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Controlled-source electromagnetic sounding of the oceanic lithosphere

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The attenuation of ionospheric signals in the frequency range 0.06–24 Hz by sea water effectively precludes using the magnetotelluric method to study the electrical structure of the upper oceanic lithosphere. We have carried out a dipole–dipole electromagnetic sounding in the North Pacific by injecting electromagnetic signals into the ocean and sea bed. The crust at the site is 25 Myr old and has a thin sediment cover. The technique, similar to that used in earlier work[1,2], involves dragging a horizontal dipole antenna along the sea floor. The electric fields that propagated through the resistive basement were detected by seafloor receivers at ranges of 10–65 km. As the ambient electric field is very small (varying from 10^{-18} V^2 m^{-2} Hz^{-1} at 0.1 Hz to 10^{-24} V^2 m^2 Hz^{-1} above 1 Hz; ref. 3), the controlled-source signals could be easily monitored. Our data are consistent with a simple one-dimensional Earth model consisting of a 3–7-km-thick crustal layer of moderate conductivity (~0.001 S m^{-1}) underlain by a thicker region of very low conductivity (<2×10^{-5} S m^{-1}). The results suggest an upper mantle water content of at most 0.1% by volume.

The experiment was performed by towing a transmitting antenna consisting of a 600-m section of insulated armoured cable coupled to the ocean by 15 m of bared cable at the ends. The antenna was energized using a waveform synthesizer supplied with 60-Hz power from a research ship's generators through a towing cable. Rectified half-cycles of the 60-Hz power were modulated by bipolar switching to produce a variable-frequency waveform having equal first and third harmonics. The current amplitude at both frequencies was 65 A. The receivers[4] were self-contained recording voltmeters which measured a single component of the horizontal electric field averaged over a 600-m antenna of insulated wire. The antennas were coupled to the ocean by low-noise, silver/silver chloride electrodes. The receivers were programmed to stack the received signal synchronously with the transmitted waveform, and the source frequency was stepped by a factor of 2 at 30-min intervals.

The transmitting antenna was towed over the sea floor so that it remained in continuous contact with the sea bed. Owing to imperfections in navigation and the effect of varying winds and currents on the towing ship and cable, it is not certain that the antenna stretched out in a straight line, but its contact with the sea floor was monitored continuously with acoustic methods. The path of the ship during the tow is illustrated in Fig. 1, which also gives the locations and orientations of the two receivers which have yielded the best data.

The propagation of electromagnetic signals at the transmitted frequencies in conducting sea water and rocks is essentially a diffusive process because the conduction current far exceeds the displacement current. The attenuation of an electromagnetic disturbance with range is very rapid in sea water, and the detection of the source at the long distances used in this experiment implies propagation through the rocks below the ocean floor. The variation of the amplitude of the electric field with range and frequency provides information on the sub-seafloor electrical conductivity as a function of depth. The theory is well-developed for one-dimensional media, where the electromagnetic fields separate into two independent modes associated with the vertical electric and magnetic fields respectively[5].

At the highest frequencies transmitted, the measured electric

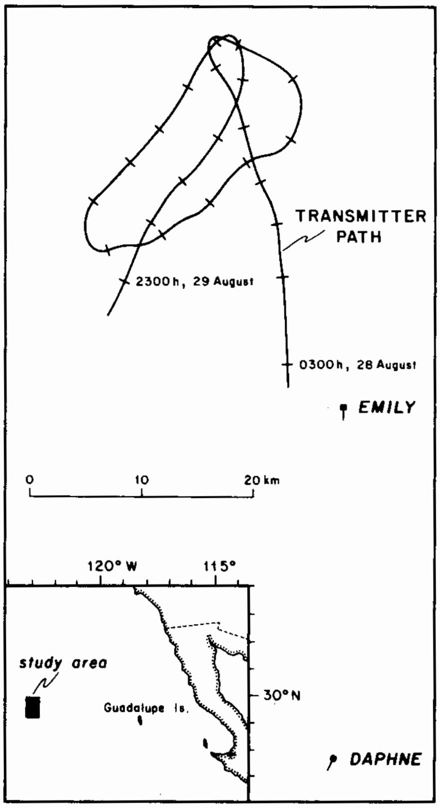

Fig. 1 The inset shows the location of the larger map of the study area, at 28° N, 122° W. The path of the ship during towing of the transmitter was derived by smoothing navigational data collected every 15 min. The cross-marks show ship positions at 2-h intervals. Emily and Daphne are the two electrical recorders. The age of the crust is 25 Myr (magnetic anomaly 6E, after ref. 11).

fields were surprisingly strong and had signal-to-noise ratios >10 (after stacking) for all separations of the transmitter and receiver. The signals at the lowest frequencies were weaker and more difficult to detect because of a higher noise level. Frequency-dependent polarization of either the transmitter or receiver electrodes would explain this type of variation. This hypothesis has been rejected on the basis of laboratory studies of the electrodes. Furthermore, the received signals were so much weaker than the broadband background noise level that polarization losses would be associated mainly with the low-frequency components of the noise. An observed independence of apparent conductivity with range despite a large decrease in amplitude of the electric field implies that any nonlinear amplitude-dependent bias is small. However, failure to stretch either the transmitter or receiver antennas to their full length would produce a systematic downward bias in the observed fields. Several independent cross-receiver checks and the reproducibility of the signal amplitude, especially at short and moderate ranges, suggests that these biases are unimportant. An earlier experiment in 1983 also gave results consistent with this experiment.

Layered structures which fit the data adequately were constructed using trial and error and Marquardt fitting techniques[6]. Model-fitting procedures of this type do not have unique solutions, and a variety of models have been constructed which fit the data equally well. Three such models are shown in Fig. 2. Although it is difficult to give bounds for all acceptable models, common characteristics of the models studied are evident. The first characteristic is a well-resolved upper layer, ~5 km thick, of moderate conductivity (10^{-3} S m^{-1}), presumably representing cool, fractured and water-saturated crust. (The seismically estimated crustal thickness at 27°24′ N, 212°35′ W is 7.4 km[7].) The

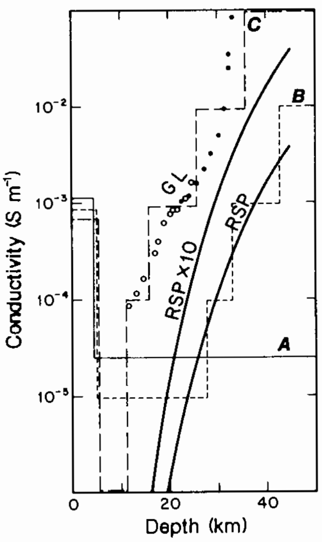

Fig. 2 Three acceptable models of conductivity, compared with laboratory data. The laboratory measurements have been interpreted as conductivity versus depth below the sea floor by assuming a uniform thermal gradient of 33 K km^{-1} (ref. 12). GL, the conductivity of a garnet lherzolite at atmospheric pressure[12]; ○, measurements of a natural rock sample; ●, measurements from a powdered, melted and recrystallized sample. Both measurements were made while the temperature was rising. RSP is the conductivity of a single crystal of Red Sea peridot[14] including a pressure correction, and RSP×10 is the 10 times larger conductivity suggested by Shankland and Waff[15] for mantle conductivity. Model *A* is the best-fitting two-layer model. Although the second layer extends to infinite depth, the data are insensitive to structure deeper than 35–40 km. In models *B* and *C* the deeper conductive layers have been constrained to fit RSP and GL laboratory data respectively, but the most resistive and the shallowest layers have been optimized during the model-fitting.

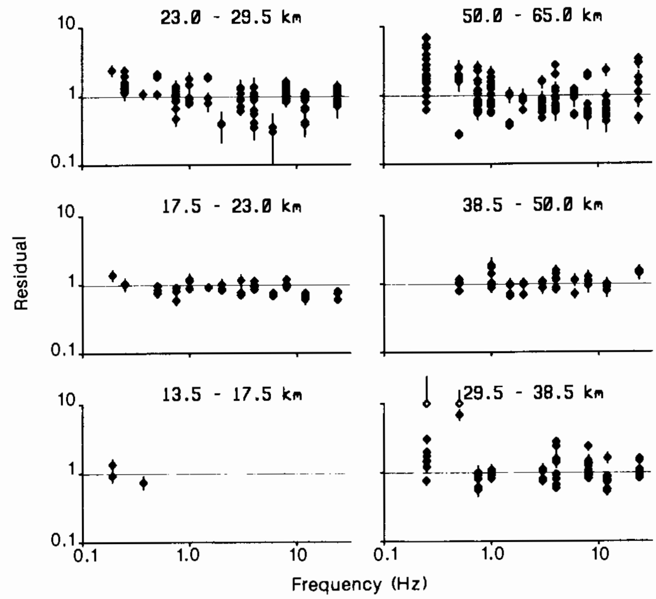

Fig. 3 The residuals (amplitude of the measured electric field component divided by the amplitude of the model prediction) are shown as a function of frequency for six different range bins for the two-layer model shown as line *A* in Fig. 2. The vertical bars represent standard errors in measuring the electric fields. The two data plotted at 10 Hz for a range of 29.5–38.5 km lie off the plot. The data vary over 4 orders of magnitude, from 10^{-17} to 10^{-13} V A^{-1} m^{-2}, although the residuals show that the model fits most of the data to within a factor of 2.

inclusion of a more conductive layer of sediment and fractured basalt at the surface has no apparent effect on the model fits.

A second characteristic is that as the deep structure below 10–20 km is made more conductive, the conductivity of the sub-crustal zone below 5 km must decrease. In the extreme case, where no increase in conductivity at depth is included (model *A*), the sub-crustal zone attains a maximum conductivity of 2×10^{-5} S m^{-1}. Model *A*, however, is impossibly resistive at depths >30 km. To illustrate this, laboratory conductivity data for upper mantle materials have been plotted in Fig. 2. The two sets of laboratory data represent fertile (GL) and completely depleted (RSP) models of mantle conductivity in the sense that partial melting of garnet lherzolite yields basalt while the pure olivine of peridot cannot. Both sets show the conductivity increase associated with the increase of temperature with depth in the mantle. As models of the conductivity in the subsolidus upper mantle, both sets of data suffer from difficulties. The defect structures in the GL sample were probably not in equilibrium at each temperature and there may have been some thermally driven disruption of the grain-to-grain cohesion. The single-crystal measurement of RSP is probably an underestimate of the mantle conductivity because grain boundary effects and impurities tend to increase the conductivity of silicates. Both of these models represent completely dry rocks. If water is present, the conductivity in the comparatively cool parts of the uppermost mantle will be increased greatly by the mobility of ions in the intergranular fluid and in adsorbed films on the grain boundaries[8].

The laboratory data may be used to constrain the deep structure of the electrical models fitting the electromagnetic data (models *B* and *C*). If the RSP curve is taken to be an underestimate of mantle conductivity, we see that the conductivity of the sub-crustal zone is now constrained to be $<10^{-5}$ S m^{-1}. We may use this constraint to place a rough upper limit on the volume fraction of water present in the upper mantle.

For this discussion we assume that the bulk conductivity of the rocks containing intergranular fluids follows Archie's law in the form

$$\sigma = \sigma_f p^2$$

where σ_f is the conductivity of the fluid and p is the volume fraction of interconnected fluid spaces. We neglect entirely surface effects of the intergranular fluid. The conductivity of 0.1 M NaCl solutions is $\sigma_f = 5.2 \pm 0.3$ S m^{-1} in the relevant temperature and pressure ranges (300–600 °C and 0.3–0.6 GPa)[9]. The upper limit on the volume fraction of water is then $p = 1.4\times10^{-3}$. For more saline water, the upper limit on water content is even lower. If the vapour pressure is lithostatic, as we have assumed, the upper limit on the mass fraction of interconnected water is then $\sim4\times10^{-4}$. These numbers show that no appreciable sea water has leaked into the upper mantle, in agreement with indications that serpentinization of mantle rocks in ophiolites occurs after obduction[10], and that juvenile water, if present in the mantle as a whole, must have been swept out during lithospheric generation.

Considerable scatter of the observations about the best-fitting model predictions is observed (Fig. 3). This is not entirely the result of errors caused by a rising noise level, as the signal-to-noise ratio remains large, even allowing for errors in the orientation of the transmitter and receiver antennas. Furthermore, the scatter is not evenly distributed across the range bins. There is evidence that the residuals are a function of transmitter position, suggesting that irregular conductivity structure in the region of the transmitter tow is causing the scatter. However, the fact that there is no systematic structure left in the residuals suggests that the one-dimensional modelling gives a reasonable picture of the gross lithospheric structure.

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