X. AQUIFER CHARACTERISTICS AND AQUIFER TESTING

Pertinent 30 CFR[1] Sections:
Description of hydrology and geology.
Ground-water information.

1. Aquifer Characteristics

The quantity and rate of ground-water discharge into a surface coal-mining excavation depend upon the hydraulic properties of the ground-water flow system near the excavation. The hydraulic properties include hydraulic conductivity (K), also called permeability, transmissivity (T), and storativity (S), which is the storage coefficient for confined aquifers and specific yield for water-table aquifers. Definitions of these properties are included in the glossary (chapter XVIII).

For steady-state ground-water flow conditions, the quantity of discharge (Q) is

$$Q = K\ I\ A \qquad \text{(x-1.0-1)}$$

where: K = hydraulic conductivity of the aquifer,
I = hydraulic gradient, and,
A = cross-sectional area of the aquifer perpendicular to the flow direction through which the ground water flows.

1.1 Saturated Thickness

The saturated area (A) of an aquifer is the product of the width and depth of the water-bearing material. The flow of ground water is commonly expressed in units related to 1 foot of width of cross section and the saturated thickness of the aquifer. This thickness, often expressed algebraically as b or m, is the distance between the water table and the bedrock surface for water-table aquifers, and for confined aquifers, as the distance between the confining beds above and below the aquifer. The vertical section in figure X-1.1-1 illustrates local variations in the saturated thickness of an unconsolidated water-table aquifer and a confined sandstone aquifer beneath it. The saturated thickness of the alluvial (water-table) deposits varies form zero, at the contact with the bedrock, to 115 ft. The sandstone thickness ranges from 100 ft at the south edge to 60 ft at the north edge.

[1]CFR= Code of Federal Regulations

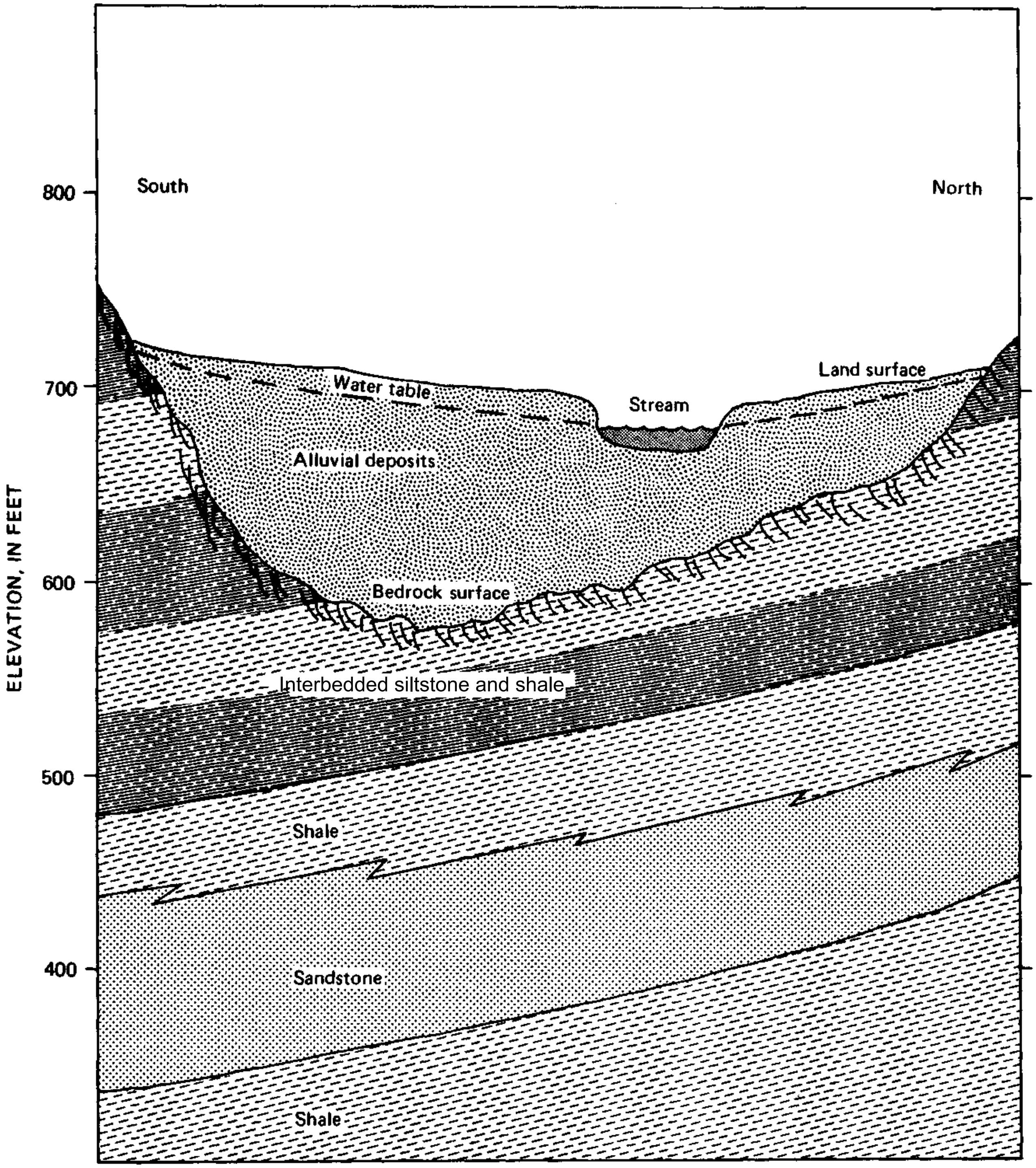


Figure X-l.1-1.— Local variations in saturated thickness (b) of unconsolidatd water-table aquifer and confined sandstone aquifer.

1.2 Hydraulic Conductivity and Transmissivity

The typical hydraulic conductivity (K) values for various geologic materials is summarized in figure X-l.2-1. For example, K for shale ranges from about 0.0000001 to 0.005 ft/d and, for sandstone, from less than 0.0001 ft/d (unfractured) to about 1 ft/d (semiconsolidated). The K values for fractured bedrock may be many orders of magnitude greater than for unfractured bedrock.

A more convenient means of calculating the steady-state volume of ground-water flow (Q) is by using transmissivity (T), which is equivalent to K multiplied by the saturated thickness (b) of the aquifer. For this case,

$$Q = T\ I\ W \qquad \text{(X-l.2-1)}$$

where W is the width of the ground-water flow cross section.

The geologic material having the greatest water-bearing capability is saturated unconsolidated alluvial sand and gravel. The hydraulic-conductivity values for unconsolidated deposits can range from about 0.0000001 to more than 100,000 ft/d. Table X-l.2-1 lists some representative hydraulic-conductivity values for specific grain-sized alluvial materials.

The units for hydraulic conductivity are feet per day. K does not represent ground-water flow velocity but is a constant of proportionality in the equation for determining ground-water discharge, Q. The K dimensions are rate per area, in $(ft^3/d)/ft^2$ (cubic feet per day per square foot), which reduces to ft/d (feet per day).

Transmissivity (T) of unconsolidated alluvial deposits can be estimated by multiplying the "handbook values" (such as in table X-l.2-1) for K.of each described material by the saturated thickness (b) of that material. As illustrated in table X-l.2-2, transmissivity at the driller's log location is the sum of these products for the saturated thicknesses, b_1, b_2, b_3, ... b_n:

$$T = \text{sum of } K_m \bullet b_m = K_1\ b_1 + K_2\ b_2 + K_3\ b_3 + ... + K_n\ b_n. \qquad \text{(X-l.2-2)}$$

The quantity of ground-water discharge into a mine excavation from an aquifer can be estimated from information given on a potentiometric map of the coal aquifer, aquifer -test results, and equation X-l.2-1. For example, the hydraulic gradient (I) for the two aquifers in the Decker mine area can be determined from the potentiometric maps of the permit and adjacent area (fig IX-4 for the clinker aquifer and fig X-l.2-2 for the D-2 aquifer). The discharge into the initial surface-mine cut is calculated as follows:

Given: 1. The two aquifers, Clinker (CD and D-2 Goal (D-2), to be excavated in surface mining

2. Potentiometric maps for both aquifers -
figure IX-4 for D-l Coal bed aquifer combined with Clinker
(also pl. 4 of VanVoast and Hedges, 1975)
and figure X-l.2-2 for D-2 Coal bed aquifer
(also pl. 5 of VanVoast & Hedges, 1975)

3. Average hydraulic gradient (I):=
I_{Cl} = 40/4827 = 0.0083; $I_{D\text{-}2}$ = 20/4224 = 0.0047

4. Saturated thickness (b)
b_{Cl} = 20 ft; $b_{D\text{-}2}$ = 15 ft

5. Average hydraulic conductibity (K)
K_{Cl} = 125 ft/d; $K_{D\text{-}2}$ =3 ft/d;

6. Discharge Formulas: Q= TIW and T=Kb

Find the average ground-water discharge per foot of width of coal aquifer that would discharge into the surface-mine excavation.

Solution:

Q_{Cl} = 125 x 20 x 0.0083 = 20.7 ft^3/d per foot of width
$Q_{D\text{-}2}$ = 3 x 15 x 0.0047 = 0.21 ft^3/d per foot of width

Total discharge (Q_T) = Q_{Cl} + $Q_{D\text{-}2}$

= 20.7 + 0.21

= 20.9 ft^3/d per foot of width of aquifer

The hydraulic-conductivity values presented in figure X-1.2-1 and table X-1.2-1 are not necessarily applicable to all permit areas. Published hydraulic-conductivity values for similar material vary by orders of magnitude. Hydraulic conductivity and transmissivity of bedrock aquifers cannot be accurately estimated, because of local variations in primary permeability and secondary permeability (fracturing). These values vary both horizontally and vertically, depending on geologic conditions.

The aquifer properties in the permit areas can be determined by aquifer testing as described in chapter X-2. Some of the published transmissivity values obtained during variious hydrologic investigations are giben in table X-2.4-1. Reference (21) also lists ranges of transmissivity for selected ground-water regions of the United States.

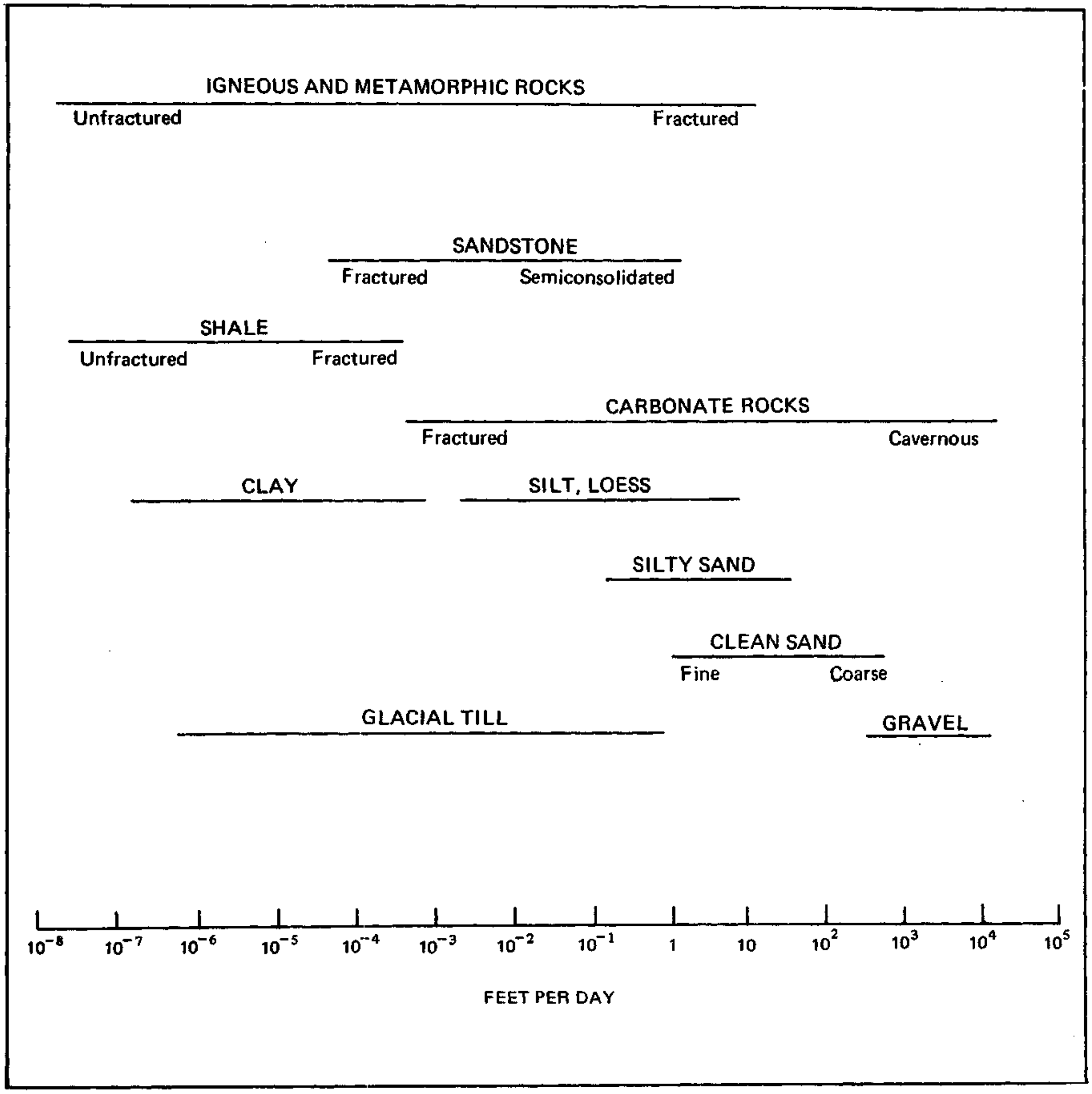


Figure X-l.2-1.— Range of hydraulic-conductivity values of selected aquifer materials. (Modified from Heath, 1983, p. 13)

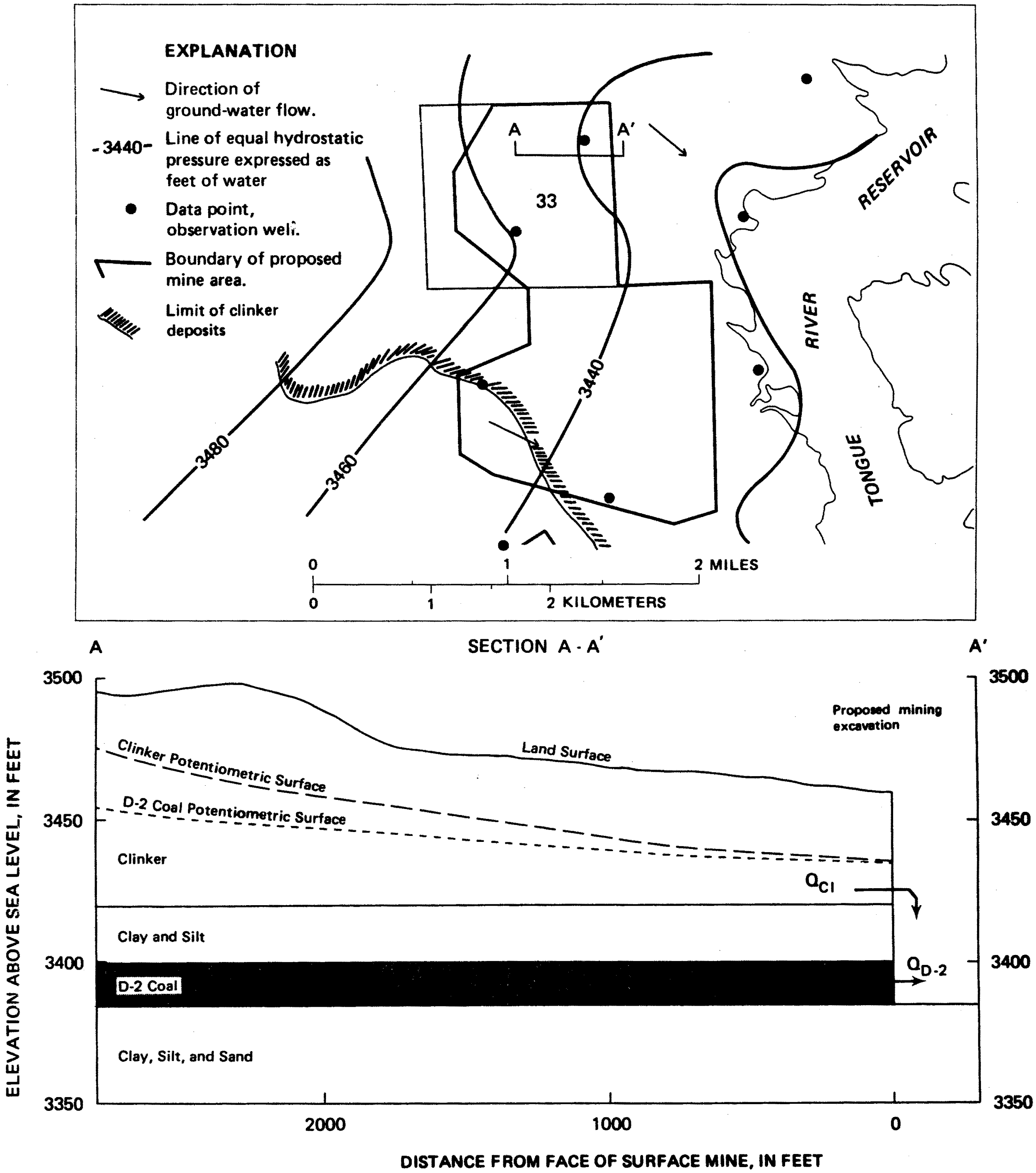


Figure X-1.2-2.— Potentiometric map of D-2 coal aquifer, near Decker, Montana, before mining began (above), and geologic section showing potentiometric surface profiles (below). (Modified from VanVoast and Hedges, 1975, pls. 5, 11)

Table X-l.2-1.– Average hydraulic conductivity values of alluvial materials (From Lohman, 1972, table 17.) [ft/d, feet per day]

Alluvial Material	Hydraulic Conductivity (ft/d)
GRAVEL:	
Coarse	1,000
Medium	950
Fine	900
SAND:	
Gravel to very coarse	800
Very coarse	700
Very coarse to coarse	500
Coarse	250
Coarse to medium.	100
Medium	50
Medium to fine	30
Fine	15
Fine to very fine.	5
Very fine	3
CLAY:	1

Table X-l.2-2. Example of transmissivity calculation from driller's log of a well in an alluvial aquifer.
[ft, feet; ft^2/d, square feet per day; ft/mi, feet per mile; gal/d, gallons per day]

Material (from driller's log)	Thickness (b) (ft)	Bottom depth (ft)	Hydraulic conductivity(K) (ft/d)	Transmissivity (T) (T = Kb) ft^2/d
Top soil	1	1		-
Clay	5	6		-
Fine sand	4	10		-
------------------------(Water level at 10 feet)------------------------				
Medium sand	5	15	50	250
Medium to coarse sand	25	40	100	2,500
Clay	20	60	1	20
Sand and gravel	10	70	800	8,000
Coarse gravel	15	85	1,000	15,000
Shale (bedrock)	5	90	-	-

T = sum of $K_m \bullet b_m$ = 25,770 ft^2/d
rounded to 26,000 ft^2/d

Ground-water discharge(Q) per foot of aquifer width at this location, initially, with T = 26,000 ft^2/d and I = 40 ft/mi would be from equation X-1.2-1

Q=T I W x 7.48= 26,000 x $^{40}/_{5280}$ x 1 = 1500 gal/d 5280

1.3 Storativity and Specific Yield

Storativity (S) indicates the capability of an aquifer to store or release water as head changes. Storativity is usually determined through analysis of data from multiple-well aquifer tests. However, rough approximations of specific yield for most types of aquifer material can be obtained from table X-l.3-1, and storativity of a confined aquifer can be calculated from the range of b values given in table X-l.3-2. The applicability of storativity to confined aquifers is increased by converting it to specific storage (Ss), which represents storativity per foot of confined aquifer thickness (b), or S/b. Table X-l.3-2 uses an Ss value of 0.000001 (or 10^{-6}) per foot of aquifer thickness.

Additional information on storage coefficients, storativity, and specific yields can be found in bibliographic references listed in chapter XVII; these include Barrett and others, 1980; Ferris and others, 1962; Freeze and Cherry, 1979; Heath, 1983; Johnson Division, 1975; McWhorter and Sunada, 1977; and U. S. Department of Interior, 198la.

TABLE X-l.3-1. Average specific yield for unconsolidated water-table aquifers (From Johnson, 1967, p. D 70.)

Alluvial Material*	Average Specific Yield
Clay □	0.02
Silt □	.08
Sandy clay □	.07
Fine sand □	.21
Median sand □	.26
Coarse sand □	.27
Gravelly sand □	.25
Fine gravel □	.25
Medium gravel □	.23
Coarse gravel □	.22

* Generally ranges from 0.1 to 0.3 depending on the size and sorting of the alluvial material.

TABLE X-l.3-2. Method of estimating storage coefficient (S) for confined aquifers (From Lohman, 1972, p. 53.)

Thickness of Confined Aquifer (b, in feet)	Storage Coefficient (S)
1	0.000001
10	.00001
100	.0001
1000	.001

Example of application: a confined aquifer with a saturated thickness(b) of 300 feet, S equals approximately 0.0003

1.4 Effect of Hydraulic Properties on Drawdown

The relationship between drawdown and time (t), distance (r), transmissivity (T), and storativity (S) under conditions of constant discharge are illustrated in figure X-l.4-1. The general pattern is that the smaller the T and S values, the larger the drawdown values near to the point of discharge and the larger the volume of aquifer affecter through time. Dewatering an excavation where T and S are small will affect a larger aquifer volume than in an area where T and S are large. The change in drawdown with distance from a well pumping from a confined aquifer after 1 year at several pumping rates with a constant T = 20 ft^2/d and S = $5X10^{-5}$ are depicted in figure X-l.4-2. The drawdowns for the same pumping rates at a single location over a period of several years are plotted in fig X-l.4-3.

An application of the information presented in figure X-l.4-2 for the stated T and S values is as follows: Consider a well pumping at a constant rate of 26 gal/min for 365 days; the drawdown in a homogeneous confined isotropic aquifer would be 50 ft at a distance of 1 mi or 95 feet at a distance of 1/4 mi.

An application of the information presented in figure X-l.4-3 for the same hydrologic conditions is as follows: For a well pumping at a constant rate of 26 gal/min, the drawdown at a distance of 1,000 feet will be 40 ft after 10 days, 90 ft after 100 days, and 135 ft after 1,000 days.

Transmissivity is locally variable; some published values for selected consolidated formations are presented in figure X-l.4-4 and table X-l.4-1. For example, transmissivity of Pennsylvanian rocks, in the border area between Tennessee and Alabama, ranges from 50 to 13,000 ft^2/d. The major bedrock types are sandstone and shale, which are fractured to varying degrees, and the T value is related to the size, density, and extent of fractures. In contrast, the T value in the Mississippian carbonate rocks ranges from 100 to 27,000 ft^2/d and is related to the size and abundance of solution openings and fractures. The drawdown distributions in the aquifer units that are affected by mining operations will be irregular and parallel the irregular distribution of hydraulic properties.

Aquifers overlying and underlying coal seams can differ in hydraulic properties because of differences in lithology and fracturing. For example, a sandstone and conglomerate rock unit would have a much higher T value than an interbedded siltstone, shale, and sandstone rock unit. Hydraulic properties can also vary laterally within the same unit. These variations are related to local differences in saturated thickness, lithology (called facies changes), solution openings, and fracture density.

The difference between transmissivity of the aquifer unit overlying a coal bed and that of a aquifer below it can be determined through aquifer tests in each water-bearing formation and in each aquifer unit that might be affected by the mining operation.

The lateral variation of T within an aquifer unit can be determined through aquifer tests at several locations within and adjacent to the proposed permit area. A map showing the variation of T in an aquifer is shown in figure X-l.4-5, where T varies from zero at the outcrop to 300 ft^2/d. The transmissivities were determined at the data points. The contours represent interpolated estimates. The best method of predicting the hydrologic affects of dewatering is through computerized ground-water modeling. Transmissivity maps such as that in figure X-l.4-5 are essential for the development of a ground-water model of a proposed permit area.

Single-well aquifer tests can be performed in conjunction with the exploratory program for coal-resource-evaluation. If the results of these tests show a range of only 1 order of magnitude in hydraulic properties of the pertinent units, the system is sufficiently uniform that additional aquifer testing is not necessary. Also, if the aquifer thickness, confining-bed thickness, and fracture spacings are consistent among units, additional aquifer testing may not be necessary. However, if aerial photographs indicate the presence of structural features such as faults, fracture traces, or lineaments, which might affect the aquifer systems in the permit area, additional aquifer testing will probably be necessary. These structural features are generally saturated, extend to hundreds of feet, and can have a hydraulic conductivity thousands of times greater than that of the adjacent unfractured rock. These water-bearing features can be particularly troublesome in deep-mine operations because they may discharge large quantities of water into underground workings.

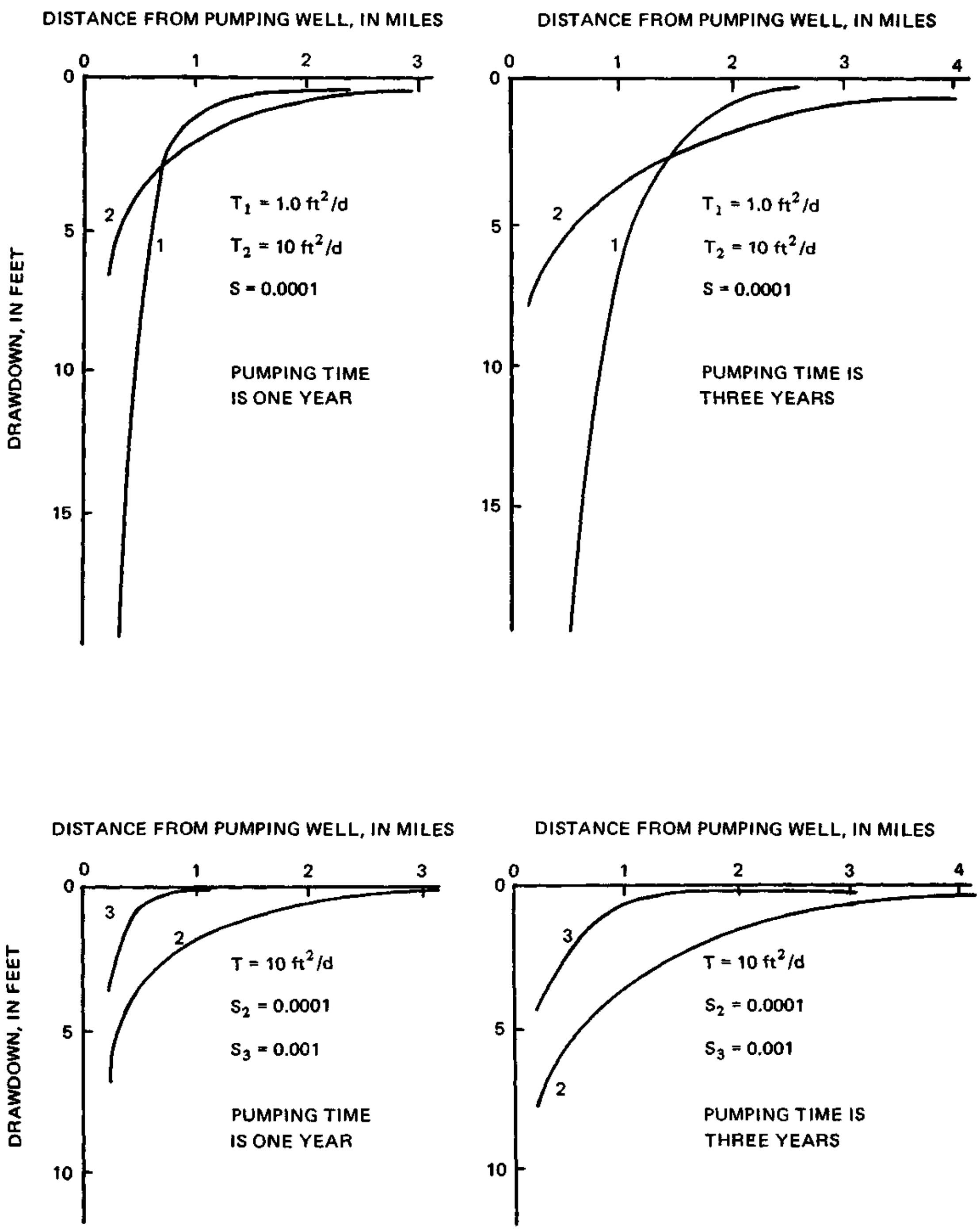


Figure X-l.4-1.— Influence of transmissivity (above) and storativity (below) on the distribution of drawdown for pooping periods of 1 year and 3 years. (Well discharge is 1 gallon per minute.)

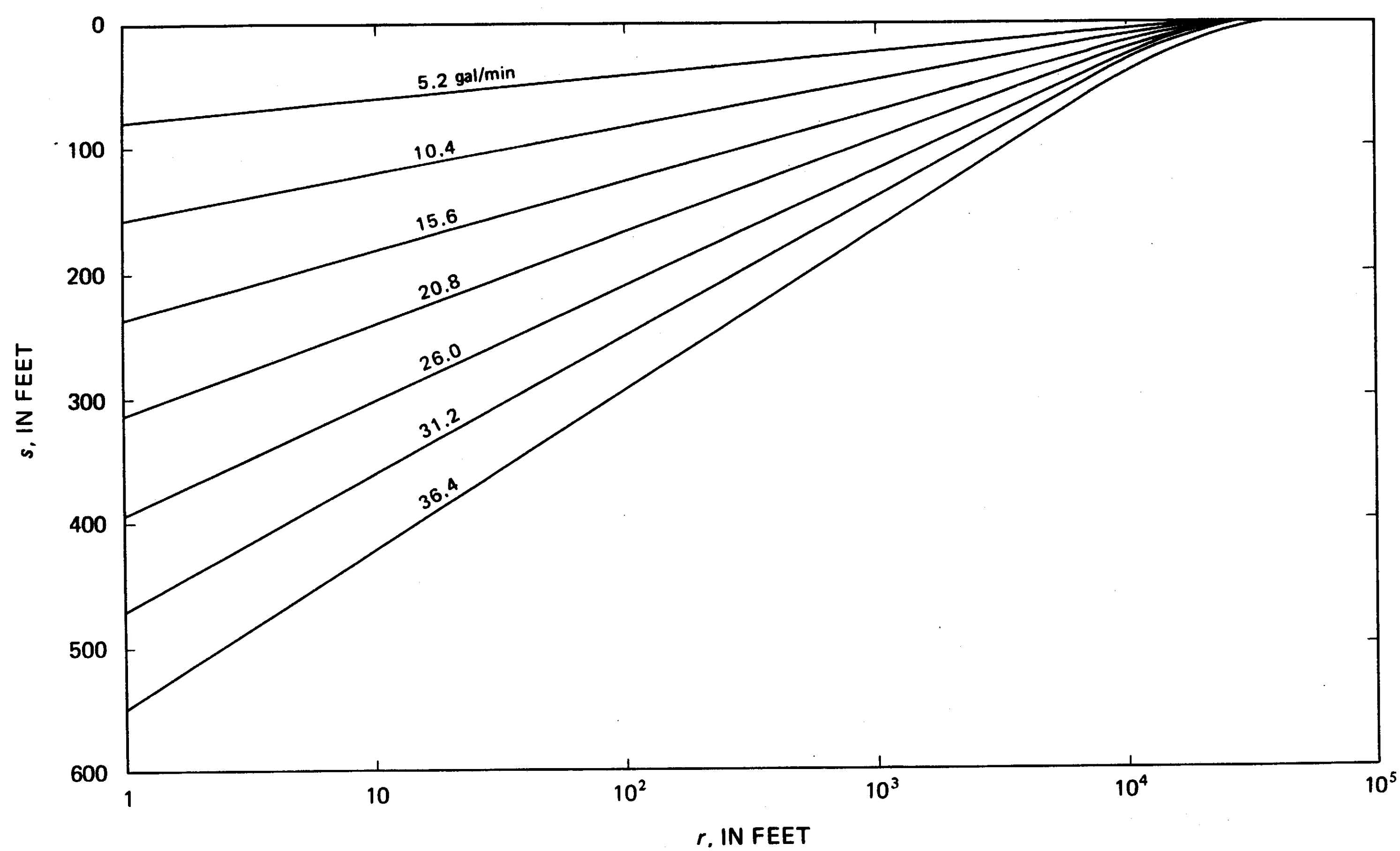


Figure X-1.4-2.— Family of curves showing drawdown produced at various distances from a well discharging at selected rates for 365 days from a confined aquifer for which transmissivity is 20 ft^2/day and storage coefficient is 0.00005. (From Lohman, 1972, p. 54)

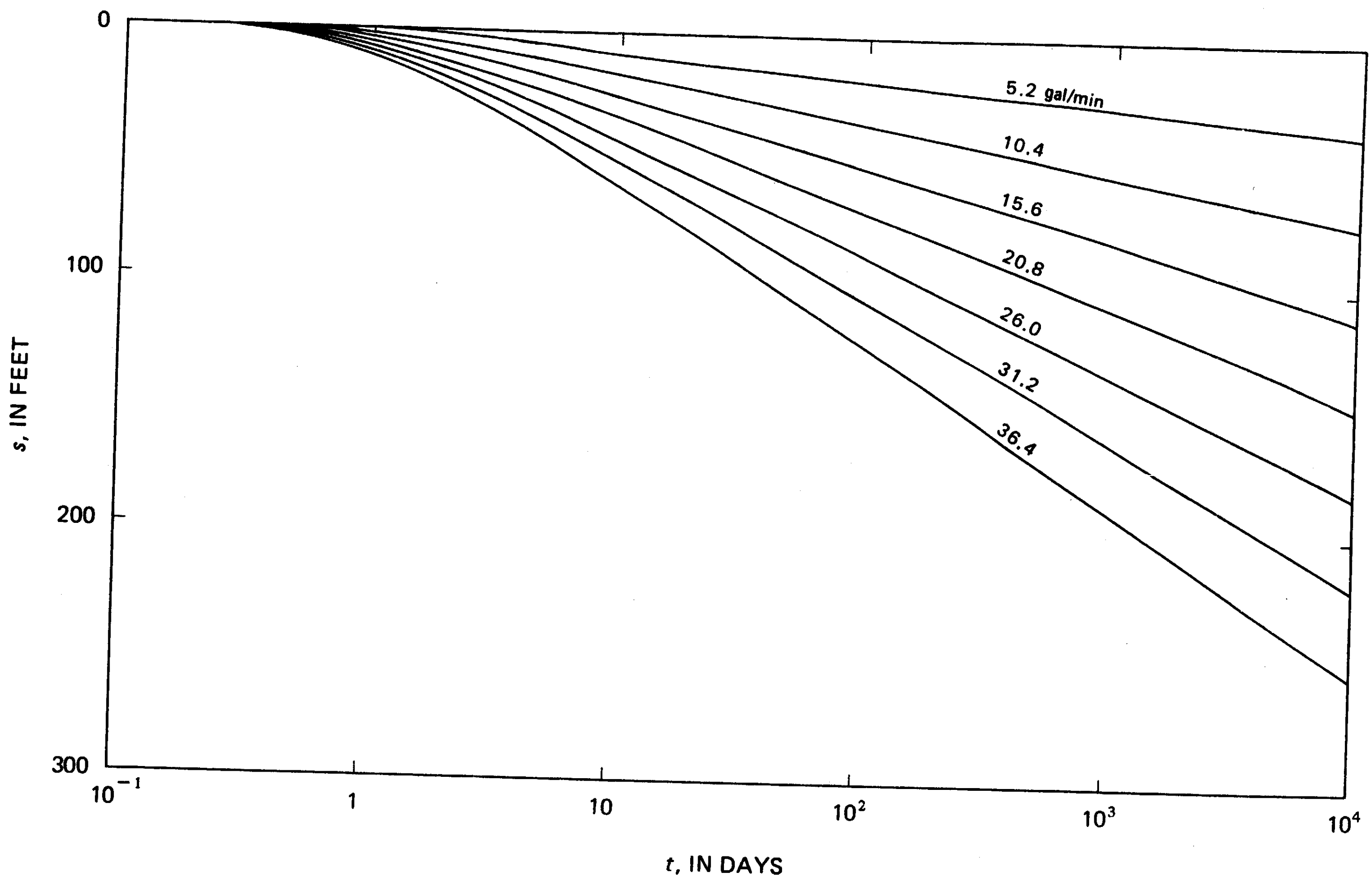


Figure X-1.4-3.— Family of curves showing drawdown produced throught time at a distances of 1000 ft from a well discharging at selected rates from a confined aquifer for which transmissivity is 20 ft^2/day and storage coefficient is 0.00005.
(From Lohman, 1972, p. 55)

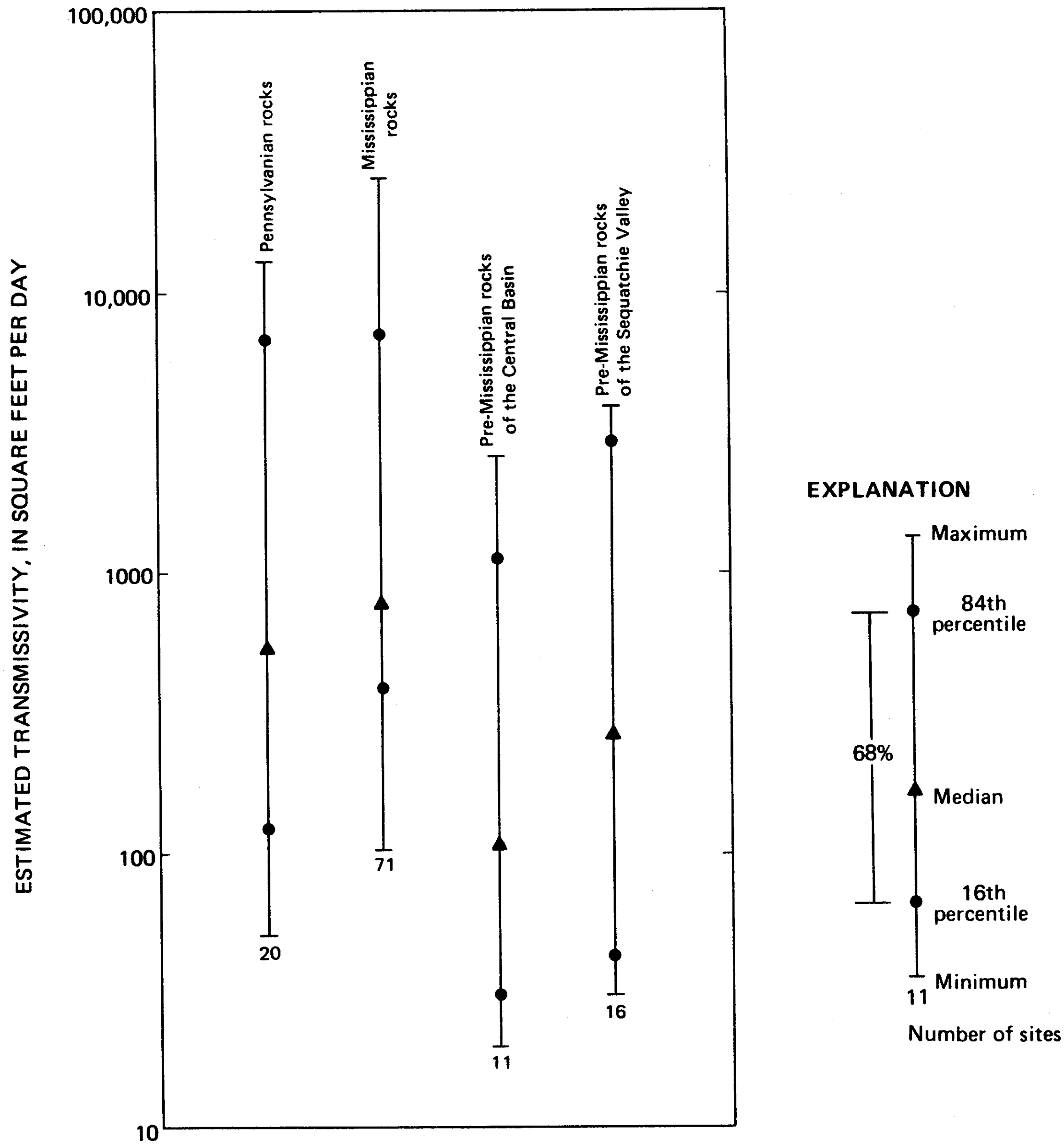
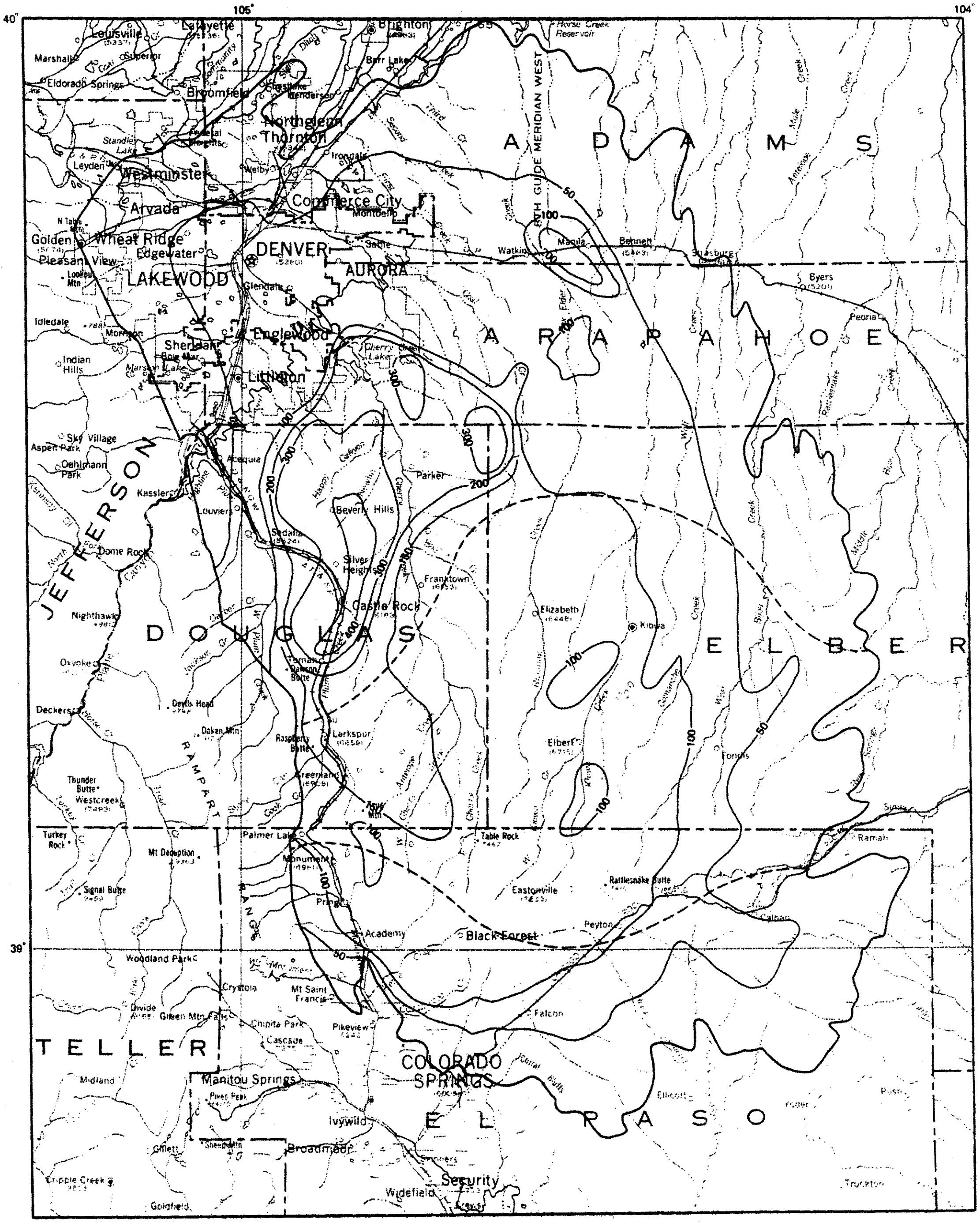


Figure X-1.4-4— Range of estimated tranmissivities of selected rock units in Eastern Coal Province, Tennessee and Alabama–Area 21.
(From May and others, 1983, p. 63)



EXPLANATION

—100— LINE OF EQUAL TRANSMISSIVITY-Interval, in feet squared per day, is variable.

——— APPROXIMATE LIMIT OF THE DENVER AQUIFER (in Denver formation)

- - - - LIMIT OF AREA IN WHICH HYDRAULIC CONDUCTIVITY WAS ESTIMATED.

Figure X-1.4-5.— Example of a transmissivity map of the Denver Aquifer in eastern Colorado.
(from Robson, 1983, fig 9)

Table X 1.4-1. Published transmissivity values of aquifers in selected permit areas.

[Rock type abbreviations: c, coal; cong, conglomerate; Is, limestone; sh, shale; sit, siltstone; ss, sandstone (listed in order of decreasing significance); frac, fractured; sol. op., solution openings; mine sub., mine subsidence. Aquifer abbreviations: Fm., formation; Ss., sandstone; Mbr., member. Geologic age abbreviations: Miss., Mississippian; Quat., Quaternary; Pa., Pennsylvanian; Paleo.-Cret., Paleocene-Cretaceous]

Coal Province	State	Aquifer	Geologic Age	Rock type	Transmissivity (feet squared per day) Range min.	max.	Median or Mean	Method of Analysis	Source of Information
Eastern	MD	Conemaugh Fm.	Pa.	frac, slt, ss, sh, & c.	3.	550.	19.	slug test (Bouwer & Rice, 1976)	(1)
	TN & KY	(not defined)	Pa.	frac, ss, sh, slt, & c.	5.	13,000.		specific capacity	(2)
			Miss.	ls, slt, & sh. sol. op.	35.	7,000.			
	TN	(not defined)	Pa.	frac. ss, sh, slt, & c.	20.	2,000.		specific capacity	(3)
			Miss.	ls, slt, & sh. sol. op.	7.	8,000.			
	TN & GA	Crab Orchard Mtns. Fm. Gizzard Fm.	Pa.	frac. ss, sh, slt, & c.	20.	750.		specific capacity	(6)
		Pennington Fm.	Miss.	ls sol. op.	30.	900.			
	TN & AL	Pottsville Fm.	Pa.	frac, ss, cong, sh, & c.	50.	13,000.	520.	specific capacity	(10)
			Miss.	ls sol. op.	20.	27,000.	800.		
	WV	Conemaugh Fm.	Pa.	frac, ss, slt, sh, & c. mine sub.	.13	41.	13.	slug test	(5)
Northern Great Plains	MT	'overburden'	Paleo-cene and Holo-cene		.008	44.	3.3	aquifer tests	(12) (13)
		(1)Anderson coal			3.2	120.	34.	aquifer tests	
		(2)Dietz 1 coal			16.	150.	(2 values)		
		(3)Dietz 2 coal			6.5	441.	40.		
		Combined aquifers (1), (2)&(3) beds			44.	84.	55.		
Rocky Mountain	UT	Ferron Ss. Mbr. Mancos Shale	Creta-ceous	ss	10.	800.	100	aquifer tests	(19)
	CO & NM	Alluvium	Quat.		0.16	570	92.9	aquifer tests	(20)
		Raton and Vermejo Formations	Paleo.	c&carb.sh	0	38.0	9.26		
		Trinidad Ss.	-	ss	0	49.2	8.11		
			Cret.	slt&sh	0	90.7	16.9		

1.5 References Cited for Aquifer Characteristics

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2. Aquifer Testing

Pertinent 30CFR[1]- Sections:
Description of hydrology and geology.
Ground-water information.

2.1 Introduction and General Procedure

Quantification of changes in aquifer storage due to recharge and discharge requires reliable estimates of transmissivity, storativity, and apparent specific-yield values as well as knowledge of hydraulic boundary conditions. Solutions to mine-related hydrology problems, such as inflow and disturbance of the potentiometric surface through time and with distance, also require accurate values of these terms.

An aquifer test is a controlled field experiment to determine the hydraulic properties of water-bearing deposits and rocks. The procedure is to cause a stress on the aquifer and to measure the observed aquifer response; the hydraulic-property values are then obtained by matching the measured response to mathematically derived relationships (type curves) between flow and aquifer pressures. Aquifer testing by pumping consists of either discharging water from or injecting water into the aquifer, and observing the response by measuring water levels in nearby wells. An example of an aquifer-test layout, with map and cross section, is shown in figure X-2.1-1.

The procedures for aquifer test design, field observation and data analysis outlined herein are modified from the published literature (9), (10), (14), (15), and (17). Elementary analytical methods commonly applied to geologic and hydro-logic settings of coal mines in the United States are listed in table X-2.1-1; detailed descriptions of these methods are presented in the following sections. Consideration of hydrologic boundary conditions and tight formations also are presented. More complete discussion of the analytical techniques for a particular problem are given in (16) and (17).

The purpose of an aquifer test design is to yield reliable values of hydraulic coefficients; therefore, the design phase is probably the most important aspect. The cost of an aquifer test ranges from a few hundred dollars for the least complicated type to thousands of dollars for a more detailed or sophisticated test. The cost depends on the number and distribution of observation wells, the duration of the test, and the manpower and equipment allocated to the field test. To increase the probability of success and to avoid unnecessary costs and waste effort, the tests should be designed carefully and followed by careful and complete data collection.

[1]CFR= Code of Federal Regulations

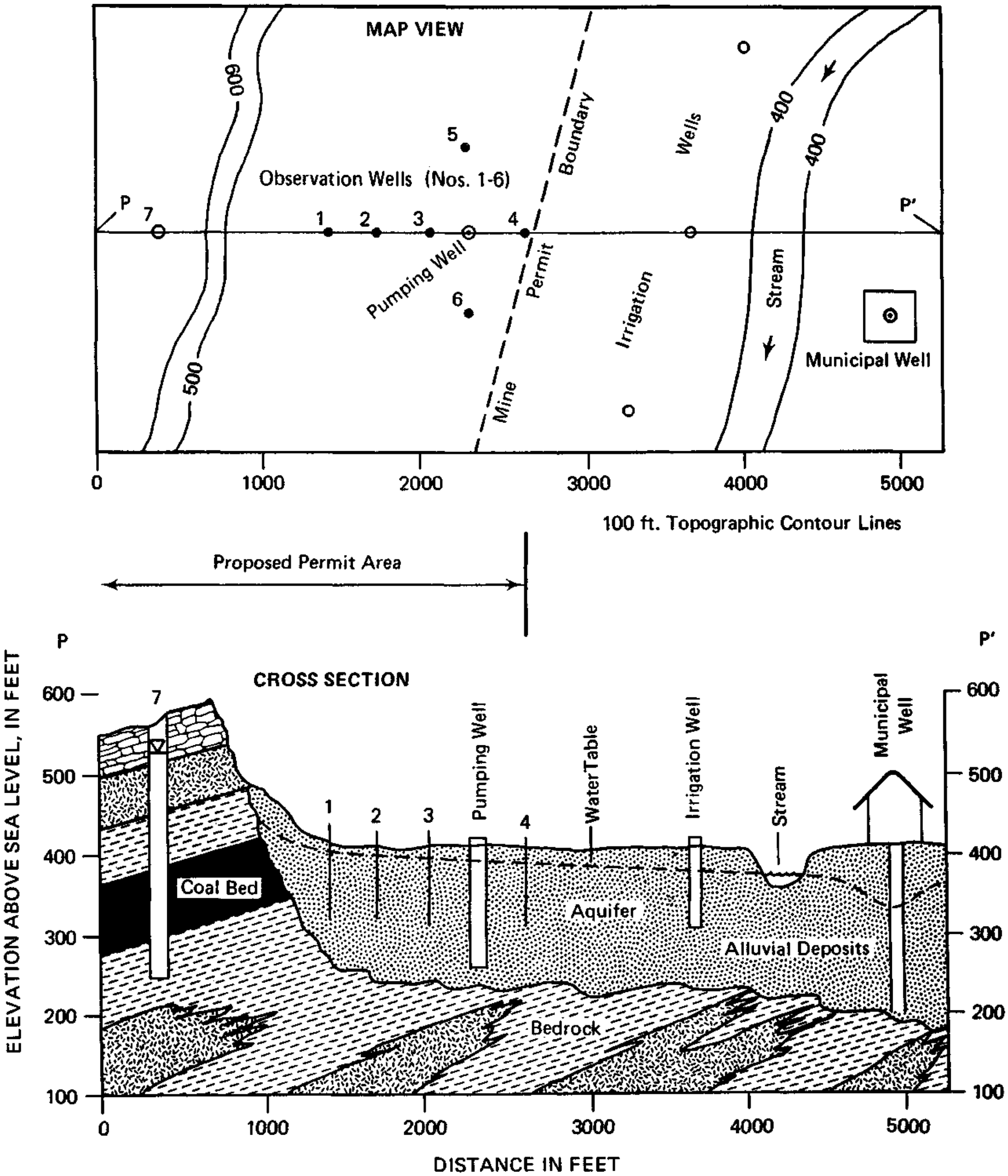


Figure X-2.1-1.— Example of an aquifer-test layout.

Information desirable in the design phase includes:

1. Knowledge of the geologic and hydrologic setting of the aquifer system in sufficient detail that the conditions controlling ground-water flow can be defined. For example, the following should be known: aquifer thickness and extent, basic lithologic composition, type of aquifer (confined or water table), approximate order of magnitude of transmissivity and storage coefficients, and magnitude of stress to be imposed on the aquifer.
2. Response (drawdown) curves, based on imposed stress and known boundary conditions, if available.
3. Type of equipment available and whether it will produce an adequate set of measurements. Some examples of equipment are devices for measuring well discharge, drawdown in observation wells, and field temperature and specific conductance.

After the geologic and hydrologic settings (item 1 above) have been evaluated, the next step is to locate and evaluate available wells in the area at which tests are proposed. The cost of testing may be reduced by using production and abandoned wells rather than installing new wells. Use of available wells is contingent upon knowing well depths, locations of intake screens, and aquifer (s) penetrated. Unfortunately, few such wells are suitable for aquifer-test purposes, and most are poorly equipped for water-level observation.

Some criteria for test-site evaluation are as follows:

<u>Pumped Well</u>

1. The pumped well must have a pump and discharge-control equipment. If the discharge is not controlled carefully at a constant value, the test results will be unreliable and thus unusable.
2. The water discharged must be conducted away from this well and not spilled on the ground where it could recharge the aquifer during the test. This is particularly important in testing shallow unconfined aquifers or deeper aquifers in fractured formations.
3. The wellhead and discharge lines should be accessible for installing discharge-regulating and monitoring equipment.
4. Measurement of depth to water in the pumped well before, during, and after pumping must be possible.
5. The diameter, depth, and position of all intervals open to the aquifer in the pumped well, as well as the total depth, should be known.

<u>Observation Well</u>

1. All observation wells must be completely developed, that is, interactive with the aquifer. Abandoned wells may tend to become clogged; consequently, they should be pumped or bailed for complete development.
2. Depth, diameter, screened or open interval, and land-surface elevation should be known for each observation well.
3. Distance from the pumped well to each of the observation wells must be determined.

Aquifer

1. Depth to, thickness of, and boundary conditions of the aquifer should be determined.
2. The aquifer and any discontinuities caused by changes in lithology or by incised streams and lakes should be mapped.
3. Hydraulic properties of the aquifer and adjacent rocks must be estimated. Estimates of transmissivity and storativity may be obtained from reports on aquifers in geologic and hydrologic settings similar to that of the mine-permit area. Estimates based on general lithology and relative degree of confinement are available in references (5), (9), and (16).

Site evaluation serves not only to determine the location and configuration of wells in the area, but to plan well installations specifically for the aquifer test. Estimating the radius of influence from the pumping well as a function of time can be determined from estimates of transmissivity and storativity based upon the Theis method of analysis (20). If only one observation well is to be installed, this procedure could be used to establish the distance from the control well within which the drawdown would be sufficient for analysis. Examples of response curves showing the relationship between drawdown and transmissivity, storativity, and distance from pumped well are depicted in figure X-l.4-1; the use of such curves for pretest evaluation and the uncertainties associated with water-table flow and partial penetration of the wells is discussed in (15).

Table X 2.1-1 - Selected aquifer test methods.

[T = transmissivity, K = hydraulic conductivity, S = storage coefficient, storativity, or apparent specific yield (S) is determined only for multi-well tests, or, under certain circumstances.]

Type of Test	Analytical method	Hydraulic Characteristics Determined	Remarks
Drawdown	Theis (1935) (curve match)	T, K, S	Multiwell: Used to determine hydraulic values averaged over a relatively large volume of aquifer; gives best reliability but is relatively expensive and requires at least two wells (preferably more), a pump, and a power source. Useful to detect uncertain boundary conditions, leakage or directional permeability. Does not work well in tight, K, formations.
	Cooper-Jacob (1946) (straight-line)	T, K, S	Multiwell: Same remarks as Theis method insofar as limiting assumptions are met, and u is less than 0.01 (See table X-2.3-2). Boundary effects may cause serious errors and must be recognized.
Recovery	Theis (1935) (straight line)	T, K, S	Multiwell: Same remarks as Theis method insofar as limiting assumptions are met. Boundary effects may cause serious errors. Single well only: Ccmmonly useful for determining T in recovering pumped well at small additional cost. Method is not recommended for determining S in tight aquifers where well-bore storage effects are evident
Slug	Cooper and others (1967) (curve match)	T, K, (S)	Single well only: Simple and inexpensive method used to estimate coefficients in a small volume of a confined aquifer. The S determination is rather insensitive to type-curve matching and is not recommended. Works well in tight aquifers. This method accounts for aquifer and well-bore storage.
	Bouwer and Rice(1976) (straight line)	T, K	Single well only: Same remarks as for Cooper and others (1967) method except that it is applied to water-table aquifers and accounts only for well-bore storage. Can also be used for partially penetrating or perforated wells
Flowing Well	Jacob and Lohman (1952)	T, K, S	Single well only: Constant drawdown and variable discharge for flowing well; artesian well must be shut-in until head is static; period of testing ranges from 2 to 4 hours; radius of well must be known to determine S.

2.2 Field Observations

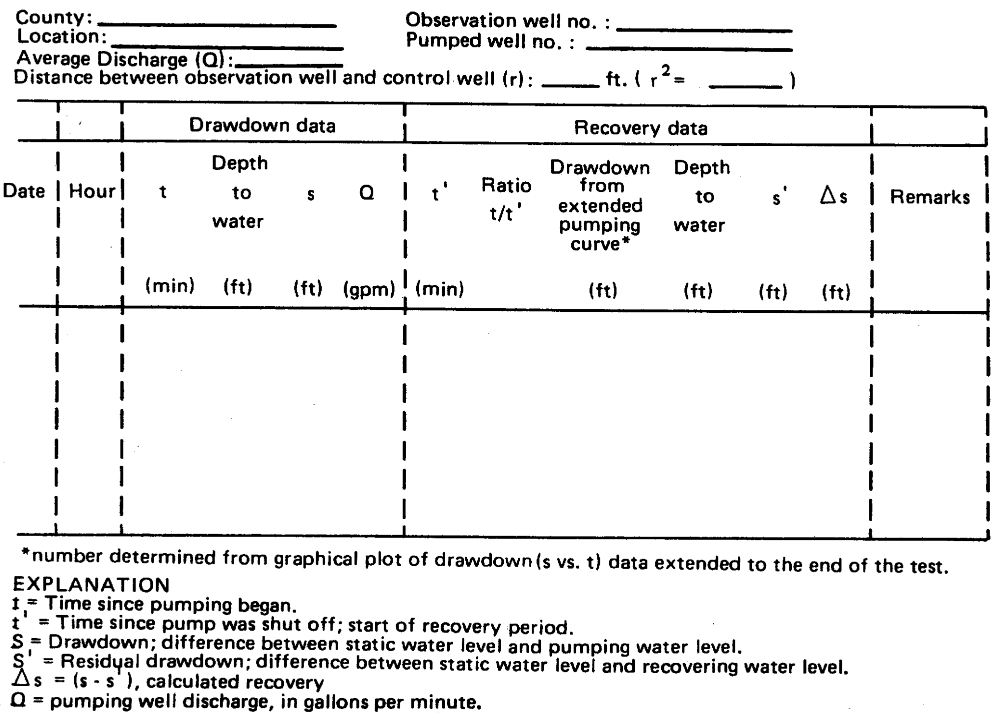
The hydrologic data required for analysis are listed below with the precision in measurement generally considered acceptable in parentheses. (See ref. 15):

1. pumped well discharge (± 10 percent),
2. depth to water below measuring point (± 0.01 ft),
3. distance from pumping well to each observation well (± 0.5 ft),
4. synchronous time (±1 percent of time since pumping started),
5. descriptions of measuring points,
6. elevations of measuring points (±0.01 ft),
7. vertical distance between measuring point and land surface (±0.1 ft),
8. measured depths of all wells (±1 percent),
9. depth and length of screened intervals of all wells (± 0.1 ft),
10. diameter, casing type, screen type, and method of construction,
11. location of all wells in plan, relative to land-survey net or by latitude and longitude, on 7½ minute U.S. Geological Survey topographic map.
12. specific conductance of discharge.

The observations needed to define boundary conditions and measure aquifer response are decided in the next phase of the aquifer test. For example, the type of geologic and hydrologic setting roughly indicates the response of the aquifer to pumping, which in turn allows the timing of water-level measurements to be estimated. Adequate attention to aquifer test design, aids the efficient allocation of the observer's time and will provide the most useful information.

Items 1 through 4, above, are documented specifically in the testing process. Some of these items are recorded on well-schedule and water-level forms, which become part of the permanent records. An example of a form for recording water-level and discharge data is shown in figure X-2.2-1. Specific conductance measurements can be recorded in the "remarks" column. An example of a drawdown and recovery curve in which the symbols and critical points are explained is shown in figure X-2.2-2.

Lithology of the aquifers and construction features of the pumping well and observation wells are generally determined by interviewing the well owners and well drillers. Field measurements of well depths and casing diameters afford a rough check on the accuracy of information obtained by the interviews. Detailed lithologic logs should be made and construction features noted during drilling of all wells installed for the test. Geophysical logs such as resistivity, self-potential, gamma, caliper, temperature, and neutron logs provide more detailed information on subsurface conditions. Accurate positioning of observation wells and pumped wells is especially important in a test of a heterogeneous or anisotropic aquifer. Position of the test site with respect to a regional land-survey net or by latitude and longitude must also be noted so that the regulatory authority can use the data to interpret the regional characteristics of the aquifer.



County: ______ Observation well no. : ______
Location: ______ Pumped well no. : ______
Average Discharge (Q): ______
Distance between observation well and control well (r): ____ ft. (r^2 = ____)

Date	Hour	Drawdown data				Recovery data							Remarks
		t	Depth to water	s	Q	t'	Ratio t/t'	Drawdown from extended pumping curve*	Depth to water	s'	Δs		
		(min)	(ft)	(ft)	(gpm)	(min)		(ft)	(ft)	(ft)	(ft)		

*number determined from graphical plot of drawdown (s vs. t) data extended to the end of the test.

EXPLANATION
t = Time since pumping began.
t' = Time since pump was shut off; start of recovery period.
S = Drawdown; difference between static water level and pumping water level.
S' = Residual drawdown; difference between static water level and recovering water level.
Δs = (s - s'), calculated recovery
Q = pumping well discharge, in gallons per minute.

Figure X-2.2-1.— Example of an aquifer-test from for drawdown and recovery. (Modified for Stallman, 1971, fig.3)

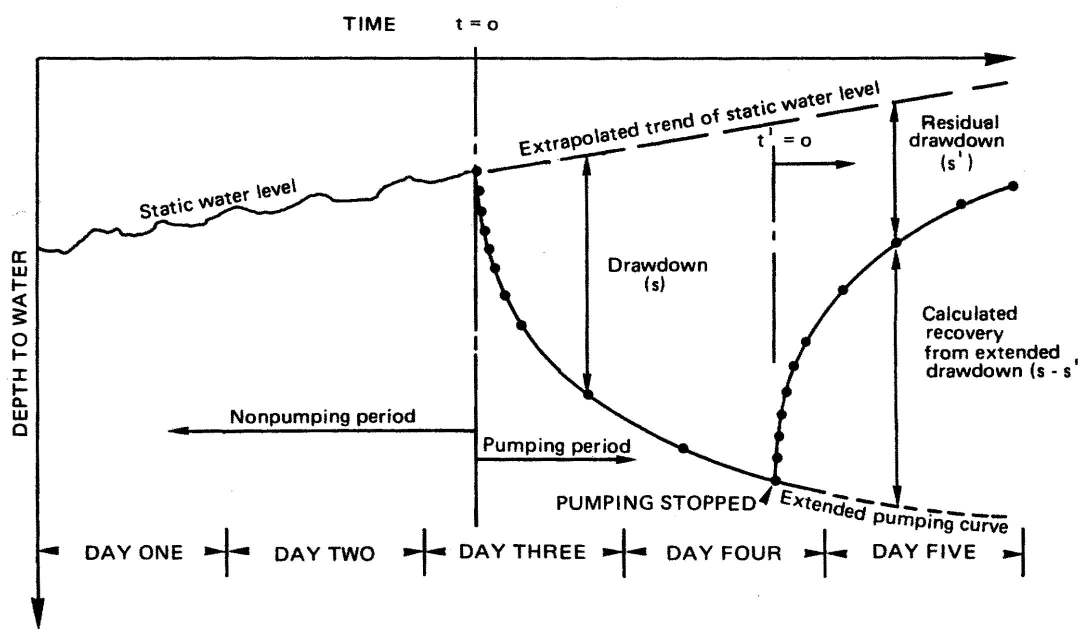


Figure X-2.2-2. Example of hydrograph of water levels in observation well before, during and after pumping test.
(Modified for Johnson Division , 1975, figs. 95 and 97; and Stallman, 1971, fig.4)

Most "type curves" indicate the change in potentiometric head, recharge, or drawdown during pumping. However, the changes in observed water level during the test may also include the effects of tides, recharge, and atmospheric pressure changes. Also, because flow in most aquifers is nonsteady; the trend of the static water level with time should be determined before testing so that drawdown measurements can be adjusted during the test. Drawdowns can be determined accurately only if background water-level trends are known or the drawdown due to pumping is large compared to other effects.

General rules regarding the length of time for aquifer tests are:
1. the minimum length of an aquifer test should be 24 hours for a confined aquifer and 3 days for a water-table aquifer (8);
2. period of pretest water-level measurements should be at least twice the length of the aquifer test. Longer aquifer test times are desirable, considering the application of the hydraulic properties to, and the drawdown effects of, long-range water-use estimates.

Water levels in many confined-aquifer wells fluctuate in response to changes in atmospheric or barometric pressure; that is, increases in atmospheric pressure cause a greater depth to water, while decreases in atmospheric pressure cause a lesser depth to water. The 6-day hydrographs in figure X-2.2-3 illustrate this relationship for two artesian wells. Changes of a foot or more in barometric pressure can cause fluctuations of a foot or more in the potentiometric surface at these wells within a few days (2).

If water-level responses to pumping in a confined aquifer test are relatively small, drawdown measurements should be corrected to remove the barometric effects before comparison with type curves. The correction procedure entails:
1. measuring the water-level variations caused by barometric pressure changes (as shown in fig. X-2.2-3) before the aquifer test;
2. determining the barometric efficiency(BE) of the well;
3. measuring barometric pressure during the aquifer-test period;
4. correcting the measured drawdowns (before applying the type-curves) accordingly by applying the barometric efficiency. The barometric efficiency of an aquifer may be expressed as shown below (6), (17), (18)

$$BE = s_w/s_b \qquad \text{(X-2.2-1)}$$

where:

s_w = the net change in water level observed in a well tapping the aquifer, in feet or centimeters, and

s_b = the corresponding net change in atmospheric pressure, in ft (or cm).

For example, a well having a 0.50 barometric efficiency would have a water-level rise of 0.05 ft for each decrease of 0.10 ft in barometric pressure. Thus, for each decrease of 0.10 ft in barometric pressure since the start of the aquifer test, the measured water levels, or drawdowns, would have to be decreased by 0.05 ft to account for the atmospheric-pressure changes.

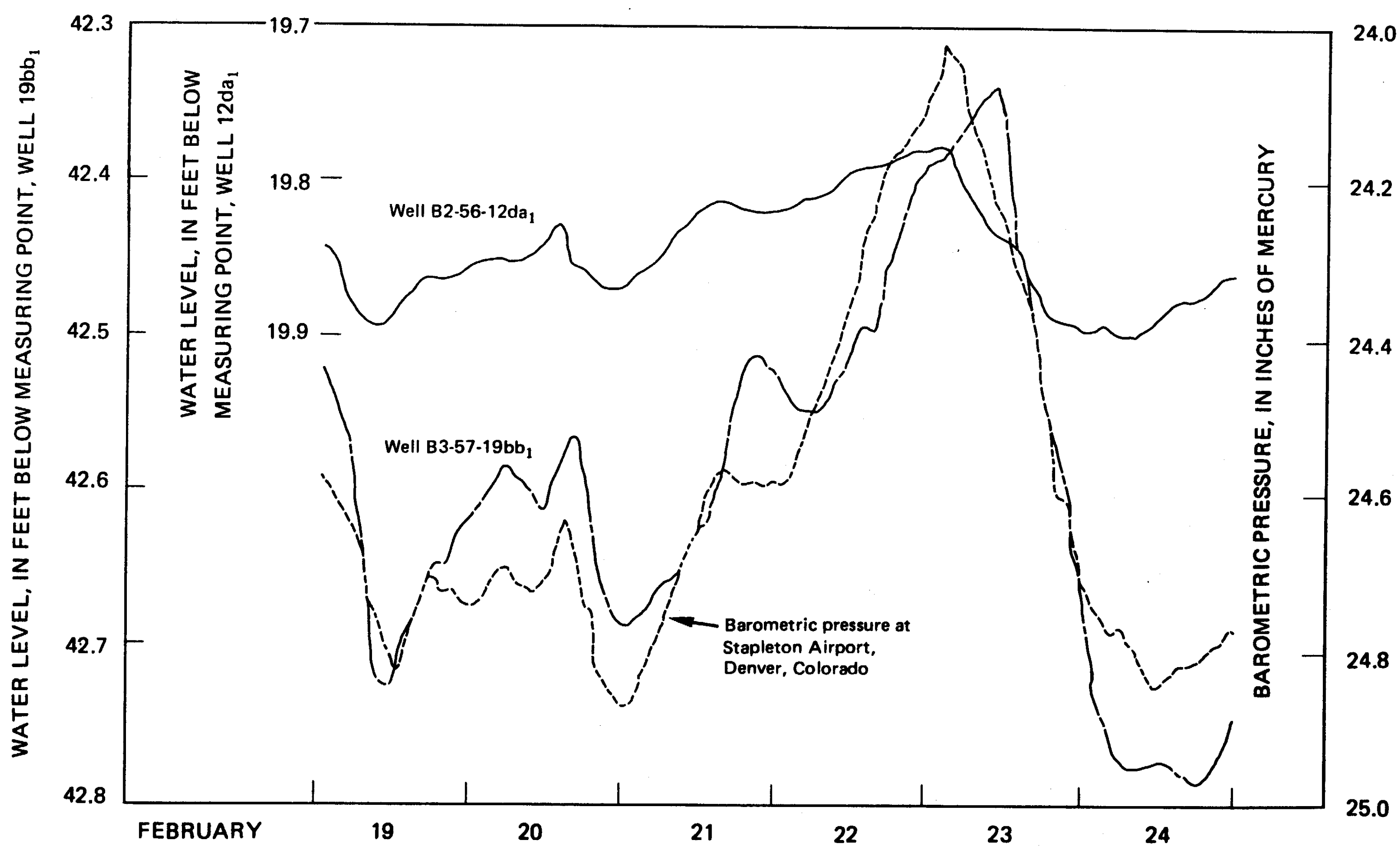


Figure X-2.2-3.— Water-level fluctuations caused by changes in barometric pressure. (Modified from U.S. Department of Interior, 1968, fig. E:1a-2)

Water levels in wells also are affected by other background factors such as the operation of nearby wells, recharge, earth and ocean tides, and pulse of force on the aquifer by trains and earthquakes. Measurements of water levels before the aquifer test will identify the extent to which such factors affect the water levels. Drawdown effects due to nearby pumping wells ordinarily can be removed from the data if the pumping times and discharge rates of these wells before and during the test are known.

Depth to water in all wells should be measured with sufficient frequency that each logarithmic cycle of time on the data plots contains at least 10 data points spread through the cycle. Thus, after t = 0, depth to water should be measured in each well at t = 1, 1.2, 1.5, 2, 2.5, 3, 4, 5, 6, 7, and 8 minutes, approximately, and all succeeding decimal multiples (10, 12, 15, 20, 25, 30, 40, 50, 60, 70, 80, 100; and 120, 150, 200, 250, 300, 400, 500, 600, etc. minutes) of these numbers until the end of the test (15). If the test design indicates that measureable drawdown is not expected at a given observation well for several hours after the test starts, the early measurements may, of course, not be necessary.

Discharge from the pumping well can be measured in several ways:
1. at wells with open discharge pipes— by the trajectory method, Hoff-meter method, and volumetrically with a tank and stopwatch method.
2. at wells with closed discharge pipes— by Pitot-tube and clamp-on ultrasonic flowmeter methods (29).
3. inline discharge measurements— by flow-rate pass-through meters and total-flow meters.
4. ditch or canal measurement— Hoff-meter, Price type M or pygmy current meters, weirs, or flumes.
The discharge is maintained at a constant rate by means of a gate-control valve. The discharge must be conducted away from the area of the pumped well.

For tests in which discharge is to be held constant throughout the test, the discharge should be measured periodically and adjusted as needed. Pumps powered by electric motors produce the most constant discharge. This discharge rate should not be allowed to vary more than $\pm$ 10 percent or it will produce aberrations in drawdowns that are difficult to analyze (15). Maintaining discharge at $\pm$ 10 percent is difficult in aquifers with low transmissivity that require flow rates less than 5 gal/min. Furthermore, discharge from tight formations must be checked and adjusted frequently, because the drawdowns are large.

2.3 General Approach to Aquifer- Test Analyses

Aquifer-test analyses involve the graphical transformation of field data into estimates of the hydraulic properties of the aquifer. These analyses include Thies, Jacob, recovery, and slug-test methods. Interpretation of aquifer tests requires a simplified concept off the aquifer, its boundaries, and the stress imposed on the aquifer. This reduction of a complex field setting to a quantifiable simplified aquifer representation is important in the determination of hydraulic coefficients. The objective is to closely simulate the observed effects of the stress imposed on the aquifer during the test.

The following examples of the use of type-curve methods apply to aquifer tests in relatively simple hydrologic settings. Analysis of more complex problems in the literature - (6), (8), (9), (3jO), (14), (15), (16), (17), (18), (22) - usually involves similar curve-matching techniques however.

The type-curve method devised by C. V. Theis (20) calculates values for two terms—transmissivity (T), and storativity (S)—in the equations:

$$s = \frac{Q\,W(u)}{4\pi T} \qquad \text{X-2.3-1}$$

and

$$u = \frac{r^2\,S}{4\,T\,t} \qquad \text{X-2.3-2}$$

where:

s = drawdown in response to the pumping
Q = pumping rate
T = transmissivity
S = storativity or storage coefficient
r = distance from the pumping well,
t = time since pumping began

In this procedure, r and t combine with T and S to define a dimensionless variable, u, and corresponding dimensionless response function W(u). Type-curve methods use W(u) in relation to u or 1/u. This; manual uses W(u) in relation to 1/u. A table of W(u) values for a range of u values is presented in table X-2.3-1; the graph of this relationship is shown in figure X-2.3-1.

Briefly, the method consists of plotting a function curve or type-curve, such as (1/u, W(u)) on logarithmic paper (figure. X-2.3-1) and plotting the time versus drawdown (t,s) data on a second sheet of logarithmic paper, having the same scales. If the two sheets are superimposed and matched with coordinate axes parallel, as shown in figure X-2.3-2, the respective coordinate axes will be related by the constant factors:

$$\frac{s}{W(u)} = CL \qquad \text{and} \qquad \frac{t}{(1/u)} = C_2$$

The values of these two constants are:

$$C_1 = \frac{Q}{4nT} \quad \text{and} \quad C_2 = \frac{R^2 s}{4T}$$

Thus, a common match point for the two curves may be chosen, and the four coordinate points W(u), 1/u, s, and t recorded for the common match point. T and S can be obtained from the following equations:

$$T = \frac{QW(u)}{4\,n\,s} \qquad \text{X-2.3-3}$$

and

$$S = \frac{4Tut}{r^2} \qquad \text{X-2.3-4}$$

where W(u), 1/u, s, and t are the match-point values.

The methods of analysis described below are used for tests in confined aquifers with fully penetrating wells, as shown in figure X-2.3-3; symbols used in these computations are listed and defined in table X-2.3-2. Water-table aquifers(fig. X-2.3-3) are analyzed by the same Methods. If drawdowns observed in thin water-table aquifers are adjusted by subtracting $s^2/2b$, equations based on the assumption of negligible dewatering and radial flow can be used (7). Where the dewatering is significant, the data plot used is s - ($s^2/2b$) and time which is matched to the 'type curves' of artesian flow. The value of S is obtained as follows (15):

$$S = \frac{(b-s)S'}{b} \qquad \text{X-2.3-5}$$

where:

- s = the appropriate drawdown at the geometric mean radius of all observation wells at the end of pumping,
- b = the original saturated thickness of the aquifer, and
- S' = the apparent storage coefficient

Jacob's correction, $s^2/2b$, is most applicable to data from observation wells fully penetrating the aquifer but may also be used for observation wells open within the bottom two-thirds of the aquifer. Moreover, the correction is applicable only after the data begin to follow the artesian response curve (11) and does not apply if effects of partial penetration are significant.

The rate of drawdown with respect to time at any point within the cone of depression is needed for the Theis, Jacob, recovery, and slug test methods. The shape and position of the cone of depression with respect to distance at some time during the aquifer test is required for the distance-drawdown method. Certain assumptions must be considered for all methods presented. The following assumptions are common to all methods (6), (14), (15):

1. The aquifer is homogeneous and isotropic.
2. The aquifer has uniform thickness.
3. The aquifer is horizontal and infinite in areal extent.
4. The well is open to the aquifer throughout the aquifer thickness.
5. Flow to the well is laminar and uniform along the length open to the well.
6. The aquifer is bounded by relatively impermeable confining layers.
7. Water is released from storage instantaneously with a decline in head.
8. The well has a reasonably small diameter.
9. Hydraulic potential at the pumped well is the only cause of flow in the aquifer system during testing.

Even though all these assumptions are rarely satisfied in any particular aquifer test, the methods are still adequate for estimating the hydraulic properties of aquifers.

In addition to these common assumptions, each analytical method entails other assunptions. Some conditions that affect the theoretical aquifer response such as vertical boundary conditions, finite well diameter, and fractured rock permeability are briefly mentioned in the Practical Considerations chapter 2.7. Treatment of leaky aquifers, confining beds with storage, anisotrqpy, and partial penetration of wells, all of which are common factors to be dealt with in real aquifer tests, are not treated in this manual, but references on these conditions include (9), (14), (16), and (17).

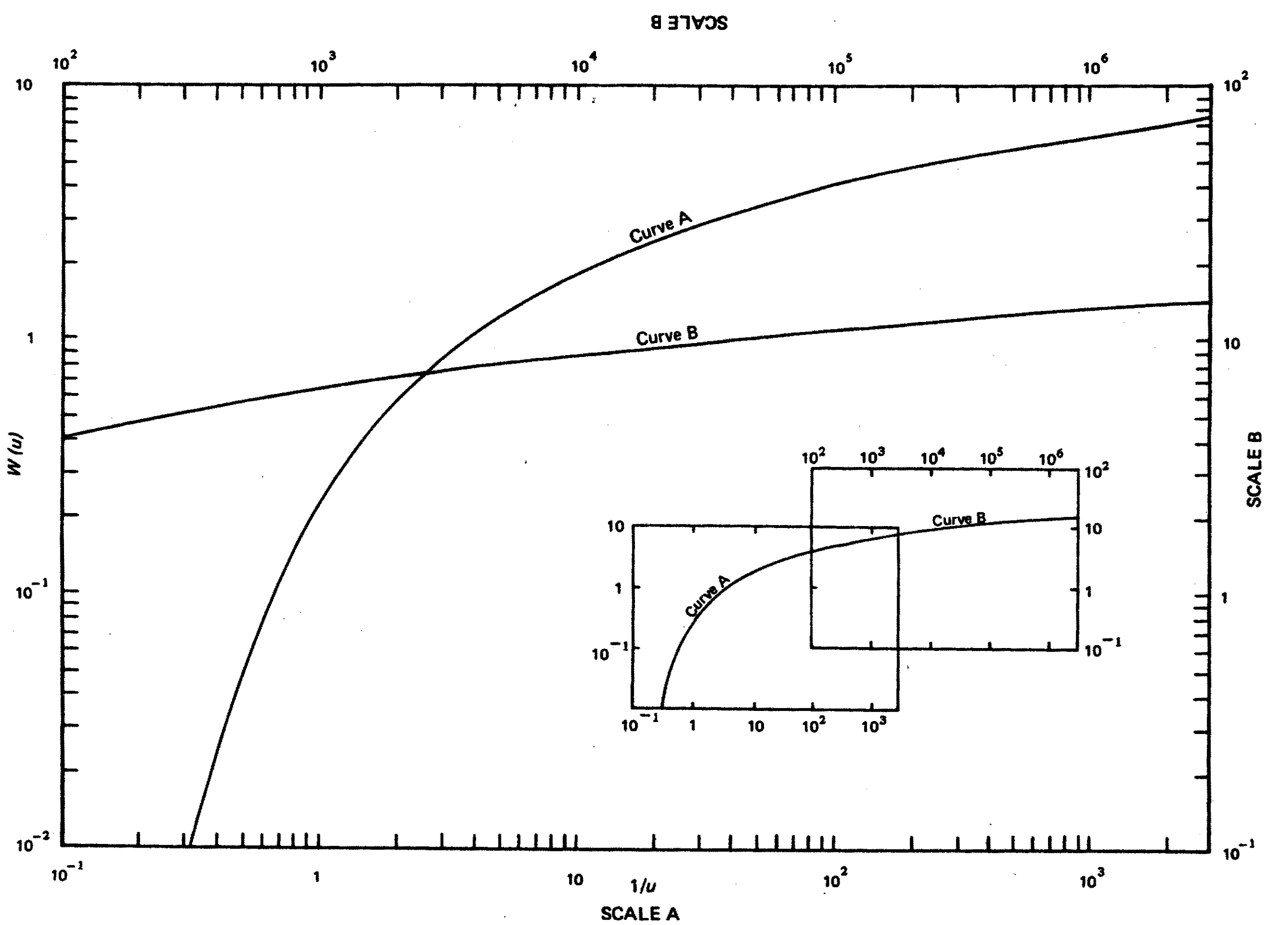


Figure X-2.3-1.— Type curve of dimensionless drawdown (W(u)) in relation to dimensionless time (1/u) for constant discharge from an artesian well (Theis curve). (Insert shows range of fields for curve A and curve B.) (Table X-2.3-1 presents W(u) versus 1/u.) (From Reed, 1980, pl. 1)

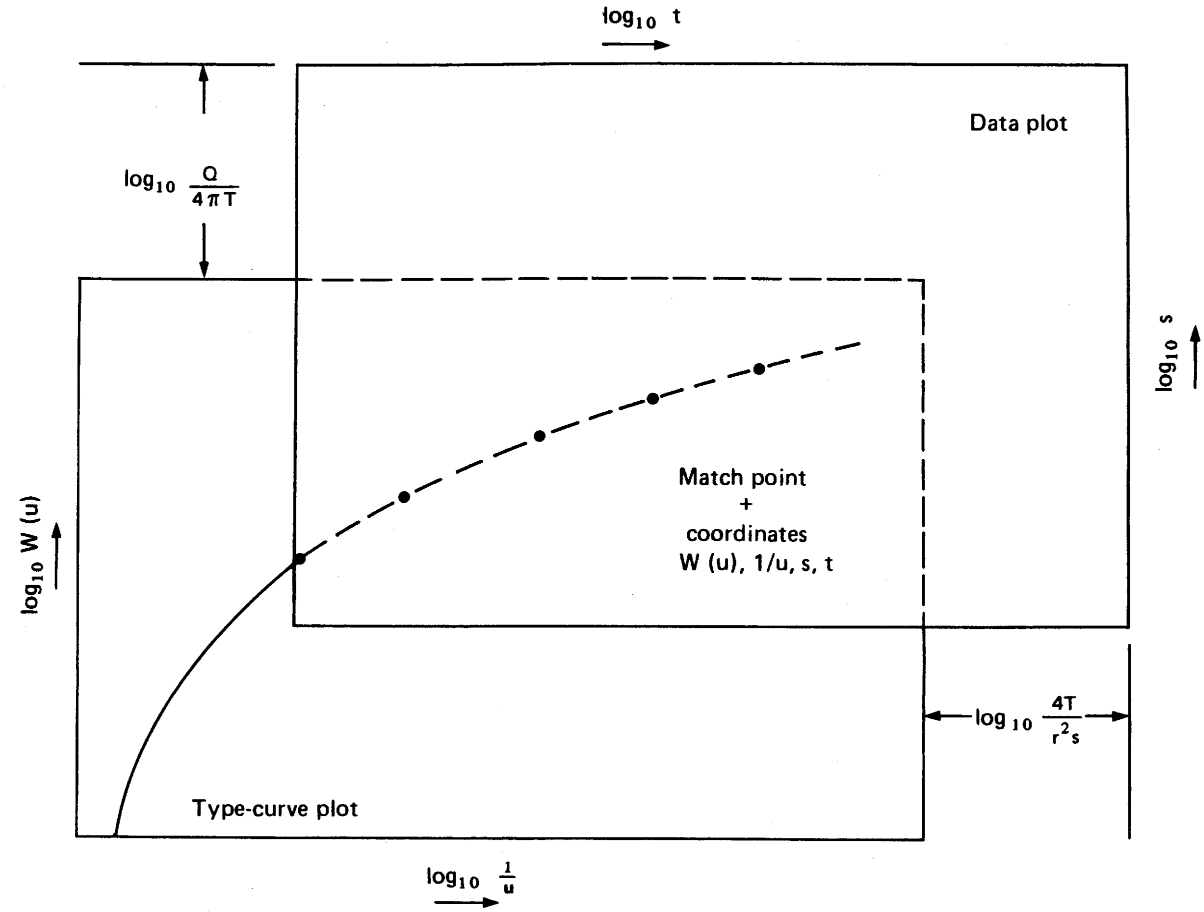


Figure X-2.3-2.— Superposition of W(u) versus l/u type curve onto B versus t data plot. (From Reed, 1980, fig. 0.1 and Stallman, 1971, fig. 1)

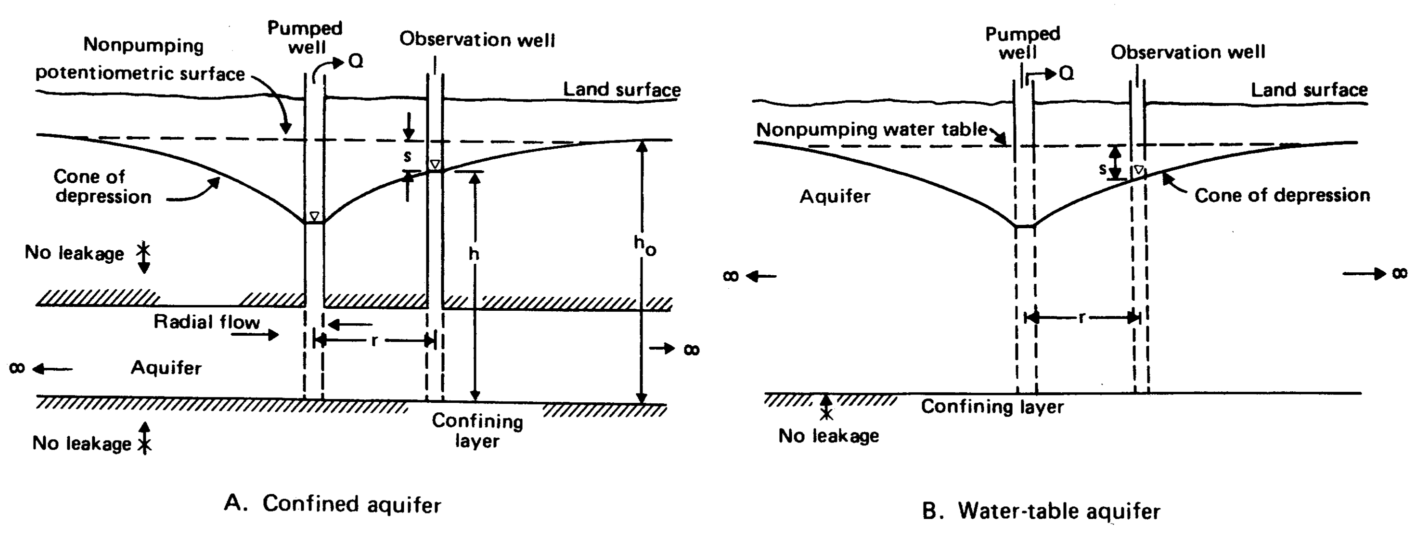


Figure X-2.3-3.— Drawdown conditions for aquifer tests with fully penetrating wells. (Modified from Walton, 1970, figs. 3.1 and 3.4)

Table X-2.3-1– Selected values of W(u) for values of 1/u. (Modified from Heath, 1983, p. 35.)

Multiplication factor for 1/u	1/u value													
	10	7.69	5.88	5.00	4.00	3.33	2.86	2.5	2.22	2.00	1.67	1.43	1.25	1.11
x 10^{-1}	0.219	0.135	0.075	0.049	0.025	0.013	0.007	0.004	0.002	0.001	0.000	0.000	0.000	0.000
x 1	1.82	1.59	1.36	1.22	1.04	.91	.79	.70	.63	.56	.45	.37	.31	.26
x 10^{1}	4.04	3.78	3.51	3.35	3.14	2.96	2.81	2.68	2.57	2.47	2.30	2.15	2.03	1.92
x 10^{2}	6.33	6.07	5.80	5.64	5.42	5.23	5.08	4.95	4.83	4.73	4.54	4.39	4.26	4.14
x 10^{3}	8.63	8.37	8.10	7.94	7.72	7.53	7.38	7.25	7.13	7.02	6.84	6.69	6.55	6.44
x 10^{4}	10.94	10.67	10.41	10.24	10.02	9.84	9.68	9.55	9.43	9.33	9.14	8.99	8.86	8.74
x 10^{5}	13.24	12.98	12.71	12.55	12.32	12.14	11.99	11.85	11.73	11.63	11.45	11.29	11.16	11.04
x 10^{6}	15.54	15.28	15.01	14.85	14.62	14.44	14.29	14.15	14.04	13.93	13.75	13.60	13.46	13.34
x 10^{7}	17.84	17.58	17.31	17.15	16.93	16.74	16.59	16.46	16.34	16.23	16.05	15.90	15.76	15.65
x 10^{8}	20.15	19.88	19.62	19.45	19.23	19.05	18.89	18.76	18.64	18.54	18.35	18.20	18.07	17.95
x 10^{9}	22.45	22.19	21.92	21.76	21.53	21.35	21.20	21.06	20.94	20.84	20.66	20.50	20.37	20.25
x 10^{10}	24.75	24.49	24.22	24.06	23.83	23.65	23.50	23.36	23.25	23.14	22.96	22.81	22.67	22.55
x 10^{11}	27.05	26.79	26.52	26.36	26.14	25.96	25.80	25.67	25.55	25.44	25.26	25.11	24.97	24.86
x 10^{12}	29.36	20.09	28.83	28.66	28.44	28.26	28.10	27.97	27.85	27.75	27.56	27.41	27.28	27.16
x 10^{13}	31.66	31.40	31.13	30.97	30.74	30.56	30.41	30.27	30.15	30.05	29.87	29.71	29.58	29.46
x 10^{14}	33.96	33.70	33.43	33.27	33.05	32.86	32.71	32.58	32.46	32.35	32.17	32.02	31.88	31.76

Examples: When $1/u = 3.33 \times 10^{-1} = 0.333$, $W(u) = 0.013$; when $1/u = 10 \times 10^{-1} = 1$, $W(u) = 0.219$.

Table X 2.3-2.3-2 –Terms used in aquifer testing.

Symbol	Definition	Units of Measure
b	Thickness of saturated part of the aquifer.	Feet.
K	Hydraulic conductivity of aquifer for horizontal flow.	Feet per day.
L	Length of open hole or screen open to aquifer.	Feet.
π	pi = 3.14159	Dimensionless.
Q	Discharge from a well.	Gallons per minute.
r	Distance from pumping well to observation point.	Feet.
r_c	Radius of casing	Feet.
r_w	Radial distance from the well center and the undisturbed aquifer, which includes the sand and gravel envelope.	Feet.
s	Change in head, or drawdown	Length.
Δs	Change in drawdown over one log cycle of t or $\frac{t}{t}$, (for straight-line solutions)	Length.
s_w	Constant drawdown (for flowing-well analysis)	Feet.
S	Storage coefficient of the aquifer(storativity).	Dimensionless.
S'	Apparent storage coefficient, observed in aquifers dewatered significantly in proportion to saturated thickness.	Dimensionless.
Sya	apparent specific yield	Dimensionless.
t	Time since pumping began.	Minutes.
t'	Time since pumping ceased, for recovery test analysis.	Minutes.
t_o	time when drawdown is zero (from extension of straight line solutions)	Minutes.
T	Transmissivity of aquifer.	Feet squared per day.
u	$\frac{r^2 s}{4Tt}$	Dimensionless.
W(u)	Theis Well function of variable (u)	Dimensionless.
y	recovery drawdown (Bouwer and Rice method)	Feet.
y_0	recovery drawdown when time is zero.	Feet.
y_t	recovery drawdown at any time, t.	Feet.

2.4 Pumping Test Methods and Analyses

Three cornmon pumping test methods used on nonflowing wells are the Theis method, Jacob straight-line method, and the recovery method. Hydrologists commonly apply all three test methods to verify results. The general procedure for each method is outlined below; assumptions beyond those mentioned earlier are included. Some data used for the examples are from Barrett and others, 1980. Metric units have been converted to inch-pound units for consistency. Aquifer-test geometry, terms, and symbols used in the sample analyses are depicted in figure X-2.3-3 and defined in table X-2.3-2.

Theis Analysis

Additional assumption:

1. Well discharge, Q, is constant starting at t = 0.

Procedure:

1. Plot s (drawdowns) against t/r^2 , such as shown in figure X-2.4-1 on transparent log-log paper[1/] having the same scale as the Theis-type curve (9) (14).
2. Superimpose this field plot over the type curve. Move the field plot over the type curve, keeping both axes parallel, until a best fit is made between the two curves.
3. Select any arbitrary match point and record the values of W(u) and 1/u from the type curve and the corresponding values of t/r^2 and s (figure X-2.4-1).
4. Insert these values of W(u), 1/u, t/r^2 , and s into equations X-2.3-3 and X-2.3-4 to determine T and S:

$$T = \frac{Q\,W(u)}{4\,\pi\,s} \qquad \text{X-2.3-3}$$

and

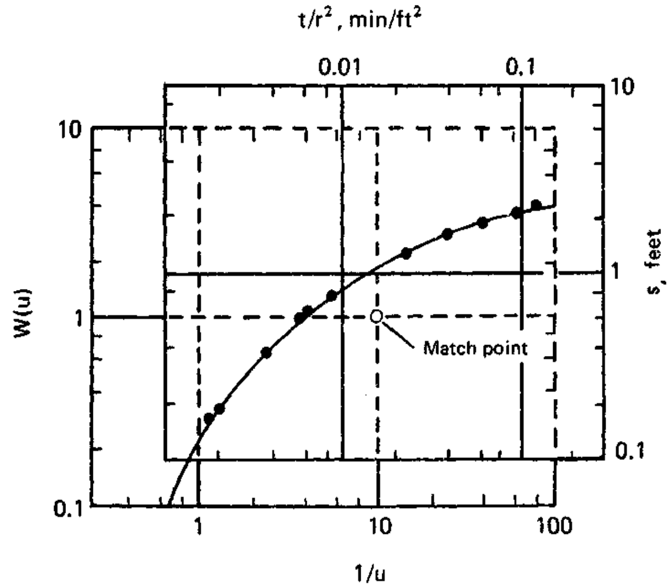
$$S = 4\,T\,u\,\frac{t}{r^2} \qquad \text{X-2.3-4}$$

Note that the field plot could be s versus t for only one observation well. For such plots, the radius can be accounted for in the formula used to compute storage coefficient. Also, the Theis analysis can be used to determine transmissivity in the pumped well.

[1/] 'K & E' - # 467323 - (Keuffel & Esser Co.) 2 cycles by 3 cycles.

(a) pumping test data.
[r, distance;
Q, discharge;
s, drawdown;
t, time;
ft, feet;
gal/min, gallons per minute;
ft^2, square feet.]

r = 65.6ft, Q = 494 gal/min



t/r^2 (min/ft^2)	s (ft)
0.00105	0.08
.00174	.16
.00197	.18
.00372	.36
.00557	.56
.00615	.59
.00837	.72
.0149	.98
.0226	1.21
.0376	1.48
.0600	1.74
.0948	2.03
.113	2.10
.119	2.13

(b) Curve matching of Theis curve to drawdown data.

(c) Calculation for Theis drawdown method.

Match point coordinates:

from fig. X-2.4-l(b), the match point coordinates:

$W(u) = 1.0$, $1/u = 10$, $s = 0.60$, $t/r^2 = 0.015$

$$T = \frac{Q\,W(u)}{4\pi s} = \frac{(494 \text{ gal/min})\,(1.0)\,(1440 \text{ min/d})}{4\pi\,(0.60 \text{ ft})(7.48 \text{ gal/ft3})}$$

$$= 12{,}600 \text{ ft}^2$$

$$S_{ya} = \frac{4T\,t}{r^2\,(l/u)} = \frac{4\,(12{,}600 \text{ ft}^2/\text{d})\,(0.015 \text{ min/ft}^2)}{(10)\,(1440 \text{ min/d})}$$

$$= 0.053$$

Figure X-2.4-1— Example of data and calculations for Theis analytical method, (From Barrett and others, 1980, p. 79)

Jacob Straight-Line Method (Cooper-Jacob Method, 1946)

Additional assumptions:

1. Well discharge, Q, is constant starting at t = 0.
2. Test must be conducted for a time sufficiently long to satisfy the condition:

$$\frac{r^2 S}{4Tt} \text{ is less than } 0.01$$

where: S = storage coefficient or apparent specific yield.

Procedure:

1. Plot drawdown on the vertical axis against time on the horizontal (logarithmic) axis of semilog paper[1/] (see figure X-2.4-2).
2. Eventually, the data should plot as a straight line. From the line, determine the change in drawdown over one log cycle.
3. Insert the change of drawdown over one log cycle (As) into the following equation to determine T:

$$T = 2.30 \frac{Q}{4\pi\Delta s} \qquad \text{X-2.4-1}$$

4. Extrapolate the straight segment of the data plot to the horizontal axis where: s = 0. Determine the time (t_o) where the line intersects the horizontal axis.
5. Insert to into the following equation to determine S:

$$S = 2.25 \frac{Tt_o}{r^2} \qquad \text{X-2.4-2}$$

6. Determine the time for which the data meets assumption 2 above by inserting the T and S values into

$$t = \frac{r^2 S}{4Tu} \qquad \text{X-2.4-3}$$

where: u = 0.01. Drawdown data collected at times less than this value of t should not be used in the straight-line plot.

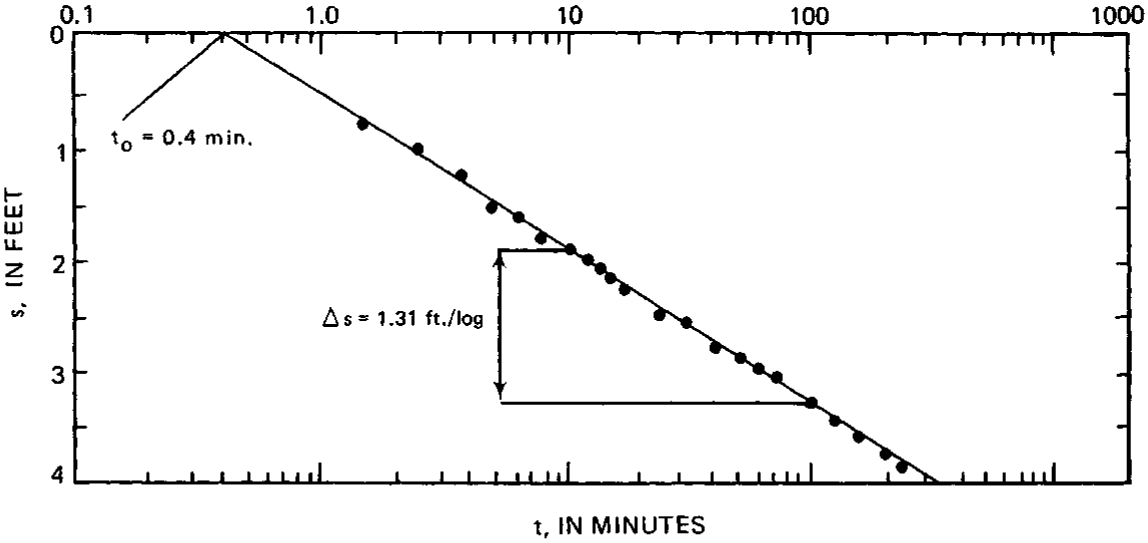
This method carmonly is used to analyze drawdown data from the pumped well. When so used, the restriction that u must be less than 0.01 is usually met. However, well-bore storage is expected to be a significant source of error in very tight aquifers. Treatment of the well storage problem is discussed in chapter X-2.6.

The coefficient of storage (storativity) from the drawdown response of a pumped well cannot be determined (6).

[1/] 'K&E'- # 466212 - 5 cyclesx70 divisions.

(a) pumping test data.
[r, distance;
Q, discharge;
s, drawdown;
t, time;
ft, feet;
gal/min, gallons per minute;
ft², square feet.]

r = 200 ft.
Q = 487 gal/min



t (min)	s (ft)
1	0.66
2	.98
3	1.21
4	1.36
5	1.48
6	1.59
8	1.74
10	1.87
12	1.97
14	2.08
18	2.20
24	2.36
30	2.49
40	2.66
50	2.79
60	2.87
80	3.03
100	3.17
120	3.28
150	3.43
180	3.51
210	3.61
240	3.67

(b) Example of Cooper-Jacob method applied to drawdown data.

(c) Calculation for Jacob drawdown method.

from fig. X-2.4-2(b), As = 1.31 ft/log cycle; t_o = 0.4 min

$$T = \frac{2.30\,Q}{4\pi\Delta s} = \frac{2.30\ (487 gal/min)\ 1440\ min/d}{4\ \pi\ (1.31)\ (7.48\ gal/ft^3)} = 13{,}100\ ft^2/d$$

$$S = \frac{2.25\ T\ t_o}{r^2} = \frac{2.25\ (13{,}100)\ (0.4)}{(200)^2\ (1440\ min/d)} = 0.00020$$

for this method to be valid, $u \leq 0.01$ or

$$t = \frac{r^2}{4}\ \frac{S}{T\ u} = \frac{(200)^2\ (0.0002)\ (1440)}{(13{,}100)\ (0.01)} = 22\ min$$

Therefore, only the data points for time greater than 22 minutes should be used to determine hydraulic coefficients by this method.

Figure X 2.4-2.— Example of data and calculations for Cooper-Jacob analytical method.
(Modified from Barrett and others, 1980, p. 80)

Theis Recovery Method

Additional assumptions:

1. Well discharge, Q, is constant starting $t = 0$. The pump is shut off some time later, $t_r = 0$.
2. Same assumption 2 as for Jacob Straight-line Method.

Procedure:

1. For each time that drawdown was measured during recovery, compute the ratio of time since pumping started, t, over time since pump was turned off t_r. (See fig. X-2.4-3).
2. Plot residual drawdown on the vertical axis against t/t_r on the horizontal axis of semilog paper. Residual drawdown is the difference between the observed water level and the non-pumping, static water level trend extrapolated from the prepumping period.
3. The data should fall on a straight line after a long time over which the change in drawdown over one log cycle (Δs) can be computed, (See fig. X-2.4-3).
4. Determine T from equation X-2.4-1:

$$T = 2.30 \frac{Q}{4 \pi \Delta s} \qquad \text{X-2.4-1}$$

5. To determine the storage coefficient, or apparent specific yield, extrapolate the straight-line segment of the data plot to the horizontal axis, and determine to at the intersection of the straight line and the horizontal axis.
6. Insert the value of t_o into the equation X-2.4-2:

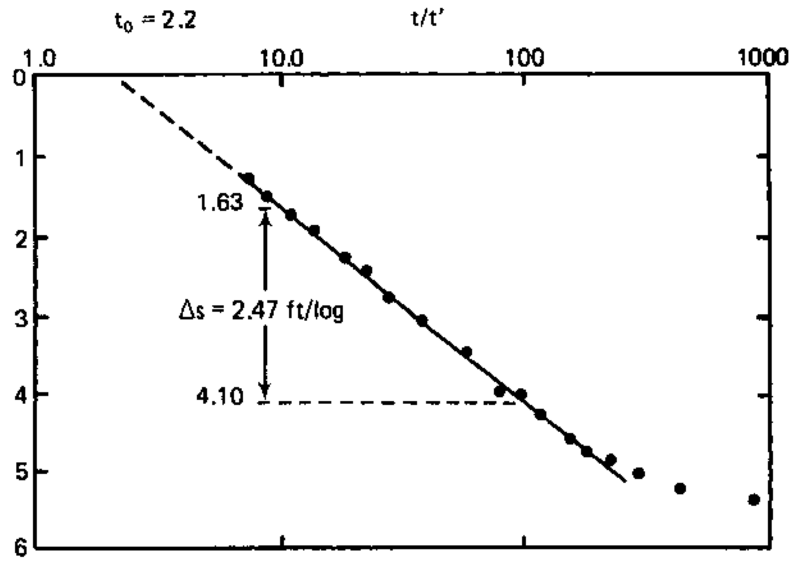
$$S = 2.25 \frac{Tt_o}{r^2} \qquad \text{X-2.4-2}$$

The precautions concerning the use of the Jacob straight-line method in the pumping well data analysis hold true for the recovery method.

(a) recovery test data.
[r, distance; Q, discharge;
s, drawdown; t, time;
t_p, pumping time;
tr, time since pump turned off;
t_o, time from graph when drawdown is zero,
ft, feet;
gal/min, gallons per minute.]

r = 15.1 ft;
Q = 473 gal/min;
t_p = 443 min

tr (min)	$\frac{t}{tr}$	s (ft)
0.5	887.	5.38
1.0	444.	5.23
1.5	296.	5.03
2.0	223.	4.89
2.5	178.	4.74
3.0	149.	4.59
4.0	112.	4.28
4.5	99.4	4.05
5.5	81.5	3.94
8.0	56.4	3.48
12.	37.9	3.05
16.	28.7	2.77
21.	22.1	2.48
26.	18.0	2.30
36.	13.3	1.94
46.	10.6	1.71
56.	8.91	1.48
71.	7.24	1.26



(b) Example of data plot for Theis recovery method.

(c) Calculations for Theis recovery method.

from fig. X-2.4-3(b) Δs = 2.47 ft/log cycle
and at s = 0 ft t/tr = 2.2

$$T = \frac{2.30\,Q}{4\pi\Delta s} = \frac{2.30\,(473)\ (1440\ \text{min/d})}{4\pi(2.47)\ (7.48\ \text{gal/ft}^3)}$$

$$T = 6{,}750\ \text{ft}^2/\text{d}$$

$$S = \frac{2.25\ T\ t_o}{r^2} = \frac{2.25\ (6{,}750)\ (2.2)}{(15.1)^2(1{,}440\ \text{min/d})}$$

$$= 0.10$$

Figure X-2.4-3.— Example of data and calculations for Theis recovery method. (Modified from Barrett and others, 1980, p. 81)

2.5 Slug-Test Method

Slug test methods are single-well tests used to determine the aquifer transmissivity of the rock material near well-bore. The basic assumptions for these methods are that (1) a known volume, V, is injected into, or removed from, the water-filled portion of the well instantaneously, at t = 0, and (2) the well is of finite diameter and fully penetrates the aquifer.

The method of Cooper and others (1967) is for application to non-leaky confined aquifers. Storage coefficient (storativity) can also be determined by this method, but with questionable reliability because of the similarity of shapes of the type curves. The determination of transmissivity, however, is insensitive to the choice of the correct curve (4).

The Bouwer and Rice method (1976) can be used to determine hydraulic conductivity of water-table aquifers with completely or partially penetrating wells, and completely or partially perforated screens. The example presented is for a fully penetrating well that is partially perforated. In addition to the assumptions mentioned above delayed yield of ground water from the unsaturated zone is ignored. Therefore, the slug volume need only be removed from the well for this water-table aquifer analysis.

Both methods provide estimates of hydraulic coefficients of aquifer material close to the well bore which commonly is altered by fracturing and (or) infiltration of drilling mud. Therefore, knowledge of near-borehole conditions, such as from drillers logs and downhole geophysical logs, is needed before values of transmissivity can be accepted as representative of the aquifer characteristics at the well site. The slug-test method has applicability when wells are not flowing or a pump is not available. The slug-test method, when applied to completely developed wells, can nonetheless provide reasonable estimates of hydraulic properties. Slug tests are most suitable for aquifers having low transmissivity, less than 7,000 ft^2/d (9).

Cooper and Others (1967) Method

Procedure:

1. Plot the ratio H/HQ on the vertical axis and time(t) on the logarithmic axis of semilog paper. (See fig. X-2.5-1). where:
   - H = head inside the well at some time, t, after injection or removal of the slug, above or below the initial head. (See figure X-2.5-1).
   - H_o = head inside the well above or below initial head at instant of slug injection or removal.
   - r_c = radius of casing in interval over which level fluctuates. rw = radius of well screen or open hole.

   For a known slug volume, V: $$H_o = \frac{V}{\pi r_c^2} \qquad \text{X-2.5-1}$$

2. Superimpose the field plot onto the suitable type curve of H/HQ versus Tt/r_c^2 (9), (14). Keeping axes parallel, adjust the field plot until a best fit is achieved (fig. X-2.5-2).

3. Select an arbitrary match point and record the value of t (from the field plot) which correlates to Tt/r_c^2 the type curve.

4. Compute T from these values of t and Tt/ r_c^2 in the equation

$$T = \frac{(Tt/r_c^2)}{t} r_c^2 \qquad \text{X-2.5-2}$$

5. Compute S by inserting oc (type-curve designation), into the following equation:

$$S = oc \frac{t_c^2}{r_w^2} \qquad \text{X-2.5-3}$$

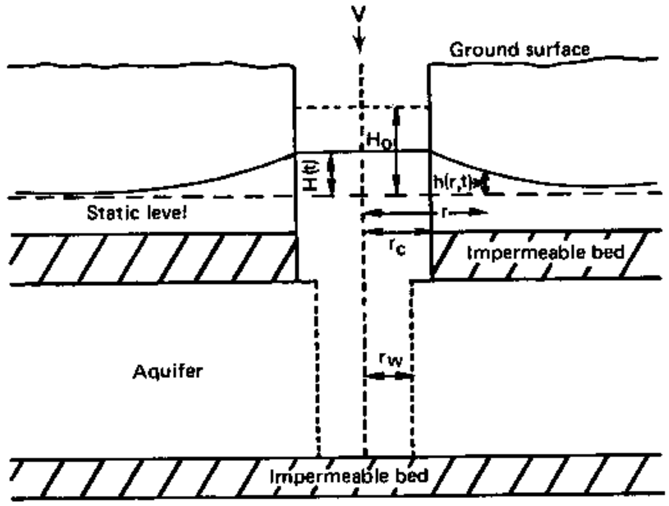


Figure X-2.5-1.— Cross section through a well in which a known volume is instantaneously injected.
(Modified from Reed, 1980, fig. 9.1)

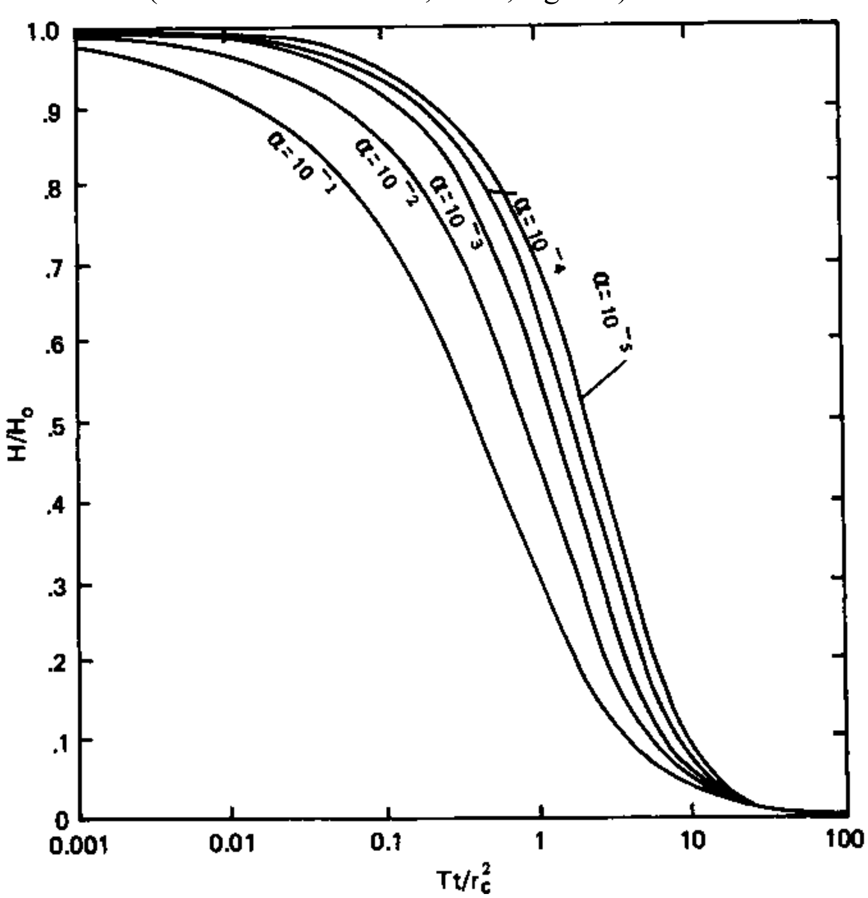
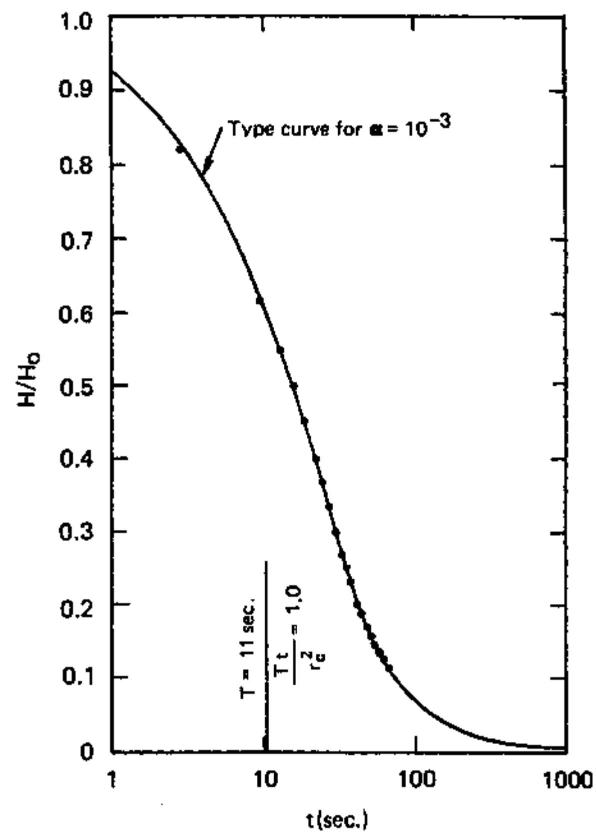


Figure X-2.5-2.— Type curves for instantaneous charge in well of finite diameter, for H/H_o versus Tt/r_c^2 for five values of *a*.
(From Lohman, 1972, pl. 2; and Cooper and others, 1967, table 1)

(a) Rise of water level in Dawsonville well after instantaneous withdrawal of weighted float.
[H, head at some time, t, after injection or removal of 'slug ';
H_0, head above or below initial head at instant of 'slug ';
(Ho=0.560m.)
m, meters; t, time]
(From Cooper, Bredehoeft, and Papadopulus, 1967, table 3)

t (sec)	Head above datum (m)	H (m	$\frac{H}{H_o}$
-1	0.896	—	——
0	.336	0.5	1.000
3	.439	.45	.816
6	.504	.39	.700
9	.551	.34	.616
12	.588	.30	.550
15	.616	.28	.500
18	.644	.25	.450
21	.672	.22	.400
24	.691	.20	.366
27	.709	.18	.334
30	.728	.16	.300
33	.747	.14	.266
36	.756	.14	.250
39	.765	.13	.234
42	.784	.11	.200
45	.788	.10	.193
48	.803	.09	.166
51	.807	.08	.159
54	.814	.08	.146
57	.821	.07	.134
60	.825	.07	.127
63	.831	.06	.116



(b) Curve matching of data from fig. X-2.5-3(a) and type curve $a = 10^{-3}$ from fig. X-2.5-2.

(c) Calculations for Cooper and others slug test method.

$r_w = r_c = 0.25$ ft

from fig. X-2.5-3(b) t = 11 sec.

and $Tt/r_c^2 = 1.0$

$$T = \frac{(1.0)\, r_c^2}{t} = \frac{(1.0)\,(0.25)^2}{11}$$

$= 0.0057\ ft^2/sec$

$= 490\ ft^2/d$

Figure X 2.5-3. Example of data and calculations for Cooper and others slug test method. (Modified from Cooper and others, 1967, p. 268)

Bouwer and Rice Method (1976)

Geometry and symbols used in this method are presented in figure X-2.5-3; a sample data set and plot of the data are included.

Procedure:

1. Plot recovery, y_t, on the logarithmic axis against time, t, on the arithmetic axis of semilog paper. (See fig. X-2.5-4).
2. Extrapolate the best-fitted straight line to intersect the t = 0 axis and determine y_o. For tests in which water-level change occurs within the casing having radius r_c, the value of y_o can be compared to the volume, V, removed from the well by:

$$y_o = \frac{V}{\pi r_c^2} \qquad \text{X-2.5-4}$$

3. From the straight-line plot, select an arbitrary time, t, and note the corresponding value of y_t Insert these values and y_o from above into the following expression:

$$\frac{1}{t} \ln \left| \frac{y_o}{y_t} \right| \qquad \text{X-2.5-5}$$

This expression is a constant for any value t and corresponding y_t from the straight-line plot.

4. Determine the coefficient C (figure X2.5-5) corresponding to the value of $\frac{LK}{r_w}$ derived from the well construction data

where: L = the length of open hole or screen open to aquifer, and

r_w = the radial distance between the undisturbed aquifer and the well center of which includes sand and (or) gravel envelopes, (See fig. X-2.5-3).

5. Solve the following equation for the natural logarithm of the ratio R_e/r_w where R_e is the effective radius of influence due to head loss y_t:

$$\ln \left| \frac{R_e}{r_w} \right| = \left| \frac{1.1}{\ln \left| \frac{H}{r_w} \right|} + \frac{C}{\frac{L}{r_w}} \right|^{-1} \qquad \text{X-2.5-6}$$

This equation is used for the case in which D = H. For partial penetration of (H less than D), use the following equation:

$$\ln \left| \frac{R_e}{r_w} \right| = \left| \frac{1.1}{\ln \left| \frac{H}{r_w} \right|} + \frac{A + B \ln \left| \frac{D - H}{r_w} \right|}{\frac{L}{r_w}} \right|^{-1} \qquad \text{X-2.5-7}$$

with an upper limit of $\ln \left| \frac{D-H}{r_w} \right| = 6$; and, where coefficients A, B, and C are obtained from figure X-2.5-5.

6. Substitute values from steps 3 and 5 along with casing radius, r_c, and screen length, L, into the following equation to determine hydraulic conductivity, K:

$$K = \frac{r_c^2 \ln \left| \frac{R_e}{r_w} \right|}{2L} \bullet \frac{1}{t} \ln \left| \frac{y_o}{y_t} \right| \qquad \text{X-2.5-8}$$

7. If needed, an approximation of transmissivity would be T = L•K.

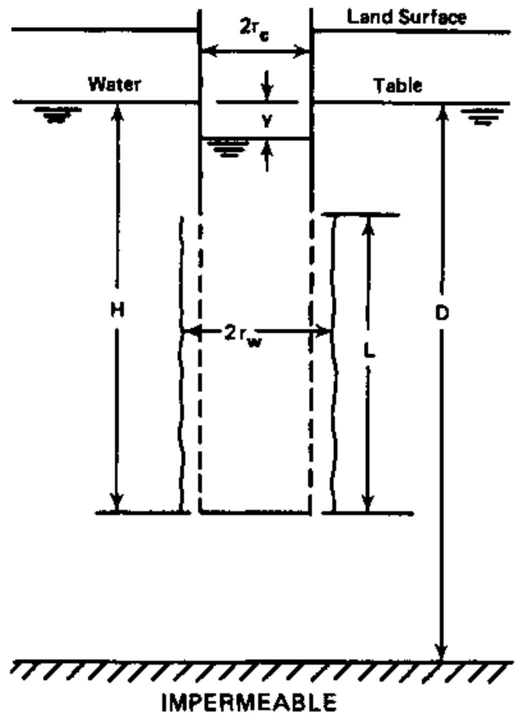


Figure X-2.5-4.— Geometry and symbols of a partially penetrating, partially perforated well in a water-table aquifer with gravel pack or developed zone around perforated section. (From Bouwer and Rice, 1976, fig. 1)

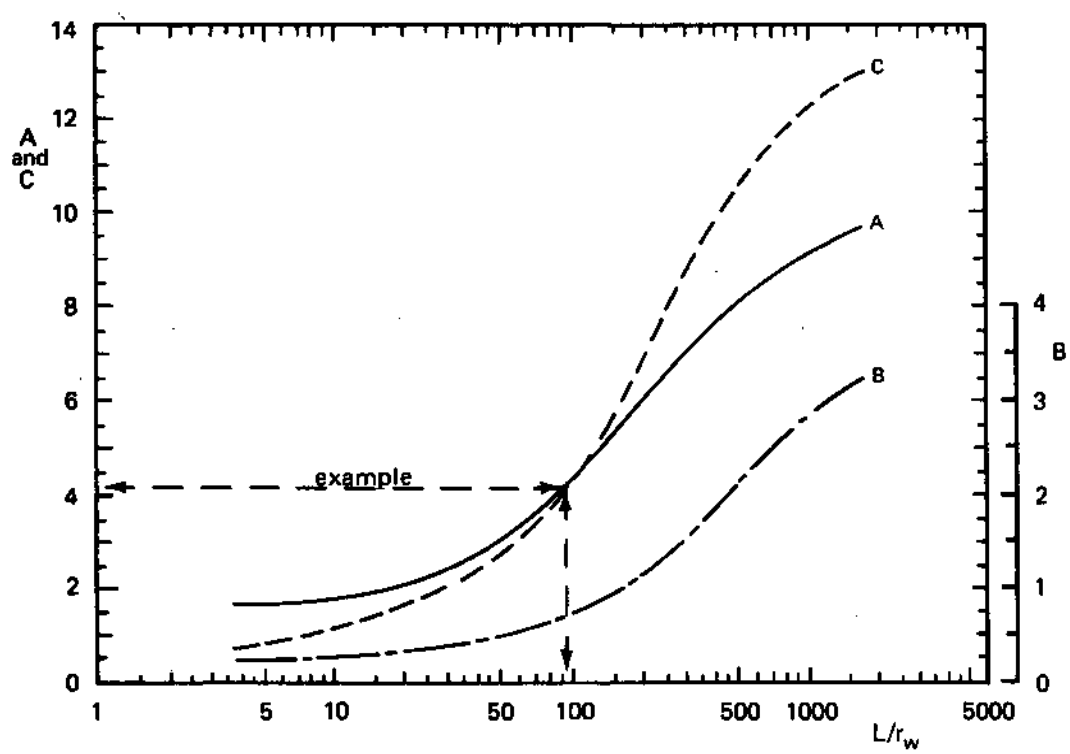
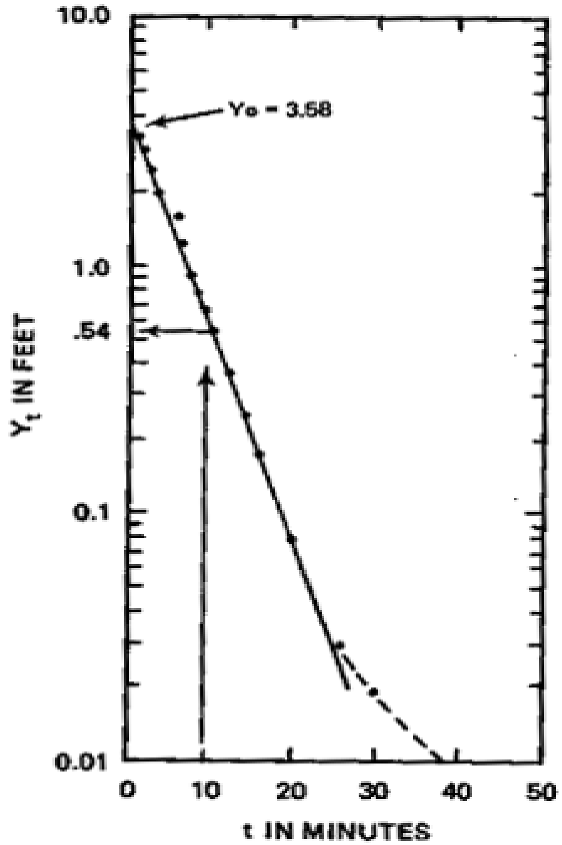


Figure X-2.5-5.— Curves relating coefficients A, B, and C, to L/r_w. (From Bouwer and Rice, 1976, fig. 3)

(a) Recover slug test data.
[r_c, radius of casing;
r_w, radius of well,
L, length of open hole;
H, distance between the water level and the bottom of the well; t, time; y_t, recovery distance of water level;
ft, feet; min., minute.]

$2\ r_c$, = 0.42 ft; L=20 ft;
r_w,= 0.21 ft; H= 94.08 ft.

t (min.)	Y_t (ft)
0.5	3.27
1.0	2.94
2.0	2.44
3.0	2.01
4.0	1.68
5.0	1.39
6.0	1.24
7.0	0.96
8.0	0.81
9.0	0.68
10.0	0.56
12.0	0.38
14.0	0.26
16.0	0.18
18.0	0.12
20.0	0.08
25.0	0.05
26.0	0.03
30.0	0.02
40.0	0.01



b) Example of data plot for Bouwer and Rice slug test method.

c) Calculations for Bouwer and Rice method.

with $r_w = r_c = 0.21$ ft, L = 20 ft, and H = D = 94.08 ft
from fig. X-2.5-6(b), intercept at t = 0, $y_o = 3.58$ ft
and with t = 10 min, $y_t = 0.54$ ft

$$\frac{1}{t} \ln\left(y_o/y_t\right) = \frac{1}{10} \ln\left(3.58/0.54\right) = 0.19\ \text{min}^{-1}$$

$L/r_w = 20/0.21 = 95.2$ and from fig. X-2.5-5, C = 4.25

$$\ln\left(R_e/r_w\right) = \left| \frac{1.1}{\ln H/r_w} + \frac{C}{L/r_w} \right|^{-1} = 4.45$$

$$K = r_c^2\ \frac{\ln(Re/r_w)}{2L}\ \frac{1}{t}\ \ln(y_o/y_t) = 0.00093\ \text{ft/min} = 1.33\ \text{ft/d}$$

$$T = K\,b - 1.33 \times 20 = 27\ \text{ft}^2/\text{d}$$

Figure X 2.5-6. Example of data calculations for Bouwer and Rice slug test method. (Modified from. Burrett and others, 1980, fig. 22)

2.6 Aquifer Test with Constant Drawdown and Variable Discharge (Plowing Well)

Transmissivity and storativity can be determined on a naturally flowing (artesian) well after the well has been shut-in for a sufficient period of time that the artesian head is virtually static. During the test, the well is allowed to flow for 2 to 4 hours, and the discharge is measured at specific time intervals. The constant drawdown, s_w, in the discharging well is the difference between the static head and the head at the discharge point. The field data collected are the time and the instantaneous discharge measurements. (See table X-2.6-1). The assumptions for the aquifer test analyses are that the aquifer is homogeneous, isotropic, and extensive laterally, and that T and S are constant at all times and all places.

Procedure:

1. Reduce the field data to $\frac{s_w}{Q}$ and $\frac{t}{r_w^2}$ as shown in table X-2.6-1.

   where: s_w = constant drawdown in the discharging well.
   r_w = radius of discharging well,
   t = time since discharging began.
   Q = instantaneous discharge.
2. Plot $\frac{t}{r_w^2}$ on the logarithmic axis against $\frac{s_w}{Q}$ an the arithmetic axis of semilog paper as shown in figure X-2.6-1.
3. Interpolate best straight line for the data and determine the change of $\frac{s_w}{Q}$ for a log cycle of $\frac{t}{r_w^2}$.
4. Compute T from the respective values of step 3 into equations, from Lohnan, 1972, p. 23, equation 71:

$$T = \frac{2.30}{4\pi\ (s_w/Q)/\ \Delta \log_{10}(t/r^2)} \qquad \text{X-2.6-1}$$

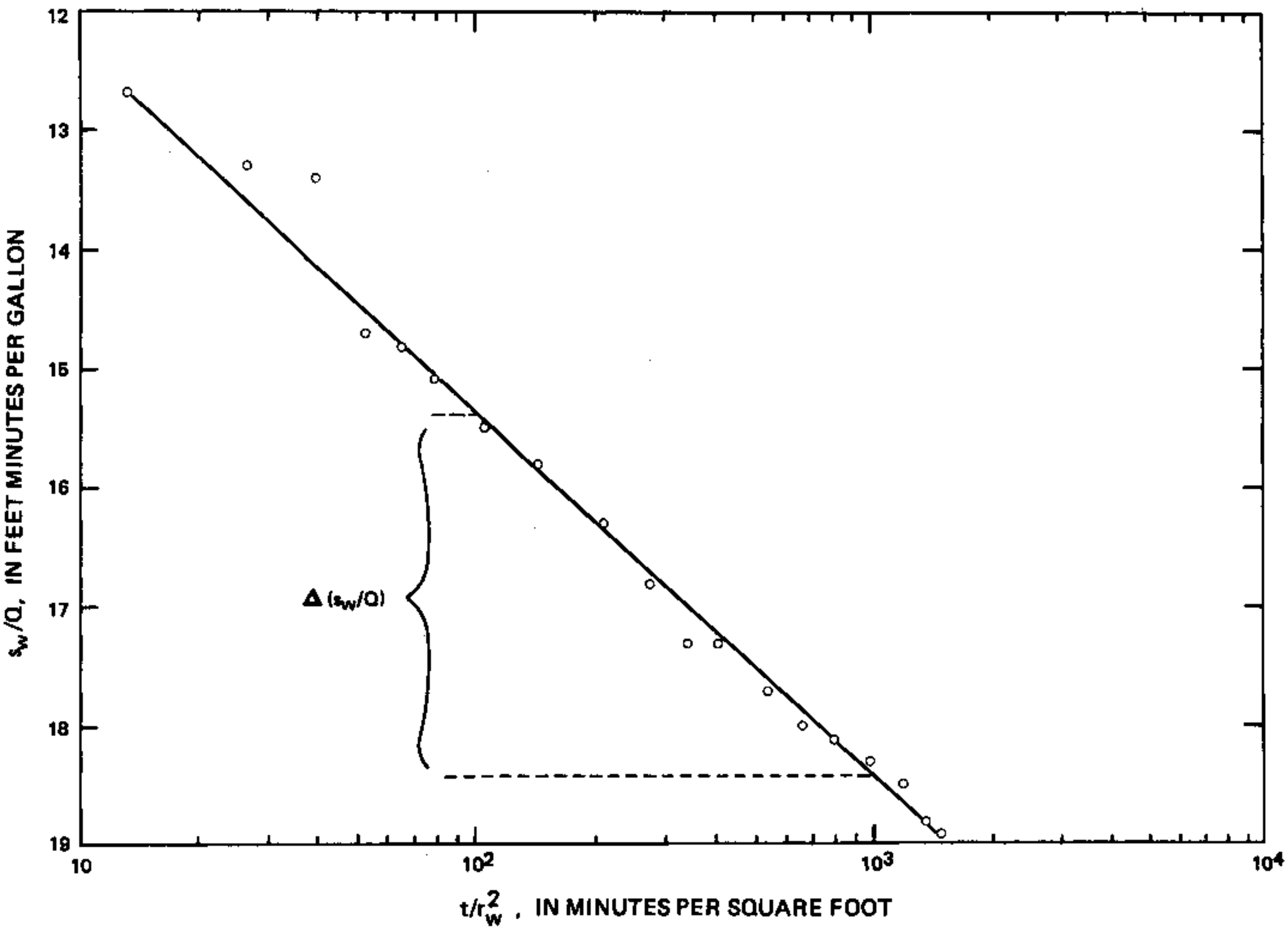
5. Compute S in the data region of the straight-line plot from Lohrnan, 1972, equation 74:

$$S = \frac{2.25\ T\ \ (t/r_w^{\ 2})}{\text{antilog}_{10} \left| \frac{s_w/Q}{\left| \frac{s_w}{Q} \right|} \right|} \qquad \text{X-2.6-2}$$

Table X 2.6-1.– Example of field data for flow test with constant drawdown and variable discharge.
(From Lohman, 1972, table 8; and Lohman, 1965, tables 6 & 7)
[gpm, gallons per minute; min, minute; ft/min, feet per minute; min/ft2, minute per square feet]
(Valve opened at 10:29 a.m.; s_w = 92.33 ft; r_w = 0.276 ft; Entrada Sandstone; well depth = 940 ft; depth of casing 936; ft; shut-in time about 12 hours.)

Time of observation	Rate of flow (gpm)	Flow interval (min)	Total flow during flow	Time since flow started	$\frac{s_w}{Q}$ (ft/min)	$\frac{t}{r_w^2}$ (min/ft^2)
10:30	7.28	1	7.28	1	12.7	13.1
10:31	6.94	1	6.94	2	13.3	26.3
10:32	6.88	1	6.88	3	13.4	39.4
10:33	6.28	1	6.28	4	14.7	52.6
10:34	6.22	1	6.22	5	14.8	65.7
10:35	6.22	1	6.22	6	15.1	78.8
10:37	5.95	2	11.90	8	15.5	105
10:40	5.85	3	17.55	11	15.8	145
10:45	5.66	5	28.30	16	16.3	210
10:50	5.50	5	27.50	21	16.8	276
10:55	5.34	5	26.70	26	17.3	342
11:00	5.34	5	26.70	31	17.3	407
11:10½	5.22	10.5	54.81	41.5	17.7	345
11:20	5.14	9.5	48.83	51	18.0	670
11:30	5.11	10	51.10	61	18.1	802
11:45	5.05	15	75.75	76	18.3	999
12: 00 (noon)	5.00	15	75.00	91	18.5	1,196
12:12	4.92	12	59.04	103	18.8	1,354
12:22	4.88	11	53.68	113	18.9	1,485
Total		114	*596.98			

* Weighted average discharge is 596.98 gallons per 114 minutes or 5.23 gpm.



(a) semilogarithmic plot of s_w/Q versus $t/\ r_w^2$ for flowing well with constant drawdown and variable discharge, data presented in table X-2.6-1.

(b) calculations

from semilog plot, $\frac{s_w}{Q} = 18.4 \quad \frac{t}{r_w^2} = 1{,}000, \quad \Delta \left|\frac{s_w}{Q}\right| - 18.40-15.38 = 3.02$

$$\Delta \ \log_{10} \left|\frac{t}{r_w^2}\right| = 3 - 2 = 1$$

$$T = \frac{(2.30X1{,}440 \text{ min/d})}{4(3.14159)\ (3.02 \text{ ft/gal/min})\ (7.48 \text{ gal/ft}^3)} = 11.7 \text{ ft}^2\text{/d}$$

$$S = \frac{(2.25X11.7 \text{ ft}^2\text{/d})\ (1{,}000 \text{ min/ft}^2)}{\text{antilog}(18.4/3.02)(l{,}440 \text{ min/d})} = \frac{(2.25)(U.7)(1{,}000)}{(1{,}230{,}000)(1{,}440)} = 0.000015$$

Figure X-2.6-1.— Example of aquifer test analysis for a flowing well. (From Lohman, 1972, p. 25 and figure 17)

2.7 Practical Considerations

2.71 Boundary conditions

At every aquifer test site, the hydrologic boundary, conditions and hydraulic properties are unknown before testing. Thus, the problem analysis in the design phase contains uncertainties (15) that must be carefully considered to avoid errors in interpreting the results of the test. Figure X-2.7-1 depicts the drawdown configuration for two different hydrologic boundary conditions— discharge boundary and recharge boundary—and compares the rates of drawdown during pumping.

The drawdown rate for the impermeable-boundary condition, figure X-2.7-la, increases with time. This response could also reflect aquifer pinchout, fault zones filled with clay gouge, or decrease in hydraulic conductivity.

The drawdown rate for the recharge-boundary condition, figure X-2.7-lb, decreases with time. This response could also be caused by increased aquifer thickness, intersection with a highly permeable water-bearing fault zone or fracture trace, increases in hydraulic conductivity, or leakage through confining layers.

The response represented by the infinite aquifer is the unchanging rate of drawdown with time. This hypothetical infinite aquifer is a water-bearing geologic unit whose hydraulic properties are constant over a long distance in all directions, which is not a naturally occurring situation. Thus, the boundary effects at each test site should be verified by further field investigations through test drilling and surface geophysical investigations. Aquifer coefficients must be determined from the test data collected before the time the boundary effects become observable.

The drawdown distribution through time during an aquifer test affected by an impermeable boundary (fig. X-2.7-2) shows the data of the first 40 minutes to match both the Theis curve and the Jacob straight line. These data are not affected by the boundary and are used to calculate the transmissivity of the valley-fill deposits aquifer.: After 40 minutes, when the drawdown effects reach the impermeable boundary, the drawdown rate increases, as illustrated by the departure of the data from the Theis curve and the straight line.

The drawdown distribution in an aquifer test affected by a recharge boundary (fig. X-2.7-3) shows the data match the Theis curve and the Jacob straight line for only the first 10 minutes. These data are not affected by the boundary and are used to calculate the transmissivity of the aquifer. After 10 minutes, however, when the discharge rate is approaches the recharge rate, the drawdown rate decreases to zero, as illustrated by the departure of the data from the Theis curve and the straight line.

2.72 Well-bore storage

In many coal fields, bedrock aquifers are tight, that is the hydraulic conductivity is low. In such aquifers, the drawdown and recovery response of a pumping well is seriously affected by well-bore-storage effects; that is, discharge from the well is derived from a depletion of storage in both the aquifer and in the well bore. For pumping times greater than $25r_c^2/T$, the drawdown response in the pumped well would be within 5-percent error of the Theis solution (12). This relationship is presented graphically in figure X-2.7-4. For example, a 4-inch diameter well that fully penetrates a confined aquifer having a transmissivity of 10 ft^2/d would have to be pumped 100 minutes before the drawdown response would be closely approximated by the Theis solution. In tight aquifers, well-bore storage is likely to be a significant source of error in the analysis of pumping well responses by the methods described previously. The slug-test methods presented, however, are not affected by well-bore storage.

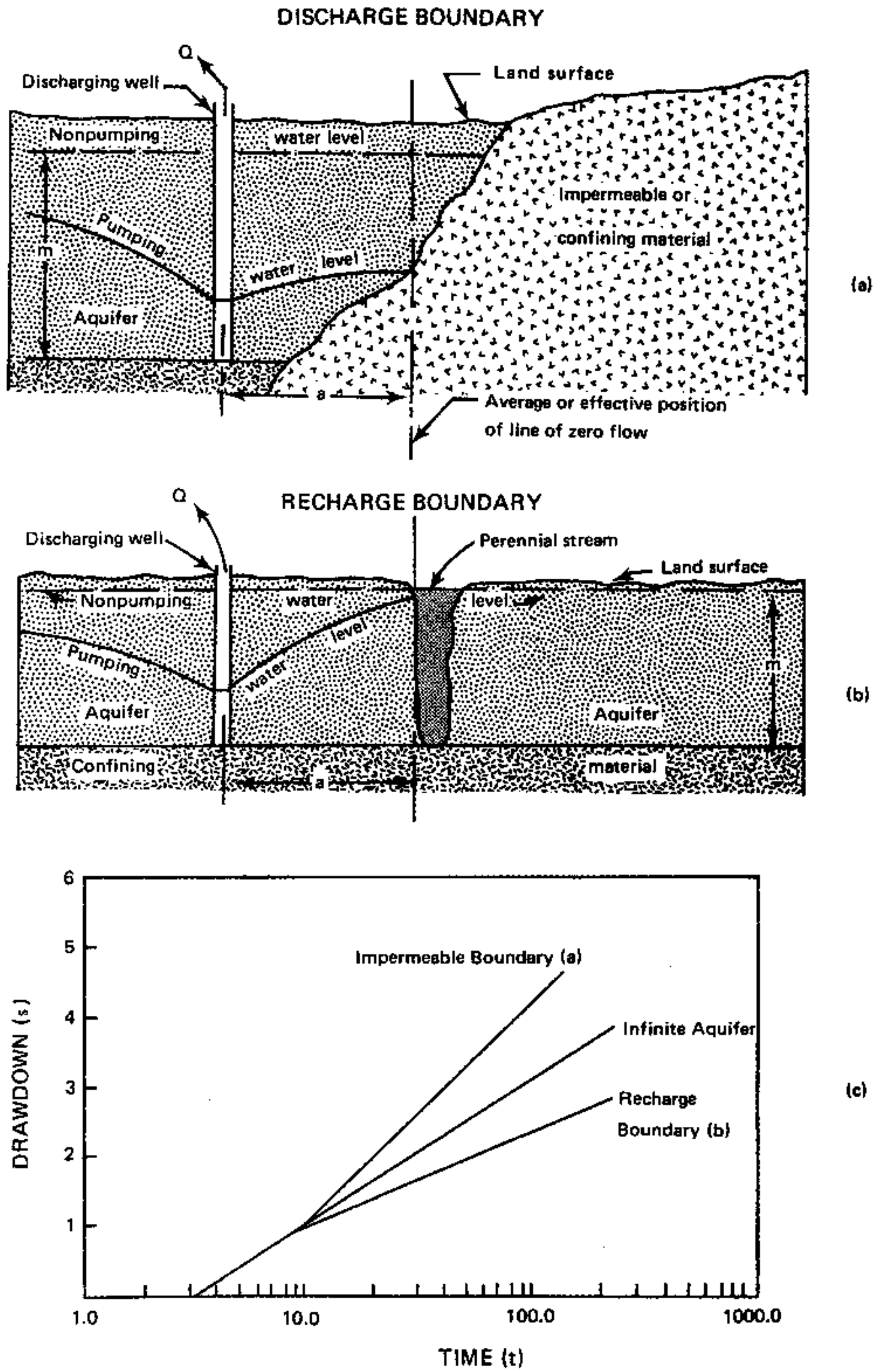


Figure X-2.7-1— Effects of recharge and discharge boundary conditions on drawdown rate.
(Parts A and B from Perris and others, 1962, figs. 35 and 37; Part C from Barrett and others, 1980, fig. 25)

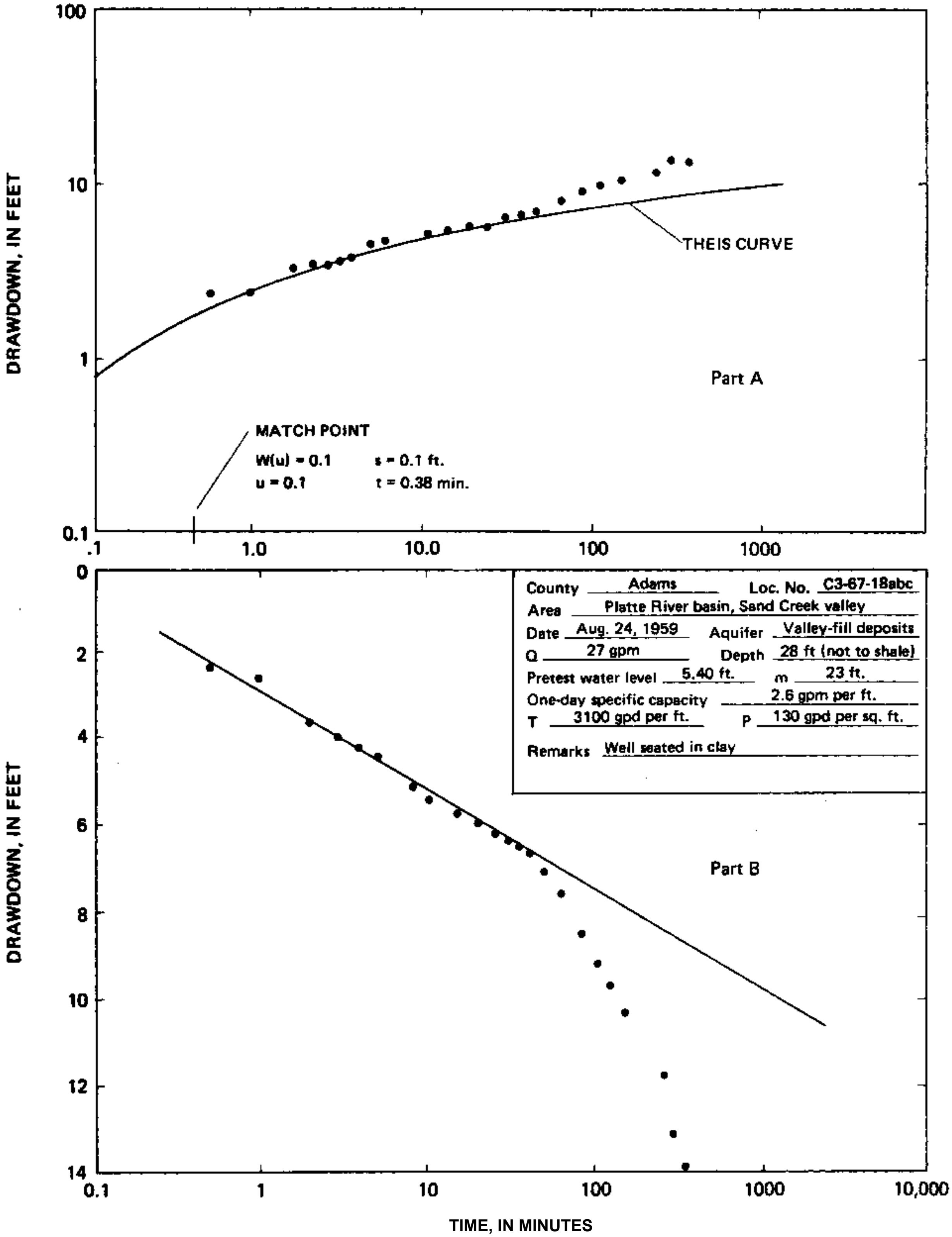


Figure X-2.7-2.— Aquifer-test results as affected by an impermeable boundary.
(Part A - drawdown increasing with increasing time in deviating from Theis curve (log s vs. log t plot)).
(Part B - drawdown increasing with increasing time in deviating from Cooper -Jacob straight-line plot (s vs. log t)).
(Data from Wilson, 1965, p. 216)

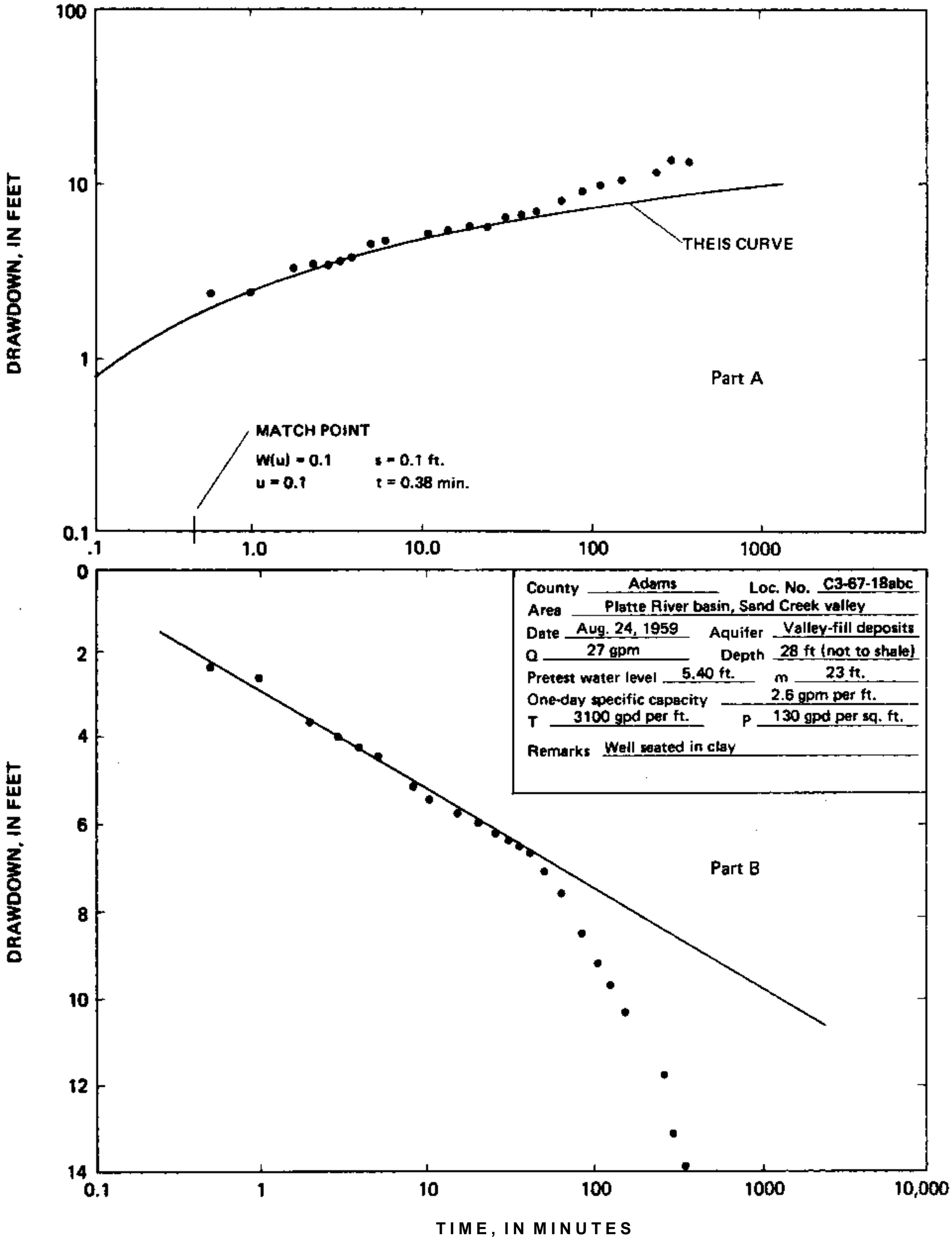


Figure X-2.7-3.— Aquifer-test results as affected by a recharge boundary.
(Part A - drawdown decreasing with increasing time in deviating from Theis curve (log s vs. lot t plot).
(Part B - drawdown decreasing with increasing time in deviating from Cooper-Jacob straight-line plot (s vs. log t).
(Data from Lang, 1960, table 1)

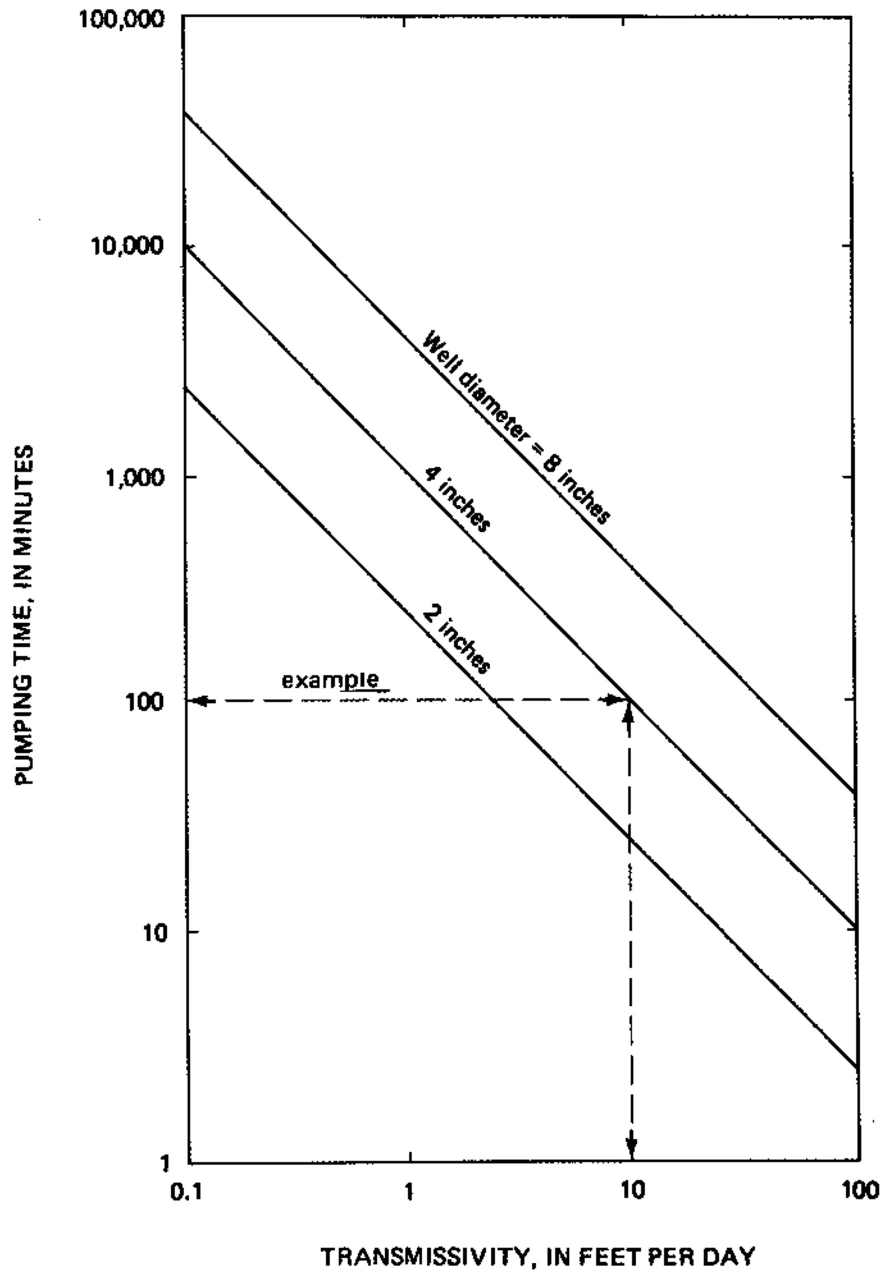


Figure X-2.7-4.— Pumping time for which the Theis solution is within 5 percent error of the theoretical drawdown response in pumping wells of indicated finite diameter.
(From Stoner, U.S. Geological Survey/ 1981, written communication)

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3.0 Qualitative Fractured Rock Hydrology

Most ground water in sedimentary terrain associated with coal deposits is related to fractures and secondary permeability. Fractures are the result of folding and faulting associated with earthquakes and mountain building processes. Fracturing commonly decreases with depth, except in the vicinity of fault structures. The greater the density of fractures, the greater the hydraulic conductivity of the bedrock unit. The hydraulic conductivity of fractured rock varies from 10 to 1000 times that of the adjacent unfractured or slightly fractured rock. The general pattern of fracture density with depth is shown in figure X-3.0-1.

Fault structures are planes of fractured rock material, and these planes vary in thickness from less than an inch to several feet. A geologic section with faults is shown in figure X-3.0-2. These planes can be identified from an interpretation of driller's logs, from geophysical investigations, and from fracture-trace interpretations (aerial photographs with a scale of 1:20,000). Fracture traces can be up to a mile long and from 5 to 65 feet wide (7). Photolinear features that are longer and wider are called lineaments. Fracture traces are commonly vertical, or near vertical, as shown in the block diagram in figure X-3.0-3. An example of photointerpreted fracture traces for a proposed permit area and adjacent area is shown in figure X-3.0-4. The high-permeability zones in the general area can discharge ground water through the adjacent area into the proposed permit area along these fracture planes. The orientation rosette in figure X-3.0-4 indicates that the two major directions of the fracture traces are north 13˚ west and east-west. Major mine adits should be oriented in directions other than these directions for maximum roof support and minimum mine inflow.

The determination of hydraulic properties of fractured rock aquifers is complex, and presentation of the analytical techniques is beyond the scope of this manual. Also, the drawdown patterns resulting from dewatering zones of fracturing are irregular because of the large variations in hydraulic conductivity (table X-l.4-1).

An aquifer test within fractured rock will initially exhibit linear flow along the high-permeability zones. With increased pumping time, however, the pumped water will be a combination of water from fractured rock, less fractured rock, and the nearby unfractured host rock. The effects of fracture-related permeability depend on the density of fractures in the aquifer volume to be tested.

Fractured-rock aquifers can be expected to follow the Theis solution more closely as fracture density increases. Data analysis by conventional methods is often possible in coal aquifers because of the closely spaced fractures in coal beds; commonly less than a few inches. However, only rough estimates of hydraulic properties of an aquifer unit can be expected from tests at wells that intersect only a few water-bearing fractures because testing in sparsely fractured aquifers commonly results in a complex drawdown response that reflects the hydraulic properties of both the fractures and the unfractured rock. For each of these components, the effective contribution to the drawdown response depends on fracture density and testing time. Other nonideal conditions commonly associated with fractured rocks are aquifer anisotropy and heterogeneity. In general, fracture density arid permeability of a particular rock type tends to decrease with depth (fig. X-3.0-1). Therefore, analyses of aquifer tests from shallow wells may indicate higher hydraulic-conductivity values than the average values for the entire thicknesses of overburden-coal aquifer systems.

Additional information on analytical aquifer testing techniques for fractured rock can be obtained from references (l), (2), (3), and (6).

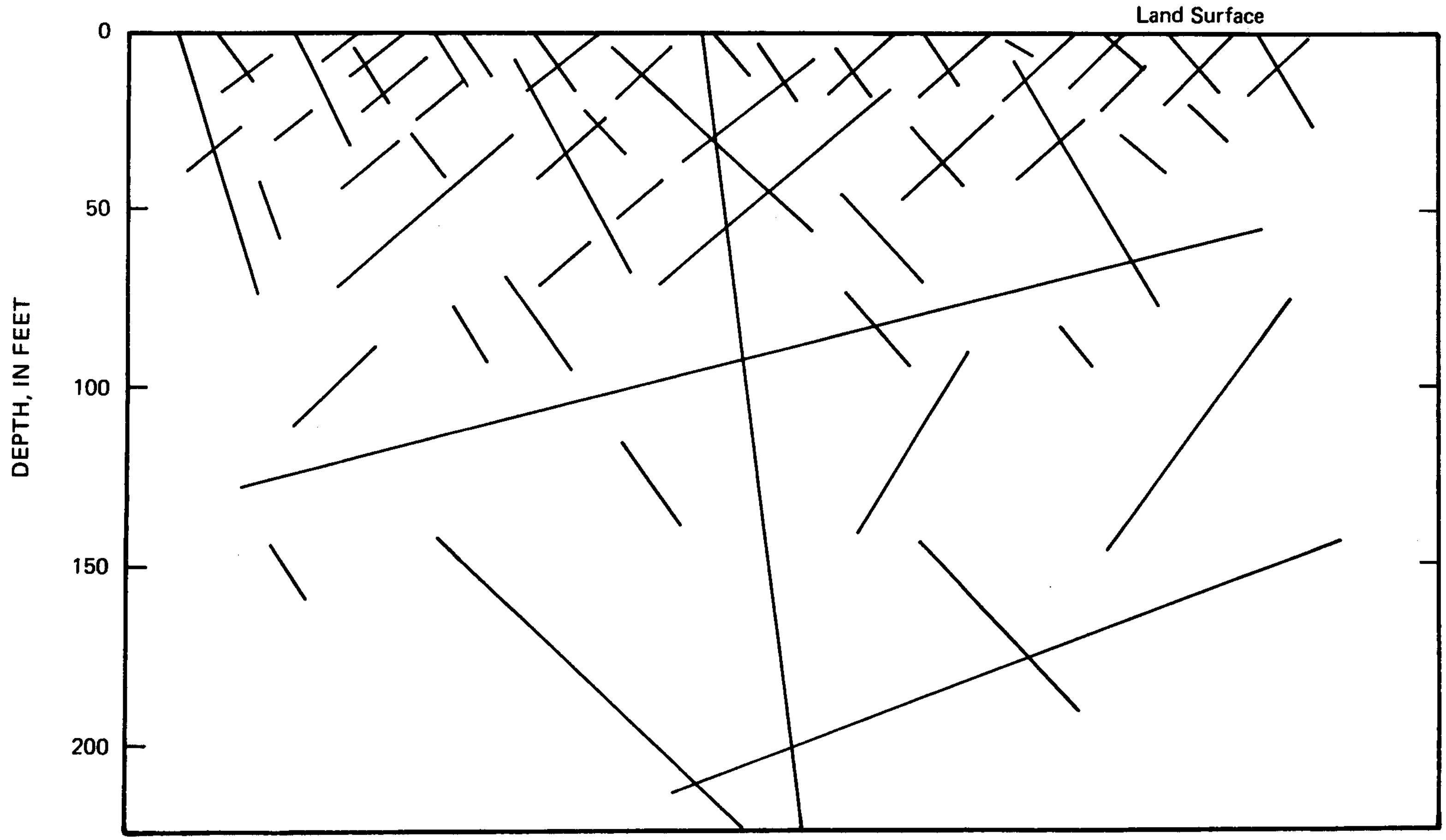

Figure X-3.0-1.— Idealized distribution of rock fractures with depth.

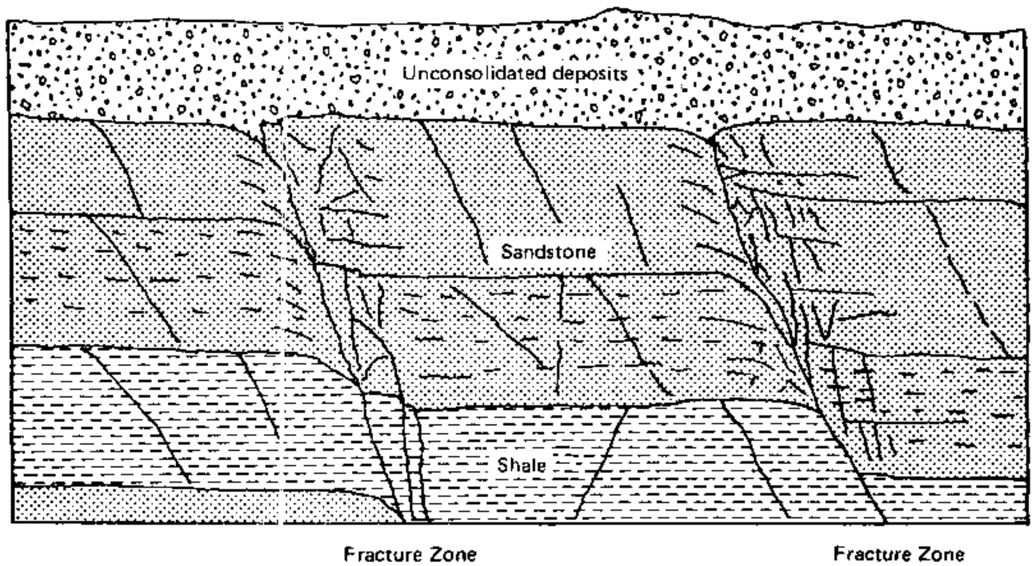

Figure X-3.0-2. — Vertical section showing buried fracture zones and joints in sedimentary bedrock,
(Modified from Meinzer, 1923a, fig. 68)

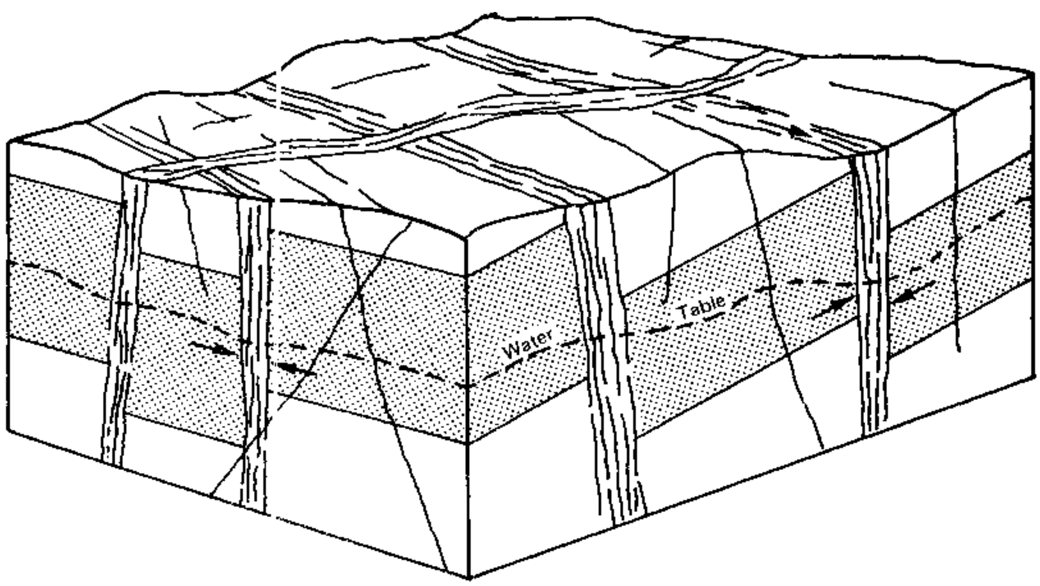

Figure X-3.0-3.— Idealized diagram of fracture traces and faults cross cutting sedimentary terrain.
(From Parizek and others, 1971)

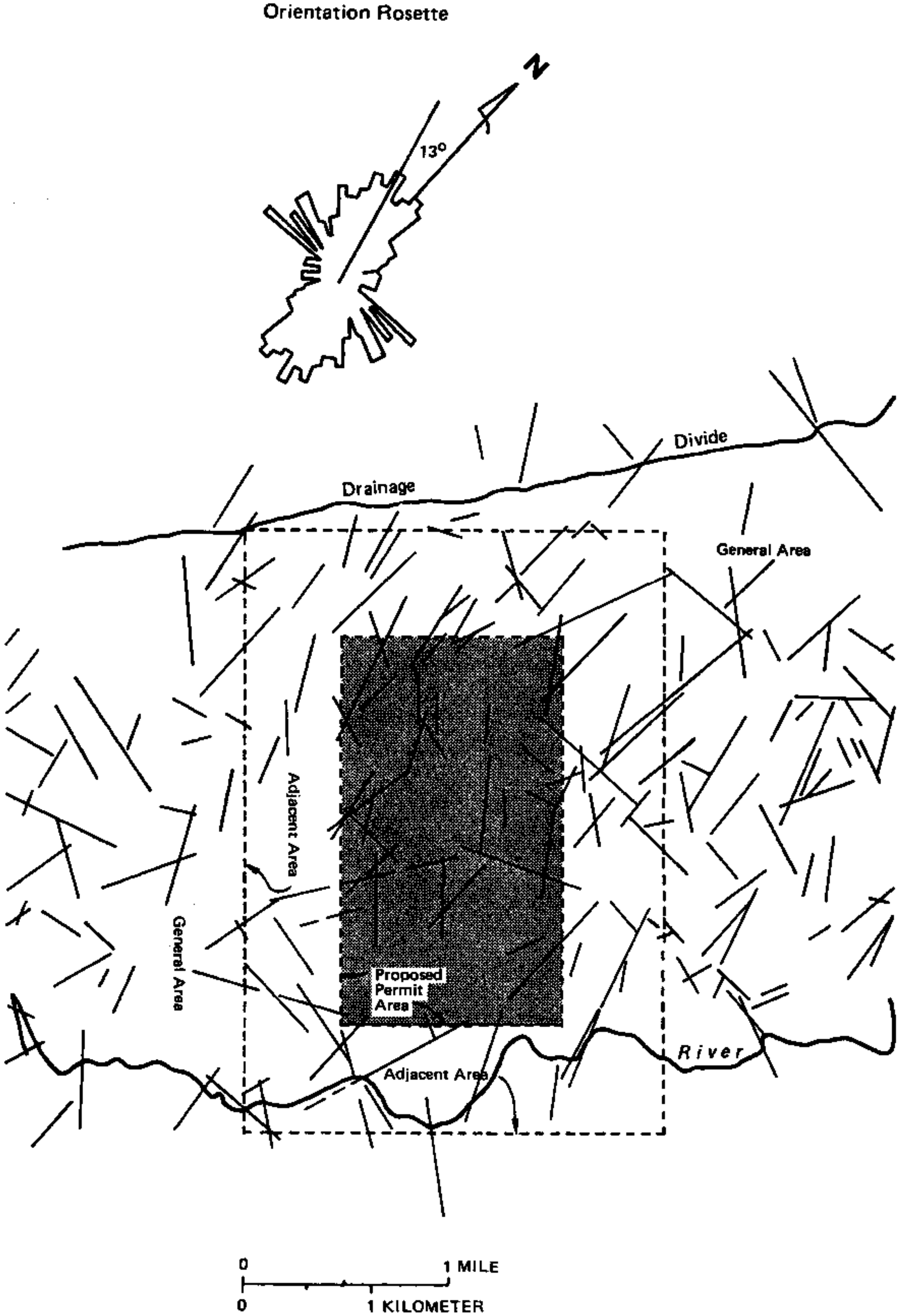

Figure X-3.0-4.— Linear fracture traces in a proposed permit area and vicinity. (Modified from Duigon, 1985, fig. 10)

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