GEOHERMAL EXPLORATION TECHNOLOGY ANNUAL REPORT 1978

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GEOHERMAL EXPLORATION TECHNOLOGY

The geothermal exploration technology program at LBL began in 1973 in response to an AEC desire to install a 10-MW geothermal demonstration plant in northern Nevada, where numerous hot springs exist. Scientists at LBL and the University of California were requested to assist with site selection by performing geological, geochemical and geophysical investigations in selected promising areas. Although the goal for a demonstration plant was later dropped by ERDA, LBL was requested to continue with the research, changing the emphasis from site selection to exploration technique development and demonstration. As a result, conventional as well as less standard exploration techniques were tested. Available state-of-the-art equipment and techniques were used, but where these were either unavailable or inadequate for the task, new hardware and computer methods were developed by LBL. Among these were instruments and techniques for sampling and analyzing spring waters, up-down counters (signal averagers) and synchronous detectors for

use in electrical-electromagnetic surveys, a modern seismic system, and various new computer techniques for data processing and interpretation. Although the northern Nevada work was concluded in fiscal year 1976 and numerous topical and site-specific reports have been issued since then, the geothermal exploration technology program continues.

The objectives of the program are:

- 1) to extend the state-of-the-art in geothermal exploration and reservoir delineation by developing new and improved exploration instruments, data processing/interpretation techniques, and exploration strategies; and
- 2) to apply the new technology by means of field surveys at geothermal areas, and to distribute the technology by means of well-documented case studies, demonstrations, and industry-coupled programs.

ELECTRICAL AND ELECTROMAGNETIC COMPUTER MODELING TECHNIQUES

K. H. Lee, D. Fridmore, M. Hoversten, and H. F. Morrison

INTRODUCTION

Since the beginning of the geothermal exploration technology project, computer modeling for electrical and electromagnetic (EM) geophysical methods has been an area of continuing interest and one in which considerable progress has been made. This work has been done by faculty, staff, and students in engineering geoscience, Department of Materials Science and Mineral Engineering, U.C. Berkeley, working through LBL.

The principal objective has been to optimize the use of electrical and electromagnetic techniques for geothermal exploration and reservoir delineation, transferring the knowledge gained to the geothermal industry through reports, journal articles, demonstrations, and consultations with industry representatives requesting advice and assistance. The basic effort has been in developing new or improved numerical methods for interpreting electric/electromagnetic survey data. Geothermal areas are often geologically complex and this has prompted the need to extend computations into the realm of two- and three-dimensional models. Our ability to deal with these more realistic models has allowed us to improve our evaluation of various techniques, to interpret survey results in a more rigorous manner than previously possible, and to apply new techniques for which adequate interpretational procedures have not been available.

Results of these studies have been directly and immediately applicable to many LBL geothermal

projects, notably the development of the controlled-source EM system (EM-60) and site-specific geothermal resource studies in northern Nevada; Cerro Prieto, Mexico; and Mt. Hood and Klamath Basin, Oregon.

Although we and others working in this area have achieved considerable success, numerous problems remain, including that of developing simpler, more cost-effective algorithms for geologically complex areas. This problem was raised in a workshop on modeling electrical and electromagnetic methods hosted by LBL in May 1978 (Lawrence Berkeley Laboratory, 1978). The workshop brought together active workers in the field of numerical and scale modeling. Participants reported on recent progress, including problems encountered in applying various modeling techniques.

Conclusions on the state of modeling techniques are as follows: The classical techniques give satisfactory solutions to the scalar electromagnetic problems and one-, two-, and possibly three-dimensional resistivity problems. That is, the solutions are relatively well documented in the literature and the problems can be solved at reasonable expense on available computers. Solutions to vector EM problems, such as dipole sources over a two- and three-dimensional earth, require significant computational resources when approached by classical techniques. Research into hybrid techniques, use of different basis functions, and special methods for certain models is needed for this class of problem.

ACCOMPLISHMENTS IN 1978

TM-Mode Plane Wave Scattering

A numerical solution for two-dimensional TM-mode plane wave scattering has been obtained using an integral equation technique (Lee and Morrison, in preparation). In conjunction with the TE-mode solution (Hohmann, 1970), the TM-mode solution obtained completes the scattering analysis for arbitrary plane waves incident on simple two-dimensional inhomogeneities.

A remarkable advantage of the integral equation approach can be found in a magnetotelluric study. For simple inhomogeneities, such as a thin dyke or some other small conductors buried in a homogeneous half-space, it is shown that the computing cost with this method is much lower than that required for the finite element solution (Ryu, 1971). However, costs of the two methods become effectively the same for a model whose cross-sectional area is approximately 10 squared skin depths, which in finite element analysis is equivalent to 40 square cells of 0.5×0.5 skin depths.

MT Modeling

Intensive MT modeling has been carried out for the interpretation of the Grass Valley MT data collected in 1977. The program we used is TEM, a finite-element algorithm written by Ryu (1971) and later modified by Lee et al. (1978).

The program first computes the scalar electric fields or scalar magnetic fields, depending upon the mode (TE or TM) of the incident plane wave. The transverse components are then numerically calculated using Maxwell's equations, from which we calculate apparent resistivity.

A short compendium of apparent resistivity pseudosections over a series of two-dimensional models have been prepared and distributed (Lee et al., 1978). Interpretation of the Grass Valley MT data is based on this compendium and a report is in preparation (Morrison et al., in preparation).

EM Scattering

A solution has been developed for the EM scattering by an arbitrary two-dimensional distribution of subsurface conductivity due to an oscillating magnetic dipole located on or above the earth's surface. The problem is essentially three-dimensional, but since the half-space geometry is two-dimensional, the electromagnetic problem can be reduced to two dimensions by Fourier transformation. A finite-element technique is used on the discretized two-dimensional grid.

Until now, important two-dimensional geologic models approximating fault contacts and large tabular bodies (dykes, ore zones, etc.) have not been successfully treated for dipolar source

fields. This is because existing three-dimensional modeling techniques, using either the finite difference or the finite element method, are extremely expensive to use. By reducing the model to a more tractable two-dimensional problem we will obtain needed solutions at lower cost.

A computer program, MDP2D (Lee, 1978), has been written and tested on a variety of simple models. The major restriction to this program is a common one for numerical techniques--the solution is somewhat questionable when high conductivity contrasts are involved in the model, i.e., $>10^3$. However, for the more realistic models where conductivity contrasts is $<10^3$, the program generates reasonably good results.

Three-Dimensional Finite Element Modeling

Work continued on the evaluation of a finite-element program for the modeling of magnetotelluric phenomena in three dimensions. The finite-element scheme used electric fields as unknowns (Pridmore, 1978). It was found that the convergence of the iterative scheme used to solve the system of equations was extremely slow. In its present form the program is satisfactory only for calculating magnetic fields over three-dimensional bodies, since the derivatives of the unknowns converge significantly faster than the unknowns themselves. A direct solution would overcome this difficulty, but would cost significantly more.

A hybrid modeling scheme that uses both finite elements and Green's functions (integral equations) is being investigated. The method, which is similar to that reported by Scheen (1978), uses finite elements only in the region around the inhomogeneities. The secondary fields from the inhomogeneities at the edge of the finite element mesh are calculated using Green's functions. The technique can handle the same restricted class of models handled by the integral equation method; namely, models for which a Green's function can be calculated. The integral equation method yields a full matrix and the hybrid method gives a sparse matrix but requires many more evaluations of the Green's function. Initial computations suggest that the hybrid scheme will be more effective than the integral equation method for earth models where a large compact region is inhomogeneous. The algorithm is being written to facilitate modeling of an extremely important geological problem, that is, a good conductor beneath an irregular conducting overburden.

Frequency and Time Domain Prospecting Systems

In electromagnetic prospecting, measurements can be made in one of two ways. One is the frequency domain, where data are measured as a function of transmitter frequency. In this case measurements are made of the total field, composed of a primary field (due solely to the transmitter) and secondary fields (due to induced currents in the ground). The second possible

measurement domain is the time domain, where secondary fields, which are produced from eddy currents induced while the transmitter is on, are measured as they decay after the transmitter has been shut off.

The question of which measurement domain is best suited for detecting and resolving particular targets in the presence of geologic noise has been and continues to be a topic of much debate within geophysical circles. In order to address this question, we have developed a computer program to calculate the time-domain response of a layered earth to various commonly used primary wave forms. Using this program, together with our earlier programs that calculate the frequency domain response of a layered earth, we are investigating how changes in an earth model translate to change in frequency and time-domain response. Particular emphasis is on defining which changes can be detected with presently available field equipment.

In addition, the usefulness of both types of data in resolving a particular model, in the generalized least-squares sense, will be investigated using inversion techniques.

ACTIVITIES PLANNED FOR 1979

Numerous MT data have been collected and processed by our group in the past. These data will be interpreted with the help of the existing TEM program and the integral equation solution TEMINTG. We also expect to document these programs for the public in the immediate future.

Another important plan for 1979 is to analyze the ground EM (EM 60) data. This requires an extensive two-dimensional as well as three-dimensional EM modeling. Program MDP2D will be used for the two-dimensional analysis.

In addition to the planned interpretation of field data, there will be a substantial effort in the development of EM modeling programs. We are working on a hybrid EM modeling technique for the three-dimensional scattering problem. The grounded electric dipole source over a two-dimensional half-space will be implemented in

the program MDP2D in order to increase the versatility of the program.

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MINICOMPUTER FOR IN-FIELD PROCESSING OF MAGNETOTELLURIC DATA

R. H. Koch, W. M. Goubau, T. D. Gamble, R. F. Miracky, and J. Clarke

INTRODUCTION

Magnetotelluric (MT) data consist of records of the natural, low-frequency time-varying magnetic $[\vec{H}(t)]$ and electric $[\vec{E}(t)]$ fields measured at the surface of the earth. From this information we hope to define the electrical properties of the subsurface through the impedance tensor $\underline{Z}(\omega)$ which is defined as

$$\vec{E}(\omega) = \underline{Z}(\omega) \vec{H}(\omega),$$

where $\vec{E}(\omega)$ and $\vec{H}(\omega)$ are the Fourier components of $\vec{E}(t)$ and $\vec{H}(t)$ at the angular frequency, ω . The determination of $\underline{Z}(\omega)$ from the raw data requires considerable data processing. In the past we acquired the raw data in analog form on FM magnetic tapes which then had to be returned to the laboratory for processing. Data tapes were first digitized and played back for visual editing so that obviously bad sections could be deleted. The acceptable data blocks were divided into time segments that were Fourier transformed, and the Fourier components averaged over many time segments. All of the average crosspowers and autopowers of the fields required for the determination of $\underline{Z}(\omega)$ and the errors in $\underline{Z}(\omega)$ were then computed. For our MT surveys we typically occupy six to eight stations, taking 6 hr of data per site. Thus, the subsequent data processing, particularly the digitizing and visual screening, become very time consuming and tedious.

ACTIVITIES IN FISCAL YEAR 1978

We have designed and field-tested a battery-operated data acquisition system (Figure 1) for on-site processing of MT data. The system, which is the first of its kind, uses an LSI-11 minicomputer to digitize data, compute Fourier transforms, and compute average crosspowers and autopowers. The analog records are digitized in real time with a 12-bit analog-to-digital converter that has a selectable sampling period ranging from 4 ms to 6000 s. After 1024 samples of each field are taken, a fast Fourier transform is performed. All of the 15 possible crosspowers and six autopowers that can be formed from the horizontal components of $\vec{E}(\omega)$, $\vec{H}(\omega)$, and the reference magnetic field $R(\omega)$, are computed for each Fourier harmonic. The number of values of other crosspower and autopower is reduced from 1024 to 15 by band averaging the products contained within neighboring frequency windows of width $\Delta f \approx f_0/3$, where f_0 is the center frequency of the band. The resultant numbers are stored in memory. The fields are then resampled and the processing repeated; the new cross- and autopowers are then averaged with the previous values. The operator can manually abort a data segment so that bad data resulting from equipment failure or excessive noise will

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Figure 1. Mobile battery-powered digital data acquisition system showing the LSI-11 minicomputer with keyboard, cassette tape recorder, and thermal printer.

not be averaged with the good data already in storage. At the end of a recording session, the average cross- and autopowers are dumped into a cassette tape that is returned to the laboratory for the final calculation of $\underline{Z}(\omega)$, the statistical errors in $\underline{Z}(\omega)$, and the functions of $\underline{Z}(\omega)$ that are relevant to interpreting MT data.

We tested the data acquisition system over a two-week period during an MT survey near Milford, Utah. The system operated flawlessly throughout the survey.

PLANS FOR FISCAL YEAR 1979

In the future we hope to upgrade the system with additional input/output devices so that complete in-field processing will be possible. In this regard, we are preparing a program to normalize or correct all average powers for system sensitivities and variable gain settings. We will also adapt an existing program for in-field calculation of various parameters of the impedance tensor, for example, tensor rotation, skewness, and error bars.

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SUPERCONDUCTING THIN-FILM GRADIOMETER AND MAGNETOMETERS FOR GEOPHYSICAL APPLICATIONS

W. M. Goubau and J. Clarke

INTRODUCTION

Superconducting thin-film magnetometers and gradiometers have applications in various geophysical techniques. Recently a number of commercial and research organizations have begun using SQUID magnetometers for magnetotelluric studies because of the higher sensitivity and somewhat easier portability of SQUIDS compared with induction coil magnetometers. A three-component SQUID magnetometer is also used as the detector in the controlled-source electromagnetic studies conducted at LBL with the EM-60 system.

A superconducting gradiometer incorporating a SQUID can be used to detect very small changes in the gradient of an external magnetic field. Such devices have applications in sorting out local magnetic signals with a high spatial gradient in the presence of a large, but more uniform, background magnetic field. These devices also have potential applications in applied geophysics and for studying the characteristics of local noise effects on natural signals from magnetospheric and ionospheric sources.

ACTIVITIES IN FISCAL YEAR 1978

To have a better SQUID magnetometer for use with the EM-60 system, we decided that high sensitivity could be sacrificed to achieve a higher slewing rate, that is, the ability to track fields at higher frequencies. For this reason a planar, thin-film device similar to that shown in Figure 1 was built, tested, and incorporated into a complete magnetometer. This SQUID has a sensitivity of about $2 \times 10^{-4} \text{ V}/\sqrt{\text{Hz}}$ * and a slewing rate of $2 \times 10^{-4} \text{ G/sec}$ at 1 kHz. It is a factor of 20 less sensitive than our cylindrical SQUIDS, but has a factor of 3 higher slewing rate. The device is easy to fabricate and three have been mounted in an orthogonal array within a cryostat to make a three-axis magnetometer.

Research continues on the use of thin-film dc SQUID galvanometers for measuring magnetic field gradients. Figures 2 and 3 show the configurations of SQUID and SQUID assembly that have been built and tested in the laboratory. The particular gradiometer shown in Figure 3 was used in two configurations. When the Pb

(XBL 7710-6155A)

Figure 1. Configuration of thin-film magnetometer.

(XBL 7710-6155B)

Figure 2. Configuration of dc SQUID galvanometer.

strips were cut at points "A" the pick-up loops were in parallel, cut at points "B" they were in series. In either case, this gradiometer measured the off-diagonal gradient $\partial H_z / \partial x$. The gradient sensitivity was determined by measuring the response to a known flux imbalance

* $1 \text{ gamma} = 10^{-5} \text{ gauss} = 1 \text{ nanotesla}$.

Figure 3. Configuration of thin-film gradiometer.

between the pick-up loops. The sensitivities of the series and parallel-loop gradiometers were $3.7 \times 10^{-4} \gamma/m \sqrt{Hz}$ and $2.1 \times 10^{-4} \gamma/m \sqrt{Hz}$ respectively. These values were in good agreement with our theoretically predicted values.

PLANNED ACTIVITIES FOR FISCAL YEAR 1979

The lower sensitivity, higher slewing rate magnetometer will be tested in the field with the EM-60 system and direct comparisons of the two devices will be made.

Continued work is needed to design better gradiometers consisting entirely of thin films. In a future design, one would use photolithographic techniques to deposit films with a linewidth of about $5 \mu m$ and with a high degree of symmetry. Gradiometers made in this way

should have an improved initial balance in the z-direction, and also in the x- and y-directions. We believe that in the present design the x- and y-balance are limited by the distortion of the magnetic field by the overlapping films in the SQUID, and a reduction in the volume of these films should correspondingly reduce the distortion. Furthermore, if the film widths were reduced by a factor of 30, the resolution of the SQUID, which is proportional to (film width) $^{1/2}$, should be improved by a factor of about 5. With the aid of photolithography, one could also make second derivative gradiometers, which would be particularly suitable for medical applications. Finally we note that this type of gradiometer may be particularly suitable for commercial production, since it involves only thin-film techniques that are well established in certain industries.

MAGNETOTELLURICS WITH SQUID MAGNETOMETERS

T. D. Gamble, W. M. Goubau, R. F. Miracky, and J. Clarke

INTRODUCTION

Magnetotellurics (MT) is a widely-used technique for subsurface mapping and has applications in fundamental geological, geothermal, oil and gas and mineral investigations (Vozoff, 1972). The goal of MT is to determine accurately the electromagnetic impedance tensor $\underline{Z}(\omega)$ at the surface of the earth. $\underline{Z}(\omega)$ relates the components of magnetic fields ($\vec{H}$) resulting from disturbances in the ionosphere, and the components of the electric fields ($\vec{E}$) induced at the surface of the earth through the relation $\vec{E}(\omega) = \underline{Z}(\omega)\vec{H}(\omega)$. In the analysis of data $\vec{E}(\omega)$ and $\vec{H}(\omega)$ are treated as two-components column vectors and $\underline{Z}(\omega)$ as a 2×2 matrix of complex terms. The matrix may be rotated analytically into a geographic orientation for which $|\underline{Z}_{xx}|^2 + |\underline{Z}_{yy}|^2$ is a minimum. The resulting off-diagonal terms, $\underline{Z}_{xy}$ and $\underline{Z}_{yx}$, are usually unequal and are assumed to describe the impedances in the principal electrical directions, which depend on local geology.

From $\underline{Z}_{xy}$ and $\underline{Z}_{yx}$ it is customary to calculate apparent resistivity, $\rho_{ij} = 0.2T |\underline{Z}_{ij}|^2$, $ij = xy, yx$. Here ρ is expressed in ohm-m, T is the period in seconds, and $\underline{Z}$ is in (mV/km)/Y, where $\gamma = 10^{-5}$ G.

One standard way to estimate $\underline{Z}$ has been to express it in terms of cross- and autopower of the horizontal, orthogonal components of $\vec{E}$ and $\vec{H}$ (Madden and Nelson, 1964). For example, one could write

$$\underline{Z}(\omega) = [EH][HH]^{-1} \quad (1)$$

where $[AB]$ is the matrix of auto- and cross-spectral terms

$$[AB] = \begin{pmatrix} \overline{A_x B_x^*} & \overline{A_x B_y^*} \\ \overline{A_y B_x^*} & \overline{A_y B_y^*} \end{pmatrix} \quad (2)$$

Unfortunately, such an expression for $\underline{Z}$ contains positive definite noise contributions which do not average to zero regardless of how many data are acquired and analyzed, and can cause substantial bias errors in estimates of $\underline{Z}$, especially at frequencies at which the magnetotelluric signals are small (Goubau, et al., 1978a).

ACCOMPLISHMENTS IN FISCAL YEAR 1978

Although various workers have recognized the problems and have proposed methods for eliminating bias errors, we have derived and tested a simple solution involving simultaneous measurements of the magnetic field at a remote site as a reference for the local signals (Gamble, et al., 1978a). We can then express the impedance tensor as $\underline{Z}(\omega) = [\underline{ER}][\underline{HR}]^{-1}$, where $\underline{R} = (R_x, R_y)$ is the remote magnetic field. Provided there is no correlation between local signals and the noises in the remote signals, this estimate for $\underline{Z}$, which we designate as $\underline{Z}^R$, will be unbiased. Any crosspowers involving noise contributions will average to zero in the limit of an infinitely large data sample.

In the absence of bias errors, the accuracy of $\underline{Z}^R$ is determined only by random errors. These can be minimized by averaging over many data sets. An estimate of these errors should be made, and confidence limits placed on all quantities derived from $\underline{Z}^R$. In the past, no rigorous error analysis applicable to MT was available. We have found that such an error analysis is in fact possible for the remote reference method (Gamble, et al., 1978b).

The error $\underline{\Delta}$ in $\underline{Z}^R$ is given by

$$\underline{\Delta} = [\eta \underline{R}][\underline{HR}]^{-1} \quad (3)$$

where η is the total noise contributed by $\underline{E}$ and $\underline{H}$, defined by

$$\underline{\eta} = \underline{E} - \underline{Z}^R \underline{H} \quad (4)$$

The variance in the ij th ($i = x, y, j = x, y$) element of $\underline{Z}^R$ can be written as

$$\langle |\Delta_{ij}|^2 \rangle = \overline{|\eta_i|^2} \overline{|\underline{A}_j|^2} / N |D|^2 \quad (5)$$

where $\langle \rangle$ denotes an ensemble average,

$$\begin{aligned} A_x &= R_x \overline{H_y^* R_y} - R_y \overline{H_x^* R_x} \\ A_y &= R_y \overline{H_x^* R_x} - R_x \overline{H_y^* R_y} \\ D &= \overline{H_x^* R_x H_y^* R_y} - \overline{H_x^* R_y H_y^* R_x} \end{aligned} \quad (6)$$

and N is the number of terms in each average crosspower. The error $\underline{\Delta}$ tends to zero as N increases. Thus, for a sufficiently large collection of data we can find the error in any function, ξ , of $\underline{Z}^R$ by carrying out a Taylor

expansion to first order in Δ_{ij} and Δ_{ij}^* . The variance in ξ can then be written simply as

$$\begin{aligned} \text{Var}(\xi) &= \sum_{k=1}^4 G_{kk} \langle |\Delta_k|^2 \rangle \\ &+ 2 \text{Re} \left[\sum_{k=1}^3 \sum_{\ell=k+1}^4 G_{k\ell} \langle \Delta_k \Delta_\ell^* \rangle \right] \end{aligned} \quad (7)$$

where

$$G_{k\ell} = G_{\ell k}^* = \frac{d\xi}{dZ_k} \frac{d\xi^*}{dZ_\ell^*} + \frac{d\xi}{dZ_\ell^*} \frac{d\xi^*}{dZ_k} \quad (8)$$

and k and ℓ are labels for the xx , xy , yx , and yy components of $\underline{\Delta}$. With the change in notation $\Delta_k (k = 1, 2, 3, 4) \leftrightarrow \Delta_{ij} (i = x, y, j = x, y)$, $\langle \Delta_k \Delta_\ell^* \rangle$ can be expressed in terms of known cross- and autopowers of the fields according to the expression

$$\langle \Delta_{ij} \Delta_{nm}^* \rangle = \overline{\eta_i \eta_n^*} \overline{A_j A_m^*} / N |D|^2 \quad (9)$$

The confidence limits for ξ depend on the distribution of errors as well as on the variance. By the central limit theorem we know that the distribution will be normal for large N , independent of the distribution of noises in the fields. Thus, confidence limits are particularly easy to compute.

Our study of the statistics of the estimation of $\underline{Z}^R$ has produced other equations that should be valuable in many other applications (Gamble, 1978). The signal power-spectral matrix of the ionospherically generated electric signals, $[\underline{E}_s \underline{E}_s]$, can be estimated by the quantity

$$\underline{Z}^R [\underline{HE}] = [\underline{ER}][\underline{HR}]^{-1} [\underline{HE}] \quad (10)$$

Similar expressions hold for the magnetic and reference signal powers. Thus the ionospherically and locally generated signals can be separated and studied independently. Our ability to separate the contributions to the total signal permits us, for instance, to measure the noise power spectra of magnetometers in an unshielded environment.

During 1978 we have applied the remote reference technique to magnetotelluric surveys in Bear Valley, California; Grass Valley, Nevada; Cerro Prieto, Mexico; and Roosevelt Hot Springs, Utah. The area surveyed near Cerro Prieto is illustrated in Figure 1 (Gamble, et al., 1978c). Magnetotelluric measurements were taken at the locations numbered 1 through 7; the remote reference fields were measured at the position labeled R . Three-axis SQUID magnetometers were used to measure the magnetic field fluctuations, and the orthogonal electric fields were measured from the electric potential between pairs of buried copper-copper sulfate electrodes, 200 m apart. The data from the remote site were telemetered to the local site and all of the remote

and local signals were recorded simultaneously on tape. The tape was subsequently analyzed using the Lawrence Berkeley Laboratory CDC 7600 computer.

As an example of the results, the apparent resistivities, skewness, and strike angle as a function of period for station 6 are plotted with their probable errors in Figures 2 and 3. The skewness W , defined as $W = |(\underline{Z}_{xx} + \underline{Z}_{yy}) / (\underline{Z}_{xy} - \underline{Z}_{yx})|$, is a measure of the complexity of the conductivity of the earth. If W is large, the structure at the site must be three-dimensional at those frequencies. The apparent strike angle is the angle by which $\underline{Z}$ is rotated to minimize the quantity $|\underline{Z}_{xx}|^2 + |\underline{Z}_{yy}|^2$. We observe that the quantities plotted here are smooth functions of the period, and exhibit no observable bias errors in the region where the signal-to-noise ratios are smallest (around 10 sec period). The errors shown as the error bars for the 50% confidence level are generally very small. For example, in the apparent resistivities, 80% of the points shown have probably errors of less than 2.5%. The largest probable errors, up to 10% occur in ρ_{yx} at long periods for which the least quantity of data are available and where we normally expect problems in estimating the average power accurately. These data are typical of the high-quality results that have been obtained using the remote reference technique.

XBL 789-11107

Figure 1. Magnetotelluric sounding sites (1-7) near the Cerro Prieto geothermal field.

Beginning with the survey at Roosevelt Hot Springs, we are now able to perform in-field A-to-D conversion of all data sets and computation of all auto- and crosspower spectra by means of an LSI-11 computer mounted in the instrument rack.

XBL 7810-5963

Fig. 2. Apparent resistivities in the coordinate system aligned with the apparent strike direction, location 6.

XBL 7810-5957

Fig. 3. Apparent strike direction (relative to magnetic north) and skewness at location 6.

PLANS FOR FISCAL YEAR 1979

The data from Nevada and Utah are still in the process of analysis; however, considerable quantities of high-quality data were obtained at both sites. Both data sets will be completely analyzed and interpreted.

Our remote reference technique has now been tested at a sufficient number of sites, and we are confident enough in the technique to recommend it to commercial users of MT. We have found that the necessity of a second magnetometer and the need to telemeter data from the remote site to the local site has neither reduced the speed of field operations nor created new technical problems. The technique has been shown to yield unbiased estimates of the apparent resistivity of the ground, and of other quantities of interest in mapping out the true resistivity of the ground. Furthermore, our new error analysis enables one to obtain reliable estimates of the random errors in the data. We believe that, as a result, it will now be possible to considerably improve the accuracy of resistivity models of the ground based on MT data.

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CONTROLLED-SOURCE ELECTROMAGNETIC SYSTEM

H. F. Morrison, N. E. Goldstein, M. Hoversten, G. Oppliger, and C. Riveros

INTRODUCTION

In 1976, during geothermal exploration investigations in northern Nevada, tests were made of a controlled-source, deep-probing, electromagnetic (EM) system pieced together mainly with components on hand. Despite the primitive nature of the equipment, data were obtained that yielded sounding results that compared well with the results from other electrical techniques (Jain and Morrison, 1976; Jain, 1978). Because a properly engineered system would fill a technological gap in electrical prospecting, we received DOE approval in 1977 to begin design and fabrication of a large-moment system. This system, designated the EM-60 because it operates from a 60-kW motor-generator, was completed and tested in 1978.

The system is shown schematically in Figure 1. It consists of two sections: (a) a transmitter section consisting of the power source, control electronics, timing and a transistorized switch capable of handling large current; and (b) a receiver section consisting of magnetic or a combination of magnetic and electric field detectors, signal conditioning amplifiers and anti-alias filters, and a multi-channel programmable receiver (spectrum analyzer). Design goals called for a system that would be relatively inexpensive to duplicate, inexpensive and easy to operate in the field, and which would

also provide useable information over a broad frequency range with transmitter-receiver separations of up to several km.

A low frequency electromagnetic prospecting system

XBL 786-2575

Figure 1. General layout of the EM-60 electromagnetic system.

Because the system does not require wires other than the transmitter loop, there would be no long wires to lay down and retrieve as there are in galvanic dc resistivity surveys. Because electrical contact to the ground is not needed, the system is suited to areas where high resistivity and high contact resistances are troublesome problems. Further, the system would also have some advantage over conventional magnetotellurics (MT) in frequency bands where MT signals are typically weak and reliable results are sometimes difficult to obtain.

ACTIVITIES IN 1978

XBC 789-12515

Transmitter Section

The transmitter section, designed by Carlos Riveros of the Field System Group at LBL, was completed and tested. It is powered by a Hercules gasoline engine linked to an aircraft 60-kW 400-Hz, 3 ϕ alternator. These two components are mounted in the bed of a one-ton, four-wheel-drive truck. The output is full-wave rectified and capable of providing ± 150 V at up to 400 A to the horizontal coil. The square wave current pulses are created by means of a transistorized switch, which consists of two parallel arrays of from 6 to 60 transistors in interchangeable modules within the "crate." The crate is one box of two, the other containing the array driving electronics and timing, which pivot out of the truck during operation (Figure 2). The transmitter is operated by one man who controls the frequency of the primary magnetic field over the range of 10^{-3} to 10^3 Hz by means of switches on a remote control box which contains a crystal-controlled oscillator and dividers (Morrison, et al., 1978).

The dipole moment is determined by the resistance and inductance of the loop. At low frequencies, below 50 Hz, inductive reactance is negligible and the dipole moment is governed by the load resistance. Four turns of #6 wire in a square or circular loop, 50 m in radius, will yield a dipole moment of about 3×10^6 mks. At frequencies above about 100 Hz the inductance causes the moment to decrease and the current waveform to become quasisinusoidal. The current in the coil and the phase reference are obtained from a 0.01 Ω , 0.1% shunt resistor in series with the coil. The reference voltage is transmitted to the receiver by a twisted pair of wires, where it is used for calibration and phase reference. The hard-wire links between transmitter and receiver through the twisted leads is a temporary feature of the system and will be eliminated.

Figure 2. The EM-60 transmitter in operation in Grass Valley, Nevada.

Receiver Section

The key component in the receiver section is a six-channel programmable multifrequency phase-sensitive receiver (Figure 3). The receiver was designed by graduate student Gary Oppliger and built in the Engineering Geoscience Group, Department of Materials Science and Mineral Engineering, U. C. Berkeley. A three-component SQUID magnetometer served as the detector in field tests. Through the keypad on the receiver the operator is able to set the parameters controlling the signal processing, such as: (a) period of the current waveform; (b) maximum number of odd harmonics to be analyzed, up to 15; (c) number of cycles, N , to be stacked prior to Fourier decomposition in increments of 2^N ; and (d) number of input channels of data to be processed.

The receiver operates as an independent unit in the sense that it does not depend on a control signal from the transmitter. The two operate asynchronously; each is driven by a separate crystal clock, having a frequency accuracy of only 10 ppm. Phase measurements relative to the transmitter current are made by processing the transmitter current waveform along with the magnetic field signals. This provides sufficiently accurate phase measurements so long as one limits the number of stacked waveforms to less than several thousand. Unlimited stacking can be obtained by locking transmitter and receiver clocks together through the external 7.68-MHz clock input on the receiver panel or by using more accurate internal clocks.

In operation the receiver provides the amplitude and phase, relative to the primary

Figure 3. The programmable microprocessor-controlled receiver for the EM-60.

field, of each component processed. These data are printed out by a battery-operated thermal printer for later analysis in the laboratory.

Data Analysis

Field tests were conducted in Grass Valley, Nevada, and Mt. Hood, Oregon, but only data from the former tests site have been fully analyzed. Analysis requires the joint inversion of the amplitude and phase spectra of the vertical magnetic (H_z) and radial magnetic (H_r) components. These values, plus the standard errors for each as determined from a statistical analysis of repeated measurements, are used to derive the best layered-earth model fitting all data sets simultaneously. Figures 4 and 5 are examples of survey results. The observed data points are shown as squares and circles, and the calculated spectra are shown as the solid and dashed lines for the best three-layer fit to the data. In this case the transmitter-receiver separation was 2 km and the data covered the range 0.02 to 200 Hz. Estimated resistivities (ρ_i) are in ohm-m and layer thicknesses (h_i) are in meters. In this area the second-layer thickness is difficult to resolve accurately because the layer is thick and conductive.

PLANNED ACTIVITIES FOR 1979

The EM-60 will be improved by substituting more accurate clocks in the transmitter and receiver sections so that the twisted lead between transmitter and receiver can be eliminated.

Displays and demonstrations of the EM-60 will be held at various sites to hasten commercialization. Among these sites will be geothermal "test areas" where good subsurface information is available. Industry will be invited to examine the EM-60 and witness the system in operation.

Figure 4. Observed and calculated amplitude spectra for a transmitter and receiver 2 km apart on line E-E', Grass Valley, Nevada.

Figure 5. Observed and calculated phase spectra corresponding to Figure 4.

Further basic research needs to be done to improve data acquisition and interpretation. We plan to investigate techniques for obtaining reliable data at low frequencies where the induction effects from the earth are very small compared with the primary field and where they are small compared with the natural, geomagnetic "noise" whose amplitude follows, roughly, a $1/f$ dependency. We will investigate whether or not it is possible to recover data at the low-frequency end of the spectrum through the use of a noisecancelling reference magnetometer, combined possibly with measurements in the time-domain.

We are also planning to continue our efforts in data interpretation, which have not been carried beyond layered-earth models. Two-dimensional modeling is possible (Lee, 1978) and needs to be exercised in conjunction with the EM-60 results.

GEOTHERMAL SEISMIC FIELD SYSTEM DEVELOPMENT

T. V. MCEvilly, E. L. Majer, J. Bartschi, and J. Heinson

INTRODUCTION

The role of seismological methods in geothermal reservoir definition has been under investigation for several years as a part of the LBL geothermal exploration technology study. A primary goal of the research was to evaluate the several seismic techniques in general use for geothermal exploration in the early 1970s. Unlike other geophysical disciplines, seismology provides a variety of methods that can be used in exploration. Chief among these were microearthquake and ground noise surveys.

A microearthquake survey involved the deployment of portable seismographs over a prospect area. Typically, 6 to 10 seismographs were spaced 5 to 10 km apart to detect and locate small earthquakes (magnitudes of 2 to -1, generally). Such earthquakes--a sign of active stress release through faulting--were considered indicative of potential fracture porosity at depth, and possibly of anomalous thermal stresses caused by the heat. The relation between anomalous seismicity and a geothermal reservoir was much discussed but not resolved because there were too few case histories during the early stages of geothermal exploration technique development.

The use of ground noise, or the ambient microseism level, in geothermal exploration involved mapping the amplitude and frequency variation of the ubiquitous background motion over the surface of the prospect, usually with a sampling density of one station per km. The physical basis for this technique rests on the possibility that the geothermal reservoir is a dynamic system radiating seismic energy generated by a hydrothermal or thermoelastic process.

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Anomalous amplitudes or frequencies would thus be detectable on the surface over the reservoir. At the onset of our study, there was also spirited controversy over the viability of this method. The few reports in the open literature gave differing viewpoints.

These two methods constituted the main thrust in what was termed "passive" seismic exploration; that is, they both involved measurements of naturally occurring seismic waves, either discrete local earthquakes or the microseismic background. A third passive approach involved recording waves from distant earthquakes or explosions above the suspected reservoir zone in a search for perturbations in amplitude or arrival time. These perturbations could then be related to local anomalies by the attenuation or velocity characteristics of the subsurface. Although we have recently shown that this approach has great potential, it was not in general use by the geothermal exploration community at the beginning of our study.

Another class of seismological methods are termed "active." In general use in petroleum exploration, engineering seismology, and crustal structure studies, these methods use controlled sources, such as precisely timed explosions, and specially placed arrays of sensors. The methods include the well-known reflection and refraction profiling techniques for subsurface structure determination. Despite their highly successful application in other studies, these methods were not felt to be demonstrably practical in geothermal reservoir delineation.

Uncertainty surrounding the use of seismology in geothermal exploration was based on a lack of case histories demonstrating success and

an absence of data on the physical properties of the reservoir, which would provide a model of the target. In addition, the high cost of seismic surveys, particularly active profiling, requires solid justification in a cost-benefit sense. In contrast to petroleum exploration, for example, there are no data on which to base an estimate of ultimate yield per dollar of exploratory effort.

Our research task thus became to evaluate seismological exploration methods for geothermal reservoir detection and delineation and to reduce and simplify field procedures so as to enhance cost-effectiveness. Technique evaluation has required the collection of high-quality data from a variety of geothermal environments. A radio-telemetered 12-station portable seismic system was fabricated. It featured battery-powered central recording on slow-speed FM magnetic tape. Detailed surveys have been conducted with this system in several geothermal regions: the northern Nevada high heat-flow region with classic basin-and-range tectonics (Beyer et al., 1976); The Geysers geothermal field in the northern California coast ranges (Majer and McEvilly, 1979); the Cerro Prieto, Mexico, geothermal field in the Imperial-Mexicali Valley (Majer et al., 1978); and the youthful volcanic caldera at Cerro la Primavera in central Mexico (Majer et al., 1978). In addition, data were processed and analyzed routinely from the U. S. Bureau of Reclamation seismic network surrounding the East Mesa geothermal field in the Imperial Valley (McEvilly and Schechter, 1977a, 1977b, 1978).

Results and analyses of these studies have been published in various forms such as Ph.D. theses (Liaw, 1977; Majer, 1978); LBL reports, UCID reports, and scientific journals (Liaw and McEvilly, 1977, 1979; Majer and McEvilly, 1979). Our general conclusions are that carefully designed field experiments have detected anomalies in microearthquake activity, seismic wave velocities, and seismic wave attenuation at one or more of the sites. At The Geysers field, anomalous velocity and attenuation appear to define the production zone. As conventionally obtained, ground noise data were found to be of little value, reflecting very shallow structure or surface hydrothermal activity.

However, an experiment for imaging on possible radiation from a deep source is being implemented by industry in at least one potential geothermal region, so the possibility of true reservoir noise should be tested soon. We have seen no indication that earthquakes associated with geothermal reservoirs possess unique characteristics. That is, we cannot define the "geothermal" earthquake. On the other hand, of the five structures surveyed, only the two producing fields exhibited seismicity clearly associated with a proven or suspected reservoir. There is some evidence that exploitation may affect seismicity, thus providing an indicator of field dynamics during production. At the same time the question of induced seismicity is introduced. At The Geysers, the earthquakes are quite shallow and appear to be associated

with the field margins, with low seismicity in the main production zone. These features, however, can be detected only with a very dense station network with spacings of 1 km or less.

ACTIVITY IN FISCAL YEAR 1978

From the evidence and conclusions presented above, it was clear that several seismological methods could indeed provide valuable information in a geothermal exploration program. Of equal or greater importance, they offered potential for precise monitoring techniques on reservoirs under development. However, from our experience in collecting and processing the data, it was clear that there is a need for a less labor-intensive (and therefore more economical) data acquisition and processing system. Working with a triggered low-power digital recorder at The Geysers demonstrated the time-saving value of recording only events of interest (up to 100 per day can occur), and discarding the remaining background data. Another two orders of magnitude reduction in processing costs would result if the field system could automatically reduce the significant events from entire time series to arrival times and amplitudes of phases. Therefore, the seismologist could spend more time improving and developing new techniques rather than reducing data.

A system concept was developed, based on available components, which would transfer the maximum possible degree of data processing and reduction to the field. Furthermore, the system could be used in any of the several seismic field methods we found useful in previous reservoir delineation studies. System conceptualization began in June 1978 and the prototype system design was completed in the fall. Figures 1 and 2 illustrate the concept. Individual low-cost microprocessors at each detector handle data acquisition, event recognition, phase identification and timing, amplitude and polarity determination, and spectral calculations and parameterization. Operated as a microearthquake array, the system could have up to 128 channels, each microprocessor simultaneously scrutinizing its individual input data. Information contained in the detected events would be reduced to less than 10 diagnostic parameters, which are transferred to the master system microcomputer for analysis. All the standard results are thus available in the field so that the seismologist can optimize the survey parameters and scope. Our initial study indicated that it should be feasible to perform hypocentral locations, rupture parameter estimations, and a modest beam-forming search—all with sufficient speed to keep up with an earthquake sequence at a rate of one event every few minutes. Power consumption can be held to about 1 W per channel.

Encouraged by these exercises using device specifications, a two-channel prototype system was designed for a feasibility study using real hardware and software. The RCA 1802 CMOS processor, operating in 16-bit precision with hardware multiply, was selected. Parts were ordered,

XBL 7812-14137

Figure 1. Block diagram, two-channel prototype seismic data acquisition/processing system, showing card structures for master and channel modules.

XBL 7812-14138

Figure 2. Generalized block diagram of 128-channel seismic field system, illustrating 16-channel modular architecture.

Figure 3. Example of detection/timing algorithm operation on six earthquakes exhibiting a range in quality of arrivals.

design of the prototype system was completed, and hardware fabrication was completed in September 1978.

PLANNED ACTIVITIES FOR FISCAL YEAR 1979

Software for channel and master processing will be developed on the two-channel prototype. Simultaneously, a 16-channel operational field system will be packaged for testing both with on-line network data at U.C. seismographic station, and with previously recorded analog tape data on hand from earlier geothermal field studies.

In the first quarter of the year, the detection, timing, and amplitude measuring algorithms will be defined and programming will begin on the 1802 channel devices. Our algorithm has been developed with simplicity in mind. Drawing on the results of Stewart (1977), Allen (1978), and Anderson (1978), and retaining the long-term to short-term average comparison (Ambuter and Solomon, 1974), a fast, effective detection and timing algorithm has been developed, which can operate in real time. Results of preliminary tests are shown in Figure 3, where event detection, phase (P and S) identification, and phase timing (the vertical bars) are reproduced for a set of average character microearthquakes. Top trace is original seismic signal, sampled at 200 sps. Middle and lower traces are the detection function with 16- and 32-point short-term averaging windows, respectively. Vertical lines show times picked for phases detected. The first phase is the P-wave arrival followed by the S-wave arrival. A 2048-point mean is removed from the time series initially. The P-wave detection threshold in all cases is three times the (2048-point) long-term average (LTA). the P-timing level is 1.25 LTA, 2.0 LTA, and 1.5 LTA in a, b, and c, respectively. P-timing in b and c is an extrapolation to LTA based on the slope of the detection function. The S-wave is detected using a level of two times PAVE, where PAVE is an average amplitude of 64 points following the P-wave detection. The S-wave is timed in several ways: by extrapolating back to LTA, to 2.0 PAVE, and to 1.0 PAVE, in a, b, and c, respectively. The parameters are selectable in the algorithms. Those results judged best for timing and detection are a for the P-wave and c for the S-wave.

During the second quarter, the basic channel device software will be implemented, the master device software will be tested, and the 16-channel system will be ordered. In the third quarter, the two-channel prototype will be exercising the entire basic software package, and fabrication of the 16-channel system will be under way.

During the last quarter, the 16-channel system will be exercised on real data. This will be the conclusive test. Final recommendations for system specifications will be made and the results disseminated to the exploration and seismic instrumentation industry. It is probable, however, that thorough testing of the system will be necessary in actual field use

before general acceptance of the system. In fiscal year 1980, it would be valuable to make a 32- or 64-channel system available for field tests at various geothermal prospects. This would also provide much needed additional high-quality seismic data for case histories.

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INTRODUCTION

of Klamath Falls. The waters range in temperature up to 113°C, and several companies and government agencies have searched for hotter water or steam suitable for electric power generation. Two unsuccessful deep holes have been drilled thus far in search of a high-temperature resource outside the immediate area of Klamath Falls, and the level of interest in the area has consequently declined over the last two or three years.

- Known geothermal resources area boundary
 --- Telluric-magnetotelluric survey area (Geonmics 1977)
 ----- Aeromagnetic and gravity survey area (Van Deusen 1978)
 ----- Roving dipole survey area (Group Seven, 1972)
 ----- Audio-magnetotelluric survey area (Senterfit & Bedinger, 1976)

In an attempt to restimulate exploration activity, LBL, working with the State of Oregon's Department of Geology and Mineral Industries (DOGAMI) and the U.S. Geological Survey (USGS), has attempted to collect, integrate and reinterpret all available exploration data from both public and private sources. Our primary goals were to develop a conceptual model for the hydrothermal system, to suggest targets for further exploration and to evaluate the exploration techniques used on a site-specific basis (Stark et al., 1979).

The Klamath graben is a dominating geological feature; it can be traced for 80 km from the Medicine Lake volcanic field in northern California to Crater Lake, Oregon, attaining a width of 15 km. The local geologic structure is characterized by numerous northwest-trending normal faults, with occasional northeast-trending cross-faults. Important rock units are late Tertiary and Quaternary volcanics and a late Tertiary lacustrine unit known as the Yonna formation.

ACCOMPLISHMENTS FOR 1978

We were able to gather, from the open literature and from companies working in the area, results of magnetotelluric (MT), audio-magnetotelluric (AMT), roving dipole resistivity, aeromagnetic, and gravity surveys. As shown in Figure 1, the survey blocks tend to cluster into two areas--the Swan Lake Valley area northeast of Klamath Falls and the Klamath Hills area, south of Klamath Falls.

For the Swan Lake Valley, we concentrated on reinterpreting the magnetotelluric data by means of two-dimensional modeling. Our models represent geoelectrical cross-sections beneath a profile line trending N. 47° E. across the northern portion of the Swan Lake Valley. The two-dimensional modeling technique is a significant improvement over conventional one-dimensional interpretation schemes for geologically complex areas such as the Klamath basin. However, it proved difficult to match the transverse electric (TE) and the transverse magnetic (TM) apparent resistivity modes with data calculated for two dimensional models. This problem, common in MT interpretation, is caused by the sensitivity of the method to lateral inhomogeneities and three-dimensional electric structure.

By trial and error, we obtained several models, each giving a reasonable fit to at least one of the modes. Moreover, these models are also consistent with interpretations of roving dipole, gravity and magnetic data in the area.

Two conductive zones, ranging in resistivity from 3 to 9 ohm-m, appear in the model (Figure 2). One is located beneath Meadow Lake Valley, about 1 km southwest of Swan Lake Valley. The other is located beneath the northwest margin of Swan Lake Valley. The depth extents of these zones are uncertain, but the proximity of the zones to major normal faults makes them possible targets for further exploration. A conductive layer lies beneath the Swan Lake Valley at a depth of 1 km, connecting the two zones.

Figure 2. Resistivity model for Swan Lake Valley area with schematic topography (values in ohm-meters).

We have recommended an exploration strategy for the Swan Lake area consisting of resistivity soundings and gravity measurements, geochemical analyses of spring and well samples, and shallow thermal gradient drilling.

To the south, the Klamath Hills area has attracted interest because of two hot wells (>90°C) and a handful of warm wells located immediately southwest of the Hills. Figure 1 shows the areas covered by AMT, roving dipole, gravity and aeromagnetic data.

We reinterpreted some of the roving dipole data by means of two-dimensional resistivity modeling in the vicinity of the hot wells. Our final models indicate the presence of a conductive zone (4 ohm-m) around the hot wells, but it appears to be limited in both lateral and depth extent. The conductive zone probably represents valley sediments saturated with hot water that has ascended from depth along the fault zone bounding the Hills. We have suggested additional resistivity studies here to determine more accurately the extent of the conductor and to detect the source of the fluid at depth.

On the northeast side of the Klamath Hills, the AMT data suggest a linear northeast-trending conductive anomaly beneath the Hills and Olene Gap. We have interpreted this as a concealed cross-fault. The gravity and aeromagnetic data are consistent with this interpretation; the contours are aligned in a northeasterly trend, forming a saddle between the Klamath Hills and Stukel Mountain to the northeast. Thus it appears that the basement on the southeast side of the inferred fault has been upthrown, forming a raised northeast-trending ridge between Klamath Hills and Stukel Mountain.

The AMT anomaly is centered near the northwestern tip of Stukel Mountain, where several warm wells (up to 42°C) exist. The extremely low (<1 ohm-m) AMT apparent resistivities measured there are encouraging signs of geothermal activity. This area lies close to an intersection between several faults and the inferred concealed cross-fault. Fault intersections appear to be favorable locations for geothermal activity in this area. However, deeper electrical soundings are needed to determine the depth extent of this anomaly. We may be seeing the effect of leakage along fractured fault zones from a larger reservoir at depth, similar to the proposed mechanism near the hot wells.

In working with these data, we have had a chance to evaluate the geophysical methods on a site-specific basis. A brief assessment of each method follows.

1. Gravity and Magnetics. These methods seem suited for resolving structure, and are best used in conjunction with other techniques. They are useful for mapping faults and for helping to determine alluvium/sediment thickness in valleys.
2. Roving Dipole Resistivity. This can be a useful reconnaissance tool, but when used

without other resistivity methods it is difficult to interpret in a meaningful, quantitative fashion. Computer modeling shows that the anomaly source may lie some distance from the anomaly.

3. AMT and MT. These methods offer more promise, but are more difficult to interpret properly than the other techniques discussed. A major advantage of MT is that it often allows crustal modeling to depths of tens of kilometers. However, owing to local surface inhomogeneities and three-dimensional deep structure, it is rarely possible to satisfy both TE and TM data sets by means of a two-dimensional model. Scalar AMT, as performed by the USGS, or scalar MT surveys are usually difficult to interpret in multidimensional geology. However, in the Lost River-Spring Lake valley, the scalar AMT data seem to reveal concealed structures.

In general, the study demonstrates the need for comparison and simultaneous interpretation of multiple data sets. In complex geology, individual geophysical data sets often cannot be interpreted clearly, and simultaneous interpretation of several types of data can lead to a clearer understanding of subsurface structure. The study also has demonstrated the importance of thorough interpretation of field data, particularly for the electrical prospecting methods. This requires computer modeling, which may involve a significant amount of time and money, but is necessary if the techniques are to be useful.

PLANS FOR 1979

In September, 1978, we received geophysical and geochemical data covering hundreds of square kilometers of land on the west and east shores of Upper Klamath Lake. The geophysical data include roving dipole, Schlumberger equatorial, dipole-dipole, and Schlumberger expander soundings and logs from an unsuccessful 2000-ft exploratory well. Preliminary work has begun with computer modeling of the electrical data. We plan to continue this type of analysis on all of our data, hoping to gain a clear understanding of the geothermal system in the entire area around Klamath Falls.

The extensive geochemical data will permit investigation of regional patterns of elements in shallow and deeper groundwaters.

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MT. HOOD GEOTHERMAL RESOURCE EVALUATION

N. E. Goldstein, H. A. Wollenberg, H. Bowman, and E. Mozley

INTRODUCTION

Mt. Hood is a young but dormant volcano located approximately 80 km east of Portland, Oregon (Figure 1). Beginning in 1977, a project began to evaluate the geothermal potential of the volcano. The DOE project is coordinated by the Oregon Department of Geology and Mineral Industries (DOGAMI) and supported in part by the U. S. Geological Survey and the U. S. Forest Service. This volcano was selected from others in the High Cascade Range because of its proximity to a major city, the abundance of federally controlled land on which to work, evidence of a heat source, and growing interest of private developers in the geothermal potential.

As part of the program to evaluate the resource, various exploration techniques were applied to determine their usefulness in this complex geological environment. LBL has the responsibility of coordinating surveys dealing with rock, gas, and water geochemistry and electrical resistivity.

ACTIVITIES IN FISCAL YEAR 1978

Geochemical Studies

Water, rock and gas samples were collected on or near Mt. Hood during fiscal year 1978, with the cooperation of DOGAMI and USGS. Periodic sampling of water from selected orifices at Swim Warm Springs (Figure 1) continued with a collection in February 1978. Results of analyses of major and trace elements indicated increased

contents of magnesium and sulfur in winter samples, compared with samples collected the previous summer. It was expected that the winter sampling would yield higher contents of all elements, because surface runoff, diluting the hot water from depth, would be diminished. However, this was not substantiated because there were significant decreases, or only slight changes, in contents of some elements in the winter samples.

Sufficient data have been collected to provide some preliminary conclusions on the geochemical setting of Mt. Hood, which are given below. Descriptions of the sampling and results of analyses are detailed in LBL-7092 (Wollenberg, et al., 1979).

1. The warm water at Swim Springs on the south flank of Mt. Hood has a large component (~90%) of near-surface cold water. The flux of cold-water runoff nearly masks the surface indications of deeper-circulating hot water. The upper-slope orifices at Swim Warm Springs are the most diluted, those down slope are successively less diluted. The oxygen-hydrogen isotope data indicate that cold-spring waters have short pathways between their sources and the springs, although a component of the warm-spring water has its origin considerably higher on the mountain.
2. Chemical geothermometers indicate that temperature at depth of water entering the Swim Springs system is within the range 104-170°C. The silica mixing model gives a range of 190-240°C for unmixed hot water.
3. Water in the geothermal test hole in Old Maid Flat (see Figure 1), most likely from the Rhododendron formation, has a different chemical character than the water at Swim Warm Springs. The relatively low silica content of Old Maid Flat well water suggests that its temperature is due to the regional geothermal gradient and not to circulation within thermal anomalies associated with Mt. Hood.
4. Iridium, a platinum-group element, is present in identifiable abundance at one orifice at Swim Warm Springs, a cold spring, and in a small lake in the summit crater area. Iridium also occurs in altered andesite at one location on the western flank of Mt. Hood. The association of Ir with an andesitic volcano is surprising; it is generally considered to be associated with basic to ultrabasic igneous rock.
5. The similarity between major- and trace-element contents of Pliocene andesite and intermediate intrusive rocks of the Laurel Hill pluton on the west side of the mountain supports Wise's (1969) contention that the pluton represents the root zone of a center of andesite extrusion.

Figure 1. Simplified location map of the Mt. Hood area.

Figure 2. The array of sensors used for one set-up of the T-MT survey. Two such set-ups with the second tensor MT station over a previous remote telluric station constituted a station cluster.

6. The basalt exposed in the Robin Hood quarry in the valley of the East Fork of the Hood River has major- and trace-element contents very similar to those of the Roza Flow of central Washington. This supports the supposition that the basalt on the east side of the valley is Columbia River basalt, and that the north-south valley marks the trace of a normal fault, upthrown side to the east.

Geophysical Studies

The telluric-magnetotelluric survey begun in fiscal year 1977 continued to early November 1978 when worsening weather conditions forced the contractor, Geonomics, Inc., to cease Phase II survey operations. During Phase II the survey aperture was opened; station density was decreased and stations were arranged in five clusters around the mountain to increase regional coverage. Each cluster was planned to consist of two complete telluric-magnetotelluric set-ups plus a reference magnetic station (Figure 2). A complete cluster yielded six data stations; two tensor MT stations plus four remote telluric stations. However, in some clusters it was convenient to put the reference magnetometer at a remote telluric site, thus obtaining another tensor MT station.

Geonomics performed the conventional data processing and we were to reprocess data using the reference magnetometer channels to obtain results in the troublesome parts of the spectrum (Gamble, 1978). Using the Geonomics results only, a preliminary interpretation report was issued in June 1978 (Goldstein and Mozley, 1978). The general conclusions reached are as follows:

1. The change in principal resistivity directions at stations around the volcano suggest that the volcano may be located at the intersection of a north-south and east-west structural trend. However, we see no evidence for the postulated north-south fault following the East Fork of the Hood River.
2. Although it was not possible to put a station close to the summit (the method is dependent on roads and easy access), encouraging results were obtained at Cloud Cap, a 12,000+ y.B.P. eruptive center in the northeast side of the mountain. Conductive zones at depths of 600 to 1000 m and >2000 m were interpreted (Figure 3).
3. Anomalous low resistivities were also observed in the vicinity of the warm water emanations at the Still Creek campground (Swim Warm Springs). However, because of apparently complex local geology and incomplete data a full interpretation was not made.
4. Stations in the Old Maid Flat area, near the geothermal test well drilled by Northwest Natural Gas (Figure 1), show no suggestion of an encouraging low conductivity anomaly. Rather, resistivities increase with period, suggesting a decrease in porosity-permeability with increasing depth. This situation was

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Figure 3. A one-dimensional interpretation of MT data at the Cloud Cap station. R_{12} and R_{21} are resistivities in the principal directions determined by rotation of the impedance tensor.

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Figure 4. The EM-60 transmitter in operation near the Timberline Lodge, south flank of Mt. Hood, elevation 5900 ft. The most distant receiver station was at an elevation of 7000 ft, at the edge of the Palmer glacier seen directly above the truck on the left.

confirmed in 1978 when the hole was deepened from 2000 to >4000 ft.

To supplement the T-MT results, the controlled-source EM system (EM-60) developed at LBL was used at selected sites during August 1978. Transmitter sites were located near Timberline Lodge (Figure 4) for coverage on the south flank, at Summit Meadow for coverage at Swim Warm Springs, at Cloud Cap to confirm the MT anomaly there, and in Old Maid Flat for information at the deep drill hole. Two or three receiver sites were occupied relative to each transmitter except at Old Maid Flat where no data were taken. The data were returned to Berkeley for post-field processing and analysis, scheduled to begin early in fiscal year 1979.

ACTIVITIES PLANNED FOR FISCAL YEAR 1979

Geochemical Studies

Firmer conclusions on the usefulness of geochemistry in evaluating the geothermal resource potential of Mt. Hood await results of analyses presently under way. These include analyses of gas samples collected in October 1978 in the summit fumarolic area. In the 1977 sampling we were not able to collect sufficient water vapor to determine isotope ratios. The volume of water vapor collected in 1978 is sufficient for these analyses. This will afford comparison with the isotope ratios of cold- and warm-spring water.

Water analyses awaiting completion include a sample collected in October 1978 from the geothermal test hole drilled near Timberline Lodge and a set of samples collected at that time from

the three orifices sampled periodically at Swim Springs. The deepened geothermal test hole in Old Maid Flat may permit us to obtain a fluid sample from a deeper aquifer (most likely within the Columbia River basalt) than the Rhododendron formation sampled in 1977.

Samples of altered and unaltered andesite from the fumarolic area in the summit crater were collected in October 1978. These are being analyzed for major and trace elements and will indicate if iridium and silver are present in the strongly altered ground, coinciding with the presence of these elements in the runoff water from the small lake.

We recommend that one or more geochemical test holes be drilled in the Swim Warm Springs area. They should be deep enough to obtain temperature measurements and fluid samples from below the zone of near-surface and shallow cold-water runoff, to permit major, trace-element, and isotope analyses of relatively unmixed hot water.

Geophysical Studies

In November 1978 the complete reference magnetic MT processing program, including the error analysis subroutine (Gamble, 1978), were checked and found to be fully operational. Edited data tapes will be reprocessed to the full extent possible; and comparisons will be made to sort out errors in the remote telluric station results due to noise (bias errors) or a nonuniform magnetic field between base and remote stations. Following this, the reprocessed data will be reinterpreted in terms of a subsurface resistivity model.

The data collected at Mt. Hood using the EM-60 system will be processed and interpreted and the results will be compared and correlated with the T-MT survey results. The EM-60 data will require additional corrections to account for the elevation differences between transmitter and receiver, and a tilted transmitter dipole vector due to sloping terrain. Interpretation of the EM-60 data will be done in terms of the layered inversion approach (Morrison, et al., 1978) and in terms of two-dimensional models (Lee, 1978).

A dc resistivity (dipole-dipole) survey will be performed in the Swim Springs area.

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EAST MESA SEISMIC STUDY

T. V. McEvelly, B. Schechter, and E. L. Majer

INTRODUCTION

Magnetic tape records collected by the U. S. Bureau of Reclamation (USBR) at their six-station seismic network around the East Mesa geothermal field were analyzed. The purposes of this study are: (a) to obtain further details of the faulting and associated stress fields as defined by local earthquakes; (b) to obtain baseline data on local seismicity before geothermal field exploitation; (c) to determine if the reservoir region causes anomalous elastic wave propagation characteristics and whether or not this information is useful for defining the reservoir; and (d) to provide USBR personnel and others with procedural guidelines for subsequent analysis of network data.

ACCOMPLISHMENTS IN 1978

Following the first report (McEvelly and Schechter, 1977a), which detailed the data analysis techniques, two subsequent reports (McEvelly and Schechter, 1977b and 1978) were issued. These reports summarize the results of observations and one (1978), gives a detailed review of past and present local seismicity.

Figure 1 shows the locations of the USBR seismograph stations. The open circles are

epicenters of the larger events recorded: the primed numbers refer to epicenters located by means of the U. S. Geological Survey (USGS)-California Institute of Technology (CIT) network in the Imperial Valley. The magnitudes of these events ranged from $M_L = 2.0$ to 4.0. The USBR and USGS-CIT epicenters agree relatively well in computed distance, but have a discrepancy in azimuth. An azimuthal distortion may be caused either by local variations in seismic velocities within or near the USBR network or by refraction across a lateral discontinuity, possibly associated with the Brawley-Imperial fault system. Every locatable event was found to originate along the Brawley-Imperial fault. The location of the smallest events, magnitudes on the order of zero, are unresolved.

The primary finding to date in the East Mesa seismic study has been the complete lack of detectable local microearthquakes to an estimated detection threshold of at least $M_L = 1.0$. The detection of smaller shocks has been hampered by the high level of low-frequency surface noise in the field. These results are consistent with previous investigations in the Imperial Valley and the East Mesa area (Hill, 1975; Iyer, 1975), but inconsistent with other microearthquake studies at East Mesa in 1973 (Combs, 1974; Combs and Hadley, 1977).

Several possible explanations might account for the discrepancies:

First, the East Mesa fault was active in the period 1973-1975 but has been inactive during the present period of study (July 1977 to present). This does not seem likely, since local activity was assumed to be tectonically linked to the broader zone of seismicity associated with the Brawley-Imperial fault system, which has continued to be active throughout the period. According to this hypothesis, one would have expected to see local events during the recent Brawley swarms.

Second, the microearthquakes used to define the East Mesa fault in 1973 were actually located along the Brawley fault. Only vertical component instruments were then available to locate these events. McEvelly and Schechter (1977b) demonstrated the problems of accurately locating events using only P arrivals, especially those events located outside the array. A comparison of station deployment and the epicenter distribution shows the station distribution heavily weighted to the

Figure 1. Regional map showing East Mesa network station, epicenters located, and USGS-CIT epicenters for larger events.

east of the epicenters, a situation which could easily lead to mislocations. Relocation of these events farther to the west would also increase the computed focal depth to values more compatible with those reported by Johnson and Hadley (1976) for Brawley events. Only a reevaluation of the actual data can resolve this problem.

Third, "geothermal events" may not be seismic in origin. Illustrations of these events in Combs (1976) appear to be more consistent with air-coupled Rayleigh waves than with microearthquakes, featuring emergent low-frequency arrivals and similar waveforms which appear simultaneously on both vertical and horizontal records. The local velocity structure (very low velocities in the near-surface) constrains the ray paths (even for local events), to propagate almost vertically near the surface, clearly separating P- and S-waves as vertical and horizontal motions, respectively. The "geothermal event" waveforms (Combs, 1976, p. 34), if body waves, are not consistent with such characteristics.

In addition, the manner in which the S-wave arrival was determined is often ambiguous. Improper determination of S-P times, more than any other error, would heavily influence the determination of local seismicity.

The observation that local seismicity remained unaffected by withdrawal and injection of geothermal fluids could be interpreted as evidence that the observed seismicity is not of local origin. A reexamination of the data appears necessary to resolve these questions. Of specific interest should be the determination of the apparent velocities associated with the "geothermal events" and the temporal correlation of these events with the records of the USGS-CIT Imperial Valley array.

The small events detected at East Mesa during the present study may be analogous to the nano-earthquakes discussed by Combs (1974 and 1976), since they were observed at roughly the same location (MBR STA 9 lies close to Combs' MGA #3). However, S-P time of 1.5 sec assigned to these events by Combs, indicating a local origin, has not been observed (no S-wave arrivals are, in fact, seen for these events). Because these events appear only on MBR 9Z and 3Z, we may speculate that they are actually the smallest members of the Brawley earthquake swarms with origins northwest of East Mesa. One would also expect the signals to show up on STA 7Z; however, a map of seismic ground noise (Combs 1974, p. 33) shows high noise levels in the vicinity of MBR 7 relative to levels around MBR 3 and 9.

Regional seismic activity (events of magnitude 2.0 or less) is well-recorded at East Mesa and is distinguishable from local events by S-P time constraints (and less accurately by relative P-wave arrival times). The quality of the data makes it possible to locate accurately larger events, although local station corrections are not determined. Several seismic wave phases, in addition to the initial P- and S-wave arrivals, are commonly observed. Most of the regional events are associated with the active Brawley-

Imperial fault system to the west and southwest. The closest events recorded, based on S-P time intervals, are located along this fault zone about 20 km west of the array. Other events recorded originated at distances exceeding 500 km and at all azimuths to the west. (A set of sources at varying distances and azimuths may provide data on subsurface properties within the network in the absence of local events.)

Other local disturbances not of earthquake origin have been detected and identified. These include meteorological disturbances, drilling activity, and weapons testing. Although these may have been previously identified as local seismic events, they are distinguishable because of waveforms, frequency content, and most important, by their slow (subseismic) apparent velocities.

The major Brawley swarms of October-November 1977 were well recorded at East Mesa. Again, no local microearthquakes were recorded during this period--a result contrary to the findings of a previous investigation in which local seismicity was reported to have peaked during swarm activity. Our results lead to the conclusion that there is no evidence that the local stress regime is associated with the Brawley-Imperial fault system.

The origin of small events remains unresolved, but they are believed to represent the smallest members of the Brawley fault swarm. Installation of a 4.5-Hz three-component geophone into a 300-ft borehole may aid in determining the source of these events and in detecting small "geothermal events" not presently observable because of the high level of surface noise in the seismic bandwidth. It is estimated that improvement of at least one order of magnitude in the threshold of detectability is required for serious study of the field seismicity.

Results of the study suggest that a reasonable and economical modification to the USBR network would provide a minimal capability of monitoring, at an improved threshold of detection, so that any seismicity within the field would be at a threshold magnitude of zero. To this end, a single three-component station should be considered with either downhole (100 m) geophones or a surface array of at least 36 geophones per channel, with a triggered digital event recorder. The downhole installation would be superior in sensitivity and probably less expensive. Such a station would provide long-term field seismicity data for induced seismicity studies.

PLANNED ACTIVITIES IN 1979

No seismic studies are planned at East Mesa. The USBR has relinquished control of the land to the Bureau of Land Management who will open the lease units to competitive geothermal leasing in 1979.

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Geophysical Studies at Cerro Prieto

The LBL geophysical field program was formulated to assist reservoir delineation and to provide baseline data against which future studies could be compared. Although standard geophysical exploration techniques were used, exploration was not the primary objective of the work. The techniques chosen were those used by LBL and U.C. Berkeley in earlier geothermal investigations, and which we had some reason to believe would provide useful information on the Cerro Prieto reservoir.

Dipole-dipole dc resistivity surveys, supplemented by magnetotellurics, were used to determine how well the reservoir or controlling geologic structures can be defined on the basis of resistivity. Permanent electrodes were emplaced so that future measurements can be made to detect resistivity changes that may occur due to changing reservoir conditions. Computer modeling was done to interpret observed data and to determine the accuracy needed to resolve changes in reservoir bulk resistivity. Because the electrical basement is difficult to resolve beneath the thick section of conductive sediments, magnetotelluric studies were conducted to provide supplementary deep structural information.

Self potential (SP), another type of electrical technique, was performed because it is relatively easy to conduct and has provided provocative and valuable results for other geothermal areas. The SP anomaly found at Cerro Prieto is a "text-book" example. Encouraged by the preliminary results, CFE has begun using self

potential for geothermal exploration outside the known field.

Passive seismic surveys were conducted in two phases. First, a microearthquake study was conducted to establish the level and nature of seismicity within the area of the production zone and for comparison to regional values. Sometimes used in the exploration phase to map faults, microearthquake survey results may be applicable to detecting changes in reservoir conditions, for example, changes in temperature, pressure, fluid saturation and, most important, subsidence.

In the second phase of work, wave propagation studies were made. Propagation characteristics, such as V_p , V_s , Poisson's Ratio and amplitude attenuations, were determined within the reservoir area and compared to values outside the reservoir boundaries. These characteristics, which were derived from recorded regional earthquake signals, help to define the field boundaries.

Precision gravity surveys, performed at regular intervals in conjunction with first-order leveling, have been started at other geothermal reservoirs. Used also with electrical resistivity and seismic propagation surveys, precision gravity could be an exceptionally promising means for monitoring changes in reservoir conditions, for example, (a) densification due to changes in temperature-dependent water-rock chemical reactions, and (b) formation of a steam zone due to lowered pressure, or fluid withdrawal without compensating recharge.

SEISMOLOGICAL STUDIES AT CERRO PRIETO

E. L. Majer and T. V. McEvilly

ACCOMPLISHMENTS IN 1978

In a preliminary investigation of the seismicity and wave propagation characteristics of the Cerro Prieto geothermal field, 17 stations were occupied with vertical and horizontal seismographs from February 14 to March 11, 1978 (Figure 1). The objectives of this field experiment were to determine the level and nature of microearthquake activity in the field area and the velocity and attenuation of P- and S-waves in the production zone relative to regional values. Similar studies of the type described here were conducted at Leach Hot Springs, Grass Valley, Nevada (Majer, 1978) and at The Geysers geothermal field in northern California (Majer and McEvilly, 1979). The previous studies indicated significant velocity and attenuation characteristics associated with both The Geysers (vapor-dominated) and Leach Hot Springs (liquid-dominated) systems. The Geysers study also revealed high microearth-

quake activity (25-30 events/day) associated with the production regions.

Continued and possibly increased fluid withdrawal and Cerro Prieto may effect changes in pore pressures, temperature gradients, volume, and stress patterns, all of which have been shown to be related to seismicity. Changes in the present level or nature of seismicity may be an indication of expansion or alteration of the production zone properties and boundaries. By using detailed microearthquake locations it may also be possible to map the location and nature of faults striking NE-SW which have been hypothesized to exist in the production zone.

The stations for this study were spaced approximately one kilometer apart across the production region and extending outside the field for adequate reference values. Because the data were collected on a 12-channel system and 17

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Figure 1. Distribution of seismic stations, seismic refraction lines and microearthquake activity during the period February-March 1978, Cerro Prieto area.

stations were occupied, the study was conducted in several different phases at sites shown in Figure 1. Stations 1, 6, 8, 10, 11 and 12 served as reference stations for the entire period with vertical, then horizontal, components. The remaining stations were occupied with vertical, then horizontal, components for periods of 4 to 8 days. At each station the signal from a 4.5-Hz geophone was radio-telemetered to a central point and recorded with time- and tape-speed compensation data at 0.24, then 0.12 ips, on 14-channel FM tape recorders with 0-80 Hz and 0-40 Hz bandwidth, respectively. Additional arrival times and first motions were provided by CICESE who maintained short-period, smoked-paper recorders at stations CPR, TLX, QKP, and SON (see Figure 1).

In addition to the data provided by this study, refraction data obtained for CFE in 1962 by Geologos Consultores Asociados (GEOCA) (line 1 and line 3, Figure 1) were analyzed and interpreted along with the available gravity

data. The combined interpretation was used to determine velocity structure and basement depths.

Microearthquake Data

During the recording period, 74 events with S-P times of less than 5 sec were observed at stations 1, 2 and 6 with a threshold of detection and location near $M_L \geq 1$ within the array area.

Of the events recorded on five or more stations, only six events were located near or within the production zone. These events occurred over a two-day period (February 18-19) and were part of a swarm of 20 events, 14 of which occurred within one hour with nearly identical S-P times. The events were located at depths between 2-3 km near stations 4 and 5 (Figure 1). The model used to locate the events was two layers over a half space with P- and S-wave velocities of (2.0, 0.8), (3.5, 2.0) and (5.5, 3.2) km/sec, and layer thickness of 1 and 2 km, respectively.

The other events detected occurred outside the array, allowing only approximate locations (shaded regions, Figure 1). Forty-eight events occurred southeast of the array in the region of a magnitude 4.7 shock which occurred one day after the completion of this study. Four events occurred just south of Cerro Prieto and 15 events originated to the northwest in the Laguna Salada region. Although accurate focal depths are not easily obtained for events outside of an array, the depths seemed to be in the 5-15 km range as contrasted to 2-3 km range for the events inside the production region.

Adequate azimuthal coverage for first motion studies was obtained for only the events occurring within the array, providing one fault-plane solution, consistent with right-lateral strikeslip motion on a NW-trending fault, similar to San Andreas fault movement. The fault plane could not be resolved unambiguously, however, between this motion or left-lateral slip on NE-trending faulting.

Considering the general tectonic nature of the Salton Trough and the complex faulting and structure within the production zone, it seems unusual that microearthquake activity should be so low. McEvilly and Schechter (1978) also report low microearthquake activity at the East Mesa geothermal field. Although low instrument magnifications were used on both surveys, it is doubtful that this can account for the apparent lack of activity. Using similar sensitivities at The Geysers field, we detected 25-30 events/day.

P- to S-wave velocity ratios (V_p/V_s) were estimated by using the Wadati diagram, where S-P time is plotted versus the P-wave arrival time at different stations for a single event, assuming the same V_p/V_s along all propagation paths. The slope of the line through the points is $k-1$, ($k = V_p/V_s$). Poisson's Ratio, σ , may then be calculated by the expression

$$\sigma = (k^2 - 2)/2(k^2 - 1)$$

The swarm of events near stations 4 and 5 provided adequate data to indicate a Poisson's Ratio of 0.4. This is an average Poisson's Ratio throughout the production zone for the upper 2-3 km. Studies at The Geysers (Majer and McEvilly, 1979) and Coso Hot Springs (Combs and Rotstein, 1975) indicated Poisson's Ratios between 0.15 and 0.20 for regions within the vapor-dominated zones. McEvilly and Schechter (1978) found Poisson's Ratio of 0.4 for the upper kilometer for sediments at the East Mesa geothermal field.

Because P-wave velocities are affected more by fluid content than S-wave velocity (Toksoz et al., 1976), the high Poisson's Ratio indicates fluid domination in the upper 2-3 km of the production zone. Somerton (1978) in a laboratory investigation of Cerro Prieto cores measured Poisson's Ratio near 0.1 for dry sandstone and 0.3 for water-saturated sandstone. These and similar laboratory studies indicate that if the reservoir changes from water domination to a two-phase fluid system, we can expect to observe a decrease in the Poisson's Ratio.

P-Wave Velocity and Attenuation Data

Because of noisy data, it was difficult to determine whether a velocity anomaly is associated with the production zone. The advances and delays in P- and S-wave arrivals may be due to faulting, varying sediment composition and thickness, or hydrothermal alteration.

To determine whether the production zone affects the amplitude and frequency of seismic waves, spectral studies were carried out using the P-wave of a regional event. The approach taken, as used at The Geysers, California (Majer and McEvilly, 1979) and Leach Hot Springs, Nevada (Majer, 1978). The method is an adaptation of a technique developed by Teng (1968) for analysis of teleseismic data. The ratio of the spectrum of the P-wave at each station to an arbitrary reference station is used to obtain the differential attenuation. It is assumed that Q is frequency-independent over the band used and that the paths to both stations are common except for the last fraction of the total path.

Stations 1, 2, 3 and 4, which are west of the production zone, exhibit slightly lower or near-normal attenuations over the 5-20 Hz range. Stations 10 and 11, which are east of the production zone, show slightly higher attenuation. Stations in the production zone exhibit significant attenuation (stations 5 and 8); however, station 6, which is the closest to the center of production, shows attenuation lower than normal. Due to high noise levels there is insufficient data at station 9 and only marginal data at station 11.

The higher attenuation noted at stations 5 and 8 may be associated with faulting, evidenced by the microearthquake activity in this region. The lack of attenuation closer to the center of the production zone (station 6) may be due to alteration by precipitation in this region. The increasing attenuation east of the field (stations 10 and 11) may only be due to increasing sediment thickness.

Refraction Data

In addition to the data obtained from the February 14-March 11 study, two lines of refraction data were analyzed to determine velocity structure and depth to basement. Unfortunately, none of the refraction lines crossed the present production zone; however, lines 1 and 3 (Figure 1) crossed near the edges of the field as presently defined by drilling.

The difference between the lines are the complexity of the travel time data and the high basement velocities (6 km/sec) along line 1 compared to those for line 3. The velocities of the layers obtained compare well with velocities for sediments in other parts of the Colorado Delta region (Kovach et al., 1962). The structure along the southern line appears much simpler than that of the northern line. There is evidence for a horst and graben structure along line 1. On line 3 there seems only to be down-faulting to the east. Although the two refraction lines

Figure 5. Two-dimensional gravity model along line 3. Values are density contrasts. Compare to Figure 3.

On the eastern edge of line 1, the northern line, the depth to basement was fixed at 2.5 km on the basis of drill cores obtained in Well #96, 1 km north of the line.

CONCLUSIONS AND WORK PLANNED FOR 1979

Seismological data indicate structural complexity associated with the Cerro Prieto geothermal field. Refraction data and microearthquakes suggest several major NW striking faults, possibly controlling the field boundaries. The refraction data also suggest high velocity sediments and shallow basement in the production zones. Compared to seismicity in surrounding areas, microearthquake activity is low in the production area. First motion studies indicate right-lateral strike-slip motion for events within the field on NW-SE trending faults. Further microearthquake data are necessary to reduce the ambiguity in earthquake source and occurrence properties for comparison to regional values.

High levels of man-made ground noise in the center of the production zone limit the amount of velocity and attenuation data that can be collected. The few data collected suggest P-wave velocity and attenuation anomalies associated with the production region. V_p/V_s values are high; further monitoring may reveal an eventual decrease, indicating a change to a two-phase reservoir. Microearthquake data may indicate steam zones if the events are related to large pressure or temperature gradients or to volume changes associated with fluid removal. There is sufficient seismic activity in the production zone to warrant continuous monitoring.

In future work, we will continuously monitor seismicity and V_p/V_s values within the produc-

tion zone by means of a five-station array centered near station 5 (Figure 1). Each station will digitally record 12-bit data on a triggered cassette system from a three-component, 4.5-Hz geophone in a 100-m-deep well. The stations will be connected to a central station that will provide a periodic time fiducial and visual monitor of the activity. These data will provide the necessary information to infer the relation between fluid withdrawal/reinjection and future seismicity.

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SELF-POTENTIAL STUDIES AT CERRO PRIETO

R. F. Corwin and H. F. Morrison

ACTIVITIES IN 1978

Self-potential anomalies are correlated with geothermal activity in many areas around the world (Corwin and Hoover, 1978). These anomalies often are associated with faults that are thought to serve as conduits for thermal fluids, or with areas of high heat flow or vigorous subsurface fluid flow. To see whether the self-potential method could prove helpful in delineating the zone of geothermal activity at the Cerro

Prieto field, about 100 line km of data were taken in the vicinity of the presently operating geothermal wells and 75-MW power plant.

Field Procedure

The self-potential survey lines in the central part of the survey area and their relation to the present geothermal production zone are shown in Figure 1. Field procedures have been described

XBL 776-1179A

Figure 1. Self-potential survey lines in the central part of the Cerro Prieto geothermal field. A complete map of the survey lines is shown in Figure 3. The dashed lines indicate the location of the charged planes used to explain the source of the anomaly. Numbered circles are geothermal production wells.

XBL 787-5327

Figure 2. Self-potential data profiles in the area of the Cerro Prieto geothermal field.

by Corwin et al. (1978). The survey data are shown in profile form in Figure 2 and contoured in Figure 3. The complete survey lines also are shown in Figure 3. Short-wavelength variations, less than 500-m wide, were ignored in the contouring.

A well-developed dipolar anomaly of about 160 mV peak-to-peak amplitude, positive to the west and negative to the east, is seen on line B-B' (Figure 2), which runs through the center of the present geothermal production zone.

XBL787-5328 A

Figure 3. Self-potential contours, Cerro Prieto area. Contour interval is 20 mV.

The contours of Figure 3 shows that the axis of the dipolar anomaly (the 0-mV contour) passes through the approximate center of the present production zone, if the zone is considered to include the presently producing well M-53 (Figure 1). The axis of the anomaly trends about N.50° W., and to the east of the axis the anomaly appears to be somewhat elongated in both the north-south and east-west directions. The contours appear to be affected by a trend running parallel to and just south of the eastern portion of line D-D'. Aside from the major dipole anomaly, strong self-potential activity is seen on line D-D' just south of the Cerro Prieto volcano. Geophysical data indicate a major fault running through this area, but the self-potential coverage is insufficient to allow any conclusions to be drawn about the relation between this fault and the self-potential activity. Smaller but still significant anomalies are seen just to the northwest of Nuevo Leon and just to the southeast of Michoacan de Ocampo.

Analysis

As the self-potential anomaly shows no correlation either to surficial or cultural features or the thermally inactive Imperial fault zone, its cause is likely to be a deep-seated mechanism related to geothermal activity. The theory of how geothermal sources may generate surface self-potential anomalies was described by Nourbehecht (1963) and developed by Fitterman (1978) of the U. S. Geological Survey.

Briefly, the mechanism is based on the fact, established by laboratory measurements, that the flow of either fluid or heat through a rock creates a measurable voltage. The ratio of the developed voltage to the pressure gradient causing fluid flow is called the electrokinetic or streaming-potential coupling coefficient (CE). The ratio of the developed voltage to an applied temperature gradient causing heat flow is called the thermoelectric coupling coefficient (C_T).

The form of the Cerro Prieto anomaly resembles a potential distribution generated by flow parallel to a vertical plane separating regions of different coupling coefficients; the potential distribution developed over only a limited length and depth extent of the plane (Fitterman, 1978). Geologically, this would correspond to a vertical flow of fluid, heat, or both, in the vicinity of a vertical fault, fracture zone, or contact separating regions of differing coupling coefficients.

There appears to be geological and geophysical evidence for the existence of a major fault or fracture zone with location, strike direction, and depth extent similar to those postulated to explain the SP anomaly. This fault is thought to provide an important conduit for thermal fluids to ascend from the basement rocks into the reservoir formation (Noble et al., 1977). The sandstone and shale reservoir formation is capped by relatively impermeable clay to a depth of about 1 km and is underlain by a granite basement at a depth of about 3 to 4 km. This region, from

about 1 to 4 km depth, corresponds well to the source region postulated for the self-potential anomaly.

Our tentative conclusion, then, is that a difference in thermoelectric or electrokinetic coupling coefficients (or both) exists across the Patzcuaro fault or fracture zone, and that a vertical component of heat or fluid flow (or both) in the presence of this vertical discontinuity generates the observed self-potential anomaly. Thus, the Patzcuaro fault or fracture zone may represent an important structural feature of the Cerro Prieto geothermal field.

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RESISTIVITY STUDIES AT CERRO PRIETO

M. Wilt and N. E. Goldstein

ACTIVITIES IN 1978

As part of a cooperative research venture between the U. S. Department of Energy (DOE) and the Comisión Federal de Electricidad de México (CFE), a series of dc resistivity investigations were performed at the Cerro Prieto geothermal field. The work has included Schlumberger and dipole-dipole field surveys, geoelectric computer modeling, and geological interpretation of results. The goals of this work are: (a) to determine whether the reservoir region or controlling geologic structure can be delineated by careful dc resistivity measurements; (b) to explore, with geoelectric modeling, the possibility of using the dipole-dipole resistivity technique for monitoring temporal changes in resistivity caused by changes in the reservoir; and (c) to perform the first set of replicate dipole-dipole measurements.

The dipole-dipole technique was chosen for temporal monitoring because it is inherently sensitive to lateral resistivity variations and because the relatively few permanent stations needed for remeasurement can be easily established.

Resistivity Surveys

During 1978, two long dipole-dipole lines were surveyed at Cerro Prieto by LBL resistivity crews each with dipole separations of 1 and 2 km (Figure 1). The northern line, D-D', is 18 km long and extends northeastward from the Cucapa Mountains into the Mexicali Valley, passing just south of the Cerro Prieto volcano. This line is roughly coincident with CFE Schlumberger-resistivity line 1, which was also used in interpretation; the two lines overlap for approximately 10 km. Line E-E' also begins at the Cucapa Mountains, is approximately 16

XBL 788-1632

Figure 1. Project location map, LBL resistivity project.

km long and trends subparallel with D-D'. It extends eastward into the Mexicali Valley passing through the central portion of the production zone.

A finite-difference computer program capable of modeling resistivity variations in two dimensions was used for interpretation of Schlumberger and dipole-dipole data (Dey, 1976; Dey and Morrison, 1976). The program computes potentials at all nodal points within and on the surface of an arbitrary two-dimensional earth model. It is capable of calculating apparent resistivity for the Schlumberger, dipole-dipole, roving dipole, and several downhole arrays. A limitation to this program is the size of the network mesh. The 58 x 16 nodal size does not allow for detailed models of both shallow and deep structures simultaneously. For example, if a model of earth structure to 3 km is desired, then the minimum size body that may be incorporated with one mesh is 100 x 100 meters. This could exclude small surficial inhomogeneities in models, even though such features are evident in data sets.

Results

Some examples of field results of Schlumberger and dipole-dipole surveys are shown in the top portions of Figures 2, 3 and 4. All plotted values are apparent resistivity in ohm-m. Schlumberger apparent resistivities (Figure 3) are plotted, by convention, below the center of the array as a function of $AB/2$ (half the separation between the current electrodes). Dipole-dipole apparent resistivities (Figures 2 and 4) are plotted as a function of distance between current and potential electrode pairs in integer multiples of the electrode separation, or N spacing. The values appear at the intersection of 45° lines drawn from the centers of the current and potential dipoles.

XBL 788-1639

Figure 2. Dipole-dipole apparent resistivity pseudosection for 1-km dipoles along line D-D'; field data, model generated data, and two-dimensional resistivity model.

OBSERVED RESISTIVITIES - SCHLUMBERGER EXPANDERS

CALCULATED RESISTIVITIES

2 D RESISTIVITY MODEL

0 1 2 3
KILOMETERS
Scale: Horizontal

XBL 788-1633

Figure 3. Schlumberger apparent resistivity pseudosection for data taken along line D-D'; field data, model generated data, and two-dimensional resistivity model.

Dipole-dipole models were first calculated for data taken with 1-km dipole lengths to obtain the near-surface features, then for 2-km dipoles to help resolve deeper boundaries. The model was considered acceptable only if it satisfied both data sets. Schlumberger data were modeled only for AB/2 spacing greater than 100 m because the grid network is insufficiently large to account for near-surface detail.

Discussion

Resistivity measurements show that the Cerro Prieto region has very low resistivities down to great depths. Surficial resistivities can be

below 0.5 ohm-m, and the majority of electrical units defined have resistivities below 2 ohm-m. In such an environment, a low resistivity region expected for a geothermal reservoir in sediments may be totally masked. However, at Cerro Prieto we find evidence that the reservoir region may actually possess a higher resistivity than the overlying rocks. Resistivity measurements were found to be very useful in delineating structure and for deducing characteristics of rocks within and outside of the geothermal field. The following is a brief summary of the principal findings:

1. The production region at Cerro Prieto is characterized by high resistivity (4 ohm-m) relative to the surrounding region (1 to 2 ohm-m).

E-E' 1km OBSERVED DATA

E-E' 1km CALCULATED DATA

2 D RESISTIVITY MODEL

XBL 788-1635

Figure 4. Dipole-dipole apparent resistivity pseudosection for 1-km dipoles along line E-E'; field data, model-generated data, and two-dimensional resistivity model.

2. Modeling studies suggest that significant increases with time of resistivity or thickness of the production zone are detectable with precision dipole-dipole measurements.
3. Several faults that may act as hydrologic barriers and are related to the production zone have been detected with dipole-dipole measurements.
4. Perturbation studies in the resistivity model over the reservoir region suggest that the northern nonproductive region adjacent to the volcano was more permeable in the past and resembled more closely the present production zone.

5. Modeling studies have shown that at Cerro Prieto, dc resistivity is not an effective tool for mapping basement surface or other deeply buried resistive horizons.

PLANS FOR FISCAL YEAR 1979

In 1979 we plan to remeasure the two established dipole-dipole lines, D-D' and E-E', extending each other 4 km to the east. In addition, a new dipole-dipole line trending northwest and connecting D-D' and E-E' will be established. The purpose of this new line is for detecting any anisotropy in the rocks and for tying the other dipole-dipole lines together for purposes

of three-dimensional resistivity modeling. All resistivity measurements will be performed with new receiver instrumentation. This new device will allow for resistivity measurements to be reproducible to within 2%, which is within the accuracy required for dipole-dipole monitoring of temporal resistivity changes.

E-field ratio telluric studies will be performed at the same time, using the same electrodes and receiver instrumentation as in the dipole-dipole study. The purpose of this study is to determine the effectiveness of the method in a highly conductive environment. This study will also test the applicability of telluric phase measurements, which are now done easily with the new general-purpose microprocessor-controlled receiver.

We will exercise for the first time a three-dimensional resistivity modeling program to investigate the effects of three-dimensional structure on results derived from two-dimensional models. The study will be based on new and existing dipole-dipole data obtained by LBL and on Schlumberger resistivity data obtained by CFE. In addition, a more complete analysis of possible resistivity perturbations caused by reservoir depletion and mineral deposition will be performed, and resistivity well logs will be correlated to surface resistivity measurements.

MAGNETOTELLURIC SURVEY AT CERRO PRIETO

T. D. Gamble, W. M. Goubau, N. E. Goldstein, and J. Clarke

ACCOMPLISHMENTS IN 1978

Magnetotelluric (MT) soundings were made at seven sites at Cerro Prieto to obtain deep resistivity information. Surveys were performed using the remote magnetic reference technique, which has recently been shown to provide more accurate results than conventional MT in areas of interfering noise or when the natural field strength is weak (Gamble 1978; Gamble et al., 1978).

The seven sites and the reference site, R, on the edge of the Cucapa mountains are shown in Figure 1. Two components of the reference signals were telemetered to the local sounding site by FM radio. Three-component SQUID magnetometers were used for magnetic field measurements, and 200-m dipoles in an L-shaped array were used for the electric field measurements. The dc telluric voltages were filtered out by means of a 0.001-Hz high-pass single-pole filter. All channels had matching 60-Hz notch filters. The seven channels of data were recorded on a Honeywell FM analog tape recorded at 1-7/8 ips. Up to 6 hours of data were recorded at each station. Data were later digitized in Berkeley using a Varian minicomputer. Subsequent digital processing and analysis were conducted on the LBL CDC-7600 system.

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The data were collected in two overlapping frequency bands but only the lower band, covering the period range of 8 to 840 sec, has been completely analyzed. Analysis and interpretation have been directed mainly to the results from stations 1 through 6, which lie along a line nearly coincident with a line of dipole-dipole and Schlumberger dc resistivity measurements. The MT results have been interpreted for comparison with the dc resistivity interpretations (Razo et al., 1978; Wilt et al., 1978).

In contrast to dc resistivity measurements, MT provides a great volume of information at each station, only a fraction of which we are capable of using in quantitative interpretations. For each frequency component, MT yields at least six independent complex values but only the real parts of two are commonly used in interpretations. These are the apparent resistivities rotated into the two principal horizontal resistivity directions (Figure 2). In this case, ρ_{xy} is the apparent resistivity in the parallel-to-strike direction (associated with the TE mode of propagation) and ρ_{yx} is in the perpendicular-to-strike direction (associated with the TM mode of propagation). The vertical bars indicate the probable errors (50% confidence limits), which become very small for the good data obtained. Unless the surface layer is anisotropic, ρ_{yx}

XBL 789-11107

Figure 1. Magnetotelluric sounding sites 1 through 7 and remote magnetometer site R at Cerro Prieto.

XBL 7810-5963

Figure 2. Rotated apparent resistivities and their probable errors for Station 6.

XBL 7810-5964A

Figure 3. Two-dimensional model of the resistivity structure derived from the ρ_{xy} data set.

and ρ_{xy} coverage at shorter period (higher frequency), but their divergence at longer periods is evidence of a multidimensional geological structure.

Because the general results indicate a two-dimensional geometry, the MT results were fitted to two-dimensional models. Starting with the model derived from dc resistivity, we obtained the model (Figure 3) that fits the ρ_{xy} (TE mode) data. The calculated resistivities for this model (Figure 4) may be compared with the observed resistivities (Figure 5).

A preliminary interpretation of the MT results shows that the model derived from the ρ_{xy} data set agrees, in general, with the model fitting the dc resistivity data along the same line. Clear evidence for both the Cucapa fault, west of Station 1, and the Cerro Prieto fault, between Stations 3 and 5, can be seen in the MT interpretation. However, the structure near Stations 2 and 3 appears to be too complex to model accurately. However, we now realize that these data are biased by noise from a CFE dc resistivity

XBL 7810-5967A

Figure 4. Pseudosection of apparent resistivities ρ_{YX} calculated from the model in Figure 3.

Figure 5. Pseudosection of observed apparent resistivities ρ_{xy} at Stations 1, 2, 3, 5, and 6.

survey. The MT data sets will have to be re-edited to remove those containing the correlated noise source.

PLANS FOR FISCAL YEAR 1979

The higher frequency band of data collected at each of the seven sites will be processed. This will provide complete coverage over the period range of 0.02 to 900 sec.

PRECISION GRAVITY SURVEYS

R. B. Grannell,* D. W. Tarman† and N. E. Goldstein

ACTIVITIES IN 1978

Fifty-five gravity stations and five alternates were established over an area of approximately 500 km² centered on the geothermal field (Figure 1). All stations are permanent, and consist of concrete pads large enough to accommodate a gravity-meter base plate. Two control stations were placed on bedrock which might be presumed stable with respect to production-induced subsidences: (a) Station 1 on granite in the Cucapa Mountains some 15 km west of the geothermal plant, and (b) Station 20 on the flanks of the Cerro Prieto volcano less than 10 km northwest of the power plant. The five alternate stations were established either to augment the bedrock stations or to provide more reliable valley stations where the initial stations were found to be unsuitable for long-term studies.

An additional group of MT stations will be occupied at selected sites over the producing part of the reservoir, including the Patzcuaro fault zone, locus of the large SP anomaly (Corwin et al., 1978). Further modeling of the results will be done and correlated with other subsurface resistivity data.

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During the year all of the stations were leveled by CFE engineers and tied into the established first- and second-order level networks around the field. Complete sets of gravity readings were made at most of the stations (Grannell et al., 1978). To maximize the precision of the gravity values, each station was occupied an average of four times, twice each with two La Coste and Romberg G model gravity meters. During each occupation, a minimum of four readings was obtained, two each by two operators. The meter set-ups were changed before each reading so that errors due to leveling or orientation could be distributed. The occupations were also distributed in time to minimize the differential effects of tides caused by vehicular noise on roads, instrument drift, external temperature changes and wind conditions.

Three base stations were used for reference, and 24 hours of tidal monitoring were conducted to obtain more accurate constants for the tidal-drift correction.

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XBL 791-162

Figure 1. Bouguer anomaly map of the Cerro Prieto geothermal field and its vicinity. Contour interval is 1 mgal, permanent gravity stations are shown as the numbered dots.

Because some sizeable long-term drift in the meters occurred, it was not possible to compare directly the gravity values for a particular station at greatly differing times. However, station data obtained in each gravity loop were subtracted from the value of internal base station 23, yielding an array of gravity differences which were comparable. The desired final observed gravity value for each station was obtained (a) by calculating the mean gravity difference with respect to station 23 and (b) by converting those values to a mean gravity difference with respect to station 1 in the Cucapa Mountains. Figure 1 shows the contoured gravity data reduced to the Bouguer anomaly. Station 1 was arbitrarily assigned a gravity of 0.0 mgal and all Bouguer values are referred to this zero base.

The standard deviations of half the station values were smaller than 15 μ gal, indicating that a fairly high degree of replicability was achieved in spite of some unfavorable field conditions such as high ambient temperatures ($>105^{\circ}\text{F}$) during part of the survey period, rough roads, and wind.

Although the gravity data obtained can be interpreted in terms of a subsurface mass distribution, it is too soon to know whether changes in gravity are occurring within the region of the geothermal field. Data collection of the first data set spanned a time of four months, during which most stations showed either little variation or gave reasonably small erratic changes. At a few stations we noticed a different and more interesting behavior. Station 33 exhibited

a large step function increase of 0.28 mgal during four occupations spanning one month. Also, the gravity between base station 23 on the valley floor and base station 1 in the Cucapa Mountains increased 0.05 mgal in four months, suggesting relative subsidence of 25 cm in the valley relative to the Sierra Cucapa.

PLANNED ACTIVITIES FOR 1979

The second cycle of complete gravity measurements and leveling will be made during the winter of 1978-1979 so that initial estimates of short-term temporal changes can be made. We hope to collect data over a shorter interval of time and during a cooler time of the year to avoid the instrumental problems encountered when ambient temperature exceeds 105°F .

Additional tidal monitoring will be done for greater accuracy of the correction factors.

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