MARINE ELECTROMAGNETIC INDUCTION STUDIES

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Abstract. In reviewing seafloor induction studies conducted over the last seven years, we observe a decline in single-station magnetotelluric (MT) experiments in favour of large, multinational, array experiments with a strong oceanographic component. However, better instrumentation, processing techniques and interpretational tools are improving the quality of MT experiments in spite of the physical limitations of the band limited seafloor environment, and oceanographic array deployments are allowing geomagnetic depth sounding studies to be conducted. Oceanographic objectives are met by the sensitivity of the horizontal electric field to vertically averaged motional currents, providing the same information, at much greater reliability and much lower cost, as an array of continuously operating current meter moorings.

The seafloor controlled source method has now become, if not routine, at least viable. Prior to 1982, only one seafloor controlled source experiment has been conducted; now at least three groups are involved in the experimental aspects of this field. The horizontal dipole–dipole configuration is favoured, although a variant of the magnetometric resistivity method utilising a vertical electric transmitter has been developed and deployed. By exploiting the characteristics of the seafloor environment, source receiver spacings unimaginable on land can be achieved; on a recent deployment dipole spacings of 90 km were used with a clear 24 Hz signal transmitted through the seafloor. This, and prior experiments, show that the oceanic upper mantle is characteristically very resistive, $10^5\,\Omega\,\mathrm{m}$ at least. This resistive zone is becoming apparent from other experiments as well, such as studies of the MT response in coastal areas on land.

Mid-ocean ridge environments are likely to be the target of many future electromagnetic studies. By taking available laboratory data on mineral, melt and water conductivity we predict to first order the kinds of structures the EM method will help us explore.

1. Introduction

This review is intended to cover the period since Law's (1983) report on marine electromagnetic research, and so the emphasis will be on papers published or presented since about 1982. Reviews from before 1982 include those by Cox (1980), Filloux (1979), and Fonarev (1982). A more recent review emphasising exploration applications is presented by Chave *et al.* (1990).

Since 1982 there has been a continued and steady interest in the use of magnetotelluric (MT) and geomagnetic depth soundings (GDS) on the seafloor. The areas of controlled source and oceanographic use of EM methods, however, have seen significant development and expansion. When Law's review was published only one controlled source experiment has been carried out, but now there are at least three groups working in this area, and most of the recent large projects have a major, if not predominant, component of oceanography, although by their very size and nature they must be multi-disciplinary in their objectives.

The largest project, EMSLAB, resulted in the deployment of 118 instruments in 1985 and involved a very large group of workers (there are 35 co-authors comprising The EMSLAB Group, 1988) from six countries. GDS and MT sites covered the US states of Oregon and Washington (spilling over into adjacent states and

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Canada) and the seafloor exposure of the Juan de Fuca plate. The main objective of this exercise was the delineation of the subducting slab, but oceanography and regional geology also motivated the work. The experiment employed mostly existing equipment, but the collection of such a large data set has supported and encouraged improvements in response function estimation and forward and inverse modelling.

In an effort to obtain detailed structure over the ridge in the EMSLAB area, a group of Japanese, Canadians and Australians have deployed 12 instruments (yielding 11 magnetometer and 2 E-field sensors) over a small area of the Juan de Fuca Ridge (unpublished EMRIDGE deployment cruise report, 1988). These instruments were recovered in November, 1988.

The Tasman project (TPSME) was a joint US/Australian deployment of seafloor pressure/MT equipment, across the Tasman Sea between SE Australia and New Zealand, for four months in 1983/1984. A landward extension of the traverse (on the Australian side) was made using Gough-Reitzel GDS instruments.

The 1986/1987 BEMPEX (*B*arotopic *E*lectro*M*agnetic and *P*ressure *EX*periment) study in the North Pacific, in which a total of 41 electric field, pressure and magnetic recorders were deployed over a region 1000 km square (Luther *et al.*, 1987; Chave *et al.*, 1989) is an ambitious project with primarily oceanographic goals. This was the longest deployment of these instruments.

The following sections of this review introduce the marine electromagnetic environment, and then consider the areas of geomagnetic and magnetotelluric sounding, controlled source methods, and oceanographic studies. Each of these sections has a brief introduction followed by a review of recent activity. Finally, the use of EM methods in the study of mid-ocean ridges is considered as a means of predicting future progress in an area of rapidly expanding research.

2. The Marine Electromagnetic Environment

For those more familiar with electromagnetic experiments conducted on, or over, the land, the marine environment is a world turned upside-down. Experiments must be carried out within a relatively good conductor, the sea, often for the purpose of studying the less conductive material of the seafloor below. This conductive environment may be advantageous in several ways, however. Movement of seawater through the Earth's magnetic field produces an electric field. While this motion constitutes a noise source for M.T. at periods longer than an hour (Chave and Filloux, 1984), measurement of the electric field can be used to infer large-scale water transport. On the deep ocean floor the large conductance of the overlying sea severely attenuates short period electromagnetic variations originating in the ionosphere. The magnetic field is attenuated more rapidly in frequency than the electric field because it is much more sensitive to the resistive seafloor (Chave *et al.*, 1990), but by 0.1 Hz both fields are reduced to 0.1% of their surface values. This results in an extremely quiet environment for controlled source experiments.

Controlled source studies also benefit from the absence of an air wave; propagation at high frequencies (or short times) is solely through the seafloor, which is usually the primary region of interest.

All marine electrical methods benefit from the ease with which contact may be made with the environment. Potential electrode noise is lower than experienced on land, as temperature, salinity and contact resistance are all nearly constant. Water choppers (Filloux, 1987) may be used to reduce low frequency electrode noise (below about 0.01 Hz) even further. Controlled source experiments may use transmission currents of up to 100 A in electric dipoles because a low impedance contact to seawater is so easily made. Both electric receivers and electric transmitters may be flown through the water, or dragged across the seabed, while making continuous electrical contact.

Figure 1 presents a resistivity-depth profile of the oceanic seafloor, based on borehole logging and soundings by controlled source and M.T. methods. These experiments will be discussed later in this review, but the figure presents an instructive summary of seafloor resistivity.

Seawater resistivity is about 0.3 Ωm throughout most of the ocean, although it is as low as half this value in warmer, surface waters. In the oceanic crust, electrical conductivity is largely controlled by the presence of pore fluids (predominantly

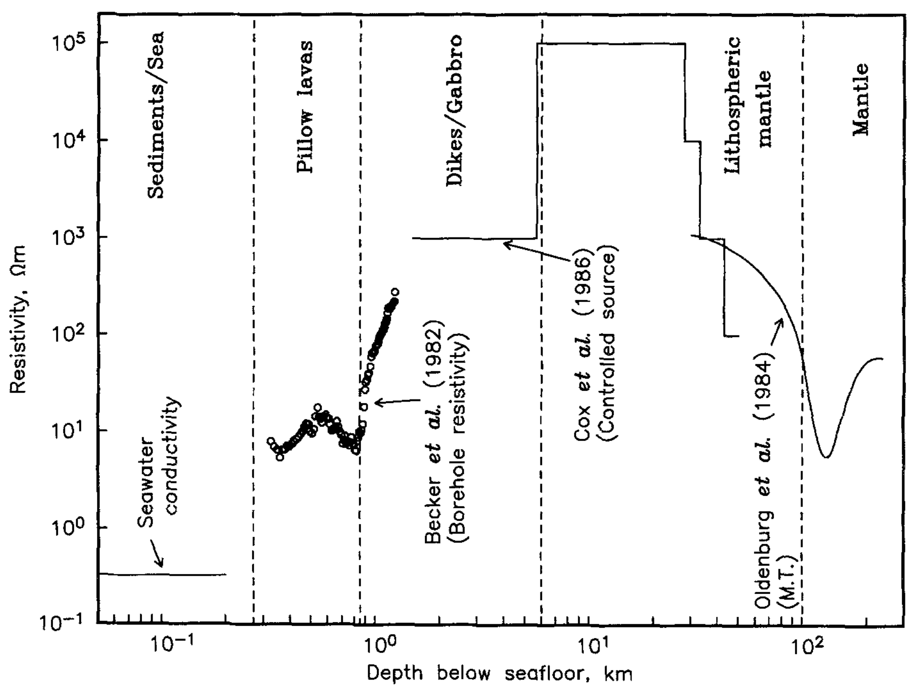

Fig. 1. Seafloor resistivity as a function of depth, based on data from large scale borehole resistivity and interpretations of controlled source and MT soundings. The lithospheric ages are 6.2, 25 and 30 My respectively.

seawater, but also magma in the ridge areas), varying both with the size and connectedness of the fluid passages and with temperature. Saltwater conductivity as a function of temperature and pressure has been measured by Quist and Marshall (1968). Near-surface, water saturated sediments have high conductivities approaching that of seawater. Pillow lava resistivities are about 10 Ωm, while the underlying intrusives are less porous and about 1000 Ωm. At the Moho porosity drops again, and the upper mantle is thought to be very dry, less than 0.1% water, and very resistive, 10^5 Ωm or more.

The mineral conductivity of cool, dry silicates is very low, but increases exponentially with temperature. The upper mantle is probably composed of a composite of olivine and pyroxene with only minor amounts of other minerals. Olivine is the dominant mineral both in terms of volume fraction and conductivity, and so has attracted considerable attention in laboratory studies of physical properties. The most reproducible data have been obtained from single crystal measurements rather than those on whole rocks, and Duba *et al.* (1974) and Schock *et al.* (1989) report conductivities for olivine crystals of mantle composition. Following the lead of Shankland and Waff (1977), it became popular to increase the conductivity estimated from single crystal studies by a factor of ten to allow for the influence of minor minerals and impurities. However, recent work by Constable and Duba (1990) indicates that single crystal conductivity may provide a good analogue for rock conductivity. These laboratory studies in combination with seafloor sounding experiments suggest that upper mantle resistivity between about 30 and 100 km drops three orders of magnitude as temperature increases from below 850 °C to around 1400 °C.

A minimum in resistivity at depths of 100–200 km is generally observed using M.T. sounding. This high conductivity zone is usually interpreted as being associated with partial melting at the base of the lithosphere. The conductive zone is also seen beneath continents, and marks the point at which oceanic and continental mantle conductivity structures are indistinguishable. Beneath the conductive zone, mantle resistivity appears to rise to about 100 Ωm.

3. Geomagnetic Depth Sounding and Magnetotelluric Studies

3.1. Introduction

Geomagnetic depth sounding (GDS) refers to the deployment of a line or array of magnetometers to observe the spatial behaviour of the vector magnetic field as a function of frequency. In its simplest form, the vertical field is considered the response to the exciting field, which is predominantly horizontal at sharp conductivity boundaries such as between the air and land or sea. In an Earth devoid of lateral variations in conductivity, there is no vertical field response to a horizontal field, and so the method is excellent at detecting and mapping 2D and 3D structure. Depth resolution is poor, however, and there are numerous cases in the literature where current being chanelled through shallow conduction paths has been interpreted as deeper conductive structure.

In magnetotelluric (MT) sounding, the horizontal electric field is measured as well as the magnetic field. Like the vertical field in GDS, the electric field can be considered the response to the exciting horizontal magnetic field, except that now there is a response from a laterally uniform Earth. The **E**/**B** response as a function of frequency may be interpreted to give vertical (1D) conductivity structure, and, if an array of stations is available, 2D or 3D structure.

On the seafloor both methods suffer from the filtering of the external field by the ocean at periods on the order of 100 s or shorter and contamination of the magnetic field by water motion, beginning with internal waves at periods of about 1 hour and extending into longer periods with tides and ocean currents. As a result, MT response functions are limited to 2 or 3 decades of frequency. However, as may be seen from the example in Figure 1, the MT method samples seafloor conductivity at a depth unattainable using other techniques.

3.2. Recent Activity

Law (1983) predicted that ring-shaped cores in fluxgate sensors would be used to improve the sensitivity of seafloor magnetometers. Many groups are indeed installing such sensors, and Segawa *et al.* (1982, 1986) describe such a device. Using the data from this test deployment, Yukutake *et al.* (1983) report preliminary results for three MT stations across the Japan Trench, modelling depths of a little over 100 km for the conductive upper mantle layer under 125 My old crust. Koizumi *et al.* (1989) deployed a proton precession magnetometer alongside the fluxgate instrument in order to establish the drift characteristics of the fluxgate. The difference between the total fields from each instrument shows a rapid drift of over 40 nT during the first four days followed by a more gentle drift of 10 nT over the remaining 75 days. The authors consider all the drift to originate within the fluxgate device.

Ferguson *et al.* (1985) present another preliminary interpretation, this time for the Tasman experiment. Models are presented, but the response function for the station under consideration exhibited a strong, frequency independent anisotropy. Analysis of further stations (Lilley *et al.*, 1989) shows that this behaviour is typical, and it is likely that current channelling, or at least the effect of 2D structure, in the restricted water of the Tasman Sea is affecting the response functions. The investigators are currently examining this possibility with the use of thin-sheet modelling.

The largest seafloor MT experiment to date has been part of the even larger EMSLAB project (The EMSLAB Group, 1988; see also the special issue of *J. Geophys. Res.* **94**, No B10). Thirty-nine of the EMSLAB instruments were deployed on the seafloor between the Washington/Oregon (U.S.A.) coast and the spreading centre at the Juan de Fuca Ridge. It will be some time before this large mass of information is fully analysed, but the magnetotelluric data and qualitative interpretations are presented by Wannamaker *et al.* (1989a). A model fitting one traverse on land (the 'Lincoln Line') and five of the seafloor sites is presented by The EMSLAB Group (1988) and Wannamaker *et al.* (1989b), in which a conductive, subducting slab is featured. Filloux *et al.* (1989) discuss the objectives of the seafloor part of the experiment and depict Parkinson vectors for the seafloor array. The induction vectors show a strong coast effect with little or no distortion associated with the ridge. Dosso and Nienaber (1986) and Chen *et al.* (1989) use an analogue model to characterize the magnetic variation pattern expected from a subducting Juan de Fuca plate.

The electrical model presented by Wannamaker *et al.* (1989b) includes structure extending 200 km each side of the coast and to a depth of 400 km. One of the principal objectives of the EMSLAB exercise was to study the entire section, from oceanic to continental regimes, so the distinction between the marine and non-marine aspects of this work becomes somewhat arbitrary. In terms of seafloor structure, the model contains a very conductive (1–2 Ωm) sedimentary wedge, which is about 2 km thick and underlain by a 3000 Ωm lithosphere extending to 35 km depth. We know from controlled source sounding that this is a simplification of the lithospheric structure; it is a little resistive for oceanic crust and is most probably too conductive for lithospheric mantle. Oceanic mantle conductivities below 35 km in the model are also unusually large; about 2 orders of magnitude more conductive than the continental mantle of the same model, and about 1 order of magnitude more conductive than the profile presented here in Figure 1 (but which has been generated from data over older lithosphere). Partial melting of 7% at a depth of 35 km is inferred from the model resistivity of 20 Ωm, with melt decreasing to zero at a depth of 200 km. This is an unusually large melt fraction; most seismologists would be uncomfortable with more than 2–3% (Sato *et al.*, 1989). Sato *et al.* also discuss the problems associated with the estimation melt fraction from electrical conductivity in this context, citing unknown pressure effects, unknown frequency dependence and the unknown effect of grain boundary phases. Of these, the unknown effect of pressure on the conductivities of olivine and melt is probably the most severe problem as it is difficult to make electrical conductivity measurements at high pressure in the laboratory whilst adequately controlling the sample environment. One might add to the list of unknowns the oxygen fugacity of the mantle, which in turn could be responsible for an order of magnitude uncertainty in olivine conductivity. The effects of grain boundary phases and frequency dependence on olivine conductivity have been demonstrated by Constable and Duba (1990) to be minor, at least for the dunite samples they studied.

Although partial melt may be invoked to explain the resistivity to 200 km depths, the EMSLAB model resistivity of 1 Ωm between 200 and 400 km presents a problem for even the authors to interpret. In view of these generally high conductivities, one is concerned that there exists a breakdown in the assumptions of two-dimensionality when interpreting the seafloor data. The compensation distance (discussed below) implied by the 3000 Ωm oceanic lithosphere of the model is 1000 km, considerably larger than the tectonic plate being studied. The effects of non-planar source-field morphology and static distortion on the MT data are shown to be minimal by Bahr and Filloux (1989), who demonstrate that estimates of the first four harmonics of the Sq response are consistent with plane-wave MT responses at similar frequencies.

In order to obtain more detailed information over the Juan de Fuca Ridge than was possible during the EMSLAB experiment, the EMRIDGE programme saw the deployment of 12 instruments to obtain 11 magnetometer sites and 2 E-field sites within the flanks and depression of the ridge (unpublished EMRIDGE cruise report, 1988). All of the instruments were recovered in November 1988, and initial results again indicate that there is no conductivity signature for the ridge (Hamano *et al.*, 1989), implying the absence of a large, continuous magma chamber. Another interesting result reported by Hamano *et al.* (1989) is that useful E-field data may be obtained within the MT frequency band without employing water choppers or very long antennae.

The estimation of lithospheric thickness and its probable increase with age has been one of the goals of marine MT sounding for some time. After Oldenburg (1981) reported a strong correlation between lithospheric thickness and age, further work by Oldenburg *et al.* (1984) showed that although the data demanded different structure beneath the different sites, the correlation with age was not as strong as previously thought.

Niblett *et al.* (1987) operated a MT station on sea ice in the Arctic Ocean for one month. During that time the ice sheet moved back and forth over the Alpha Ridge, a topographic high on the Arctic seafloor. Apart from the technical difficulty of operating MT equipment in subzero temperature, the experiment is novel because in many respects it is equivalent to a seafloor sounding, except of course that the measurements were made on the sea surface. The problems of contamination by water motion and insensitivity to shallow seafloor structure are the same as those for seafloor measurements. The results were very sensitive to seafloor topography, and 2D modelling of bathymetry accounted for the anisotropy observed in the data. It should be noted that seafloor topography up to 300 km from the measurement site was considered to influence the data. No lateral structure in seafloor conductivity was required, and the data were fit by a 1000 Ωm lithosphere underlain by a 10 Ωm mantle at a depth of 85 km. As noted by the authors, this electrical asthenosphere is shallow for the inferred age of the seafloor in this region (100 My).

Berdichevsky *et al.* (1984) used 2D finite difference modelling of a horst type structure to conclude that seafloor MT and GDA experiments (in contrast to observations at the sea surface) are relatively insensitive to such topographic irregularities.

3.3. The coast effect

It was stated above that the GDS method is particulary sensitive to lateral variations in conductivity. The largest variation of this kind is, of course, the junction between land and sea, or coast, and indeed the shorelines of the world produce a marked distortion of the magnetic field at periods of about an hour. The distortion is most easily seen in the vertical field, which is of comparable magnitude to the horizontal field at the coast and extends inland for many hundreds of kilometres. Kendall and Quinney (1983), Singer *et al.* (1985) and Winch (1989) consider the theory of modelling induction in the world ocean.

Neumann and Hermance (1985) deployed magnetometers at 9 sites perpendicular to the Pacific coast in Oregon, U.S.A., and observed a coast effect that was too large to be explained by the ocean alone. They concluded that currents induced in a thick sedimentary wedge on the continental shelf were required to satisfy the data. Kellett *et al.* (1988a) presented three Parkinson vectors, from part of a larger experiment, to demonstrate that the coast effect off SE Australia was largest at the middle of the continental slope, as one would expect. Kellett *et al.* (1988b) present Parkinson vectors for an observatory in New Zealand at periods of 200–10 000 s.

In a modelling study of the coast effect, Fischer and Weaver (1986) examined the ability to detect lateral variations in lithospheric conductivity in the presence of the strong ocean response. They note that MT methods do not perform very well in these circumstances, but that GDS experiments are well suited to the detection of lateral conductivity changes, and, in principle, could discriminate lateral lithospheric structure in spite of the strong oceanic response. However, they show that the presence of even 100 km of 250 m deep water over the continental shelf prevents this discrimination, even when seafloor measurements on the shelf are included.

Similar modelling was used in a study supported by DeLaurier *et al.* (1983) to interpret 3 seafloor and 2 land magnetometer stations on and off Vanvouver Is., Canada. They produce a model featuring a sedimentary wedge and a downward dipping conductive slab. Although much of this model is generated by the workers preconceptions of the structure, in a later MT analysis of three land stations, Kurtz *et al.* (1986) clearly observe a conductive feature at a depth of about 20 km, associated with the down-going slab and coincident with a strong seismic reflector. This study must be considered the first unambiguous detection of a subducting slab using EM methods.

In a review of Australian conductivity studies, Constable (1990) compiled coast effect data for Australia, reproduced here as Figure 2, and noted that the distinction between shield regions and younger regimes is not marked, as has been commonly proposed. The data are fit reasonably well by a model of Cox *et al.* (1971) in which an ocean of realistic thickness and conductivity is embedded in an infinitely resistive lithosphere. The lithosphere is terminated by a conductor at a depth between 240 and 420 km. The conclusion is that without data beyond the continental shelf there is no sensitivity to continental resistivity in this type of experiment.

It is encouraging to see coast effect studies being supported by quantitative modelling, although the non-uniqueness of EM interpretation is a persistent problem and the modelling has to rely heavily on structure presumed *a priori.* Such studies have a great advantage in that the predominant feature, the ocean, is of known shape and conductivity. Indeed, by utilising analogue modelling the

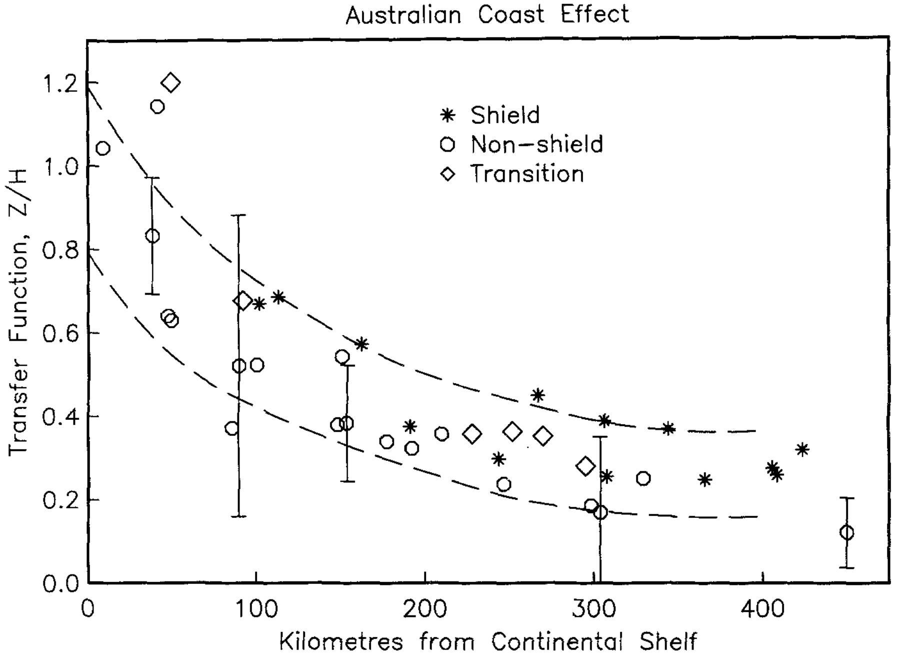

Fig. 2. Compilation of Australian coast effect data from Constable (1990), categorized according to geological regime. While the coast effect for shield areas is generally larger than for non-shield areas, there is some overlap. The broken lines are models of Cox *et al.* (1971) with conductors at depths of 240 (lower curve) and 420 km.

response of arbitrarily complex coastal boundaries can be obtained (Herbert *et al.*, 1983; Chan *et al.*, 1983; Dosso *et al.*, 1985; Hu *et al.*, 1986; Dosso *et al.*, 1986). None of the coast effect studies have incorporated the very resistive oceanic lithosphere, and one of the most interesting goals of future work will be to examine the rôle of continental margins in the leakage of current past this layer.

4. Controlled Source Methods

4.1. Introduction

The use of a controlled EM source to study the seafloor has at last come of age. There are groups at Cambridge University (U.K.), Toronto University/Pacific Geoscience Centre (Canada), and Scripps Institution of Oceanography (U.S.A.) all active in this area. The importance of this method lies in the ability to fill the gap between deep boreholes a kilometre or two deep and MT sounding, which on the deep seafloor is not sensitive to structure shallower than about 50 km. The gap is filled by generating at the seafloor the high frequency source field that is removed from the natural spectrum by the overlying water, that is at frequencies of 1 Hz plus or minus a few decades. In addition to the necessity of providing a high frequency signal, there are three elements which make seafloor controlled source EM a very attractive method:

Firstly, the absence of a natural signal at the frequencies of operation combined with an isothermal and isosaline environment for the receiver electrodes results in a very low noise level for electric field receivers; as low as $10^{-24}\,V^2\,m^{-2}\,Hz$ above 1 Hz for a 1 km antenna (Webb *et al.*, 1985). Magnetic noise is also lower than on land, but magnetometers are vulnerable to motion caused by water currents and lack the resolution afforded by E-field receivers employing long antennae. As a consequence, magnetic receivers have only been used for relatively shallow sounding.

Secondly, since the seawater absorbs controlled source signals as effectively as ionospheric signals of similar frequency, receivers sufficiently far from the source never detect energy which has propagated through the seawater. Instead the measured signals must travel through the more resistive seafloor, and since the seafloor is usually of interest, this is a great advantage. In contrast, on land the primary signal propagating through the resistive atmosphere is much larger than the signal through the earth. Viewed in terms of time domain sounding, on the ocean bottom the signal at early time is sensitive to seafloor conductivity and at late times is a measure of seawater conductivity. On land, the early time signal has propagated through the atmosphere.

Thirdly, electrical contact is easily made with seawater, allowing large transmitter currents on the order of 100 A and allowing both E-field sources and receivers to be dragged through the water during operation.

We may quantify the comparison between magnetic and electric receivers for controlled source sounding. The noise level for the E-field receiver reported by Webb *et al.* (1985) corresponds to a controlled source signal of about $10^{-18}\,V\,m^{-1}$ per unit dipole moment after typical source strength (2×10^4 Am for a horizontal electric dipole) and stacking time (30 min) are taken into account. As illustrated by the models of Chave and Cox (1982), this corresponds to a magnetic field on the order of 10^{-12} nT, which may be converted to an instrument noise level of $10^{-12}\,nT^2/Hz$. Noise estimates for seafloor magnetometers in the relevant frequency band are difficult to find in the literature, but Wolfgram *et al.* (1986) report a resolution of 10^{-3} nT after stacking, which corresponds to $10^{-7}\,nT/Am^2$ for the source dipole assumed above. From this value it may be concluded that the effective range for a horizontal E-field transmitter/B-field receiver combination would be about 5 km at 1 Hz. The E-field receivers, on the other hand, are capable of measuring the signal to ranges of many tens of kilometers. This having been said, it may be noted that below 1 Hz the seafloor E-field spectrum rises more rapidly than the magnetic spectrum, and that magnetometers do not suffer from the $1/f$ noise which is unavoidable using electrodes until frequencies are low enough for water choppers to be practicable. It may be that for controlled source experiments in the 10–100 s period range, which might be required for a relatively conductive target, magnetometers may be useful in deep controlled source soundings.

4.2. Recent activity

There have been several controlled source system geometries deployed in the marine environment. The Scripps work has concentrated on the use of seafloor horizontal electric dipoles for both the source and receivers, with a fully inductive AC signal being broadcast from the transmitter dipole. This geometry is also used by the Cambridge group.

The Toronto/PGC approach has been a variation on the magnetometric resistivity method of Edwards (1974), and is described by Edwards, Law and DeLaurier (1981). A vertical electric bipole is suspended from a ship to the seafloor; the receivers are magnetometers measuring the azimuthal magnetic field generated either by an essentially DC current flow in the transmitter or a transmission at frequencies high enough to induce secondary currents in the seafloor.

After the initial successful, but limited, experiment in 1980 by Young and Cox (1981) the Scripps group developed the forward modelling theory for the seafloor 1D Earth (Chave and Cox, 1982; Chave, 1983a) and a more sensitive receiver (Webb *et al.*, 1985). In 1984 an experiment was conducted over 25 My old seafloor in the northeastern Pacific in which source-receiver spacings of 60 km were achieved (Cox *et al.*, 1986). This experiment showed that the bulk resistivity of the crust was about 1000 Ωm, in good accordance with borehole measurements of Becker *et al.* (1982), and that the upper mantle resistivity was 10^5 Ωm or greater. A second experiment recently extended the maximum spacing to 90 km in an older (45 My) area of oceanic crust, and qualitatively supported the earlier result, since frequencies as high as 24 Hz were again detected at the longest range. Such a low conductivity for the upper mantle is a little surprising, but reasonable for a model in which the mantle is swept free of volatiles during the formation of crust at mid-ocean ridges and then cooled away from the ridges, with very little water circulation from the crust into the mantle.

The vertical electric bipole method has been called MOSES by Edwards *et al.* (1985). The theory for the method is contained in Edwards *et al.* (1981) and Edwards *et al.* (1984). It is essentially a DC method, generating apparent resistivity curves similar to those of resistivity soundings, but Edwards *et al.* showed that if the frequency is made high enough to cause induction in the seafloor, the coefficient of anisotropy as well as the average resistivity can be obtained. The first experiment using this method was carried out in 640 m of water in Bute Inlet, British Columbia, Canada (Edwards *et al.*, 1985). Source-receiver separations up to 2000 m were achieved, and a geologically reasonable model of 560 m of 1.9 Ωm of sediment underlain by relatively resistive basement was produced to fit the data.

The second MOSES deployment was a novel experiment near a polymetallic sulphide deposit on the Juan de Fuca ridge. Rather than suspending the bipole from a ship, a short (100 m) bipole was floated above a seafloor transmitter unit. A second seafloor unit was moved to ranges of 30 and 85 m by the submersible

Alvin before other experimental priorities halted the exercise. The data yield half-space resistivities of 13 and 20 Ωm; in excellent agreement with established values for seafloor basalt. It is unfortunate that this experiment could not be completed and data collected over the sulphide deposit.

In what may be regarded as a related experiment, Francis (1985) deployed a short (50 m) Wenner resistivity array from a submersible over a sulphide deposit on the East Pacific Rise. Resistivities of around 10 Ωm for the pillow basalts and as low as 0.17 Ωm for the sulphide deposit were recorded. Although the experiment was extremely simple, these represent the only *in situ* conductivity measurements of seafloor sulphides, and indicate what a useful tool resistivity mapping will be if these deposits are to be exploited.

Having suspended a transmitter from a ship and floated a second from within a hydrothermal vent field, it was clear that the next exercise for the Toronto/PGC group would be to suspend one from a hole drilled through an ice sheet. This was done through ice covering 18 m of water in the Beaufort Sea, in order to map the electrically resistive permafrost zone which lies beneath the unfrozen seafloor sediments (Edwards *et al.*, 1988). Transmitter-receiver separations of 20–200 m were obtained. The data did indeed show evidence of a buried zone at a depth of 15 m or so with a resistivity an order of magnitude greater than the sediments. An attempt was made to study anisotropy by operating at a second, higher (39 Hz) frequency, but the researchers were thwarted by noise.

The Toronto school has been prolific in the generation of background theory for various aspects of underwater controlled source EM sounding, including the response to transient systems and extension to simple 2D models (Edwards and Chave, 1986; Edwards and Cheesman, 1987; Cheesman *et al.*, 1987; Edwards, 1988a, b; Everett and Edwards, 1989). Most of this theory is directed to a completely new area of marine research; that of exploring the continental shelves and mid-ocean ridges for natural resources using EM methods. Although early attempts have been made using DC methods and a recent dipole–dipole IP system was used in search of seafloor mineral sand deposits (Wynn, 1988) both Toronto/PGC and Scripps have recently tested inductive systems, using different geometries. The electric dipole-dipole system that works so well in deep water can be modified to operate in shallow water in search of shallower targets by increasing the frequency of operation to hundreds of hertz (Constable *et al.*, 1986). A magnetic dipole-dipole system working in the time domain was operated by Cheesman *et al.* (1988).

5. Oceanographic Studies

5.1. Introduction

There has been an increased usage of electromagnetic instruments for oceanographic purposes. Measurement of the seafloor horizontal electric field gives a measure of (orthogonal) horizontal water motion, based on the simple principle that motion of velocity $\mathbf{v}$ through a magnetic field $\mathbf{B}$ produces an electric field $\mathbf{E} = \mathbf{v} \times \mathbf{B}$. One of the great advantages of this method is that it is mainly sensitive to the barotropic (depth independent) component of water flow. Traditional oceanographic techniques must employ current meters distributed throughout the ocean depth to separate the barotropic and baroclinic components. Recent treatments of the theory of oceanographic induction are given by Chave and Cox, (1983), Chave (1983b) and Chave and Filloux (1984), and Filloux (1987) presents a good overview of the method and describes some of the instrumentation involved.

There are two basic types of motional induction experiments which have been performed. In the first type the voltage across a submarine cable is recorded. Under favourable circumstances the voltage is a measure of the total water transport above the cable. Often the cable spans a channel through which the seawater is constrained to flow. If the seafloor is flat and there are no lateral variations in seabed conductivity, the potential across the cable is a linear integral of the water motion. That is, the voltage is proportional to the total water transport regardless of the spatial distribution of water flow. However, if the seafloor is not flat, or the seabottom conductivity varies laterally, the integral is not linear, and the voltage across the cable will vary if the distribution of water flow varies, even if the total transport stays the same. Thus meandering of a water current can corrupt the cable voltages as a measure of transport, and is responsible for several failed experiments. However, Sanford (1982) demonstrated that measurements of voltage across a cable spanning the Florida Straits were consistent with known total water transport and seasonal variation, probably because of little lateral motion in the pattern of water currents.

The second type of experiment utilises seafloor electric field recorders, in which the voltage is measured across antennae of only a few metres span, effectively giving a point measurement of the electric field rather than the integral along a cable. The electric field is proportional to the barotropic flow above the instrument, with a lateral resolution comparable to the water depth. By deploying an array of seafloor recorders, the spatial as well as temporal variations in water transport may be monitored. One of the earliest demonstrations of the viability of electric field on the deep seafloor as a measure of water transport was made by Cox *et al.* (1980).

In both the above methods, the electric field is reduced if significant return currents flow through a conductive seafloor. This need not affect the viability of the method if the seafloor conductivity is 1D, as the electric field or voltage for a resistive seafloor (sometimes called the 'open circuit' voltage) is simply scaled by a constant less than or equal to one. The usual practice is to estimate this constant by making measurements of water velocity using current meters operated simultaneously with the electrical measurements.

5.2. Recent activity

Larsen and Sanford (1985) estimated water transport of the Florida Current using both velocity profiling and potentials measured across a submarine cable. The agreement between the methods is excellent, and the nearly 2 y of continuous electric field data demonstrate that temporal variations in water transport can be studied. Such continuous observations would be almost impossible using conventional profiling. Baines and Bell (1987) attempted to interpret voltages measured across the Tasman Sea between Australia and New Zealand, but found no systematic variation with transport. The data series was correlated with sea level in Sydney, Australia, and is probably mainly sensitive to local currents associated with eddies near the Australian end of the cable.

The motional fields recorded by the EMSLAB experiment are presented by Chave *et al.* (1989). It is shown that the magnetic fields are mainly ionospheric in origin for periods below 10 days, but that the electric fields are contaminated (from a GDS and MT point of view) by oceanographic signals beyond periods of 2–4 days. One of the interesting features reported is a southward propagating wave attached to the topography of the Juan de Fuca Ridge.

Oceanographic aspects of the Tasman project are the subject of papers by Mulhearn *et al.* (1986), Bindoff *et al.* (1986), and Lilley *et al.* (1986, 1989). In particular, they document the seafloor response of a warm-core ring generated by the southward-flowing eastern Australian current.

Early results from the BEMPEX experiment in the North Pacific (personal communication with Chave and Luther) indicate that there is negligible electric current leaking into seafloor, as predicted by the controlled source observations of the crust and upper mantle. A resistive seafloor is desirable on two counts. The lateral resolution of an ocean bottom measurement will be improved, and the calibration of the experiment with data from current meter moorings become less critical. This latter point can be very important when one considers how notoriously unreliable current meters are at measuring water motions which are small but not necessarily atypical or of insignificant contribution to total transport. However, an interesting reversal of the situation occurs when one considers cable measurements, which monitor an integral of the electric field. Spain and Sanford (1987) show that a large but uniform seafloor conductivity can lessen the detrimental effects of an irregular seabed on cable measurements, making it desirable to have a very conductive seafloor if the water currents meander in an irregular channel.

In an unusual application of motional E-field studies, Korotayev *et al.* (1986) estimated water velocities associated with spring water flow from a fracture in the Black Sea, by towing a 1500 m electric bipole through the water. In another Black Sea study, Korotayev *et al.* (1985) used the leakage of motionally induced currents to estimate the conductivity-thickness product of the seafloor sediments, as well as collecting enough magnetic field data to compute an MT response for the area.

We have only considered the case where the water moves past the sensor. If the electric field sensor also moves within the Earth's magnetic field, this motion is also the cause of an electric field. Webb and Cox (1982) showed how horizontal electric field antennae could detect seismic waves in the frequency range 0.05 to 1 Hz, and papers by Webb and Cox (1984) and Webb and Constable (1986) presented data to support this method. In this frequency band the technique has an advantage over accelerometers in that the antenna is better coupled to the seafloor, and unlike measurements of pressure the E-field method is sensitive to horizontal motion of the seafloor.

6. Discussion

6.1. THE RESISTIVE LITHOSPHERIC MANTLE

The high resistance of mature lithosphere determined from controlled source studies has wide implications for seafloor experiments. In order for currents induced in the ocean to be able to leak through the resistive region into the conductive mantle below, a horizontal scale or compensation distance of $L = \sqrt{ST}$ is required (Cox, 1980), where S is the longitudinal conductance of the ocean and T is the transverse resistance of the lithosphere. From controlled source studies $T = 2.5 \times 10^9\ \Omega m^2$, yielding 6000 km for the compensation distance. Singer *et al.* (1985) used a similar value for integrated lithospheric resistance ($7 \times 10^9\ \Omega m^2$) in their model of global induction and concluded that leakage through the lithosphere did occur, and that currents induced in the world ocean could go under, rather than around, continents. They did observe, however, that compensation distances of 4–5000 km were required. In a recent paper, Mackie *et al.* (1988) modelled the transverse resistance of the north-eastern Pacific oceanic crust-mantle to be $1 \times 10^9\ \Omega m^2$ from long period MT sounding on the adjacent land. This is in excellent agreement with the controlled source experiment, and also serves to illustrate the dangers of ignoring oceanic induction when analysing coastal MT data. Another estimate of the compensation distance was made by Lilley *et al.* (1989) from a plot of the anisotropy of seafloor MT sites as a function of distance from the SE Australian coast. Their data are fit very well by a compensation length of 400 km, but this implies a T of only $2 \times 10^7\ \Omega m^2$, 2 orders of magnitude smaller than observed in the other experiments. However, in the region of their experiment there is only about 800 km of oceanic seafloor between Australia and the continental crust of the Lord Howe Rise (and even this narrows to the north), and so the seafloor geometry is not ideal for a determination of this type.

For an MT experiment to be interpretable using the 1D approximation, electrical structure must be 1D over the compensation scale, or charge will build up on the ocean boundaries (the coastlines), reducing the electric field perpendicular to the coast. The result will be anisotropic sounding curves even when the seafloor is otherwise one-dimensional, with apparent resistivities being too low for the component derived using the E-field perpendicular to the coast. This imposes severe restrictions if scales of thousands of kilometres are involved. Motional induction studies, however, benefit from the fact that little leakage of induced currents into the lithosphere is occurring, and so measurements of the electric field accurately reflect horizontal water motion. This is evident in the early results of the BEMPEX experiment.

The horizontal scales of the controlled source experiments were only 60–100 km, and it is reasonable to suppose the current may leak into the conductive asthenosphere through paths at ridges, continental margins, fracture zones or transform faults. However, the study of Mackie *et al.* (1988) is responsive to a larger region, and in particular suggests that no great leakage is occuring at the continental margin. It should also be noted that the transverse resistance of the lithosphere is probably a function of temperature, and therefore age, of the upper mantle. Controlled source soundings have been on 25 My and 45 My old seafloor, but the EMSLAB site, for example, is much younger (0–10 My).

6.2. The electromagnetic study of mid-ocean ridges

There is a growing interest world wide in the study of mid-ocean ridge systems. To understand how ridges behave is to understand how 60% of the Earth's lithosphere is generated. Although many of the geophysical studies conducted in the past have concentrated on the various seismic techniques, and this trend will undoubtably extend into the future, electromagnetic methods have a great advantage in their sensitivity to the physical properties under consideration; namely temperature and fluid content. Changes which produce only subtle variations in acoustic velocities generate orders of magnitude changes in electrical conductivity.

At the mid-ocean spreading centres upwelling mantle partially melts as a result of pressure relief, and this melt then migrates and solidifies to create new oceanic crust. Little is known about the mechanisms by which the melt moves from the region of melting to crustal depths and then differentiates to form the cumulates, dikes, and pillow basalts which make up the crust. Many geometries have at some time or other been proposed for the crustal magma chamber which is assumed to form, at least at rapid spreading rates and possibly intermittently. Although some proportion of melt is required to produce the cumulates we see in ophiolite analogues, no one knows if the hypothesised chamber is fully or partially molten.

Detrick *et al.* (1987) present a picture of the east Pacific rise (EPR) between about 9° and 14° N, based on seismic imaging, in which a liquid upper surface of an axial magma chamber is a nearly continuous feature, about 2 km below the seafloor and with a width of 2–3 km. The walls and floor of the chamber cannot be resolved, but the Moho may be traced to within a few kilometers of the ridge axis, suggesting that the width of a magma chamber at this depth (about 6 km) is no more than 4–6 km. Expanding spread profiles suggest that only a narrow (about 1 km) section of the upper surface of the chamber has the unusually low velocities associated with mostly molten rock (Harding *et al.*, 1990), but that low velocities on the flanks are suggestive of a wider region of hot, mostly solid, rock. Thus we have a picture of a narrow (both laterally and vertically) melt accumulation at the top of what may be a larger volume of partially molten, or merely hot, material.

The top of the liquid magma chamber might thus be between 30% melt at a temperature of 1185 °C (Sleep, 1978) and 100% melt at a temperature of 1270° or so. The resistivity of tholeiitic melt under these conditions is between 0.5 Ωm

(1200 °C) and 0.25 Ωm (1300 °C) (Waff and Weill, 1975; Tyburczy and Waff, 1983). The dependence of conductivity on melt fraction is less clearly defined. Taking the mathematical model presented by Shankland and Waff (1977) which assumes melt connectivity, we modify the above figure to get about 2 Ωm for a 30% melt in the magma chamber. Thus we see that the resistivity of the magma chamber is likely to vary over one order of magnitude as the melt fraction varies from 30 to 100%.

Some models for the spreading center have demanded hydrothermal circulation through the crust to cool the lower sides of the magma chamber. The results of Becker *et al.* (1982) and Cox *et al.* (1986) show that older, cooler crust has a resistivity of around 1000 Ωm between about 1 and 6 km deep, indicating a porosity of 1% or so. It is probable that crust close to the ridge axis, thermally cracked and with pores not yet plugged with alteration minerals, will have greater porosities. However, one notes with interest that Caress (1989) presents a seismic tomography model for the EPR at 13° N which features anomalously fast velocities for the upper 500 m of the crust within a kilometre or two of the axis. This suggests that immediately above the magma chamber cooling has not yet produced significant cracking, or that cracking is quickly plugged by intrusives. If this is indeed the case, this low porosity cap will be evident in controlled source soundings. The potential for electromagnetic experiments to address the question of porosity variations is improved by saltwater conductivity not being as sensitive to temperature variations one might expect. The average temperature of the older crust is about 100 °C. Seawater has a maximum conductivity at 300–400 °C which is only double its conductivity at 100° (Quist and Marshall, 1968), so water saturated rock with porosity similar to that of the old crust, no matter how hot, should not have a resistivity much above 500 Ωm (assuming, of course, that it does not become hot enough for silicate conduction to dominate).

Extrusive lavas at the very top of the crust are much less resistive, about 10 Ωm based on the measurements of Francis (1985) and Becker *et al.* (1982) as well as reasonable estimates at porosity (10–15%) and fluid conductivity. Primitive magma feeding any magma chamber is probably not much hotter than the magma chamber. The different chemistry of this material will not alter its conductivity significantly from that of the tholeiitic melt described above (Waff and Weill, 1975). Estimates of bulk conductivity for this region obtained from EM studies are admirably suited to discriminate between porous flow models and models of dike or conduit injection. For the 2% porosities required by the porous flow models, Shankland and Waff's model predicts resistivities of 10 to 20 Ωm. Olivine resistivity at 1300 °C is at least 100 Ωm, so assuming an episodic injection model, or a system of cracks which are not interconnected all the time, this higher resistivity would be observed.

Both controlled source and GDS methods may prove useful in discriminating between the various tectono-physical models. The controlled source experiment of Cox *et al.* (1986) on older crust was sensitive to mantle conductivity to at least 20 km depth. As argued above and also indicated by the Young and Cox (1981) experiment, the conductivity of the crust near the ridge is not likely to be more than a few times more conductive than at older sites, so similar penetration depths are possible a few 10's of kilometres from the ridge. GDS experiments are admirably suited to mapping lateral variations in conductivity, and although they have no resolution in the resistive crust and upper mantle, conductive melt regions make an ideal target. GDS and MT studies to date, however, have failed to detect any continuous conductive zone along a ridge (Filloux and Tarits, 1986; Filloux *et al.*, 1989; Hamano *et al.* 1989).

It could be very important for the purposes of a ridge GDS or MT study to know where the resistive upper mantle layer terminates. Currents channelled along the surface of this layer and then penetrating the mantle at a ridge may have a marked effect on the electromagnetic signature; the anomalous electric field perpendicular to the EPR seen by Filloux and Tarits (1986) might be such an effect.

Of as much interest as the fluids in the ridge system is the off-axis temperature distribution, both in the crust and the upper mantle. Our knowledge of the conductivity of saltwater and olivine as a function of temperature should allow suitably placed controlled source soundings to give an indication of the rate of cooling within the lower crust/upper mantle, which in turn will indicate the extent of hydrothermal circulation.

In June of 1989 a joint Cambridge/Scripps expedition to the EPR at 13° N collected controlled source data for several days, using a specially constructed (by Cambridge) transmitting antenna designed to be neutrally bouyant and flown a few tens of metres above the rugged seafloor. At the time of writing the data had only been examined to the extent that the operation of the transmitter and seven receivers was verified, but a few simple 1D models illustrate the sensitivity of the dipole–dipole method to hypothetical ridge structures. The basic model we have taken consists of an ocean 2.5 km deep, a layer of extrusives on the seafloor which is 500 m thick and has a resistivity of 10 Ωm, beneath this a crustal layer of intrusives and gabbros which is 5.5 km thick and 250 Ωm, and finally a mantle region of 100 Ωm (olivine at 1300 °C).

Figure 3(a) shows the response of this basic model (circles) at a source-receiver spacing of 40 km, and a second model in which the mantle is replaced with partially molten material of 10 Ωm resistivity (diamonds). The broken line represents an estimated noise level for the Scripps controlled source system as deployed in the 1984 experiment. (The solid lines are merely smooth curves joining the actual computations of the electric field denoted by the symbols). Although 40 km is the largest range at which the signal is above the noise, this long range presents an interesting picture of EM propagation. At a frequency of 0.25 Hz we observe that the model with the more conductive mantle actually results in a *larger* signal. This is a consequence of the resistive, crustal, layer acting in a manner which may be loosely called a waveguide. At certain frequencies the efficiency of this waveguide is improved by the greater conductivity contrast presented by the more conductive mantle, and so a larger signal is seen. At frequencies lower than 0.25 Hz we see the effect of energy being lost more readily to the partially molten mantle material of the second model, and so depressing the observed signal.

Figure 3(b) presents the situation closer to the ridge where the lower crust may still be above 1000 °C and so below about 100 Ωm. A source-receiver spacing of 20 km is considered. The curves show the response of such an isotherm at a depth of 2.5 km (circles) and 4.5 km (diamonds). Finally, Figure 3(c) shows a sounding

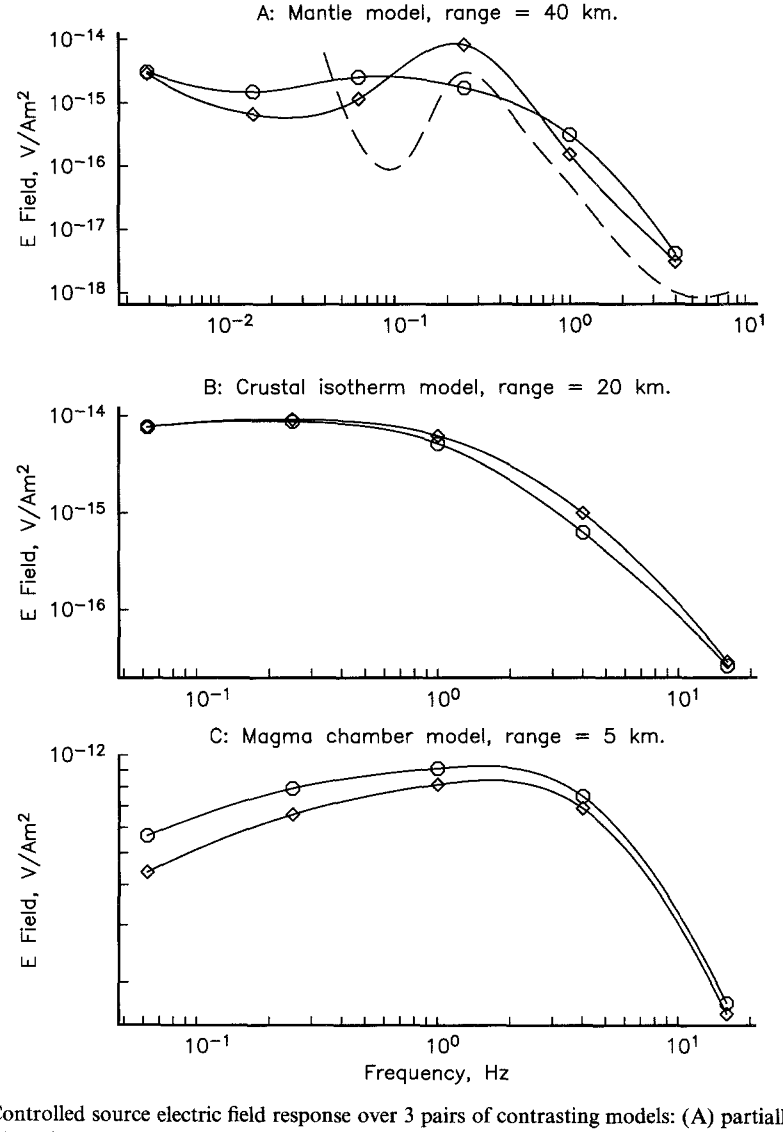

Fig. 3. Controlled source electric field response over 3 pairs of contrasting models: (A) partially molten (diamonds) and merely hot (circles) upper mantle, (B) a crustal 1000° isotherm at depths of 2.5 km (circles) and 4.5 km (diamonds), and (C) a crustal magma chamber which is fully (diamonds) or 30% (circles) molten. The broken line in (A) represents a typical noise level for the Scripps sounding equipment. The notch in the noise at 0.1 Hz is a permanent part of the natural spectrum.

immediately over the ridge axis with the source-receiver spacing reduced to 5 km. The two models purport to show a magma chamber at a depth of 2 km containing partially molten (2 Ωm, circles) and fully molten (0.3 Ωm, diamonds) rock.

Although the 1D models considered above provide an indication of the sensitivity of the controlled source method to ridge structures, they are clearly inadequate for quantitative modelling of what is obviously not a 1D structure.

Ridges are good candidates for 2D modelling, and indeed one is struck by how remarkably 2D they are, with hundreds of kilometres between significant offsets. While 2D MT modelling is becoming commonplace, the difficulty associated with controlled source modelling is that the source-field is 3D. However, progress is being made, and forward models of ridges in which the approximation that the source is 2D have been presented by Everett and Edwards (1989) for time-domain sounding and Flosadottir and Cox (1989) for frequency-domain sounding. The models of Everett and Edwards show that the maximum response to a crustal magma chamber for a time-domain system occurs at delay times of about 1 s.

6.3. Future Progress

We are seeing diminishing returns from the deployment of single-station MT sites on the seafloor. There is little prospect of significantly expanding the limited frequency band available between long period contamination by oceanic effects and short period cutoff due to screening of the source field, and so resolution is necessarily limited to a narrow depth range in the upper mantle. On the other hand, the expanding use of long period E-field instruments for the study of oceanography is likely to support the continued deployment of seafloor EM arrays, from which we shall continue to extract MT and GDS data. Magnetic measurements are much easier to collect than electric field data, which to date have required either sophisticated water choppers or long antennae, but GDS offers possibilities in the discrimination of lateral changes in structure across features like ridges; thus one expects to see expansion in this area. Already there is a lot of international cooperation, reflecting the fact that once a ship is available for an experiment, more instruments can be deployed than are usually available to one institution. There will be reduced motivation for one facility to maintain a large suite of instruments if this continues.

Controlled source experiments are more difficult (and so more costly and with a higher instrument loss) than GDS and MT studies, but offer the chance to study the first 50 km or so which is invisible to MT sounding. Now that ground has been broken in this area, we will see an expanding use of this method. Although the oil and mining industries are currently depressed world-wide, there is academic activity in developing EM exploration techniques for the seafloor which might well blossom during the next rejuvenation of the industry, whenever that may be.

A scientific/political emphasis on exploring mid-ocean ridge environments has improved the prospects for a vigorous programme of EM experiments in the intermediate future, and an EM component is being considered important by many of the groups coordinating ridge studies. However, these are not easy areas to work in and represent challenges to the various disciplines; problems include conductive material (controlled source method), dimensionality problems (MT), and topographical inhospitability (all methods). We must be cautious not to promise more than we can achieve.

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