6089

G-000-307 .8

REPORT TO THE SOUTHWESTERN OHIO WATER COMPANY ON EVALUATION OF THE POTENTIAL YIELD OF THE AQUIFER IN THE BIG BEND AREA, OHIO (USED AS A REFERENCE IN THE OU1, OU2, OU4 AND OU5 RI REPORTS)

/ /56

45
REPORT

Report to
The Southwestern Ohio Water Company
on
Evaluation of the potential
yield of the aquifer in the
Big Bend area
by
Schaefer & Walton
Consulting Ground Water Hydrologists

Contents

	Page
Synopsis	1
Introduction	4
General review of ground-water conditions.....	5
The hydraulic properties of the aquifer	11
Controlled pumping tests.....	12
The water table of the aquifer.....	17
Recharge by the induced infiltration of water in the Miami River.....	20
Infiltration rates.....	22
Potential yield of the aquifer.....	28
Conclusions.....	43
References.....	46
Appendix.....	47

Illustrations

Figure		Page
1.	Map showing contours on the bedrock surface.....	8
2.	Map showing locations of wells used during controlled pumping tests.....	14
3.	Distance-drawdown semilog graphs for test 1 and test 2.....	15
4.	Profiles showing the configuration of the bed of the Miami River in the Big Bend area, A-A', B-B', J-J' and K-K'.....	24
5.	Profiles showing the configuration of the bed of the Miami River in the Big Bend area, C-C', D-D', H-H' and I-I'.....	25
6.	Profiles showing the configuration of the bed of the Miami River in the Big Bend area, E-E', F-F' and G-G'.....	26
7.	Mean daily gage heights of the Miami River at Hamilton, Ohio during 1954.....	30
8.	Profiles of the river channel along line F-F'....	31
9.	Graphs showing the average depth and width of the Miami River for various stages of the river.....	32
10.	Graphs showing the average depth of the Miami River and the average area of the river bed submerged for various stages of the river.....	33
11.	Water levels in observation well K-2,1952-1956...	39
12.	Theoretical Distance-drawdown graph.....	41

Plate

I. Map showing water table contours on August 31,1956 In pocket

Synopsis

A hydrogeological survey was made to determine the potential yield of the Southwestern Ohio Water Company well field. The well field is located along the Miami River near Venice, Ohio about 14 miles northwest of Cincinnati. At the present time an average of about 13 mgd. of ground water is pumped from two Water Collectors in the permeable sand and gravel deposits which underlie the area.

Geologic studies show that permeable sand and gravel deposits averaging about 150 feet in thickness occur in an area about 1-3/4 miles wide at the well field. These glacial deposits partially fill the deep buried valley of the ancestrial Miami River.

From the results of two controlled pumping tests and a study of the water table, the average coefficients of permeability and transmissibility of the sand and gravel aquifer were computed to be 2560 gpd./sq.ft. and 315,000 gpd./ft. respectively. The coefficient of storage of the aquifer was estimated to be 0.20. These hydraulic properties of the glacial deposits filling the wide buried valley are high enough to permit the development of a large supply of ground water.

Under heavy pumping conditions, recharge to the valley-fill deposits will occur chiefly by the induced infiltration of surface water from the Miami River. Potential recharge from this source was estimated by correlating infiltration and streamflow data for the record dry period 1952, 1953

and 1954.

The water levels in about 40 observation wells were used to prepare a water-table map showing the cones of depression created by pumping from the existing collectors. The cones of depression, distorted by recharge from the river between the two collectors, were analyzed and the average infiltration rate of the river bed was determined to be 168,000 gallons per day per acre of river bed per foot of head. Computations show that 62 percent of the pumpage from the collectors, about 10,550,000 gpd., was recharged by the induced infiltration of river water on August 31, 1956. Studies of ground-water levels at eleven measured sections of the river bed show that most of the infiltration occurred in the deep narrow channel of the river and that only a small amount of infiltration occurred in the shallow part of the river where fine silty materials have been deposited. Two natural dams in the river (rapids occur up and downstream from the collectors) adversely affect the infiltration capacity of the river bed.

The magnitude of potential recharge by the induced infiltration of surface water was estimated by correlating data collected on the stage of the Miami River during the period 1952 through 1954 (an extended dry period) with data on the average infiltration rate of the river bed.

The studies show that the potential yield of the well field area is about 40 million gallons a day in dry periods. In the record dry period studied (1952, 1953, 1954) with-

drawals of 40 mgd. would have been balanced by recharge, chiefly by induced infiltration from the Miami River. In this period the recharge at times would have averaged as little as 24 mgd. However, computations show that the storage available in the aquifer would have been more than adequate to sustain the 40 mgd. rate through the minimum infiltration rate periods, and until replacement of the lost storage in higher infiltration rate periods.

The studies also show that an extensive, relatively flat cone of depression will be created when large amounts of water are pumped from storage in the dry periods. For example, it was found in checking the available storage, that in these dry periods ground-water levels 1 mile northeast of existing Collector 2 would be lowered as much as 29 feet and that the lowering 1-1/4 miles southwest of Collector 2 would be as much as 23 feet.

All predictions in the report were made on the assumption that the well field will be limited to the area south of the Butler-Hamilton County line. It should be emphasized that the predicted 40 mgd. yield is the minimum yield to be expected. In years of normal or above normal precipitation the yield of the aquifer will be much greater and the declines in ground-water levels for the 40 mgd. pumping rate will be less than those mentioned above.

The potential yield of the well field area computed in this report (40 mgd.) is much less than the yield (63 mgd.) given in the progress report previously submitted by the writers, because the adverse effects of the natural dams in the river, which were unknown before the present investigation, were not considered in computations made in the previous report.

INTRODUCTION

The Southwestern Ohio Water Company is now pumping ground water at an average rate of about 13 mgd. from two Ranney Water Collectors in the northwest corner of Colerain Township in northern Hamilton County, Ohio. The well field is located along the oxbow meander of the Miami River near Venice, Ohio about 14 miles northwest of Cincinnati. The Water Company plans to increase withdrawal rates in the near future and is conducting an investigation to determine the potential yield of the ground-water reservoir in the well field area (Big Bend area).

In January, 1956 the writers were authorized to supervise the investigation mentioned above and to periodically report on the results of the study. A report entitled "First Progress Report on Results of Test-Well Drilling and Observation-Well Program" was prepared by the writers and presented to The Southwestern Ohio Water Company on June 6, 1956. This report described the information obtained from six exploratory test holes and seven observation wells which were drilled or installed during the winter and spring months of 1956. The nature of the studies to be made during the summer and fall months of 1956 was discussed and recommendations were made as to the number and location of additional observation wells needed to accurately define the water table. In addition, upon request, the writers estimated the sustained yield of the well field by studying the meager geologic and hydrologic data available at that time. The writers expressed

the opinion, based chiefly on available infiltration data collected in other parts of Ohio, that the potential recharge to the aquifer in the Big Bend area by the induced infiltration of river water during extended dry periods could be expected to be about 63 mgd.

During the summer months of 1956 fourteen 2-inch diameter observation wells were installed and the water levels in these wells together with 26 other observation wells were measured each month. Eight of the above mentioned observation w ells were driven in the river bed. The configuration of the river bed was determined in August by taking soundings at 11 points along the river. In October and November studies were made of existing controlled pumping test and streamflow data. This report describes the information obtained to date and gives the results of an interpretation of the data with regard to the evaluation of the well field capacity.

GENERAL REVIEW OF GROUND-WATER CONDITIONS

The Big Bend area is underlain by approximately 150 feet of unconsolidated deposits consisting mostly of sand and gravel. These materials partly fill a deep valley now occupied in part by the Miami River. The thick deposits of sand and gravel were for the most part deposited during two periods of Pleistocene glaciation. The deep valley was incised in bedrock of sedimentary origin and of Ordovician age. The rocks that crop out along the sides of the valley are relatively impermeable shales and limestones.

Before the Illinoian stage of glaciation, the rocks of the Cincinnati region were planed to a fairly smooth, gently rolling surface into which deep valleys had been carved by the cutting action of surface streams. One of the principal streams of the region was the ancestral Miami River (see Klaer and Thompson, 1948) which occupied approximately the same course as the present Miami River as far south as Venice in Hamilton County. At Venice, in the Big Bend area, the ancestral Miami River turned west and flowed past New Haven and joined the ancestral Whitewater River near Harrison. The deposits under study occur in the buried valley which marks the approximate course of the ancestral Miami River. The buried valley is about 1-3/4 miles wide and is probably U-shaped, having a broad, relatively flat bottom and steep valley walls.

The sand and gravel materials were deposited in the buried valley by the meltwaters from the glaciers of Pleistocene age. The meltwaters, by washing action prior to final deposition of the glacial materials which they carried, removed a high percentage of fine sand and silt, and sized and sorted the remaining sand and gravel into a highly permeable mixture.

A careful study of the logs of exploratory test wells recently drilled and the logs of all other wells drilled to date reveals that in general a sandy zone containing clay lenses exists between the approximate depths of 105 to 125 feet below land surface. These deposits limit to some ex-

tent the vertical movement of ground water in the valley-fill deposits.

As shown in figure 1, the bedrock surface in the Big Bend area slopes generally towards the northwest to a deep buried channel which trends northeast to southwest and underlies the village of Venice and portions of the Hamilton Cleves Road. The axis of the deep channel turns west in the vicinity of well O-2. The elevation of the floor of the narrow deep channel is probably about 360 feet above mean sea level and the elevation of the land surface is about 540 feet. The bedrock surface rises abruptly east of Collector 1 near the East Miami River road and in the vicinity of well C-1 as the edge of the buried valley is approached. Collector 1 is located over a bedrock terrace which has an elevation of about 400 feet above sea level.

The average thickness of the sand and gravel beneath the bed of the Miami River is over 100 feet. Bedrock was encountered at a depth of 113 feet below the land surface in well WK-1, (see figure 1) north of the Colerain Pike bridge. The glacial materials underlying the Miami River near and south of well C-1 are thin, and are composed chiefly of clayey materials.

Ground water in the glacial deposits in the Big Bend area occurs in the irregular spaces or pores that exist between the individual particles of sand and gravel. The pores are relatively small, irregular in shape and interconnected so that water can move under the action of gravity

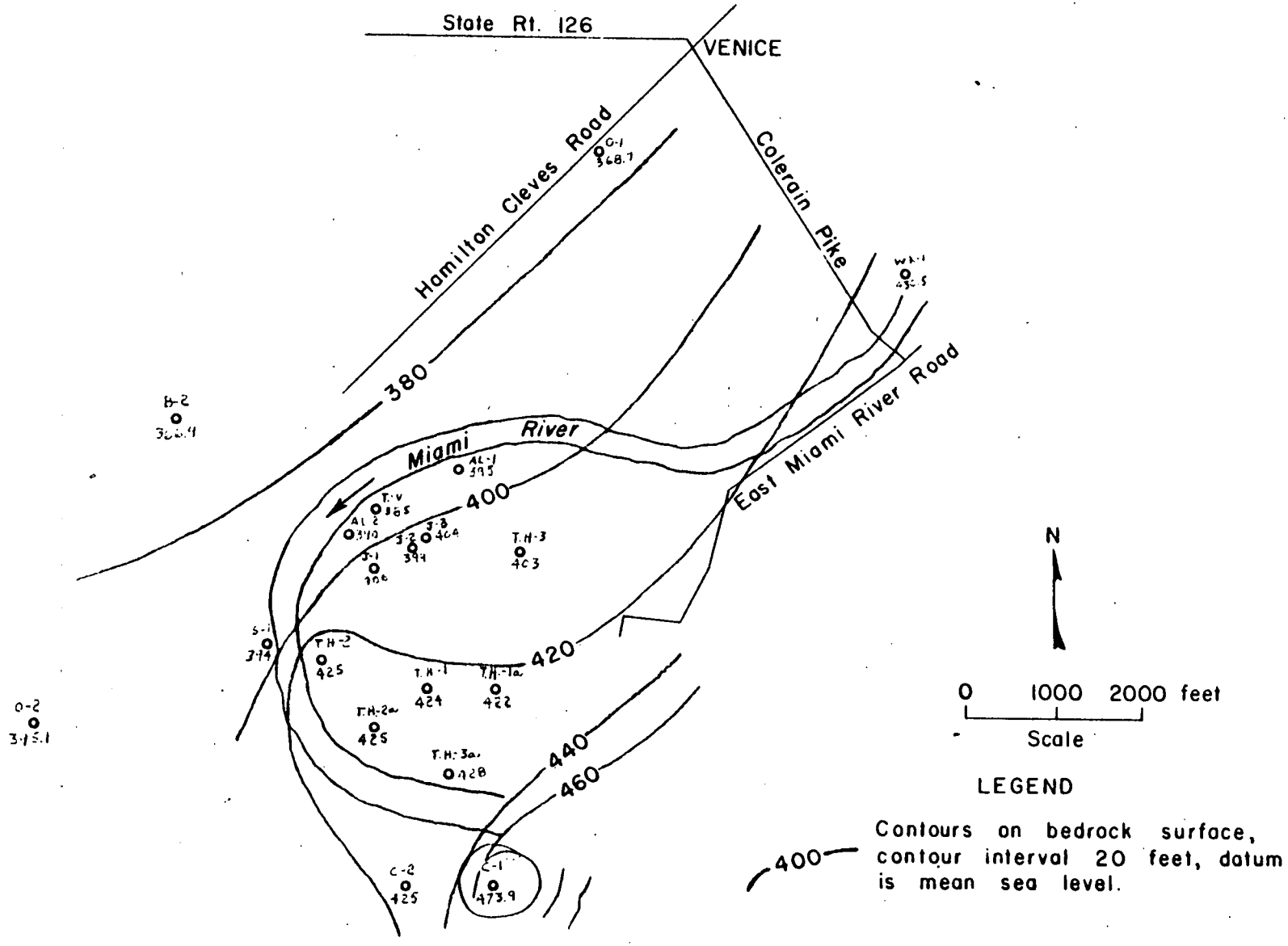

FIGURE 1.-MAP SHOWING CONTOURS ON THE BEDROCK SURFACE.

000011

6089

by percolating from one opening to another.

The percentage of pore spaces that a water-bearing formation (aquifer) contains is indicated by the porosity of the formation. Porosity is expressed quantitatively as the ratio of the void space to the total volume of the aquifer. The porosity of glacial outwash deposits generally ranges from 25 to 40 percent.

Not all of the water in the pore spaces of an aquifer will drain into wells. A part of the stored water is retained against the force of gravity by the forces of adhesion and cohesion. The amount of water that an aquifer will yield by gravity drainage is indicated by its specific yield. Specific yield is expressed quantitatively as the ratio of the volume of water yielded by an aquifer to the total volume of the aquifer. The average specific yield of the outwash sand and gravel in the B ig Bend area is estimated to be 0.20.

The source of ground water is precipitation. Of the precipitation that falls on the land surface a part runs off overland to streams, a part evaporates or is transpired by plants, and the remainder seeps into the ground to recharge the ground-water reservoir. A reasonable estimate of the average annual rate of recharge from precipitation in the Big Bend area is about 22 percent of the average annual precipitation (38.55 inches) or about 8-1/2 inches per year. On this basis, considering a square mile of catchment area, the recharge under natural conditions, would amount to approximately 400,000 gpd.

Under heavy pumping conditions, large amounts of recharge are contributed to the aquifer in the Big Bend area by percolation of water in the Miami River through the river bed and into the aquifer. The cones of depression created by the pumping of the collectors may be considered to be pirating agents. When a collector is first pumped water is withdrawn from storage in the aquifer in the immediate vicinity of the collector. The cone then spreads, drawing from storage continuously, and as water levels are lowered in the discharge area (Miami River), more and more of the natural discharge is diverted to the pumping collector and less and less is discharged by natural methods, which include evaporation from the water table, transpiration through plants, and ground-water runoff to the Miami River. Ultimately, when withdrawal rates exceed natural discharge, ground-water levels decline below the level of the river and surface water is induced to flow into the aquifer.

If infiltration from the reach of the river in the immediate vicinity of the collector does not keep pace with pumping rates, the cone of depression spreads up and down stream until it intercepts sufficient area of the river-bed to balance induced infiltration with pumpage. During extended dry periods, pumpage may exceed the maximum infiltration rate and the cone of depression deepens and expands as water is taken from storage within the aquifer. In wet periods, the cone shrinks as large amounts of recharge are received from high stream flows.

Recharge to an aquifer by the influent seepage of surface water depends upon many factors. Important among these are (1) the permeability of the river-bed and underlying materials; (2) the position of the water table with respect to the bed of the river; and (3) streamflow characteristics. Recharge from streamflow is directly proportional to the difference between ground-water levels beneath the river and the river level, until the water table has declined below the river-bed. Thus, the infiltration rate of the river during high flows may be several times that experienced during low-flow periods because of the change in the depth of water in the river. In addition, the permeability of the river-bed is temporarily increased by the scouring action in high flow periods and is temporarily decreased by the deposition of fine silty materials during low-flow periods

THE HYDRAULIC PROPERTIES OF THE AQUIFER

The hydraulic properties of the aquifer, commonly expressed in terms of the coefficients of permeability, transmissibility and storage, were used in making a quantitative estimate of the water available in the Big Bend area. These coefficients were computed from the results of two controlled pumping tests.

The <u>coefficient of permeability (P)</u> is defined as the rate of flow of water in gallons per day through a vertical cross-sectional area of 1 sq.ft. of the aquifer under a

hydraulic gradient of one foot per foot at the prevailing temperature of the ground water. The coefficient of transmissibility (T), is defined as the rate of flow of water in gallons per day through a vertical strip of the aquifer 1 foot wide and extending the full saturated thickness under a hydraulic gradient of one foot per foot and at the prevailing temperature of the water. T = Pm, where m is the saturated thickness of the aquifer in feet. The coefficient of storage, S, is defined as the change in the stored volume of water in the aquifer for each unit change in head on each unit surface area of the aquifer. (Under water-table conditions, S is for practical purposes equal to the specific yield of the aquifer).

Controlled pumping tests

Under favorable conditions, the hydraulic properties of an aquifer can be determined by means of a controlled pumping test. In this type of a test a well drilled into the aquifer is pumped at a known constant rate, and the effects of this pumping on water levels at other points in the aquifer are determined by obtaining water-level measurements in other wells drilled into the aquifer for observation purposes. Measurements are collected in the observation wells prior to and during the pumping period. The data so obtained is then processed graphically, and used in hydraulic formulas to determine the hydraulic properties of the aquifer.

Controlled pumping tests of the sand and gravel aquifer underlying the Big Bend area were made in 1950 and 1954,

using the wells shown in figure 2. Pumping test 1 was conducted August 31, 1950 to September 6, 1950 (see Kazmann,1950) using one 12-inch pumping well and eight 6-inch observation wells, equipped with automatic water stage recorders. The pattern of wells was located south of the Miami River. Pumping was started at 1:45 pm., September 1, 1950, and was continued at a constant rate, 1350 gpm., for a period of 96 hours ending at 1:45 pm., September 5, 1950. Pumping test 2 was conducted from January 25, 1954 to February 2, 1954 using one 12-inch pumping well and nine 6-inch observations, equipped with automatic water stage recorders. The pattern of wells was located north of the river as shown in figure 2. Pumping was started at 12:45 pm. on January 27, 1954, and was continued at a constant rate, 1175 gpm., for a period of about 5 days ending at 3:09 pm., February 1, 1954.

Values of drawdowns in the observation wells were determined by comparing the extrapolated graphs of water levels before pumping started with the graphs of water levels during the pumping periods. The drawdowns in the observation wells close to the pumped wells were adjusted (see computation sheet in appendix) for the effects of partial penetration using the method outlined by Jacob (1945).

Adjusted values of drawdowns in the observation wells parallel to the river were plotted against values of the distances from the pumped wells to the respective observation wells, on semilog paper as shown in figure 3. Straight lines were drawn through the data for each test and the slopes of

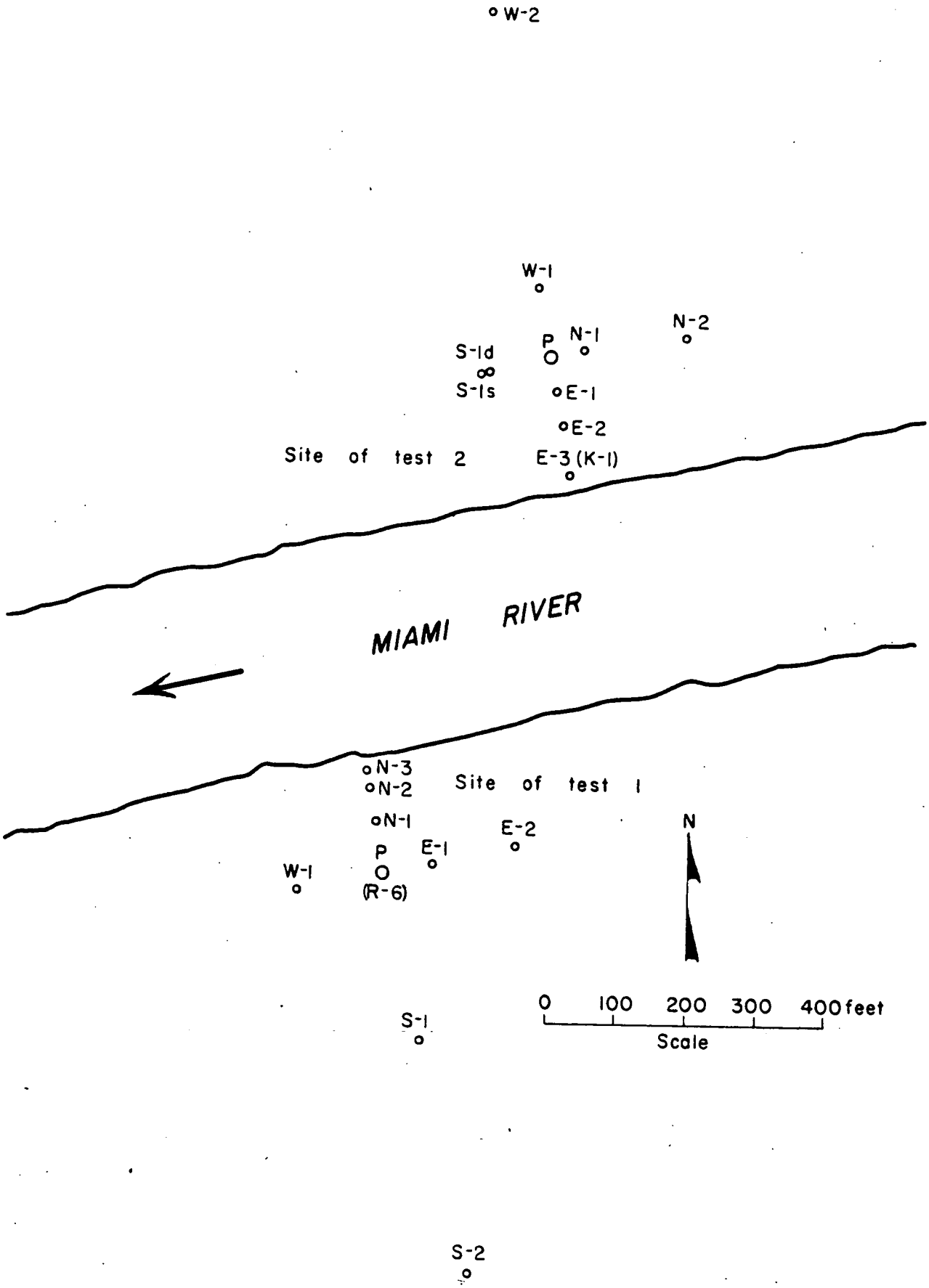

FIGURE 2.- MAP SHOWING LOCATIONS OF WELLS USED DURING CONTROLLED PUMPING TESTS.

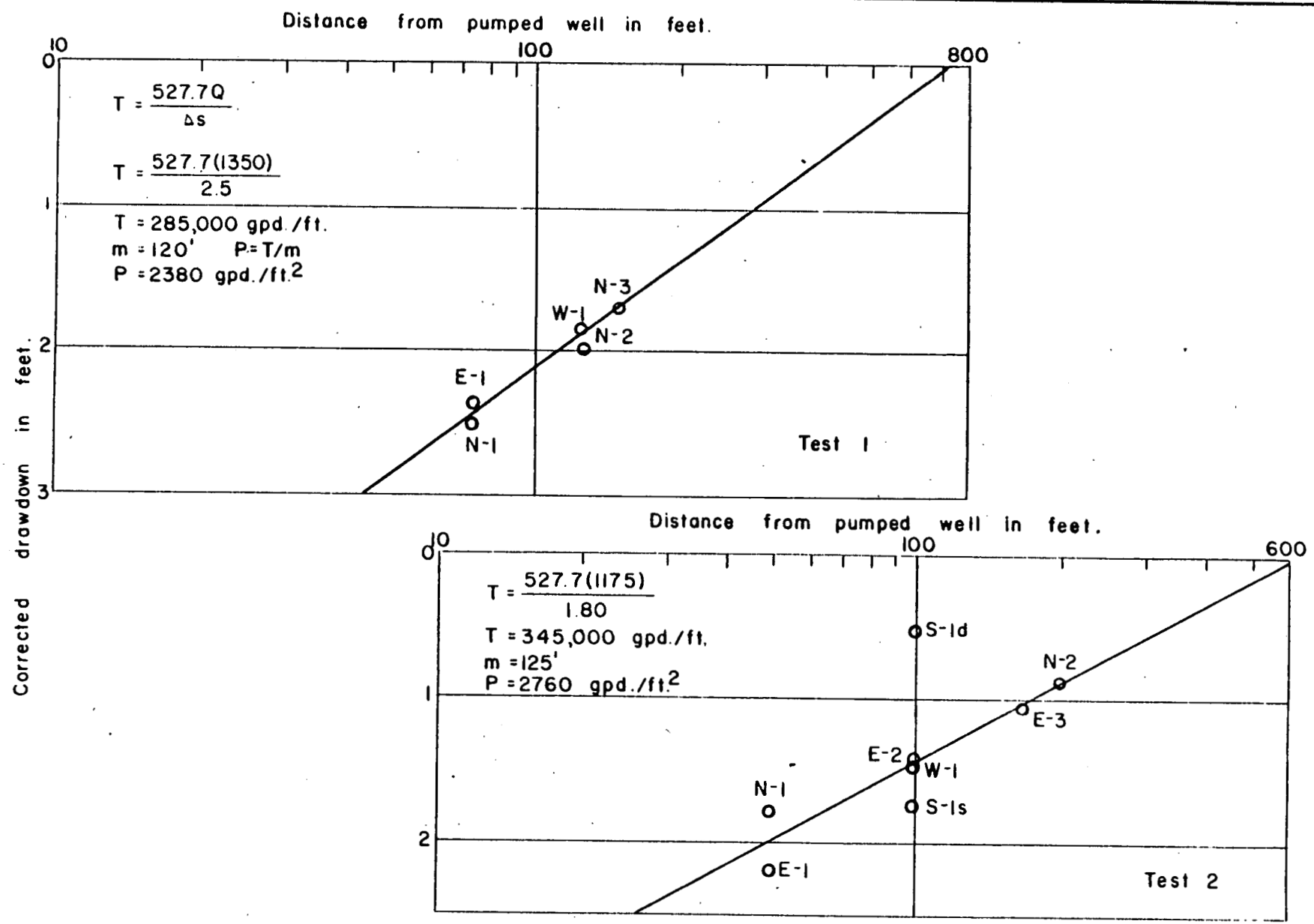

FIGURE 3.-DISTANCE-DRAWDOWN SEMILOG GRAPHS FOR TESTS 1 AND 2.

the straight lines were used to compute the coefficients of transmissibility of the aquifer at the two test sites by using the following equation:

$$T = \frac{527.7Q}{\Delta s} \qquad (1)$$

where:

T = coefficient of transmissibility in gallons per day per foot

Q = discharge of pumped well in gallons per minute

Δs = drawdown difference per log cycle, in feet as determined from semilog graph.

For complete theoretical analyses of ground-water hydraulic formulas see references.

The distance-drawdown data is somewhat scattered, suggesting that the aquifer is heterogeneous and stratified and that the permeability of the sand and gravel deposits varies within the areas of influence of the pumping.

The data obtained during the tests, when analyzed using procedures mentioned above, yielded the following results:

(1.) The coefficients of transmissibility of the aquifer at sites 1 and 2 are 285,000 gpd./ft. and 345,000 gpd./ft. respectively. The coefficients of permeability of the aquifer at sites 1 and 2, are 2380 gpd./sq.ft. and 2760 gpd./sq.ft. respectively. The saturated thickness of the aquifer was about 120 feet at site 1 and about 125 feet at site 2.

(2.) The permeability of the aquifer increases to the north as does the saturated thickness of

the aquifer. The deep channel of the buried valley lies north of site 2 and it is probable that the permeability of the valley-fill deposits increases to a maximum value in the deep channel.

The coefficient of storage of the aquifer can not be determined from the test data because the total values of the drawdowns, which are used in computing S, were affected by recharge from the induced infiltration of river water. A reasonable estimate, based on experience in other areas, of the coefficient of storage of the aquifer is 0.20.

The existence of river infiltration in the areas tested is conclusively shown in the results of the pumping tests. Water levels in the observation wells rose rapidly in response to rises in the stage of the Miami River. In addition, the rate of the decline of water levels decreased more rapidly during the tests than would have been expected if the water had been pumped entirely from storage in the aquifer.

THE WATER TABLE OF THE AQUIFER

The irregular surface below which all the pore spaces of the sand and gravel aquifer are saturated is called the water table. Water levels in 40 observation wells in the Big Bend area were measured on August 31, 1956. The elevations of the water surface in each well was computed to sea level (see appendix sheet) and plotted on a base map prepared by Ayers and Graf, Civil Engineers and Surveyors. The water-

table map shown in plate 1 was then prepared by drawing contours through points having the same water-level elevation. A contour interval of 1 foot was used.

From the water-table map the direction of movement of ground water, the slope of the water table, areas where infiltration of surface water is occurring, and cones of depression caused by pumping water from the two collectors may be determined. Ground-water movement is always in a downslope direction at right angles to the water-table contours as indicated by the arrows in plate 1.

The cones of depression around each collector were studied to check the coefficients of permeability computed from the results of the controlled pumping tests. The contours having elevations of 512 and 513 feet above msl. were selected for analysis because they do not intercept the river and they are not in the zone of the projected laterals of the collectors.

Consider the approximate cylinders of the aquifer as outlined by the two contours with an elevation of 512 feet above msl. All of the water pumped from Collectors 1 and 2 moved through the walls of the two cylinders in accordance with Darcy's law which may be stated as follows:

$$Q = PIA \qquad (2)$$

where:

Q = quantity of water flowing through a given cross-section of an aquifer in gallons per day (discharge of collector)

P = coefficient of permeability, in gallons per day per square foot

I = hydraulic gradient, in ft./ft.

A = cross-sectional area of the saturated portion of an aquifer measured at right angles to the direction of flow, in square feet.

The coefficient of permeability of the aquifer in the vicinity of each collector was determined by substituting appropriate data in equation 2. The hydraulic gradients governing the movement of water through the aquifer were computed by using the following formula:

$$I = \frac{C}{A'/L} \qquad (3)$$

where:

C = water-table contour interval, in feet (1 foot)

A' = area, in square feet, between the 512 and 513 water-table contours

L = average length of the 512 and 513 water-table contours, in feet.

The cross-sectional areas A_1 and A_2 of the aquifer through which the water moved to collectors 1 and 2 were computed by using the following formula:

$$A = Lm \qquad (4)$$

where:

L = average length of the 512 and 513 water-table contours, in feet.

m = average saturated thickness of aquifer, in square feet, at section midway between the 512 and 513 water-table contours as determined from the water-table and bedrock-surface maps of the area.

The coefficients of permeability of the aquifer in the vicinity of Collectors 1 and 2 were computed (see appendix) to be 2440 gpd./sq.ft. and 2780 gpd./sq.ft. respectively by using the formulas mentioned above. These values agree closely

with the coefficients of permeability determined from the results of the two controlled pumping tests and were used to study infiltration as described below.

RECHARGE BY THE INDUCED INFILTRATION OF WATER IN THE MIAMI RIVER

The water-table contours on the map (plate 1) bend in an upstream direction and in a downstream direction, and roughly parallel the river in the oxbow area in a pattern which shows that a ground water mound exists beneath the bed of the river. The slope of the water table between the collectors and the river is much greater than the slopes of the water table landward from the collectors. The water-table map is distorted because the aquifer was receiving large amounts of recharge from the river. Pumping from the collectors lowered ground-water levels below stream level which in turn induced surface water to flow downward through the river bed and into the aquifer.

Contours of the water table north of the Colerain Pike bridge and near the south end of the oxbow bend in an upstream direction, creating valleys in the water table corresponding to the channel of the river. This shows that ground water is moving toward and discharging into the Miami River through seeps and springs north of the bridge and near the south end of the oxbow. Thus, recharge by the induced infiltration of surface water was confined to the reach of the river south of the Colerain Pike bridge and north of the south end of the oxbow.

Flow lines (the path followed by a particle of water as it moves through the aquifer in the direction of decreasing head) were drawn from the river at the north and south ends of the oxbow towards the collectors as shown in plate 1. The four flow lines delimit the section of the aquifer through which the infiltrated water from the river moved towards the areas of withdrawal.

The two sections of the aquifer included between the four limiting flow lines and the 512 and 513 water-table contours were studied in detail to determine the amount of induced infiltration occurring on August 31, 1956.

The hydraulic gradients involved were computed by use of equation 3, in which A' is the area, in square feet, between the two limiting flow lines and the 512 and 513 water-table contours, and L is the average of the lengths of the 512 and 513 water-table contours included between the flow lines, in feet. A planimeter was used to measure the area between the flow lines and contours. Permeability, hydraulic gradient, and cross-sectional data were then substituted in equation 2 and the amount of infiltration was determined. Computations (see appendix) show that approximately 10,550,000 gpd., 62 percent of the amount of water pumped, was infiltrated from the Miami River in the Big Bend area on August 31, 1956.

Flow lines were also drawn from the river at stream bed profiles E-E' and G-G', E-E' and D-D', and G-G' and I-I', as shown on plate 1, towards the collectors. Computations

(see appendix) similar to those outlined above show that recharge on August 31, 1956 by the induced infiltration of surface water from the reaches of river included between profiles E-E and G-G', E-E' and D-D', and G-G' and I-I' amounted to 6,250,000 gpd., 1,280,000 gpd., and 965,000 gpd. respectively.

Infiltration rates

The average infiltration rates of the river bed along the oxbow and along the reaches of river (sections E-E' to G-G', E-E' to D-D', G-G' to I-I', and B-B' to K-K') where recharge was investigated as explained in the preceeding section of this report were computed by using the following formula:

$$I_h = \frac{Q_T}{H\ (L_r W)(43,560)} \qquad (5)$$

where:

I_h = infiltration rate in gallons per day per acre of river bed per foot of head loss

Q_T = quantity of water infiltrated from stretch of river under investigation, in gallons per day.

H = average head loss associated with the infiltration, in feet

L_r = length of the stretch of river under investigation, in feet.

W = average width of the stretch of the river under investigation, in feet.

The soundings made to determine the configuration of the river bed at the eleven sections shown in plate 1 were plotted

on standard graph paper and profiles of the river bed were prepared (see figures 4, 5 and 6). Ground-water levels beneath the river bed, as indicated by the river wells, and the surface of the river, both computed to sea level elevations, were drawn on the profiles. The average head loss (difference between the surface of the river and ground-water levels, or the bed of the river in the case where the water table is below the bed of the river) and the average width of the river at each profile were then determined. The lengths of river between sections were measured by means of a calibrated map measuring instrument.

Computations (see appendix) show that the average infiltration rate of the river bed as determined from a study of infiltration between sections B-B' and K-K' is 168,000 gpd./acre/ft. The average infiltration rate of the river bed varies from one location to another and decreases downstream from the collectors. The infiltration rates for the reaches of river included between sections E-E and G-G', E-E' and D-D', and G-G' and I-I' were computed as 245,000 gpd./acre/ft., 229,000 gpd./acre/ft., and 63,500 gpd./acre/ft. respectively.

The infiltration rates given above were determined by using the average width of the river assuming that the infiltration rate was uniform over the entire river bed. However, the infiltration rate varies across the river bed. Most of the flow of the river during dry periods is concentrated in a deep channel less than 100 feet wide as shown by the river-bed profiles in figure 4, 5 and 6. The velocity of the

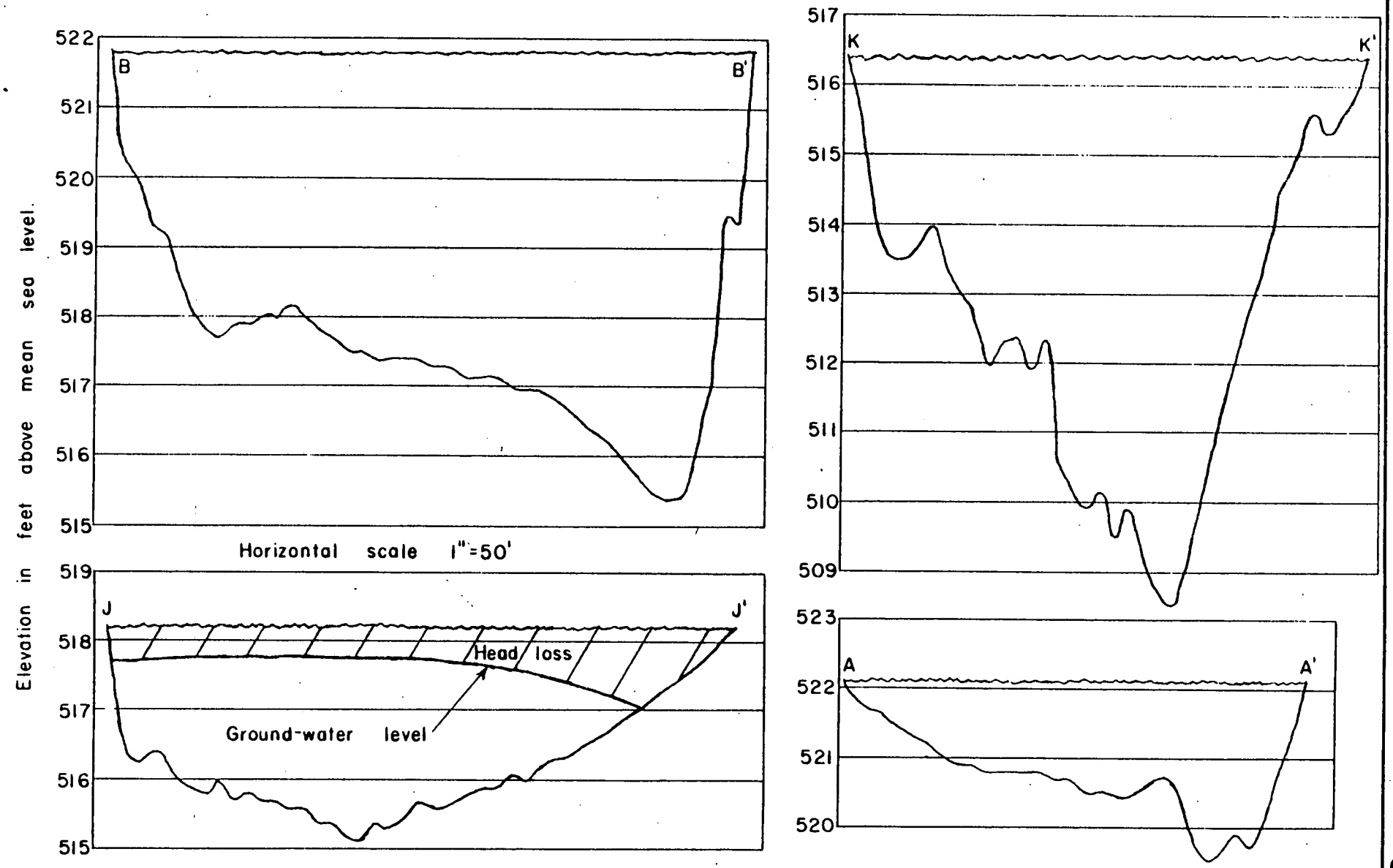

FIGURE 4.- PROFILES SHOWING THE CONFIGURATION OF THE BED OF THE MIAMI RIVER IN THE BIG BEND AREA, A-A', B-B', J-J' AND K-K'.

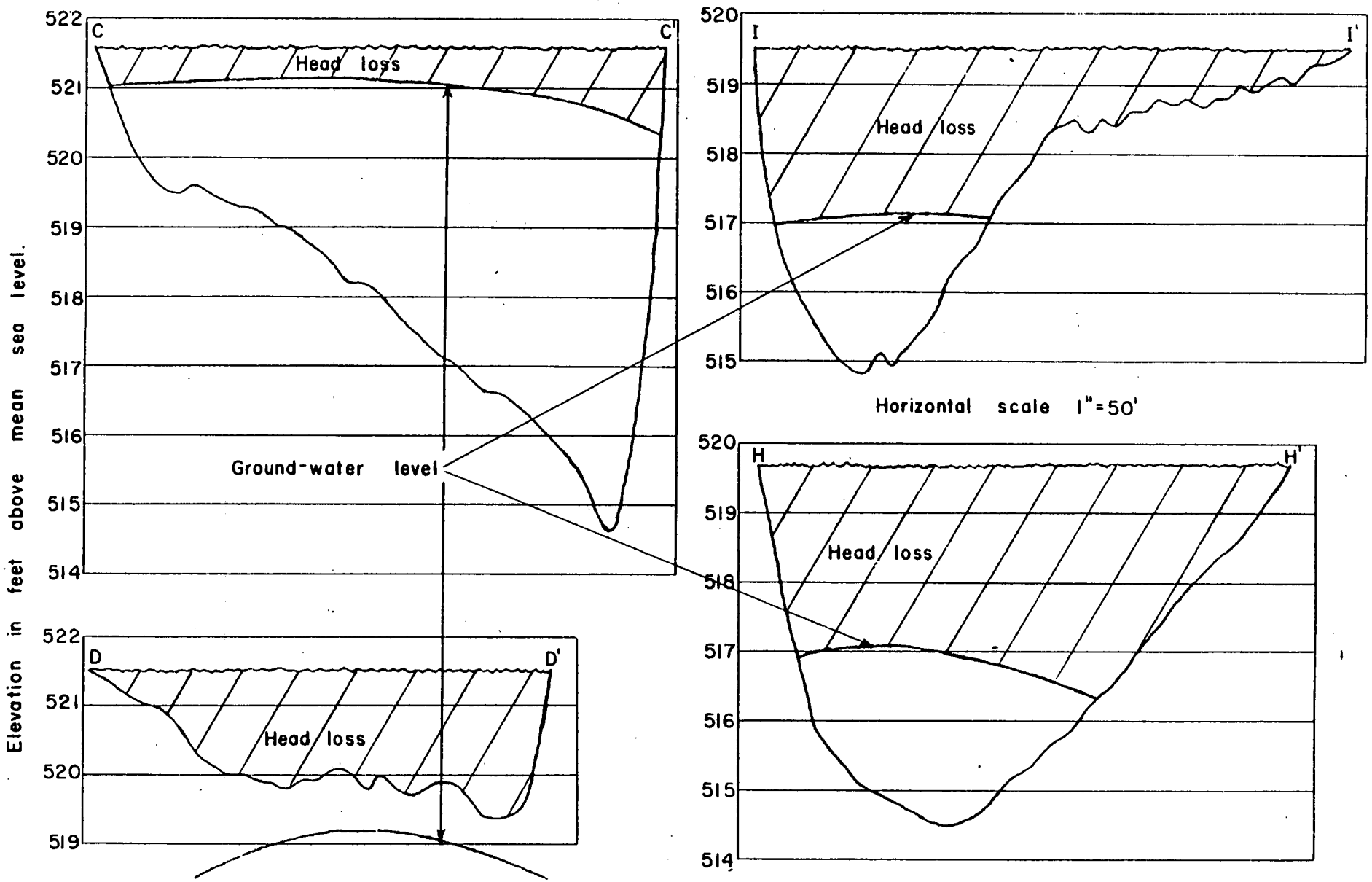

FIGURE 5.- PROFILES SHOWING THE CONFIGURATION OF THE BED OF THE MIAMI RIVER IN THE BIG BEND AREA, C-C', D-D', H-H' AND I-I'.

6089

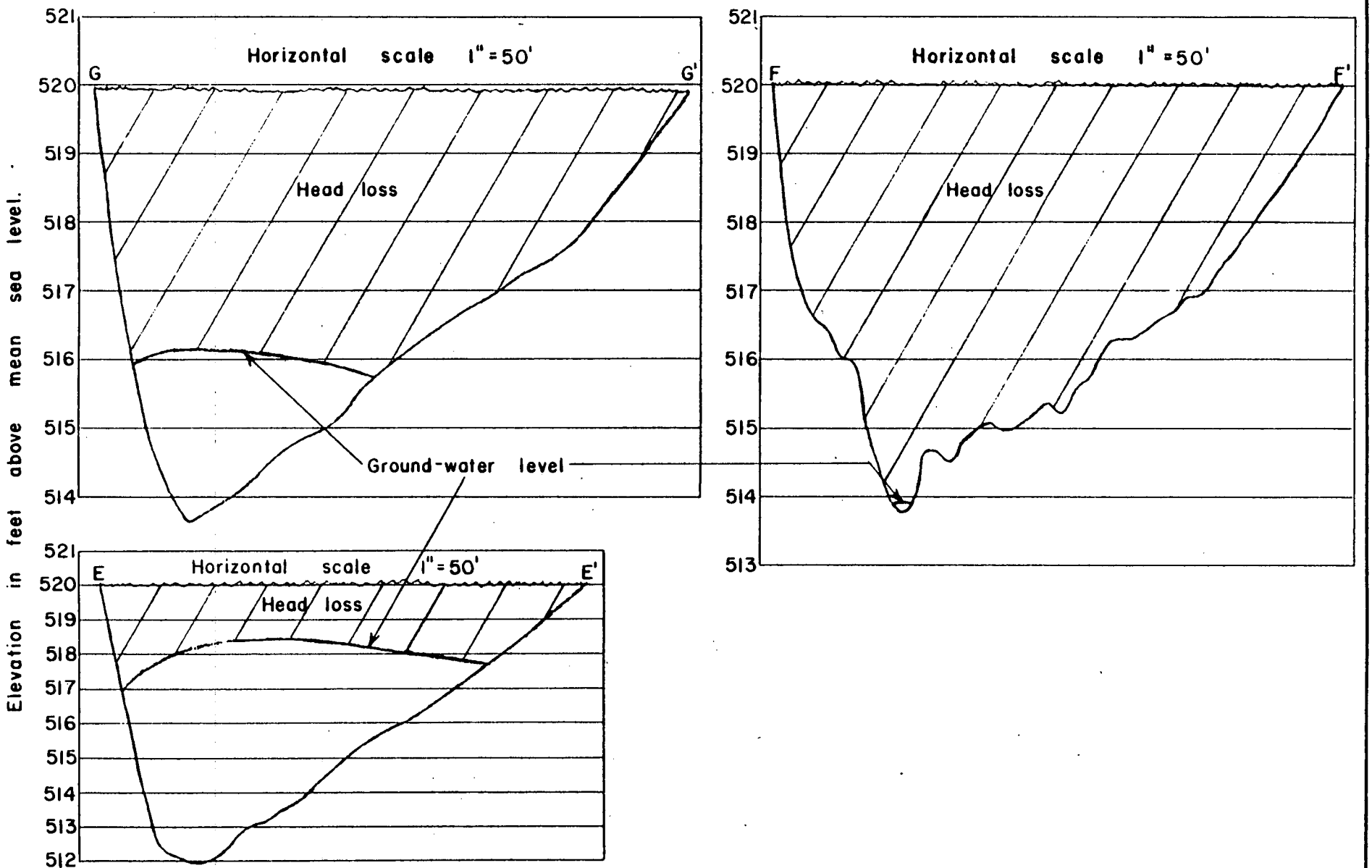

FIGURE 6.- PROFILES SHOWING THE CONFIGURATION OF THE BED OF THE MIAMI RIVER IN THE BIG BEND AREA, E-E', F-F' AND G-G'.

000029

6089

flow in the deep channel is high enough to prevent large amounts of silt and sediment from settling from the moving waters. Conditions are therefore favorable for a high infiltration rate in the deep channel. However, the flow is sluggish in the shallow part of the river, which is over 100 feet wide and about 2 feet deep. Fine materials settle from the slowly moving waters and these materials become more or less consolidated through the interlocking of particles and the cementing action of colloidal materials. The deposition of fine materials on the river bed in the shallow part of the river tends to reduce its permeability. The infiltration rate of the river bed in the shallow part of river is therefore only a fraction of the infiltration rate of the river bed in the deep channel.

The stretch of the river between sections E-E' and G-G' was studied to determine the infiltration rates of the shallow and deep parts of the river channel. Most of the flow at sections E-E', F-F', and G-G' is concentrated in a deep channel approximately 95 feet wide. The amount of water infiltrated between sections E-E' and G-G' was computed, as stated in the preceeding section of this report, to be about 6,250,000 gpd. It is estimated that about 5,890,000 gpd. was infiltrated from the deep channel. The average head loss associated with the above mentioned infiltration is 3.4 feet. Computations (see appendix) made using equation 5 show that the infiltration rate of the river bed in the deep channel is about 415,000 gpd./acre/ft. The infiltration rate of the

river bed in the shallow part is estimated to be about 40,000 gpd./acre/ft.

The depth and width of the river in the Big Bend area during periods of low streamflow is controlled by two natural dams (rapids occur in the vicinity of sections D-D' and I-I'). The relatively low average infiltration rate of the river bed results from the ponded condition behind these rapids.

POTENTIAL YIELD OF THE AQUIFER

The magnitude of potential recharge by the induced infiltration of surface water was estimated by correlating data on the average depths of water in the river with data on the average infiltration rate of the river bed. The results of the controlled pumping tests, and of geologic and hydrologic studies made in the Big Bend area, were used to predict future pumping levels in collectors resulting from assumed pumping conditions.

The maximum amount of infiltration will occur along the portions of the river where the water table has been lowered below the river bed and all of the head of the surface water has been used. The maximum potential recharge by induced infiltration is the product of the infiltration rate per foot of head (gpd./acre/ft/), the area of river bed being infiltrated (acres), and the average depth of water in the river (head in feet). The smallest amounts of recharge will occur during extended dry periods, therefore, the average depths of water in the river which occur in prolonged low-flow

periods must be used to compute the potential infiltration.

Fortunately, streamflow records for the Miami River were collected during the recent unusual dry period, 1952 through 1954, by The Southwestern Ohio Water Company at the Colerain Pike bridge. In addition, daily gage height readings for the dry period are available for the gaging station at Hamilton, Ohio which is operated by the Miami Conservancy District in cooperation with the U.S.Geological Survey. The data collected in the period 1952 through 1954 at the Colerain Pike bridge, is given in the appendix. Daily gage height readings for the year 1954 at Hamilton are shown in figure 7.

The profiles of the river channel shown in figures 4, 5, 6 and 8 were used to determine the average depth of water in the river and the average width of the river for different stages of the Miami River as recorded at the Colerain Pike bridge. Graphs of the elevation of the river at the Colerain Pike bridge versus the average depth of water in the river and the average width of the river were then prepared (see figure 9). The relation between gage height readings at Hamilton and the elevation of the river at the Colerain Pike bridge was established by comparing the data collected by The Southwestern Ohio Water Company and by the U.S.Geological Survey. The graph of gage readings at Hamilton versus the average depth of the river in the Big Bend area was then prepared (see figure 10) so that the more complete records collected at Hamilton could be used to compute potential infiltration. Considering the length of river south of the

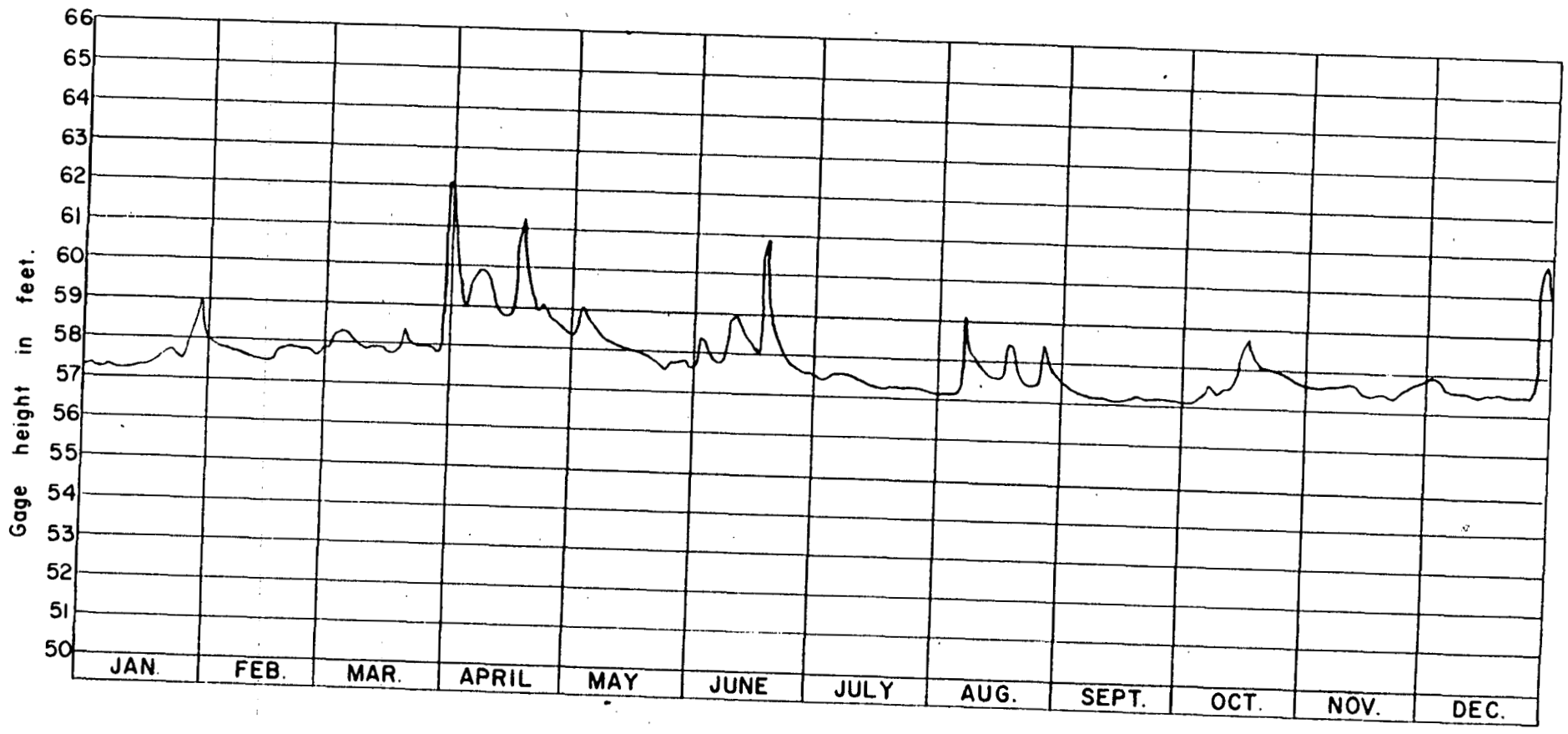

FIGURE 7.-MEAN DAILY GAGE HEIGHTS OF THE MIAMI RIVER AT HAMILTON, OHIO DURING 1954.

000023

6089

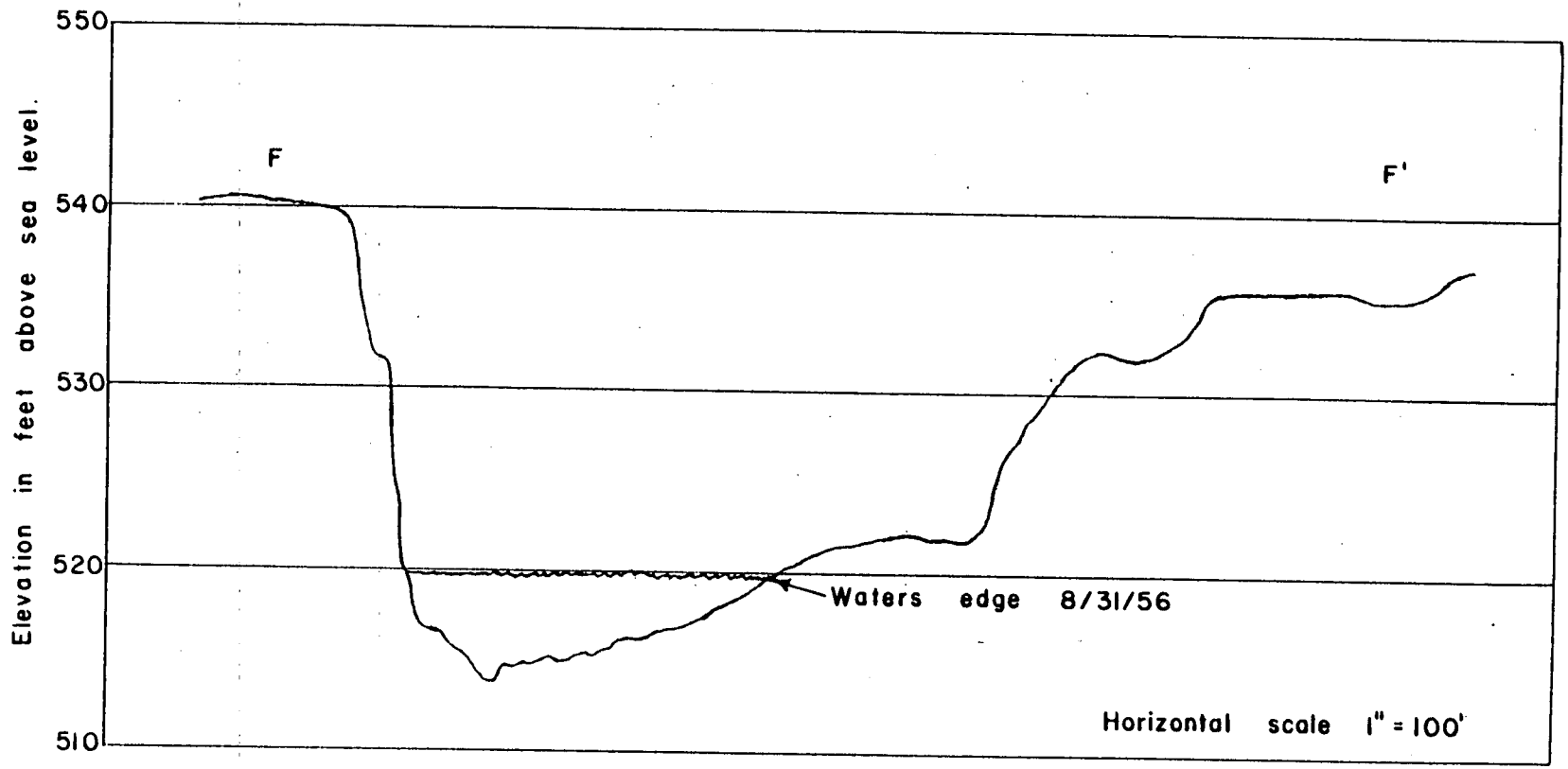

FIGURE 8.-PROFILE OF RIVER CHANNEL ALONG LINE F-F'.

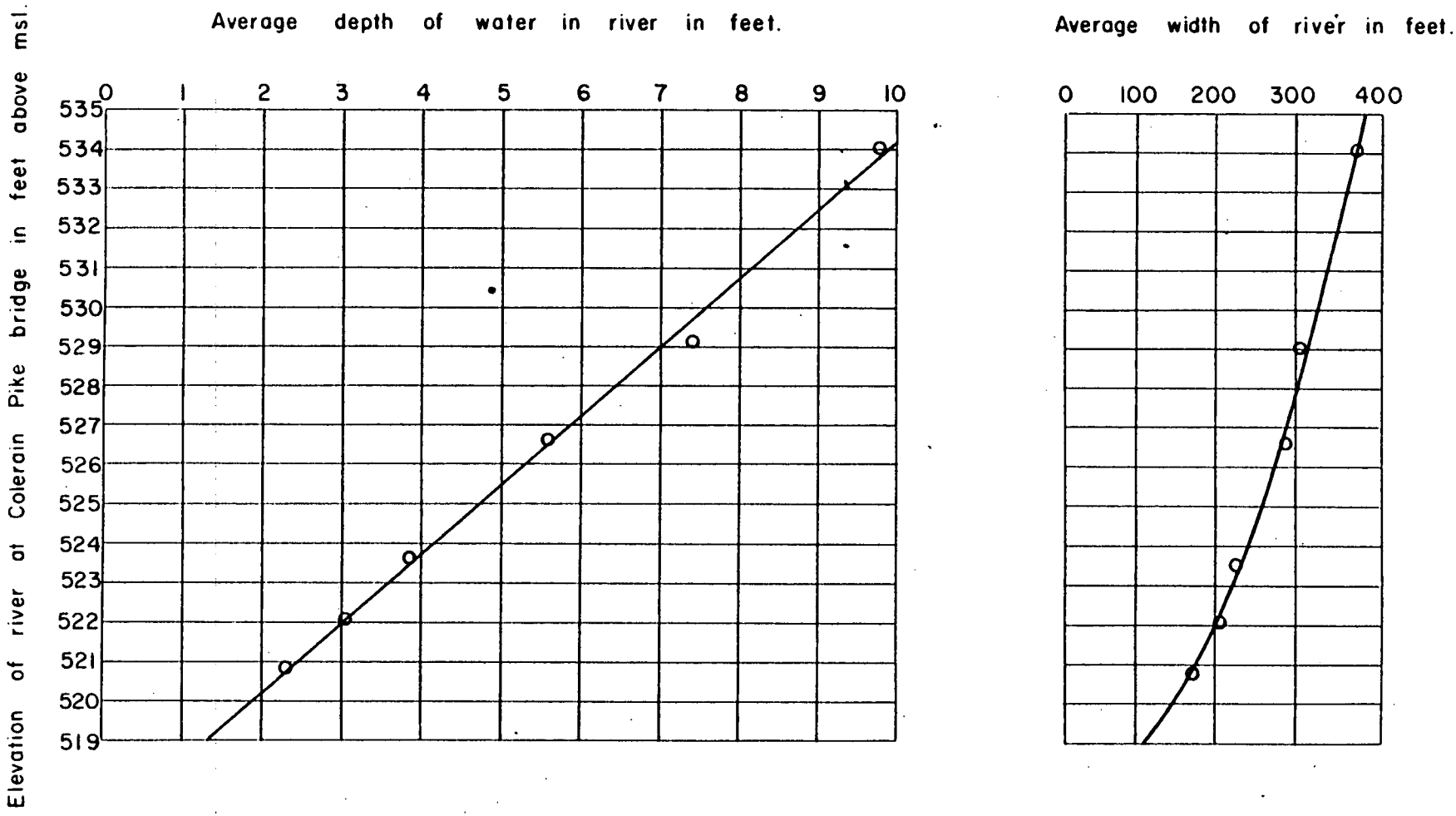

FIGURE 9.-GRAPHS SHOWING THE AVERAGE DEPTH AND WIDTH OF THE MIAMI RIVER FOR VARIOUS STAGES OF THE RIVER.

000035

6089

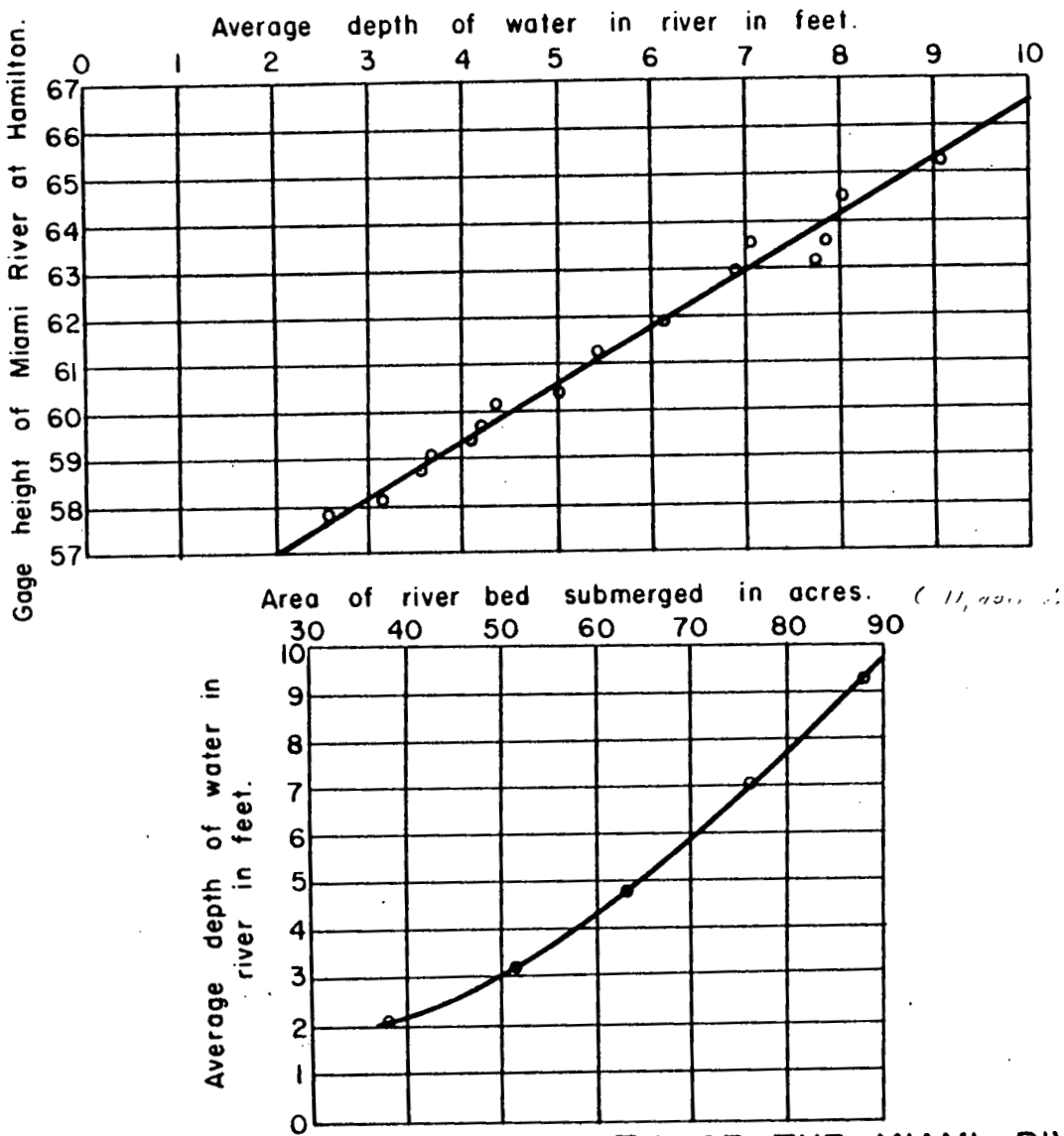

FIGURE 10.-GRAPHS SHOWING THE AVERAGE DEPTH OF THE MIAMI RIVER AND THE AVERAGE AREA OF THE RIVER BED SUBMERGED FOR VARIOUS STAGES OF THE RIVER.

000036

6089

Colerain Pike bridge available for infiltration, 11,000 feet based on geologic data, a graph of average depth of water in the river versus acres of river bed submerged was prepared as shown in figure 10.

Infiltration varies with the viscosity of the surface water. A decrease in the temperature of the water increases its viscosity and thereby decreases infiltration. A decline in temperature of one degree will decrease infiltration about 1.5 percent. The infiltration rate computed from test data was determined in August when the average temperature of the river water was about 82°F. During the winter months the temperature of the river water declines to about 39°F according to records collected by the Champion Paper and Fibre Company at Hamilton (see Rose, 1951 in references). Thus, considering only the change in viscosity, the infiltration rate per foot of head on the river bed during the winter months will be only about 52 percent of the infiltration rate per foot of head measured during the summer months.

Streamflow records, the graphs in figures 9 and 10, and temperature data were used to compute the maximum potential recharge by the induced infiltration of surface water from the reach of the Miami River south of the Colerain Pike bridge for each day of the years 1952 through 1954 (see appendix). The data mentioned above were substituted in the following formula:

$$R = I_h \left(\frac{\mu_a}{\mu_t}\right) H_a A \qquad (6)$$

where:

R = potential recharge by the induced infiltration of surface water in gallons per day

I_h = average infiltration rate of riverbed as determined from data collected August 31, 1956, 168,000 gpd./acre/ft.

μ_t = coefficient of viscosity at temperature of river during period of computation

μ_a = coefficient of viscosity at temperature of river during August, 1956

H_a = average depth of water at stage of river being investigated in feet

A = area of river bed submerged during period of computation, in acres

Computations show that the average maximum amount of recharge to be expected during a prolonged dry period similar to 1952 through 1954 from the two-mile reach of the river south of the Colerain Pike bridge in the Big Bend area is about 24 mgd. (see appendix)

The cone of depression created by pumping large quantities of water from collectors located south of the Colerain Pike bridge would spread at least 1-1/2 miles upstream from the bridge. Ground-water levels would be lowered beneath the level of the river and infiltration of surface water would occur upstream from the bridge. Maximum infiltration would probably be limited to the reach of the river about 3/4 mile northeast of the bridge where the water table would decline below the bed of the river. Infiltration in reduced amounts would occur further upstream where the cone of depression would be relatively shallow. A reasonable estimate

of the potential recharge from the stretch of the river mentioned above during extended dry periods is about 12 mgd., 1/2 of the computed potential recharge for the stretch of the river south of the bridge.

The average amounts of recharge to be expected in the Big Bend area during dry periods are given below:

Induced infiltration from stretch of river southwest of Colerain Pike bridge.....	24 mgd.
Induced infiltration from stretch of river northeast of Colerain Pike bridge.....	12 mgd.
Recharge from precipitation.....	3 mgd.
Flow from valley walls (7 miles of valley wall. Flow estimated at 150,000 gpd. per mile.....	1 mgd.
Total	40 mgd.

Recharge by the infiltration of water from Dry Run during dry periods would be very small and because it can not be estimated with any degree of accuracy it is omitted from the inventory. The potential yield of the aquifer in the Big Bend area during extended dry periods is therefore estimated to be in the magnitude of 40 mgd.

In computing potential infiltration it was assumed that the infiltration rate remained constant for low and high flows and that the infiltration rate computed from data collected in August 1956 would apply to the dry periods recorded 1952 through 1954. Because most of the flow in the river is concentrated in a deep channel less than 1/2 the width of the

river and the flow in river during August 1956 was almost as low as the flows recorded during the dry period, the writers believe that the opportunity for fine silty materials to settle on the river bed reducing its permeability was about as great in August 1956 as it was during 1952, 1953 and 1954. During dry periods the velocity of flow in the deep channel is sufficient to prevent deposition of large amounts of sediment. The river bed in the shallow part of the river where most of the deposition of fine materials occurs with low flows contributes very little recharge during most summer months. Thus, even if the infiltration rate of the shallow part of the river was lower in 1952, 1953 and 1954 than it was in August 1956, no serious error would be introduced in the computations made in this report. At times of high river flows fine sediments are scoured from the river-bed and the infiltration rate is increased. However, very few high river flows occurred during the extended dry period. The computations made by using a low-flow infiltration rate are therefore not overly conservative.

Streamflow records show that the discharge of the Miami River is more than enough to replenish the large withdrawals of ground water which can be obtained from induced infiltration through the river bed in the Big Bend area. A summary of the streamflow records (Norris, Cross and Goldthwait, 1948) of the Miami River at Venice as given below shows that even on the day of lowest recorded flow (9/8/25) the flow was 232 cfs. which is about 150 mgd.

Drainage area (square miles)	3,790	
Period of record	1915-1927	
Mean Discharge (cfs.)		
Highest day	55,600	(4-22-20)
Highest month	16,900	(Jan.1916)
Highest year	5,880	(1927)
Lowest day	232	(9-8-25)
Lowest month	480	(Oct.1923)
Lowest year	1680	(1925)
Minimum Momentary Discharge (cfs)	225	(9-8-25)

The water-level data collected to date show that the yield of the aquifer in the Big Bend area is much greater than present pumping rates. The difference between the maximum and minimum water levels in observation well H-2 in the period 1952 to 1956 (see figure 11) is less than 9 feet which is small indeed when compared to the average saturated thickness of the aquifer, over 100 feet. Water levels in well H-2 leveled off during the dry period as recharge by the induced infiltration of surface water balanced discharge.

Determination of drawdowns to be expected in a system of collectors designed to pump the amount of water available in the Big Bend area were made by substituting a hydraulic system amenable to mathematical treatment for the complicated hydrogeologic system of the area under investigation and by using the image-well theory (Ferris, 1951). The yield of the aquifer during dry periods was used in the computations.

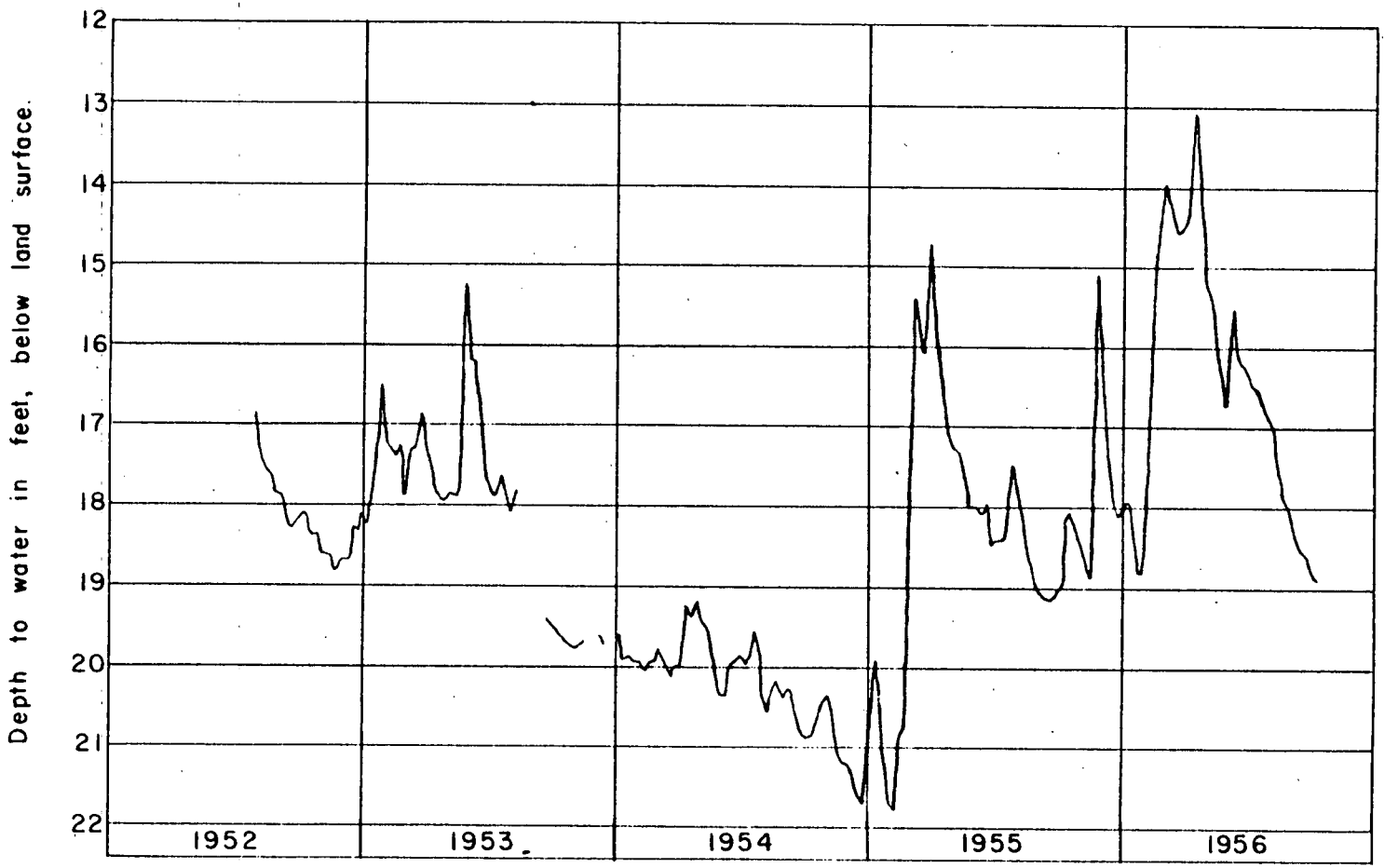

FIGURE 11.-WATER LEVELS IN OBSERVATION WELL H-2, 1952 TO 1956.

000042

6089

Recharge during the fall and winter months of years similar to 1952, 1953 and 1954 (see appendix) will be less than the average recharge, 40 mgd., computed for the entire dry period and will smount to about 24 mgd. Thus, approximately 16 mgd. must be withdrawn from storage during the fall and winter months of each dry year to balance a pumping rate of 40 mgd.

Assume that two additional collectors are constructed at distances of 4000 feet northeast and southwest of existing Collector 2, as shown on the computation sheet in the appendix, and that 40 mgd. are pumped from four collectors. The magnitude of the drawdown in Collector 2 at the end of a prolonged dry period as a result of the withdrawal of 40 mgd. was computed by using the hydraulic system shown in the appendix. Two discharging image wells with strengths of 16 mgd. each were used to imitate the combined effects of the two valley walls (limited areal extent of the aquifer) and partial recharge (24 mgd.) on the drawdown in Collector 2.

The nonequilibrium formula (Theis, 1935), the average value of T computed from the controlled pumping test data described earlier in this report, and the estimated value of S were used to construct the theoretical distance-drawdown curve given in figure 12. A value of 400 days, the approximate accumulative number of days when discharge exceeds recharge during an extended dry period similar to 1952 through 1954, was substituted for t in the nonequilibrium formula given below.

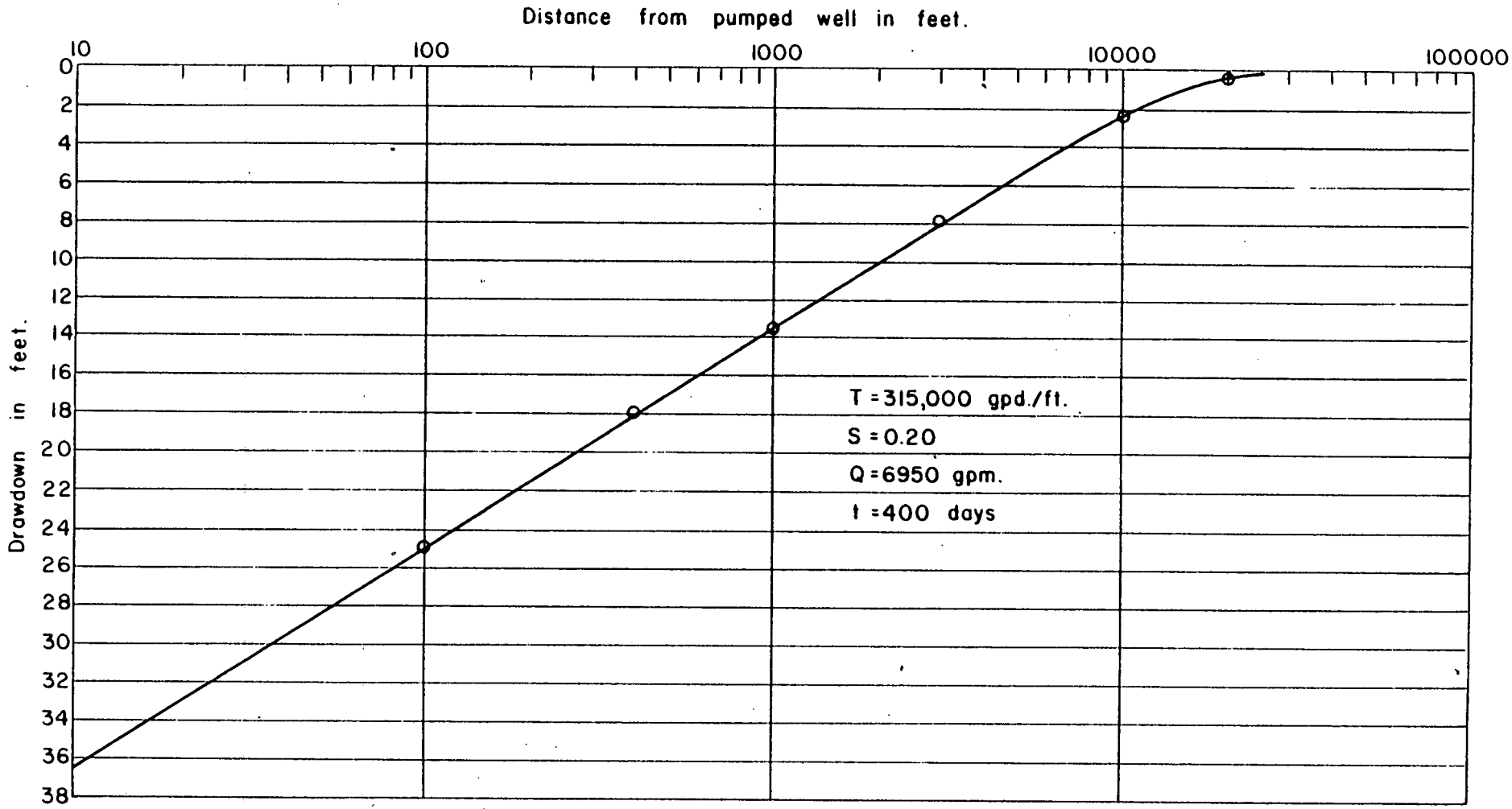

FIGURE 12.- THEORETICAL DISTANCE-DRAWDOWN GRAPH.

000044

6089

$$s = \frac{114.6Q}{T} W(u)$$

where:

$$W(u) = \int_u^{\infty} \frac{e^{-u}}{u} du$$

and $$u = \frac{1.87r^2S}{Tt}$$

t = time in days since pumping started

s = drawdown at any point in area of influence of pumping, in feet

Q = discharge of pumped well, in gallons per minute

T = coefficient of transmissibility in gallons per day per foot

S = coefficient of storage, as a ratio or decimal

r = distance of point in area of influence from pumped well in feet, or if the analysis is being made for a pumped well, r is the effective radius of the pumped well in feet.

The net drawdown, s_T, in Collector 2 as a result of pumping 40 mgd. from 4 collectors during a dry period (includes the withdrawal of water from storage in the aquifer as mentioned earlier) is equal to the drawdown due to pumping the collector itself, plus the interference due to pumping the 3 other collectors, plus the net drawdown due to the two discharging image wells associated with the valley walls

minus the recharge effects. The net drawdown, s_T, must be increased to compensate for the dewatering of the aquifer by the quantity $s_T^2/2m$, where m is the original saturated thickness of the aquifer.

Computations (see appendix) made by using figure 12 show that the drawdown in Collector 2 caused by development of a 40 mgd. water supply in the Big Bend area under adverse conditions will be about 73 feet. The pumping level would be about 30 feet above the top of the screens in the collector.

The coefficient of transmissibility of the aquifer is high and as a result, the cone of depression created by pumping (see figure 12) has a relatively flat slope. Large quantities of water can be transmitted from distant points to a pumping collector under the influence of a low hydraulic gradient. Thus, during dry periods when large amounts of water are taken from storage, the cone will spread to relatively great distances from the collectors. For example, computations show that water levels would be lowered about 29 feet one mile northeast of Collector 2 and about 23 feet 1-1/4 miles southwest of Collector 2.

CONCLUSIONS

Based on the study of existing geologic and hydrologic data described in the report the potential ground water yield of the Big Bend area is estimated to be 40 million gallons a day in dry periods. The record dry period studied in detail as described in the report included the years 1952, 1953 and 1954. The computations described in the report

show that withdrawals from collectors averaging 40 mgd. would have been balanced by recharge in this period, chiefly by induced infiltration from the Miami River. In this period the recharge at times would have averaged as little as 24 mgd. However, the computations described in the report show that the storage available in the aquifer would have been more than adequate to sustain the 40 mgd. pumping rate through the minimum infiltration rate periods, and until replacement of the lost storage in higher infiltration rate periods.

For computation purposes, a four-collector system was used to check the adequacy of the available storage, as described in the report. Further study should be given, however, to proper location of new collectors to insure that efficient operation of the collector field as a whole will be obtained.

As large amounts of water are pumped from storage in the aquifer in dry periods the combined cone of influence of the pumping will extend over a relatively large area. The computations show for example that water levels in the aquifer will be lowered as much as 29 feet one mile northeast of existing Collector 2 and about 23 feet at a distance 1-1/4 miles southwest of Collector 2.

All predictions in the report were made on the assumption that the well field will be limited to the area south of the Butler-Hamilton County line. It should be emphasized that the predicted 40 mgd. yield is the minimum yield which will be obtained in very dry periods. In years of normal or

above normal precipitation the yield of the aquifer will be much greater and the declines in ground water levels for the 40 mgd. pumping rate will be less than given in the report.

It should be noted that the potential yield of the well field area computed in this report (40 mgd.) is much less than the yield (63 mgd.) estimated in the progress report previously submitted by the writers, because the adverse effects of the natural dams in the river, which were unknown before the present investigation, were not considered in computations made in the previous report.

Columbus, Ohio
November 24, 1956

Schaefer & Walton
Consulting Ground Water Hydrologists
by

William C. Walton

William C. Walton

and

Edward J. Schaefer

Edward J. Schaefer

References

Darcy, Henri, 1856, Les Fontaines publiques de la ville de Dijon, Paris

Ferris, J.G., 1951, in Wisler, C.O. and Brater, E.F., Hydrology, Chap. VII, Ground Water: John Wiley and Sons, Inc., New York.

Jacob, C.E., 1945, Partial penetration of pumping well, adjustments for: U.S.Geological Survey Water Resources Bulletin, p. 169-175, August.

Kazmann, R.G., 1950, Memorandum report on Hydrogeological Survey for Southwestern Ohio Water Company Big Bend area, Miami River Hamilton County, Ohio

Klaer, F.H. Jr., and Thompson, D.G., 1948, Ground-water resources of the Cincinnati area Butler and Hamilton Counties, Ohio: U.S.Geological Survey Water-Supply Paper 999.

Norris, S.E., Cross, R.P., and Goldthwait, R.P., 1948, The water resources of Montgomery County, Ohio: Bulletin 12, Ohio Water Resources Board, Columbus, Ohio.

Rose, N.A., 1951, Results of investigation for Ground-Water development at Big Bend well field Hamilton County, Ohio

Schaefer & Walton, 1956, First progress report on results of test-well drilling and observation well program: Columbus, Ohio.

Theis, C.V., 1935, The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using ground-water storage: Am. Geophys. Union Trans., pt.2, p. 519-524.

Wenzel, L.K., 1942, Methods for determining permeability of water-bearing materials: U.S.Geological Survey Water-Supply Paper 887.