

Electrical conductivity models for the continental crust based on laboratory measurements on high-grade metamorphic rocks

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Summary. Deep electrical conductivity soundings are increasingly being used as an additional source of information regarding the nature of the lower continental crust. However, a lack of relevant laboratory-based conductivity measurements makes interpretation of such soundings difficult. Laboratory measurements have been made on saturated and unsaturated samples of possible lower crustal rock types subjected to confining pressures up to 0.4 GPa, temperatures up to 300°C, and with variable pore fluid pressure up to the confining pressure. Extrapolation of these results suggests that the surprisingly high conductivities deduced for depths of approximately 20 km in certain stable continental areas may result from a combination of basic rock type and high pore fluid pressures, for whereas the conductivities measured in acid rock types can be explained in terms of conduction through the pore fluid alone, the conductivities measured in basic rock types imply enhanced conduction presumably through the matrix or along grain boundaries. The lower conductivities deduced from field experiments for the upper crust may be due to more acid rock types and/or lower pore fluid pressures, perhaps due to hydration reactions. In areas where the high conductivity layer is coincident with a low velocity layer an explanation in terms of changing pore fluid pressure, i.e. low to high, with increasing depth seems more likely.

Introduction

Although there have been a considerable number of laboratory studies of the electrical conductivity of rocks and minerals, the majority have been made on nominally dry materials subjected to the very high pressures and temperatures associated with the Earth's mantle (Parkhomenko 1967; Keller 1966). However, similar measurements on rocks subjected to crustal conditions are complicated by the possible presence in the Earth's crust of saline pore fluids which can significantly increase the measured conductivity (Brace 1971;

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Parkhomenko, Stafankevich & Vysokova 1972; Drury 1979). On the basis of electrical conductivity measurements on rocks saturated with different pore fluids, Brace (1971) proposed a model in which the mineral grains act as an insulating matrix and conduction occurs solely through the pore fluid between the mineral grains. He then interpreted the pressure dependence (at room temperature) of the conductivity of saturated rocks in terms of the progressive closure of fractures, cracks and pores with increasing pressure. In order to make an estimate of the conductivity within the Earth's crust, the effect of temperature as well as pressure has to be taken into account. Since conduction on Brace's model was assumed to occur through the pore fluid alone, the effect of temperature was incorporated by using the known variation of the conductivity of the pore fluid with temperature (Quist & Marshall 1968, 1969). Thus in this way attempts have been made to predict the variation in conductivity with depth in the Earth's crust by combining the measured pressure dependence of the conductivity of saturated rocks at room temperature with the known temperature dependence of the conductivity of their pore fluids (Brace 1971; Dvorak 1975).

This report describes laboratory experiments designed to measure the conductivity of saturated rocks as a function of both pressure and temperature. In particular, it was hoped to determine whether simultaneous application of temperature, in addition to that of pressure, would lead to any physical/chemical changes to saturated rocks which had not previously been taken into account. The experimental results are presented, together with a physical interpretation of the conduction mechanism, and a brief discussion of the possible geological implications of these results.

2 Sample preparation and description

The rocks studied were very low porosity (less than 1 per cent), high grade metamorphic gneisses from the Lewisian of NW Scotland. This rock type was chosen as being representative of the lower crust, a depth to which we hoped to extrapolate our results.

Samples were cored, cut and ground into accurate right cylinders of diameter 24.5 mm and length 10–20 mm, using water as the grinding fluid to avoid contamination. They were saturated by placing them in an evacuation chamber for 24 hr, at a pressure of 10^{-4} Pa and a temperature of 75°C, in an attempt to remove the air trapped in their pores, and then breaking the vacuum by allowing a saline solution to enter the chamber to fill the accessible pores of the sample. The composition of this pore fluid was a 0.5 M NaCl solution (a concentration which is approximately equivalent to that of seawater) except for one sample which was saturated with a 0.05 M NaCl solution. Density and porosity measurements were carried out on an adjacent section of the core by weighing before and after saturation. The actual sample was not used for these measurements because they necessitated baking at 110°C to obtain the 'dry' weight before saturation. A description of the samples used is given in Table 1.

3 Experimental details

Fig. 1 shows the measurement configuration employed. Two stainless steel electrodes permitted measurement of the AC conductance between the sample faces by means of a transformer ratio-arm bridge with an accuracy of ± 0.1 per cent (Wayne-Kerr, model B 224). A 3-terminal measurement was made, the third terminal, or neutral line, on the measuring bridge being connected to the shield of a co-axial cable from the lower (un-earthed) electrode to the bridge. With this arrangement leakage conductance, of the total measurement system, was kept below $10^{-11} \Omega^{-1}$ at 20°C, rising to $2 \times 10^{-11} \Omega^{-1}$ at 250°C, several

Table 1. Description of samples.

Sample no.	Rock name	Porosity (per cent)	Density kg m ⁻³	Mineralogy	Texture
A1(a), (b), (c)	Biotite granite gneiss	0.37	2670	Incl. bleb perthite and approx. 55 per cent quartz	Finely banded; medium to coarse grained
A2	Biotite granite gneiss	0.19	2670	Incl. approx. 40 per cent quartz	Coarsely banded; medium grained
A3(a), (b) ³	Granite gneiss	0.32	2680	Orthoclase, plagioclase, biotite, apatite and approx. 50 per cent quartz	Banded; medium to coarse grained
B1(a) ¹ , (b), (c) ² , (d) ³ , (e) ⁴	Amphibolite gneiss	0.07	3120	50 per cent epidote; 40 per cent hornblende, minor chlorite, quartz and magnetite	Very finely banded; fine grained
B2(a) ¹ , (b) ^{1,3}	Amphibolite gneiss	0.07	3000	Hornblende, plagioclase, orthoclase, quartz chlorite and magnetite	Very finely banded; coarse grained
B3	Basic gneiss	0.06	3120	Plagioclase, orthoclase, hornblende, chlorite, magnetite	Indistinct banding; variable grain size

Notes:

- (1) Drilled perpendicular to layering.
- (2) Saturated with 0.05 M NaCl solution.
- (3) Unsaturated.
- (4) Pre-baked to 750°C.

NB: With the exception of three samples (see note 1) all samples were drilled such that the axis of the cylinder (and hence the measurement direction) was parallel to the metamorphic foliation; and with the exception of four samples (see notes 2, 3, and 4) all samples were saturated with 0.5 M NaCl solution.

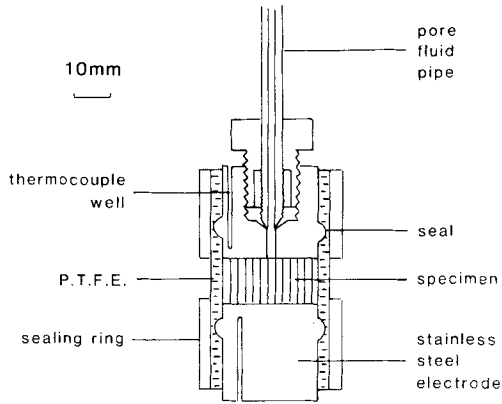


Figure 1. Specimen assembly.

orders of magnitude less than the minimum rock conductance of $10^{-8} \Omega^{-1}$. Throughout, only the real, in phase, part (σ) of the complex conductivity was measured.

The confining pressure acting on a rock within the Earth, due to the column of rocks above it, was simulated by hydrostatically pressurising the sample (up to a maximum of 0.4 GPa) in a pressure vessel. This pressure was measured to an accuracy better than ± 2.5 per cent by a manganin coil sensor attached to a digital display. To prevent contamination from the pressure transmitting medium (Plexol 262 oil), the sample and electrodes were sheathed in a PTFE (Teflon) tube. PTFE is particularly suitable for this purpose because of its high operational temperature, excellent electrical insulation properties, and flexible nature.

Most rocks are believed to have sufficient strength to prevent their pores from closing completely when subjected to the confining pressures appropriate to the Earth's crust (Brace, Orange & Madden 1965), and the pressure of the pore fluid can therefore differ considerably from the confining pressure. To allow for this, a hole passing through the upper electrode to the surface of the rock was connected to a pipe so that the pore fluid pressure could be varied independently of the confining pressure.

The maximum temperature to which the samples could be heated ($\approx 275^\circ\text{C}$) was determined by the upper temperature of operation of the oil pressure medium and the PTFE sleeve. A furnace around the sample enabled this temperature to be reached, and thermocouples situated in holes drilled into each electrode allowed the temperature to be measured as close to the sample faces as possible.

In order to be sure that the measured conductance was that of the rock sample, and was not resulting from conduction through a film of pore fluid between the rock and its sleeve, preliminary measurements were made using non-porous, electrically insulating plugs of PTFE and sapphire as test samples. Results showed that although such a conduction path did exist on assembly (i.e. at atmospheric pressure) it was eliminated with the application of a very low confining pressure (approximately 10 MPa) which forced the sleeving tightly on to the surface of the sample.

For all of the saturated samples studied, an initial pressure cycle was necessary before reversible results were obtained. Such an irreversible behaviour on the first application of a confining pressure is a common phenomenon in the measurement of physical properties, e.g. compressibility (Walsh 1965) and acoustic velocity (Simmons 1964), which are sensitive to the presence of cracks in rocks. It can be attributed to irreversible movements at grain boundaries as adjacent grains are forced into a closer packing arrangement. The procedure

followed therefore was to cycle the confining pressure (P_c), at zero pore fluid pressure (P_p), several times until repeatable conductivity values were obtained, after which measurements were made at different values of P_p .

The measurements reported here were all made at the internal frequency of the measurement bridge, 1592 Hz, but an external source enabled exploratory measurements to be made in the frequency range 200 Hz–50 kHz. The variation in the measured conductivity with frequency was found to be independent of rock type, or pressure and temperature conditions, and only dependent on the magnitude of the conductivity measured, the amount of dispersion decreasing with increasing conductivity. The results were in close agreement with the work of Parkhomenko and coworkers (Parkhomenko & Dvorak 1971; Valeev & Parkhomenko 1965; Volarovich, Valeev & Parkhomenko 1965). Thus for rock conductivities greater than approximately $10^{-3.5} \Omega^{-1} \text{m}^{-1}$ dispersion is unimportant. However, conductivities measured at 1592 Hz which are lower than this value are somewhat greater than would be obtained at the lower frequencies which are more characteristic of field measurements. Since it is the higher conductivities measured in the field and in the laboratory which are of particular interest this is not considered to be a serious limitation of this work.

4 Results and discussion

4.1 SATURATED SAMPLES

4.1.1 Conductivity as a function of confining pressure (at room temperature)

According to the Brace model, for rocks saturated with pore fluids conduction occurs through the electrolyte present in the pore network between essentially insulating mineral grains. Since it is known that the conductivity of such electrolytes is approximately independent of pressure (Quist & Marshall 1968, 1969), it appears reasonable to assume that any variation of σ with P_c can only be due to changes in the shape or distribution of the pores. Furthermore, if such a variation is reversible, it would be expected to indicate an elastic response of the rock to the applied stress.

Fig. 2 shows the variation of σ (at room temperature) observed in all of the saturated samples studied. In each case one or more pressure cycles had been carried out to ensure that the variation was reproducible. It would appear from these results that the samples can be divided into two groups on the basis of their pressure dependence. Group A contains the acidic rock types, i.e. the granite gneisses, which show a non-linear decrease of $\log \sigma$ with P_c . Group B contains the basic rock types, i.e. the basic and amphibolite gneisses, which show a smaller, linear decrease of $\log \sigma$ with P_c .

It can be shown (Brace *et al.* 1965; Lee 1977) that an elastic reduction in the cross-section of cylindrical pores would only result in a very small (~ 1 per cent) decrease of conductivity in the relevant pressure range, and it has been suggested (Brace *et al.* 1965) that the observed decrease in σ can only be attributed to the complete cutting-off of conduction paths as the confining pressure increases. Such a cutting-off, however, would have to result from an elastic deformation of the mineral grains surrounding a pore if it was to account for the reversible nature of the conductivity change. This idea has been theoretically evaluated by Walsh (1965), who concluded that for any pore there is a critical pressure required to just bring its walls into contact and therefore cut it off. This critical pressure was found to be approximately proportional to the aspect ratio (the ratio of the minor to major axes) of the pore. Thus long, thin planar pores (i.e. cracks) with low aspect ratios would be closed at low values of P_c while the more circular high aspect ratio pores would remain open to much higher pressures.

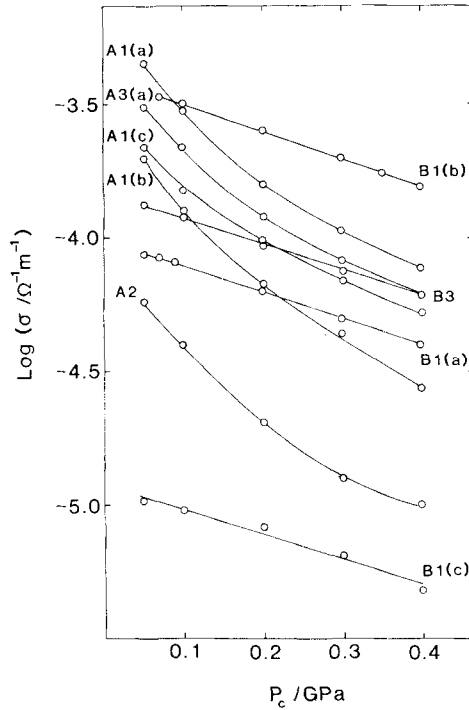


Figure 2. Electrical conductivity of the saturated samples as a function of confining pressure (at room temperature, and with the pore fluid pressure equivalent to atmospheric pressure). All samples were saturated with 0.5 M NaCl solution with the exception of sample B1(c) which was saturated with 0.05 M NaCl solution. The size of the data points indicates the error limits on the measurements.

We therefore attribute the observed decrease of σ with increasing P_c to a successive cutting-off of conduction paths as a result of the elastic deformation of the pore walls. Although it might appear that the different pressure dependence of group A and B samples could be explained by the simple model of different distributions of pore shape, factors such as the degree of interconnection of the pores (which determines the decrease in the total conductance resulting from the loss of any given conduction path) make such an interpretation uncertain without additional information on the detailed pore structure.

4.1.2 Test of Archie's law at room temperature

It was of interest to see whether the magnitude of the measured conductivity was consistent with the predicted values using the empirical relationship known as Archie's Law (Archie 1942). This links the conductivity of a saturated rock σ_R , to its porosity η , through the equation

$$\sigma_R = \sigma_S \eta^m \quad (1)$$

where σ_S is the conductivity of the pore fluid, and the index m has a value between 1 and 2. Brace *et al.* (1965) and Brace & Orange (1968) have demonstrated that for a wide range of rock types the conductivity measured at 0.4 GPa and the estimated 'pore' porosity at this pressure, i.e. after the closure of the low aspect ratio pores or 'cracks', satisfy the form of Archie's Law in which the index m assumes the value 2, i.e.

$$\sigma_R = \sigma_S \eta^2. \quad (2)$$

(Following Brace *et al.*, crack and pore porosity are defined here as the porosity decrease and the porosity remaining on subjecting a rock sample to a confining pressure of 0.4 GPa.)

A crude test of this empirical relationship was carried out using the present data by substituting the measured values of the bulk porosity (Table 1) and an appropriate value for the conductivity of the pore fluid (i.e. $6 \Omega^{-1} \text{m}^{-1}$) in equation (2) and comparing the results with the measured conductivities at 0.4 GPa. Clearly this procedure is not entirely satisfactory in that it differs from that of Brace *et al.* on a number of counts. In this case the porosities were measured on adjacent samples rather than the samples through which the conductivity was measured, it was not possible to estimate the crack porosity because no information was available on the linear compressibility or seismic velocities for the samples, and the conductivity measurements were made at a higher frequency than those of Brace *et al.* The latter two points tend to produce calculated and measured values respectively which are somewhat higher than those which would have been obtained by Brace *et al.* on the same samples. It is difficult to assess the precise effect of these various discrepancies, but in terms of departures from the validity of equation (2) they are likely to be quite small and certainly not greater than an order of magnitude.

On comparing the observed and predicted conductivities it was found that in the case of the rocks of group A, the calculated values were approximately in agreement with the values measured at 0.4 GPa whereas the rocks in group B all had measured values of σ that were between 1 and 2 orders of magnitude higher than the calculated values. Although there are inherent difficulties in the measurement of the very low porosities of these samples, the porosity required to account for the measured conductivity on the basis of equation (2) would have to be considerably greater than that actually measured. For example, in sample B1(b), the measured porosity was 0.07 per cent, compared with the value of 0.5 per cent which would be required to satisfy equation (2), and it is difficult to see why such an inaccurate measurement in the case of the group B samples would not be apparent in the group A samples which also had low porosities.

Evans (1980) has carried out a more rigorous investigation of equation (2) having measured both seismic velocities and electrical conductivity as a function of pressure on samples from a much greater variety of igneous and metamorphic rock types than in the present work. Thus he was able to estimate the crack and pore porosity of each sample and carry out an analysis which was much more akin to that of Brace *et al.* Evans demonstrated that, as here, the more basic rock types showed significant departures from equation (2) and he suggested that for rocks in which the major constituent is free quartz the measured conductivity satisfies this relationship, whereas for rocks in which free quartz is a minor constituent or absent the measured conductivity is always higher than predicted.

Departures from Archie's Law implying enhanced conduction are well known from studies of (much higher porosity) sedimentary rocks saturated with low conductivity electrolytes (Barker 1979). The additional conduction mechanism is presumably associated with the matrix and is typically attributed to the presence of clay minerals in an otherwise non-conducting matrix. Here also it is thought that the enhanced conduction may be associated with certain hydrous layered silicates which are characteristic of basic but not acidic rocks, as discussed below.

4.1.3 Conductivity as a function of pore fluid pressure (at room temperature)

Fig. 3(a) shows the effect of pore fluid pressure on σ in a typical saturated sample. It is clear that at any given value of P_c , the effect of increasing P_p is to increase σ . The two pressures P_c and P_p therefore act in opposition to one another, and it would seem reasonable to

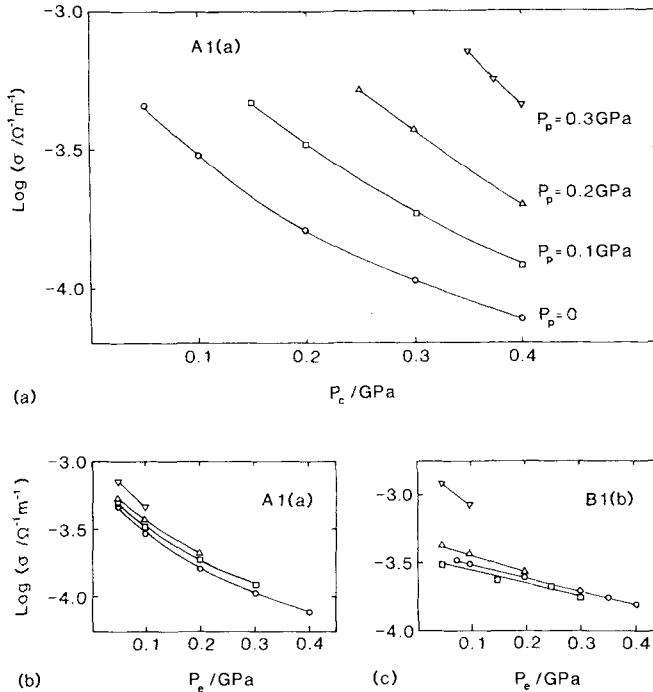


Figure 3. (a) Electrical conductivity of sample A1(a) as a function of confining pressure, at room temperature and for various values of the pore fluid pressure as indicated. (b), (c) Electrical conductivity of samples A1(a) and B1(b) as a function of effective pressure (at room temperature). The size of the data points indicates the error limits on the measurements.

assume that the net deformation of the pore walls (and hence the net change to the measured conductivity) will be determined by the difference between the two pressures. This pressure difference is commonly referred to as the effective pressure, P_e , given by

$$P_e = P_c - P_p \quad (3)$$

although strictly this is an oversimplification (Nur & Byerlee 1971). There is however empirical evidence to suggest that this expression is appropriate in the case of electrical conductivity (Robin 1973).

Fig. 3(b) shows the data of Fig. 3(a) replotted as a function of P_e as defined above. If equation (3) were obeyed exactly, all the points should lie on a single curve, and, in the case of this sample (A1a), deviation from this prediction is slight. About half of the samples studied exhibit similar behaviour, in agreement with this model. Fig. 3(c) shows the behaviour observed in sample B1(b), where departure from a single curve is very much more marked. By regarding departure from equation (3) as implying a non-elastic response of the rock, we interpret this latter behaviour as evidence of fresh cracking when the pore pressure was increased. The samples in which this latter behaviour occurred showed no correlation with the division into groups A and B. One possibility is that fresh cracking is caused in some of the samples studied by a non-equilibrium pressure wave resulting from the movement of pore fluid when P_p was suddenly changed. Such an explanation would imply that the opening of fresh cracks was an artefact of the experimental conditions of the measurement, and, as such, would not be directly relevant to conditions within the Earth.

4.1.4 Conductivity as a function of temperature

As described elsewhere (Lee, Ross & Vine 1977), we found that when the saturated samples were first heated under constant pressure conditions an irreversible behaviour on cycling was observed, after which a reversible temperature dependence was obtained. This phenomenon is thought to result from thermally activated creep processes and gives rise to a drop in the measured conductivity of approximately one-half of an order of magnitude from the value obtained before thermal cycling. Again further experiments by Evans (1980) have confirmed the existence of this transient in relation to a wider variety of rock types than considered here.

Fig. 4(a) shows the reversible variation in conductivity with temperature relative to the conductivities at 20°C for the saturated samples studied here. In order to make a direct comparison of these results with the relative variation in pore fluid conductivity, a correction has to be made for the effect of thermal expansion on the conduction paths through the pores of the samples (the effect of thermal expansion on the external dimensions of the sample having already been taken into account). This can be estimated by assuming that the effect of volume change on conductivity is the same whether due to a change in temperature or a change in pressure. For a given increase in temperature, the decrease in pressure which would produce the same change in volume may be estimated from the expression:

$$\frac{\delta V}{V} = -\beta \delta P = \alpha \delta T \tag{4}$$

where β is the isothermal compressibility and α is the thermal expansivity. Using typical values of α and β for these rock types (Clark 1966) the volume change resulting from a given

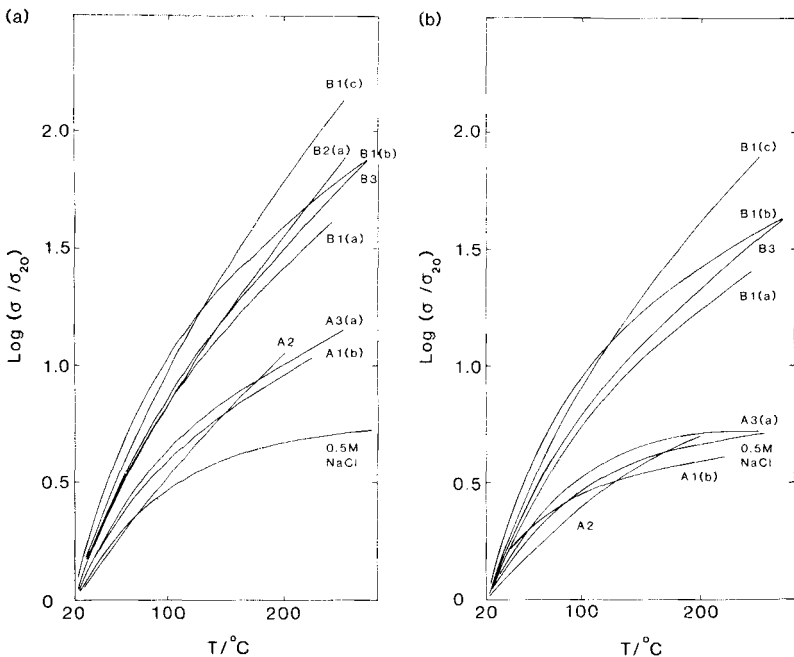


Figure 4. (a) Relative change in the conductivities of the saturated samples as a function of temperature, at a constant effective pressure of 0.35 GPa in all cases except for samples A3(a) and B3 for which it was 0.2 GPa. (b) the same results having removed an estimate of the effect of thermal expansion of the matrix (see text). The errors in the measurement of conductivity are comparable to those shown in Fig. 5.

increase in temperature was calculated, as was the change in pressure required to produce the same change in volume. The measured pressure dependence of the conductivity (Section 4.1.1) was then used to estimate the change in conductivity for such a volume change. Fig. 4(a) was then replotted to take account of these estimated conductivity corrections for pore geometry changes resulting from thermal expansion (see Fig. 4b).

As with the pressure dependence, the samples fell into two distinct groups, A and B, on the basis of their temperature dependence. On the Brace model, of conduction entirely through the pore fluid, these corrected curves should coincide with the relative change in conductivity of the pore fluid with temperature. From the variation in conductivity of 0.5 M NaCl solution, also included in Fig. 4(b), it can be seen that for group A samples this was in fact the case. In this group of samples, therefore, both the magnitude of the measured conductivity at room temperature, and the temperature dependence, are consistent with the idea that conduction occurs entirely through the pore fluid with which they are saturated. Fig. 4(b) shows that in the group B samples there is a much greater variation of σ with temperature than expected. It has already been noted, in Section 4.1.2, that these samples had room temperature values of σ that were greater than could be accounted for on the basis of Archie's Law. Both these facts indicate that, in this group of samples, conduction does not occur through the pore fluid alone. In order to examine this possibility, measurements were made on samples that had not been saturated with a pore fluid.

4.2 UNSATURATED SAMPLES

Two samples of amphibolite gneiss from group B (samples B1(d) and B2(b)) were prepared as described in Section 2, except that after grinding they were allowed to dry in air before assembly. However, no other attempt was made to remove any moisture content before the measurements were made. It was expected that their conductivities would be very low and show a negligible pressure dependence, i.e. the same properties as the majority of the materials from which they were composed (Parkhomenko 1967; Keller 1966). Somewhat surprisingly, we found that their conductivities were higher than expected, and furthermore, that they exhibited a similar pressure dependence to that of the saturated group B samples. On heating, these samples had their pores open to the atmosphere, and it was expected that if any natural pore fluids were present, the value of σ would drop significantly at 100°C as the pore fluid evaporated. This was not found to be the case, and the same behaviour on cycling that occurred in the saturated samples was observed.

Fig. 5 shows that the temperature dependence (after thermal cycling) of the conductivity of the two unsaturated group B samples is similar to that of the saturated samples. One possible explanation of these results could be the presence of conducting minerals, possibly in the form of hydrous alteration products, within (or lining) the pores. In such minerals, water can be tightly bound, and dehydration may not occur until temperatures considerably in excess of 300°C are reached (Turner 1968). Thin sections were examined to try to identify such minerals. Unlike the granite gneisses of group A, which had highly idiomorphic crystalline structure with high quartz content, the basic and amphibolite gneisses of group B were typically lineated, and composed of hornblende, plagioclase, orthoclase and epidote. In all of them there was evidence that the hornblende was partially altered to the hydrous mineral chlorite, which together with the hornblende and/or epidote might provide additional conduction paths associated with the bound and/or adsorbed water which characterizes these minerals.

To test this explanation further, a group B sample was pre-baked to 750°C before assembly in order to dehydrate any alteration products which might be present and to

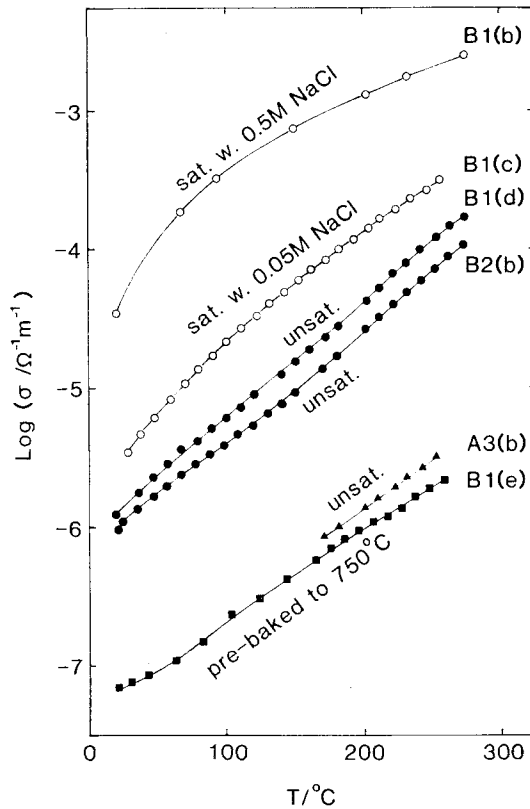


Figure 5. Reversible electrical conductivity of selected samples as a function of temperature following thermal cycling at a confining pressure of 0.35 GPa. Data for sample A3(b) were only available over a restricted temperature range due to instrument failure. The size of the data points indicates the error limits on the measurements.

remove any adsorbed water. In this case the measured conductivity was independent of pressure. On heating there was no evidence of the behaviour on cycling observed in all of the saturated and unsaturated samples, and, as Fig. 5 shows, the conductivity was considerably lower than the unsaturated group B samples. Only the conductivity of the unsaturated group A sample (also included in Fig. 5) was comparable to the pre-baked group B sample, and it was concluded that the adsorbed water and mineral or minerals providing the additional conduction paths were not longer present in the baked sample, presumably because of dehydration reactions which occur in the range 300–750°C.

5 Summary of results

Our principal conclusion is that the saturated samples measured can be divided into two groups on the basis of the presence (in the basic and amphibolite gneisses) or absence (in the granite gneisses) of an additional conduction path in parallel to that through the pore fluid. We suggest that this additional conduction mechanism could be due to the presence of certain minerals in or lining the pores.

Evidence for the presence or absence of such an additional conduction path was based on both the magnitude of the room temperature conductivity, and on the temperature dependence of σ . The samples divided into the same two groups on the basis of their pressure

dependence at room temperature, and, although it was suggested in Section 4.1.1 that the different behaviour of the two groups could be associated with the nature of the pore structure, it is also possible that the different pressure dependent behaviour could be due to the presence or absence of certain hydrous minerals, i.e. it might be more difficult to close off a conduction path which contains such minerals, resulting in a slower decrease of σ with P_c .

Although there have been comparatively few studies of the conductivity of saturated crustal rocks as a function of pressure and temperature, the results available appear to divide into the two groups reported here. In measurements on granites up to temperatures of approximately 200°C, Parkhomenko *et al.* (1972) found that the conductivity could be accounted for by conduction through the pore fluid alone, and a similar conclusion was reached by Brace & Orange (1968) in measurements on a great variety of rocks. Hyndman & Drury (1976) and Drury (1979), however, in studying seawater-saturated basalts have reported a much greater than expected variation of σ with temperature which they attributed to the presence of clay minerals. Stesky & Brace (1973) have also reported anomalously high values of σ at room temperature in serpentinites, i.e. altered ultramafic rocks, which they suggest might be due to the presence of magnetite along the grain boundaries.

Further studies are necessary therefore in order to identify unequivocally the additional conduction mechanisms associated with certain, typically basic, rock types.

6 Possible geophysical implications

On the basis of the results presented above it is possible, although somewhat speculative, to make estimates of the variation of electrical conductivity with depth in the continental crust. Needless to say any such estimate is hedged around with a great number of assumptions and uncertainties but in view of the apparent novelty of these experiments in terms of the physical conditions attained, the rock types studied and the results obtained, this exercise would appear to be warranted. Certainly the implications are quite unlike those suggested previously on the basis of laboratory measurements, and are of considerable interest in relation to the results of recent controlled source, geomagnetic and magnetotelluric field experiments in continental areas.

In order to make such estimates one must make assumptions about the geothermal gradient, the presence or absence of a pore fluid, the pore fluid pressure and pore fluid composition if present, and the porosity and composition of the rocks throughout the crust.

In the first instance the following assumptions will be made, that:

- (1) the rocks throughout the crust are saturated with pore fluid, the composition of which is 0.5 M NaCl solution;
- (2) the pore fluid pressure remains hydrostatic through the crust, i.e. a continuous pathway exists from any depth to the surface;
- (3) the various transient effects, which were observed in the laboratory studies, i.e. before the rock samples responded reversibly to changes in temperature and pressure, all occur at comparatively shallow depths given a geological time span;
- (4) the results obtained for upper crustal pressures and temperatures can be extrapolated to lower crustal conditions;
- (5) an 'effective pressure model' is obeyed;
- (6) the acceleration due to gravity, g , and the density of rocks and pore fluid remain constant with depth;
- (7) the limited size and number of samples studied in the laboratory are representative of large rock masses in the Earth.

Predicted profiles for three regions of the United States have been calculated, in part to enable a comparison to be made with the predicted profiles of Brace (1971), but also because the geothermal gradients and the structure of these provinces have been widely studied.

The variation in conductivity with depth for the acidic and basic groups (Fig. 6) was calculated as follows:

(1) The confining pressure, P_c , at any depth, h , due to the column of rocks above, was calculated using $P_c = \rho gh$, where ρ was the mean density of the samples studied.

(2) The pore fluid pressure, P_p , at any depth was calculated in the same way, using a value of $\rho = 1000 \text{ kg m}^{-3}$. Strictly speaking the specific volume and hence density of the pore fluid will vary considerably with temperature and pressure; see, for example, the values for pure water given by Burnham, Holloway & Davis (1969). In the cases considered here, however, the effects of elevated temperatures and pressures tend to cancel out unless the

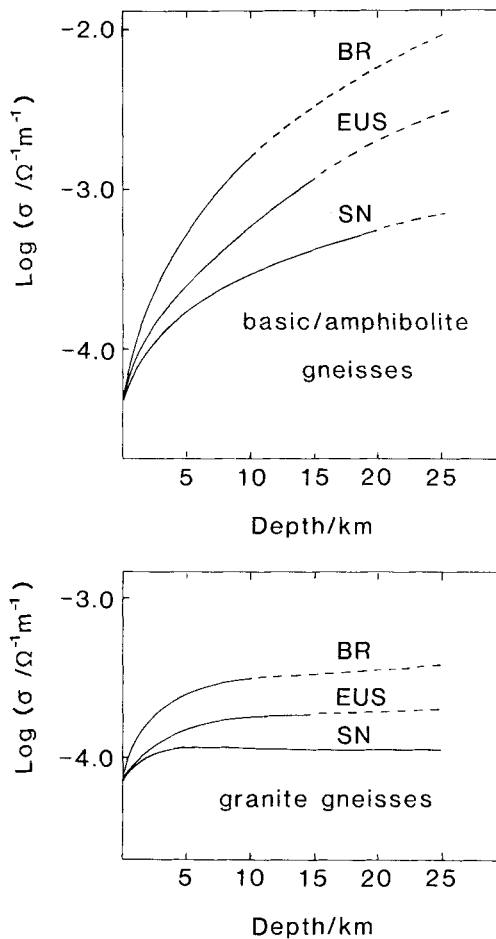


Figure 6. Predicted variation in conductivity with depth for the three heat flow provinces identified by Roy *et al.* (1968) i.e. the Basin and Range (BR), the Eastern United States (EUS), and the Sierra Nevada (SN). The profiles have been derived from the experimental data by assuming that crust composed of the appropriate rock type is saturated with a pore fluid of electrolytic composition equivalent to 0.5 M NaCl solution, that the pore fluid pressure throughout the crust is hydrostatic and that the geothermal gradients are as given by Roy *et al.*

pore fluid pressure is very low or the geothermal gradient very high in which case the density can be as little as 500 kg m^{-3} or even less; see Brace (1971).

(3) The value of the effective pressure at any depth was found using equation (3) and from the observed variation of conductivity with effective pressure, the variation of conductivity with depth, due to pressure effects alone, was calculated for both rock types.

(4) The variation of temperature with depth in the Basin and Range (BR), Eastern United States (EUS), and Sierra Nevada (SN) heat flow provinces was taken from the analysis of Roy, Blackwell & Birch (1968), although clearly this is not strictly appropriate for the hypothetical and uniform crustal models being assumed at this stage. These temperatures were then combined with the observed variation of conductivity with temperature (taken from the data in Fig. 4a) to produce the variation of conductivity with depth due to temperature effects alone.

(5) The pressure and temperature effects were then combined to produce an overall variation of conductivity with depth for each rock type in the three regions, the absolute conductivity values being derived from the reversible conductivity/temperature curves taking into account the effective pressure at which these were obtained.

In an attempt to satisfy both the heat flow data and the seismic refraction data Blackwell (1971) made estimates of the depth to a 'granitic' rock/'basic' rock interface within the crust in each of these three provinces. His model assumed depths to this interface of 19, 15 and 20 km in the BR, EUS and SN regions respectively, and depths to the base of the crust of 30, 35 and 40 km respectively. Using calculated conductivity/depth data for the granitic gneisses above the 'granitic'/'basic' interface, and data for the basic gneisses below, the profiles shown in Fig. 7(a–c) were constructed for the three regions. Below the Moho, the dry rock conductivity assumed by Brace (1971) has been added. A typical value for the conductivity found near the Earth's surface (Keller, Anderson & Pritchard 1966), due to the presence of major fractures and cracks, has also been included, and it has been assumed that such near surface effects only occur to a depth of about 1 km. The profiles predicted by Brace (1971) for the same regions are shown as dashed lines in Fig. 7(a–c). As indicated above, the temperature profiles assumed in these calculations were those given by Roy *et al.* (1968), since these were the temperatures assumed by Brace. Clearly in view of the specific crustal models assumed here it would have been more appropriate to have used the temperature profiles deduced by Blackwell (1971) for these particular models. Below depths of about 15 km these are systematically higher than the temperatures assumed here, thus as noted by Blackwell, in relation to Brace's results, both the 'wet' and 'dry' conductivities predicted here represent a lower bound for this particular set of assumptions.

It is perhaps useful to state the differences and similarities between the assumptions made here and those made by Brace (1971) in predicting conductivity as a function of depth in the crust. Clearly the thermal gradients assumed are identical, as indicated above, and the assumption of hydrostatic pore fluid pressure is also common to both analyses. Similarly the rock porosities assumed are very comparable. Of the differences between the two models the two major ones are the higher salinity assumed in this study for the pore fluid permeating the acidic rocks, approximately a factor of 5 higher, and the higher conductivities obtained and hence assumed here for the basic rocks. Differences are also produced by the fact that Brace's predictions do not take into account thermal transients or the effect of thermal expansion on the effective porosity. Dvorak (1975), using Brace's technique, has predicted a variation in conductivity with depth throughout the continental crust which is very analogous to that deduced here. Within the limitations of this technique however this has only been possible because the author has assumed a porosity of approximately 3 per cent throughout the middle and lower crust and an order of magnitude difference in the

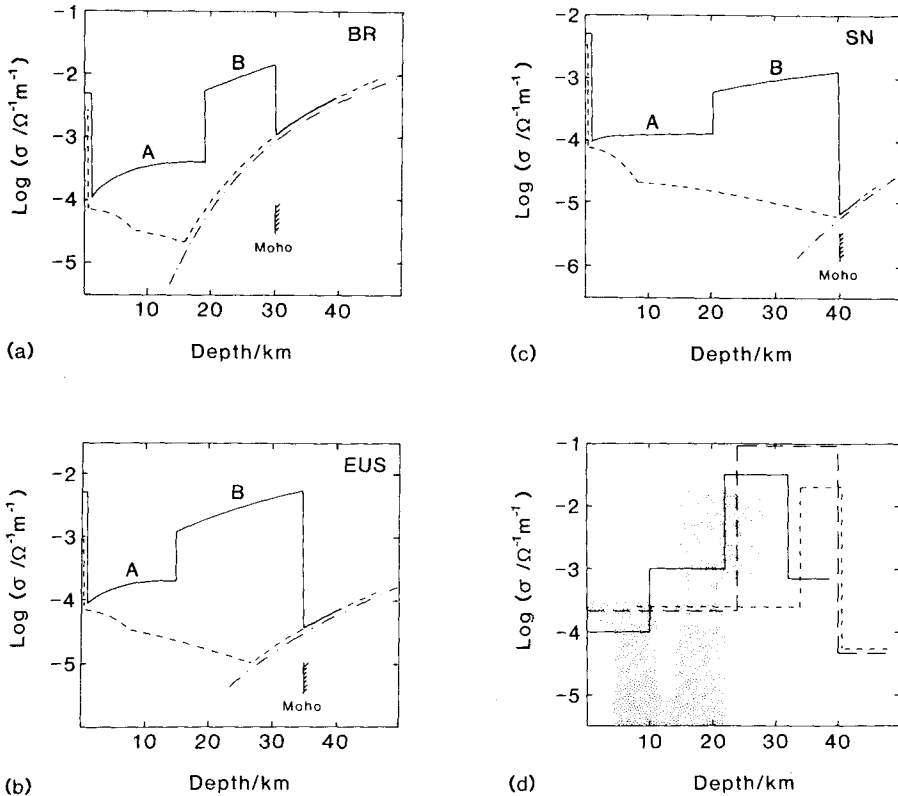


Figure 7. (a–c) Solid curves indicate the predicted variation in conductivity with depth in the Basin and Range, Eastern United States and Sierra Nevada regions respectively assuming that the upper, ‘granitic’ part of the crust can be represented by the rocks of group A, that the lower ‘basic’ part of the crust can be represented by the rocks of group B, and that the physical and chemical conditions in the crust are as given in the caption to Fig. 6 and in more detail in the text. Below the Moho the mantle is assumed to be ‘dry’ and typical dry rock conductivities are indicated by the dash-dot line. The conductivity/depth curves suggested by Brace (1971) for the same three heat flow provinces are indicated by dashed lines. (d) Electrical conductivity as a function of depth in various stable Precambrian shield areas as deduced from controlled-source field experiments. The solid curve is for the Adirondacks (Connerney *et al.* 1980); the curves defined by the long and short dashes are for the Limpopo belt and the Kaapvaal craton respectively (Van Zijl 1977) and the shaded area summarizes the range of inversions obtained for the Wisconsin arch area (Sternberg 1979).

salinity of the pore fluids permeating an upper ‘granitic’ crust and a lower ‘basic’ crust. We consider both of these assumptions, and particularly the former, to be geologically implausible.

Because of the relatively high salinity assumed for the pore fluid it might be thought that the conductivities predicted in Fig. 7(a–c) represent upper limits on the basis of plausible assumptions. However, it must be emphasized that this is not so because the pore fluid pressure could be considerably greater than the hydrostatic pressure and conceivably comparable to the confining pressure. The low effective pressures implied by higher pore fluid pressures could give rise to appreciably higher conductivities, as illustrated in Fig. 8. It will be clear from this diagram that major changes in the pore fluid pressure with depth in the crust could also give rise to major changes in conductivity; that is, in addition to or in place of changes due to changes in the bulk mineralogical composition of the crust. Mitchell &

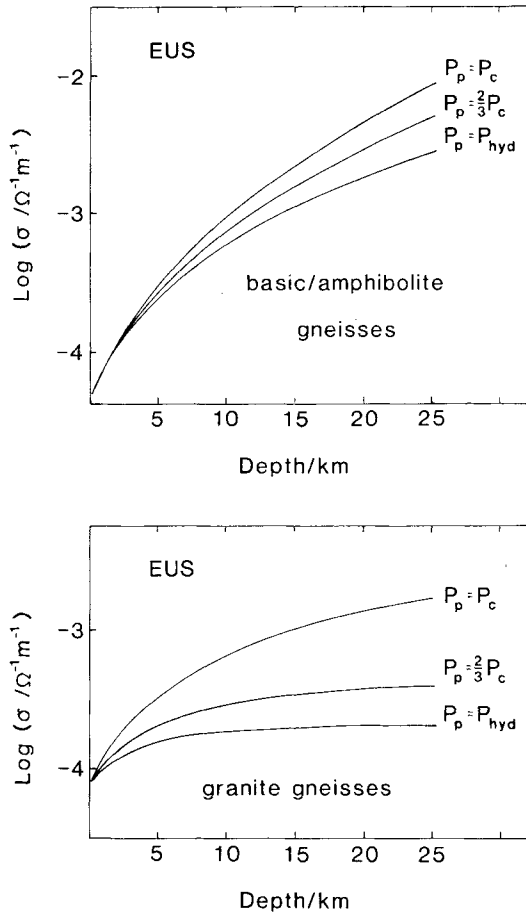


Figure 8. Curves illustrating the effect of different pore fluid pressures (P_p) on the predicted variation in conductivity with depth for the Eastern United States heat flow province. Other assumptions as for Fig. 6.

Landisman (1971) invoked water saturation and increased pore fluid pressure due to dehydration reactions to explain essentially coincident high electrical conductivity and low seismic velocity layers at a depth of approximately 20 km beneath eastern New Mexico. If such layers are indeed coincident then a change in pore fluid pressure and hence effective pressure would appear to be the only plausible explanation in areas of near or below average geothermal gradient. The compositional changes which have been suggested are not capable of producing the required effect in both parameters. In areas of above average geothermal gradient however, such as the Basin and Range province of the western United States, an explanation in terms of partial melting is more likely, as suggested by Lienert & Bennett (1977).

There is now mounting field evidence from geomagnetic, magnetotelluric and controlled-source experiments that high conductivity layers are a common but not a ubiquitous feature of the lower continental crust. Connerney, Nekut & Kuckes (1980), Kay & Kay (1981), and Jones (1981) have recently reviewed this evidence. Some of the results of controlled-source experiments in stable Precambrian shield areas are summarized in Fig. 7(d) on the assumption that controlled source experiments give the best estimates of the depth to and conductivity of

the high conductivity lower crustal layer (Garland 1981). From this it will be seen that the inferred depth to this high conductivity layer varies from 14 to 34 km and that its conductivity is of the order of $10^{-2} \Omega^{-1} \text{m}^{-1}$. It will be seen from Figs 7(b) and 8 that in these areas, with a normal or somewhat below average geothermal gradient, as in the eastern United States, such high conductivities imply a combination of high pore fluid pressures *and* basic rock types on the basis of the results presented here. It is noteworthy that in the Basin and Range province the conductivity of the high conductivity layer deduced from the field data is of the order of $1 \Omega^{-1} \text{m}^{-1}$ (Lienert & Bennett 1977) which perhaps favours the alternative explanation of partial melting, for at the temperatures which characterize the lower crust in these regions, i.e. in excess of 400–500°C, the increase in conductivity of the electrolyte with increasing temperature begins to drop off (Quist & Marshall 1968, 1969) and many hydrous rock forming minerals take part in dehydration reactions. In shield areas however temperatures at the *base* of the crust probably do not exceed 400–500°C (Blackwell 1971).

Our results on 'unsaturated' basic rock samples suggest that the pore fluid in the lower crust need not necessarily be wholly liquid or electrolytic. Chemically bound or adsorbed water in certain minerals may play a crucial role. Such an interpretation of the high conductivities deduced for the lower crust was first discussed by Hyndman & Hyndman (1968), although these authors thought that it might only be applicable to geosynclinal areas and young mountain belts, the crust beneath stable shield areas being dehydrated and 'granulitic'. However, the electrical conductivity results suggest that pore fluids which give rise to low effective pressures and contain at least some water may be more widespread in the lower continental crust than has previously been assumed. In this context it should be borne in mind that the information on the presence and composition of pore fluids in the crust inferred from metamorphic petrology and fluid inclusions typically relates to metamorphic events which took place in the distant geological past and is not necessarily relevant to the conditions prevailing at the present day which determine the geophysical results.

It is evident that, if the results of electrical soundings are to be correctly interpreted, a much clearer understanding of the role played by water, bound or otherwise, is needed, particularly in rocks of basic composition and at high temperatures and pressures.

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