STATE OF OHIO
DEPARTMENT OF NATURAL RESOURCES
DIVISION OF WATER

VERTICAL LEAKAGE THROUGH TILL AS A SOURCE OF RECHARGE TO A BURIED-VALLEY AQUIFER AT DAYTON, OHIO

TECHNICAL REPORT No. 2

Prepared in cooperation with the
Water Resources Division, U.S. Geological Survey,
and the Miami Conservancy District

Columbus, 1959

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By

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ABSTRACT

The rate of leakage through a bed of glacial till, which separates two sand and gravel aquifers in a buried valley at Dayton, Ohio, was calculated from a pumping test run in 1953, before pumping from the lower aquifer became appreciable. The median value (0.003 gpd/ft^3) of the leakage coefficient $\left(\frac{P'}{m'}\right)$ is checked (1) by relating it to the quantity of water subsequently pumped from the well field, the estimated areal extent of the cone of depression, and the difference in water levels between the upper and lower aquifers and (2) by computing the drawdown at a point where water-level data are available and comparing the computed value with the observed drawdown.

INTRODUCTION

The principal source of large ground-water supplies in Ohio is sand and gravel of glacial origin, deposited as outwash in bedrock valleys and traversed by modern streams. Of prime consideration in an evaluation of water-supply potential at selected sites is the rate of percolation through the stream bed and various methods have been used to determine this rate under field conditions. Not generally considered, however, by those responsible for the design of water-supply facilities is the fact that in most buried valleys the outwash deposits were laid down during more than one glacial stage. As a consequence, these generally permeable deposits are commonly interbedded with layers of till which for practical purposes separate the sand and gravel into two or more aquifers. Where the bedrock is relatively impermeable most recharge to the deeper aquifers must come from leakage through the till layers, and, therefore, the rate of percolation through the till is highly important. Very little information on such leakage is available, partly because the effects of leakage closely resemble the boundary effects produced by an infiltrating stream, and often are mistaken for the latter in pumping-test analysis. Thus, leakage may go unrecognized and its effects become the basis for serious misinterpretation.

The importance of leakage through glacial till, as a source of water to a sand and gravel aquifer, is illustrated by the results of an investigation at Dayton, Ohio, made by the U. S. Geological Survey in cooperation with the Miami Conservancy District and the Ohio Department of Natural Resources, Division of Water. At the Dayton municipal well field, where the geology of the deposits in the Mad River Valley is fairly typical of outwash-filled valleys in Ohio, recharge to an important aquifer is derived principally by leakage through a thick layer of till. The rate of leakage through this till layer is critically important to Dayton's available ground-water supply.

The Dayton Municipal Water Supply

The Dayton municipal water supply is pumped from wells drilled in glacial outwash deposits in the Mad River valley at a site about 5 miles northeast of town, centering around Rohrers Island, a 180-acre tract which splits the stream into two principal channels (fig. 1). Pumpage in 1957 averaged about 46 million gallons a day (mgd), making this the largest municipal ground-water supply in Ohio and one of the largest in the Midwest.

The Mad River in the vicinity of the municipal well field flows in a preglacial valley more than 2 miles wide, cut into shale of the Richmond group, of Ordovician age. The valley fill includes two layers of sand and gravel having a maximum total thickness of 200 feet, separated by a layer of till ranging in thickness from about 11 to 50 feet, which extends all the way across the valley as a continuous sheet (fig. 2). The till layer acts hydraulically as a confining bed and separates the sand and gravel deposits into two aquifers (Norris and others, 1948, p. 52). At the well field the average thickness of the upper and lower aquifers is about 65 and 50 feet, respectively. The confining bed is commonly reported by drillers as "clay" or "clay and gravel," and sometimes "hardpan." The upper aquifer is interbedded with discontinuous lenses of till, but on the basis of well performance is known to have a somewhat greater permeability than the lower aquifer. The lower aquifer is underlain generally by thin till, which separates it from the bedrock. Locally, the sand and gravel lies directly on the bedrock.

The upper aquifer is the main source of the Dayton municipal supply. Most of the water is pumped from 24 wells on Rohrers Island which average 75 feet in depth

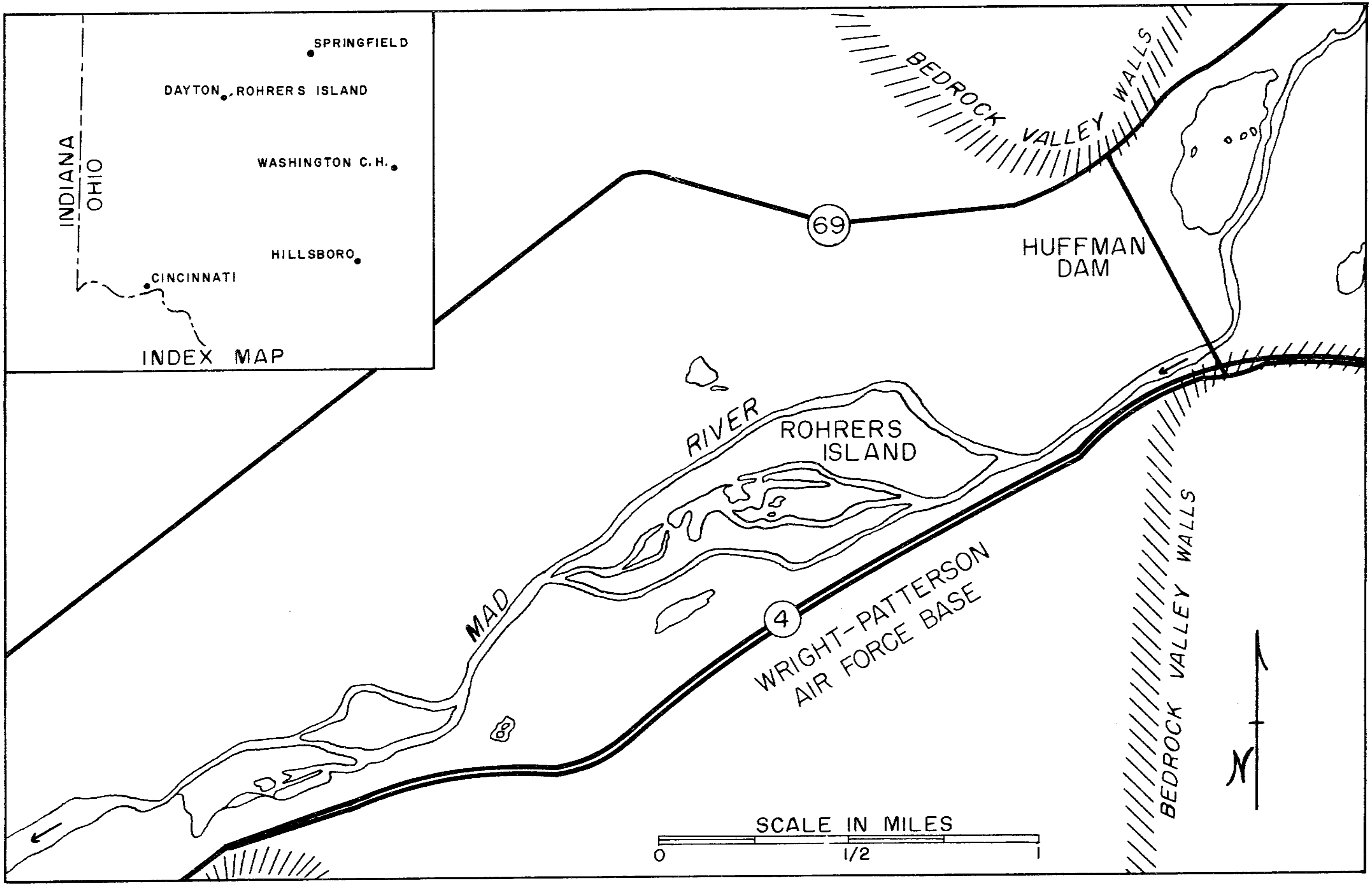

Figure I. Map of the Mad River in the vicinity of Rohrers Island, showing approximate boundaries of the bedrock valley.

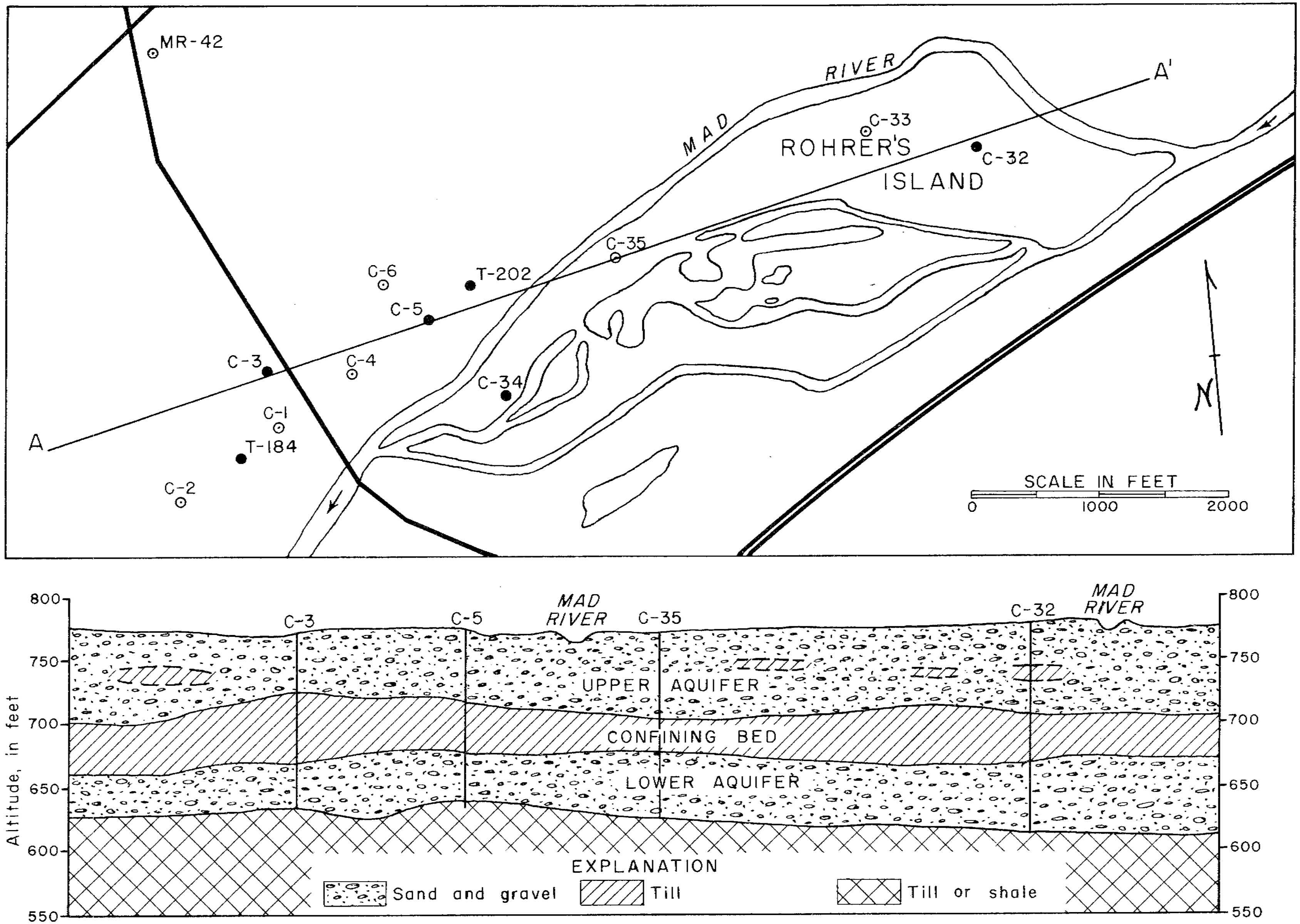

Figure 2. Map showing location of wells tapping lower aquifer, and generalized cross section through the Dayton municipal well field. (Wells are designated by numbered circles. Numbers prefixed by letter C denote supply wells; prefixes MR and T denote observation wells. Solid circles indicate wells used in October, 1953 pumping test.)

and are recharged principally by induced infiltration from the Mad River, through the river bed or through the beds of lagoons and ditches which have been dug on Rohrers Island. The lagoons and ditches are cleaned periodically by dredging, and the water level in the upper aquifer, which remains relatively constant, is a few feet below the level of the river.

Development of the lower aquifer was delayed until recent years because the water is harder and contains more iron than water from the upper aquifer. Prior to the construction of a softening plant in 1953, only 4 wells had been drilled into the lower aquifer, and these wells were kept chiefly in reserve. After the softening plant was built 6 additional wells were drilled into the lower aquifer, which then became important as a source of supply. By 1957 between 12 and 15 mgd was being pumped from the lower aquifer.

The lower aquifer is recharged principally with water which leaks downward through the confining bed and, to a lesser extent, by underflow moving down the valley in response to natural hydraulic gradients. No significant amount of recharge is contributed by the relatively impermeable shale which forms the walls and floor of the buried valley.

HYDRAULIC PROPERTIES OF THE LOWER AQUIFER

In October 1953, personnel of the U. S. Geological Survey made a pumping test, using 6 wells drilled into the lower aquifer. Figure 2 shows the location of the 6 wells used in the pumping test; 4 of the wells are on the north side of the Mad River, opposite the lower end of Rohrers Island, and the other 2 are on Rohrers Island. Table 1 lists the drillers' logs of the wells.

Pumping from the lower aquifer was stopped approximately 18 hours before the test started to allow water levels to stabilize. Pumping from well C-5 was begun at 10:43 a.m. on October 28, 1953, and was continued at an average rate of about 2,300 gpm until 1:43 p.m. on October 29, a period of 27 hours. Recorders were operated on three wells, C-3, T-184, and T-202. Water levels in two other wells, C-32 and C-34, were measured periodically with a steel tape.

Graphs of water-level fluctuations in four of the observation wells caused by pumping from well C-5 are shown in fig. 3. The graphs reveal that several hours after pumping started there was no further drawdown. This leveling-off resulted from recharge which effectively balanced the pumpage. A slight rising trend, shown in most of the graphs in the latter part of the pumping period, was produced by interference effects resulting from operation of the shallow wells, screened above the confining bed.

The coefficients of transmissibility and storage were computed by means of the Theis nonequilibrium formula (Theis, 1935; Brown, 1953) and averaged about 125,000 gpd per foot and 0.0001, respectively. A logarithmic graph of drawdown vs. distance, after 2 hours of pumping (fig. 4), is the basis of calculation of the aquifer constants. A logarithmic graph of time vs. recovery of water levels in observation well C-3 (fig. 5) shows the effect of leakage.

Effect of Leakage

Recharge by leakage produced an effect on water levels in the observation wells similar to that which would have occurred had the cone of depression reached a finite line source, such as an infiltrating stream. That the Mad River did not affect the wells, however, is shown by the results of an analysis of boundary effects based on images. Using a method described by Ferris (1948), hypothetical distances between each observation well and the apparent source of recharge were calculated, and found to range from 1,230 feet, for observation well T-184, to 7,450 feet for observation well C-32. The real distances between the observation wells and the stretch of the stream nearest the pumped well range from about 300 feet to a little more than 4,000 feet. It can be argued that recharge might not necessarily have occurred along the stretch of stream nearest the pumped well, but at a greater distance, where streambed conditions conceivably are different. However, when the various calculated distances are plotted on a map as radii of circles whose centers are located at the respective wells, the resulting arcs do not intersect in the same area, as would be true were the recharge from a comparatively ideal finite line source. In fact, two of the radii form roughly concentric circles which do not intersect at all. This random distribution of the arc intersections also rules out the possibility that recharge occurred through some particular channel in the confining bed which acted as a conduit between the upper and lower aquifers. Channels of this type probably do not occur in the vicinity of the well field, so far as the geologic evidence shows.

The quantity of water that leaks downward into the lower aquifer depends upon the thickness and permeability of the confining bed and the average difference in

Table 1.--Logs of wells, Dayton municipal well field

Well No.	Casing diameter (inches)	Screen length (feet)	Material	Thick-ness (feet)	Depth (feet)	Remarks
C-3	26	30	Top soil and clay	4	4	
			Gravel, dry	13	17	
			Gravel	18	35	
			Sand and gravel	8	43	
			Clay, streaks of gravel	22	65	
			Clay, mucky	19	84	till, confining bed
			Clay, hard	17	101	
			Sand	11	112	
			Gravel, good	25	137	
			Clay	?		
C-5	26	30	Soil	1	1	
			Gravel, dry	16	17	
			Gravel, muddy	6	23	
			Sand, brown, some gravel	7	30	
			Sand and gravel	10	40	
			Sand and gravel, coarse	18	58	
			Clay, sandy	12	70	till, confining bed
			Clay, blue	26	96	
			Sand, fine	4	100	
			Sand and gravel, coarse	35	135	
			Clay	5	140	till?
C-32	17½	33	Top soil	5	5	
			Gravel, dry	5	10	
			Gravel, wet	3	13	
			Gravel, large, coarse, and boulders	20	33	
			Clay and gravel	13	46	till?
			Gravel, coarse, water-bearing, and boulders	13	59	
			Gravel, boulders and sand	11	70	
			Boulders, large	1	71	
			Clay, blue	2	73	till, confining bed
			Clay and gravel	30	103	
			Sand and gravel, with boulders	48	151	
			Gravel and boulders	12	163	
			Clay	?		
C-34	17½	25	Top soil	5	5	
			Clay, gravel and boulders	36	41	till?
			Gravel, coarse, and clay	4	45	
			Clay, gravel and boulders	5	50	
			Clay, hard	4	54	till, confining bed
			Clay and coarse gravel	31	85	
			Sand, fine	19	104	
			Sand and gravel	10	114	
			Clay	1	115	
			Gravel, coarse	3	118	
			Clay and coarse gravel	6	124	
			Gravel, coarse	6	130	
			Gravel, coarse, and clay	5	135	
			Boulders	2	137	
			Sand, fine	4	141	
T-184	6	None	Top soil and clay	5	5	
			Gravel, dry	13	18	
			Gravel	6	24	
			Clay	12	36	till
			Sand and gravel	18	54	
			Clay	41	95	till, confining bed
			Sand	5	100	
			Sand, dirty	2	102	
			Sand and gravel	30	132	well plugged back to 120 ft.
			Hardpan	4	136	till
			Gravel	10	146	
			Clay	?		till?
T-202	6	None	Soil	2	2	
			Gravel, dry	16	18	
			Sand and gravel, coarse	8	26	
			Sand	25	51	
			Sand and gravel, coarse	10	61	
			Clay, blue, some gravel	11	72	till, confining bed
			Quicksand	28	100	
			Sand and gravel	35	135	
			Clay, blue	8	143	till? (well plugged back to 135 ft.)
			Rock, soft	4	147	Ordovician shale
			Clay, blue	90	237	

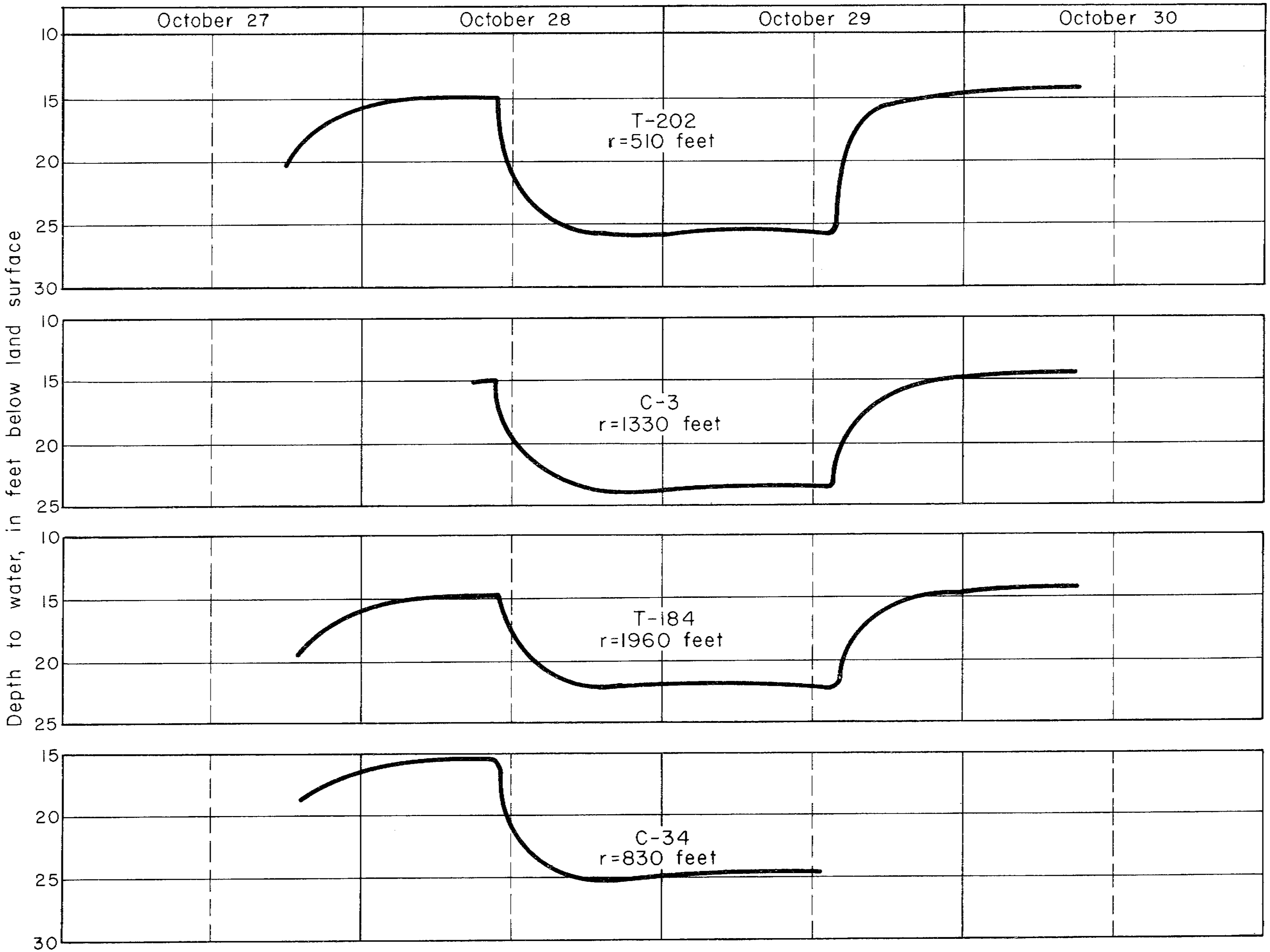

Figure 3. Fluctuation of water levels in observation wells during pumping test of October 28-29, 1953.

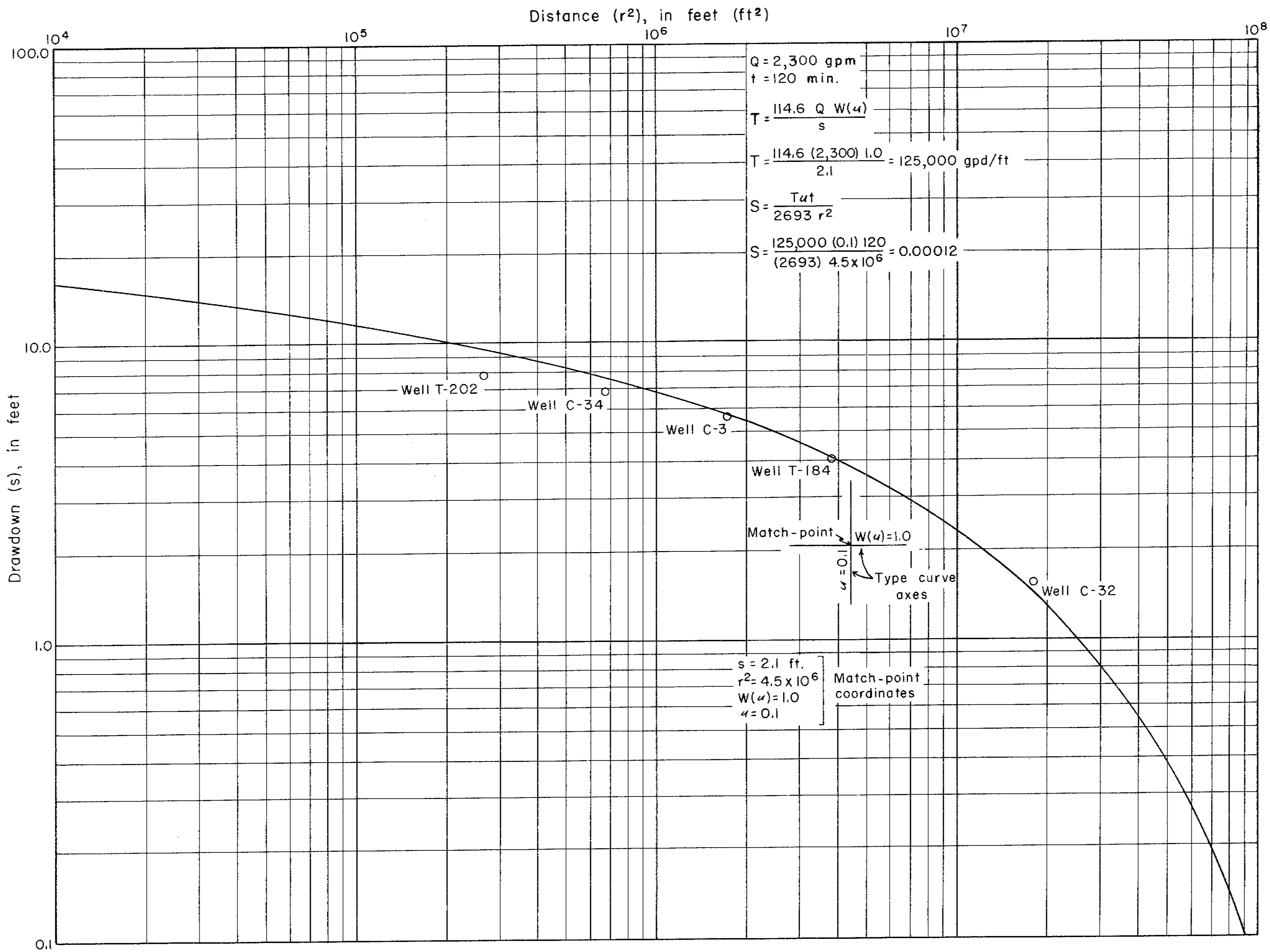

Figure 4. Distance-drawdown log graph, with Theis type curve superposed, after 2 hours of pumping from well C-5.

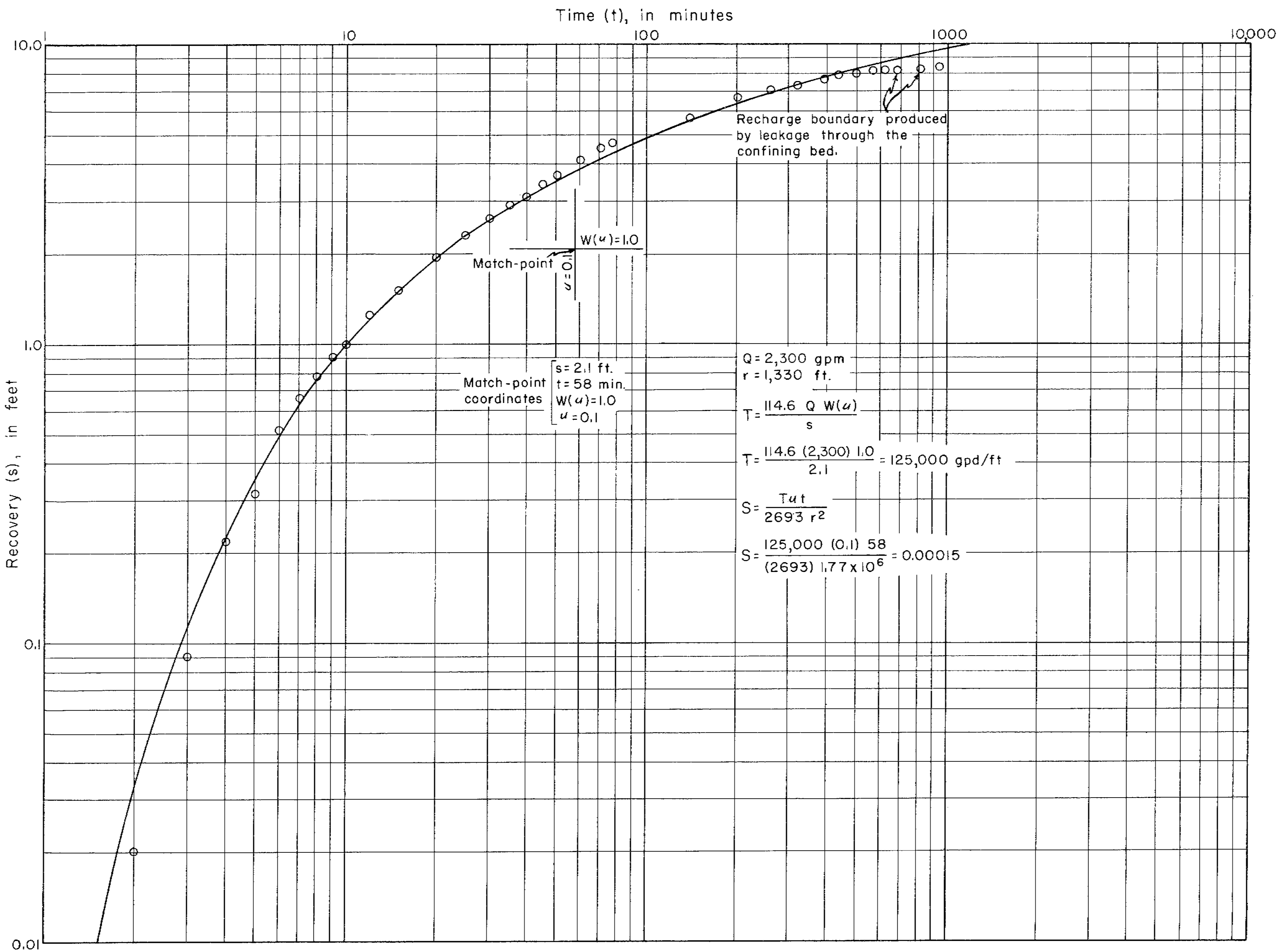

Figure 5. Time-recovery log graph of water level in well C-3, with Theis type curve superposed, showing recharge effects.

water levels in the upper and lower aquifers. The relationship is expressed as follows:

$$Q = \left(\frac{p'}{m'}\right) A \quad \Delta h \qquad (1)$$

where:

Q = leakage, gal/day.
p' = permeability of confining bed, gpd/ft^2.
m' = thickness of confining bed, ft.
A = area over which leakage is occurring, ft^2.
Δh = difference in average water levels in the aquifers, ft.

The leakage coefficient $\left(\frac{p'}{m'}\right)$ (Hantush, 1955) was calculated by use of a method developed by H. H. Cooper, Jr., of the U. S. Geological Survey (written communication). The values obtained for each observation well are listed in table 2.

Table 2.--Leakage coefficient $\left(\frac{p'}{m'}\right)$ and related values for observation wells.

Well	$\left(\frac{p'}{m'}\right)$ gpd/ft^3	Thickness of confining bed (ft)	Permeability of confining bed (gpd/ft^2)
T-202	0.012	11	0.13
T-184	.003	41	.12
C-34	.002	40	.08
C-32	.001	32	.03
C-3	.002	58	.11

The coefficients of transmissibility and leakage were determined also by a type-curve method developed by Ferris (see Knowles, 1955), for the solution of a steady-state equation of Jacob (1946). Fig. 6 is a logarithmic graph of drawdown vs. distance after 10 hours of pumping from well C-5, after leakage became effective in all observation wells. Enough time had elapsed in 10 hours, and the observation wells were far enough from the pumped well, to get good definition of the leaky aquifer type curve. The leakage coefficient was determined from the relationship:

$$x = r \sqrt{\frac{p'}{m'}(T)} \qquad (2)$$

where:

T = coefficient of transmissibility of the artesian aquifer, gpd/ft.
r = distance from pumped well to the observation well, in feet.

and:

x is related to the solution of Jacob's steady-state equation in the same manner that the distance (r) is related to drawdown. This method yielded a value for the leakage coefficient $\left(\frac{p'}{m'}\right)$ of 0.002 gpd/ft^3, which is an average value and numerically the same as that found for 2 of the 5 observation wells listed in table 2. The average of the values for the leakage coefficient listed in table 2 (0.004 gpd/ft^3) is thought to be somewhat high, owing to the weight given it by the single value, 0.012 gpd/ft^3, which is outside the general range of values in table 2. The median value, 0.003 gpd/ft^3, is selected as representative of conditions generally in the well-field area.

Check of Results

When pumping from the lower aquifer had become appreciable and steady, and after it had gone on long enough for hydraulic conditions to stabilize, it became evident

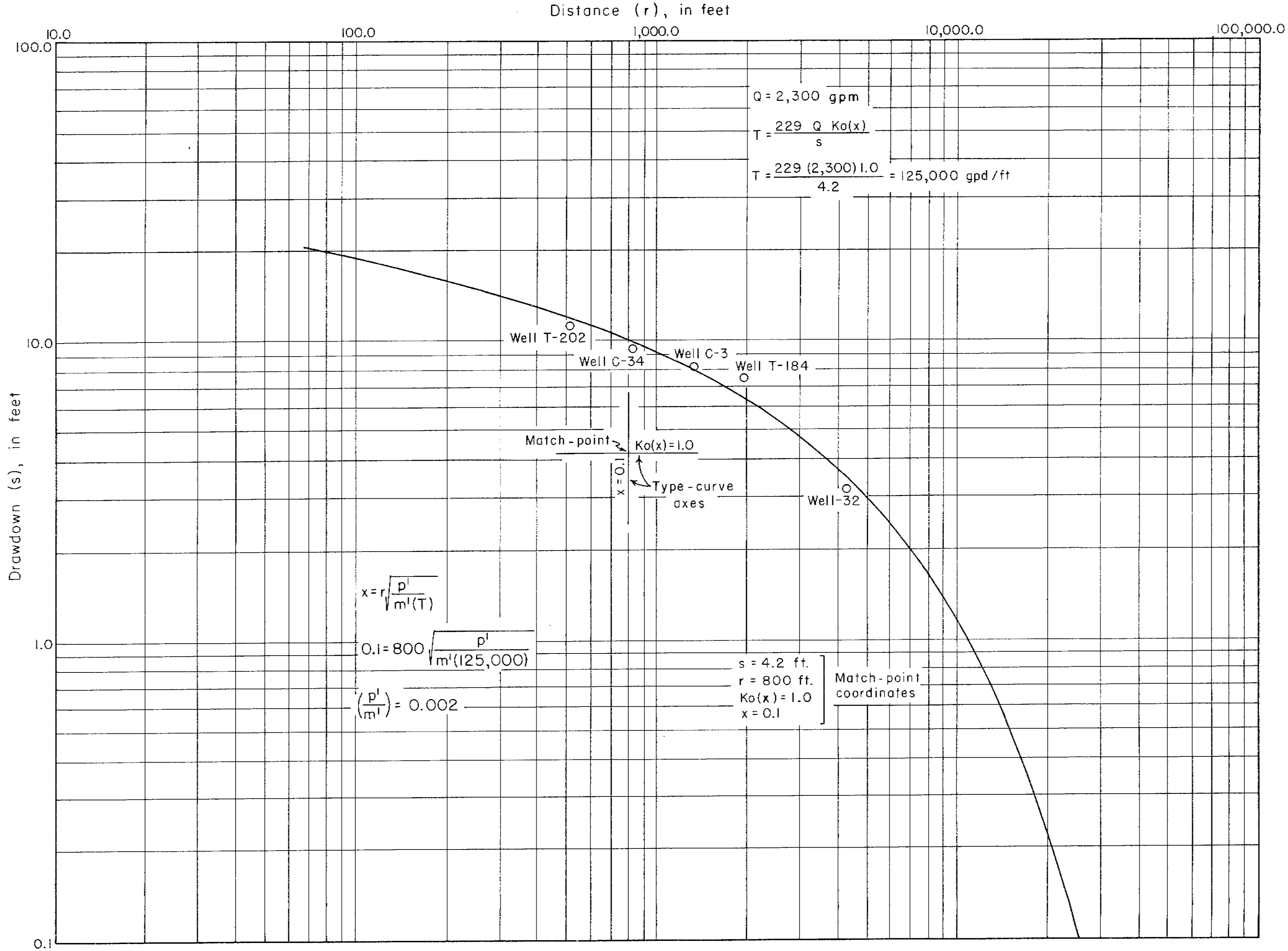

Figure 6. Distance-drawdown log graph, with leaky-aquifer type curve superposed, after 10 hours of pumping from well C-5.

tha a unique opportunity was afforded to check the results of the pumping test. This could be done in either of two ways (1) by relating the known quantity of water withdrawn at the well field to the selected value of the leakage coefficient, together with the areal extent of the cone of depression in the lower aquifer and the average difference in water levels between the upper and lower aquifers, or (2) by computing the theoretical drawdown at points where water-level data are available and comparing the computed drawdown with the lowering that has taken place since pumping became appreciable. If essential agreement between the observed facts and the calculations could be obtained by both methods, it would constitute strong evidence for the validity of the value selected for the leakage coefficient.

Pumping from the lower aquifer is nearly constant at a rate between 12 and 15 mgd. The 10 supply wells are divided arbitrarily into 2 groups, of 4 and 6 wells. The wells in one group are pumped at night, for a period of 12 hours, and the wells in the other group are pumped during the corresponding daylight hours. Each well is pumped at the rate of about 2,000 gpm, making a daily average rate from all wells of about 10,000 gpm or 14 mgd. This pumping procedure has gone on with little change since about 1955. Periodic measurements made during the past several years show that water levels in the wells are essentially stable and that pumpage is effectively balanced by recharge. For the purpose of this discussion we will select a somewhat conservative figure of 13.5 mgd as the amount of water withdrawn perennially at the well field.

Not all the water pumped, however, is replenished by leakage through the confining bed. A small amount comes from natural underflow in the lower aquifer which brings water into the cone of depression. Water moves towards the well field principally in two directions - southwestward, along the axis of the Mad River, and northward in the buried valley that underlies the Wright-Patterson Air Force Base. Calculations, based on a transmissibility of 125,000 gpd per foot for the lower aquifer, indicate the underflow component to be roughly 1.5 mgd. When this quantity is deducted from the average amount pumped from the lower aquifer at the well field it leaves 12 mgd to be accounted for by leakage through the confining bed.

The areal extent of the cone of depression in the lower aquifer is controlled largely by the configuration of the walls of the bedrock valley in which the outwash deposits occur. The well field is located at a bend in the preglacial channel at a point where the ancestral river, flowing north from the vicinity of the Wright-Patterson Air Force Base, turned west towards Dayton. The river was joined in the area of the well field by a large tributary which flowed southwest to its confluence through the col at Huffman Dam (fig. 1). The tributary stream established the course in that area now followed by the Mad River. Thus, the outwash deposits are bounded on the north, east, and southwest of the well field by the valley walls, at distances ranging from a little more than half a mile to about a mile. If lines were drawn across the buried valley so as to form roughly a square, connecting points where the valley walls project closest to the well field, they would enclose an area of about 4 square miles. However, in broad areas west and south of the well field, and to the northeast in a more limited area in the direction of Huffman Dam, the cone of depression can expand far beyond the boundaries of such a square.

Evidence of the development of the cone of depression southward, beneath the Wright-Patterson Air Force Base, is lacking owing to the absence of observation wells. West toward Dayton, however, about 2 miles from the well field, the water levels in a group of wells drilled to the lower aquifer in 1956-57 were reported to be only about 15 feet below the land surface, showing virtually no effects of pumping at the well field. On the basis of this, and the geologic evidence, the areal extent of the cone of depression is estimated to be approximately 5 square miles.

If we substitute in equation (1) the selected value of the leakage coefficient (0.003 gpd/ft^3), together with the estimated values of (a) the quantity of water that leaks through the confining bed (12 mgd) and (b) the areal extent of the cone of depression (5 sq. mi.), and solve for Δh, the average difference in water levels between the upper and lower aquifers, the result is 29 feet. In the absence of conclusive data the problem now becomes one of deciding whether this figure is in the correct order of magnitude.

Water levels in the 10 supply wells, measured when they are not being pumped, average about 46 feet below the land surface. The pumping levels in these wells average about 69 feet below the surface. Before pumping was started the water level in the wells averaged about 15 feet below the surface; therefore, there has been a general lowering of water levels in the well field of 31 to 54 feet. Aside from the supply wells, only one well tapping the lower aquifer can be observed in the vicinity of the well field. That is an abandoned well (MR-42, fig. 2) at a school about 2,500 feet from the nearest supply well, and about 4,000 feet northwest of the center of pumping. The water level in the school well, in the spring and summer of 1958, ranged from about 33 to 43 feet below the surface, indicating a lowering caused by pumping at the well field on the order of 20 or 25 feet. The amounts that water levels have been lowered in the well field and in the abandoned well at the school are close enough to the figure of 29 feet, computed above, to suggest that the latter probably is not far from the actual average difference between water levels in the upper and lower aquifers.

As a further check on the value selected for the leakage coefficient, the theoretical drawdown in well C-5 was computed and compared to the lowering that has taken place in the well since pumping became appreciable. Using the leaky-aquifer formula of Jacob (1946), and taking into account the boundary effects imposed by the bedrock valley walls, the drawdown in well C-5 was computed to be 27 feet, when its pump is idle, and about 57 feet when it is being pumped. The measured drawdown in well C-5 is 35 feet when its pump is idle and 55 feet when the pump is running. The agreement between the computed and observed values of drawdown in well C-5, and the results of the analysis based on the probable area of influence and average head difference between the upper and lower aquifers, strongly support the validity of the value (0.003 gpd/ft^3) selected for the leakage coefficient.

CONCLUSIONS

One of the prime advantages of knowing the leakage coefficient and the hydraulic regimen of the aquifer systems at the Dayton municipal well field is that calculations can now be made involving future large-scale withdrawals of water. Moreover, the average permeability of the till, determined from the pumping test, should prove useful in that this value probably can be applied elsewhere, in a general way at least, in quantitative investigations of geologically similar areas. This investigation, by focusing attention on the importance of leakage through glacial till as a source of water to buried-valley aquifers, may open the door to future studies in other areas which will ultimately make possible much closer estimates of the ground-water resources in glaciated regions.

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