ELECTROMAGNETIC STUDIES IN GEOTHERMAL REGIONS

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Abstract. During the past 25 yr, nearly all available electromagnetic and geoelectric techniques have been tested for their usefulness in geothermal exploration and exploitation. Dipole-dipole profiling, audiomagnetotellurics and controlled source electromagnetic methods are examples of those which have proven to be rather efficient for geothermal exploration. From the hundreds of field surveys which have been performed in many geothermal regions of the world, a large variety of geothermal regions and local geothermal systems, with different geological, hydrological and heat transfer characteristics, has been found to exist. Depending on the combination of these different characteristics each geothermal region or system presents a new problem which may need a different field technique or group of field techniques for optimal exploration. Despite these problems, new geothermal regions have been detected and structures and processes in geothermal systems are now much better understood. For example, advances have been made in the study of (a) the characteristics of porous/permeable hot water/vapor reservoirs and of fractioned zones for hot water/vapor circulation and production (b) the distribution and movement of cold meteoric and of hot water (c) the thermal insulation of reservoirs by cap-rocks (d) convective and/or conductive heat transfer and (e) the thermal influence of magma intrusions to high crustal levels.

New exploration techniques, data analysis procedures and model calculations have been developed in the course of research in geothermal areas. They include the controlled source electromagnetic methods, the remote reference field technique and the development of better and faster algorithms for direct and inverse model calculations. Problems for the future are (a) the development and improvement of equipment and field techniques for more precise delineation and resolution of the conductivity distribution in geothermal areas especially those with productive zones of high porosity/permeability and fracturing, (b) the improvement of computerised data analysis in the field to optimise progress during the field measurements and (c) the development of more efficient interpretation procedures for the rather inhomogeneous conductivity distribution which exists in most geothermal areas.

1. Introduction

The growing interest in geothermal energy as one of the 'alternative' energies has stimulated geoscientists to test methods and to develop them for its exploration and exploitation. Hundreds of geoscientific field and laboratory measurements have been carried out in the last 25 yr. Because the transfer of the geothermal energy from the Earth's interior to the surface by hot water and/or vapor must be cost effective mainly the uppermost km of the Earth (< 5–10 km) have been investigated. However, to understand geothermal systems better their roots also (e.g. magma intrusions into the crust, etc.,) have had to be examined.

Many different geophysical methods have been used and are still used for the study of geothermal regions, e.g. heat flow measurements, active and passive seismic observations (microseismicity, teleseismic P-wave delays, etc.,) airborne infrared surveys, electric and electromagnetic measurements including self potential and induced polarization, gravity and magnetic methods (depth of the Curie isotherm) and geophysical borehole measurements. In this paper only investigations with electromagnetic techniques and to some extent electric methods will be discussed. These techniques have proven to be among the most useful geophysical methods for the

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investigation of geothermal regions. This is due to the fact, that the spatial distribution of conductivity in a geothermal area is not only determined by the host rock distribution, but is also directly related to the distribution of the actual exploration target – hot water.

As will be discussed in Section 4, different electromagnetic and geoelectric techniques have been tested in geothermal areas of different geologic-tectonic environment and of different dimensions. Much has been learned about the usefulness of the different field techniques in the different geothermal systems (e.g. reconnaissance, semi-detailing, detailing techniques, resolution of small geothermal anomalies, influence of a well conducting sedimentary cover or of lateral variations of conductivity on natural and artificial current systems, distortion effect of topography on current systems, etc.). Some of the new techniques such as audio-magnetotellurics and controlled source electromagnetic methods seem to be promising.

Advances in field technique and data analysis (e.g. highly sensitive induction coil magnetometers and SQUID-magnetometers, suppression of artificial electromagnetic noise, remote reference and multivariate data analysis) have improved the quality of the analysed data to a point where random errors in the data may be smaller than the uncertainty in their interpretation. Therefore the main problem in the future will be to find better and faster algorithms to calculate the response of more complicated two- and three-dimensional models to homogeneous and inhomogeneous source fields.

The conductivity distribution within geothermal areas has often been found to be rather inhomogeneous and hard to map in detail. Despite all these difficulties new geothermal regions and systems have been found and known geothermal systems are now better understood.

2. The Electrical Conductivity of Fluids and Rocks in Geothermal Regions

The conductivity of fluids in the subsurface depends on several physical parameters such as

- the concentration of ions (amount of salt/unit solvent, degree of dissociation);
- the interaction between ions in the solution;
- the charge number of the ions (kind of substance);
- the mobility of the ions in the solution;
- as well as on the density, viscosity, dielectric constant, pressure, and temperature of the fluid.

The viscosity, density and dielectric constant of water (as the solvent for salt) decrease with increasing temperature. Viscosity decreases exponentially with increasing temperature, resulting in an increase in the mobility of the ions. At low temperatures this causes a strong increase in conductivity but at higher temperatures ($T > 200\,°C$) this increase diminishes sharply. The density of water and thus the number of ions per litre solution decreases continuously with increasing temperature (at constant pressure), causing a reduction in conductivity. The decrease in dielectric constant with increasing temperature causes an increasing apposition of positive and negative ions and hence a decrease in conductivity. For temperatures below 300–400 °C, the temperature dependence of conductivity is mainly determined by the viscosity and thus the conductivity increases considerably with increasing temperature – Figure 1. For temperatures above 300–400 °C, the effects of decreasing density and dielectric constant predominates resulting in a decrease in conductivity with increase in temperature.

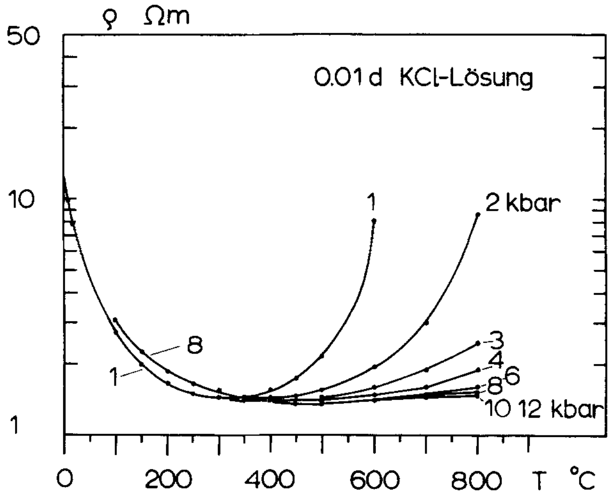

Fig. 1. Variation of the resistivity ϱ of a 0.01 demal KCl solution (0.745 g KCl/1000 g solution) as a function of temperature and pressure (redrawn after Quist *et al.*, 1970).

The dependence of the conductivity of salt solutions on temperature and pressure is known from laboratory measurements within a wide range of concentrations, up to five molar solutions (Franck, 1956, 1961, 1973; Franck *et al.*, 1962; Hartmann *et al.*, 1969; Holzapfel *et al.*, 1966; Hwang *et al.*, 1970; Klostermeier, 1973; Mangold *et al.*, 1969; Quist *et al.*, 1963, 1966, 1968, 1969, 1970; Renkert *et al.* 1970; Ritzert *et al.* 1968; Berktold, 1982). For example, the resistivity of a one molar NaCl solution (58.44 g NaCl/1 l solution) is 0.11 Ohm m at $T = 26$ °C and 0.025 Ohm m at $T = 300$ °C ($p = 1000$ bar) (Klostermeier, 1973).

The fluid concentrations in geothermal reservoirs vary widely, with reported values of 3.5 molar for the Salton Sea brine (California), 0.35 molar for the East Mesa brine (California), 0.3 molar for the Cerro Prieto brine (Mexico) and 10^{-2} molar for the Broadlands (New Zealand), (Ucok *et al.*, 1980; Moskowitz *et al.*, 1977). Typical resistivities of geothermal reservoir fluids range from 0.01 to 10 Ohm m. For comparison the resistivity of seawater is about 0.2 Ohm m at a concentration of about 0.5 molar.

As the electrical conductivity of vapor decreases strongly with decreasing density of the vapor, vapor dominated fracture systems generally have a very high resistivity. Sometimes the steam zone may be overlain by a layer of condensate which may be detected by electrical or electromagnetic methods.

The conductivity of porous and permeable rocks saturated with fluids can be described by

$$\sigma_r = \frac{1}{F}\sigma_f + \sigma_{is}$$

where σ_r is the rock (bulk) conductivity, σ_f the conductivity of the pore fluid, σ_{is} the conductivity due to the boundary layer at the internal surface of the rock and F the formation factor.

For rocks not completely saturated with fluids the degree of saturation has to be considered.

The formation factor F is closely related to the porosity ϕ through

$$F = c \cdot \phi^{-m}$$

where $c \approx 1$ and $1.3 \leqslant m \leqslant 2.5$.

F is a constant of the rock material describing macroscopically the geometric structure of its pore space (e.g. the ratio of cross section to length of the available ducts).

The electrical conductivity of moisture-containing rocks as a function of pressure and temperature has been investigated by several authors (Brace, 1971; Hermance, 1973; Shankland and Waff, 1974; Drury and Hyndman, 1979; Rai and Manghnani, 1981; Olhoeft, 1981, etc.). As shown in Figure 2 the conductivity of a rock increases by several orders of magnitude as water is added to the rock. The amount and chemistry of free water in a rock also significantly changes the temperature dependence of conductivity. Compared to the many orders of magnitude of conductivity change due to temperature variation the effects of pressure on rock conductivity are negligible.

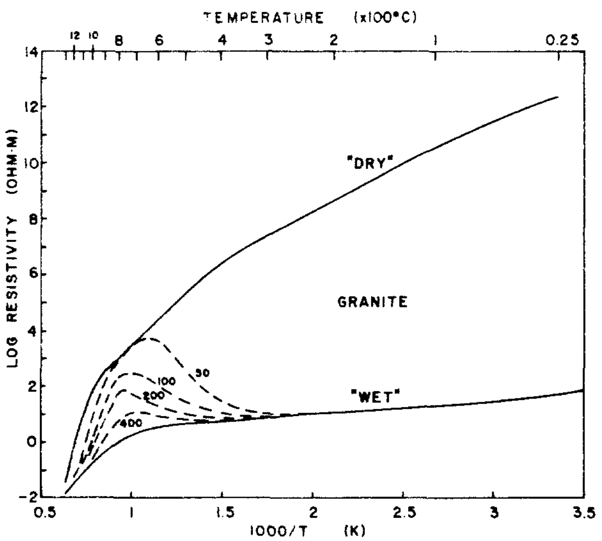

Fig. 2. Summary plot of the best available estimate for electrical resistivity of granites versus temperature, pressure and water content. The dashed lines are for various water pressures in MPa for water salinities less than 0.1 molar NaCl. (Olhoeft, 1981).

The electrical conductivity of a rock depends not only on the matrix permeability due to pore space but also on the permeability due to fractures, fissures and joints. The latter permeability is commercially of more interest as fluid circulation is much higher in fractures etc. than in the porous permeable channels within the rock matrix. The distribution of fractures, fissures and joints within a geothermal system, however, is hard to map in detail by geophysical methods.

The investigations of the last 25 yr have shown that the conductivity distribution within most of the geothermal anomalies is a complex function of porosity, permeability and fracturing of the host rock, fluid composition and distribution of fluids within the rock as well as of the pressure and temperature of the fluids and rocks.

3. The Geothermal Regions and Systems

In terms of plate tectonics, geothermal regions occur mainly at or near divergent and convergent plate boundaries (spreading ridges and subduction zones), at intraplate rifts (continental rifts, and thermal anomalies (hot spots)). Plate movement may be

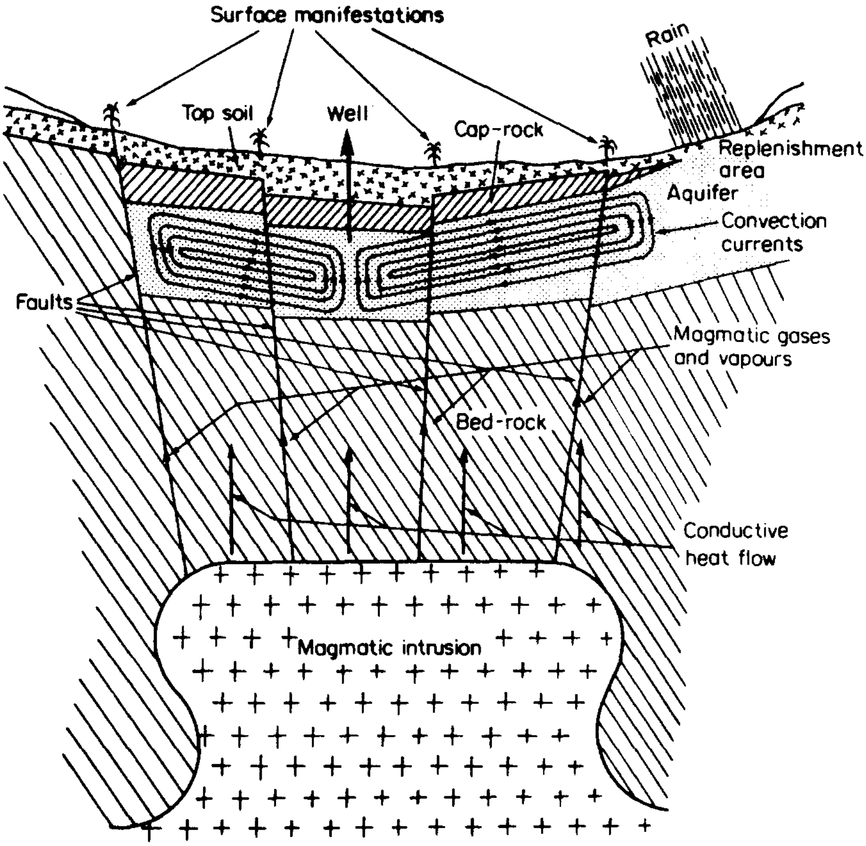

Fig. 3. Idealized model of a hyper-thermal field.

accompanied by magma intrusions into the crust and by volcanism. Tectonic activity may cause deep faults and fractures in the sedimentary cover and in the upper crust. In those areas heat is transferred from the Earth's interior not only by conduction but also by vertical mass transfer, like upwelling of magma or deep water circulation and heat flow is often increased regionally with local extrema. To create a thermal anomaly of recent activity the igneous intrusions should not be older than $10^4 - 10^6$ yr.

The geothermal systems which are most interesting for electric power production, occur where magma intrudes into high crustal levels ($< 10\,\mathrm{km}$) and thermal water convection can take place above the intrusive body as shown in Figure 3. The heat stored in the rocks above the magma intrusion must be collected by water and transferred to a high porosity/permeability reservoir at shallower depth – shallow enough, to be tapped by drill holes. For optimal thermal insulation the reservoir rocks should be covered by an impermeable layer. The water circulating in geothermal systems is normally of meteoric origin. The surface area where cool meteoric water percolates underground generally extends much more than the area where hot water moves upwards.

This type of geothermal system is the most important but not the only one. There exists a great variety of different geothermal systems. Depending on their geological, hydrological and heat transfer characteristics, viz. structure, stratigraphy and type of rocks, presence or absence of permeable reservoirs and insulating cap-rocks, chemistry and distribution of fluids, temperature range, convective and/or conductive heat transfer, etc., these geothermal systems can be subdivided into the following broad groups (Rybach and Muffler, 1981):

(1) Geothermal systems with predominantly convective heat transfer (in natural fluids).

These consist of:

(a) hydrothermal systems in environments of high porosity and permeability, which are related to high heat flow from shallow, young intrusions as illustrated in Figure 3. This group includes practically all geothermal systems which have actually been developed for commercial electric power production.

Sometimes the group is further subdivided into:

(α) vapor dominated systems (e.g. The Geysers, California; Larderello, Tuscany/Italy);

(β) liquid dominated systems (most other hot water systems, e.g. Imperial Valley, California; Broadlands, Wairakei, New Zealand); and

(b) hydrothermal circulation systems in environments of low porosity but sufficient fracture permeability for deep circulation of meteoric waters in fault and fracture zones. These circulation systems may occur also in areas of normal to slightly increased regional conductive heat flow.

(2) Geothermal systems in the thermal regime of conductive heat transfer. These can be classified as:

(a) low temperature aquifers ($T < 150\,^{\circ}\mathrm{C}$) in environments of high porosity and permeability (e.g. aquifers, sometimes of high salinity, in the deeper layers of

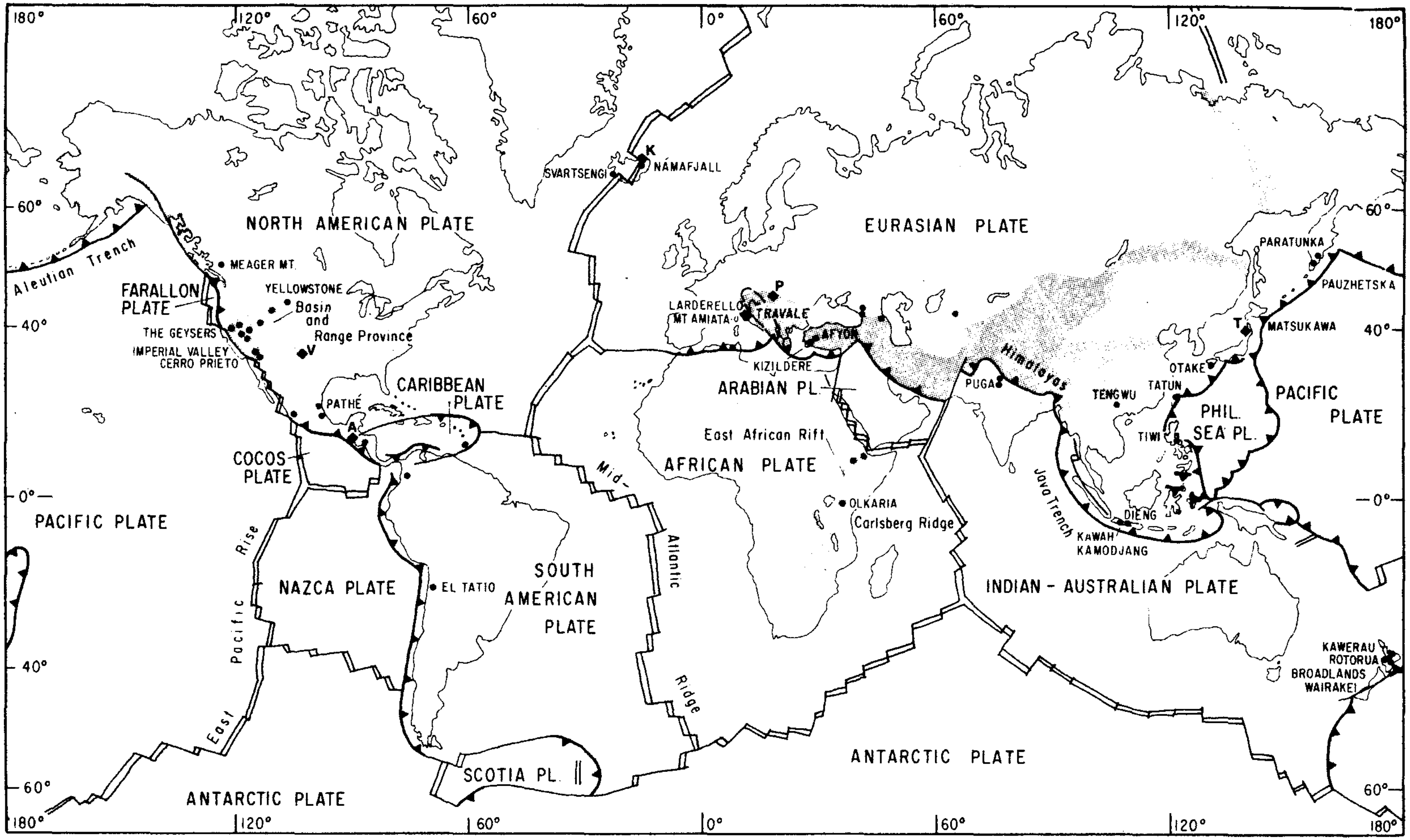

Fig. 4. The global distribution of major geothermal systems in the framework of lithospheric plate boundaries (Rybach and Muffler, 1981).

sedimentary basins, and geopressurised reservoirs, where the pore fluids are under pressure exceeding the hydrostatic pressure); and

(b) 'Hot Dry Rock' systems, in which the rocks near the Earth's surface have low natural porosity and permeability but high temperatures. In these systems cold water is pressed through a borehole into an artificially created small and extended fracture within the rock mass. While moving through the fracture the water is heated up to a temperature at which vapor is produced which can – after being transferred back to the Earth's surface through a second borehole – be used for electric power production.

The major known geothermal regions of the world are shown in Figure 4.

4. The Electrical and Electromagnetic Methods

A whole family of electrical and electromagnetic techniques is available for the investigation of the different geothermal systems. These techniques can be subdivided into groups in different ways depending on:

(a) whether the source field is natural or artificial;

(b) whether the spatial geometry of the source field is homogeneous or inhomogeneous;

(c) the frequency or frequency range of the measurements (direct current (dc) or alternating current (ac) techniques); and

(d) the kind of field survey i.e. reconnaissance, semi-detailed or detailed technique.

In this paper the following subdivisions are used:

(1) techniques with an artificial source field and a dc or quasi-dc galvanic current injection;

(2) borehole electrical and electromagnetic measurements and electromagnetic measurements between boreholes (tomography);

(3) techniques with natural source fields of homogeneous or quasi-homogeneous spatial distribution;

(4) techniques with artificial source fields and an alternating current flow i.e. controlled source electromagnetic methods.

These are discussed in Sections 4.1 – 4.4 respectively. Which of these many techniques should be used for a particular field project depends on many different factors such as the availability of a type of equipment, experience with the adopted field technique, data analysis and model calculation procedures, the dimensions of the geothermal regions and systems, the intensity and kind of artificial electrical noise, distortion of the source field by topography or near and distant lateral inhomogeneities, etc.,

4.1. The Electrical Methods

This group of methods includes vertical electric soundings (VES) such as Schlumberger or Wenner soundings, dipole-dipole methods (mainly collinear dipole-dipole profiling), bipole-dipole mapping (total field resistivity mapping), and quadripole resistivity mapping. Electrical and electromagnetic measurements have been performed in many geothermal regions all over the world. As, in the opinion of the author the most modern methods have been utilised in the U.S.A., many of the field techniques and results described here have been undertaken in the U.S.A..

As indicated by the title of this paper, mainly electromagnetic techniques will be discussed. Nevertheless, it seems useful to give a brief overview of the four-point-techniques done in geothermal areas.

A major advantage of the four-point-techniques – which applies to all other controlled source techniques – is the limited extent of the current system created. As a result therefore, field distortions caused by distant lateral conductivity anomalies are not as strong as is often the case with natural source methods. Furthermore the dimensions of the source field can be adapted optimally to the dimensions of the structures being investigated. Disadvantages of the four-point-techniques, compared to the natural source methods, are the limited depth of current penetration and the interpretational difficulties which arise when investigating a three-dimensional structure by a three-dimensional source field.

Schlumberger and Wenner methods were used for most of the early resistivity surveys, especially in New Zealand (Banwell and Macdonald, 1965; Hatherton *et al.*, 1966; Macdonald and Muffler, 1972), but have also been used in some recent surveys (e.g. Meidav and Furgerson, 1972; Zohdy *et al.*, 1973; Stanley *et al.*, 1976; Tripp *et al.*, 1978; Razo *et al.*, 1980). The main disadvantages of Schlumberger and Wenner profiling are of a logistical nature since to lay out cables of 10 km length and more, is neither time- nor cost-effective.

In contrast to Schlumberger and Wenner profiling, dipole-dipole collinear profiling needs only short cables and is said to provide good resolution of conductive structures as well as good depth of information. Dipole-dipole measurements have been made and are still made in many geothermal areas (e.g. Klein *et al.*, 1975; McNitt, 1976; Garcia, 1975; Jiracek *et al.*, 1976; Patella *et al.*, 1979; Baudu *et al.*, 1980; Wilt *et al.*, 1979).

Bipole-dipole mapping has been used in several geothermal areas as a reconnaissance method (e.g. Risk *et al.*, 1970, 1976 a, b; Bibby and Risk, 1973; Jiracek and Gerety, 1978; Williams *et al.*, 1976; Beyer *et al.*, 1976; Soutu, 1978). As the bipole of several km length remains fixed during each survey the field measurements are relatively fast. Theoretical investigations and numerical model calculations, however, have shown (Dey *et al.*, 1977) that bipole-dipole results easily lead to misinterpretations when the conductivity pattern is not simple since the measurements are strongly affected by the position of the current bipole relative to lateral boundaries. Conductive structures which are not near the Earth's surface, or near the bipole or dipole, are hardly detected.

The quadripole resistivity mapping technique (Tasci, 1975; Doicin, 1976; Bibby, 1977) has been developed to eliminate the disadvantages of the bipole-dipole mapping method. By using two source bipoles instead of only one (approximately perpendicular to each other) and two rectangular receiver dipoles, two independent resistivity maps can be produced. The quadripole method has been found to be superior to the bipole-dipole method particularly in defining true resistivities and delineating boundaries. A quadripole survey was undertaken by Harthill (1978) in the geothermal regions of the Imperial Valley.

4.2. Borehole Measurements

Little has been published about electromagnetic studies in boreholes and between boreholes in geothermal regions. One technological problem is the high temperature in the boreholes. Electromagnetic probing in the MHz range between pairs of boreholes (Lytle *et al.*, 1979) may be useful in delineating fracture zones under certain conditions, associated with the conductivity contrast between the host rocks and the hot water/vapor.

4.3. Techniques with Natural Source Fields

These techniques include geomagnetic deep sounding – magnetovariational sounding – differential geomagnetic sounding, the telluric and telluric-magnetotelluric mapping methods, magnetotellurics and audio-magnetotellurics.

In geomagnetic deep sounding surveys, the three components of the natural geomagnetic variations are measured at roving field stations as well as at a base station. The transfer function between the vertical component at a field station and the maximum correlated horizontal component at the field or a base station is calculated as a function of the period T. The transfer functions for the different field sites may be plotted on a map as induction arrows (or tipper). Sometimes also the transfer functions between correlated horizontal components at field and base stations are computed. These may be plotted as perturbation arrows (Schmucker, 1970).

Geomagnetic deep sounding surveys of local geothermal systems, which are of interest for exploitation, are scarce. This can be explained by the fact that the induced magnetic field, being an integral over the current density distribution, is not very sensitive to small variations of amplitude and direction of current density. In combination with magnetotelluric surveys, however, geomagnetic depth sounding surveys may provide additional information, which may help to explain the influence of lateral conductivity distribution on the magnetotelluric sounding curves, especially when the conductivity distribution is not too inhomogeneous. Combined geomagnetic deep sounding/magnetotelluric measurements have been made in the region of the geothermal anomaly of Urach in SW-Germany (Richards *et al.*, 1982; Berktold and Kemmerle, 1982). The period dependence of the length and direction of the perturbation arrows as well as of the preference directions of the induced electric field at the different measuring sites suggested a superficial conductivity anomaly in the area of this geothermal anomaly. The maximum distortion effect of the induced current system within the conductivity anomaly occurred at periods of 25 – 150 s. The lengths and directions of the induction arrows were rather inhomogeneous within the geothermal anomaly for short periods ($T < 100$ s). Since the sedimentary cover is more or less plain layered in the area this variability may be caused partially by tertiary volcanic intrusions of basalt and partially by the hot water distribution. With increasing period the lengths and directions of the induction arrows becomes more and more stable. The apparent resistivity and phase curves in parts of this geothermal area were strongly influenced by the inhomogeneous lateral conductivity distribution.

Differential geomagnetic soundings have been done by Mosnier and Babour near the geothermal anomaly of Pechelbronn in the Upper Rhine Valley (Babour and Mosnier, 1977; Mosnier and Babour, 1980). Variometers with an accuracy of better than 0.05 nT (Mosnier, 1970) were used to measure the horizontal components of the geomagnetic variations. The NS-component H_0 (t) and the EW-component $D_0(t)$ of the magnetic field variations at a base station outside the Rhinegraben were subtracted in the time domain from the corresponding magnetic field components $H_i(t)$ and $D_i(t)$ at each of the 'i' field stations. These magnetovariational field differences $H_i(t) - H_0(t)$ and $D_i(t) - D_0(t)$ represent the anomalous horizontal magnetic field correlated with a conductivity anomaly. These anomalous magnetic field variations were found to be linearily polarized and independent of the period. A local maximum of the anomalous horizontal magnetic field was found in the geothermal area of Pechelbronn. This was superimposed on the anomalous field of the Rhinegraben.

The telluric mapping method has been used in several geothermal areas for reconnaissance measurements (Bodvarsson *et al.*, 1974; Maas, 1975; Mabey *et al.*, 1978; Jackson and O'Donnell, 1980; Long *et al.*, 1980; Berktold *et al.*, 1982). In this method two perpendicular components of the induced electric field are measured at roving field stations within the geothermal anomaly. The electric field components at the roving stations are compared with the corresponding components at a base station. In most of these surveys the telluric fields have been measured over a limited period range, e.g. $T \approx 20$–50 s or $T \approx 150$ s.

The lateral distribution of the induced electric field for a given period T is caused by the superposition of the effects of conductive structures inside and outside the area of interest. The effect of the geothermal anomaly alone is often difficult to separate from the effect of other conductive structures in the area. The later undesired effect may be due to the direct current distortion of the induced electric field by small conductive structures at or near the measuring site. On the other hand – as the telluric current system is of great lateral extent – the electric field distribution within the geothermal anomaly also depends on the conductivity distribution outside the geothermal anomaly. This is a function of the dimensions of the external conductive structure and the given period and is clearly a disadvantage of the telluric method compared to the controlled source methods, where the dimensions of the source field can be adapted to the dimensions of the geothermal anomaly of interest. The telluric mapping method, however, does have the following advantages: (1) Because of the large period range available, deeper parts of a geothermal anomaly can also be mapped. (2) As the method is fast and inexpensive in its field application, a dense net of stations can be set up. This is an important fact in an area of inhomogeneous conductivity distribution as is the case for geothermal anomalies.

Most of the telluric field measurements in geothermal areas have been analyzed using the 'vectogram' method of Yungul (1968). In Yungul's vectogram method the two components of the horizontal electric field are amplified, bandpass filtered and fed into the two channels of an x-y recorder. This is done simultaneously at a field station and at the base station. Depending on the polarization of the inducing magnetic field and on the lateral conductivity distribution, the measured electric field variations are quasi-elliptically polarized loops. For corresponding field and base loops (vectograms) the ratio of the areas of the loops can be calculated. This ratio is called the J-value. To obtain a representative J-value for a site several pairs of loops caused by inducing magnetic fields with different directions must be averaged. The J-values for a limited period range are then plotted on a map.

Maas (1975) has reported on telluric mapping measurements in the vicinity of the Mesa geothermal anomaly in the Imperial Valley of California. The period range of measurement was around 150 s. The most striking feature of the J-value map was a large negative J-anomaly coinciding in position with the Mesa geothermal anomaly. Comparison of the J-contours with the shallow thermal gradient data shows a strong agreement in the position and shape of both anomalies. Long and Kaufman (1980) have measured the telluric field at 31 sites in the known geothermal resource area of Weiser, Idaho and Vale, Oregon. The period range of measurement was 10–50 s. The survey covered an area on the western edge of the Snake River plain where highly conducting sediments (1–20 Ohm m) occur in the uppermost 1000–1500 m. The field data were presented as J-value maps. The telluric currents were found to be strongly influenced by the highly conductive near-surface layers which also affected the AMT results made in the same area. The telluric current results however also reflect the presence of deeper structures and the basement topography. Used in combination with AMT measurements they located high conductivity anomalies near prominent hot springs. The limited extent of these anomalies suggests that near surface thermal waters are restricted to a few narrow fault zones in the immediate vicinity of the hot springs. Jackson and O'Donnell (1980) made telluric measurements in the Coso Range in east central California. The period range of measurements was again 10–50 s. Data were presented in the J-value map shown in Figure 5. This telluric map was found to delineate the major resistivity features of the Coso Range and Rose Valley to the west and Coso Basin to the south.

Yungul's vectogram method is a qualitative data reduction method and does not make the best use of telluric field data. Nowadays – as fast data analysis by minicomputers is available in the field – an intensive and quantitative data analysis does not contradict the advantages of the telluric mapping method.

Additional information is available if at the base station a full set of magnetotelluric equipment is used to measure also the two horizontal components of the magnetic field. This method has been called the telluric-magnetotelluric method by Hermance and Thayer (1975). The authors calculated the magnetotelluric impedance tensor at the base station and in addition a telluric transfer tensor between the electric fields at the base station and at the different field stations. If the magnetic fields at the base and field site are equal – which, of course, is not often true – then the impedance at the remote site can be calculated by multiplying the base impedance tensor with the telluric transfer tensor.

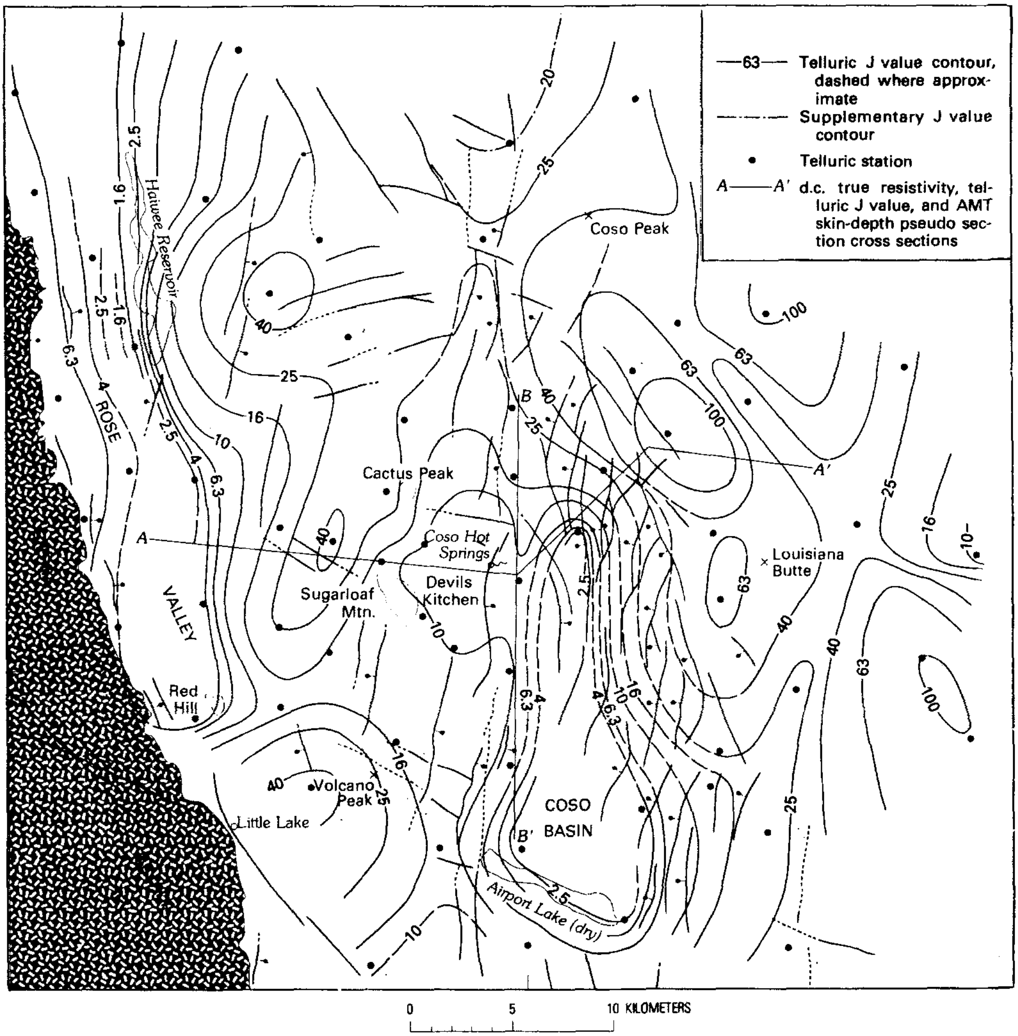

Fig. 5. Telluric J-value map contoured at a logarithmic scale of five intervals per decade. Low and high values of J are equivalent to low and high apparent resistivities with respect to the base station 3.5 km east of Coso Hot Springs (Jackson and O'Donnell, 1980).

Fischer (1982) has argued that this method is closely related to the Gamble *et al.* (1979) remote reference method which will be discussed later. The main similarity between both methods is that observations at distant sites are correlated. As a consequence the local perturbations, which are assumed to be uncorrelated will not affect the determination of the telluric transfer tensor, the scattering range of which should be quite small. If the base station is a good site, then the base station impedance tensor can be determined with accuracy and the scatter in the approximate transfer impedance tensor will be small too.

Hermance and his coworkers have done telluric-magnetotelluric measurements in geothermal regions of Iceland and in other areas (Hermance *et al.*, 1976; Thayer *et al.*, 1981). Measurements of a similar kind have also been made by Bodvarsson *et al.*, (1974) in geothermal areas of Oregon. Goldstein and Mozeley (1978) have reported on telluric-magnetotelluric measurements at Mount Hood, Oregon, in combination with remote reference measurements. In this study, the instruments were set up in clusters. A complete cluster yielded six data stations, two being magnetotelluric stations and four being remote telluric stations.

The next two methods to be discussed are the magnetotelluric method (MT) and the audio-magnetotelluric method (AMT), which just differ in the period range of measurement. Periods of about one second to several thousands of seconds are measured in MT while frequencies of about 1 Hz to several kHz are measured in AMT.

In the last few years major advances have occurred in instrumentation, field techniques and data processing. Josephson effect magnetometers (SQUIDS) seem to have become the standard for reliable low noise magnetic field sensors (Clarke, 1980). For higher frequency applications highly sensitive induction coil magnetometers are available. Minicomputers are used more and more in the field.

After SQUIDS had been used in some field surveys it soon become obvious that the use of the SQUIDS could not significantly improve the quality of the MT data when artificial electromagnetic fields produced strong data scattering. This led Gamble *et al.* (1979) to introduce a second remote magnetometer as a reference – a method capable of yielding high quality sounding curves. The reference magnetometer must be far enough from the survey area so that the artificial noise in both areas is uncorrelated, with one or more MT systems being set up in the survey area. If conventional MT data processing is used the impedance tensor elements are biased upwards or downwards by the noise according to which component the noise is associated and which autopower occurs in the formulae. In this case the noise does not average to zero regardless of how much data are acquired and analyzed. This bias has been shown to be the major source of error in the determination of impedance tensor elements. Using a remote magnetic reference the noise in both the auto and crosspower spectra can be reduced provided there is no correlation between the noise in the remote channels and the noise in the local channels. Unbiased estimates of the auto- and crosspower of the signals in each channel as well as of the noise in the different channels can be obtained in the remote reference method.

Remote reference measurements have been done in geothermal areas such as High Cascade/Oregon, Klamath Basin Oregon/California, Grass Valley/Nevada, Roosevelt Hot Springs/Utah and Cerro Prieto/Mexico (Lawrence Berkeley Laboratory Annual Reports, 1978–1980, Gamble *et al.*, 1980).

In the Cerro Prieto geothermal area, up to 1980 the method had been used at more than 20 stations along three NE-SW lines, an example of the sounding curves at one site being shown in Figure 6. The MT-measurements indicated a zone of relatively high resistivity enclosing the region of current brine production. The results of the MT-measurements were compared with those of Schlumberger and dipole-dipole sound-

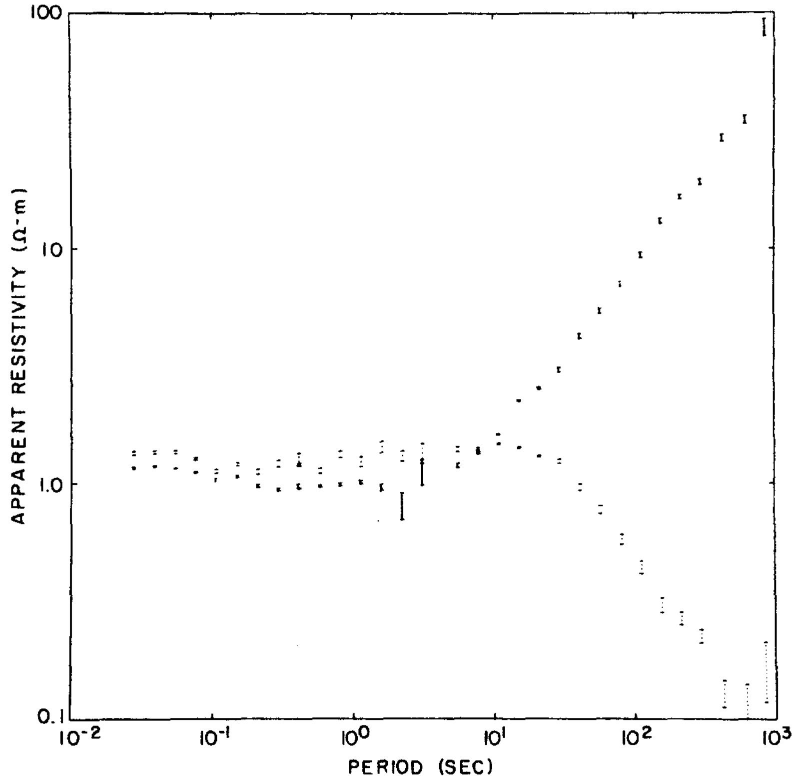

Fig. 6. Rotated apparent resistivities and their probable errors as a function of the period for station 2 of the remote reference magnetotellurics at Cerro Prieto/Mexico (Gamble *et al.*, 1980).

ings previously made in the area (Wilt *et al.*, 1979). Two-dimensional model calculations were done using a simplified version of the electrical model suggested by Wilt *et al.* as an initial guess. The final model fitting the TE-mode differs only slightly from the model proposed by Wilt *et al.*, 1979.

Magnetotelluric measurements have been made in many geothermal areas of regional and local scale such as in the Pannonian Basin of Hungary and in other geothermal regions of East-Central Europe and Soviet Asia (Adam, A. (ed.), 1976); in the Larderello/Travale area of Tuscany in Italy (Celati *et al.*, 1973; Berktold *et al.*, 1982; Haak *et al.*, 1982; Hutton *et al.*, 1982), in France (Malerque, 1980; Dupis *et al.*, 1980), in SW-Germany (Richards *et al.*, 1980; Berktold *et al.*, 1980), in Iceland (Hermance and Thayer, 1975; Hermance *et al.*, 1976; Thayer, 1975; Björnsson, 1976; Hersir, 1980; Beblo *et al.*, 1980; Arnason, 1981; etc.), in the United States (Gamble *et al.*, 1980; Boehl *et al.*, 1977; Aiken *et al.*, 1981; Goldstein and Mozeley, 1978; Goldstein *et al.*, 1982) as well as in western parts of Canada, S-America, East-Africa, Indonesia, Japan, New Zealand, etc.

Audio-magnetotelluric (AMT) measurements have also been done in many geothermal areas, mainly in the western United States. The natural signals measured in AMT originate in worldwide lightning storms, most of the energy being produced by storm cells in tropical regions. During the summer-time local thunderstorms also contribute to the AMT signals. These tend to peak in their activity in the early afternoon, local time. On the average, AMT field strengths are greater at low latitudes, in the afternoon hours and in summer season (Strangway *et al.*, 1973). The sources for AMT contain a wide spectrum of frequencies, but due to the characteristics of the wave guide formed between the Earth's surface and the ionosphere, many of these frequencies tend to be attenuated. Reasonably strong energy peaks are observed at the Schumann wave-guide resonances of 8, 14, 20, and 25 Hz. Below 200 Hz the spectrum is fairly flat and the signal strength is sufficient for field operations throughout the year. Above 200 Hz the absorption in the Earth-ionosphere wave-guide increases and it has a maximum near 2 kHz. This makes data acquisition more difficult in the frequency range from about 200 – 5000 Hz.

Data scatter in AMT has sometimes been reported to be rather high. This is partially due to the variation of source location during a given recording period, e.g. two or more separated storm centres may supply energy at different times during the recording period. Furthermore, AMT-data are sometimes not repeatable at a measuring site. This may be due to the fact that the source locations have been different at the different recording times. These effects are strongest near lateral inhomogeneities of conductivity. Other sources of strong scattering are artificial current systems and the harmonics from local 50 or 60 Hz power lines. These harmonics can be found up to more than 2 kHz and still present problems for optimal filtering. On the other hand, effective use can be made of the power line harmonics as an artificial source. This is particularly attractive in the range of the absorption band where the natural signals are weak and the minimum source-receiver distance of 3 skin-depths for plane wave interpretation is easily reached.

Although many sources of data scattering exist, the AMT technique has proven to be a major one for reconnaissance work in geothermal areas as shown by the AMT surveys which have been made by the US Geological Survey in more than 40 geothermal areas of the western United States (Hoover *et al.*, 1978). Twelve selectable and fixed frequencies between 7.5 Hz and 18.6 kHz were used for these AMT field measurements. In all the areas surveyed, a consistent correlation was observed between low resistivity regions and known surface, or near surface, geothermal activity (hot springs, hot water wells, etc.,). To examine possible resistivity differences between different types of geothermal regions or geologic provinces, resistivity histograms were prepared for typical geothermal regions (Basin and Range, Columbia Plateau, and Vale-Weiser region) using apparent resistivity values obtained at 7.5 and 27 Hz. These histograms, however, reflect mainly the different types of rocks occurring preferably in the different regions. For each geothermal area a correlation of average apparent resistivity values at 7.5 Hz with major areas where the heat flow is over 2.5 hfu was attempted. Comparison of the average of the measured apparent resistivities of each known geothermal resource area and the heat flow shows an inconclusive relationship but a closer association is obtained with hydrothermal convection systems of high temperature and large stored heat capacity, such as the Vale-Weiser region of Oregon and Idaho and in the NW-part of Nevada.

A 7.5 Hz apparent resistivity map of the Rose-Valley/Coso-Range area of east-central California was published by Jackson and O'Donnell (1980). This looks similar to the J-value map shown in Figure 5. A similar 7.5 Hz apparent resistivity map has been published for the Weiser/Idaho and Vale/Oregon region (Long and Kaufman, 1980). Large conductivity anomalies were found to be located near prominent hot springs in these AMT contour maps. MT-measurements have also been made along a profile extending from the Raft River geothermal area in southern Idaho to Yellowstone National Park in Wyoming (Stanley *et al.*, 1977) These MT-soundings revealed a highly anomalous crustal structure involving a conductive zone at depths between 18 km in the central part of the eastern Snake River Plain, 7 km beneath the Raft River thermal area and only about 5 km in Yellowstone. Resistivities in this conductive zone are less than 10 Ohm m and at some sites less than one Ohm m.

4.4. Controlled Source Electromagnetic Methods

For controlled source electromagnetic methods some of which are also called active or controlled source audio-magnetotellurics, the subsurface may be energized in two

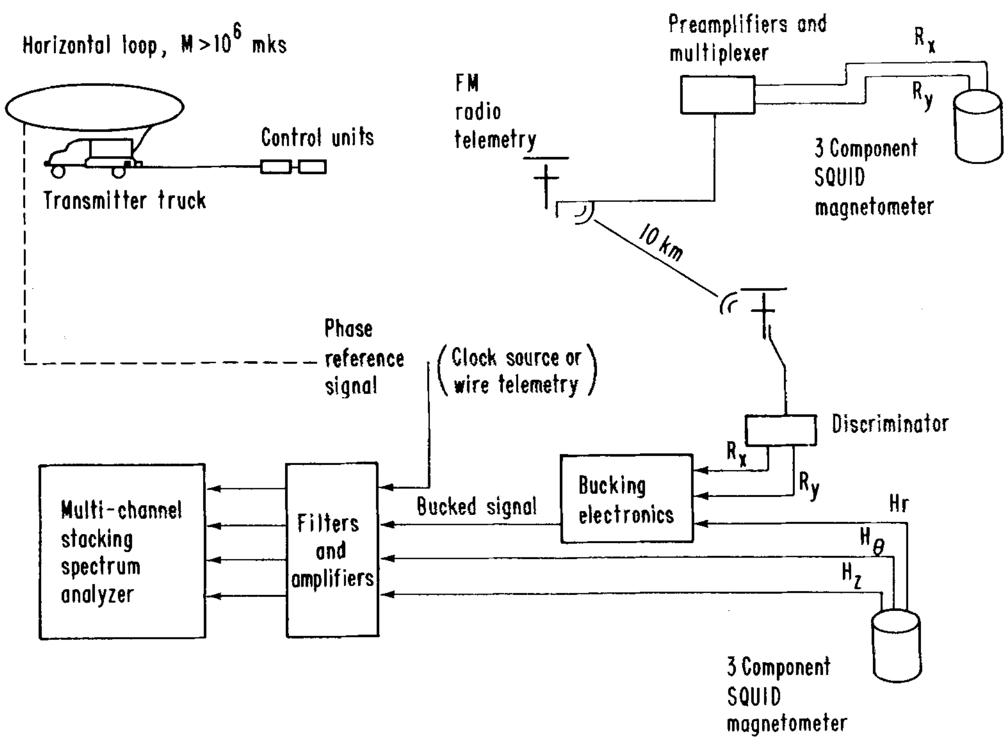

Fig. 7. Schematic diagram of the EM-60 horizontal-loop prospecting system (as used 1979), (Wilt *et al.*, 1981).

ways. A grounded wire may be used for galvanic current injection or a closed current loop for an inductive energizing of the subsurface.

Such a horizontal loop electromagnetic system (EM-60) has been described by Wilt *et al.* (1981), as shown in Figure 7. A square wave current is created by an alternator (max. 150 V, 400 A) Typically, $\pm$ 65 A are applied to a 100-m-diameter four-turn horizontal loop. This provides adequate signal for transmitter-receiver separations less than about 4 km. Maximum depth of information is thus also about 4 km. The frequency range of measurement is 10^{-3} to 10^3 Hz. The magnetic fields at the receiver stations are measured with a three-component SQUID-magnetometer (vertical, radial and tangential components with respect to the transmitter loop). To reduce the effect of natural geomagnetic noise, a second (reference) magnetometer is placed far enough from the transmitter loop (usually about 10 km) so that the observed remote fields will consist only of the natural fluctuations. The remote reference signals are transmitted to the receiver stations using FM radio telemetry. Before the source loop is energized, the remote signals are inverted, adjusted in amplitude, and then added to the receiver station geomagnetic signals to produce essentially a null signal. Once the loop is energized, the resulting receiver magnetic signal is essentially free of geomagnetic noise. The resulting signal to noise improvement of roughly 20 dB enables reliable data to be obtained up to 0.05 Hz.

The EM-60 method is believed by the authors to be a significant improvement in geothermal exploration over dc resistivity and magnetotellurics for three reasons: (1) The maximum depth of exploration with the EM-60 is approximately equal to the distance between the transmitter and receiver. For dc resistivity measurements, almost five-times the source-receiver separation is required for the same depth of exploration; (2) the EM-60 method can provide comparable field data in less time and at less expense than dc resistivity or magnetotellurics; and (3) distant lateral inhomogeneities, which often affect MT data, have a relatively minor influence on EM-60 data because the strength of the fields decreases sharply with increasing distance from the transmitter.

In mountainous field areas loops must sometimes be placed on inclined surfaces since level areas are often not available. Where this occurs, the source dipole must be treated as the sum of a vertical and a horizontal dipole. Theory and field measurements show how ignoring even small inclinations at the source dipole (e.g. one degree) can give misleading results (Haught *et al.*, 1980). This is particularly true in regions of high resistivity, where small secondary magnetic fields may easily be distorted by tilting the source dipole.

For data interpretation, layered model forward solutions may be calculated for a finite-loop source or for a point-dipole source (Ryu *et al.*, 1970; Anderson, 1979). The loop-source solution is perfectly general and is more accurate when soundings are made close to the source. The point-dipole solution is calculated using digital filters and is identical to the loop solution for transmitter-receiver separations greater than 10 loop radii.

For quantitative interpretation a one-dimensional inversion program is available. The inversion program uses the Marquardt least-squares algorithm to fit amplitude, phase and/or ellipse polarization parameters jointly or separately to layered models (Inman, 1975). The author's experience indicates that one-dimensional interpretation seems to give adequate results because of the fast spatial decay of the dipole fields.

Field measurements with the EM-60 system have been made at several geothermal prospects in northern Nevada at Panther Canyon (Grass Valley), near Winnemuka; Soda Lakes, near Fallon; McCoy, west of Austin; Mount Hood, Oregon, etc., (Lawrence Berkeley Laboratory, Annual Reports 1978-1980; Wilt *et al.*, 1981; Goldstein *et al.*, 1981). Results from EM-60 work at Panther Canyon compare very favourably with earlier dipole-dipole resistivity surveys. Both methods adequately outlined an irregularly shaped, buried conductive body associated with a region of high heat flow. The area was covered with EM-60 measurements in just over half the field time required for the dipole-dipole resistivity survey. At Soda Lakes the depth to and inclination of a buried conductive body associated with an area of high subsurface temperature, could be mapped. In this case, the EM-60 results confirmed an earlier MT survey interpretation and gave additional detailed near-surface information. At the remote and mountainous McCoy site, data interpretation was complicated because of the rugged terrain. The EM-60 soundings detected a conductive zone at a depth of 200 m at the south end of the prospect. In addition, EM-60 soundings at McCoy provided information on a deep conductor below 2 km. In the Mount Hood/Oregon area, one-dimensional interpretations of EM-60 and MT data show a similar subsurface resistivity pattern, viz. a resistive surface layer 400–700 m thick, underlain by a conductive layer with variable thickness and resistivity of less than 20 Ohm m. The surface layer is assumed to consist of volcanic rocks partially saturated with cold meteoric water. The underlying conductive zone is presumed to be volcanic material saturated with hot water.

Controlled source electromagnetic measurements using a long grounded wire for galvanic current injection have been made in several geothermal areas, including Hawaii where the length of the grounded wire was some hundreds of m to some km and the injected current was 1–15 A. Measurements have been made partially in the time domain, using a square wave current of 10–15 s period. The vertical magnetic field, produced by the horizontal grounded wire, was measured as a function of time by recording the induced voltage in a coil laid horizontally on the ground (Klein and Kauakikaua, 1975; Keller *et al.*, 1977; Kauakikaua and Mattice, 1981; Kauakikaua, 1981).

A 'Megasource' electromagnetic survey has been carried out in the Dixie Valley/Carson Sink area of north western Nevada by Keller *et al.* (1982). The electromagnetic field was generated by passing a square wave current of 3000 A peak to peak amplitude along a 1 km grounded wire, providing a source moment of 3×10^6 mks. The source was located at a distance of about 50 km from the target areas. Observations of transient electromagnetic coupling using a vertical axis simulated-loop receiver were made at nearly 400 receiver locations distributed at about 1 km distance along the accessible roads in the survey area. The time interval over which the transients were recorded was from a minimum of 30 ms to a maximum of 30 s. Many of the curves were interpreted using a one-dimensional inversion approach. In the upper crust at depths as shallow as 7 km anomalously conductive rocks were detected. Resistivity in the anomalous zone in the upper crust is reasonably uniform, ranging from 5–15 Ohm m. A partially molten zone at shallow depth in the crust may occur in the measuring area, but it seems more likely that the high conductivity results from extensive fracturing and high saturation of hot water in the crust.

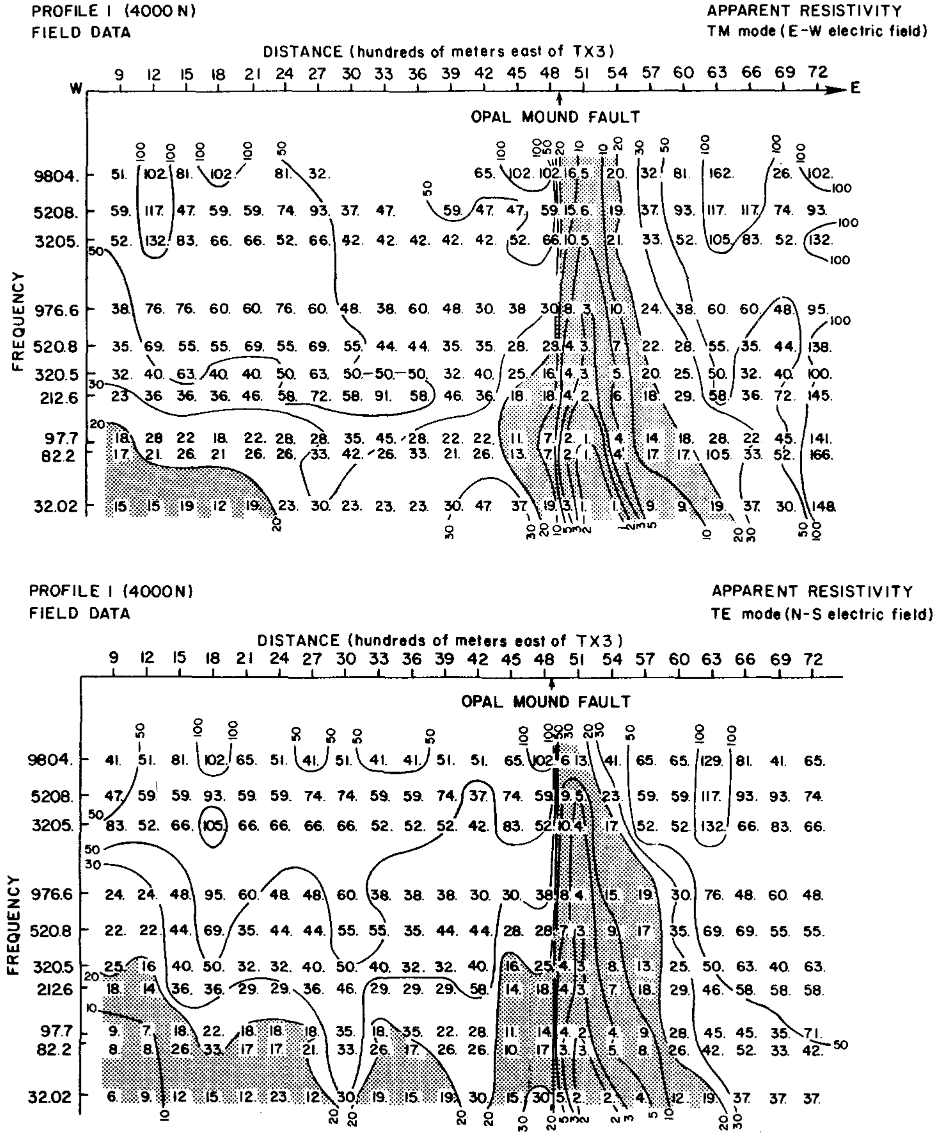

Fig. 8. Controlled-source audio-magnetotellurics at the Roosevelt Hot Springs geothermal area in Utah. Pseudo-sections from profile 1, apparent resistivity TE-mode and TM-mode field data. Apparent resistivity less than 20 Ohm m is shaded (Sandberg and Hohman, 1982).

Sandberg and Hohman (1982) have reported on controlled-source audio-magnetotellurics at the Roosevelt Hot Springs geothermal area in Utah. The Roosevelt Hot Springs are a structurally controlled geothermal reservoir. Geothermal exploration targets are the faults and fractures which control the movement of fluids. Due to alteration minerals and brine, the fault zones respond as low resistivity anomalies in an otherwise moderate to high-resistivity background. Two east-west profiles across the low-resistivity zone were measured with the controlled source equipment (as well as dipole-dipole resistivity mapping). The transmitter consisted of an orthogonal pair of bipoles (609.6 m long), allowing apparent resistivity measurements with the electric field perpendicular and parallel to the traverse. Two sets of data were thereby obtained, corresponding to electric field orientations perpendicular (TM) and parallel (TE) to the geologic strike. The TE- and TM-mode field data of profile 1 are shown in Figure 8. For an initial model, one-dimensional MT inversion of the TM- mode data was undertaken for each station along the profile. This model was then refined by utilizing a two-dimensional MT finite-element forward computer program. Because of the complexity of the area two-dimensional model calculations were made only for the TM-mode. The TM-mode modelling produced interpretations similar to those derived by modelling the dipole-dipole resistivity data. Additionally, however, the controlled source audio-magnetotellurics resolved details not shown by the resistivity modelling. Apparent resistivity mapping was undertaken with the controlled source audio-magnetotellurics at 4 frequencies: 32, 98, 977, and 5208 Hz. The general contour trend and the position of the low resistivity zone compare well with a first separation dipole-dipole resistivity sounding. The low resistivities also coincide with mapped near-surface alteration and brines associated with the geothermal system. Controlled source audio-magnetotelluric measurements have also been made in the Travale geothermal test area in Tuscany/Italy (Otten and Musmann, 1982).

5. Summary

Nearly all available electromagnetic and geoelectric techniques have been tested in the last 25 yr for their usefulness in geothermal exploration and hundreds of field measurements have been carried out in many geothermal regions of the world. Several techniques such as controlled source electromagnetic methods, have been developed especially for the investigation of the uppermost kilometers of geothermal systems. Some of the methods such as magnetotellurics have proven to be more useful for regional studies while others such as dipole-dipole profiling, audio-magnetotellurics and controlled source electromagnetic methods are more efficient for local studies. A great variety of geothermal regions and local geothermal systems have been found to exist with different geological, hydrological and heat transfer characteristics. Examples of the varying characteristics are different host rocks with different structure and stratigraphy, the presence or absence of porous/permeable reservoirs and fracture zones for hot water/vapor storage and circulation, spatial distribution of cold meteoric and of hot water, the existence of thermally insulating cap-rocks, the existence or absence of young magma intrusions to high crustal levels, the kind of heat transfer (convective and/or conductive), etc. Depending on the combination of these different factors each geothermal region or system to be investigated presents a new problem which often is not comparable to that of the geothermal areas already investigated. Nevertheless new geothermal regions and systems have been detected and much detailed information has been extracted from them.

New exploration techniques, data analysis procedures and model calculations have been developed during research of geothermal areas. These are – among others – the controlled source electromagnetic methods, the remote reference field and data analysis technique and the development of better and faster algorithms for direct and inverse calculations of two- and three-dimensional models.

The main problems to be solved in the future are:

– the development and improvement of equipment and field techniques for a more precise delineation and resolution of the conductive structures – mainly of productive zones with high porosity/permeability and fracturing;

– the improvement of computerized data analysis in the field for faster adaptation of the field strategy to the observed conductivity distribution;

– the improvement of interpretation procedures for a more detailed interpretation of the normally rather inhomogeneous conductivity distribution in geothermal systems;

– the provision of a more meaningful and reliable interpretation of the conductivity distribution in terms of geologic models and of a deeper understanding of the structures and processes in geothermal systems.

Controlled source electromagnetic methods with deep current penetration have proven to be rather effective for the investigation of conductive structures and layers at different levels of the Earth's crust – both within and outside geothermal regions. These methods in combination with audio-magnetotellurics and/or four-point-techniques can be jointly inverted to provide optimal resolution of such conductive structures and layers within the crust.

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