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**Analysis of Temperature-Time Data From 3 m
Drillholes at Crystal Hot Springs, Utah**

by

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ABSTRACT

A method for determining background geothermal gradient values through the analysis of temperature measurements at multiple depths to 3 m and recorded over a time span of several days is presented. The analysis is based on the amplitude decay and phase shift of temperature waves with depth. Diurnal and other high frequency temperature variations are used to compute thermal diffusivities which in turn are used to model and remove the effect of the annual temperature wave. The analysis considers both a homogeneous half space and a two layer medium consisting of an overburden of finite thickness overlying a semi-infinite substratum.

The method was tested in three holes in the Crystal Hot Springs geothermal field. Temperatures in each hole were recorded once a minute over a period of three days with a probe containing thermistors at eight different depths. Five of the thermistors were positioned at shallow depths (≤ 0.5 m) to monitor diurnal and other high frequency waves and three at greater depths (≥ 1 m) to measure lower frequency variations. Since measurements were recorded at only three sites, the accuracy and reliability of the method is not fully evaluated. Potential problems to the method resulting from inaccurate model parameters and convective heat transport are investigated.

INTRODUCTION

Shallow temperature surveys, at depths less than 5 m, have been used as a geophysical prospecting tool in a variety of applications. These include: mapping of subsurface fissures (Kappelmeyer, 1957), groundwater exploration (Birman, 1969), subsurface structure mapping (Poley and Van Steveninck, 1970), and geothermal exploration (Kintzinger, 1956; Olmstead, 1977; Le Shack et al., 1979; Lange et al., 1982).

The principal advantages of shallow temperature surveys over surveys in deeper conventional drillholes are the lower cost per station and the better spatial coverage of the prospect area accomplished at lesser environmental impact. The disadvantages of the method are related to problems of separating the thermal signal of structural, hydrologic or geothermal origin from background noise. Noise in this sense includes the transient temperature fields associated with time dependent fluctuations of the surface temperature, variations in surface conditions which lead to spatial variations in the amplitude of these transient effects, and lateral variations in subsurface thermal properties which affect the propagation of surface fluctuations into the ground. Some of these perturbing factors are listed explicitly by Le Shack et al. (1977) as follows: (a) diurnal solar heating variations, (b) annual solar heating variations, (c) aperiodic solar heating variations, (d) variations in surface albedo, (e) variations in surface roughness, (f) variations of soil diffusivity, (g) changes in slope and exposure of the terrain, (h) variations in elevation, and (i) variations in level of groundwater and groundwater movement. Factors a, b, and c govern the time dependence of surface temperature fluctuations, factors d, e, g, and h govern their amplitude, while f and i govern propagation of the noise. Discussions on the effects of these perturbing factors are given by Poley and Van Steveninck (1970), Kappelmeyer and Haenel (1974), and Le Shack et al. (1977).

Any successful application of the shallow temperature prospecting tool must overcome the noise problem which is especially significant in the uppermost 10 m of the

earth. In this paper we present a new approach to this problem based on the analysis of temperatures measured at multiple depths in a 3 m probe and recorded over a time span of several days. The method is tested in a geothermal prospect area where the temperature field is known from deeper drillholes.

Previous Studies

Use of shallow temperature surveys as a geophysical prospecting method can be divided between geothermal and non-geothermal applications. Kappelmeyer (1957) was one of the first to explore the possibilities of temperature measurements at depths between one and two meters for discovering anomalies due to causes at greater depths. Using thermistors emplaced at a depth of 1.5 m, he was able to locate a subsurface fissure which formed a conduit for the flow of hot water from great depth to a groundwater stream near the surface. With thermistors at a depth of 3 m, Birman (1969) measured temperature variations ascribed to groundwater flow in southern California and found this method a valuable tool in groundwater exploration. The application of shallow temperature measurements to locating salt domes and shallow faults was investigated in detail by Poley and Van Steveninck (1970).

A number of shallow temperature surveys have been conducted in an attempt to develop a reliable exploration tool for locating geothermal anomalies. Two of the earlier examples of shallow temperature surveys at geothermal areas made little or no attempt to correct for the perturbing factors mentioned above. Kintzinger (1956) surveyed an area of hot ground near Lordsburg, New Mexico using thermistors installed to a depth of 1 m. Due to the proximity of the heat source to the surface, he was able to observe an anomaly of about 10°C around a hydrothermal area. Olmstead (1977) also reported temperature measurements conducted in 1973 at a depth of 1 m to outline an anomaly surrounding a fumerole in the Soda Lakes area of Nevada. He interpreted this substantial anomaly to be caused by ascending steam. Later, in 1975, Olmstead surveyed an area at

Upsal Hogback, Nevada, where temperature data from many 30 m deep holes were available. In comparing his uncorrected 1 m temperatures with the 30 m data, he found no meaningful correlation between the data sets.

Some of the more recent examples typify efforts in which corrections for some of the perturbing factors were applied. Le Shack et al. (1977) conducted surveys at Long Valley and Coso Hot Springs KGRAs, California. Although they made measurements for albedo, surface roughness, percentage of soil moisture, and thermal diffusivity, no corrections were applied due to the high degree of spatial homogeneity they found in these parameters coupled with strong thermal anomalies in each of the fields. However, corrections for variations in elevation were applied. In a later report, Le Shack et al. (1979) demonstrated the importance of corrections for variations in thermal diffusivity with analysis of Olmstead's (1977) 1 m data at Upsal Hogback. By eliminating 3 temperature data points, considered to be measured in areas with significantly higher thermal diffusivity than the remaining sites, and recontouring the temperature maps with elevation corrections, Le Shack et al. (1979) suggested that comparison of the corrected 1 m data with the 30 m temperature data showed good correlation, in contradiction to Olmstead's earlier conclusion.

Using data from somewhat deeper drillholes (7 to 15 m), Lee (1977) obtained values for the geothermal gradient and thermal diffusivity through two mathematical formulations based on nonlinear least squares fitting. The first method requires temperature measurements at two given depths over several months to determine the temperature distribution in the ground as a function of depth and time. This distribution is a function of thermal diffusivity and the thermal gradient, and thus provides a means for calculating these parameters. In the second method, temperature measurements are made at more than five depths at one given time. Thermal diffusivity is obtained by scanning all possible values of this parameter and finding a minimum in the standard deviation between observed and calculated temperature curves. The thermal gradient is

one of four parameters then determined from a least-squares method assuming the thermal diffusivity is correct. Lee (1977) tested these methods with a survey in the Salton Sea geothermal field. The gradients from both methods showed good agreement with those determined from much deeper holes in the area.

Most recently Lange et al. (1982) have further demonstrated the suitability of using 3 m deep temperature surveys by tests at geothermal prospects McCoy and Dixie Valley in Nevada. They measured temperatures at seven depths spaced 0.25 m apart between 1.5 and 3 m depth. An analytical innovation included in this study was the use of the depth to a temperature extremum (summer maximum or winter minimum of the subsurface temperature wave) to calculate in situ thermal diffusivity values. Consequences of mapping 3 m temperature values with and without corrections for areal differences in thermal diffusivity were demonstrated by a comparison between contour maps of each of these quantities. A high correlation existed between low 3 m temperatures and low thermal diffusivities as is expected from theory. Shallow temperature logs should be recomputed and normalized to a common diffusivity in such cases. A limitation of the Lange et al. (1982) method is that the survey must be conducted in the periods for which the temperature extremum is at depth between 1.5 and 3 m, a time window of about 23 to 44 days in Spring and Fall depending on thermal diffusivity values.

In view of previous work, we have examined a slightly different procedure for conducting shallow temperature surveys using measurements to a depth of 3 m. The procedure includes determining thermal diffusivity through analysis of the direct effects this quantity has on the decay and phase lag with depth of the diurnal and other high frequency temperature waves. The geothermal gradient can then be determined by removal of the annual wave assuming that thermal diffusivity is constant below the detectable depth of the diurnal wave. Unlike previous work, this method makes use of temperatures at several depths sampled every minute over a period of days for the direct

computation of thermal diffusivity. The method can be used any time of year without snowcover.

THEORY

The temperature distribution within a homogeneous, isotropic semi-infinite medium with no heat sources or sinks and negligible convective heat transport is governed by the equation

$$\frac{\partial T}{\partial t} = \alpha \frac{\partial^2 T}{\partial z^2} \quad (1)$$

where T is temperature, t is time, z is depth into the medium, and α is thermal diffusivity. The solution of this heat conduction equation with appropriate initial and boundary conditions yields the time dependent thermal regime within the medium.

Suppose a temperature variation about the mean surface temperature T_0 , described by a Fourier series, is applied to the surface of this half-space. For the case of the earth's surface, the two largest amplitude components of this series consist of waves with annual and diurnal periods; the other components are smaller amplitude waves needed to describe aperiodic temperature variations as well as deviations from sinusoidal waveforms in the annual and diurnal inputs. If, in addition, we consider a uniform heat flux q_0 from the interior, the temperature distribution in the half-space can be expressed as

$$T = T_0 + \frac{q_0}{k} z + \sum_{n=1} A_n e^{-K_n z} \cos(\omega_n t - K_n z - \epsilon_n) \quad (2)$$

$$K_n = (\pi / \alpha \tau_n)^{1/2}$$

In this equation, k is the thermal conductivity of the medium, ω_n are the angular frequencies of the waves with surface amplitudes A_n , wave numbers K_n , and arbitrary phase shifts ϵ_n . The wave numbers K_n are functions of the periods of the waves τ_n and the thermal diffusivity α of the medium.

Figure 1 illustrates representations of a single periodic surface temperature variation in both temperature - time coordinates and temperature - depth coordinates. For waves with several frequency components, the amplitude of each component decays separately with depth. Each component is also subjected to a phase lag with depth. The decay is stronger and the phase shift greater for a short period wave than for a long period wave. Poley and Van Steveninck (1970) have shown that with constant thermal diffusivity and equal surface amplitudes, the annual wave will penetrate $\sqrt{365}$ or about 19 times the depth of the diurnal wave.

Equation (2) expresses the temperature distribution within the near surface region of the earth only if the assumptions of homogeneity, isotropy, absence of heat sources and sinks, and conductive heat transport are valid. It is appropriate to examine some of these assumptions in detail in terms of complications to this ideal temperature distribution.

Thermal Diffusivity

Heat transfer by conduction is controlled by two independent thermal properties, namely the thermal conductivity k , and the heat capacity per unit volume C . The quotient of these two quantities appears in many heat flow problems giving rise to the parameter $\alpha = k/C$ which is termed thermal diffusivity. Thermal diffusivity is also commonly expressed in terms of the specific heat c and density ρ in which case $\alpha = k/\rho c$. The range of thermal diffusivity in rocks is small because the properties k and c generally change in the same direction; however, soils have a much larger range due to fluid filling of pore spaces (see Table 1). Thermal diffusivity of rocks varies by about a factor of two; soil diffusivities vary by an order of magnitude. This variation gives rise to a serious complication in thermal gradient determinations from shallow temperature measurements.

TABLE 1

Thermal diffusivity of soils, rocks, and water.

Material	Thermal Diffusivity (mm^2s^{-1})
Water	1.4
Bog soil	0.1 - 0.2
Limestone	0.4 - 0.8
Tuff	0.4 - 0.9
Basalt	0.5 - 0.9
Loamy Soil	0.8
Sand, wet	0.9
Sandy, dry	1.1 - 1.3
Shale, slate	0.8 - 1.6
Sandstone	1.1 - 2.3
Granite	1.4 - 2.1
Rocksalt	1.1 - 3.4

Sources: Kappelmeyer and Haenel (1974), Table 3.6, p. 60; Roy et al. (1981), Tables 12.4, p. 429 and 12.5, p. 430.

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A measurable influence of the annual surface temperature wave extends to a depth of about 10 m or greater as seen in Figure 2. Since this influence is superimposed on, and in cases of normal heat flow dominates the geothermal gradient, it is necessary to remove the effects of the annual wave on the temperature distribution when shallow temperature measurements are made for the purpose of geothermal prospecting. This situation is aggravated by variations in thermal diffusivity across the prospect area.

Spatial variations in thermal diffusivity cause a change in the propagation of a thermal wave into the earth by altering the decay (as seen in Figure 2) and phase shift of the wave with depth. Similar to a change in period, the decay is more pronounced and the phase shift greater for a small value of diffusivity than for a large value. Figure 3 illustrates the effect of different diffusivities on the propagation of the annual wave for several times of the year. It is clear from Figure 3 that measuring and mapping temperatures at single depths down to 10 m, especially from 0 to 5 m, without regard for possible diffusivity variations could lead to spurious results. Thus, it is desirable to determine and remove the effect of the annual wave at each measurement site as it is conceivable that no two sites are affected in the same way.

One approach to solving this problem is to monitor the annual wave at a number of depths over a long period of time. However, this would prove time consuming and costly. An alternative approach is to monitor diurnal and other high amplitude, high frequency waves over a shorter period of time. For example, the amplitude of the diurnal wave decays rapidly with depth as seen in Figure 2. If this amplitude is determined at certain discrete points over an appropriate depth range, the exponential decay can be determined which in turn yields a value for thermal diffusivity. Similarly, a value for diffusivity can also be obtained by determining the phase shift at each of these discrete depth points. The same information is available in other high amplitude harmonics of comparable frequency.

So far our formulation considers areal variations in thermal diffusivity only.

Vertical variations complicate the problem considerably as shown by Parasnis (1976) in a study of the effect that a uniform overburden has on the propagation of a thermal wave into the underlying rock. He considered a uniform overburden with thickness h , thermal conductivity k_1 , and thermal diffusivity α_1 , overlying a semi-infinite substratum with conductivity k_2 and diffusivity α_2 , and the temperature at the surface varies with time t as $T_0 \cos \omega t$. The temperatures T_1 in the overburden and T_2 in the substratum are given by Parasnis (1976) as follows:

for $0 \leq z \leq h$ (overburden)

$$T_1(z, t) = \frac{T_0 B}{R} \left[\exp(-(\omega/2 \alpha_1)^{1/2} (2h - z)) \cos(\omega t - (\omega/2 \alpha_1)^{1/2} (2h - z) + \varphi) \right. \\ \left. - \exp(-(\omega/2 \alpha_1)^{1/2} (2h + z)) \cos(\omega t - (\omega/2 \alpha_1)^{1/2} (2h + z) + \varphi) \right] \\ + T_0 \exp(-(\omega/2 \alpha_1)^{1/2} z) \cos(\omega t - (\omega/2 \alpha_1)^{1/2} z) \quad (4)$$

for $h \leq z \leq \infty$ (substratum)

$$T_2(z, t) = \frac{T_0(1+B)}{R} \exp(-(\omega/2 \alpha_1)^{1/2} ((\alpha_1/\alpha_2)^{1/2} (z-h) + h)) \cos(\omega t - (\omega/2 \alpha_1)^{1/2} ((\alpha_1/\alpha_2)^{1/2} (z-h) + h) + \varphi) \quad (5)$$

where z is depth from the surface and

$$R = (1 + 2B e^{-u} \cos u + B^2 e^{-2u})^{1/2}$$

$$\varphi = \tan^{-1} (B e^{-u} \sin u / (1 + B e^{-u} \cos u))$$

$$u = 2h (\omega/2 \alpha_1)^{1/2}$$

$$B = [(k_1/k_2) ((\alpha_2/\alpha_1)^{1/2} - 1)] / [(k_1/k_2) ((\alpha_2/\alpha_1)^{1/2} + 1)]$$

Important features of this temperature distribution discussed by Parasnis (1976) are: (a) the attenuation of temperature with depth in the overburden is not exponential, (b) the attenuation in the substratum is exponential with the attenuation constant $(\omega/2 \alpha_2)^{1/2}$ depending on the diffusivity of the substratum alone, and (c) the phase displacement of the substratum temperature consists of three part, a constant that depends on the overburden thickness and is independent of the substratum, a second constant part that depends on the substratum, and a variable part that depends on the path length.

In the case under consideration, removal of the annual effect becomes a five variable problem. We must determine conductivities (k_1, k_2), diffusivities (α_1, α_2), and the thickness of the overburden h to model accurately the propagation of the annual wave. Since conductivities and thickness of the overburden can be measured, we are left with determining the diffusivities. If diffusivities are calculated at discrete depth intervals with the method outlined above for the case of no vertical variation, then α_1 and α_2 can be found by matching these calculated diffusivities, which vary with depth, to diffusivities calculated in the same way using data generated by a computer model of the two layer case. However, this procedure can only be implemented in cases where the thickness of the overburden is less than the detectable depth of penetration of the diurnal wave.

Cases involving more than two layers have been considered by Lindquist (1983) and Clauser (1984). They found solutions to the heat diffusion equation through use of the Laplace transform. However, values for thermal conductivity and thermal diffusivity were found independently. Due to the increasing difficulty in constraining cases involving more than two layers without thermal diffusivity measurements, we will consider the two layer case only.

Rainwater Percolation

A further complication to shallow geothermal prospecting is perturbation of the conductive thermal regime caused by migration of fluids. The most prevalent type of heat transport by migration of fluids in the uppermost 3 m of the earth results from the flow of rainwater into the ground. Since this flow is generally sporadic and occurs in unsaturated sediments down to the water table, modeling this effect on shallow temperatures is difficult and uncertain. However, some insight can be gained by examining the one dimensional, conduction-advection, heat transport equation presented by Andrews and Anderson (1979) with some added simplifying assumptions.

Consider a constant vertical water flow with specific discharge Q in a homogeneous, isotropic, semi-infinite, saturated medium. If we assume no thermal dispersion, the equation governing the temperature within this medium is

$$k \frac{\partial^2 T}{\partial z^2} - \rho_f c_f Q \frac{\partial T}{\partial z} = (n \rho_f c_f + (1-n) \rho_s c_s) \frac{\partial T}{\partial t} \quad (6)$$

where T is temperature, t is time, z is depth into the medium, k is thermal conductivity, ρ_f and ρ_s are densities of the fluid and the saturated medium respectively with respective specific heats c_f and c_s .

With boundary conditions:

$$T(0,t) = T_0 \cos \omega t$$

$$T(\infty,t) = 0$$

we obtain as a solution to this equation

$$T(z,t) = T_0 \cos(\omega t - \xi z) e^{-\Phi z}$$

where ξ controls the phase change with depth and Φ is the attenuation constant.

Solving for ξ and Φ one obtains the relation

$$\xi = \frac{(n \rho_f c_f + (1-n) \rho_s c_s)}{(2 k \Phi + \rho_f c_f Q)} \quad (7)$$

In equation (2), the case of no fluid migration, the factors ξ and Φ are equivalently K . Therefore, the significance of equation (7) is that in the presence of fluid migration, ξ and Φ are no longer equal as in equation (2) and thermal diffusivities as calculated from phase changes with depth will not be the same as those calculated from amplitude changes with depth. We will not pursue the specific solution to this problem but will have occasion to remark on the influence of rainwater percolation in interpreting our observations.

DATA ACQUISITION

In testing the applicability of the method, we conducted a shallow hole temperature survey at Crystal Hot Springs geothermal area (Figure 4) at the south end of Salt Lake Valley, Utah, during the month of November, 1982. This site was chosen on the basis of (1) its proximity to our laboratory as field data were sent to the laboratory computer by telemetry, and (2) the existence of a geothermal survey conducted in the area by Murphy and Gwynn (1979) which serves as a test for the shallow drill hole results. Because of time limitations and the completion of only one working temperature probe, our survey was limited to monitoring shallow temperatures at three sites. The three shallow holes were drilled next to the deep wells labeled D, E, and F in Figure 4. These three sites are characterized by a range of thermal gradients from a value of $0.32^{\circ}\text{C m}^{-1}$ outside the geothermal field at well F to the anomalously high near surface value of approximately $4.3^{\circ}\text{C m}^{-1}$ in the middle of the field at well E. Each shallow hole was cased with 1 1/4 inch PVC pipe sealed at the bottom and capped at the top to prevent water from seeping into the temperature probe. Samples of the sediments brought up during drilling were kept for thermal conductivity measurements.

The temperature probe consisted of a 3/4 inch diameter PVC pipe with eight thermistors embedded in the side with thermally conductive epoxy. Yellow Springs Instruments YSI 44004 thermistors with a resistance of about 2000 ohms at room temperature were used. In the temperature range of interest (-10 to $+50^{\circ}\text{C}$), the sensitivity of this thermistor varies from 650 ohms/ $^{\circ}\text{C}$ at -10°C to 31.5 ohms/ $^{\circ}\text{C}$ at 50°C . This sensitivity was partially linearized by connecting a temperature stable resistor with a resistance equal to the thermistor resistance at 25°C in parallel with each thermistor. This gave sensitivities of 16.0 ohms/ $^{\circ}\text{C}$ at -10°C , 25.0 ohms/ $^{\circ}\text{C}$ at 25°C , and 16.9 ohms/ $^{\circ}\text{C}$ at 50°C . The thermistors were placed to record temperatures at depths of 0, 0.05, 0.10, 0.25, 0.50, 1.0, 2.0, and 3.0 meters. The four shallowest thermistors were required to monitor changes with depth of the amplitude and phase of the diurnal and other temperature waves with comparable frequency. The deeper thermistors provided information for determining the thermal gradient once corrections were made. Once the circuitry was in place, the pipe was filled with an electrically nonconductive sealant for waterproofing.

Associated with the temperature probe was an instrument package referred to as the data logger. The data logger was connected to the probe by cable to measure the resistance of the thermistors at a specified time interval. We chose the time interval of one minute between readings; however, this interval can be lengthened for increased battery life. The thermistor resistances were measured with a four wire technique; two wires supplied the drive current and the remaining two wires were used for voltage measurements. After each measurement was made, the information was converted to digital form and sent to our laboratory computer by radio. Features incorporated in the design of the data logger to increase the accuracy and precision of the measurement include (1) the measurement was ratiometric or independent of the drive current, (2) voltage offsets caused by thermocouple effects, amplifiers, etc., were nulled out, and (3) 60 Hz, power line noise was eliminated by integration of the measurement over a full

cycle.

The temperature probe was calibrated in a constant temperature bath specially designed to hold it, over a temperature range from 0°C to 44°C. Within this range, an increment of about 5°C was used between 0°C and 20°C and increment of about 8°C was used between 20°C and 44°C. Temperatures were measured with a Hewlett Packard 2804A quartz thermometer. To fit the points, we used the calibration curve $\ln R = A + B/(T + C)$ where T is temperature in °C, R is resistance in ohms, and A, B, C are coefficients to be determined. A set of coefficients was determined for each of the eight thermistors and used to convert resistance readings to equivalent temperatures. With individually calibrated thermistors the probe resolution was better than 0.002°C and the accuracy about 0.02°C.

DATA ANALYSIS

The temperature data collected over a period of three days for sites E, F, and D are plotted in Figures 5, 6, and 7 respectively. The beginning times for these records were 12:00 noon November 4 for well E, 3:40 p.m. November 10 for well F, and 5:50 p.m. November 15 for well D. An initial inspection of the zero depth record for each well reveals a roughly periodic waveform with short day time peaks and long night time troughs. The uneven peaks and troughs result from the short days and long nights of late Fall. The zero depth record for site E is relatively consistent from day to day as a result of calm weather and sunny days. This is not the case for site F where measurements occurred during a period of cloudy days and blustery weather. Thus, the diurnal amplitude is subdued and the waveform less regular. The records for site D were taken during a two day warming trend following a storm with a third day of partly cloudy weather.

The temperature-time records for the deeper thermistors reveal the amplitude

attenuation and phase lag with depth expected from theory. For each site, the surface temperature variations become increasingly subdued with depth until they are no longer visible at the 0.5 m depth. The stronger decay of the higher frequencies is also apparent in Figures 5, 6, and 7 as the curves become increasingly smooth with depth. The phase lag can be seen by inspecting the shift of the diurnal peak with depth. For example, the temperature peaks at time 25 hours in the zero depth record for site E (Figure 5) but peaks at about 26 hours and 28 hours in the 0.05 m and 0.10 m depth records respectively.

Amplitudes and phases for the frequencies with periods of 3, 1 and 1/2 days calculated from the 4320 point discrete Fourier transform of the temperature-time series obtained for each site are listed in Table 2. The number of points used in the transform results from three days of recording with one temperature measurement per minute. Information from depths greater than 0.25 m was not included as the thermal waves with these frequencies decayed too rapidly to be recognized below the 0.25 m depth. Other frequencies were not listed because they were either too weak or decayed too rapidly to be sufficient for thermal diffusivity calculations.

Plots of temperature averaged over the three day measurement period vs. depth for each site are shown in Figure 8. The expected late Fall annual temperature effect is quite apparent in the plots for all the sites. At site D, the presence of an additional temperature effect is noted by the break in gradient at the 1 m depth. This effect is superimposed on the annual signal causing it to be less apparent. The break in geothermal gradient suggests that the additional temperature effect could be due to a large change in thermal conductivity at about 1 m depth. Thermal conductivity measurements using drill chips collected at the site indicate an increase of about 50% between samples above and below 1 m. However, an increase of about 360% is needed to explain the observed change in gradient. Further complications may be attributed to groundwater disturbance. The temperature log from the deep hole D in the study conducted by Murphy and Gwynn (1979) (see Appendix A) shows an extremely high

TABLE 2

Amplitudes and phases of selected frequencies for depths down to 0.25 m as determined from the temperature-time measurements shown in Figs. 5, 6 and 7.

Frequency (cycles/day)		1/3		1		2	
Site	Depth (m)	Amp. (°C)	Phase (rad)	Amp. (°C)	Phase (rad)	Amp. (°C)	Phase (rad)
D	0.00	2.42	-2.35	6.95	-1.24	3.49	-2.41
	0.05	1.37	-1.95	2.68	-0.77	1.20	-1.84
	0.10	0.94	-1.69	1.31	-0.34	0.50	-1.31
	0.25	0.29	-1.02	0.17	1.11	0.29	0.20
E	0.00	2.83	-2.12	8.08	0.36	4.00	0.45
	0.05	1.57	-1.87	3.68	0.86	1.51	1.14
	0.10	1.13	-1.74	1.96	1.30	0.66	1.80
	0.25	0.38	-1.21	0.33	3.02	0.08	4.14
F	0.00	2.00	-0.72	3.25	-0.40	1.86	-1.30
	0.05	0.91	-0.18	1.36	0.10	0.66	-0.57
	0.10	0.65	0.25	0.74	0.56	0.29	0.03
	0.25	0.50	1.16	0.23	1.80	0.07	1.70

gradient down to about 15 m and a temperature reversal below 20 m. This behavior is indicative of groundwater movement, most likely a shallow lateral outflow. Since the water table at the site of well D is only about 1 m deep (the depth at which the break in gradient occurs in shallow well D), it is conceivable that the inferred groundwater disturbance observed in deep well D has an effect up to this depth. Due to the complex nature of this site we exclude it from further shallow temperature analysis.

Thermal diffusivities as calculated from changes in amplitude and phase between consecutive depth measurements down to 0.25 m for sites E and F are listed in Table 3. Consecutive depths were used so that changes in the calculated thermal diffusivity with depth could be monitored and the two layer model for removing the annual temperature effect could be applied. Since only one period of the wave with frequency of one cycle per three days is represented in the data, thermal diffusivities calculated using this frequency are considered less reliable. Therefore, we have excluded some of these values from the calculated averages in Table 3. The value obtained from the change in amplitude of the wave with frequency of 2 cycles per day between the depths of 0.10 m and 0.25 m at site F was also deleted from calculation of the average as unreliable. The amplitude of this wave, which is weak at the surface, becomes nearly undetectable at the 0.25 m depth.

Inspection of the mean thermal diffusivities listed in Table 3 reveals two major deviations from the results we expect under ideal conditions. First, the thermal diffusivity generally increases with depth indicating that the wave amplitude attenuation is not purely exponential and the phase lag is not a linear function of depth. This result can be explained by the existence of layers of sediment with differing thermal properties. Drill records, which show the existence of a thin layer of topsoil overlying clay, substantiates this observation. Second, the mean thermal diffusivities as calculated from phase are larger than those calculated from amplitude with the exception of the deepest value. Although our analysis of convective heat transport did not indicate how

TABLE 3

Thermal diffusivity results calculated from amplitude and phase changes with depth at sites E and F.

Site	Frequency (cycles/day)	Amplitude (°C)	α from amplitude			α from phase		
			Δz_1	Δz_2	Δz_3	Δz_1	Δz_2	Δz_3
E	1/3	2.83	0.09	0.28	0.24	0.48	1.82*	0.97*
	1	8.08	0.15	0.23	0.26	0.38	0.45	0.28
	2	4.00	0.19	0.27	0.38	0.38	0.41	0.30
	Average		0.14	0.26	0.29	0.41	0.43	0.29
F	1/3	2.00	0.05	0.27	3.75*	0.11*	0.16*	0.33*
	1	3.25	0.12	0.25	0.59	0.37	0.42	0.53
	2	1.86	0.17	0.27	0.86*	0.34	0.49	0.59
	Average		0.11	0.26	0.59	0.36	0.46	0.56

notes: Δz_1 refers to the depth interval 0.00 m to 0.05 m, Δz_2 = 0.05 m to 0.10 m, and Δz_3 = 0.10 m to 0.25 m; α = thermal diffusivity in units of $\text{mm}^2 \text{s}^{-1}$, amplitude refers to the surface amplitude of the wave in °C.

* Less reliable diffusivity result

thermal diffusivities determined from the two methods would differ, we believe this discrepancy is caused by the downward migration of rainwater as the survey took place during a period of rainy weather. Further, we feel the deepest values show better agreement because the thermistors at 0.10 m and 0.25 m depth were in the clay below the layer of topsoil for both wells.

In correcting for the annual temperature effects at sites E and F, we also considered a two layer medium to model the propagation of the annual wave. Parameters necessary for this model, which consists of a topsoil overburden overlying a clay substratum, are listed in Table 4. Thickness of the topsoil was measured in the field and thermal conductivities were measured on a divided bar using drill chip samples and the technique of Sass et al. (1971). To increase the accuracy of thermal conductivity calculations, porosity measurements were made by determining the amount of water in a fully saturated sample. The samples were obtained from the field in a relatively undisturbed state using a soil sampler. Thermal diffusivity for the clay was assumed to be the average of reasonable values obtained from both amplitude and phase determinations between the 0.1 m and 0.25 m depths. Thermal diffusivity for the topsoil at each site was obtained by the method outlined in the theory. That is, the topsoil diffusivity was adjusted until the diffusivities at the three shallowest depth intervals as calculated from data generated with the two layer model matched those observed in the field. We chose to match the observed thermal diffusivities as determined from amplitude variations as we felt they were more reliable than those determined from phase variations.

Since the survey took place during the month of November, 1982, we obtained temperature records from the Department of Meteorology at the University of Utah for the entire year of 1982 to determine the annual air temperature time series. The average temperature for each month was calculated, and listed in Table 5. The discrete Fourier transform of these 12 points was computed to obtain amplitude and phase information for the six frequencies need to match these points. This information is listed

TABLE 4

Thermal conductivities (k_1 , k_2), thermal diffusivities (α_1 , α_2), and thickness (h) of topsoil overlying clay at sites E and F.

Site	k_1 ($\text{Wm}^{-1}\text{K}^{-1}$)	k_2 ($\text{Wm}^{-1}\text{K}^{-1}$)	α_1 (mm^2s^{-1})	α_2 (mm^2s^{-1})	h (m)
E	0.83	1.08	0.17	0.29	0.06
F	1.19	1.46	1.14	0.57	0.10

TABLE 5

Monthly temperature averages for the year of 1982 calculated from University of Utah meteorology data.

Month	T ($^{\circ}\text{C}$)
January	-1.1
February	1.1
March	5.6
April	8.3
May	14.4
June	20.0
July	23.9
August	25.6
September	17.8
October	9.4
November	3.6
December	0.0

in Table 6. The continuous temperature function produced from the Fourier series containing these frequencies was obtained and is shown in Figure 9.

Assuming the ground surface temperature at the survey site could be represented by this function, we calculated that part of the measured temperature due to the effect of this temperature wave for each thermistor depth at the time of measurement for both sites. We then subtracted these temperatures from the three day average to arrive at the corrected temperatures. Results of this process are found in Table 7.

For comparison, annual temperature corrections and corrected temperatures which were calculated assuming a homogeneous half-space are also listed in Table 7. The thermal diffusivities for these calculations were determined to be $0.23 \text{ mm}^2 \text{ s}^{-1}$ and $0.35 \text{ mm}^2 \text{ s}^{-1}$ for sites E and F respectively. These values were obtained by finding a single diffusivity value which provided a least squares best fit to the amplitude decay of the diurnal wave with depth.

Plots of corrected temperature versus depth for the two layer and homogeneous half-space models are shown for sites E and F in Figures 10 and 11 respectively. If in either the two layer or homogeneous half-space case we have properly corrected past surface temperature history and the subsurface thermal properties, then the respective plots should show the background thermal gradients caused by quasi steady state heat sources at each site of measurement. The plot of temperature-depth for the two layer case for site E in Figure 10 is nearly linear. However, the general increase in gradient with depth indicates we have slightly overcorrected for the annual temperature effect. The plot for the homogeneous half-space case in Figure 10 shows decreasing gradient down to the 2 m depth which changes to a large increase in gradient from 2 to 3 m depth. This behavior may indicate inadequacies in the homogeneous half-space model. Plots from both diffusivity models for site F are shown in Figure 11. The extreme non linearity of curve (a) corrected according to the two-layer model suggests that in this case the more complicated model may be inappropriate. Corrected curve (b) is

TABLE 6

Results from the Fourier Transform of the annual temperature data.

Frequency (cycles/year)	Amplitude (°C)	Phase (rad)
Mean	10.72	
1	12.52	3.08
2	2.17	-0.93
3	0.66	2.10
4	0.62	-2.83
5	0.49	0.81
6	0.03	3.14

TABLE 7

Results from the removal of the annual effects in wells E and F.

Site	Depth (m)	Mean T (°C)	Two Layer Model		Homogeneous 1/2 space	
			Annual T correction (°C)	Corrected T (°C)	Annual T correction (°C)	Corrected T (°C)
E	0.00	4.72	8.04	12.75	8.04	12.75
	0.05	4.38	7.38	11.75	7.45	11.83
	0.10	5.48	6.61	12.09	6.87	12.35
	0.25	8.15	4.43	12.58	5.40	13.55
	0.50	11.96	1.60	13.56	3.05	15.01
	1.00	17.09	-1.50	15.60	0.07	17.16
	2.00	22.16	-1.85	20.31	-2.20	19.96
	3.00	25.98	-0.51	25.47	-1.59	24.39
F	0.00	1.56	8.88	10.44	8.88	10.44
	0.05	1.83	8.41	10.24	8.43	10.26
	0.10	2.90	7.97	10.87	7.96	10.86
	0.25	4.84	4.28	9.12	6.7	11.54
	0.50	6.35	0.26	6.61	4.85	11.20
	1.00	8.97	-2.03	6.94	2.17	11.14
	2.00	12.21	-0.27	11.94	-1.34	10.97
	3.00	13.52	0.08	13.60	-2.02	11.50

reasonably linear.

Linearity of a shallow temperature-depth profile, however, does not guarantee that the corrected gradient is a true representation of the actual stationary temperature field. As a further check on our results, temperature-depth plots for the deep wells E and F from Murphy and Gwynn (1979) (presented in Appendix A) are compared to the shallow site results. The plot for deep well E shows a steep gradient of approximately $4.3^{\circ}\text{C}/\text{m}$ above 10 m depth. Below 25 m depth, the gradient is greatly reduced to about $0.32^{\circ}\text{C}/\text{m}$. The plot for well F shows a near surface gradient of approximately $1.2^{\circ}\text{C}/\text{m}$ decreasing to about $0.32^{\circ}\text{C}/\text{m}$ below 10 m depth. The geothermal gradients in shallow sites E and F, as calculated between the 2 and 3 m depths, are $5.2^{\circ}\text{C}/\text{m}$ and $1.6^{\circ}\text{C}/\text{m}$ respectively. The 2 and 3 m depths were used in the calculation of the gradient in the shallow wells because the temperatures at these depths are the least affected by the annual temperature variation. The difference between the shallow well and near surface deep well gradients is 20% for well E and 35% for well F.

Geothermal gradients were also calculated between the 2 and 3 m depths in wells E and F using the corrected temperatures obtained assuming a homogeneous half-space. These were found to be $4.4^{\circ}\text{C}/\text{m}$ and $0.63^{\circ}\text{C}/\text{m}$ for wells E and F respectively. The difference between these values and those from the near surface deep well gradients is 3% for well E and 47% for well F. All shallow gradient values and percent differences between deep hole and shallow hole determinations are listed in Table 8.

TABLE 8

Geothermal gradients comparison between shallow temperature results determined from two layer and the homogeneous half-space models with results from deep wells reported by Murphy et al. (1979)

Site	Grad. ($^{\circ}\text{C}/\text{m}$) Murphy et al.,	Grad. ($^{\circ}\text{C}/\text{m}$) (2-layer)	% diff.	Grad. ($^{\circ}\text{C}/\text{km}$) (homo. 1/2-space)	% diff.
(a) gradients determined using data between 2 and 3 m.					
E	4.6	5.16	-12%	4.43	+4%
F ₁	0.23	1.62	-604%	0.63	-187%
F ₂	0.30	1.62	-4.40%	0.63	-110%
(b) gradients determined using surface temperature of 10.7 $^{\circ}\text{C}$ and corrected 3m temperature					
E	4.6	4.93	-7%	4.57	+1%
F ₁	0.23	0.97	-322%	0.27	-17%
F ₂	0.30	0.97	-223%	0.27	+10%

DISCUSSION

The main problem to be overcome in shallow temperature surveys is the separation of the quasi steady state natural geothermal gradient and heat flux arising from a geothermal source from the principally transient noise. The largest contribution to the noise in this sense is the propagation of surface temperature fluctuations into the ground. This propagation is controlled by the thermal diffusivity of the medium which may vary both vertically and horizontally.

The general approach adopted in this study is to determine thermal diffusivity at each site by analysing a time series of temperature data collected at multiple depths over a recording period of several days. Observations were analysed in terms of two models: a homogeneous half space model and a two layer half space model. Results were tested against gradient deduced from nearby deeper wells at each site. The extremely high thermal gradient between 5 and 6 °C/m at site E was reproduced by the shallow measurements using both models. In contrast, lower thermal gradient of about 0.3 °C/m at site F was only reproduced satisfactorily by one variant of the homogeneous half space model.

The net effect of using the homogeneous half-space model in place of the two-layer model was to give lower corrected geothermal gradients. These lower gradients gave a better match between deep well and shallow well geothermal gradient determinations at well site E but increased the difference in these determinations at well site F. However, since thermal conductivity measurements and calculated thermal diffusivities indicated the existence of at least two distinct layers at both sites, we feel the homogeneous half-space model is less appropriate. Therefore, it is necessary to examine the two-layer approach to this problem in terms of possible sources of error.

The sources of error affecting the outcome of our results consist of two types: those that arise from inaccuracies in the model parameters and those that arise from the model being insufficient to describe adequately the propagation of the temperature

variations. We found by means of a sensitivity analysis that inaccuracies in any of the model parameters could produce significant errors in the results. Results for the 3 m corrections were found to be relatively insensitive to changes in the thickness of the top layer and the thermal conductivity contrast between the two layers. However, our method for determining the diffusivity of the top layer is quite sensitive to changes in these parameters. A change of 10% in thermal conductivity contrast produces a 20% change in the determined value of the thermal diffusivity for the top layer. Similarly, a 10% change in the thickness of the top layer gives a change of about 8% in diffusivity. This is unfortunate as we found the model results to be quite sensitive to changes in the diffusivity of both layers. We feel that our measurements for the thickness of the top layer were quite accurate; however, it is conceivable that we have 10% or possibly greater error in the measurement of thermal conductivities.

Further, it is unlikely that the thermal diffusivities determined from amplitude changes accurately represent the diffusivities we would observe with purely conductive heat transport which is a basic assumption to the development of our model. It is this assumption that leads to one of the sources of error resulting from insufficient modeling of the annual temperature wave propagation. We have shown in our analysis of convective heat transport that changes in phase with depth are no longer controlled by the same factor as changes in amplitude with depth. Diffusivities calculated from changes in either of these parameters represent the values we would obtain in a non-convective medium and should be viewed with caution if advective heat transport is suspected. Although the extent of this type of error cannot be accurately estimated without a better understanding of the effect convective heat transport has on the calculation of thermal diffusivities, we feel it is probably the most significant source of error for our results. Our measurements were made during a rainy season when the downward migration of rainwater, and therefore, convective heat transport was at a peak.

Another source of error results from the assumption that the second layer is semi-

infinite. Drill records suggest that there are more than two layers in each of the shallow wells. It is unlikely that these layers have equal values of thermal diffusivity; thermal conductivities from drill chip measurements vary between the layers.

Further errors of the second type could result from the assumption that the ground temperature at the site is the same as the air temperature about 50 km to the north and the fact that other long term temperature variations caused by changes in the weather are unaccounted for. Although we have made a correction for the annual surface temperature variation using 1982 weather records we have not tested whether the years prior to 1982 depart significantly from the assumed annual variation and whether they have a measurable effect on our analysis.

CONCLUSION

Three day temperature time series at multiple depths have been collected at three sites within a geothermal system at Crystal Hot Springs, Utah. The thermal data exhibit general features expected when the earth's surface is subjected to a roughly periodic temperature variation with time. The thermal signals observed below the surface were attenuated and phase shifted with respect to the surface temperature wave. For example the daily wave, which during the survey period had an amplitude between 3 and 8 °C at the surface, was attenuated to about 5% of its surface value and phase shifted by nearly 180° at a depth of 0.25 m.

An analysis of the temperature-time series was performed in order to determine in situ thermal diffusivity values and their possible lateral and vertical variations. These values are necessary to correct the shallow temperature-depth survey data for annual temperature effects. Thermal diffusivity values deduced from the time series data ranged mostly between 0.1 and 0.6 mm² s⁻¹, reasonable values for the particular soils encountered. Diffusivities determined from phase information were generally higher than diffusivities determined from amplitude information. There is a strong, but not unequivocal, suggestion in the diffusivity data of both vertical and horizontal heterogeneity.

Measured temperatures at two sites, averaged over the three day surveys, were corrected for the 1982 annual temperature wave. The temperature correction was between +8 and +9 °C for the surface temperature and between +0.1 and -2.0 °C at 3 m depth. Corrected thermal gradients between 2 m and 3 m depth were between 4 and 5 °C/m at site E and 0.3 and 1.6 °C/m at site F depending on the detailed method of analysis. These values compare reasonably well with the thermal gradients of 4.6 °C/m and 0.3 °C/m for sites E and F respectively determined from deeper drillhole data.

A more detailed survey using ten or more sites and longer monitoring times was originally planned, but delays in constructing the equipment together with an early snowfall precluded the completion of those objectives. Thus the applicability of using temperature-time series at multiple depths in order to determine in situ diffusivity patterns and to make the necessary large annual temperature wave corrections to raw survey data has not been fully evaluated. Comparison with thermal gradient data from deeper drillholes at Crystal Hot Springs indicates a good correspondence between the shallow and deep results, although the most sensitive test on the edge of a geothermal system where heat flow is about 0.5 W m^{-2} is yet to be made. Several sources of error including possible advective heat transport by percolating rainwater have been discussed; these should be investigated systematically in future studies.

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REFERENCES

- Andrews, C. B., and Anderson M. P., 1979. Thermal alteration of groundwater caused by seepage from a cooling lake. *Water Resources Research*, v. 15, p. 595-602.
- Birman, J. H., 1969. Geothermal exploration for ground water. *Geol. Soc. Am. Bull.*, v. 80, p. 617-630.
- Clauser, C., 1984. A climatic correction on temperature gradients using surface-temperature series of various periods. *Tectonophysics*, v. 103, p. 33-46.
- Kappelmeyer, O., 1957. The use of near-surface temperature measurements for discovering anomalies due to causes at depths. *Geophysical Prospecting*, v. 5, p. 239-258.
- Kappelmeyer, O., and Haenel, 1974. *Geothermics with Special Reference to Application*. Gebruder Borntraeger, Berlin, pp. 238.
- Kintzinger, P. R., 1956. Geothermal survey of hot ground near Lordsburg, New Mexico. *Science*, v. 124, p. 629-630.
- Lange, A. L., Pilkington, H. D., and Deymonay, J., 1982. Comparative studies of geothermal surveys in 3-meter and temperature-gradient holes. *Geothermal Resources Council Transactions*, v. 6, p. 1-4.
- Lee, T-C., 1977. On shallow-hole temperature measurements - a test study in the Salton Sea geothermal field. *Geophysics*, v. 42, p. 572-583.
- Lee, T-C., and Cohen, L. H., 1979. Onshore and offshore measurements of temperature gradients in the Salton Sea geothermal area, California. *Geophysics*, v. 44, p. 206-215.
- Le Shack, L. A., Lewis, J. E., Chang, D. C., 1977. Rapid reconnaissance of geothermal prospects using shallow temperature surveys. U.S. Dept. of Energy, Semi-Annual Tech. Rept. for contract EG-77-C-01-4021.
- Le Shack, L. A., Lewis, J. E., Chang, D. C., Lewellen, R. I., and O'Hara, N. W., 1979. Rapid reconnaissance of geothermal prospects using shallow temperature surveys. U.S. Dept. of Energy, Second Tech. Rept. under contract EG-77-C-01-4021. (also report for contract N00014-78-C-0535 for the Office Naval Research).
- Lindqvist, G., 1983. Heat flow density measurements in the bottom sediment of some lakes in north Sweden. Ph.D. dissertation, Lulea University of Technology, Sweden.
- Murphy, P. J., and Gwynn, J. W., 1979. Geothermal investigations at Crystal Hot Springs, Salt Lake County, Utah. Utah Geological and Mineral Survey Rept. for contract DE-A507-77ET28393 for the Dept. of Energy.
- Olmstead, F. H., 1977. Use of temperature surveys at a depth of 1 meter in geothermal exploration in Nevada. U.S. Geol. Surv. Professional Paper 1044-B.
- Parasnis, D. S., 1976. Effect of a uniform overburden on the passage of a thermal wave and the temperatures in the underlying rock. *Geophys. J. R. astr. Soc.*, v. 46, p. 189-192.

55

Poley, J. Ph., and Van Steveninck, J., 1970. Geothermal prospecting - - delineation of shallow salt domes and surface faults by temperature measurements at a depth of approximately 2 meters. *Geophysical Prospecting*, v. 18, p. 666-700.

Sass, J. H., Lachenbauch, A. H., and Munroe, R. J., 1971. Thermal conductivity of rocks from measurements on fragments and its application to heat flow measurements. *J. Geophys. Res.*, v. 76, p. 3391-3401.

Van Wiyk, W. R., (ed.), 1963. *Physics of plant environment*. North Holland Publishing Co., Amsterdam.

Appendix A

Temperature depth profiles from drillholes at Crystal Hot Springs.

Figure 19 from Murphy and Gwynn, 1979.

TEMPERATURE - °C

Figure 19

Appendix B

SENSITIVITY ANALYSIS

A sensitivity analysis of the two-layer model was conducted by perturbing each of the principal parameters in turn from a standard original model. The original model parameters are listed in Table B1. Table B2 gives the results of this analysis including: actual and percentage changes in parameter values, the temperature corrections (ΔT) for the 1982 annual wave computed at 2 m and 3 m in depths, and the effect of those temperature shifts on the gradient correction between 2 m and 3 m. Recall that the general magnitude of the gradient correction (i.e., 1.5°C/m) is smaller than the thermal gradient of 5°C/m at site E in the center of the geothermal system, but is greater than the thermal gradient of 0.3°C/m at site F. Thus the consequence of errors in parameter determination on the ultimate goal of accurately mapping the stationary geothermal gradient depends on the part of the field being mapped.

Several observations can be made from the sensitivity analysis results given in Table B2. First, errors of 10% in thermal diffusivity values lead to comparable order errors in the gradient correction between depths 2 to 3 meters. Equal negative and positive perturbations to the diffusivities produce different changes in the gradient correction because of the exponential attenuation factor and the trigonometric dependence on phase. Second, both the thickness h of the first layer and the ratio of thermal conductivities seem to be relatively insensitive parameters, at least in this particular simulation.

Appendix B

Table B1 Original Parameter Values

Parameter	Value
α_1	$0.23 \text{ mm}^2 \text{ s}^{-1}$
α_2	$0.58 \text{ mm}^2 \text{ s}^{-1}$
	0.08 m
k_1	$0.8 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$
k_2	$1.6 \text{ W m}^{-1} \text{ }^\circ\text{C}^{-1}$
k_1/k_2	0.5
	Nov. 1 (300 days)

Table B2 Sensitivity Analysis Result

perturbed parameter value	change from original	T @ 2 m	T @ 3 m	gradient correction 2 to 3 m	change in gradient correction
	(%)	($^\circ\text{C}$)	($^\circ\text{C}$)	($^\circ\text{C}/\text{m}$)	(%)
original model	—	-1.709	-0.253	1.46	—
$\alpha_1 = 0.21 \text{ mm}^2 \text{ s}^{-1}$	-10	-1.312	-0.039	1.27	-13
$\alpha_1 = 0.25 \text{ mm}^2 \text{ s}^{-1}$	+10	-2.035	-0.522	1.51	+3
$\alpha_2 = 0.52 \text{ mm}^2 \text{ s}^{-1}$	-10	-1.888	-0.390	1.50	+3
$\alpha_2 = 0.64 \text{ mm}^2 \text{ s}^{-1}$	+10	-1.538	-0.147	1.39	-5
$k_1/k_2 = 0.8$	+60	-1.754	-0.272	1.48	+1
$h = 0.1 \text{ m}$	+25	-1.716	-0.257	1.46	0

CAPTIONS FOR FIGURES

Figure 1. Computed subsurface temperatures resulting from a sinusoidal surface temperature variation with time: (a) temperature versus time at various depths, (b) temperature versus depth at various times. The plots are normalized in terms of A_0 (surface temperature wave amplitude), τ (period of the wave) and λ (wavelength; $\lambda = (\alpha \tau / \pi)^{1/2}$).

Figure 2. Attenuation of a sinusoidal thermal wave with depth and its dependence on thermal diffusivity α . Left scale shows annual wave properties; right scale diurnal wave.

Figure 3. Effect of thermal diffusivity on the propagation of an annual wave with mean value 10°C and surface amplitude 10°C . Background thermal gradient is 0.05°C/m . Three different thermal diffusivities 0.2 , 0.8 and $1.6 \text{ mm}^2 \text{ s}^{-1}$ are represented by curves a, b and c. Sets of curves are shown for time increments of $\tau/8$.

Figure 4. Location map of Crystal Hot Springs, Utah, showing well sites D, E and F together with temperature contours for a depth of 61 m. (from Murphy and Gwynn, 1979).

Figure 5. Temperature-time data for 8 depths at site E. Curve parameters represent results from the following depths: (1) 0.00 m, (2) 0.05 m, (3) 0.10 m, (4) 0.25 m, (5) 0.50 m, (6) 1.00 m, (7) 2.00 m, (8) 3.00 m. Start time is 12.00 hrs, Nov. 4, 1982. Temperatures were recorded every minute.

Figure 6. Temperature-time data for 8 depths at site F. Curve parameters represent results from the following depths: (1) 0.00 m, (2) 0.05 m, (3) 0.10 m, (4) 0.25 m, (5) 0.50 m, (6) 1.00 m, (7) 2.00 m, (8) 3.00 m. Start time is 15.40 hrs, Nov. 10, 1982. Temperatures were recorded every minute.

Figure 7. Temperature-time data for 8 depths at site D. Curve parameters represent results from the following depth: (1) 0.00 m, (2) 0.05 m, (3) 0.10 m, (4) 0.25 m, (5) 0.50 m, (6) 1.00 m, (7) 2.00 m, (8) 3.00 m. Start time is 17.50 hrs, Nov. 15, 1982. Temperatures were recorded every minute.

Figure 8. Mean temperature during the survey period versus depth for sites D, E and F at Crystal Hot Springs. Site locations are shown on Figure 4.

Figure 9. Air temperature from the University of Utah weather station for the year 1982.

Figure 10. Temperature versus depth at site E. Open circles represent mean temperatures observed during survey period; triangles (curve a) represent temperatures corrected for a two layer model; closed circles (curve b) represent temperatures corrected for a homogeneous half-space model.

Figure 11. Temperature versus depth at site F. Open circles represent mean temperatures observed during survey period; triangles (curve a) represent temperatures corrected for a two layer model; closed circles (curve b) represent temperatures corrected for a homogeneous half-space model.

Fig. 1b

Fig. 2

Fig. 3

Fig. 4

Fig. 5

Fig. 6

Fig. 7

Fig. 8

Fig. 9

Fig. 10

Fig. 11