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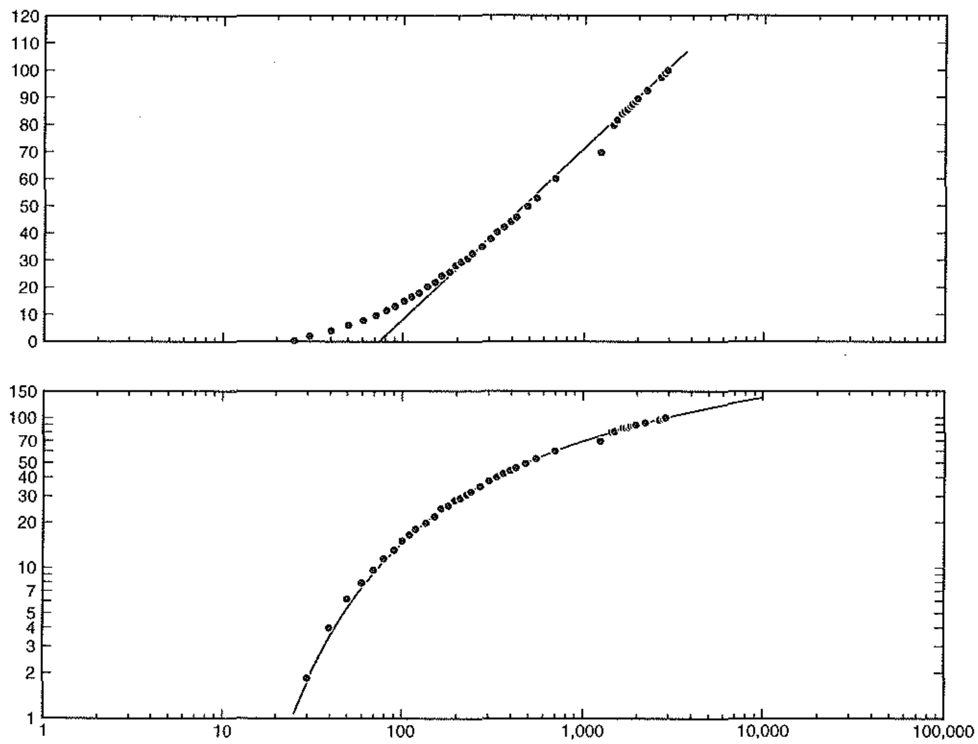
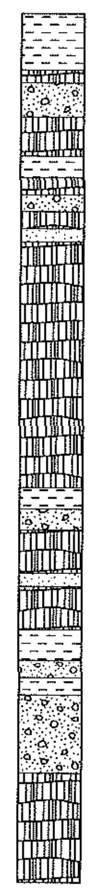
Estimating Transmissivity and Storage Properties from Aquifer Tests in the Southern Lihue Basin, Kauai, Hawaii

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 99-4066





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By Stephen B. Gingerich

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Honolulu, Hawaii
1999

U.S. DEPARTMENT OF THE INTERIOR
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CONTENTS

Figures

Tables

Conversion Factors

Multiply	By	To obtain
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
cubic foot per minute (ft^3/min)	0.02832	cubic meter per minute
mile (mi)	1.609	kilometer

Estimating Transmissivity and Storage Properties from Aquifer Tests in the Southern Lihue Basin, Kauai, Hawaii

By Stephen B. Gingerich

Abstract

Three to four different analysis methods were applied to the drawdown or recovery data from five constant-rate aquifer tests of 2 to 7 days in length to estimate transmissivity of rocks in the southern Lihue basin, Kauai, Hawaii. The wells penetrate rocks of the Koloa Volcanics and the underlying Waimea Canyon Basalt. Because the wells are located far apart and in previously unexplored areas, it is difficult to accurately define the aquifer or aquifers penetrated by the wells. Therefore, the aquifer tests were analyzed using a variety of curve-matching methods and only a range of possible values of transmissivity were determined. The results of a multiple-well aquifer test are similar to a single-well aquifer test done in the same area indicating that the single-well aquifer-test results are reasonable.

The results show that transmissivity in the Lihue basin ranges over several orders of magnitude, 42 to 7,900 square feet per day, but is generally lower than reported values of transmissivity of other basaltic aquifers in Hawaii. Estimates of confined-aquifer storage coefficient range from 1.3×10^{-4} to 8.2×10^{-2}. The hydraulic conductivity estimates obtained using an elliptical-equation method compare favorably with the results obtained from the generally more-accepted curve-matching methods. No significant difference is apparent between the estimated transmissivity of the Koloa Volcanics and the Waimea Canyon Basalt in the study area. An analysis of the lithology penetrated by the wells indicates the transmissivity is probably controlled mainly by the stratigraphic position of the layers penetrated by the well. The range of transmissivity values estimated for the southern Lihue basin is lower than reported values from aquifer tests at wells penetrating postshield-stage or rejuvenation-stage lava flows on other Hawaiian islands. This range is one to four orders of magnitude lower than most reported values for dike-free basalt aquifers in Hawaii.

INTRODUCTION

Kauai is the most geologically complex of the eight main Hawaiian islands (Macdonald and others, 1960) and the Lihue basin (fig. 1) is one of the most geologically complex areas on Kauai, yet little subsurface geohydrologic information is available compared with more developed areas in Hawaii. Geohydrologic information, such as aquifer transmissivity and storage coefficient, is necessary for developing conceptual and numerical ground-water flow models of ground-water movement in the basin. Most of the currently available geohydrologic information about Kauai was presented by Macdonald and others (1960) who readily admitted the occurrence of ground water in the Lihue area was practically unexplored. Currently, no published reports exist which describe transmissivity estimates of the rocks forming the Lihue basin.

In 1991, the County of Kauai Department of Water (Kauai DOW) and the U.S. Geological Survey (USGS) began a cooperative study of the ground-water resources of Kauai, a study which also will increase what is known in general about ground-water occurrence on eroded volcanic islands. The study included an existing-data review, a water-budget computation (Shade, 1995a) and a 1990 water-use summary (Shade, 1995b). In 1995, the studies were focused on the

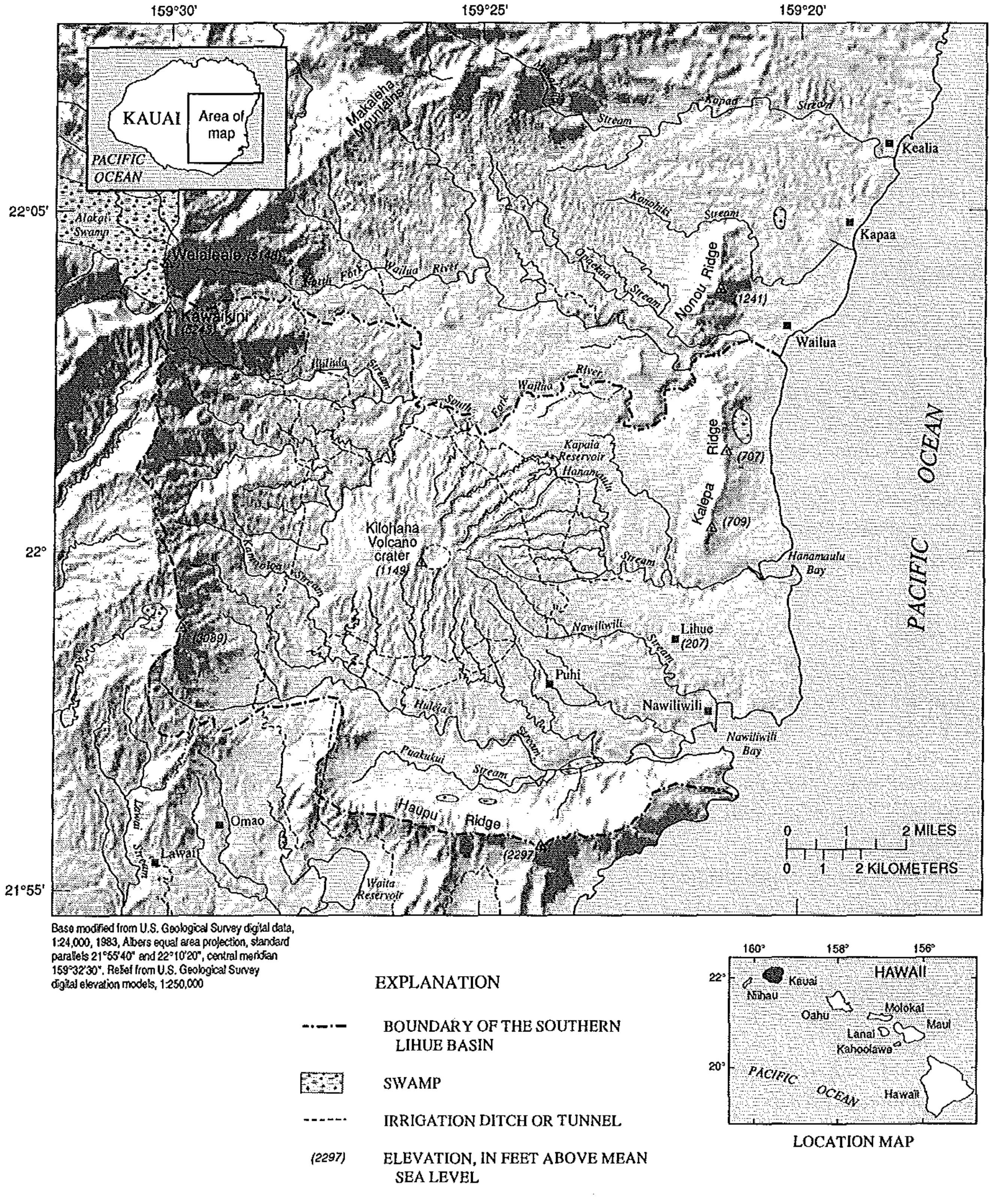


Figure 1. Location of the southern Lihue Basin, Kauai, Hawaii.

southern Lihue basin, which extends from the south fork of the Wailua River to the base of Haupu Ridge (fig. 1), where the need for resource assessment was considered most critical. New ground-water data were collected and analyzed, including drilling, lithologic descriptions, and aquifer tests of new monitor wells in previously unexplored areas of the southern Lihue basin (Gingerich and Izuka, 1997a, 1997b; Izuka and Gingerich, 1997a, 1997b, 1997c, 1997d). The study, which included a numerical ground-water flow model, produced a comprehensive description of ground-water occurrence and movement in the southern Lihue basin (Izuka and Gingerich, 1998).

Purpose and Scope

The purpose of this report is to present aquifer transmissivity, hydraulic conductivity, and storage coefficient estimates that were made on the basis of the analyses of aquifer-test data collected from wells drilled into the rocks of the southern half of the Lihue basin, Kauai, Hawaii. Five single-well aquifer tests and one multiple-well aquifer test were done and several different analysis methods were applied to the drawdown or recovery data including the methods of Moench (1985), Cooper and Jacob (1946), Theis (1935), Neuman (1974), and Harr (1962)/Polubarinova-Kochina (1962). (The Harr/Polubarinova-Kochina method will hereafter in this report be referred to as the Harr method.) Included is a discussion of the appropriateness of the different methods used to analyze the aquifer-test data and the advantages and disadvantages of applying each method to aquifer tests done in thick basalt aquifers. The results of these aquifer-test analyses are compared with published results from some of the other Hawaiian islands and the differences and similarities between these results are considered.

Hydrogeologic Setting

The Lihue basin is a large semicircular depression in the eastern half of Kauai, the fourth-largest island in the Hawaiian archipelago. The western margin of the basin is formed by the high central mountains of Kauai, including Mt. Waialeale, which is at 5,480 ft altitude (fig. 1). The northern boundary of the basin is formed by the Makaleha Mountains and the southern margin of the basin is formed by Haupu Ridge. Kalepa Ridge and Nonou Ridge form a line of smaller mountains near the eastern coastline. In the south-central part of the basin lies the broad low dome of the Kilohana Volcano, a rejuvenated-stage shield volcano. In this report, the southern Lihue basin is considered the part of the basin south of the South Fork of the Wailua River (fig. 1).

The rocks of the Lihue basin are divided into two geologic formations which are separated by erosional unconformities (Macdonald and others, 1960; Langenheim and Clague, 1987). Kauai is composed primarily of the tholeiitic Waimea Canyon Basalt (fig. 2) formed during the shield-volcano building period of Kauai's geologic history. In the Lihue basin, the Waimea Canyon Basalt forms the basement on which younger sediments and volcanic rocks lie, but crops out only in the ridges and high central mountains surrounding and within the basin (fig. 2). Most of the Waimea Canyon Basalt in the Lihue basin belongs to the Napali Member, which consists of thick accumulations of thin lava flows. The Napali Member is classified as highly permeable by Macdonald and others (1960). Numerous volcanic dikes cut vertically across the lava flows in the ridges where the Waimea Canyon Basalt is exposed and dikes also may be present in the Waimea Canyon Basalt beneath the Lihue basin although there is no drilling information confirming the latter. Volcanic dikes are commonly considered barriers to ground-water flow because of their relatively low permeability. The dike-intruded rocks of Haupu Ridge are classified as moderately to poorly permeable (Macdonald and others, 1960).

Sediments and volcanic rocks of the Koloa Volcanics rest unconformably on the eroded surface of the Waimea Canyon Basalt (Macdonald and others, 1960). The rocks of the Koloa Volcanics include thick, massive lava flows of highly alkalic rocks including alkalic olivine basalt, nephelinite, melilitite, and basanites. These mafic igneous rocks were erupted during a period of rejuvenated-stage volcanism from vents scattered over the old, eroded shield volcano and fill valleys, gorges, and depressions in the Waimea Canyon Basalt. The Koloa Volcanics is a heterogeneous unit which includes weathered lava flows, ash, tuff, cinder, and sediments (Macdonald and others, 1960). Some of the sediments have been divided into the Palikea Breccia Member.

The Koloa Volcanics in the Lihue basin accumulated to greater than 1,000 ft thick. A geologic cross

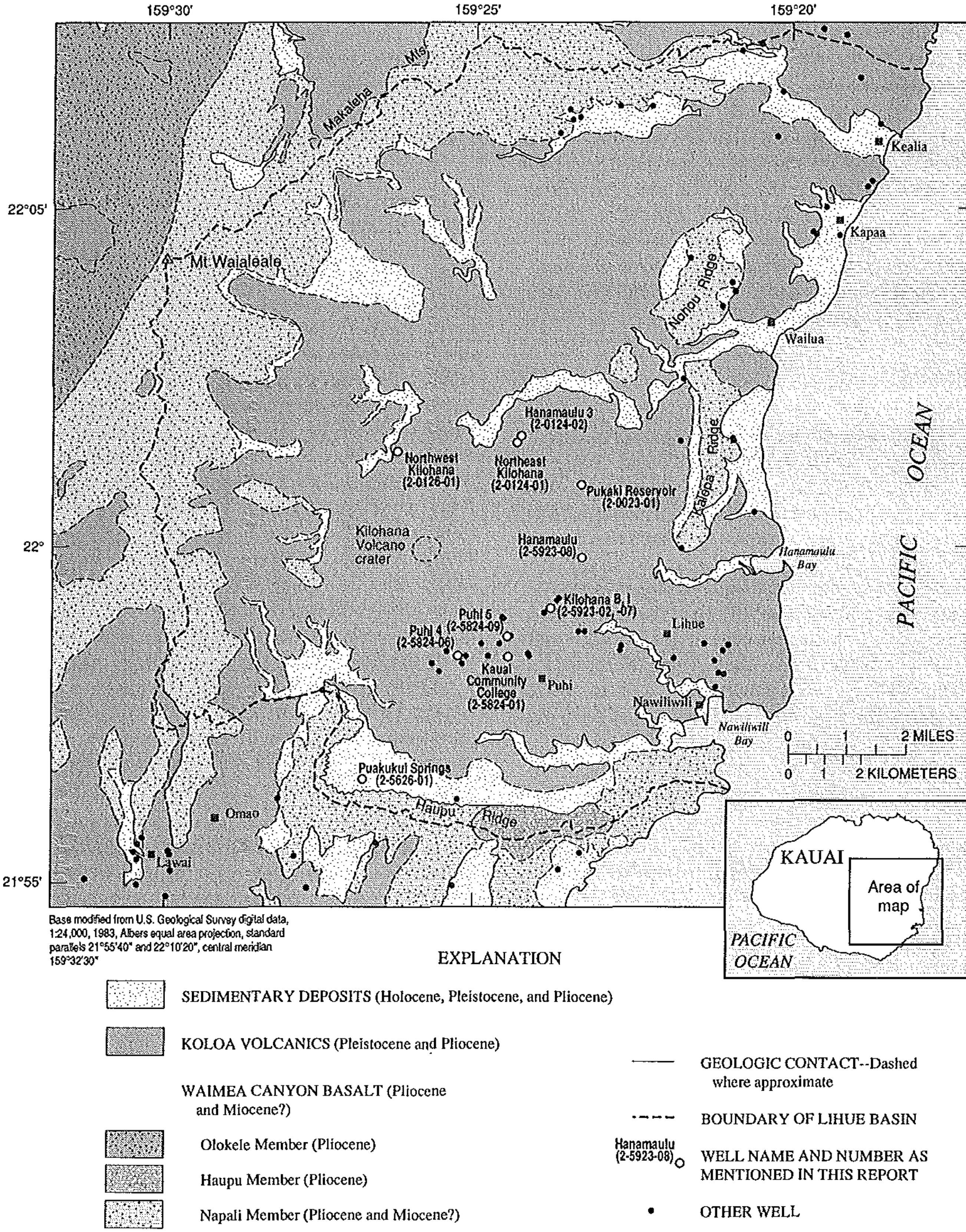


Figure 2. Geology and well locations in the Lihue basin area, Kauai, Hawaii (modified from Macdonald and others, 1960).

section in the geologic map of Macdonald and others (1960) shows the contact between Koloa Volcanics and the underlying Waimea Canyon Basalt at about -500 ft elevation. However, exploratory wells indicate the Koloa Volcanics is thicker in some places (Izuka and Gingerich, 1997a, 1997b, 1997c, 1997d; Reiners, P.K., and others, Univ. of Washington, written commun., 1997) and thus the contact is probably deeper. The thickness of the Koloa Volcanics is variable and depends mainly on the shape of the eroded surface that these rocks overlie. The thick, dense, lava flows and intercalated sediments of the Koloa Volcanics are classified as poorly to moderately permeable by Macdonald and others (1960) although no transmissivity data substantiate this classification.

Acknowledgments

The author is grateful to current Kauai DOW Chief Manager and Engineer Ernest Lau, and the staff of the Kauai DOW for their cooperation and assistance.

AQUIFER-TEST ANALYSES

Five single-well aquifer tests and one multiple-well aquifer test were analyzed using several different analysis methods. The single wells and the methods used to analyze the data, listed in order from north to south (fig. 2) are: the Northeast Kilohana monitor well (Jacob, Moench, Theis recovery, and Harr), the Northwest Kilohana monitor well (Jacob, Moench, and Harr), the Pukaki Reservoir monitor well (Jacob, Theis, Moench, and Harr), the Hanamaulu monitor well (Jacob, Theis, Theis recovery, and Harr), and the Puakukui Springs monitor well (Jacob, Theis, and Harr). In addition, the Northeast Kilohana monitor well was used as an observation well for an aquifer test of a new nearby production well and the data were analyzed using the methods of Neuman and Theis recovery. The monitor wells were drilled for an exploratory study in areas where no other geologic or hydrologic information previously existed. Because the wells are far apart and in previously unexplored areas, it is difficult to accurately define the extent and thickness of the aquifer penetrated by the wells. All of the wells were assumed to penetrate layers of volcanic rocks that are fully saturated from the water table at altitudes of several hundred feet above sea level to the base of the well below sea level. Although the wells penetrate multiple layers of basalt flows and sedimentary deposits, the aquifer-test methods used assume a single aquifer because scant lithologic information is available to accurately define the thickness or extent of any one layer penetrated by a well. The aquifer tests were analyzed using a variety of methods selected on the basis of the descriptions available from the drilling records and on the type of drawdown response recorded. Because of the uncertainty of the aquifer conditions and the variety of the methods used, reporting only a range of possible values of transmissivity and storage coefficient is appropriate.

Aquifer-Test Methods

For the single-well aquifer tests, three or four different analysis methods were applied to the drawdown or recovery data measured in the pumped well; Moench (1985), Cooper and Jacob (1946), Theis (1935), and Harr (1962)/Polubarinova-Kochina (1962). In addition, the Neuman (1974) and Theis methods were applied to the drawdown data measured in Northeast Kilohana monitor well during several multiple-well aquifer tests.

All of the methods used in this report, with the exception of the Moench and Theis recovery methods, require values of aquifer drawdown which are usually measured in an observation well near a withdrawal well. In the single-well aquifer tests, drawdown is measured in the pumped well and therefore must be corrected before the methods are applied because the pumped-well drawdown is a combination of aquifer drawdown (aquifer loss) and well loss (Jacob, 1947). The drawdown data is corrected by subtracting the well loss from the total measured drawdown to calculate the aquifer drawdown.

Aquifer loss, which varies linearly with the withdrawal rate, represents the loss in head caused by the friction of water moving through the aquifer material. Well loss is defined by Jacob (1947) to be the loss of head as water flows turbulently at high velocities through the well screen and upward inside the well casing to the pump intake. Jacob (1947) also states that well loss is approximately proportional to the square of the withdrawal rate. In a review paper on the well-loss function, Ramey (1982) indicates that the nonlinear head losses also include the effects of high velocity nonlaminar or non-Darcian flow in the aquifer adjacent to the well. One cause of nonlaminar flow effects is partial penetration of the aquifer which causes the total flow to pass through limited openings at the well face (Ramey, 1982). Much of the water probably enters the uncased well bore through openings between individual lava flows which, in the Koloa Volcanics, may be many tens of feet thick in places. In addition, Cooley and Cunningham (1979) show head losses are minimized if most of the flow into the well bore is near the pump intake. Conversely, if the pump intake is not directly opposite the section of the aquifer containing openings capable of producing significant flow, well loss will be higher. The pump-intake location was not determined on the basis of well lithology in any of the aquifer tests analyzed in this report, therefore well losses are expected to be greater. Overall, a significant amount of well loss may be attributed to turbulent flow near and in the well bore even though the wells are uncased or unscreened.

Some of the assumptions used in the aquifer-test methods and the appropriate equations necessary for applying the methods are described below.

Moench Method

The Moench (1985) method is an analytical solution developed for analyzing drawdown data from a single large-diameter pumped well completely penetrating an infinite aquifer with semi-confining units above and below the aquifer. This method is an extension of the method described by Hantush (1960) that proposed three idealized systems containing an aquifer between various combinations of low-permeability units. This analysis assumes the aquifer is bounded above and below by semi-confining units separating the aquifer from units that may act as constant sources of water (fig. 3). Although evidence for this layered configuration is conjectural, it is possible given the multiple lava flows that make up the aquifer in the Lihue basin.

The measured drawdown data (not corrected for well loss) is plotted against time on a log-log plot and a type curve is fit to the data. The equations for the type curves are complex and are usually solved with the aid of a computer. The reader is referred to the original publication for a complete description of the development of the analytical solution (Moench, 1985). The shape of the type curve depends on several factors: well-bore storage effects, the transmissivity and storage coefficient of the aquifer, and the transmissivity and storage coefficient of the confining unit(s). The coefficients used to generate the type curve are shown in figure 3.

A best-fitting match between the data and a type curve is determined and a match point (t_D, h_D, t, s) is obtained. Aquifer transmissivity is estimated from:

$$T = \frac{Qh_D}{4\pi s}, \qquad (1)$$

where:

T = transmissivity, in feet squared per minute,
Q = withdrawal rate, in feet cubed per minute,
h_D = dimensionless drawdown, determined from the match point,
π = the number pi, 3.14159, and
s = drawdown at the match point, in feet.

An advantage of this method is that it allows the aquifer storage coefficient to be calculated from a single-well test using:

$$S = S_s b = \frac{Tt}{t_D r_w^2}, \qquad (2)$$

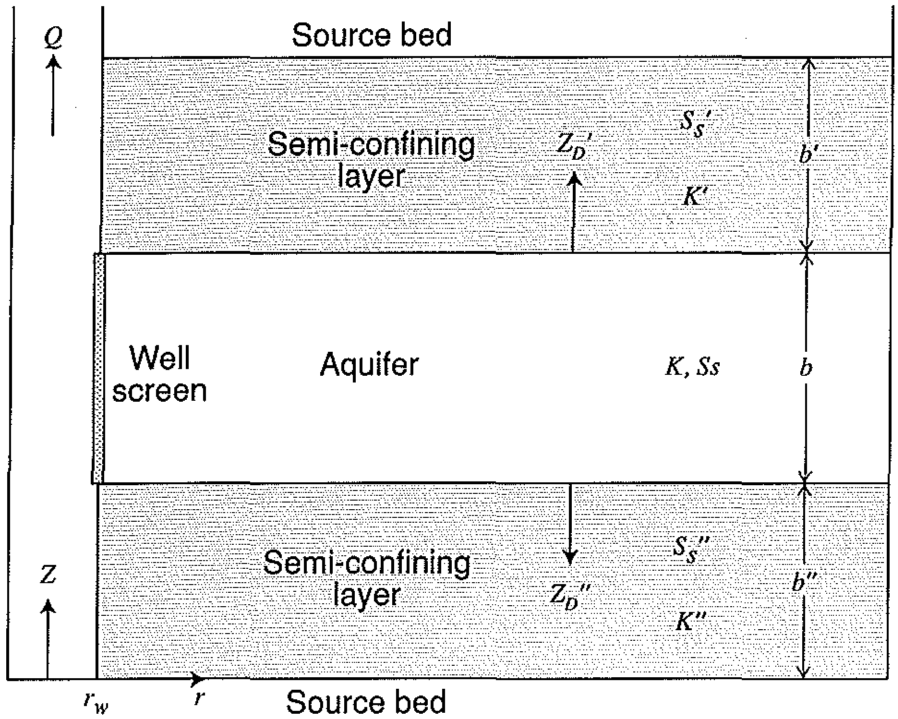
where:

S = aquifer storage coefficient, dimensionless,
S_s = aquifer specific storage, in inverse feet,
b = aquifer thickness, in feet,
t = time since withdrawal began at the match point, in minutes,
t_D = dimensionless time, determined from the match point, and
r_w = the radius of the pumped well, in feet.

When the pumping well diameter is small, effects of well-bore storage may not be apparent in the drawdown data and care must be taken not to overanalyze the data. For aquifers with the appropriate geometry, the Moench method is useful for three main reasons: (1) it was developed specifically for single-well tests, (2) it can be used on data that do not have to be corrected for well loss, and (3) it provides an estimate of aquifer storage coefficient.

Jacob Method

The Jacob method (Cooper and Jacob, 1946) is a simple method for analyzing drawdown data measured in an observation well located a distance, r, from a pumped well in a confined infinite aquifer. For single-well tests, the drawdown data must be corrected for well loss and the determination of aquifer storage coefficient is not possible. This method can be used for an



$$\gamma' = r_w \left(\frac{K'}{K\,b\,b'} \right)^{1/2}$$

$$\gamma'' = r_w \left(\frac{K''}{K\,b\,b''} \right)^{1/2}$$

$$\sigma' = \frac{S_s'\,b'}{S_s\,b}$$

$$\sigma'' = \frac{S_s''\,b''}{S_s\,b}$$

Q = well withdrawal rate, ft^3/min

Z_D = dimensionless vertical distance

b = thickness of a given layer, ft

K = hydraulic conductivity of a given layer, ft/min

r = radial coordinate originating at the center of pumped well, ft

r_w = effective radius of pumped well, ft

S_s = specific storage of a given layer, ft^{-1}

$'$, $''$ = denotes upper and lower confining unit, respectively

Figure 3. Schematic diagram of a leaky aquifer system (modified from Moench, 1985).

unconfined aquifer when aquifer drawdown is insignificant compared to the saturated thickness of the aquifer or if a correction to the aquifer drawdown data, attributed to C.E. Jacob by Kruseman and de Ridder (1991), is made as follows:

$$s_c = s - \left(\frac{s^2}{2b}\right), \tag{3}$$

where:

s_c = corrected drawdown, in feet.

Aquifer transmissivity is estimated by fitting a straight line through the data on a semi-log plot of drawdown on the linear axis against time since withdrawal began on the logarithmic axis.The amount of drawdown per log cycle, Δs, is determined and used in:

$$T = 2.3\frac{Q}{4\pi\Delta s}. \tag{4}$$

This method is considered acceptable for drawdown data from single-well tests after:

$$t > \frac{25r_w^2}{T}. \tag{5}$$

According to Kruseman and de Ridder (1991), the effects of well-bore storage can be neglected after this time.

The semilog plot of the drawdown data is useful for demonstrating the presence of one or more barriers to ground-water flow. Theoretically, when the effect of low-permeability barrier is observed at the observation well, the slope of the time-drawdown data on the semilog plot will double and image well theory can be used to determine the barrier locations if at least three observation wells are available. In practice, the geometry of the ground-water barrier(s) is rarely known and usually only one or no observation wells are available, making it difficult to analyze the aquifer-test data.

Theis Method

The Theis (1935) method also is used for analyzing drawdown data measured in an observation well located a distance, r, from a pumped well in a confined infinite aquifer. Hantush (1961) presented a modification to the Theis method that allows for wells that only partially penetrate an aquifer.

As with the Jacob method, this method can be used to analyze data from wells in an unconfined aquifer when aquifer drawdown is insignificant when compared to the aquifer thickness or if a correction is made to the drawdown data using equation 3.

Transmissivity is estimated by matching a log-log plot of drawdown against time with a theoretical type curve of the Theis well function, $W(u)$, plotted against $1/u$. Most ground-water texts contain the derivation of these terms and usually present a table of $W(u)$ for values of u over a wide range (see Lohman, 1972). A match of the drawdown data to the type curve is obtained, either through visual inspection or with the aid of computer software, a match point (s, t, $W(u)$, $1/u$) is determined and the values are entered into the following equation:

$$T = \frac{Q}{4\pi s}W(u). \tag{6}$$

The use of the Theis curve-matching method for single-well tests can be difficult because aquifer storage coefficient cannot be determined. The aquifer storage coefficient mainly controls the type-curve position on the time axis. Because the storage coefficient estimate is meaningless for single-well tests, the type-curve position has no limit in the time dimension. Therefore, the type-curve fit can be ambiguous and should not be accepted without comparison to results from other appropriate methods.

Neuman Method

The Neuman (1974) method is an analytical solution developed for analyzing drawdown data from an observation well located a distance, r, from a pumped well in an anisotropic, unconfined infinite aquifer. The measured drawdown data in the observation well is plotted against time on a log-log plot and a type curve is fit to the data. The equations for the type curves are complex and are usually solved with the aid of a computer. The reader is referred to the original publication (Neuman, 1974) for a complete description of the development of the analytical solution. The shape of the type curve depends on several factors: the aquifer transmissivity, aquifer anisotropy, the storage coefficient controlling the early-time drawdown, and the specific yield controlling the late-time drawdown.

A best-fitting match between the data and both early-time and later-time type curves is determined and match points (t, s, t_y, s_D) are obtained. Aquifer transmissivity is estimated from:

$$T = \frac{Qs_D}{s}, \tag{7}$$

where:

T = transmissivity, in feet squared per minute,
Q = withdrawal rate, in feet cubed per minute,
s_D = dimensionless drawdown, determined from the match point,
s = drawdown at the match point, in feet.

Aquifer specific yield is calculated using:

$$S_y = \frac{Tt}{t_D r^2}, \tag{8}$$

where:

S_y = aquifer specific yield, dimensionless,
t = time at the match point since withdrawal began, in minutes,
t_D = dimensionless time, determined from the match point, and
r = the distance from the pumped well to the observation well, in feet.

This method can also be used on data from single-well tests but is not recommended because the effects of well loss and well-bore storage produce drawdown data that are commonly ambiguous. Therefore, the investigator may be tempted to overanalyze the data and attribute more reliability than is warranted to a good fit of a type curve using the Neuman method.

Theis Recovery Method

The Theis (1935) recovery method is used to analyze recovery data from an observation well that fully penetrates a confined infinite aquifer. For unconfined aquifers, corrections to the drawdown data should be made using equation 3. The pumped well, also fully penetrating, is assumed to have had a constant withdrawal rate and then shut off at time t. The residual drawdown data, are plotted on a semi-log graph against t/t' (on the logarithmic axis) with t' being the time since withdrawal stopped. A straight line is drawn through the late-time data and the drawdown, Δs, over one log cycle is determined from the graph. A transmissivity estimate is determined using equation 4. No well-loss corrections to the observed recovery data are made for this analysis.

For multiple-well tests, this method is acceptable for unconfined conditions if:

$$t' > \frac{r^2 S}{4T(0.1)}. \tag{9}$$

For a single-well test in a confined or unconfined aquifer, Kruseman and de Ridder (1991) suggest that the following condition must be met before this method is applied:

$$t' > \frac{500 r_w^2}{T}. \tag{10}$$

The effects of well-bore storage are assumed to have completely dissipated by time t'.

Harr/Polubarinova-Kochina Method

The method presented by Harr (1962) and Polubarinova-Kochina (1962) estimates the conductivity of a thick, unconfined aquifer that is penetrated only partially by a pumped well. Well-construction and aquifer-test information are used in the following equation to estimate aquifer hydraulic conductivity:

$$K = \frac{Q}{2\pi L s_s} \ln\left(1.6 \frac{L}{r_w}\right), \tag{11}$$

where:

K = hydraulic conductivity, in feet per minute,
Q = withdrawal rate, in feet cubed per minute,
L = length of open interval of pumped well, in feet,
s_s = steady-state drawdown in pumped well, in feet, and
r_w = radius of pumped well, in feet.

Transmissivity is related to the aquifer hydraulic conductivity using:

$$T = Kb. \tag{12}$$

The Harr method is useful for obtaining hydraulic conductivity estimates from aquifer tests that are poorly run or for which only sparse data are available. A drawback to this method is the ambiguity of the value of s_s used in the analysis. The term "steady-state drawdown" described by the authors does not include a definition at which time this condition is met. One approach is to assume "steady-state" conditions are met when the drawdown per unit time has become relatively small

Table 1. Summary of aquifer-test data, Kauai, Hawaii
[min, minutes; ft, feet; ft^3/min, cubic feet per minute; --, not applicable]

Pumped-well State well number	Pumped well name	Date of test	Duration of test (min)	Radius of well, r_w (ft)	Pumping rate, Q (ft^3/min)	Estimated well loss at Q (ft)	Assumed aquifer thickness, b (ft)	Open or screened interval of pumped well, L (ft)
2-5626-01	Puakukui Springs	12/4/95	10,080	0.50	40.9	3.6	486	486
2-5923-08	Hanamaulu	10/6/95	2,880	0.17	10.2	12	943	878
2-0023-01	Pukaki Reservoir	4/2/96	5,700	0.42	38.0	50[a]	975	991
2-0124-01	Northeast Kilohana	7/1/95	10,080	0.42	42.3	14	924	307
2-0124-02	Hanamaulu 3	10/12/98	351	146[b]	13.4–26.7	--	924	--
		10/15/98	5,759	146[b]	21.6–42.8	--	924	--
		10/26/98	6,082	146[b]	19.4–20.1	--	924	--
2-0126-01	Northwest Kilohana	1/24/96	10,080	0.42	41.8	56	916	806

[a] Well loss estimated on the basis of drawdown after 1 minute of withdrawal

[b] Distance to observation well

compared with the early-time drawdown. The time at which this condition has been met varies and must be determined on a case by case basis.

The drawdown per log cycle can be estimated from a semi-log plot of drawdown and time (using a straight-line fit as in the Jacob method) and extrapolated to determine the expected drawdown after a time at which "steady-state" conditions have been approached. For an aquifer test that lasts several days in an aquifer with a relatively high hydraulic conductivity, the drawdown per unit time becomes small relative to the early time drawdown shortly (100 to 1,000 min) after the beginning of the test. The extrapolated drawdown will not be much different from the drawdown at the end of the test (see the analysis of the Northeast Kilohana monitor well aquifer test [fig. 11] for an example). But in an aquifer with a relatively low hydraulic conductivity, the drawdown per unit time can be high even after 10,000 min (7 days) of withdrawal so the extrapolated drawdown may vary significantly depending on how far into the future the extrapolation is made (the Hanamaulu monitor well aquifer test [fig. 7] is an example of this case).

Two values of steady-state drawdown were used for each test to get a range of possible hydraulic conductivity values using the Harr method. The drawdown was extrapolated to 1×10^4 minutes and 1×10^6 minutes (about 2 years) and equation 11 was solved for each case.

Aquifer-Test Analyses

For each well, the discussion includes the well location, details on the aquifer tests made at the well, information about the lithology penetrated by the well, and an analysis of the drawdown data using various aquifer-test methods described above.

Puakukui Springs Monitor Well Aquifer Test

The Puakukui Springs monitor well (State well 2-5626-01) is about 5.7 mi southwest of Lihue and about 0.4 mi north of Haupu Ridge (fig. 2). The well construction, well lithology, and constant-rate and step-drawdown aquifer-test data are documented in Gingerich and Izuka (1997b). The constant-rate aquifer test, which began December 4, 1995, lasted for 10,080 minutes at an average withdrawal rate of 40.9 ft^3/min (table 1). The withdrawal rate fluctuated by no more than 1 percent throughout the test. Well recovery was monitored for 8,940 minutes after withdrawal was stopped.

The open interval of the well is from 228 to -317 ft altitude (fig. 4) and the well penetrates basaltic lava flows intermixed with several 10- to 20-ft thick layers of basaltic cinders (Gingerich and Izuka, 1997b). On the basis of geochemical analysis of the drill cuttings, the well penetrates alluvium and underlying tholeiitic lavas of the Waimea Canyon Basalt (S.K. Izuka, U.S. Geological Survey, 1997, oral commun.). The well is within

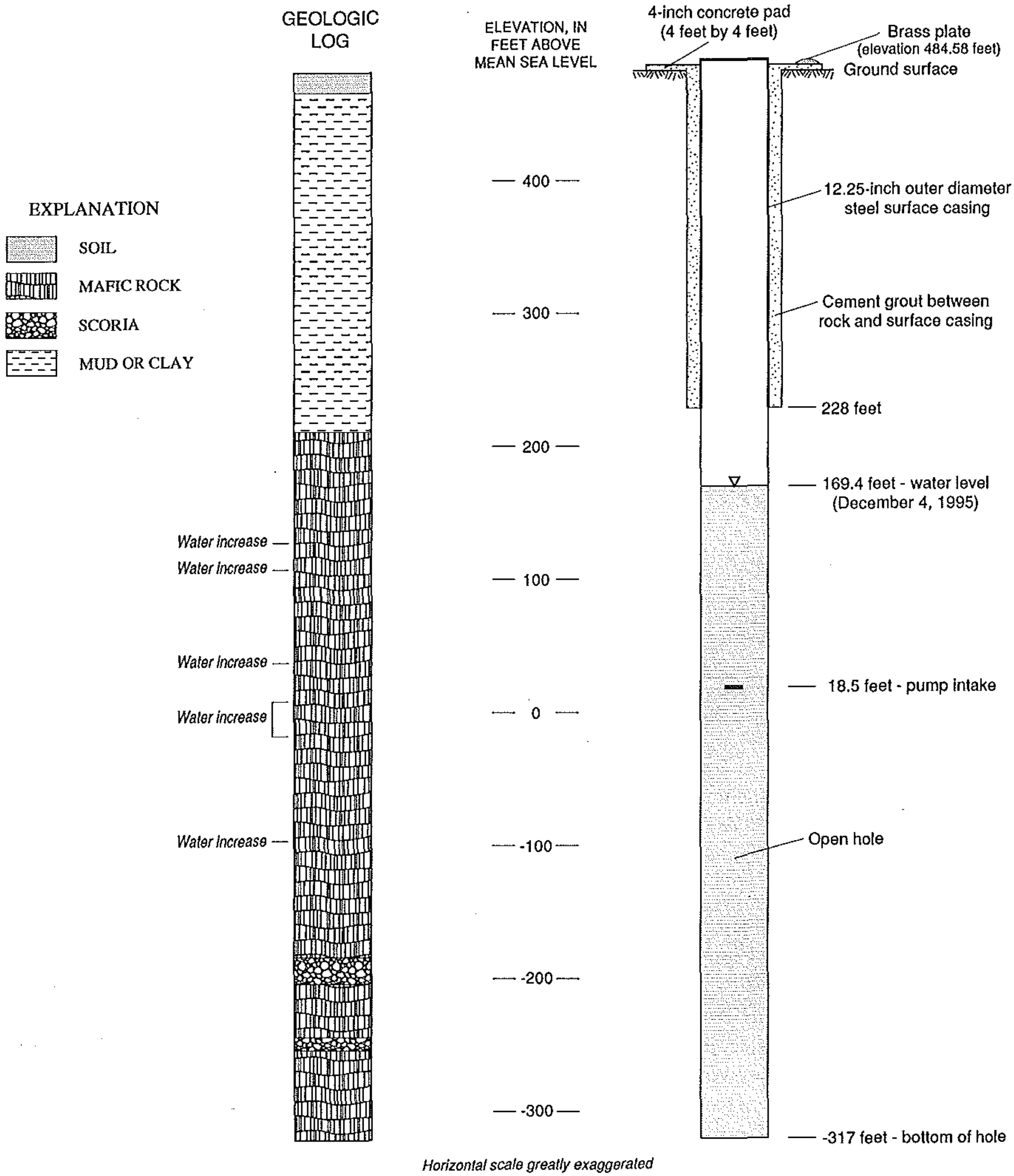


Figure 4. Geologic log and construction details of the Puakukui Springs monitor well (2-5626-01), Kauai, Hawaii.

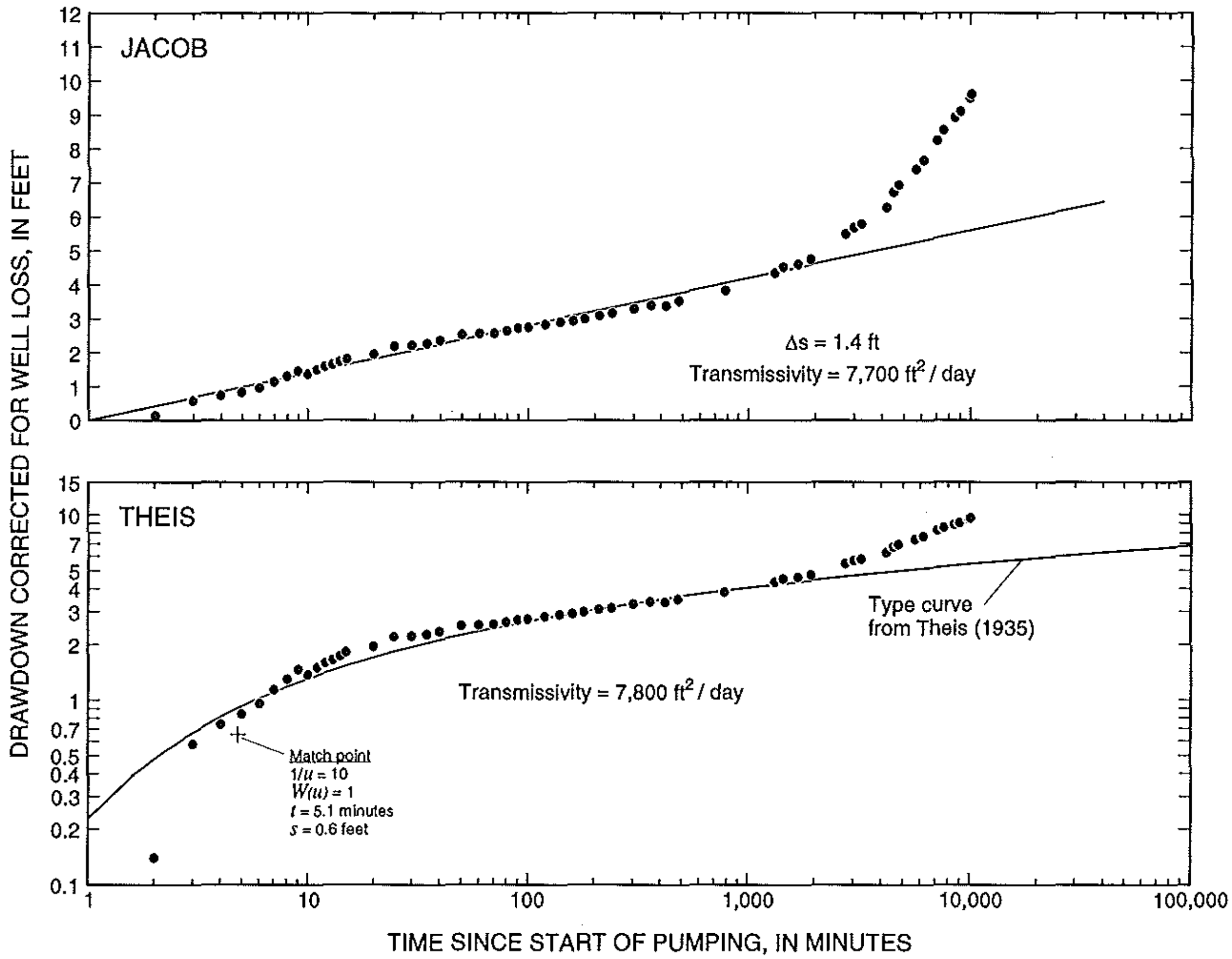


Figure 5. Analysis by Jacob and Theis methods of aquifer-test drawdown data for Puakukui Springs monitor well (2-5626-01), Kauai, Hawaii.

1,000 ft of at least five mapped volcanic dikes (Macdonald and others, 1960). Other dikes not visible at the ground surface are undoubtedly also within this distance. The water-table altitude at the start of the test was 169.4 ft and the aquifer was assumed to extend from the water table to the base of the well, a distance of 486 ft.

At the average withdrawal rate of 40.9 ft^3/min used in the constant-rate test, well loss was 3.6 ft (Gingerich and Izuka, 1997b). Well-loss corrections were applied to the drawdown data analyzed by using the Jacob, Theis, and Harr methods.

The well is assumed to have penetrated an unconfined aquifer; therefore, the data should be corrected using equation 3 to allow the application of a confined-aquifer method. The maximum aquifer drawdown in the well was about 9.6 ft. The correction for this drawdown, with an assumed aquifer thickness of 486 ft, is less than 0.1 ft. This correction was assumed to be insignificant (less than 1 percent of the aquifer drawdown) and the confined aquifer methods were applied with no correction made to the drawdown data.

The Jacob method requires a semi-log plot of drawdown against time for the constant-rate test (fig. 5). The slope of the straight line, Δs, fit through the drawdown data prior to 2,000 minutes of withdrawal, is 1.4 ft per log cycle. After 2,000 minutes, the effect of one or more boundaries is apparent in the data because the slope increases to about 8 ft per log cycle, six times the slope of the prior drawdown data. Multiple boundary effects are expected because the well is in an area that contains numerous mapped dikes. Using equation 4, the transmissivity estimate is 5.3 ft^2/min (7,700 ft^2/d). From equation 5 with r_w equal to 0.5 ft, the method is considered valid using data points later than 1.1 minutes after the beginning of withdrawal.

The Theis curve is fit to the data that fall between 3 and 2,000 minutes of withdrawal (fig. 5), after the effects of well-bore storage are dissipated and before the boundary effects are apparent. The transmissivity estimate for this type-curve match is 5.4 ft^2/min (7,800 ft^2/d).

The range of transmissivity from the two analysis methods is 7,700 to 7,800 ft^2/d (table 2). From equation 12 and with an assumed aquifer thickness of 486 ft, hydraulic conductivity is 15.8 to 16.1 ft/d.

Hydraulic conductivity values estimated using the Harr method are obtained from applying equation 11 to the Puakukui Springs monitor well data (table 1). The expected drawdown after 1×10^4 minutes and 1×10^6 minutes was estimated by extending the straight line fit through the first 2,000 minutes of the well-loss-corrected drawdown data (fig. 5). The resulting hydraulic conductivity estimates are 25 ft/d and 17 ft/d, respectively (table 2).

Hanamaulu Monitor Well Aquifer Test

The Hanamaulu monitor well (State well 2-5923-08) is about 1.5 mi northwest of Lihue and about 2.8 mi east of Kilohana Crater (fig. 2). Details of the well construction, lithology, step-drawdown and constant-rate aquifer-test data are in Izuka and Gingerich (1997a). The constant-rate test, which began October 6, 1995, lasted for 2,880 minutes (2 days) at a average withdrawal rate, of 10.2 ft^3/min (table 1). The withdrawal rate decreased by about 20 percent over the length of the test. The aquifer test was terminated when the water level in the pumped well approached the pump intake. Recovery data were collected for 1,260 minutes after withdrawal stopped.

The Hanamaulu well has a slotted screen and gravel pack between 148 and -730 ft altitude (fig. 6). The screened interval of the well penetrates lava flows of the Koloa Volcanics, alluvium, and marine mud and gravel deposits (Reiners, P.K., and others, Univ. of Washington, written commun., 1997). The water-table altitude was 213.5 ft at the time of the aquifer test. The aquifer thickness was assumed to be 943 ft, the distance from the water table to the base of the well.

For the average withdrawal rate used in the constant-rate test, well loss is estimated to be 12 ft (Izuka and Gingerich, 1997a). Well-loss corrections were made to the drawdown data before the Jacob, Theis, and Harr methods were applied. When well loss was removed from the measured data, the corrected drawdown was negative until 25 minutes into the test, indicating well loss and well-bore storage were significant factors during the early part of the test.

The slope of the straight line through the drawdown data, Δs, is 64.0 ft per log cycle using the Jacob method (fig. 7). A correction for unconfined conditions was not made because the maximum correction was only about 5 ft, an amount considered insignificant for this analysis. Using equation 4, the transmissivity estimate from the Jacob method is 0.029 ft^2/min (42 ft^2/d). The method is considered valid using data points after 25 minutes of withdrawal on the basis of equation 5 with r_w equal to 0.17 ft.

The match of the Theis type curve to the well-loss-corrected data for the Hanamaulu monitor well provides an estimate for transmissivity of 0.031 ft^2/min (40 ft^2/d) (fig. 7).

The Theis recovery method is applied directly to observed recovery data without a correction for well loss. The slope of the best-fit line is 61.1 ft per log cycle which is entered into equation 4 along with the average withdrawal rate producing a transmissivity estimate of 0.030 ft^2/min (44 ft^2/d) (fig. 7). On the basis of equation 10, this method is valid for t' greater than 473 minutes or t/t' less than 7. At least 10 data points are after this point, so the effects of well-bore storage should have dissipated and the method can be considered valid.

The range of transmissivity on the basis of the three analysis methods is 40 to 44 ft^2/d (table 2). For the assumed aquifer thickness of 943 ft, hydraulic conductivity ranges from 0.042 to 0.047 ft/d.

Equation 11 is applied to obtain a hydraulic conductivity estimate using the Harr method. The expected drawdown was estimated by extending the best straight-line fit through the well-loss-corrected drawdown data between 200 and 2,000 minutes after the start of withdrawal (fig. 7, Jacob). The resulting hydraulic conductivity estimates are 0.18 ft/d and 0.089 ft/d, using the extrapolated drawdown estimated at 1×10^4 minutes and 1×10^6 minutes, respectively (table 2).

Table 2. Summary of aquifer-test results, Kauai, Hawaii

[ft^2/d, feet squared per day; ft/d, feet per day; ft, feet; --, no analysis done]

Pumped-well State well number	Pumped well name	Transmissivity, *T*, (ft^2/d), calculated from curve-matching methods					Range of equivalent hydraulic conductivity, *K*, from curve-matching methods (ft/d)	Harr (1962)/Polubarinova-Kochina method: Estimated steady-state drawdown, s_s (ft)		Harr (1962)/Polubarinova-Kochina method: Hydraulic conductivity, *K* (ft/d)		Storage coefficient, *S*
		Moench[a] (1985) method	Jacob[b] (1946) method	Theis[c] (1935) method	Neuman[d] (1974) method	Theis[e] (1935) recovery method		Extra-polated to 1×10^4 minutes	Extra-polated to 1×10^6 minutes	Extra-polated to 1×10^4 minutes	Extra-polated to 1×10^6 minutes	
2-5626-01	Puakukui Springs	--	7,700	7,800	--	--	15.8–16.1	5.4	8.2	25	17	--
2-5923-08	Hanamaulu	--	42	40	--	44	0.042–0.047	142	270	0.18	0.089	--
2-0023-01	Pukaki Reservoir	420	730	680	--	--	0.43–0.75	97	125	0.70	0.54	8.2×10^{-2}
2-0124-01	Northeast Kilohana	810	1,200	--	--	1,400	0.9–1.5	40	58	5.5	3.8	1.6×10^{-2}
2-0124-02	Hanamaulu 3	--	--	--	1,050–1,450	1,080–1,260	1.1–1.6	--	--	--	--	--
2-0126-01	Northwest Kilohana	220	200	--	--	--	0.22–0.24	184	294	0.52	0.32	8.5×10^{-4}

[a] equation 1 in text

[b] equation 4 in text

[c] equation 6 in text

[d] equation 7 in text

[e] equation 4 in text

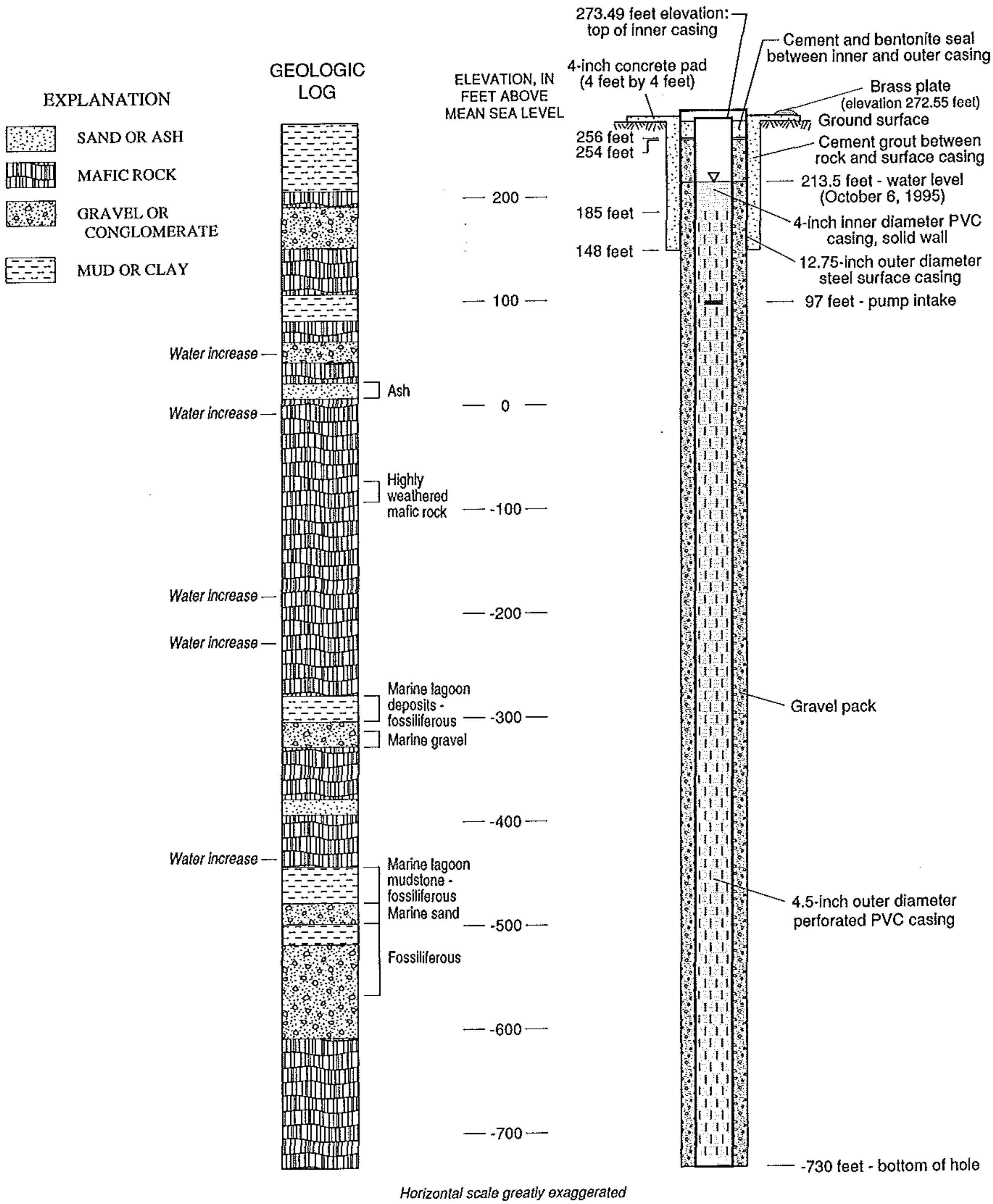


Figure 6. Geologic log and construction details of the Hanamaulu monitor well (2-5923-08), Kauai, Hawaii.

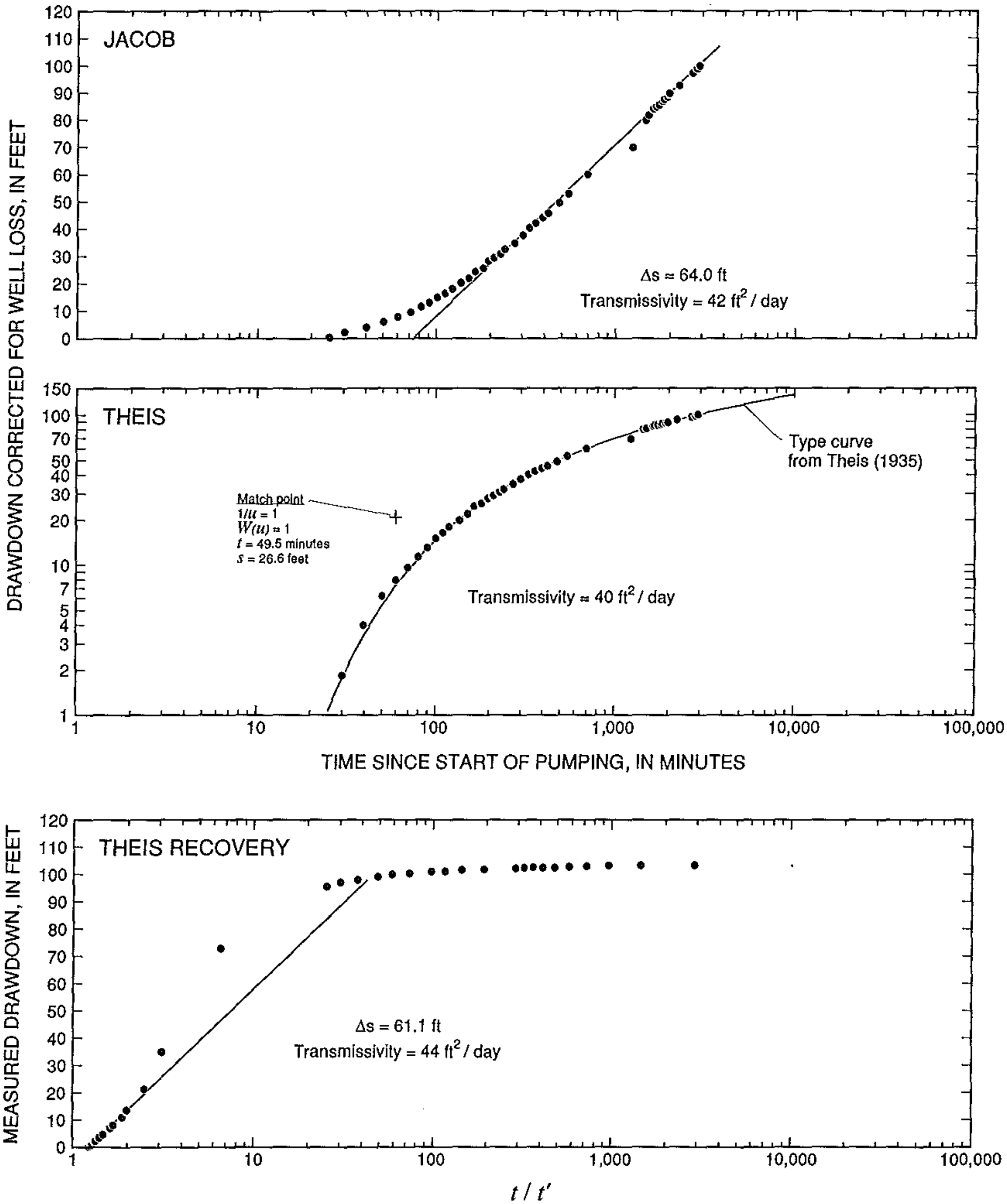


Figure 7. Analysis by Jacob, Theis, and Theis recovery methods of aquifer-test drawdown data for Hanamaulu monitor well (2-5923-08), Kauai, Hawaii.

Pukaki Reservoir Monitor Well Aquifer Test

The Pukaki Reservoir monitor well (State well 2-0023-01) is about 2.5 mi northwest of Lihue and about 2.8 mi northeast of Kilohana Crater (fig. 2). Details of the well construction, the lithology penetrated by the well, constant-rate and step-drawdown aquifer-test data are described in Izuka and Gingerich (1997c). The constant-rate aquifer test, which began April 2, 1996, lasted for 5,700 minutes (4 days) at an average withdrawal rate of 38.0 ft^3/min (table 1). Withdrawal rates fluctuated by less than 4 percent during most of the constant-rate test. But after 3,400 minutes into the constant-rate test, withdrawal stopped for 120 minutes because of a failure in the generator powering the pump, causing a momentary recovery in the well until the pump was restarted. The well recovery was monitored for only 270 minutes after withdrawal was stopped at the end of the test.

The well has an open interval from 163 to -828 ft altitude and the water-table altitude was 147.4 ft at the time of the aquifer test (fig. 8). The well penetrates a thick sequence of lava flows of the Koloa Volcanics, alluvial layers, marine sediments and lava flows of the top of the Waimea Canyon Basalt (Reiners, P.K., and others, Univ. of Washington, written commun., 1997). The aquifer thickness was assumed to be 975 ft, the distance from the water table to the base of the well.

Analysis of the step-drawdown data was attempted by Izuka and Gingerich (1997c) but suitable results were unobtainable. For an unknown reason, analysis of the step-drawdown data produced a calculated well-loss coefficient that was negative. Therefore, well-loss corrections were applied assuming the measured drawdown in the well after 1 minute of withdrawal was equivalent to the well loss. Well loss calculated from step-drawdown tests for the other four wells investigated in this study (table 1) averaged about 106 percent of the drawdown measured after 1 minute of withdrawal so this approximation should be acceptable. Therefore, a well-loss correction of 50 ft was subtracted from the drawdown data analyzed using the Jacob and Harr methods.

The match between the observed drawdown data measured in the well and the type curve generated using the Moench method is reasonable (fig. 9). The coefficients used to generate the type curve are as follows:

$$\gamma' = 0.001, \sigma' = 30, \gamma'' = 0.002, \sigma'' = 30, \text{ and } W_D = 5{,}000.$$

Entering the match-point values into equations 1 and 2 produces transmissivity and aquifer storage coefficient estimates of 0.29 ft^2/min (420 ft^2/d) and 8.2×10^{-2}, respectively (table 2).

If the well is assumed to be in an unconfined system, the drawdown data should be corrected using equation 3 to allow the use of the Jacob method. The maximum well-loss-corrected drawdown in the well was about 99 ft and the correction for this drawdown assuming an aquifer thickness of 975 ft is only about 5 ft. This correction was assumed to be insignificantly related to the total drawdown and the method was applied with no correction made to the drawdown data. Using the Jacob method, the slope of the straight line through the drawdown data, Δs, is 13.8 ft per log cycle (fig. 9). Using equation 4, the transmissivity estimate is 0.51 ft^2/min (730 ft^2/d). The method is considered valid using data points after 8.7 minutes of withdrawal.

The match between the Theis type curve and the well-loss-corrected drawdown data provides a transmissivity estimate of 0.47 ft^2/min (680 ft^2/d) (fig. 9). The type curve was fit to the drawdown data after 9 minutes of withdrawal.

The range of transmissivity on the basis of the three analysis methods is 420 to 730 ft^2/d (table 2). For the assumed aquifer thickness of 975 ft, hydraulic conductivity ranges from 0.43 to 0.75 ft/d and the storage coefficient is 0.08. The methods used to analyze this well seem to be inconsistent because two methods require the system to be unconfined and one method assumes a semi-confined aquifer with sources above and below. Because of the many uncertainties about the configuration of aquifers penetrated by the well, it is not possible to say which assumption is most like the actual field situation. But the range of transmissivity estimates from the different methods indicates that a reliable estimate of the aquifer transmissivity can be obtained without knowing all of the specific details of the aquifer geometry.

The estimates of expected drawdown after 1×10^4 minutes and 1×10^6 minutes needed in equation 11 were determined by extending the best-fit line though the well-loss-corrected drawdown data between 5 minutes and about 5,000 minutes (fig. 9, table 2). The resulting hydraulic conductivity estimates are 0.70 ft/d and 0.54 ft/d.

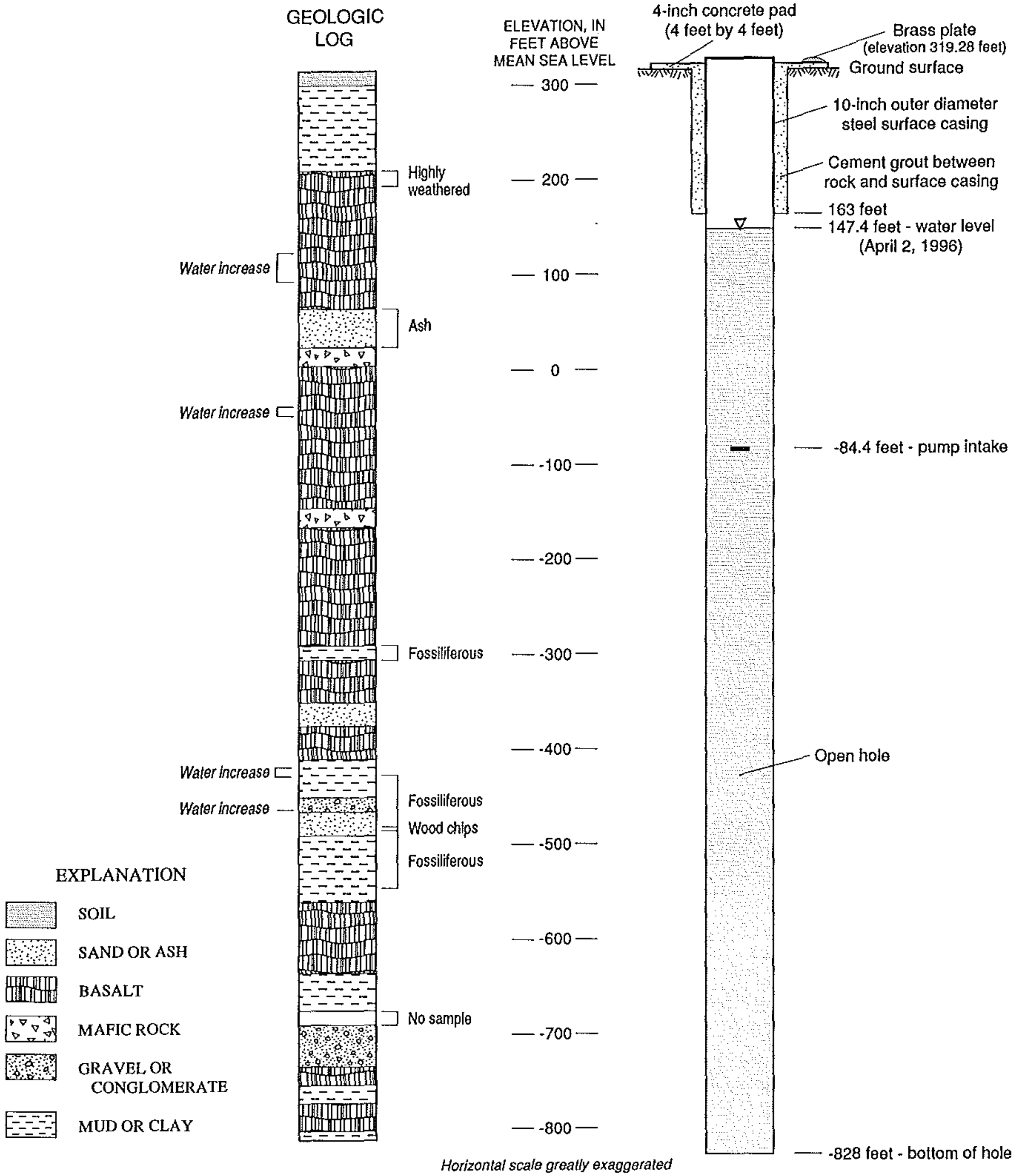


Figure 8. Geologic log and construction details of the Pukaki Reservoir monitor well (2-0023-01), Kauai, Hawaii.

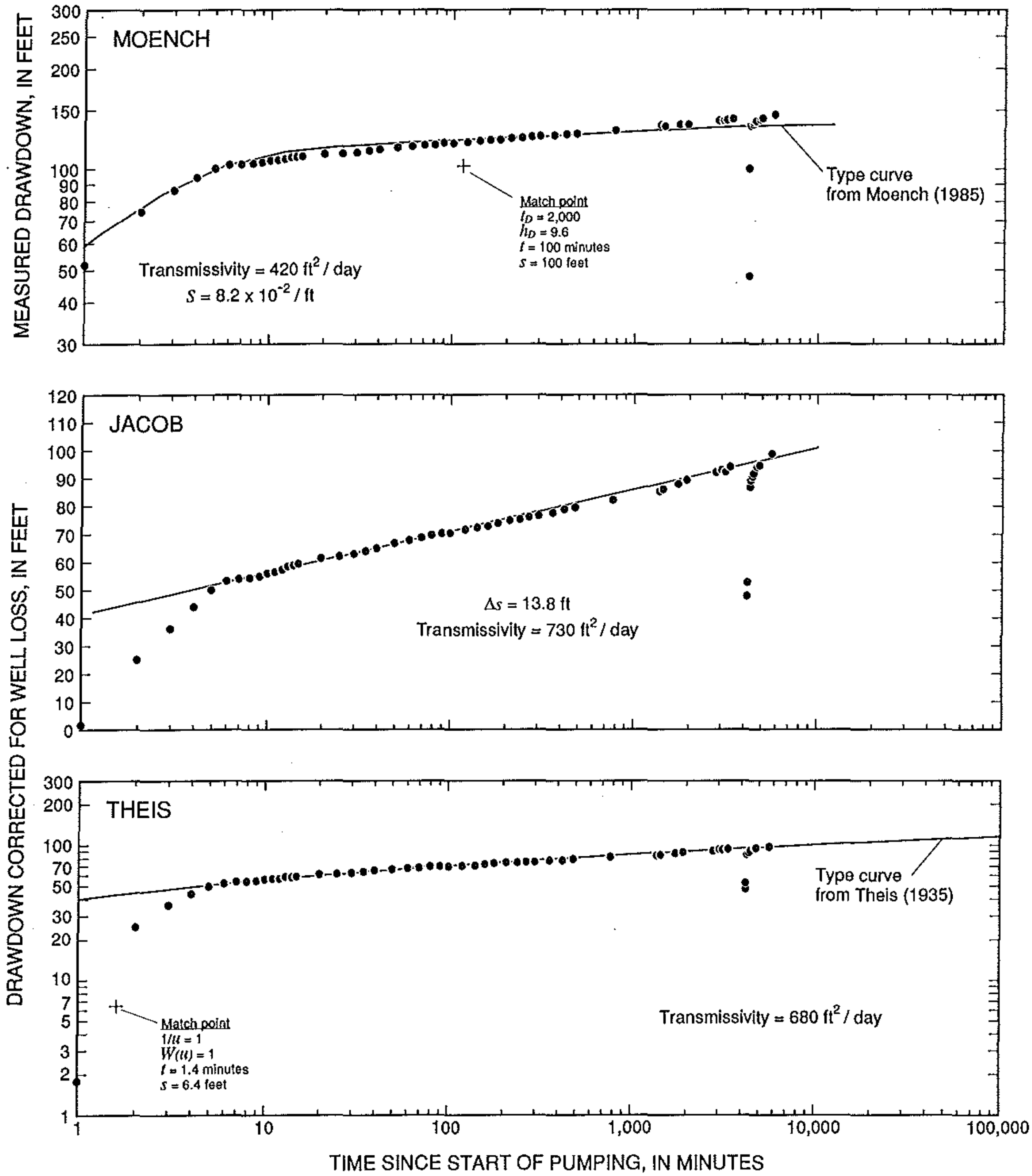


Figure 9. Analysis by Moench, Jacob, and Theis methods of aquifer-test drawdown data for Pukaki Reservoir monitor well (2-0023-01), Kauai, Hawaii.

Northeast Kilohana Monitor Well Aquifer Tests

The Northeast Kilohana monitor well (State well 2-0124-01) is about 3.7 mi northwest of Lihue and about 2.3 mi northeast of Kilohana Volcano Crater (fig. 2). The details of the well construction, the lithology penetrated by the well and the aquifer tests are provided in Izuka and Gingerich (1997b). A production well (Hanamaulu 3, State well 2-0124-02) was drilled in 1998 about 146 ft southwest of Northeast Kilohana monitor well. Step-drawdown and single-well constant-rate tests were made on the Northeast Kilohana monitor well before the production well was drilled. After Hanamaulu 3 was drilled, several step-drawdown and constant-rate aquifer tests were done and the drawdown was measured at the Northeast Kilohana monitor well (unpub. data, U.S. Geological Survey, Hawaii District aquifer-test archive).

The single-well aquifer test was made before the Northeast Kilohana well was drilled to its final depth. At the time of the test, the well had an open interval from 306 to 0 ft altitude (fig. 10) and the water-table altitude was 374.9 ft. The open interval of the well penetrated a sequence of lava flows of the Koloa Volcanics, alluvial layers, and an ash layer (Izuka and Gingerich, 1997b; Reiners, P.K., and others, Univ. of Washington, written commun., 1997). The single-well constant-rate test, which began July 1, 1995, lasted for 10,080 minutes (7 days) at an average withdrawal rate of 42.3 ft^3/min (table 1). The withdrawal rate fluctuated by less than 5 percent during the test, and recovery data were collected for 300 minutes after withdrawal stopped.

The well-loss coefficient for the Northeast Kilohana monitor well was 8.03×10^{-3} min^2/ft^5 and at the average withdrawal rate (42.3 ft^3/min) the well loss is estimated to be 14 ft (table 1) (Izuka and Gingerich, 1997b). The well-loss corrections were subtracted from the single-well-test drawdown data analyzed using the Jacob and Theis methods.

After about 6,000 minutes of withdrawal during the single-well aquifer test, the drawdown stopped increasing, indicating that a source of water to the aquifer began to influence the test (fig. 11, Moench). One possible explanation for this is that the pumped aquifer is semi-confined, and an overlying and/or underlying bed is supplying water through the semi-confining unit(s) (fig. 3) although there is little data to confirm or deny this explanation. The Moench method is appropriate for this situation and the type-curve fit to the data is reasonable (fig. 11). The coefficients used to generate the type curve are as follows:

$$\gamma' = 0.019,\ \sigma' = 95,\ \gamma'' = 0.009,\ \sigma'' = 95,\text{ and } W_D = 160.$$

Entering the match-point values (fig. 11) into equations 1 and 2 produce transmissivity and aquifer storage-coefficient estimates equal to 0.56 ft^2/min (810 ft^2/d) and 1.6×10^{-2}, respectively.

The correction for an unconfined aquifer using equation 3 is about 1.6 ft and was assumed to be insignificant; thus the confined-aquifer method was applied with no correction made to the single-well-test drawdown data. The Jacob method straight line is fit through the drawdown data prior to about 6,000 minutes of withdrawal, when the possible water-source effects are first apparent (fig. 11). The slope of the straight line through the drawdown data, Δs, is 9.0 ft per log cycle. Using equation 4, the transmissivity estimate is about 0.83 ft^2/min (1,200 ft^2/d). On the basis of equation 5 with r_w equal to 0.42 ft, the method is considered valid using data points after 4.8 minutes of withdrawal.

For the Theis recovery method, the slope of the best-fit line is 8.0 ft per log cycle which is entered into equation 4 along with the average withdrawal rate to produce an transmissivity estimate equal to about 0.97 ft^2/min (1,400 ft^2/d) (fig. 11). On the basis of equation 10, this method is valid for t' greater than 129 minutes or t/t' less than 80. Because there are only three data points with values less than t/t' = 80, the results could be considered questionable but the resulting value for transmissivity compares favorably with the results from the other two methods.

The range of transmissivity on the basis of the three analysis methods of the single-well test data is 810 to 1,400 ft^2/d (table 2). For an assumed aquifer thickness of 924 ft (the distance from the water table to the final depth of the well), hydraulic conductivity ranges from 0.9 to 1.5 ft/d.

For the Harr method, the drawdown was estimated after 7 days and 2 years on the basis of drawdown of 9.0 ft/log cycle. The resulting hydraulic conductivity estimates from equation 11 are 5.5 ft/d and 3.8 ft/d, respectively.

Hanamaulu 3 was drilled to an altitude of -94 ft, screened between 336 ft and 196 ft altitude, and is an open hole below 196 ft altitude (fig. 12). At the time of

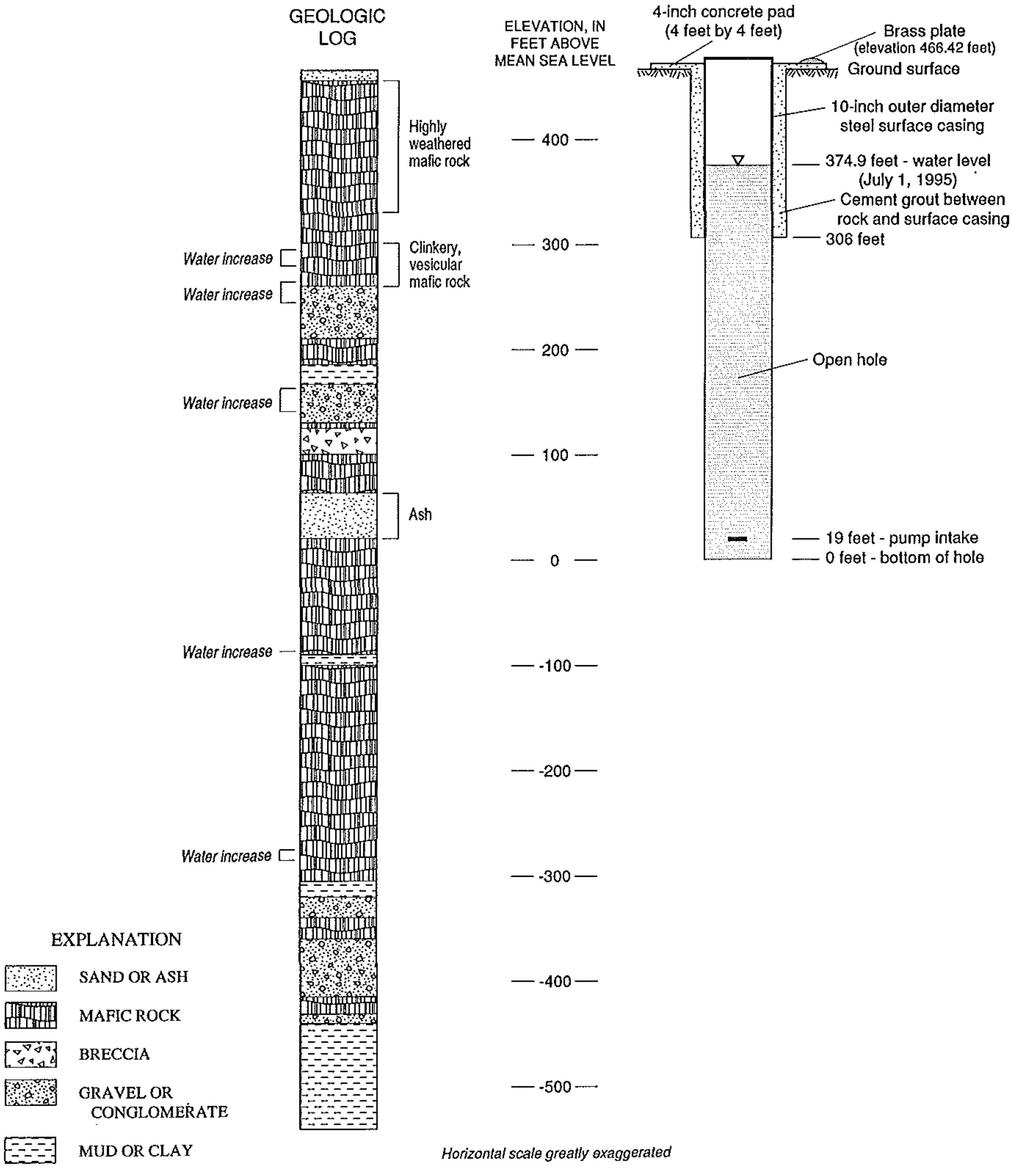


Figure 10. Geologic log and construction details of the Northeast Kilohana monitor well (2-0124-01), Kauai, Hawaii.

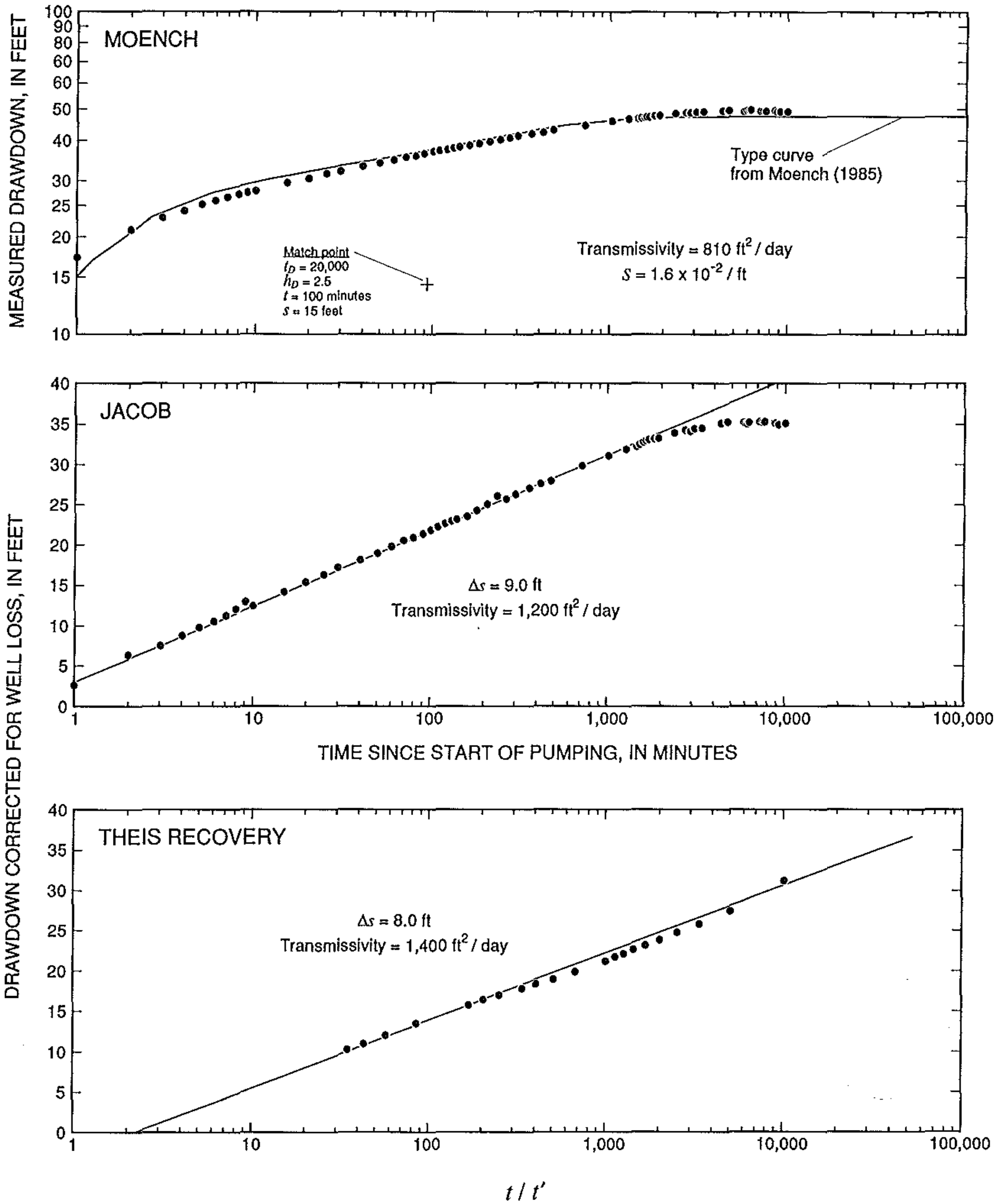


Figure 11. Analysis by Moench, Jacob, and Theis recovery methods of aquifer-test drawdown data for Northeast Kilohana monitor well (2-0124-01), Kauai, Hawaii.

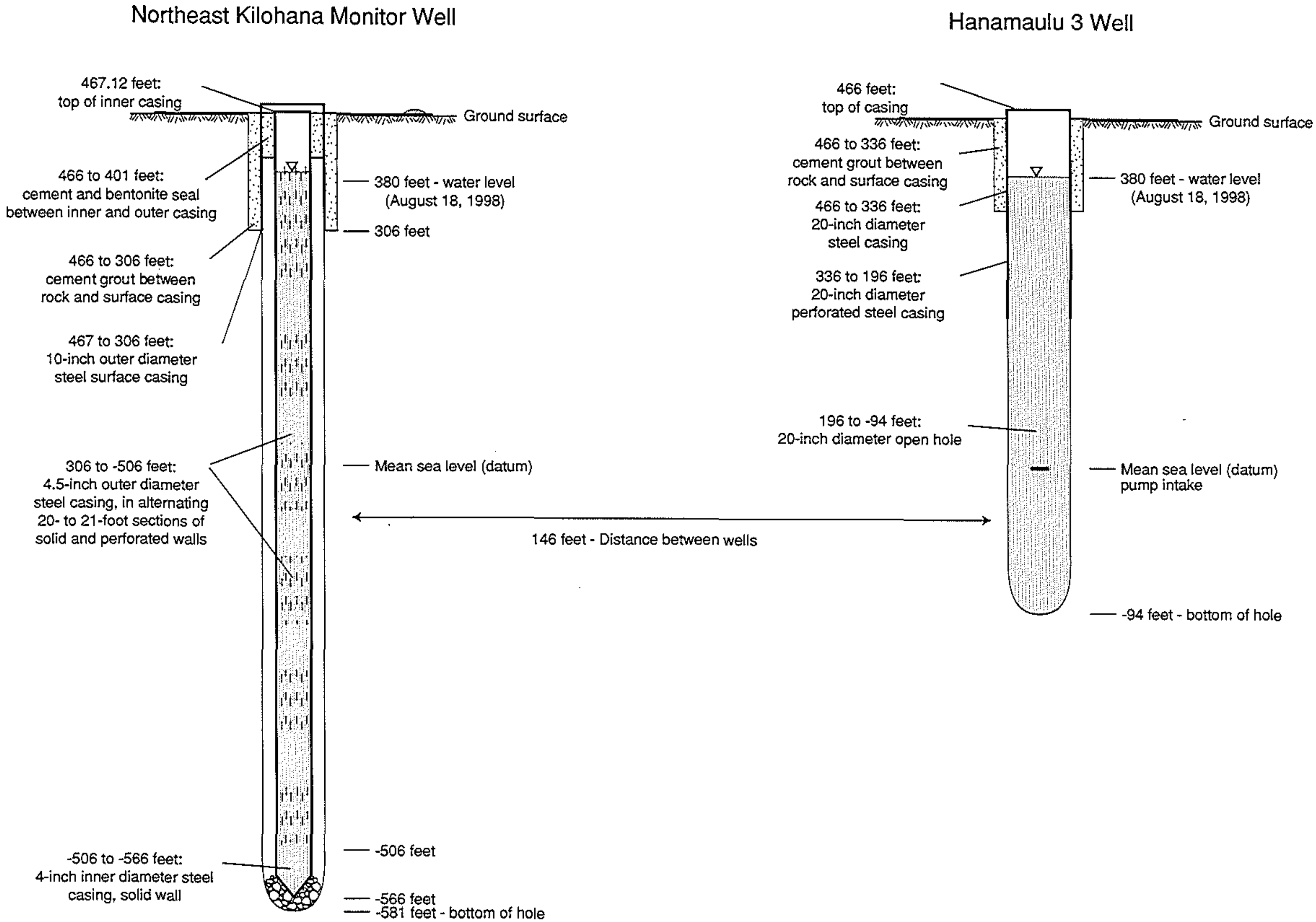


Figure 12. Construction details of the Northeast Kilohana monitor well (2-0124-01) and the Hanamaulu 3 well (2-0124-02), Kauai, Hawaii.

the aquifer tests using this well, the Northeast Kilohana monitor well was screened between 386 ft and -506 ft altitude (fig. 12). The analyses of these tests where an observation well is present is useful to compare to the results of the single-well aquifer tests.

Three aquifer tests were analyzed using the Neuman (1974) method for unconfined aquifers exhibiting delayed gravity yield. These analyses were made with the aid of a commercial software package, Aqtesolv®, which has the capability of analyzing aquifer tests having variable withdrawal rates. Estimates of transmissivity using the Neuman method range from 0.73 to 1.0 ft^2/min (1,050 to 1,450 ft^2/d) for these multiple-well tests (fig. 13). Specific yield estimates, ranging from 1.3×10^{-4} to 3.2×10^{-4}, are unreasonably small and are not considered valid.

Two of the three multiple-well aquifer tests had recovery data which were analyzed using the Theis recovery method (fig. 13). From these analyses, transmissivity ranged from 0.75 to 0.88 ft^2/min (1,080 to 1,260 ft^2/d). Using equation 9, the lowest T, and the highest S estimated, these results are valid for t/t' less than 110.

The hydraulic conductivity estimated from the multiple-well tests ranges from 1.1 to 1.6 ft/d when the aquifer thickness is assumed to be 924 ft. The single-well tests results are similar for hydraulic conductivity (0.9 to 1.5 ft/d) but are several orders of magnitude too high for estimates of storage coefficient.

Northwest Kilohana Monitor Well Aquifer Test

The Northwest Kilohana monitor well (State well 2-0126-01) is about 5.3 mi northwest of Lihue and about 1.5 mi northwest of Kilohana Volcano Crater (fig. 2). Details of the well construction, lithology, step-drawdown and constant-rate aquifer tests can be found in Gingerich and Izuka (1997a). The constant-rate test, which began January 24, 1996, lasted for 10,080 minutes (7 days) at an average withdrawal rate, Q, of 41.8 ft^3/min (table 1). The withdrawal rate decreased by as much as 17 percent from the beginning of the test until the end. Recovery data were collected for 10,080 minutes after withdrawal stopped.

The well, with an open interval from 480 to -326 ft altitude (fig. 14), penetrates lava flows of the Koloa Volcanics, Palikea Breccia Member debris-flow deposits, and tholeiitic lava flows below the Koloa Volcanics (Reiners, P.K., and others, Univ. of Washington, written commun., 1997). The water-table altitude at the time of the test was 590.3 ft and the aquifer thickness was assumed to be 916 ft.

Analysis of the step-drawdown data provided a well-loss coefficient of 3.22×10^{-2} min^2/ft^5 (Gingerich and Izuka, 1997a). For the constant-rate-test withdrawal rate of 41.8 ft^3/min, well loss is estimated to be 56 ft. The well-loss corrections were applied to the drawdown data analyzed using the Jacob method.

The drawdown data using the Moench method display similar properties as the Northeast Kilohana monitor well test and a possible source of water to the aquifer is apparent after about 3,000 minutes of withdrawal (fig. 15). The coefficients used to generate the type curve matched to the observed drawdown data are as follows:

$$\gamma' = 0.01,\ \sigma' = 300,\ \gamma'' = 0.02,\ \sigma'' = 300, \text{ and } W_D = 500.$$

The analysis assumes the aquifer, which is 916 ft thick, is bounded above and below by semi-confining units separating the aquifer from units which act as constant sources of water (fig. 3).

Entering the match-point values (fig. 15) into equations 1 and 2 produces transmissivity and storage coefficient estimates equal to 0.15 ft^2/min (220 ft^2/d) and 8.5×10^{-4}, respectively.

If the well is assumed to be in an unconfined aquifer, the data should be corrected using equation 3 to allow the use of a confined-aquifer method. The maximum well-loss-corrected drawdown in the well was about 158 ft and the correction for this drawdown assuming an aquifer thickness of 916 ft is only about 14 ft. This correction was assumed to be insignificant relative to the total drawdown and the confined-aquifer method was applied with no correction made to the drawdown data.

For the Jacob method, the slope of the straight line through the drawdown data, Δs, is 55 ft per log cycle (fig. 15). Only data between about 200 minutes and 3,000 minutes of withdrawal were used in fitting the best-fit straight line. Using equation 4, the transmissivity estimate is 0.14 ft^2/min (200 ft^2/d). On the basis of equation 5 with r_w equal to 0.42 ft, the method is considered valid using data points after 32 minutes of withdrawal.

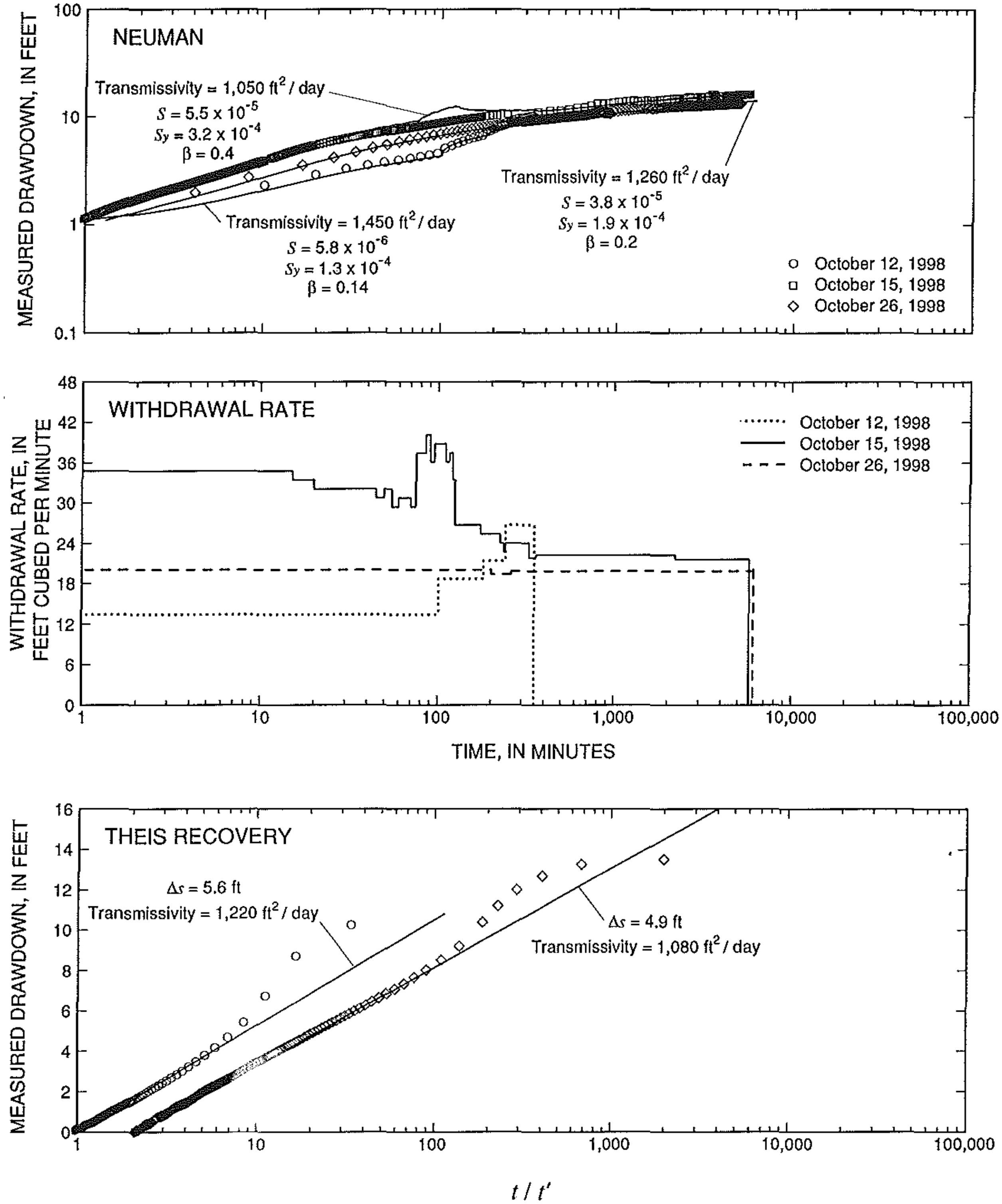


Figure 13. Analysis by Neuman and Theis recovery methods of aquifer-test drawdown data for Northeast Kilohana monitor well (2-0124-01) during aquifer test at Hanamaulu 3 well (2-0124-02), Kauai, Hawaii.

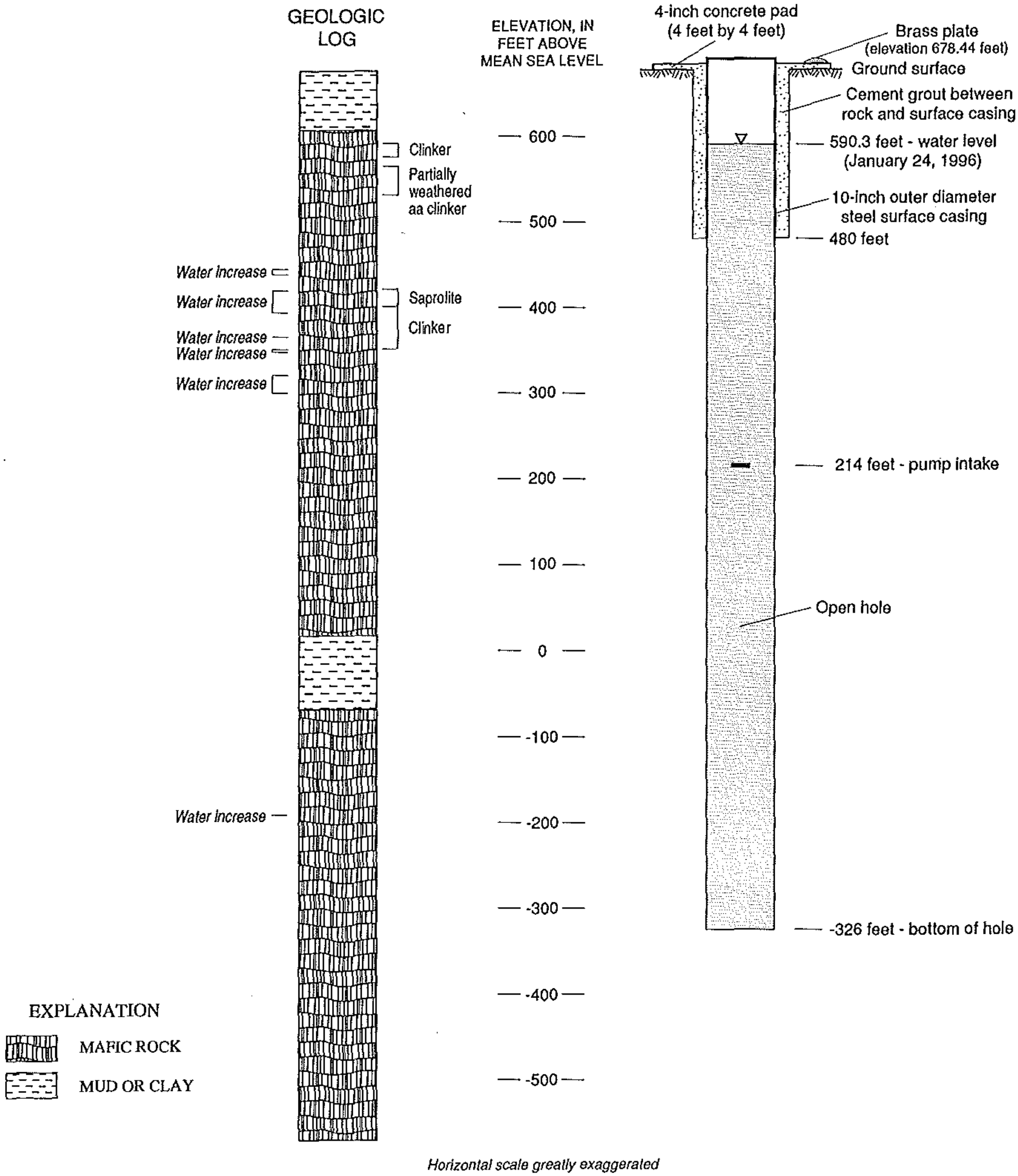


Figure 14. Geologic log and construction details of the Northwest Kilohana monitor well (2-0126-01), Kauai, Hawaii.

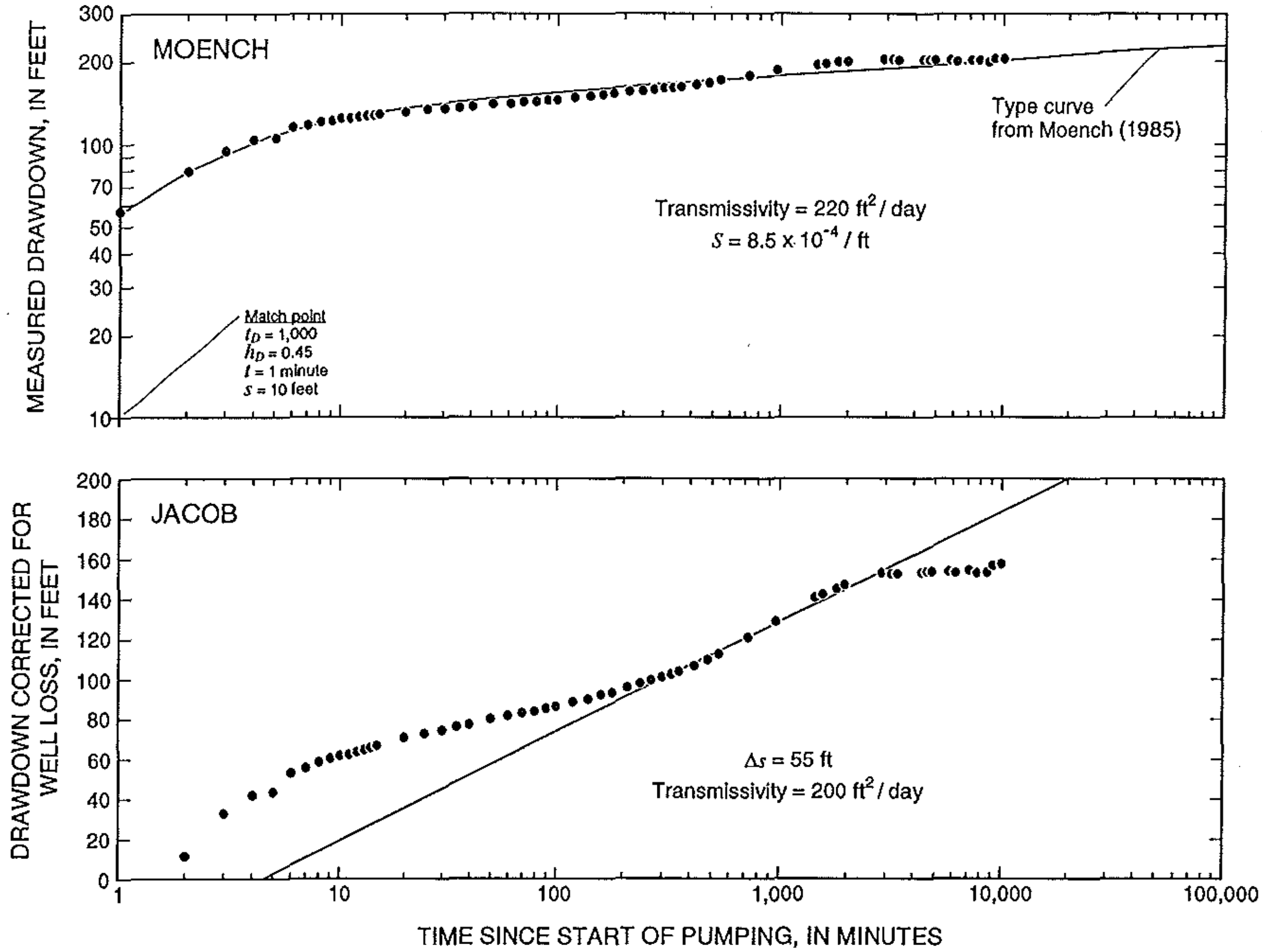


Figure 15. Analysis by Moench and Jacob methods of aquifer-test drawdown data for Northwest Kilohana monitor well (2-0126-01), Kauai, Hawaii.

The range of transmissivity from the two analysis methods is 200 to 220 ft^2/d (table 2). For the assumed aquifer thickness of 916 ft, hydraulic conductivity ranges from 0.22 to 0.24 ft/d.

The straight line used in the Jacob method to estimate the expected values of drawdown in the Northwest Kilohana monitor well after 1×10^4 minutes and 1×10^6 minutes fits the data between about 300 and 3,000 minutes (fig. 15). Applying equation 11 provides hydraulic conductivity estimates of 0.52 ft/d and 0.32 ft/d for the two estimates of expected drawdown using the Harr method (table 2).

DISCUSSION OF METHODS AND RESULTS

The results of the above analyses confirm the importance of using a variety of methods to estimate the transmissivity of a thick basaltic aquifer. The results demonstrate that transmissivity in the Lihue basin ranges over three orders of magnitude. The range of transmissivity and hydraulic conductivity values estimated from the different aquifer-test methods for each individual well is generally small. The largest range is for the Northeast Kilohana monitor well where the highest estimate is about 23 percent greater than the average of all of the estimates for this well. For the dikeity estimates are about one to four orders of magnitude lower than other reported values for dike-free shield-building-stage basaltic aquifers in the State of Hawaii.

Results of aquifer tests in wells that penetrate the Waimea Canyon Basalt do not differ markedly from tests in wells that penetrate just the Koloa Volcanics or both the Koloa Volcanics and the underlying Waimea Canyon Basalt. The transmissivity estimated in aquifer tests in the study area is probably controlled mainly by the stratigraphic relationships and areal extent of the layers penetrated by the wells.

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