

# BEYOND KTB - ELECTRICAL CONDUCTIVITY OF THE DEEP CONTINENTAL CRUST

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**Abstract.** Great strides have been made in understanding the upper part of the crust by in-situ logging in, and laboratory experiments on core recovered from super-deep bore-holes such as the KTB. These boreholes do not extend into the lower crust, and can contribute little to the elucidation of mechanisms that produce the high electrical conductivities that are commonly observed therein by magneto-telluric (MT) methods. Laboratory studies at simulated lower crustal conditions of temperature, pressure and saturation, on electrolyte saturated rocks thought to have been derived from the lower crust, have not been possible up until now due to their experimental difficulty. It is necessary to subject electrolyte-saturated rock samples to independently controlled confining and pore-fluid pressure, which implies that the rock be sleeved in some impermeable but deformable material, that can withstand the very high temperatures required. Metals are the only materials capable of being used, but these cause great difficulties for cell sealing and conductivity measurement. In this paper we describe recent breakthroughs in experimental work, specifically the development of two new types of sophisticated metal/ceramic seal, and a conductivity measurement technique that enables the measurement of saturated rock conductivity in the presence of a highly conducting metallic sleeve. The advances in experimental technique have enabled us to obtain data on the electrical conductivity of brine saturated basic, acidic and graphite-bearing rocks at lower crustal temperatures and raised pressures. These data have facilitated the comparison of MT derived crustal electrical conductivity profiles with profiles obtained from laboratory experiments for the first time. Initial modelling shows a good agreement between laboratory derived and MT derived profiles only if the mid-crust is composed of amphibolite pervaded by aqueous fluids, and the lower crust is composed of granulite that is saturated with aqueous fluids and/or contains interconnected grain surface films of graphite. The experimental data are consistent with a three layer crust consisting of an aqueous fluid saturated acidic uppermost layer, above an aqueous fluid saturated amphibolite mid-crust, and a granulite lowermost crust, which may or may not be saturated with aqueous fluids, but if not, requires the presence of an additional conduction mechanism such as conduction through thin graphite films.

**Key words:** continental crust, electrical conductivity, high pressure, high temperature

## 1. Introduction

A decade ago Earth scientists who sought to discover the secrets of the Earth's continental crust had only surface methods to resort to. The existing boreholes rarely penetrated the Earth more than a few kilometres into relatively young sedimentary basins. Since that time, a set of super-deep boreholes have been drilled, the most recent of which is the KTB borehole in the GFR. Data from in-situ logging studies in these boreholes and experiments on the recovered core at simulated conditions of pressure, temperature and saturation have led to a much greater understanding of the geological processes occurring in the upper parts of the crust.

The ability to sample the crust at great depths directly provides a snap-shot of a wide range of geochemical and geophysical properties but only at the location of the borehole. These data can, however, be used to place constraints on the structural data for a wider area obtained from surface-bound geophysical sounding techniques. The geophysical techniques for the determination of the electrical properties of the continental crust are those of magneto-telluric and controlled source surveying (Jones, 1987). These methods have revealed that the Earth's continental crust contains zones of high electrical conductivity that are inexplicable in terms of pressure reducing the porosity, and hence, electrolytic conduction through the fluid-saturated rock pores (Haak and Hutton, 1986). The cause of these zones of high conductivity has been the focus of many theories that have invoked the presence of graphite (Frost *et al.*, 1989; Mareschal *et al.*, 1992; Glover and Vine, 1992, Duba *et al.*, 1988, Duba and Shankland, 1982, Alabi *et al.*, 1975), partial melt (Hermance, 1979), hydrated minerals (Stesky and Brace, 1973), metal oxides (Parkhomenko, 1982) and brines (Marquis and Hyndman, 1992; Shankland, 1989; Hyndman and Shearer, 1989; Yardley, 1986; Shankland and Ander, 1983; Lee, Vine and Ross, 1983; Olhoeft, 1981); but the unravelling of the actual physical processes that cause the dramatically increased conductivities is still elusive. It is difficult even to make confident inferences from compilations of field and laboratory data that have been collected (Shankland and Ander, 1983; Marquis and Hyndman, 1992; Kariya and Shankland, 1983) to help understand the physical processes leading to high electrical conductivities at depth. Progress towards the solution of this problem has been hampered by two problems: firstly, our lack of accurate knowledge of the composition of the crust; and secondly the lack of relevant laboratory data on rocks derived from or thought to be derived from the lower crust. The latter is mainly due to the difficulty in attaining simulated lower crustal conditions especially on saturated samples with controlled pore fluid pressure. Great advances in experimental technique have taken place since the original ground-breaking studies of Brace, Orange and Madden (1965) and Brace and Orange, (1968), that enable us to treat the electrical conductivity of fluid saturated rocks as the consequence of many variables, not the least of which are confining pressure, temperature, pore fluid pressure and fluid composition.

The measurement of the physical properties of rocks subjected to simulated crustal conditions of pressure, temperature, saturation and deformation in the laboratory has two main purposes: (i) to provide data that can be used to place constraints on larger scale geophysical soundings, and (ii) to make the high quality measurements on rocks subjected to changes in their environmental conditions that enable the physical mechanisms that are operating in rocks at depth to be understood and modelled in a rigorous manner.

Measurements of rock physical properties from cores derived from super-deep boreholes have shed a great deal of light upon the physical properties and mechanisms occurring in the upper crust, and have produced some surprising results. The boreholes, however, only penetrate the upper crust. The KTB, Kola, and Saatly

boreholes, for example, have reached 7.2 km, 12 km, and 8.5 km (Kozlovsky, 1984) respectively. For greater depths, and greater lateral coverage at all depths, we must still rely upon surface-based geophysical remote sounding techniques and laboratory measurements which are made on rocks that we assume to be present at depth. Rock samples for such work are obtained from areas which are thought to represent exposures of the lower crust at the surface, such as the Ivrea Zone in northern Italy (Brodie and Rutter, 1987) and rare crustal xenoliths (Downes, 1993). Such rocks may be the best opportunity we have of obtaining middle and lower crustal rock samples but there remains the problem that by definition these exposed rocks are not typical of the bulk crust. This sample problem will not improve until we have yet deeper boreholes to provide material from even deeper in the crust.

The purpose of this paper is to look at the types of electrical property data that might be obtained if it became possible to extend the KTB borehole to the full depth of the crust. The measurements that are presented have been made on rock samples from the Ivrea Zone, Morocco, and Norway. They have been subjected to simulated crustal conditions of confining pressure, pore-fluid pressure, temperature and fluid saturation (Glover, 1989; Glover, Ross and Jolly, 1990; Glover and Ross, 1990). The electrical conductivity of a range of samples shows that the nature of the fluid saturating the rock pores at depth plays a key role in the control of the rock electrical conductivity (Glover and Vine, in press). The presence of aqueous fluids provides an electrolytic conduction mechanism that controls the bulk electrical conductivity in the upper parts of the crust whereas electronic mineral conduction begins to dominate as the temperature increases at depth. The presence of a carbon dioxide rich pore-fluid in the lower crust may have a very important role in the expulsion of water, isolation of remaining water, stabilisation of an anhydrous mineralogy and as a source for the deposition of highly conductive graphite films coating grain surfaces (Glover and Vine, 1992).

## 2. Electrical Conductivity in the Deep Crust

The last twenty five years have seen the widespread use of large scale geophysical sounding methods to probe the structure of the continental crust in such studies as the European Geo-traverse (Blundell, Freeman and Mueller, 1992). These studies have provided evidence that the crust is not the simple two layered structure that it was once thought to be, but has a much more complex structure varying both laterally and with depth. The three most surprising results are (1) the presence of a highly seismically reflective lower crust (e.g. Allmendinger *et al.*, 1987), (2) velocities of elastic waves in the lower crust that are significantly lower than those expected from assumed composition (e.g. Hyndman and Klemperer, 1989) and (3) the presence of high conductivity zones in the middle and lower crust (Haak and Hutton, 1986). These results seem to apply to most areas of the continental crust and are not explained by current hypotheses. There needs to be a theory to explain these findings and correlate them with the seismicity, rheology and heat-

flow in the crust. There is a profusion of theories that purport to do so but they do not explain all the observed physical properties. This may be because there has been insufficient recognition of the lateral variability of the crust, and of the fact that the cause of electrical conductivity in the upper, middle and lower crust and from location to location is also likely to vary i.e. electrolytic, or electrolytic and electronic through the rock matrix minerals, or graphite and electronic through the rock matrix minerals.

The remainder of this paper will concentrate upon the high conductivity zones present in the continental crust and their elucidation. Firstly, it is important to review the limitations of the geophysical sounding measurements from which these zones are deduced. The main method is that of magneto-telluric surveying (Kaufmann and Keller, 1981) although some work has been done with controlled source methods (Thompson *et al.*, 1983) and the analysis of three component magnetic field variations. The main draw-back with these methods is that it is very difficult for them to resolve whether layers in the deep crust are very conductive and thin or are moderately conductive and thicker. The parameters that the methods measure with the greatest accuracy are the depth to the top of a more conductive layer and the thickness-resistivity ratio of the layer. The limitations are described further in a paper by Jones (1987).

Whether high conductive zones in the continental crust existed at all was a subject for debate until improved experimental and inversion techniques produced data that were incontrovertible. The presence of continental high conductivity zones is now well accepted, for instance Hyndman and Shearer (1989) have said: "... with modern data, little doubt now remains that the lower crust is commonly conductive". A recent paper by Marquis and Hyndman (1992) has made a compilation of magneto-telluric data that shows crustal high conductivity zones to be ubiquitous. Figure 1 shows some typical electrical conductivity inversions made from magneto-telluric measurements in the U.K., by way of an example (Beamish, 1984). Analysis of such data shows that the highly conductive zones have variable characteristics, leading to the possibility that they are caused by differing mechanisms in different locations.

These data contradict the expected scenario, which involves the decrease in electrical conductivity with depth as rocks undergo pressure-induced compaction reducing the connectivity of conductive pore-fluid pathways until deep levels are reached at which conductivities recover due to high temperature, electronic conduction through the mineral matrix. There have been many theories that have sought to explain this discrepancy some of which have already been mentioned in the introduction. It is likely that many of them will have some input into the explanation of the high conductivity in each particular area. Laboratory experiments on rocks thought to be derived from the deep crust at conditions of confining pressure, pore-fluid pressure, temperature and saturation present at those levels can, in principle, indicate whether they are capable of providing the observed high conductivities, investigate the mechanisms of conduction, and suggest which are

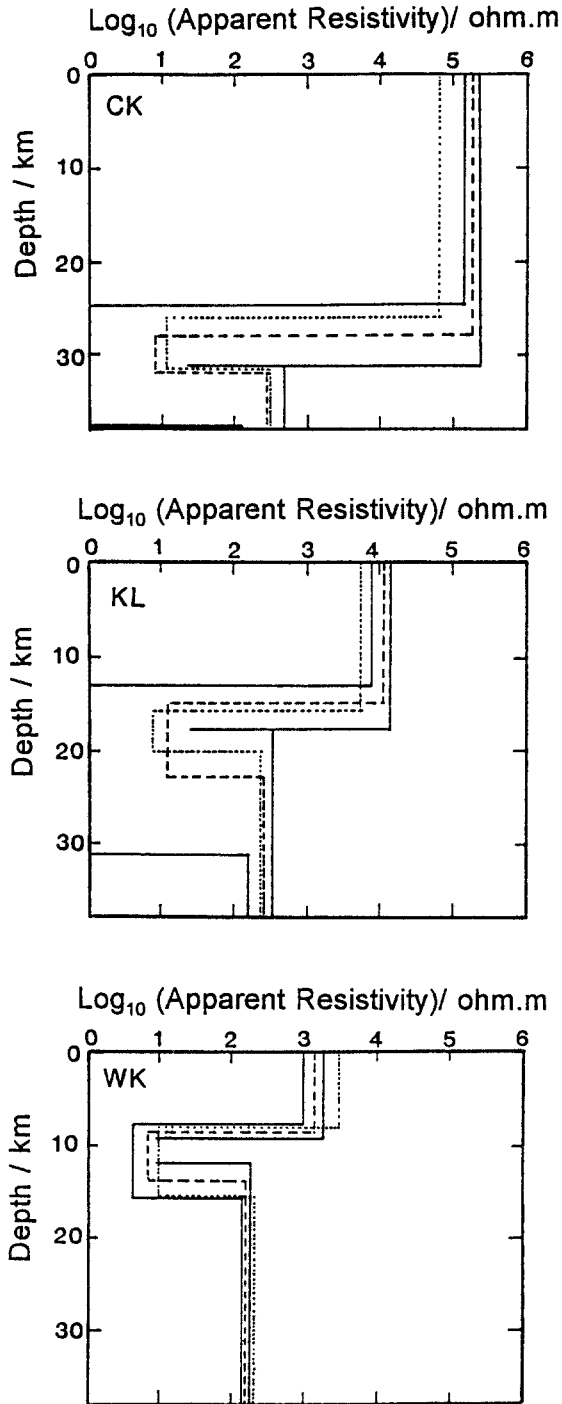


Fig. 1. Resistivity/depth profiles for three areas in northern England (After Beamish 1986).

most likely to cause the observed high electrical conductivity zones. It is extremely important that these rocks should be saturated with brine as the results from all the deep bore-holes show active, high salinity fluids present in the crust, and laboratory measurements on dry rocks show that dry rocks cannot provide sufficient conductivities.

The arrangement of a cell to measure the electrical conductivities of electrolyte saturated rocks at very high temperatures and pressures is very difficult. We have, however, completed these experiments on a range of rocks thought to be representative of the crust (Glover, Vine and Ross, 1990; Glover, Ross and Jolly, 1990; Glover and Ross, 1990; Glover and Vine, 1992; Glover and Vine, in press). The results, presented in the remainder of this paper, are unique and enable an electrical pseudo-section of the continental crust to be constructed from them. This section seems to agree well with the results of the geophysical sounding methods.

### 3. Experimental Methodology

Measurement of the electrical conductivity of electrolyte saturated rocks at temperatures up to 900 °C raises some particularly difficult experimental problems. Such experiments require that the saturated rock sample is sleeved in a deformable jacket. This enables the sample to be placed under confining pressure as well as allowing the pressure of the fluids occupying the rock pores to be raised and controlled independently. It also serves to separate the confining pressure medium (high pressure argon) from the electrolytic pore-fluid saturating the sample. In the past experimental studies on saturated rocks have been made up to about 270 °C as this is the temperature limit of the high temperature plastics that were used as sleeving material. The extension to lower crustal temperatures has required the abandonment of this method and the adoption of a metal sleeve. The sleeve was made from 250  $\mu\text{m}$  thick stainless steel tube machined directly from stainless steel bar. The use of a metal sleeve results in the presence of large leakage currents through the sleeve which would make the true rock electrical conductivity unmeasurable if conventional two or four electrode measurement techniques were used. The errors induced by the presence of sleeve leakage currents were reduced to an acceptable level by the use of a guard-ring electrode arrangement and setting the electrodes in cold isostatically pressed pure recrystallised alumina.

The measurement arrangement is shown schematically in Figure 2. It consisted of an emitter electrode, collector electrode and a guard-ring electrode. A conventional two electrode conductivity measurement is shown in Figure 2a. The leakage currents in this set-up were approximately 3 times that flowing purely through the rock for an average electrolyte saturated rock. Their inclusion in the conventional measurement would have resulted in a gross overestimation of the rock conductivity. The guard-ring configuration shown in Figure 2b collects the leakage currents and disposes of them to earth without their inclusion in the conductance measurement. The measured conductance is that of a barrel shaped volume within

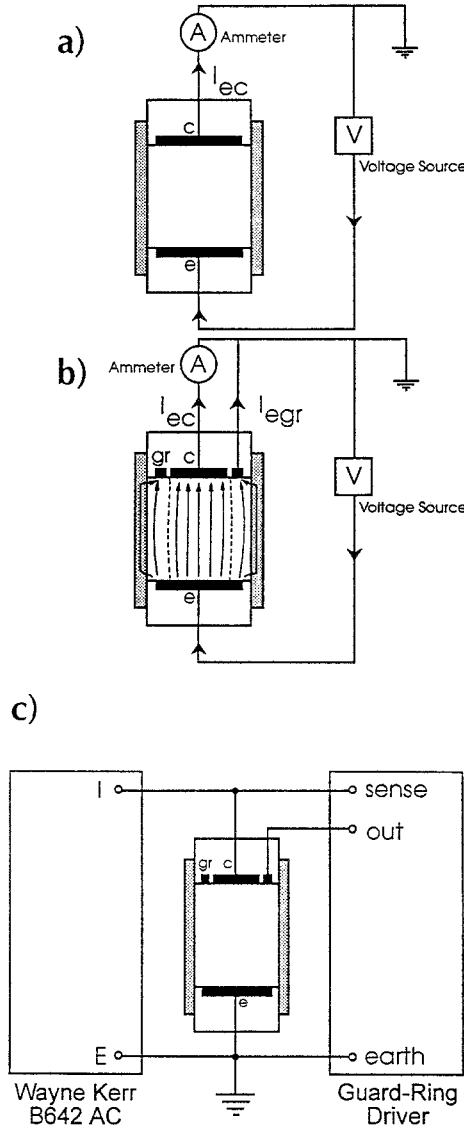


Fig. 2. Schematic diagrams showing the use of the guard-ring technique for making electrical conductivity measurements; a) measurements made with a conventional electrode system with a highly conducting sleeve, b) measurements made with a guard-ring electrode system and a highly conducting sleeve, c) the practical measuring arrangement. Notation; e, emitter; c, collector; gr, guard-ring;  $i_{ec}$ , current flowing from emitter to collector;  $i_{egr}$ , current flowing from emitter to guard-ring,  $V_{egr}$ , instantaneous potential difference between the emitter and guard-ring, and  $V_{ec}$  instantaneous potential difference between the emitter and collector.(After Glover and Ross 1990).

the sample which does not include the sleeve or parts of the rock that are close to the sleeve and containing current paths that are distorted by the presence of the conducting sleeve. The rock conductivity can be obtained from the measured

conductance of this barrel shaped volume by calibration against a cell of simpler geometry at low temperatures and by numerical modelling at higher temperatures. The criterion that needs to be fulfilled to ensure effective removal of the leakage currents is that the instantaneous potential between the guard-ring electrode and the emitter electrode must be identical to the instantaneous potential between the collector electrode and the emitter electrode (i.e.  $V_{ec} = V_{egr}$  at all times in the AC cycle, for all frequencies). A special guard-ring driving circuit had to be designed to provide this matching, while ensuring that the guard-ring connection had a sufficiently low impedance to accept a large leakage current, and the collector connection had a sufficiently high impedance not to remove the current that had flowed through the measured portion of the rock. The final arrangement is shown in Figure 2c (Glover, 1989; Glover, Vine and Ross, 1990; Glover, 1989; Glover, Ross and Jolly, 1990; Glover and Ross, 1990).

The majority of measurements were made with a Wayne Kerr B642 AC transformer bridge at a frequency of 1592 Hz. Some measurements were made across a range of frequencies at low pressures and temperatures to check that the measurements were not being affected by electrode polarisation (<100 Hz) or capacitive coupling with the pressure vessel (>200 kHz). The guard-ring driving system used in this work automatically compensates for stray conductances and AC fields. This avoids the problem of guard-ring synchronisation drift that has been a difficulty encountered by researchers using a manually balanced guard-ring system (Cemic and Jansen, 1975).

The experimental cell was designed to take samples 21.6 mm in diameter and 27.8 mm long. Two new types of seal were used. The first is a matched expansion self locking taper seal and was used to seal the electrical connections and pore fluid connections to the ceramic electrode holders. These were based upon a self-locking taper seal design and were made by finely diamond grinding the surface and grommets with a fine copper powder that was oxidised by heating in air. The metal part of each seal was made from niobium as this has a coefficient of linear expansion that matched that of the ceramic. The second is an swaged annular ring-groove seal and was used to lock the metal sleeve to the ceramic electrode-holder. This seal is initially made by swaging the thin metal sleeve into an annular groove in the ceramic. The swaging is done by inserting the sleeved ceramic into a slowly rotating lathe and using a wheel tool to gradually press the metal into the groove. A copper circllet is subsequently laid in the resulting metal groove and held in place by an arrangement of metal jaws and a collar. The thermal expansivities of all parts of the seal are arranged so the seal becomes tighter at high temperatures. The cell is shown in Figure 3 (Glover, 1989).

The cell was arranged in a furnace and attached to the pressure vessel end-cap (Figure 4). The experiments were carried out by remote control in a 6 litre hydrostatic gas pressure vessel held within a large safety room with 6 cm thick steel blast walls.



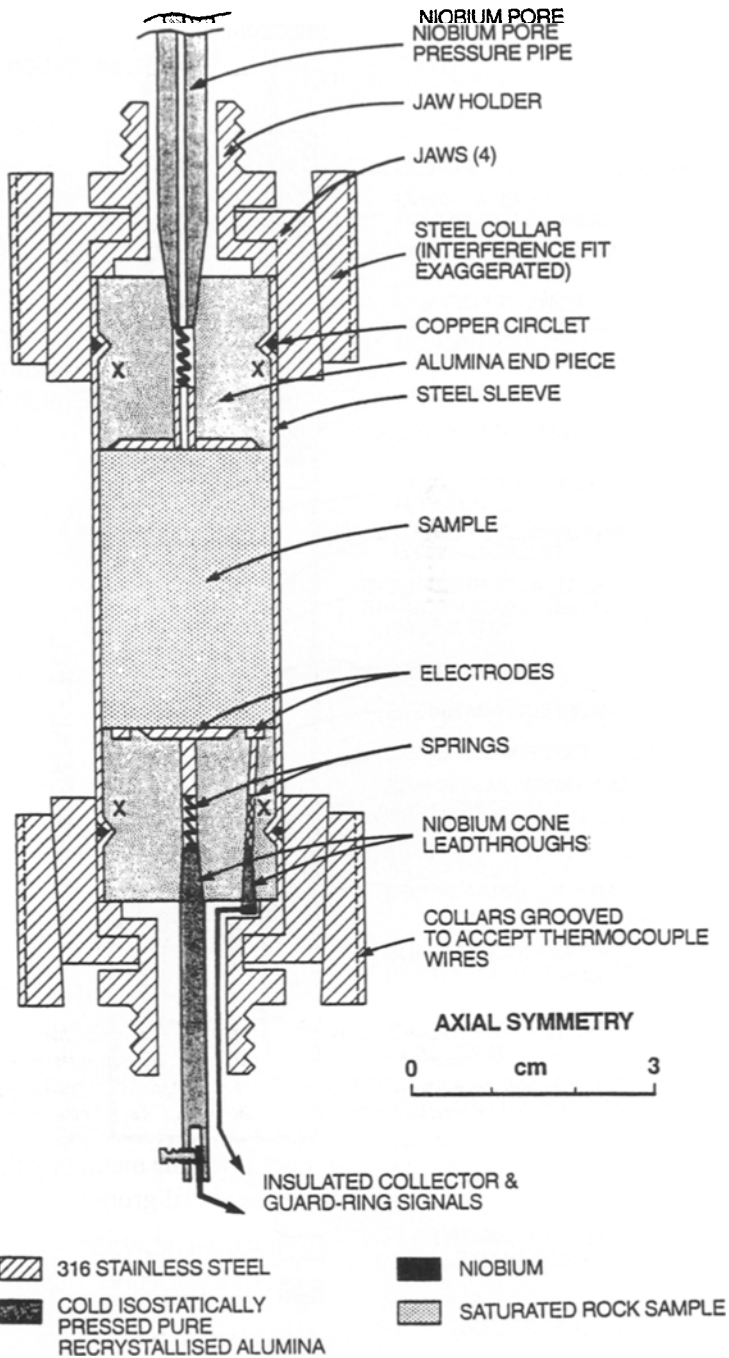


Fig. 3. The experimental cell (After Glover and Ross, 1990).

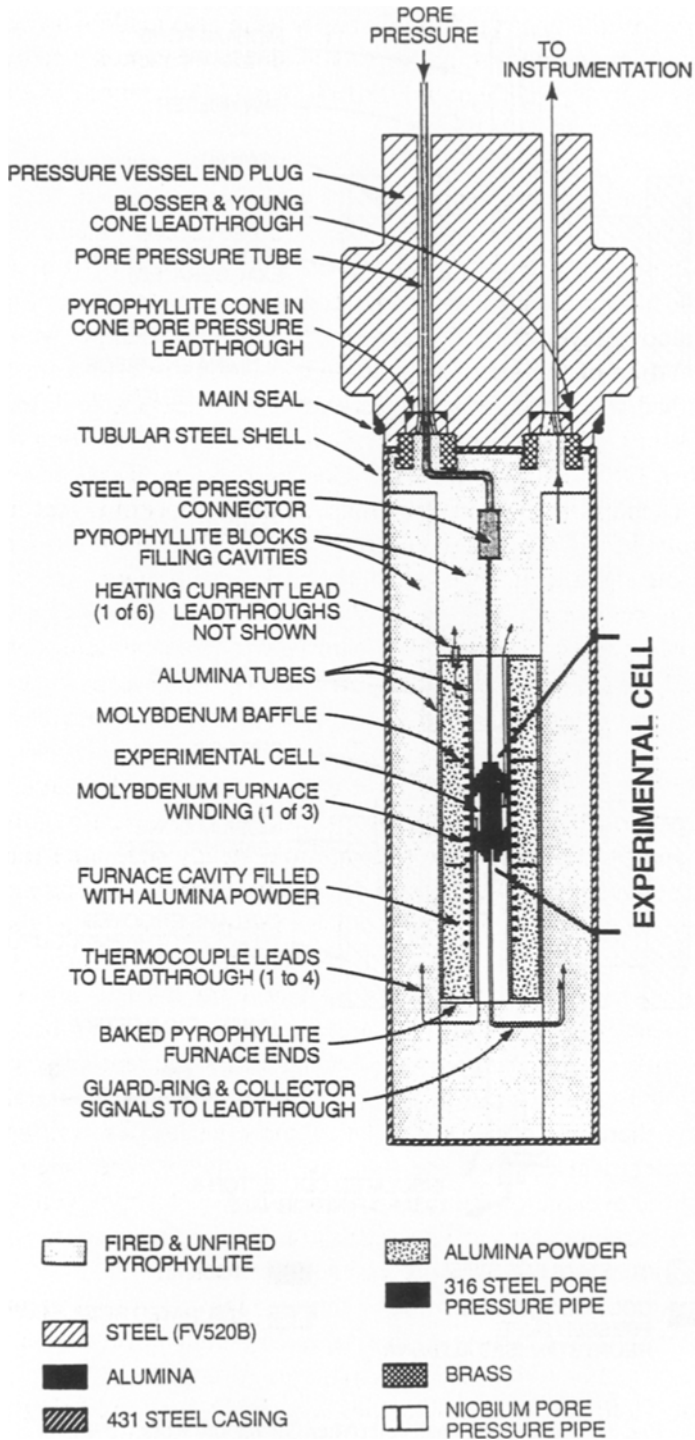


Fig. 4. The arrangement of the experimental cell inside its furnace, attached to one end plug of the pressure vessel and ready for loading (After Glover and Ross 1990).

Some of the results presented in this paper were also made with conventional two and four electrode measuring systems at single frequencies or over a range of frequencies at lower temperatures and pressures. Differences in experimental method will be noted with the data.

It has been noted that the use of stainless steel in this work could lead to measurements that are unrepresentative of real rocks at depth. The electrodes were made from stainless steel which has good corrosion resistance, and very little corrosion of the electrodes was noted even when it had been in contact with electrolytes at high pressure and temperature for up to 24 hours. The presence of a thin insulating layer of oxide between electrode and sample would result in an underestimation of the sample conductivity. Comparison of a fresh electrode with one that had been used for a full test showed no measurable difference when measuring a saturated granite at laboratory temperatures, indicating that the thin oxide layer has little effect in this case. Despite these findings, a new electrode was used for each of the experimental runs. Another concern is that iron leached by the electrolyte from the electrodes and jacket will react with minerals in the rocks and produce phases not typical for the rock type, temperature and pressure. If this were the case one might expect the conductivity to progressively change with time, however, once the saturated sample pore microstructure had come to an initial steady state, it was found that the electrical conductivity as a function of temperature and pressure were extremely repeatable. Additionally, observation of the sample after a run showed no difference in mineralogy either near to the electrodes and jacket, or on mineral surfaces with respect to their interiors, which one might expect if reaction with externally generated iron were a problem.

The rock and fluid require time to come to a steady state after the sample is taken to a new temperature or pressure, as the pore micro-structure stabilises at the new conditions. The samples were always left to attain a steady state before conductivity measurements were taken, in practice this meant invoking the criterion that there was less than a 1.5% change in electrical conductivity in five minutes. The confining pressure and pore-fluid pressure were raised and then the temperature was independently cycled in a step-wise manner until three repeatable curves were obtained with no hysteresis; this occurred in the first three cycles for all the rocks studied. It has therefore been assumed that the criterion for waiting for pore-structure and electrical conductivity stability is sufficient. The data presented in this paper are the average of the three experimental cycles that were found to be repeatable; i.e. measured only after all transient changes in conductivity had taken place and the rock was in micro-structural equilibrium.

The rock samples subjected to high temperatures and pressures in this work were all saturated with electrolyte. The saturating fluid in each case was 0.5 M aqueous NaCl based on the fluid measured from inclusions of deep crustal origin (Crawford and Hollister, 1986). The consequence of the rock being completely saturated with aqueous fluids, is that the electrolyte controls the oxygen fugacity of the system, as well as buffering any iron leached from the stainless steel electrodes

and jacket. Comparative observation of samples before and after being subjected to high temperatures and pressures showed very little additional oxidation.

#### 4. Results

The electrical conductivity of electrolyte saturated rocks is usually considered to be primarily due to conduction through the electrolyte at low temperatures (up to 200 °C). At higher temperatures electronic conduction through the rock forming minerals making up the rock matrix becomes important in some rocks. The following results show clearly which saturated rocks are the better conductors at high temperatures and at what temperature their electronic conduction becomes significant. The results also demonstrate that although the electrical conduction through the pore fluid increases with temperature, it reaches a peak and then declines with increasing temperature after a peak at about 350 °C. Other sources of increased conductivity are conduction through accessory minerals such as metal oxides and sulphides, graphite films, surface conduction on all minerals but especially phyllosilicates and, at very high temperatures, partial melting. When considering the electrical conductivity of the crust or of saturated samples in laboratory measurements, it is necessary that one considers all these possible sources of conduction and all possible mechanisms for the variation of their contribution to the bulk conductivity.

##### 4.1. ELECTROLYTE SATURATED ACIDIC ROCKS

The variation of the electrical conductivity of electrolyte saturated rocks with applied pressure is usually ascribed to confining pressure reducing the porosity of the rock during compaction. The pore-fluid pressure can mitigate against this closure if sufficiently high. Figure 5 shows the variation of the electrical conductivity of saturated rocks with confining pressure. The sensitivity of the electrical conductivity to increases in confining pressure is ascribed to the geometries of the cracks and pores in the rock; long thin cracks close under pressure much more easily than more rounded pores. The shape of these curves can, therefore, be used to gain a quantitative understanding of the pore shape or aspect ratio of the cracks as they are closed during rock compaction by the use of the appropriate inversion technique.

The progressive closure of pores also has the effect of reducing the overall connectivity of the electrolyte conduction paths and, hence, increasing their tortuosity. The reduction in electrical conductivity due to this mechanism may be very important in some rocks, especially those for which the majority of their porosity resides in the form of pores with a small range of aspect ratios. If these are connected by narrow pore throats that can close under pressure this will reduce the pore connectivity and resulting electrical conductivity.

Bulk rock electrical conductivity decreases with the application of increasing confining pressure until a limiting value of electrical conductivity occurs. This value

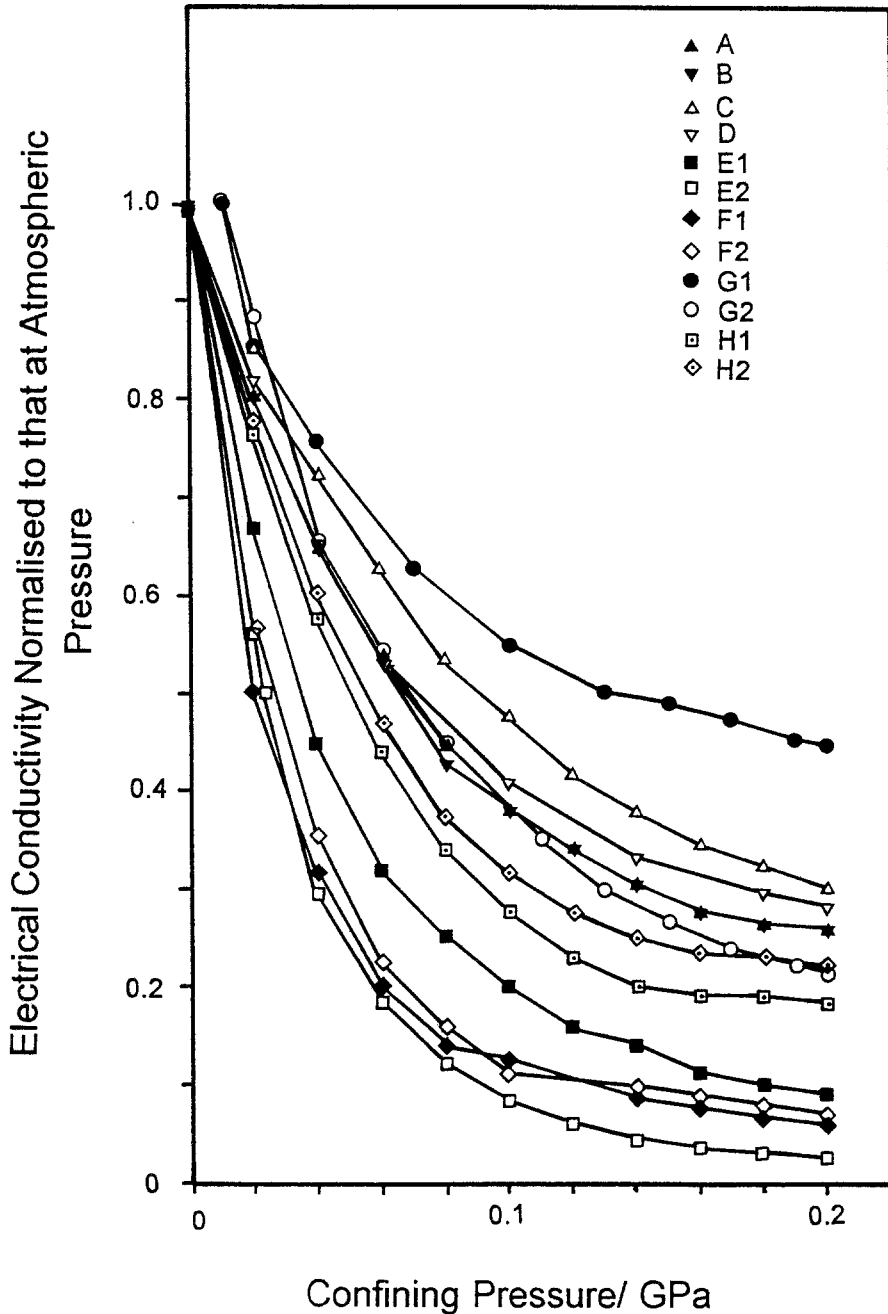


Fig. 5. The electrical conductivity normalised to that at atmospheric pressure of a range of rock samples which were measured saturated with 0.5M NaCl brine [except where noted] when subjected to raised confining pressures up to 0.2 GPa. The size of symbols indicates the approximate error associated with each individual reading. A: Granite (0.5% porosity), B: Granodiorite (0.7%), C: Amphibolite gneiss (1.02%), D: Amphibolite gneiss (0.89%), E1: Basic amphibolite (0.68%), E2: Basic amphibolite [Saturated with Pure Water] (0.65%), F1: Basic 2-pyroxene granulite (0.78%), F2: Basic 2-pyroxene granulite [Saturated with Pure Water] (0.72%), G1: Graphite-bearing granulite (0.88%), G2: Graphite-bearing granulite (0.93%), H1: Orthopyroxenite (0.4%), H2: Orthopyroxenite (0.39%) [Saturated with Pure Water].

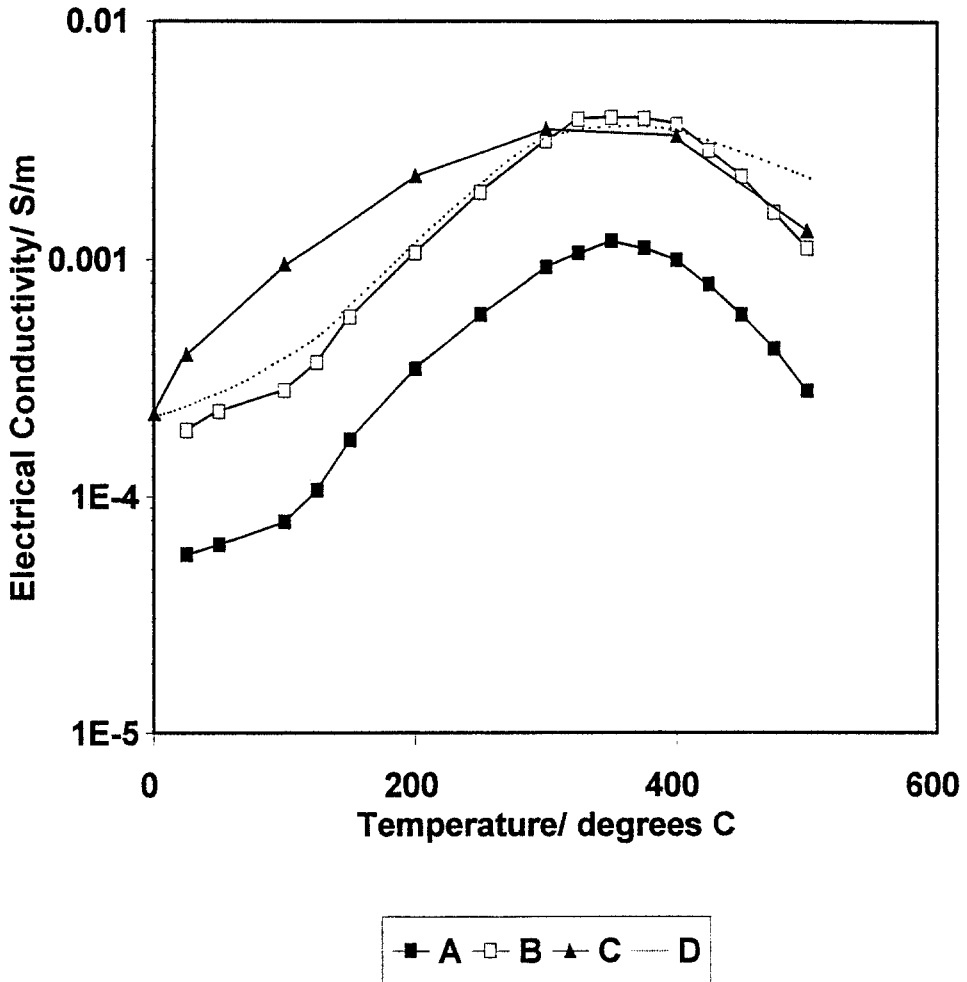


Fig. 6. The electrical conductivity of a range of rock samples which were measured saturated with 0.5M NaCl brine when subjected to a constant raised confining pressure (0.2 GPa), pore fluid pressure (0.18 GPa) and temperatures (up to 500 °C). The size of symbols indicates the approximate error associated with each individual reading. A: Granite (0.5%), B: Granodiorite (0.7%) obtained from the Ivrea Zone, N. Italy. Profile C is the electrical conductivity of a model rock with conduction occurring through a 0.5M NaCl pore-fluid with no matrix conduction. It has been generated using the empirical relationship by Archie (1942):  $\sigma_{rock}(T) = \sigma_{fluid}(T)\phi^n$ , where  $\sigma_{fluid}(T)$  is the electrical conductivity of the electrolyte fluid (obtained from Quist and Marshall, 1968),  $\phi$  is rock porosity (arbitrarily chosen as 0.008) and the exponent  $n$  is 2. Profile D has been adjusted to account for the porosity loss at  $T < 250$  °C due to mineral grain expansion and new cracking at temperatures greater than 250 °C.

is higher than the electrical conductivities associated with mineral conduction and does not, therefore, represent the case of a completely pore and crack free rock due to compaction. This effect has been explained by Gavrilenko and Gueguen (1989) by considering the observed electrical conductivity limit to represent the conductivity when asperities on the surface of cracks come into contact. Further increases in confining pressure lead only to small changes in the remaining porosity as the asperities are crushed and the remaining electrical conduction paths become more tortuous, but the pore space remains open.

The variation of electrical conductivity in saturated rocks with temperature becomes more important than the value of confining pressure at depth, partly due to the limiting nature of the confining pressure variation as mentioned above and partly because the electrical conductivity of pore fluid conductivity increases strongly with temperature up to 350 °C. The variation of the electrical conductivity of NaCl solutions with temperature is a characteristic singly peaked curve with the maximum occurring at about 350 °C at low fluid pressures (0.1 to 0.3 GPa) (Quist and Marshall, 1968). At higher fluid pressures the decrease in electrical conductivity past the peak becomes progressively less marked. It would be expected that a saturated rock, with no other form of conduction mechanism, would mirror the behaviour of the saturating electrolyte. The difficulties in making this type of measurement had resulted in this hypothesis only being tested to 270 °C (Lee, Vine, and Ross, 1983). Recently, however, measurements have been extended to higher temperatures (Glover and Vine, 1992; Glover and Ross, 1990; Glover, Ross and Jolly, 1990; Glover and Vine, in press). The results show a clear single peak variation in the acidic rocks studied up to 500 °C (Figure 6). The peak in bulk rock electrical conductivity occurs at about  $350 \pm 10$  °C which agrees well with the behaviour of the electrolyte alone. The electrical conductivity of the acidic rocks studied under relatively low pore fluid pressures (0.18 GPa) is, therefore, consistent with the rock acting as an insulating matrix with conduction being entirely attributable to the saturating fluid. The maximum electrical conductivity of these rocks ( $3\text{--}4 \times 10^{-3} \text{ Sm}^{-1}$  at 350 °C) is insufficient to explain the high conductivity zones ( $0.02$  to  $0.1 \text{ Sm}^{-1}$ ) even if the fluid pressure was sufficient to ensure that the post peak decrease in electrical conductivity did not occur.

The shape of the electrical conductivity variation of saturated acidic rocks with temperature differs slightly from that of the pore-fluid. This has been attributed to the effect of mineral expansion reducing the porosity and hence electrical conductivity at temperatures up to 200 °C, with confining pressure, and pore-fluid pressure held constant at 0.2, and 0.18 GPa respectively. At higher temperatures mineral expansion continues but the thermally induced stresses are sufficiently large to cause the formation of new cracks which balance the effect of the porosity loss due to the initial mineral thermal expansion. The way in which thermal expansion of mineral grains and the development of thermal stresses affects porosity will be addressed in more detail later in the paper.

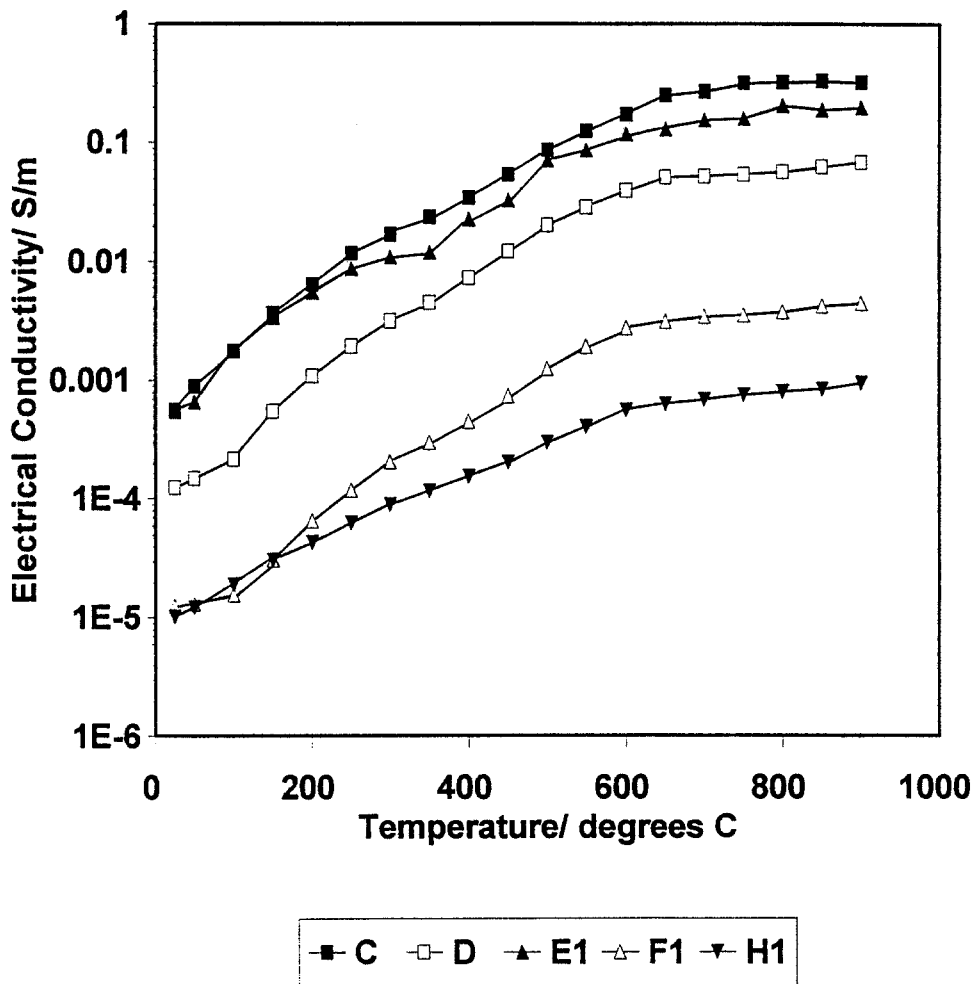


Fig. 7. The electrical conductivity of a range of rock samples which were measured saturated with 0.5 M NaCl brine when subjected to a constant raised confining pressure (0.2 GPa), pore fluid pressure (0.18 GPa) and temperatures (up to 900 °C). The size of symbols indicates the approximate error associated with each individual reading. C: Amphibolite gneiss (1.02% porosity), D: Amphibolite gneiss (0.89%), E1: Basic amphibolite (0.68%), F1: Basic 2-pyroxene granulite (0.78%), H1: Orthopyroxenite (0.4%). Rock types E1, F1 and H were obtained from the Ivrea Zone, N. Italy, all others were from the Lofoten-Vesteralen area, N. Norway.

#### 4.2. ELECTROLYTE SATURATED BASIC ROCKS

The variation of the electrical conductivity of basic rocks with confining pressure was found to be similar to that of the acidic rocks (Figure 5). The variation of



electrical conductivity with temperature, however, showed some clear differences. The measured electrical conductivity of electrolyte saturated basic rocks behaved in a similar manner to that of the saturated acidic rocks up to about 300 °C but then diverged. The conductivity of the basic rocks became ever larger with further increases in temperature whereas that of the saturated acidic rocks was declining (Figures 6 and 7). This is clear evidence for additional, non-electrolytic, conduction mechanisms becoming dominant in basic rocks post-350 °C. This is mainly due to trans-mineral electronic conduction which increases rapidly with temperature at temperatures greater than 300 °C in many semi-conducting minerals (Toussaint-Jackson, 1984). It is particularly effective in hydrous mineralogies such as those present in the basic amphibolite (Stesky and Brace, 1973). There may also be some contribution to the conductivity from clay minerals coating pore surfaces (Etheridge *et al.*, 1984). Conduction in these clays can involve ions moving over the large surface areas of their crystals and electronic conduction through the body of the clay. The latter mechanism may be very important in certain rocks with grain surface clay deposits, especially if the pores are closed by the overburden pressure leaving only clay conduction pathways where formerly connected cracks existed (Toussaint-Jackson, 1984). Further discussion of the processes leading to electrical conduction in these rocks will be left to the discussion of the electrical conductivity of the Phanerozoic crust.

#### 4.3. GRAPHITE-BEARING ROCKS

Measurements on graphite-bearing rocks that might be present in the continental crust deserve special treatment for three reasons. Firstly, there have been very few experimental measurements made on such rocks (e.g. Glover and Vine, 1992; Duba *et al.*, 1988). Secondly, the mechanisms of conduction through graphite in rocks enable the bulk electrical conductivity to be very high even with very small proportions of graphite in the rock by weight. Graphite films of only 1000 Å thickness have been discovered by auger spectroscopy in otherwise apparently graphite-free rocks (Frost *et al.*, 1989; Mareschal *et al.*, 1992) and it has been calculated that graphite films of this order of thickness (Duba and Shankland, 1982) can produce bulk electrical conductivities in the range of those observed in the lower continental crust providing the films are well inter-connected and the grain size of the rock is less than 10mm (Frost *et al.*, 1989). Thirdly, the occurrence of graphite in deep crustal rocks may have been greatly underestimated in the past due to the lack of rock samples from the deep crust and the fact that the rocks we do have may contain unrecognised graphite films (the rocks in which graphite has been identified by Frost's team were apparently graphite-free when analyzed with conventional techniques). There are several other clues that point to the Earth having a greater proportion of carbon than has traditionally been assumed. It has been noted that carbonaceous chondrites, which commonly contain a few percent carbon, have a chemical composition which otherwise strongly mirrors the Earth's bulk composition (Taylor and McLennan, 1985). Carbon also occurs naturally in

rocks in the form of graphitic shales and sandstones and has been noted in lower crustal xenoliths (Downes, 1993).

We have made electrical conductivity measurements on saturated samples of graphite-bearing granulite rocks using the guard-ring electrode method at temperatures up to 800 °C, confining pressures up to 0.2 GPa and pore-fluid pressures up to 0.18 GPa, and with a conventional electrode arrangement on unsaturated samples up to 0.4 GPa at laboratory temperatures (Glover and Vine, 1992). Measurements were made on samples cut parallel and perpendicular to the foliation and main graphite distribution. The graphite-bearing rocks came from the Beni Bousera massif in Morocco which is known to contain rocks with significant proportions of graphite (Nicolas, 1985).

There are two important results from this study; firstly that graphite films ruptured by decompression as the rock was exhumed will reconnect when re-subjected to the pressures that they were at in the deep crust, resulting in an increase in the electrical conductivity, and secondly that these graphite-bearing rocks had very high electrical conductivities at temperatures corresponding to the middle and lower crust.

The overall reduction of conductivity due to the application of pressure can be lessened if there is significant conduction through the rock matrix (Hyndman and Shearer, 1989; Drury and Hyndman, 1979). The electrical conductivity of the electrolyte-saturated graphite-bearing granulite samples (G1, G2) decreased with the application of confining pressures as expected. The decrease was, however, less than that for the graphite-free granulites, F1 and F2 (Figure 8a). This can be explained if the graphite-free granulites have a higher porosity, weaker matrix or a greater proportion of long thin cracks than the graphite-bearing granulites. The opposite was the case: the graphite-bearing granulite samples had the higher porosity and a more friable nature. The highly foliated nature of the carbon-bearing granulites also suggested that a large proportion of their porosity was in the form of long thin cracks rather than less compressible, more rounded pores. One would expect that the presence of flat thin cracks in the carbon-bearing granulites would lead to a greater sensitivity of the electrical conductivity to applications of pressure than that observed, if the pressure variation were solely due to electrolytic conduction through the pore fluid. In terms of these three points, it is clear that sample pore-space/matrix properties alone cannot account for the observed differences in the electrical conductivity variations.

The conductivity of unsaturated samples (Figure 8b) showed clearly that the effect noted in the saturated samples was caused by some conduction mechanism that increased the bulk rock conductivity on the application of pressure. The conductivity of unsaturated carbon-bearing samples (G3, G4) at near-surface conditions was less than that when saturated with electrolyte as might be expected, but increased with the application of confining pressure. The graphite-free samples (F3, F4) showed lower conductivities than the saturated case that underwent small decreases due to the effect of pressure on the small amounts of aqueous fluids

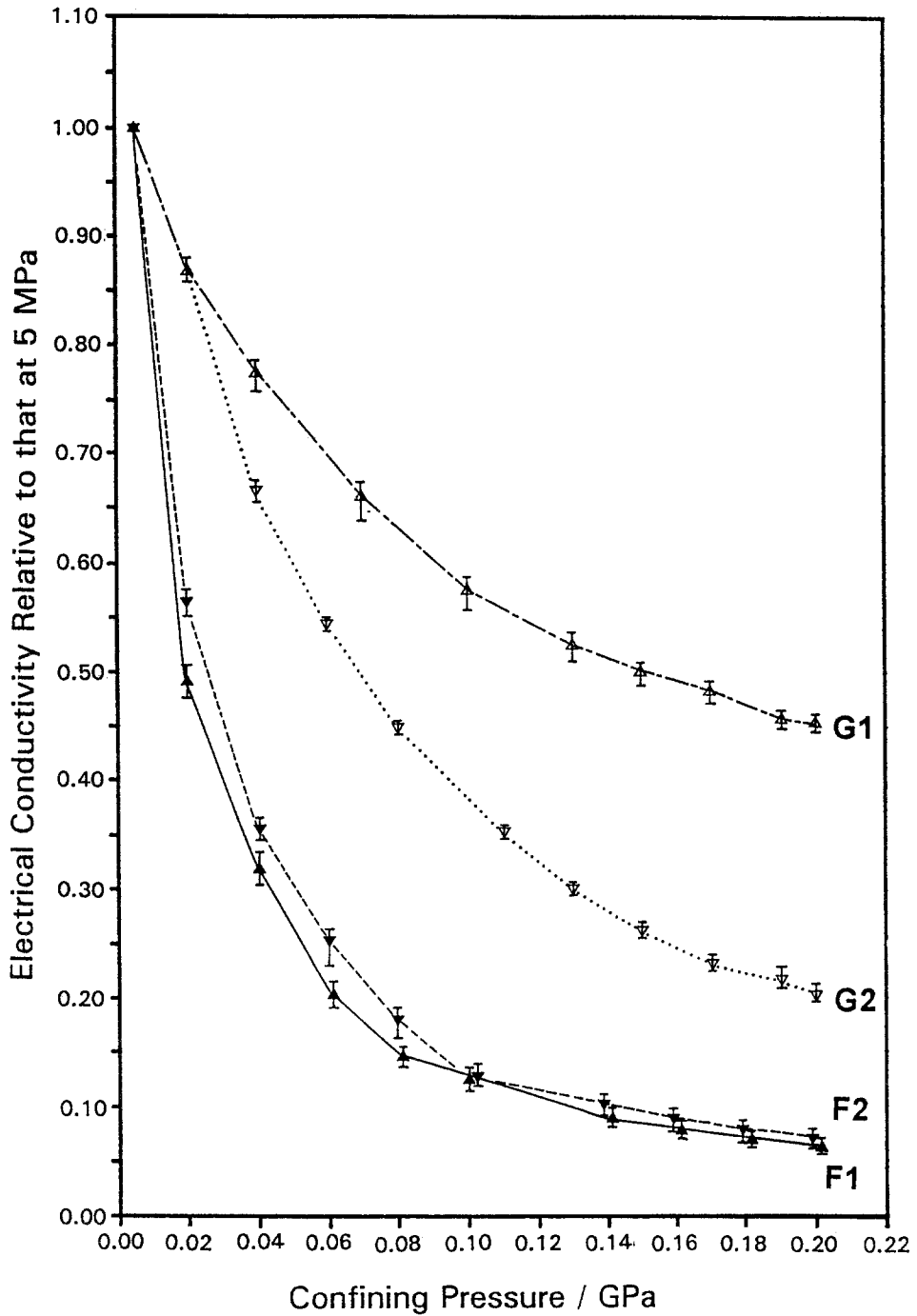


Fig. 8a. The variation of the electrical conductivity of water and brine saturated carbon-free and carbon-bearing granulites when subjected to raised confining pressures. G1/G2 – carbon-bearing, electrolyte saturated, measurement direction parallel and perpendicular to planes of foliation respectively. F1/F2 – carbon-free, no foliation, saturated with electrolyte and distilled water respectively. The data are normalised to that at 5 MPa; absolute values for each of the samples at this pressure were as follows: G1,  $5.01 \times 10^{-4} \text{ Sm}^{-1}$ ; G2,  $2.51 \times 10^{-4} \text{ Sm}^{-1}$ ; F1,  $1.995 \times 10^{-4} \text{ Sm}^{-1}$ ; F2,  $2.24 \times 10^{-4} \text{ Sm}^{-1}$  (After Glover and Vine 1992).

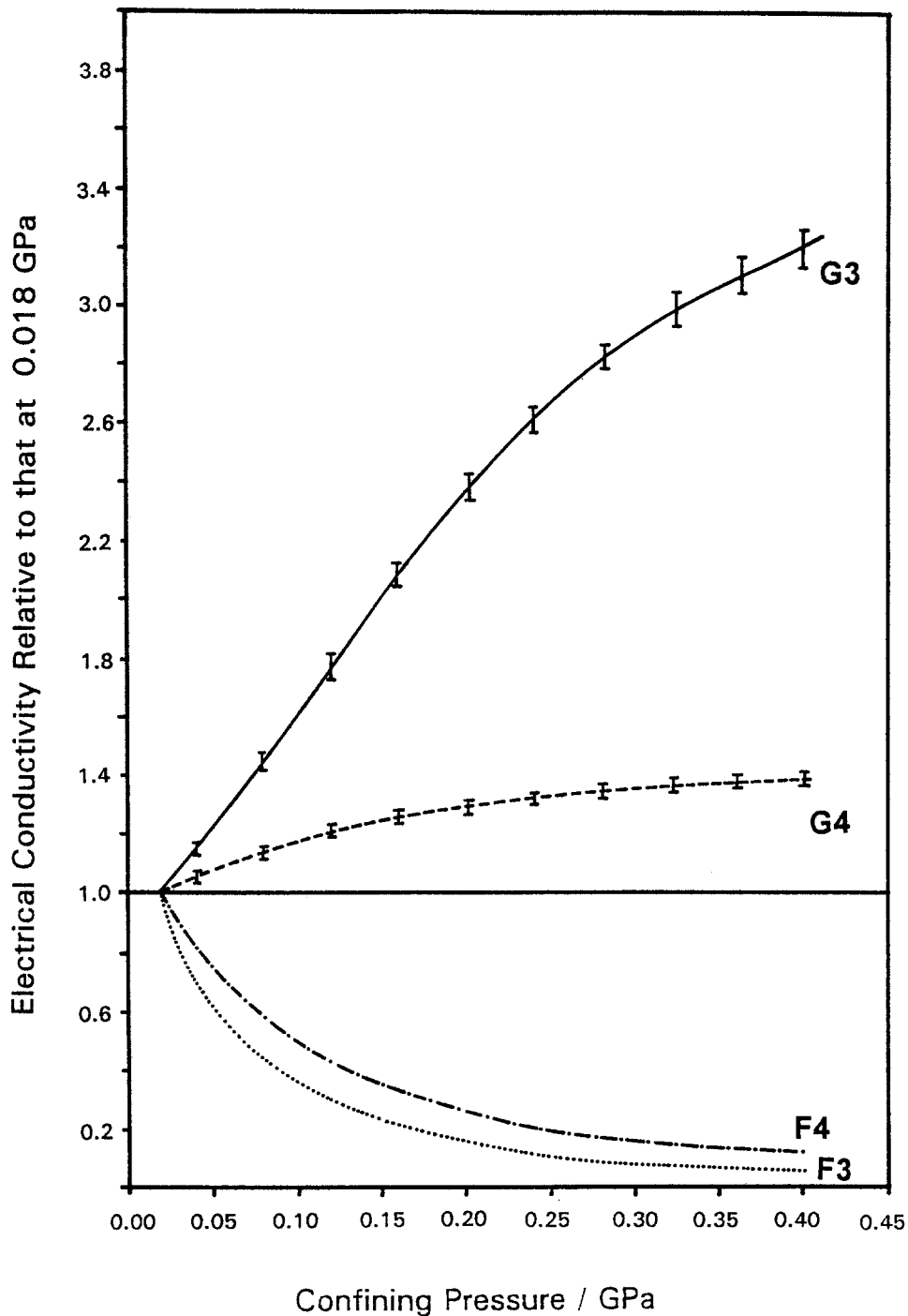


Fig. 8b. The variation of the electrical conductivity of unsaturated samples of carbon-free and carbon-bearing granulites when subjected to raised confining pressures. G3/G4- carbon-bearing, unsaturated, measurement direction parallel and perpendicular to planes of carbon foliation respectively. F3/F4 -carbon-free, unsaturated, no foliation. The data are normalised to that at 18 MPa; absolute values for each of the samples at this pressure were as follows: G3,  $2.24 \times 10^{-5} \text{ Sm}^{-1}$ ; G4,  $3.55 \times 10^{-6} \text{ Sm}^{-1}$ ; F3,  $2.51 \times 10^{-6} \text{ Sm}^{-1}$ ; F4,  $2.3 \times 10^{-6} \text{ Sm}^{-1}$  (After Glover and Vine, 1992) with data from Evans 1980).

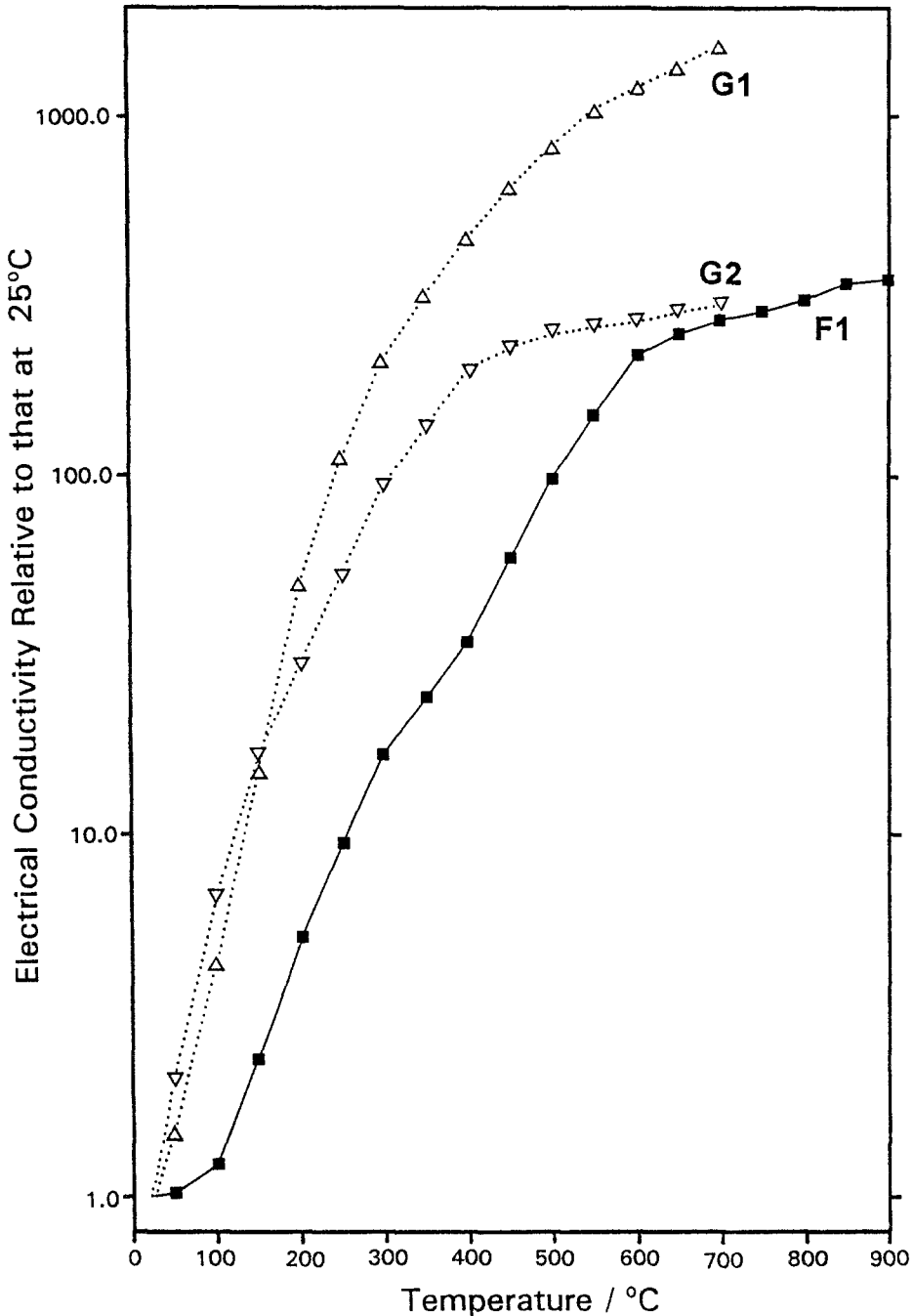


Fig. 9. The variation of the electrical conductivity of 0.5M NaCl brine saturated carbon-free and carbon-bearing granulites when subjected to a constant raised confining pressure (0.2 GPa), pore fluid pressure (0.18 GPa) and temperatures (up to 800 °C). G1/G2 -carbon-bearing, measurement direction parallel and perpendicular respectively to planes of carbon foliation. F1- carbon-free, no foliation. The data are normalised to that at 25 °C; absolute values for each of the samples at this temperature were as follows: G1,  $1.26 \times 10^{-4} \text{ Sm}^{-1}$ ; G2,  $0.631 \times 10^{-4} \text{ Sm}^{-1}$ ; F1,  $1.26 \times 10^{-5} \text{ Sm}^{-1}$  (After Glover and Vine 1992).

still present in the dried rock. The increasing conductivity of the graphite-bearing samples with the application of high confining pressures has been attributed to the increasing connectivity of inter-granular and grain surface graphite as the pore volume is lost and pore throats close down, leaving only compacted carbon conduction pathways. This reconnection of graphite pathways provides another mechanism for the increase of rock conductivity with depth and also demonstrates that electrical conductivity measurements on graphite-bearing rocks without confining pressures cannot give a true indication of their conductivity at depth.

Measurements on electrolyte-saturated graphite-bearing samples (G1, G2) at constant raised confining and pore-fluid pressures, and a range of temperatures up to 800 °C, (Figure 9) have shown that the rate of increase in the electrical conductivity of these samples is greater than that for the electrolyte-saturated graphite-free samples (F1 - Figure 9). This increased temperature dependence cannot be explained merely by the presence of additional graphite conduction as this is nearly independent of temperature. Clearly there must be some other effect to consider. Further work needs to be done on the carbon-electrolyte system to ascertain whether these observations are connected with a carbon-water interaction or possibly the formation of carbon-sodium intercalation compounds at high temperature and pressure (Glover and Vine, 1992). Graphite-alkali metal intercalation compounds are well known to have conductivities many times greater than pure graphite but they also have a negative coefficient of electrical conductivity with temperature. The observed increase in conductivity with temperature, however, may be explained by the increased formation of such compounds as temperature increases.

It is also interesting to note that the graphite-bearing granulite samples showed considerable electrical conductivity anisotropy (samples G1 and G3 being measured parallel to the well developed graphite foliation, compared to samples G2 and G4 which were measured perpendicular to the foliation). The effect of the foliation can be seen in the form of higher absolute values of conductivity and the greater sensitivity of electrical conductivity increases due to greater reconnection of graphite pathways when the direction is parallel to the foliation.

Complex electrical conductivity measurements made from 5 Hz to 20 MHz showed an increased out-of-phase conduction in the graphite-bearing samples. This was ascribed to the way in which charge carriers move between the graphite crystal planes. The manner in which charge carriers build-up on alternate crystal planes produces a time-varying secondary electric field that induces out-of-phase current flow.

## 5. Electrical Conductivity Modelling

The measurements shown in Figures 5 to 9 can be used to construct electrical conductivity profiles of the phanerozoic and pre-cambrian continental crust (Figures 10 and 11). In these Figures a three-fold division of the crust has been assumed.

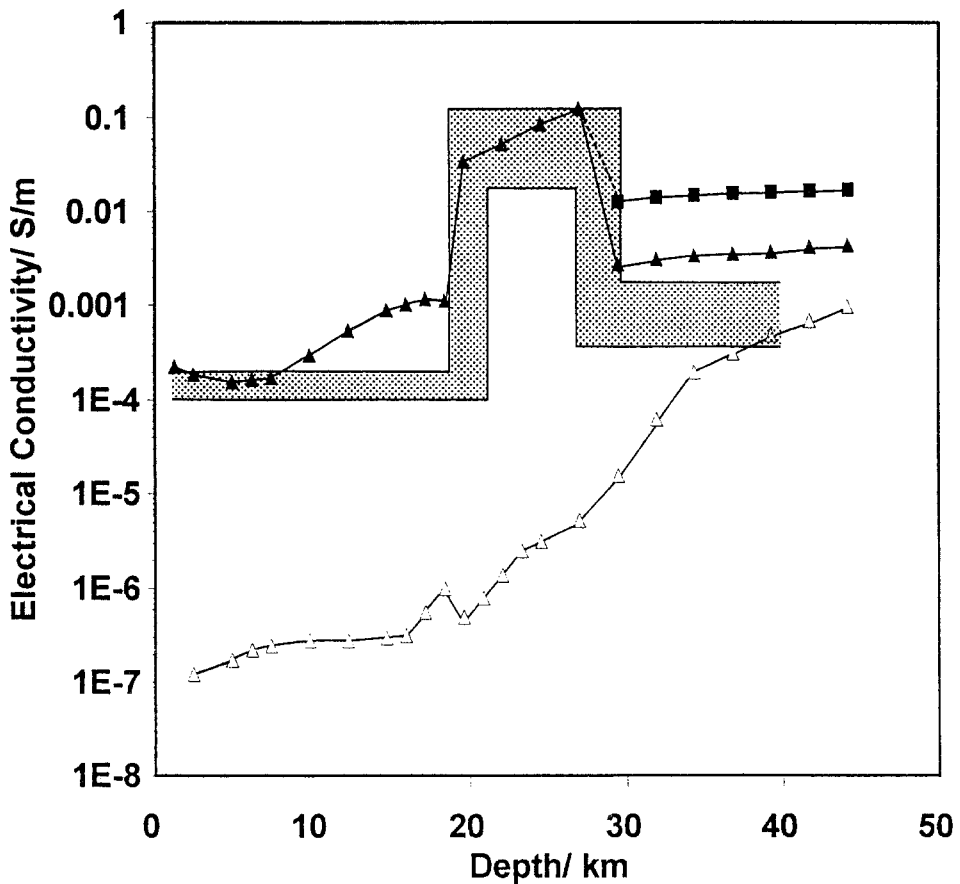


Fig. 10. Predicted conductivity/depth profiles for the Phanerozoic continental crust based on assumed crustal structure and mineralogies (filled symbols show profiles for 0.5M NaCl saturated rock, open symbols show profiles for dry rock; triangles show graphite-free rocks and squares show graphite-bearing samples). All data obtained from Figures 5 to 9 except the graphite-free dry rock data which is from Toussaint-Jackson (1984) and Čermák and Laštovičová (1987). The geotherm was 20.4 °C/km calculated from data in Marquis and Hyndman (1992).

This is based upon the typical electrical conductivity behaviour of the crust inferred from a review of available magneto-telluric data (Marquis and Hyndman, 1992; Jones, 1981; Beamish, 1986). These profiles show that measurements of the electrical conductivity of electrolyte saturated rocks at simulated crustal conditions are able to recreate the electrical conductivities observed by magneto-telluric surveys. The highest conductivities in the deep crust were attributed to brine saturated basic amphibolites at mid-crustal depths and have contributions from electrolytic con-

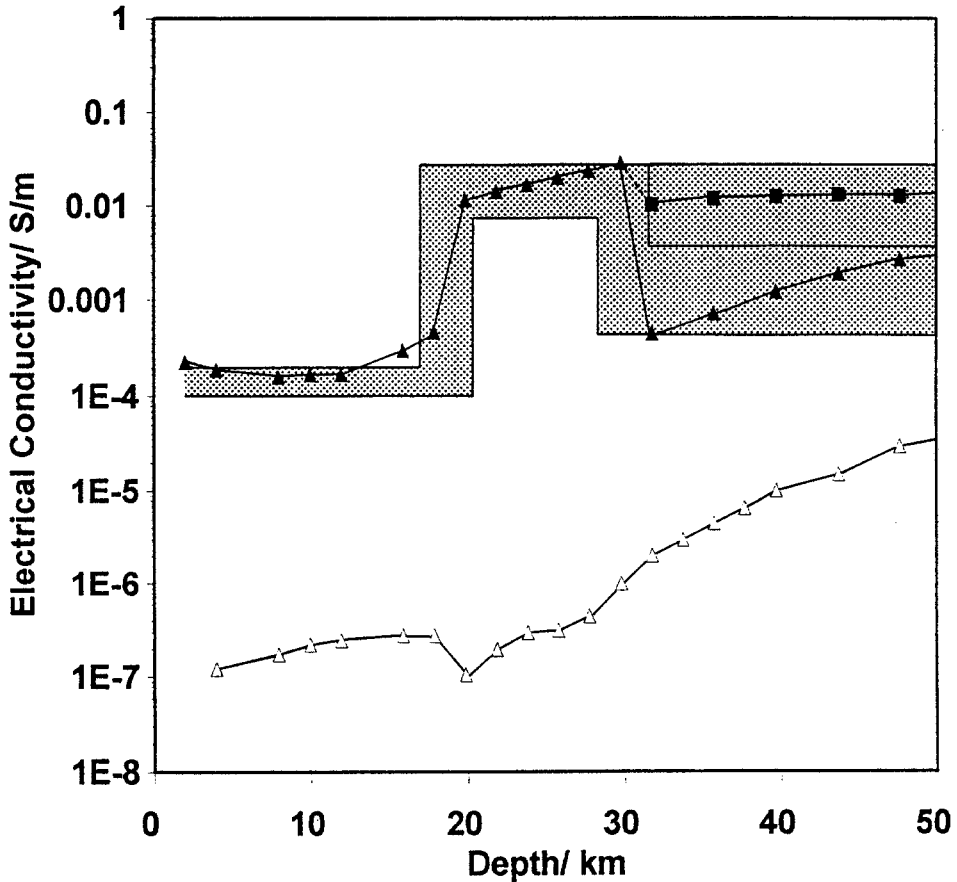


Fig. 11. Predicted conductivity/depth profiles for the Pre-Cambrian continental crust based on assumed crustal structure and mineralogies (filled symbols show profiles for 0.5M NaCl saturated rock, open symbols show profiles for dry rock; triangles show graphite-free rocks and squares show graphite-bearing samples). All data obtained from Figures 5 to 9 except the graphite-free dry rock data which is from Toussaint-Jackson (1984) and Čermák and Laštovičová (1987). The geotherm was 12.7 °C/km calculated from data in Marquis and Hyndman (1992).

duction through the saturating electrolyte and also from electronic conduction via the mineral matrix.

The profiles have been calculated from the data presented in Figures 5 to 9, associated data in Glover (1989), Glover, Vine and Ross (1990), and Glover and Vine (1992), and from data in Toussaint-Jackson (1984) and Čermák and Laštovičová (1987). Each profile for saturated rocks was constructed by using the laboratory measurements directly. The electrical conductivities measured in the laboratory



as a function of temperature (Figures 6, 7, and 9) were converted to depth using geotherms calculated by finding the mean of the dataset provided in Marquis and Hyndman (1992) with a few additions (Beamish, 1986). The geotherm for Phanerozoic crust was  $20.4\text{ }^{\circ}\text{C km}^{-1}$ , and that for Pre-Cambrian crust was  $12.7\text{ }^{\circ}\text{C km}^{-1}$ . Both of these are within the range of representative geotherms for regions of these ages ( $19\text{--}26\text{ }^{\circ}\text{C km}^{-1}$  for phanerozoic crust (meso-Europe; Cermak, 1982), and  $8\text{--}16\text{ }^{\circ}\text{C km}^{-1}$  for Pre-Cambrian crust (Baltic and Ukrainian Shields; Cermak, 1982). Geotherms from the Marquis and Hyndman dataset were used to derive electrical conductivity/depth profiles from laboratory measurements in order that they can be compared directly with the average electrical conductivity/depth profiles derived from geophysical field measurements also provided by Marquis and Hyndman (1992).

Where adjustment for the effects of pressure were necessary (very low pressures and very high pressures) the calculation used data from the pressure variation of the same rock sample (given in Figures 5 and 8, and Glover (1989)), and taking into account the effect of pore-fluid pressure using measured data from the same rock sample, given in Glover (1989). Any extrapolation at high confining pressures took full account of the way in which the variation of the electrical conductivity was less sensitive to changes in confining pressure at higher temperatures. The crustal profile was split into layers, each of which was assigned the electrical variation of a particular rock type. Three layers have been assumed; an upper crust represented by a granodioritic composition, an uppermost part of the lower crust consisting of amphibolites, and a lowermost crust of basic granulites. The depth to the transitions between these were taken from an average of the magneto-telluric data in Marquis and Hyndman (1992), and represents the only point where the experimentally derived conductivity/depth profiles and those derived from the field observations are not independent.

The experimentally derived profiles, shown in Figures 10 and 11, are compared with shaded areas that indicate the approximate bounds of the electrical conductivity in the phanerozoic and Pre-Cambrian crust. The bounds of electrical conductivity have been calculated from data relating to over 65 magneto-telluric and controlled source studies at 24 phanerozoic and 11 pre-cambrian locations, obtained mainly from a compilation by Marquis and Hyndman (1992) but with a few additions (Beamish, 1986).

## 6. The Phanerozoic Crust

The following discussion relates to Figure 10. The upper crust is quite well represented by electrolytic conduction through predominantly acidic rocks with conduction occurring solely through the pore-fluid. The decrease in electrical conductivity with depth due to pressure compaction is initially important. As depth, pressure and temperature increase, the porosity becomes more insensitive to changes in confining pressure as crack surface asperities meet but the cracks do not completely

close, and the increase in the electrical conductivity of the saturating electrolyte begins to control the bulk conduction in the rock.

Mineral expansion, at temperatures up to about 100 °C, also leads to a small reduction in porosity (Darot *et al.*, in press) and hence electrical conductivity. At temperatures greater than 100 °C mineral expansion induces sufficient thermal stresses to cause new cracks to occur in most rocks. The associated dilatancy in porosity compensates for the loss of porosity due to mineral expansion and can lead to an increase in electrical conductivity resulting from the presence of new conduction pathways. It is important that a distinction is made between the samples measured in the laboratory, and the rock in the Earth. The laboratory samples were all rocks recovered from field locations, and were relatively unaltered. They were subjected to progressively higher confining pressure, pore fluid pressure and temperature, simulating increasing depth, and this was done very rapidly compared with geological timescales. This is very unlike the experience of saturated rocks deep in the Earth. It is possible that the relatively rapid temperature rises experienced by the laboratory samples might result in the decrease of porosity, and hence electrical conductivity, by thermal expansion into voids, followed by new crack growth and dilatancy due to the thermal stresses as temperatures are increased yet further, resulting in increases in electrical conductivity due to increased porosity and pore connectivity. However, in the crust, where temperature changes are vastly slower, the texture of the porosity is likely to be in equilibrium, and fresh cracking, if any occurs, is more likely to stem from tectonic stresses at virtually isothermal conditions.

The profile calculated from electrical conductivity measurements on dry rocks of granodioritic composition shows that they are three to four orders of magnitude too low to account for the observed upper crustal conductivities as expected. In the upper crust temperatures are insufficient to enable solid state mineral conduction to provide any contribution to the bulk rock conductivity, except in the case of highly localised ore bodies. Clearly the upper crust needs to be saturated with a saline aqueous fluid to provide the observed high electrical conductivities, this is confirmed by fluid measurements in super-deep bore-holes such as the KTB and the Kola hole (Kozlovsky, 1984).

The profile constructed from electrolyte-saturated amphibolite facies rocks is very important. It shows directly from laboratory measurements, that the electrical conductivity of these rocks is sufficient to account for the very high electrical conductivities at these levels, for the first time. The conduction in amphibolites under these conditions has contributions from electrolytic pore-fluid conduction with additional and significant mineral conduction. The profile constructed from measurements on dry amphibolites, by comparison, shows conductivities that are insufficient to explain the high conductivities observed at these levels by geophysical field measurements.

The lowermost part of the crust has a more ambiguous interpretation. Electrolyte saturated, dry graphite-free and dry graphite-bearing granulites all have conductiv-

ities of the correct order of magnitude. If the crust is saturated with aqueous fluids at these depths, their presence does not add to the overall conductivity because they have a relatively low electrical conductivity compared with the electronic matrix conduction at lower crustal temperatures. The lowermost crust could be saturated with aqueous fluids or be anhydrous. Either scenario would produce a similar bulk electrical conductivity. It may very well be the case that the Earth's lower continental crust is saturated in some locations and anhydrous in other areas depending on the availability of water and its local mobility in the crust.

The electrical conductivity/depth profiles derived from experimental measurements presented in this paper (i.e. Figure 10) indicate that whatever other conduction mechanisms may be present, the occurrence of aqueous electrolytes in the upper and middle crust seems to be necessary to obtain the electrical conductivities that are inferred from geophysical field measurements. Certainly, laboratory measurements on dry samples (Toussaint-Jackson, 1984; Čermák and Laštovičová, 1987; Kariya and Shankland, 1981) show that the electrical conductivity of dry mineralogies is too small to explain the field measurements. The presence of a particular mineralogy is more ambiguous. This work indicates that only electrolyte-saturated amphibolite rocks containing significant proportions of hornblende, of those tested, have sufficiently high electrical conductivities. This is supported by the compilation of dry rock experiments by Kariya and Shankland (1981), which indicate that significant conduction through the rock matrix is unlikely under crustal conditions with the possible exception of amphibolites. It will only be possible to be sure of the full implications of the presence of either aqueous fluids or a particular mineralogy in the deep crust when it is possible to carry out experiments such as those reported here in this work, upon fresh samples obtained directly from bore-holes penetrating the deep Earth.

If aqueous fluids pervade the crust, as increasingly seems to be the case, their origin is not yet fully clear. There are several possible sources of deep crustal water. Water may be derived from the dehydration of subducted slabs of oceanic crust, as the water of dehydration of deeper rocks that have undergone granulite facies metamorphism, from the accretion of saturated rock of oceanic crustal origin, and finally as water derived from the mantle either directly or dissolved in magma. It is likely that the source of aqueous fluids in the crust will change from location to location, depending upon the local crustal structure and dynamics. Whatever the source of the deep water, it is important that the permeability of the rocks in the crust is sufficiently low to trap the electrolyte in place. While this criterion needs to be fulfilled, the porosity must also be sufficiently inter-connected to allow there to be pore-fluid paths for conduction. The deposition of clay minerals from hot, mineralised brines, of possibly metamorphic origin, at mid-crustal depths is one mechanism that has been invoked to help reduce the permeability (Etheridge *et al.*, 1984). In this case the deposited minerals may also enhance the bulk rock conduction by increased surface conduction and bulk conduction through the body of the clay minerals (Toussaint-Jackson, 1984).

Permeability may also be controlled by the mineral/fluid dihedral angle (Watson and Brennan, 1987). This angle describes the wettability of the fluid on the mineral surface. If the value of the dihedral (wetting) angle is high, the fluid does not wet the surface well. In this case small cracks and pore-throats can remain unsaturated with the aqueous fluid, reducing pore fluid connectivity even when the pores are physically inter-connected (Hyndman and Shearer, 1989). If the dihedral angle is low, however, the fluid wets the minerals so well that it can form a thin coating that remains inter-connected even through thin cracks and pore-throats. The increase in conductivity in the middle crust has been ascribed to the connection of previously isolated pockets of fluid (either physically isolated pores or unconnected fluids due to high dihedral angles in connected pores) by the process of lower crustal shear to produce a series of electrolyte saturated horizontal layers with high horizontal connectivity and low vertical permeability (Marquis and Hyndman, 1992). A scenario similar to this has been numerically modelled by Merzer and Klempner (1992). Such horizontal layers have also been invoked as the source of the strong sub-horizontal seismic reflectors present in the lower crust (Hyndman and Shearer, 1989). There is good evidence for the presence of lower crustal shear zones in the Ivrea Zone lower crustal exposure (Brodie and Rutter, 1987) but they are anhydrous. These shear zones could have been saturated at depth but the free water would have had to be expelled from the rocks before uplift of the zone occurred.

There is an intrinsic uncertainty in the interpretation of magneto-telluric data between thick moderately highly conducting crustal layers or thin very highly conducting layers (Beamish, 1986). In most locations it is not clear whether high conductivity layers extend to the mantle or whether a thin, highly conducting electrolyte saturated layer in the mid-crust conceals a moderately high conductivity lowermost crust consisting of dry or electrolyte saturated high grade rocks. One source of aqueous fluids in the mid-crust is dehydration reactions at deeper levels. In this scenario there would be a deep layer of crustal rock which is anhydrous and consisting of high grade, granulite facies. Granulites, such as those from the Ivrea Zone lower crustal exposure are very unstable in the presence of a free aqueous phase at high temperatures. They rapidly undergo retrograde reactions to form hydrated minerals (Wood and Walther, 1983; Gough, 1986). If the aqueous fluids produced in the initial dehydration reactions were mobile and percolated to shallower layers, a stable high grade, dry lower layer may be formed. In locations where there is a high flux of  $\text{CO}_2$  from the mantle (Harris, 1989) the presence of  $\text{CO}_2$  may play an extremely important part in the lower crust. It has at least five roles. Firstly, it has been shown to promote the initial granulite facies metamorphism that produces the free aqueous fluids and dry, high-grade mineralogy (Peterson and Newton, 1989; Wendlandt, 1981). Secondly, the flux of  $\text{CO}_2$  can help move the fluids to shallower levels. Thirdly, the presence of  $\text{CO}_2$  can help stabilise the high grade mineralogy (Newton, Smith and Windley, 1980). Fourthly, evidence is growing that the presence of  $\text{CO}_2$  increases the aqueous fluid-mineral dihedral angle leading to the further isolation of whatever aqueous fluids remain (Watson

and Brennan, 1987; Holness and Graham, 1991). Finally the presence of CO<sub>2</sub> provides a source of carbon for the deposition of conducting graphite films (Glover and Vine, 1992; Mathez and Delaney, 1981; Frost *et al.*, 1989; Fuhrman *et al.*, 1988).

In other areas there may be a very abundant source of aqueous fluids. In these locations it is unlikely that an anhydrous high grade lowermost crust will be formed, with the crust remaining saturated for its whole depth. Examples where this might occur are 1) where there is a high flux of water vapour from the mantle, 2) where there is water of dehydration from subducted oceanic plates, 3) where there has been accretion of saturated oceanic material at a plate boundary, and 4) where a high flux of water is associated with partial melting and volcanism. In these areas the lower crustal conductivity will rely on solid state mineral conduction through matrix minerals such as Fe/Ti oxides, and what little pore-fluid conduction is possible at the temperatures at these depths. Graphite is unlikely to have been deposited if the crust has always been saturated with aqueous fluids, but may be present if the rocks were once dry and carbon dioxide rich and subsequently re-saturated. Differences in the availability of aqueous fluids and CO<sub>2</sub> and their relative mobility from location to location for these and other reasons may have a controlling influence on the presence, depth and size of high conductivity zones in the continental crust.

## 7. The Pre-Cambrian Crust

Figure 11 shows that the electrical conductivity of the upper part of the Pre-Cambrian crust behaves in a similar manner to that of the Phanerozoic crust. The mechanisms of pore closure, mineral expansion and increasing electrolyte conductivity with rising temperature are all applicable. The lower geotherm ensures that electrical conductivity does not increase as rapidly with depth as in the Phanerozoic case.

The electrical conductivity of electrolyte-saturated amphibolite facies rocks under Pre-Cambrian mid-crustal conditions is similar to that in the Phanerozoic crust with lower absolute conductivities. This is mainly due to the lower ambient temperatures at these depths resulting from a lower geotherm. This profile agrees very well with the magneto-tellurically observed crustal conductivities.

The magneto-telluric data for the lowermost part of the Pre-Cambrian crust sometimes show a distinct layer of moderate conductivity, otherwise it shows high conductivities of the middle crust extending to the full depth of the crust. This ambiguity may be due to the inherent difficulty to probe a layer of moderate conductivity beneath a layer of higher conductivity with the magneto-telluric method, or the effect of graphite conduction in some areas and not in others. The bounds given in Figure 11 are those for all cases. It can be seen that dry or electrolyte saturated, graphite-bearing or graphite-free rocks all fall within the observed conductivity bounds.

## 8. Conclusions

The laboratory study of saturated rocks under controlled conditions of pressure and temperature has the ability to help us improve our understanding of the physical processes occurring in crustal rocks. In this case experiments on saturated rocks at lower crustal temperatures and raised pressures have shown that certain rocks (saturated amphibolites) have electrical conductivities consistent with those observed magneto-tellurically. Other experiments have shown that graphite-bearing rocks have the ability to contribute extensively to the electrical conductivity of the Earth's continental crust. Both of these important results imply that the electrical conductivity of the crust is inextricably linked with the presence and mobility of aqueous fluids and carbon dioxide in the continental crust. The concept of fluid controlled continental crustal electrical properties fits in well with the observed variations of electrical conductivity laterally and with depth, ascribing these changes to differences in the availability and mobility of fluids in different parts of the crust.

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## References

- Alabi, A. O., Camfield, P. A. and Gough, D. I.: 1975, 'The North American Central Plains Conductivity Anomaly', *Geophys. J. R. Astr. Soc.* **43**, 815–833.
- Allmendinger, R. W., Nelson, K. D., Potter, C. J., Barazangi, M., Brown, L. D. and Oliver, J. E.: 1987, 'Deep Seismic Reflection Characteristics of the Continental Crust', *Geology* **15**, 304–310.
- Archie, G. E.: 1942, 'Electrical Resistivity Log as an Aid in Determining Some Reservoir Characteristics', *Am. Inst. Min. Met. Eng.* **146**, 54–62.
- Beamish, D.: 1986, 'Deep Crustal Geoelectric Structure Beneath the Northumberland Basin', *Geophys. J. R. Astr. Soc.* **84**, 619–640.
- Blundell, D., Freeman, R. and Mueller, S.: 1992, *A Continent Revealed: The European Geotraverse*, Cambridge University Press, Cambridge, U.K.
- Brace, W. F., Orange, A. S. and Madden, T. R.: 1965, 'The Effect of Pressure on the Electrical Resistivity of Water-Saturated Crystalline Rocks', *J. Geophys. Res.* **70**(22), 5669–5678.
- Brace, W. F. and Orange, A. S.: 1968, 'Further Studies of the Effects of Pressure on Electrical Resistivity of Rocks', *J. Geophys. Res.* **73**(16), 5407–5420.
- Brodie, K. H. and Rutter, A. H.: 1987, 'Deep Crustal Extensional Faulting in the Ivrea Zone of Northern Italy', *Tectonophysics* **140**, 193–212.
- Cemic, L. and Jansen, E.: 1975, 'Measurement of the Electrical Conductivity in a Solid Medium High Pressure Cell by A. C. Bridge', *Hi. Temp. Hi. Press.* **7**, 295–298.
- Čermák, V.: 1982, 'Crustal Temperature and Mantle Heat Flow in Europe', *Tectonophysics* **83**, 123–142.
- Čermák, V. and Laštovičová, M.: 1987, 'Temperature Profiles in the Earth of Importance to Deep Electrical Conductivity Models', *Pure Appl. Geophys.* **125**, 255–284.

- Crawford, M. L. and Hollister, L. S.: 1986, 'Metamorphic Fluids: The Evidence from Fluid Inclusions', in Walther, J. V. and Wood, B. J. (eds.), *Advances in Physical Geochemistry*, Vol. 5. Springer-Verlag, Berlin, pp. 1–35.
- Darot, M., Ruffet, C. and Gueguen, Y.: 1994, 'Experimental Measurements of Transport Properties in Thermally Cracked Granites', in Meridith, P. G., Glover, P. W. J. and Sammonds, P. R. (eds.), *Proceedings of the Crack Systems in Rocks Meeting*, 1st–2nd April 1993, University College London. (in press).
- Downes, H.: 1993, 'The Nature of the Continental Crust of Europe: Petrological and Geochemical Evidence from Xenoliths', *Phys. Ea. Planet. Int.* **79**, 195–218.
- Drury, M. J. and Hyndman, R. D.: 1979, 'The Electrical Resistivity of Oceanic Basalts', *J. Geophys. Res.* **18**(B9), 4537–4545.
- Duba, A. Huenges, E., Nover, G., Will, G. and Jödicke, H.: 1988, 'Impedance of Black Shale from Münsterland, 1. Bore-Hole: An Anomalously Good Conductor?', *Geophys. J.* **94**, 413–419.
- Duba, A. G. and Shankland, T. J.: 1982, 'Free Carbon and Electrical Conductivity in the Earth's Mantle', *Geophys. Res. Lett.* **9**(11), 1271–1274.
- Etheridge, M. A., Wall, V. J., Cox, S. F. and Vernon, R. H.: 1984, 'High Fluid Pressures During Regional Metamorphism and Deformation; Implications for Mass Transport and Deformation Mechanisms', *J. Geophys. Res.* **89**, 4344–4358.
- Evans, C. J.: 1980, *The Seismic and Electrical Properties of Crystalline Rocks under Simulated Crustal Conditions*, PhD. Thesis, University of East Anglia, Norwich.
- Frost, B. R., Fyfe, W. S., Tazaki, K. and Chan, T.: 1989, 'Grain-Boundary Graphite in Rocks and Implications for High Electrical Conductivity in the Lower Crust', *Nature* **340**, 134–136.
- Fuhrman, M. L., Frost, B. R. and Lindsley, D. H.: 1988, 'Crystallisation Conditions of the Sybille Monzosyenite, Laramie Anorthosite Complex, Wyoming', *J. Petrol.* **29**, 699–729.
- Gavrilenko, P. and Gueguen, Y.: 1989, 'Pressure Dependence of Permeability: A Model for Cracked Rocks', *Geophys. J. Int.* **98**, 159–172.
- Glover, P. W. J., Vine, F. J. and Ross, R. G.: 1990, 'Measurements of Saturated Basic and Acidic Rock Electrical Conductivities at Lower Crustal Temperatures and High Pressures', *Hi. Press. Res.* **5**, 735–737.
- Glover, P. W. J., Ross, R. G. and Jolly, H.: 1990, 'The Measurement of Saturated Rock Electrical Conductivity at Lower Crustal Temperatures and High Pressures', *Hi. Press. Res.* **5**, 705–707.
- Glover, P. W. J. and Ross, R. G.: 1990, 'A Method of Measuring the Electric Conductivity of Saturated Rocks in a High-Pressure Cell Subjected to High Temperatures', *Hi. Temp. Hi. Press.* **22**, 533–539.
- Glover, P. W. J. and Vine, F. J.: 1992, 'Electrical Conductivity of Carbon-Bearing Granulite at Raised Temperatures and Pressures', *Nature* **360**, 723–726.
- Glover, P. W. J.: 1989, *Electrical Conductivity of Rock Samples Subjected to High Temperatures and Pressures*, PhD Thesis, University of East Anglia, Norwich, 2 volumes, 408 pp.
- Glover, P. W. J. and Vine, F. J.: 1993, 'Electrical Conductivity of the Continental Crust', In press.
- Gough, D. I.: 1986, 'Seismic Reflectors, Conductivity, Water and Stress in the Continental Crust', *Nature* **323**, 143–144.
- Haak, V. and Hutton, V. R. S.: 1986, 'Electrical Resistivity in the Continental Lower Crust', in Dawson, J. B., Carswell, D. A., Hall, J. and Wedepohl, K. H. (eds.), *The Nature of the Lower Continental Crust*, Spec. Pub. Vol. 24, Geol. Soc., London, pp. 35–49.
- Harris, N.: 1989, 'Carbon Dioxide in the Deep Crust', *Nature* **340**, 347.
- Hermance, J. F.: 1979, 'The Electrical Conductivity of Materials Containing Partial Melt', *Geophys. Res. Lett.* **6**, 613–616.
- Holness, M. B. and Graham, C. M.: 1991, 'Equilibrium Dihedral Angles in the System H<sub>2</sub>O-CO<sub>2</sub>-NaCl-Calcite and Implications for Fluid Flow during Metamorphism', *Contrib. Mineral. Petrol.* **108**, 368–383.
- Hyndman, R. D. and Klempner, S. L.: 1989, 'Lower Crustal Porosity from Electrical Measurements and Inferences about Composition from Seismic Velocities', *Geophys. Res. Lett.* **16**, 255–258.
- Hyndman, R. D. and Shearer, P. M.: 1989, 'Water in the Lower Continental Crust: Modelling Magneto-Telluric and Seismic Reflection Results', *Geophys. J. Int.* **98**, 343–365.

- Jones, A. G.: 1981, 'On a Type Classification of Lower Crustal Layers under Precambrian Regions', *J. Geophys.* **49**, 226–233.
- Jones, A. G.: 1987, 'MT and Reflection: An Essential Combination', *Geophys. J. R. Astr. Soc.* **89**, 7–18.
- Kariya, K. A. and Shankland, T. J.: 1983, 'Electrical Conductivity of Dry Lower Crustal Rocks', *Geophysics* **48**(1), 52–61.
- Kaufman, K. and Keller, G. R.: 1981, *The Magnetotelluric Sounding Method*, Elsevier, Amsterdam, 595 pp.
- Kozlovsky, Y.A.: 1984, 'The World's Deepest Well', *Sci. Am.* **251**, 106–112.
- Lee, C. D., Vine, F. J. and Ross, R. G.: 1983, 'Electrical Conductivity Models for the Continental Crust Based on Laboratory Measurements on High-Grade Metamorphic Rocks', *Geophys. J. R. Astr. Soc.* **72**, 353–371.
- Mareschal, M., Fyfe, W. S., Percival, J. and Chan, T.: 1992, 'Grain Boundary Graphite in Kapukasing Gneisses and Implications for Lower Crustal Conductivity', *Nature* **357**, 674–676.
- Marquis, G. and Hyndman, R. D.: 1992, 'Geophysical Support for Aqueous Fluids in the Deep Crust: Seismic and Electrical Relationships', *Geophys. J. Int.* **110**, 91–105.
- Mathez, E. A. and Delaney, J. R.: 1981, 'The Nature and Distribution of Carbon in Submarine Basalts and Peridotite Nodules', *Earth. Planet. Sci. Lett.* **56**, 217–232.
- Merzer, A. M. and Klemperer, S. L.: 1992, 'High Electrical Conductivity in a Model Lower Crust with Unconnected, Seismically Reflective Layers', *Geophys. J. Int.* **108**, 895–905.
- Newton, R. C., Smith, J. V. and Windley, B. F.: 1980, 'Carbonic Metamorphism, Granulites, and Crustal Growth', *Nature* **288**, 45–50.
- Nicolas, A.: 1989, *Structures of Ophiolites and Dynamics of Oceanic Lithosphere*, Kluwer Academic, Dordrecht.
- Olhoeft, G. R.: 1981, 'Electrical Properties of Granite with Implications for the Lower Crust', *J. Geophys. Res.* **86**(B2), 931–936.
- Parkhomenko, E. I.: 1982, 'Electrical Resistivity of Minerals and Rocks at High Temperature and Pressures', *Rev. Geophys. Sp. Phys.* **20**(2), 193–218.
- Peterson, J. W. and Newton, R. C.: 1989, 'CO<sub>2</sub>-Enhanced Melting of Biotite-Bearing Rocks at Deep-Crustal Pressure-Temperature Conditions', *Nature* **340**, 378–380.
- Quist, A. S. and Marshall, W. L.: 1968, 'Electrical Conductances of Aqueous Sodium Chloride Solutions from 0 to 800 °C and at Pressures to 4000 Bars', *J. Phys. Chem.* **72**, 684–703.
- Shankland, T. J.: 1989, 'A Case of Two Conductors', *Nature* **340**, 102.
- Shankland, T. J. and Ander, M. E.: 1983, 'Electrical Conductivity, Temperatures, and Fluids in the Lower Crust', *J. Geophys. Res.* **88**(B11), 9475–9484.
- Stesky, R. S., and Brace, W. F.: 1973, 'Electrical Conductivity of Serpentinised Rocks to 6 kbars', *J. Geophys. Res.* **78**, 7614–7621.
- Taylor, S. R. and McLennan, S. M.: 1985, *The Continental Crust: Its Composition and Evolution*, Blackwell Scientific, London.
- Thompson, B. G., Nekut, A. and Kuckes, A. F.: 1983, 'A Deep Crustal Electromagnetic Sounding in the Georgia Piedmont', *J. Geophys. Res.*, **88**, 9461–9473.
- Toussaint-Jackson, J. E.: 1984, *Laboratory and Theoretical Studies of Electrical Conductivity in Crystalline Rocks*, PhD Thesis, University of East Anglia, Norwich.
- Watson, E. B. and Brennan, J. M.: 1987, 'Fluids in the Lithosphere, 1. Experimentally-Determined Wetting Characteristics of CO<sub>2</sub>-H<sub>2</sub>O Fluids and their Implications for Fluid Transport, Host Rock Physical Properties, and Fluid Inclusion Formation', *Earth Planet Sci. Lett.* **85**, 497–515.
- Wendlandt, R. F.: 1981, 'Influence of CO<sub>2</sub> on Melting of Model Granulite Facies Assemblages: A Model for the Genesis of Charnockites', *Am. Miner.* **66**, 1164–1174.
- Wood, B. J. and Walther, J. V.: 1983, 'Rates of Hydrothermal Reactions', *Science* **222**, 413–415.
- Yardley, B. W. D.: 1986, 'Is There Water in the Deep Continental Crust?', *Nature* **323**, 111.