GEOPHYSICAL RESEARCH LETTERS, VOL. 18, NO. 10, PAGES 1917-1920, OCTOBER 1991

UPPER CRUSTAL RESISTIVITY STRUCTURE OF THE EAST PACIFIC RISE NEAR 13° N

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Abstract An active source electromagnetic (EM) sounding has been conducted on the axis of the East Pacific Rise (EPR) at 13° 10′ N. 1D inversion and modelling techniques, seeking resistivity as a function of depth, have been applied to 8 Hz amplitude data collected along the ridge crest. Resistivity is seen to increase monotonically between 50 m and 1 km below the seafloor, increasing from ~1Ωm to around 90Ωm. We observe no intrinsic difference in upper crustal resistivity structure between the rise axis and 100 000 year old crust. Inferred surface porosities of 20% are larger than those recorded in 5.9 my old crust in DSDP hole 504B. Our data do not require, and lack sufficient information for, the reliable inclusion of a conductive termination to the model below 1.2 km.

Background

This letter outlines initial results from cruise CD39, carried out in June 1989 on board RRS Charles Darwin. The objective was to identify the crustal electrical structure at the axis of the EPR (figure 1.) using active source EM techniques developed separately at the Bullard Labs, Cambridge University [Sinha et al., 1990] and at Scripps Institution of Oceanography [Webb et al., 1985]. This area has been the subject of recent seismic surveys seeking to resolve an axial magma chamber from the presence of a seismic low velocity zone (LVZ) [Harding et al., 1989]. There is compelling evidence for a large discontinuity in seismic velocity at a depth of 1.2 km although, in contrast to the EPR at 9°30′ N, no bright crustal reflector is seen on across strike profiles. Present opinion at 9°30′ N points towards a melt lens on the order of 100's of metres thick and less than 1500 m wide flanked by a larger zone of lower melt fraction [Kent et al., 1990]. Similar models are proposed for 13° N but with the melt lens more intermittent along strike. Harding et al. [1989] report a 50-250 m thick surface layer exhibiting low P-wave velocities, which is interpreted to consist of heavily fractured extrusive basalts.

The EM skin depth characterizes the decay of the electric field and, at 8 Hz, is 100 m in seawater: the generally more resistive crust is characterised by longer skin depths. Therefore, when a source and receiver, which are both close to the seafloor, are separated by a distance of more than a few oceanic skin depths, a received signal will be dominated by the field that has diffused through the crust. By measuring the seafloor electric field, at a large number of source-receiver separations, it is possible to infer the sub-seafloor resistivity structure [Cox et al., 1986]. The seafloor is extremely quiet in the controlled source band, mainly because of screening by the conductive seawater. [Chave & Cox, 1982, Constable, 1991].

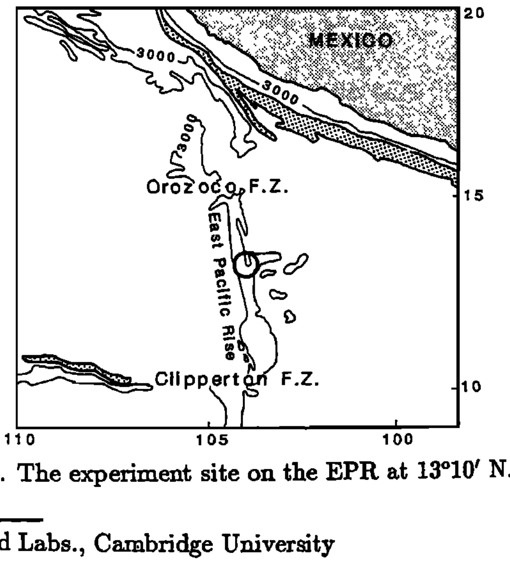

Fig. 1a. The experiment site on the EPR at 13°10′ N.

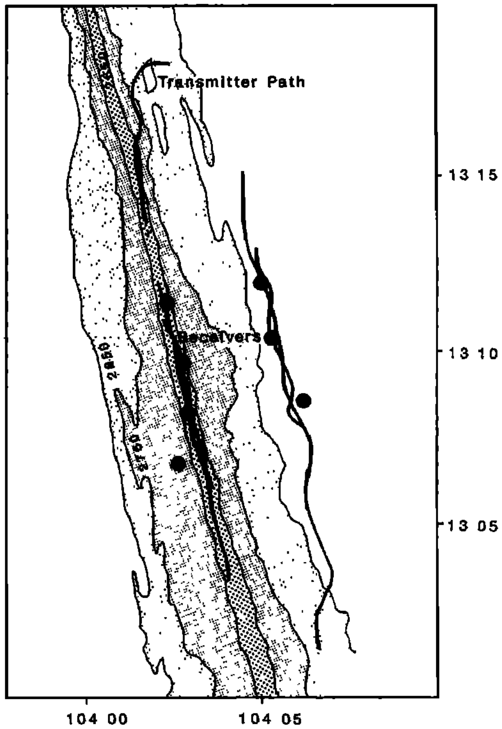

Fig. 1b. Bathymetry of the experiment site after Monti et al. [1987]. Water depths are in metres. Shown are the positions of the seven receivers (circles) and the path of the deep tow (solid line).

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Paper number 91GL02305
0094-8534/91/91GL-02305$03.00

The transmitting source used in the present experiment consists of a 100 m horizontal electric dipole of moment 6000 Am which is part of a neutrally buoyant streamer. This is flown at heights of the order of 30 m above the sea floor. Although this reduces the energy coupled to the seafloor, it allows transmission directly over the ridge crest where the topography is rough and might cause damage to a transmitter placed on the sea floor. An in-phase stacking procedure is employed by the receivers to reduce the volume of data and so it is important to maintain as constant a height as possible in order to minimise phase variations in the received signal. The height is monitored by a transducer mounted on the front of the deep tow package and is controlled by paying in or hauling out wire. Variations in height of less than ±10 m over a 15 minute interval were recorded, during which typical heights above the seafloor of 30 m were attained. A 10 m change in height would shift the electric field by ~10% at 8 Hz transmission.

The receivers are independent packages utilizing 12 m electric antennae to measure the horizontal electric field at the sea floor. Each has an orthogonal pair of such dipoles equipped with low noise Ag-AgCl electrodes.

Our experiment involved the deployment of receivers in two lines (figure 1.); one along the ridge crest, with four instruments placed ~2.5 km apart and the other, of three instruments, ~5 km to the east and parallel to strike on 100 000 year old crust. In all, 36 hours of coverage were obtained along the ridge crest and along the strike-parallel line. Initial transmission of 15 minute blocks at frequencies of 1/4, 1/2, 1, 2, 4 and 8 Hz was completed along both tow lines, and was followed by continuous 8 Hz transmission also along both lines.

Data Analysis and Inversion

The data were examined in the frequency domain. Due to the geometry of the tow lines, most of the received field is in the strike parallel direction, or radial to the transmitter. Amplitudes of the transmitted dipole field were therefore expressed as radial (E_{rad}) and azimuthal (E_{az}) components.

Signal from the transmitter is recorded to source-receiver separations of 7 km before the noise level dominates. Ambient noise parameters were determined from recordings made before and after transmission. The data were screened for a signal to noise ratio of greater than 3. Typical noise values, at 8 Hz, are of the order of 10^{-19} $V^2m^{-2}Hz^{-1}$ although this figure varies between instruments.

In this paper the 8 Hz data set, which was the most substantial, is examined (figure 2.). Curves showing the decrease in amplitude with range for various uniform half-spaces are shown superimposed which clearly do not explain our data and which indicate that the experiment is sensitive to more complex structure. The amplitudes exhibit the degree of scatter expected from a region with significant surface structure and heterogeneity. The sim-

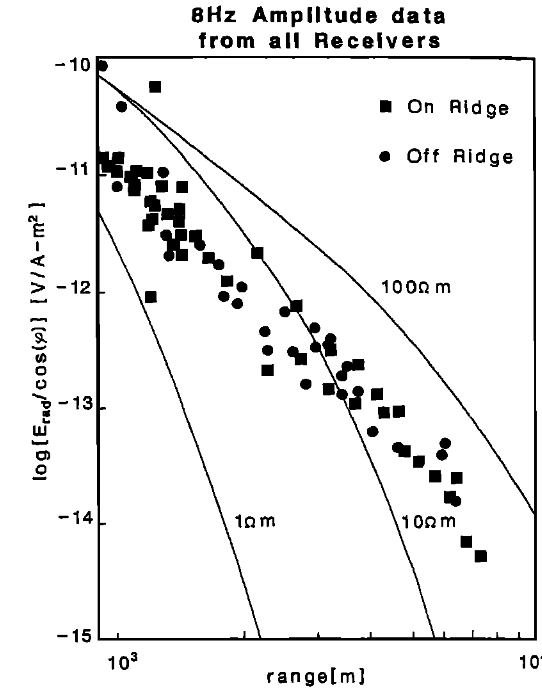

Fig. 2. The data collected at 8 Hz transmission frequency divided into those data collected on (squares) and 5 km off (circles) the ridge crest. Note the similarity between the two groups. Superimposed are amplitude curves for uniform half spaces of 1, 10 and 100Ωm which do not explain our data, indicating that the sounding has sampled a range of structures. The scatter in amplitude is expected from a region with significant near surface structure.

ilarity between amplitudes recorded along both lines, indicating a lack of large scale lateral variation in resistivity structure, supports our approach of modelling the resistivity solely as a function of depth. Furthermore, the geometry of the experiment precludes us from modelling resistivities which vary across strike.

The E_{rad} amplitudes have been partitioned into range bins and means and standard errors in the means computed for each range in the log-amplitude domain (figure 3a.). The errors so calculated attempt to describe the scatter in the data caused by surface heterogeneity. This procedure assumes that the amplitude data in a given range bin are distributed in a Gaussian manner and that bias from systematic sources of error are significantly smaller than random errors. Geometric sources of error, such as changes in transmitter height, errors in source-receiver separation and bias between receivers, have also been considered as explanations for the scatter but are found to be smaller than observed variations. Height errors have been minimized by normalizing the amplitudes to zero transmitter height by accounting for attenuation through the water column. The averaged data were subjected to a regularized inversion seeking maximally smooth 1D models for the expected RMS misfit of 1.0 [Constable et al., 1987] defined in a standard weighted least squares sense. Although better fits may be obtained, decreasing the misfit below the expected value can lead

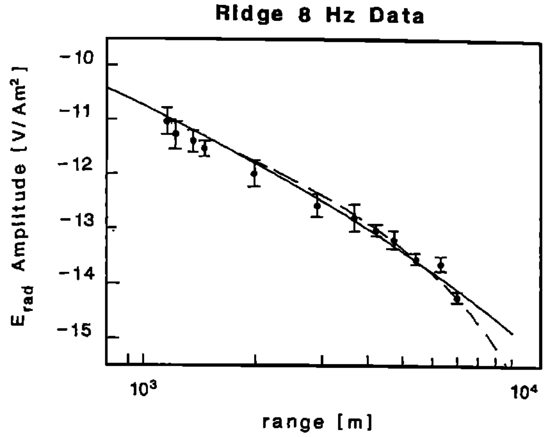

Fig. 3a. The result of averaging the 8 Hz data collected along the ridge crest in range bins. Error bars reflect the scatter in the data caused by near surface structure. Shown also is the response to the optimally smooth model which fits the data to an RMS misfit of 1.0 (solid). The dashed line is the response of the smooth model incorporating a conductive termination beneath 1.2 km.

to large perturbations in the resulting model, introducing structure not required by the data. Figure 3b. shows models which are smooth in the 1st and 2nd derivative sense for both the ridge axis and for 100 000 year old crust. All show a monotonic increase in resistivity with depth. The divergence points between models with the two roughness measures indicate the loss of resolution in our data at depths shallower than 50 m and deeper than 1 km. The response of the optimally smooth models is also shown in figure 3a. Changes in surface resistivity of an order of magnitude either way extending for a depth of 50 m and with a horizontal wavelength of a few hundred metres could cause the scatter observed.

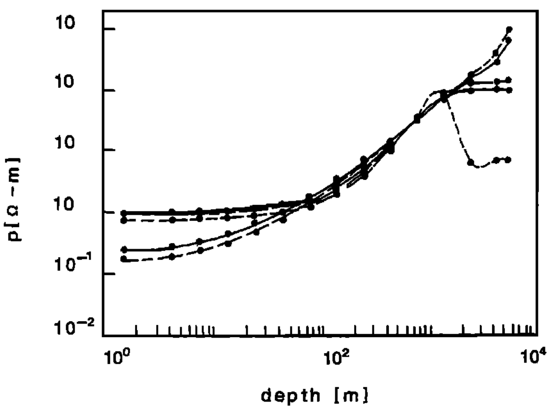

Figure 3b. The two optimally smooth models in a first and second derivative sense for the binned 8 Hz data collected along the rise axis (solid curves). shown also (dashed curves) are the results for inverting 5 km off-ridge data which has been similarly treated. Note the divergence points between the two models indicating loss of resolution at <50 m and >1 km. The smooth model which shows a decrease in resistivity beneath 1.2 km also fits the data to RMS 1.0 and is shown to demonstrate the large variations possible in the terminating half-space.

It is unlikely that the true resistivity structure varies smoothly across geological and seismic boundaries but the diffusive nature of the EM field means that we are sensitive to the bulk properties of the Earth and not to sharp physical boundaries. However, the inclusion of a-priori information allows us to test likely Earth models against our data for consistency. Of particular interest is whether our data support the existence of both the reported seismic boundary at 150 m and the presence of an axial LVZ below 1.2 km. These hypotheses have been tested by means of forward modelling also incorporating DSDP bore hole resistivity values between 150 m and 1.2 km. The data are seen to be consistent with a model with a conductive upper 150 m underlain by a sharp boundary to resistivities similar to those of 6 my old crust. A melt layer underlain by a conductive half space is also consistent with the data although we are unable to specify the values of resistivity in this region. A smooth model in the first derivative sense has been generated with the constraint that the resistivity tends towards an arbitrary conductive value beneath 1.2 km (figure 3b.). The two extremes of acceptable solution - resistive and conductive - correspond geophysically to the absence or presence of a region of partial melt and demonstrate the lack of sensitivity in the data presented to the presence of the seismic LVZ inferred in the region. 1D modelling and inversion studies, as well as sensitivity analyses reveal that 8 Hz amplitude data to ranges greater than 10 km, with smaller errors, must be obtained in the present experimental configuration in order to identify the presence of a broad melt layer.

Interpretation

We relate the values of resistivity obtained to porosities in the upper crust by means of the empirical Archie's law: $\rho_m = \rho_f a \phi^{-t}$, where ρ_m is the measured resistivity, ρ_f is that of the penetrating fluid and ϕ is the porosity. This is possible because the electrical resistivity of dry basalt is in excess of 10^6 Ωm at seafloor temperatures and the low resistivities observed are due solely to the distribution of the conductive seawater within the crust. Although Archie's law was developed for sedimentary rocks it has been shown reliable for basalts [Shankland & Waff, 1974]. The choice of the exponent, t, is dependent on the environment. Recognizing that the upper 200 m are likely to exhibit extensive cracking and that tectonic extension is prevalent, we choose an exponent of 1 to describe our results. An exponent of 1 is also supported by examination of laboratory samples from DSDP hole 504B basalts [Pezard, 1990]. The resistivity of seawater, extrapolated from measurements of saltwater conductivity by Quist and Marshall [1968], is a strong function of temperature but reaches a minimum at 350° C at pressures relevant to the experiment. Choosing the seawater temperature to be the ambient sea floor temperature will over estimate the porosity. Porosity is estimated to be greater than 20% at 50 m depth.

The increase in resistivity with depth reflects a decrease in fluid content with depth which, in turn, is related to the lithological and tectonic structure of oceanic crust. Our data support the existence of a highly fractured layer of extrusives and rubble extending to a depth of around 200 m. Extensive fluid penetration is required to explain the resistivities measured. This is in contrast to measurements made on DSDP bore hole 504B [Becker et al., 1985, Pezard, 1990] where average porosities of 7% are reported in the pillow section. The similarity between data on the ridge crest and on 100 000 year old crust requires a fractured surface layer in both areas which is in broad agreement with expanding spread profile results. The large degree of scatter observed in the amplitudes, particularly among those data collected along the ridge crest, is indicative of heterogeneity. Unfractured sheet flows and near surface dyke intrusions, both of which are resistive features, may be expected on axis. At around 200 m depth, the extent of water penetration has decreased significantly as the basalts become less fractured and below, our data are consistent with the resistivity structure measured in DSDP hole 504B. Our models infer that the largest changes in physical properties occur over in the surface regions. The similarity in resistivity structure below 200 m on and off the ridge is in contrast to seismic observations [Harding et al., 1989]. Higher axial P-wave velocities between 200 m and 400 m depth are recorded in comparison to those of 70 kyr crust, while the trend is reversed below this depth. The changes in seismic velocity have been explained by the rapid burial of near-surface axial dyke intrusions by successive layers of extrusives as the crust migrates away from the neovolcanic zone. The resistivity structure is sensitive to the amount and degree of interconnection of pore fluid in an ambiguous manner. Temperature effects will complicate matters so that a unique transform from resistivity to porosity is not possible. Changes in the large scale crack structure of the basaltic fabric clearly affect the seismic velocity but, the change in resistivity structure, if any exists, is hidden beneath the scatter induced by surface features.

Additional low frequency data is required if information is to be gleaned regarding the presence of a melt body. Identification of the electrical nature of the surface structure appears to be of great importance since the scatter caused by shallow features greatly influences the data and its interpretation.

Acknowledgements We would like thank all who participated in cruise CD39/89 on board the RRS Charles Darwin. Special thanks are due to P. Patel, M. MacCormack, J. Leonard and J. Lemire without whom there could have been no experiment. A. Chave is thanked for the use of his forward code and A. Flosadóttir for implementing the derivatives. The project was supported by the NERC and by the NSF.

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(Received: June 10, 1991;
accepted: July 31, 1991.)