GEOPHYSICS, VOL. 63, NO. 3 (MAY-JUNE 1998); P. 816–825, 10 FIGS.

Marine magnetotellurics for petroleum exploration
Part I: A sea-floor equipment system

Steven C. Constable*, Arnold S. Orange‡, G. Michael Hoversten**, and H. Frank Morrison**

ABSTRACT

Induction in electrically conductive seawater attenuates the magnetotelluric (MT) fields and, coupled with a minimum around 1 Hz in the natural magnetic field spectrum, leads to a dramatic loss of electric and magnetic field power on the sea floor at periods shorter than 1000 s. For this reason the marine MT method traditionally has been used only at periods of 10^3 to 10^5 s to probe deep mantle structure; rarely does a sea-floor MT response extend to a 100-s period. To be useful for mapping continental shelf structure at depths relevant to petroleum exploration, however, MT measurements need to be made at periods between 1 and 1000 s. This can be accomplished using ac-coupled sensors, induction coils for the magnetic field, and an electric field amplifier developed for marine controlled-source applications. The electrically quiet sea floor allows the attenuated electric field to be amplified greatly before recording; in deep (1-km) water, motional noise in magnetic field sensors appears not to be a problem. In shallower water, motional noise does degrade the magnetic measurement, but sea-floor magnetic records can be replaced by land recordings, producing an effective sea-surface MT response. Field trials of such equipment in 1-km-deep water produced good-quality MT responses at periods of 3 to 1000 s; in shallower water, responses to a few hertz can be obtained. Using an autonomous sea-floor data logger developed at Scripps Institution of Oceanography, marine surveys of 50 to 100 sites are feasible.

INTRODUCTION

Originally proposed in Tikhonov (1950) and Cagniard (1953), the magnetotelluric (MT) method has been used to map sedimentary structure as an aid to petroleum exploration for several decades (e.g. Vozoff, 1972; Orange, 1989). The essence of the MT method is the computation of an electromagnetic earth impedance from measurements of orthogonal horizontal magnetic and electric fields at the surface. Estimates of impedance magnitude (transformed to an apparent resistivity) and phase at various frequencies allow investigation of electrical conductivity as a function of depth. Impedance measured at several locations allows investigation of conductivity as a function of horizontal position.

The reliability and usefulness of the MT method has improved greatly over the past few years as a result of progress in several areas. These include improvements in data acquisition technology (e.g., Nichols et al., 1988), the introduction of a remote reference to reduce bias associated with noise in the magnetic field measurements (Gamble et al., 1979), robust response function estimation methods (e.g., Egbert and Booker, 1986), and improved 1-D, 2-D, and even 3-D forward and inverse modeling codes (e.g., Wannamaker et al., 1986; Constable et al., 1987; Smith and Booker, 1991). As a consequence, the MT method represents an important nonseismic exploration tool, particularly for reconnaisance surveys and in areas where the seismic reflection method performs poorly. The latter include buried salt, carbonate, and volcanic horizons that efficiently reflect and scatter acoustic energy. In a companion paper (Hoversten et al., 1998 this issue), 2-D and 3-D modeling demonstrate the utility of the MT method in delineating subsalt structure.

Many petroleum prospects are offshore; but because seawater attenuates the magnetic source field at frequencies above

Manuscript received by the Editor May 31, 1996; revised manuscript received April 15, 1997.
*Scripps Institution of Oceanography, IGPP 0225, 8604 La Jolla Shores Drive, La Jolla, CA 92093-0225. E-mail: sconstable@ucsd.edu.
‡Arnold Orange Associates, 8806 Point West Drive, Austin, TX 78759. E-mail: 73321.3045@compuserve.com.
**Lawrence Berkeley Laboratory, University of California, 1 Cyclotron Rd., Bldg. 90, Rm 1116, Berkeley, CA 94270. E-mail: mhovers@socrates.berkeley.edu.; hfmengeo@socrates.berkeley.edu.

0.01 Hz, the marine MT method traditionally has been considered sensitive only to great depths (e.g., reviews by Law, 1983; Constable, 1990; Palshin, 1996). Indeed, the focus of marine MT instrumentation has been drift-free measurement of electromagnetic (EM) fields at essentially dc frequencies by the use of torsion fiber and fluxgate magnetometers and water-chopped electric field sensors (e.g., Filloux, 1987). Deployed mainly on the deep sea floor to study mantle conductivity structure, these instruments rarely provide responses at periods shorter than a few hundred seconds.

Investigations of shallow conductivity have been undertaken using a controlled-source method, replacing the lost power at high frequencies using a man-made transmitter positioned on or close to the sea floor (e.g., Constable and Cox, 1996; Edwards et al., 1981). However, the use of a marine controlled-source method is technologically challenging, and the method favors the more resistive hard-rock sea floor of the deep ocean over the conductive sediments of petroleum targets on the continental shelf. Modeling a 3-D source field also presents a greater difficulty than modeling the MT plane-wave source.

A companion paper (Hoversten et al., 1998, this issue) shows that data in the frequency band 0.001–1 Hz are sensitive to structures typical of those encountered in the exploration for oil and gas. Our paper demonstrates that if instrumentation and processing are optimized for this frequency band, then the marine MT method is indeed viable. This work resulted from field trials conducted to provide the petroleum industry with a usable offshore natural-source electromagnetic exploration method (Constable et al., 1994).

SEA-FLOOR MAGNETOTELLURIC FIELDS

Simple 1-D theory demonstrates that for downward-propagating energy, electric and magnetic fields measured on the sea floor provide an MT impedance for the subsea section that is independent of the overlying conductive layer. It appears, then, that one may neglect the effect of the sea-water layer and approach sea-floor MT just as for land MT. However, although the field ratio is unchanged, the fields themselves will be attenuated by induction in the seawater layer, so the instrumental and logistical impact of smaller fields must be considered.

The decay of external electric and magnetic fields through seawater depends both on attenuation in seawater and on the reflection coefficient of the sea floor and so cannot be estimated in detail prior to carrying out an electromagnetic survey. However, general predictions can be made. To model the effect of the seawater layer, we use the theory developed by Schmucker (1970, 61–65) for a layered (1-D) structure, in which N layers are numbered downward starting at layer one (the surface layer). The MT ratio of electric field E to magnetic induction B at the surface of this model can be written as

$$\frac{E}{B} = \frac{i\omega}{k_1 G_1}.$$

The complex skin depth or propagation constant k for zero wavenumber (plane waves) is given for each layer of conductivity σ_j at a frequency f by

$$k_j = \sqrt{\omega\mu_o i\sigma_j} = \sqrt{2\pi f\mu_o i\sigma_j}\,, \qquad j = 1,\ldots N.$$

G is a dimensionless quantity defined for the top of each layer of thickness h_j and can be obtained from the recurrence relation

$$G_j = \frac{k_{j+1}G_{j+1} + k_j \tanh(k_j h_j)}{k_j + k_{j+1}G_{j+1}\tanh(k_j h_j)},$$

started by setting $G_N = 1$. (The familiar MT apparent resistivity and phase are related to G by $\rho_a = \omega\mu_o/|k_1G_1|^2$ and $\phi = \arctan[\mathrm{Imag}(k_1G_1)/\mathrm{Real}(k_1G_1)]$.)

Schmucker's formulation allows one to write the ratio of the electric field at the bottom of any layer E_{j+1} to the electric field at the top of that layer E_j as

$$\frac{E_{j+1}}{E_j} = \cosh(k_j h_j) - G_i \sinh(k_j h_j)$$

and the magnetic field ratio as

$$\frac{B_{j+1}}{B_j} = \frac{k_{k+1}G_{j+1}}{k_j G_j}(\cosh(k_j h_j) - G_j \sinh(k_j h_j)).$$

Figure 1 presents an example of the attenuation through 100 m and 1000 m of seawater over half-spaces of 1 and 100 Ωm resistivity. The effect of the resistive sea floor is to enhance slightly the magnitude of the electric fields over their half-space values. The effect of the sea floor on the magnetic field, on the other hand, is much larger, and increasing the resistivity of the sea floor greatly attenuates the fields over their half-space values.

The impact on sea-floor MT measurements is clear: for a moderately resistive seabed, the magnetic field is attenuated one to two orders of magnitude across the usable MT band. Fortunately, the resistivity of sea-floor sediments is unlikely to exceed 100 Ωm. The sea-floor electric field shows no attenuation—even an enhancement—but short-period sea-surface electric fields are already 17 times smaller than they would be over a 100-Ωm half-space because of the highly conductive seawater. The effect of a seawater layer gives rise to quite different field behavior than found on land. The electric field varies significantly from place to place on land as the subsurface resistivity changes, while the magnetic field varies much less (and not at all for 1-D structures). On the sea floor, the magnetic field varies markedly with subsurface structure, and both the magnetic and electric fields contain information on structure, even for the 1-D case. A companion paper (Hoversten et al., 1998 this issue) quantifies this effect for 2-D structures and considers the effects of vertical magnetic fields and currents associated with lateral conductivity contrasts.

INSTRUMENTATION

Marine MT studies have traditionally used fluxgate or torsion fiber magnetometer sensors and dc coupled electric field sensors (e.g., Filloux, 1987), both because upper mantle structure has been a target of interest and because generally it was considered that sea-floor fields below periods of 100–1000 s were too small to measure. On land, however, MT exploration for shallower targets is conducted with induction coils that respond to the time derivative of the magnetic field. This ac coupling removes the 50 000-nT main component of the field; but because the natural field spectrum is very red at frequencies below 1 Hz, coils can be used effectively to periods in excess of 1000 s. We use the BF-4 magnetometer coil manufactured by Electromagnetic Instruments Inc. (EMI), which has a frequency response shown in Figure 2.

It follows that ac coupling is also appropriate for measuring electric fields, and the amplifier described by Webb et al. (1985), developed for sea-floor controlled-source applications, is well suited to sea-floor MT. This amplifier is designed to take advantage of the low noise and low impedance of the marine silver-silver chloride electrodes described by Webb et al. (1985). After a large (20 000-μF) series coupling capacitor, the first stage is a power field effect transistor bridge, which chops the signal at 2 kHz. The second stage is a 30:1 step-up transformer, which provides transformer isolation for the electrodes and impedance matching prior to further amplification. Only after the transformer stage does conventional amplification using active components occur. The amplifier has a total gain of 10^6 and a noise level lower than Johnson noise from the 6-Ω electrode impedance. Although the two-pole, low-frequency roll off of this original amplifier is somewhat severe (Figure 2), a new amplifier developed specifically for MT applications with a higher input impedance (2700 Ω versus 58 Ω for the original version) has a low-frequency roll off very similar to that of the BF-4 coils, at the cost of slightly higher amplifier noise.

The MT method requires time series of electric and magnetic fields that are many times longer than the longest period of interest; a typical recording time of 10 hours would be used to obtain reliable MT responses at 1000-s periods or greater. It is impractical to moor a ship for this length of time and somewhat difficult to do in deep water, so we chose autonomous

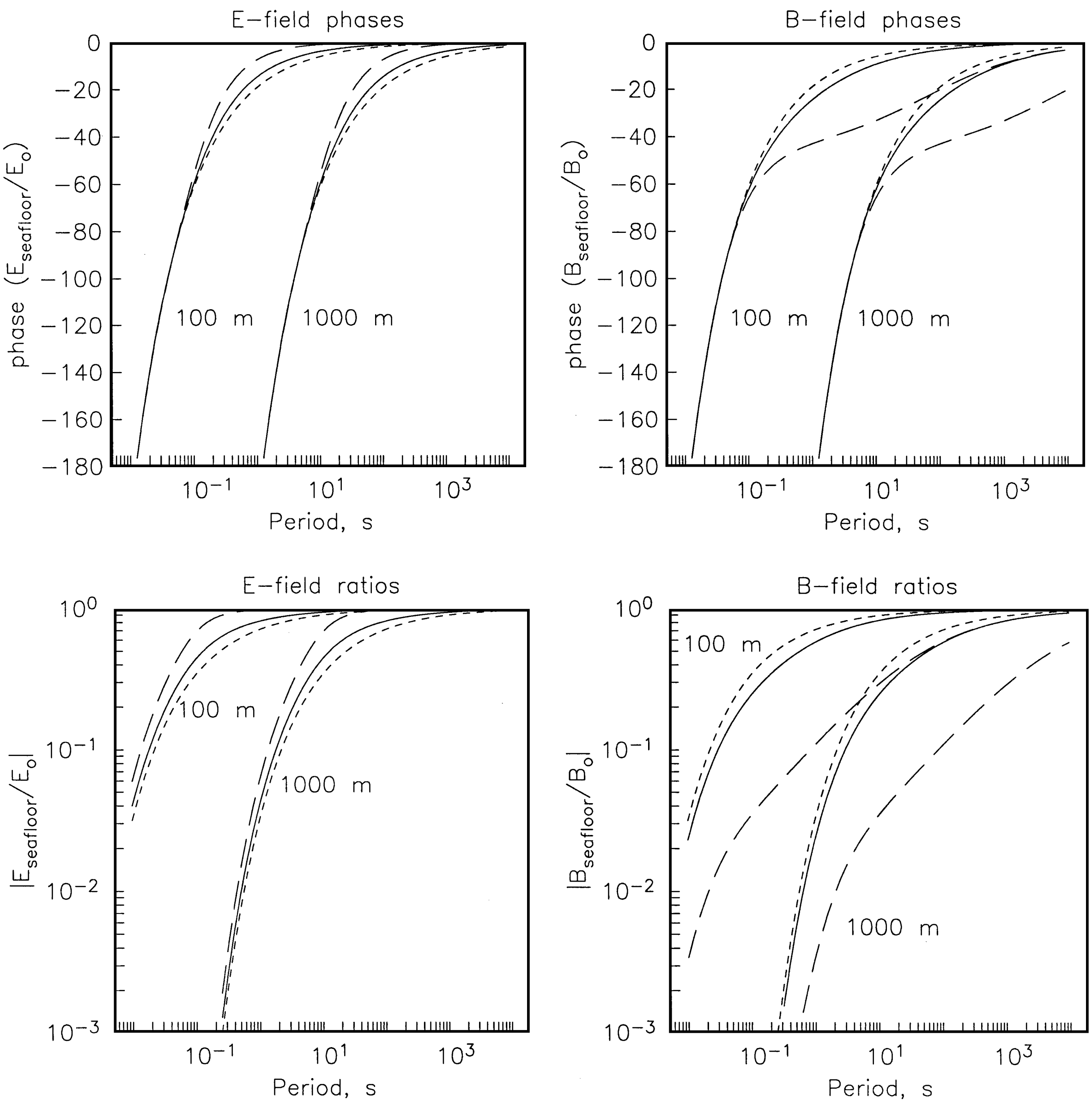

FIG. 1. Ratios and phase leads of sea-floor electric and magnetic fields to sea-surface fields. Solid lines are for a 1-Ωm sea floor, coarsely broken lines are for a 100-Ωm sea floor, and finely broken lines represent simple half-space attenuation (i.e., attenuation with a sea floor of seawater conductivity). Water depths of 100 m and 1 km are modeled.

data recorders rather than wireline instrumentation. Sea-floor data loggers have been developed for a variety of applications at Scripps Institution of Oceanography (SIO). Two generations of instrument have been tested for sea-floor MT: an older unit based on Onset Corporation's C44 bus and CPU-88 computer, described by Webb et al. (1985) and Constable and Cox (1996), and a newer unit based on Onset's Tattletale 8 (TT8) computer/logger. The capabilities of the two instruments are similar: 6–16 channels of 16-bit data, 256 Hz to 2 kHz maximum sampling rate, 4–40 Gbytes of data storage, up to 12 months' endurance, 1–10 ms/day crystal clock drift. The most significant difference is that the newer instrument is smaller, lighter, and thus better suited to operations where equipment is air freighted to the field area and instruments are deployed and redeployed daily.

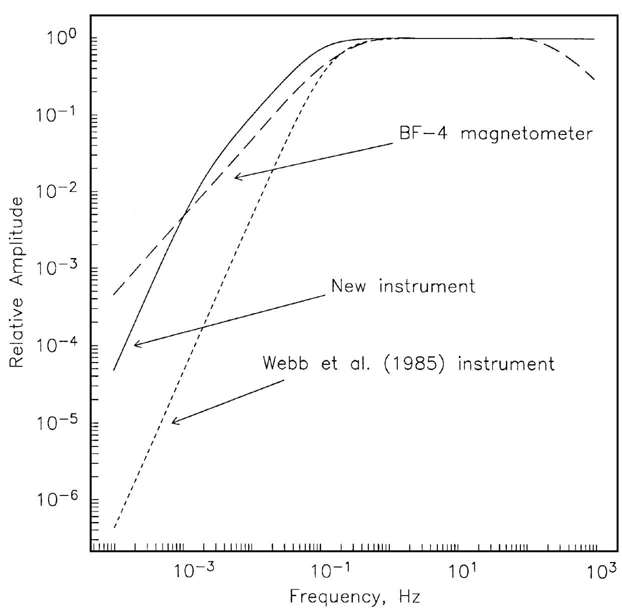

FIG. 2. Frequency response of the BF-4 induction coil and the older and newer versions of the sea-floor electric field amplifier described by Webb et al. (1985).

Figure 3 shows a block diagram of the second-generation data logging electronics, configured for four-component MT acquisition. The TT8 logger unit has been augmented by a small computer systems interface, allowing two disk drives of up to 10 Gbytes each to be installed. A software data compression algorithm allows about a factor of two reduction in data volume. An external crystal oscillator with a drift rate of about one in 10^8 provides timing accurate to about 1 ms during a one-day deployment, and the TT8's on-board 12-bit analog-to-digital (A/D) converter has been replaced by an external 16-bit A/D converter that can multiplex up to 16 channels of data.

The new logger was developed in conjunction with a marine seismology group at SIO, so attributes such as the massive data storage and highly accurate crystal clocks have been driven by the need to collect long-time series of seismic data. These features represent no disadvantage for MT data acquisition. Although still under some development to reduce power consumption, at a typical sample rate of 25 Hz/channel the logger consumes under 500 mW, allowing it to run for about a week from a pack of D-cell nickel-cadmium batteries. During this time the logger collects 120 Mbytes of data, easily accommodated on a modern disk drive.

Figure 4a shows a unit being deployed from a 40-m-long ship, and Figure 4b shows a drawing of the same instrument with components labelled. Logger electronics reside in a 15-cm-ID 7075-T6 aluminum tube, anodized and painted to resist corrosion by seawater and terminated by two endcaps sealed with O-rings. One endcap has ports to start the computer and purge damp air from the instrument, and the other has high-pressure underwater connectors for linking the sensors to the logger inputs. The entire system is capable of resisting water pressure to depths of 6000 m. A second, smaller pressure case houses acoustic release and navigation electronics. The acoustic unit locates the instrument underwater and releases the package from the sea floor at the end of the recording period. Acoustic pings at around 12 kHz frequency and 20 ms duration are generated by a ship-board system, and the sea-floor unit replies with a similar ping immediately on receiving the outgoing signal. The two-way traveltime can then be used to determine the range of the instrument from the ship, allowing sea-floor locations accurate to a few meters to be determined

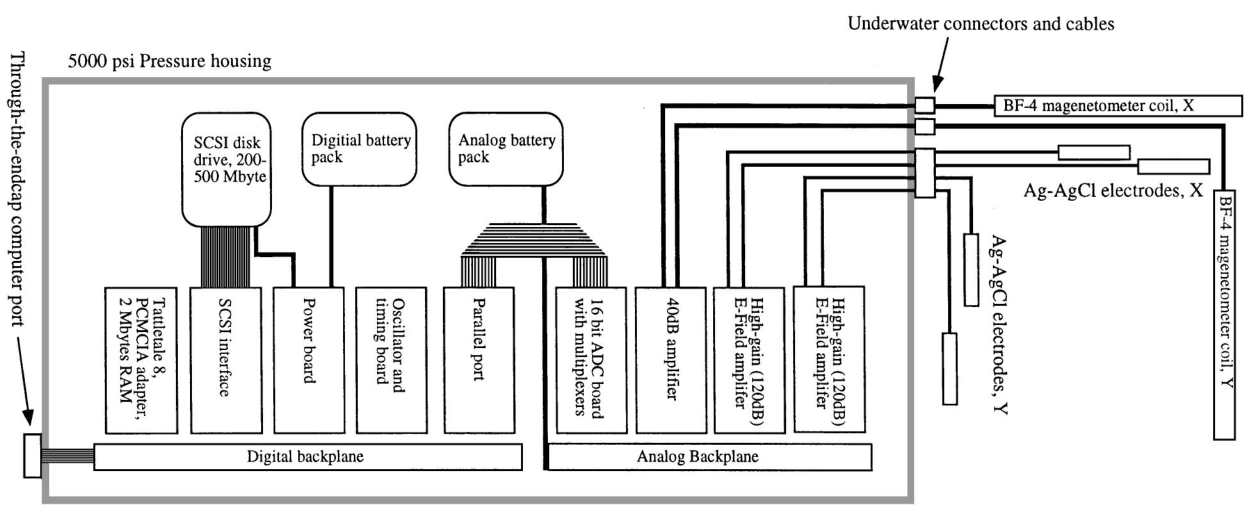

FIG. 3. Block diagram of SIO data logger, configured to collect four-component MT data.

from differential Global Positioning System (GPS) ship positions.

The logger and release pressure cases are supported in a polyethylene framework, which protects the instrument from damage during handling and supports the four glass flotation spheres, two magnetometer coils, four 5-m electrode arms, and concrete anchor. The Webb et al. (1985) instrument has a recording magnetic compass inside the logger pressure case. Not only is this undesirably close to magnetic components of the instrument, but also loading and developing photographic film becomes tedious when several instruments are deployed each day. For the new instrument an external compass

a)

b)

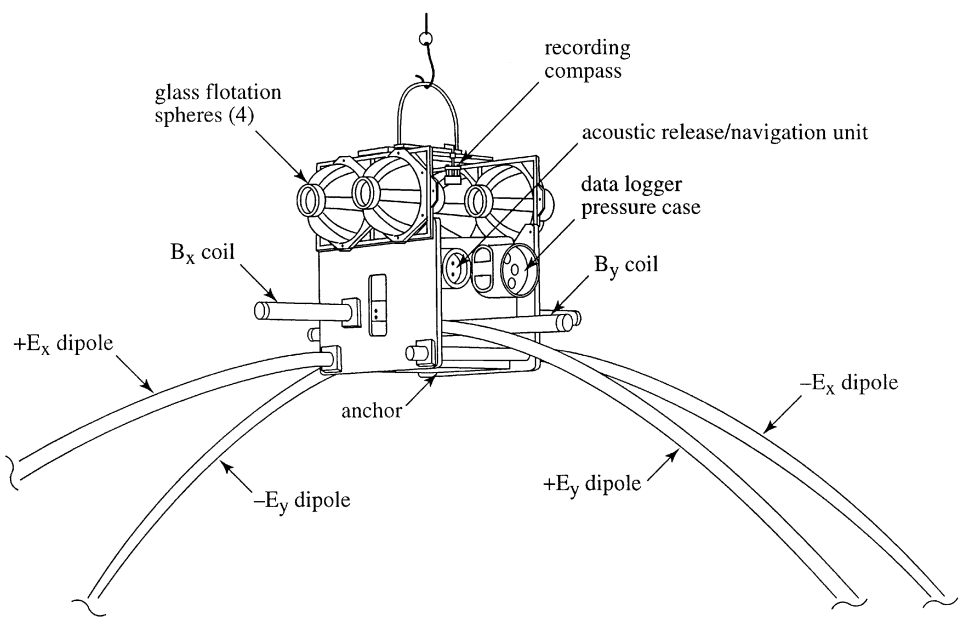

FIG. 4. (a) Sea floor MT instrument being deployed from a 40-m-long ship. (b) Line drawing of sea-floor MT instrument. A 60-kg concrete anchor held beneath the center of the package sinks the device to the sea floor. The anchor is released by the acoustic unit on receipt of a command code, and the device rises to the surface with the help of the glass flotation spheres. The electric dipole arms are 5-m lengths of 5-cm-diameter polypropylene pipes terminated with silver–silver chloride electrodes. Dipole cables run along the insides of the tubing.

equipped with a timed release based on dissolving sugar crystals is used. About one hour after deployment, this device locks a compass needle mechanically to record the orientation of the system on the sea floor.

Each acoustic unit also responds to three different coded sequences of timed pings and, on command, will go quiet, wake up, or release the instrument. During marine operation, a 60-kg (air weight) concrete anchor is attached to the plastic frame by means of a short, nylon-insulated stainless steel wire (burn wire). The insulation is cut to bare a 2-mm section of wire, and the wire is electrically connected to the release unit. On receipt of the release command, the acoustic unit supplies the cut wire with +18 V from internal batteries. Within about 15 minutes, this positive voltage causes the steel wire to electrolyze away, releasing the anchor from the instrument and allowing the positively buoyant package to float to the surface for recovery.

Remote reference processing requires that data at the base and reference sites be acquired simultaneously with synchronized acquisition systems. More importantly, processing sea-floor data with land magnetometer components requires precise timing; phases accurate to 5° at a frequency of 2 Hz require timing to 7 ms or better. Since the sea-floor instruments are autonomous and beyond the reach of radio communication, accurate timing must be accomplished by on-board quartz clocks. The on-board clocks are started using a GPS time standard with initial timing accurate to 1–10 μs. After recovery, clocks are again checked against the GPS standard to estimate drift or error. Drift rates are typically less than 1 ms/day.

OFFSHORE TESTING

Although sea-floor electric fields are very small compared to their terrestrial counterparts, the sea floor is a benign and quiet environment for measuring electric fields. By contrast, land E-field measurements are plagued with all manner of difficulty: 60- or 50-Hz power, active corrosion protection systems, stray currents from electric trains and other industrial infrastructure, large temperature variations, streaming potentials in soils, mineral redox reactions, and variations in soil dampness and permeability.

Figure 5 presents an estimate of electrode, amplifier, and instrument noise obtained by taking the difference between the signals recorded on two 300-m-long parallel, adjacent, and identical electric field antennas. These data came from an instrument 1500 m deep in the North Atlantic, deployed as part of the controlled-source EM experiment reported by Sinha et al. (1997). Although electrode noise dominates the system below about 1 Hz, the flattening of the curve above this frequency is likely to be associated with noise from the recording system. For comparison, the laboratory results of Petiau and Dupis (1980) are shown. Even in the controlled environment of the laboratory, electrode noise is over two orders of magnitude worse in parts of the spectrum. This is explained by the isothermal and isosaline conditions on the sea floor, as well as the large (20-cm-long) electrodes used. Not only is Johnson noise minimized by using large, low-impedance electrodes, but there is evidence that larger electrodes are intrinsically quieter. The curves converge at low frequency, but in the field the factors cited above would severely degrade electrode performance over the laboratory values.

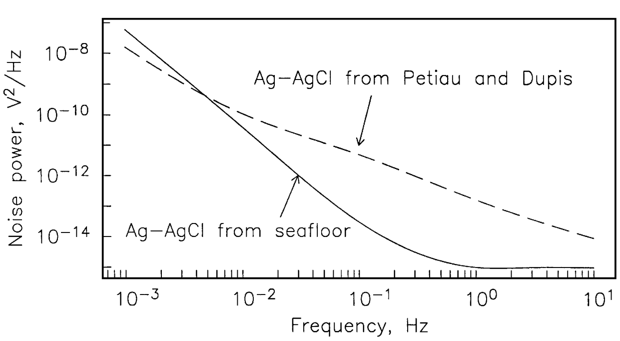

FIG. 5. Sea-floor electrode and instrument noise, obtained by spectral analysis of the difference between the signals on two parallel 300-m electric field antennas. For comparison, see laboratory measurements of Ag-AgCl electrode noise made by Petiau and Dupis (1980).

To evaluate the operation of the MT instrument system, we have tested it in various water depths off San Diego. The first test was conducted April 7–16, 1994, by fitting three orthogonal BF-4 to coils one of the original ELF-type SIO electric field recorders described by Webb et al. (1985) and Constable and Cox (1996). This instrument was deployed in 1000 m of water in the San Diego Trough, approximately 30 km offshore and due west of SIO (Figure 6). The sea-floor instrument was also fitted with a pressure sensor of the type described by Cox et al. (1984), allowing pressure variation measurements between the Nyquist frequency and about a 1000-s period.

A second sea-floor data logger was deployed at Pinyon Flat Geophysical Observatory about 150 km northeast of the offshore site. This instrument was connected to two horizontal BF-4 sensors and to 75-m electric dipoles terminated by conventional lead-lead chloride electrodes. An Electromagnetic Instruments (EMI) electric field signal conditioner was used with 40 dB of gain and 10 Hz low-pass filtering. Crystal oscillators in both instruments were started from a precision clock at SIO prior to deployment. Sampling rate was continuous at 8 Hz during the 10-day deployment.

Figure 7 shows an example 500-s time series from this experiment on all ten data channels. The recording compass shows that the sea-floor instrument was fortuitously aligned north–south, the same direction as the remote reference installation. Even visual inspection shows high correlation at periods around 10 s. The sea-floor fields are attenuated over the land recordings (10^{-8} V/m on the sea floor and 10^{-5} V/m on land), partly as a result of the severe roll off of the sea-floor amplifier and partly as a consequence of the lower apparent resistivities that would be observed on the sea surface. Magnetic field attenuation at periods longer than 10 s is modest—barely more than a factor of two and about that predicted on a conductive sea floor. The "fuzz" on the sea-floor magnetic field, most apparent on the east sensor, occurs at about 3 s and is associated with the sensor rocking or vibrating in the Earth's magnetic field. The main earth field couples into the induction coils as $\cos\theta$, where θ is the angle between the coil axis and the main field. Rocking of the instrument can be expressed as $d\theta/dt$, where t is time; so noise associated with rotational movement goes as $\sin\theta$, which is maximum for $\theta = 90°$, or the east–west component. The rocking is calculated to be about one microradian in magnitude.

The data were reformatted from the SIO binary format used in the sea-floor instruments to EMI time series format and written to optical disk using the ISO 9660 standard. This allows considerable volumes of data (up to 700 Mbytes at a time) to be read directly into MTR-93, a commercial MT software package for computation of MT response functions. A total of nine days of data were processed in this way.

Figure 8 shows coherency between the unrotated magnetic field at the remote site and the sea-floor site. Coherency is high at periods longer than 10 s, showing that the magnetic fields are spatially homogeneous over these distances, that the orientation of the sea-floor instrument has been correctly estimated, and the sea-floor magnetic field is being measured with fidelity. The dramatic drop in coherency at periods shorter than 7 s comes from a combination of the red source-field spectrum and seawater attenuation.

Figure 9 (solid symbols) shows the computed MT response computed using both E and B fields from the sea-floor instrument, remote reference processed using Pinyon Flat data. At high frequencies we see the sea floor response for both modes asymptote to seawater resistivity (0.3 Ωm), as one would expect for water-saturated surface sediments. At low frequencies a depression of the XY mode creates considerable anisotropy, as one might expect from a coast effect that depresses the electric field perpendicular to the coastline. Data quality is excellent down to periods of about 8 s, whereupon the loss of field strength increases error considerably, although the data are still consistent. At periods shorter than 3–5 s, errors exceed 100%.

To attempt a quantitative interpretation of a single MT site having such a complicated response would be ill advised. However, it is instructive to find the best possible fit to the two components that could be derived from 1-D structure. Such a fit is provided by the ρ^+ algorithm of Parker and Booker (1996), which is similar to the well-established D^+ algorithm of Parker and Whaler (1981) but operating on resistivity and phase rather than complex admittance. The results are shown by the solid lines in Figure 9. These fits tell us three things.

1) On land, experience shows that the two components of a 2-D MT response can be fit separately using 1-D theory. This does not hold for the sea floor. The departures from the best 1 D fits, evident as overly steep transverse electric (TE) resistivities at periods longer than 100 s and low, even negative, phases at the 40-s period, are characteristic of working within the conductive seawater near a coastline. Indeed, simple modeling of the sea-floor bathymetry and coastline reproduces both these features using a 2-D model.
2) The ability to fit the broad features of apparent resistivity and phase indicates the data are meaningful and causal (that is, instrumental problems are not evident).
3) The one sigma error bars plotted for the data are derived from the response function estimation process and are only approximately related to true data quality. We can use the scatter of the apparent resistivity data around

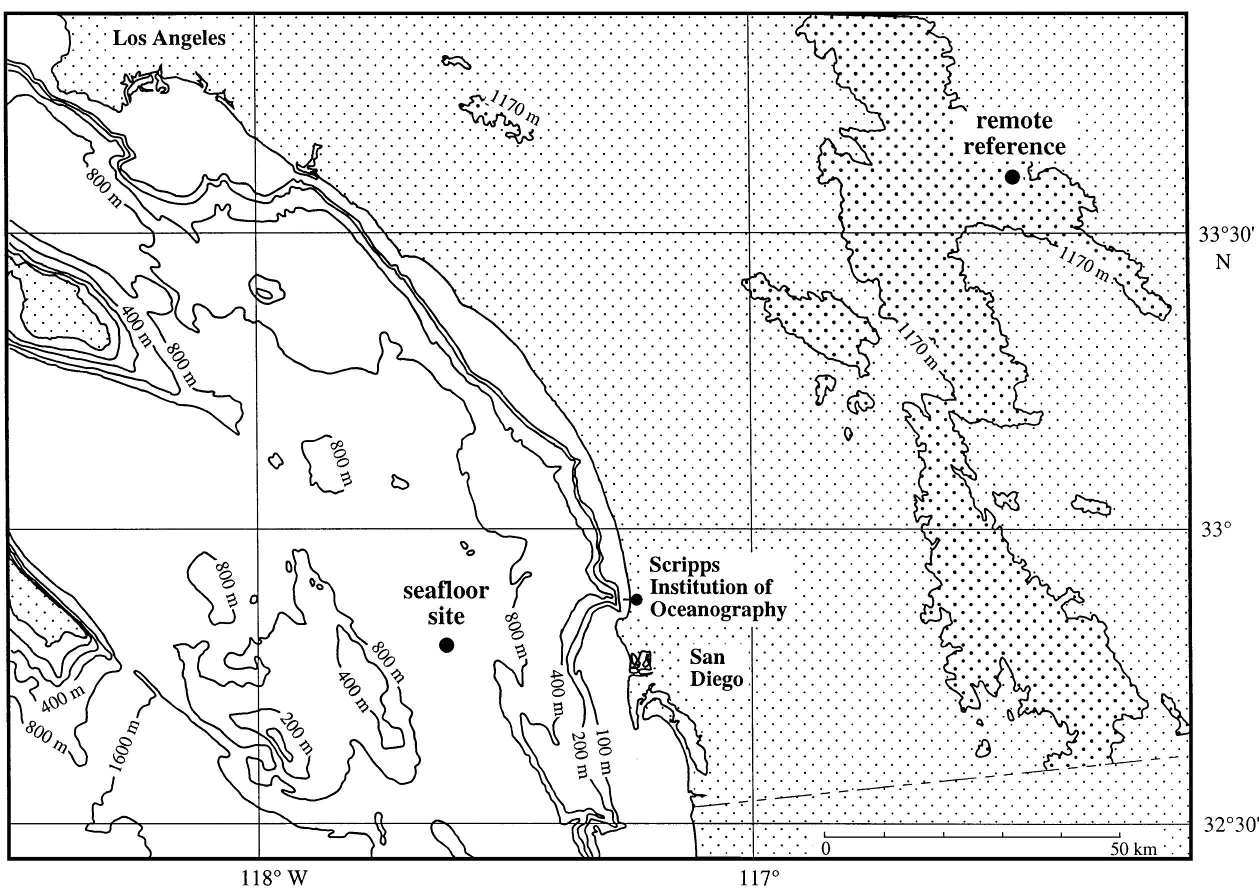

FIG. 6. Map showing location of sea floor and remote reference sites.

the optimal fits in the band 10–100 s (where 3-D effects appear minimal) to make a more realistic estimate of data quality. The scatter measured in this way suggests an average error of 12% for the *YX* mode and 8% for the *XY* mode. With 11 data per decade of frequency, we have considerable redundancy; stacked down to 3 data points per decade (as is used for the modeling in the companion paper), the appropriate errors would be 6% and 4%.

Shallow-water deployments (around 100 m) have also been carried out off San Diego. In these deployments the quality of the electric field measurements is still good, but motional noise associated with wave activity and water currents degrades the quality of the magnetic field significantly. Sea-floor magnetic recordings can be replaced with nearby land magnetic recordings to produce a hybrid MT response constructed from sea-floor electric fields and land magnetic fields. Using 1-D theory described earlier, it can be shown that this hybrid response is similar to a sea-surface MT response (Figure 10). Since the difference between the (effective) sea-surface response and the sea-floor response is not great for the thinner seawater layer, this approach is quite tractable, and excellent MT responses to frequencies of 10 Hz have been obtained in this way. Both 1-D and 2-D forward and inverse modeling codes can be modified to deal with such a hybrid MT response in a rigorous fashion, or conventional 1-D interpretation can be used on the longer period part of the response curve, assuming the response to be a sea-surface MT curve. In a compromise approach, 1-D inversion of the hybrid MT curve can be used to correct the data for the effect of seawater, thus constructing a sea-floor response for conventional 2-D or 3-D interpretation.

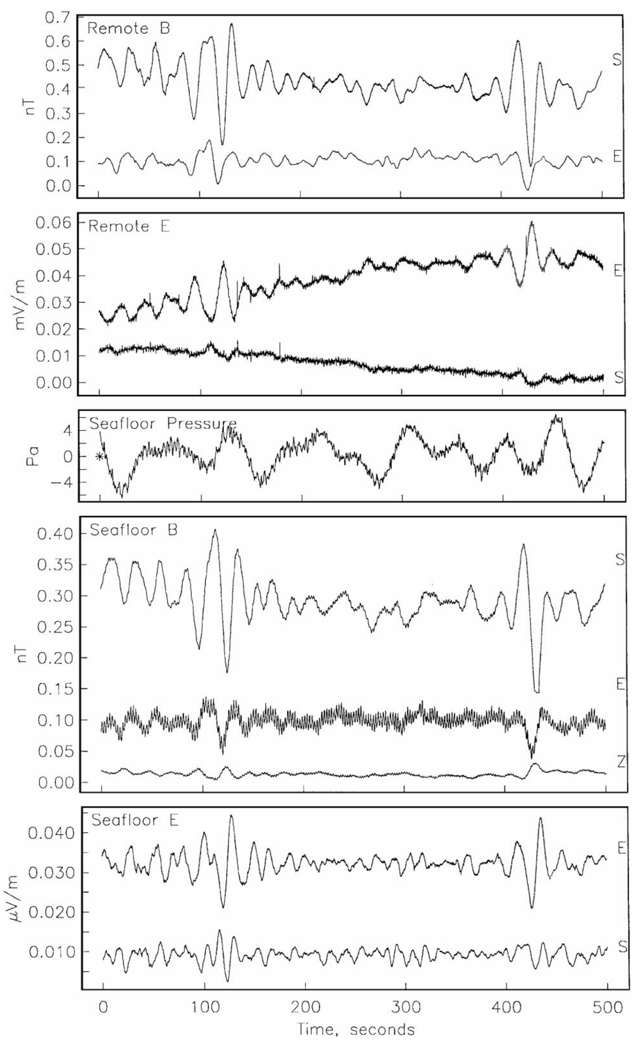

FIG. 7. Sample time series collected in 1-km water depth off San Diego and the remote reference site at Pinyon Flat Geophysical Observatory.

CONCLUDING REMARKS

Using a novel combination of induction coil sensors and ac-coupled electric field amplifiers, good-quality sea-floor magnetotelluric responses at periods of 3 s to 1000 s have been obtained in water 1 km deep. In the past, a short period limit of 300 s would be typical for marine MT in water of this depth. In shallower water, responses to several hertz can be obtained. This increased bandwidth allows the MT method to be used on the continental shelf to map sedimentary structure as an aid to petroleum exploration. While MT lacks the resolution of the seismic reflection technique, it can be used in situations where the seismic method performs poorly, such as imaging beneath salt, volcanics, or carbonates. Also, because electrical conductivity is a strong function of porosity, it can be used alone or in conjunction with seismic velocities to interpret porosity and permeability (e.g., Evans, 1994). There are also many academic applications of these instruments because the increased bandwidth extends the resolution of the MT method from the mantle into the oceanic crust. For example, model studies show that crustal magma chambers at midocean ridges are resolvable with 10–1000-s data, easily obtained over the relatively shallow ridge bathymetry. Structure associated with porosity variations

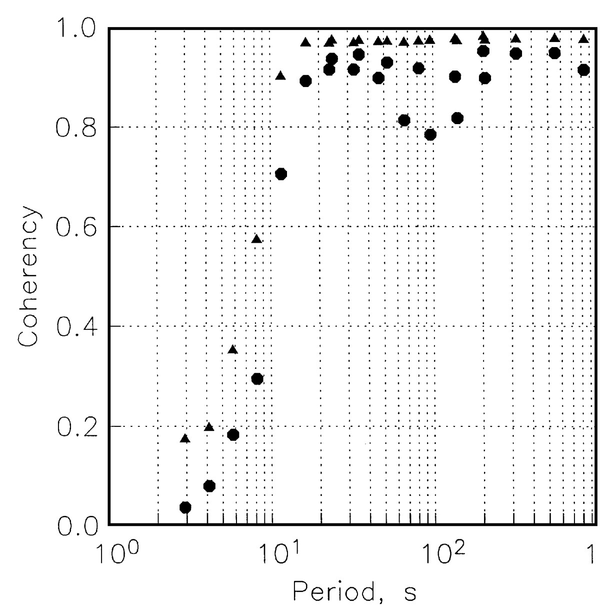

FIG. 8. Unrotated coherency between magnetic fields at Pinyon Flat and the sea-floor site in the N–S (circles) and E–W directions (triangles).

in the accretionary wedges of subduction zones also represents a viable target.

In the companion paper (Hoversten et al., 1998, this issue), 10–20 stations of synthetic data are inverted to recover accurate 2-D structure for a variety of subsalt geometries. In practice, several times this number of stations might be needed to characterize a prospect, and a modern MT land survey can consist of up to 50 or 100 stations. The new SIO MT system has been designed with operations of this size in mind. Once a field project is mobilized, it takes about an hour to prepare an instrument for deployment and perhaps another hour to recover a deployed instrument (depending on water depth and station spacing). Thus, about five instruments can be recovered and deployed in a 10-hour work day and left overnight to record. At this rate a

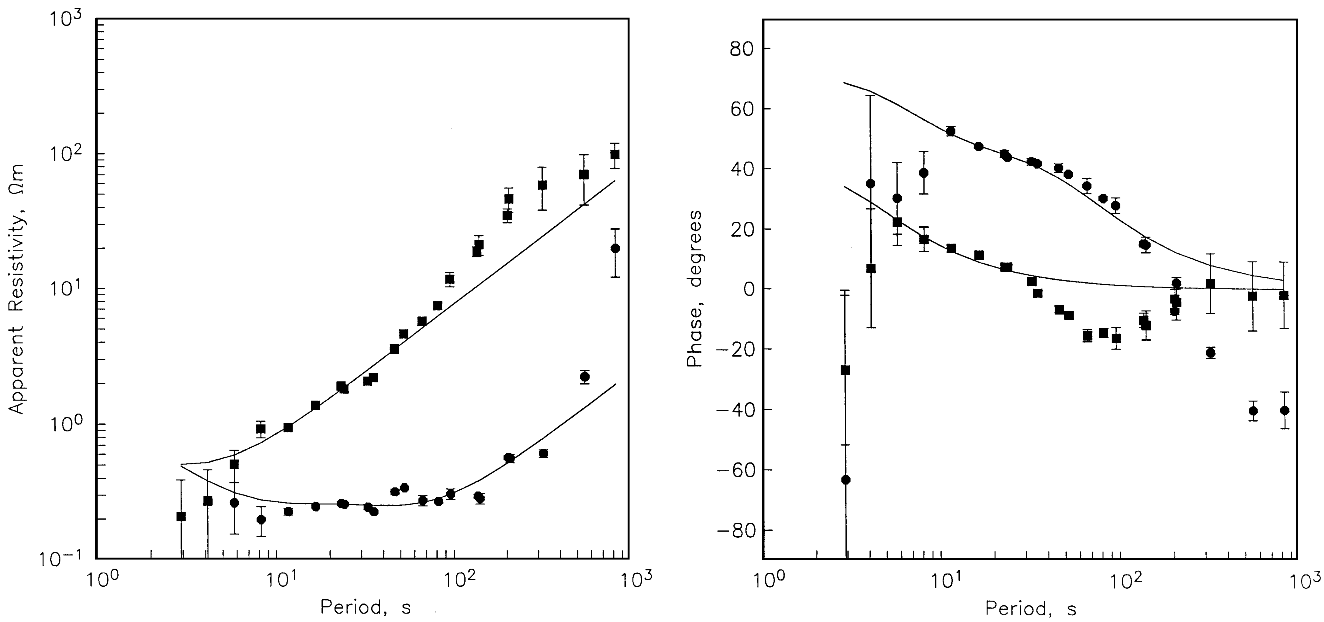

FIG. 9. Apparent resistivity and impedance phase of rotated sea-floor MT response computed using the Pinyon Flat remote reference. The YX component is plotted as squares and the XY component as circles, where YX is associated with an electric field oriented at 5°. The error bars are one standard error derived from the response function estimation. As a measure of internal scatter and as an indicator of 3-D effects, optimal 1-D fits to each component have been made using the ρ^+ algorithm of Parker and Booker (1996) (solid lines).

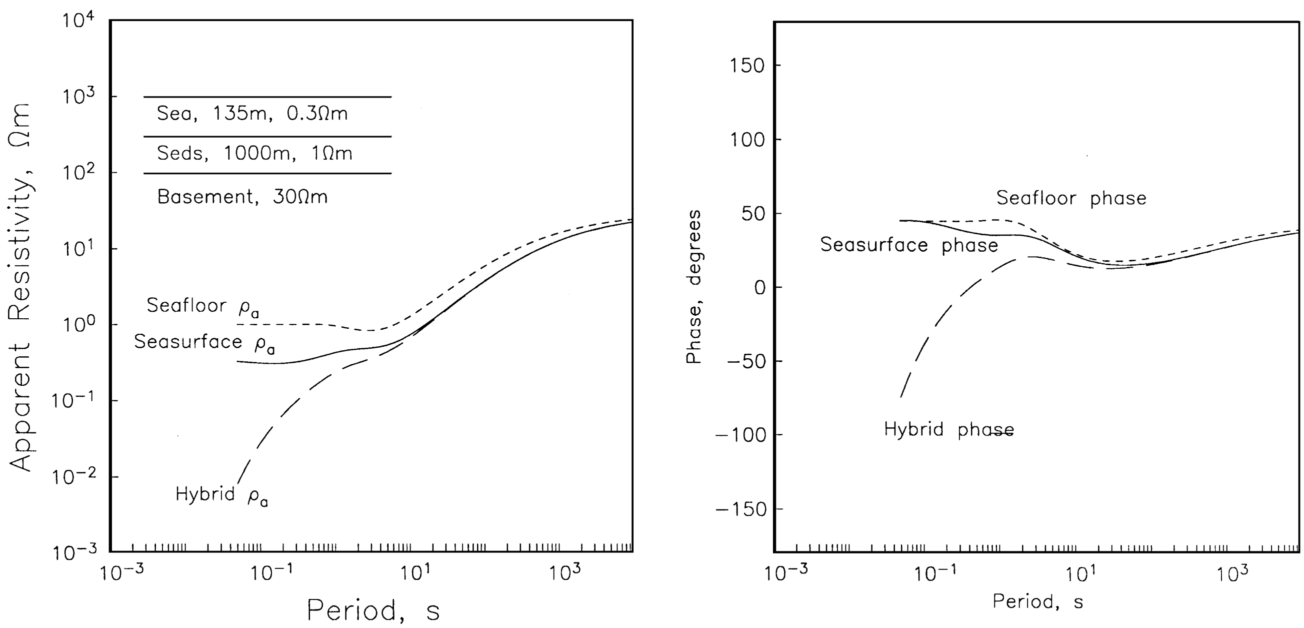

FIG. 10. The effect of constructing an MT impedance using the sea surface (land) magnetic field with a sea-floor electric field to produce a hybrid response, compared with conventional MT responses on the sea floor and sea surface for the simple two-layer model shown.

100-station survey can be completed in three weeks. Loss rates vary between about 1 and 5% per deployment, so a fleet of ten instruments at the start should ensure completion of the marine operation. While the costs of the ship and instrument replacement do not have to be borne by land surveys, the daily productivity is greater for marine MT; it is estimated that the per-station cost of a marine survey would not greatly exceed the cost of an expensive land operation. Indeed, a small proprietary survey was conducted during the development of the equipment and methodology.

ACKNOWLEDGMENTS

An engineering team consisting of T. Deaton, C. Hollinshead, D. Jacobs, J. Lemire, A. Nance, D. Willoughby, and P. Zimmer helped develop the sea-floor MT instrument. We thank Captain L. Zimm and the crew of the R.V. *Sproul* for assistance with the marine experiment and Electromagnetic Instruments Inc. for the loan of equipment. We thank Ransom Reddig for running 2-D model simulations of the experimental data, and an anonymous reviewer, John Stodt, and associate editor Richard Sigal for helpful comments on the manuscript. Part of this work was conducted under sponsorship from Chevron, Texaco, and Agip.

REFERENCES

Cagniard, L., 1953, Basic theory of the magneto-telluric method of geophysical prospecting: Geophysics, **18**, 605–635.

Constable, S. C., 1990, Marine electromagnetic induction studies: Surv. Geophys., **11**, 303–327.

Constable, S., and Cox, C. S., 1996, Marine controlled-source electromagnetic sounding 2. The PEGASUS experiment: J. Geophys. Res., **101**, 5519–5530.

Constable, S. C., Parker, R. L., and Constable, C. G., 1987, Occam's Inversion: a practical algorithm for generating smooth models from EM sounding data: Geophysics, **52**, 289–300.

Constable, S., Hoversten, M., Morrison, F., and Orange, A., 1994, Seafloor MT for petroleum exploration: Contributed paper at 12th workshop on electromagnetic induction, Brest, France.

Cox, C. S., Deaton, T., and Webb, S., 1984, A deep-sea differential pressure gauge: J. Atmos. Oceanic Tech., **1**, 237–246.

Edwards, R. N., Law, L. K., and DeLaurier, J. N., 1981, On measuring the electrical conductivity of the oceanic crust by a modified magnetometric resistivity method: J. Geophys. Res., **86**, 11609–11615.

Egbert, G. D., and Booker, J. R., 1986, Robust estimation of geomagnetic transfer functions: Geophys. J. R. Astr. Soc., **87**, 173–194.

Evans, R. L., 1994, Constraints on the large-scale porosity and permeability structure of young oceanic crust from velocity and resistivity data: Geophys. J. Int., **119**, 869–879.

Filloux, J. H., 1987, Instrumentation and experimental methods for oceanic studies, *in* Jacob, J. A., Ed., Geomagnetism: Academic Press, Inc., 143–248.

Gamble, T. D., Goubau, W. M., and Clarke, J., 1979, Magnetotellurics with a remote reference: Geophysics, **44**, 53–68.

Hoversten, G. M., Morrison, H. F., and Constable, S. C., 1998, Marine magnetotellurics for petroleum exploration, Part 2: Numerical analysis of subsalt resolution, **63**, 826–840.

Law, L. K., 1983, Marine electromagnetic research: Geophys. Surv., **6**, 123–135.

Nichols, E. A., Morrison, H. F., and Clarke. J., 1988, Signals and noise in measurements of low-frequency geomagnetic fields: J. Geophys. Res., **93**, 13 743–13 754.

Orange, A. S, 1989, Magnetotelluric exploration for hydrocarbons: Proc. IEEE, **77**, 287–317.

Palshin, N. A., 1996, Oceanic electromagnetic studies: A review: Surv. Geophys., **17**, 455–491.

Parker, R. L., and Booker, J. R., 1996, Optimum one-dimensional inversion and bounding of magnetotelluric apparent resistivity and phase measurements: Phys. Earth Planet. Inter., **98**, 6–282.

Parker, R. L., and Whaler, K. A., 1981, Numerical methods for establishing solutions to the inverse problem of electromagnetic induction: J. Geophys. Res., **86**, 9574–9584.

Petiau, G., and Dupis, A., 1980, Noise, temperature coefficient, and long time stability of electrodes for telluric observations: Geophys. Prospect., **28**, 792–804.

Schmucker, U., 1970, Anomalies of geomagnetic variations in the southwestern United States: Bull. Scripps Inst. Oceanogr., **13**, Univ. Calif. Press.

Sinha, M. C., Navin, D. A., MacGregor, L. M., Constable, S., Pierce, C., White, A., Heinson, G., and Inglis, M. A., 1997, Evidence for accumulated melt beneath the slow-spreading mid-Atlantic ridge: Phil. Trans. Roy. Soc. London, Ser. A, **355**, 233–254.

Smith, J. T., and Booker, J. R., 1991, Rapid inversion of two- and three-dimensional magnetotelluric data: J. Geophys. Res., **96**, 3905–3922.

Tikhonov, A. N., 1950, Determination of the electrical characteristics of the deep strata of the earth's crust: Doklady Akadamia Nauk, **73**, 295–297.

Vozoff, K., 1972, The magnetotelluric method in the exploration of sedimentary basins: Geophysics, **37**, 98–141.

Wannamaker, P. E., Stodt, J. A., and Rijo, L., 1986, A stable finite element solution for two-dimensional magnetotelluric modelling: Geophys. J. Roy. Astr. Soc., **88**, 277–296.

Webb, S. C., Constable, S. C., Cox, C. S., and Deaton, T. K., 1985, A seafloor electric field instrument: Geomag. Geoelectr., **37**, 1115–1129.