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ELECTRICAL METHODS FOR THE STUDY OF REGIONAL CRUSTAL CONDUCTIVITY ANOMALIES

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Some of the electrical methods that may be used to map the electrical conductivity of the uppermost 20 km of the earth's crust are described. They include direct current galvanic methods and controlled source and natural field inductive methods operating in the frequency band .01—100 Hz. The galvanic and controlled source inductive methods are based on a fixed grounded long-wire transmitter.

The galvanic dipole mapping method is applied in the prediction of earthquakes and the evaluation of hydrothermal reservoirs. The magnetometric resistivity method, in which the magnetic field of the galvanic current flow is measured, may be used to locate deep structures in the presence of superficial topography or conductive overburden.

Using source currents of the order of 100 A and frequencies of 45 and 76 Hz, the effective conductivity in the vicinity of the Project Sanguine transmitter in Wisconsin may be determined.

Although large source currents are desirable, controlled source currents may be limited to 5 A, provided on-line digital cross-correlation techniques are used to improve the ratio of signal to noise by averaging over time.

Natural field magnetotelluric methods can operate efficiently in the frequency band. .01—.2 Hz and the audio band 7—100 Hz. Soundings at audio frequencies have revealed unexpected conductive regions in the Precambrian crust.

Introduction

This review is concerned with the electrical methods that may be used to detect lateral changes in the electrical conductivity of the upper 20 km of the earth's crust. The object of a survey using any of the described methods shall be to produce a map of electrical conductivity as a function of position and depth over an area perhaps 20 km square.

Two direct current methods, the dipole mapping method and the magnetometric resistivity method, are discussed initially. Their inclusion may be justified on two grounds. First, if a direct current method is sufficient for the mapping problem presented, it should take precedence over a controlled source inductive method as the data are much easier to interpret. Second, in some cases a direct current method may give different and complementary data from an inductive method. For example, a vertical electric field cannot be induced within a plane layered earth by a purely inductive source no matter what the shape or orientation of the source. Estimating the resistivity of a highly resistive horizontal slab in a conductive medium is difficult as no induced current flows in the slab, although the thickness of the slab may be determined. In contrast, the direct current flow from a point electrode certainly has a vertical component. The galvanic methods are sensitive to the transverse resistance (resistivity-thickness product) of the slab [20]. The resistivity and thickness may be determined separately by combining the results of the inductive and galvanic methods. A joint inversion scheme has recently been proposed and demonstrated by VOZOFF and JUPP [31].

The current electrodes in both selected direct current methods are located a large distance apart and form a current bipole. The magnetometric resistivity method differs from the dipole mapping method in that the short dipole used to measure the local electric field is replaced by a magnetometer and a component of the magnetic field due to the current flow in the ground is measured.

The flexibility of the dipole method is illustrated from two contrasting published applications. In the first, changes in resistivity of several orders of magnitude associated with a geothermal zone were mapped. In the second, subtle changes in apparent resistivity of a few per cent attributed to changes in porosity in an earthquake zone were presented.

The magnetometric resistivity method has an important advantage over the dipole mapping method. The electric field at the surface of the earth can be severely distorted by local, insignificant inhomogeneities in conductivity or local fluctuations in surface topography. The magnetic field, being an integral over a volume distribution of current, is not greatly influenced by a surface layer, provided most of the current is not trapped in the layer. This advantage is illustrated from a published example in which a contact at a depth of almost 1 km was mapped beneath a volcanic layer of rugged topography.

The electromagnetic methods fall into two categories: controlled source methods, where a time varying source field is generated by a current in a wire resting on the surface of the earth, and natural field methods, which rely on source fields generated by ionospheric and magnetospheric currents. Both forms of source field induce eddy currents in the earth and the mapping of these currents with magnetic and/or electric field detectors indirectly maps the conductivity anomalies.

Controlled source methods have principally been used by the prospector to locate base metal mineral deposits whose electrical conductivity may be vastly greater than that of the host rock which contains them. The methods generate anomalies which are successfully interpreted in terms of the strike, dip, depth extent and, perhaps, the conductivity of the deposit. The more modern multiple frequency methods [33; 34] are used in more complex terrains and they enable the interfering effects of conductive overburden, current channelling and induced polarization to be discriminated.

For crustal mapping on the scale proposed here, these methods have to be modified, the principal modification being the alteration of the operating frequency band to 0.01—100 Hz. The band 0.4—7 Hz is most important as in this band there is very little natural geomagnetic activity. The next step is to increase the sensitivity of the magnetic and electric field detectors and over the past few years there have been major advances in this direction [25; 2; 36]. If a continuous signal is transmitted, the separation between the transmitter and the receiver must be increased, else the low level secondary fields generated at depth may be buried in the large primary field. The alternative is to use an impulsive current in the transmitter or to switch off rapidly a constant current. The transient secondary field can then be measured even near the transmitter as the primary field is absent. Indeed, in a transient profiling method in common use in the Soviet Union, the same large wire loop is used both as a transmitter and as a magnetic field detector.

Controlled source experiments in this frequency band have been attempted by several groups. Some of the more successful of these have been conducted using very high current bipole transmitters. Included are determinations of the effective earth conductivity in the vicinity of the transmitters constructed as part of Project Sanguine [32]. In each of these experiments a single frequency was transmitted continuously and the three components of the magnetic field were recorded along radial profiles. Project Sanguine signals may be detected at distances of several thousand kilometers from the transmitter as they are channelled in the waveguide formed between the conductive earth and the conductive ionosphere. Some of the data from this work are presented in this review.

Very high current transmitters, while desirable, are not necessary for controlled source mapping on this scale. The alternative is the use of cross-correlation techniques, improving the ratio of signal to noise by averaging over time. The recent development of sophisticated micro-processors means that this technique can be used in the field to process data on line. One such system has been developed at the University of Toronto and some data collected with it are shown.

Many research groups are using natural field electromagnetic methods for crustal mapping. In the GDS method, natural magnetic fields are recorded using an array of magnetometers [11]. The method has been employed successfully to probe the upper mantle and to map surficial conductors whose integrated conductivity is very large, such as the oceans and the deeper conductive sedimentary basins. It has also detected anomalous conductors in the crust [12]. However, as the frequencies recorded are lower than 0.01 Hz, the interpreted parameters are often poorly determined. A crustal conductor is often electromagnetically thin, even at the highest frequencies used, implying that the thickness is uncertain. The depth to the top of the conductor is invariably much smaller than the lateral extent of the anomaly, implying that this depth is uncertain. The cause of the magnetic variation anomaly is sometimes attributed to current channelling, implying that the absolute resistivity of the conductor is uncertain, only an estimate of the resistivity contrast between the conductor and its environment being obtained.

The GDS method could be extended to higher frequencies than 0.01 Hz but an array of three component magnetometers or coils suitable for recording the lower geomagnetic activity is prohibitively expensive. Instead, some researchers prefer to construct one very good system, occupying with it a large number of stations, perhaps as many as twenty in one day, concentrating on the audiofrequencies above 7 Hz. The two components of the electric field are invariably measured at the same time as this can be done with little additional effort. Hence, non-simultaneous magnetotelluric array studies are common at audio frequencies and tend to take the place of GDS studies.

Magnetotelluric data can of course be obtained throughout the band 0.01—100 Hz, even in the band 0.4—7 Hz, where there is little geomagnetic activity [7; 35]. However, the collection of sufficient data for interpretation purposes is a laborious process. The rapid collection of data must be an important consideration in selecting an electrical method because a high density of recording sites is required for adequate lateral resolution.

Two examples from the literature have been chosen to illustrate the magnetotelluric method at audio frequencies. The first is a survey over the same geothermal area as that mentioned earlier, surveyed with the dipole mapping method. Apparent resistivities determined by the two methods may be compared. The second example is a survey to measure the thickness of the permafrost in the Arctic. This is an important crustal mapping problem on a regional scale. The frequencies used in this survey are much greater than 100 Hz because the permafrost is very resistive.

Mapping methods using a long wire grounded source

Introduction

The long, grounded source or "bipole" consists of two current electrodes separated by up to 100 km and connected to a current generator or transmitter with two straight insulated wires. A current field is set up in the ground by the source. In part, this is a galvanic, return current flow between the electrodes to which may be added a transient, induced current flow depending on the temporal variation of the transmitted current. The source is not moved for the duration of the survey, but the electric or magnetic fields are mapped in detail around the source. Variations in the behaviour of these fields may be identified with vertical or lateral changes in the resistivity of the ground.

The dipole mapping method

The dipole mapping method was first described by ALPIN [1] and it has recently been reviewed by KELLER et al. [18]. It is a direct current method even though alternating or commutated currents are often used. Using two closely spaced potential electrodes, a dipole, the magnitude of the horizontal electric field, or a component of it, is measured at a field point remote from the bipole source. An apparent resistivity may be determined at the field point by dividing the measured electric field by the source current and multiplying the result by a factor which depends on the location of the field point with respect to the source and the electric field component measured. For a uniform Earth, the apparent resistivity is constant and has a value equal to the resistivity of the ground.

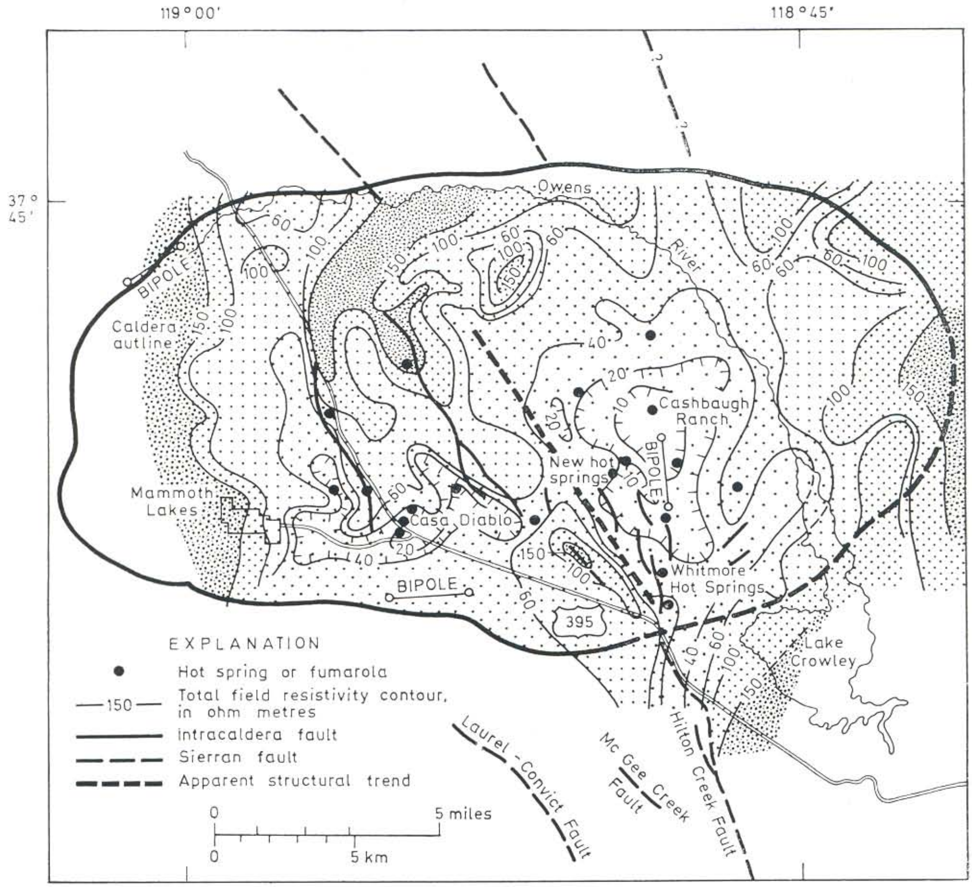

Fig. 1. A composite total field resistivity map for the Long Valley caldera, compiled from data obtained using the three bipoles shown (after [26])

The method has many applications in geothermal exploration, mining exploration and geotechnical engineering. RISK et al. [24] located the lateral boundaries of a conductive region associated with a geothermal system in New Zealand. STANLEY, JACKSON and ZOHDY [26] have studied the Long Valley Geothermal Area, California to evaluate its hydrothermal potential. Whole rock resistivities in a hydrothermal system may be decreased by several orders of magnitude because of a number of factors. Ionic concentration and

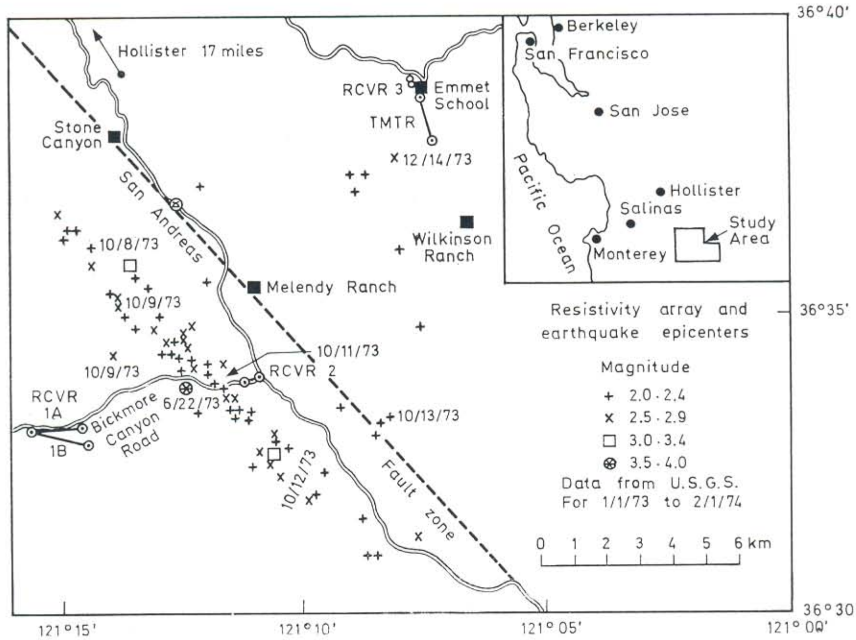

Fig. 2. The locations of the current transmitter bipole (TMTR) and the three receiver dipoles (RCVR) used for monitoring deep resistivity changes across the San Andreas Fault (after [22])

ionic mobility in the rock fluid increases with increasing temperature. The porosity and permeability of the rock is increased when the rock fluid is active.

A composite apparent resistivity map for the Long Valley Area is shown in Fig. 1 based on measurements by STANLEY, JACKSON and ZOHDY from the three bipoles shown. Two significant zones of low resistivity, Cashbaugh Ranch and Casa Diablo, outlined on the map are attributed to conductive layers of resistivity 1 to 10 ohm.m at depths of the order of 200 m. The zones follow fracture systems, especially those related to regional faulting.

MAZELLA and MORRISON [22] used the technique to record electrical resistivity variations associated with earthquakes on the San Andreas Fault, California. The locations of their transmitting bipole and receiving dipoles are shown in Fig. 2. The variation of the resistivity at each of the dipoles is plotted

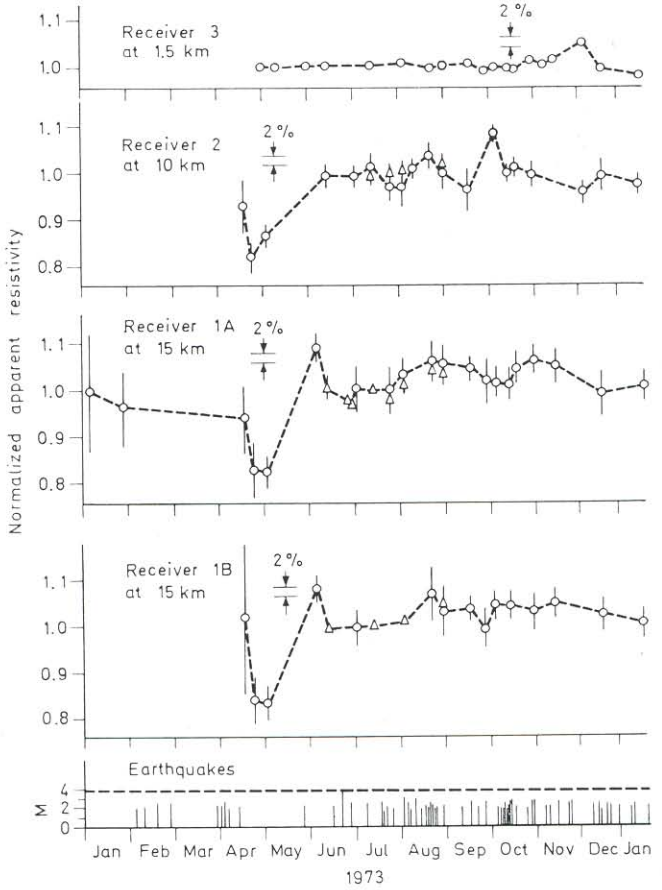

Fig. 3. The normalized apparent resistivity as a function of time at each of the three receiver dipoles. The error bars are plus and minus two standard deviations (after [22])

in Fig. 3. Prior to a magnitude 3.9 earthquake on the 22nd June, 1973, the resistivity rises then falls quickly. The changes could be accounted for by an 80% change in the intrinsic resistivity in a region 4 km wide and 2 to 6 km deep in the fault zone. They could reflect stress induced porosity changes in the zone and are consistent with a dilatational theory of earthquake mechanisms.

Based on their experience of a large number of dipole surveys, KELLER et al. comment that the method is more sensitive to lateral changes in resistivity beneath the area being surveyed rather than to vertical changes. This is particularly evident in regions bounded at depth by a resistive layer. They interpret many of their data in terms of an integrated resistivity, assuming that the earth is represented by a conducting sheet.

In most of the surveys reported in the literature, the length of the transmitter bipole was at least 1 km; the transmitted current amplitude ranged from 5 to 200 A and the period of the current was typically 10 sec, which is low enough to prevent the effects of electromagnetic induction reducing the penetration of the current field. Useful estimates of the electric field can be made with 100 m dipoles to a range of the order of 5 km. An improvment in signal recovery is possible by transmitting the current waveform through a telemetric link and cross correlating this signal with the measured field at the recording site.

The magnetometric resistivity (MMR) method

The MMR method differs from the dipole method in that the dipole is replaced by a sensitive coil or a magnetometer and a component of the horizontal magnetic field due to the current field in the ground is recorded. This method of exploration and mapping was patented by JAKOSKY as early as 1933 [15], but it met with little success at that time. Part of the problem was undoubtedly with the instrumentation and only recently have robust instruments, capable of recording magnetic fields of the order of a few hundred milligamma, been developed and field data of good quality obtained. However, the main reason for the failure of the initial experiments was due to a lack of understanding of the basic principles involved. The method cannot be used for depth sounding as an Earth composed of uniform, horizontal layers generates no MMR anomaly. The current is altered by the layering from that of a uniform earth but the horizontal magnetic field produced is exactly the same as for a uniform Earth.

Despite the early problems with the method, S. S. STEFANESCU and his students did calculate the theoretical responses due to simple geometric structures, such as the contact and the outcropping dike, starting in the late fifties. STEFANESCU [27] also formulated an important algorithm which enables the anomalous vertical magnetic field to be calculated using Ampere's circuital theorem.

The first successful field surveys using the method were published by SEIGEL [25] and by EDWARDS [9], who developed the methodology and normalizing procedures as presently used. The results of a second field survey by EDWARDS and HOWELL [10] proved conclusively the validity of the method as a mapping tool. The survey was conducted on a plateau in the western United States, where the topography is characterised by steep hills, bold ridges, gullies and narrow canyons. A steep faulted contact between basement rocks of differing resistivity is exposed on one flank of the plateau, beneath over 500 m of tertiary volcanics and sediments as shown in Fig. 4.

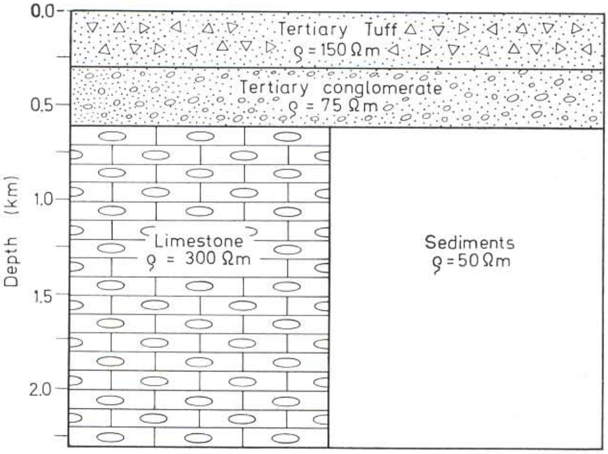

Fig. 4. A geological section through the faulted contact (after [10])

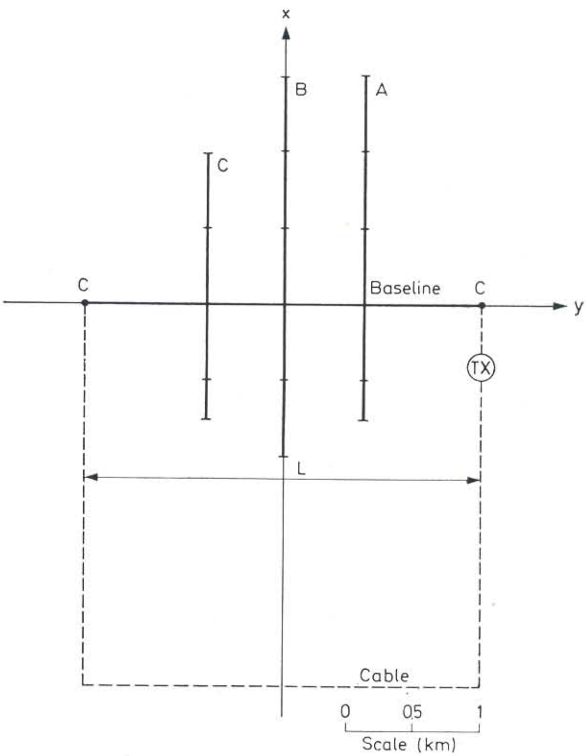

Fig. 5. The geometry of the MMR array. The current electrodes are at points *C* on the baseline. The measurements were made along lines *A*, *B* and *C* (after [10])

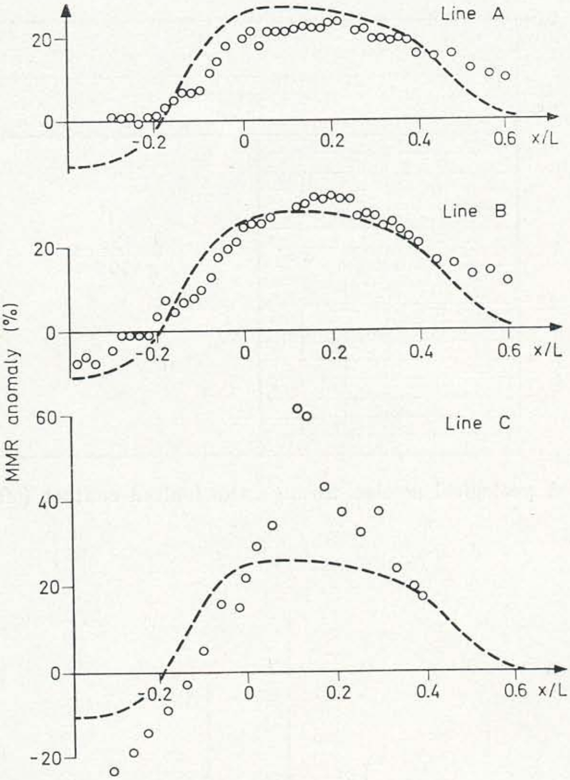

Fig. 6. The measured anomalous horizontal x-component of the magnetic field along lines A, B and C compared with the theoretical anomalies due to a dike model. The vertical boundaries of the dike are the planes $x = -0.16L$ and $x = 0.45L$. The conductivity contrasts across these boundaries are respectively 1 : 6 and 10 : 1 (after [10])

The object of the test was to determine if the basement contact could be mapped by the method, working entirely on top of the plateau. Two current electrodes were placed immediately above the expected strike of the structure, which is the baseline and y-axis of Fig. 5. A current of 2.5 A at a frequency of 3 Hz was transmitted into the ground. Measurements of the x-component of the magnetic field were made along lines A, B and C. The data were reduced by first subtracting the normal horizontal field expected for a uniform earth at each measurement point and then expressing the result as a percentage of the normal field at the center of the array — the point 0 (this was only 330 milligamma !).

The reduced data are plotted in Fig. 6 and are compared with theoretical curves for a two-dimensional outcropping vertical dike, or double contacts. The contacts are respectively at $x = -0.16\,L$ and $x = +0.45\,L$, where L is the separation of the electrodes. It seems the method has indeed located the contact at depth and also delineated a second contact.

The model is a limited one in that it neglects the presence of the overburden. Further modelling pinpointed an anomalous region of unusually high conductivity in the vicinity of $x = 0.15\,L$ on line C.

Bipole electromagnetic sounding and crustal mapping

The direct current methods described earlier are most useful for the mapping of shallow regional conductivity anomalies. However, the presence of the resistive second layer in the Earth's upper crust limits their usefulness for mapping any conductive structures at depth. The resistive layer screens the deeper structures and very large current electrode separations must be used if the direct current is to penetrate it. A time varying current in the bipoles overcomes this problem by producing a vertical magnetic field which induces eddy currents in the conductive layer. The channelling of these currents by a conductivity anomaly causes a perturbation of the observed electric and magnetic fields and enables the anomaly to be mapped.

Although extensive small scale surveys, principally in sedimentary basins, have been conducted successfully by scientists in the Soviet Union [29], very few large scale crustal surveys have been reported in the literature. The difficulties are again mostly of a technological nature: the transmission of large currents, typically 100 A, and the detection of small magnetic fields, typically 1 milligamma, or small electric fields, typically 10 μV/km, in a broad frequency band from 0.01 to 100 Hz. A further difficulty is obtaining a suitable long wire transmitter inexpensively!

Over the past five years, a number of crustal sounding experiments have been conducted using the U.S. Navy's Test Facilities in Wisconsin. Here there are two orthogonal bipoles, NS and EW, each 22.5 km long, and each capable of transmitting currents in excess of 300 A. The effective earth resistivity in the immediate vicinity of the array has been measured by a number of electrical methods and the results are summarized by BANNISTER [4]. The radial H_ϱ, and the tangential, H_φ, components of the horizontal magnetic field were measured along radial profiles from each bipole at frequencies of 45 and 76 Hz. These frequencies are much too high for deep crustal sounding. Nevertheless, the data obtained are worth looking at because they are of very high quality, even out to distances of 80 km, and they demonstrate just what can be done. The profile of H_ϱ, broadside on to the NS antenna at a frequency of 45 Hz is shown in Fig. 7, from an earlier paper by BANNISTER and WILLIAMS [5]. The three solid curves were determined from the theoretical model of a uniform earth of resistivities 4000, 5900 and 8300 ohm. m respectively. The central value is clearly one of the better estimates of the average resistivity of the first 10 km of the crust. The bedrock of Wisconsin is primarily a Precambrian metamorphic–igneous complex consisting mainly of granites. There may also be more conductive rocks such as schists in addition to local mineralization. Magnetotelluric data presented later will show that the crust does not have constant resistivity as a function of depth in this region.

In more geologically complex areas, the transmission of a single frequency is insufficient to determine the electrical structure. The measurements have

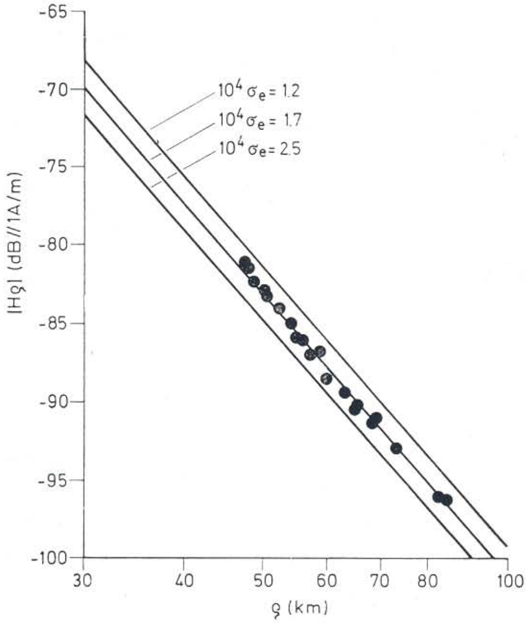

Fig. 7. A plot of the radial magnetic field, H_ϱ, against the radius, ϱ, at 45 Hz for a traverse broadside to the NS antenna in Wisconsin (after [5])

to be repeated perhaps 16 times or more at different frequencies. This is an inefficient, time-consuming procedure which almost necessitates some form of verbal communication between the transmitter and receiver locations. It is possible to simultaneously transmit a number of frequencies, replacing the analogue-tuned filters commonly used in a single frequency system with some form of digital filter to improve the ratio of signal to geomagnetic and instrumental noise. Another alternative is to transmit a repeated step in time and to record and stack the shape of the transient [17]. The amplitude spectrum of the step is inversely proportional to frequency, f. The magnetic field is often detected by a coil which has an inherent response proportional to f. Hence the stacked output is the impulse response of the earth.

The same repeated step recorded with many of the newer sensitive magnetometers would yield the Heaviside response of the earth as these instruments have a response independent of f. The Heaviside step response emphasizes the low frequencies or long times and it is preferable for crustal sounding because the response at longer times is due to the deeper structures and the secondary fields from these structures are weaker [30]. If an impulse response were required for crustal mapping, a sharp impulse of charge might seem to be an appropriate form for the transmitted current. But this results in the peak current transmitted being very much larger than the average

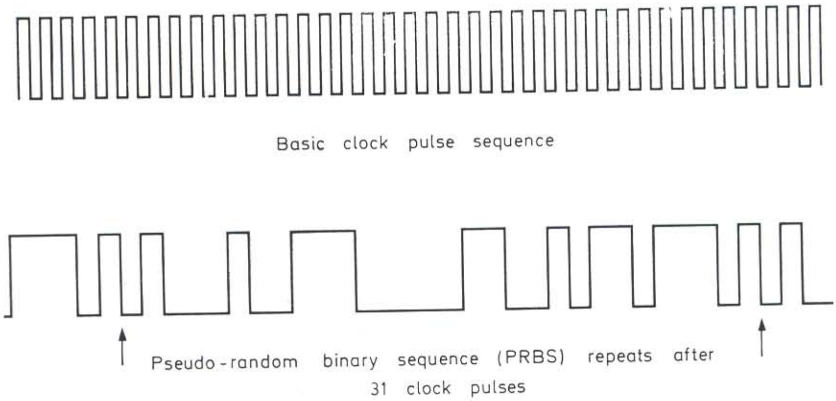

Fig. 8. An example of a pseudo-random signal generated using a shift register with programmed feedback from the clock signal shown above it. The signal repeats after 31 clock pulses (i.e. $n = 5$)

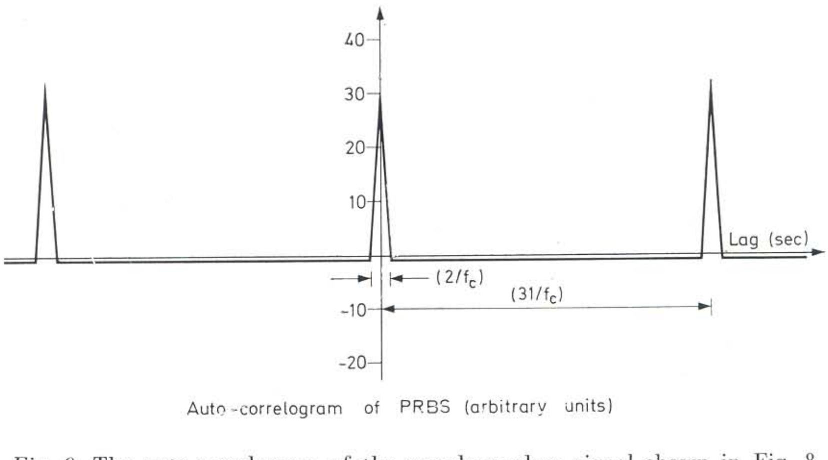

Fig. 9. The auto-correlogram of the pseudo-random signal shown in Fig. 8

current and it leads to an overdesign of the bipole system which clearly has to handle the peak current. A possible compromise is a signal of constant power whose autocorrelation is an impluse in lag domain. This is the basis of the VIBROSEIS technique for seismic profiling.

An example of such a signal is shown in Fig. 8. The signal is generated easily and accurately by pulsing a shift register which has a specified, programmed feedback from certain of its outputs into its input [6]. The resulting sequence produced at any output of the register is a pseudo random binary sequence (PRBS) and it is used to operate a double pole, double throw switch between a direct current generator and the bipole. The transmitted current

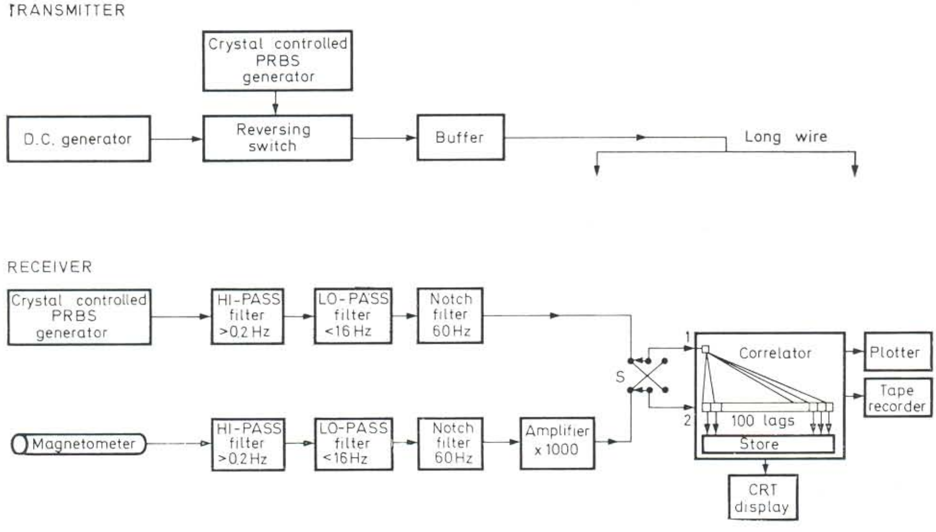

Fig. 10. A block diagram of the controlled source electromagnetic system. The correlator samples channels 1 and 2 every 10 mS. The current value in channel 1 is multiplied in turn by each of the 100 previous values of channel 2. The products are added into the store. The interchange switch S enables the correlograms for positive and negative lags to be obtained sequentially. The PRBS parameters n and f_c are 7 and 16 Hz

has a flat amplitude spectrum from half the clock frequency, f_c, to the repetition frequency $f_c/(2^n-1)$, both f_c and n being selectable variables. The autocorrelation of the current function displayed in Fig. 9 is a comb of triangles separated in time by $(2^n-1)/f_c$.

A block diagram of an experimental system which was developed at the University of Toronto and which uses two PRBS generators is shown in Fig. 10. The two generators are locked in phase at the beginning of the day. One remains at the transmitter controlling the current, the other is carried with a recording magnetometer. At the recording site, the measured magnetic field is cross correlated digitally in real time with the output from the local generator. The cross correlogram builds up in time and the site is occupied until an adequate signal-to-noise ratio is obtained. We have successfully recorded magnetic fields whose peak amplitudes are of the order of a few microgamma.

In one experiment, the horizontal magnetic field, H_ϱ, was recorded along

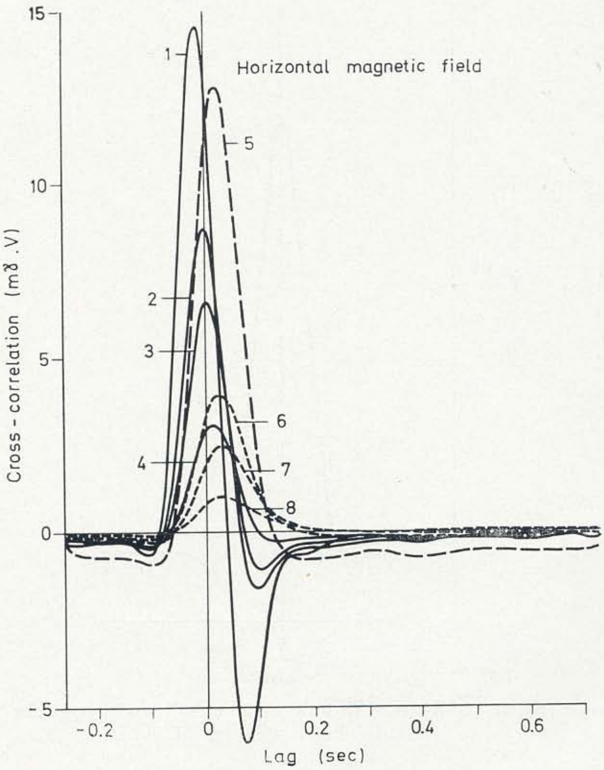

Fig. 11. The radial magnetic field, H_ϱ, as a function of "time" for the eight stations in southern Ontario. The vertical scale for stations 5 through 8 should be divided by a factor of 10

a profile broadside on to a 17 km bipole located in Southern Ontario. The measured magnetic fields at stations 1 through 8, respectively $\simeq 1$, $\simeq 3$, 5.7, 10.8, 18.0, 23.7, 32.0 and 40.1 km from the bipole, are shown in Fig. 11. Consider them as magnetic fields produced as a function of *time* due to a current in the bipole, Fig. 12, also varying as a function of time. The current function differs from a perfect triangle because the correlogram is limited to a finite frequency window, from 0.2 to 16 Hz. The data resemble standard interpretation curves for a thin conductive (Paleozoic) layer, of integrated resistivity 0.12 ohms, overlying a relatively insulating (Precambrian) crust.

I should add that these data were obtained using a 500 W generator and an average transmitter current of only 4.5 A. The effectiveness of using cross-correlation techniques to improve the ratio of signal to noise cannot be overemphasized.

The current flow set up in the ground by a bipole transmitter is in part a galvanic return current flow and in part a divergence free induced current

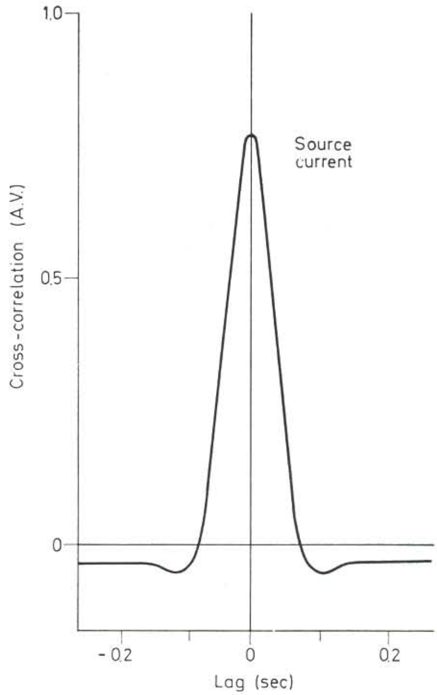

Fig. 12. The current in the bipole as a function of "time" corresponding with the magnetic fields shown in Fig. 11

flow. The reduction of data can be quite a complicated procedure. There are two exceptions. First, if the frequency is low enough, the induced currents may be neglected and MMR interpretation curves may be used. Second, over a uniform layered earth, the horizontal magnetic fields due to the return current in the ground are independent of frequency. They may be computed trivially and subtracted from the observed fields and the resulting secondary induced fields may be matched against type curves for induction in a layered earth by a (non-physical) finite, ungrounded current carrying wire. The fact that the galvanic and inductive fields separate in this way reveals that the bipole source really behaves like a purely external source over a layered earth, as far as the anomalous magnetic fields it generates are concerned. The secondary magnetic fields are only due to currents which flow in closed horizontal loops [23].

The natural field audio-magnetotelluric (AMT) method

The magnetotelluric method has been in common use for determining the electrical structure of the Earth since about 1950. The electromagnetic impedance — the ratio of a horizontal electric field component (Ex) to the orthogonal corresponding horizontal magnetic field component (Hy) — is measured at a number of frequencies to yield the apparent resistivity of the ground as a function of frequency, resulting in a form of depth sounding.

At frequencies below 0.4 Hz, the sources of the fields are currents in the magnetosphere and the ionosphere whereas at frequencies above about 7 Hz, the main source of energy is thunderstorm activity which tends to propagate around the world trapped in the waveguide formed between the ground surface and the ionosphere. Certain frequencies are preferentially propagated by the waveguide, notably the Schumann resonances at 8, 14, 20 and 25 Hz whereas others, at about 2 KHz, are strongly absorbed by it [21]

A small number of papers reporting low frequency, 0.0001 to 0.4 Hz, magnetotelluric depth sounding appears in the literature every year. Each paper may describe a survey comprising no more than a dozen temporary observatories which have often taken a whole field season to occupy. The limited number of observatories is a direct result of two factors. Firstly, the average research group cannot afford more than two or three complete magnetotelluric systems making consecutive installation of the observatories necessary. Secondly, each site must be occupied for at least two weeks for adequate data at all frequencies to be obtained.

For crustal mapping, as opposed to depth sounding, the same equipment could be used more efficiently by limiting the bandwidth of the recorded signals to 0.01 to 0.2 Hz. This would enable a new observatory to be occupied about every other day. Over the season, an increase in lateral resolution is obtained at the expense of frequency information. I know of no group who have considered it worth while to adopt this kind of approach at these frequencies — something of a surprise, since in exploration, the single frequency electromagnetic system is still the most common type. No doubt there is a compromise.

At audio frequencies above 7 Hz, the collection of data is more rapid. Strangway et al. [28] describe an AMT system in which analogue filters are used to select a narrow frequency band. The amplitudes of the electric and magnetic field components are determined by averaging over at most 3 mins. The ratio of the fields and the apparent resistivity is determined immediately. It takes no more than one hour at each station to cover the whole audio spectrum from 10 Hz to 10 KHz

Similar systems have been developed in France [16] and in the United States [14]. A block diagram of the latter system is shown in Fig. 13. Notice that phase information may be preserved by means of a phase-locked loop and synchronous detectors. Nevertheless, in comparison with the sophisticated data processing techniques that may be applied to recorded data, the on line processing of AMT data is presently quite crude. In the vicinity of lateral conductivity contrasts, apparent resistivities determined from average ampli-

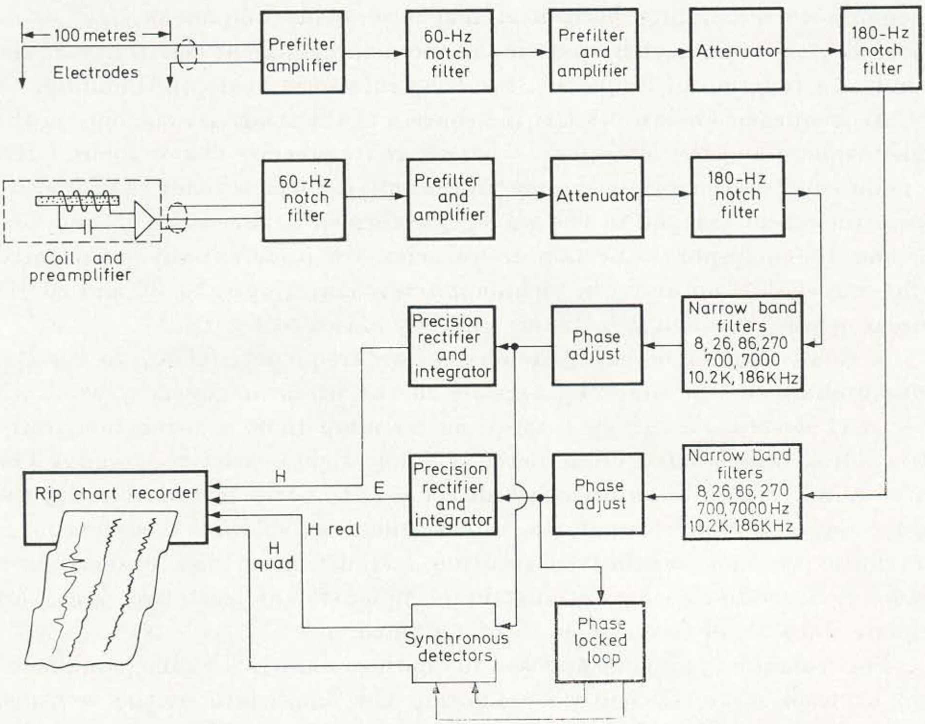

Fig. 13. A block diagram of an audiomagnetotelluric system (after [14])

tude and phase difference information only may be systematically in error unless the direction of regional current flow in the earth is known. The latter is highly variable depending, at audio frequencies, on the location of thunderstorm activity. These errors may be overcome to first order by measuring both components of the horizontal electric and magnetic fields and computing the tensor relationship between them. A digital processor capable of doing this on line at the field station is being developed.

In defence of the AMT systems presently used, it should be added that the errors in the apparent resistivities are unlikely to be greater than one order of magnitude whereas the variation in apparent resistivity in many crustal mapping problems can span many orders of magnitude.

The financial limitation of using a large array of instruments operating simultaneously and the consequent trend towards an array of stations occupied consecutively poses an important question. Can data collected by the two methods ever be considered equivalent? The simultaneous array certainly provides information about the geometry of the source field which is unobtainable with a single station. Into an average apparent resistivity at a single station (or a transfer function in the case of GDS sounding) go source fields of all descriptions. Yet BAILEY and EDWARDS [3] have shown that transfer function GDS data can be processed to look like data collected with an array subject to the condition that the source fields are relatively uniform or of large spatial wave length. Hypothetical events of given polarizations can even be simulated.

HOOVER et al. developed the AMT technique for use as a reconnaissance geothermal exploration method. Their philosophy was that a survey using a rapid, inexpensive technique such as AMT would be followed up by a more definitive electrical surveying programme in promising areas. The Long Valley Geothermal Area was chosen as a test area so that the AMT results could be compared with apparent resistivities determined by STANLEY et al. using the dipole mapping method. Their apparent resistivity maps determined at 26 Hz for two polarizations of the electric field are shown in Fig. 14a and 14b. Clearly the correlation between these maps and the map of STANLEY et al. shown in Fig. 1 is very good. It is remarkable that the whole AMT survey took only 2 man-weeks of work.

The AMT method has been used extensively in North America but the results of many of the surveys are either awaiting publication or are in contract reports. Although primarily a crustal mapping technique, the method has revealed a consistent, unexpected layer of high conductivity in the upper crust. D. W. STRANGWAY (personal communication) used the method to map a region of Wisconsin, around the U.S. Navy's Test Facility. As part of the data analysis, the apparent resistivities obtained as a function of frequency at about 20 sites were averaged together. The resulting curves are plotted in Fig. 15 for two polarizations of the electric field.

The curves are interpreted in terms of a four-layer structure. The uppermost, surface layer is strongly anisotropic and is only a few meters thick. The second thin conductive layer has a conductivity-thickness product of about 0.35 mhos. The third resistive layer is about 4.5 km thick and has a resistivity of more than 6600 ohm m. The lower fourth layer has a resistivity much less than 1000 ohm m. but its thickness cannot be determined. It is this fourth layer that is unexpectedly conductive. It clearly lies beneath a layer of very dry crystalline rock.

The apparent resistivity values at 45 Hz do not differ greatly from the effective resistivities measured by BANNISTER and WILLIAMS, who also report that the NS resistivity is a factor of 1.65 larger than the EW resistivity at this frequency.

DOWLING [8] conducted a series of magnetotelluric experiments over a large area of Wisconsin at frequencies from 10 to 0.001 Hz. The apparent resistivity curves he obtained are presented in Fig. 16. Although the frequencies he used are too low to discriminate the layers in the uppermost crust, many of the curves show a low resistivity at the highest frequencies which is again

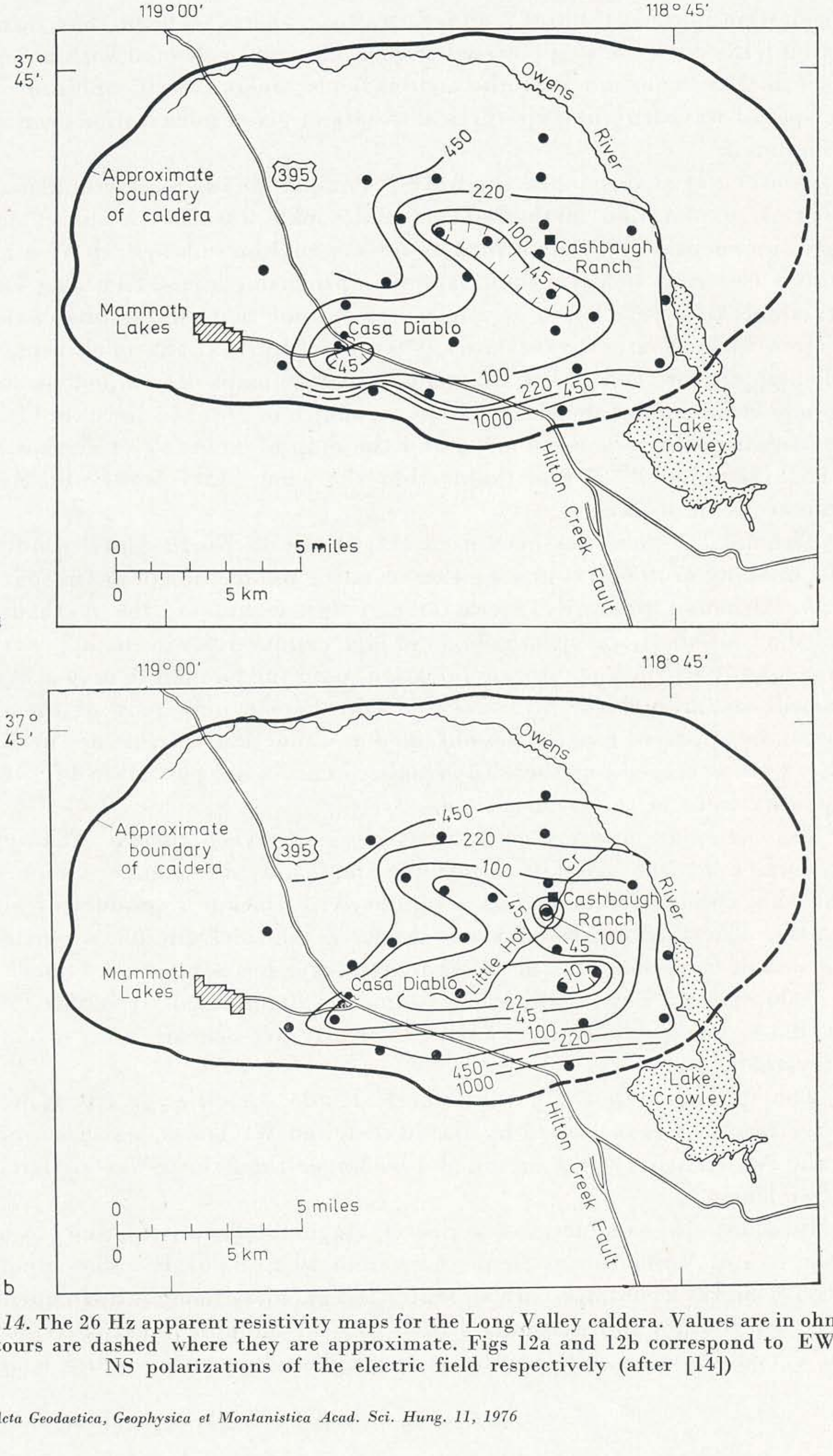

Fig. 14. The 26 Hz apparent resistivity maps for the Long Valley caldera. Values are in ohm. m. Contours are dashed where they are approximate. Figs 12a and 12b correspond to EW and NS polarizations of the electric field respectively (after [14])

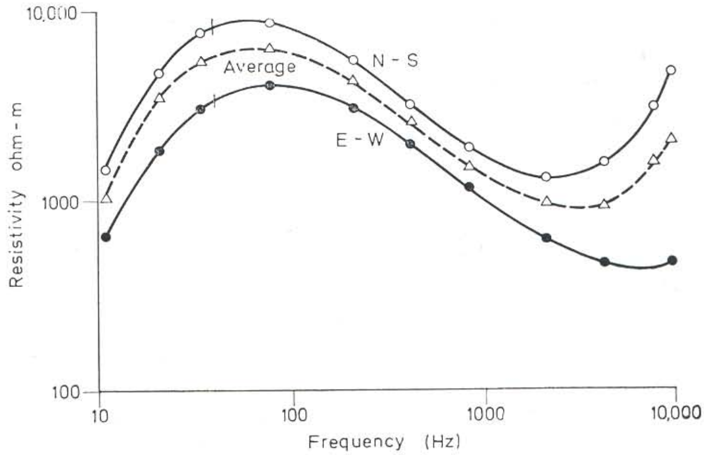

Fig. 15. The apparent resistivity against frequency for two polarizations of the electric field obtained by the AMT method in Wisconsin (after STRANGWAY, personal communication)

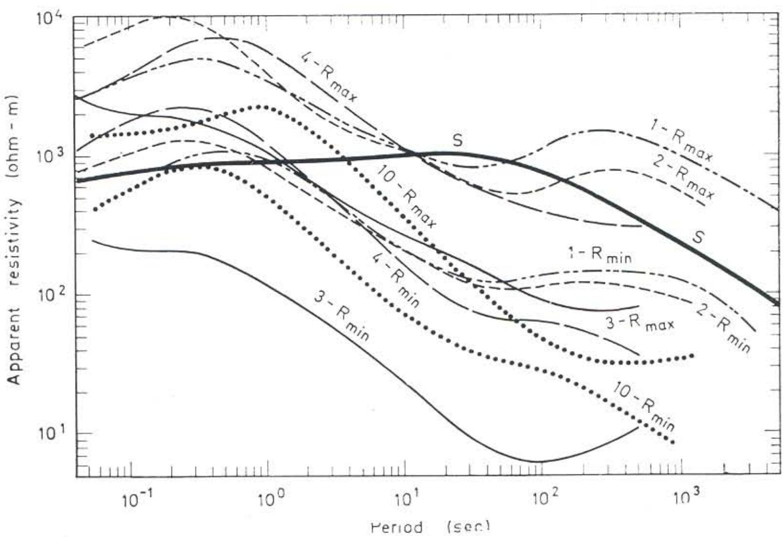

Fig. 16. The apparent resistivity (maximum and minimum orientations) for a set of stations in Wisconsin. The curve *S* is for a "postulated, resistive Precambrian crust" (after [8])

consistent with the AMT data. The crustal geoelectric structures derived by DOWLING are shown in Fig. 17. Many of them include a conductive layer in the lower crust.

A geoelectric model for the Precambrian crust in Wisconsin may be obtained by combining DOWLING's data with the AMT data. Clearly, at least three conductive layers are present; a surface layer, a layer in middle crust and a layer in the lower crust. Further experiments by STRANGWAY in Michigan and North Ontario support this model.

The use of AMT to determine the thickness of permafrost in the Arctic is an important applied crustal mapping problem which has applications ranging from the construction of pipe lines to the interpretation of reflection seismic data. The definition of permafrost, or perennially frozen ground, is

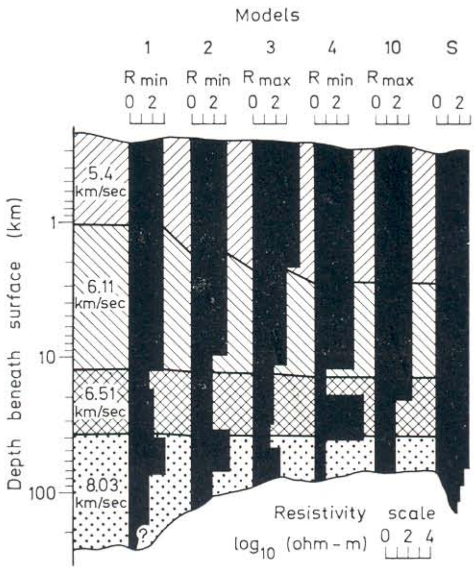

Fig. 17. The preferred resistivity models for Wisconsin derived from the MT data and the resistivity model for a "postulated, resistive Precambrian crust" (after [8])

based on temperature. When the temperature of a section of ground is below 0 °C for longer than one year, permafrost is said to exist.

The variation of resistivity with temperature for several saturated soil types and one rock type is shown in Fig. 18, from HOEKSTRA et at. [13]. The resistivity does not increase abruptly at the freezing point, but gradually with decreasing temperature as the ionic mobility of the charge carriers in unfrozen solution decreases.

Generally, the permafrost layer encountered in the Arctic winter is much more resistive than the unfrozen ground beneath it and the problem of determining permafrost thickness becomes one of measuring the depth to this electrical interface. KOZIAR and STRANGWAY [19] used the AMT method to map a region of the MacKenzie Delta, NWT, Canada. Frequencies in the broad band from 10 Hz to 10 KHz were measured. The apparent resistivities calculated along one profile, are shown in Fig. 19c as a 'pseudo-section' which

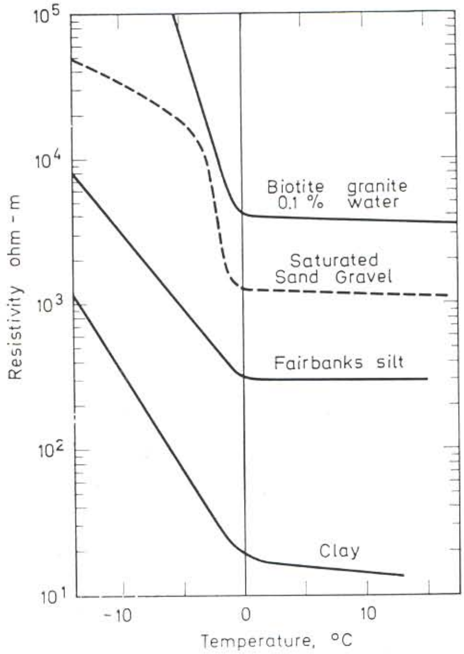

Fig. 18. The resistivities of several saturated soil types and one rock type as a function of temperature (after [13])

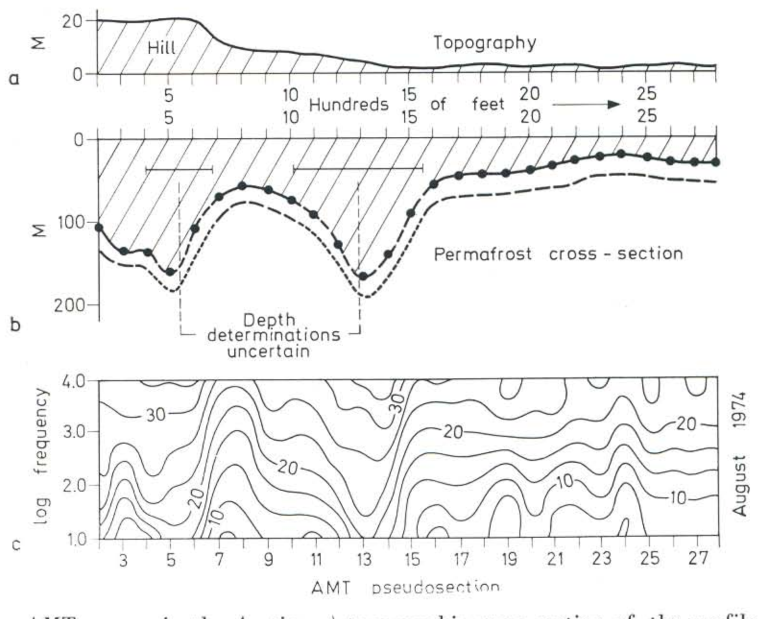

Fig. 19. The AMT survey in the Arctic; *a)* topographic cross section of the profile. The hill is a vestige of an areal ice sheet; *b)* permafrost thickness determined at 10 KHz using a two-layer model; *c)* pseudosection representation of the data. The contours are at intervals of 10 log ϱ_a (after [19])

enables both the lateral variation and the frequency dependence of the apparent resistivity to be described simultaneously.

The pseudo-section is a favourite method of displaying AMT data. Over a layered earth, the log frequency axis of the pseudo-section can be converted directly to a log depth axis. In more complex situations, pseudo-sections derived from theoretical models are compared with those displaying field data. The effects of current channelling can often be recognized on pseudo-sections as short wave-length features present at all frequencies.

The resistivity data at every station were compared with the theoretical response of a highly resistive layer over a half space of finite resistivity (50 ohm m). In this case, it is possible to derive a very simple asymptotic function which relates the depth of the layer, h, to the frequency, f, the measured apparent resistivity, ϱ_a, and the skin-depth d_2 in the half space

$$\varrho_a = \varrho_2 \left(1 + \frac{2h}{d_2} + \frac{2h^2}{d_2^2}\right)$$

or

$$h = 356\,(\varrho_a/f)^{1/2} \text{ at high frequencies.}$$

The interpreted depths of the permafrost along the profile using these two formulae at 10 KHz are shown in Fig. 19b as dotted and dashed lines respectively.

Conclusions

In this review, some of the electromagnetic methods that may be used to map crustal conductivity anomalies have been described. The ideal inductive method should operate at frequencies in the band 0.01 to 100 Hz. In the field, all the raw data should not be recorded, but the measured electric or magnetic fields should be processed to obtain an appropriate response function for the ground within this frequency band. A crustal survey using the method where 40 or 50 stations are occupied should be completed within one field season.

The inductive methods fall into two categories: controlled source and natural field. Only the controlled source methods using a transmitting bipole were discussed. They can operate throughout the required range of frequencies. Although source currents in excess of 100 A are desirable, currents may be limited to 5 A provided on line cross-correlation techniques are used to improve the ratio of signal to noise.

The natural field magnetotelluric methods can operate efficiently in two frequency bands, 0.01 to 0.2 Hz and 7 to 100 Hz respectively. For crustal mapping, it is necessary to limit the band-width recorded at low frequencies to obtain a higher density of stations when only one or two sets of equipment are available.

Two controlled-source direct-current methods were also described. The dipole method can give different and complementary data from any inductive method. The MMR method is superior to the dipole method in rejecting noise generated by local, insignificant conductivity anomalies or topographic features, but it cannot be used for crustal sounding.

Some of the many applications of crustal mapping were presented. They include the prediction of earthquakes, the evaluation of hydrothermal reservoirs, the determination of permafrost thickness and the mapping of regional, structural geology. The AMT soundings have revealed unexpected conductive regions in the crust. The mapping of these layers is a brand new field which might provide fundamental information about the nature of the continental crust.

Acknowledgements

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ЭЛЕКТРИЧЕСКИЕ МЕТОДЫ ДЛЯ ИЗУЧЕНИЯ АНОМАЛИЙ РЕГИОНАЛЬНОЙ ПРОВОДИМОСТИ КОРЫ

Р. Н. ЭДУАРДС

РЕЗЮМЕ

В статье рассматриваются некоторые электрические методы, которые могут быть использованы для картографирования проводимости верхних 20 км-ов земной коры. Сюда принадлежат гальванические методы постоянного тока и индуктивные методы, применяющие искусственные и естественные поля и работающие на частотах 0,10—100 Гц. Гальва-

нические методы и индуктивные методы с искусственным полем основываются на зафиксированном и заземленном передатчике с длинныи кабелем.

Гальваническое диполное картографирование пригодно для прогноза землетрясений и исследования термальных резервнаров. Изменение сопротивления магнитометро, при котором измеряется магнитное поле гальванического тока, можно применять при выявлении глубинных структур в случае поверхностной топографии или проводящей кровли. Применяя источник тока в 100 А с частотами 45 и 76 Гц, в окрестности передатчика Project Sanguine можно было определить эффективную проводимость в Висконизне.

Хотя желательно иметь большой ток питания, все же целесообразно ограничивать его до 5 А, если применяют технику дигитальной поперечной корреляции для улучшения отношения сигнала к шуму образованием среднего во времени.

Мегнитутеллурические методы с естественным полем эффективно могут использоваться в диапазоне 0,01 – 0,2 Гц, а также в звуковом диапазоне 7 – 100 Гц. Зондирования, проведенные в звуковом диапазоне, открыли территории с неожиданной проводимостью в прекамбрийском слое.