Subsurface Hydraulics in the Area of the Gila River Phreatophyte Project, Graham County, Arizona

GEOLOGICAL SURVEY PROFESSIONAL PAPER 655-F

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By RONALD L. HANSON

With a section on AQUIFER TESTS

By S. G. BROWN

GILA RIVER PHREATOPHYTE PROJECT

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An evaluation of the aquifer constants and ground-water movement in the alluvial and basin-fill deposits of the Gila River Phreatophyte Project area

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SUBSURFACE HYDRAULICS IN THE AREA OF THE GILA RIVER PHREATOPHYTE PROJECT, GRAHAM COUNTY, ARIZONA

By RONALD L. HANSON

ABSTRACT

Along a 15-mile reach of the Gila River valley upstream from the San Carlos Reservoir in south-central Arizona, the flood plain and its adjacent terraces are underlain by basin fill and alluvial deposits. The basin fill consists of silt, sand, and clay and is estimated to be more than 1,000 feet thick. The alluvium consists of as much as 60 feet of gravel, sand, and silt and fills a 6,000-foot-wide valley incised in the basin fill.

Results obtained by using several analytical techniques indicate that the average storage coefficients for the basin fill and alluvium are 0.0005 and 0.15, respectively. The average transmissivities for the basin fill and alluvium are 15 cubic feet per day per foot (110 gallons per day per foot) and 28,000 cubic feet per day per foot (210,000 gallons per day per foot), respectively. Estimates of hydraulic conductivity are 0.015 foot per day (0.11 gallon per day per square foot) for the basin fill and 695 feet per day (5,200 gallons per day per square foot) for the alluvium.

Downvalley ground-water flow through the alluvium averages 5.1 acre-feet per day. Estimates of flow from the basin fill into the overlying alluvium range from 0.14 to 2.4 acre-feet per day per mile of valley length.

INTRODUCTION

The primary objective of the Gila River Phreatophyte Project is to evaluate evapotranspiration from an analysis of all the significant components comprising the hydrologic system (Culler and others, 1970). Because ground water is a principal component in the hydrologic system, a relatively accurate evaluation of its amount and rate of movement through the project area is a prerequisite to making reliable estimates of evapotranspiration.

This ground-water flow is a function of the hydraulic properties of the water-bearing material underlying the flood plain and terraces. The two principal hydraulic properties evaluated in this study are the storage coefficient and the transmissivity for both the basin fill and the overlying alluvium. Included in this investigation are estimates of (1) the areal extent of the saturated alluvium, (2) the spacial variability of the storage coefficient, transmissivity, and diffusivity of the alluvium and basin fill, (3) the upward vertical velocity of ground water from the basin fill to the overlying alluvium, and (4) the rate of ground-water movement in the alluvium and basin fill. Where sufficient data were available, estimates were also made of the accuracy of these determinations. Only the basic equations used in the solutions and the significant constraints unique to each method are presented. A more detailed discussion of the development of each method is available in the indicated references. All equations are expressed in a general form, hence; they are applicable where consistent units are used.

DESCRIPTION OF STUDY AREA AND WATER-BEARING DEPOSITS

The project area covers a 15-mile reach of the Gila River valley immediately above the San Carlos Reservoir in south-central Arizona (fig. 1). The oldest rock units in the area are igneous and metamorphic and are exposed only in the adjoining mountains. In the valley these units are overlain by a fine-grained material of low permeability referred to as basin fill. This material, which consists primarily of clay, silt, and sand, is more than 1,000 feet thick and 10 miles wide (Davidson, 1961).

Alluvial deposits with relatively high permeability fill a valley incised in the basin fill. These deposits, as much as 60 feet thick and averaging 6,000 feet in

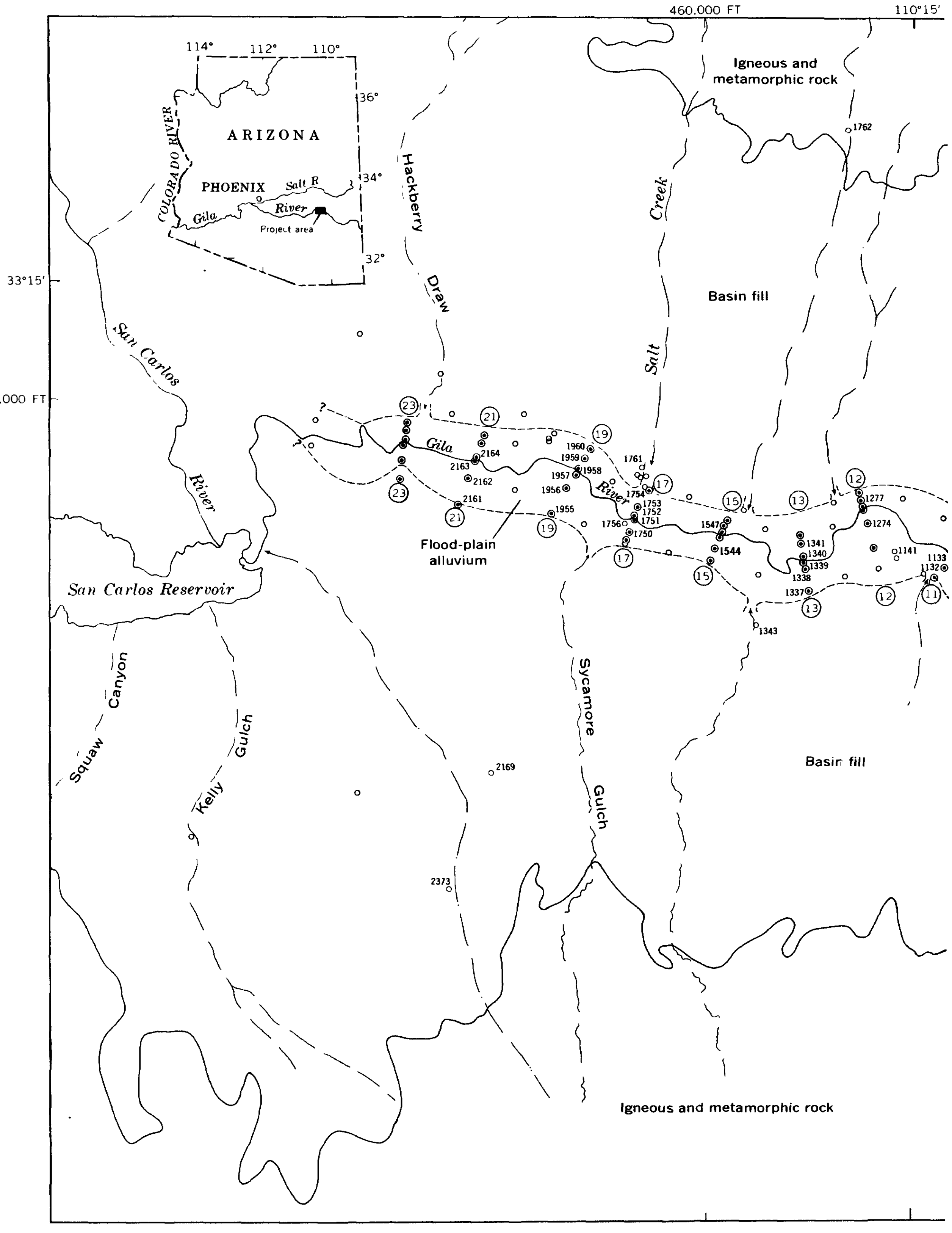
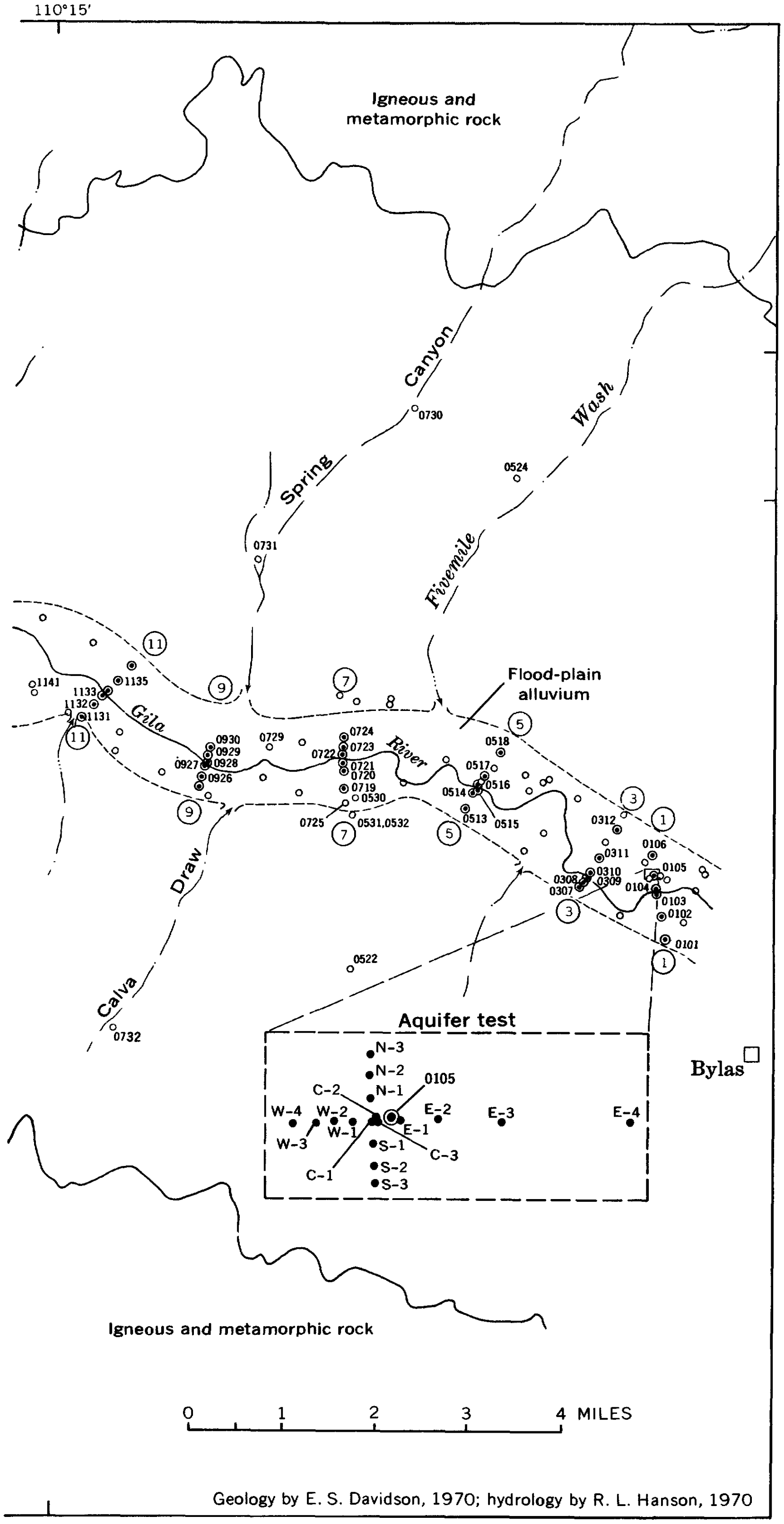


FIGURE 1.—Location of cross sections, data-collection sites, boundaries of saturated alluvium,



EXPLANATION

Contact
Between basin-fill deposits and igneous and metamorphic rocks

Contact
Between basin-fill deposits and saturated alluvium

1

Cross section and number

Wells
Well numbers (0926) *given only for wells discussed in this report*

0106
Observation well and soil-moisture access pipe in cross section

0522
Observation well at miscellaneous site

E-4
Observation well used in aquifer test at cross section 1

and the contact between basin fill and igneous and metamorphic rock in the project area.

width, form the flood plain and lower terraces of the study area. They consist of lenticular gravel, sand, and silt beds. The present Gila River flows across the alluvium in a channel 100 to 400 feet wide and 5 to 8 feet deep.

Figure 2 shows a typical geologic section through the flood plain and lower terraces near the middle of the study area. A more detailed description of the geology of the area was discussed by Weist (1971).

To study ground-water movement in the area, a program was initiated in 1963 to record ground-water levels in 164 wells and soil-moisture contents at 77 access tubes. The locations of these sites are shown in figure 1. Well numbers are indicated on the map for only those wells discussed in this report.

During most of the year the average depth to ground water in the alluvium ranges from 5 to 8 feet below land surface near the river and from 15 to 20 feet below land surface near the outer boundaries of the flood plain. On the terraces adjacent to the flood plain, however, the depth to water ranges from 20 to 40 feet below land surface. Ground-water levels in the basin-fill wells 2 to 4 miles distant from the flood plain range from 100 to more than 300 feet below land surface.

Wells on the flood plain which penetrate through the alluvium into the underlying basin fill indicate that the ground water in the basin fill is under artesian head. Water levels in this unit at well 0105 (fig. 1) were observed to be 6.1 feet higher than those in the alluvium, indicating flow from the basin fill to the overlying alluvium.

An estimate of the areal extent of saturated alluvium in the flood plain was made by Weist (1971, pl. 2). Additional ground-water-level data, well-log information, and field investigations of the surface contact between the basin fill and alluvium were used to define the revised saturated-alluvium boundaries, as shown in figure 1. The average width of the alluvium between these boundaries at the water table is about 5,500 feet, and the average saturated thickness of the alluvium is about 40 feet.

DETERMINATION OF AQUIFER CONSTANTS

STORAGE COEFFICIENT

"Storage coefficient [S] is defined as the volume of water that an aquifier releases from or takes into storage per unit surface area of aquifer per unit change in the component of head normal to that surface" (Todd, 1963, p. 31). For an unconfined aquifer, estimates of S may be obtained from the ratio $\Delta c/\Delta h$, where Δc is the change in soil-moisture content resulting from a ground-water altitude change, Δh (Stallman, 1967, p. 183).

Figure 3 shows moisture-content profiles and corresponding ground-water levels at well 1958 for October 26, 1964, and April 13, 1965. The change in moisture content (Δc) for this period, indicated by the shaded area in figure 3, includes only that change in the zone arbitrarily selected to represent the capillary zone. This zone extends from about 2 feet above the top of the maximum-recorded water table to about 1 foot below the minimum-recorded water table. Soil moisture above the capillary zone is assumed to be unaffected by changes in water level and, therefore, is not included in the storage-coefficient computation. Any change in soil-moisture content below the minimum water table, as indicated in figure 3, is due to errors in measurement of the moisture content. Studies by this writer indicate that measurement errors of moisture content in the capillary zone are generally less than 5 percent of the total moisture content of the zone.

The storage coefficient is generally considered to reflect soil-moisture change due to gravity drainage alone. In areas of phreatophyte cover, however, the

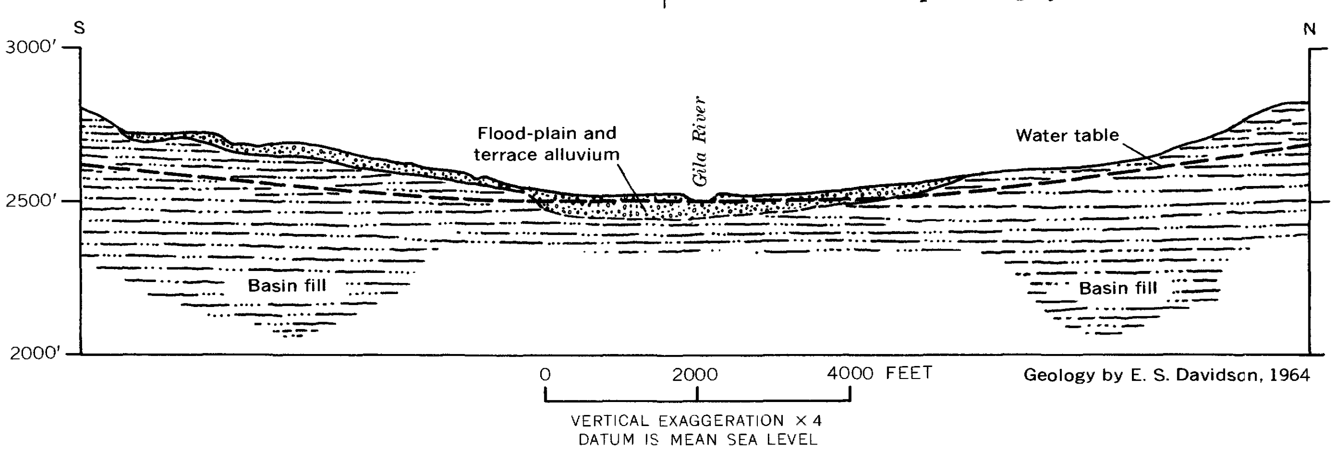


FIGURE 2.—Geologic section across the study area.

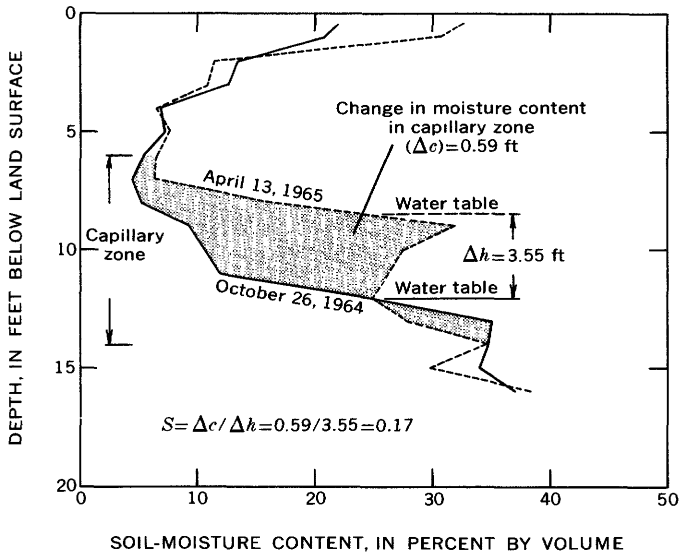


FIGURE 3.—Profiles of soil-moisture content at well 1958, showing increase in water level (Δh) and change in moisture content (Δc) in the capillary zone.

soil-moisture change results not only from gravity drainage, but also from transpiration.

In an attempt to evaluate the effect of transpiration on the storage-coefficient determination, values of S determined from soil-moisture and water-level data obtained before clearing phreatophytes from the area were compared with those values from data obtained after the phreatophytes had been cleared.

Table 1 shows calculated S values at eight wells in sections 1 through 7 for "before-clearing" conditions during the winter of the 1965 water year, and for "after-clearing" conditions during the winter of the 1969 water year. The average storage coefficient value, $\bar{S}$, for these wells decreased from 0.34 before phreatophyte clearing to 0.19 after clearing. This change in $\bar{S}$ suggests that transpiration by phreatophytes accounts for about 45 percent of the soil-moisture change in those regions having phreatophyte cover. The $\bar{S}$ for the total saturated thickness of the alluvium is probably less than 0.19 because some of the soil-moisture change in the cleared areas is caused by evaporation from bare ground and by transpiration from deep-rooted grasses that have replaced the phreatophytes.

An analysis of the spacial variability of S throughout the study area was made by determining an average storage coefficient, $\bar{S}_i$, at each of 34 well sites in the flood plain. Each $\bar{S}_i$ value was computed from the arithmetic average of the storage coefficients determined from n selected periods of water-level and corresponding moisture-content change. In order to analyze several periods without introducing the variability caused by phreatophyte clearing, the analysis was confined to the period, before clearing during water years 1964–67. Values of Δh and Δc were selected from relatively long periods (40 days or more) because the measurement error of these parameters is frequently a significant part of their total change for shorter periods.

Table 2 lists the average storage coefficient, ($\bar{S}_i$), number of determinations of S per site, (n), and standard deviation of $\bar{S}_i$, (σ), for each of the 34 sites analyzed. The weighted-average storage coefficient, ($\bar{S}_w$), was computed from the relation

$$\bar{S}_w = \sum_{i=1}^{34} \frac{n_i \bar{S}_i}{\Sigma n_i}.$$

The weighted-average storage coefficient for these 34 sites is 0.30 with a standard deviation of ±0.07, indicating that the spacial variability in S over the area is about 25 percent.

The pooled standard deviation of S_i is ±0.10, in-

TABLE 1.—*Comparison of storage-coefficient determinations before and after clearing phreatophytes from the flood plain*

[S, storage coefficient; $\bar{S}$, average storage coefficient]

Well No.	Before clearing; 1965 water year: Period	S	After clearing; 1969 water year: Period	S
0102	Nov. 10–Feb. 23	0.36	Nov. 11–Feb. 3	0.15
0103	Nov. 10–Feb. 24	.25	Nov. 11–Feb. 3	.08
0308	Nov. 10–Feb. 23	.39	Nov. 11–Feb. 3	.21
0311	Nov. 9–Feb. 23	.34	Nov. 12–Feb. 3	.27
0514	Nov. 10–Feb. 23	.37	Nov. 12–Feb. 4	.07
0517	Nov. 9–Feb. 23	.36	Nov. 13–Feb. 4	.30
0720	Nov. 10–Feb. 24	.28	Nov. 12–Feb. 4	.15
0723	Dec. 3–Mar. 17	.36	Nov. 13–Feb. 4	.27
	$\bar{S}$ =	0.34	$\bar{S}$ =	0.19

TABLE 2.—*Average storage coefficient* ($\bar{S}_i$) *at each well, number of determinations of* S *per well* (n), *and standard deviation of* $\bar{S}_i$ *per well* (σ), *for 34 wells in the flood plain of study area*

Well No.	$\bar{S}_i$	n	σ	Well No.	$\bar{S}_i$	n	σ
0102	0.32	6	0.10	1274	0.30	2	----
0105	.20	5	.10	1277	.33	4	0.19
0307	.30	2	----	1337	.26	2	----
0308	.36	9	.15	1338	.26	4	.10
0309	.37	2	----	1340	.25	5	.03
0311	.36	5	.10	1341	.27	5	.03
0312	.19	2	----	1544	.31	5	.08
0514	.46	4	.01	1750	.32	6	.18
0517	.31	4	.12	1753	.33	3	.06
0720	.26	7	.07	1956	.24	6	.05
0721	.12	2	----	1957	.20	4	.07
0723	.30	8	.09	1958	.18	7	.04
0926	.29	7	.09	1959	.32	6	.10
1131	.26	2	----	1960	.34	4	.10
1132	.27	5	.08	2162	.42	3	.15
1133	.43	2	----	2163	.32	2	----
1135	.38	5	.12	2164	.26	2	----

Weighted-average storage coefficient, $\bar{S}_w$	0.30
Standard deviation of $\bar{S}_w$	±.07
Confidence interval $\bar{S}_w$ at 95 percent level	±.02
Pooled standard deviation of $\bar{S}_i$	±.10

dicating that two-thirds of the storage coefficient determinations at a particular site range from 0.20 to 0.40. However, by assuming that the $\bar{S}_i$ in table 2 defines a t distribution, one can say with 95-percent confidence that the true weighted-average storage coefficient lies between 0.28 and 0.32. The season of the year, location of the wells in relation to the river, and the hysteretic effect, due to a rising versus a falling water table, do not appear to explain the relatively high standard deviation in S. Much of this variability may be real, however, because of the heterogeneity of these alluvial deposits.

Estimates of S were also obtained from relatively short term aquifer tests in the alluvium and basin fill. Analysis of a constant-discharge aquifer test in the alluvium near well 0105 (inset in fig. 1) gave a storage coefficient of 0.16. A similar test of shorter duration in the basin fill at well 0532 gave a storage coefficient of 0.0005. Discussion of the analytical techniques used to derive these values is presented in the next section.

On the basis of the preceding analysis, an average storage coefficient for the zone of water-level change of 0.30 is considered to be representative for the alluvium with phreatophyte cover. This value should be reduced to about 0.20 for alluvium after clearing. A storage coefficient of 0.15 is considered to be a reasonable value for the total thickness of saturated alluvium and should be used when predicting the behavior of wells or drains in the alluvium.

The single storage-coefficient determination of 0.0005 for the basin fill may differ substantially at other sites and is here considered to be only an approximate value.

AQUIFER TESTS OF THE ALLUVIUM AND BASIN FILL

Transmissivity (T) is defined as the volume rate of flow of water through a vertical strip of aquifer of unit width extending the full saturated thickness of the aquifer under a hydraulic gradient of 1 : 1. Diffusivity (D) is defined as the ratio T/S. Hydraulic conductivity (K) is defined as the volume rate of flow of water through a unit cross-sectional area under a hydraulic gradient of 1 : 1. The relation between transmissivity and hydraulic conductivity is $T = Km$, where m is the saturated thickness of the aquifer.

These aquifer constants are commonly derived from an analysis of the drawdown and recovery of water levels in a pumped well and adjacent observation wells. This method is restrictive, however, in that it defines these constants for only the region in which drawdown and recovery occurs. Furthermore, this technique is time consuming and expensive to perform. Various other techniques for determining aquifer constants have been developed which consider natural or suddenly induced changes in ground-water levels. Some of these techniques were employed in this analysis in an attempt to verify the aquifer constants derived from drawdown and recovery tests and to evaluate the areal variation in these constants.

Most of these analytical techniques assume that the aquifer is homogeneous and isotropic, and that the observation wells fully penetrate the aquifer. These assumptions, along with those constraints unique to each method, limit their application in the field and reduce the reliability of the results. However, where different methods give consistent results based on different sets of data, the results are considered to be representative.

The following sections discuss the various aquifer tests used to determine transmissivity, diffusivity, hydraulic conductivity, and storage coefficient. Determinations for the alluvium are presented first, followed by those for the basin fill.

ALLUVIUM

The aquifer constants of the alluvium were determined by analyzing the following aquifer conditions:

1. Drawdown and recovery in the aquifer due to constant discharge from a well penetrating the aquifer.
2. Propagation of a flood wave through the aquifer in response to fluctuations in river stage.
3. Cyclic fluctuations of ground-water levels due to sequential flood stages in the adjacent stream.
4. Decline in artesian head of a well using the constant-head "drain function."
5. Response of a well following an instantaneous charge of water (slug test).

AQUIFER TEST AT SECTION 1

By S. G. Brown

Estimates of transmissivity and storage coefficient were obtained from analysis of an aquifer test of the alluvium on the right bank of the Gila River at section 1.

Well C–1, the pumped well, was drilled near observation well 0105, and 16 other observation wells were drilled nearby. (See inset in fig. 1.) Well C–1 was 161 feet deep and 8 inches in diameter. It completely penetrated the flood-plain alluvium and 110 feet of the underlying basin-fill unit. The flood-plain alluvium at well C–1 is 51 feet thick and consists of sand, gravel, and some silt. The basin-fill unit that underlies the flood-plain alluvium consists mainly of silt and clay and some thin layers of fine to very fine sand. The water level is 12 feet below the land surface in observation wells that were completed in the alluvium near well C–1. Initially, casing in well C–1 was set to a depth of 81 feet, which is about 30 feet below the base of the alluvium. The well then tapped only water in the basin fill, causing the water level in the well to rise to about 6 feet higher than the water level in nearby wells that tapped only water in the alluvium.

A wire-wrapped well screen—6 inches in diameter and 30 feet long with slot openings of 0.060 inch—was set in the well with the bottom of the screen 49½ feet below the land surface and about 2 feet above the alluvium–basin-fill contact. A neoprene packer at the top of the screen fit snugly inside the 8-inch-diameter casing. Below the alluvium-basin-fill contact, the well was filled with sand, gravel, and silt and was capped with five sacks of quick-setting cement. The cement was allowed to harden, and the bottom of the 6-inch-diameter screen was set on the concrete. The 8-inch-diameter casing was withdrawn until about 1 foot extended below the packer at the top of the screen. The well was treated with a dispersing agent to aid the removal of clay-size particles from the aquifer during development, and the well was developed by bailing and surging with a surge block and, later, by surging with a pump. As the fine material was removed from the aquifer materials surrounding the screen during the development process, the rate of pumping was increased slowly until the well was discharging 0.50 cfs (cubic feet per second) or 225 gpm (gallons per minute). At this maximum rate of pumping, very fine sand and silt were drawn into the well; however, when the discharge of the well was regulated to 0.35 cfs (160 gpm), the sand problem was eliminated. The aquifer test was therefore made at a regulated discharge of 0.34 cfs (152 gpm).

The test was begun on April 16, 1964, at 2 p.m. and was continued until 1:41 a.m. on April 20, 1964, when the test was unintentionally terminated during an attempt to repair a leaky fuel line. The length of the pumping test was 5,021 minutes or 3.487 days.

Depth to water was measured systematically in 18 wells, including the pumped well. Nine observation wells were equipped with digital recorders. Of the nine wells, six were equipped with a 5-minute punch interval, and three were equipped with a 1-minute punch interval. Water levels in eight observation wells were measured frequently with a chalked steel tape, and the water level in the pumped well was measured with a mercury manometer.

Well discharge was regulated by a gate valve, measured with a 6-inch throat Parshall-type flume, and carried to the Gila River, about 1,500 feet to the south. Leakage from the plastic-film-lined ditch was negligible.

The data were analyzed by use of the modified nonequilibrium equation (Ferris and others, 1962, p. 100), in which

$$T = \frac{2.3Q}{4\pi\Delta s}, \tag{1}$$

where

T = transmissivity, in cubic feet per day per foot of width of aquifer;

Q = rate of discharge of the pumped well, in cubic feet per day; and

Δs = change in the drawdown over one log cycle of time, in feet.

$$S = 2.25\ T\ \frac{t_0}{r^2}, \tag{2}$$

where T is as previously defined, and

S = storage coefficient (dimensionless);

t_0 = time intercept, in days, where the plotted straight line intersects the zero-drawdown axis; and

r = distance, in feet, from the pumped well to the observation well.

In general, this modification of the Theis nonequilibrium equation is valid when the value u is ≤ 0.01 (Todd, 1963, p. 94), in which

$$u = \frac{r^2S}{4Tt}, \tag{3}$$

where

t = the time, in days, since pumping began; and r, S, and T are as previously defined.

Figure 4 and table 3 summarize the results of the analysis of the aquifer test for all observation wells in which u is ≤ 0.01 within the 3.5-day test period.

In all instances the calculated value of the storage coefficient fell well within the range of values characteristic of nonartesian, water-table conditions. The average transmissivity, calculated by eliminating data from all wells at which u was greater than 0.01, was somewhat lower and the storage coefficient somewhat higher than if the data for all wells had been used. The rate of drawdown per log cycle for wells where u was equal to or less than 0.01 is somewhat steeper than that for wells farther away with a u value greater than 0.01. Therefore, the calculated storage coefficient is greater for the observation wells

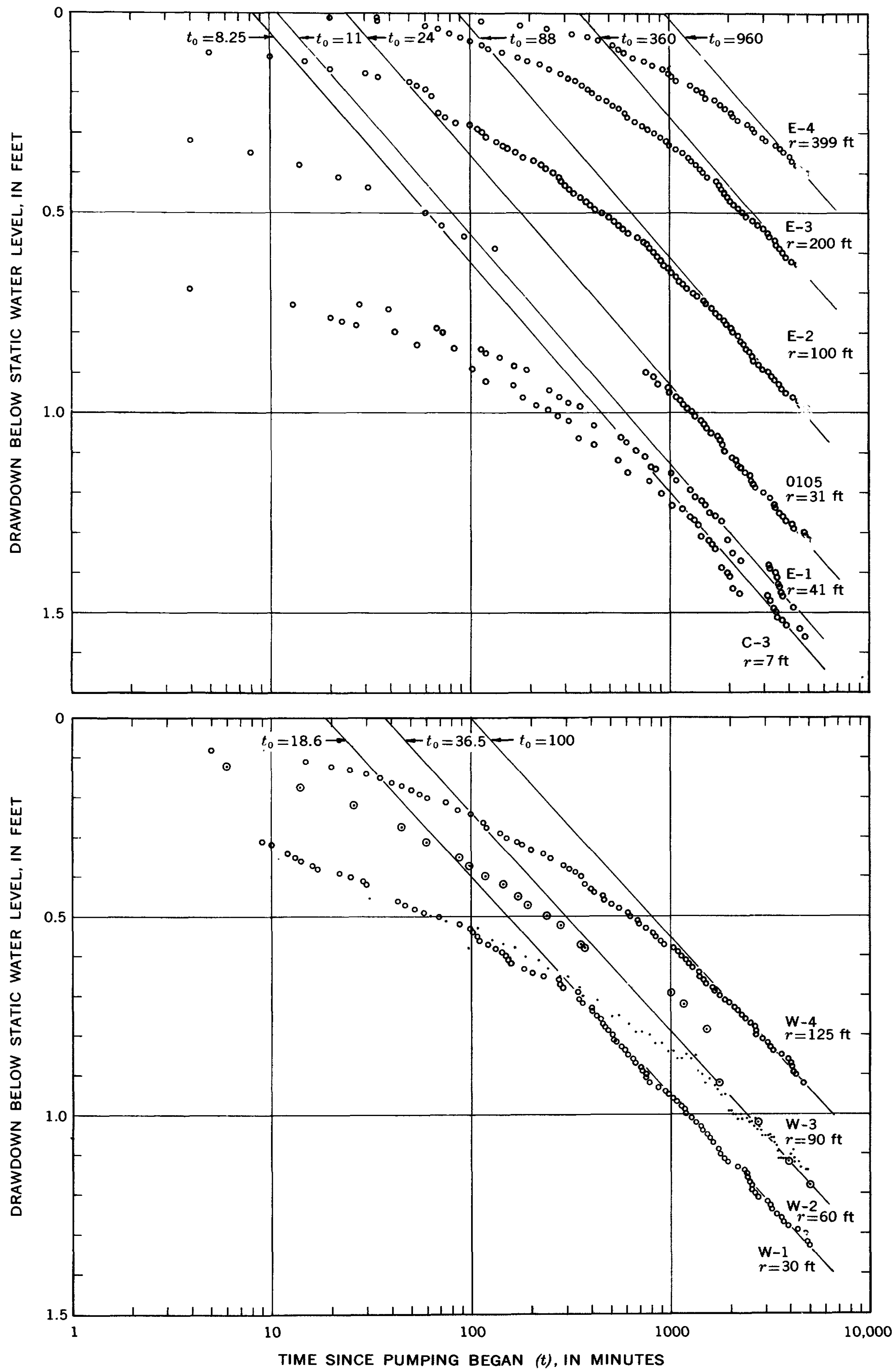
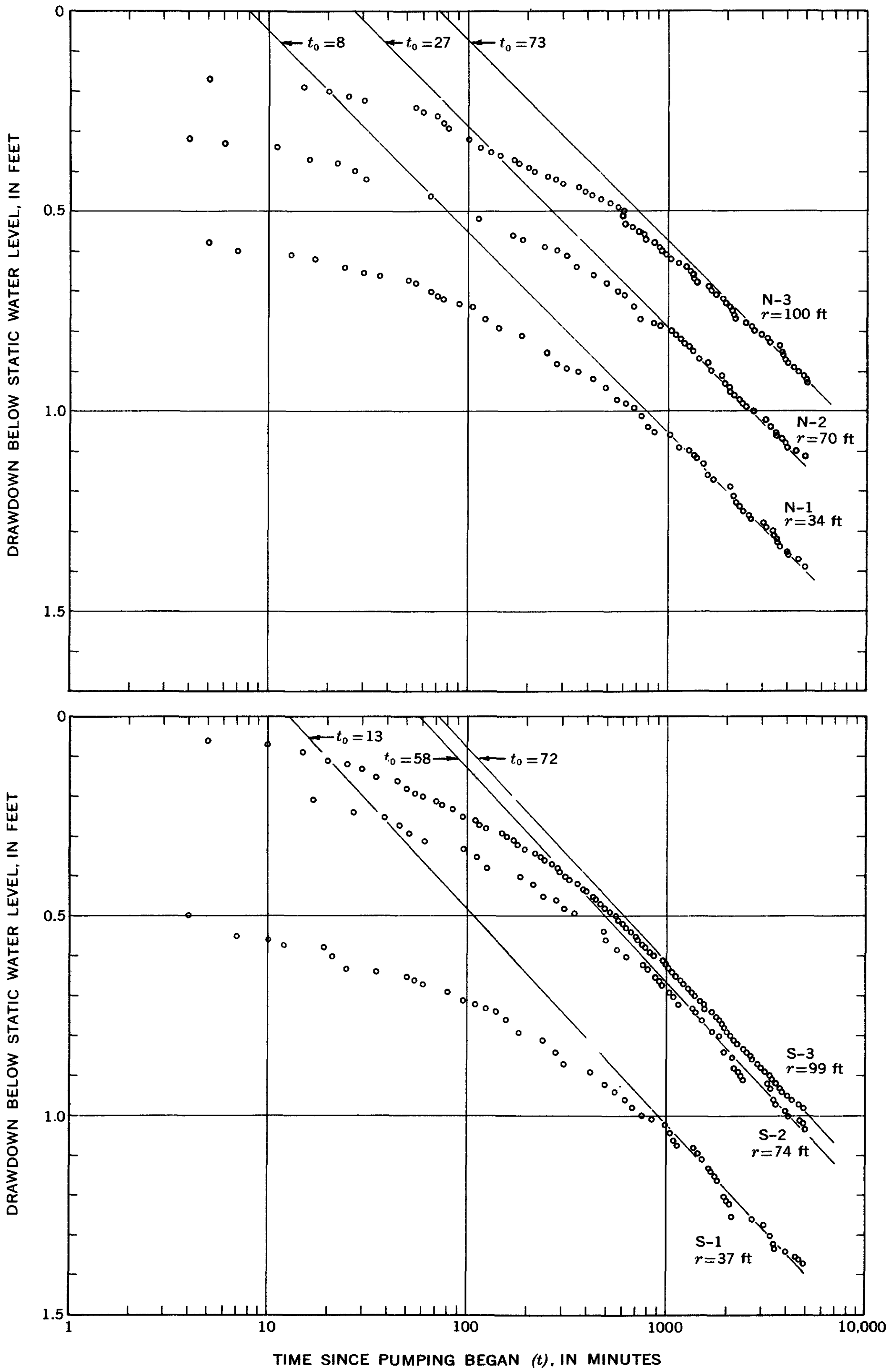


FIGURE 4.—Drawdown in observation wells near well C-1,



showing results of the modified Theis type analysis.

TABLE 3.—*Summary of transmissivities* (T) *and storage coefficients* (S) *at observation wells used in the aquifer tests at section 1*

Well No.	Distance from pumped well (ft)	T (cu ft per day per ft)	S
N–1	34	10,700	0.12
N–2	70	10,700	.09
N–3	100	10,700	.12
C–3	7	9,200	.23
0105	31	9,200	.36
E–1	41	9,200	.09
E–2	100	9,200	.12
S–1	37	9,900	.15
S–2	74	9,900	.16
S–3	99	9,900	.11
W–1	30	9,800	.31
W–2	60	9,800	.15
W–3	90	9,800	.05
Mean		9,800	0.16
Maximum		10,700	.36
Minimum		9,200	.05

near the pumped well because the aquifer had been draining for a longer time.

The range of T determined at this site—from 10,700 cu ft per day per ft (cubic feet per day per foot) (80,000 gpd per ft (gallons per day per foot)) to 9,200 cu ft per day per ft (69,000 gpd per ft)—is reasonable and probably can be accounted for by local variations within the aquifer. The range of S—0.36 to 0.05—is much greater in relation to the average S, but is characteristic of water-table (nonartesian) conditions. The S calculation is sensitive to t_0, which depends on the slope and placement of the straight line through the drawdown data. Some variation can be expected because of lithologic variations within the aquifer.

The mean transmissivity of 9,800 cu ft per day per ft (73,000 gpd per ft) and the mean storage coefficient of 0.16 can be used with confidence to calculate ground-water underflow and change in ground-water storage in the area sampled by the aquifer test.

FLOOD-WAVE-PROPAGATION ANALYSIS

A comprehensive investigation of the diffusivity (D) of the alluvial material underlying the flood plain and lower terrace was made by using a technique that describes the response of an aquifer to the propagation through it of a flood wave originating from a hydraulically connected stream. Pinder, Bredehoeft, and Cooper (1969) developed the technique based on the theory of heat flow and adapted it to the digital computer. The computer calculates a synthetic flood hydrograph (type curve) of ground-water heads at a distance x_0 from the river, using as input: (1) the "reference hydrograph" of ground-water heads at a distance x_r from the river, where $x_r < x_0$; (2) the aquifer width, defined as the distance between the point of the reference hydrograph and the boundary of saturated alluvium; and (3) an assumed diffusivity for the aquifer. A visual comparison between the computed hydrograph and corresponding observed hydrograph is provided by a plotting routine included in the computer program. Additional hydrographs are generated for selected diffusivities until a reasonable match is obtained between the computed and observed hydrographs.

The assumptions underlying this technique are that the aquifer is confined, homogeneous, and isotropic and is bounded on one side by impermeable material and on the other side by a hydraulically connected straight stream which fully penetrates the aquifer.

Diffusivity determinations were made at sections 1, 3, 5, 7, 9, 11, and 13. (See fig. 1.) There are three wells on each side of the river at each section. Except for those in section 1, the wells in these sections are in a line approximately normal to the river. All wells extend below the water table, but only a few penetrate the full thickness of the alluvium. The wells were cased with 4½-inch-diameter lightweight steel tubing into which a 4¼-inch-diameter plastic liner was inserted. The casings were not perforated, but were left open at the bottom. Digital recorders recorded water levels at 1-hour intervals at each well.

At each section the well adjacent to the river is designated "river well," the outermost well is designated "terrace well," and the middle well, between the river and terrace wells, is designated "flood-plain well," as indicated in figure 5. Diffusivity was determined by using one of the three combinations of reference and observation wells.

Figure 6 compares two computed hydrographs with the observed hydrograph of the July 16 to September 24, 1967, flood period for observation well 0307 (terrace well), using well 0308 as a reference well (flood-plain well). The diffusivity corresponding to the computed hydrograph of best fit in figure 6 is 122,000 cu ft per day per ft (916,000 gpd per ft).

Seldom could the computed hydrograph be matched exactly with the observed one. At several well sites, large changes in the diffusivity determination changed the computed hydrograph only slightly. Occasionally, a reasonable match of the computed with the observed hydrograph could be obtained by changing the aquifer width. No reason-

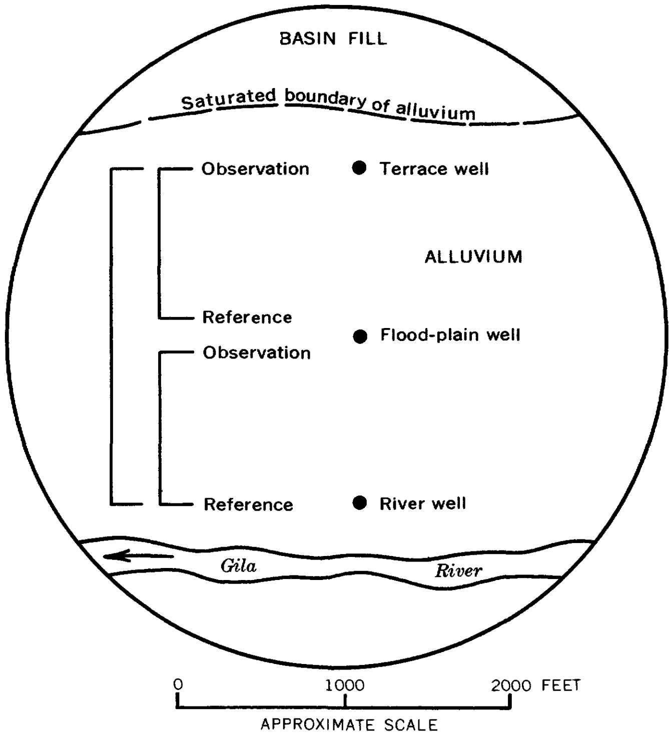


FIGURE 5.—Position of terrace, flood-plain, and river wells, and their designation when used as reference and observation wells in the flood-wave propagation analysis.

able match was obtained at a few sites, regardless of the aquifer width or diffusivity used.

An attempt to verify the diffusivity determinations was made by analyzing two or three different flood peaks at each site. For several sites, however, water-level data were available for only one flood peak. The flood periods analyzed included August 1 to September 30, 1963; March 8 to April 21, 1966; July 16 to September 24, 1967; and December 16, 1967, to February 10, 1968. Table 4 summarizes the results of this analysis and groups the data according to the combination of terrace, flood-plain, and river wells used.

The diffusivities in table 4 are consistent and of a reasonable magnitude at some sites, but are unrealistically low or high at others. The large discrepancy in D values at a given site may be attributed to (1) slow response of the reference well or observation well, as a result of partial plugging; (2) a change of diffusivity, due to a change in saturated thickness; and (3) use of a partial-penetrating standpipe as a piezometer. In some instances, those values obtained by using the river well as a reference well may reflect vertical-flow components and, thus, represent erroneous results.

The extreme variability in D values between sites

TABLE 4.—*Diffusivities at the sites of selected wells in the alluvium*

Reference well	Observation well	Water year of flood	Flood amplitude at reference well (ft)	Diffusivity (cu ft per day per ft)
		River–flood-plain wells		
0103	0102	1963	2.95	26,700
		1967	6.60	2,251,000
0104	0105	1963	2.75	21,400
		1968	2.95	147,000
0309	0308	1966	4.30	245,000
		1967	6.00	338,000
		1968	6.40	245,000
0310	0311	1963	1.90	96,200
		1967	5.30	9,625,000
		1968	6.55	9,625,000
0515	0514	1963	3.10	40,000
		1967	5.90	250,000
		1968	5.45	444,000
0516	1517	1963	3.15	78,600
0721	0720	1963	2.65	17,100
		1968	[1]4.5	107,000
0928	0929	1963	3.10	719,000
1133	1132	1966	3.90	722,000
1339	1338	1966	5.30	225,000
		River–terrace wells		
0103	0101	1963	2.95	44,400
		1967	6.60	379,000
		1968	3.75	327,000
0104	0106	1963	2.75	89,800
		1968	2.95	561,000
0309	0307	1966	4.30	305,000
		1967	6.00	197,000
		1968	6.40	197,000
0310	0312	1963	1.90	57,700
		1967	5.30	497,000
		1968	6.55	230,000
0515	0513	1963	3.10	44,400
		1967	5.90	73,800
		1968	5.45	88,200
0516	0518	1963	3.15	103,000
0721	0719	1963	2.65	1,070
		1968	[1]4.5	107
0722	0724	1968	[1]4.5	5,880
0927	0926	1966	3.90	17,600
0928	0930	1963	3.10	123,000
		1967	6.70	1,361,000
		1968	4.90	766,000
1133	1131	1966	3.90	112,000
1339	1337	1966	5.30	64,200
		Flood-plain–terrace wells		
0102	0101	1967	6.85	94,600
0308	0307	1967	4.95	122,000
		1968	5.70	245,000
0514	0513	1967	5.15	81,800
		1968	4.70	62,600
0517	0518	1967	5.40	358,000
		1968	4.60	160,000
0723	0724	1968	2.00	36,900
1132	1131	1966	4.00	40,100
1338	1337	1966	4.10	64,200

[1] Estimated.

may be due to the above-mentioned problems and may reflect the effect of using incorrect aquifer widths. Some of this variability probably reflects the true areal distribution of D throughout the area; however, a wide variation in the values at a given site precludes the use of these results to evaluate their distribution.

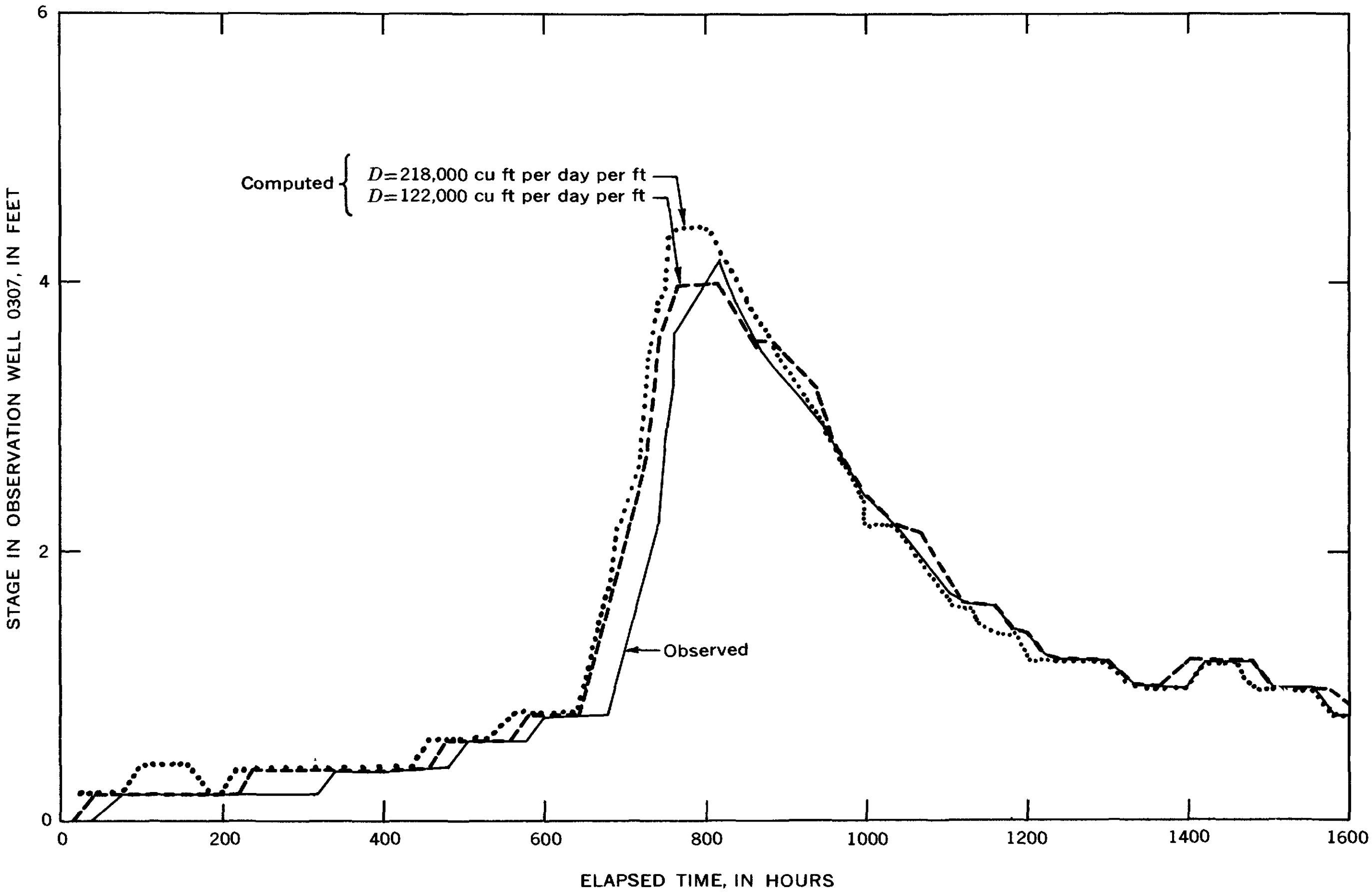


FIGURE 6.—Observed and computed hydrographs at observation well 0307, July 16 to September 24, 1967. Computed graphs were derived from flood-wave propagation analysis, using well 0308 as reference well.

Figure 7 shows a plot of flood amplitude at the reference well versus the corresponding diffusivity at the obervation well. The plot indicates a slight direct correlation that may reflect a change in diffusivity, due to a corresponding change in saturated thickness.

An estimate of the median diffusivity for each combination of wells listed in table 4 was made by plotting a frequency distribution of diffusivities corresponding to a given combination of wells and flood period. All diffusivities for a given well combination were delineated by flood period and assigned probabilities by using the relation

$$P\,(D_m \geq \frac{m}{n+1}\,, \qquad (4)$$

where D_m is the mth largest diffusivity value, and n is the total number of values. The largest diffusivity was, therefore, assigned $m = 1$; the second largest, $m = 2$, and so forth. A log-normal probability plot with D_m on the logarithmic scale against $P(D_m)$ on the probability scale was then made, as illustrated in figure 8. The diffusivity at the 50-percent probability level was selected from the line averaging the data points on this plot. (Outlying points were not considered in drawing this line.) Similar plots were prepared for each of the three combinations of wells for each of the four flood periods. The diffusivities corresponding to the 50-percent probability level for each of these plots is given in table 5.

The weighted-mean diffusivity, $\bar{D}_{50}$, was computed for each well combination from the relation:

$$D_{50} = \frac{\sum_{i=1}^{k} n_i\, h_i\, D_{50,i}}{\sum_{i=1}^{k} n_i\, h_i}\,, \qquad (5)$$

where k is the number of flood periods, n_i is the number of diffusivity determinations used to define the frequency distribution plot for the ith flood period, h_i is the average change in water level at the observation wells associated with the ith flood period, and $D_{50,i}$ is the 50-percent-diffusivity value selected from

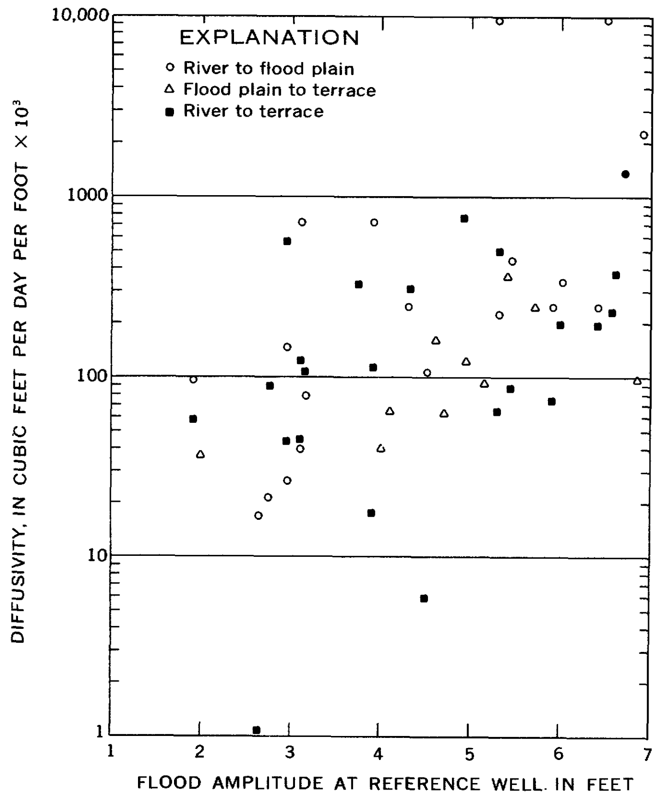


FIGURE 7.—Relation between flood amplitude at the reference wells and computed diffusivity at the observation wells. (Data are from table 4.)

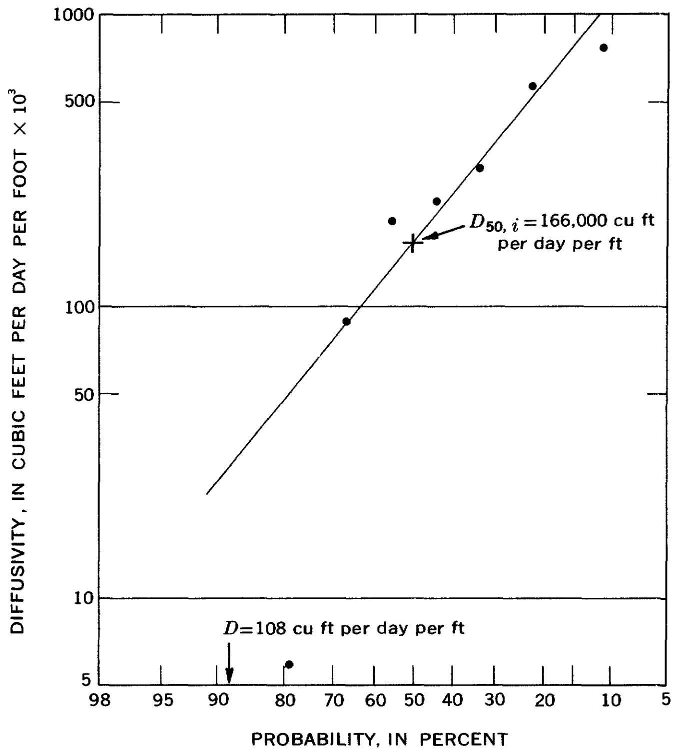


FIGURE 8.—Frequency distribution of diffusivity, D, derived from the flood-wave propagation analysis by using the combination of river and terrace wells at eight sites during the 1968 flood period. (Data from table 4.)

the frequency distribution plot for the ith flood period. This relation gives the greatest weight to those median diffusivities derived from the greatest number of determinations and highest flood peaks.

Table 5 shows that the weighted-mean diffusivity is greater for the combination of river and flood-plain wells than for the flood-plain and terrace wells, which suggests that D decreases with the increase in distance from the river.

Selection of that $\bar{D}_{50}$ value in table 5 which best represents the alluvium is questionable because none of the wells used in the analysis fully meet the well-construction requirements desired for this type of test—that is, wells that penetrate the full depth of the alluvium and that are screened. An observation well slow to respond because of partial plugging will result in a low estimate of D. Conversely, a reference well slow to respond will give an estimate of D which is too high. Some of the shallow river wells may respond too rapidly because of vertical-flow components, and the resulting estimates of D are, thus, too low. Despite these problems, the magnitude and range of the $\bar{D}_{50}$ values in table 5 are thought to be reasonable.

The weighted-mean diffusivity of 185,000 cu ft

TABLE 5.—*Summary of alluvium diffusivities derived from flood-wave propagation analysis, using various combinations of river, flood-plain, and terrace wells*

Flood period	Number of diffusivity determinations (n)	Average flood amplitude at observation well (h, in ft)	Diffusivity at 50-percent probability level $D \times 10^3$ (cu ft per day per ft)
	River–flood-plain wells		
1963	7	1.5	42.8
1966	3	3.9	34.2
1967	5	4.8	337
1968	6	4.2	112
Weighted-mean diffusivity, $\bar{D}_{50}$			215
	Flood-plain–terrace wells		
1963	----	----	----
1966	2	1.8	50.3
1967	4	3.4	134
1968	4	3.0	96.3
Weighted-mean diffusivity, $\bar{D}_{50}$			108
	River–terrace wells		
1963	7	1.0	64.2
1966	4	2.1	80.2
1967	5	3.6	305
1968	8	3.0	166
Weighted-mean diffusivity, $\bar{D}_{50}$			185

per day per ft derived from the combination of river and terrace wells is considered to be representative of the total width of saturated alluvium. If the storage coefficient is 0.15, the corresponding weighted-mean transmissivity is 28,000 cu ft per day per ft (210,000 gpd per ft).

These results are considered to be significant because they were derived from different flood periods and from several observation sites dispersed over a large part of the study area.

ANALYSIS OF CYCLIC FLUCTUATIONS OF WATER LEVELS

Estimates of diffusivity at several sites in the project area were obtained by analyzing the cyclic fluctuation of water levels in response to sinusoidal changes in stage of the Gila River. Ferris (Ferris and others, 1962, p. 132–135) showed that diffusivity may be determined from the relation

$$D = \frac{0.189\pi(\Delta x)^2}{t_0}, \tag{6}$$

where D is diffusivity, t_0 is the average period of stage fluctuation, and Δx is the distance corresponding to one log cycle of a plot of the stage ratio, $s_r/2s_0$, on logarithmic coordinates versus distance, x, between the river and a given well on rectangular coordinates. The values, s_r and s_0 are the total range of the ground-water stage and one-half the range of the surface-water stage, respectively. (See inset in fig. 9.)

This solution assumes an isotropic semi-infinite artesian aquifer of uniform thickness. A change in water storage within the aquifer is assumed to occur instantaneously with a change in pressure in the aquifer. Further, the bounding stream is assumed to fully penetrate the aquifer. Because the saturated alluvium of the study area is not under artesian conditions and the Gila River only partially penetrates the aquifer, application of this technique is limited to those observation wells sufficiently distant from the stream to be unaffected by vertical components of flow within the aquifer. Also, the cyclic fluctuations in the observation wells must be of fairly uniform duration, and their amplitude must be small relative to the saturated thickness of the alluvium.

Figure 9 shows a plot of x versus the average stage ratio, $s_r/2s_0$, for nine flood-plain wells. The average stage ratio for each well is based on four flood cycles that occurred sequentially from July 14 to August 10, 1964. Table 6 shows an example of the tabulations used to determine the average stage ratio and average period of flood-stage fluctuation for well 0105. The average stage ratio of 0.047 is an average of the rise and fall over the four flood cycles and the average period of flood-stage fluctuation of 7 days is the average duration of each of these four flood cycles. This average t_0 of 7 days also applies

TABLE 6.—*Example of tabulations used to define the average period of flood-stage fluctuation,* t_0, *and the average stage ratio,* $s_r/2s_0$, *at well 0105 for four flood cycles during July and August 1964*

	1964 flood cycles							
	July 14–20		July 20–24		July 24–30		July 30–Aug. 10	
	Rise	Fall	Rise	Fall	Rise	Fall	Rise	Fall
s_r (ft)	0.14	0.04	0.03	0.06	0.07	0.05	0.26	0.23
$2s_0$ (ft)	3.16	2.80	1.06	1.04	2.16	1.90	2.80	2.90
$s_r/2s_0$*	.044	.014	.028	.058	.032	.026	.092	.080
t_0† (days)	6		4		6		11	

*Average $s_r/2s_0 = 0.047$.
† Average $t_0 = 7$ days.

to the other wells in this analysis because these same flood cycles were used in determining their stage ratios. The straight line averaging the points in figure 9 gives a Δx value of 1,460 feet per log cycle of the stage ratio. The resulting diffusivity, computed by using equation 6, is 179,000 cu ft per day per ft. If the storage coefficient of the alluvium is 0.15, the transmissivity is 27,000 cu ft per day per ft (200,000 gpd per ft).

These results compare favorably with those obtained using the river wells in the flood-wave propagation analysis, but are considered as only approximate because of the fairly wide scatter of points in the plot of figure 9. The fact that the flood cycles used in this analysis are not of uniform duration and magnitude and that vertical-flow components may exist at some of the wells near the river probably accounts for much of this scatter.

CONSTANT-HEAD DRAIN-FUNCTION ANALYSIS

Stallman (Ferris and others, 1962, p. 126–131) showed that when a change in artesian head occurs as a result of a sudden change in river stage, diffusivity, D, can be estimated from the relation

$$D = \frac{x^2}{4tu^2}, \tag{7}$$

where x is the distance between the river and a point at which the decline in the head of the aquifer, s, is observed; t is the time from a sudden change in the river stage; and $u = x^2S/4Tt$, where S and T are as previously defined. Equation 7 assumes that:

1. The aquifer is confined, homogeneous, isotropic, and semi-infinite in extent and is bounded on one side by a straight stream which fully penetrates the aquifer,
2. The head in the stream suddenly changes at time $t = 0$,
3. The direction of ground-water flow is perpendicular to the direction of streamflow,
4. The change in the rate of discharge from the

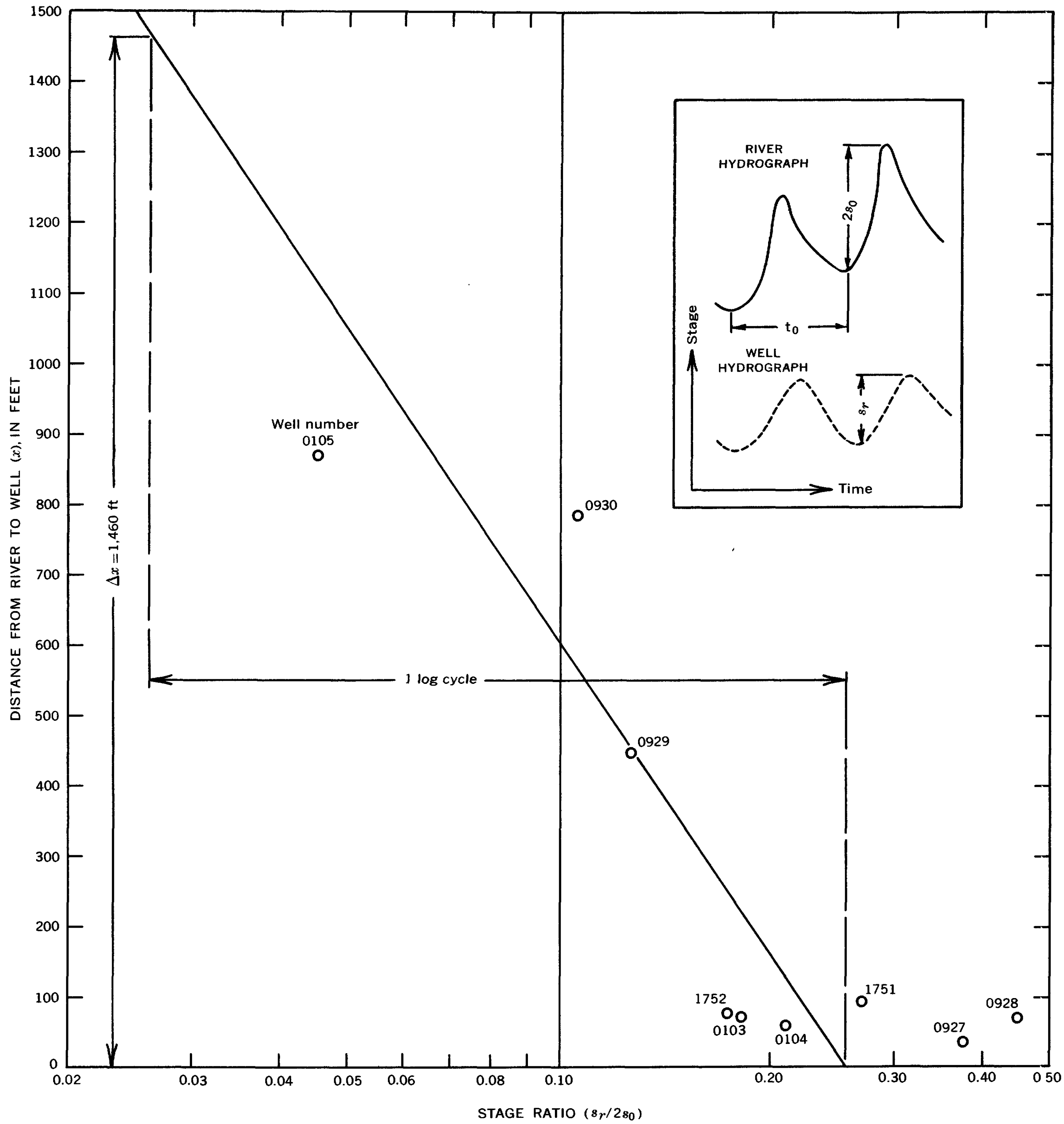


FIGURE 9.—Relation between stage ratio, $s_r/2s_0$, and distance, x, from the river to the well for nine wells in the flood plain, where $s_r/8s_0$, is an average of four flood cycles during July and August 1964.

aquifer is derived from changes in storage by drainage after $t = 0$.

A solution for equation 7 is obtained by first matching the drain-function type curve of the relation between $\phi(u)_h$ and u^2 (Ferris and others, 1962, fig. 31) to a logarithmic data plot of x^2/t versus s. Any value of u^2 (usually $u^2 = 1.0$ is selected for convenience of computation) from the type curve and corresponding value of x^2/t from the data plot may then be substituted in equation 7 to solve for D.

Figure 10 shows data plots of x^2/t versus s for the left-bank and right-bank wells in section 5 of the Gila River, at $t = 2.4$ days following a fairly sudden rise in river stage on July 17, 1967. Included in the figure are the drain-function type curves adjusted to the position that best fits the data points, the corresponding match-point values of u^2 and x^2/t, and computed diffusivities. If the storage coefficient is 0.15, the corresponding transmissivities are 16,000 cu ft per day per ft (120,000 gpd per ft) for the left bank and 56,000 cu ft per day per ft (420,000 gpd per ft) for the right bank in section 5.

A check on these results was made by applying this same analysis to the recovery of five wells in sections 1 and 5 following the fairly rapid rises in river stage. For each well, a plot was made of s versus $1/t$ defined by the ground-water recessions following selected river rises. As before, the drain-function type curve was matched to each plot, values of u^2 and corresponding t selected from each plot, and substituted in equation 7 to solve for D. The results of these determinations are listed in table 7.

The diffusivities in table 7 give a median value of 16,000 cu ft per day per ft when their frequency distribution is plotted on log-normal probability paper. For a storage coefficient of 0.15, the corresponding transmissivity is 2,400 cu ft per day per ft (18,000 gpd per ft).

The wide range in diffusivities in table 7 makes these results questionable. The assumptions of a semi-infinite artesian aquifer, an instantaneous change in river stage, and a river which penetrates the entire thickness of the alluvium are not fully satisfied in this problem. The particularly low diffusivities for those wells near the river (wells 0103 and 0104) may not be representative because of possible vertical flow in that region of the alluvium.

SLUG TESTS

Slug tests described by Cooper, Bredehoeft, and Papadopulos (1967), which induce a sudden change

TABLE 7.—*Diffusivities for five wells at sections 1 and 5 derived from constant-head drain-function analysis*

Well No.	Distance from stream to well (ft)	Flood period	Diffusivity (cu ft per day per ft)
0103	81	July 1964	4,500
0104	61	July 1964	3,420
0105	890	July 1964	23,900
0105	890	Sept. 1966	103,000
0514	280	July 1964	11,600
0514	280	July 1967	7,490
0517	744	July 1964	155,000

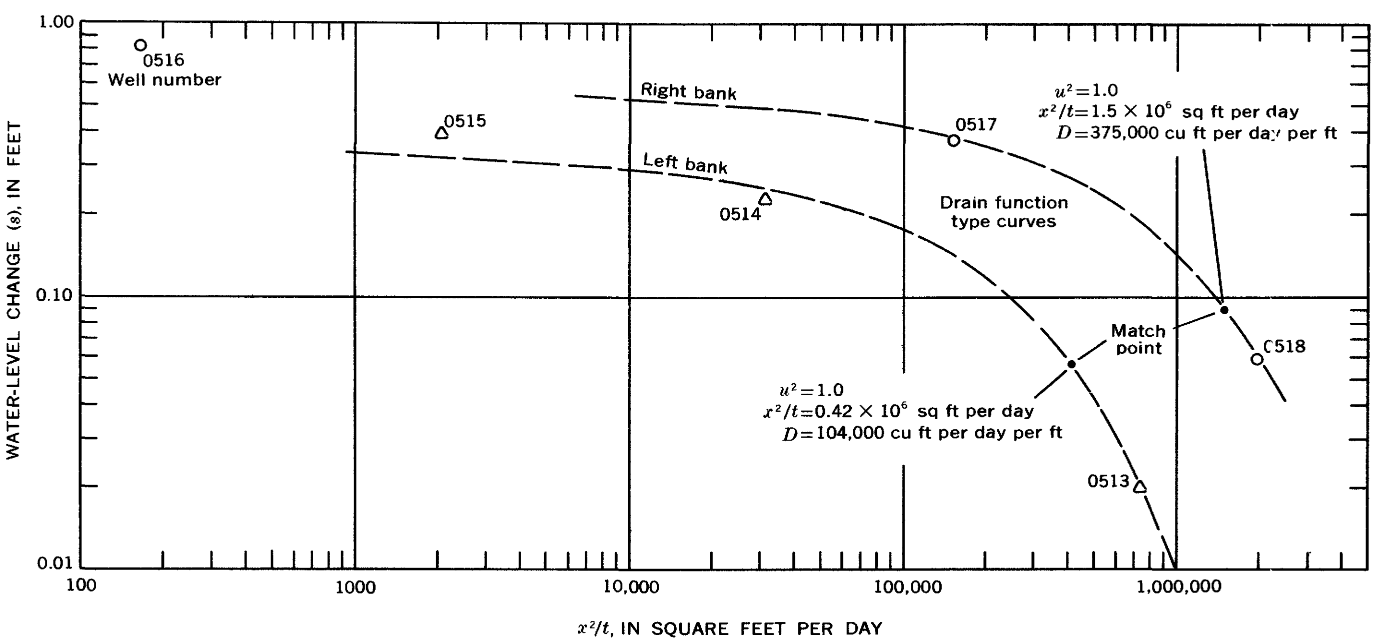


FIGURE 10.—Plots of x^2/t versus s, for the left-bank and right-bank wells in section 5, at $t = 2.4$ days following a rise in river stage on July 17, 1967.

in the water level of the aquifer, were performed on eight wells in the alluvium in a further attempt to define the approximate magnitude of transmissivity and its variation throughout the study area. Transmissivities were determined from an analysis of the recovery of the wells following instantaneous submergence or removal of a mass (slug) of known volume. The resulting transmissivities ranged from 0.4 to 150 cu ft per day per ft (3 to 1,100 gpd per ft).

These transmissivities are substantially lower than those values determined by the other methods previously discussed. The wells used in this analysis are cased to the bottom of the hole with no screening or perforations in the well casing. As a result, only a relatively small volume of the aquifer is stressed during these short-term tests, and movement of water into and out of the well is primarily vertical rather than horizontal. Therefore, these determinations are not considered to be representative, and this method of analysis is probably not applicable to most wells in the study area.

BASIN FILL

Estimates of the transmissivity and the storage coefficient of the basin fill were obtained from an analysis of the drawdown and recovery of the ground-water level following a constant discharge from the aquifer. Estimates of diffusivity were obtained from an analysis of the seasonal ground-water-level recessions. Estimates of the upward vertical flow from the basin fill to the overlying alluvium were obtained from an analysis of geothermal gradients in the basin fill.

AQUIFER TESTS AT WELLS 0531 AND 0532

An estimate of transmissivity and storage coefficient for the basin fill was obtained from an aquifer test performed at wells 0531 and 0532 in the basin fill outside the boundaries of the alluvium. (See fig. 1.) The pumped well (0531) is 6 inches in diameter and about 200 feet deep. The observation well (0532), 62 feet away from the pumped well, is 16 inches in diameter and 160 feet deep. Whether the bottom of these wells are perforated, finished with a screen, or are open ended is unknown, as they were drilled for stock and domestic purposes before inception of this study.

Figure 11 shows a time-drawdown plot of the water level in observation well 0532 during 24 hours of pumping at an average rate of $Q = 0.0194$ cfs (8.7 gpm). An estimate of transmissivity was made from this data plot by using Theis' nonequilibrium equation (Theis, 1935), in which

$$T = \frac{Q}{4\pi s} W(u), \tag{8}$$

where T and Q, are as defined in equation 1, s is the drawdown in the observation well, and $W(u)$ is the well function for constant discharge (or recharge) of the aquifer without vertical leakage. A type curve of the relation $W(u)$ versus u (Ferris and others, 1962, fig. 23) was matched to the data points in figure 11, and match-point values for u from the type curve and s from the data plot were substituted in equation 8 to give a transmissivity of 15.5 cu ft per day per ft (116 gpd per ft). Equation 3 was used to compute a storage coefficient of $S = 0.00048$ for the basin fill.

Discharge from the pumped well during this test varied by about 6 percent. This is the probable cause for the deviation of data points from the type curve in figure 11.

A check on this estimate of T was made by applying equation 1 to the data points defining the rate of recovery of the water level in observation well 0532. A convenient procedure for obtaining Δs in equation 1 is to plot recovery, s', on rectangular coordinates against t/t' on logarithmic coordinates, where t is the time since pumping started, and t' is the time since pumping stopped. After t' becomes sufficiently large, the data points approach a straight line, and $\Delta s'$ may be selected over one log cycle of t/t'.

Figure 12 shows a data plot of the recovery of well 0532. A straight-line projection of the data points over one log cycle of t/t' defines $\Delta s' = 23$ feet. The resulting transmissivity computed from equation 1 is 13.4 cu ft per day per ft (100 gpd per ft).

The relatively close agreement between the drawdown and recovery determinations of transmissivity at this well (16-percent difference) suggests that a value of about 15 cu ft per day per ft (110 gpd per ft) may be representative for the basin fill at this test site.

ANALYSIS OF WATER-LEVEL RECESSIONS

Rorabaugh (1960) described a method for determining diffusivity from the seasonal recession of the water table. This method of analysis is applicable when sufficient time has elapsed following recharge to the aquifer for the ground-water profile to reach a stable condition. The profile is considered to be stable when the water table falls exponentially with time. For this condition, diffusivity, D, may be expressed as

$$D = 4\frac{a^2}{\pi^2} 2.3 \log\left(\frac{h_1}{h_2}\right) / (t_2 - t_1), \tag{9}$$

where a is the distance from the ground-water di-

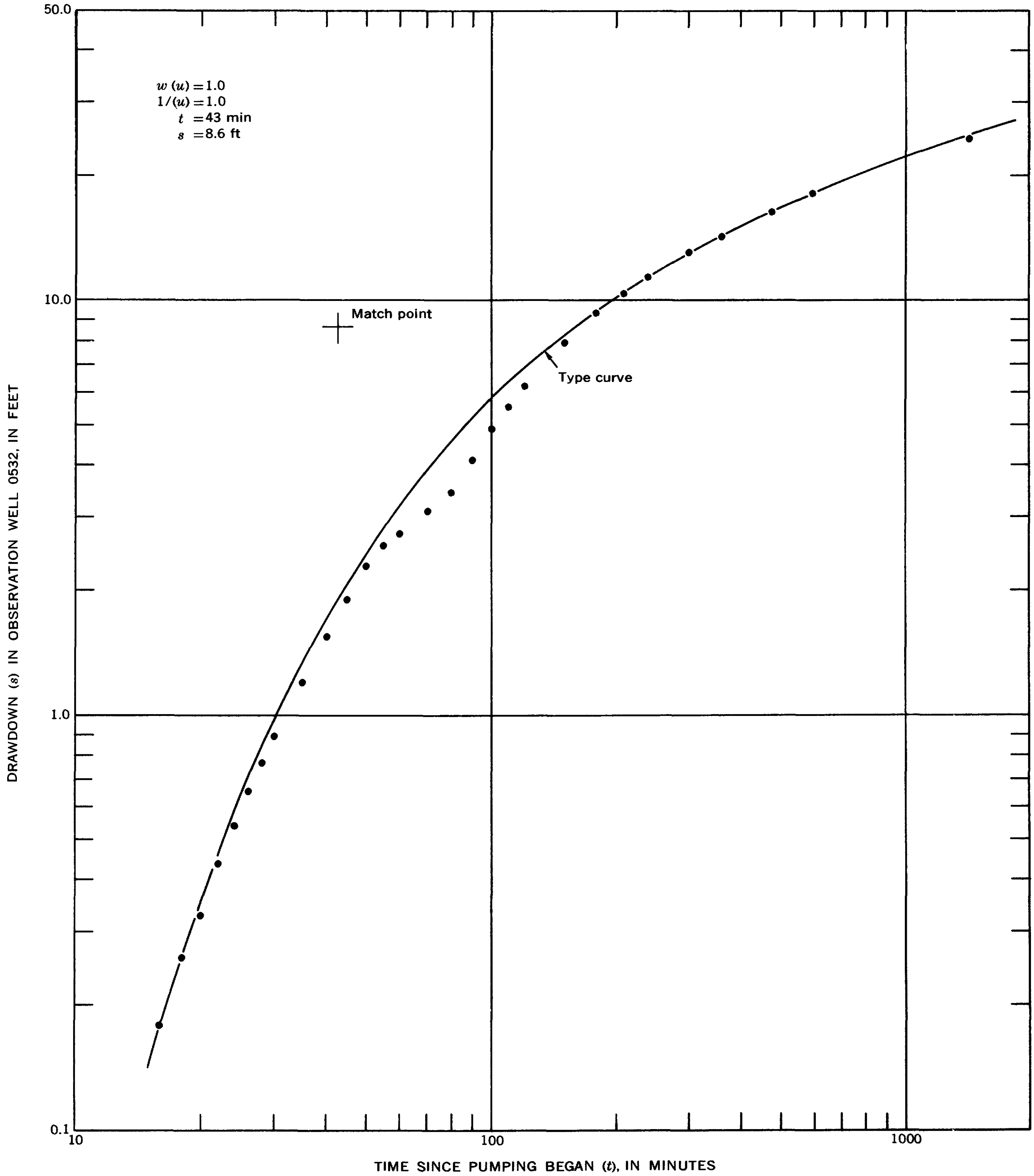


FIGURE 11.—Time-drawdown plot and matching nonleaky type curve for aquifer test of basin fill at well 0532.

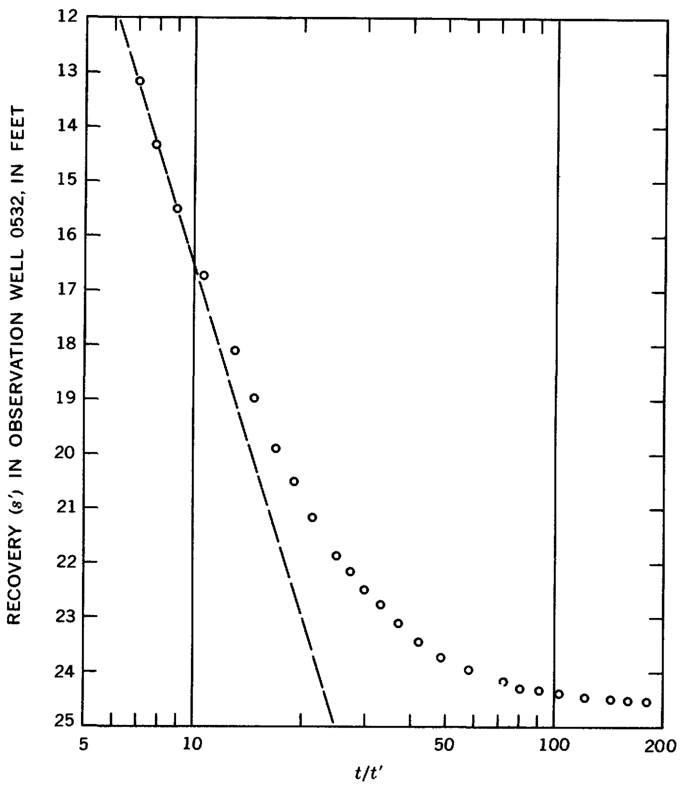


FIGURE 12.—Time-recovery plot of aquifer test in basin fill at well 0532.

vide to the discharge boundary, and h_1 and h_2 are the water-table heads at any point in the aquifer at times t_1 and t_2, respectively, following a sudden recharge at time $t = 0$. Equation 9 assumes that the water table is horizontal prior to recharge and that the water-table heads represent the rise in water above this initial horizontal level.

To obtain a solution for D, a semilog plot is made of the water-level recession at any given well, and the time interval $(t_2 - t_1)$, in days, over one log cycle of head is selected from the straight-line portion of the recession. This time interval may then be substituted for $(t_2 - t_1)/\log(h_1/h_2)$ in equation 9.

For this analysis, a was taken as the width of the basin fill between its contact with the saturated alluvium, near the edge of the flood plain, and its contact with the igneous and metamorphic rock, approximately 5 miles from the flood plain. (See fig. 1.)

Equation 9 was used to compute diffusivity from selected water-level recessions at each of five wells located in the basin fill. A description of these wells is summarized in table 8. Figure 13 shows graphs of the recessions for each well. The resulting diffusivities computed for each recession are listed in table 9. The median diffusivity for these five wells, ob-

TABLE 8.—*Description of basin-fill wells used in water-level recession analysis*

Well No.	Total depth (ft)	Depth cased (ft)	Diameter (in.)	Type of finish
0524 [1]	300	Unknown	8	Unknown.
0530 [1]	66	do	16	Do.
0725	92	44	1	Well point.
0730 [1]	226	226	6	Casing perforated in lower 50 ft.
0731 [1]	340	Unknown	6	Unknown.

[1] Occasional pumping for stock use.

TABLE 9.—*Diffusivities at five wells in the basin fill computed from selected water-level recessions, using equation 9*

Well No.	Basin-fill width (ft)	Recession period	Time interval per log cycle (days)	Diffusivity (cu ft per day per ft)
0524	27,000	Aug. 1967–May 1968 [1]	425	1,600,000
0530	24,000	Apr. 1965–Oct. 1965 [1]	425	[2] 1,070,000
0530	24,000	Apr. 1969–Aug. 1969 [1]	565	
0725	24,000	June 1968–Nov. 1968	385	[2] 1,280,000
0725	24,000	Feb. 1969–May 1969	450	
0730	27,000	Oct. 1968–June 1969 [1]	470	1,440,000
0731	28,000	Sept. 1966–June 1967 [1]	1,370	535,000
	Diffusivity at 50-percent-probability level			1,200,000

[1] Affected by periodic pumping.
[2] Average of two recession periods.

tained from a log-normal probability plot of the values in the table, is 1.2×10^6 cu ft per day per ft (9.0×10^6 gpd per ft). If the storage coefficient of the basin fill is 0.0005, then the transmissivity is 600 cu ft per day per ft (4,500 gpd per ft).

Rorabaugh (1960, p. 315) stated that the period required for the water-table profile to stabilize following recharge to the aquifer may be approximated by $t = 0.15\ a^2/D$). Assuming a = 25,000 feet and $D = 1.2 \times 10^6$ cu ft per day per ft, then t = 78 days. Thus, the water-table profile in the basin fill would not be expected to stabilize until about 2½ months after it had been recharged. Most of the recessions used in this analysis define a continuous water-level decline for at least 6 months and, therefore, should satisfy this criteria. The range in diffusivities shown in table 9 may indicate actual variability over the study area, but it may also reflect erroneous estimates of the basin-fill width. All the wells, except well 0725, may define recessions that are too steep because of periodic pumping; resulting D values would then be too high. The assumptions of an initial horizontal water table and an instantaneous recharge to the aquifer are not satisfied in this problem, and their effect on the results is unknown. These determinations are also dependent on the altitude of the water table prior to recharge, which is questionable in this analysis. However, the standard deviation of the computed diffusivities in table

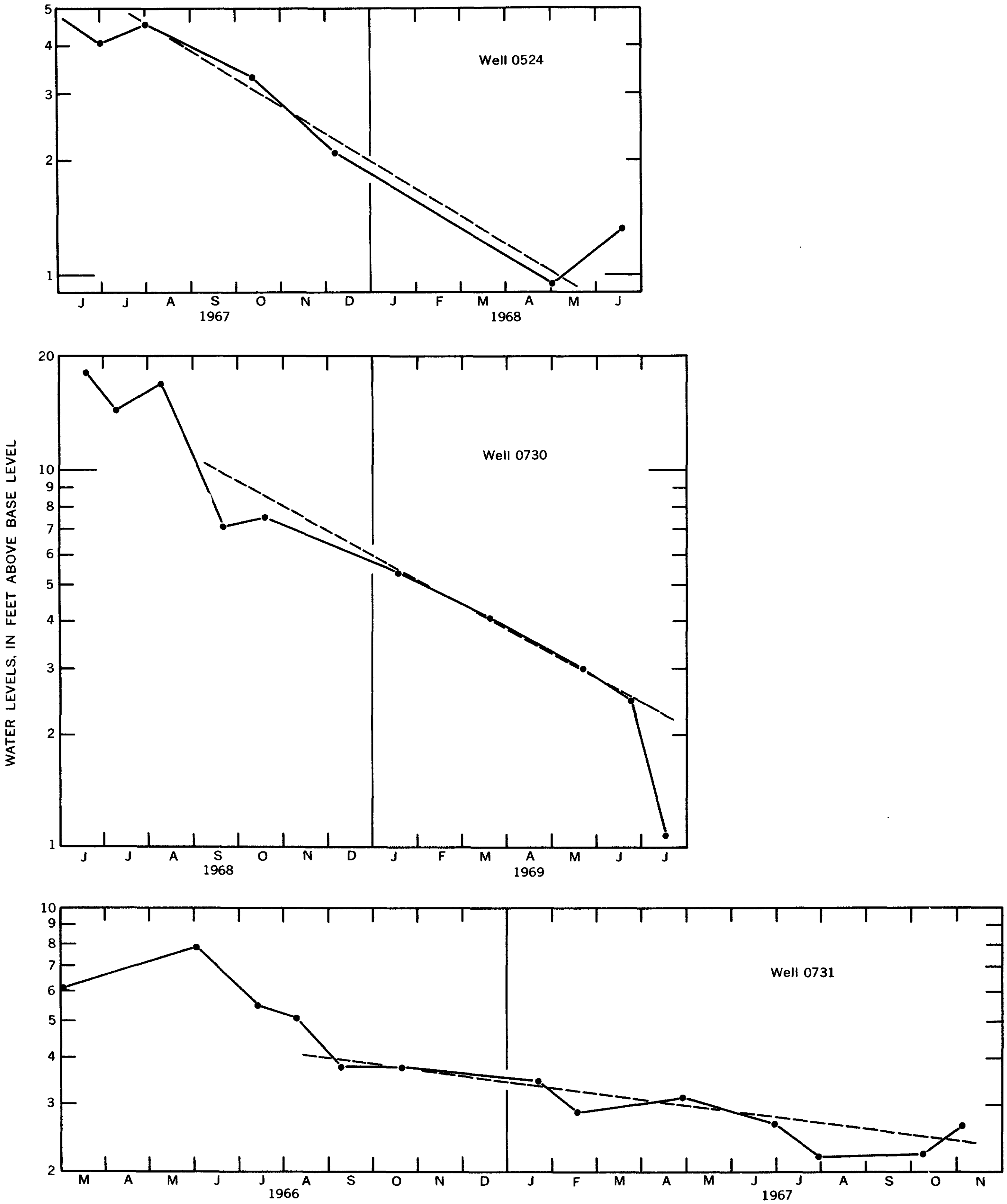


FIGURE 13.—Seasonal water-level recessions in five basin-fill wells, and straight lines averaging their slopes for obtaining the time interval per log cycle of water-level change.

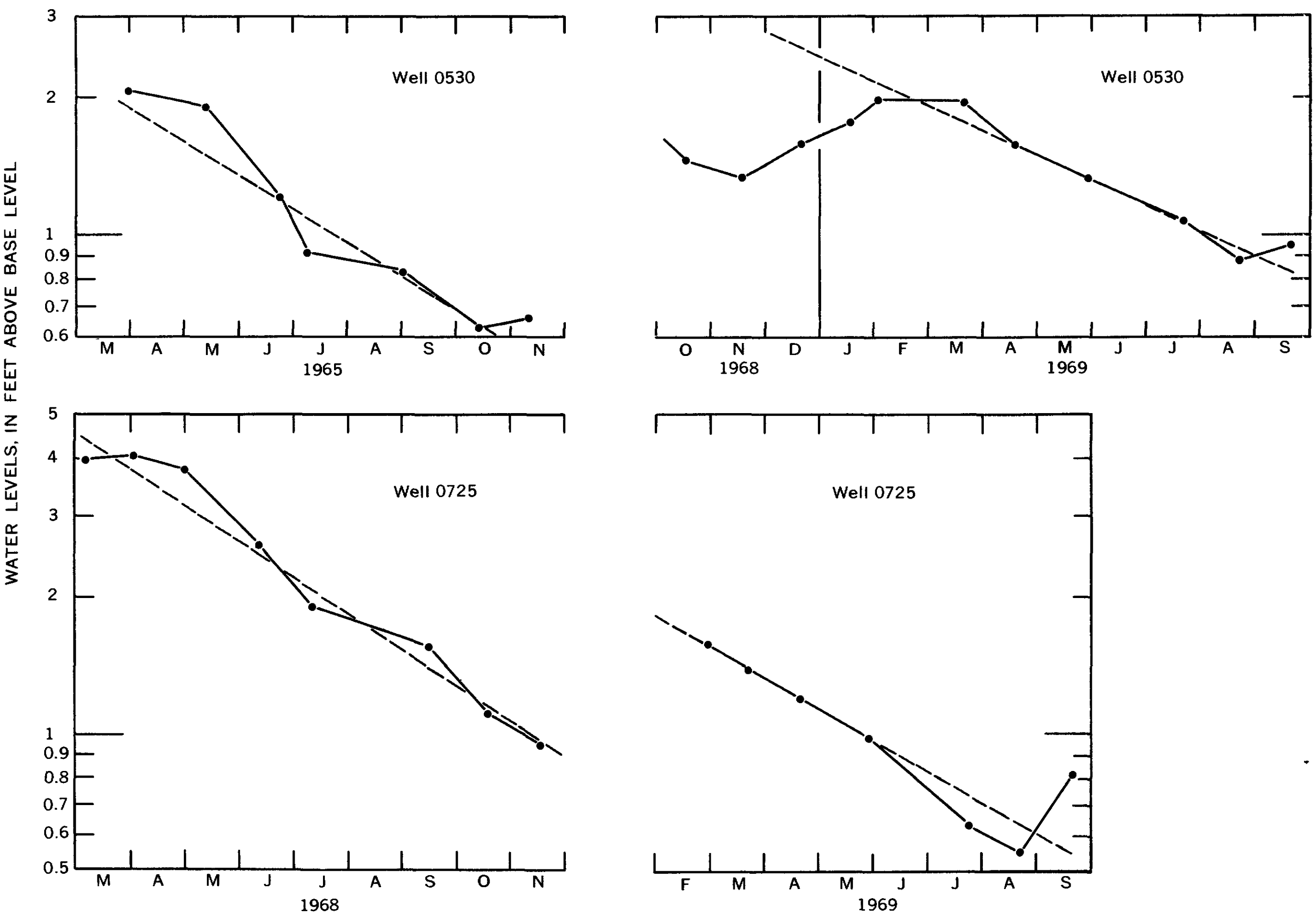


FIGURE 13.—Continued.

9 is relatively small (±31 percent), considering the numerous limitations and unknowns associated with this analysis.

Rorabaugh (1960) also described a method of estimating diffusivity from the decline in the groundwater profile during a recession period, using three wells in a line normal to the flood plain. The recession period is assumed to represent steady-state conditions, so that the shape of the profile is stable throughout its decline. An expression for diffusivity in this instance is

$$D = \frac{\Delta h}{\Delta t}[cd(c-d)/2(bc-ad)], \qquad (10)$$

where Δh is the drop in the water-table profile during the period Δt; a and b are the differences in the head of the profile between the upper and middle wells, and between the middle and lower wells, respectively; and c and d are the horizontal distances between these respective wells.

Three wells (0730, 0731, and 0729), on the right bank of the Gila River approximately on a line perpendicular to the flood plain near section 9 (fig. 1) were used to estimate diffusivity of the basin fill by this method. Wells 0730 and 0731 are in the basin fill and are described in table 8. Well 0729 is in the flood plain; it was augered to a depth of 42 feet, cased with a 1-inch-diameter pipe to a depth of 32 feet, and finished with a 2½-foot well point. Whether this well penetrates the basin fill is questionable.

Computations were made for two recession periods—June 2 to August 8, 1966, and January 17 to March 19, 1969. Wells 0730 and 0731 may have been pumped occasionally during these two periods, but drawdown does not appear to have been significant, judging from the shape of their water-level hydrographs. Average water-table profiles for these recession periods are shown in figure 14. The data obtained from these profiles and used in equation 10 to compute diffusivity are listed in table 10. The resulting diffusivities are 516,000 cu ft per day per ft for the 1966 profile and 62,900 cu ft per day per ft for the 1969 profile. If the storage coefficient of the basin fill is 0.0005, the transmissivities for these two profiles are 258 cu ft per day per ft (1,930 gpd per ft) and 31.5 cu ft per day per ft (235 gpd per ft). The large discrepancy between these two determinations suggests that these profiles are not sufficiently stable for this method to be applicable. This is apparent from the change in shape of the profiles during each recession period (indicated by the range in Δh values between wells) and from the change in shape between each recession period. The water levels in wells 0730 and 0731 may also be affected by periodic pumping and well 0729 may reflect water levels in the alluvium, rather than in the basin fill. These results are therefore considered to be unreliable.

GEOTHERMAL TESTS AT WELLS 1141 AND 1756

Bredehoeft and Papadopulos (1965) described a technique for determining the vertical velocity of ground water in a fully saturated porous medium from observations of the temperature profile in the medium. By assuming that the aquifer is isotropic and homogeneous, the vertical steady movement of ground water may be expressed as

$$v_z = \frac{\beta k}{c_0 \rho_0 L}, \qquad (11)$$

where v_z is the apparent vertical velocity, k is the thermal conductivity of the water-mass complex, c_0 is the specific heat of water, ρ_0 is the density of water, L is the vertical distance over which the temperature profile is measured, and β is a dimensionless parameter obtained from type curves used to match the measured temperature profile. Velocities computed from equation 11 account for the porosity of the medium, and, therefore, represent the "apparent" velocity through the gross cross-sectional area of flow. The corresponding velocity through the pore spaces is $v_p = v_z/\alpha$, where v_p is the pore velocity, and α is the porosity of the medium.

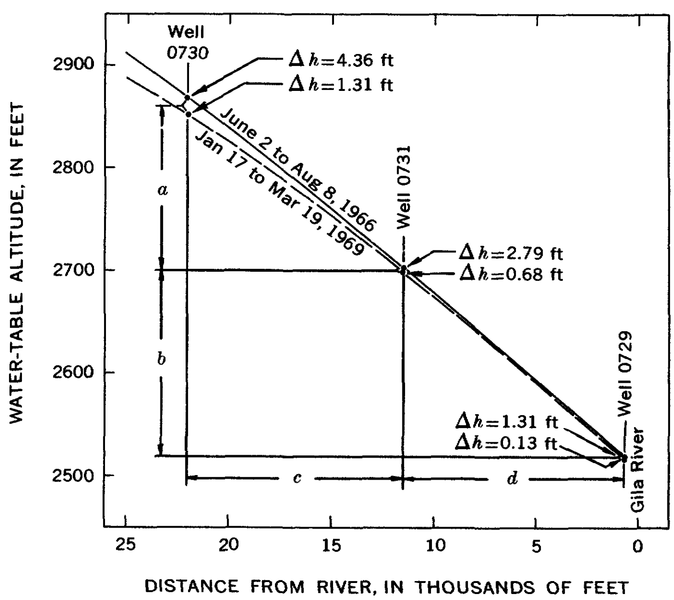


FIGURE 14.—Average water-table profiles in the basin fill near cross-section 9 during the periods June 2 to August 8, 1966, and January 17 to March 19, 1969.

TABLE 10.—*Water-level heads* (a *and* b), *water-level changes* (Δh), *distances between wells* (c *and* d), *time increments* (Δt), *and corresponding diffusivities* (D), *computed from equation 10 using water-level recessions at wells 0730, 0731, and 0729 during 1966 and 1969*

[See fig. 14]

Period	*a* (ft)	*b* (ft)	*c* (ft)	*d* (ft)	Average Δh (ft)	Δt (days)	*D* (cu ft per day per ft)
June 2–Aug. 8, 1966	168.80	182.94	10,500	10,800	2.8	67	516,000
Jan. 17–Mar. 19, 1969	152.45	181.98	10,500	10,800	0.84	61	62,900

The values for c_0 and ρ_0 in equation 11 are 1 cal/g-°C (calorie per gram per degree Celsius) and 1 g/cm^3 (gram per cubic centimeter), respectively. Birch (Birch and others, 1942, p. 259) gave values of k for various soils ranging from 0.8×10^{-3} cal/cm-sec-°C (calorie per centimeter per second per degree Celsius) for muck soil at 40-percent moisture by weight to 5.5×10^{-3} cal/cm-sec-°C for fine sandy loam at 21-percent moisture by weight. A k of 3×10^{-3} cal/cm-sec-°C is considered to be a reasonable value for the basin fill—deposits of saturated sand, silt, and clay.

A solution for equation 11 is obtained by first plotting the measured temperature profile of the saturated medium in the nondimensional form $(z/L) - (T_z - T_0)/(T_L - T_0)$ versus z/L (Stallman, 1967, p. 185), where

T_z = temperature at any depth, z, below the uppermost temperature measurement;
T_0 = temperature at the top of the measured profile ($z = 0$); and
T_L = temperature at the bottom of the measured profile ($z = L$).

This plot is then compared with type curves of $z/L - f(\beta)$ versus z/L prepared from tabulated values in Bredehoeft and Papadopulos (1965, table 1, p. 327) and the β value associated with the type curve which best fits the data points on the plot is substituted in equation 11 to obtain v_z.

R. W. Stallman applied this technique in estimating the rate of upward flow from the basin fill to the overlying alluvium at wells 1141 and 1756.

(See fig. 1.) Both wells were drilled 120 feet deep, and the upper 50 feet in each was cased with an 8-inch-diameter pipe. The total depth in each was then cased with a 2-inch-diameter pipe placed inside the 8-inch-diameter casing. A temperature-measuring device sensitive to small temperature changes was lowered into the 2-inch-diameter pipes, and water-temperature readings were obtained at intermittent depths. Table 11 lists these readings for both wells.

Well-log data indicate that the basin-fill deposits occur 50 feet below land surface at well 1141 and 60 feet below land surface at well 1756. The temperature profile for well 1756 indicates a downward movement of water above the 80-foot depth. To insure that the velocity determinations were representative of the region of upward movement in the basin fill, the analysis considered only that part of the temperature profile below 60 feet at well 1141 and below 85 feet at well 1756. The temperature measurements at the bottom of each well (120-ft depth) were also omitted because of possible adverse effects that might result from the temperature sensor being buried in silt and clay.

Figure 15 shows the measured-temperature profiles for both wells and the type curves with their corresponding β values. The β values for the type curves which provide the best fit to the data plots are -0.2 for well 1141 and -0.7 for well 1756. The computed velocities corresponding to these β values are 0.4 foot per year for well 1141 and 2.4 feet per year for well 1756.

TABLE 11.—*Water-temperature observations in wells 1141 and 1756 from the geothermal test*

[----, No observation]

Depth below measuring point[1] (ft)	Water temperature (°C)	
	Well 1141	Well 1756
40	17.36	--------
45	17.43	--------
50	--------	20.62
55	17.55	20.68
60	--------	20.74
65	17.72	20.81
70	--------	20.88
75	17.96	20.93
80	--------	21.00
85	18.19	21.04
90	--------	21.10
95	18.48	21.15
100	--------	21.22
105	18.72	21.27
110	--------	21.32
115	18.90	21.38
120	18.92	21.43

[1] Measuring point 1 ft above land surface for well 1141, and 2 ft above land surface for well 1756.

A brief discussion of this test by Weist (1971) indicates apparent velocities of 0.8 foot per year at well 1141 and 1.1 feet per year at well 1756. The revised values presented in this report result from a slightly different interpretation of the temperature profiles and a reevaluation of the value for k.

Thermal instability within the aquifer caused the turnover of water within the observation wells, resulting in temperature readings at any given depth to vary about $\pm 0.01°C$. However, the rate of change in temperature with depth—an indicator of vertical flow—was generally greater than this measurement variation. Similar tests by Edwin P. Weeks (written commun., 1970) suggest that the metal well casing may act as a thermal short circuit and distort the thermal gradient. Thus, the actual temperature profiles may depart from the measured profiles. The value of k for the basin fill is also questionable and undoubtedly varies with depth. Because the true average of k may range from 1×10^{-3} to 5×10^{-3} cal/cm-sec-°C, the corresponding value computed for v_z would accordingly range from 0.1 to 4.1 feet per year.

Because of these uncertainties, the vertical velocities derived in this analysis are considered as only approximate. The results do indicate, however, that the movement of water from the basin fill into the overlying alluvium is very small and may average about 1.5 feet per year, or 0.004 foot per day.

EVALUATON OF RESULTS

The average storage coefficient, $\bar{S}$, for the full saturated thickness of alluvium was estimated to be 0.15. The $\bar{S}$ for the alluvium in the zone of water-table fluctuations with phreatophyte cover is 0.20. The pooled standard deviation of these estimates is 33 percent, whereas the areal variability in the storage coefficient throughout the alluvium is about 25 percent.

The $\bar{S}$ for the basin fill has been estimated as 0.0005. Insufficient data are available to evaluate the error in this estimate or define the areal variation of the storage coefficient.

Several analytical techniques were employed in this analysis to determine the diffusivity and transmissivity of the alluvium and basin fill. Table 12 summarizes the results of these determinations. Where more than one determination was made for a particular location, the indicated diffusivity and the transmissivity represent median values taken from their probability distribution. The table indicates a wide range in values for both the alluvium and the basin fill, which may, in part, reflect the true areal

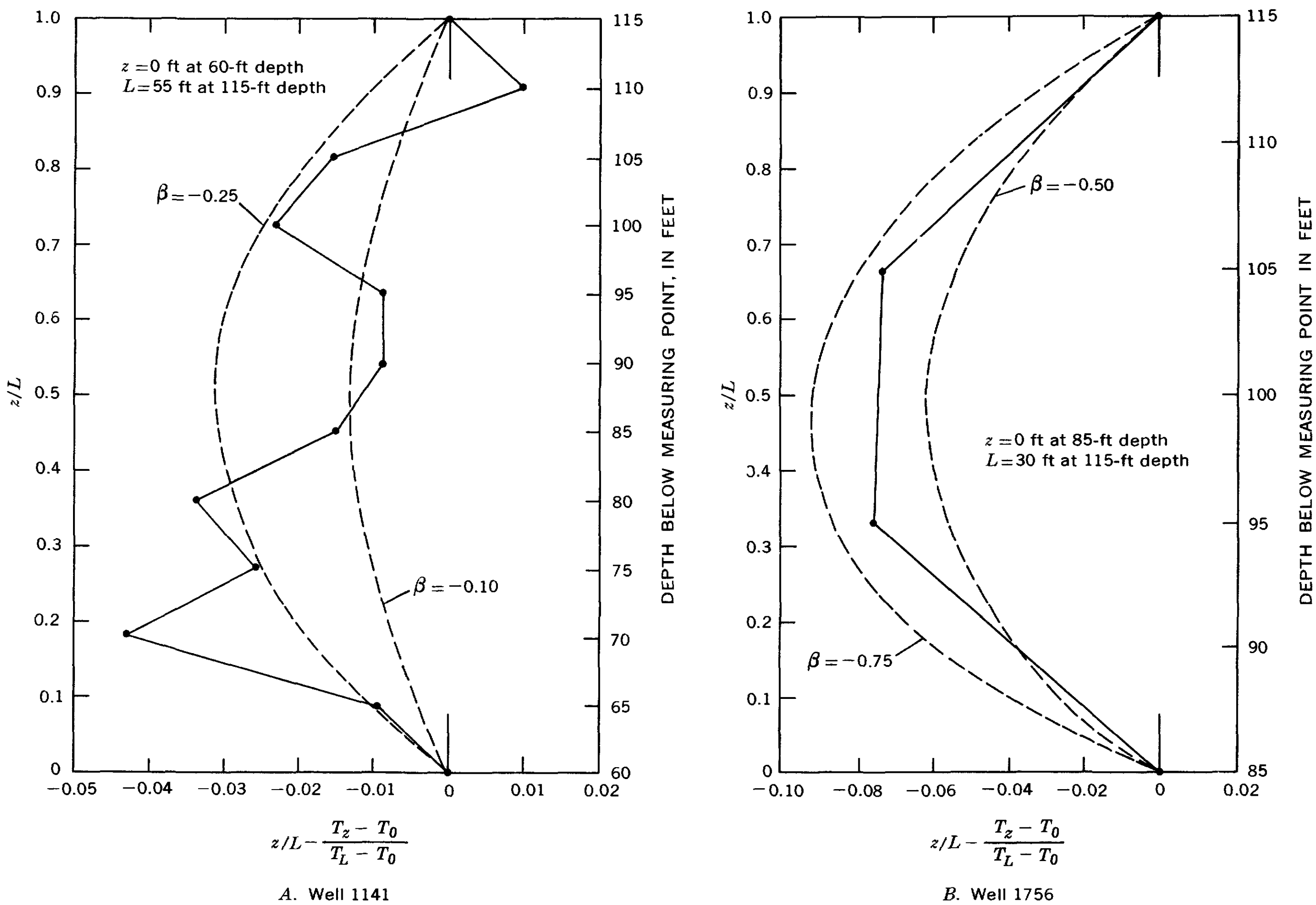


FIGURE 15.—β type curves and temperature profiles in the basin fill at wells 1141 and 1756.

variation in the water-carrying capacity of each of these aquifers. Some of these values are questionable, however, because the assumptions underlying their solution may not have been sufficiently satisfied. In general, those values derived from several determinations at one site or from data at several wells are considered to be the most representative.

Of the methods used in this study, the drawdown-recovery tests are considered to give the best estimate of transmissivity at a particular site, while the flood-wave propagation analysis is considered to give the best estimate of the average diffusivity for the area.

On the basis of these criteria, the best estimate of transmissivity for the alluvium at a particular site is 9,800 cu ft per day per ft (73,000 gpd per ft) obtained from the drawdown-recovery test at section 1. Assuming the saturated thickness of alluvium, m, is 40 feet, the corresponding hydraulic conductivity, K, is 245 ft per day (1,830 gpd per sq ft). The storage coefficient at this site was computed as 0.16.

A representative average value of diffusivity for the alluvium is 185,000 cu ft per day per ft, as obtained from the flood-wave propagation analysis. If the average storage coefficient is 0.15, the corresponding average transmissivity for the study area is 27,800 cu ft per day per ft (208,000 gpd per ft). Assuming m = 40 feet, the corresponding average K is 695 feet per day (5,200 gpd per sq ft). This value of K would classify the alluvium as a good aquifer consisting of mixtures of clean sands and gravels (Todd, 1963, figure 3.4).

The drawdown-recovery tests at wells 0531 and 0532 were used to obtain an average transmissivity for the basin fill of 15 cu ft per day per ft (110 gpd per ft). The significantly higher transmissivities obtained from the ground-water recession analysis in the basin fill are considered to be less reliable because of the effect that periodic pumping may have had on the recessions.

Assuming that the basin fill is 1,000 feet thick, the corresponding horizontal hydraulic conductivity

TABLE 12.—*Summary of diffusivities and transmissivities for the alluvium and basin fill*

[Asterisk (*) indicates value has been calculated from its indicated storage-coefficient value]

Location	Number of wells	Number of determinations	Storage coefficient	Diffusivity (cu ft per day per ft)	Transmissivity: Cubic feet per day per foot	Transmissivity: Gallons per day per foot	Method of analysis
Alluvium							
Section 1, right bank	13	13	0.16	*61,300	9,800	73,000	Drawdown-recovery.
River and flood plain wells	21	21	.15	215,000	*32,200	*241,000	Flood-wave propagation.
Flood-plain and terrace wells	10	10	.15	108,000	*16,200	*121,000	Do.
River and terrace wells	24	24	.15	185,000	*27,800	*208,000	Do.
Sections 1, 9, and 17	9	1	.15	179,000	*26,800	*201,000	Cyclic fluctuation.
Section 5, left bank	3	1	.15	104,000	*16,000	*120,000	Constant-head drain function.
Section 5, right bank	3	1	.15	375,000	*56,000	*420,000	Do.
Sections 1 and 5	5	7	.15	16,000	*2,400	*18,000	Do.
Basin fill							
Wells 0531 and 0532	2	2	0.0005	*27,000	15	110	Drawdown-recovery.
Basin-fill wells, sections 5–7	5	7	.0005	1,185,000	*600	*4,500	Ground-water recession.
Wells 0729, 0730, and 0731	3	1	.0005	516,000	*258	*1,930	Ground-water recession.
Do	3	1	.0005	62,900	*31.5	*235	Do.

is $K = 0.015$ foot per day (0.11 gpd per sq ft). This K value classifies the basin fill as a poor aquifer, consisting of sand, silt, and clay (Todd, 1963, fig. 3.4).

The upward vertical velocity determinations of 0.4 foot per year at well 1141 and 2.4 feet per year at well 1756, based on the geothermal tests in the basin fill, are questionable, but are considered to be reasonable estimates of vertical movement of water from the basin fill into the overlying alluvium.

GROUND-WATER MOVEMENT

The aquifer constants derived in this analysis may be used to compute ground-water movement in the study area. Two components of flow are considered here—downvalley flow through the alluvium, and basin-fill flow into the overlying alluvium. Figure 16 illustrates the direction of both of these ground-water-flow components.

Downvalley flow through the alluvium is computed from

$$q = Tiw, \tag{12}$$

where q is the downvalley flow, T is the transmissivity of the alluvium, i is the downvalley hydraulic gradient, and w is the alluvium width. By substituting in equation 12 the representative average values of $T = 28{,}000$ cu ft per day per ft, $i = 0.0016$, and $w = 5{,}000$ feet, the average downvalley flow through the alluvium is computed to be 224,000 cu ft per day (5.1 acre-ft per day).

Virtually all ground water flowing from the basin fill into the alluvium moves vertically upward, as indicated in figure 16. The contribution to the alluvium may be computed from

$$q_b = v_z wL, \tag{13}$$

where q_b is the total vertical upward flow through an area of alluvium of width w and length L, and v_z is the apparent upward vertical velocity through the basin fill, as defined in equation 11. Assuming that $w = 5{,}000$ feet, $L = 5{,}280$ feet, and $v_z = 0.004$ foot per day, equation 13 gives a vertical upward flow of 106,000 cu ft per day per mile of valley length (2.4 acre-ft per day per mile).

Rorabaugh (1964) showed that basin-fill flow into the alluvium may be computed from

$$q_b = 2Tha/x(2a-x), \tag{14}$$

where q_b is the flow per unit strip of aquifer normal to the downstream direction, T is the transmissivity of the basin fill, a is the width of the basin fill between its contact with an impermeable boundary (ground-water divide) and its contact with the saturated alluvium, and h is the hydraulic head in the basin fill at a distance, x, from the basin-fill–alluvium contact. This relation assumes that recharge to the basin fill is under steady state conditions and is considered applicable where the water-level fluctuations are small compared to the hydraulic head. If the basin fill is recharged at a constant rate over a prolonged period, equation 14 gives an upper limit of discharge.

Equation 14 was used to compute q_b at five well sites on the right bank (north bank) and six well sites on the left bank (south bank) of the study reach of the Gila River. A transmissivity of 15 cu ft per day per ft (110 gpd per ft) was assumed for use in all computations. Table 13 lists the values of h, x, a, and corresponding q_b for each site.

The median flows in table 13 are 0.42 cu ft per day per foot of valley length on the right bank, and

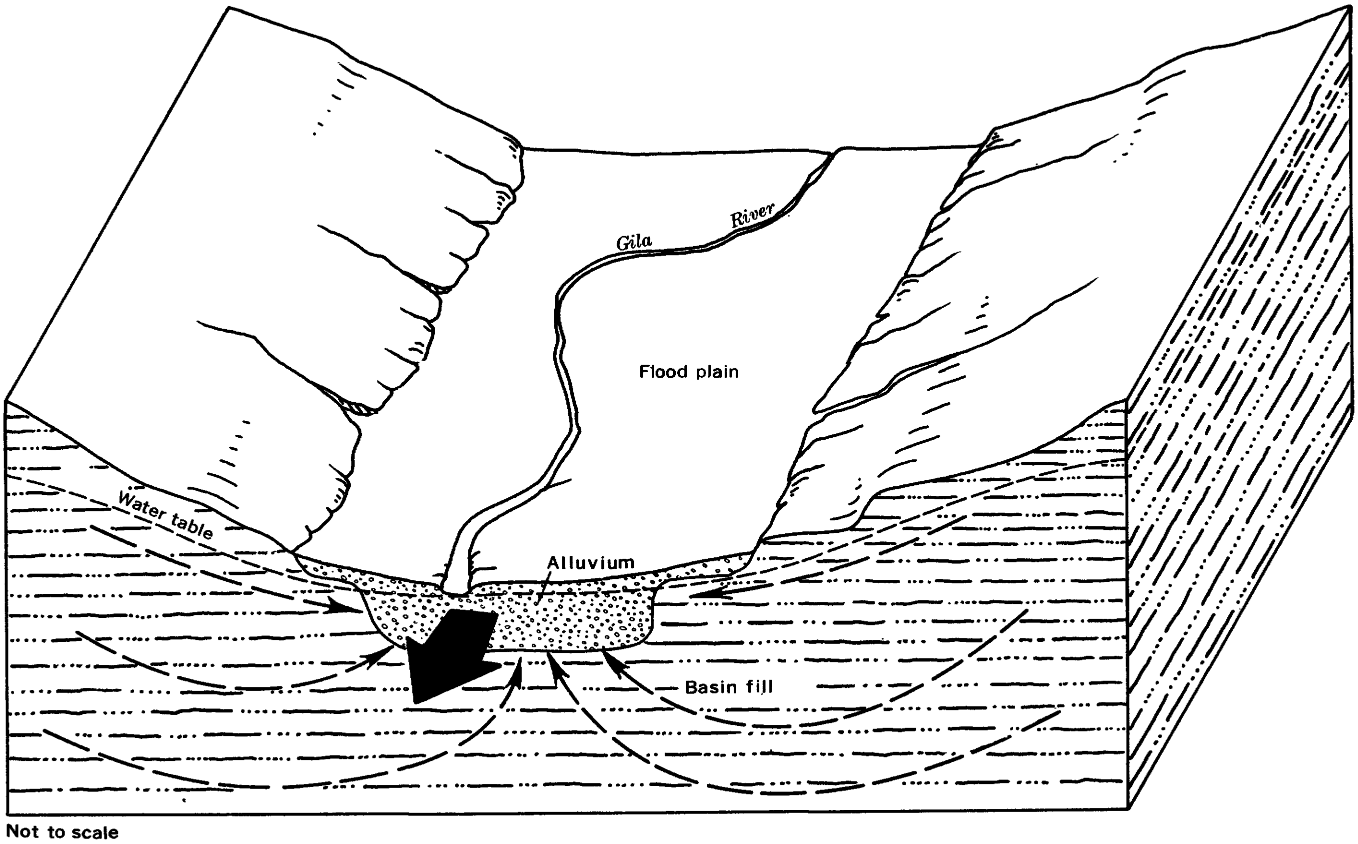


FIGURE 16.—Ground-water-flow pattern (arrows) in the alluvium and the basin fill.

0.72 cu ft per day per foot of valley length on the left bank. The total average flow from the basin fill into the alluvium is, therefore, 1.14 cu ft per day per foot of valley length or 0.14 acre-ft per day per mile of valley length.

TABLE 13.—*Hydraulic head* (h), *distance between well and outflow boundary* (x), *width of basin fill* (a), *and corresponding basin-fill flow* (q_b) *computed from equation 14 for five wells on the right bank and six wells on the left bank*

Well No.	h (ft)	x (ft)	a (ft)	q_b (cu ft per day per ft)
		Right bank		
0524	325	14,000	26,000	0.47
0730	329	17,000	24,500	.45
0731	181	7,500	27,500	.42
1761	16	1,000	21,500	.24
1762	451	26,000	26,000	.52
Median q_b				0.42
		Left bank		
0522	423	9,200	21,200	0.89
0531	25	500	23,300	.76
0732	460	13,700	21,700	.74
1343	34	2,400	22,400	.22
2169	416	18,800	33,800	.46
2373	822	27,500	36,500	.72
Median q_b				0.72

If virtually all this water moves vertically upward into the overlying alluvium, having a width of 5,000 feet, the corresponding apparent vertical velocity is $v_z = 1.14/5{,}000 = 0.0002$ foot per day. This velocity is 6 percent of the average apparent velocity of 0.004 foot per day obtained from the geothermal test. (See p. F23.) Additional aquifer tests in the basin fill would be required to explain the large discrepancy between these two average values of v_z. However, these determinations do provide an estimate of the possible magnitude of basin-fill flow into the overlying alluvium.

CONCLUSIONS

Various analytical techniques were used to evaluate the aquifer characteristics of the alluvium and basin fill. Extensive use was made of existing soil-moisture, ground-water, and streamflow data in applying these techniques.

The reliability of the resulting determinations of storage coefficient, diffusivity, transmissivity, hydraulic conductivity, and vertical velocity of ground water is questionable in many of the determinations because of the constraints and assumptions associated with the techniques used in the analysis. However, where relatively consistent aquifer constants were obtained by using different methods of analysis or different sets of data, the results are considered to be representative.

This study indicates that the average storage coefficient is 0.15 for the alluvium and 0.0005 for the underlying basin fill. The average transmissivity is 28,000 cu ft per day per ft (210,000 gpd per ft) for the alluvium and 15 cu ft per day per ft (110 gpd per ft) for the basin fill. The corresponding average horizontal hydraulic conductivity of the alluvium is 695 feet per day (5,200 gpd per sq ft). The horizontal hydraulic conductivity in the basin fill is estimated at about 0.015 foot per day (0.11 gpd per sq ft). Geothermal measurements of the upward vertical flow of water from the basin fill to the overlying alluvium average 1.5 feet per year per square foot of area. Because of the wide range in determinations of storage coefficient and transmissivity at any given site, no attempt has been made to define their spacial distribution. These aquifer constants will be used, however, in analog and digital groundwater models to test their reliability and approximate their spacial distribution.

The aquifer constants derived in this analysis indicate that the average downvalley ground-water flow through the alluvium is 5.1 acre-feet per day. Estimates of flow from the basin fill into the overlying alluvium range from 0.14 to 2.4 acre-feet per day per mile of valley length.

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