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Electromagnetic Induction Studies

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INTRODUCTION

This report constitutes an attempt to review the major developments and identify important trends in the broad field of geophysical electromagnetic induction and related phenomena over the past four years. Following in the spirit of previous reports of this type [e.g., *Filloux*, 1979; *Hermance*, 1983*b*], the work of US researchers will be emphasized, although we will cover foreign research when appropriate. Many of the recent theoretical developments and the largest EM field program ever (EMSLAB) are the direct result of international cooperation, and strict adherence to the concept of national boundaries would result in an uninformative and incomplete review.

Due to the fact that readers of this paper have diverse interests ranging from theory through field to laboratory studies, we have attempted to treat a variety of topics in EM induction and electrical geophysics. We begin by reviewing the state-of-the-art in data collection, including new instrumentation. We continue by examining data analysis methods, with an emphasis on noise and bias reduction in the computation of the magnetotelluric and magnetic variations response functions. We then treat forward modelling developments, especially for two- and three-dimensional induction problems. Recent progress has been made in EM induction inverse problems, and we assess the impact of this on the field. An overview of field measurements in North America is given, including the recent EMSLAB experiment which was carried out in 1985—1986 in the northwest US, southwest Canada, and contiguous offshore regions. This is followed by a review of developments in oceanic applications of EM principles.

The reference list is believed to be complete through June 1986. In the interest of brevity, only refereed publications or works in press are included, and meeting abstracts or technical reports are generally not cited. Nevertheless, the number of references exceeds 600, attesting to the health of the discipline. Additional information on EM induction research may be found in the proceedings of the most recent semiannual Workshops on EM Induction held in Victoria, Canada, in 1982, Ile-Ife, Nigeria, in 1984, and Neuchâtel, Switzerland, in 1986.

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DATA COLLECTION AND INSTRUMENTATION

To a large extent, recent improvements in EM data have come about through more sophisticated time series analysis methods rather than from changes in instrumentation. Data analysis is covered in the next section. Progress in the quality of the sensors themselves has been more gradual.

It has been the general experience of US research groups that SQUID magnetometers do not achieve their laboratory potential in field situations, and this has led to a trend back to induction coils for wide-band MT work. Coils have become much lighter as amplifier technology has improved, and are easily constructed. However, developments in fluxgate design may soon produce a sensor that is comparable in sensitivity to coils in the dead band around 1 Hz and that has a far better long-period response. A fluxgate instrument would also be easier to deploy in the field. Theoretical analyses that elucidate some important design criteria for sensitive, low noise fluxgates are given by *Russell et al.* [1983] and *Narod and Russell* [1984]. Three companies are now manufacturing ring-core fluxgates similar to those flown in MAGSAT. One obstacle to making these instruments much better than their predecessors is the limited availability of the best core materials. *Narod et al.* [1985] present an experimental study of amorphous alloys which shows that fluxgate noise depends on material properties that have not been considered before. This gives some hope that materials can be found that are easily manufactured and that will deliver very good performance.

As in most branches of the physical sciences, microcomputer technology is having an enormous impact on induction work. Long-period data are now almost always collected digitally, and solid-state data loggers and high density redundant tape recording is offering much higher reliability and larger capacity than before. Even the venerable Gough-Reitzel magnetometer is receiving new life through image processing which allows easy digitization of the film records. Wide-band MT data are now routinely processed on site. It is possible to build and operate large arrays of sensors that can map the electric field in great detail and overcome the effects of local distortion. It is also feasible to make wide-band MT (and controlled source) equipment so portable that it can be taken virtually anywhere.

One of the most significant developments of the past four years has been the realization by US and Canadian workers that group field efforts are essential to the study of many relevant large-scale problems. The desire to upgrade field equipment for academic research is also supplying an impetus to work in groups. The equipment that is currently available to academic scientists lags far behind what is possible with current technology. New equipment will require a substantial capital investment which can be minimized by careful sharing of resources.

RESPONSE FUNCTION ESTIMATION

The estimation of EM response functions or impedances from data is of central importance to the natural source EM methods, and especially for MT. Increasingly more sophisticated ways to reduce the influence of noise on the response functions have evolved over the past four years. In addition, new ways to interpret the full response tensor are being developed.

Most of the methods for computing response functions in current use are based on least squares principles, and share the inherent advantages and limitations of that technique. An important requirement for the proper operation of least squares is that the residuals or errors from the fit be uncorrelated and of equal variance or power. Data that produce residuals which fail to meet this condition may be termed outliers, and least squares estimates are very sensitive to their presence. This type of outlier, as well as ordinary Gaussian noise, can induce serious bias and distortion into EM response functions. Outliers in EM data may be caused by a variety of instrumental, cultural, and natural processes, many of which are not well understood, and a myriad of procedures to reduce their impact continue to be proposed.

Recognition that the predominant source of outliers in MT data was inherent in the measurements rather than in the measuring devices led to the development of the remote reference method. The success of remote reference methodology is attested to by its nearly universal adoption in terrestrial MT, and improvements continue to be introduced. *Clarke et al.* [1983] give a recent review of remote reference equipment and procedures. *Kröger et al.* [1983] discuss the bias effects of coherent and incoherent noise on local and remote estimates of the response function. *Goubau et al.* [1984] conducted an experimental investigation of the correlation scale of MT noise by comparing a standard remote reference response function, where the separation between measurements was several km, to a local reference result using a third magnetometer and shorter spacings. They obtained the surprising result that separations of as little as 200 m were adequate to remove outlier bias at periods longer than 1 s, suggesting local nonplanar source fields as the contaminant. At shorter periods, the noise appears to be instrumental, originating in the shields of the SQUID magnetometers, and could be removed by a reference only 2 m from the base. This study indicates how little is known about how and why the remote reference method works. While it is clear that the technique is effective against many types of MT outliers, its limitations are not so obvious, and further work like that of *Goubau et al.* [1984] should be encouraged.

Chave et al. [1986] and *Egbert and Booker* [1986] have proposed new procedures to eliminate the effect of outliers on response functions. Such methods are modifications of proven ones from the field of robust statistics, and are based on iterative re-weighting of the data during regression. The weights are automatically chosen by comparing the regression residuals to a predicted value obtained from the appropriate statistical distribution, and the influence of data corresponding to large residuals is reduced. This gives unbiased response functions as well as meaningful estimates of their error. These methods are still undergoing development and testing using a variety of EM data. However, we predict that robust estimation of the response functions will yield substantially smoother and more precise results under contamination by a broad class of outliers. When combined with the remote reference technique and instrument arrays, this method should prove very powerful.

Park and Chave [1984] have presented a rigorous derivation of the singular value decomposition (SVD) method for estimating the response function and its associated errors. While other least squares approaches assume that Gaussian noise is present in only part of the data (e.g., the electric or magnetic field), SVD treats the more realistic case where noise is distributed among all of the data. Since the relative amounts of noise in the data are rarely known a priori, *Park and Chave* [1984] derive a statistical test that helps to establish the correct relative noise level, and show how the response function varies as this quantity is altered. This approach is useful in dealing with the well-known bias effect of Gaussian noise in MT data, and could easily be combined with robust statistical methods to remove additional Gaussian or non-Gaussian outliers.

Gamble et al. [1982] address the problem of defining a unique coordinate system over three-dimensional (3D) structures, and note that the usual practice of finding individual strike directions using separate principal axis transformations at each frequency often breaks down. They propose a new empirical method, called regional strike determination, that is based on minimization of weighted sums of squares of the response functions over all frequencies, with the weights chosen to accentuate the most precisely known data. They also present examples which demonstrate the consistency of this approach and the inconsistency of the more standard one.

Eggers [1982] gives a thorough discussion of the parameters studied in MT, and shows that the conventional ones—the amplitude and phase of the off-diagonal tensor components, principal direction of the response tensor, skew, and ellipticity—are incomplete, since the full tensor possesses 8 degrees of freedom while these quantities have only 6.5. He then derives an alternate and complete set of parameters using an eigenvalue-eigenvector decomposition of the tensor, and shows how they provide additional insight for interpretation purposes. In particular, the polarization ellipse display of the eigenstates can indicate the form of the 3D structure. *Spitz* [1985] addresses the problem of defining coordinate systems for this type of response function formulation.

Response functions also are important in the GDS method. Gough and Ingham [1983] present a thorough review of single- and multiple-station methods to get the GDS response, and give a variety of ways to present the results. *Beamish and Banks* [1983] discuss the use of a common reference site to study regional structure using a limited set of instrumentation, and claim to get results comparable to those from larger arrays. *Richmond and Baumjohann* [1983] present a new method to treat magnetometer array data, and address the more general problem of inferring spatially continuous patterns from finite sets of point observations. In contrast to most earlier studies, in which a small set of parameters are fit to a large set of data (e.g., spherical harmonic fits with truncation at low order), they propose the use of a large set of interpolating functions and apply constraints from the governing physics to regularize the result. Their examples of field mapping and internal/external part separation are quite encouraging, and this paper deserves serious attention. Future progress in GDS requires the application of more sophisticated analysis techniques, especially frequency-wavenumber and polarization processing to better quantify source field structure.

FORWARD MODELLING

Forward modelling—the prediction of an EM response for a specified earth model—is of central importance in all of the EM disciplines. A notable amount of progress in handling two-dimensional (2D) and (3D) models has been made over the past four years. This has led to improved insight into the effects of complex sub-surface structures on the observed response. In addition, better ways of computing and viewing one-dimensional (1D) models are evolving.

Many types of EM problems require the numerical approximation of integral transforms. Most 1D controlled source models use the Hankel transform. *Anderson* [1982] has produced software based on the digital filter method for the computation of Hankel transforms that is substantially faster and more accurate than previous implementations. *Chave* [1983*b*] has reported a direct numerical integration scheme with Padé convergence acceleration that is generally slower than a digital filter formulation, but that is very accurate and capable of handling integrals having formally divergent integrands. Time domain EM computations require the numerical inversion of the Laplace transform. *Knight and Raiche* [1982] discuss the Gaver-Stehfest algorithm, a procedure which is simple, more computationally efficient than discrete Fourier transform approaches, and which requires a knowledge of the integrand only for real values of the transform variable. An alternate view of Laplace transform inversion based on first kind Fredholm integral equation theory is given in *Pike et al.* [1984], and deserves greater attention.

Forward modelling in 1D is straightforward for MT and GDS, and little purpose is served by the continued publication of analytic solutions for specialized conductivity profiles. However, considerable insight into controlled source and time domain applications continues to come from 1D models, but even relatively simple cases can rarely be expressed in analytic form. While many mathematical approaches to 1D problems exist, the use of a formulation involving poloidal and toroidal modes is especially useful. *Backus* [1986] gives a thorough and rigorous derivation of this Mie representation of the EM field on a sphere that is readily extended to the plane.

Several investigators have stressed the importance of placing equal emphasis on the theoretical behavior of the EM field and on the theoretical resolution of a given measurement. The 1D Fréchet derivatives of the fields are especially useful in this context. The use of Fréchet derivatives as sensitivity functions is discussed by *Gómez-Treviño* and *Edwards* [1983], *Chave* [1984*a*], and *Edwards et al.* [1984].

Most of the effort in 2D and 3D modelling of EM phenomena over the past four years is based on the integral equation (IE), finite element (FE), or thin sheet approaches. The IE method is the most widely used and thoroughly developed EM modelling technique for 3D media. It is especially well-suited for treating isolated bodies embedded in a simpler substrate, since the numerical complexity is limited to the body itself. *Hohmann* [1983] reviews the formulation of IE problems and computational procedures for their solution. *Wannamaker et al.* [1984*a*] describe an IE algorithm for MT problems that can handle a 3D body in an arbitrary layered 1D medium.

The FE method is also receiving increasing attention. *Lee and Morrison* [1985*a*] derive the FE equations for a 2D problem with a finite (controlled) source from a variational principle. P.E. Wannamaker (private communication, 1986) has distributed a 2D FE code for MT, and has applied it to a study of topographic effects on MT data [*Wannamaker et al.*, 1986]. Hybrid methods which combine the FE and IE methods are also in use [*Best et al.*, 1985].

Over the past quadrennium, thin sheet modelling has grown from a mathematical curiousity to a very viable means for treating surface inhomogeneities. *Dawson et al.* [1982] treated TM mode induction with two thin sheets over a halfspace, where the thin sheets represent respectively a conducting ocean adjoining a continent and a resistive crust. *Dawson* [1983] extended this to include the TE mode. Both of these models are substantially more realistic than earlier ones. *McKirdy and Weaver* [1984] developed the theory for a 2D variable conductance sheet over a layered medium, and *McKirdy et al.* [1985] generalized this to the 3D case. Applications of thin sheet techniques include studies of regional induction in Scotland [*Weaver*, 1982] and of current channeling between two oceans [*McKirdy and Weaver*, 1983].

The further development of 2D and 3D modelling codes must involve checking for internal consistency and cross validation with other algorithms or analytic solutions. *Weaver et al.* [1985] have proposed a standard 2D MT model consisting of three adjoining blocks of different conductivities overlying a perfect conductor, along with a closed form expression for the TM mode response function. *Weaver et al.* [1986] present a similar 2D control result for the TE mode. There is a definite need for similar analytic or quasi-analytic solutions for simple 3D bodies.

Numerical modelling has been applied to the study of 3D effects on the MT response functions. This is important both to determine the possible biases caused by multidimensionality and to ascertain the limits where 1D or 2D models are suitable approximations to the 3D earth. *Hermance* [1982*d*, 1983*a*] has used DC thin sheet models to model telluric distortion effects from surface inhomogeneities. *Park et al.* [1983] and *Park* [1985] have identified three distortion mechanisms caused by 3D media: coupling between the upper crust and mantle across a resistive lower crust, resistive coupling of conductive features within the upper crust, and local induction of current cells within finite-size, good conductors. The first two of these produce telluric distortion that is frequency-independent at low frequencies, while the latter is usually frequency-dependent. In many cases, these mechanisms can be differentiated by examining the spatial variation of the response functions. *Wannamaker et al.* [1984*b*] used an IE code to study the effect of a 3D body in a layered host. They present a thorough discussion of the bias effects on the MT apparent resistivity and phase, and conclude that, under certain circumstances, 1D or 2D modelling techniques are suitable for the study of real 3D structures. *Newman et al.* [1985] modelled crustal magma chambers using the IE technique, and showed that the effect of such a body was often surprisingly limited. For a thorough review of 3D current channeling effects in MT, see *Jones* [1983*a*].

As supercomputers and advanced computational algorithms become more widely available, 2D and 3D modelling will become more common. However, it should be remembered that approximate solutions to EM problems are often as useful as the more complex, full solutions; this type of approach has been emphasized and justified by *West and Edwards* [1985].

INVERSION

Fischer and LeQuang [1982] state that the 1D magnetotelluric inverse problem is essentially understood. Despite this optimism and the fact that a 1D model is not appropriate for most MT data, the 1D problem continues to receive substantial attention, judging from the number of papers devoted to it. To some extent, the MT inverse problem is relatively easy when compared to other geophysical inverse problems. EM data are Fréchet differentiable [*Parker*, 1983; *Chave*, 1984*a*; *Abramovici and Baumgarten*, 1985; *MacBain*, 1986]. Furthermore, there are existence and uniqueness theorems for various sorts of ideal data; the most recent of these is due to *MacBain and Bednar* [1986]. Clever schemes to directly invert ideal induction data continue to appear [e.g., *Barcilon*, 1982; *Coen et al.*, 1983].

Real MT data are always discrete and have errors associated with them. Most workers are aware of Parker's work on this type of data [*Parker*, 1983]. He has shown conclusively that when no 1D model fits a data set exactly, then the conductivity model with the smallest least squares or χ^2 misfit will always consist of a set of infinite spikes or delta functions in conductivity. He called this the *D*+ case. *MacBain and Bednar* [1986] claim that Parker's result is not rigorous, but this does not alter the fact that practical schemes for inverting noisy data which do not exclude the possibility of delta function models will converge to these models as the misfit decreases. This applies in particular to most least squares-based layered model fitting routines. For a graphic example, see *Smith and Booker* [1986]. It is extremely important to note that this behavior requires that the data actually contain noise. Inverse schemes which are expected to operate with real data simply cannot be tested with artificial data which equal the response functions for any 1D model even when the data are assigned an error. Synthetic data must have noise added to them after they are generated.

Parker's delta function models usually grossly overfit the data, in the sense that χ^2 is much smaller than its expected value. It is then statistically valid to relax the fit and achieve a larger χ^2. There are an infinite number of possible models between the best-fitting *D*+ type and one with any larger value of the misfit. *Parker* [1983] reviews several ways of constructing families of models in which the model space has been defined to exclude delta functions. *Hooshyar and Razavy* [1982*b*] present related results for the case where the data cover a range of spatial wavenumbers at a fixed frequency. To choose among these models, it is necessary to add a side or regularization condition. Most regularization conditions in current use involve some sort of smoothing. In the past, the most common smoothing criteria involved expanding the model in some finite parameterization (e.g., a small number of layers). The criterion for choosing a specific model is still minimization of the least squares misfit, as was the case for *D*+, but the side condition prevents it from approaching a global optimal fit. The best that can be said for this approach is that, if the parameterization is essentially the same as the truth which generated the data, the inversions seem to work and will recover a reasonable facsimile of the truth; see *Fischer and LeQuang* [1982] and *Pedersen and Hermance* [1986] for examples. The interpretation of such models is not clear if the truth happens to be parameterized in some substantially different way.

Recently, several groups have focused on choosing models which are extreme in some sense. A particularly fruitful criterion is the flattest or smoothest model fitting the data within some prescribed χ^2 [*Marchisio and Parker*, 1984; *Constable et al.*, 1986; *Smith and Booker*, 1986]. These minimum structure inverses allow one to ascertain which features are actually required by a given data set. Furthermore, they turn out to be remarkably good at recovering the structure of synthetic models from noisy data for reasons that are not yet entirely clear.

Whittal and Oldenburg [1986] present another type of extremal inversion in which the problem is cast in terms of inverse scattering. Estimation of the impulse response is analogous to the deconvolution of a seismogram and is a linear inverse problem which is solved by minimizing various norms of the response. This effectively limits the possible structure in the model. The conductivity itself is recovered from the impulse response by solving a nonlinear Fredholm integral equation. Other types of extremal inversions are due to *Oldenburg* [1983] and *Weidelt* [1985], who bound functionals of the conductivity structure.

Linearized Backus-Gilbert types of inversions have been used by *Hobbs* [1983], *Hobbs et al.* [1985], and *Abramovici and Baumgarten* [1985]. *Oldenburg et al.* [1984] have attacked this type of inversion on the grounds that models exist which fit the data and are not linearly close to those produced by the Backus-Gilbert technique. *Smith and Booker* [1986] show that a proper choice of datum and model variable can lead to a nearly linear inverse problem.

Unfortunately, the papers cited do not cast the problem in its nearly linear form, so that nonlinear errors may invalidate their conclusions.

Two- and three-dimensional inversion of electromagnetic induction data is ultimately a more important problem for investigating the earth, but is less advanced than its 1D counterpart. Most groups still rely on forward modelling to interpret data. We believe that rapid advances in 2D and 3D inversion techniques will be made in the next few years as modern algorithms and high speed processors become more widely available.

Inverse schemes for 2D data fall into two classes. The first expands the model using a limited parameterization (i.e., a small number of conductive blocks), and adjusts the values of the parameters to fit the data to within some prescribed tolerance. All existing schemes involve linearization about a starting model and some form of least squares fitting. Two strategies for parameterization are current. For instance, *Zhdanov and Golubev* [1983] advocate a particular function set which allows only seven parameters to describe a wide variety of shapes for an anomalous body. This means that the inverse problem will almost always be grossly overconstrained. The problem with this strategy is that one has no rigorous way of examining nonuniqueness. The alternative is to use a large number of parameters. The most advanced scheme of this type is a proprietary program called ESP/MT developed by W.L. Rodi and colleagues at S^3 in San Diego and described by *Jiracek et al.* [1986]. The forward calculation for the Fréchet derivatives uses self-adjusting finite elements. The normal equations for an updated model are then inverted using a damping method which minimizes spatial derivatives of the model as well as the size of the perturbation. Other programs, such as the widely used Jupp-Vozoff code, minimize only the size of the perturbation.

Minimum structure models are likely to be at least as beneficial in 2D as in 1D, and are essential if 3D inverses are to be obtained. Although existing and potential algorithms discretize the structure, the resolution matrix can be interpreted as digitized Backus-Gilbert windows if the discretization is on a finer scale than the true structure. The nonlinear errors inherent in interpreting the results in this way are just as important in 2D and 3D as in 1D.

The second type of parameterization strategy picks one of the infinite number of possible 2D solutions by finding the one that is closest to some prescribed structure. The method of tightening of surfaces, applied to GDS data by *Zhdanov and Varentsov* [1983] and described in more detail by *Berdichevsky and Zhdanov* [1984], can be useful if the background conductivity is well-known, but could be quite misleading otherwise, especially if the measurements do not include the electric field. Solving the inverse problem for the flow of electric current [*Berdichevsky and Zhdanov*, 1984] is also helpful, but displays similar dangers. These methods all involve some form of spatial filtering of the data. A serious problem with much existing work is that the fields, and especially the electric field, vary on scales shorter than the station spacing. This means that filtering operations are applied to aliased data sets. F.X. Bostick (private communication, 1986) has suggested that the electric field should be profiled, with each successive leg touching the previous one, and calls the method EMAP. With this type of data, prescribed structure methods such as the method of tightening of surfaces, may become quite useful.

A final approach to 2D interpretation which shows some promise, especially for developing starting models for further 2D inversion, is the plotting of pseudo-sections of invariants of the MT response tensor [*Ranganayaki*, 1984]. In particular, she found that the pseudo-section of the phase of the determinant looked remarkably like the structure.

Developments in inversion of controlled source data parallel those for induction by natural fields to some extent. *Parker* [1984] has derived a result for DC resistivity methods which is analogous to the delta function or $D+$ model in MT. He finds that the best-fitting 1D model always contains an arbitrarily thin but complex surface layer. Work by *Lang* [1986] relevant to borehole resistivity concludes that resistivity variability explodes as the layers are allowed to become thin near the hole. This is presumably closely related to Parker's result. *Smith and Vozoff* [1984] have developed a 2D inverse code for dipole-dipole resistivity data which expands the model in boxes and is generically related to earlier MT work. *Tripp et al.* [1984] used a similar philosophy for their 2D DC inverse.

The inversion of time-domain electromagnetic data is in a relatively crude state. Virtually all existing algorithms are based on the assumption of an extremely limited parameterization which forces the problem to be overconstrained. Although perhaps useful in an exploration context, these inverses hardly qualify for the name since they allow no exploration of model space. In contrast to this type of modelling, there is the electromagnetic migration technique of *Zhdanov and Frenkel* [1983]. This technique, which is analogous to seismic migration and closely related to the analytic continuation of fields in the frequency domain, is actively being pursued by groups in the US and Canada. It remains to be seen whether migrated data can be reliably inverted for material properties. In any case, it is likely to give useful structural information.

EARTH STRUCTURE

Deep Sounding

Over the past decade, the ELAS program has focused the efforts of the international EM community on determining electrical properties below the lithosphere. Following on earlier work, several recent papers have examined global averages of deep conductivity. *Campbell and Anderssen* [1983] analyzed the harmonics of the solar daily variation S_q. Their results appear to imply conductivity increases which correlate with the seismic discontinuities at about 400 and 600 km, but no resolution or uniqueness analysis was presented. *Winch* [1984] also looked at S_q and included corrections for a highly conductive ocean. His results are not clearly interpretable in terms of any single model, although the principal concern of the work was possible contamination of the internal part of the magnetic field by the dynamo effect of ocean tides. *Jady and Patterson* [1983] applied three inversion schemes to the construction of models using disturbed time data in the frequency range 0.07–2 cpd. They conclude that a steep conductivity increase occurs near a depth of 1000 km. A subsequent paper by *Jady et al.* [1983] generated a large family of models using Parker's layered and continuous inverses fitting essentially the same data. This time, they concluded that the sharp conductivity increase was more probably at about 700 km. In a different approach to global sounding, *Didwall* [1984] used OGO satellite data from disturbed times to derive a transfer function which is broadly averaged in both time and space. However, she was only able to interpret it in terms of a constant conductivity shell of prescribed thickness.

A dominant trend in recent deep soundings has been the search for lateral conductivity variations in the mantle. Since conductivity is highly temperature dependent, global tectonics virtually guarantees lateral changes in the conductive structure of the earth. *Roberts* [1983, 1986*a*, 1986*b*] reviews a variety of evidence for lateral conductivity changes in the upper mantle. *Vanyan* [1984] argues for deep differences between cratons and younger zones based simply on gross differences in the long period response. More detailed studies are beginning to appear. *Schultz and Larsen* [1983, 1986*a,b*] find equivalent MT responses for a variety of three component magnetic observatories assuming a P_1^0 source. They find that many of these responses can individually be fit with a 1D model within the expected value of χ^2. However, there exist pairs of these stations whose data cannot be jointly fit by any 1D model and must have different local structures.

We expect to see significant progress in this area in the near future. It is probable that electrical structure information in the upper 1000 km of the earth comparable in resolution to seismic normal modes will shortly be available.

Regional Studies

Induction and related techniques have been used in virtually every area of North America and on scales ranging from magnetometer arrays covering 100 square degrees in the EMSLAB project to outlining the building foundations at an archaeological site using DC methods [*Young and Droege*, 1986]. The largest concentration of effort has occurred in the northwestern US and southwestern Canada. EM induction offers a tool which may provide information about the structure and physical properties of the active subduction zone in the region which have eluded seismologists because of the generally low historic seismicity. The largest coordinated EM induction experiment ever, EMSLAB (*E*lectro*M*agnetic *S*tudy of the *L*ithosphere and *A*sthenosphere *B*eneath the Juan de Fuca Plate), has as its major goal the delineation of the complete conductivity structure of the Juan de Fuca plate and underlying asthenosphere from its birth at the ridge to its consumption under North America. The EMSLAB main experimental phase occurred in the summer of 1985. The land-based part of the experiment involved a 67-station magnetometer array stretching from northern California to southern British Columbia and from the coast eastward to Idaho and Nevada, a 15-station MT array on a profile stretching 170 km in from the central Oregon coast which extends a similar offshore profile described under Oceanic Studies, a large number of wide-band MT sites along or near the same profile, and 75 very closely spaced ($\approx$3 km) magnetometer sites along a similar parallel profile. Most of the equipment ran for the months of August and September. A second phase of the field project in the summer of 1986 involved 4 wide-band MT systems using in-field processing and occupying many additional sites along and near the central EMSLAB profile. The total data set is of unprecedented size and quality, and its full analysis will occupy several years.

Earlier work offers supporting data for the EMSLAB goals. *DeLaurier et al.* [1983] used magnetometer data from Vancouver Island (VI) and the adjacent seafloor to construct a model with a good conductor at depth, a thick sedimentary wedge at the coast, and a mid-crustal conductor under the British Columbia (BC) mainland. A conducting slab dipping eastward under VI is consistent with, but not required by, the data. Land-based magnetometer data further south in Oregon [*Neumann and Hermance*, 1985] also require a thick sedimentary wedge, but do not extend to long enough periods to provide any information on the existence of a conducting slab. *Nienaber et al.* [1982] used only land magnetometer data and analog modelling to place a dipping conductor under VI which subsequently rises under the mainland. One could simply interpret their model as a resistive root for VI. However, in recent work performed in conjunction with the Canadian Lithoprobe program, which did detailed seismic profiling across VI, *Kurtz et al.* [1986*a*] collected some very exciting MT data. A 1D inversion of their most isotropic station shows a conducting layer whose top is coincident with the seismic reflector that has been interpreted as the upper surface of the subducted Juan de Fuca plate. They also present a 2D model which is consistent with their MT and earlier magnetometer array data, and interpret the results as strong evidence that substantial sediment is being subducted.

The mid-crustal conductor under BC extends eastwards as far as the Rocky Mt. Trench, where it terminates sharply [*Bingham et al.*, 1985; *Gough et al.*, 1982]. A conductive ridge rising to the shallow crust lies just east of the Rocky Mt. trench. Its relationship to the mid-crustal conductor is uncertain, but its structure is quite sinuous, and it passes close to a known geothermal area studied with a concentrated magnetometer array by *Ingham et al.* [1983]. The southern extent of the mid-crustal conductor in BC may be determined by EMSLAB.

Proceeding eastwards, the next major conductive structure in North America is the Central Plains Anomaly (NACP), which begins in southwest Wyoming and proceeds northwards up the Montana-Dakota boundary into Saskatchewan. *Handa and Camfield* [1984] trace it into northern Saskatchewan, where it bends eastwards, and interpret it as a manifestation of a Proterozoic convergent plate boundary. *Gupta et al.* [1985] track the NACP further eastwards into the Hudsons Bay region.

Another region of probable ancient convergence occurs in the Grenville province of eastern Canada. Again, there is a deep crustal conductor which *Kurtz* [1982] ascribes to pore fluids. This conductor may extend down the Appalachians. *Mareschal et al.* [1983] found that a major conductor paralleling the trend of the mountains must exist west of a magnetometer profile collected in northwestern Georgia. An interesting active source experiment in the same area by *Thompson et al.* [1983] using a 1 km diameter loop source reported a conductor beneath the station at depths coincident with the base of the megathrust discovered by COCORP. This lends support to the COCORP interpretation of a sedimentary structure beneath the crystalline rocks of the overthrust. An example of a conductor in a Tertiary convergent zone is presented in *Stanley's* [1984] interpretation of the Cascade geomagnetic anomaly.

An ancient divergent plate boundary sometimes called the Keweenawan Rift is responsible for the mid-continent gravity high several hundred km east of the NACP. *Young and Rogers* [1985] and *Young and Repasky* [1986] have used MT to investigate small scale structures associated with this ancient rift. However, *Prugger and Woods* [1984] reexamined old magnetometer array data over this feature, and concluded that no major conductivity structure was involved. It is probable that virtually all of the deep, cratonic conductivity structures are associated with old convergent boundaries [*Gough*, 1983]. It is presumably only there that conductive sediments and pore fluids can be carried to great depths.

Modern rifts are quite different from the ancient ones, and most of the induction research in the southwestern US has focused on the Rio Grande Rift and associated structures. *Ander et al.* [1984] briefly outline a large MT data base collected under the auspices of DOE in New Mexico and Arizona, and then present a detailed discussion of 119 audiomagnetotelluric (AMT) and 25 MT stations in a 161 km^2 region of the Jemez Lineament. A 2D modelling effort leads them to the conclusion that a highly conductive body rises to within 20 km of the surface. They interpret this as evidence for partial melt. However, *Jiracek et al.* [1983] argue against partial melt as the direct cause of high crustal conductivity in the nearby Rift, and find that the crust is less conductive in a zone interpreted as containing partial melt by seismic reflection profiling than in nearby regions which appear not to have melt. They suggest that the conductor is probably hot water and that partial melt has actually disrupted a cap rock which traps the hot water. A final paper on the Rio Grande Rift by *Keshet and Hermance* [1986] reconciles older magnetometer array data which were previously interpreted as requiring a deep conductor with more recent MT data which require a shallower structure.

Another large AMT data set in the Questa Caldera of northern New Mexico is presented by *Long* [1985]. It consists of stations every 3 km in a 318 km^2 region which are interpreted by patching together and contouring 1D Bostick inversions of the logarithmic average of the response functions. This paper, as well as *Ander et al.* [1984], demonstrate the need to find better ways to fully present the information contained in very large data sets.

Most of the work in the western US and particularly in the Great Basin between the Sierras and the Rockies reported in the literature has been concentrated on geothermal targets. It ranges from the reconnaissance study of the Long Valley caldera and environs by *Hermance et al.* [1984] and work at Coso Hot Springs reviewed by *Wright et al.* [1985] through controlled source work at Roosevelt Hot Springs, UT, covered by *Ward* [1983]. Other geothermal work in the general area includes the MT survey at Cerro Prieto just south of the California-Mexico border by *Araki* [1982] and a variety of other examples treated by *Berktold* [1983]. Examples of non-geothermal work in the western US are given by *Frischknect and Raab* [1984], who demonstrate the superiority of time-domain EM over conventional resistivity techniques to detect fault structures at the Nevada Test Site, a magnetometer array study by *Towle* [1984], which demonstrates the existence of a conductive zone associated with the Mesa Butte fault system in north central Arizona, and the examination of the channeling of current at tidal periods in the San Andreas fault zone [*Johnston et al.*, 1983]. *Prieto et al.* [1985] present an interesting study in which MT and potential field data are integrated to produce a regional model of the Columbia River basalt plateau.

OCEANIC STUDIES

Over the past four years, the nature of oceanic EM induction studies has undergone some substantial changes, and new directions and applications for this type of research are now reaching fruition. The use of controlled sources to sound the sediments, crust, and uppermost mantle beneath the sea is yielding unique information about the electrical conductivity in this virtually unexplored region of the earth. The application of EM principles to the study of ocean water motions holds the promise of new insight into heat transport and barotropic flow. In addition, the more traditional MT method continues to be applied in new locales, giving valuable measurements of deeper structure.

The Scripps MT results from the Marianas region and on the East Pacific Rise were summarized in the last quadrennial report, and have subsequently been published [*Filloux*, 1982*a*, 1982*b*]. Other recent seafloor MT work has been performed east of Japan in 1981, in the Bay of Plenty near New Zealand in 1982, in the Tasman Sea off of Australia in 1984, and in conjunction with EMSLAB in 1985.

The Japan MT profile, located between the island and the Japan Trench was reviewed by *Yukutake et al.* [1983]. Four magnetometer-electrometer pairs were deployed by the Scripps group for two months at distances of up to 600 km from Honshu, while new seafloor fluxgate magnetometers [*Segawa et al.*, 1982, 1983] were placed nearer the Japanese coast. A notable feature of the data is the strong coast effect, marked by large vertical magnetic fluctuations on the shelf and slope and very small ones on the deep seafloor. Parkinson vectors with an amplitude of 1.9 were seen on the slope, and the peak values occurred at periods near 50 minutes. It is probable that this is the result of electric currents flowing both above and beneath the seafloor observation point. It is interesting to note that a typical oceanic conductivity profile with a rise in conductivity below 100 km is seen at the deepest site, yet a tectonically-similar location in the Marianas [*Filloux*, 1982*a*] does not contain this feature.

In 1984, a set of eight sites in the Tasman Sea were occupied by the Scripps group during a joint investigation with the Australian National University. The results from a single site have been published [*Ferguson et al.*, 1985]. The Tasman Sea electric field data are contaminated by a large component of oceanic origin, presumably associated with nearby western boundary currents that are dynamically analogous to the Gulf Stream. As a result, good MT response functions could be obtained only in the period range of 20 minutes to 10 hours, reducing the resolving power of the data. The response functions also exhibit substantial skew and anisotropy. Inversion of the response functions suggests unusually high conductivity at shallow depths, although the lack of any resolution analysis makes this result difficult to assess.

In the summer and fall of 1985, an oceanic component of EMSLAB involving 40 seafloor pressure, vertical and horizontal electric, and magnetic instruments from the US, Canada, Japan, and Australia was deployed between the coast of Oregon and Washington and the Juan de Fuca Ridge about 500 km offshore. Three east-west lines of instruments were laid out, and the middle one coincided in latitude with the MT profile in central Oregon. The seafloor data are being analyzed in conjunction with the land array described earlier.

A number of workers have suggested a correlation of the depth to conductor inferred from seafloor MT and lithospheric plate age, usually based on linearized model fitting or inversion of the data. *Oldenburg et al.* [1984] reanalyzed the response functions from three seafloor sites of different age using the nonlinear inversion algorithms of Parker. They showed conclusively that distinct models were required by the data from different age regions of the plate, but the monotonic trend of increasing depth to conductor with age could not be fully supported. This was due in large part to unexpectedly low resolving power for the data, as evidenced by the diversity of models that fit them equally well. *Oldenburg* [1983] used a new extremal inversion method to further quantify the low resolving power of seafloor MT data. This problem is due to the narrow, two decade range of usable frequencies in seafloor MT. It is not likely that improvements in instrumentation will dramatically improve this situation, and other methods will be required to investigate shallow electrical conductivity in particular. Future applications of seafloor MT in the oceans will probably be aimed at the delineation of tectonic structure using arrays of instruments in the spirit of EMSLAB. Array deployments also allows the use of GDS, which is not as limited as MT by low frequency oceanic noise.

Geomagnetic induction in transoceanic telecommunications cables has been studied extensively by a group at AT&T Bell Labs [*Lanzerotti et al.*, 1985, 1986; *Meloni et al.*, 1983, 1984; *Thomson et al.*, 1986]. A review of the subject appears in *Meloni et al.* [1983]. In the most recent of these papers, *Lanzerotti et al.* [1986] note a high correlation of the voltage in a nearly E-W cable with the E-W magnetic field. They suggest a N-S flowing telluric current off of the coast to explain the data. This is probably another manifestation of the enhanced coast effect noted by *Yukutake et al.* [1983], with electric current flowing in both the ocean and underlying rock on the continental shelf and slope. Contemporaneous seafloor magnetic and cable observations would be invaluable in sorting this out.

New information on the conductivity of the oceanic crust has come from the application of controlled source induction methods. *Becker et al.* [1982] and *Becker* [1985] describe several experiments using a large scale resistivity method in a deep ($\approx$1500 m) DSDP borehole on the Costa Rica Rift. The method is useful for the inference of conductivity in a zone of 20–50 m radius about the hole. Conductivities of $\approx$0.1 S/m were found in the upper pillow lavas of the oceanic crust, decreasing sharply to $\approx$0.002 S/m in the underlying dike complex near the base of seismic layer 2. Using Archie's Law, the inferred apparent porosity varies from 10% in the pillow lavas to about 2% at depth, and three porosity zones were observed which correspond roughly in location to seismic layers 2A, 2B, and 2C.

Frequency-domain controlled source measurements in the sea are being performed by groups in both Canada and the US. The former work is based on a vertical wire source extending from seafloor to sea surface and energized by a surface ship together with a series of seafloor horizontal magnetic receivers. The method is a variant of the magnetometric resistivity method. *Edwards et al.* [1985] describe the first use of the method in an inlet off of British Columbia, in which a conductivity profile through a thick sedimentary section was obtained. *Nobes et al.* [1986] give results from a similar sounding in the Middle Valley of the Juan de Fuca Ridge through a thick hemipelagic sequence overlying basaltic basement.

Cox et al. [1986] present some preliminary results from a deep controlled source sounding using a seafloor horizontal electric dipole source and a series of horizontal electric field receivers placed up to 70 km away. Signals were quite identifiable at the longest ranges at frequencies up to 24 Hz. A series of simple models could be fit to the data and are typified by a 5 km crustal layer of moderate ($\approx$0.001 S/m) conductivity overlying a resistive halfspace of conductivity 5×10^{-5} S/m. The low conductivity in the uppermost mantle requires a low volatile content in the rocks to be consistent with laboratory data. However, the conductivity of the oceanic lithosphere cannot be this small everywhere, or the resulting electrical isolation of the ocean from the conductive deeper mantle would produce large electrostatic fields at the ocean boundaries extending well into the ocean basins that are not observed. *Chave and Cox* [1983] used a simple model of this effect and measured oceanic MT responses to show that the average conductivity of the oceanic lithosphere is $\approx$.001 S/m. This suggests that high conductivity paths must exist within the ocean basins which short circuit a resistive ocean-deep mantle path, assuming the *Cox et al.* [1986] results are typical of the oceanic lithosphere away from tectonic complications. These high conductivity pathways are probably associated with mid-ocean ridges or continental shelves.

There has also been a substantial rise in interest in EM induction by ocean water currents in recent years, both due to its possible role as a noise source for seafloor MT and for oceanographic applications. *Chave* [1984*b*] investigated EM induction by oceanic internal waves. Oceanic internal wave model spectra are similar in magnitude to seafloor magnetic field spectra at frequencies between 0.2 and 1 cph, depending on ionospheric activity and latitude, and could serve as a source of contamination in seafloor data. The effect is more severe in the vertical magnetic component and at high latitudes, hence may be more serious in a GDS than in an MT context.

Chave and Filloux [1985] and *Bindoff et al.* [1986] have examined a usually neglected portion of the seafloor EM field, the vertical electric component. In the absence of marked structural heterogeneity, this part of the EM field is entirely of oceanic origin, reflecting the east-west water velocity at the point of measurement, and has no counterpart on land. Both of these studies showed that the vertical electric field spectrum can be explained by the internal wave model of *Chave* [1984*b*] between about 1 cph and 1 cpd. At longer periods, mesoscale oceanic motions dominate the data, and the ocean tides are also prominent. This type of measurement will undoubtedly find increased application in oceanography, particularly in the study of long-period, bottom-trapped wave phenomena.

The induction of electric currents in submarine cables by ocean flows, and especially intense western boundary currents like the Gulf Stream, has been known for many years. *Sanford* [1982*a*] provides a thorough review of theoretical and observational aspects of cable measurements. *Larsen and Sanford* [1985] report on the analysis of long-term measurements collected on a cable under the Florida Current. After correction for geomagnetic and tidal induction, they found agreement of the cable and more conventional oceanographic measurements of transport to within 2%.

At periods of several days to months, the baroclinic (i.e., depth-dependent) variability of the ocean is larger than the barotropic (i.e., depth-independent) variability, and hence dominates conventional point measurements made in the deep ocean. The seafloor horizontal electric field yields a depth-averaged estimate of the water velocity, and is well-suited to studies of the poorly understood barotropic component. *Sanford* [1986] reviews the use of EM principles to examine barotropic flow. A major experiment to use EM methods for oceanographic purposes is now being conducted by Scripps. In the summer of 1986, 44 seafloor pressure recorders, magnetometers, and horizontal and vertical electric field instruments were deployed in a 1500 km by 800 km array for one year to study the wavenumber structure of barotropic wind-forced flow, as well as pursue a variety of other oceanographic and geophysical objectives. This experiment, called BEMPEX (*B*arotropic *E*lectro*M*agnetic and *P*ressure *EX*periment), is the first use of EM techniques in the deep ocean basins for oceanographic purposes at long periods.

MISCELLANEOUS TOPICS

Most natural source EM studies are concerned with induction in the conducting earth by external current systems. At very long periods, induction from the core dynamo below the earth's surface may also be important. *Backus* [1983] determined the weighted averages of mantle conductivity that can be inferred by considering the earth as a linear filter, with a geomagnetic jerk as input at the core-mantle boundary and an output at the earth's surface. *Lanzerotti et al.* [1985] used a $\approx$4500 km telecommunications cable to determine the DC component of the earth potential, obtaining a nearly null result. This may require nearly equal toroidal and poloidal parts for the geomagnetic field at the core-mantle boundary. However, *Backus* [1982] showed that a critical layer will exist in the mantle that screens out an internal electric field if a conductivity minimum occurs between the earth's surface and the core-mantle boundary, complicating the interpretation of the cable data.

Time domain or transient EM methods have received an increasing amount of attention, mostly concentrated on shallow exploration targets of industrial interest. The advantages of time domain over frequency domain EM include reduced sensitivity to near-surface lateral heterogeneity and freedom from contamination by the portion of the signal travelling through air, since measurements are typically made when the transmitter is off. *Hoversten and Morrison* [1982] derived the transient magnetic fields of a loop source inside of a 1D layered medium, demonstrating graphically the "smoke ring" diffusion form of the induced fields and giving a simple picture of the effects that structure has on surface observations. *Oristaglio and Hohmann* [1984] give a similar view of some 2D time domain problems. *Keller et al.* [1984] describe an electric dipole source, loop receiver system designed for deep sounding. *Fitterman and Stewart* [1986] present a time domain model study of four groundwater exploration situations. *Edwards and Chave* [1986] and *Cheesman et al.* [1986] suggest some systems and applications for transient EM on the seafloor. A variety of other time domain problems are covered in a special issue of *Geophysics* [*Nabighian*, 1984]. Numerical models for 2D/3D time domain EM are also appearing [*Adhidjaja et al.*, 1985; *SanFilipo and Hohmann*, 1985; *SanFilipo et al.*, 1985; *Newman et al.*, 1986], and will provide insight for the interpretation of field data, although the difficulty of obtaining such solutions cannot be overemphasized.

Laboratory measurements of crust and mantle materials are reviewed by *Hinze* [1982], *Duba* [1982], and *Lǎstovičková* [1983]. The complicating effects of inadequate sample characterization and physiochemical changes during the measurement process are emphasized by *Duba* [1982]. Recent work on olivine has revealed that point defects play a crucial role in determining its electrical conductivity [*Schock et al.*, 1984; *Schock and Duba*, 1985; *Sato*, 1986].

Kariya and Shankland [1983] compiled laboratory conductivity measurements for dry mafic and silicic lower crustal rocks as a function of temperature. Using best-fitting curves of conductivity against temperature, they showed that the results could be used to infer an upper bound to in situ temperature from MT measurements. Building on this study, *Shankland and Ander* [1983] expanded the data base and compared the results to field EM and heat flow measurements. They showed that plots of conductivity against reciprocal temperature were reasonably ordered, but that all of the field data had conductivity values orders of magnitude above the laboratory ones, suggesting the presence of volatiles. They also found that the inferred temperatures for tectonically-active areas were systematically above those under shields, and suggested that EM surveys could be used to predict regional geotherms. These results provide considerable encouragement that EM field data can be interpreted in terms of fundamental physical parameters.

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