TRANSMISSIVITY OF PERCHED AQUIFERS AT THE IDAHO NATIONAL ENGINEERING LABORATORY, IDAHO

By D.J. Ackerman

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 91-4114

Prepared in cooperation with the

U.S. DEPARTMENT OF ENERGY

Idaho Falls, Idaho

1991

U.S. DEPARTMENT OF THE INTERIOR

MANUEL LUJAN, JR., Secretary

U.S. GEOLOGICAL SURVEY

Dallas L. Peck, Director

For additional information write to:

U.S. Geological Survey
INEL, MS 4148
P.O. Box 2230
Idaho Falls, ID 83403-2230

Copies of this report can be purchased from:

Books and Open-File Reports Section
Western Distribution Branch
Box 25425, Federal Center
Denver, CO 80225

CONTENTS

ILLUSTRATIONS

TABLES

CONVERSION FACTORS AND ABBREVIATIONS

Multiply	By	To obtain metric unit
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.09290	meter squared per day
gallon per minute (gal/min)	0.06309	liter per second
gallon per minute per foot [(gal/min)/ft]	0.2070	liter per second per meter
inch (in.)	25.4	millimeter
square mile (mi^2)	2.590	square kilometer

TRANSMISSIVITY OF PERCHED AQUIFERS AT THE IDAHO NATIONAL ENGINEERING LABORATORY, IDAHO

By D.J. Ackerman

ABSTRACT

Aquifer-test data of 43 single-well tests at 22 wells in perched aquifers at the Idaho National Engineering Laboratory were analyzed to estimate values of transmissivity. Estimates of transmissivity for individual wells ranged from 1.0 to 15,000 feet squared per day, more than 4 orders of magnitude. These data were determined in a consistent manner and are useful for describing the distribution of transmissivity at the Idaho National Engineering Laboratory.

The results of type-curve analysis of eight tests at six wells were used to verify a regression relation between specific capacity and transmissivity. This relation, in turn, was used to analyze all specific-capacity data. Values of relative uncertainty for estimated values of transmissivity generally ranged from 0.1 order of magnitude for type-curve analysis to 0.5 order of magnitude for regression analysis and measured drawdown of less than 0.1 foot. The values of transmissivity given in this report represent the transmissivity near the test wells and within the test interval. Due to the high degree of heterogeneity of the basalt and the unknown thickness of the aquifers, it is more likely the transmissivity of the whole basalt sequence is different from those values given in this report.

INTRODUCTION

The INEL (Idaho National Engineering Laboratory), which includes about 890 mi^2 of the eastern Snake River Plain in southeastern Idaho (fig. 1), is operated by the U.S. Department of Energy primarily to build, test, and operate nuclear reactors, and to process spent fuel from government-owned

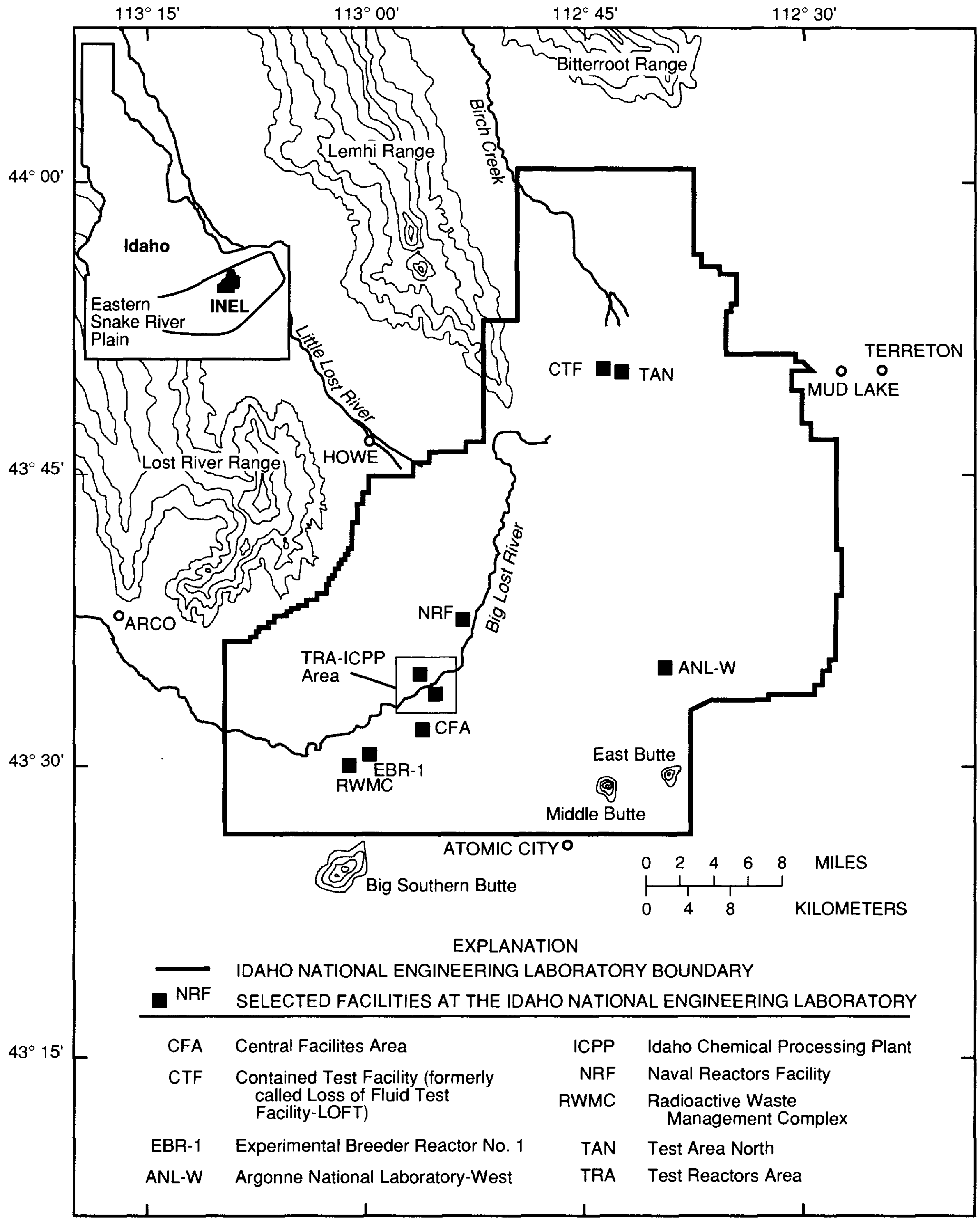


Figure 1.--Location of the Idaho National Engineering Laboratory.

reactors. The INEL also supports other government-sponsored projects such as energy, defense, medical, and environmental research. The entire water supply for the INEL is obtained from the Snake River Plain aquifer. Ground water is used at the INEL as the source of noncontact cooling water, process water, and drinking water.

Aqueous chemical and radioactive wastes have been discharged to deep wells and shallow ponds at the INEL since 1952. Wastewater from the ponds percolates downward forming perched aquifers above the Snake River Plain aquifer. Some of these wastes have also moved into the Snake River Plain aquifer. Wastes that reach the aquifer move downgradient toward the southern boundary of the INEL. The effects of waste disposal have been investigated by the U.S. Geological Survey, in cooperation with the U.S. Department of Energy, since disposal began.

Purpose and Scope

The purpose of this report is to present a consistently determined set of transmissivity values estimated from aquifer-test and specific-capacity data for perched aquifers at the INEL. These transmissivity values are needed for studying ground-water flow at the INEL.

Because of the paucity of multiple-well aquifer tests near the INEL, the scope of interpretation is limited to single-well tests or to specific-capacity data. Much of the data are simply specific capacities, which were used to estimate transmissivities. These estimates are useful for evaluating regional differences in transmissivity and preparing transmissivity maps for use in models of ground-water flow systems (Heath, 1983, p. 60). Single-well tests allow the estimation of transmissivity only; the interpretation of all possible hydrologic properties from multiple-well tests is not part of the scope of this report.

Data and interpretations are given for 43 tests at 22 wells at the INEL. An estimate of transmissivity is made for perched aquifers at each well. This report relies on and is an extension of a previous report

(Ackerman, 1991), which gives transmissivity estimates for the Snake River Plain aquifer at and near the INEL.

Geohydrologic Setting

Aquifers at the INEL consist of layered sequences of basaltic-lava flows and cinder beds intercalated with sedimentary deposits mainly made up of fluvial and lacustrine deposits. Individual lava flows typically are 20 to 25 ft thick and 50 to 100 mi^2 in areal extent. Rubble, clinker zones, fractures, and vesicular zones (collectively referred to as interflow zones) are prevalent near the surfaces of flows. Subsequent lava flows or sedimentary deposits may partly fill fractures and vesicles. The centers of individual flows, especially thick flows, are typically less vesicular and more massive, although they may be characterized by vertical fractures.

The geology and hydrology of the Snake River Plain at the INEL describe a water-table aquifer of large areal extent (Garabedian, 1989) with overlying perched aquifers near waste disposal ponds (Robertson, 1977; and Pittman and others, 1988). Perched aquifers form when downward flow from waste ponds is impeded by silt and clay in sedimentary deposits or by dense basalt flows. Well yields are large because of the highly transmissive nature of interflow zones. The aquifer framework results in a complex, heterogeneous, and anisotropic medium at the scale aquifer tests stress the system. If a mathematical treatment is to be workable, certain idealizations regarding homogeneity, well construction, and aquifer extent are imperative.

The water table for the Snake River Plain aquifer at the INEL ranges from about 200 ft below land surface in the north-central part to about 900 ft in the southeastern part. Perched water tables are usually between 50 and 140 ft below land surface. Ground-water levels are relatively stable at the INEL, although they respond to climatic trends and, locally, to recharge from intermittent streams. The bases of the perched aquifers are uncertain but may be at two levels near TRA (Test Reactors Area), at the upper surface of the first layer of basalt (about 50 ft below land surface) and within the basalt (about 140 ft below land surface) (Robertson, 1977, p. 9). Hydrologic conditions for perched aquifers at the ICPP (Idaho Chemical Processing Plant) have not been documented.

During 1982-85, 276 million gallons per year of wastewater were disposed to infiltration ponds at the TRA (Pittman and others, 1988, p. 20). During 1986-88, approximately 259 million gallons per year of waste-water were disposed to infiltration ponds at the TRA (Orr and Cecil, 1991, p. 15). The average discharge to ponds at the ICPP during 1986-88 was about 580 million gallons per year (Orr and Cecil, 1991, p. 18).

For additional discussions of the geology of the Snake River Plain and the hydrology of aquifers at and near the INEL, the reader is referred to Garabedian (1989) and Robertson and others (1974).

DESCRIPTION OF AQUIFER-TEST DATA AND WELLS

Transmissivity is a measure of the ability of an aquifer to transmit water to hydraulically downgradient areas and to pumping wells. Transmissivity is defined as the rate at which water of prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient (Lohman, 1972, p. 6).

The transmissivity of an aquifer can be estimated from aquifer-test and specific-capacity data. An aquifer test consists of pumping a well at a constant rate for a specified time and measuring the resultant water-level declines in the pumped well and in nearby wells that are not pumped during the testing period. After the pumping is stopped, measurements are made to define the rate of recovery of water levels. The specific capacity of a well is the ratio of the pumping rate to the resultant water-level decline measured in the pumping well. Specific capacity can also be used to estimate transmissivity although there is a lesser degree of certainty when compared to estimates made using aquifer-test data.

Time-drawdown data have been collected at 116 wells at and near the INEL for a variety of purposes, but rarely for the express purpose and design of calculating aquifer hydrologic properties. Many tests conducted before 1980 primarily were part of well-acceptance tests for production or injection wells. After 1986, most tests were made to obtain transmissivity estimates but often were constrained by pumping system capacity and design.

Records for 43 aquifer tests at 22 wells completed in perched aquifers at the INEL (table 1 and fig. 2) were reviewed and analyzed as part of this investigation. These included only one multiple-well test that was inadequate for multiple-well analysis. Type-curve analysis was possible for eight tests at six wells. The remaining 35 tests at 21 wells provided little more than specific-capacity data.

All methods of analyzing aquifer-test data are developed using simplifying assumptions. When these assumptions are not met, the ability to interpret the data is limited. Improper well construction and test procedures can also diminish the usefulness of the data. Problems in the execution of these aquifer tests that limit the ability to interpret the data are:

1. nonsteady discharge rates;
2. inadequate discharge rates;
3. inefficient production wells (well loss);
4. well-bore storage effects;
5. filling of well bores with sediment;
6. decrease in saturated thickness with pumpage;
7. filling and draining of pump columns (small discharge rates coupled with great depths to water);
8. inability to quantify partial penetration effects;
9. lack of record of prior trends in water levels;
10. lack of records of either barometric fluctuations or efficiency;
11. interference from other pumping or injecting wells;
12. observation wells too far from pumping well to clearly quantify the effects of partial penetration, release of water from elastic storage, delayed water-table response, and anisotropy; and

Table 1.--Data on specific-capacity tests of wells in perched aquifers at and near the Idaho National Engineering Laboratory

[locations of wells shown on figure 2; Date, year-month-day of beginning of test; gal/min, gallon per minute; (gal/min)/ft, gallon per minute per foot; Remarks: C, type curve analysis; R, regression analysis; a, multiple well test; d, estimated decrease in saturated thickness greater than 10 percent; h, large barometric change; s, drawdown measurable but less than 0.1 foot; x, pumped dry or no recovery]

Well identifier	Test number	Date	Length of test, minutes	Dis-charge, gal/min	Draw-down, feet	Specific Capacity, (gal/min)/ft	Remarks
CWP-8	1	86-05-21	63	1.1	0.06	18.0	R,s
CWP-9	1	86-05-21	37	1.2	26.89	0.045	R,d
PW-1	1	87-08-21	40	0.90	0.29	3.1	R
PW-2	1	87-08-21	50	0.55	1.39	0.40	C
PW-3	1	87-08-24	10	0.40	3.06	0.13	R,x
PW-4	1	87-08-24	60	0.80	0.03	27.0	R,s
PW-6	1	87-08-24	20	0.50	5.31	0.094	R,x
PW-8	1	87-08-25	30	0.80	0.13	6.2	R
	2	88-06-16	120	9.4	2.22	4.2	R
53	1	86-05-27	130	1.1	0.01	110.0	R,s
	2	87-09-11	35	0.70	<0.01	>70.0	R,s
	3	88-06-24	60	0.80	<0.01	>80.0	R,s
	4	88-06-30	120	1.5	0.01	150.0	R,s
54	1	86-05-14	87	0.91	0.03	30.0	R,h,s
	2	87-08-25	120	0.90	0.08	11.0	R,s
	3	88-06-23	120	9.5	0.78	12.0	R
55	1	86-05-15	150	1.1	0.52	2.1	R
	2	87-09-11	60	0.70	0.33	2.1	R
56	1	86-05-30	120	1.2	0.16	7.5	R
	2	88-07-01	42	1.1	4.37	0.25	R,d,x
60	1	86-04-04	171	1.1	0.06	18.0	R,s
	2	87-08-25	30	0.85	0.03	28.0	R,s
	3	88-06-15	120	9.1	0.55	17.0	C,a
61	1	86-04-01	195	0.83	4.48	0.19	C
	2	87-09-04	90	0.40	1.34	0.30	R
62	1	86-03-31	5	0.44	0.03	15.0	R,s
	2	86-08-24	74	10.	0.28	36.0	R
	3	88-06-22	120	8.5	0.32	27.0	R
63	1	86-05-20	103	1.11	0.02	56.0	R,s
	2	87-08-27	30	0.70	0.01	70.0	R,s
	3	88-06-20	120	9.3	0.28	33.0	C
68	1	86-06-03	155	1.0	21.98	0.045	R,d
	2	88-07-06	120	1.1	24.27	0.045	R,d
69	1	86-05-23	160	0.95	7.96	0.12	C,d
	2	87-08-31	75	0.65	3.82	0.17	C
70	1	86-05-16	148	0.95	0.71	1.3	C
	2	87-09-11	50	0.70	0.53	1.3	C
71	1	88-06-27	5	9.0	0.39	23.0	R
73	1	86-05-19	127	0.91	0.09	10.0	R,s
	2	87-09-01	55	0.60	13.17	0.046	R,d
	3	88-06-21	5	9.3	17.34	0.54	R,d
	4	88-06-30	110	0.88	30.56	0.029	R,d
78	1	86-05-28	125	0.86	0.15	5.7	R

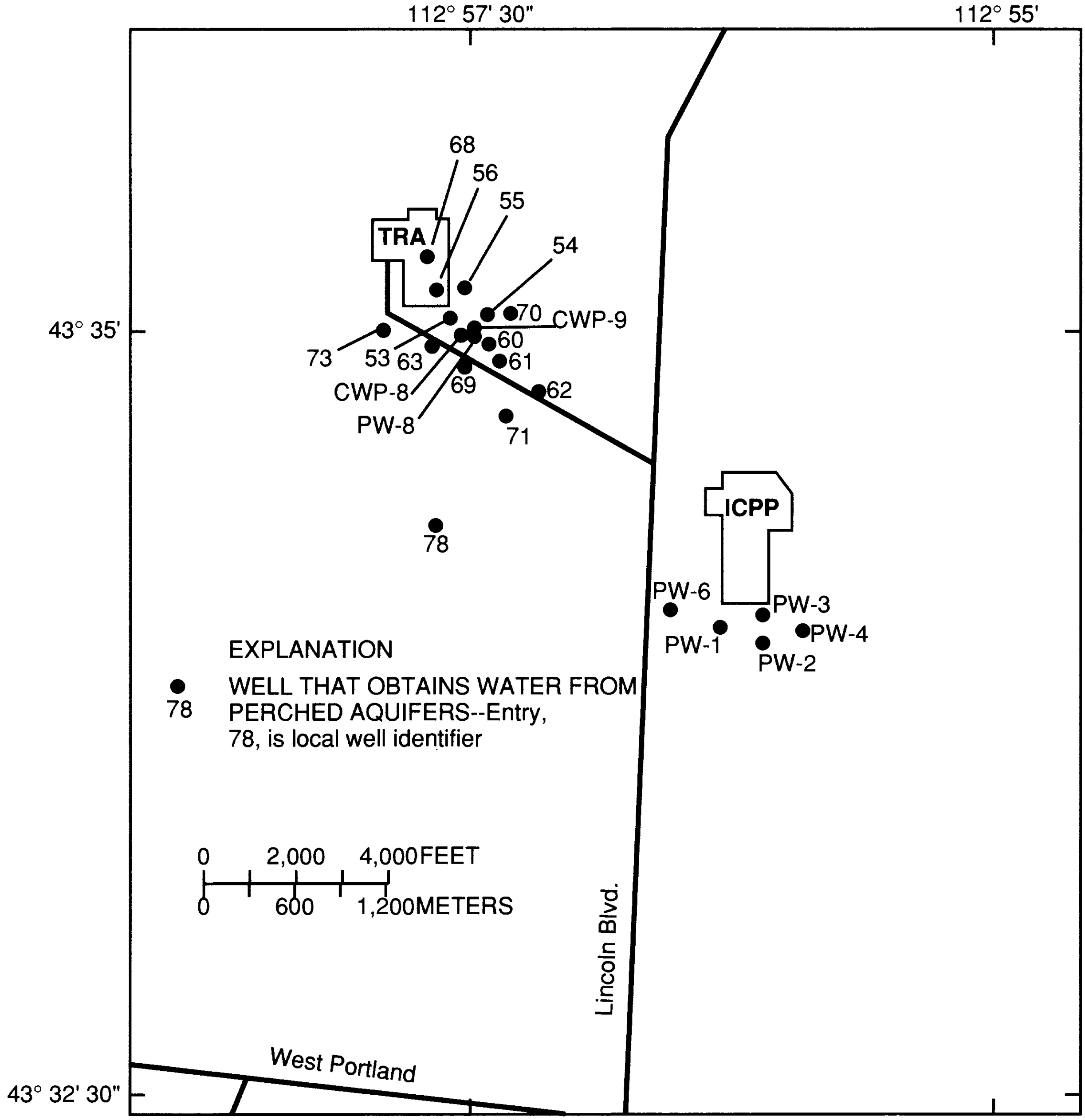


Figure 2--Locations of wells completed in perched aquifers with aquifer tests in the TRA-ICPP area.

13. insufficient early- or late-time observations of drawdown to fully utilize an applicable method of interpretation.

The general method of well construction used for perched aquifer observation wells at the INEL consists of installing 4- to 8-in. diameter casing above the water level and leaving a 6- to 10-in. open hole below. Perforated or torch-cut steel casing is sometimes hung in the open hole below the water level and is gravel packed. Pre-made screens are rarely used at the INEL.

Water levels in wells generally were measured with electric tapes or wetted steel tapes. The resolution of water levels and accuracy of drawdown measurements was usually about 0.01 ft but was sometimes 0.1 ft. Most measurements of discharge were made volumetrically or by observations of total flow on an inline meter. Some checks on the flow meter were made volumetrically.

Names and locations of wells are given in tables 1 and 2 and figure 2. The well identifier used for reference between tables and figures is a form of the local name in use at the INEL. A second local well identifier, the local well number, also is given on table 2. This identifier is derived from the township, range, and section location of the well and is useful for plotting map locations and for cross reference with other data bases. For an explanation of the well numbering system and an example of another useful data base for the project area see Bagby and others (1984, p. 12).

ANALYSIS OF AQUIFER-TEST DATA

The interpretation of the response of water levels in wells to pumping withdrawals is the most common method of determining hydrologic properties of aquifers. In general, the drawdown, s, in a pumped well or an observation well is measured at regular intervals during constant-rate pumping and is compared with predicted drawdowns from well-hydraulics equations. A large number of combinations of aquifer conditions, geometry, and aquifer-test designs can be accommodated by various analytical and graphical

Table 2.--Estimates of transmissivity from tests of wells in perched aquifers at and near the Idaho National Engineering Laboratory

[Locations of wells shown on figure 2; ft^2/d, feet squared per day; Relative uncertainty in orders of magnitude; Well depth, below land surface; Water level, at beginning of test below land surface; Penetration, below water table]

Well identifier	Transmissivity, ft^2/d	Relative uncertainty	Test number	Local well number	Well depth, feet	Water level, feet	Pentration, feet	Perforated or open interval, feet below land surface
CWP-8	1,300	0.5	1	03N 29E 14DDB3	60	49	11	20-59
CWP-9	1.0	0.4	1	03N 29E 14DDB4	64	39	25	20-43
PW-1	160	0.4	1	03N 29E 25AAA1	120	72	48	100-120
PW-2	9.4	0.1	1	03N 30E 30BBD1	133	120	13	121-131
PW-3	3.6	0.4	1	03N 30E 30BBA1	127	118	9	103-123
PW-4	2,000	0.5	1	03N 30E 30BAC1	123	83	40	110-123
PW-6	2.5	0.4	1	03N 29E 25AAB1	125	119	6	105-125
PW-8	220	0.4	2	03N 29E 14DDA4	178	70	108	140-165
53	15,000	0.5	4	03N 29E 14DAC1	88	69	19	44-88
54	790	0.4	3	03N 29E 14DAC2	91	65	26	60-91
55	99	0.4	1	03N 29E 14DAB1	80	58	22	45-80
56	440	0.4	1	03N 29E 14DAB2	80	56	24	62-80
60	880	0.1	3	03N 29E 14DDA3	120	69	51	59-120
61	10	0.1	1	03N 29E 13CCC1	123	79	44	89-123
62	2,800	0.4	2	03N 29E 24BBA1	165	133	32	145-165
63	3,500	0.1	3	03N 29E 14DCA1	110	78	32	62-110
68	1.0	0.4	1	03N 29E 14ACD2	128	58	70	50-128
69	5.6	0.1	1	03N 29E 14DDC1	115	64	51	95-115
70	51	0.1	1	03N 29E 13CBC1	100	66	34	50-100
71	1,700	0.4	1	03N 29E 24BBC1	184	155	29	60-184
73	1.0	0.4	2	03N 29E 14DCB1	127	91	36	62-127
78	320	0.4	1	03N 29E 23DAC1	204	91	113	66-204

methods. However, a survey of the various analytical treatments reveals the similarity of time-drawdown response for differing aquifer conditions. Interpretations of hydrologic properties are therefore not unique because of differences in analytical treatments, complications introduced by field conditions, and uncertainty in hydrologic conditions.

The methods of analyses for aquifer-test data at the INEL were chosen on the basis of the conceptualization of the perched aquifers. The analytical treatments were applied as uniformly as possible while remaining consistent with the conceptual model and assumptions of the analysis method to increase the significance of comparisons of results.

The analytical treatment chosen as most representative of the perched aquifers was that of Neuman (1972, 1974, 1975) for an anisotropic, unconfined aquifer considering delayed (water-table) gravity response, vertical components of flow, specific storage, specific yield, and partial penetration. Neuman's solution reproduces the typical response of a water-table aquifer as indicated on time-drawdown curves (Freeze and Cherry, 1979, p. 326). The method of Neuman was advanced from the work of Boulton (1954, 1963), and uses a graphical method of solution involving type-curve matching. The type curves given by Neuman (1975) include as a subset the type curves most commonly used in aquifer-test analysis, those of Theis (in Lohman, 1972, p. 17).

The Theis method has an advantage of greater simplicity; however, it has the disadvantage of simplifying some of the physics of the perched aquifers. For unconfined aquifers, the more rigorous treatments of boundary conditions and the more complete consideration of hydrologic properties used by Neuman are preferred from a theoretical standpoint. Aquifer-test data support type-curve matches with both the Theis and Neuman methods of analysis.

Other methods could have been chosen for analysis of these data. Some methods not chosen and the reasons for not choosing them were:

The Thiem equation (in Lohman, 1972, p. 11) was not used as it requires steady-state and isotropic conditions.

The Jacob straight-line (or semilogarithmic) method (in Lohman, 1972, p. 19) as modified for use with the conditions to which Neuman's method apply (Neuman, 1975, p. 331) could not be used for nearly all of the data at the INEL because the early-time data were poor or not discernible from intermediate-time data. More will be said concerning early-, intermediate-, and late-time response in the section on type-curve analysis by the Neuman method.

The specific-capacity methods for estimating transmissivity (Theis and others, 1963, p. 331-341) could have been used for those observations having time-drawdown data that followed the Theis curve. The method was not used because estimations by the linear-regression method gave results of acceptable accuracy for all data.

The type-curve analysis methods of Boulton (1963) and Stallman (in Lohman, 1972, p. 35) were probably adequate for the purpose of this study but are not as rigorous, flexible, or complete as those of Neuman.

Numerical-model analysis, such as used by Lindner and Reilly (1983) or Prince and Schneider (1989), is perhaps the most rigorous method and can account for field conditions most accurately. However, data and available models were not sufficient to use this method efficiently.

Three methods were used to estimate transmissivity from drawdown data for single-well tests at the INEL. Two methods of type-curve analysis were used to analyze the most complete tests. The type curves used were those of Neuman (1975) and Theis (in Lohman, 1972, p. 15). A simple analytical method was used to analyze all other tests. Transmissivity was estimated from specific capacity by use of a simple linear-regression equation. The estimates of transmissivity calculated from type-curve analyses were regressed with corresponding specific capacities. Transmissivity was estimated for all remaining tests by application of the linear-regression equation.

The transmissivity derived by any of these methods from a single-well test has a wider confidence interval than a transmissivity derived by more rigorous analysis of a properly-designed and well-executed multiple-well aquifer test. Comparison of estimates of transmissivity determined from data at observation wells with those determined from data at the pumping well indicated a possible bias of 0.5 to 1.5 orders of magnitude with an average bias of about 1 order of magnitude. The single-well tests gave lower values.

Type-Curve Analysis

Field data were matched with theoretical type curves to determine aquifer transmissivity. The type curves used were those of Neuman (1975, as given in Freeze and Cherry, 1979, fig. 8.12) for fully penetrating wells in an anisotropic unconfined aquifer. A distinctive S-shaped curve with three distinct segments results (fig. 3) when drawdown and time for an aquifer test in an unconfined aquifer are plotted on a logarithmic coordinate scale graph. During the first or early-time segment, which only covers a very short period, the aquifer shows a typical confined response. Water is released from storage as a result of aquifer compaction and the expansion of water. The time-drawdown response may follow a Theis nonequilibrium type curve for a typical confined aquifer storage coefficient. During the second or intermediate-time segment, a distinct departure from the Theis curve results in response to the effects of water delivered to the well by dewatering at the water table. This decrease in the rate of drawdown is called either delayed gravity response, delayed yield, or delayed drainage. This intermediate response may produce a definite flat or nearly horizontal part of the curve.

Given enough time, a third segment may be recognized after the effects of delayed gravity response have dissipated. During the third or late-time segment, time-drawdown response will gradually start to follow the Theis

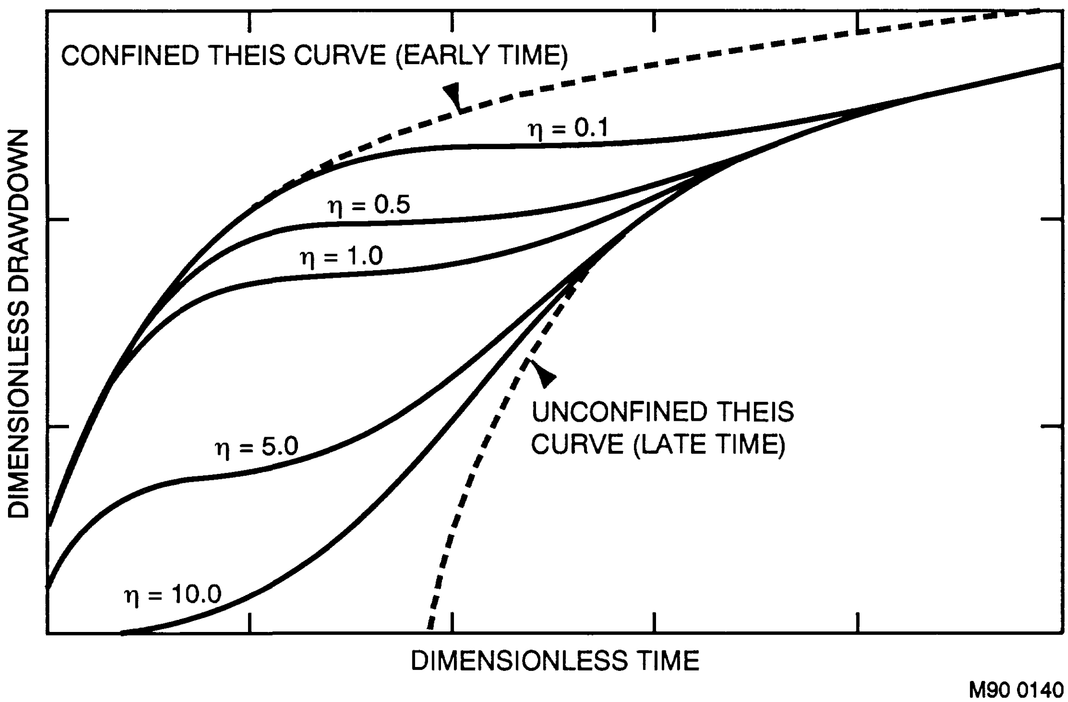


Figure 3.--Relation of drawdown and time for a well completed in an unconfined aquifer considering the effects of delayed gravity response and vertical components of flow. Both scales are logarithmic. The dimensionless parameter η relates anisotropy, radius, and aquifer thickness. Adapted from Neuman (1972, fig. 8).

nonequilibrium type curve for an unconfined aquifer. Neuman's curves reproduce all three segments of the time-drawdown response and allow the determination, with adequate field data, of horizontal and vertical hydraulic conductivity, specific (elastic) storage, and specific yield. Because aquifer thickness and effective radius are unknown, the analysis of time-drawdown data can yield only transmissivity and the dimensionless parameter η relating anisotropy, aquifer thickness, and radius. Freeze and Cherry (1979) use the notation η, which is used in this report; Neuman (1975) used β for the same parameter.

The type curves given by Neuman are for fully penetrating wells. Neuman (1975) has provided a computer program to develop additional theoretical curves for partially penetrating wells. Because data were lacking for aquifer thickness and (effective) radius of the pumping well, the special curves were not developed.

For the single-well aquifer tests at the INEL, most time-drawdown data complete enough for analysis show parts of the first two segments of the typical delayed gravity response in an unconfined aquifer. An example of the response and interpretation is given on figure 4. The data for 8 tests at 6 wells in perched aquifers and 27 tests at 20 wells in the Snake River Plain aquifer (Ackerman, 1991) matched type curves with a value of η between 0.001 and 0.4, generally between 0.004 and 0.1. The time-drawdown data for 11 tests at 7 wells, however, did not show a definite flat segment for intermediate time response. These data can be best matched to the Theis segment of the type curves. Those tests for which transmissivity was calculated by the type-curve method are noted on table 1.

Specific-Capacity Method

The specific-capacity method was used to estimate transmissivity from single-well tests. The advantages of this method are its simplicity and flexibility. The method does not require as much data as type-curve methods. However, the results may only represent the transmissivity near

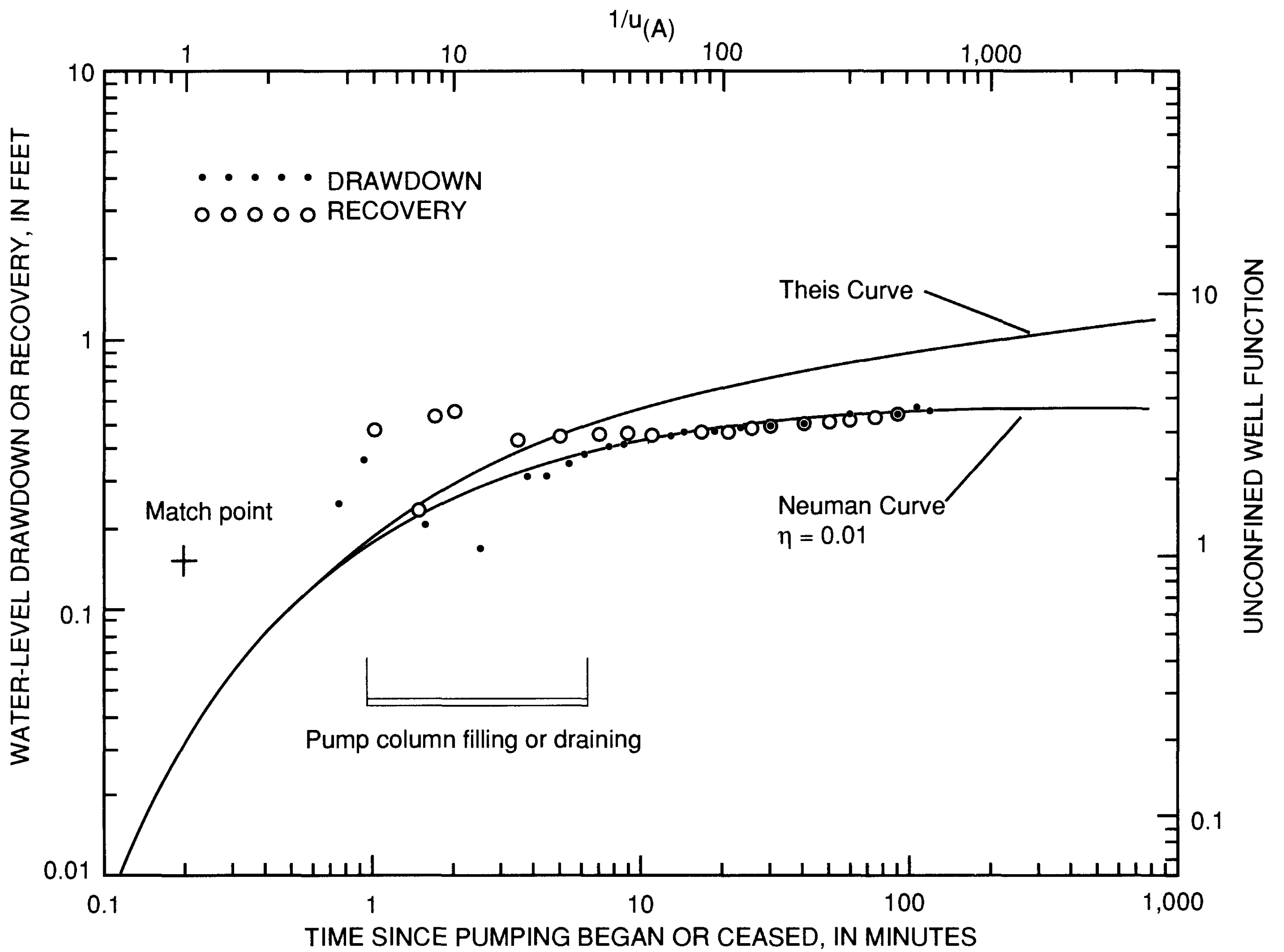


Figure 4.--Relation of observed drawdown to drawdown computed using Neuman's (1975) delayed gravity response method. Well is 03N 29E 14DDA3, well 60. Date of test is June 15, 1988.

the tested wells. Nevertheless, due to the wide distribution of these types of data, it is useful for studying the ground-water flow in the INEL area.

One of the most common and useful types of data available for describing the hydrologic properties of the aquifers at the INEL is the specific capacity of wells. Specific capacity is an expression of the productivity of a well and is commonly expressed as the ratio of the pumping rate, Q, in gallons per minute, to the total measured drawdown, s_T, in feet. The total drawdown, s_T, in a pumped well is the sum of all or some of the following components, which depend on hydrologic and well construction conditions:

> the drawdown s (aquifer loss), in hydraulic head in the aquifer at the well screen or borehole boundary due to laminar flow of water through the aquifer; plus the drawdown s_{wL} (well loss), due to turbulent flow of water through the screen or well face and inside the casing into the pump intake; plus the drawdown s_p, due to partial penetration of the aquifer by the pumped well; plus the drawdown s_d, due to dewatering part of the aquifer; plus or minus the drawdown or buildup s_b, due to boundaries of the aquifer; minus the buildup s_r, due to recharge boundaries of the aquifer.

Stated as an equation (Walton, 1970, p. 311),

$$s_T = s + s_{wL} + s_p + s_d \pm s_b - s_r. \quad (1)$$

In general, aquifer loss is by far the largest component of total drawdown. With proper design of aquifer tests and construction of wells, aquifer loss becomes the only measurable term.

A simple linear-regression analysis was used to empirically predict transmissivity from specific capacity. Because the data take on values covering more than 4 orders of magnitude, the regression analyses were performed on logarithmic-transformed values. The logarithmic-transform procedure minimized the overweighting of the largest values. The resulting regression equation relating transmissivity from type-curve analyses to specific capacity for single-well tests of perched aquifers with a correlation coefficient of 0.99 ($r^2 = 0.98$) is:

$$\text{Log } T = 1.1169 \text{ Log } Q/s + 1.6603 \qquad (2)$$

or

$$T = (Q/s)^{1.1169} \times 45.74 \qquad (3)$$

where Q/s = specific capacity of a well in gallons per minute per foot of drawdown and

T = transmissivity in feet squared per day.

A similar relation (Ackerman, 1991, equations 3 and 4) was reported for the type-curve analysis of 37 single-well tests for the Snake River Plain aquifer at and near the INEL. The equation derived for the 37 values was:

$$\text{Log } T = 1.1853 \text{ Log } Q/s + 1.6087 \qquad (4)$$

or

$$T = (Q/s)^{1.1853} \times 40.62. \qquad (5)$$

Because the same materials comprise the Snake River Plain aquifer and the perched aquifers, and because the regression for Snake River Plain aquifer tests had more degrees of freedom, equation 4 or 5 was used to estimate transmissivity. The use of the relation as determined from the Snake River Plain aquifer data allows for a consistent estimate to be made. The average difference between the residuals of the two relations was about 0.06 orders of magnitude. The data for the perched aquifers and the relation used are shown on figure 5.

Estimates of transmissivity calculated from equations 4 or 5 were regressed with corresponding estimates determined from type-curve analysis of aquifer tests on wells completed in perched aquifers. Results of a

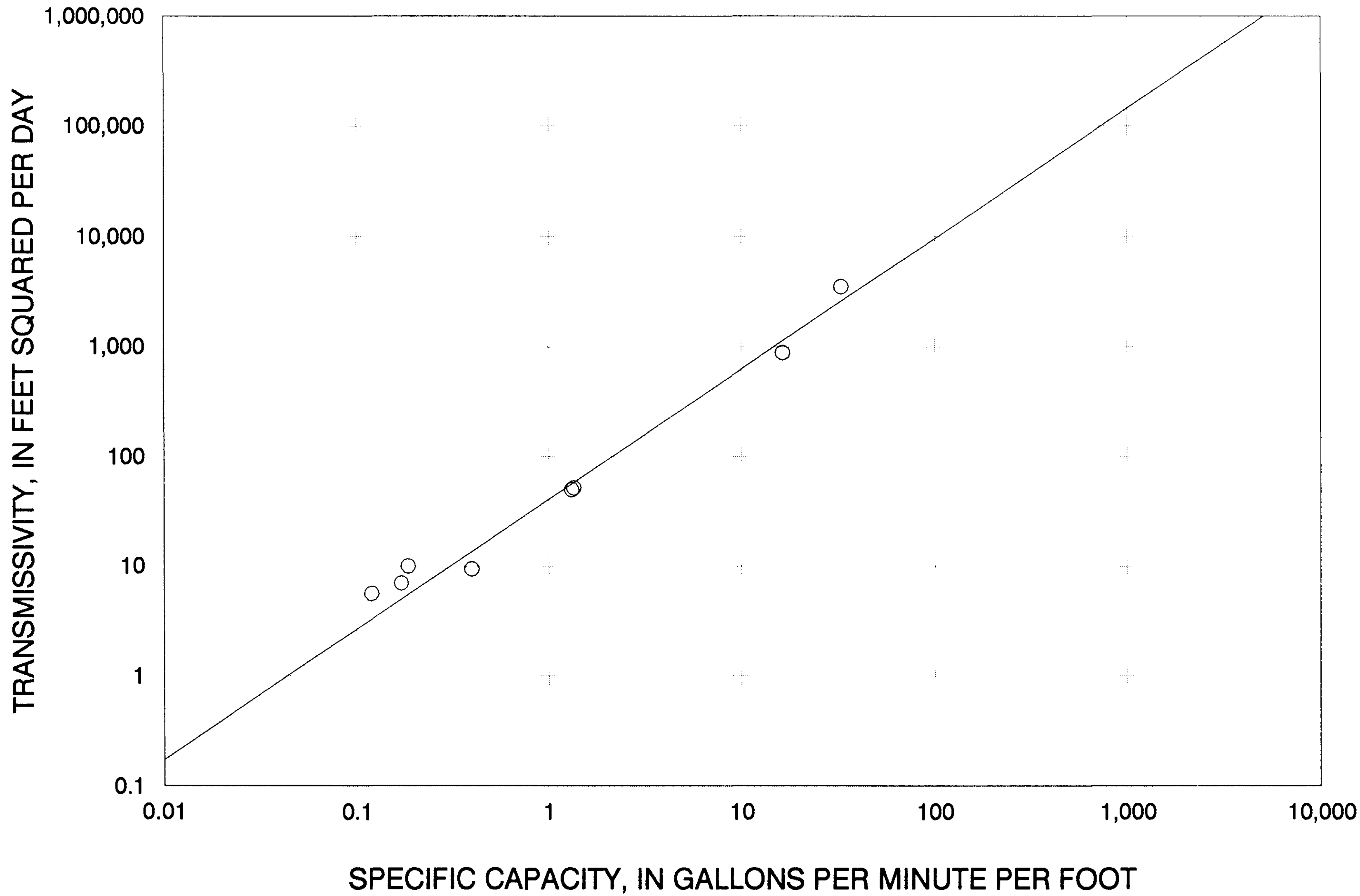


Figure 5.--Transmissivity estimates from specific-capacity data. Observations are from type-curve analysis. Regression equation is:

$$\text{Transmissivity} = \text{specific capacity}^{1.1853} \times 40.62.$$

simple least-squares regression of logarithmic-transformed values gave a correlation coefficient of 0.99 ($r^2 = 0.98$). Residuals of predicted transmissivity were evenly distributed (fig. 6) and had a maximum, minimum, and average of 0.26, -0.16, and 0.05 orders of magnitude, respectively.

RESULTS OF AQUIFER-TEST ANALYSES

Analyses of aquifer-test data by the type-curve method were used to judge the relative accuracy of estimates of transmissivity from the specific-capacity data. Because the data span more than 4 orders of magnitude, 0.029 to 150 (gal/min)/ft for specific capacity and 1.0 to 15,000 ft^2/d for transmissivity, uncertainties are expressed as orders of magnitude. This method of expressing uncertainty is a convenient normalization of data with a wide range of values. To convert the uncertainty to engineering units, subtract or add the uncertainty to the logarithm of the value and take the antilogarithm.

For those tests where data were insufficient to use the type-curve method, the established regression of transmissivity and specific capacity model was used to estimate transmissivity. If the drawdown was less than the limit of detection for the measurement method used, it is listed in table 1 as less than (<) the detection limit. Specific capacity calculated from these values is given as greater than (>) values.

The estimates of transmissivity in table 2 were chosen as the best or most representative of type-curve analyses or from specific-capacity data at individual wells. The tests listed in table 2 are cross-referenced by test number to those of table 1.

On the basis of the repeatability of transmissivity determinations for individual wells and on the agreement between type-curve and specific-capacity data analyses, relative uncertainties tabulated for estimates of transmissivity were assigned as follows:

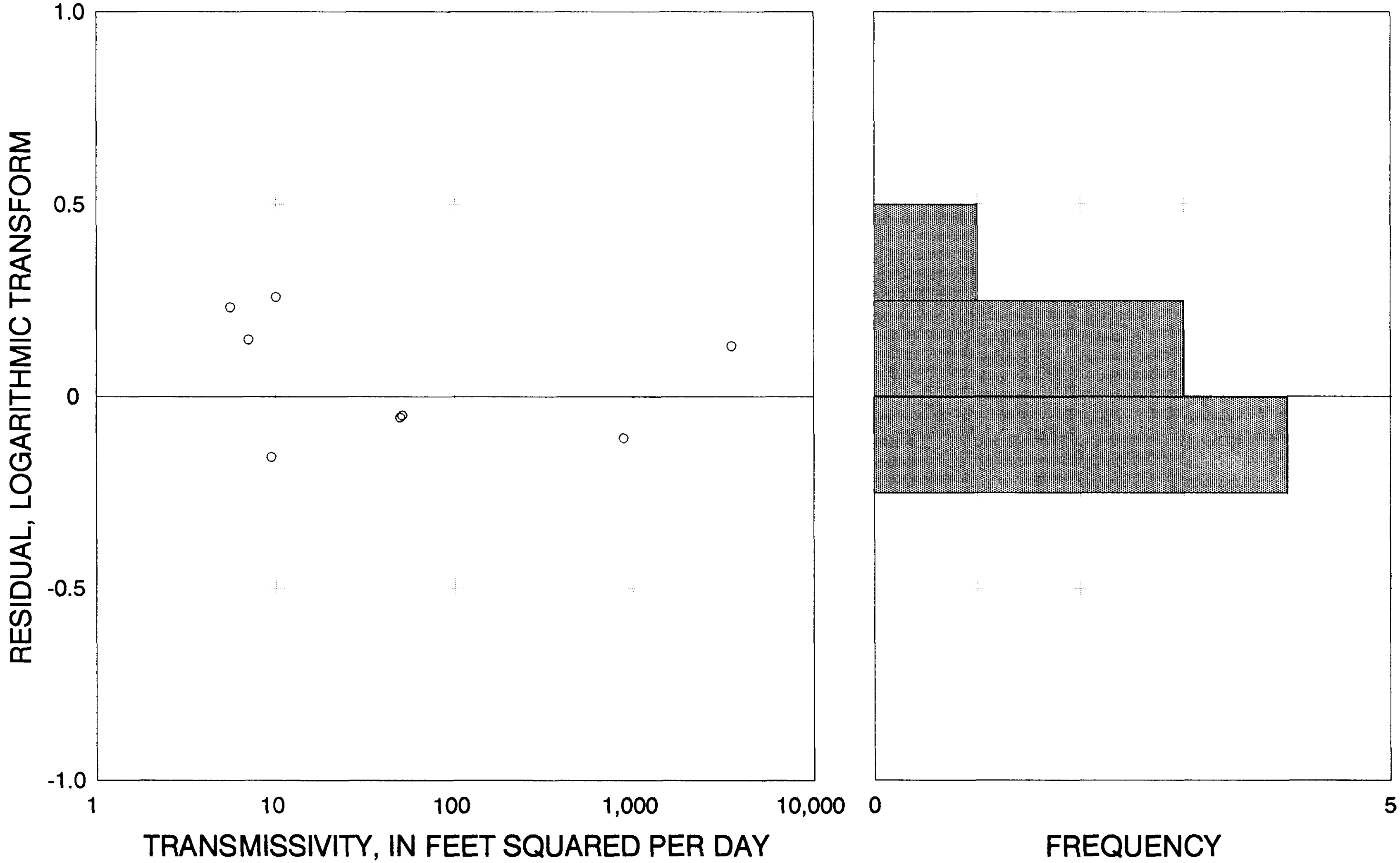


Figure 6.--Residuals of transmissivity predicted by simple linear-regression analysis.

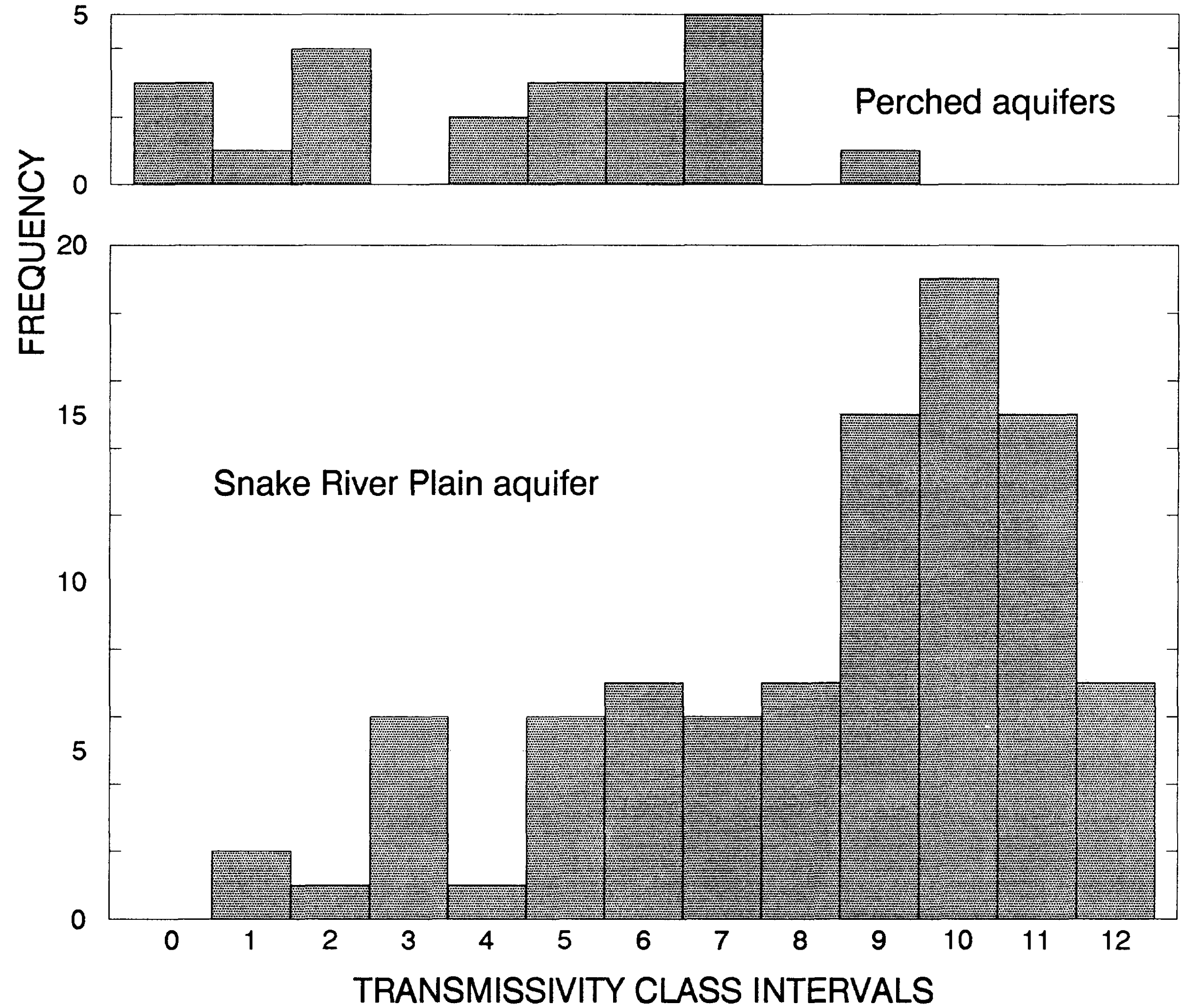
±0.1 order of magnitude, type-curve analysis;

±0.4 order of magnitude, specific-capacity analysis, drawdown greater than 0.1 ft;

±0.5 order of magnitude, specific-capacity analysis, drawdown observed less than 0.1 ft.

The range, distribution, and measures of central tendency for transmissivity estimates are shown on figure 7 and table 3 for aquifers at the INEL with definite (not based on drawdown less than detection limit) values. The transmissivity estimates for perched aquifers span more than 4 orders of magnitude (1.0 to 15,000 ft^2/d). Most measures of central tendency are close to 3,000 ft^2/d. Because of the complex nature of individual well completions, no attempt was made to present a hydraulic conductivity. A rough estimate of hydraulic conductivity can be made by dividing the transmissivity by the penetration of the well below the water table (table 2). The range of penetrations was 6 to 113 ft and the arithmetic mean 38 ft. Most values were between 20 and 50 ft. The range of values for hydraulic conductivity calculated using penetration and estimated transmissivity was nearly 5 orders of magnitude (0.014 to 790 ft/d). These values are consistent with the hydraulic conductivity of fractured basalts and lava flows as given by Heath (1983, p. 13), and Freeze and Cherry (1979, table 2.2). The range of values is also within the range of values (0.0086 to 5,500 ft/d) for the Snake River Plain aquifer at the INEL as given by Ackerman (1991, p. 30).

The large range of values for transmissivity or hydraulic conductivity has profound implications concerning flow in perched aquifers and the Snake River Plain aquifer at the INEL. Parts of the aquifers having a hydraulic conductivity of about 8 ft/d would be more than 2 orders of magnitude less permeable than parts of the aquifers with the greatest hydraulic conductivity. In like manner, that same part of the aquifer would be more than 2 orders of magnitude more permeable than the parts of the aquifer with the smallest hydraulic conductivity. Heath (1983, p. 24) stated that aquifers are 1 to 3 orders of magnitude more permeable than confining beds. If a



EXPLANATION

Units are feet squared per day; intervals are 0.5 order of magnitude wide

Class interval	Center of interval	Extremes of interval		
0	0.6	0.36	-	1.1
1	2	1.1	-	3.6
2	6	3.6	-	11
3	20	11	-	36
4	60	36	-	110
5	200	110	-	360
6	600	360	-	1,100
7	2,000	1,100	-	3,600
8	6,000	3,600	-	11,000
9	20,000	11,000	-	36,000
10	60,000	36,000	-	110,000
11	200,000	110,000	-	360,000
12	600,000	360,000	-	1,100,000

Figure 7.--Distribution of estimated transmissivity at and near the Idaho National Engineering Laboratory.

Table 3.--Central tendencies of transmissivity estimates for wells at and near the Idaho National Engineering Laboratory

[ft^2/d - feet squared per day]

Perched aquifers

Number of observations	22
Range of values	1.0 - 15,000 ft^2/d
Measures of central tendency	
Arithmetic mean	1,300 ft^2/d
Root mean square	3,500 ft^2/d
Geometric mean	98 ft^2/d
Median	19 ft^2/d
Mode	1,500 ft^2/d

Snake River Plain aquifer
(from Ackerman, 1991, table 3)

Number of observations	91
Range of values	1.1 - 760,000 ft^2/d
Measures of central tendency	
Arithmetic mean	93,000 ft^2/d
Root mean square	180,000 ft^2/d
Geometric mean	9,600 ft^2/d
Median	25,000 ft^2/d
Mode	60,000 ft^2/d

All aquifers

Number of observations	113
Range of values	1.0 - 760,000 ft^2/d
Measures of central tendency	
Arithmetic mean	76,000 ft^2/d
Root mean square	160,000 ft^2/d
Geometric mean	4,100 ft^2/d
Median	11,000 ft^2/d
Mode	58,000 ft^2/d

criterion of 2 orders of magnitude difference in hydraulic conductivity is sufficient to differentiate aquifers and confining beds, then some parts of the aquifer may be at once an aquifer and a confining bed relative to flow in other parts of the aquifer.

The estimates of transmissivity for perched aquifers provided in this report and in the companion report for the Snake River Plain aquifer (Ackerman, 1991) were determined in a consistent manner and are useful for describing the three dimensional distribution of aquifer properties. Such information is useful for evaluating regional differences in transmissivity of ground-water flow systems. Because the values were not determined from properly designed and well executed multiple-well tests, they are only of limited use for estimating well-field performance.

SUMMARY

Aquifer-test data of 43 single-well tests at 22 wells in the perched aquifers were analyzed to estimate values of transmissivity. Estimates of transmissivity for individual wells ranged from 1.0 to 15,000 ft^2/d, more than 4 orders of magnitude. These data were determined in a consistent manner and are useful for describing the distribution of transmissivity at the INEL.

The results of type-curve analysis of eight tests at six wells were used to verify a regression relation between specific capacity and transmissivity. This relation, in turn, was used to analyze all specific-capacity data. An estimate of transmissivity is made for the aquifer at each well. Values of relative uncertainty for estimated values of transmissivity generally ranged from 0.1 order of magnitude for type-curve analysis to 0.5 order of magnitude for specific-capacity data analysis with measured drawdown of less than 0.1 ft. Because of the paucity of adequate multiple-well aquifer tests the scope of interpretation is limited to single-well tests or specific-capacity data.

The values of transmissivities given in this report represent the transmissivity near the test wells and within the test interval. Due to the high degree of heterogeneity of the basalt and the unknown thickness of the aquifers it is more likely the transmissivity of the whole basalt sequence is different from those values given in this report.

REFERENCES

Ackerman, D.J., 1991, Transmissivity of the Snake River Plain aquifer at the Idaho National Engineering Laboratory, Idaho: U.S. Geological Survey Water-Resources Investigations Report 91-4058 (DOE/ID-22097), 35 p.

Bagby, J.C., White, L.J., and Jensen, R.G., 1984, Water-quality data for selected wells on or near the Idaho National Engineering Laboratory, 1949 through 1982: U.S. Geological Survey Open-File Report 84-714 (DOE/ID-22068), 800 p.

Boulton, N.S., 1954, The drawdown of the water table under nonsteady conditions near a pumped well in an unconfined formation: The Institution of Civil Engineers, Proceedings [London], pt. 3, p. 564-579.

----- 1963, Analysis of data from nonequilibrium pumping tests allowing for delayed yield from storage: The Institution of Civil Engineers Proceedings [London], v. 26, p. 469-482.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Englewood Cliffs, N.J., Prentice-Hall, 604 p.

Garabedian, S.P., 1989, Hydrology and digital simulation of the regional aquifer system, eastern Snake River Plain, Idaho: U.S. Geological Survey Open-File Report 87-237, 151 p.

Heath, R.C., 1983, Basic ground-water hydrology: U.S. Geological Survey Water-Supply Paper 2220, 82 p.

Lindner, J.B., and Reilly, T.E., 1983, Analysis of three tests of the unconfined aquifer in southern Nassau County, Long Island, New York: U.S. Geological Survey Water-Resources Investigations Report 82-4021, 46 p.

Lohman, S.W., 1972, Ground water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Neuman, S.P., 1972, Theory of flow in unconfined aquifers considering delayed response of the water table: Water Resources Research, v. 8, no. 4, p. 1031-1045.

Neuman, S.P., 1974, Effects of partial penetration on flow in unconfined aquifers considering delayed gravity response: Water Resources Research, v. 10, no. 2, p. 303-312.

----- 1975, Analysis of pumping test data from anisotropic unconfined aquifers considering delayed gravity response: Water Resources Research, v. 11, no. 2, p. 329-342.

Orr, B.R., and Cecil, L.D., 1991, Hydrologic conditions and distribution of selected chemical constituents in water, Snake River Plain aquifer, Idaho National Engineering Laboratory, Idaho, 1986 to 1988: U.S. Geological Survey Water-Resources Investigations Report 91-4047 (DOE/ID-22096), 86 p.

Pittman, J.R., Jensen, R.J., and Fischer, P.R., 1988, Hydrologic conditions at the Idaho National Engineering Laboratory, 1982 to 1985: U.S. Geological Survey Water-Resources Investigations Report 89-4008, (DOE/ID-22078), 73 p.

Prince, K.R., and Schneider, B.J., 1989, Estimation of hydraulic characteristics of the upper glacial and Magothy aquifers at East Meadow, New York, by use of aquifer tests: U.S. Geological Survey Water-Resources Investigations Report 87-4211, 43 p.

Robertson, J.B., 1977, Numerical modeling of subsurface radioactive solute transport from waste-seepage ponds at the Idaho National Engineering Laboratory: U.S. Geological Survey Open-File Report 76-717 (IDO-22057), 68 p.

Robertson, J.B., Schoen, Robert, and Barraclough, J.T., 1974, The influence of liquid waste disposal on the geochemistry of water at the National Reactor Testing Station, Idaho, 1952-70: U.S. Geological Survey Open-File Report (IDO-22053), 331 p.

Theis, C.V., Brown, R.H., and Meyer, R.R., 1963, Estimating the transmissibility of aquifers from the specific capacity of wells, in Bentall, Ray, Methods of determining permeability, transmissibility, and drawdown: U.S. Geological Survey Water-Supply Paper 1536-I, p. 331-341.

Walton, W.C., 1970, Groundwater resource evaluation: New York, McGraw-Hill, 664 p.