CYCLIC-FLUCTUATION METHODS FOR DETERMINING PERMEABILITY AS APPLIED TO VALLEY-TRAIN DEPOSITS IN THE MAD RIVER VALLEY IN CHAMPAIGN COUNTY, OHIO[1]

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INTRODUCTION

As a part of the cooperative program with the Ohio Department of Natural Resources, Division of Water, the U. S Geological Survey is making an investigation of the ground-water resources of Champaign County, Ohio It was deemed important to future water-resources development in the Mad River valley in Champaign County to have information on the hydraulic properties of the valley-train deposits Unfortunately, so few wells exist in the area that ordinary pumping-test methods seldom can be used to obtain this information In the vicinity of Urbana, however, there is a shallow observation well, screened in the valley-train deposits. The fluctuations of water level in this well reflect corresponding changes in stage in the nearby Mad River. Ferris (1952) shows that observations of the lag, velocity, and wave length of sinusoidal changes in ground-water level, such as are produced by similar changes in stream stage, can be used to determine the transmissibility coefficients of aquifers.

I compared the results of the cyclic-fluctuation methods with pumping-test data from the Urbana well field and laboratory determinations of permeability. The pumping-test data were made available by E. J. Schaefer, Consulting Hydrologist, Columbus, Ohio. The laboratory determinations were made by the U. S. Geological Survey's Hydrologic Laboratory, Denver, Colo.

The purpose of this paper is to demonstrate how, by using only one observation well and assuming a single point of offshore recharge, a reasonably accurate value of transmissibility can be obtained.

Geography

The Mad River flows from north to south across the approximate center of Champaign County, Ohio. The valley in the vicinity of Urbana is bounded on the west by till plains. On the east, "high level" outwash-plain deposits join the "low level" valley-train deposits at several points. The relief between the two deposits is approximately 40 ft. Both deposits are considered together and are hereafter referred to as the valley-train deposits (fig. 1). Both deposits are of Wisconsin (Glacial) age, the lower level deposits consisting primarily of reworked material from the slightly older higher level deposits.

Definition of Terms

The field coefficient of permeability may be defined as the number of gallons of water a day that percolates, under prevailing conditions of water temperature, etc., through each mile of water-bearing bed under investigation (measured at right angles to the direction of flow), for each foot of thickness of the bed, and for each foot per mile of hydraulic gradient (Wenzel, 1942).

The coefficient of transmissibility is the field coefficient of permeability times the thickness of the aquifer, in feet, it may be defined as the number of gallons of water that will move in 1 day through a vertical section of the aquifer 1 mile wide, having a height equal to the full thickness of the aquifer, under a hydraulic gradient of 1 ft per mile (Meinzer, 1923).

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The coefficient of storage under water-table conditions is very nearly, though not exactly, equal to the specific yield, which is defined as the ratio of the volume of water that will drain by gravity from a saturated rock to the total volume of the rock (Meinzer, 1923).

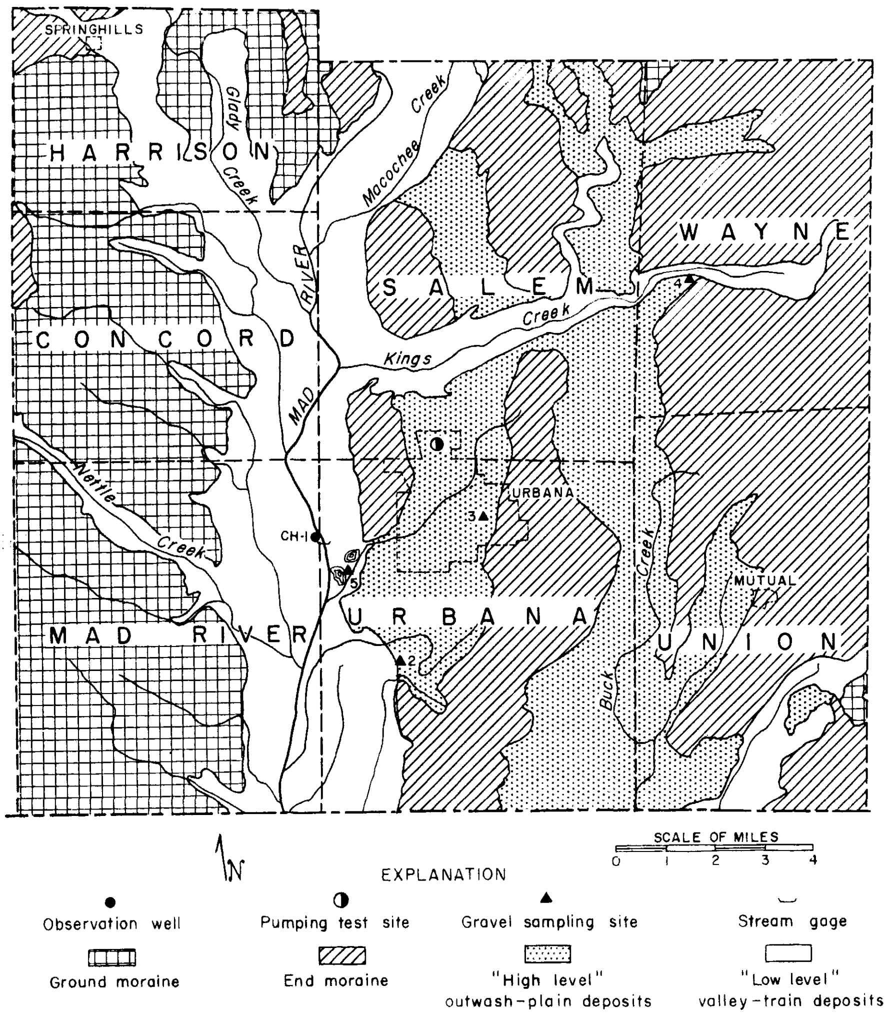

FIGURE 1. Map of the central part of Champaign County, Ohio, showing the extent of the valley-train deposits and location of wells, gravel-sampling sites, and stream gage. (Pleistocene geology generalized from unpublished map by R. P. Goldthwait.)

CYCLIC-FLUCTUATION METHODS

The methods used here were adapted from the work of Ferris (1952), who states: "As the stage of the surface water rises, the head upon the subaqueous outcrop of the aquifer increases and thereby either increases the rate of inflow to the aquifer or decreases the rate of outflow therefrom. The increase in recharge or reduction in discharge results in a general recovery of water level in the aquifer."

Ferris applied the cyclic-fluctuation method to fluctuations of the river stage and observation-well levels along the Platte River at Lincoln, Nebr. In Ferris' example three observation wells were used and fluctuations on both the rising and the falling stage were used. In this study the cyclic fluctuations of water levels in the Mad River and one observation well, Ch-1, were tabulated. The method of Ferris was applied only to the ratio of rise of water level in the observation well to the corresponding stream rise. This was done because the stream was not controlled as it was in Ferris' original work; instead, the stream stage rose quickly but took a relatively long period to decline.

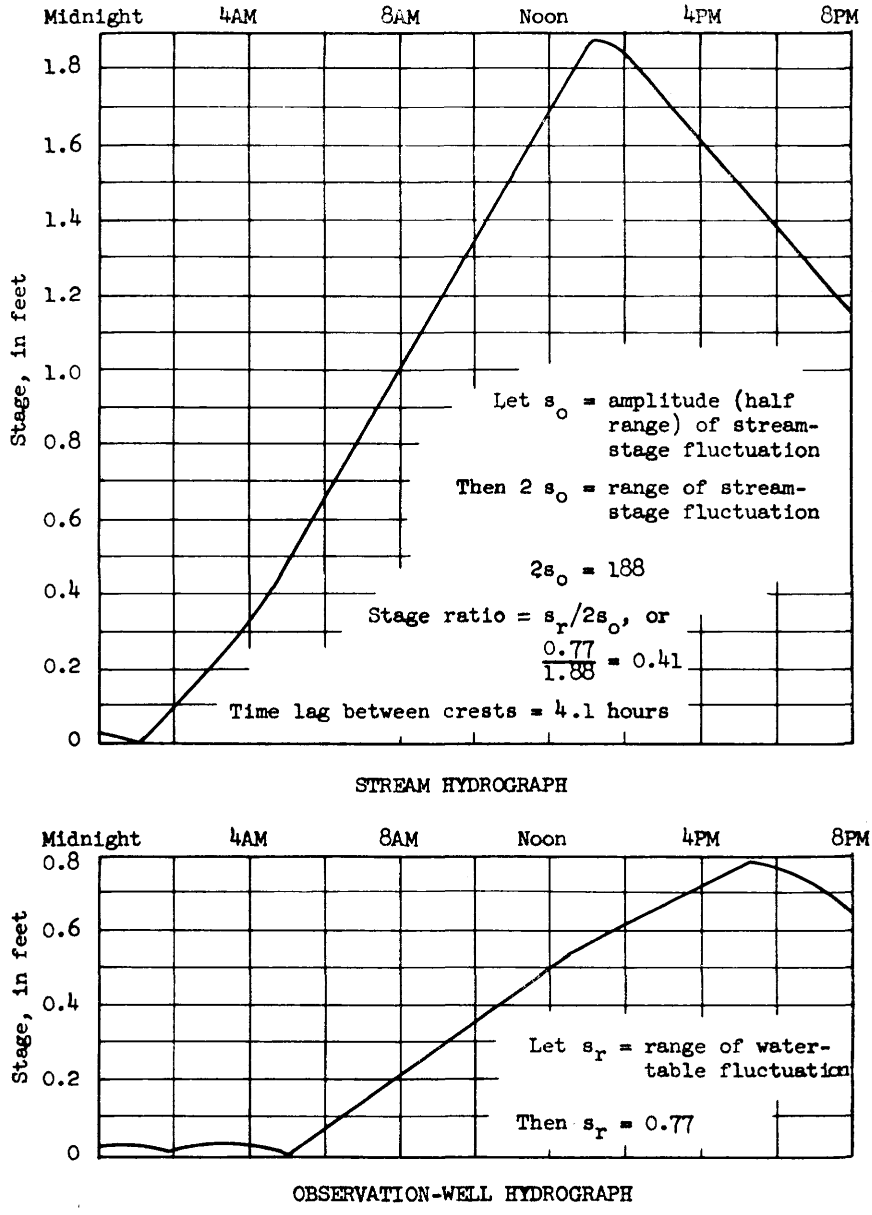

FIGURE 2. Generalized and expanded graph showing stream-stage fluctuation, observation-well fluctuation, and methods used to obtain stage-ratio and time-lag figures.

The observation well, Ch-1, is 155 ft west of the Mad River and approximately 100 ft upstream from a gaging station operated by the U. S. Geological Survey. The well, which was constructed as an observation well by the Ohio Division of Water, is 45 ft deep, is cased to the bottom, and is drilled in valley-train deposits.

Ten periods in which there were rises in stream stage and corresponding rises in water level in the observation well were chosen. These periods were selected when there was no appreciable rainfall in the area and when the river was free of ice. The rises resulted from rainfall in the headwaters area of the river. The average ratio of ground-water fluctuation to river change was computed, as was the average time lag between corresponding crests in the stream and in observation well Ch-1. Figure 2 shows, on an expanded scale, a typical stream rise, the corresponding rise in the observation well, and the methods used to obtain figures used in the calculations.

Stage-Ratio Method

In the stage-ratio method, use is made of the ratio of ground-water fluctuation to stream-stage change. This ratio, as worked out for the stream gage on the Mad River near Urbana and observation well Ch-1, averages 0.41 (see example on fig. 2). As stated previously, this ratio is based on the rising stage and not the falling stage. Ferris' original work averaged both the rise and fall, since flow in the stream used in the Platte River tests was controlled by discharge from a dam upstream. As indicated in the equation below it is necessary that S, the coefficient of storage, be known in order to evaluate T, the coefficient of transmissibility. A reasonable estimate of S can be made because it is known that the aquifer is nonartesian. The formula, in units of gallons per day per ft, as set up by Ferris is:

$$4.4\left(\frac{S}{t_oT}\right)=\left[\frac{-\log_{10}(s_r/2s_o)}{x}\right]^2$$

The logarithmic quantity $(s_r/2s_o)$ is in effect the ratio of the range of ground-water stage to the range of river or tide stage. Thus the right-hand member of the above equation represents the slope of this plot, and if the change in logarithm of the range ratio is selected over one log cycle then the numerator of this slope expression reduces to unity. Then:

$$4.4\left(\frac{S}{t_oT}\right)=\frac{1}{\Delta x} \qquad \text{or:}$$

$$T=\frac{4.4\ \Delta x^2}{t_o}S$$

Where:

T = coefficient of transmissibility [gal/(day·ft)]
S = coefficient of storage (dimensionless)
t_o = period of uniform tide or stage (days)
x = distance (ft) from suboutcrop to observation well
Δx = change in logarithm of the range ratio over one log cycle

The period of river fluctuation on the rising limb of the cycle ranged from 9 to 18 hr and averaged 12 hr. Assuming that the falling limb of the cycle would have been the same if streamflow could have been controlled, the period of river fluctuation would equal approximately 24 hr.

Using the average figure of 0.41 for the ratio of ground-water stage to river stage, we can construct a semilogarithmic plot of this ratio versus distance from the river (fig. 3). Using the Δx value indicated for one log cycle, and $t_o = 1$ day, and assuming various values of S which are reasonable for water-table conditions, the equation is:

$$T = \frac{4.4\ (470)^2}{1} S \text{ or } 970{,}000\ S$$

$T = 97{,}000$ if $S = 0.10$
$T = 150{,}000$ if $S = 0.15$
$T = 190{,}000$ if $S = 0.20$
$T = 240{,}000$ if $S = 0.25$

The negative value shown on figure 3 represents the effective distance to the offshore outcrop, which is assumed to be half the stream width, 27 ft. This assumption introduces a possible factor of error into the calculations, which in this investigation is thought to be less than 10 percent.

Time-Lag Method

Because there is no large withdrawal of water from the valley-train deposits along the Mad River or any other distorting factors to affect the observation well, variations in the lag between the peak period of stream rise and the peak period of well rise can be attributed to difficulty in reading the compressed time scale of the water-level recorders or to inadequate knowledge of the exact effective distance

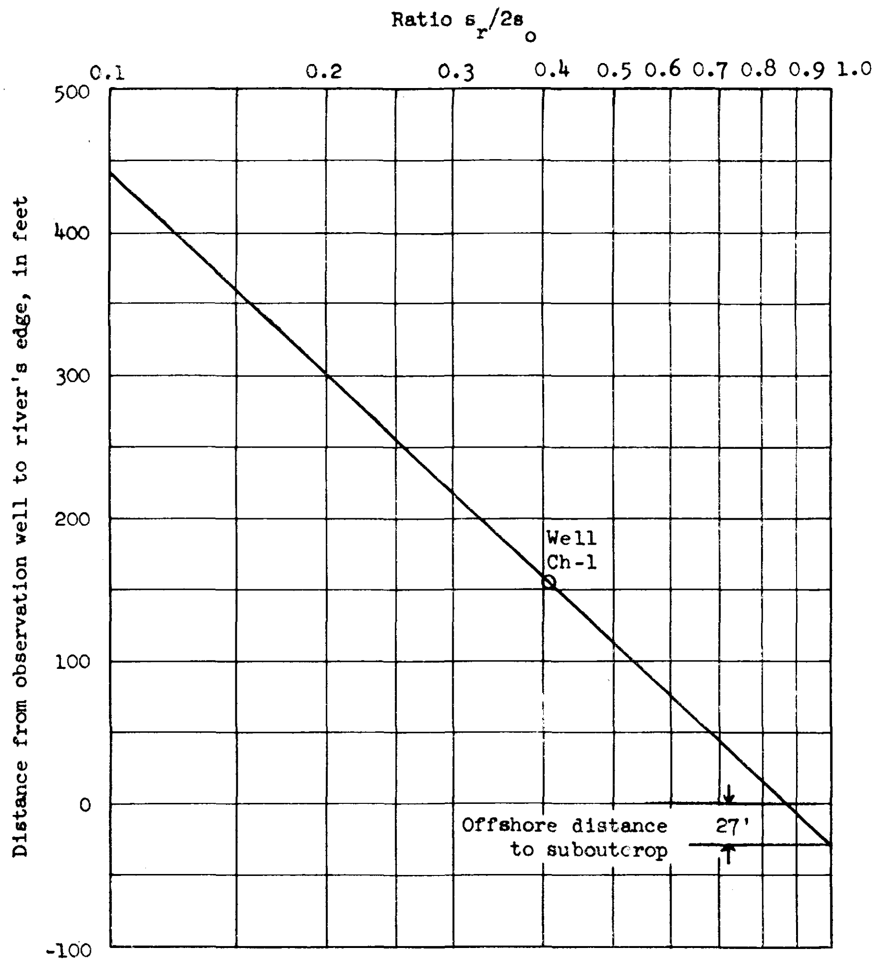

FIGURE 3. Semilogarithmic plot of the ratio of ground-water stage to stream stage versus the distance of the observation well from the cyclic stream.

to the offshore outcrop. The average period of time lag, t_1, between the maximum stage of the Mad River and the maximum ground-water stage is 4.1 hr (fig. 2). This time lag is plotted against distance, in ft, from the observation well to the river's edge (fig. 4). Again assuming that the effective distance to the offshore outcrop is 27 ft, and using the formula of Ferris we obtain T in units of gallons per day per ft, using t_o and t_1 in days.

$$T = \frac{0.60\ x^2\ t_o}{t_1^2}\ S$$

Substituting values obtained from figure 4, the equation becomes:

$$T = \frac{0.60\ (155)^2\ 1}{(3.5/24)^2}\ S \text{ or } 690{,}000\ S$$

Again assuming the various values for S under water-table conditions we get:

$$T = 69{,}000 \text{ if } S = 0.10$$
$$T = 100{,}000 \text{ if } S = 0.15$$
$$T = 140{,}000 \text{ if } S = 0.20$$
$$T = 170{,}000 \text{ if } S = 0.25$$

The difference in the coefficients of transmissibility indicated by the two methods used is probably due to a lack of sufficient information for proper appraisal of the suboutcrop distance or to inadequacies in reading the compressed scale of the recorder graphs.

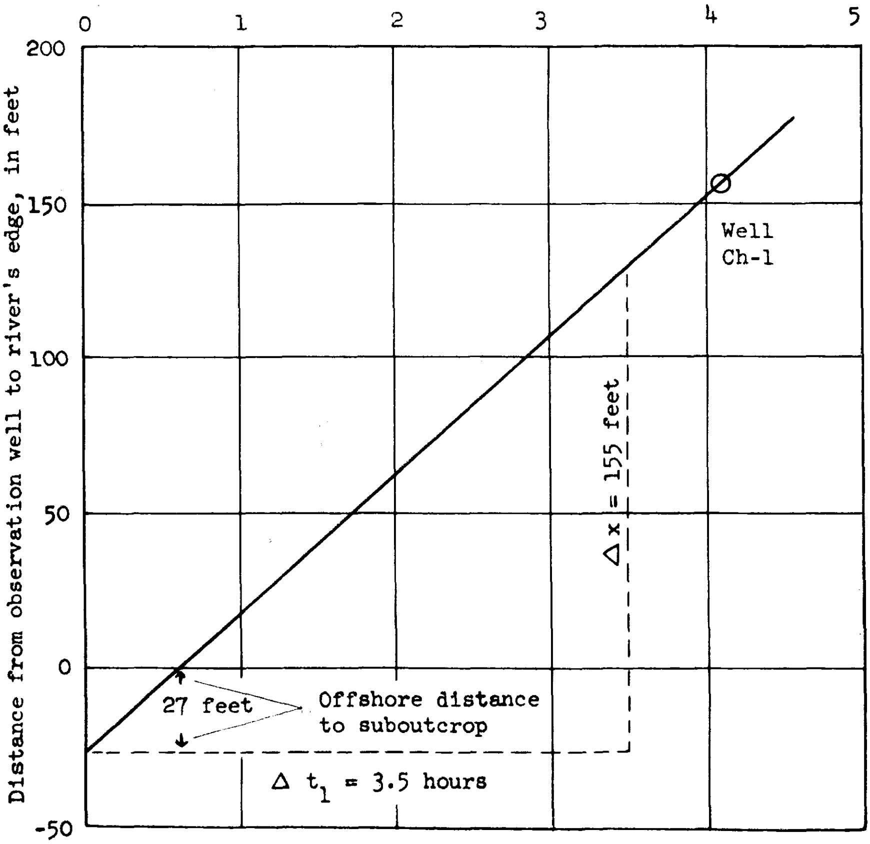

FIGURE 4. Plot of time lag versus distance of observation well from suboutcrop.

Average of Cyclic-Fluctuation Methods

The results of the two methods are summarized as follows:

Method	T, coefficient of transmissibility, in gal per day per ft			
Assumed coefficient of storage	0.10	0.15	0.20	0.25
Stage-ratio method	97,000	150,000	190,000	240,000
Time-lag method	69,000	100,000	140,000	170,000
Average (rounded)	83,000	120,000	160,000	200,000

The saturated thickness of valley-train deposits at observation well Ch-1 is approximately 40 ft. Using this thickness and the average values from the data given above, the average field permeability, in gallons per day per ft^2, is found by dividing the coefficient of transmissibility, in gallons per day per ft, by the saturated thickness of the aquifer, in feet.

The field-permeability values, using each of the average values given previously, are:

Coefficient of storage	Field coefficient of permeability in gal per day per ft^2
0.10	2,000
0.15	3,000
0.20	4,000
0.25	5,000

PUMPING-TEST DATA

A pumping test was made in 1956 for the city of Urbana by E. J. Schaefer, Consulting Hydrologist, Columbus, Ohio. The municipal well field is just north of Urbana in the "high level" outwash-plain deposits approximately 3 miles east of the Mad River (fig. 1).

Schaefer's computed value of T, the coefficient of transmissibility, using one pumped well and one observation well, was 222,000 gal per day per ft. He computed the value of S, the coefficient of storage, in the same test as 0.14. The saturated thickness of material in the area of the well field is approximately 50 ft which, divided into the coefficient of transmissibility calculated from the test data, gives a field coefficient of permeability of approximately 4,400 gal per day per ft^2.

LABORATORY ANALYSES

Five samples of the valley-train and outwash-plain deposits were collected and sent to the U. S. Geological Survey's Hydrologic Laboratory, Denver, Colo., where determinations were made of porosity and permeability.

Figure 1 shows the location of four of the sampling sites. The fifth sampling site, No. 1, was a gravel pit just west of the Mad River, near Eagle City, in Clark County, approximately 3 miles south of the Champaign County border.

Presented below in tabular form are the sample number, field description of location, and the porosity and coefficient of permeability as determined in the laboratory. The laboratory, or standard, coefficient of permeability is the same as the field coefficient except that it is determined at 60° F. The average temperature of the ground water in the Urbana area is a little less, but not enough to make a substantial difference when comparing field and laboratory data.

TABLE 1

Sample number	Field description of location	Porosity (% voids)	Coefficient of permeability gal/[day·ft²]
1	Gravel pit near Eagle City, Ohio	23.7	4,500
2	Valley wall south of Urbana, Ohio	32.6	6,000
3	Gravel pit, east edge of Urbana, Ohio	36.3	4,500
4	Gravel pit near Cable, Ohio	26.4	3,400
5	Gravel pit, west edge of Urbana, Ohio	40.3	2,000

Averages of all values give a porosity of 31.8 percent and a coefficient of permeability of 4,100 gal per day per ft^2.

SUMMARY AND CONCLUSIONS

The cyclic-fluctuation methods, using only one well and assuming the coefficient of storage to be about 0.2, gives values of the coefficient of permeability that are in reasonable agreement with values obtained by both of the other methods used. Because the coefficient of storage under water-table conditions usually is between 0.1 and 0 3, it is probable that the value of the valley-train deposits is not far from 0.2. Analysis of the pumping-test data collected by E. J. Schaefer gives a permeability of 4,400 gal per day per ft^2 for what probably are very similar deposits. The average figure for the laboratory determinations of the coefficient of permeability, also for probably similar deposits, is 4,100 gal per day per ft^2. Each of these independent methods of arriving at the coefficient of permeability agrees well with the cyclic-fluctuation method, which gave a figure of 4,100 gal per day per ft^2 using a storage coefficient of 0.2.

The permeability of the valley-train deposits in Champaign County, Ohio, may vary over a wide range but probably is in the general vicinity of 4,000 gal per day per ft^2 wherever well-sorted sand and gravel make up the deposits.

The cyclic-fluctuation method as described here promises to be useful in water-resources investigations, in that it can be applied to stream fluctuations and their effect on ground-water levels in areas having only a single observation well. An error, of course, is introduced by assuming the position of the single line of recharge, but if the stream is relatively narrow and the observation well is located at a comparatively large distance the error involved is small. The agreement among the three independent methods of analysis of the hydrologic data indicates that this method is applicable in the area studied, and by inference in other areas of similar hydrology. Checking of the method in other areas would be desirable, but even without checking it appears that the method is useful in situations where pumping tests are impractical.

REFERENCES

Ferris, J. G. 1952. Cyclic fluctuations of water level as a basis for determining aquifer transmissibility. U. S. Geol. Survey open-file rept.

Meinzer, O. E. 1923. Outline of ground-water hydrology, with definitions. U. S. Geol. Survey Water-Supply Paper 494. pp. 28, 44.

Wenzel, L. K. 1942. Methods for determining permeability of water bearing materials, with special reference to discharging-well methods. U. S. Geol. Survey Water-Supply Paper 887. p. 7.