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# Water in the lower continental crust: modelling magnetotelluric and seismic reflection results*

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## SUMMARY

Magnetotelluric and multichannel seismic reflection measurements indicate that the Phanerozoic lower continental crust is commonly electrically conductive and reflective, in contrast to a more resistive and transparent middle to upper crust. A few per cent free saline water can provide an explanation for both results along with the apparent requirement that neither the conductive nor the reflective properties are retained when lower crustal rocks are brought to the upper crust. Common 10 km thick and 20–30 Ωm resistivity layers can be explained with 0.5–3 per cent pore water, if there are equilibrium pore geometries and the salinity is close to that of sea-water as suggested by lower crust fluid inclusions. Seismic velocities and impedances must be affected if such porosity exists. Seismic reflectors with reflection coefficients of 5–10 per cent can be explained by layers or lamellae with porosity contrasts of 1–4 per cent and reasonable effective pore aspect ratios of 0.1–0.03. A minimum temperature of 350 °C is estimated from a correlation between heat flow and depth to the top of conductive and reflective layers. The upward limit in the crust may occur at an impermeable boundary formed by hydration reactions at the top of greenschist facies conditions or by precipitation of silica. It also may be associated with the minimum temperature for ductile behaviour and equilibrium grain boundary pore configurations. The maximum temperature is about 700 °C according to the evidence indicating that there is no free water in granulite facies conditions. Areas that have been subject to such high temperature conditions without the subsequent addition of water, i.e. the lower crust of shields, are generally non-reflective and electrically resistive.

**Key words:** deep crustal reflections, electrical resistivity, lower crust, magnetotellurics, porosity

## INTRODUCTION

The lower continental crust is commonly electrically conductive and seismically reflective. Why should this be so? In this paper we discuss the possibility that these properties have a common origin in the presence of free water (e.g. Gough 1986).

Direct information on the composition, structure and state of the lower continental crust comes primarily from magnetotelluric and other electrical sounding techniques, and from seismic reflection and refraction. The geophysical techniques have brought surprising results. Magnetotelluric measurements have shown that the Phanerozoic lower continental crust is commonly very electrically conductive and multichannel seismic reflection surveys have defined widespread regions of lower crustal reflectivity in contrast to usually transparent upper crust. In comparison, the lower crust in shield areas is generally resistive and has few reflectors. The results from both measurement techniques in the younger areas were unexpected. The high electrical conductivity was surprising because most common minerals when dry are very electrically resistive, and fluids were not expected in the lower crust since at high pressure it was thought that the porosity would be small (e.g. Brace 1971). Also, increasing temperature usually results in dehydration. The lower crustal reflectors were also unexpected, since it was thought that a variety of geological processes would at higher temperatures tend to homogenize the lower crust compared with the upper crust. In addition, exhumed very high grade metamorphic rocks such as granulites generally considered to be of lower crustal origin are rarely conductive or reflective and usually have a mineralogy indicating dry conditions.

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A wide variety of explanations have been suggested for both the conductive and reflective properties of the lower crust. In this article we briefly review the various explanations, but concentrate on the hypothesis that these two properties and a number of other more indirect data can be explained by a few per cent porosity in the lower crust. This hypothesis can explain the observation that the conductive and reflective properties of lower crustal rocks do not generally appear to be retained when the rocks are brought to the upper crust. Both properties appear to be restricted to depths where the temperature is above about 400 °C, and there may also be an upper temperature limit. Thus heat flow and inferred crustal temperatures are important to the discussion.

A major part of the paper presents the results of our attempt to model as quantitatively as possible the porosity and pore structure that would be required to produce the observed conductive zones and seismic reflectivity in the lower crust. In an associated article Hyndman & Klemperer (1989) discuss the effect of possible lower crustal porosity on seismic refraction velocities and inferred lower crustal compositions.

## DIRECT CORRELATION OF REFLECTIVITY AND RESISTIVITY ON SPECIFIC PROFILES

A central feature of this hypothesis is that low resistivity and deep crustal reflectors have a common origin. If this is the case, we should expect to see a strong correlation between seismic and resistivity data at lower crustal depths. Gough (1986) noted that there was often such a good correlation and that fluids might be the origin. Jones (1987) pointed out that the depths to the reflectors and conductors often did not correspond exactly, but with the depth uncertainty in magnetotelluric models (see discussion below) they could be coincident. Of course low resistivity may occur without seismic reflectivity if the required layering is not present. Unfortunately, only infrequently have coincident high-quality seismic reflection and magnetotelluric surveys been carried out; this is important work yet to be done. However, there are a number of areas where at least rough associations can be made.

Jones (1987) discussed examples in six areas: Vancouver Island (western Canada), Sweden, Germany, France, NE United States and Scotland. A particularly interesting correlation over a large region is in eastern North America (e.g. Haak & Hutton 1986). The region of Scotland that adjoined eastern North America prior to the Atlantic opening has given similar results. A conductive lower crust has been established over much of the Appalachian region by magnetotelluric, controlled source and geomagnetic deep sounding measurements. In a few areas controlled source measurements have been made that give accurate depths to conductive lower crust, such as in Georgia (S Appalachians) and Adirondacks (N Appalachians), and some of the results have been confirmed by magnetotelluric results (e.g. Connerney, Nekut & Kuckes 1980; Thompson, Nebut & Kuckes 1983; Haak & Hutton 1986, and references therein). The regional distribution of the lower crustal conductor over most of the Appalachians has been established by geomagnetic deep sounding (e.g. Greenhouse & Bailey 1981; Mareschal, Musser & Bailey 1983). The general regional reflectivity of the lower crust in the Appalachians has been discussed extensively (e.g. Cook *et al.* 1979; Ando *et al.* 1984). More specific correlations with electrical resistivity are possible in the Adirondack Mountains (e.g. Klemperer *et al.* 1985; Connerney *et al.* 1980) and in Georgia (Cook & Oliver 1981; Thompson *et al.* 1983). The Scottish magnetotelluric data is also close to a number of BIRPS seismic lines that exhibit strongly reflective lower crust (e.g. Matthews 1986).

A good correlation between seismic reflectivity (e.g. McCarthy & Thompson 1988) and low electrical resistivity (e.g. Jiracek, Rodi & Vanyan 1987) occurs in the lower crust of Nevada and other parts of the Basin and Range area of the SW United States. However, we have excluded such areas of Tertiary tectonic activity in the modelled average lower crust data since, although saline fluids provide an attractive explanation for the observations, partial melt may be present.

Another special area is the continental crust above subducting oceanic crust. Kurtz, DeLaurier & Gupta (1986) have found a conducting layer dipping inland at depths of about 20–35 km beneath the Cascadia subduction zone at Vancouver Island. It corresponds to a well-defined dipping reflective band 5–10 km thick (e.g. Clowes *et al.* 1987; Green *et al.* 1987). Hyndman (1988) has modelled both results in terms of water driven off the downgoing oceanic plate by dehydration reactions and trapped in the overlying continental crust. A similar conductor has been found dipping inland beneath Oregon by Waff *et al.* (1988).

## MEASUREMENT OF LOWER CRUSTAL ELECTRICAL RESISTIVITY

In the analysis and interpretation of lower crustal electrical resistivity estimates, it is important to recognize the limitations of the measurements. Most information on the electrical conductivity structure of the lower continental crust comes from the magnetotelluric method (e.g. Kaufman & Keller 1981). Some confirmation of lower crustal high conductivity layers has also come from analysis of three component magnetic field variations (geomagnetic depth sounding) and by controlled source techniques (e.g. Connerney *et al.* 1980; Thompson *et al.* 1983), but the depth resolution of the former is poor, and the latter has been carried out on the scale required to reach lower crustal depths at only a few sites.

Electromagnetic induction in the earth is a diffusion process with long wavelength. The response at the surface depends on the electrical conductivity structure down to the maximum depth of induced currents at a particular frequency. Thus, the magnetotelluric technique has inherently poor depth resolution compared with seismic reflection. The limitations and problems have been well summarized by Jones (1987). It is for example very difficult to distinguish between a very conductive thin layer and a moderately conductive thicker layer. Only the depth to the top and the conductance or the thickness–resistivity ratio of the layer are well defined unless more than one method is employed at the same site (Edwards, Bailey & Garland 1981). It is also difficult to obtain deep information beneath surface conductors such as sediments. Of critical importance is the phenomenon of DC shifts of apparent resistivity curves by local near surface inhomogeneities (often called 'static shift', e.g. Jones 1987) that result in depth and resistivity offsets in the resulting one-dimensional conductivity–depth models. Much progress has been made during the past 10 yr in the collection of high quality broad frequency band data, in processing techniques, in model inversion and in treating 2-D and 3-D structures. For older data in which the uncertainties in depth and resistivity could be at least a factor of 2, there was considerable doubt as to the occurrence of high conductivity in the lower crust (e.g. Porath 1971). However, with modern data, little doubt now remains that the lower crust is commonly conductive (e.g. Edwards *et al*. 1981).

The observed resistivities are very different from those predicted from laboratory measurements on probable deep crustal rocks. For example Brace (1971) predicted pore fluid conduction in the upper crust and the resistivity increasing with depth as pressure closes the pore spaces. The predicted trend reverses in the lowermost crust as the temperature becomes high enough for semiconduction processes to become important. However, at temperatures below the minimum melting point, i.e. less than about 700 °C, semiconduction processes cannot give resistivities below about $10^3\,\Omega$m, which is much higher than the values generally observed.

While the resistivity in the middle to upper crust is relatively high, it is still much too low to be explained by conduction in any common dry rocks by several orders of magnitude. Small amounts of fluid appear to be required (e.g. Shankland & Ander 1983). The uppermost mantle exhibits resistivities however close to those expected from dry mineral conduction, based on laboratory measurements simulating *in situ* temperature and pressure. Field data for the upper mantle also exhibits the much more rapid temperature dependence of semiconduction mechanisms compared with saline fluids. Partial melting may be important in a few areas of the upper mantle.

## GEOLOGICAL ENVIRONMENT OF CONDUCTIVE LOWER CRUST AND AVERAGE LOWER CRUSTAL RESISTIVITY

A number of compilations of lower crustal resistivity data have been carried out that permit estimates of average resistivity and correlations with geological environment (e.g. Shankland & Ander 1983; Adam, 1976, 1987; Haak & Hutton 1986; Jones 1981; Hjelt 1988). The primary geological association is with geological or tectonic age. The highest resistivities in the lower crust, $10^3\,\Omega$m or higher, are associated with the cores of the oldest Precambrian cratons ('type I' shield areas of Jones 1981; 'anomalously resistive areas' of Haak & Hutton 1986); moderate resistivities with adjacent stable areas; and the lowest resistivities, i.e. less than 100 $\Omega$m, with Phanerozoic mobile belts and those areas tectonically active at present. This association is similar to that for lower crustal reflectors as discussed below.

It is evident that all areas have deep crustal resistivities lower than that of dry rocks and fluid is inferred to be present in variable amounts. However, for comparison with areas of high reflectivity in the lower crust, we focus on those areas where the last thermotectonic event was in the Phanerozoic. For modelling, we have identified the

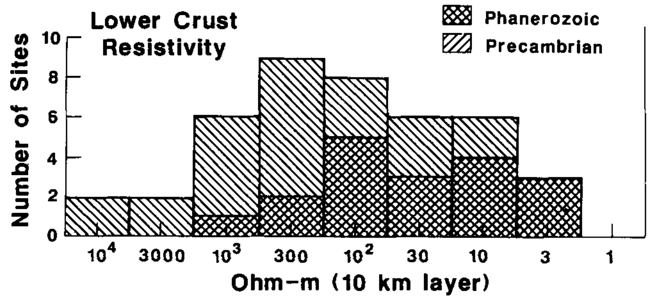


**Figure 1.** A histogram of lower crust resistivity for Phanerozoic and Precambrian areas from Shankland & Ander (1983) and Haak & Hutton (1986). A 10 km layer thickness is assumed.

Phanerozoic sites in the compilations of Haak & Hutton and Shankland & Ander. Most of these sites are included in their categories of 'anomalously low resistivities' and 'high conductivity layers', respectively. They roughly correspond in frequency of occurrence and area represented, to the reflective regions. Fig. 1 presents a histogram of the inferred resistivities for Phanerozoic and Precambrian sites. Areas with Tertiary tectonic activity have been excluded. As discussed above, only the conductance is well determined, so the results are best compiled as frequency of occurrence versus conductance. The 18 Phanerozoic sites have a geometric mean conductance of about 400 Siemens (or km $\Omega$m$^{-1}$) with most values lying between 40 and 4000. The average layer thickness reported is about 10 km, similar to that for reflective lower crust. Thus, the mean resistivity is 20–30 $\Omega$m, with most values between 3 and 100 $\Omega$m. In contrast, the Precambrian shield areas have a mean resistivity of about 500 $\Omega$m.

The seismic reflection data indicate that lower crustal porosity may be in roughly horizontal lamellae. Thus, the resistivity of the high porosity individual layers may be lower by roughly a factor of 2 compared with the overall average measured by electrical sounding.

## ORIGIN OF CONDUCTIVE LOWER CRUST

As discussed above the resistivity of the Phanerozoic lower crust is usually several orders of magnitude lower than that predicted from laboratory measurements on dry probable rock types (e.g. Brace 1971) and some special conduction mechanism is required. Only Precambrian shields have observed lower crustal resistivities of over $10^3\,\Omega$m, while common dry rocks at lower crustal temperatures have resistivities over $10^4\,\Omega$m.

Saline fluids are more generally agreed to be the origin for conductive lower crust than for lower crustal seismic reflectors, but other origins have been suggested. The primary alternative candidate is graphitic shales, but metallic oxides, sulphur, carbon dioxide, and hydrated minerals such as serpentinite and clays have also been suggested. Partial melting can give very low resistivities, but it is a viable mechanism in only a few areas where lower crustal temperatures exceed at least 650 or 700 °C. We have excluded these areas in the data averages.

Black shales are relatively common and their association with localized crustal conductors has been made in several areas including the Rhenish Massif and central North

America (see discussions in Haak & Hutton 1986; Shankland & Ander 1983; Alabi 1984). As pointed out by Shankland & Ander, only a few per cent of graphite is required for interconnected high conductivity channels (e.g. Keller & Frischknecht 1966) and graphite has been observed in some lower crustal xenoliths (e.g. Padovani & Carter 1977). However, this mechanism, and most of the other alternates, does not explain the apparent temperature dependence of resistivity in the lower crust of Shankland & Ander (1983) which is that of saline fluids. As pointed out by Haak & Hutton, it also requires that graphite or the other conductive material be more common in younger areas compared with shields and that it be concentrated in the lower crust (above about 400 °C) compared with the upper crust. There are no obvious mechanisms for such distributions, nor field evidence that they occur.

In the discussion on porosity below we have assumed that the fluid is located mainly in grain boundary porosity, although it is possible for it to be located in larger lenses (e.g. Fyfe, Price & Thompson 1978; Fournier 1988).

## MATRIX AND PORE FLUID RESISTIVITIES

Modelling the porosity needed to produce the observed lower crustal resistivities requires three parameters, the pore fluid resistivity, the matrix resistivity and the pore geometry. We discuss each of these parameters and their effect on the estimated porosity in the next sections.

### Pore fluids

Three main sources of information on lower crustal pore fluid salinity are available, (1) the initial composition from models of the fluid source, (2) from experimental or theoretical reconstruction of the compositions of fluids in equilibrium with particular metamorphic mineral assemblages, and (3) from fluid inclusions estimated to have been formed under lower crustal conditions. The latter source appears to give the most quantitative information.

The initial composition of pore fluids at the time of formation of most sedimentary and many igneous rocks is probably close to that of the sea-water in which they were deposited. This is also the initial composition of fluids that have their origin in the dehydration of downgoing sediments and oceanic crust in subduction zones. It may provide a minimum salinity for estimates of *in situ* composition, since a number of deep metamorphic processes tend to increase the salinity. However, substantial changes in composition occur during emplacement and alteration in the lower crust and the pore fluids of lower crustal rocks now exposed at the surface are rarely representative of their former composition at depth, so other estimates must be sought.

Fluid inclusions are very small volumes of fluid trapped within mineral crystals by a variety of processes such as recrystallization and the healing of microfractures. They are surprisingly unaffected by subsequent processes and often appear to be representative of the fluids present at great depth. Fortunately, it is also possible to estimate the temperature at the time of formation, and to a lesser accuracy the pressure (cf Hollister & Crawford 1981; Roedder 1984) and thus identify inclusions formed at lower crustal conditions.

The main solvent found in deep fluid inclusions is water, with $CO_2$ of secondary importance. There is an important exception that will be returned to later, a change from primarily aqueous solutions to primarily $CO_2$ in very high grade metamorphism, i.e. granulite, over about 700 °C. The primary solute is NaCl, and to an adequate approximation for estimating the electrical resistivity the total concentrations of dissolved constituents may be taken as equivalent NaCl, although the effect of other constituents can be estimated (Quist & Marshall 1969; Ucok 1979). Fluid inclusion data from intermediate grade metamorphic rocks suggest a salinity and ionic concentration generally similar or somewhat greater than sea-water, although occasionally less. For example, the range for pelite schists and gneisses is about 2–6 per cent, while much higher values are indicated for calcarious rocks (Crawford & Hollister 1986; Roedder 1984; Hollister & Crawford 1981; Fyfe *et al.* 1978; Touret 1977). In the analysis below, an average salinity equal to that of sea-water (about 3 per cent or 0.5 M) has been used although it is recognized that different, particularly higher, values are likely to occur. Fortunately, as we show below, the calculated lower crustal porosity from the electrical resistivity is not very sensitive to the value chosen.

An estimate of pore fluid electrical resistivity at lower crustal conditions for this average concentration can be obtained from the data given by Quist & Marshall (1968), Parkhomenko (1967), and Ucok (1979). Resistivity decreases with increasing temperature to about 300 °C, because of increasing ionic mobility. At higher temperatures, this effect is balanced by decreases in ionic concentration due to density and association effects (Quist & Marshall 1968). At low pressures the resistivity increases with temperatures above about 300°, but at pressures of the lower crust, extrapolation of the available data suggests a slight decrease (Olhoeft 1981). A plot of the salinity effect on resistivity from Quist & Marshall (1968) is given in Fig. 2. The maximum pressure of 4 kbars (0.4 GPa) of this data only represents middle crustal conditions. At lower crustal pressures of 6–10 kbars (0.6–1.0 GPa) extrapolation of the data suggests slightly lower resistivities. In the lower crust at an average temperature of about 500 °C, this salinity has a resistivity of about 0.02–0.04 Ωm depending on the magnitude of the pressure effect. We have used 0.02 in the

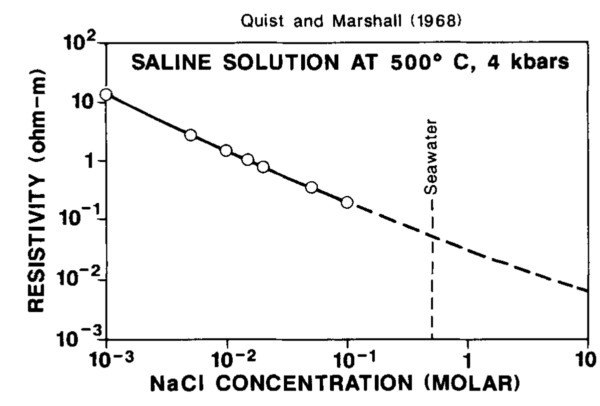


**Figure 2.** Resistivity as a function of NaCl concentrations at 500 °C and 4 kbars (0.4 GPa) pressure, from Quist & Marshall (1968) and Ucok (1979). The resistivity is probably slightly less at lower crustal pressures of 6–12 kbars (0.6–1.2 GPa).

models below. Reasonable limits seem to be 0.05–0.1 Ωm for a concentration of 0.1 M or 0.5 per cent NaCl equivalent, and 0.005–0.01 Ωm for a concentration of 2 M or 10 per cent NaCl. An error of a factor of 2 in pore resistivity gives an error of 10–30 per cent in the estimated porosity depending on the model used. A pore resistivity error of a factor of 10 gives a porosity error of about a factor of 3 (see below).

### Matrix resistivities

Numerous laboratory measurements have been made of dry rocks that might occur in the lower crust (compilations in Kariya & Shankland 1983; Olhoeft 1981, 1980; Haak 1980). These results probably provide reasonable estimates of matrix resistivities although surface conduction and other grain boundary conduction effects may be important, particularly for low porosities and poor interconnection. At surface temperatures the resistivities of most completely dry rocks are very high, over $10^{10}$ Ωm, but the measured values decrease rapidly with increasing temperature. The mean for andesites and basalts is about $10^4$ at 500 °C, decreasing and increasing by factors of 2 and 3 at 600 and 400 °C, respectively (Kariya & Shankland 1983). For gabbros, diabases and diorites the resistivities are slightly higher, $5 \times 10^4$ at 500 °C with a similar temperature dependence. Granites and granodiorites have again slightly higher resistivities, about $10^5$ Ωm at 500 °C. The average results from Kariya & Shankland (1983) are shown in Fig. 3. The actual dry resistivities may be slightly higher, since for many of the measurements very small amounts of water may be present. However, as noted above this error may be balanced by the effective matrix resistivities being lower because of surface conduction effects. A few metamorphic rocks have been suggested to be moderately conductive but this appears to be a consequence of small amounts of free water (Shankland & Ander 1983). Olhoeft (1981) found under carefully controlled dry conditions that a hornblende schist had a resistivity close to that of granite at lower crustal temperatures. In the analysis below a matrix resistivity of $10^4$ Ωm has been used. Values of $10^3$ and $10^5$

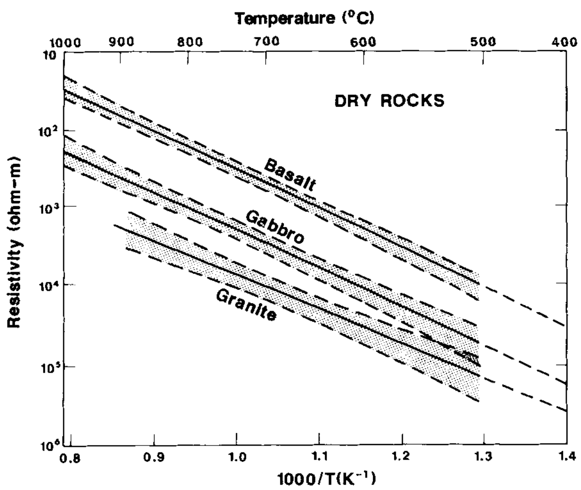


**Figure 3.** Average laboratory resistivities for several types of dry rocks as a function of temperature from Kariya & Shankland (1983).

change the calculated porosities by less than 20 per cent for most models (see below).

## ELECTRICAL RESISTIVITY AS A FUNCTION OF POROSITY

### Theoretical models of pore geometry

The pore geometry, particularly the degree of pore interconnection, is of critical importance in modelling bulk resistivity because the resistivities of the solid and fluid phases differ by six orders of magnitude. This is in contrast to the seismic models where the differences in properties are only factors of 2–4. Reviews of previous theoretical models for the effect of conducting fluid porosity on bulk resistivity have been given by Waff (1974) and Schmeling (1985b, 1986). The limiting cases for an isotropic medium are isolated spherical pores for which the bulk resistivity is close to that of the dry grains (e.g. Waff 1974; Schmeling 1986), and thin completely interconnected films that wet the grain boundaries, for which the resistivity is very low (Fig. 4) (i.e. the Hashin–Shtrikman bounds) (Schmeling 1986). These two extreme models are reasonably excluded by the estimated equilibrium pore shapes and by laboratory data. Thin films require a dihedral angle of 0° and spherical pores, a dihedral angle of 180° (see below). A number of more reasonable intermediate models have been examined.

A network of interconnected tubes (Grant & West 1965; Schmeling 1985a) represent a reasonable maximum degree of pore interconnection (Fig. 4) since grain boundary tubes probably provide the main conduction paths. However, only a small part of the porosity is located in the tubes in the expected equilibrium pore geometry, a fraction that is dependent on porosity and the fluid wetting angle (see below).

A model of ellipsoidal pores (Fig. 5) permits a direct

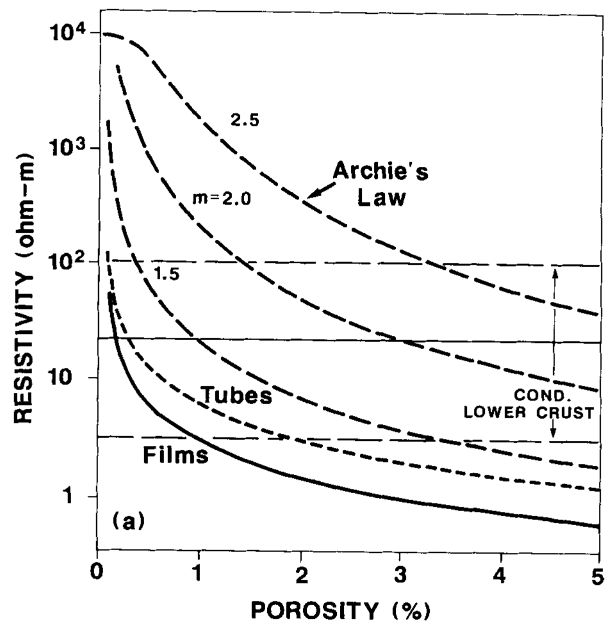


**Figure 4.** Electrical resistivity as a function of porosity for the models of thin films, tubes, and modified Archie's law with different exponents.

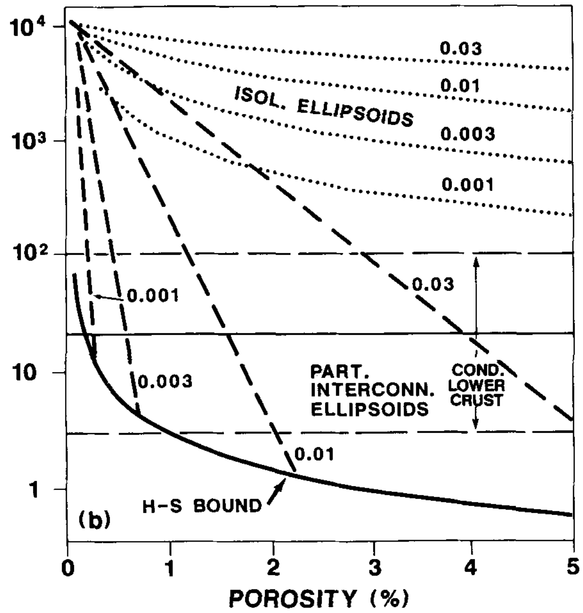


**Figure 5.** Electrical resistivity as a function of porosity for isolated and partially interconnected ellipsoidal pores.

comparison with the theoretical seismic velocity models discussed below, but has limited applicability since the important conduction paths are probably in grain boundary tubes. This model does however provide an estimate of the upper limit of porosity. If there is no interconnection between pores, very high porosities are required to produce the observed layer resistivities (Fig. 5). A modification was developed by Schmeling (1986) that allowed for increasing interconnection with increasing crack or pore density, based on the statistical probability that pores will intersect. The resistivities approach those for thin films at large porosities, and those for isolated pores at low porosities. For the minimum aspect ratio of about 0.03 suggested from the seismic data, the porosity estimate is about 4 per cent. Larger aspect ratios require very high porosity since the statistical model predicts little interconnections.

Archie's law relating bulk resistivity to the inverse of a power of porosity is widely used in sedimentary rocks, and it appears to have general applicability for many crystalline rocks at high confining pressures (e.g. Brace, Orange & Madden 1965; Madden 1976; Hermance 1979). For crystalline rocks measured in the laboratory under high confining pressures, the exponent is usually close to 2. Madden (1976) reviewed the theoretical requirements for this behaviour in terms of random networks and emphasized that Archie's law with exponent 2 represents rather inefficient conduction. Hermance (1979) suggested the applicability of the model for fluid distributions intermediate between thoroughly interconnected and located in isolated pockets (also Schmeling 1986). In his modified form, the bulk resistivity asymptotically approaches that of the solid phase at very low porosities. Higher and lower degrees of interconnection can be modelled by smaller and larger exponents. The results for several values are given in Fig. 4. For an exponent of 2 and a layer resistivity of 20–30 Ωm, the porosity estimate is about 3 per cent. The exponent of 1.5 representing better pore interconnection is suggested by some high temperature and pressure sample data. It gives a porosity of about 1 per cent while a less likely high exponent of 2.5 gives 6–8 per cent.

**Equilibrium pore geometry**

A more direct constraint on *in situ* pore shapes in the lower crust is obtained if textural equilibrium conditions are assumed between the fluid and host rock. At temperatures in the lower crust of above 350–400 °C such equilibrium pore shapes should be reached in geologically short times (e.g. Watson & Brenan 1987; Meissner & Kusznir 1987). In contrast, porosity may be dominated by cracks and non-equilibrium pore shapes in the cool upper crust. Theoretical and experimental results have shown that under equilibrium conditions the shape of the fluid-filled pores is largely determined by just two parameters, the porosity and the dihedral or grain wetting angle (e.g. Waff & Bulau 1979). At angles less than the critical value of 60°, the fluid pores are interconnected along grain edges at all porosities. At angles greater than 60° the fluid forms isolated pores located at the grain corners for small porosities, only becoming interconnected at a few per cent porosity. This latter behaviour was seen experimentally in studies of calcite hot pressing by Bernabe, Evans & Brace (1982) who found that pore interconnection and permeability virtually disappeared below a particular porosity, about 4 per cent in that material. In real rocks a range of dihedral angles is expected (Jurewicz & Jurewicz 1986) so that the permeability transition will be over a broader porosity range.

The dihedral angle for water at lower crustal conditions depends upon the surface free energies of the fluid and the host rock. Watson & Brenan (1987) examined dihedral angles produced experimentally with aqueous fluids. For a few per cent porosity of pure water in a variety of rock types, they found average angles close to the critical value of 60°. The addition of $CO_2$ generally increased the angle. The effect of adding salts was variable, decreasing the angle for quartz and having no effect on olivine; unfortunately the effect of NaCl was not tested for other more common crustal minerals. Since at least small amounts of $CO_2$ are undoubtedly present (e.g. Abbott & Lyle 1984; Fyfe 1986), the available data suggest average dihedral angles for lower crustal rocks that are close to or just above 60°.

Theoretical calculations of pore shape are practical using equilibrium pore theory if a simple grain geometry for the host matrix is assumed and a number of other simplifying assumptions are made (i.e. identical grain size, shape and composition, no wetting angle dependence on crystal face, only hydrostatic stress). For example, von Bargen & Waff (1986) determined the equilibrium pore shapes for fluid within a solid matrix composed of tetrakaidecahedral grains of equal size for a variety of dihedral angles and porosities. They did not calculate the resistivity of their equilibrium pore geometries, but they did plot the minimum channel cross-sectional area of the grain boundary tubes as a function of porosity and dihedral angle. This information is sufficient for an approximate calculation of resistivity, since the bulk resistivity is largely determined by this minimum area.

Consider a lattice of tubes along the grain edges of a

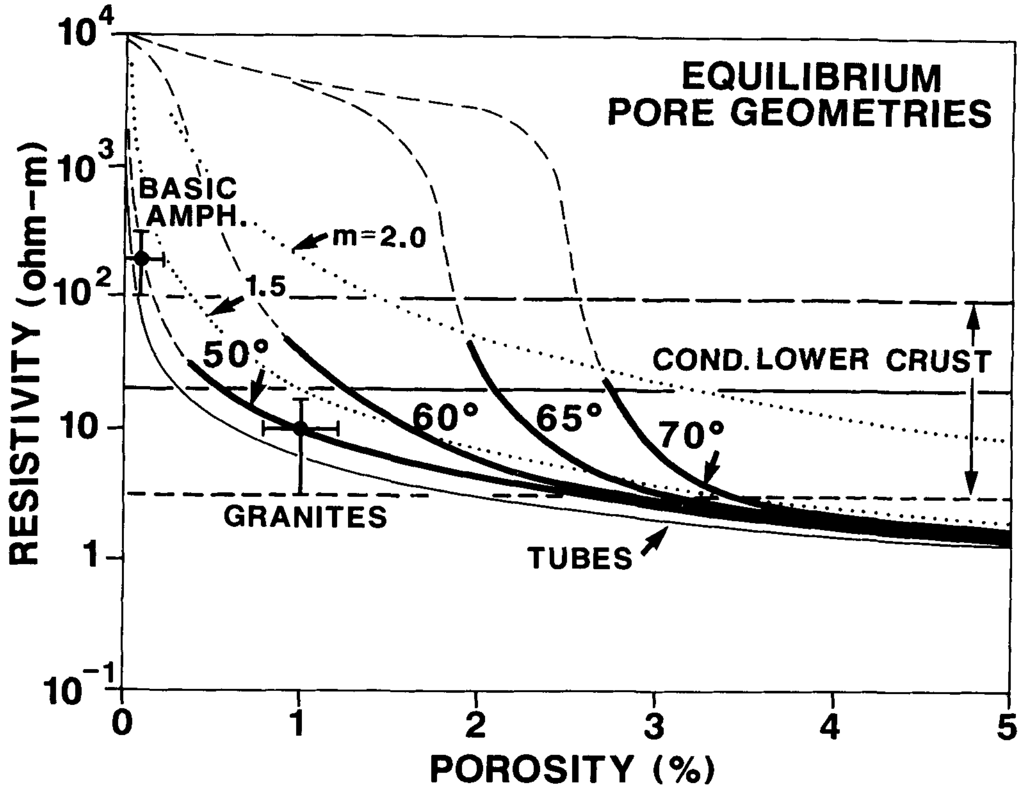


**Figure 6.** Electrical resistivity as a function of porosity for equilibrium pore geometries, with several values of dihedral angle. The curves for Archie's law with exponents 1.5 and 2 are shown as dotted lines for comparison. Several laboratory measurements are also shown.

tetrakaidecahedron matrix. Assume that the radius of the tubes, $r$, is much smaller than the distance between grain faces, $a$ ($r \ll a$), and that the conductivity of the fluid $c_f$, is much greater than the matrix. For a direction perpendicular to the square faces, one-third of the tubes will be normal to this direction and not contribute to the conduction, while two-thirds will be at an angle of 45°. The approximate conductivity (1/resistivity) of the tube system is then given by: $c = 1/3\ (\sqrt{2}\,Pc_f)$, where $P$ is the porosity. Note that the conductivity is higher than the result for a cubic lattice by a factor of $\sqrt{2}$ (Grant & West 1965).

This result assumes that the cross-sectional area of the tubes is constant along their length. However, in the equilibrium pore shapes, the porosity is concentrated near the grain corners and pinches to a minimum in the middle of the grain edge. As a first-order approximation to these shapes, we calculate the conductivity of a truncated cone with shorter radius, $r_1$, determined by the von Bargen & Waff (1986) results for the minimum tube cross-sectional radius, and larger radius, $r_2$, determined so as to give the correct porosity. The conductivity of this cone is a factor $E = r_1/r_2$ less than the conductivity of a cylinder of radius $r_2$, which gives us the correction factor needed to use the above equation. Fig. 6 shows the results of this calculation using the von Bargen & Waff data and assuming a fluid resistivity of 0.02 Ωm. The pinch-off of the porosity is clearly evident. For dihedral angles of 60° or greater there is an abrupt transition from connected to unconnected as the porosity decreases. The average conductivity lower crust resistivity occurs at the transition. A bulk resistivity of 10–30 Ωm is predicted at porosities of 0.5–3 per cent for dihedral angles between 50° and 70°.

### Laboratory data

Laboratory data have the potential of providing direct calibration of pore geometry models. However, very few measurements have been made at simulated lower crustal conditions on saline fluid saturated rock samples. For extrapolation from lower pressures, the relevant laboratory measurements are probably those made with substantial confining or effective pressures, i.e. low pore pressure. High confining pressure is necessary to close the extensive microcracks characteristic of surface samples that will probably be annealed at the temperatures of the lower crust. However, this pressure may also reduce the grain boundary pore interconnection compared with *in situ* conditions in the lower crust where the effective confining pressure is inferred to be near zero.

Oholeft (1981) and Lebedev & Khitarov (1964) measured the resistivity of granite at lower crustal conditions. While granites are unlikely to be important in the lower crust, the porosity and resistivity are close to that we estimate for conductive lower crust. The resistivity is about 10 Ωm for a 0.4 M (near sea-water) salinity and low pressure (Fig. 6). The porosity is not well determined but probably is about 1 per cent. This resistivity is too low for Archie's law with exponent 2, but fits with an exponent of 1.5 indicating a relatively high degree of pore interconnection. It also fits an equilibrium pore dihedral angle of about 50°. However, the presence of quartz is probably important because the laboratory data show that this mineral has an unusually low dihedral angle for a saline fluid. There are probably higher average dihedral angles in the lower crust where quartz is undoubtedly rare, and thus higher estimated porosity.

Measurements on three samples of amphibolite and basic gneiss which are particularly relevant in composition were reported by Lee, Vine & Ross (1983). However, the maximum temperature was only about 270 °C and pore pressures were not controlled, so equilibrium pore geometry is not expected. They gave resistivities extrapolated to lower crustal conditions and a near sea-water salinity (0.5 M) fluid of 100–200 Ωm. The measured porosities were only roughly determined at about 0.1 per cent, but the porosity in such samples is very difficult to measure and the real value may be somewhat higher. These results are substantially more conductive than the standard Archie's law with exponent 2, and again fit an exponent of about 1.5 or a dihedral angle of about 50° (Fig. 6). This indicates a relatively high degree of pore interconnection. Evans (1980) measured a greater variety of mafic rocks and found a similar result.

Thus, while laboratory measurements may not simulate *in situ* conditions accurately, and laboratory measurement durations are probably too short for equilibrium pore shapes to be ensured, these results suggest more pore interconnection than Archie's law with exponent 2 and give a better fit with an exponent of about 1.5. They also fit an equilibrium pore geometry dihedral angle of about 50°. The relatively good pore interconnection inferred for the granites probably reflects the presence of quartz, and for the basic amphibolites, that the laboratory temperatures were too low to anneal the extensive cracks usually present in surface samples.

### Estimated porosities in conductive lower crust

The analysis above suggests that Archie's law with exponent 2 provides a reasonable upper limit, and an exponent of 1.5 a lower limit to the porosity in conductive lower crust, i.e. about 3 and 1 per cent, respectively, for 20–30 Ωm average layer resistivity. A porosity range of 0.5–3 per cent is indicated by the equilibrium pore model, with values over 1.0 being preferred. The best estimate is about 2 per cent. The effect of different fluid matrix resistivity on the Archie's law model with exponent 1.5 is illustrated in Fig. 7. The effect of changing the matrix resistivity by a factor of 10 is negligible for most reasonable models.

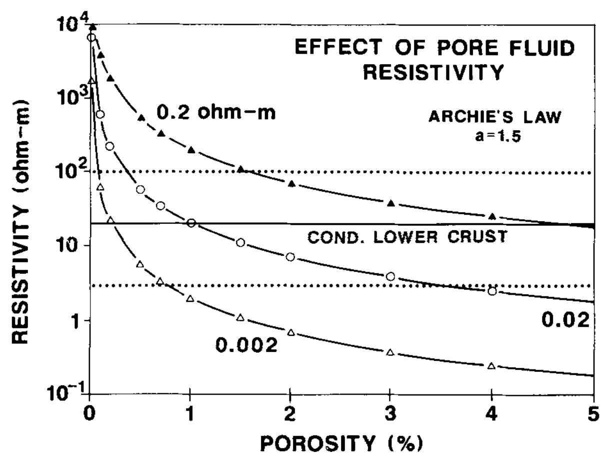


**Figure 7.** Electrical resistivity as a function of porosity for Archie's law with exponent 1.5, showing the effect of different fluid resistivities.

## MAINTENANCE OF POROSITY IN THE LOWER CRUST

### Maximum permeability required

A critical question in the model for lower crustal porosity is how the fluid is kept from moving upward. Although an impermeable layer in the mid-crust can contain rising fluid as discussed below, a mechanism is required to maintain it distributed in the underlying region. The density difference between the fluid and surrounding matrix provides a difference in gravitational potential that should tend to concentrate the fluid at the top of the layer, probably to eventually break through, unless the effective permeability is sufficiently low. This is a similar problem to that of keeping a few per cent of partial melt in a mantle seismic low velocity layer, as discussed by Waff (1980). A critical element of the model is how to reconcile the degree of pore interconnection required to explain the electrical resistivity data and the requirement that the permeability be sufficiently low for the fluid to remain in the layer. We discuss two approaches: the first is to estimate the bulk permeability limit for fluid stability and then to see if this value is consistent with the electrical resistivity data. The second, in a later section, is to examine models where vertical movement of the fluid is restricted by the small scale layering required by the seismic data, while the electrical resistivity in the horizontal direction remains low.

One estimate of the minimum permeability for assumed uniform properties, can be obtained using the theory for compaction of a layer of porous media, developed to model the expulsion of partial melt (e.g. McKenzie 1985, 1987; Scott & Stevenson 1986). The high permeability solution to this problem is controlled by the effective viscosity, i.e. the rate of deformation of the solid grains, and may not be relevant here. With water present, the grains will probably readjust their shape rapidly by recrystallization (e.g. Watson & Brenan 1987; Etheridge *et al.* 1984) rather than by viscous or ductile deformation. The low permeability solution depends on the fluid viscosity and the permeability of the matrix to fluid percolation. The fluid viscosity may be taken as that of water at lower crustal conditions, i.e. about $10^{-4}$ Pa s$^{-1}$ (compilations in Walther & Orville 1982; Walder & Nur 1984). The compaction time to $1/e$ of original porosity for a 10 km thick layer, and for a range of permeabilities is shown in Fig. 8. As an example, if the layer has 2–3 per cent porosity, a compaction time of 100 Ma requires that the permeability be less than about $10^{-21}$ m$^2$. If there is a source of lower crustal water on a shorter time scale, say a few Ma, a permeability of about $10^{-19}$–$10^{-20}$ m$^2$ is required.

If the porosity must be restricted to the 100 m thick layers suggested by the seismic lamellae, about a factor of 100 lower permeability is required (Fig. 8). These low permeabilities are at the extreme of those found in

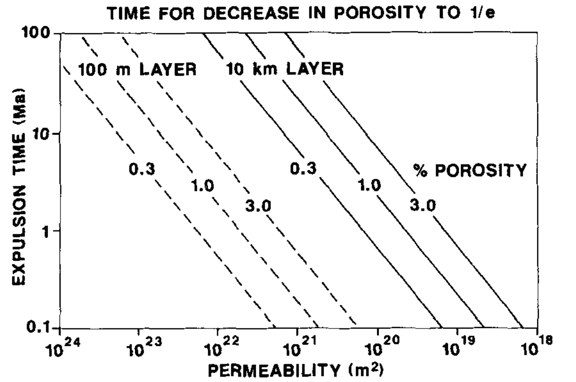


**Figure 8.** *Estimated time for water to move out from a lower* crustal layer 10 km thick layer and from 100 m thick lamellae, for a range of permeabilities and initial porosities.

laboratory measurements of metamorphic and igneous rocks (e.g. about $10^{-16}$–$10^{-21}$ m$^2$; Brace 1980; Bredehoeft & Hanshaw 1968; Trimmer *et al.* 1980), but these laboratory samples are unlikely to have had equilibrium pore configurations and as discussed in the next section, lower permeabilities can readily occur under the special conditions of the lower crust.

## Permeability and resistivity for equilibrium pores

The relation between resistivity and permeability can best be seen by considering equilibrium pore geometry models. For a grain size of 1 mm appropriate for many metamorphic rocks and dihedral angles just over 60°, the equilibrium pore model predicts very high permeabilities of about $10^{-13}$ m$^3$ for over 2–3 per cent porosity. Below this porosity the permeability rapidly decreases to very low values (e.g. less than the required $10^{-21}$ m$^2$) as porosity decreases (von Bargen & Waff 1986) (Fig. 9). Pore fluid is readily lost until this porosity is reached. This critical porosity is in agreement with the estimated average values for the reflective and conductive lower crust. Such a transition between interconnected and isolated pores is consistent with the observation of Etheridge *et al.* (1984), White & White (1981) and Hall (1986a) that the pore spaces in surface

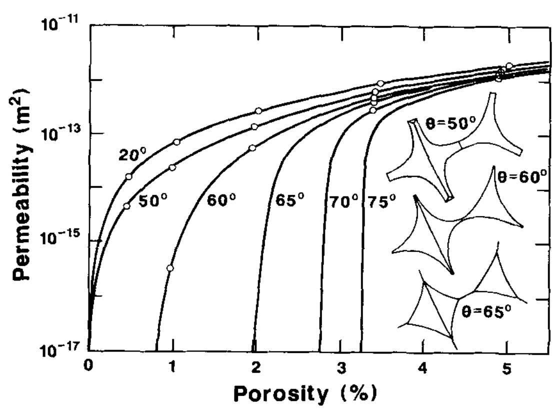


**Figure 9.** Permeability as a function of porosity for a range of dihedral angles from von Bargen & Waff (1986). The grain size assumed is 1 mm. The pore shapes illustrated are for 1 per cent porosity; note the complete pinch-off of pore interconnection at dihedral angles above 60°.

metamorphic rock samples are primarily unconnected bubbles and cavities. The final state of metamorphic rocks at depth after peak metamorphism and before uplift and exposure is largely unconnected. At the same time it predictes the very high permeabilities of at least $10^{-15}$ m$^2$ that are required during active metamorphism (e.g. Etheridge *et al.* 1984), if as expected the porosity is higher at that time. Thus, if the dihedral angle in the lower crust is over 60°, water can readily be held for geological lengths of time.

The difficulty comes in reconciling the low permeabilities with the pore interconnection required for the observed low electrical resistivities. For example, Fig. 6 shows that the lower crustal resistivity can be explained by equilibrium pore conditions with a dihedral angle of 60° and a porosity of 1.5 per cent. However, von Bargen & Waff (1986) show that for a typical metamorphic rock grain size of 1 mm, the corresponding permeability is $10^{-14}$, at least 5 orders of magnitude greater than that required to hold the water in the lower crust assuming a simple homogeneous layer (see above). The two simple solutions to this problem are to make the grain size smaller, or to make the fluid resistivity lower. The other solution is to appeal to some form of loose bonding that physically holds the fluid without reducing the conduction. However, we are not aware of any such mechanism.

The permeability of a channel varies as the fourth power of the channel radius, while the resistivity varies as the second power. Thus, in the above example the permeability could be reduced to $10^{-21}$ m$^2$ if the grain size were reduced to 0.003 mm, while the resistivity would be unchanged. However, such small grain sizes seem unrealistic for the lower crust. Very fine grained minerals such as quartz deposited along the primary grain boundaries might have such small grain sizes but they must make up only a small part of the cross-section.

A second approach is to keep the grain size fixed but decrease the fluid resistivity, so that very small pore interconnection, close to pinch-off could still provide the required low resistivity. At dihedral angles above 60° the permeability decrease will be dominated by the channel radius at the pinch off point. von Bargen & Waff (1986) do not provide results at such low permeabilities but an approximate scaling can be used to extrapolate their data. For a truncated cone model with minimum cross-sectional area $A_1$, and maximum cross-sectional area $A_2$ ($A_1 \ll A_2$), the permeability will vary as $A_1^{3/2}$ while the conductivity will vary as $A_1^{1/2}$. For example, at a dihedral angle of 60°, a porosity of 1 per cent, and a grain size of 1 mm, von Bargen & Waff calculate a permeability of $3 \times 10^{-16}$ m$^2$. This can be reduced to $10^{-21}$ m$^2$ if the minimum cross-sectional area is reduced by a factor of 4500. Our calculated bulk resistivity then becomes 2300 Ωm for the preferred fluid resistivity of 0.02 Ωm. Decreasing the fluid resistivity by a factor of 230 to about 0.0001 Ωm gives the required bulk resistivity of about 10–30 Ωm. However, such low resistivities cannot be achieved in a saline pore fluid even at very high salinities, and some other more conductive fluid would be required.

Thus, it is difficult to explain the required low permeability in the lower crust together with the observed low resistivity using simple homogeneous layer models of equilibrium pore shape and reasonable parameter values.

For this reason Warner (1988) argues that the low resistivity must result from thin conductive fluid layers shallower in the crust. Although as noted above, electrical conductivity inversions have poor depth resolution, we feel that on average they are of sufficient accuracy that confining the conductive zone to the upper crust is not an acceptable option. Also, the porosity required for a thin very conductive layer would be very high. The solution probably lies with more realistic and thus complex pore models.

The mechanisms by which the effective vertical permeability is reduced may be related to the processes that produce the strong seismic layering discussed below. The abrupt transition from interconnected to pinched-off pores predicted for dihedral angles above 60° allows pronounced changes in permeability with small changes in any one of a number of parameters. These include rock compositional layering and non-equilibrium stresses. Waff (1980) has described a model for the stabilization of partial melt in the mantle in layers based on the effect of the gravitational field on pore geometries. Stevenson (1986) has argued that this process will work only if there is also thin-scale compositional layering. However, such layering is very common in metamorphic rocks so this may be a viable mechanism.

Scott & Stevenson (1986) have shown that for dihedral angles above 60° pore fluid will tend to concentrate in larger volumes, in contrast with smaller angles for which inhomogeneities will tend to smooth out. Such larger fluid volumes could be horizontally aligned by lower crustal shear processes.

To resolve this problem, there is clearly a need for combined electrical resistivity and permeability measurements under lower crustal conditions on the same samples of probable lower crustal rocks, if possible under varying stress fields.

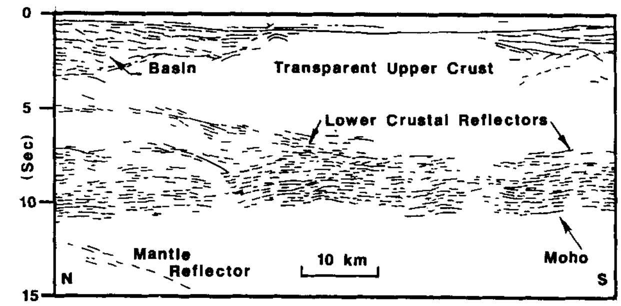


**Figure 10.** Example of a multichannel seismic reflection line drawing section showing strong lower crustal reflectivity (BIRPS SWAT profile from Matthews 1986).

## LOWER CRUSTAL REFLECTORS

### Nature of reflectors

Strong reflections distributed throughout the lower crust and a generally transparent upper crust have been recognized in a few areas for many years (Dix 1965; Dohr & Fuchs 1967; Meissner 1967a; Clowes, Kanasewich & Cumming 1968; Fuchs 1969). It is now established as the most common multichannel reflection response, evident to varying degrees in about one half of the deep crustal surveys carried out, especially in Phanerozoic areas (e.g. the surveys of BIRPS; DEKORP; ECORS; COCORP; ACORP; LITHOPROBE etc., see symposium volume edited by Matthews & Smith 1987). Concentration of reflectors are sometimes found in the mid-crust (e.g. Conrad discontinuity) and near the base of the crust, or they may be uniformly distributed (e.g. Wever, Trappe & Meissner 1987). They are typically found below 5–7 s or 15–21 km depth. There are many striking examples; particularly clear cases are in many of the BIRPS lines (e.g. Matthews 1986), in the Black Forest area surveyed by the German DEKORP (e.g. Luschen *et al.* 1987; Wenzel, Sandmeier & Walde 1987), and in the Basin and Range of the U.S.A. (e.g. McCarthy & Thompson 1988). Fig. 10 shows a BIRPS example from SW of the British Isles. Other reflection patterns do occur (see example from BIRPS data by McGeary 1987), but only infrequently is the upper crust more generally reflective than the lower. The uppermost mantle is usually, with a few very notable exceptions, without reflections. A number of tentative associations of strong lower crustal reflectivity with tectonic history or crustal stress regime have been suggested. They are discussed below.

The numerous short subhorizontal reflectors give a character to the lower crustal seismic sections that is unlike those observed in the upper crust or in the exhaustive surveys of sedimentary basins by the petroleum industry, which suggests a different origin. The lower crustal reflectors are usually subhorizontal although occasionally dipping as much as 20°. They may become flatter towards the bottom of the crust, and where crossing lines are available the reflector lengths appear to be longer in one direction, for example parallel to strike in extensional regimes (e.g. Reston 1987; Wever *et al.* 1987). They are sometimes associated with relatively high seismic refraction (i.e. horizontal) velocities (e.g. Mooney & Brocher 1987). There also may be a decrease in reflector length with geological age (Wever *et al.* 1987). They have rather short coherent lengths; apparent lengths range from 1 to 10 km and occasionally to 35 km, but generally about 2–4 km (e.g. Matthews & Cheadle 1986; Blundell & Raynaud 1986; Reston 1987; Sandmeier, Walde & Wenzel 1987; Meissner 1986).

### Inferred seismic impedance structure

Care must be taken in inferring seismic velocity and density structure from seismic sections, both because of the inherent limitations of imaging deep structures and the effects of different processing. The commonly observed reflector image lengths of 2–4 km are similar to and sometimes shorter than the Fresnel zone at lower crustal depths (3.5 km at 20 km depth and 6 km $s^{-1}$), and as pointed out by Hurich & Smithson (1987) and Reston (1987) shorter reflector horizons should appear with lengths comparable to the Fresnel zone, and with reduced amplitude. They also point out that closely spaced reflectors smaller than the Fresnel zone give spatial interference between small body diffractions and can have complex images. Thus, the reflective lower crust may consist of numerous small reflectors and be more complex than suggested by the seismic sections.

Constructive interference between wavelets reflected from thin rock layer impedance contrasts is also probable (Fuchs 1969; Blundell & Raynaud 1986; Hurich & Smithson 1987; Christensen & Szymanski 1988). Such constructive interference from thinly layered sequences is well recognized from experience with sedimentary sections where well logs provide control (e.g. O'Doherty & Anstey 1971; Hughes & Kennett 1983; Hurich & Smithson 1987). Blundell & Raynaud (1986) showed that layers as thin as 1/32 of a wavelength $L$, i.e. only a few metres thick, could produce the observed reflections. Very thin layers, however, require large impedance contrasts, and increasing the number of layers increasingly limits the frequency band of the resulting wavetrain (Fuchs 1969; Hurich & Smithson 1987). High amplitudes can be generated from multiple thin layers spaced at $L/2$ or $L/4$, etc. Layers spaced at $L/4$ can as much as double the reflection amplitude (e.g. Meissner 1967b).

As an alternative to multiple layers, Blundell & Reynaud (1986) and Klemperer *et al.* (1986) showed that some of the characteristics of deep crustal reflectors can result from a single 2D undulating surface (e.g. 'egg carton' surface) that generates off-section reflections. Matthews (1986) argues that even with such an effect, at least four and probably more surfaces are required to match the observed records. This process also cannot give the observed abrupt reflection cut-off at the Moho; the upper and lower boundaries of the reflectivity would be diffuse. Thus, it is reasonable to conclude that the seismic sections provide at least a rough representation of the lower crust impedance structure and that multiple lamellae are required.

Synthetic reflection images that approximate those observed have been given by Wenzel *et al.* (1987) and Sandmeier *et al.* (1987) using randomly spaced horizontal layers about 120 m thick and 400–1200 m long and by Reston (1987) using randomly spaced horizontal lens-shaped bodies ranging from 100 by 500 m to 500 by 4000 m. Multiple lamellae of approximately equal thickness of high and low velocity material are suggested. The velocity variation with depth may be very irregular but it must contain a strong component with the order of 100 m vertical variation. For example Wenzel *et al.* (1987) found that their models for very strong reflectors required alternating approximately 100 m layers or lamellae with the low velocity elements at 5.6–6.1 km $s^{-1}$ and the high velocity elements at 6.3–7.3 km $s^{-1}$. Changing the lamellae thickness to 50 or 200 m substantially degraded the fit to the observed data.

The reflection coefficients required to give the observed reflections are difficult to estimate accurately, but appear to be commonly a few per cent (e.g. Klemperer *et al.* 1987; Matthews 1986). For the highest amplitude reflectors, modelling suggests reflection coefficients of at least 5 and probably 10 per cent (e.g. Wenzel *et al.* 1987; Sandmeier *et al.* 1987; Reston 1987; Jones 1983). Warner (1989) has estimated even higher reflection coefficients. Even with amplification from optimum lamellae spacing, these bright reflectors require at least 10 per cent impedance contrasts. However, for the typical more uniform reflective lower crust with multiple reflectors, the reflection coefficients cannot be much more than 10 per cent or the deeper reflectors would appear reduced in amplitude. For example, in the model of Wenzel *et al.* (1987) and Luschen *et al.* (1987) the impedance contrast of the thin lamellae was increased by a factor of 3 from top to bottom of the 10 km thick lower crust. Even so, the decrease in amplitude with depth in the modelled reflectors appears greater than that in the data. We will assume that typical lower crustal reflectors require impedance contrasts between lamellae of 5–10 per cent, i.e. reflection coefficients of 2.5 to 10 per cent depending on the lamellae spacing, although we recognize that impedance contrasts as large as 15 to 20 per cent may be required for individual bright reflectors.

## GEOLOGICAL ENVIRONMENT OF LOWER CRUSTAL REFLECTORS

The primary association with lower crustal reflectivity is geological age, the time of the last thermal–tectonic events. Precambrian shields have few lower crustal reflections, particularly the Archean cores (e.g. Gibbs 1986; Wever *et al.* 1987; Allmendinger *et al.* 1987; Green *et al.* 1986). The shield margins are often more reflective (e.g. Nelson *et al.* 1987). The foreland and mountain belts of Phanerozoic orogens or collision zones have a complex reflection character with many reflectors usually dipping toward overthrust mountain belts (Allmendinger *et al.* 1987). The lower crust in these areas is not systematically reflective. Younger regions, the hinterlands of compressional mountain belts (Zone 4 of Allmendinger *et al.* 1987), often areas of multiple accreted terranes, commonly have numerous reflections in the lower crust. These latter areas have often been subject to extension, and extension may be an important common factor, e.g. off eastern North America (Phinney 1986), most BIRPS profiles in the area of the British Isles (Matthews & Cheadle 1986), and the presently active Basin and Range area (e.g. McCarthy & Thompson 1988). However, profiling towards the ocean on rifted margins has not shown the expected increase in lower crustal reflectivity as the degree of extension increases (e.g. Cheadle *et al.* 1987; Peddy & Hobbs 1987; Hobbs *et al.* 1987; de Voggd & Keen 1987), although reflectivity continues to the continent–ocean transition. In addition, the relative importance of horizontal extension and contraction is difficult to determine because most of these areas have been subject to both types of deformation. Probably of more importance is that strong lower crustal reflectors generally occur where the last tectonic event was a thermal event, often associated with extension and rifting. The role of temperature in generating and preserving lower crustal fluids is discussed below.

## ORIGIN OF LOWER CRUSTAL REFLECTORS

This paper focuses on the model of fluids producing lower crustal reflectors, but a number of other explanations have been proposed. It is not the intention of this paper to show that the other processes cannot produce lower crustal reflections, but to argue that fluids are an important mechanism to be considered. Undoubtedly all of the other processes do produce crustal reflectors, but none provides an entirely satisfactory explanation for the widespread characteristic reflective zones. Some of the alternative explanations are (see reviews in Wever *et al.* 1987; Matthews 1986):

(1) Pervasive lower crustal mafic (or ultramafic) sills, perhaps from underplating by hotspot volcanism (e.g. Meissner 1973; McKenzie 1984; Finlayson & Mathur 1984; Cheadle *et al.* 1987; Serpa & de Voogd 1987). Mafic surface volcanism associated with hotspots is widespread and lower crustal mafic underplating along rifted continental margins near hotspots is inferred from recent seismic refraction data (White *et al.* 1987).

(2) Layer cumulates such as anorthosites, gabbros, etc., in large igneous bodies (e.g. Klemperer *et al.* 1985; Meissner 1986).

(3) Horizontal foliation in gneisses, mylonites, etc., perhaps generated by horizontal shear that draws pre-existing structures into parallel horizontal zones (e.g. Phinney & Jurdy 1979; Hurich *et al.* 1985; Fountain, Hurich & Smithson 1984; Bois *et al.* 1987) or by boudinage and shear pods (Reston 1987). For example, Reif & Robinson (1981) described bright near-surface reflections that on drilling proved to be amphibolite-grade gneisses.

(4) Overthrust metasedimentary and metavolcanic layering (e.g. Cook 1986; Ando *et al.* 1984; Klemperer *et al.* 1985). Continental collision zones may produce extensive overthrusting, such as postulated for the eastern US Appalachians. Of course fluids trapped by this mechanism rather than composition differences may be the source of the reflections.

Related to these explanations are computer simulations, where exposed former deep crustal sections such as the Ivrea zone in Italy and the Kapuskasing zone in Ontario have been used to generate synthetic seismograms that have some similarity to those commonly observed from the lower crust (Hale & Thompson 1982; Hurich & Smithson 1987; Fountain, McDonagh & Gorman 1987). However, with some notable exceptions, where data are available the exposed sections are not generally reflective at present.

Some of the difficulties with these explanations are:

(1) While mafic or ultramafic sills and horizontally foliated structures do occur in exposed formerly lower crustal rocks, the required extensive layering is uncommon. It is also unclear why underplated mafic rocks should have the required layered impedance contrasts.

(2) They do not explain the surprisingly consistent pattern of reflectors in many different areas.

(3) The velocities that appear to be required for the lower velocity component of the layering are too low for rocks (with low porosity) thought to be commonly present in the lower crust, particularly under deep crustal conditions. Only high silica rocks containing quartz commonly have such velocities, and substantial quartz predicts a lower Poisson's ratio than that observed (e.g. over 0.25; Luschen *et al.* 1987; 0.25, Hall 1986a).

(4) These explanations do not explain why the reflector zones are usually restricted to the lower crust when formerly lower crustal rocks now often occur at or near the surface; neither do they explain the temperature of over about 400 °C that appears required for pervasive crustal reflectors.

(5) High electrical conductivity in the lower crust is not generated by these models and an independent explanation must be found.

The explanation of fluids in the lower crust has been notably proposed by D. H. Matthews (e.g. Matthews 1986; Matthews & Cheadle, 1986) (also Hall 1986a), but it has not received much support, mainly because of the perceived difficulty of maintaining substantial porosity in the lower crust for geologically long periods of time. This latter problem is discussed above. It is important to note that even if fluids are the primary cause of lower crustal reflectors, their detailed distribution and pattern may be structurally controlled. We also note that it appears to be difficult to model the brightest reflectors with any of the proposed mechanisms including fluid porosity.

## SEISMIC IMPEDANCE CONTRASTS FROM POROSITY IN THE LOWER CRUST

### Porosity models

An interpretation of the seismic evidence for fluid in the lower crust requires an estimate of the porosity and pore structure necessary to produce the observed impedance contrasts. The effect of porosity on elastic properties can be estimated from theoretical models and from laboratory studies on samples of the rock types that might be present. In the discussion below we assume that the fluid is located mainly in grain boundary porosity. However, it is possible that the fluid is contained in larger subhorizontal bodies (e.g. Fyfe *et al.* 1978; Fournier 1988).

The importance of pore structure is seen from theoretical models (reviews by Watt, Davies & O'Connell 1976; Schmeling 1985a,b; Shearer 1988). The maximum effect is for 3D thin films, i.e. complete wetting of grain boundaries, for which the shear velocity approaches zero. The minimum effect is for spherical pores. A wide range of models lie between these limits which approximates the Hashin–Shtrikman bounds (Hashin & Shtrikman 1963).

The ellipsoidal or penny-shaped pore model which has been well studied is a particularly useful approximation. The shape is described by a single parameter, the aspect ratio giving the ratio of the minor to the two major axes. It can be treated theoretically and the behaviour of a wide range of more realistic pore shapes can be approximated by ellipsoids with an equivalent or 'effective' aspect ratio. For example, the results for ellipsoids are not greatly different from those for 'penny'-shaped pores that taper to a thin edge (Mavko & Nur 1978) or triaxial ellipsoids with three different axes, as long as one axis is much smaller than the other two (O'Connell & Budiansky 1974; Schmeling 1985a).

The theoretical models for ellipsoidal pores are strictly valid only for low pore or 'crack' densities for which there is limited pore interaction. Self-consistent theories are believed to be more accurate at high crack densities than other theories, but this has not been proved rigorously. The results for the different theories are similar at low crack densities (see Shearer 1988); the analysis below uses the self-consistent theory as given by Schmeling (1985a). The laboratory results given below indicate that the theory is applicable to real pore structures and is probably valid to quite high crack densities. Most theoretical treatments of cracks assume that they have random orientation. The theory of Hudson (1980, 1981) gives results for aligned ellipsoidal pores, in which case the material properties are anisotropic.

For comparison we also give results for tubular pores and the simple Wyllie time-average equation. The effect of tubes on velocity has been given by Mavko (1980) for three cross-sections, circular, rounded corner triangular, and thin corner triangular (his $k = 0$, 1, $\infty$). The thin corner triangular cross-section ($k = 0$) most closely approximates the equilibrium pore cross-section except at high dihedral angles. The Wyllie time-average equation ($1/V = P/V_f + (1-P)/V_s$) where $V_f$ and $V_s$ are the velocities of the fluid and solid respectively and $P$ is the porosity, is widely used for simple downhole log interpretations in sedimentary sections. It may be a poor model for crystalline rocks, but is given here for comparison.

For modelling purposes we assume that the fluid properties are those of water at lower crustal conditions, about 500 °C and 8 kbars (20–25 km depth), i.e. approximately $V_p = 2.2\ \mathrm{km\,s^{-1}}$, $V_s = 0$, and density $0.9\ \mathrm{g\,cm^{-3}}$ (Burnham, Holloway & Davis 1969; Hall, 1986a; Warner 1989). At lower temperatures the velocity is somewhat higher, i.e. $2.5\ \mathrm{km\,s^{-1}}$ at 300 °C, and at high salinities the density will be higher. The zero porosity elastic parameters of the rock matrix are assumed to be $V_p = 7.0\ \mathrm{km\,s^{-1}}$, $V_s = 4.04\ \mathrm{km\,s^{-1}}$ (Poisson's ratio of 0.25) and density $2.9\ \mathrm{g\,cm^{-3}}$, but the fractional changes in impedance and Poisson's ratio with porosity are not sensitive to the exact values chosen.

The seismic impedance dependence on porosity for randomly orientated water-filled ellipsoids with a range of aspect ratios is shown in Fig. 11, and for tubes and the Wyllie time-average relation in Fig. 12. These plots also approximate the effect on velocity since the density change compared with velocity is generally small. For the ellipsoids the solid lines extend to pore densities of 0.1 for which the theory is probably a reasonable approximation. The dashed lines extend to densities of 0.5. At high pore densities such as required for a greater than 10 per cent impedance

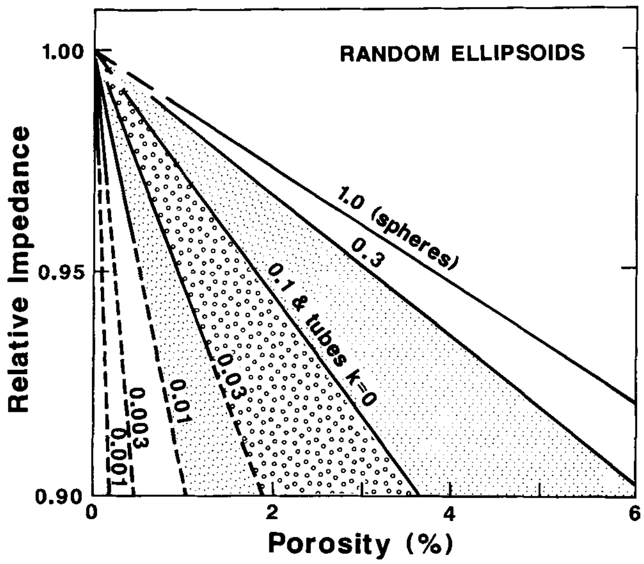


**Figure 11.** Seismic impedance decrease with porosity for the model of randomly orientated ellipsoids with a range of aspect ratios (after Schmeling 1985a). The lines are solid for crack or pore densities less than 0.5.

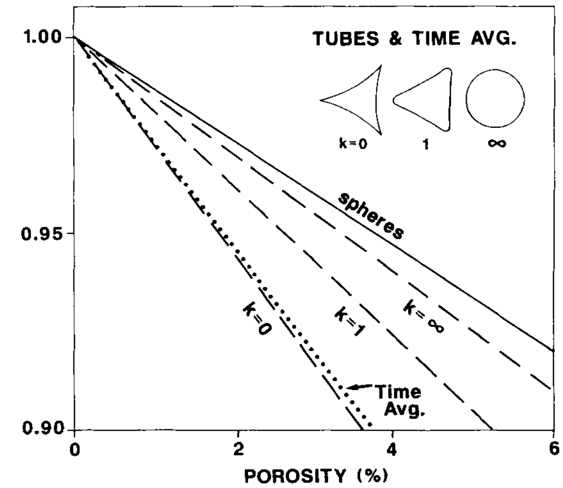


**Figure 12.** Seismic impedance decrease with porosity for the tube model with the various cross-sections shown, i.e. $k = 0$, 1, $\infty$ (after Mavko 1980), and the Wyllie time-average equation.

decrease at small aspect ratios, the results should be interpreted with caution. The laboratory data described below provide some assurance that the theory is a good approximation up to crack or pore densities of at least 0.5.

The case of horizontally aligned pores and vertically incident waves is shown in Fig. 13. The effect is seen to be about double that of the isotropic case, so the porosity estimates are about half for a given impedance reduction.

The above analysis does not consider variations in pore pressure, only that all pores are fluid filled. Changes in pore pressure can produce velocity changes of 10–20 per cent even in low porosity rocks (Christensen 1984; Jones 1983; Todd & Simmons 1972), presumably through changes in pore shape and size. One mechanism for generating variable pore pressure is variable degrees of metamorphic dehydration. For example, laboratory measurements show a velocity decrease of about 10 per cent for dehydration of a basalt and a greywacke, confined at 5–6 kbars. It is concluded that high pore pressures were produced by the fluid generated, and the effective confining pressure (Robin 1973; Nur & Byerlee 1971) was decreased to near zero. Jones (1983) has shown through synthetic seismograms that deep crustal

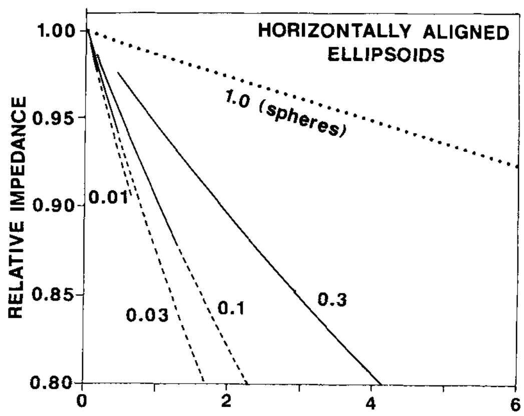


**Figure 13.** Seismic impedance decrease with porosity for horizontally aligned ellipsoidal pores with a range of aspect ratios, and vertically incident waves (after Hudson 1980).

reflectors can be modelled assuming local variations in pore pressure but it is difficult to maintain substantial variations in pore pressure over the required distances of a few tens of metres unless the permeability is unreasonably low. As well, it is expected that the lower crust porosity will everywhere be at or near lithostatic pressure (very small confining pressure), since over geological periods of time at over 400 °C the grains have negligible strength and will undergo ductile deformation. This conclusion is well supported by numerous studies in metamorphic geology. Etheridge, Wall & Vernon (1983) and Fyfe *et al.* (1978) document the evidence for fluid pressures being equal to or greater than the lithostatic pressure (minimum principal compressive stress) during most crustal metamorphism. Thus, the pore pressure in the lower crust is probably uniform and close to lithostatic, and the required layered impedance contrasts are due to simple variations in porosity rather than variations in pore pressure.

**Shear-wave constraint on pore shape**

The above analysis shows that for the ellipsoidal pore model, a wide range of porosity contrasts and pore aspect ratios can provide the required impedance contrasts of 5–10 per cent within the lower crustal lamellae and some further constraint is needed. Shear-wave data provide one such constraint. Fig. 14(a) shows Poisson's ratio as a function of porosity for ellipsoids with different pore aspect ratios and for tubular pores. Fig. 14(b), which shows Poisson's ratio as a function of aspect ratio for several amounts of P-wave impedance decrease, illustrates the strong dependence. 'Thin' cracks with aspect ratios less than about 0.03 cause an increase in Poisson's ratio, 'thick' cracks with aspect ratios of around 0.1 cause a decrease, while spherical pores cause little change.

There are very few direct measurements of Poisson's ratio for the lower crust, and even these are usually from refraction studies for which the wave propagation is near-horizontal, so they must be interpreted with caution. Most studies suggest ratios between 0.25 and 0.27. Examples are in SW Germany where lower crustal reflectors are particularly strong and the estimated Poisson's ratios are 0.24–0.30, averaging 0.26 (Holbrook, Gajewski & Prodehl 1987; Luschen *et al.* 1987), and in Britain where Poisson's ratio is estimated to be about 0.25 near highly reflective regions (e.g. Hall, 1986a). Frequency and amplitude variations with reflection angle have the potential of giving constraints on Poisson's ratio, but the data so far reported have been inconclusive (e.g. Louie & Clayton 1987; Luschen *et al.* 1987).

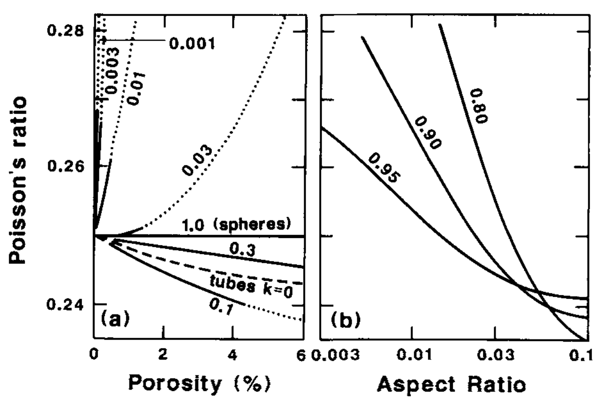


**Figure 14.** (a) Poisson's ratio as a function of porosity for different ellipsoidal pore aspect ratios. (b) Poisson's ratio as a function of pore aspect ratio for several values of impedance decrease.

The zero porosity Poisson's ratio is poorly known, but most shield areas where lower crustal reflectors are infrequent and porosity is inferred to be low from electrical data, have ratios of about 0.25 with occasional higher values (e.g. Jones, 1981). Laboratory data indicate that the observed seismic velocities of the lower crust allow a wide range of compositions even without the effect of porosity (e.g. Christensen 1982; Hall 1986b; Rybach 1987), but rocks that are likely to occur have zero porosity Poisson's ratio values in the range 0.25–0.27. Even mafic rocks, which have common laboratory Poisson's ratios of 0.30, usually have lower zero-porosity values. Substantially lower values are common only for rocks containing quartz (e.g. Kern 1982) which is not expected in the lower crust. Thus, no change in Poisson's ratio, or a small increase in the lower crust due to porosity is suggested. A 5 per cent decrease in impedance in a porous layer and a limit of Poisson's ratio change between −0.1 and +0.3 requires the aspect ratio to be above 0.003. For an impedance decrease of 10 per cent or more, the allowed aspect ratio range is narrower, i.e. greater than about 0.01. There is also a weak constraint to less than about 0.06 provided by the decrease in Poisson's ratio predicted at ratios around 0.1.

**Laboratory measurements and pore shapes**

Empirical estimates of the effect of small-scale porosity on the velocity of crystalline rocks can be obtained from laboratory measurements. Unfortunately, we have found no reported measurements under simulated lower crustal conditions with controlled porosity variation. However, even though low temperature laboratory sample pore structure may not be representative of that in the lower crust, such laboratory data provide an important test of the theoretical models for the effect of porosity on seismic properties.

There have been only a few studies in which rock samples with constant matrix composition and significant accurately measured porosity variation have been studied under high confining pressure. Confining pressure is required to close the extensive microcracks characteristic of surface rocks, since, although the effective or confining pressure in the lower crust may be small (see below), such cracks will probably be annealed. Most measurements with a porosity range are on marine basalts.

As an example, a collection of subaqueous basalts with up to 15 per cent, mostly vesicular porosity (i.e. numerous roughly equidimensional pores), exhibits a porosity effect on both velocity and Poisson's ratio similar to that predicted for an effective aspect ratio of about 0.15 (Fig. 15a) (Hyndman, Christensen & Drury 1984; Hyndman 1979). Poisson's ratio decreases with increasing porosity as predicted for ellipsoidal pores with that aspect ratio (Fig. 15b). It is important to note that for this suite of rock samples that has most porosity in apparently nearly spherical vesicular pores, the effective aspect ratio is only 0.15. Presumably this is

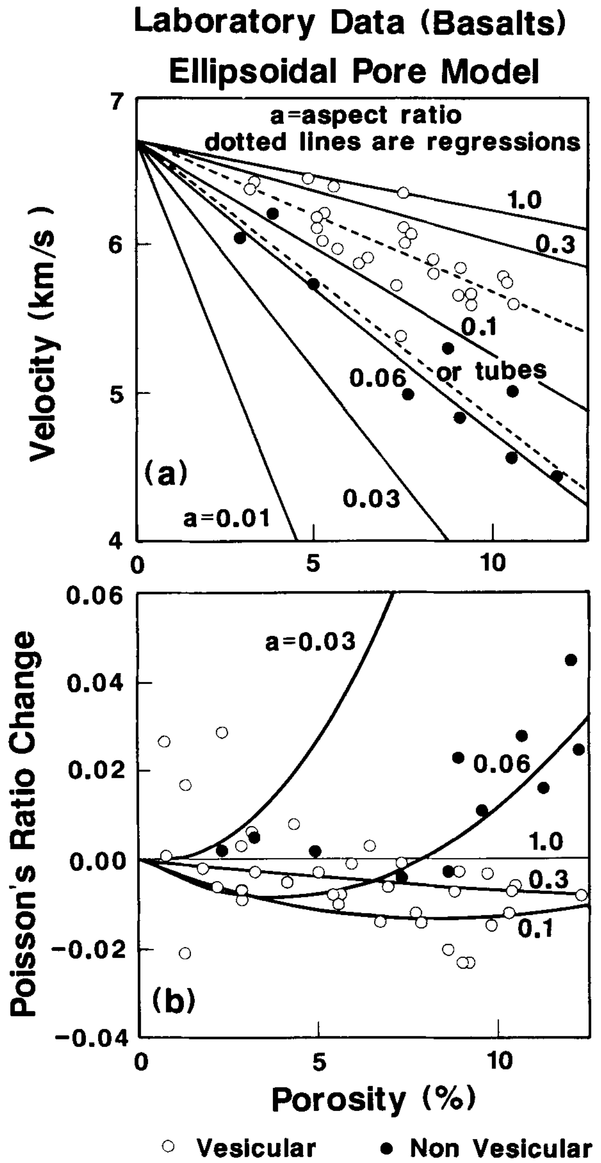


**Figure 15.** (a) Velocity as a function of porosity for subaqueous vesicular basalts (open circles), and for subaerial basalts with primarily grain boundary porosity (solid dots). The ellipsoidal pore model with several pore aspect ratios is shown for comparison. (b) Poisson's ratio change as a function of porosity.

because most of the irregularities and deviations from ideal shapes have smaller effective aspect ratios. Representing quite a different pore structure, a suite of subaerial basalts with less than 10 per cent primarily grain boundary porosity, fit an aspect ratio of 0.06 (Fig. 15a). Poisson's ratio increases rapidly with porosity, again as predicted (Fig. 15b).

On both accounts, velocity and Poisson's ratio, these examples of laboratory data fit the ellipsoidal pore model with effective aspect ratios of 0.06–0.15. Thus, while the laboratory measurements do not well simulate deep *in situ* conditions, they have effective pore geometries similar to that concluded for the lower crust. The laboratory data also provide important evidence that the theoretical relations for ellipsoidal pores describe the behaviour of real porosity up to very high crack densities, at least to 0.5.

### Equilibrium pore shape

An important estimate of the effective pore aspect ratio can be obtained from the expected equilibrium pore shape. Given the pore shapes, it should be possible to determine the effect of porosity on seismic properties, but the exact calculation has not yet been made. Schmeling (1985a,b) approximated an equilibrium pore geometry by a model composed of both spherical and tubular pores, representing porosity at grain corners and grain edges, respectively. However, at dihedral angles near 60°, the theoretical equilibrium pore shape resembles a tetrahedron with concave faces (e.g. von Bargen & Waff 1986). This shape is not well approximated by the Schmeling model which probably overestimates the equivalent ellipsoid pore aspect ratio.

To obtain a first-order approximation, we assume that the equilibrium shapes have a seismic behaviour similar to ellipsoids, with an equivalent aspect ratio having the same surface area and volume. A single pore per grain corner is assumed. This effective aspect can be calculated using the plot of pore surface area as a function of dihedral angle and porosity given by von Bargen & Waff (1986). We recognize that this approach may not give an accurate approximation; in particular, since pore shapes are concave outward for dihedral angles less than 70°, this calculation will probably underestimate the true effective aspect ratio. However, we also have obtained similar results from a rough approximation of the ratio of pore cross-section to volume. At the porosity of 1 per cent for which von Bargen & Waff provided data, the effective aspect ratio is about 0.1 for a dihedral angle of 40°, 0.15 for an angle of 60° and 0.25 for an angle of 70°.

As seen by microscopic examination of samples treated to achieve equilibrium pore shapes, and of natural metamorphic rocks, there undoubtedly are significant deviations from this equilibrium pore model. First, a rather wide spectrum of dihedral angles are measured. Secondly, non-uniformities, particularly much elongated pores are observed (e.g. Watson & Brenan 1987; Hall 1986b), probably in part due to grain anisotropy and to the multiple mineral phases present. As well, large isolated pores of arbitrary aspect ratio and volumes greater than that of individual grains in the matrix are observed. The effective pore aspect ratio thus may be significantly different, probably smaller, than that estimated.

### Estimated porosities in lower crustal lamellae

The effective pore aspect ratio in the lower crust is not well constrained, but three independent approaches, Poisson's ratio data, laboratory sample data and equilibrium pore shape provide some information. The available lower crust Poisson's ratio data indicate that aspect ratios less than about 0.01 are unlikely, and that the values are probably above 0.03. The upper limit is very poorly constrained, but ratios much over 0.1 result in quite low density if there is substantial porosity. Perfect equilibrium pore geometries have estimated aspect ratios of 0.1–0.3, but as indicated above, real pores may have smaller average values. The available laboratory data for basalt samples indicate ratios of 0.06–0.2.

In Fig. 11 we show a reasonable allowable range from 0.01 to 0.3. The narrower range from 0.03 to 0.1 indicates values for randomly orientated pores that give the estimated reflector impedance contrasts with porosities in agreement with those estimated from the electrical resistivity data. If the pores have a preferential horizontal alignment, higher aspect ratios can give agreement. It is interesting to note that the inferred pore aspect ratio is very similar to that estimated for the lower crustal laminae themselves, i.e. a few kilometres in length and roughly 100 m in thickness, or aspect ratios of about 0.1–0.01. As shown above, the tube model with $k = 0$ that appears to most closely approximate the grain edge pore shapes for dihedral angles near 60°, also gives a porosity effect close to that for ellipsoids of 0.1 aspect ratio. With the narrower aspect ratio range, a decrease in seismic impedance of 5–10 per cent requires from 1 to 4 per cent porosity contrast. With the wider range, 0.5–6 per cent is required. For brighter reflectors with estimated impedance contrasts of over 10 per cent, the porosity contrast must be larger, or the effective pore aspect ratios must be smaller, or the pores must be horizontally aligned.

The intervening higher velocity layers or medium surrounding the lamellae undoubtedly has some porosity which must be included in an average porosity for comparison with the results from the electrical resistivity data. A very rough idea of the porosity can be obtained from the resistivity data itself for lower crustal areas where there are few reflectors and no good conductors. These areas commonly have resistivities of about $10^3\,\Omega$m (Shankland & Ander 1983), which by the analysis below gives a porosity in the range of 0.1–0.5 per cent. The porosity estimated for the high porosity layers is thus slightly higher. Our preferred estimate for the overall average in a 10 km lower crust is then 2–3 per cent, assuming that there is half high and half low porosity, and that the impedance contrasts are in the range of 5–10 per cent. Thus, while the uncertainties are large, there is general agreement between the electrical resistivity estimates of lower crustal porosity and that required to explain lower crustal reflectors.

## DEEP FAULTS, BRIGHT SPOTS AND VELOCITY IN THE LOWER CRUST

Three other important features of crustal multichannel reflection profiles relevant to this discussion are deep dipping reflectors usually interpreted as faults, mid-crust seismic 'bright spots' and low velocity in the mid to lower crust. A detailed discussion of these features is beyond the scope of this article, but their possible relation to lower crustal reflectors as generated by fluids is relevant.

### Deep reflective faults

Two things about major fault zones are particularly relevant to the hypothesis of lower crustal porosity. The first is that exhumed deep crustal fault or shear zones that are vein filled have transmitted enormous amounts of water. This conclusion comes from the very large water–rock ratios required to explain the volumes of silica and other minerals deposited (e.g. discussion in Fyfe *et al.* 1978). Frequently multiple phases of fluid flow are inferred, separated by considerable periods of time. It is generally concluded that the fluid has moved upward from great depths rather than circulating from the surface (e.g. Kerrich, LaTour & Williams 1984). The relevant question is, have such fault zones tapped and drained pervasive lower crustal fluid?

The second feature is the difficulty of explaining why mid-crustal fault zones are so strongly imaged. There are many possibilities including juxtaposition of different rock types and mylonites in the fault zone (e.g. Jones & Nur 1982, 1983, 1984; Jones 1983; Christensen & Szymanski 1988) but the inferred reflectivities are near the reasonable limits for these mechanisms. Chemical alteration within the fault zone associated with the concentrated fluid fluxes may be an important source of the required impedance contrasts (e.g. Etheridge & Vernon 1983). It is also possible that some mid-crust reflective fault zones are at present filled with free fluid or have high porosity (Jones & Nur 1982; Jones 1983).

### 'Bright spots'

Isolated seismic reflection 'bright spots' in the middle crust shallower than the generally reflective lower crust, are infrequent but are rather striking. They have very high reflectivity (reflection coefficients may be over 10 per cent) are subhorizontal and have lateral extents of only a few kilometres. They are generally mid-crustal; Rio Grande rift at 18–24 km, Death Valley (California) at 15 km, eastern Nevada at 11–17 km, Surrency (southern Appalachians) at 15 km, and Haslach (Black Forest, Germany) at 10 km (e.g. Brown 1987; Brown *et al.* 1987; Luschen *et al.* 1987; Wille *et al.* 1989). Some of the areas have had recent volcanic activity and crustal magma chambers are inferred, and some are in tectonically active high heat flow areas where magma within the crust is possible. However, others such as the Surrency area in very stable areas of low crustal temperature, and large lenses of very high porosity or even volumes of free fluid have been hypothesized as the only satisfactory explanation (Wille *et al.* 1989). Luschen *et al.* (1987) also suggest fluids as the cause of the Haslach bright spot. Again the relevant question is, are these reflectors concentrations of fluid, perhaps rather transient and short lived, that have moved up from the lower crust? The fluid porosities required are high, probably over 5 per cent (see models above).

### Seismic velocity in the deep crust

Seismic refraction velocities provide an important constraint on the composition of the lower crust. The velocities found are commonly in the range of 6.5–6.8 km s$^{-1}$. Most rocks that have this velocity range measured in the laboratory under simulated deep crustal conditions are of intermediate composition. However, an intermediate composition is inconsistent with other estimates of a largely mafic lower crust, such as obtained from xenoliths and the composition of rocks involved in crustal accretion. The discrepancy is particularly evident in Phanerozoic areas. A mafic composition can be reconciled with the refraction data if the *in situ* velocities are commonly reduced from the mafic composition values of 7.0–7.2 km s$^{-1}$ to the observed lower values by porosity of a few per cent. The required porosity of several per cent is close to that estimated from magnetotelluric data. Hyndman & Klemperer (1989) discuss this model in more detail.

Low velocity layers in the mid-crust have also been explained by high porosity (e.g. Berry & Mair 1977). In several areas the low velocities have been associated with low electrical resistivity perhaps produced by high porosity (e.g. Mitchell & Landisman 1970; Wenzel *et al.* 1987) and with high Poisson's ratio (e.g. Jones 1981). We note that the effect of porosity may also be important to gravity models where a generalized relation between velocity and gravity is used as a constraint.

## TEMPERATURE CONTROL OF LOWER CRUST REFLECTIVE AND CONDUCTIVE ZONES

### Upper boundary–lower temperature limit

The general observation that broad reflective zones and thick high-conductivity layers are restricted to the lower crust has been quantified by Klemperer (1987) and Adam (1976, 1987). In a global compilation Klemperer found the depth to the tops of reflective zones to be generally correlated with heat flow and a nearly constant temperature of 300–400 °C was estimated (see also Wever *et al.* 1987; Meissner & Kusznir 1987; Royer & Danis 1988). Similarly Adam has found a correlation between the top of high conductivity lower crustal layers and heat flow with a similar estimated temperature limit (see also Shankland & Ander 1983; Haak & Hutton 1986). One interpretation of these boundaries is that they represent the top of a porous zone of free water. The seismic and electrical data are combined in Fig. 16. Constant temperature curves using the generalized relation between heat flow and crustal temperature by Champan (1986) are also shown. A variation of at least 50 °C is to be expected in such temperature estimates because of variations in crustal radioactive heat production and thermal conductivity. Although the global data are scattered, the tops of most reflective bands and conductive layers are about 350–400 °C over a range of a factor of 3 in depth and in surface heat flow. Why substantial pore fluid should be limited to temperatures above 350–400 °C is discussed in the next section. An upper temperature limit of about 700 °C for fluid in the crust may also apply based on the evidence for the crust being dry under granulite facies conditions as discussed in the following section.

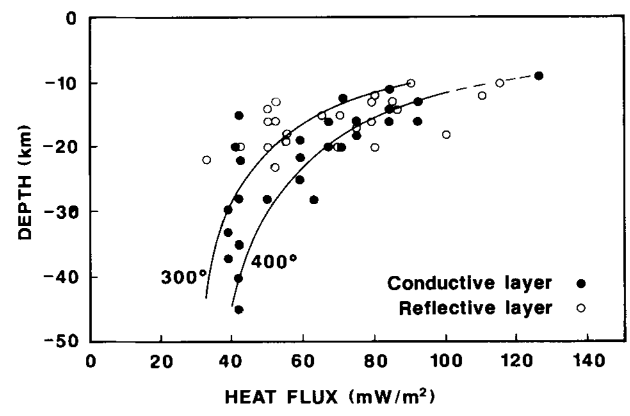


**Figure 16.** A global compilation of estimated depths to the top of seismically reflective (from Klemperer 1987) and low electrical resistivity (from Adam 1987) lower crust. The standard isotherms are from Chapman (1986).

### Mechanisms for porosity being limited to the lower crust

Two mechanisms are suggested for significant porosity being restricted to the lower crust where the temperatures are above about 400 °C; metamorphic hydration reactions and silica precipitation that form an impermeable layer, and a change from brittle to ductile deformation that anneals cracks and restricts porosity to low permeability grain boundaries.

Free water either generated by dehydration reactions in the lower crust or from the mantle (discussed below) can move up until reduced temperature results in hydration reactions and mineral precipitation, forming an impermeable layer that traps the fluid horizon below (e.g. Jones 1987; Etheridge *et al.* 1983, 1984). The fluid beneath the impermeable horizon may be in irregular layers as required by the seismic reflectivity. It may be significant that the critical temperature of about 400 °C is near the top of the greenschist metamorphic facies boundary at lower crustal pressures. Above the boundary are blueschist facies conditions in regions of low temperature gradients, and prehnite–pumpellyite or zeolite facies in regions of normal to high temperature gradients. A simplified metamorphic facies diagram from Hyndman (1985) is shown in Fig. 17, with geotherms typical of high, average and low heat flow areas (50, 60 and 80 mW m$^{-2}$). The pore pressures are assumed to be lithostatic, equal to the overlying formation load. Of course substantial variation in the hydrated mineral stability fields occurs with different rock and fluid compositions. Substantial solid phase volume changes due to hydration should occur at the greenschist facies boundary. In areas of particularly low heat flow, the region above the boundary may have the high seismic velocities and high densities that appear to be characteristic of blueschist facies rocks.

Water may also be restricted to the lower crust by precipitation of silica along grain boundaries that sharply

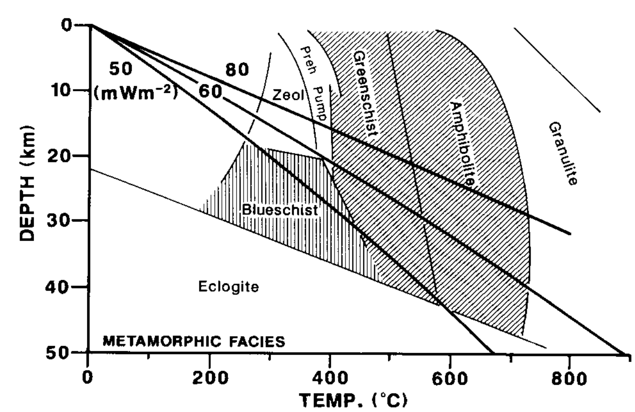


**Figure 17.** Simplified metamorphic facies diagram from Hyndman (1985) with geotherms from Chapman (1986) superimposed, for surface heat flows of 50, 60 and 80 mW m$^{-2}$.

reduces permeability (e.g. Walder & Nur 1984). Since fluid in the lower crust should be saturated in silica, a decrease in temperature and pressure in upward moving fluid will result in substantial amounts of precipitate. For example, in the lower crust a decrease in temperature from 500 to 400 °C decreases the solubility markedly, from about 1 to about 0.3 per cent (e.g. Fyfe *et al.*, 1978).

Another process that may be important is the decrease in crustal viscosity and grain strength with depth. For example, Meissner & Kusznir (1987) have concluded that the lower crustal reflective zones only occur where the effective bulk viscosity is less than about $10^{22}$ Pa s$^{-1}$. They also noted that the depths to the brittle–ductile transition estimated from earthquakes (e.g. Sibson 1982) were similar to those for the tops of reflective zones. The relevant constraint may be that stable grain boundary fluid is restricted to temperatures at which the grains can readily deform and adjust their shape to the equilibrium boundary configuration. This grain shape adjustment could be through recrystallization, but there will still be a temperature control. In the upper crust, significant permeability will also occur in fractures, while in the lower crust at high temperature they will rapidly heal unless kept open with continuing pore pressure exceeding lithostatic.

### Lower boundary–maximum temperature limit

It is important to recognize that the model for fluid in the lower crust probably applies only to areas which have not been subject to temperatures above about 700 °C without the subsequent addition of water, as well as to a minimum temperature of about 400 °C. This range covers the greenschist to amphibolite facies conditions, for which metamorphic and fluid inclusion studies indicate chemical equilibrium with free saline water at close to lithostatic pressure. At higher temperatures in granulite facies conditions (e.g. Fig. 17) there is much evidence for there being little or no free water. Studies of geochemical equilibria as first partial melting is approached indicate no free water; the water in the melt must come almost entirely from the breakdown of hydrated minerals such as biotite and hornblende (Wannamaker 1986; Clemens & Vielzeuf 1987).

Fluid inclusion studies (e.g. Crawford & Hollister 1986; Hollister & Crawford 1981; Roedder 1984, pp. 368, 578; Touret 1977) indicate few inclusions in granulite facies rocks, and that those that do occur are primarily $CO_2$. Dehydration may occur through the pore water being taken up by partial melts, or by flushing by $CO_2$. This dehydration may explain the evident lack of pore water in the lower crust in shield areas. They have been dehydrated during high temperature granulite facies metamorphism and have had no subsequent water input (e.g. Hyndman & Hyndman 1968). Of course subsequent input of water may occur, either from below or through continued dehydration of hydrated minerals. Feldman (1976) also proposed that water in the lower crust lies above the granulite–amphibolite facies boundary.

## LAYERING IN THE LOWER CRUST

The preceding calculation of the maximum permeability for water to be held in the lower crust assumed that the layer was homogeneous. In fact, the seismic data appear to require interleaved high and low porosity lamellae. The velocity variation with depth may be very irregular but it must contain components with the order of 100 m vertical scale. The abrupt transition from interconnected to pinched-off pores predicted for dihedral angles just above the critical value of 60° allows pronounced changes in permeability with small changes in any one of a number of parameters, and thus to permeability and porosity layering. These include (a) changes in fluid composition, (b) changes in rock composition, (c) non-equilibrium stresses such as the gravitational effect or lower crustal shear, and (d) instabilities during the expulsion of excess fluid from the region that are preserved as porosity variations. No evidence for layered changes in fluid composition have been reported but there is some indication for the other three mechanisms.

Layered variations in composition are very commonly observed in greenschist and amphibolite facies metamorphic rocks, usually associated with shear deformation (e.g. Etheridge *et al.* 1983; summary in Hyndman 1985), but usually on a scale of millimetres to centimetres rather than than tens to hundreds of metres as required by the seismic reflection data. Such fine-scale layering may be important in stabilizing vertical fluid motion since pore surface tension effects can resist the gravitational effects at this scale (von Bargen & Waff 1986; Stevenson 1986). No reports of metamorphic rocks with systematic variations over the dimension of the inferred seismic lamellae have been found but this may reflect a lack of appropriate observation.

Waff (1980) has developed a model for the stabilization of partial melt in the mantle in layers based on the effects of the gravitational field on pore geometries. If the fluid is interconnected, a pressure difference between the fluid and solid grains will develop vertically because of the difference in their density. The pressure difference may change the pore fluid geometry such as to cut off pore fluid interconnection and produce an impermeable layer. Above this layer the pressure difference will again increase upward until the pores are once more pinched off. The result is permeability and porosity layering. Compositional layering may be necessary for this mechanism to be effective. Stevenson (1986) has also argued that the gravitational forces are only adequate to oppose the surface energies over vertical distances of centimetres.

Stress field anisotropy could also be important for the horizontal orientation of porosity, for example related to extensional regimes. Ductile shearing may also play a role, providing a horizontal foliation which gives a preferential horizontal porosity orientation.

Another mechanism is instability in the rising fluid while the porosity is high. Solitary waves, a phenomenon similar to zone refining, may occur (e.g. Scott & Stevenson 1986; McKenzie 1985, 1987) because of the rapid variation of permeability with porosity suggested at 2–3 per cent porosity. Such waves might leave a residual porosity variation with the required wavelengths.

## ORIGIN OF FLUID IN THE LOWER CRUST

The sources of water in the lower continental crust can be grouped into three main types (e.g. Hall 1986a; Kay & Kay

1986; Fyfe 1986): (a) wet rocks carried to the lower crust by geological processes, (b) water introduced into the overlying continental crust in subduction zones from dehydrating oceanic crust and sediments, and (c) water from the mantle generated by devolatilization processes. It is important to recognize the different time-scales involved, and thus the time that the fluid must have been held in the lower crust.

The processes that have formed the continental crust are complex and may be different for Archean shields. If the latter are excluded, subsequent crust appears to have come largely from accretion of oceanic terranes, including ocean plateaux, island arcs and marginal basins that have been thickened by tectonic, depositional and igneous processes. Most such rocks start wet, with from several to about 10 per cent porosity for volcanic rocks, to tens of per cent for sediments. With increase of temperature and pressure on burial, this water is partly expelled and partly taken up in hydrated minerals, the amount in the latter decreasing with increasing metamorphic grade. Up to amphibolite facies there must be excess free water as long as temperatures are increasing.

In a previously stable environment, an increase in crustal temperature will result in further metamorphic mineral dehydration and generate free water. Following the equilibrium pore model discussed above, this free water will be expelled until pinch-off porosity (several per cent) is approached. The remaining water will be trapped by very low permeability. Decreasing temperature will also facilitate trapping of the remaining fluid (e.g. Hall 1986a). Hall has also discussed why free water may remain trapped in pores on cooling and not be taken up in retrograde metamorphic reactions. Processes that increase crustal temperature and thus result in new free water include crustal extension or rifting, mantle upwelling in back arc regions, and continental hotspots.

A special case of wet rock introduction into the deep crust is that of major overthrust sheets in continent collision zones such as the Appalachians of the eastern United States (Cook *et al.* 1979, 1986; Ando *et al.* 1984). Sediments and upper crustal rocks with high porosity may have been over-ridden.

Subduction provides several obvious mechanisms for introducing water into the deep continental crust. The first is the dehydration of downgoing crust and sediments (e.g. Fyfe *et al.* 1978; Kay & Kay 1986). This model has been developed to explain seismic reflectors and conductive layers beneath the Cascadia subduction zone (Hyndman 1988). The oceanic crust starts with perhaps 5 per cent water as porosity and hydrated minerals (e.g. Anderson, Uyeda & Miyashio 1976; Ito, Harrison & Anderson 1983). It is inferred to reach at least amphibolite facies under the continental crust in forearc areas (e.g. Hyndman 1988) so most of this water must be lost upward. While an unknown amount goes back up the main subduction thrust to the ocean, a major part probably goes into the overlying continental crust. There is a continuous supply so that present forearc reflectors do not require fluid to be held for a long period, i.e. no more than several Ma. The required residence times are longer if this mechanism is invoked for former forearc areas now incorporated within a continent.

Arc volcanism also provides a mechanism for the introduction of water into the deep crust (e.g. Fyfe 1986). The distribution to the base of the crust could be over a much wider area than just that of the volcanic arc. However, the volume is probably much less than shallower dehydration of downgoing crust and sediments, since only the highest temperature hydrated minerals will remain in the subducted crust as far as beneath the volcanic zones.

Mantle devolatilization processes are a poorly defined but potentially large source of water for the lower crust (e.g. Kay & Kay 1986). Localized zones of mantle upwelling and higher than normal mantle temperature where water is carried in partial melts are the prime candidates. These include extensional and rift regions and continental hotspot areas. Water may be removed from much larger mantle volume and over a much larger area than expressed by the observed surface volcanism because it is more mobile than partial melts. Water from the mantle may also be associated with widespread crustal underplating.

## CONCLUSIONS

Strong geophysical evidence for fluids in the lower crust comes from observations of low electrical resistivity. Many younger areas have resistivities in the range of 10–30 Ωm in contrast to an average of about $10^3$ Ωm for the lower crusts of shields. The low average resistivity in Phanerozoic areas can be explained by a porosity of about 0.5–3 per cent assuming a fluid salinity close to that of sea-water as suggested by fluid inclusion data, and reasonable pore geometries such as to satisfy Archie's law with exponent 1.5–2.0, or equilibrium pore geometry. It is difficult to explain the electrical results with less than about 0.5 per cent porosity assuming 10–15 km thick layers. Higher porosities are required if the conductive layers are postulated to be thinner. The main difficulty with the simple pore fluid models of electrical resistivity is that the required pore interconnection appears to be inconsistent with the low permeability required to hold the fluid in the crust for geologically long periods of time, assuming reasonable values for grain size and fluid resistivity. Either very small effective grain sizes or more complicated geometries than the simple isotropic equilibrium model are required.

The frequently observed lower crustal seismic reflectivity has a number of reasonable explanations. However, the other explanations do not readily explain why the reflective structures appear to be restricted to the lower crust and why they rarely occur in shield areas. Fluid in the lower crust can explain most of the observed seismic reflections with layered porosity contrasts of 1–4 per cent if the effective pore aspect ratios are between 0.03 and 0.1 per cent. These aspect ratios are somewhat lower than those predicted by the simplest models of equilibrium pore shape, but they are within the reasonable range for equilibrium conditions and they are in agreement with several other loose constraints. The seismic data for reflective crust require subhorizontal lamellae roughly 100 m thick and a few kilometres long with impedance contrasts of 5–10 per cent. The formation of the layering is an important unresolved question in this model although a number of possible mechanisms for producing porosity layering are suggested.

If fluids are the primary cause of lower crustal reflectors, a correlation should be seen between reflective and conductive lower crusts. There are a number of examples of reflective and conductive lower crust coinciding, but the data are inadequate to determine whether or not low resistivity generally occurs with reflectivity as this model suggests. Conductive zones that are not reflective of course can occur if there is no layering of the required vertical scale, but such zones should be characterized by low velocity and possibly higher than normal Poisson's ratio.

Reflective and conductive zones appear to be restricted to temperatures above 350–400 °C, approximately at the top of the greenschist facies conditions. This may be the result of trapping below an impermeable layer formed by hydration reactions or silica precipitation, or by the minimum temperature for ductile behaviour such that equilibrium pore shapes which have the required low permeability occur. The cores of shields characteristically have transparent and resistive lower crusts and are postulated to have been heated to granulite facies conditions (about 700 °C) and dehydrated, with no subsequent addition of water.

A number of sources of water for the lower crust are also suggested. These include wet rocks carried to the lower crust by geological processes such as margin sedimentary basin of terrane overthrusting and accretion, dehydration of downgoing plates in subduction zones and water from the mantle released in devolatilization processes, particularly in extensional regimes and over hotspots.

The primary conclusions of this study are that electrical sounding results require at least 0.5 per cent and probably 1–3 per cent free water porosity in the lower crust of many geologically younger areas. Such porosity must have an important effect on seismic velocities, and if layered, it can readily produce strong lower crustal reflectors.

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