The RAMESSES experiment—III. Controlled-source electromagnetic sounding of the Reykjanes Ridge at 57°45′N

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SUMMARY
A controlled-source electromagnetic sounding survey centred on an axial volcanic ridge (AVR) segment of the Reykjanes Ridge at 57°45′N was performed as part of the RAMESSES experiment. Low-frequency (0.35–11 Hz) electromagnetic signals were transmitted through the crust to an array of horizontal electric field recorders at the seafloor to ranges of 15 km from the source, which was a 100 m long horizontal electric dipole towed at heights of 50–80 m from the seafloor. Coincident seismic and magnetotelluric studies were conducted during the rest of the RAMESSES experiment.

Data were interpreted using a combination of 1-D forward modelling and inversion, and iterative forward modelling in two dimensions. On the axis of the AVR, the resistivity at the seafloor is 1 Ω m. There is a steep resistivity gradient in the upper few hundred metres of the crust, with the resistivity reaching approximately 10 Ω m at a depth of 500 m. In order to explain the low resistivities, the upper layer of the crust must be heavily fractured and saturated with sea water. The resistivity increases with distance from the axis as the porosity decreases with increasing crustal age.

The most intriguing feature in the data is the large difference in amplitude between fields transmitted along and across the AVR axis. A significant zone of low-resistivity material is required at approximately 2 km depth beneath the ridge crest in order to explain this difference. It is coincident with the low-velocity zone required by the seismic data, and has a total electrical conductance in excellent agreement with the results of the magnetotelluric study. The low-resistivity zone can be explained by the presence of a body of partially molten basalt in the crust. Taken together, these results provide the first clear evidence for a crustal magma chamber at a slow spreading mid-ocean ridge. The data constrain the melt fraction within the body to be at least 20 per cent, with a melt volume sufficient to feed crustal accretion at this segment of the ridge for of the order of 20 000 years. Since this body would freeze in the order of 1500 years, this finding lends support to the hypothesis that, at slow spreading rates, crustal accretion is a cyclic process, accompanying periodic influxes of melt from the mantle to a crustal melt reservoir.

Key words: electrical resistivity, electromagnetic surveys, magma, mid-ocean ridges, oceanic crust.

INTRODUCTION
The electrical resistivity of solid, dry basalt exceeds that of molten basalt or sea water by orders of magnitude, so seawater penetration into cracks, the presence of hydrothermal systems, or the presence of melt will all decrease crustal resistivity. Electrical exploration methods, sensitive to these resistivity variations, thus provide information on the amount, distribution and temperature of fluid present, all of which are important parameters in understanding the processes occurring at mid-ocean ridges.

Controlled-source electromagnetic (CSEM) methods utilize time-varying electric and magnetic fields from an artificial source. At frequencies sufficiently high that electromagnetic...
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Fields are attenuated rapidly in the sea water, energy detected by a receiver remote from the transmitter follows diffusion paths through the crust, and is therefore sensitive to its resistivity structure. The CSEM method provides resolution of resistivity structure on a crustal scale by using higher frequencies than is possible in conventional seafloor MT sounding, in which high frequencies from natural ionospheric sources are strongly attenuated by the water layer. Accounts of the theory and practice of marine electromagnetic methods are given by (for example) Cox (1980), Chave & Cox (1982), Edwards & Chave (1986) and Constable & Cox (1996).

Several CSEM experiments using a horizontal electric dipole source operated in the frequency domain have been performed to study the resistivity structure of normal oceanic crust (Young & Cox 1981; Cox et al. 1986; Constable & Cox 1996). The first experiment to be performed over the axis of a mid-ocean ridge was centred at 13°N on the fast spreading East Pacific Rise (Evans et al. 1994). No difference was observed between data collected at the ridge crest from zero-age crust and those collected on the ridge flank crust from over 100 000 years old. Evidence against the presence of a low-resistivity anomaly in the crust led to the conclusion that any melt present must be in the form of small isolated pockets, suggesting that at 13°N the East Pacific Rise is in a state of magmatic quiescence compared with other parts of the ridge.

The experiment described in this paper formed part of the RAMESSES (Reykjanes Axial Melt Experiment: Structural Synthesis from Electromagnetics and Seismics) study (Sinha et al. 1998), which was centred on an axial volcanic ridge segment of the Reykjanes Ridge at 57°45′N. The Reykjanes Ridge forms the northern part of the slow spreading Mid-Atlantic Ridge, extending from the Reykjanes peninsula on Iceland at 63°30′N to the Bight fracture zone at 56°50′N (inset). The full spreading rate is 20 mm a\(^{-1}\) along 096° (DeMets et al. 1990), oblique to the overall trend of the ridge.

The experiment was designed to study the processes of crustal accretion at a slow spreading mid-ocean ridge, especially melt delivery to and storage within the crust. CSEM sounding, magnetotelluric (MT) sounding (Heinson, White & Constable, in preparation) and both wide-angle and normal-incidence seismic profiling (Navin, Peirce & Sinha 1998) were performed. With its sensitivity to crustal water and melt, the CSEM experiment was a key component of the study, and was centred on an AVR segment characterized by evidence of recent volcanism.

THE CSEM METHOD

The decay of electromagnetic fields in a medium is governed by both the resistivity of the medium and the frequency of the signal. The electromagnetic skin depth defines the distance over which the amplitude of an electromagnetic field decays by a factor e\(^{-1}\) and the phase of the signal is shifted by \(\pi\) radians:

\[
\delta_s = \frac{\sqrt{2\mu}}{\sqrt{\rho}} \approx 500 \frac{\nu}{f}.
\]

where \(\delta_s\) is the skin depth, \(\rho\) is the resistivity of the homogeneous medium through which the electromagnetic field is diffusing, \(\omega\) is the angular frequency of the signal and \(\mu\) is the magnetic permeability, usually assumed to take its free-space value everywhere. Although this expression is derived assuming simple plane wavefields, it provides a useful guide to the attenuation of more complicated dipole fields in the crust. The

Figure 1. The CSEM study centred at 57°45′N on the Reykjanes Ridge. The inset shows the location of the RAMESSES experiment (square), with bathymetric contours at 800 m intervals from sea level. Bathymetry of the work area, shaded at depths greater than 1800 m, is derived from the gridded swath bathymetry data collected during the RAMESSES experiment. The CSEM instrument positions and DASI tow lines are shown. Tow 1 ran down the crest of the AVR. Tow 2 was 4 km to the west of the AVR. Instruments were deployed along and across the AVR coincident with the wide-angle seismic lines. Symbols show the instrument positions and types. The arrows on the LEM instruments show the directions of their 300 m long antennae. Also shown are the positions of the navigation transponders used in determining the source position.

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diffusive nature of the fields means that sharp boundaries and anomalies with a scale-length much less than a skin depth are not well resolved.

Our experiment involves the transmission of an electromagnetic signal at discrete frequencies (in contrast to the time-domain system of Edwards & Chave 1986) from a horizontal electric dipole (HED) source to an array of remote seabed-bottom receivers which detect the horizontal electric field. The frequency at which signals are transmitted must be carefully chosen, taking into account the possible resistivity and scale of structures that might be encountered. If the frequency is too low, the skin depth in the seafloor is very long, so the signal is not inductively attenuated between the source and the receiver and cannot resolve crustal-scale structure. If the transmission frequency is too high, the skin depth is very small, and signals only penetrate the shallowest part of the crust. In this case, all but the very shortest-range signals are attenuated to such an extent that they can no longer be detected above the noise (e.g. Flosádóttir & Constable 1996). Another consideration is the noise spectrum. At frequencies below approximately 0.3 Hz, microseism noise is significant (Webb & Cox 1986), and, at lower frequencies still, ionospheric noise starts to leak through the conductive ocean. The range of frequencies that can be usefully employed for probing crustal structure to depths of a few kilometres or more is therefore between a few tenths and a few tens of hertz.

THE EXPERIMENT

The source, DASI (Deep-Towed Active Source Instrument), consists of a deep-towed vehicle which in turn tows a neutrally buoyant streamer that forms a 100 m long HED (for details see Sinha et al. 1990). The DASI has a source moment of approximately 10^4 A.m, and is towed at heights of 50–80 m above the sea bottom, monitored acoustically by a 3.5 kHz echo-sounder mounted on the deep-tow vehicle. The source is kept as close as possible to the seafloor to ensure good coupling of electromagnetic energy into the crust, without its coming into contact with the rugged ridge topography and risking damage. A high-voltage shipboard power-supply unit generates a 256 Hz signal which is passed via an armoured coaxial cable to the DASI package, where it is transformed to a high-current output, typically 300 A peak-to-peak. Four ancillary electrodes spaced along the source dipole are connected to a data logger mounted on the deep-tow vehicle (the DASI logger) and are used to monitor the transmitted fields. The location of the source is determined by acoustic ranging between the ship, the deep tow and the seafloor instruments, augmented by an array of navigation transponders.

Three types of receiver were deployed. ELF (ELectric Field) and LEMUR (Low Frequency ElectroMagnetic Underwater Receiver) instruments measure two components of the horizontal electric field at the seafloor, using HED receivers (Young & Cox 1981; Sinha et al. 1990; Constable & Cox 1996). Each 10 m receiver dipole consists of a pair of silver/silver chloride electrodes supported at the ends of orthogonal plastic arms. These instruments can detect signals to a source–receiver separation of approximately 15 km. LEM (Long-wire ElectroMagnetic) instruments have a single long antenna (300 m in this case) deployed on the seafloor behind the instrument package (Webb et al. 1985; Constable & Cox 1996). Although only able to detect one component of the seafloor electric field, the long receiver dipole makes these instruments more sensitive, and less susceptible to internal instrumental noise than the short-arm ELFs and LEMURs, extending the source–receiver range to around 90 km. The ELF instruments were modified from earlier experiments by replacing tape drives with high-capacity disk drives, allowing the stacking algorithm used previously to be replaced with the collection of continuous time series at 64 Hz sampling. All the LEMs and several of the ELFs were equipped with MT amplifiers for use in the MT experiment described by Heinson et al. (in preparation), as well as CSEM amplifiers. Instrument positions were determined to within 50 m by acoustic surveying.

The experimental geometry is illustrated in Fig. 1. Instruments were deployed in lines along and across the AVR axis, coincident with the seismic wide-angle profiles (Navin et al. 1998). The short-arm ELFs Quail, Noddy and Kermit were deployed close to the axis, with the long-wire LEMs, Macques, Rhonda, Lolita and Pele at longer range to the east of the axis. Modelling by Unsworth (1991) suggests that trans-ridge transmission into an array of seafloor instruments on the opposite side of the axis from the source is the optimum geometry for detecting sub-axial crustal resistivity structure. The geometry of the experiment was designed with this in mind, to maximize the sensitivity of the data to both 1-D and 2-D structure.

The first source tow was down the crest of the AVR and was cut short due to source failure after three hours of transmission. However, the source passed directly over ELF Quail, providing valuable short-range data capable of constraining the shallow resistivity structure on-axis. Frequencies of 0.35 Hz and 11 Hz were transmitted in a pattern of low- and high-frequency bursts. The second tow, 4 km west of the AVR axis, was completed successfully, with seven hours of 0.75 Hz transmission both along and through the axial zone. DASI tows are carried out at a speed of 1–2 knots (approximately 1 m s^-1) to minimize the phase shift at the receivers during synchronous detection of the signal, reduce wear on the deep tow and maintain small tow cable angles. It thus takes a considerable amount of time to cover significant distances with the source, and instrument failure coupled with unfavourable weather conditions limited total transmission time to 10 hours.

DATA REDUCTION

Except at the closest ranges the signal is at or below the receiver noise level and synchronous detection or stacking must be employed. Source fields must therefore be well characterized. The intention was to transmit frequencies of 0.25, 8 and 64 Hz but a transmitter malfunction modified this.

High-voltage 256 Hz power from the ship is fully rectified, and the resulting half sinusoids switched in DASI to give blocks of positive and negative polarity. The transmission frequency is controlled by switching the polarity of the rectified signal after a specified number of zero crossings of the 256 Hz carrier have been detected. A preset program within DASI determines the specific frequencies and duration of each frequency burst. Owing to instabilities in the 256 Hz carrier signal and errors in the part of DASI controlling the zero crossing count, the polarity of the rectified sinusoids was switched too frequently, thus increasing the frequency of the signal and shortening the length of single-frequency bursts. The switching error was larger for the positive-going half sinusoids, making the transmitted
The waveform slightly asymmetric.

The DASI logger, which monitored the transmitted signal, logged for at most three minutes in every half hour, and so provided an incomplete record of transmission. A source model was constructed by interpolating between points at which logger records were available, supplemented by short-range data from instrument Quail during the first tow. Transmission frequencies during the first tow were 0.35 and 11 Hz, in burst lengths of 92 s. During the second tow the transmission frequency drifted between 0.69 and 0.76 Hz, with an average of 0.75 Hz, in bursts of 44 s.

A least-squares fit at the transmission frequency during each burst was used to extract the amplitude of the fundamental sinusoid from the raw time series recorded by each receiver channel, and to estimate the error on this value. The phase of the transmitter changed between bursts by an unknown amount, making it impossible to stack several bursts to improve the signal-to-noise ratio. Because of this, none of the long-range data from the LEM instruments that was at or beneath the ambient noise level could be extracted. The LEMURs employed a stacking algorithm to reduce data volume, so no data could be recovered from these instruments.

**NOISE EVALUATION**

Time series recorded during the 96 hour break in transmission between tows allowed background noise levels on the instruments to be obtained by applying the same least-squares fitting procedure as was used to extract data amplitudes. Although there is an implicit assumption that the noise level remained constant over the four days of the CSEM experiment, this approach allows direct comparison with the data. Noise levels are plotted in Fig. 2 at frequencies chosen to cover the full range transmitted during the experiment. The noise values determined for each receiver have been normalized by the source dipole moment to provide estimates that can be compared directly with the data amplitudes shown in later figures.

Sources of noise can be divided into two categories: external and instrumental. External sources include electromagnetic fields from ionospheric and cultural sources, and motionally induced fields caused by microseisms, water currents and tectonic activity. Electric fields decay rapidly in the conductive sea-water layer, with the result that ionospheric noise at

![Figure 2](https://example.com/figure2.png)

**Figure 2.** Noise levels for (a) the ELF and (b) the LEM instruments, defined as the RMS electric field in a bandwidth of 0.01–0.02 Hz (the inverse of the time series length used). The noise levels are derived from time series recorded during the break in transmission between source tows, by applying the same least-squares fitting procedure as was applied to the data. The electric field values have been normalized by 10^700 A m, the average source dipole moment during the experiment.
frequencies greater than 0.03 Hz is effectively screened by the water depths encountered on the Reykjanes Ridge. Similarly, high-frequency noise from cultural sources is also removed from the spectrum. Between 0.1 and 2 Hz the dominant sources of external noise are microseisms generated by the non-linear interference of sea-surface gravity wave trains ( Webb & Cox 1986). These produce pressure fluctuations which can propagate to the seafloor even in deep water, resulting in water and seafloor motions which generate electromagnetic fields. In addition, tectonically generated seismic activity can cause ground and water motions which also induce electromagnetic fields ( Webb & Cox 1982, 1986).

Noise generated in the instrument itself by the amplifiers and electrodes is independent of the receiver dipole length, and so the signal-to-noise ratio increases with increasing receiver dipole length. The ELFs were substantially noisier than the LEMs due to their shorter receiver dipole length, although the factor of 30 difference in antenna length suggests that the difference between ELF and LEM noise levels should be larger if the observed values were due only to instrumental noise. Kermit was nearly an order of magnitude more noisy than the other ELF instruments because of contamination by noise from an MT magnetometer deployed in tandem with the instrument. Quail, which was re-deployed after releasing prematurely, was much noisier during its second deployment than its first. This could be due either to an external noise source associated with the change in location or to instrumental noise introduced between deployments (it was modified to record MT as well as CSEM data).

Electric field decomposition

The electric field amplitudes recovered from the raw time series using the least-squares fitting procedure were screened for a signal-to-noise ratio greater than two. Most of the data passed this screening, the exceptions being from the particularly noisy ELF Kermit: no signals from tow 1 were recovered, while only a small number of the data from tow 2, recorded at the shortest range (3–4 km), were included in the modelling. After screening for signal-to-noise ratio, the remaining amplitudes were corrected for the height of the transmitter above the seafloor assuming simple exponential decay of the fields, corrected to take into account the frequency response of the instruments, and finally normalized by the source dipole moment, calculated from DASI logger data and calibrations during pre-deployment deck tests.

The data recorded by the two orthogonal channels of the ELF instruments were further decomposed to yield polarization ellipse parameters (Smith & Ward 1974; Constable & Cox 1996). The maximum axis of the polarization ellipse is a more robust measure of the seafloor electric field than either of the separate components. Since it depends only on the phase difference between orthogonal components, it is unaffected by errors in the absolute phase of the source signal. It is independent of the receiver orientation and varies more slowly with the azimuth from the source than the separate components, and is therefore less affected by geometrical errors. The orientation of the polarization ellipse depends more strongly on the source and receiver orientation, both of which are subject to error, than the seafloor resistivity structure, and for this reason was not included in the analysis. The complete data set is shown in Fig. 3. Error bars show one standard deviation errors associated with the least-squares fitting procedure. The minimum error was set to 20 per cent for the 0.35 and 0.75 Hz data, and 30 per cent for the 11 Hz data, based on an internal scatter in the data similar to that observed by Evans et al. (1991) in the CSEM data collected at 13°N on the East Pacific Rise. It is assumed that this scatter is caused by lateral resistivity variations in the crust over which the source is moving (Evans 1991; Unsworth 1994) due to water- or sediment-filled cracks, metalliferous deposits, or variations in temperature or porosity. The larger scatter in the higher-frequency 11 Hz data is consistent with this assumption.

MODELLING THE DATA

The Reykjanes Ridge data extend from source–receiver offsets of 0 km to 15 km, covering a variety of geometries. Data at ranges of less than 1 km were not analysed because they have only a limited sensitivity to crustal structure, and are affected to a greater extent by errors in source and receiver position. Although 2-D inversion is in principle feasible (Unsworth & Oldenburg 1995), existing implementations are not able to handle realistic experimental geometries and seafloor topography. Our approach is therefore to invert subsets of the data in 1 dimension using the Occam algorithm implemented for seafloor CSEM data by Flosadottir & Constable (1996), and
use the results as a guide to iterative 2-D forward modelling using the finite-element modelling code of Unsworth (1991) (described also in Unsworth et al. 1993), modified for the purpose of this study to include realistic seafloor topography.

Inversion of the data in one dimension
Data will be divided into two subsets (Fig. 3): on-axis data consisting of 0.35 and 11 Hz signals detected by Quail during tow 1, covering source–receiver offsets of 1–5 km; and off-axis data, collected with the source, receiver or both away from the AVR crest, covering source–receiver ranges of 3–15 km and consisting of signals recorded by ELF Noddy during tow 1 and Quail, Noddy and Kermit during tow 2.

On-axis data
The Occam inversion process requires an a priori estimate of the data error, but quantification of these errors can be difficult. Although geometrical errors associated with the source–receiver position and geometry can be estimated, scatter in the data caused by small-scale heterogeneity in the crust over which the source is moving is much harder to quantify (Evans et al. 1994; Unsworth 1994). We used a method of determining the appropriate level of misfit by successively reducing the required tolerance until the structure is removed from the residuals, as described in Constable & Cox (1996).

The variation of the model resulting from joint inversion of the 0.35 and 11 Hz data as the misfit is reduced is illustrated in Fig. 4. As the misfit decreases, the largest variations in the model are at depths greater than around 1 km. Shallower than this, the resistivity contours are nearly parallel to the misfit axis, implying that in this region the resistivity structure is well constrained and the steep resistivity gradient is required. Below 1 km there is less constraint, so the variation in misfit is accommodated by resistivity variations in this region. The effect of over-fitting the model is clear. The resistivity contours bend back on themselves at very low misfit, evidence of the rough and oscillatory structures that are produced.

The vertical dashed line in Fig. 4 marks the level of misfit deemed appropriate on the basis of the structure in the residuals. The corresponding model is plotted in Fig. 5 along

(a) Resistivity (Ω m)

(b) Log10(PEmax (VA m^-2))

Figure 5. (a) The models resulting from 1-D inversion of the on-axis data. Joint inversion of the 0.35 and 11 Hz data to RMS 1.3 gives the model shown by the solid line, equivalent to the resistivity depth structure along the dashed line in Fig. 4. The long-dashed line shows the model resulting from inversion of the 11 Hz data alone, which constitute the most 1-D part of the data set, and are sensitive to resistivity structure in approximately the top 1 km of the crust. The responses of these models are plotted in (b). The jitter in the response results from the varying source–receiver geometry during the tow.

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with its response, which fits to RMS 1.3. The CSEM method by itself cannot resolve sharp resistivity discontinuities, and the Occam inversion models any such discontinuities as a rapid but smooth variation in resistivity. However, the seismic model of the upper crust (Navin et al. 1998) features steep velocity gradients but no discontinuities, suggesting that the smooth electrical model is geologically appropriate. The response of the model fits the 11 Hz data and the shortest-range 0.35 Hz data well, but there is a small bias in the fit to the longer-range 0.35 Hz data. We show later that this mismatch can be resolved if the seafloor topography around Quail is included. The resistivity in the upper 10–100 m of the model is much lower than that of sea water, and is likely to be the result of the 1-D inversion trying to fit the effect of the known 2-D seafloor topography. The 11 Hz data are much less affected by topography on the scale of the AVR than the 0.35 Hz data, because the skin depth at 11 Hz is much shorter so the induction is more local. The model resulting from inversion of the 11 Hz data alone to RMS 1.5 (Fig. 5) therefore gives a better estimate of the resistivity beneath Quail to a depth of around 1 km.

**Off-axis data**

The noticeable feature of the off-axis data (Fig. 3) is the large difference in amplitude between the 0.75 Hz data recorded by Noddy during tow 2 and the 0.35 Hz data recorded by the same instrument during tow 1. It can be seen from the dashed line in Fig. 8 that the difference in frequency alone has only a small effect on the response. However there is a pronounced difference in source-receiver geometry: during tow 1, Noddy was along-strike from the source so the fields detected were predominantly radial (parallel to the line joining the source and receiver). During tow 2, the source was on the opposite side of the AVR axis from the instrument so the fields detected were predominantly azimuthal (perpendicular to the line joining the source and receiver). One would expect some level of anisotropy caused by ridge-parallel cracks and fissures, but the results of Yu & Edwards (1992) suggest that, because the transmitter is always parallel to the strike of the AVR, the effect of any anisotropic structure should be similar for both these geometries.

The large difference in amplitude between these two groups of data can be explained by the geometrical effect on the response of buried conductive layers. The magnitude of the radial fields is increased by the presence of buried conductive layers, an effect described in terms of galvanic current channelling by Unsworth (1991) or a lithospheric waveguide by Chave, Flosadottir & Cox (1990). In contrast, azimuthal fields are more strongly affected by the attenuative effects of a conductive layer. If there is any increase in the field magnitude, it is much less than that observed in the radial component. This results in a distinctive radial/azimuthal field split. Inverting the off-axis ELF data to RMS 2.2 produces the model shown in Fig. 6. As expected, the model features a downturn in resistivity at a depth of 1–5 km. The lack of off-axis short-range data means that structure much shallower than this is poorly resolved, but resistivities are considerably higher than in the on-axis model. Since a simple increase in resistivity with depth cannot explain the radial/azimuthal amplitude split in the data, it is suggested that structure varying in two dimensions at least is required to satisfy the complete data set simultaneously.

**Forward modelling in two dimensions**

We have undertaken further and more detailed analysis of the CSEM data using finite-element modelling based on the assumption that the resistivity structure is 2-D, with an axis of symmetry parallel to the trend of the AVR. The 1-D inversion

![Figure 6](image-url)
results suggest that the data set cannot be simultaneously satisfied by a 1-D model, while all viable geodynamical models of mid-ocean ridge systems indicate that we should expect large variations in structure to occur in a plane parallel to the spreading direction. Although the extent to which along-axis as well as across-axis variations in structure are important at ridges remains to some extent unresolved, a 2-D approach as a first approximation is justified by two factors. First, all of our data come from within one spreading segment, as defined by the AVR. Second, data collected on Quail during the first source tow show no variation between the signals detected when DASI was to the north and those when DASI was to the south of the instrument, suggesting that, in the vicinity of this instrument at least, there is along-axis uniformity. We will return to the question of possible along-axis variations in the deeper crustal structure in a later section.

In the 2.5-D finite-element modelling code of Unsworth (1991), the 2-D region of interest is discretized into area elements each associated with a number of nodes. The solution for the electromagnetic fields is approximated by the values at the nodes, and within each area is linearly interpolated using a series of basis functions. Unsworth (1991) examined the effect of seafloor topography of the type that is encountered at fast spreading mid-ocean ridges such as the East Pacific Rise. The seafloor was allowed to slope away from the axis at angles of between 2.5° and 10°. The seafloor topography at the Reykjanes Ridge (Figs 8a and b) is more complicated than the rather subdued topography of the East Pacific Rise. The effect of this rugged topography on the response must therefore be included in the modelling.

The resistivity structure is defined at the mesh generation stage, by assigning a resistivity to each rectangular region within a rectilinear mesh. These regions are then subdivided into triangular elements by the mesh generator itself. Using a structure definition of this sort, sloping seafloor topography can only be modelled as a series of small steps within the rectilinear mesh. This is both inaccurate and unrealistic. Wannamaker, Stodt & Rijo (1986) studied the effect of 2-D topography on MT data, and solved this problem by making the earth–air interface follow the sides of triangular elements within a finite-element mesh.

Here the problem is solved by distorting a rectilinear mesh to follow a specified seafloor topography (Fig. 7). An upper and lower level, \( z_u \) and \( z_l \) are defined, between which the mesh is distorted. The position of the row of seafloor nodes, \( z_{sf} \), is also specified. New node positions are calculated so that

\[
z_{new}^{\text{old}} = z_{new} - z_{sf}
\]

for rows of nodes above the seafloor, and

\[
z_{new}^{\text{old}} = z_{new} - z_{sf}
\]

for rows of nodes below the seafloor, where \( z_{new}^{\text{old}} \) and \( z_{new}^{\text{old}} \) refer to an undistorted part of the mesh. There must be enough rows of nodes between \( z_u \) and \( z_l \) to ensure that when the mesh is stretched, as it would be under a hill, the vertical mesh spacing does not become too wide to represent the fields accurately. Model parametrizations including topography are therefore extremely computationally expensive, especially for models containing low-resistivity zones. Because of this, it proved impossible to model 11 Hz data with the bathymetry included. A flat seafloor approximation was therefore used to explore, by forward modelling, large regions of model space. Topography was then included during the final stages of modelling to refine and validate the conclusions.

Shown in Fig. 8 is the real bathymetry along a line perpendicular to the AVR axis and passing through its centre, and the final model of ridge resistivity structure and topography. The response, which fits the data to RMS 2.3, is plotted in Fig. 8(d). The resistivity structure to a depth of 1 km on the axis of the AVR is constrained by the on-axis data recorded by Quail during the first tow. Below this, the resistivity must be increased to fit the off-axis data (the 40 \( \Omega \) m zone in the model). The resistivity outside the axial region is not well constrained by the data and is therefore chosen to be broadly consistent with the results of previous CSEM experiments (Young & Cox 1981; Cox et al. 1986; Evans et al. 1994; Constable & Cox 1996) and the borehole measurements of Becker (1985). However, a simple increase in crustal resistivity

![Figure 7](https://example.com/figure7.png)

**Figure 7.** The central portion of the mesh used to model realistic seafloor topography. The full mesh contains 3869 nodes, and extends from −25 km to 25 km horizontally and from −23 km to 2.5 km vertically to avoid edge effects contaminating the response. The node spacing is reduced in the region where the topography is varying fastest, and also at a depth of 2–3 km below the seafloor to allow accurate modelling of the low-resistivity anomaly.

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with distance from the AVR axis cannot explain the data. The dashed line in Fig. 8(d) shows the response of the model in (c) with the 1 Ω m lens and surrounding 2.5 Ω m region removed. Although the amplitudes of the data are of the right order, the large split between the predominantly radial Noddy tow 1 and predominantly azimuthal Noddy tow 2 data is not reproduced. Sinha et al. (1997) demonstrated that when the seafloor is flat, a zone of low resistivity must be included beneath the axis to explain this feature of the data. Here we show that this conclusion is not altered by the inclusion of realistic seafloor topography.

The resistivity of the mid-crustal low-resistivity anomaly is chosen to be as high as possible while still reproducing the features seen in the data with some degree of accuracy. The splitting effect between the radial and azimuthal fields is governed by the 2.5 Ω m region. Increasing the resistivity of this region decreases the degree of enhancement of the radial fields, and consequently degrades the fit of the model. Although the dimensions of this region can be constrained by the CSEM data, its exact shape cannot. Since the seismic data have a greater structural resolution than the diffusive electromagnetic fields, the shape of the 2.5 Ω m region is chosen to be coincident with the region in which the P-wave velocity anomaly is greater than −0.3 kms⁻¹ (Navin et al. 1998). However, there is little difference in the response between the anomaly shown in Fig. 8 and a simple rectangular anomaly which also satisfies the constraints on dimension discussed in the next section. A 100 m thick, 4 km wide lens of melt is included at a depth of 2.2 km below the axis, and has a resistivity of 1 Ω m, a value near the upper end of possible resistivities of a pure basaltic melt (Waff & Weill 1975). This feature is included for consistency with the structure detected by the seismic experiment only, and cannot be constrained independently of the surrounding 2.5 Ω m region. The effect of the melt lens on the response is small.

The seafloor topography produces a small downward shift in 0.35 and 0.75 Hz amplitudes, but the character of the response is governed by the crustal resistivity structure, and therefore the main conclusions of the modelling with a flat seafloor are valid. For this experiment the effect of the rugged seafloor topography is to increase the need for a zone of low resistivity below the axis. A resistivity of 2.5 Ω m should therefore be regarded as the highest resistivity in the mid-crustal anomaly that is capable of producing the required split between the azimuthal and radial data.

The main area of misfit is in the on-axis 0.35 Hz data recorded by Quail during the first tow. The data can be fitted, however, if the seafloor topography appropriate to this instrument is used. Shown in Fig. 8(a) is a bathymetric profile along a line perpendicular to the AVR axis and passing through the position of instrument Quail. The on-axis response at 0.35 Hz of the model with this more northerly topography represented is plotted in Fig. 8(d). Although computer memory limitations prevented modelling of the high-frequency data using a distorted mesh, results using models in which the topography of the AVR was represented by a simple elevated block suggest that the effect of the topography on the 11 Hz data would be minimal. In order to satisfy the entire data set simultaneously, topography varying in three dimensions would be required.

Constraints on the model

We ran a variety of models to test the limits that can be placed
on the axial resistivity structure. Of particular interest are the constraints that can be placed on the mid-crustal low-resistivity anomaly. The lowest reasonable resistivity for the anomalous body is 0.2 Ω m, corresponding to pure basaltic melt at approximately 1300 °C (Waff & Weill 1975). The effect of varying the dimensions of a molten body is shown in Figs 9(a) and (b). A melt lens 100 m thick and with a width of 4 km, consistent with the lens of very low-velocity material detected seismically (Navin et al. 1998), cannot by itself explain the data. Although there is a radial–azimuthal split for the data from Noddy, it is not as large as that observed in the data. The fit of the response is improved if the thickness of the melt lens is increased to 400 m, corresponding to one skin depth at the lowest transmitted frequency. The magnitude of the radial field enhancement caused by a low-resistivity body increases quickly as the thickness of the body increases, levelling off for thicknesses greater than a skin depth. Increasing the width of the melt lens up to 8.7 km improves the fit to the radial fields; however, the fit to the azimuthal fields is degraded, resulting in an overall increase in the misfit.

Although the data can be fitted using a 400 m thick melt lens, such a large body of melt is incompatible with the seismic data, which suggest the presence of a much smaller melt lens, surrounded by a wider low-velocity zone (Navin et al. 1998). Since there is a trade off between the size of the low-resistivity zone and the resistivity within it, a body with a higher resistivity distributed over a wider zone is also possible, and more likely given the seismic constraints. Instead of a thin body of very low resistivity, a thicker zone with a higher resistivity can be used to explain the data so long as it is at least a skin depth thick. Figs 9(c) and (d) show the effect of varying the resistivity of such a body. Although a radial–azimuthal field split starts to develop for a resistivity of 10 Ω m, resistivities of 2.5 Ω m or lower are required to reproduce the degree of splitting observed, requiring a thickness of at least 1300 m. The maximum thickness of the low-resistivity anomaly cannot be constrained with the CSEM data, although the MT results are fit best when the low axial conductance is confined to the upper

Figure 9. The effect of varying the properties of the mid-crustal low-resistivity zone. For simplicity a flat seafloor is assumed. (a) and (b) demonstrate the effect of varying the size of a melt lens. Solid line: \( W = 4 \) km, \( T = 100 \) m, corresponding to a melt lens of the size detected by the seismic data. The response fits to RMS 2.7. Short dashed line: \( W = 4 \) km, \( T = 400 \) m. An increase in the thickness of the body to 400 m, one skin depth at the lowest transmitted frequency, improves the fit to RMS 2.1. Long-dashed line: \( W = 8.7 \) km, \( T = 400 \) m. Increasing the width of the body further improves the fit to the radial, although the fit of the azimuthal response is degraded. The response of this model fits to RMS 2.6. (c) and (d) show the effect of adding a wider mush zone round the melt lens and varying its resistivity. The shape of the zone is chosen to be coincident with the zone in which the seismic velocity is depressed by more than 0.3 km s\(^{-1}\) relative to normal oceanic crust. Long-dashed line: \( \rho_{LRZ} = 10 \) Ω m, and the response fits to RMS 2.5. A split has started to develop; however, it is not large enough to explain the data. Short-dashed line: \( \rho_{LRZ} = 5 \) Ω m. The fit of the response is improved to RMS 2.4. Solid line: \( \rho_{LRZ} = 2.5 \) Ω m, for which the response fits to RMS 2.3. This is the highest resistivity for which the radial–azimuthal split is reproduced with any degree of accuracy.
few kilometres of the crust (Sinha et al. 1997).

The fit of the response is not adversely affected by varying the depth past 2.5 km results in a significant decrease in the 0.75 Hz response between 5 and 9 km source–receiver range, worsening the fit. When the top of the anomaly is shallower than 1.8 km, the low-resistivity zone severely attenuates fields diffusing through the axis, again worsening the fit at ranges between 9 and 15 km. These limits bracket the seismically determined depth to the top of the low-velocity zone.

The best results are obtained when the width of the 2.25 Ω m region is in the range 7–9 km. If the region is too narrow there is insufficient enhancement to explain the magnitude of the predominantly radial Quail tow 2 and Noddy tow 1 data. Increasing the width of the 2.25 Ω m region too far results in response amplitudes that overestimate the magnitude of the Noddy data from the second source tow.

The minimum width of the structure above the low-resistivity anomaly is constrained by the short-range data from Quail. The 11 Hz data in particular can be adequately fitted with a 1-D resistivity structure, implying that they are not affected by lateral changes in resistivity. In order for the 11 Hz data to be unaffected by the increase in resistivity with distance from the axis, the width of the axial region above the mid-crustal anomaly must be greater than 2 km. Constraint on the maximum width of the structure above the anomaly comes from examining the off-axis ELF data. Increasing the width of the shallow axial structure results in an increase in attenuation of fields diffusing through the axis. Since the fields detected by the ELF instruments have all diffused through the axial zone, this results in a decrease in all amplitudes. Similarly, decreasing the width results in an increase in the amplitudes. However, a decrease in attenuation will increase the azimuthal fields more than the radial fields, reducing the split between the two components. A balance must therefore be found between the width of the shallow axial structure, which controls the overall level of the response, and the magnitude of the sub-axial anomaly, which controls the splitting between the radial and azimuthal components. For the anomaly shown in Fig. 8, chosen to have as high a resistivity as possible, the width of the shallow axial zone is fairly well constrained at 4–5 km.

ALONG-AXIS VARIABILITY

The bathymetric map of the AVR and surrounding area (Fig. 1) shows that, although the main bathymetric features are elongated parallel to the strike of the AVR, neither the AVR itself, nor the surrounding topographic features are strictly 2-D. It was shown earlier that the broad shape of the central AVR, which has only a small effect on the off-axis ELF data, cannot explain the on-axis 0.35 Hz data, which require the more pointed topography seen at the northern end of the AVR. In addition, the MT data from instrument Kermit at the southern end of the AVR show no evidence for the low crustal resistivities that are required by data from the more northerly sites (Sinha et al. 1997; Heinson et al., in preparation).

It is therefore important to examine what the effect on the response would be if the mid-crustal low-resistivity anomaly were discontinuous. There is as yet no code available for modelling the effects of 3-D resistivity structure; however, a qualitative understanding of the effects of along-axis resistivity variation can be gained from 2-D modelling studies. The axis of the AVR is oriented perpendicular to the invariant direction of a 2-D model with the source dipole in the same direction. The effect of along-axis variability in the structure can then be assessed by examining the behaviour of the source-parallel components of the electric field.

Fig. 10 shows the effect on the response of a gap in the low-resistivity layer. Also shown are the responses of two 1-D models, one with a continuous layer of low resistivity, and one without the low-resistivity layer. The effect of the gap in the 0.2 Ω m layer is to interrupt the channelling of current, and therefore the degree of enhancement of the fields along the axis of the dipole is severely reduced. Since the low-resistivity layer can no longer efficiently channel current, its effect for a receiver on the far side of the gap from the transmitter is to attenuate fields relative to the model without the 0.2 Ω m layer. It can also be seen from Fig. 10 that the reduction in enhancement begins before the gap itself is reached. Although qualitative, these results suggest that, in order to see an enhancement effect to a source–receiver range of 15 km, the low-resistivity anomaly must be continuous, at least along the northern half of the AVR.

IMPLICATIONS FOR THE FLUID CONTENT OF THE CRUST

The overall resistivity of a two-phase medium consisting of a liquid phase and a solid phase of much higher resistivity depends on the proportion of each present and on their distribution. A medium with a high fluid content distributed in isolated pockets will appear much more resistive than a medium in which the fluid content is much lower but distributed in a connected network throughout the solid. The Hashin–Shtrikman (HS) bounds describe an upper and lower limit on the conductivity of an isotropic two-phase medium in which β is the volumetric fluid fraction (Schmeling 1986):

\[
\sigma_{HS^+} = \sigma_s + \beta \left( 1 - \frac{1}{3\alpha_l} \right)^{-1},
\]

\[
\sigma_{HS^-} = \sigma_s + (1 - \beta) \left( 1 - \frac{1}{3\alpha_l} \right)^{-1},
\]

where \( \sigma_s \) is the conductivity of the solid and \( \alpha_l \) is the conductivity of the liquid. The upper HS bound, \( \sigma_{HS^+} \), describes the conductivity of a medium in which the liquid forms a completely connected network. In the case of a liquid with a conductivity much higher than that of the solid, the conductivity of the medium is dominated by conduction within the liquid phase (Waff 1974). The lower HS bound describes the situation in which the liquid phase is included in isolated pockets, and therefore does not contribute significantly to the overall conductivity of the medium.

The shallow axial crust

In the upper oceanic crust, electrical conduction occurs predominantly within sea-water-filled cracks. Drury & Hyndman (1979) observed a decrease in resistivity of about two orders of magnitude between dry and sea-water-saturated basalt samples. The overall resistivity of the medium therefore depends strongly on the resistivity of sea water, which itself varies with temperature. At ambient deep-ocean temperatures
(approximately 2°C), sea water has a resistivity of 0.3 Ω m. This falls to a minimum of about 0.04 Ω m at 350°C and then rises slightly for higher temperatures (Nesbitt 1993). Up to temperatures of about 350°C, the resistivity of sea water, $\rho_{sw}$, at a temperature $T$ is given to a good approximation by

$$\rho_{sw}(\Omega \text{m}) = \left( \frac{3 \times 10^{-2}}{T(°C)} \right)$$

(e.g. Becker 1985). There is an ambiguity in the porosity required to explain a given measurement of resistivity, with a high porosity and ambient seafloor water temperature giving the same resistivity as a much lower porosity but higher seawater temperature (Evans 1994). Although conduction in sea-water-filled cracks is the predominant mechanism, the presence of alteration products precipitated in cracks also affects the conductivity (Drury & Hyndman 1979; Pezard 1990). Such alteration products in general have a greater resistivity than sea water, and therefore as cracks are in-filled there is an increase in the resistivity of the crust (Becker 1985; Pezard 1990).

The shallow resistivity structure of the AVR is fairly well constrained around Quail by the short-range data collected during the first source tow. The upper 700 m to 1 km of the structure are constrained primarily by the 11 Hz data, which are only slightly affected by the 2-dimensionality of the ridge. Use of the simple 1-D model is therefore valid providing the study is limited to the upper 1 km of the crust. The steep resistivity gradient in the upper 500 m of the structure is the best-constrained feature of the shallow structure, and is consistent with the steep seismic velocity gradient in the upper few hundred metres of the crust (Navin et al. 1998) equated with seismic layer 2A.

The upper HS bound can be used to estimate the porosity, taking the resistivity of the solid phase to be $10^4$ Ω m, much larger than that of sea water, so that conduction within the solid is negligible. This parametrization gives a good estimate of the volume of fluid contributing to conduction, but may underestimate the total porosity if the medium contains isolated pores. The resistivity is approximately 1 Ω m close to the seafloor. Assuming ambient ocean-floor temperatures, the porosity inferred from this measurement is approximately 50 per cent in the upper 100 m of the crust. Values of around 40 per cent are inferred from seismic P-wave velocities (Navin et al. 1998). Purdy (1987) obtained slightly lower seafloor
porosities of around 30 per cent using seismic data from the Mid-Atlantic Ridge at 23°N. Similar values were obtained by Evans et al. (1994) using CSEM data from the crest of the East Pacific Rise at 13°N. Submersible surveys reveal significant fracturing and fissuring of the neovolcanic zone, with fissure widths on the scale of centimetres to metres (Ballard & van Andel 1977). Such fissuring could account for the high porosities observed in the upper crust.

The results from the Reykjanes Ridge can be compared directly to the CSEM results of Evans et al. (1994) from 13°N on the East Pacific Rise. At 1 km below the seafloor the resistivity at the Reykjanes Ridge is 15 Ωm, five times less than the resistivity at equivalent depths on the East Pacific Rise at 13°N. Temperature/porosity trade-off curves are shown in Fig. 11 for these two bulk resistivities. If the porosity were 0.5 per cent at both ridges, a temperature difference of approximately 200 °C could explain the observed difference in resistivity. Such a temperature difference would be consistent with the lack of evidence for a large crustal melt body at 13°N on the East Pacific Rise, which is inferred to be in a state of relative magmatic quiescence (Evans et al. 1994), compared to the Reykjanes Ridge, where the RAMESSES experiment has detected a large crustal magma chamber. It is probable that the difference results from a combination of temperature and porosity effects. The seismic results (Navin et al. 1998) show a zone of depressed P-wave velocity in the upper crust on the AVR axis, which is not observed at the East Pacific Rise and could indicate increased fracture porosity at the Reykjanes Ridge. Results from Deep Sea Drilling Project hole 409, drilled in crust estimated to be 2.2–3.2 Ma on the western flank of the Reykjanes Ridge, indicate that the layer 2 basalts have a high vesicularity, averaging 27 per cent, caused by the low confining pressure during eruption at this shallow ridge (Duffield 1979).

Since a connected porosity of at most 3 per cent in layer 2 is inferred from the CSEM measurements (Fig. 11), such vesicles must be predominantly isolated or in-filled so that their contribution to conduction in the basalt is small.

### The mid-crustal low-resistivity anomaly

The most significant result of this study is the identification of a substantial low-resistivity zone beneath the axis of the AVR segment at 57°45′N on the Reykjanes Ridge. The anomaly constrained by the CSEM data is consistent in position and geometry with the low-velocity zone detected by the wide-angle seismic component of the integrated geophysical study (Navin et al. 1998) beneath the same AVR segment. It is also consistent with the high crustal conductance required to explain the MT data (Sinha et al. 1997; Heinson et al. in preparation). Resistivities in the crust that are as low as those observed, in a zone that is coincident with a significant reduction in seismic velocity, can be most easily explained if there is melt in the crust.

Laboratory measurements (Waff & Weill 1975; Tyburczy & Waff 1983) show that the resistivity of pure basaltic melt is in the range 1 to 0.1 Ωm, for temperatures between approximately 1200 °C and 1500 °C. In partially molten regions, the resistivity depends much more strongly on the resistivity of the molten phase than that of the solid, and the resistivity of melt is affected much more by variations in temperature than by other parameters such as pressure or oxygen fugacity. Since the resistivity of molten basalt increases with temperature, a low bulk resistivity in a medium can be achieved either by a large melt fraction and a relatively low temperature, or by a lower melt fraction and a higher temperature (Shankland & Waff 1977; Tyburczy & Waff 1983). As before, this makes inference of the melt fraction from a resistivity measurement ambiguous.

However, for a given observation of electrical resistivity, the possible range of temperatures and melt fractions may be examined. It has been shown experimentally that, for melt fractions as low as 3 per cent, the melt forms an interconnected network along the grain boundaries (Tyburczy & Waff 1983). The upper HS bound (eq. 5) can therefore be applied to obtain a lower bound on the possible melt fraction. Fig. 12 shows curves of constant effective resistivity in a two-phase medium consisting of a basaltic melt and a solid phase, calculated by Shankland & Waff (1977), using the experimentally determined variation of basaltic melt resistivity with temperature of Waff & Weill (1975). The resistivity of the solid phase was taken to be much higher than that of the melt so that it contributed relatively little to the conduction process. The results were calculated at zero pressure. Tyburczy & Waff (1983) showed that the resistivity of basaltic melt increases slightly with pressure for pressures up to 5–8 kbar. The pressure at 2 km depth in the crust beneath 2 km of sea water is approximately 1 kbar, and therefore including the pressure effect would increase the melt fraction required to explain an observed resistivity. The low pressure curves are appropriate for minimum melt fraction calculations.

Geologically reasonable estimates of the melt fraction required to explain an observed resistivity may be obtained by assuming that the basaltic melt in a crustal magma chamber is at a temperature between its solidus and liquidus. The solidus and liquidus temperatures of basalt, determined experimentally by Presnall, Simmons & Porath (1972) at atmospheric pressure

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**Figure 11.** The trade-off between porosity and temperature of seawater-saturated basalt for two bulk resistivity values, 70 and 15 Ωm, which are appropriate for the CSEM results from the East Pacific Rise (Evans et al. 1994) and the Reykjanes Ridge respectively, at 1 km depth below the seafloor. The curves were calculated using the upper HS bound and the variation of sea-water resistivity with temperature given in the text.

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are plotted in Fig. 12. Bender, Hodges & Bence (1978) studied the melting of basalts collected from the median valley of the Mid-Atlantic Ridge at pressures from atmospheric to 15 kbar. The liquidus temperature rises by approximately 20 °C between 0 and 8 kbar. The error introduced to the inferred melt fraction by assuming the solidus and liquidus at atmospheric pressure is therefore small.

The upper bound on the resistivity of the sub-axial anomaly is 2.5 Ω m, constrained by the CSEM data. Fig. 12 shows that for this resistivity the melt fraction is constrained to be at least 20 per cent, and may be as high as 30 per cent. Even a resistivity in the mid-crustal anomaly of 5 Ω m, which Fig. 9(d) shows is barely low enough to explain the data, would require a melt fraction of 10–15 per cent. These values are much larger than estimates of a few per cent in the low-velocity zones detected seismically at the East Pacific Rise (Caress, Burnett & Orcutt 1992; Wilcock et al. 1995), suggesting a significant difference in melt content between the crustal magma reservoirs detected at fast and slow spreading rates.

**DISCUSSION**

Despite the wide variation in mid-ocean-ridge morphology, the basic structure of mature oceanic crust is remarkably consistent, regardless of the spreading rate at which it formed (Raitt 1963). At fast and intermediate spreading ridges, crustal magma chambers have been imaged seismically, and provide an explanation of the method of formation of new oceanic crust (Detrick et al. 1987; Vera et al. 1990; Collier & Sinha 1992). Until now, there has been no evidence of similar melt accumulations in the crust beneath a slow spreading mid-ocean ridge. Thermal arguments suggest that no steady-state magma body can exist at spreading rates less than 20–40 mm a\(^{-1}\) (Sleep 1975; Kuszni & Bott 1976; Phipps-Morgan & Chen 1993). Two alternative hypotheses to explain the formation of crust at slow spreading ridges have been proposed.

In the first scenario, melt rises in small discrete packets, which feed individual seamounts and flows on the seafloor through a series of dykes (Nisbet & Fowler 1978; Smith & Cann 1992; Magde & Smith 1995). Each small melt body is likely to solidify before the next one is emplaced, with the result that a magma chamber as such never develops.

In the alternative hypothesis, melt injection into the crust is periodic. Periods of magmatic quiescence are punctuated by influxes of melt from the mantle. This view is supported by the observations of Parson et al. (1993), who used deep-towed sidescan sonar data to infer an AVR lifecycle of volcanic activity and quiescence at the Reykjanes Ridge. In this case, substantial melt accumulations may exist beneath the axes of slow spreading mid-ocean ridges, but would be highly transient features, accompanying fresh melt inputs from the mantle.

A melt fraction of 20 per cent suggests a total volume of 3 km\(^2\) of melt per kilometre beneath the AVR. A crustal thickness of 7.5 km is formed at a rate of 20 mm a\(^{-1}\) at this point on the Reykjanes Ridge, so the mid-crustal low-resistivity zone contains enough melt to feed crustal accretion at this AVR segment for 20 000 years or longer. Thermal arguments make it likely that the melt body is a transient feature. In addition, if such large magma chambers were typical features of slow spreading ridges, it is unlikely that they would have escaped detection for so long. The discovery of a melt accumulation of this size lends support to the model of cyclic crustal accretion accompanying periodic influxes of melt from the mantle at slow spreading rates.

It is probable that volcanic activity at the seafloor accompanies fresh inputs of melt from the mantle. Since a melt input of the size detected is only required once in every 20 000 years to produce the 7.5 km thickness of crust observed seismically, this implies that the period between episodes of constructive volcanism is also of the order of 20 000 years. This is of the same order as the periodicities of active volcanism on slow spreading ridges, estimated to be about 10 000 years, based on the spatial density of volcanoes in the crust and the spreading rate (MacDonald 1982). Between these times the AVRs are faulted and broken up by tectonic extension (Murton & Parson 1993).

A useful quantity to estimate is the time taken for a partially molten body of the type detected beneath the Reykjanes Ridge to solidify. It is thought that hydrothermal circulation is the dominant mechanism of heat loss, and therefore it may be assumed that the latent and specific heat of injected melt is convected upwards through the hydrothermal system (Sleep & Wolery 1978; Henstock, Woods & White 1993). If it is assumed that, at the fast spreading East Pacific Rise (EPR), crustal accretion is a steady-state process (Phipps-Morgan & Chen 1993) then the amount of heat dissipated by hydrothermal circulation is equal to the heat input by the melt. For a ridge with a spreading rate \(v\), where a thickness of crust \(t\) is formed, then the heat dissipated per unit length of the ridge in a year is

\[
\Delta H = \gamma L C_p \Delta T, \tag{7}
\]

where \(\gamma\) is the density, \(L\) is the latent heat of crystallization, and \(C_p\) is the heat capacity.
C_p is the heat capacity and ΔT is the drop in temperature as the melt cools. For a melt body of width w, the heat loss per unit area, ΔH, across the top of the body is given by

\[ \Delta H = \frac{1}{2} \rho C \frac{\Delta T}{\Delta t} \]

where \( \rho \) is the density, \( C \) is the heat capacity, and \( \Delta t \) is the time between successive influxes of melt.

In order to explain the data recorded at longer range by the ELF instruments, the resistivity must increase with distance from the axis. However, the details of the resistivity structure outside the axial region are poorly constrained. This is primarily because of a lack of short-range data collected with both the source and receiver off-axis.

Values of resistivity used in the 2-D modelling are broadly consistent with the results of previous electromagnetic experiments elsewhere on oceanic crust.

CONCLUSIONS

(1) The shallow resistivity structure on-axis is well constrained to a depth of 1–1.5 km by the short-range data recorded by Quail during the first source tow, at least in the vicinity of the instrument. Resistivities rise steeply from 1 Ω m at the seafloor, reaching approximately 10 Ω m at 1 km depth. The low resistivities suggest an interconnected porosity in excess of 40 per cent in the shallow crust on-axis, falling to below 3 per cent at 1 km depth.

(2) In order to explain the data recorded at longer range by the ELF instruments, the resistivity must increase with distance from the axis. However, the details of the resistivity structure outside the axial region are poorly constrained. This is primarily because of a lack of short-range data collected with both the source and receiver off-axis. The data are sensitive to the average off-axis resistivity, rather than to the details of its distribution.

(3) The most striking feature of the model is the presence of a large zone of anomalously low resistivity at mid-crustal levels beneath the AVR axis. This is required to explain the large difference in the amplitudes of predominantly radial and predominantly azimuthal electric fields detected by the ELF instruments at source–receiver offsets of 5–15 km.

(4) The resistivity of the sub-axial anomaly must be less than 2.5 Ω m in order to produce an adequate fit to the data. Although a melt lens similar to that imaged seismically (Navin et al. 1998) is consistent with the data, it cannot be constrained independently of the surrounding 2.5 Ω m region. The minimum thickness of the low-resistivity zone is constrained to be greater than approximately 1.3 km, corresponding to a skin depth at the lowest transmitted frequency. No maximum value can be placed on the thickness using this CSEM data set. The across-axis extent of the anomaly is in the range 7–9 km. Although the shape of the low-resistivity anomaly is not constrained by the CSEM data, a body coincident with the zone in which the seismic P-wave velocity anomaly is depressed by more than 0.3 km s⁻¹ is compatible with the data.

(5) The mid-crustal low-resistivity anomaly can be explained by the presence of basaltic melt in the crust. In order to produce a resistivity of 2.5 Ω m, the connected melt fraction in the low-resistivity zone must be greater than 20 per cent. A 7.5 km thickness of oceanic crust is formed at a rate of 20 mm a⁻¹ at this point on the Reykjanes Ridge. There is therefore enough melt in the low-resistivity anomaly to feed this crustal accretion for approximately 20 000 years. However, it is probable that the melt body will solidify in less than a tenth of this time, supporting the hypothesis that at slow spreading rates crustal accretion is a cyclic process, accompanying periodic influxes of melt from the mantle.

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REFERENCES


