CONTROLLED SOURCE ELECTRICAL METHODS FOR DEEP EXPLORATION

STANLEY H. WARD

Earth Science Laboratory, University of Utah Research Institute and *Department of Geology and Geophysics, University of Utah*

Abstract. Application of controlled source electrical methods (CSEM) is impeded by natural field, electrification, geological, cultural, and topographic noise. Lateral resolution of parameters of adjacent steeply dipping bodies and vertical resolution of parameters of adjacent beds in a flatly dipping sequence are concerns with any CSEM method. Current channeling into a localized good conductor from a surrounding, overlying, or underlying conductor poses problems for the interpreter.

A summary of the results of several recent experiments with CSEM techniques illustrates that with care and difficulty they can be used to depths on the order of 20 km. If measurements are made on a relatively uniform resistive surface, as can be found in glaciated Precambrian terranes, then either a grounded bipole or a loop source is acceptable. Most of the recent CSEM experiments were made over resistive Precambrian rocks and all were directed toward detecting a conductive layer near 20 km depth.

For exploration beyond this depth, however, the MT/AMT method would seem to be preferred. The rationale behind this conclusion is largely contained in consideration of the ratio of signal to natural field noise. Where thick irregular surficial overburden of low resistivity occurs, two- and three-dimensional modeling is necessary to stripp off the effects of the shallow layers. This may not be possible for CSEM and then MT/AMT becomes the only alternative.

1. Introduction

Controlled source electrical methods (CSEM) have been used in a limited way in exploration of the deep crust and upper mantle. Problems encountered with CSEM in this application include lateral and vertical resolution, current channeling, and natural field, electrification, geological, cultural, and topographic noise.

After briefly reviewing the above problems, a summary is given of some recent results using CSEM in exploration of the deep crust and upper mantle. A conductive horizon at a depth near 20 km is commonplace in these studies, yet it is difficult to be certain of its geological meaning. Most of these studies have been conducted over resistive Precambrian terranes.

When one attempts to apply CSEM to unravel geological problems in terranes where heterogeneous conductive geological materials exist near surface, deep crustal and upper mantle exploration by CSEM becomes problematical. Two- and three-dimensional earths must be used in interpretation in such areas. We conclude by demonstrating, via some simple models, the difficulty of applying CSEM to regions in which the earth is 3D near surface.

2. Problems with CSEM

We shall briefly summarize each of the problems encountered in application of controlled source electromagnetic methods.

Geophysical Surveys **6** (1983) 137-152. 0046-5763/83/0062-0137$02.40.

2.1. Natural field noise

The slope of the natural field spectrum below 1 Hz is as steep as f^{-4} where f is the frequency of observation (SanFilipo and Hohmann, 1982). Such a steep spectrum places serious constrants on measurements below about 0.03 Hz (Connerney *et al.*, 1980; SanFilipo and Hohman, 1982). Coherent stacking of data for days at a time is commonplace. SanFilipo and Hohmann (1982) demonstrate that high pass filtering prior to stacking reduces the time required for stacking. Ideally, one would use a remote reference (Gamble *et al.*, 1979) to cancel coherent noise, but as SanFilipo and Hohmann (1982) demonstrate, the separation of the remote reference from the transmitter must be at least five times the separation of the receiver from the transmitter. Since separation of the receiver from the transmitter ranges up to 100 km, a serious problem is posed in transmitting, reliably, the remote reference information to the receiver site. Hence, remote reference has not been used in deep crust/upper mantle CSEM studies.

2.2. Cultural noise

Cultural developments create active and passive noise. Circuits completed through fences, pipelines, power lines, telephone lines, rails, and other conductive cultural structures produce anomalies largely unrelated to subsurface geology. These sources of noise can, in rare instances, be reduced by removing the offensive structure, but, by and large, they can only be avoided by placing transmitters and receivers well away from cultural structures. This is not always possible in areas of concentrated industrialization, and hence important geological problems simply cannot be attacked in such areas.

Some of these cultural developments also serve as sources for narrow or broad-band electric and magnetic field noise, especially power lines, telephone lines, and electrified rails. Further compounding the problem is the fact that active sources of cultural noise induce eddy current noise in the passive cultural noise sources, such as fences and pipelines, which then reradiate the noise.

2.3. Overburden and lateral inhomogeneities

Overburden can be described variously as unconsolidated sediments, weathered rock, or consolidated sediments above crystalline basement. The overburden may be resistive (e.g., glacial gravels) or conductive (e.g., unconsolidated valley sediments or deep basin consolidated sediments).

In glaciated terrane both the overburden and the weathered layer may be totally removed as is typical of areas of extensive Precambrian exposure in North America and Scandinavia. When present the weathered rock is invariably conductive because the geological process of weathering leads to (a) increased porosity, (b) increased presence of clay minerals with their attendant surficial electrical conduction, and (c) increased concentrations of ions in the pore waters of the weathered rocks. In dry climatic environments, evaporation strongly increases the concentration of ions in the pore waters, on average. Dilution of these ions takes places during the rainy season in dry or wet invironments. A worldwide study of these factors points out that the shallow resistivity profile is closely related to local climatology, to glaciation, and to tectonic style. The depth of overburden related to weathering seldom exceeds 100 m but can easily reach 2 km for valley fill in the Great Basin of the U.S.A. (Eaton, 1982).

Consolidated sediments in basins often will include saline evaporites or shales while unconsolidated sediments in deep valleys may include saline evaporites and lacustrine clay-rich deposits. Sedimentary rocks in such areas as the Gulf Coast oil-producing region of Texas and Louisiana in the U.S.A. commonly exhibit resistivities as low as 1 to 10 Ω-m due to interstitial brines (Pirson, 1963). This is also true of brine-saturated consolidated and unconsolidated sediments in some geothermal areas such as the Imperial Valley in California, U.S.A. (Meidav and Furgerson, 1972) and of deep valley fill containing evaporites and lacustrine deposits, in Nevada and Utah, U.S.A. (Ward and Sill, 1976; Stewart, 1980; Hintze, 1980). Usually such resistivities are relatively uniform within the basins or valleys in which they occur.

Whatever its origin, overburden of irregular shape or resistivity will produce geologic noise which may render difficiult the recognition of vertical variations in the resistivity profile of the deep crust and upper mantle. Wannamaker *et al.* (1982a, b) illustrate how two and three-dimensional modeling can be applied to strip off the effects of irregular overburden from magnetotelluric data. The section on current gathering will elaborate on this problem.

2.4. Resolution and the Effect of Other Geological Noise

To facilitate vertical resolution, i.e., resolution of the resistivities and thicknesses of horizontally layered media, an inductive electromagnetic system would need to sample at three or four frequencies per decade (Glenn and Ward, 1976). At least four decades of spectrum are required if one wishes to explore both shallow and deep layers with parametric sounding only. With combined parametric-geometric sounding this requirement may be relaxed (op. cit.). If lateral inhomogeneities are superimposed on the layering, then an adequate spatial density of receiving stations must also be assured over a distance sufficiently large to permit delineation of all inhomogeneities of interest so that the geological noise they contribute can be removed and the layering more accurately delineated.

It is well known that inductive techniques, passive or active, usually provide information on conductivity-thickness products of conductive layers, whereas they usually provide only thickness information on resistive layers (e.g., Madden, 1971). On the contrary, resistivity techniques usually provide information on resistivity-thickness products for resistive layers and conductivity-thickness products for conductive layers (op. cit.). Vertical resolution of resistive and conductive layers is well illustrated via inversion (e.g., Fullagar and Oldenburg, 1982). Joint inversion of inductive and resistive data sets can markedly improve the resolution (Petrick *et al.*, 1977).

2.5. Effects of Topography

Variations in elevation of the receiver relative to the transmitter will produce *elevation* *errors* in electric or magnetic fields along a traverse, relative to the fields that would be observed over a flat surface. An article by Tripp *et al.* (1978), which considers a problem of smaller scale, is a good study of this effect. These elevation errors can be severe for short separations between transmitter and receiver (op. cit.), but they are not expected to be large for the large separations associated with deep crust/upper mantle studies.

If topographic relief is large, one seeks to assure that a square coil or bipole source is horizontal and that measurements are made of horizontal and vertical magnetic and horizontal electric fields. Alternatively, the plane of the transmitting coil must contain the axis or the plane of the receiving coil and orthogonal magnetic field components are measured relative to this axis or plane. If either one of these alternatives is ignored, *alignment* errors will result.

A third effect of topography must also be considered. If, for example, a transmitter is located below and adjacent to a ridge, *induced currents* will occur in the ridge at the higher frequencies and will contribute a source of noise which may obscure the anomalies due to subsurface features.

2.6. Current channeling

Current channeling occurs due to the discontinuity in electric field, and a consequent charge accumulation, at a boundary between media of different resistivities. The normal component of current density J_n is continuous across an interface while, by definition, the resistivity ρ is discontinuous. Hence at an interface between media 1 and 2 we find

$$J_{n1} = \frac{E_{n1}}{\rho_1} = \frac{E_{n2}}{\rho_2} = J_{n2},$$

forcing E_n to be discontinuous at the interface. If $\rho_1 > \rho_2$ then $E_{n1} > E_{n2}$ in order to satisfy this relation; normal electric fields are larger in the medium of highest resistivity adjacent to the boundary. Therefore, at an interface, dielectric displacement $D = \varepsilon E$ must be discontinuous and the fourth Maxwell equation must be written as $\nabla \cdot D = \rho_s$ where ρ_s is a surface charge accumulation. Since two interfaces bound the sides of a two- or three- dimensional body in a homogeneous exterior, a dipolar charge distribution occurs across the body. When the electric field of this body is added to the inducing electric field, the net result is an electric field distribution that causes the current flow in the external medium to be channeled, or 'short-circuited' into the body.

For two- or three-dimensional sources, current channeling may occur along one or all axes of the body, depending upon the direction of propagation of the plane wavelets associated with the source. Interpretation of CSEM data obtained over an inhomogeneous earth must be made with mathematical techniques which include current gathering in the formulation, as was done at the Roosevelt Hot Springs, Utah, U.S.A. for magnetotelluric data (Wannamaker *et al.*, 1982b).

2.7. Lack of interpretational aids

One of the greatest hindraces to application of CSEM is the lack of interpretational aids. One cannot stress too much the need for cost-effective analytical or numerical solutions to problems involving three-dimensional sources over two-dimensional structures (2D–3D) and three-dimensional sources over three-dimensional structures (3D–3D). Since Hohmann (this volume) treats this problem in detail, I shall not dwell on it here.

3. Applications to the Deep Crust and Upper Mantle

3.1. Some recent deep crustal CSEM studies

Table I summarizes the results of five recent CSEM field experiments. Two experiments used grounded dipole sources of about 20 km length, one used a grounded power line of 1400 km length, one used power and telephone lines in numerous Schlumberger arrays, while one used a loop source of 4.5 km diameter. The other pertinent details of the experiments are included in Table I. Even the most casual inspection of Table I will reveal that these deep crustal studies are logistically difficult. Lienert and Bennett (1977) and Lienert (1979) performed their sounding in the western Basin and Range Province, U.S.A.; Sternberg (1979) performed his in Precambrian terrane of Wisconsin, U.S.A.; Connerney *et al.* (1980) worked in the Precambrian of the Adirondack Mountains, U.S.A.; Duncan *et al.* (1980) made measurements in the Precambrian of eastern Canada; while Van Zijl (1977) made his measurements in the Precambrian of southern Africa. Each of the five experiments detected a relatively conductive (10 to 1000 Ω-m) layer in the deep crust at depths varying from about 20 to 40 km.

A widely cited explanation for low resistivity at such great depths is hydrated rock in the form of serpentinite or amphibolite. This explanation is invoked because crustal temperatures inferred on the basis of anhydrous crustal rock conductivities seem unrealistically high (Kariya and Shankland, 1982; Shankland and Ander, 1982). Parkhomenko (1982) and Olhoeft (pers. comm., 1982) confirm that serpentine at 500 °C and riebeckite (an uncommon amphibole) at 400 °C can exhibit resistivities as low as 10 Ω-m. Serpentine at 500 °C can show values as low as 100 Ω-m (Olhoeft, 1979) or even lower if interconnected magnetite is present (Stesky and Brace, 1973). However, Olhoeft (1981) shows that biotite and hornblende are not exceptionally conductive at such temperatures. The existence of a free water phase in the lower crust as suggested by Olhoeft (1981), Shankland and Ander (1982), and others is appealing in that both low values of resistivity and seismic low volocities at 15 to 25 km depths can be explained by it. Unfortunately water in a gabbroic lower crust is not expected to be free, but rather to be bound in hornblende and perhaps biotite (Burnham, 1979), minerals previously noted as not contributing to low resistivities. In fact, Kay and Kay (1981) find that in the southwestern U.S.A., at least, hydrous minerals of any sort are rare in crustal xenoliths. Only in the upper mantle would the transition from peridotite to serpentinite be accommodated, so that widespread serpentine in the lower crust in general is unlikely. Explanation of low resistivities in the deep crust by means of hydrous minerals remains problematical.

TABLE I

Recent experiments using CSEM for deep crustal studies

Investigators	Source type	Source current	Bandwidth	T_x–R_x Separation	Waveform	Stacking time	Crustal model
Lienert and Bennett, 1977	1400 km grounded power line	300 A	0.001–0.01 Hz	5–55 km	square	20 stacks	0 20 10^2–10^3 Ωm 35–45 1–10 Ωm km 100 Ωm
Sternberg, 1979	22–24 km grounded bipole	70 A	0.5–10 Hz	5–40 km	square	100 stacks	0 10 10^3 Ωm 14–22 10^5 Ωm km 50–1200 Ωm
Connerney *et al.*, 1980	4.5 km dia. loop	65 A	0.05–400 Hz	20–65 km	square	to days	0 10 10^4 Ωm 20 10^3 Ωm 30 10–25 Ωm km 10^4 Ωm
Duncan *et al.*, 1980	20.5 km grounded bipole	5 A	1.0–50 Hz	35–85 km	PRBS	to 10 hr	0 17–29 1.5×10^4 Ωm km 270 Ωm
Van Zijl, 1977	Schlumberger	72 A	D.C.	AB to 1000 km	square	174 stacks	0 28–30 2–10×10^5 Ωm 30–40 10–60 Ωm km highly resistive

Partial melting in the lower crust is the alternate explanation preferred by Lienert and Bennett (1977) for low resistivities in the northwestern Basin and Range Province in the U.S.A. Nevertheless, if free water is absent in the lower crust, then it is uncertain whether or not deep temperatures are sufficiently high for this explanation (ibid.).

Stesky and Brace (1973) note that serpentine with abundant magnetite, as is common, exhibits resistivities less than 10 Ω-m at 6 kbar (15 to 20 km) due to electrical contact between the magnetic mineral grains. The notion of conducting mineral grains in contact to form a low resistivity network, as found by Stesky and Brace (1973), provides stimulus for finding conducting minerals other than magnetite to serve this purpose. Graphite is abundant in Archean rocks, especially schists (e.g., Nash *et al.*, 1981). If the film of graphite was sufficiently thin and was connected around all silicate mineral grains, much less than 1% graphite would be sufficient (Olhoeft, pers. comm., 1982; also see Shankland and Duba, 1982).

One might question the assumption of an infinite line source and the neglect of near-surface geologic noise in the experiment performed by Lienert and Bennett (1977). Duncan *et al.* (1980) admitted that lateral variations in resistivity were probably affecting their interpretations. Van Zijl (1977) performed numerous detailed Schlumberger soundings in order to assess the effects of lateral inhomogeneities. Connerney *et al.* (1980) and Sternberg (1979) seemed to have found locations largely devoid of lateral inhomogeneities.

3.2. Deep Crustal Exploration in the Eastern Basin and Range Province, Utah, U.S.A.

The Basin and Range Province is characterized by NNE-trending mountain ranges and deep sediment-filled valleys (Eaton, 1982). Regional lithospheric extension commencing about 20 m.y. ago yielded the elongate uplifted mountain blocks bounded by normal faults. Erosion filled the intervening valleys with conductive sediments

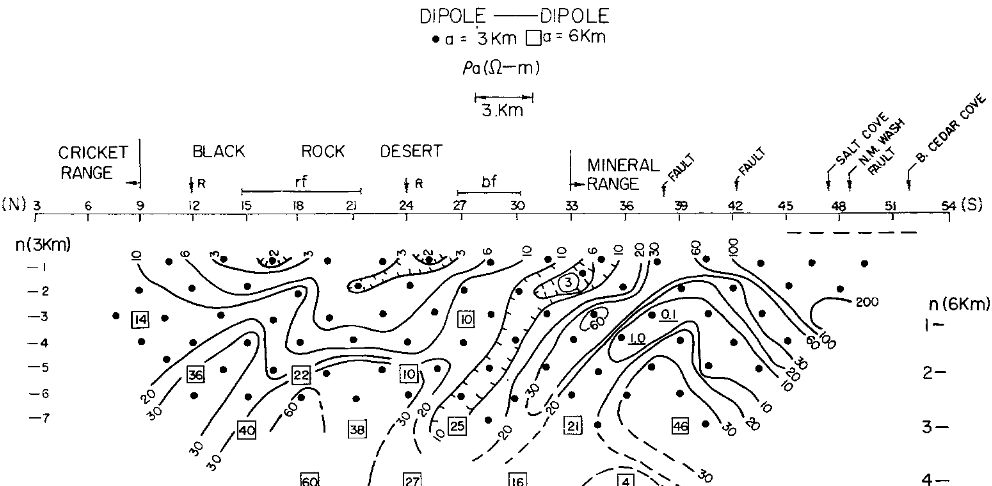

Fig. 1. Resistivity pseudosection from combined 3 and 6 km dipole-dipole survey near Roosevelt Hot Springs, Utah, U.S.A.

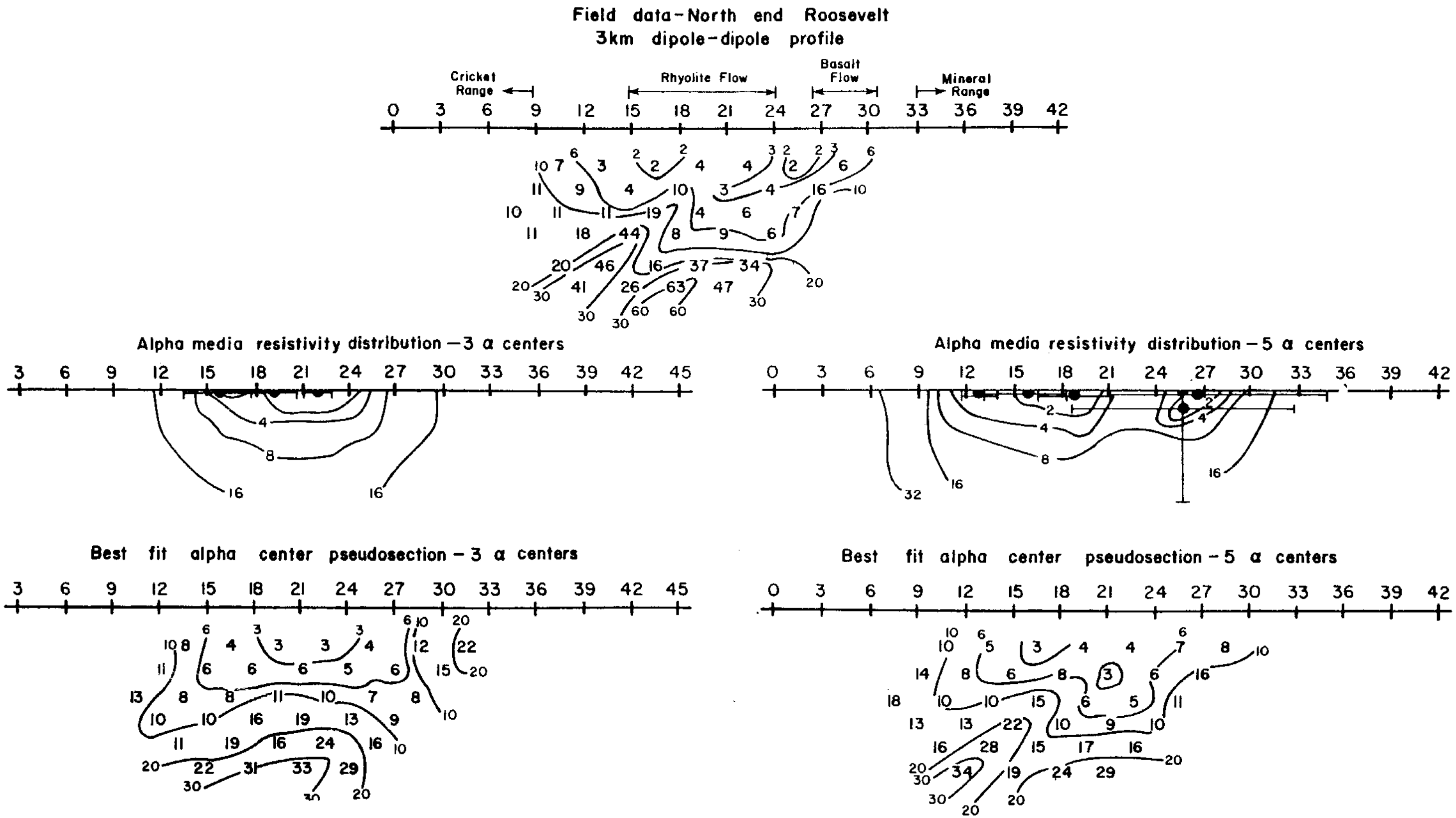

Fig. 2. Dipole-dipole resistivity data and two sets of inversion results illustrating the trade-off between data fit and parameter resolution when more α centers are used in the inverse solution. The plots to the left show the results when only three α centers are used to define the conductivity distribution, while those on the right show the results for five centers. If any deep conductors exist in this region, the α center inversion scheme failed to discriminate between their response and the near-surface conductive areas. The dipole length is 3 km. Data acquired near Roosevelt Hot Springs, Utah, U.S.A. (After Petrick *et al.*, 1981).

commonly as deep as 2 km. The mountains, in contrast, are electrically resistive. Performing electrical exploration of the deep crust and upper mantle beneath this surficially complex geoelectric section requires the use of two- and three-dimensional numerical modeling algorithms.

In 1977 we experimented with crustal exploration in the Basin and Range Province using the dipole-dipole array. The site we chose was near a high-temperature geothermal resource located at Roosevelt Hot Springs near Milford, Utah. Figure 1 shows the pseudosection resulting from this survey. Both 3 km and 6 km dipoles were used and implied, according to criteria employed by Roy and Apparao (1971), depths of exploration of 3 km and 6 km respectively. However, when the 3 km data were inverted to three or five alpha centers, the locations of the alpha centers (Figure 2) were estimated reliably only at depths of less than 1 km (Petrick *et al.*, 1981). The near-surface resistivity inhomogeneities precluded exploration to depths predicted on the basis of a plane-layered earth model. The effects of near-surface inhomogeneities on electrical exploration of the crust are dramatically illustrated by this example.

In 1976, 1977, and 1978 we carried out extensive magnetotelluric soundings in an area of the eastern Basin and Range near Milford, Utah (Wannamaker *et al.*, 1982b). Although the geoelectric section in the top 2 km was clearly three-dimensional, we approximated it by two-dimensional models since the mountain and valley involved in our studies were quite elongate. However, we restricted our interpretations to the TM mode of excitation (or E perpendicular to the axes of mountain and valley) since it is the only 2D mode that includes current channeling (Wannamaker *et al.*, 1982a). The ultimate earth model, which facilitated a virtually perfect match between the resistivity and impedance phase pseudosections of the model and those observed, included a deep valley fill of low resistivity, a geothermal system of low resistivity, and mountain ranges of high resistivity, all underlain by a layered earth to 100 km.

The algorithms we use for modeling MT or CSEM with 3D bodies are modifications of an integral equation formulation attributable to Hohmann (1975). They are expensive to utilize for all but the simplest of bodies. Further, the 3D body cannot be at the surface if convergence of the numerical approximation is to be within time and cost bounds. Still further, they are limited to a resistivity contrast between the 3D body and its surroundings of about 1000 : 1. Because of excessive costs, one cannot consider modeling more than one valley and one mountain at a time. This latter constraint forces us to limit the source dimensions to the valley dimensions for CSEM.

From a practical viewpoint one could not lay out a grounded bipole in the mountains due to lack of access. Even if access were not a problem, the source would assume a very irregular shape if it were draped along or across a mountain range. Thus the source must sensibly be kept in the valley or in the low foothills of the mountains.

All of the above explains our reasons for selecting the source dimensions and the simple generic model of conductive valley fill in the study which follows.

3.3. A GENERIC BASIN MODEL STUDY OF SURFACE GEOLOGIC NOISE

Figure 3 shows the simple model of a conductive valley fill simulating the Milford

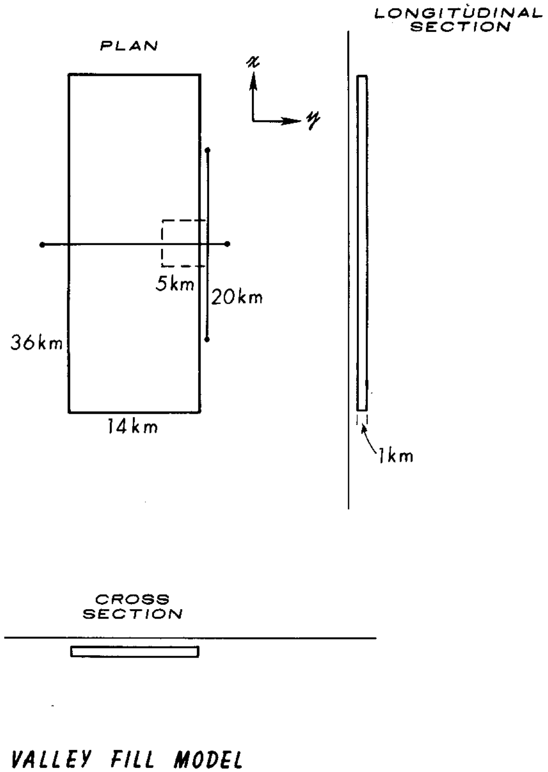

Fig. 3. Model used to calculate approximate response of a valley fill to excitation of a NS 20 km grounded bipole (x directed), an EW 20 km grounded bipole (y directed), and a 5 km square loop.

valley; it is 36 km NS, 14 km EW, and 1 km deep and is buried 1 km beneath resistive overburden. This valley may or may not be typical of Great Basin valleys, but we choose it as a model for these generic studies because of our experience with it. It is excited at 0.03 Hz in turn by a 20 km NS grounded bipole, i.e., parallel to the long axis of the valley fill, by a 20 km EW grounded bipole, and by a square loop 5 km to the side. We chose 0.03 Hz as the lowest practical frequency one would attempt to use with these sources since low ratio of signal to noise would virtually prohibit use of lower frequencies. A somewhat larger NS grounded bipole and a somewhat larger square loop would be used for deep crust/upper mantle exploration, but a larger EW grounded bipole could not sensibly be used because of the mountains. A Schlumberger array is out of the question at this locality, for if AB was set equal to 36 km, the maximum length that could be deployed in the valley, then the depth of exploration would only be 4.5 km.

Figure 4 shows the amplitude and phase of the total horizontal magnetic (EW) field normalized by the primary field values for the north-south grounded bipole. The amplitude of the horizontal magnetic field over the body rises to more than 3 times the primary field. Off the body it falls to 60 % of the primary field. Channeling of the bipole

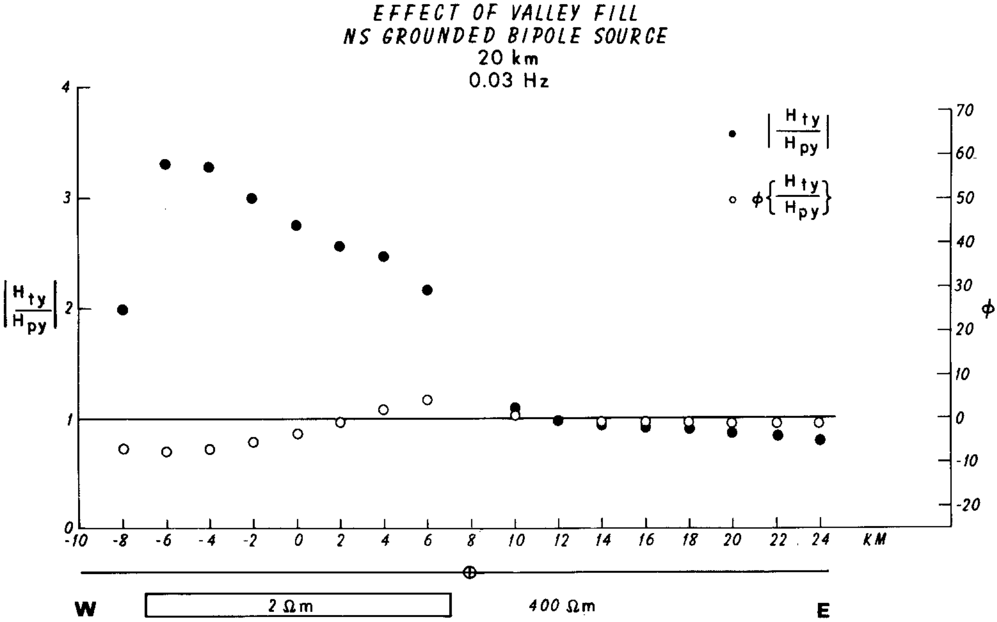

Fig. 4. Amplitude and phase of the normalized horizontal magnetic field for the NS grounded bipole adjacent to the valley fill model shown in Figure 3.

return current through the valley fill would explain the above observations. As far as deep crust and upper mantle exploration is concerned, the source looks like a loop and not a bipole. One could calculate whether or not the various crust and mantle layers found with MT here could be detected with such a complex source. We have not done this. It seems problematical to us that we could model the geoelectric section in three dimensions to a sufficient degree of precision to permit detection of layering at depths greater than the 2 km depth of the near-surface inhomogeneities. In this example we set the NS bipole 1 km to the east of the valley fill because in the Milford Valley a north-south access road occurs in this locality. The bipole could not be moved further east because of the mountains.

Figure 5 portrays the behavior of the amplitude and phase of the normalized east-west horizontal magnetic field for a 5 by 5 km square loop. The distortions in the horizontal magnetic field are every bit as severe as for the NS grounded bipole. The importance of this particular result cannot be overstressed; distortions in $|H_{ty}|/|H_{py}|$ occur out to extremely large distances even for an ungrounded source at low frequencies. The influence of the valley fill on a loop source (i.e., an ungrounded source) is profound.

Figure 6 exhibits the bahavior of the amplitude and phase of the normalized horizontal magnetic field for a 20 km grounded bipole placed across the valley fill model of Figure 3. The distortions are not as great for this source type as they are for the other two but they are still significant. This reduction in distortion probably occurs

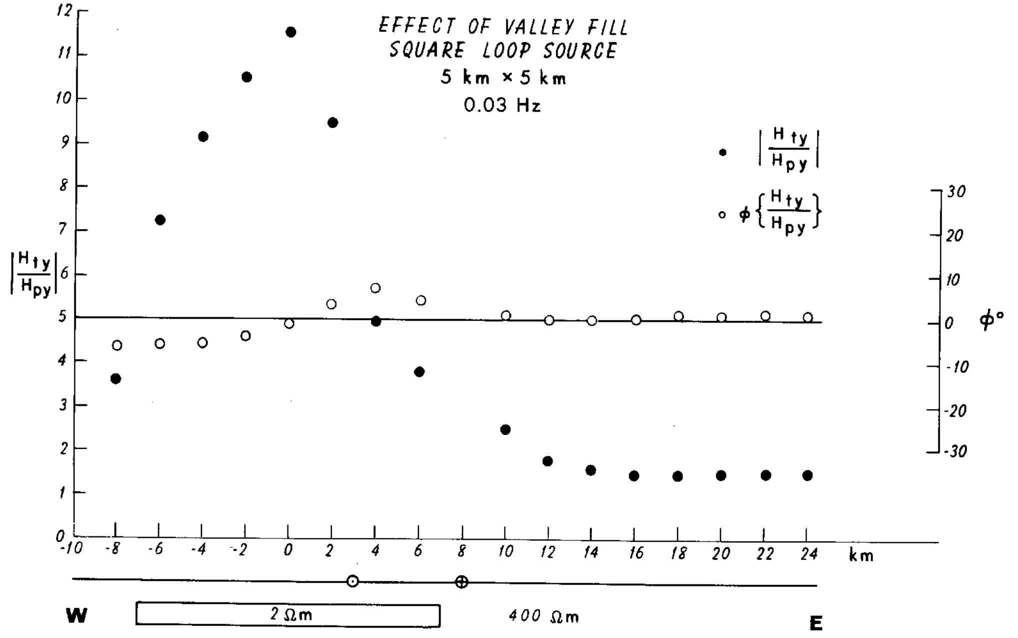

Fig. 5. Amplitude and phase of the normalized east-west horizontal magnetic field for the 5 km square loop overlapping the valley fill model shown in Figure 3

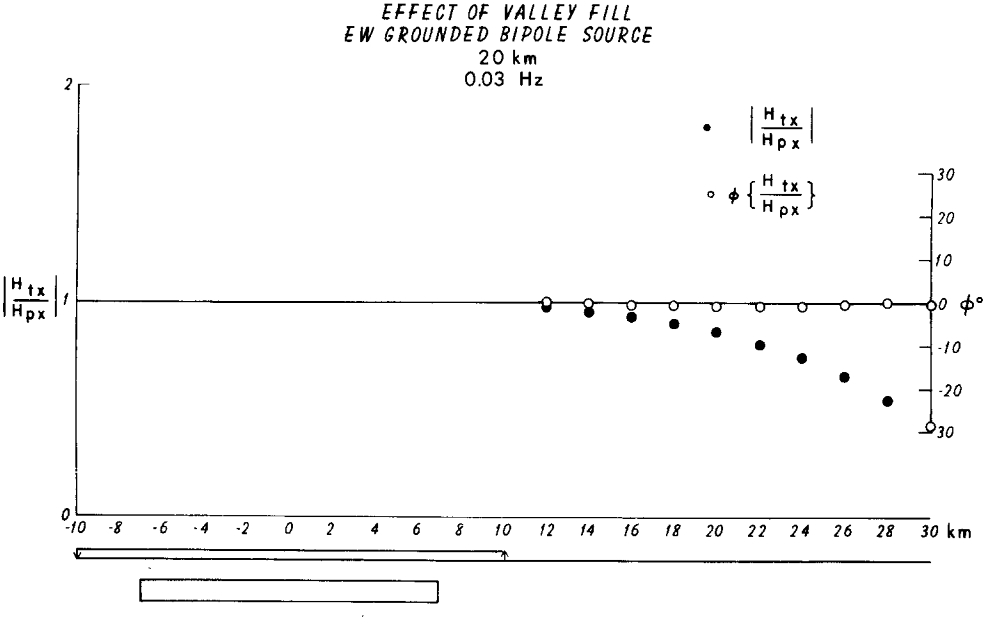

Fig. 6. Amplitude and phase of the normalized NS magnetic field for the EW grounded bipole straddling the valley fill model shown in Figure 3.

because the grounds of the bipole were 3 km on either side of the valley and hence the bipole was not directly 'shorted' by the valley. Such a placement is possible in the Milford Valley but the source would have the appearance of a catenary because of the topographic relief of the sloping valley floor and the mountains. However, the ends would be only 1.5 km above the center of the catenary. This seems to be the most promising source type to use, and since it excites the TM mode primarily, it seems possible that 2D–3D modeling rather than 3D–3D modeling could be used if an algorithm existed for this case. Cost and accuracy improvements are expected with 2D–3D compared to 3D–3D modeling.

3.4. Selection of a Source for Exploration of the Deep Crust and Upper Mantle

For deep exploration one would certainly elect to use transmission lines and/or telephone lines as Lienert and Bennett (1977) and Van Zijl (1977) have done, if such lines were available. They both afford adequate ratio of signal to noise without excessive stacking. Further, they both explore to upper mantle depths. Without such long line sources, which only occasionally become available, one is then left with MT/AMT, grounded bipole, or loop sources. In terranes where thick irregular conductive materials exist at surface it would appear prudent, from our computations, to use MT/AMT. However, if no overburden exists then either a loop source or a grounded bipole source may be used provided their moments are equal and provided ground contacts are not a problem for the bipole source. Unfortunately, as stated before, the extreme requisite separations of remote and base sites make data quality improvement via remote reference impractical.

In Table II we list the minimum characteristics a CSEM system would need for

TABLE II

CSEM for deep crustal probing

- T_x– Bipole 20 km length
  Loop 20 km circumference
- R_x– E_x, E_y, H_x, H_y, H_z, (H sensor RMS noise $< 10^{-4}\gamma$)
- 3×10^{-2} Hz to 10^4 Hz, 4 frequencies per decade
- 50 to 100 AMPS
- R_x distances 1–100 km 4 stations per decade
- 12 bit a to d.
- Coherent detection
- High pass filter plus stacking
- Crystal clock reference

complete crustal sounding and profiling. This system would be used to detect both lateral and vertical variations in resistivity from surface to depths of order 30 to 40 km. The system used by Connerney *et al.* (1980) meets most of the characteristics listed in Table II. These authors prewhitened their data but did not describe otherwise the filter characteristics used to reject low-frequency noise. Their highest frequency was 400 Hz since they did not attempt to map the near surface.

One might want to use both a grounded bipole and a loop source and employ joint inversion to improve the resolution of the layered media in the geoelectric section.

4. Conclusions

If a CSEM experiment is to deduce the lateral and vertical variation in resistivity to depths on the order of 30 to 40 km, it should be placed to avoid regions of electrification, geological, cultural and topographic noise. The main problem then met is dealing with natural field noise; a lower frequency limit near 0.03 Hz probably would be dictated by this noise. While most deep crust/upper mantle experiments performed so far have concentrated on layered resistivity distributions, experiments of the future should be designed to ensure resolution of both lateral and vertical variations in resistivity from surface to the maximum depth of the experiment.

Controlled source electrical methods offer an advantage over MT/AMT in that they do not suffer nearly as much from current channeling in distant overburden or oceans. On the other hand, frequencies as low as 10^{-4} Hz are readily available for MT and the source field has no geometric decay so that deeper exploration is possible with this latter method. Either a grounded bipole or a loop source is satisfactory for deep crustal exploration if a controlled source is warranted. Use of both source types in the same experiment offers the improved resolution of layered models available via joint inversion of the two resulting data sets. CSEM experiments do not seem well suited to delineating deep crust and upper mantle resistivity distributions in regions where irregular overburden exists; the MT/AMT method is preferred there, because of the availability of 2-D modeling algorithms and because of the availability of ultra-low frequencies.

Five recent experiments each detected a low-resistivity zone in the crust at depths ranging from about 15 to 30 km. The geological explanations most commonly offered for this zone is either surficial conduction in hydrated rocks or conduction via free water. Other explanations should be sought. Graphite is a logical alternative. Of the five experiments, four were performed in Precambrian terranes with two using grounded bipoles, one using a loop source, and one using a Schlumberger array via transmission and telephone lines. The fifth experiment was conducted in the western Basin and Range Physiographic Province in the U.S.A. using a long transmission line. The logistics involved in all five experiments were formidable. The uncertainty in physiochemical explanations of deep, low-resistivity layers will never be completely overcome until the uncertainty in the effects of lateral inhomogeneities is removed.

Acknowledgements

I wish to thank W. A. SanFilipo and P. E. Wannamaker for computations of the effects of three-dimensional bodies on the fields of finite sources. Further, I wish to thank W. A. SanFilipo, W. R. Petrick, and especially P. E. Wannamaker for critically reviewing this manuscript. Joan Pingree typed the manuscript while Sandra Bromley and Doris Cullen supervised the preparation of the illustrations; their efforts are sincerely appreciated.

Preparation of this manuscript has been made possible by a grant from the Mobil Foundation, Inc.

References

Burnham, W. C.,: 1979, 'Magmas and Hydrothermal Fluids', in H. L. Barnes (ed.), *Geochemistry of Hydrothermal Ore Deposits*, John Wiley and Sons, New York, pp. 71–136.

Connerney, J. E. P., Nekut, A., and Kuckes, A. F.: 1980, 'Deep Crustal Electrical Conductivity in the Adirondacks', *J. Geophys. Res.* **85**, 2603–2614.

Duncan, P. M., Hwang, A., Edwards, R. N., Bailey, R. C., and Garland, G. D.: 1980, 'The Development and Applications of a Wide Band Electromagnetic Sounding System Using a Pseudo-Noise Source', *Geophysics* **45**, 1276–1296.

Eaton, G. P.: 1982, 'The Basin and Range Province: Origin and Tectonic Significance', *Ann. Rev. Earth and Plan Sci.* **19**, 409–440.

Fullagar, P. K. and Oldenburg, D. W.: 1982, 'Inversion of Horizontal Loop Electromagnetic Soundings', *Geophysics* (in press).

Gamble, T. D., Goubau, W. M., and Clarke, J.: 1979, 'Magnetotellurics with a Remote Reference', *Geophysics* **44**, 53-68.

Glenn, W. E. and Ward, S. H.: 1976, 'Statistical Evaluation of Electrical Sounding Methods Part I: Experiment Design', *Geophysics* **41**, 1207–1221.

Hintze, L. F.: 1980, *Geologic Map of Utah*, Utah Geological and Mineralogical Survey, Salt Lake City.

Hohmann, G. W.: 1975, 'Three-Dimensional Induced Polarization and Electromagnetic Modeling', *Geophysics* **40**, 309–324.

Kariya, K. A. and Shankland, T. J.: 1982, *Interpretation of Electrical Conductivity of the Lower Crust*, Los Alamos National Laboratory Report LA–9050–MS–UC–11, 47 p.

Kay, R. W. and Kay, S. M.: 1981, 'The Nature of the Lower Continental Crust: Inferences from Geophysics, Surface Geology, and Crustal Xenoliths', *Reviews of Geophysics and Space Physics* **19(2)**, 271–297.

Lienert, B. R.: 1979, 'Crustal Electrical Conductivities Along the Eastern Flank of the Sierra Nevadas', *Geophysics* **44(11)**, 1830–1845.

Lienert, B. R. and Bennett, D. J.: 1977, 'High Electrical Conductivities in the Lower Crust of the Northwestern Basin and Range: An Application of Inverse Theory to a Controlled-Source Deep-Magnetic-Sounding Experiment', in J. G. Heacock (ed.), *The Earth's Crust: AGU Mono.* **20**, 531–552.

Madden, T. R.: 1971, 'The Resolving Power of Geoelectric Measurements for Delineating Resistive Zones in the Crust', in J. G. Heacock (ed.), *The Structure and Physical Properties of the Earth's Crust: AGU Mono.* **14**, 95–105.

Meidav, T. and Furgerson, R.: 1972, 'Resistivity Studies of the Imperial Valley Geothermal Area, California', *Geothermics* **65**, 47–62.

Nash, J. T., Granger, H. C., and Adams, S. S.: 1981, 'Geology and Concepts of Genesis of Important Types of Uranium Deposits', in B. J. Skinner (ed.), *Economic Geology 75th Anniversary Volume*, pp. 63–116.

Olhoeft, G. R.: 1979, 'Electrical conductivity from 200 °C to 1000 °C for 264 Rocks and Minerals from Data in Parkhomenko and Bondarenko, 1972', *U.S.G.S. Open File Report* 79–846, 23 p.

Olhoeft, G. R.: 1981, 'Electrical Properties of Granite with Implications for the Lower Crust', *J. Geophys. Res.* **86(B2)**, 931–936.

Parkhomenko, E. I.: 1982, 'Electrical Resistivity of Minerals and Rocks at high Temperature and Pressure', *Reviews of Geophysics and Space Physics* **20(2)**, 193–218.

Petrick, W. R., Pelton, W. H., and Ward, S. H.: 1977, 'Ridge Regression Inversion Applied to Crustal Resistivity Sounding Data from South Africa', *Geophysics* **42**, 995–1005.
Petrick, W. R., Sill, W. R., and Ward, S. H.: 1981, 'Three-Dimensional Resistivity Inversion Using Alpha Centers', *Geophysics* **46**, 1148–1162.
Pirson, S. J.: 1963, *Handbook of Well Log Analysis*, Prentice-Hall, Inc. New Jersey, 326 p.
Roy, A. and Apparao, A.: 1971, 'Depth of Investigation in Direct Current Resistivity Prospecting', *Geophysics* **36**, 943–959.
SanFilipo, W. A. and Hohmann, G. W.: 1982, *Computer Simulation of Low frequency Electromagnetic Data Acquisition*, Earth Science Lab., Univ. of Utah Research Institute, report 62, 33 p.
Shankland, T. J. and Duba, A. G.: 1982, 'Carbon-Enhanced Electrical Conductivity in Rocks', *EOS Transactions*, **63(18)**, 438.
Shankland, T. J., and Ander, M. E.: 1982, 'Electrical Conductivity, Temperatures, and Fluids in the Lower Crust', Unpublished manuscript Geophysics Group MSJ980, Los Alamos National Laboratory.
Sternberg, B. K.: 1979, 'Electrical Resistivity Structure of the Crust in the Southern Extensions of the Canadian Shield-Layered Earth Models', *Geophys. Res.* **84**, 212–228.
Stesky, R. M. and Brace, W. F.: 1973, 'Electrical Conductivity of Serpentinized Rocks to 6 kbars', *J. Geophys. Res.* **78**, 7614–7621.
Stewart, J. H.: 1980, *Geology of Nevada*, Nevada Bureau of Mines and Geology Special Pub. 4, 136 p.
Tripp, A. C., Ward, S. H., Sill, W. R., Swift, C. M., Jr., and Petrick, W. R.: 1978, 'Electromagnetic and Schlumberger Resistivity Sounding in the Roosevelt Hot Springs KGRA', *Geophysics* **43(7)**, 1450–1469.
Van Zijl, J. S. V.: 1977, 'Electrical Studies of the Deep Crust in Various Tectonic Provinces of Southern Africa', in J. G. Heacock (ed.), *The Earth's Crust: AGU Mono.* **20**, 470–500.
Wannamaker, P. E., Ward, S. H., and Hohmann, G. W.: 1982a, *Magnetotelluric Responses of Three-Dimensional Bodies in Layered Earths*, Earth Science Lab., DOE/ID/12079–87, 80 p.
Wannamaker, P. E., Ward, S. H., and Hohmann, G. W., and Sill, W. R.: 1982b, *Deep Resistivity Structure in Southwestern Utah and its Geothermal Significance*, Earth Science Lab., DOE/ID/12079–89, 89 p.
Ward, S. H. and Sill, W. R.: 1976, *Dipole-Dipole Resistivity Surveys, Roosevelt Hot Springs KGRA*, N.S.F. final report, N.S.F. Grant GI-43741, Dept. Geol. and Geophys., Univ. of Utah, 29 p.