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(1) B 7806 Dr 4

LBL-~~12869~~ Revised
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A SIMPLE MODEL FOR FAULT-CHARGED
HYDROTHERMAL SYSTEMS

Gudmundur S. Bodvarsson, Constance W. Miller,
and Sally M. Benson

August 1981

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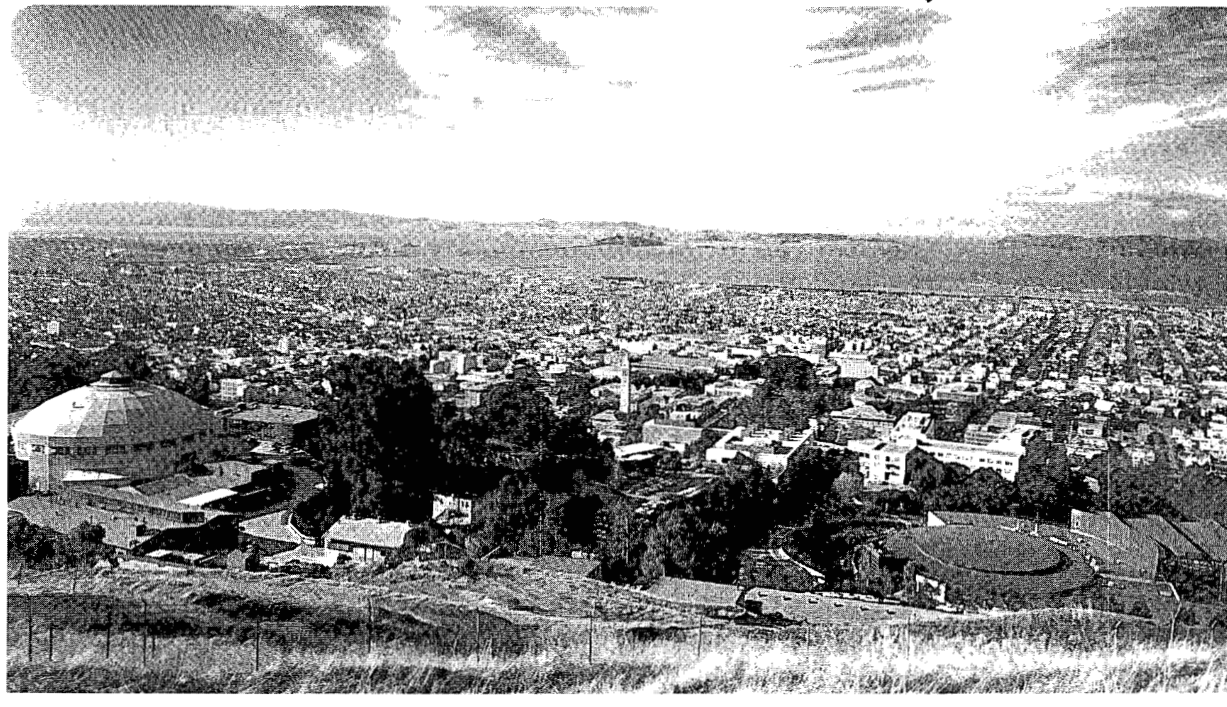

Prepared for the U.S. Department of Energy under Contract W-7405-ENG-48

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Printed in the United States of America
Available from
National Technical Information Service
U.S. Department of Commerce
5285 Port Royal Road
Springfield, VA 22161
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LBL-12869
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A SIMPLE MODEL FOR FAULT-CHARGED HYDROTHERMAL SYSTEMS

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This work was supported by the Assistant Secretary for Conservation and Renewable Energy, Office of Renewable Technology, Division of Geothermal and Hydropower Technologies of the U. S. Department of Energy under Contract No. W-7405-ENG-48.

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ABSTRACT

A two-dimensional transient model of fault-charged hydrothermal systems has been developed. The model can be used to analyze temperature data from fault-charged hydrothermal systems, estimate the recharge rate from the fault, and determine how long the system has been under natural development. The model can also be used for theoretical studies of the development of fault-controlled hydrothermal systems. The model has been tentatively applied to the low-temperature hydrothermal system at Susanville, California. A reasonable match was obtained with the observed temperature data, and a hot water recharge rate of 9×10^{-6} $m^3/s \cdot m$ was calculated.

INTRODUCTION

One of the most important tasks in geothermal reservoir engineering is to predict the useful lifetime of the resource for a given exploitation scheme. In order to make these predictions, reliable estimates must be available of the amount of hot water in place, the rate at which it can be extracted (transmissivity of the reservoir), and the rate and extent of hot water recharge into the system. The first two estimates can often be readily obtained from simple volumetric calculations and well-test analysis, respectively; reliable estimates of the recharge are often harder to get. This paper describes a simple model for calculating the rate of recharge into a fault-charged hydrothermal reservoir.

All geothermal reservoirs are controlled to some extent by faults and fractures; in some, however, a single fault or the intersection of two or more

major faults is believed to act as the main conduit for recharge. Examples of these include high-temperature systems such as Roosevelt Hot Springs, Utah; and East Mesa, California; and low-moderate temperature systems such as the Klamath Falls and Vale geothermal fields in Oregon and the Susanville system in California. This paper discusses the model developed for evaluating such systems and illustrates its applicability by estimating hot water recharge into the Susanville, California, geothermal resource--a shallow low-temperature hydrothermal system.

In contrast to the temperature logs from most geothermal wells, those from wells in fault-charged geothermal reservoirs often display anomalous behavior. One such profile, shown in Figure 1, was obtained from a well in the "steamer district" of the Klamath Falls KGRA (Benson and O'Brien, 1980). The profile shows the typical linear characteristics associated with conductive heat transfer in the top 200 ft, then a typical convective type profile down to 250 ft. At a depth of 250 ft the profile displays a definite reversal (below that the temperature profile reflects downflow in the well). One possible explanation for the behavior shown in Figure 1 is that a fault recharges an aquifer located at a depth of 200-250 ft below the ground surface. The relatively hot water travels up the fault until it intersects the permeable aquifer; it is then transported laterally in the aquifer. As the hot water moves through the aquifer, heat is lost mainly by conduction to the overlying and underlying strata. Variations in the temperature profiles between wells at different distances from the recharging fault can be used to estimate the recharge rate.

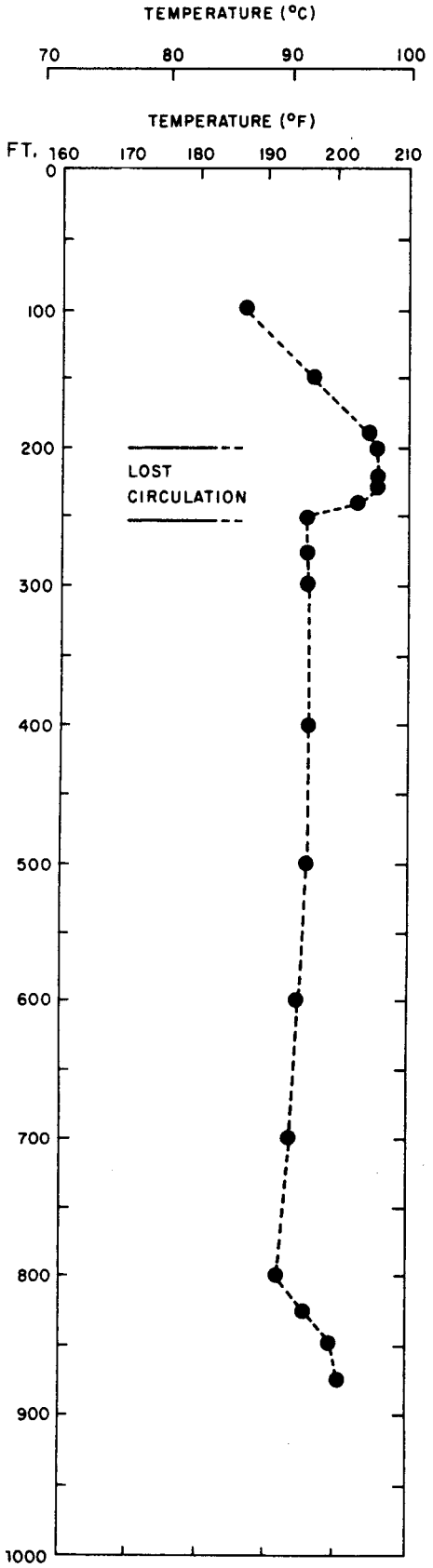

Figure 1. Temperature profile from a well at Klamath Falls, Oregon. [XBL 8011-2979]

Various mathematical models applied to fault-charged hydrothermal systems are cited in the literature. Kilty et al. (1978) and Goyal and Kassoy (1981) developed two-dimensional models (semi-analytic solutions) of the Monroe hydrothermal system, Utah, and the the East Mesa field, respectively. Sorey (1976) and Riney et al. (1979) applied numerical models to Long Valley Caldera and the East Mesa system, respectively.

In contrast to these models, the model presented here does not consider vertical temperature variations within the aquifer, but calculates the transient heat losses to the caprock and the bedrock. The model may therefore be quite useful in analyzing relatively young fault-charged thin hydrothermal systems, as well as in theoretical studies of the development of such systems.

MATHEMATICAL MODEL

Figure 2 shows the reservoir system on which the mathematical model is based. Hot water flows up the fault and feeds a shallow aquifer. The fault is shown by broken lines to illustrate that no heat losses are considered when the fluid is flowing up the fault. Initially the temperature in the system is linear with depth (normal geothermal gradient) as controlled by the constant-temperature boundaries at $z = D$ (ground surface) and $z = -H$. At time $t = 0$ hot water starts to flow into the reservoir at a temperature T_f. The primary assumptions employed are listed below:

1. The mass flow is steady in the aquifer, horizontal conduction is neglected, and temperature is uniform in the vertical direction (thin aquifer). Equilibrium between the fluid and the solids is instantaneous.

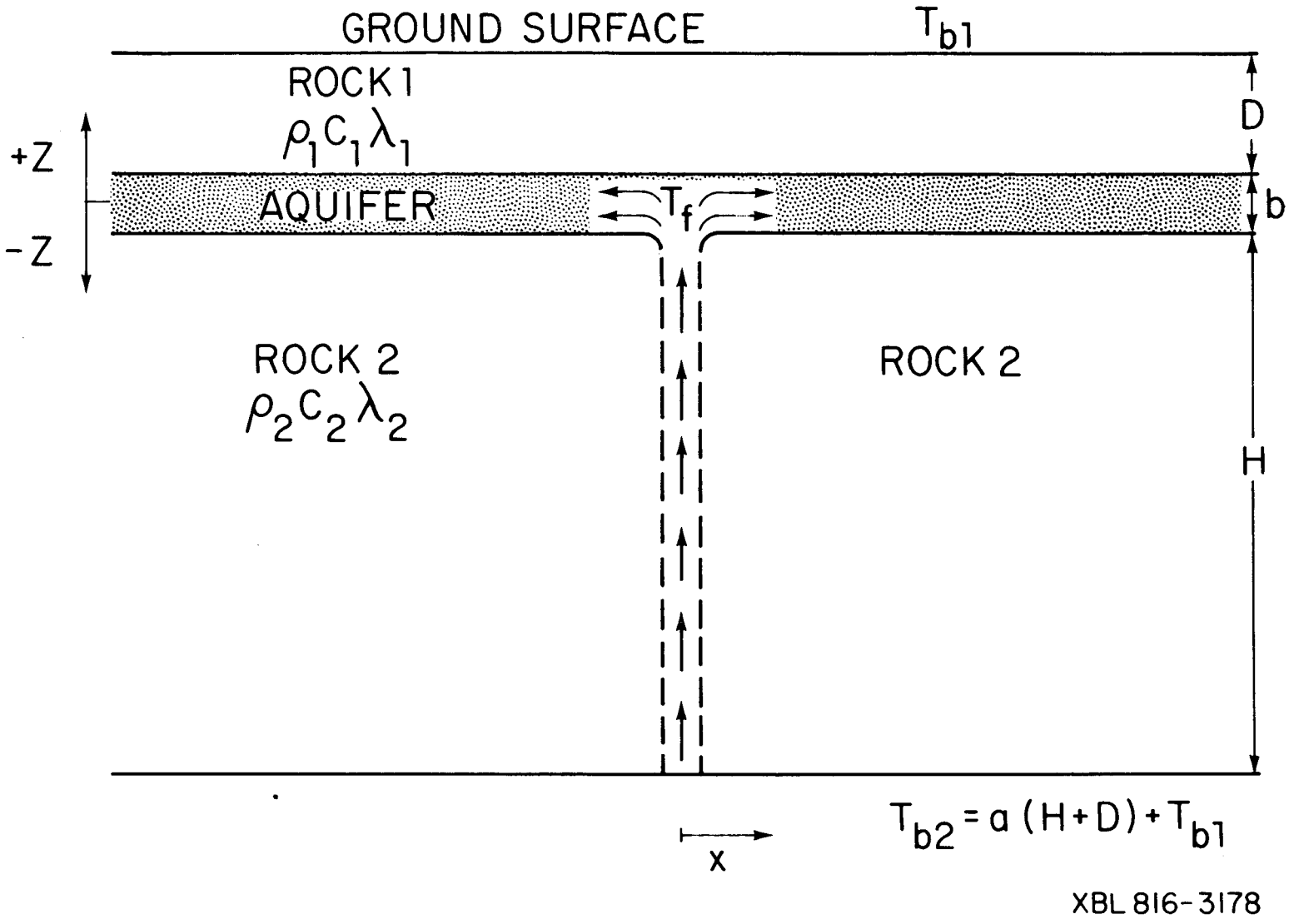

Figure 2. The mathematical model considered.

2. The rock matrix above and below the aquifer is impermeable. Horizontal conduction in the rock matrix is neglected.

3. The energy resistance at the contact between the aquifer and the rock matrix is negligible (infinite heat transfer coefficient).

4. The thermal properties of the formations above and below the aquifer may be different, but all thermal parameters for the liquid and the rocks are constant.

The differential equation governing the temperature in the aquifer at any time t can readily be derived by performing an energy balance on a control volume in the aquifer:

$$z = 0: \quad \frac{\lambda_1}{b} \frac{\partial T_1}{\partial z}\bigg|_{z=0} - \frac{\lambda_2}{b} \frac{\partial T_2}{\partial z}\bigg|_{z=0} - \frac{\rho_w c_w q}{b} \frac{\partial T_a}{\partial x} - \rho_a c_a \frac{\partial T_a}{\partial t} = 0. \tag{1}$$

The symbols are defined in the nomenclature. In the caprock and the bedrock the one-dimensional heat-conduction equation controls the temperature:

$$z > 0: \quad \lambda_1 \frac{\partial^2 T_1}{\partial z^2} = \rho_1 c_1 \frac{\partial T_1}{\partial t}, \tag{2}$$

$$z < 0: \quad \lambda_2 \frac{\partial^2 T_2}{\partial z^2} = \rho_2 c_2 \frac{\partial T_2}{\partial t}. \tag{3}$$

The initial conditions are:

$$T_a(x,0) = T_1(x,z,0) = T_2(x,z,0) = T_{b1} - a(z - D). \tag{4}$$

The boundary conditions are:

$$T_a(0,t) = T_f, \; t > 0, \tag{5a}$$

$$T_a(x,t) = T_1(x,0,t) = T_2(x,0,t), \tag{5b}$$

$$T_1(x,D,t) = T_{b1}, \tag{5c}$$

$$T_2(x,-H,t) = T_{b2} = T_{b1} + a(H + D). \tag{5d}$$

The following dimensionless parameters are introduced:

$$\xi = \frac{\lambda_1 x}{\rho_w c_w q D}, \tag{6a}$$

$$\tau = \frac{\lambda_1 t}{\rho_1 c_1 D^2}, \tag{6b}$$

$$\theta = \frac{b}{D} \frac{\rho_a c_a}{\rho_1 c_1}, \tag{6c}$$

$$\eta = \frac{z}{D}, \tag{6d}$$

$$\gamma = \frac{\rho_2 c_2}{\rho_1 c_1}, \tag{6e}$$

$$\omega = \frac{\lambda_2}{\lambda_1}, \tag{6f}$$

$$T_D = \frac{T - T_{b1}}{T_f - T_{b1}}, \tag{6g}$$

$$T_g = \frac{aD}{T_f - T_{b1}}, \tag{6h}$$

$$\alpha = \frac{H}{D}. \tag{6i}$$

Substitution of Equations (6a)-(6i) into Equations (1)-(3) yields

$$\eta = 0: \quad \left.\frac{\partial T_{D_1}}{\partial \eta}\right|_{\eta=0} - \omega \left.\frac{\partial T_{D_2}}{\partial \eta}\right|_{\eta=0} - \frac{\partial T_{D_a}}{\partial \xi} - \theta \frac{\partial T_{D_a}}{\partial \tau} = 0, \tag{7}$$

$$\eta > 0: \quad \frac{\partial^2 T_{D_1}}{\partial \eta^2} = \frac{\partial T_{D_1}}{\partial \tau}, \tag{8}$$

$$\eta < 0: \quad \frac{\partial^2 T_{D_2}}{\partial \eta^2} = \frac{\gamma}{\omega} \frac{\partial T_{D_2}}{\partial \tau}. \tag{9}$$

The initial conditions become

$$T_{D_a}(\xi,0) = T_{D_1}(\xi,\eta,0) = T_{D_2}(\xi,\eta,0) = -T_g(\eta - 1). \tag{10}$$

The boundary conditions become

$$T_{D_a}(0,\tau) = 1, \ \tau > 0, \tag{11a}$$

$$T_{D_a}(\xi,\tau) = T_{D_1}(\xi,0,\tau) = T_{D_2}(\xi,0,\tau), \tag{11b}$$

$$T_{D_1}(\xi,1,\tau) = 0, \tag{11c}$$

$$T_{D_2}(\xi,-\alpha,\tau) = T_g(\alpha + 1). \tag{11d}$$

The solution of Equations (7)-(11) can be easily obtained in the Laplace domain as (see Bodvarsson, 1981):

$$\eta = 0: \quad \mu = \frac{1}{p}\,[1 - T_g] \exp - \left[\theta_p + \frac{\sqrt{p}}{\tanh\sqrt{p}} + \frac{\omega\sqrt{q}}{\tanh\sqrt{q}}\right] + \frac{T_g}{p}. \tag{12}$$

$$\eta > 0: \quad v = \left[u - \frac{T_g}{p}\right] \cosh\sqrt{p}\eta - \frac{\left[u - \frac{T_g}{p}\right]}{\tanh\sqrt{p}} \sinh\sqrt{p}\eta - \frac{T_g}{p}(\eta - 1). \tag{13}$$

$$\eta < 0: \qquad w = \left[u - \frac{T_g}{p}\right] \cosh\sqrt{\frac{\gamma}{\omega} p}\, \eta + \frac{\left[u - \frac{T_g}{p}\right]}{\tanh\sqrt{\frac{\gamma p}{\omega}}} \sinh\sqrt{\frac{\gamma p}{\omega}}\, \eta - \frac{T_g}{p} (\eta - 1). \qquad (14)$$

In Equations (12)-(14), u, v, and w represent the temperature in the Laplace domain of the aquifer, the rock above the aquifer, and the rock below the aquifer, respectively.

As Equations (12)-(14) cannot easily be inverted from the Laplace domain, a numerical inverter developed by Stehfest (1979) was used. The results obtained by using the inverter are discussed below.

THEORETICAL STUDIES

The model has been employed to study the evolution of fault-charged hydrothermal systems. Figure 3 shows a plot of dimensionless temperature T_D versus depth at a given location for several different values of dimensionless time τ. All of the dimensionless parameters are defined in the nomenclature. The figure shows that initially ($\tau = 0$) the system is in equilibrium with a linear geothermal gradient. At $\tau = 0$ the hot water starts to flow into the permeable aquifer; in the early stages of development, only the aquifer is being heated. Later on, however, the conductive heat transfer between the aquifer and the adjacent rocks increases, causing the surrounding rock to be heated and the temperature in the aquifer to stabilize.

The temperature in the aquifer and the caprock reaches steady state at a dimensionless time, τ, between 1 and 10. At this time the temperature in the

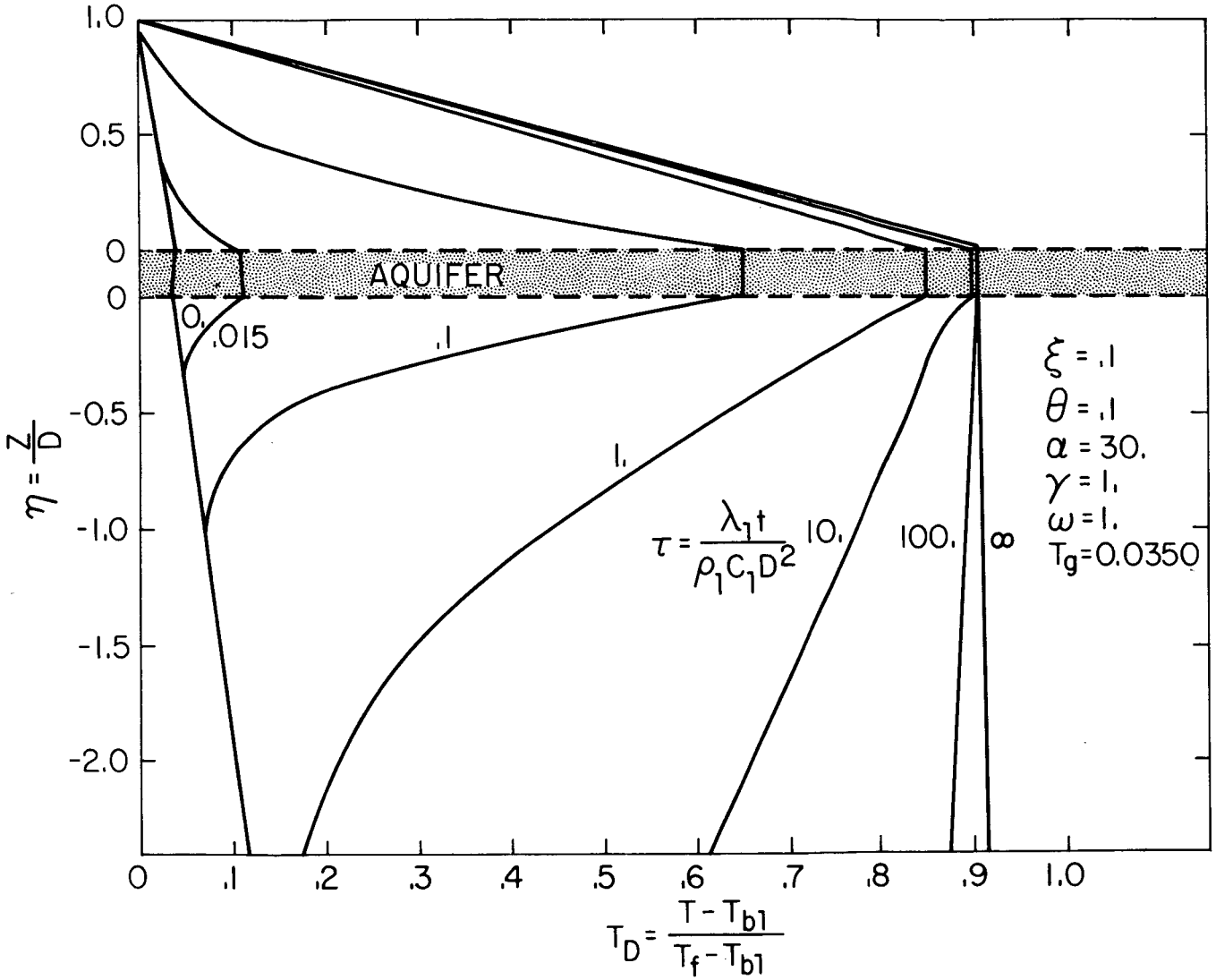

Figure 3. Evolution of a fault-charged hydrothermal system.

XBL 816-3179

rock formation below the aquifer is nowhere near a steady-state condition. The high value of $\alpha = 30$ shows that the constant-temperature boundary at the ground surface is much closer to the aquifer than the deep boundary and should therefore control the thermal response. In the example shown in Figure 3, the steady-state temperature of the aquifer at the location in question is approximately $T_D = 0.91$.

The temperature distribution along the aquifer is shown in Figure 4 for similar parameters as were used in Figure 3. The figure shows that close to the fault (small ξ) the temperature rises almost immediately to the temperature of the recharging water. The figure also shows that a steady-state temperature distribution is reached at a dimensionless time τ between 1 and 10. The steady state temperature distribution is independent of θ. The effects of other parameters on the development of a fault-charged hydrothermal system are given by Bodvarsson (1981).

APPLICATIONS

As a first attempt to verify the usefulness of this model for fault-charged hydrothermal systems, it was applied to data from the geothermal system at Susanville, California. The more than 20 exploration wells in Susanville have located a low-temperature (< 80°C) shallow geothermal aquifer of limited areal extent (Benson et al., 1980). Figure 5 shows the location of the wells and the temperature contours at an elevation of 1150 m, which corresponds to a depth of 125 m, where the primary aquifer is found. The temperature contours shown in Figure 5 suggest that the reservoir is charged by a fault with

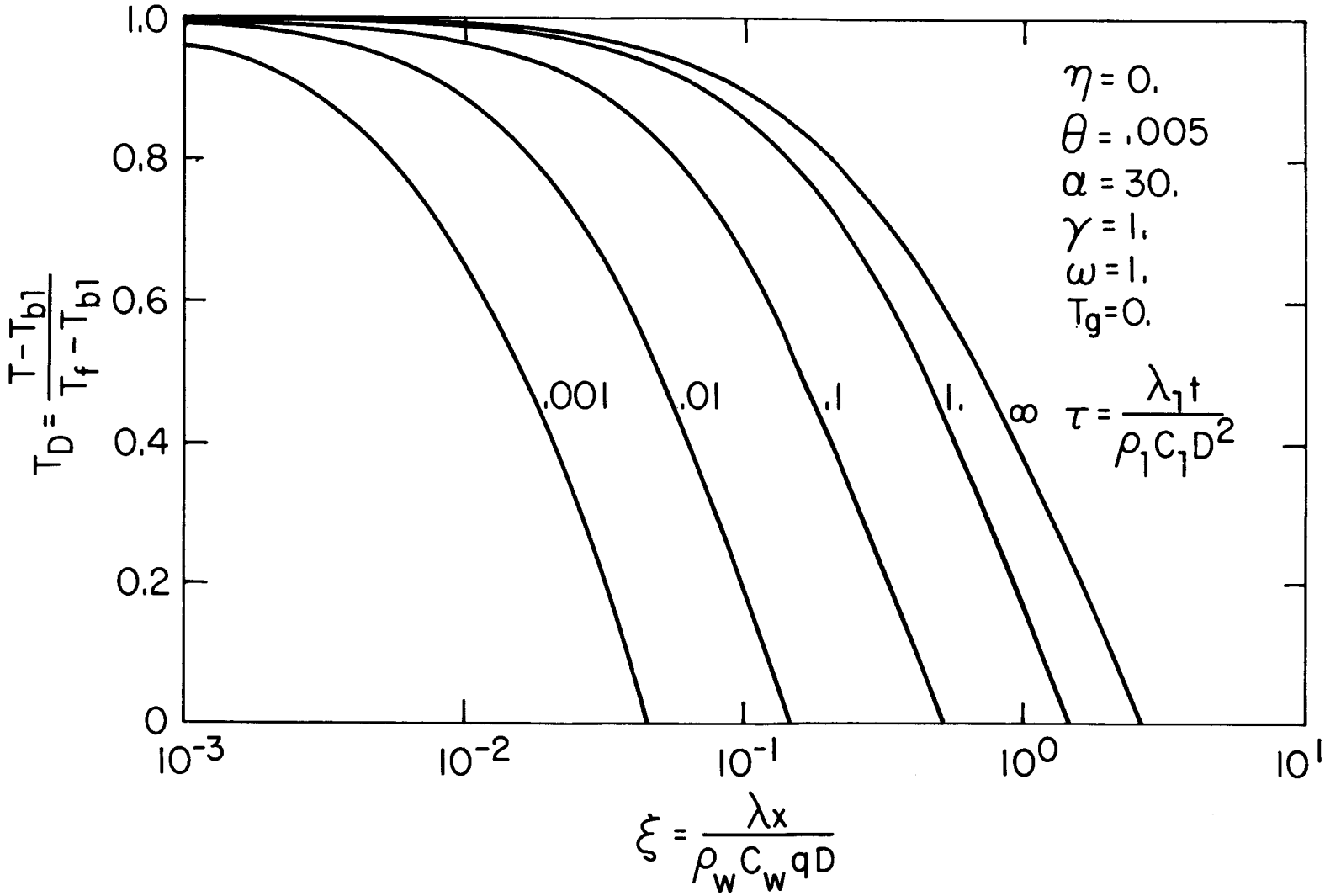

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Figure 4. Dimensionless temperature profiles away from the fault at different dimensionless times.

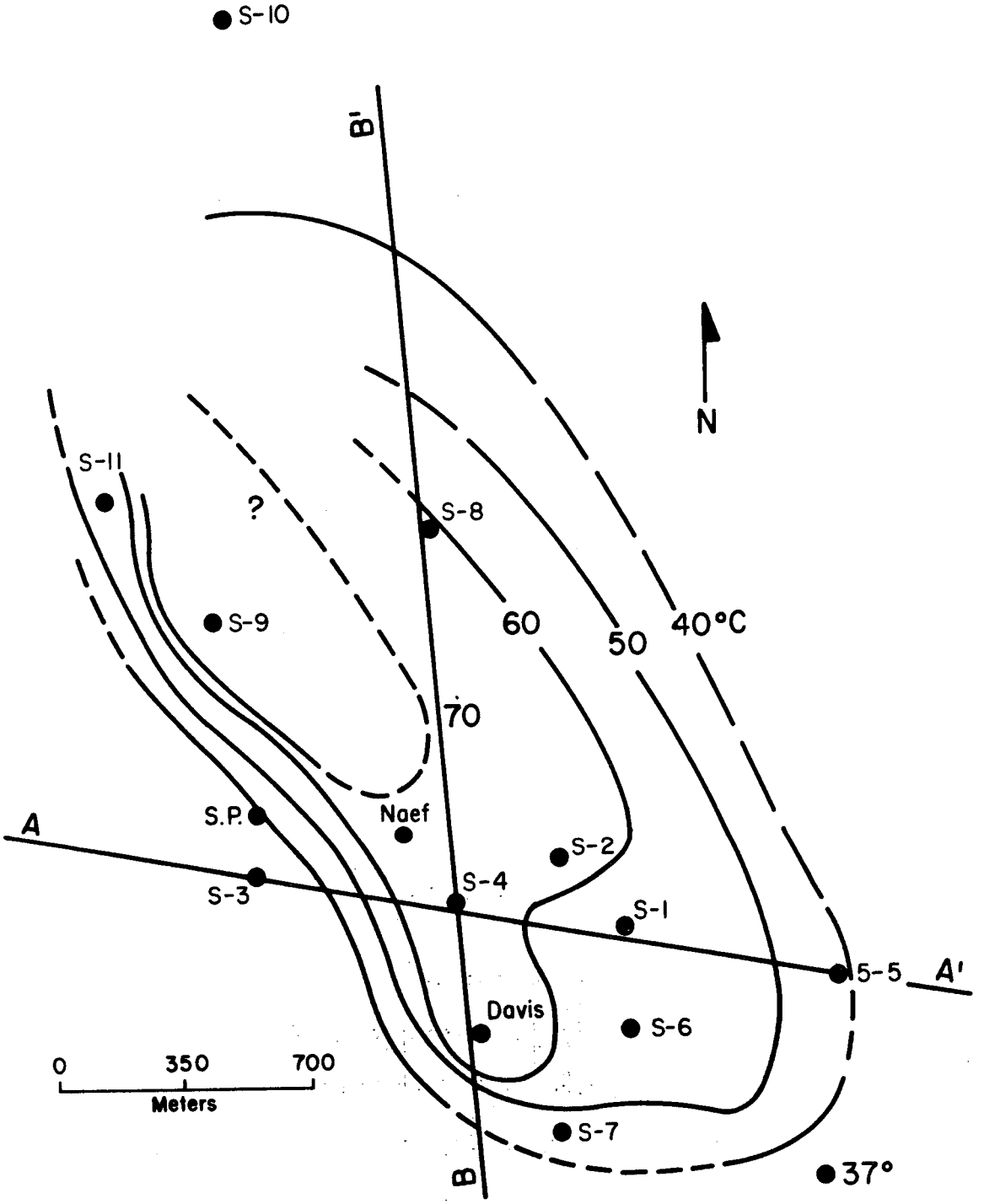

Figure 5. Temperature contours at 1150 m elevation at Susanville.
[XBL 807-7247]

a NW strike. The steep temperature gradients to the west of the proposed fault illustrate that the fault is recharging the aquifer only to the east. Temperature contour maps at different depths show fault-related characteristics similar to those shown in Figure 5. Furthermore, many of the wells at Susanville show a reversal with depth similar to that shown in Figure 1 for the Klamath Falls well.

One potential use for the hydrothermal energy at Susanville is space heating. However, the limited areal extent of the hydrothermal system (Figure 5) indicates that the mass of hot water (the limiting temperature taken as 60°C) amounts to only 1-3 x 10^7 m^3 (depending upon the aquifer thickness selected). Current plans (Department of Energy, 1980) call for an extraction rate of approximately 0.035 m^3/s (550 gpm) for space heating of 14 public buildings. If recharge is neglected, this corresponds to a lifetime of 9-27 years. If the project is intended for 20 years, its success will depend greatly upon the recharge rate. A reliable estimate of the recharge into the Susanville hydrothermal system is therefore greatly needed. Application of our model to the Susanville anomaly will give the first estimate of the recharge rate.

Table 1 shows the parameters selected from the well data. The maximum temperature measured in the field is approximately 80°C in well S-9a, which is located very close to the proposed fault (see Figure 5). The temperature of the water recharging the aquifer is therefore fixed at 80°C. Picking 60°C as the average aquifer temperature, the fluid parameters can be obtained, ρ_w = 983 kg/m^3, C_w = 4179 J/kg °C. It is now possible to calculate the appropriate value of θ, θ = 0.31 (Equation 6c).

TABLE 1. Parameters used for the Susanville model.

Parameter	Value
Aquifer thickness, b	35 m
Depth to aquifer, D	125 m
Aquifer porosity, ϕ	0.2
Thermal conductivity of rock, λ_1	1.5 J/m·s·°C
Rock heat capacity, c_1	1000 (J/kg·°C)
Rock density, ρ_1	2700 (kg/m^3)

The objective of this exercise is to use the model to match the temperature contour data shown in Figure 5 and the temperature profiles from individual wells in an attempt to estimate the hot water recharge. After a number of computer runs, the match shown in Figures 6 and 7 was obtained. As Figure 6 shows, the calculated temperature contours compare very well with the observed ones in the hottest region of the field, close to the proposed fault. Further away, however, there are large differences between the calculated and the observed temperatures. There are many possible reasons for the discrepancy. First, only limited data are available away from the fault (only wells S-5 and S-10), so that temperature contours are not accurately known. Second, evidence shows that there is a high regional flow of ground-water to the southeast and that mixing of the colder shallow groundwater with the hot fluids is taking place. Third, the subsurface geology is considerably more complex than can be accounted for by the simple model we have used here. In any case, the model matches the temperature profiles of wells close to the proposed fault very well, as shown in Figure 7.

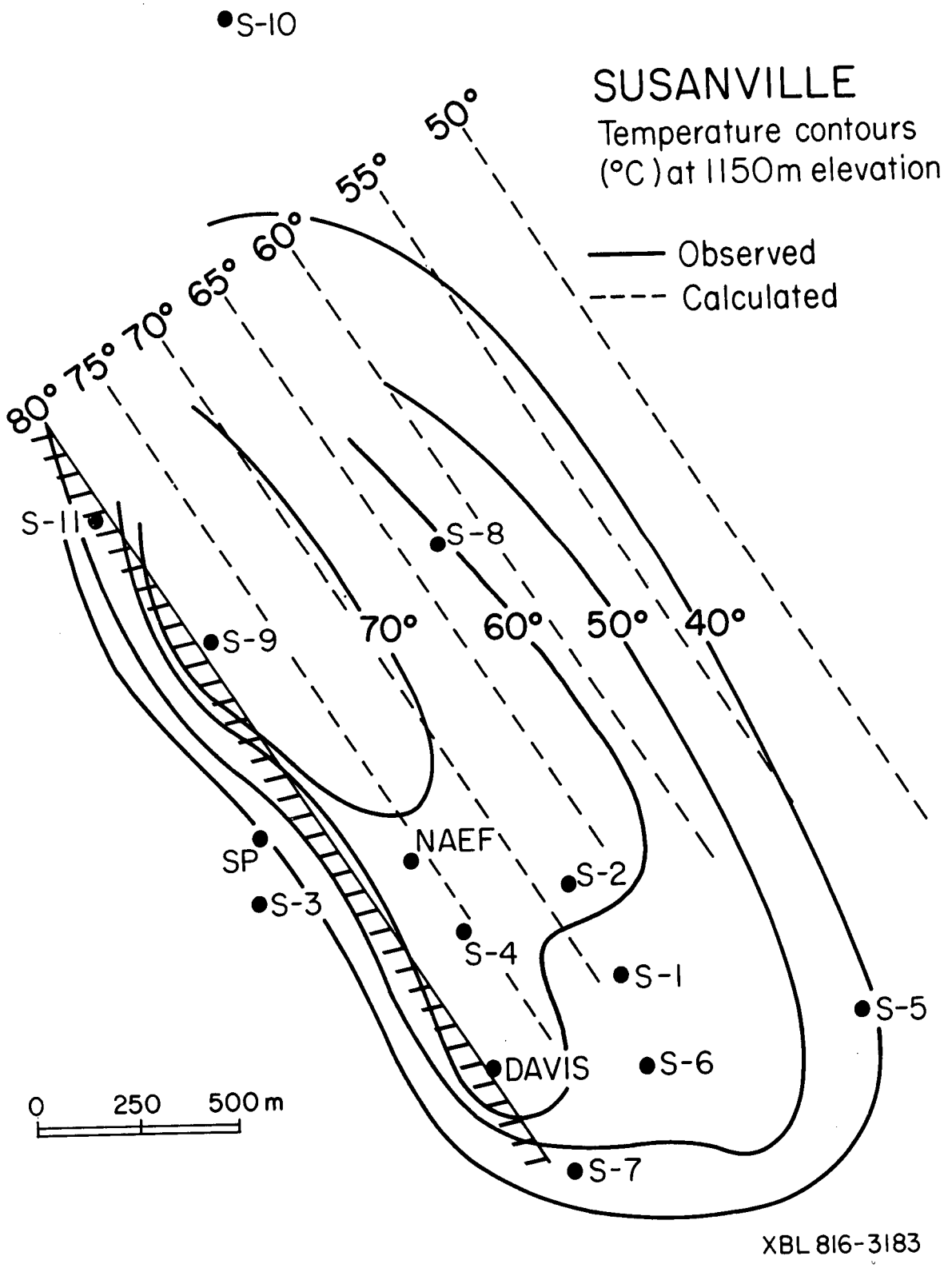

Figure 6. Comparison between observed and calculated temperature contour data.

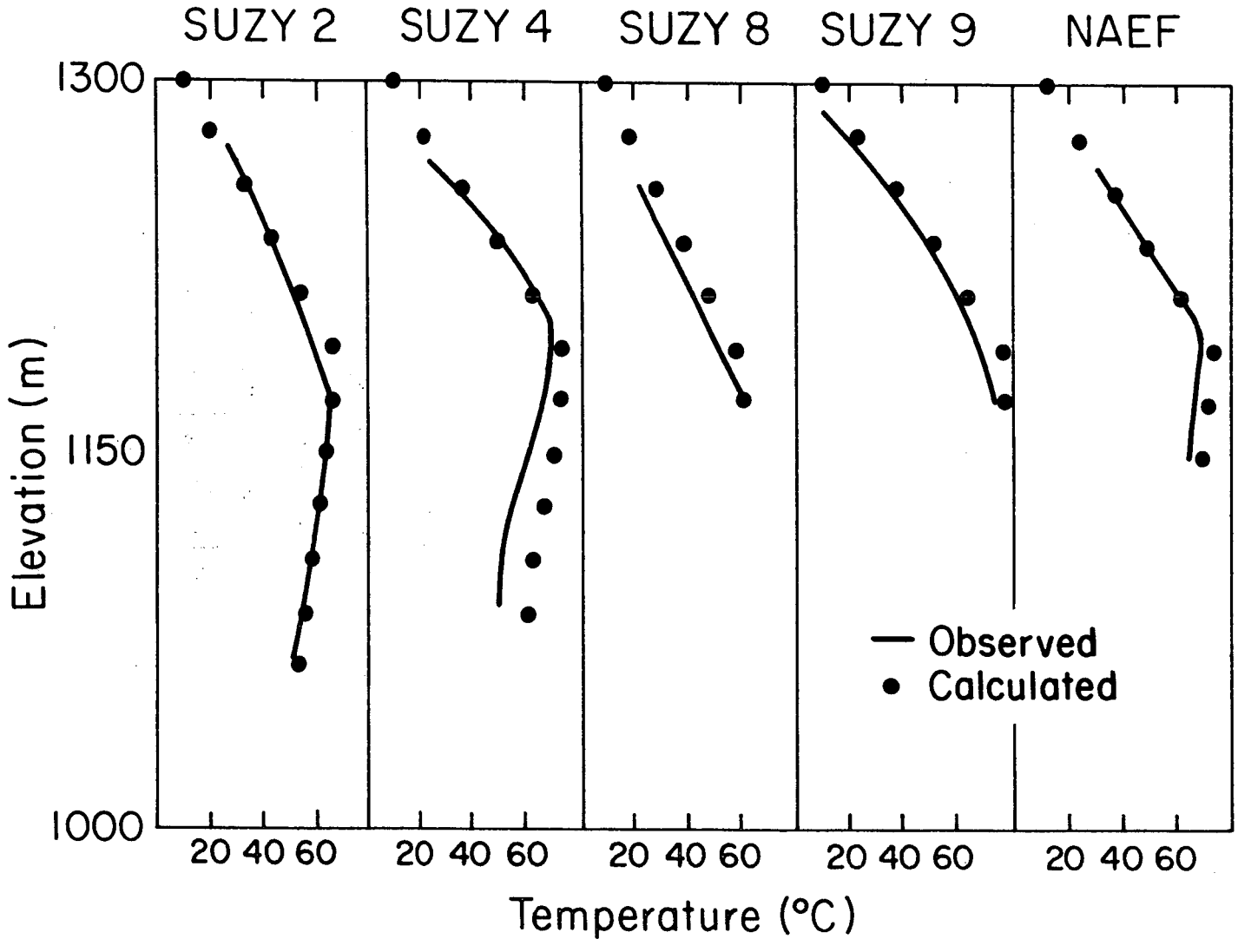

Figure 7. Comparison between calculated and observed temperature profiles in wells. [XBL 816-3181]

The match shown in Figures 6 and 7 was obtained using two different sets of parameters. First, if the lower constant temperature boundary is placed very deep ($H \gg D$), the parameters obtained indicate that the hydrothermal system has been under development approximately 2000 years and that the fault charges the system at a rate of 9×10^{-6} $m^3/s \cdot m$. Second, a very similar match is obtained if the constant temperature boundary is placed at a depth of about 400 meters ($\alpha = 2.0$); in this case the parameters obtained show that steady-state temperature conditions are reached (consequently the evolution time cannot be determined except that it exceeds 10,000 years) but the calculated recharge rate is the same as in the first case (9×10^{-6} $m^3/s \cdot m$). If one considers the age of the subsurface formations at Susanville, the second case seems more likely. Also it is not unlikely that a deeper permeable aquifer with circulation of colder water is present at the site, and this would act as a constant temperature boundary.

In any case, the accuracy of the calculated recharge rate is of more concern to the developers of the Susanvillle hydrothermal system than the time of evolution. If the heat losses from the aquifer are controlled by heat conduction as we have assumed in the present model, the calculated recharge rate should be reasonably accurate. However, in the model horizontal conduction is neglected, and this may make the actual recharge rate greater than what we have calculated.

If we assume that the calculated flow rate is correct and that the fault recharges over a distance of 2500 m, the total rate of recharge is approximately 0.0225 m^3/s. This recharge rate corresponds to approximately 70% of

the proposed extraction rate; consequently a project lifetime of 25-75 years could be expected. It should be emphasized, however, that the simplicity of the present model does not warrant conclusive interpretations. The results presented here should be considered as rough first estimates.

Unfortunately, detailed heat flow data over the Susanville anomaly are not available at present; such data would have been useful in confirming the accuracy of the model. Figure 8 shows the calculated heat flow values plotted against distance from the proposed fault.

SUMMARY

A simple two-dimensional model has been developed for fault-charged hydrothermal systems. The model has been used for some theoretical studies on the development of such systems. Furthermore, the model has been tentatively applied to the hydrothermal system at Susanville, California. A reasonable match with temperature data from the field allowed approximate calculations of the recharge rate from the fault into the hydrothermal system.

ACKNOWLEDGMENTS

Valuable discussions with C. Goranson during the course of this study and critical reviews by C. Doughty and M. Wilt are most appreciated. This work was supported by the Assistant Secretary for Conservation and Renewable Energy, Office of Renewable Technology, Division of Geothermal and Hydropower Technologies of the U. S. Department of Energy under Contract No. W-7405-ENG-48.

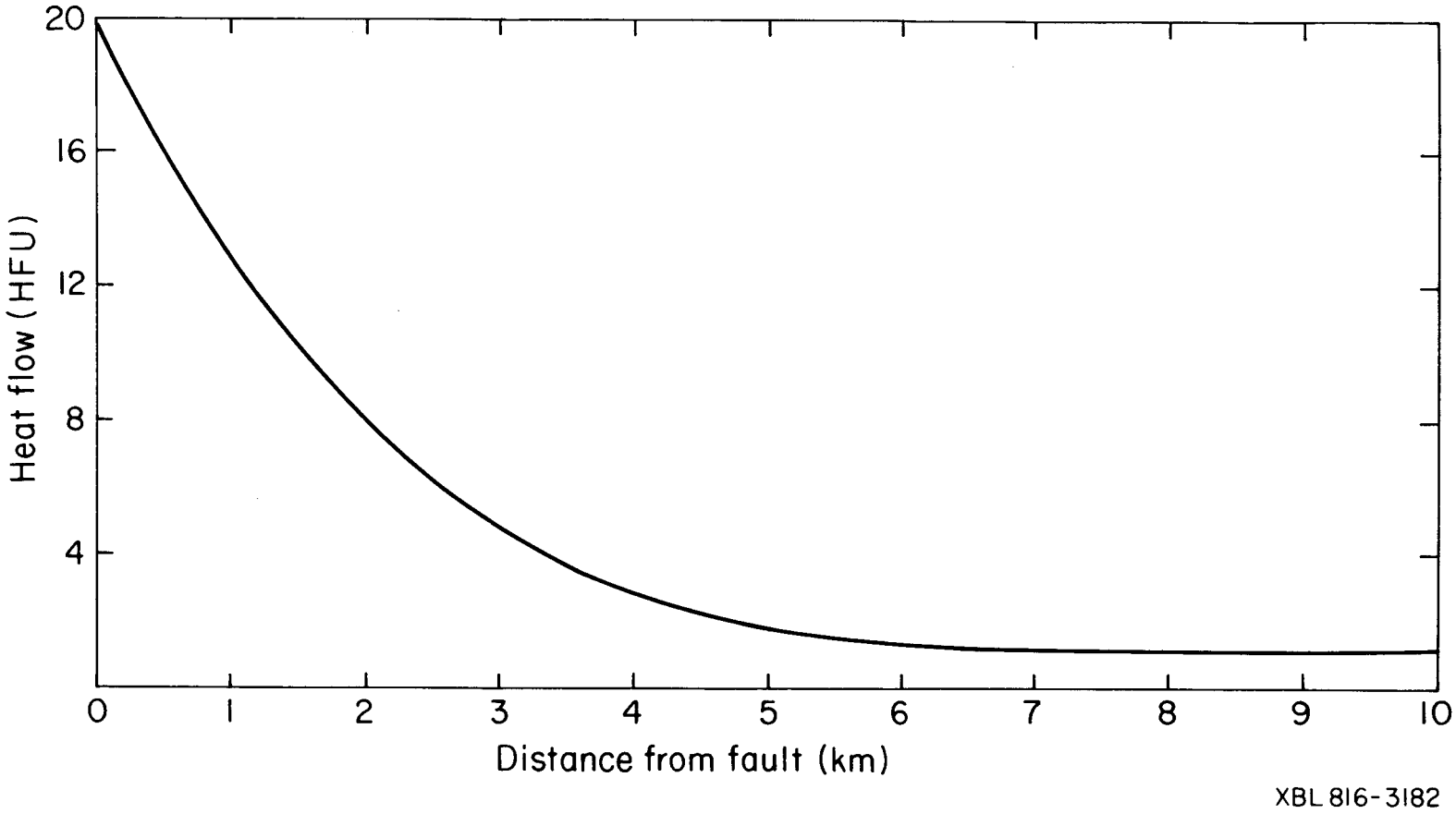

Figure 8. Calculated heat flows at Susanville.

NOMENCLATURE

a	geothermal gradient (°C/m)
b	aquifer thickness (m)
D	thickness of caprock (m)
H	thickness of bedrock (m)
p	Laplace parameter
ϕ	porosity
q	the recharge rate ($m^3/s \cdot m$)
t	time (sec)
T	temperature (°C)
T_{b1}	temperature at ground surface (°C)
T_f	temperature of recharged water (°C)
u	temperature in aquifer in Laplace domain
v	temperature in rock matrix above aquifer in Laplace domain
w	temperature in rock matrix below aquifer in Laplace domain
x	lateral coordinate (m)
z	vertical coordinate (m)
λ	thermal conductivity (J/m•s•°C)
ρc	volumetric heat capacity ($J/m^3 \cdot$°C)

Subscripts

a	aquifer
1	rock matrix above aquifer
2	rock matrix below aquifer
w	liquid water

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This report was done with support from the Department of Energy. Any conclusions or opinions expressed in this report represent solely those of the author(s) and not necessarily those of The Regents of the University of California, the Lawrence Berkeley Laboratory or the Department of Energy.

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