33	36	39,22	75
22,24	50	26,33	8	33,34	22	39,24	50
22,25	50	27,18	50	34,20	80	39,26	30
22,26	50	27,19	50	34,21	56	39,27	48
22,27	50	27,25	42	34,22	48	39,28	48
22,28	50	27,26	46	34,31	74	39,29	48
22,29	50	27,27	96	34,32	66	39,30	44
22,30	50	27,28	42	34,33	62	39,31	20
22,31	25	27,29	56	34,34	62	39,32	16
22,32	25	27,30	42	34,35	62	39,33	16
22,33	25	27,31	16	34,36	28	40,23	50
22,34	25	27,32	16	35,21	98	40,24	50
23,20	50	27,33	22	35,22	48	40,27	18
23,21	50	28,18	50	35,23	42	40,28	12
23,22	50	28,19	50	35,30	6	40,29	6
23,23	4	28,26	12	35,31	50	40,30	24
23,25	12	28,27	18	35,32	50	40,31	34
23,26	32	28,28	44	35,33	50	40,32	16
23,27	16	28,29	50	35,34	50	41,23	50
23,28	12	28,30	44	35,35	50	41,24	50
23,29	4	28,31	16	36,21	70	41,30	6
24,19	50	28,32	16	36,22	56	41,31	36
24,20	50	28,33	16	36,23	52	41,32	40
24,22	4	29,19	48	36,28	24	42,23	50
24,23	16	29,20	37	36,29	50	42,24	50
24,24	16	29,27	24	36,30	50	42,30	6
24,25	36	29,28	48	36,31	66	42,31	56
24,26	62	29,29	48	36,32	62	42,32	30
24,27	32	29,30	54	36,33	62	43,24	50
24,28	40	29,31	50	36,34	48	43,25	50
24,29	20	29,32	16	36,35	52	43,30	40
24,30	6	29,33	14	36,36	16	43,31	48
25,18	50	30,19	50	37,21	98	44,24	50
25,19	50	30,20	48	37,22	98	44,25	50
25,22	4	30,29	42	37,23	50	44,30	12
25,23	28	30,30	48	37,24	48	44,31	22
25,24	46	30,31	22	37,26	24	45,26	50
25,25	56	30,32	32	37,27	30	45,27	50
25,26	68	30,33	16	37,28	48	45,28	50
25,27	52	31,20	50	37,29	50	45,29	50
25,28	32	31,21	50	37,30	114	45,30	50
25,29	32	31,30	48	37,31	120	45,31	50
25,30	54	31,31	78	37,32	134	45,32	42
25,31	56	31,32	16	37,33	132	46,30	50
25,32	28	31,33	20	37,34	96	46,31	50
		32,21	70	37,35	60	46,32	42
Total discharge							9,850

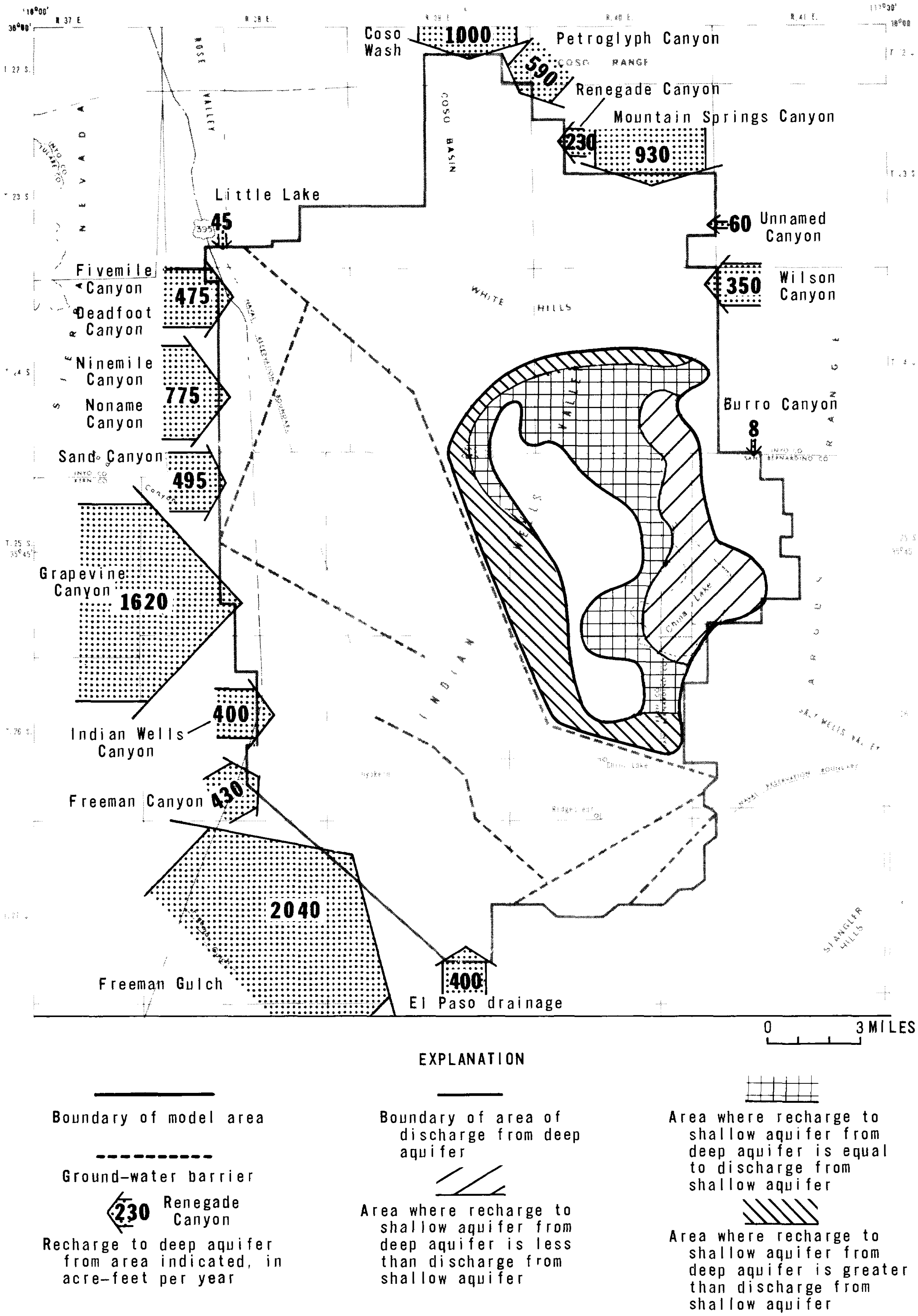


FIGURE 9.--Areas of recharge and discharge for deep and shallow aquifers.

TABLE 3.--*Net steady-state recharge for the shallow aquifer, in acre-feet per year*

[Recharge values are recharge less discharge at each node. The total recharge for the aquifer is 9,850 acre-feet per year]

Node number	Recharge	Node number	Recharge
20,28	50	36,21	8
21,23	50	36,22	22
21,24	50	37,21	26
21,25	50	37,23	50
21,26	50	37,24	50
21,27	50	38,22	20
21,28	50	38,23	50
21,29	50	38,24	50
21,30	50	39,22	44
21,31	25	39,23	50
21,32	25	39,24	75
21,33	25	39,25	25
21,34	25	40,23	50
22,21	50	40,24	50
22,22	50	41,23	50
23,20	50	41,24	50
24,19	50	42,23	50
25,18	50	42,24	50
26,18	50	43,24	50
27,18	50	43,25	50
29,18	50	44,24	50
30,19	50	44,25	50
31,19	50	45,26	50
31,20	50	45,27	50
32,19	70	45,28	50
32,20	54	45,29	50
33,20	30	45,30	50
33,21	70	45,31	44
34,20	48	46,30	50
34,21	80	46,31	50
35,22	50	46,32	42
Total net recharge			2,910

TABLE 4.--*Net steady-state discharge for the shallow aquifer, in acre-feet per year*

[Discharge values are discharge less recharge at each node. The total discharge for the aquifer is 9,850 acre-feet per year]

Node number	Discharge	Node number	Discharge
24,30	30	35,31	6
24,31	36	35,32	6
25,30	60	35,33	50
25,31	60	35,34	50
26,30	102	35,35	30
26,31	102	35,36	16
26,32	98	35,37	34
26,33	50	36,23	20
27,31	80	36,24	6
27,32	116	36,31	10
27,33	90	36,32	20
28,31	58	36,33	50
28,32	44	36,34	60
28,33	38	36,35	50
29,31	46	36,36	50
29,32	46	36,37	14
29,33	30	37,30	10
30,31	40	37,31	20
30,32	40	37,32	30
30,33	40	37,33	60
31,32	40	37,34	60
31,33	40	37,35	50
32,22	24	37,36	16
32,32	28	37,37	8
32,33	40	38,31	30
33,22	24	38,32	40
33,23	30	38,33	76
33,32	30	38,34	66
33,33	30	38,35	58
34,22	36	38,36	30
34,23	48	39,31	30
34,33	50	39,32	50
34,34	30	39,33	50
34,35	10	40,31	6
35,30	6	40,32	50
		40,33	26
Total net discharge			2,910

The selection of the 4,500-foot altitude in the Sierra Nevada and the 5,000-foot altitude elsewhere for effective precipitation was arbitrary and was shown to be in error by a series of model runs. This assumption resulted in too much recharge emanating from the Coso and Argus Ranges and too little recharge emanating from the Sierra Nevada. Evidently the orographic effects are much greater in the Sierra Nevada than originally assumed. However, no attempt was made to adjust the original altitude-area relation; instead, a simple trial-and-error process was used to make changes in the recharge values until the head configuration determined by the model was in agreement with the 1920-21 water-level contour map drawn from available water-level measurements of wells. This assumes that the 1920-21 water-level data, which are the oldest data available in sufficient quantities to construct a contour map, accurately represent steady-state conditions.

The determination of individual steady-state nodal recharge and discharge quantities for the deep and shallow zones was a trial-and-error process, in which the two zones were checked for internal consistency, and in which computed steady-state head configurations were checked for agreement with available historical water-level data.

Non-Steady-State Ground-Water Recharge and Discharge

Most of the principal water users in the valley have metered their ground-water pumpage so that pumpage for the periods of ground-water development (non-steady-state conditions) are readily available. The Navy is the largest water user in the valley followed by Indian Wells Valley County Water District, Stauffer Chemical Co., and American Potash and Chemical Co. Major ground-water developments are near Ridgecrest, near Inyokern, and in the area midway between Ridgecrest and Inyokern.

The metered pumpage can be assumed to be 100 percent consumptively used, at least in relation to the nodal area from which the water was pumped. Almost all the water pumped by the four principal users is either piped out of the valley or used as a source of municipal supply, which prevents significant ground-water recharge from occurring except near the sewage-treatment facilities. Table 5 lists the modeled pumpage by node for the period 1930-68, in acre-feet per year. These quantities of discharge are used in the non-steady-state model for the deep aquifer.

TABLE 5.--*Pumpage by node for the deep aquifer, for the period 1930-68, in acre-feet per year*

Node	1930	1931	1932	1933	1934	1935	1936	1937	1938	1939
29,15										
30,15										
31,21										
36, 9										
36,19										
37, 8										
38, 9										
39,17										
40,12										
41,24										
42,18										
43, 9		5	5	5	5	5	5	5	5	5
46,10										
46,19										
46,20										
47,10										
47,14										
47,19										
47,20										
47,21										
47,22										
48,10	20	20	20	20	20	20	20	20	20	20
48,17										
48,19										
48,21										
48,22										
48,25								25	25	25
48,27						25	25	25	50	50
49,20										100
49,21				50	100	150	200	250	300	300
49,27										
50,23										
50,26		75	175	225	275	300	350	400	500	700
51,26	105	205	205	205	205	205	205	205	255	305
51,27										
52,26										
Total	125	305	405	505	605	705	805	930	1,155	1,505

TABLE 5.--*Pumpage by node for the deep aquifer, for the period 1930-68, in acre-feet per year*--Continued

Node	1940	1941	1942	1943	1944	1945	1946	1947	1948	1949
29,15										
30,15										
31,21							2	2	2	2
36, 9										
36,19										
37, 8										
38, 9										
39,17										
40,12										
41,24										
42,18										
43, 9	5	5	5	5	5	5	5	5	5	5
46,10										
46,19										
46,20						72	208	277	276	283
47,10						182	123	44	246	742
47,14										
47,19				300	300	300	300	300	300	300
47,20						7	180	401	644	685
47,21						149	220	500	482	534
47,22							180	271	302	257
48,10	25	25	25	25	25	25	25	28	35	35
48,17	2	2	2	2	2	2	2	2	2	2
48,19										
48,21										
48,22										
48,25	25	25	25	25	25	25	25	25	25	25
48,27	50	50	50	50	50	50	50	50	50	50
49,20										
49,21	300	300	300	300	300	300	300	300	300	300
49,27										
50,23										
50,26	700	700	700	700	700	700	700	700	700	700
51,26	355	405	455	505	540	650	775	850	925	1,000
51,27						190	19	243	236	301
52,26										
Total	1,462	1,512	1,562	1,912	1,947	2,657	3,114	3,998	4,530	5,221

TABLE 5.--*Pumpage by node for the deep aquifer, for the period 1930-68, in acre-feet per year*--Continued

Node	1950	1951	1952	1953	1954	1955	1956	1957	1958	1959
29,15				1	1	1	1	1	1	1
30,15				1	1	1	1	1	2	2
31,21	2	2	2	2	2	2	2	2	4	4
36, 9										
36,19				5	5	5	5	5	6	6
37, 8			350	350	350	350	350	350	385	385
38, 9							35	35	35	35
39,17	5	5	5	5	5	5	5	5	9	9
40,12				1	1	1	1	1	1	1
41,24	1	1	1	1	1	1	1	1	1	1
42,18			5	5	5	5	5	5	5	5
43, 9										
46,10										
46,19										
46,20	227	128	111	132						
47,10	829	1,441	2,069	2,007	3,221	2,720	3,473	4,017	3,079	3,780
47,14									35	35
47,19	300	200	211	187	100	100	100	111	99	103
47,20	712	592	536	475	311	314	180	251	747	583
47,21	677	416	547	558	411	384	297	441	338	237
47,22	340	263	237	222	130	100	86	50	115	50
48,10	36	1,034	738	1,192	1,040	1,847	1,628	1,308	1,475	1,570
48,17	5	5	5	5	5	5	5	5	5	5
48,19		100	100	100	100	100	98	100	99	103
48,21					100	100	100	111	99	103
48,22										
48,25	25	25	25	25	25	25	25	25	25	25
48,27	50	50	50	50	50	50	50	50	50	50
49,20										
49,21	300	300	300	300	300	300	250	250	250	250
49,27		35	40	40	45	45	50	50	30	30
50,23									35	35
50,26	600	500	400	350	300	250	250	150	100	50
51,26	1,070	935	960	987	1,132	1,246	1,319	1,319	1,714	1,896
51,27	613	139	69	147	37	123	165	139	313	245
52,26		140	150	160	170	175	185	205	261	295
Total	5,792	6,311	6,911	7,306	7,848	8,255	8,697	9,018	9,353	9,929

TABLE 5.--*Pumpage by node for the deep aquifer, for the period 1930-68, in acre-feet per year*--Continued

Node	1960	1961	1962	1963	1964	1965	1966	1967	1968
29,15	1	1	1	1	1	1	2	2	3
30,15	2	2	2	2	2	2	12	13	9
31,21	4	4	4	4	4	1	9	7	5
36, 9	35	20	20	20	20	20	20	20	20
36,19	6	6	6	6	6	6	31	31	18
37, 8	385	280	280	280	280	280	280	280	280
38, 9	35	20	20	20	20	20	20	20	20
39,17	9	9	9	9	9	9	15	15	11
40,12	1	1	1	1	1	1	3	2	2
41,24	1	1	1	1	1	1	10	11	11
42,18	5	5	5	5	5	5	18	26	29
43, 9									
46,10				765	904	459	1,107	1,241	964
46,19				4,026	3,387	3,565	2,933	3,665	3,161
46,20									
47,10	3,989	4,034	3,741	372	766	1,237	1,186	864	1,727
47,14	35	35	35	35	35	35	35	35	35
47,19	92	117	150	162	165	168	198	225	251
47,20	1,009	716	971	1,303	1,450	1,447	1,081	1,259	887
47,21	142	167	200	212	215	218	248	275	251
47,22	50	50	50	50	50	50	50	50	
48,10	1,279	1,426	1,792	138	635	282	917	199	822
48,17	5	6	7	8	9	10	11	14	16
48,19	92	117	150	162	165	168	198	225	251
48,21	92	117	150	162	165	168	198	225	251
48,22						154	662	735	914
48,25		25	25	25	25	25	25	25	
48,27	25	50	50	50	50	50	50	50	
49,20									
49,21	250	250	250	250	250	250	250	250	
49,27	30	32	34	34	36	28			
50,23	35	35	35	35	35	35	35	35	
50,26	50								
51,26	1,911	1,980	2,122	2,214	2,059	2,164	2,012	2,014	1,998
51,27	45	254	368	2					
52,26	298	318	338	379	431	481	449	403	466
Total	9,913	10,078	10,817	10,733	11,181	11,340	12,065	12,216	12,437

Estimates of recharge of sewage effluent to both the shallow and deep aquifers were made. Estimated ground-water recharge from the sewage ponds is the difference between the flow into the sewage ponds and the potential evapotranspiration from the pond area. Metered sewage-effluent data were available for 1953-67. A potential evapotranspiration of about 610 acre-feet per year was computed for the average surface area of the sewage ponds, and about 980 acre-feet per year was computed for the surface area of the Navy golf course, which is watered by sewage effluent. No recharge was assumed from the effluent applied to the golf course because total applied effluent never exceeded the potential evapotranspiration in any year. Recharge to either the shallow or the deep aquifer from the Ridgecrest and Inyokern sewage-disposal sites was assumed to be negligible because the estimated potential evapotranspiration was greater than the total annual sewage effluent.

In modeling the recharge from the Navy sewage ponds (fig. 7), it was assumed that half the total recharge from the ponds percolated to the shallow aquifer and half percolated to the deep aquifer. The direction of this water movement between aquifers is opposite to that of steady-state conditions because of non-steady-state head changes in the shallow and deep aquifers. Table 6 shows the sewage-effluent recharge used in the non-steady-state model for the deep aquifer.

TABLE 6.--*Sewage-effluent recharge by node for the deep aquifer*

Node	1954	1955	1956	1957	1958	1959	1960	1961	1962	1963	1964	1965	1966	1967	1968
43,28													20		
44,28	39	45	50	50											
44,29	80	90	95	100											
45,28	39	45	45	48	30	28	28		8	10		10		15	22
45,29	75	90	95	95	60	55	55	35	15	20	30	20	40	30	43
45,30	80	90	95	100											
45,31	80	90	95	100											
45,32	39	45	50	50											
46,29	43	49	49												
46,30	75	90	95	95	60	55	55	35	15	20	30	20	40	30	43
46,31	75	90	95	95	60	55	55	35	15	20	30	20	40	30	43
46,32	39	45	45	48	30	28	28	18	8	10	15	10	20	15	22
Total:	664	769	809	781	240	221	221	123	61	80	105	80	160	120	173

Capture of natural ground-water discharge from the deep aquifer began about 1963. Capture is the reduction of natural discharge from one area of ar aquifer because of an increased artificial discharge from another area of the aquifer. Discharge from the deep aquifer near China Lake playa is being reduced by the increase in ground-water pumping near Ridgecrest and the increase of water levels in the shallow aquifer near China Lake because of recharge from the Navy sewage ponds. This caused a decrease, or in some areas a reversal, of the steady-state head differential between the two aquifers.

No special programing was done to automatically handle the capture in the model. The assumption was made that if there was a 10-foot head decline at a node in the deep aquifer with steady-state discharge the effect of capture became significant. Half of the steady-state nodal discharge was eliminated during the first year after capture began. In the second year the entire nodal discharge was eliminated if the head declines were not reduced by the capture assumed for the first year. Table 7 shows the distribution and quantity of capture used in the model for the period 1963-68.

TABLE 7.--*Capture of ground-water discharge from deep aquifer, for the period 1963-68, in acre-feet per year*

Node	1963	1964	1965	1966	1967	1968
43,24	0	25	25	25	25	25
43,25	0	25	25	25	25	25
43,30	0	20	20	20	20	20
43,31	0	24	24	24	24	24
44,24	0	25	25	25	25	25
44,25	0	25	25	25	25	25
44,30	0	6	6	6	6	6
44,31	0	11	11	11	11	11
45,26	25	50	50	50	50	50
45,27	25	50	50	50	50	50
45,28	25	50	50	50	50	50
45,29	25	50	50	50	50	50
45,30	0	25	50	50	50	50
45,31	0	25	50	50	50	50
45,32	0	21	42	42	42	42
46,30	50	50	50	50	50	50
46,31	50	50	50	50	50	50
46,32	42	42	42	42	42	42
Total:	242	574	645	645	645	645

THE GROUND-WATER MODEL

Assumptions Required for Modeling

If the actual aquifer system in Indian Wells Valley were to be precisely modeled, a three-dimensional nonlinear flow equation would be necessary to define the flow within the system. Solutions to such an equation are not readily nor economically available except for simple cases. Hence, simplifying assumptions of the structure of the Indian Wells Valley ground-water basin and of the flow in the basin are necessary if a solution to the flow equation is to be obtained.

The simplifying assumptions made are:

1. There are only two aquifers in the valley, deep and shallow. Where the shallow aquifer occurs, the underlying deep aquifer is confined.

2. Flow between the aquifers under steady-state conditions is in one direction, from deep to shallow.

3. Where the deep aquifer is confined, it is of uniform thickness.

4. Where either aquifer is unconfined, the drawdown with respect to the saturated thickness is small (transmissivity is constant with time).

5. The storage coefficient in either aquifer is constant with time.

6. Vertical flow components within either aquifer are negligible compared with horizontal flow components.

Kunkel and Chase (1969, p. 39) showed the deep aquifer as being confined east of a confining clay bed. This probably was the case under steady-state conditions. However, model runs suggest that in the area between the edge of the confining clay and the fault zones trending northwest-southeast, from node 23,15 to 48,34, a confined condition could not persist in the deep aquifer when pumping occurred. This is because pumping in this area produces large head declines and the potentiometric surface falls below the confining zone, thus producing water-table conditions. To abide by the assumption of a constant storage coefficient with time, a storage coefficient of 0.05 was used in the area of question. This yielded reasonable head-decline results in the model verification.

The above modification also defined the western extent of the shallow aquifer under steady-state conditions as being along the trace of the fault zones trending from node 23,15 to node 48,34 (fig. 2). Kunkel and Chase (1969, figs. 3 and 4) showed the thickness of the shallow aquifer diminishing from east to west. For the model study, the shallow aquifer was assigned zero thickness adjacent to the fault zones mentioned above and increasing thickness to the east.

Even if the assumed zero thickness boundary is in error, the saturated thickness of the shallow aquifer approaches zero at some point near the center of the valley. The significance of zero or small saturated thickness is that any change in head in the shallow aquifer cannot be small in relation to the saturated thickness. Hence, assumption 4 is violated, and the shallow aquifer could not be modeled under non-steady-state conditions.

Verification of the Model

Before using a model to predict future ground-water levels, the model parameters must be verified or checked against available geologic and hydrologic data. When the model-generated water levels for a particular set of conditions approximate the historical water levels within some predetermined limit of accuracy, the model is considered verified and ready for predictive use.

The continuous form of the two-dimensional differential equation used to describe the flow conditions in the nonhomogeneous anisotropic aquifer of Indian Wells Valley is:

$$\frac{\partial}{\partial x}\left(T_{xx}\frac{\partial h}{\partial x}\right)+\frac{\partial}{\partial y}\left(T_{yy}\frac{\partial h}{\partial y}\right)=S\frac{\partial h}{\partial t}+W$$

where $\frac{\partial}{\partial x}$ and $\frac{\partial}{\partial x}$ are first partial derivatives,

T_{xx} is transmissivity in the x direction,
T_{yy} is transmissivity in the y direction,
h is the hydraulic head,
S is the storage coefficient,
W is the net rate of pumping per unit area.

The discrete form of the above equation was solved by the alternating-direction method (Peaceman and Rachford, 1955). The solution to the equation is the hydrologic model.

Using the estimates of T, S, and W and the historical head values, the verification of the Indian Wells Valley model proceeded in three steps:

1. Simulation of steady-state conditions in the deep aquifer.

2. Simulation of steady-state conditions in the shallow aquifer.

3. Simulation of non-steady-state conditions in the deep aquifer.

The computers used in the study were an IBM-360 model 50 and an IBM-360 model 65. The digital program is in FORTRAN IV language. The program is usable on any machine capable of compiling FORTRAN IV and possessing a storage capacity of at least 150,000 bits.

Steady-State Water Levels

The steady-state water-level contours (fig. 10) constructed from model-generated head values for the deep aquifer compare favorably with the water-level contours constructed by L. C. Dutcher and W. R. Moyle, Jr. (written commun., 1970) from water-level data for 1920-21 (fig. 4) when conditions were assumed to approximate steady-state conditions. The model-generated heads for the northern half of the model area cannot be directly compared with historical data for that period because water-level data are not available. However, the model verification assures that the computed head values for the entire model area are compatible with all hydrologic information about the basin. In the area where water-level data are available, the model-generated heads are within about 5 feet of the measured heads.

The computed steady-state data suggest that the natural flow of ground water in the deep aquifer was from the north, west, and southwest towards the depression at China Lake playa (fig. 10). The highest indicated heads are in the extreme northern part of the valley along row 2, and the lowest indicated head is at node 37,33 in China Lake playa.

Because few historical water-level data were available for the shallow aquifer, the model-generated steady-state water-level contour map for the shallow aquifer (fig. 11) is a derivative or by-product of the computations of head in the deep aquifer. The actual points of recharge, the amounts of recharge, and the boundary head values were all fixed by the results of the deep aquifer head computations. This sort of derivative or by-product process is a valuable part of the model study. Assumptions are made and then verified or modified as necessary until verification is obtained. Additional assumptions or estimates based upon the earlier verified assumptions can then be made and so on as in a building process. The model then generates results for the entire model area. This technique assures that the results are internally consistent. The reasonableness and accuracy of the original assumptions are the responsibility of the hydrologist.

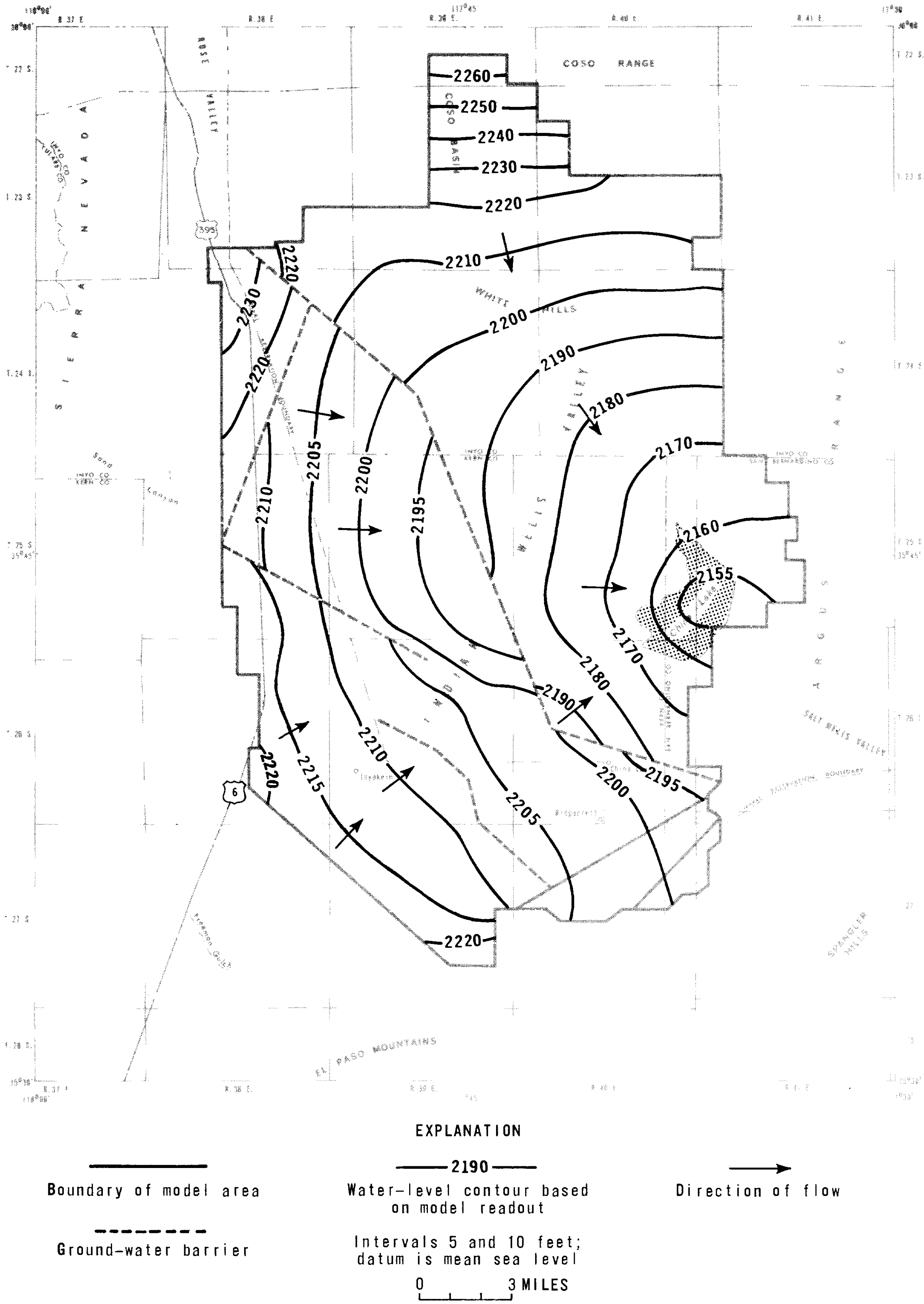
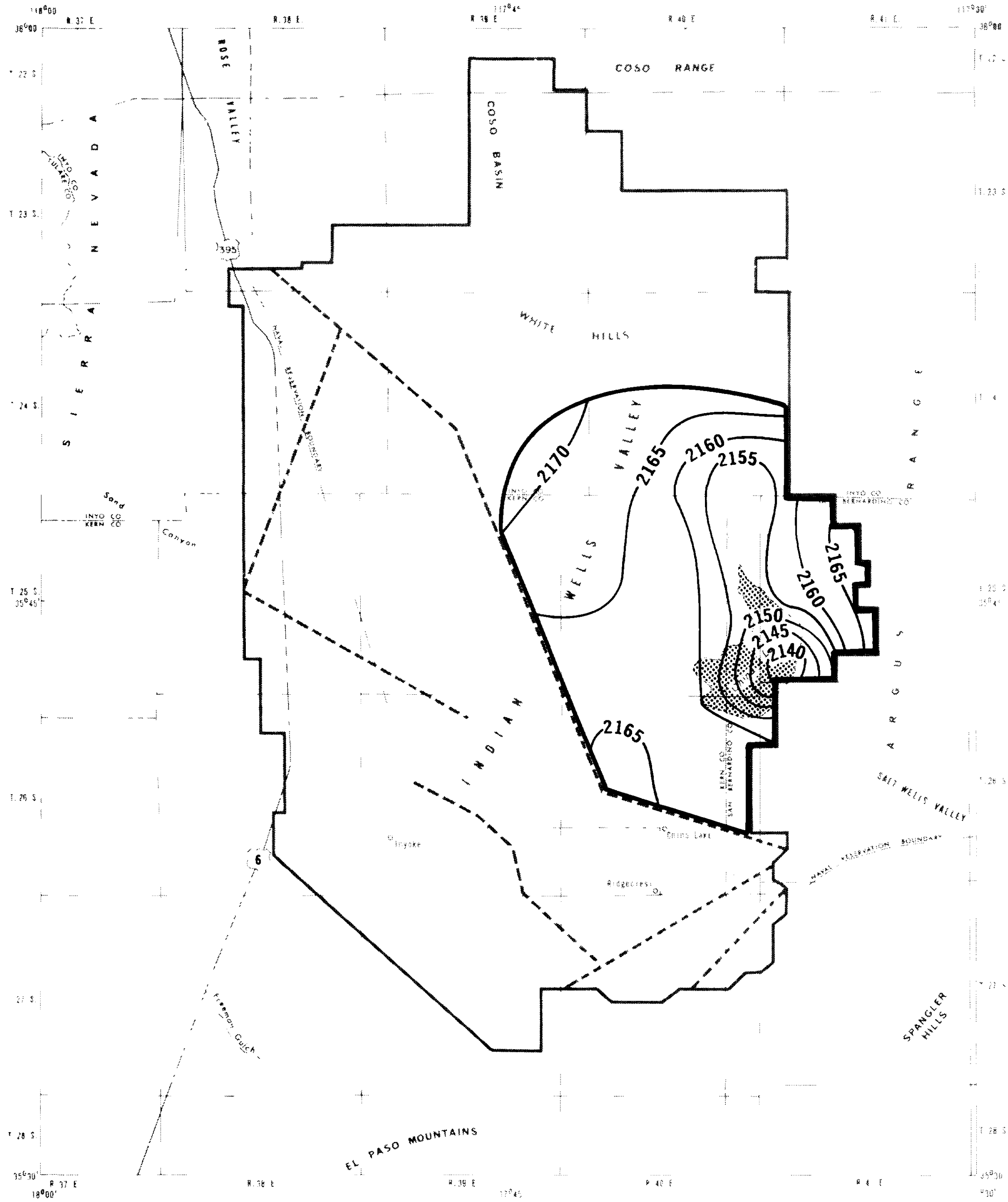


FIGURE 10.--Model-generated steady-state water-level contours for deep aquifer.



EXPLANATION

Boundary of model area

Ground-water barrier

Estimated boundary of shallow aquifer

——2170—— Water-level contour

Interval 5 feet; datum is mean sea level

0 3 MILES

FIGURE 11.--Model-generated steady-state water-level contours for shallow aquifer.

Non-Steady-State Water Levels

Non-steady-state conditions were assumed to begin in Indian Wells Valley in 1931. Although ground-water pumping began before 1931, pumpage was not significant until then and major pumping did not begin until the 1940's.

Ground-water pumping fluctuates seasonally because much more water is pumped during the summer than during the winter. An example of this fluctuation is the Indian Wells Valley County Water District pumpage, which is similar to the requirements of most other water users in the valley (fig. 12). In 1967 more than 50 percent of the total annual pumpage by the water district was in June through September.

The fluctuation in ground-water levels caused by seasonal fluctuations in pumping rates is shown by the water-level records of typical wells in the major pumping areas. Hydrographs of these wells exhibit saw-toothed curves with a decline to a low point each autumn, a rise to a high point each spring, and a general decline in water levels from year to year. In verifying the model under non-steady-state conditions, no attempt was made to model the seasonal fluctuations of the hydrographs because the model is based on annual decline in water levels. All pumpage was modeled as an average annual rate for each pumping node.

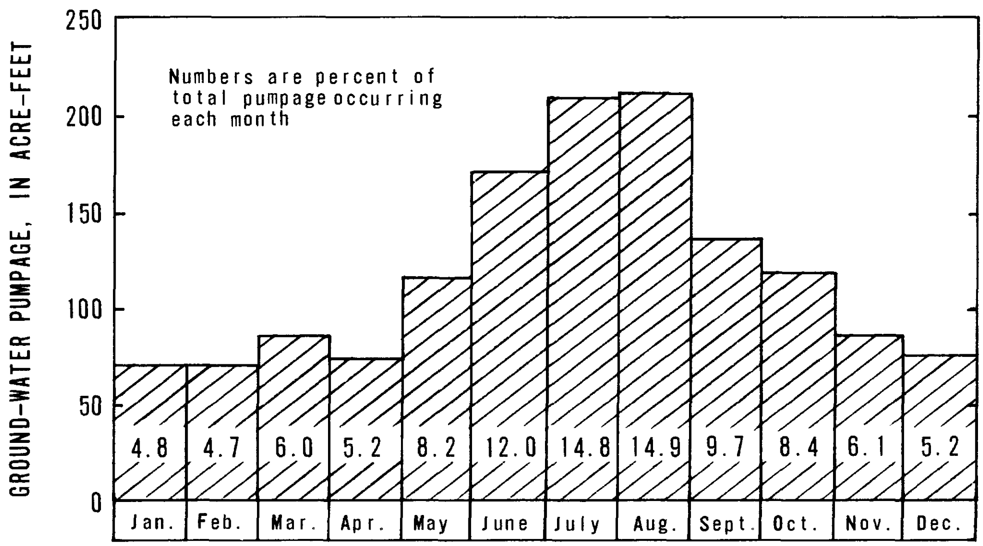


FIGURE 12.--Monthly pumpage of ground water by Indian Wells Valley County Water District for 1967.

Care must be taken in comparing model-generated water levels with measured water levels because of this modeling procedure. Model-generated water levels for 1966 probably would be higher, or exhibit less drawdown, than measured water levels for autumn 1966. By contrast, model-generated 1966 water levels would probably be lower than measured water levels for spring 1967. The model-generated water levels shown on the series of contour maps are for the last day of the year being considered and represent the average water level for that year.

The 1953 water-level contour map (fig. 3) constructed from measured water levels compares favorably with the 1953 water-level contour map based on model readouts (fig. 13). In general, the two maps agree within about 5 feet, except for two local areas. The areas immediately northeast and south of China Lake playa show discrepancies of about 10 feet. This may be due to an incorrect distribution of steady-state recharge at the model boundaries near those areas. Because of the lack of 1920-21 water-level data in those areas, the steady-state model cannot indicate the discrepancy.

The model-generated water levels show declines of more than 10 feet between 1931 and 1953 in the Ridgecrest-China Lake area and in the Inyokern area. The model indicates that water levels declined in about half of the model area between 1931 and 1953, although pumping was concentrated in the southern part of the area. The area north of row 21 did not have significant water-level declines and can be considered to be in a steady-state condition in 1953.

The 1968 water-level contour map based on water-level data (fig. 5) compares favorably with the model-generated 1968 water-level contour map (fig. 14).

Two main pumping depressions are in evidence, one near Ridgecrest and one about midway between Ridgecrest and Inyokern. A smaller pumping depression is in the Inyokern area. The general movement of ground water in the northern part of the valley is toward China Lake playa and in the southern part toward the two main pumping depressions. A reversal of flow is in evidence across the fault zone trending northwest-southeast near China Lake. Whereas under steady-state conditions the flow across the fault was from south to north, the flow in 1968 was from north to south.

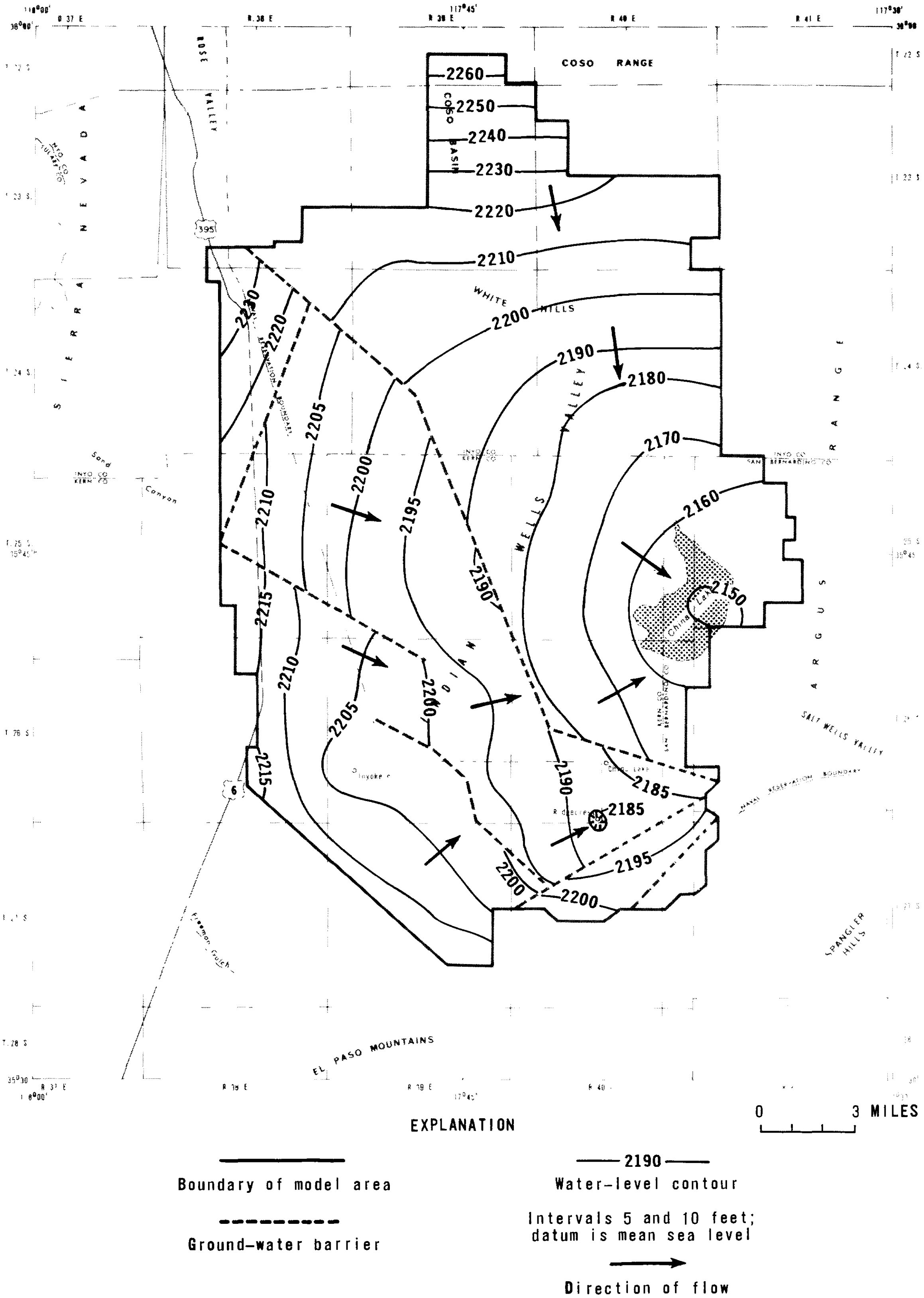


FIGURE 13.--Model-generated 1953 water-level contours for deep aquifer.

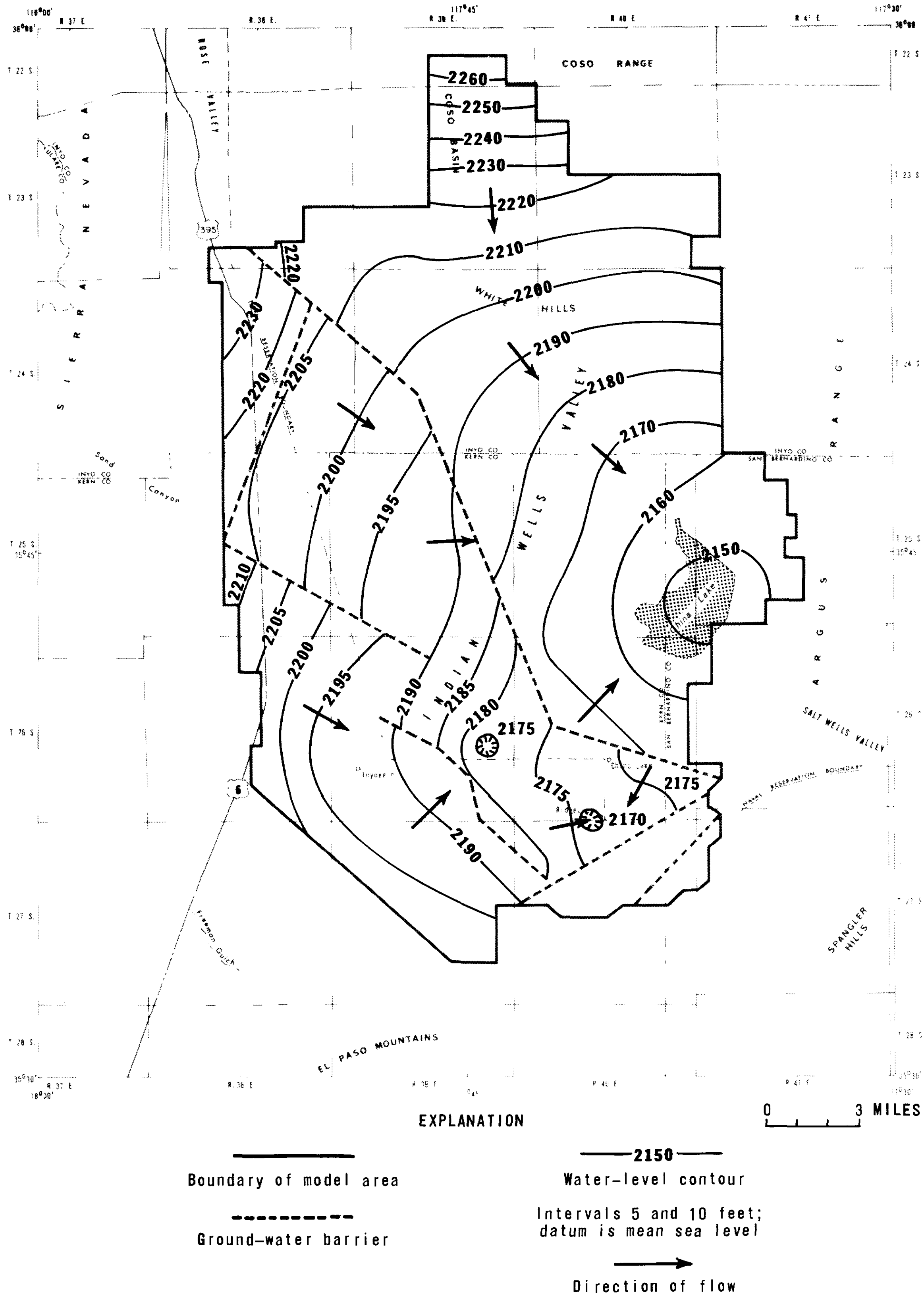


FIGURE 14.--Model-generated 1968 water-level contours for deep aquifer.

A difference of 10 to 15 feet exists between the water-level contours in figures 5 and 14 in the area southeast of Inyokern and in the area east of Ridgecrest and China Lake. These differences are probably due to localized discrepancies in the data used to construct the model and should not significantly impair the usefulness of the model as a predictive tool. In general the water levels shown in figures 5 and 14 agree within about 5 feet. This degree of similarity is adequate verification of the ability of the model to compute head configurations from given parameters. If future parameters are similar in magnitude, distribution, and duration to the historic parameters used to verify the model, the future water-level configuration generated by the model should be about as accurate as that shown by comparison of figures 5 and 14. Whether or not this predicted water-level configuration truly represents the water-level conditions that will exist in the future depends on how accurately the parameters used in the model represent the actual future parameters.

Although model-generated water-level contour maps for the deep aquifer are presented for only 3 different years, several others were drawn and analyzed during the study. From data available for the southern part of the area contour maps for the late 1950's through 1967 could be constructed. In addition to the water-level contour maps and water-level change maps, a series of simultaneous hydrographs for wells in the major pumping areas were compared with computed water-level declines at the nearest nodes in the model. When comparisons of measured water-level decline and computed drawdowns are reasonable, the model is considered to be verified.

The results of the model study suggest that the fault shown by figure 3 as trending southeast from sec. 10, T. 26 S., R. 39 E., to sec. 18, T. 26 S., R. 40 E., is not as effective a hydrologic barrier as is implied by the water-level contours. However, the computed head declines suggest that the postulated fault zones or low permeability zones used in the hydrologic model are effective ground-water barriers. This is also borne out by water-level measurement data. The fault zone which trends southwest-northeast from node 56,21 (fig. 2) to near node 49,33 and the one which trends northwest-southeast from node 45,24 to node 48,34 seem to be especially effective barriers.

PREDICTIONS BY THE MODEL

The main use of the Indian Wells Valley ground-water model will be for simulation of future ground-water levels under various patterns and quantities of pumping. The question naturally arises as to how accurately the model can make these simulations.

The user must realize that the model is a dynamic tool. Perfect results from the present model, which is based on an incomplete knowledge of the hydrology of Indian Wells Valley, should not be expected. However, as more data become available, refinements in the model can be made and greater accuracy can be achieved. The present model should be adequate to determine general water-level patterns and approximate drawdown values. With no actual field data available for almost half of the valley, it would be unwise to expect accuracies greater than plus or minus a few feet from model simulations extending into the long-term future.

However, in the Ridgecrest and Inyokern areas, where the most data are available, historical ground-water levels have been simulated within a few feet in most places. Therefore, the implication is that in these areas the model is more precise. If the model is refined as new data become available, it should become a more valid and increasingly valuable tool.

The hydrologic model of Indian Wells Valley is now available to the cooperating agencies as a predictive tool. The end products of each interrogation run will be water-level decline values or head values for each node. Therefore, a question whose answer is expressible in terms of a water-level or head value or is directly related to water-level values is a reasonable question to pose to the model. This assumes that for any question the proper input data are made available by the poser of the question.

Examples of questions that may be posed are:

1. Assuming the present pumping patterns are continued for a specified number of years, what will the water-level configuration in Indian Wells Valley be after the specified number of years?

2. With a specified distribution and quantity of pumping, and an economic pumping limit, when will the economic pumping limit be reached in time?

3. How much ground water can be pumped from a particular area at a predetermined pumping rate before some specified water-level decline occurs at a specified node or series of nodes?

4. What effect will a change in pumping patterns have on the present ground-water levels?

A first estimate of ground-water levels in 1983 in Indian Wells Valley (fig. 15) was produced by the model to give the cooperators an idea of the configuration of future water levels in the valley. The computations were based on an initial set of projected pumpage figures. With this first estimate as a guide various alternative pumping patterns will be considered and then tested with the model.

Because the initial water-level prediction run does not incorporate additional capture after 1968, the model output for 1983 depicts what is probably the maximum possible water-level decline for the period 1930-83 for the given withdrawal figures. Additional capture was not considered at this time because the main purpose of the initial model run was to supply the cooperating agencies with an initial estimate of future declines based on withdrawal figures comparable to 1968-69 values. Naturally, additional capture will be considered in future runs.

ADDITIONAL DATA REQUIREMENTS

Additional water-level and pumpage data should be collected in NE¼ T. 26 S., R. 39 E. This study suggests a greater decline of ground-water levels in this area by 1968 (fig. 5) than shown by L. C. Dutcher and W. R. Moyle, Jr. (written commun., 1970). The need for additional data for this locality is not yet critical. However, as drawdowns become greater (fig. 15) in the area between Ridgecrest and Inyokern, additional data may be necessary to better define the extent of drawdowns caused by pumping.

The scope of this investigation did not include the evaluation and analysis of ground-water chemical-quality data. Future studies should include the potential for degradation of the chemical quality of the local ground-water supply because chemical quality is an important consideration in long-term basin management decisions.

To facilitate future investigations, ground-water quality data should be collected in the area adjacent to and north of the postulated fault trending northwest-southeast from node 43,23 to node 47,32. This is the area in which a ground-water quality monitoring network may become vital because of the possibility of poor-quality ground water migrating into the pumping depression in the fresh-water aquifer near Ridgecrest.

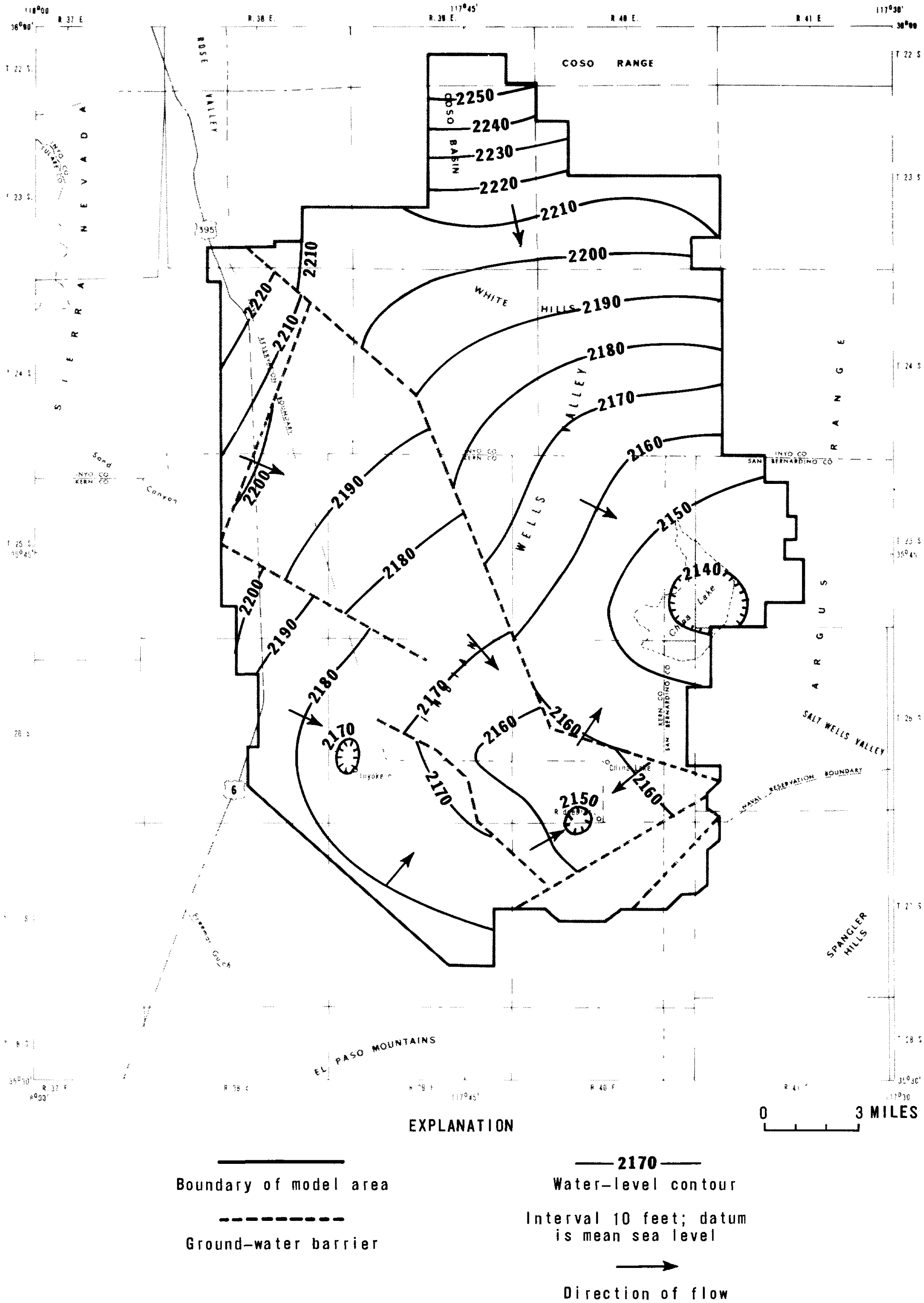


FIGURE 15.--Model-generated 1983 water-level contours for deep aquifer.

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