STRESS AND SEISMICITY IN THE LOWER CONTINENTAL CRUST: A CHALLENGE TO SIMPLE DUCTILITY AND IMPLICATIONS FOR ELECTRICAL CONDUCTIVITY MECHANISMS

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Abstract. The lower crust is generally considered to be an aseismic, weak zone where fluid distribution might be governed by textural equilibrium geometries. Saline fluids below the transition from brittle to ductile rheology have been advanced as a joint explanation for deep crustal conductivity and seismic reflectivity, the depth of onset of both phenomena being apparently bounded by isotherms in the 300–450 °C temperature range. Some petrologists, meanwhile, contest that the deep crust should be devoid of extensive fluid networks. This review exposes some geophysical exceptions to the statistical norm suggested by global geophysical data compilations and presents counter-arguments that the lower crust in places may be both dry and strong, that fluids if at all present at such depths may not necessarily be connected and that fluid mobility in the lower crust may be more limited and heterogeneous than commonly assumed.

Laboratory data on crustal rocks implies that the transition from brittle to ductile rheology actually occurs over a much broader range of temperatures than 300–450 °C, and the apparent association of deep crustal conductive horizons with a temperature field of 300–450 °C may be interpretable in terms of formation temperatures of graphite, rather than fluids and brittle-ductile transition rheology.

High v_P/v_S ratios from a 6 km thick, seismically layered zone below the Weardale granite, NE England can be explained by underplated mafic material. They are unlikely to be explained by fluids in an area where deep crustal conductance has been shown to be relatively low, unless conventional assumptions regarding deep crustal fluid distribution are inadequate or false.

Perusal of the literature reveals that lower crustal seismicity is less seldom than generally appreciated. Interpretation of earthquakes nucleating at lower crustal depths is ambiguous, but in some tectonic regimes may indicate preservation of brittle rheology to the Moho and a lower crust that is predominantly mafic and dry.

A better understanding of lower crustal deformation mechanisms and history may provide better insight into deep crustal conductivity mechanisms. Recent rock mechanical experiments suggest that permeability (and thus fluid connectivity) may be decreased by ductile shearing, whereas ductile shearing may aid graphitisation at lower crustal temperatures. If the lower crust in some regions is strong, this may explain the apparent preservation of both extant- and palaeostress orientations in interpretations involving electrical anisotropy.

Keywords: brittle-ductile transition, lower crust, magnetotellurics

1. Introduction

Lower crustal conductivity sections have been reviewed by Haak and Hutton (1986), Jones (1987) and Jones (1992). It is, by now, well established that the deep

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crust frequently – though not ubiquitously – and to varying degrees manifests a region of enhanced conductivity (e.g. Shankland and Ander, 1983; Haak and Hutton, 1986). In this review I will concentrate on likely mechanisms for explaining this conductivity enhancement and for explaining the growing number of observations of deep crustal anisotropy (e.g. Cull, 1985; Rasmussen, 1988; Kellet et al., 1992). In doing so it will be necessary to draw on data from other geophysical methods, laboratory data, and petrological data.

The lower continental crust is generally envisaged to be that part of the crust which is aseismic (e.g. Meissner and Strehlau, 1982), representing a weak layer that can transfer stress between the brittle upper crust and mantle via ductile flow. Such a weak zone is generally incorporated as a ubiquitous feature into geodynamic models aiming to explain tectonic evolution (e.g. Schmeling and Marquart, 1990; Kruse et al., 1991; Kaufman and Royden, 1994), whilst the brittle-ductile transition zone is often cited as a convenient permeability pinch-off mechanism for preserving fluids in the deep crust (e.g. Bailey, 1990). The general lack of earthquakes emanating from lower crustal depths, coupled with the assumption that earthquakes occur exclusively by 'stick-slip' brittle failure in a frictional regime provides the general argument in favour of a ductile lower crust. Correlations of the depths to deep crustal conductors with the limits of crustal seismicity (e.g. Stanley et al., 1990) and with temperatures believed to be representative of the transition from brittle to ductile rheology (Ádám, 1978; Hyndman and Shearer, 1989) arouses speculation that conductivity anomalies in the deep crust are images of processes (often involving fluids) confined below the mechanical bottom of the upper crust, the generally more resistive character – despite its high fluid content (as demonstrated by both the Kola and KTB boreholes) – of which has been attributed to its brittle nature and low pore fluid pressures inhibiting extensive, lateral, fluid connectivity (e.g. Gough, 1986). The physical explanation of deep crustal conductivity relating to fluids below the onset of the brittle-ductile transition has been extensively adopted in the electromagnetic literature, where a standardised temperature field of 300-450°C is often cited, with little regard to tectonic province or likely lithology. In this paper I will reassess and challenge the basis for the hypothesis that brittle-ductile transition rheology controls the onset of enhanced conductivity due to fluids, consider how rheology is really likely to affect the conductivity of conductive media such as saline fluids and graphite, and hypothesize that a rather more complex view of lower crustal rheology than the current view of the lower crust as a uniformly weak, ductile layer may help to reconcile the preservation of tectonic markers in the lower crust and the fact that anisotropy axes sometimes appear to be better aligned with the palaeostress field rather than the present day stress field.

2. Electrical Conductivity and Temperature: Fluids or Graphite in the Lower Continental Crust?

Anhydrous granulites and other high grade metamorphic rocks from exhumed terrains of presumed lower crustal provenance exhibit laboratory conductivities several orders of magnitude less than those typically observed from field electromagnetic studies probing the deep crust (e.g., Haak and Hutton, 1986). The enhanced conductivities in the deep crust are generally attributed to the presence of saline fluids (e.g., Connerney et al., 1980; Shankland and Ander, 1983; Gough, 1986; Hyndman and Shearer, 1989; Marquis and Hyndman, 1992), black shales and/or graphite (e.g., Camfield and Gough, 1977; Brown,1979; Stanley, 1989; Mareschal et al., 1992; Korja and Hjelt, 1993; Stanica and Stanica, 1993; Haak et al., 1997) and less commonly (in regions of recent tectonic activity) partial melt (e.g., Pous et al., 1995; Chen et al., 1996).

Possible mechanisms for generating saline fluids at lower crustal depths include dehydration of hydrated minerals (Hyndman, 1988), diffusion of volatile content from mafic magmas (e.g., Jones, 1987), dewatering of subducted or orogenically underthrust sediments (Kurtz et al., 1986; Jones, 1987; Newton, 1990) or degassing of the upper mantle (Kay and Kay, 1986). Consumption of fluids by retrogressive dehydration reactions (e.g., Sanders, 1991), their localisation in dislocated brittle fractures (e.g., Gough, 1986) or their loss during uplift (e.g., Shankland and Ander, 1983; Haak and Hutton, 1986) could readily explain the lack of high electrical resistivities manifested by former lower crustal rocks uplifted to the surface.

In order for saline fluids to enhance conductivities an interconnected network is required. Physical problems arise with the fluids hypothesis when the (seemingly mutually exclusive) criteria of extensive interconnectivity, implying high permeability versus residence timescales requiring low permeability are considered. Calculations based on the principles of Darcian flow in a ductile crust (e.g., Bailey, 1990; Warner 1990b) suggest evacuation of hot, low-density fluids from the lower crust within geologically short timescales, implying that although fluids may contribute to high electrical conductivities in active regions, where fluids in the lower crust may be replenished, a convincing mechanism for extending the residence time of free, interconnected, fluids in the lower crust to geologically significant time periods is required if saline fluids are to be the primary cause of enhanced conductivities in stable regions (though fluid transport may be coupled to deformation mechanisms in a more complex manner (e.g., Goldfarb et al., 1991; Wong et al., 1997) than simple Darcy's Law calculations assume). An impermeable layer created by mineral precipitates at mid-crustal depth (e.g., Etheridge et al., 1983), the brittle-ductile transition zone (e.g., Bailey, 1990), an impermeable zone associated with hydration reactions at the top of the greenschist facies (Hyndman and Shearer, 1989), permeability layering arising from differential pressure gradients (Hyndman and Shearer, 1989) or from tectonic stresses (Goldfarb et al., 1991) or vertically overlapping, laterally unconnected, highly conductive lamellae (Merzer and Klemperer, 1992) have been suggested as possible mechanisms. Impermeable barriers are unlikely to serve as more than transitory barriers to arrest ascent of fluids, since pressure on such a boundary could be expected to intensify sufficiently for the fluid to egress (Hyndman, 1988), and a rheological trap formed by the brittle-ductile transition zone (e.g., Bailey, 1990) has been more widely supported in the electromagnetic literature.

Compilations relating the depths to the tops of deep crustal conductors imaged in stable regimes to surface heat flow suggest that they are bounded by isotherms spanning 300–450 °C (e.g. Ádám, 1978). Such a temperature field has been associated with the transition from brittle-elastic to ductile crustal rheology and has thus been taken to lend support to the hypothesis of fluids trapped below the onset of the brittle-ductile transition (possibly as a thin layer within this zone (e.g., Etheridge et al., 1983; Jones, 1987)). The apparent correlation tends to be disrupted in tectonically active regions, where the conductive zones are elevated (Ádám, 1987), and also in central Finland where deep crustal conductors appear to straddle the brittle-ductile transition zone (Korja and Hjelt, 1993).

Even if the brittle-ductile transition is introduced as a means of maintaining fluids in the deep crust for geologically significant timescales, a number of problems remain in hypothesising fluids as the principal cause of deep crustal conductivity. Some of these – the petrological arguments and the actual occurrence of deep crustal seismicity – are dealt with later in the paper.

Whilst moderately raised conductivities may be explicable in terms of fluids, this seems an unlikely explanation where conductances of several 1000 S to tens of 1000 S (e.g., Korja and Hjelt, 1993) are involved. A major contender to fluids as the cause of such enhanced conductivities is carbon (either mantle-derived as a product of CO_2 outgassing (Newton et al., 1980) or biogenic (e.g., Korja et al., 1996)) in the form of graphite and/or black shales. It has been suggested that under strongly reducing conditions (Touret, 1986) graphite might form as a grain boundary precipitate from CO_2-rich or hydrocarbon bearing metamorphic fluids (Frost, 1979; Glassley, 1982; Rumble and Hoering, 1986 and references therein) or CO_2 outgassed from the mantle (Newton et al., 1980). Artificial manufacture of graphite via carbon vapour diffusion requires formation temperatures of between 600 °C (with nickel as a catalyst) and 2000+ °C. In contrast, Léger et al. (1996) find that graphitisation of low grade metamorphic rocks from the Waits River Formation, NE Vermont occurred below temperatures of 450 °C and pressures of 450 MPa, with high-grade metamorphic rocks from the same complex devoid of graphite. Higher grade metamorphic conditions, with temperatures in the 500–700 °C temperature range, may have played a role during graphitisation of deposits in New Hampshire (Rumble and Hoering, 1986).

C_{13} isotope studies have shown that graphite naturally occurring in outcrops is formed from biogenic sources (Rumble and Hoering, 1986; Korja et al., 1996; Large et al., 1994; Mathez et al., 1995) and that graphitisation depends on the crystalline grade of the source hydrocarbons (e.g., Large et al., 1994) and strain energy (e.g., Ross and Bustin, 1990), as well as temperature. Shear stress facilitates completion of graphitisation in the 400–500 °C temperature range (Ross and Bustin, 1990; Large et al., 1994). Roberts et al. (1999) also report the formation of graphite along new mineral surfaces during brittle fracturing in the presence of carbon gases at temperatures above 400 °C, though they acknowledge that the strain rates applied to the laboratory specimens (10^{-6}–10^{-5} s^{-1}) were orders of magnitude greater than typical in nature (10^{-16}–10^{-10} s^{-1}), raising the question if and how such experimental data should be extrapolated to geological conditions. Reconciliation of the timescales on which laboratory measurements can be conducted with geological timescales is a general scaling problem pertinent to all experiments involving a dynamic element.

Graphitic horizons have been interpreted as tectonic markers possibly related to old collision zones (e.g., Korja and Hjelt, 1993). Stanley (1989) documents extensive outcrops of graphitic shales apparently associated with active and fossilised subduction zones, and exposures of black shales have also been mapped along the Grenville Front (Mareschal et al., 1991). Graphitic films have been observed in anorthosite rocks derived from the mid crust in Wyoming (Frost et al., 1989), whilst cores samples from the KTB borehole and proximate surface exposures contain graphite both as disseminated flakes and concentrated along shear zones (Haak et al., 1991; Haak et al., 1997), to which formation temperatures of 250 °C to 340 °C have been assigned (Kontny et al., 1997). Shankland et al. (1997) report electron probe and scanning electron microscopy analyses revealing the occurrence of graphite on the internal cleavage planes of amphibole. Graphite has also been detected in (mafic) xenoliths of recent crustal origin (Padovani and Carter, 1977).

In none of the low- to medium-grade rocks from the aforementioned Waits River Formation was the graphite found to form interconnected networks, but the electrical conductivity of wet rocks was found to be significantly higher than that expected from the fluid content alone (Léger et al., 1996). Duba et al. (1994) have reported an unexpected, anisotropic (factor 3.4 in conductivity at 225 MPa, and too small in itself to explain the anisotropy seen on the field scale) increase in the conductivity of wet rock samples from the KTB borehole with increasing pressure. Their interpretation involves interconnected fluids providing a network for intergranular, solid phase conduction (principally by graphite), so that fluid loss associated with increasing pressure is more than counter-balanced by reconnection of the conductive solid phase (Shankland et al., 1997). Mathez et al. (1995) consider the effect of brittle deformation in disrupting carbon connectivity. A similar loss of connectivity may occur during uplift (Katsube and Mareschal, 1993), and Shankland et al. (1997) consider the pressure-induced reconnection of solid phases to better represent the likely fabric of the rock at depth.

It is inappropriate to pose the question ‘fluids *or* graphite?’ in a global sense. In some cases the amount of conductance may help to distinguish between fluids or graphite as *principal* causes of conductivity (though both fluids *and* graphite may still be present). Bahr (1997) has used fractal random networks of crack distributions to explain the scale dependence of electrical anisotropy between laboratory and field scales. Small biases in the tendency for cracks to form in one direction can reproduce the observed scale dependence and anisotropy is thus interpreted in terms of low degree connectivity, possibly associated with the network lying close to the percolation threshold. This may imply small amounts of a conducting phase and might help to distinguish between significant quantities of fluids or minor quantities of graphite where relatively high conductances are observed in one direction. The percolation threshold and what qualifies as a 'small' amount of a conducting phase can be expected to be influenced by rheology. A more simple explanation for the scale dependence of anisotropy might be that whereas on the field scale the range of fracture sizes is sampled in its entirety, laboratory measurements are generally performed on intact samples, which do not usually contain fractures of dimensions greater than the size of the sample, but rather microcracks and small-scale, grain boundary heterogeneities. If the stress field imprints preferred directions, for example in the form of graphite in shear zones, then there is no need for a system to be close to the percolation threshold to produce the anisotropy observed on the field scale. Furthermore, connectivity in the vertical and not just horizontal plane must be considered (Merzer and Klemperer, 1992).

3. Lower Crustal Seismic Layering, Electrical Conductivity and Magnetisation

Deep crustal seismic profiling in Europe (e.g., BIRPS and ECORS, 1986; DEKORP, 1991), America (e.g., Allmendinger et al., 1983) and Australia (e.g., Mathur, 1983) has often (though not ubiquitously (e.g., Tewari et al., 1997)) imaged a laminated lower crustal structure of alternating high- and low-velocity 'layers' (manifest as extensive sequences of short, brightly reflective elements) with vertical scales of the order 80-200m (e.g., Wenzel et al., 1987; Paul and Nicollin, 1989) and impedance contrasts of the order 20% (Paul and Nicollin, 1989; Warner, 1990a). Some seismic sections also evince mid crustal banded reflections (DEKORP, 1991). Analogous structure in the oceanic crust is uncommon.

Preferred mechanisms for explaining lower crustal reflectivity involve ductile shear (e.g., Reston, 1988; Christensen and Szymanski, 1988; Jones and Nur, 1984), underplating of mafic magma (e.g., McKenzie, 1984; Furlong and Fountain, 1986; Nelson, 1991; de Franco et al., 1997) or stratified zones of free fluids (e.g., Hall, 1986; Hyndman and Shearer, 1989; Merzer and Klemperer, 1992). Singh and McKenzie (1993) have shown that with typical sampling frequencies only layers that are thicker than 50 m should have any effect on seismograms. Magnetotelluric (MT) measurements are inherently unable to resolve such horizontal to vertical anisotropy.

Concomitance has been inferred between deep crustal high conductivity zones and zones of seismic layering (which have also been linked to brittle-ductile transition temperatures (e.g., Hyndman and Shearer, 1989)), and free aqueous fluids appear in the literature as a favoured mechanism for explaining both phenomena jointly (e.g., Hyndman, 1988: Hyndman and Shearer, 1989; Marquis and Hyndman, 1992; Merzer and Klemperer, 1992). Significant quantities of fluids were logged in the deep upper crust perforated by the KTB and Kola boreholes. Fluids pumped from the KTB deep borehole in Bavaria down to a depth of 8.9 km have salinities of 68 gl^{-1} NaCl (1.7 M), corresponding to an approximate resistivity of 0.1 Ωm at 20 °C (Huenges et al., 1997). Ca-Na-(Mg)-Cl fluids of even higher salinity (300 gl^{-1}) were sampled within the Kola borehole (Digranes et al., 1996). Laboratory data on KCl brines compiled by Nesbitt (1993) suggests an overall temperature dependent decrease in resistivity from 20 °C to 350 °C of the order 3.5. Thus, saline fluids at deep crustal depths can be expected to have conductivities exceeding 35 Sm^{-1}. Current hypotheses concerning pore geometry and interconnectivity at lower crustal depths tend to assume that fluid distribution is governed by textural equilibrium geometries (Bulau et al., 1979; Watson and Brenan, 1987), and that cracks of lower aspect ratio than 0.03 (e.g., Hyndman and Shearer, 1989) are not maintained. High v_P/v_S ratios (1.86 $\pm$ 0.05 (Ward et al., 1992)) from a 6 km thick, seismically layered zone below the late-Caledonian Weardale granite, NE England are unlikely to be explained by fluids in an area where deep crustal conductance has been shown to be relatively low ($<$200 S), unless conventional assumptions regarding deep crustal fluid distribution are inadequate or false (Simpson and Warner, 1998). The lid of a relatively weak, deep crustal conductor below the Weardale granite is modelled at 12 $\pm$ 4 km, and may correspond to temperatures in the 350 to 450 °C range. In contrast, P- (Evans et al., 1988) and S-wave (Ward et al., 1992) seismic sections evince sharp onsets of seismic layering at an unusually deep depth of 24 $\pm$ 2 km (Ward et al., 1992). High surface heat flow (of the order 95 m Wm^{-2} (Evans et al., 1988)) over the granite suggests that temperatures of 565 $\pm$ 25 °C (calculated according to Chapman, 1986) could be reached at this depth, implying that the processes governing the onset of seismic layering are not necessarily governed by temperatures in the 300–450 °C range.

Shear zones have been extensively documented in terrains which may represent exhumed lower crust (e.g., the Giles Complex, central Australia, Etheridge, 1975; the Ivrea Zone, Brodie and Rutter, 1987), whilst Green et al. (1990) document an intensely sheared upper crust in Canada, manifesting similar seismic reflection characteristics to those displayed in the lower continental crust. Reflectivity induced by shear deformation would be expected to be characterised by elongation of reflectors parallel to strike of upper crustal extensional structures, which Reston (1987) demonstrates to be the case for lower crustal reflectivity imaged by SWAT (BIRPS and ECORS, 1986) in the extensional regime offshore of SW England. However, the nucleation of earthquakes from within a zone of seismic layering extending from approximately 15 km depth beneath the Black Forest (Fuchs et al., 1987; Brun et al., 1992) seems to be incompatible with the accommodation of ductile shear, unless the layering is a relic of shearing in a ductile regime (for example prior to uplift). Furthermore, the frequent lack of enhancement of crustal layering in regions subjected to greater crustal extension seems inconsistent with lamination caused by regional extensional strain and associated shearing (Nelson, 1991). Thus, whilst shear deformation of the lower crust may contribute to lower crustal reflective fabric, it is unlikely to provide the primary explanation thereof.

High pressure, ultrasonic pulse measurements of the elastic properties of mineral assemblages from ultramafic complexes in Scotland and Pakistan and single crystals from a carbonatite complex in Norway indicate that dry, amphibole – feldspar aggregates with a quartz content of less than 5% could satisfy both the high v_P/v_S ratios and high impedance contrast (Domnesteanu, 1997) displayed by seismic layering below the Weardale granite. High P-wave velocities from seismic refraction data concentrated mainly on N. America and Europe (Rudnick and Fountain, 1995) also tend to support the hypothesis of a mafic lower crustal composition, though not below E. China, where a bulk intermediate composition is indicated (Gao et al., 1998). Based on data on felsic granulite terrains, mafic xenoliths and a 3000 km long, refraction seismic profile traversing western Europe, Wedepohl (1995) has calculated a felsic to mafic compositional ratio for the lower crust of 0.6:0.4. The analysis also indicates a high bulk crustal content of tonalite, requiring partial melting of mafic rocks and the presence of aqueous fluids, possibly derived from mafic intrusions.

Mafic dykes have been mapped in Precambrian shield regions and many of the brightest, continuous reflections from granulite terrains of inferred mid to lower crustal origin and from the crystalline upper crust have been disclosed by drilling to be mafic sills (e.g., Christensen, 1989; Juhlin, 1990). Of the three most favoured hypotheses for explaining lower crustal reflectivity, the hypothesis that underplating of mafic magmas is the primary cause of seismic layering seems to be the least problematic and at the same time provides a mechanism for cratonic growth (Nelson, 1991). The high density of mafic magma should render its containment at lower crustal depths gravitationally stable (Herzberg et al., 1983), whilst rheological properties and shear strain could promote extensive distribution of the magma within the lower crust, effecting a stratified fabric (Waff, 1980). In the mafic magmatism (underplating) model, association of seismic reflectivity and electrical conductivity could arise from outgassed brines trapped over the underplate (Nelson, 1991).

Lee et al. (1983) and Glover and Vine (1994) have presented laboratory data which appear to suggest that mafic rocks are more conductive than felsic: whereas electrolyte saturated, felsic rocks show peak conductivities at 350 °C, the conductivity of electrolyte saturated, mafic rocks shows a clear increase (and departures from Archie's Law) above 350 °C. Their interpretation involves electronic semiconduction through hydrated amphiboles. This temperature-dependent mechanism appears to become dominant above 350 °C, a temperature with which the depth to the tops of many conductors has been associated. At 450 °C, the electrical conductivity is apparently raised by an order of magnitude (from approximately 0.003 Sm^{-1} for pure electrolytic conduction (with 0.5M NaCl fluid saturation) to approximately 0.03 Sm^{-1}). Duba (pers. comm., 1998) attributes the actual cause of the raised conductivities to the formation of grain boundary magnetite as a result of oxidation reactions arising from failure to control oxygen fugacity, the importance of which is well documented (e.g., Duba and Nicholls, 1973; Duba et al., 1974; Duba, 1976).

Modelling of magnetic data suggests that the lowermost continental crust supports significant volumes of highly magnetic material, with magnetite being a preferred candidate (e.g., Shive et al., 1992; Nolte and Hahn, 1992). Consideration of the stability field of magnetite under lower crustal conditions implies a dominantly mafic lower crust (Shive et al. 1992 and references therein). Interconnected magnetite could also enhance conductivity, though to date magnetite has only been observed in disseminated form in field samples.

4. Petrologic Evidence for a Partially Hydrated, Mafic Lower Crust?

Yardley (1986), Yardley and Valley (1997) and Frost and Bucher (1994) argue that the preservation of anhydrous granulites in exposures of assumed lower crustal provenance (Percival et al., 1992 and references therein) is inconsistent with infiltration of lower crustal rocks with aqueous fluids. They argue that the fluids would be taken up in metamorphic reactions (e.g., pyroxene + quartz + water $>$ amphibole) on a time scale even quicker than that required for them to be flushed out due to mechanical instability. Sanders (1991), drawing on data revealing the sensitivity of metamorphic equilibria reactions to NaCl fluid concentrations compiled by Bowers and Helgeson (1983), hypothesises that NaCl brine may buffer amphibolite to greenschist facies dehydration reactions (the onset of which release substantial amounts of water) to depths corresponding with temperatures in the 350–400 °C range. He advances the hypothesis of ‘self-sealing hydration’, in which marginal hydration reactions form impermeable envelopes around granulite lozenges, arresting further hydration and allowing the fluids to coexist in a network of anastomosing, brine-soaked, ductile shear zones, as a means of reconciling the presence of free fluids with the preservation of anhydrous rocks. However, Frost and Bucher (1994) consider fluid-amphibole equilibria reactions at lower crustal temperatures and pressures and conclude that hydration should cease only when all free fluid has been incorporated. To what extent exhumed granulite terrains can be generalised to represent typical lower crust and what is meant by ‘typical’ lower crust is, however, controversial. Exposures of presumed lower crustal provenance in cratonic regions may not be representative of cratonic lower crust now in situ (e.g., Brodie and Rutter, 1987). Percival et al. (1992) suggest that exhumed granulite terrains may be representative of the mid crust rather than the lower crust, whereas the denser lower crust is less prone to tectonic exhumation and may only be represented in xenoliths. In the case of the Ivrea Zone, a thrust fault cutting the upper section of the unit is not in fact mapped to lower crustal depths on seismograms (Nelson, 1991). During uplift, erstwhile interconnected conductive fabrics may be dislocated, whilst retrograde metamorphism, oxidation and weathering of exposed conductors may further reduce conductivities of composite minerals, such that the electrical properties of sections of uplifted crust are unlikely to represent those of their counterparts at depth.

Contrary to the arguments presented by Yardley (1986), Frost and Bucher (1994) and Yardley and Valley (1997), eclogite (ultramafic) facies metamorphic rocks having significant hydrous phases are preserved along an approximately 2000 km long Uralian suture zone (Dobretsov and Sobolov, 1970) and eclogite facies shear zones, presumed to have been formed in the deep crust have been found to coexist with granulite facies rocks in the Bergen Arcs, Norway (Boundy et al., 1992). The fine-scale isotopic and petrologic heterogeneity of the Bergen Arcs eclogites and their formation adjacent with anhydrous granulites implies locally heterogeneous fluid infiltration and limited fluid mobility, whilst local buffering similar to that suggested by Sanders (1991) is suggested by the appearance of granulite lenses surrounded by anastomosing eclogites (Boundy et al., 1997).

Yardley and Valley (1997) also reject graphite as a cause of lower crustal conductivity on the basis that it would be unlikely to form and maintain extensive, laterally interconnected networks. However, as illustrated by Merzer and Klemperer (1992), in a model initially advanced to explain deep crustal seismic layering and conductivity jointly by fluids, absolute lateral connectivity is not a pre-requisite for imaging by magnetotelluric methods if connection is provided by plates which overlap in the vertical plane, whilst laboratory studies on graphite-bearing rocks reveal restoration of connectivity under pressures representative of lower crustal conditions (Duba et al., 1994).

In contrast to exposed granulite terrains of presumed lower crustal provenance, the diversity of chemical and isotopic compositions (Kay and Kay, 1981), metamorphic grades (Fountain and Salisbury, 1981) and ages (McCulloch et al., 1982) exhibited by lower crustal xenoliths, suggests a highly heterogeneous lower crust, of predominantly mafic composition (Kay and Kay, 1981). The random and incomplete crustal sampling afforded by analysis of xenolith suites may, however, bias compositional models of the lower crust, which may be more felsic than xenolith data alone tend to credit. Based on seismic velocity data from lower crustal, mafic xenoliths from the Chudleigh volcanic province, north Queensland, Rudnick and Jackson (1995) suggest that lower crustal reflectivity probably arises from interlayering of mantle-derived basalts and pre-existing felsic crust. Fountain and Salisbury (1981) explain the laminated seismic nature of the lower crust similarly, with a model incorporating amphibolite facies rocks interleaved with lower velocity, quartz-feldspathic gneisses. Cox (1980) suggested on the basis of the extensive fractionation interred in flood basalts that a similar process to that engendering their formation could produce lower crustal layering. Singh and McKenzie (1993)

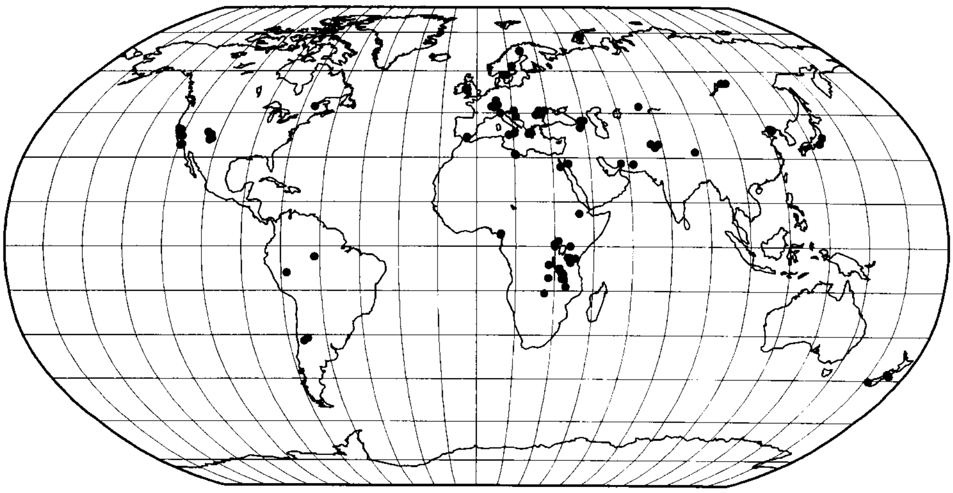

Figure 1. Global map with black dots showing regions of inferred lower crustal seismicity (references in Appendix 1).

demonstrated that the layered, ultramafic, igneous sequence exposed on the island of Rum, Scotland could reproduce seismograms with similar length scales and reflection coefficients to those observed from the lower crust.

5. Lower Crustal Seismicity: Geophysical Evidence for a Strong Mafic Lower Crust?

It is often assumed that crustal seismicity is confined to a brittle, upper crustal layer and that deep crustal earthquakes occur only seldomly in cold shield regions (e.g., Meissner and Strehlau, 1982). Contrary to this wide spread prejudice, perusal of the literature reveals that earthquakes have been suggested to have nucleated from lower crustal depths in over 30 geographical regions, encompassing volcanic (e.g., Ambeh and Fairhead, 1991), rift (e.g., Nyblade and Langston, 1995), basin (e.g., Bryant and Jones, 1992), shield (e.g., Arvidsson and Kulhanetz, 1994) and convergent (e.g., Anderson et al., 1993) tectonic regimes (Figure 1).

The occurrence of lower crustal earthquakes is often interpreted to imply a region of anomalously strong lower crust (e.g., Shudovsky et al., 1987; Cloetingh and Banda, 1992), coupled with an anomalous heat flow regime (e.g., Nyblade and Langston, 1995). Mafic rocks are more resistive to plastic failure than felsic – whereas wet quartz deforms plastically from approximately 275 °C, feldspar deforms by brittle failure to temperatures exceeding 450 °C, and pyroxenes can maintain brittle strength to temperatures exceeding 600 °C (Rutter and Brodie, 1992, and references therein) – and mafic petrology is thus seen as one potential factor in maintaining lower crustal brittle strength (e.g., Shudovsky et al., 1987).

Lower crustal seismicity below the Rukwa, Rhinegraben and Baikal rifts has been associated with deeply penetrating faults in a brittle, lower crust (Brun et al., 1992; Deverchere et al., 1993; Camelbeeck and Iranga, 1996). Armbruster et al. (1978) attribute deep crustal seismicity in the Himalayan region to a deep fault zone, possibly related to the Indus suture. A deep crustal earthquake in North Wales has also been attributed to brittle failure (Ansell et al., 1986).

Interpretation of deep crustal seismicity is ambiguous. At present, there is no way of distinguishing between earthquakes nucleating in a brittle regime or a ductile regime from their seismic signatures. The style of deformation of continental crustal rocks depends on petrology, temperature and strain rate (e.g., Meissner and Strehlau, 1982). Localised strain acceleration can cause fine-grained, quartz-controlled material to behave in a brittle manner above temperatures at which ductile behaviour would normally be expected (Cloetingh and Burov, 1996). Seno and Saito (1994) attribute deep crustal seismicity in Zaire to the existence of high pore pressures and strain rates.

Calculations of yield strength envelopes involve extrapolation of mm scale measurements on rock failure (Byerlee, 1978) and intersection with extrapolated steady-state flow laws, and often incorporate poorly constrained heat production data and geothermal gradients (as emphasised by deep crustal drilling (e.g., Jobmann and Clauser, 1994; Kukkonen and Clauser, 1994)). The KTB borehole was inferred to have penetrated into the brittle-ductile transition zone within quartz-rich gneiss at 265 °C and 9.1 km depth (Dresen et al., 1997) and Byerlee's Law was found to hold (Zoback et al., 1993). However, laboratory measurements by Blanplied et al. (1988) with displacements of up to 400 mm reveal that for displacements above 70 mm, rock friction is effected by the competing processes of velocity strengthening and slip weakening associated with grain size reduction, and extrapolation of laboratory measurements to realistic geological strain rates and timescales is clearly problematic.

A brittle regime may not be a pre-requisite for earthquake nucleation (e.g., Hobbs et al., 1986). Earthquakes in active regions with high heat flow have been ascribed to instabilities associated with magma movements (e.g., Young et al., 1991). The presence of fluids is expected to reduce the yield strength and the temperature at which plastic flow occur (Tullis and Yund, 1980; Kirby, 1984 and references therein). Thus, a dry lower crust is likely to be stronger than a wet lower crust. Compositional heterogeneities, such as garnet inclusions (Rutter and Brodie, 1992) and unstable phase changes (as have been suggested for deep mantle events) may also play a role. Strain hardening, generally involving grain size reduction may cause a return to brittle character (e.g., Rutter and Brodie, 1992 and references therein). In such cases evidence of shearing (as observed in exhumed sections of assumed lower crustal provenance) and intense mineral alignment may be preserved. Whilst rheological properties are generally considered as bulk properties, if seismic layering is caused by interleaving of felsic and mafic material, then differential stresses may arise from differences in yield strengths, and the crust may support more than one brittle-ductile transition zone. The presence of brittle-ductile-brittle transitions have been inferred to depths of 40 km from studies on eclogisation within palaeo-alpine subduction zones (Philippot, 1993). Such rheological layering cannot be resolved using MT methods, because of the inherent lack of resolution of horizontal to vertical anisotropy.

6. Stress, Rheology, Temperature and Electrical Conductivity

The way in which the deep crust deforms has a bearing on the chosen mechanisms for interpreting its electromagnetic properties. Brittle-ductile transition conditions have, for example, been seen as favourable to the preservation of interconnected, saline fluid networks. In contrast, conditions in the brittle, crystalline, upper crust are not considered to favour laterally extensive connectivity of fluids. In the Black Forest, a deep crustal conductor (Tezkan, 1988) has been imaged from 12 km depth (though lateral distortion effects from the highly conductive surface sediments of the Rhinegraben may not be adequately controlled), apparently corresponding to the felsic brittle-ductile transition range (275–325 °C), yet within a deep crustal seismogenic zone. Seismic layering is also imaged in this region from a depth of 14–15 km (Lüschen et al., 1987; Fuchs et al. 1987; Brun et al., 1992). Seismicity persists to at least 20 km (Ebel and Bonjer, 1990; Koch, 1993). Lower crustal reflectors at 28–36 km depth below Lake Vänern, Sweden, a region where earthquake hypocentres have been located at depths of 20–35 km (Arvidsson and Kulhanetz, 1994), have also been reported (Juhlin, 1990). 1D conductivity models of the Lake Vänern region indicate a fairly resistive lower crust with a conductance of less than 50 S (Korja and Hjelt, 1993). In the Baikal region, enhanced conductivities have also been reported from depths of 12–15 km (Berdichevsky et al., 1980), to the south of the deep crustal seismogenic zone (20–30 km) reported by Deverchere et al. (1993).

Whether permeability increases or decreases with deformation appears to bear a complex relationship to rock type (for example porous and coarse-grained rocks are expected to be less brittle) and stress regime. Zoback and Byerlee (1975) observed that the permeability of samples of Westerly granite samples increased by approximately 300% in response to stresses applied in the brittle regime. However, based on laboratory experiments on porous sandstones (porosity 15–35%) with stress states spanning the brittle-ductile transition range, Zhu and Wong (1997) report a several orders of magnitude decrease in permeability 'triggered by the onset of shear enhanced compaction caused by grain crushing and pore collapse' and 'the development of a relatively impermeable shear band' such that 'permeability consistently decreases with increasing strain'. They suggest that interconnected porosity may, ultimately, be eliminated in the ductile regime, producing impermeable rocks close to the percolation threshold at porosities as high as 18%. This counter-balances the argument based on dihedral angles that saline fluids in the ductile regime should form interconnected networks (Watson and Brenan, 1987). It should also be noted that the assumption of grain-boundary wetting in the ductile regime is often based on measurements of the dihedral angle formed between aqueous fluids and quartzite rocks at temperatures more representative of mantle conditions. In contrast to quartz-rich rocks which may, at temperatures of 950–1150 °C, form wetting angles as low as 40° with brines, wetting angles for mafic rocks may be greater than 60°, though this is not conclusive, particularly since in some assemblages dihedral angles may be highly anisotropic (Watson and Brenan, 1987). Furthermore, experiments by Holness (1993) on the effect of temperature on quartz-aqueous fluid wetting angles demonstrate the inappropriateness of extrapolating the Watson and Brenan (1987) results for quartz-aqueous fluid assemblages to temperatures more representative of lower crustal conditions. In the 450–1000 °C temperature range, quartz-aqueous fluid wetting angles are above 60 °. No data are available for temperatures below 450 °C.

Strain energy has been shown to reduce the activation energy required for the formation of graphite. Thus, whilst shearing may reduce permeability and hence fluid connectivity and electrical conductivity where fluids are the primary cause, the same deformation mechanism may increase conductivity if graphitisation occurs. A similar range of temperatures (350 °C–450 °C) to that which, on the basis of often misleading statistical compilations, has been associated with the transition from brittle to ductile rheology is associated with graphitisation of shear zones.

In the brittle regime, fractures rather than dihedral angle dictate fluid distribution, whilst in the ductile regime, shear stress may play a dominant role in controlling connectivity, particularly close to the brittle-ductile transition, where textural equilibrium is unlikely to be reached. In the brittle regime, microcracks do not only evolve parallel to the direction of maximum stress, but also aligned at approximately 30° to maximum compressive stress. Experiments on fractured rock masses reveal a large azimuthal deviation (of the order 35°) between axes of maximum permeability and applied stress directions (Zhang and Sanderson, 1996), wherein pre-existing network geometry strongly influences the development of permeability anisotropy.

Nyblade and Langston (1995) have suggested that brittle character may be preserved to Moho depth if the lower crust is mafic and heat production is dominated to a greater extent by upper crustal sources than is usually assumed. Lateral heterogeneities, for example, may lead to a bias in the allocation of crustal heat sources, thus causing overestimation of lower crustal temperatures (Furlong and Chapman, 1987). The temperatures recorded within both the KTB and Kola boreholes were hotter than predicted (e.g., Burkhardt et al., 1989). Cores from the KTB deep borehole exhibited no depth dependent decay of radiogenic elements – an assumption in the Chapman (1986) model. However, this result must be set against the fact that the crust through which the borehole was drilled has been thrice overstacked tectonically. Another assumption of the Chapman model is that lateral heat flow can be neglected. In the KTB pilot borehole, strong, unexpected, vertical variations in heat flow density were observed, with a sharp increase at 500 m depth raising the average heat flow density from 52 m Wm^{-2} to 83 m Wm^{-2} in the intervals above and below 500 m respectively (Jobmann and Clauser, 1994). Jobmann and Clauser (1994) have shown that lateral refraction of heat due to thermal conductivity contrasts, structural heterogeneities and advection do play a role in determining the temperature profile within the borehole, wherein the presence of fault zones containing graphite in layers up to several cm thick is a significant factor given that the thermal conductivity of graphite is approximately 60 times that of crystalline rock. Large variations in vertical heat flow density were also observed within the Kola borehole, situated on the northern rim of the Baltic Shield. Here, the heat flow density was measured to rise from 30 m Wm^{-2} at 1 km depth to a peak value of 67.5 m Wm^{-2} at 4 $-$5 km, diminishing again to 50 m Wm^{-2} at 5–6 km depth (Kukkonen and Clauser, 1994). Advective heat transport was suggested as the most dominant mechanism giving rise to this depth dependent heat flow variation, with refraction of heat into inclined strata and heat production variations also playing roles (Kukkonen and Clauser, 1994), again drawing into question the assumptions commonly applied when calculating temperature-depth profiles. Heat transport properties can also be expected to be effected by the presence of fluids and by deformation style. If the lower crust has been extensively underplated and is not merely the residue remaining after differentiation of the granitic, lithophile enriched upper crust (Taylor and McLennan, 1985), then the distribution of lower crustal radiogenic elements may be falsely estimated. A knowledge of the geothermal gradient is important for reliably determining yield strength envelopes.

Given the ambiguity in lower crustal rheological properties and in estimating deep crustal temperatures, interpretation of deep crustal electrical conductivities in terms of fluids at depths below the onset of brittle to ductile rheology should not be so extensively invoked in the literature with so little regard to tectonic regime. Scepticism should meet any argument that the lower crust must behave as a ductile layer on the basis of its aseismic character. The lower continental crust may in places be brittle, in places ductile. Fluids may not form extensively interconnected networks in either regime, but may exist as a thin layer close to the percolation threshold (e.g., Zhu and Wong, 1997) within brittle-ductile transition zones (e.g., Etheridge et al., 1983; Jones, 1987).

The deep crust is increasingly reported to exhibit electrical anisotropy (e.g., Cull, 1985; Rasmussen, 1988; Kellet et al., 1992). If the lower crust does not always form a rheologically weak layer, but in some environments can sustain brittle stress, then this may help to reconcile the maintenance of electrical anisotropy over geological time scales, such that both present day and palaeo-stress fields can control preferred directions of anisotropy. For example, based on a compilation of preferred directions of electrical anisotropy calculated from MT data, Bahr (1998) observes that maximum conductance is generally aligned perpendicular to maximum extant stress, though exceptions occur in which the alignment may relate to a palaeostress field. One such exception is within the Baltic Shield, where based on lower crustal seismicity the entire crust may be brittle. If this is so, a 'frozen in' palaeostress field may be easier to reconcile. At least two different azimuths of electrical anisotropy are preserved in the lower crust/mantle of the Canadian Shield (Kellet et al., 1992 and references therein).

The deepest boreholes in the world – KTB and Kola – are 8.9 and 12.3 km respectively – less than one third of the crustal thickness (approximately 35 km (Franke, 1988) and 40 km (Kukkonen and Clauser, 1994)) in their respective locations. The direction of crack alignment within the Kola borehole, where fractures have been logged at all depths, does not correspond with the direction of the present day stress field, but may be consistent with the direction of palaeostress. Within the KTB deep borehole, the principal horizontal stress was found to be uniform (N160 $\pm$ 10E) and consistent with the average orientation identified throughout western Europe to a depth of 7.2 km, where the borehole intersects a major fault and the principal stress axis reorientates by 60° (Brudy et al., 1997), indicating the preservation of a palaeostress direction. The direction of upper crustal electrical anisotropy (Eisel and Haak, 1999) in the vicinity of the KTB site also seems to mimic the direction of palaeostress from 300 Ma (Zulauf, 1992), and may relate to graphite formed in the mid crust, but now uplifted and preserved in relic shear zones (Haak et al., 1997).

Rutter and Brodie (1992, and references therein) report modification of the rheological properties of continental crustal rocks via strain hardening and shearing processes in which ductile flow and dynamic recrystallisation (generally with reduced grain size) occurs in continental crustal rocks over a wide range of pressure and temperature environments and often produces intense alignment of constituent minerals along shear axes. As such the lower crust may preserve markers of its complex tectonic history.

7. Conclusions

A better understanding of lower crustal deformation properties may help in the interpretation of electrical anisotropy and in understanding electrical conductivity mechanisms in general. Deep crustal rheology will effect how deep crustal fluids distribute and the presence or absence of deep crustal fluids will have an effect on deep crustal rheology. If the lower crust can be brittle, this may explain how both stress and palaeostress fields or directions of plate motion can be 'frozen in' in the form of preferred directions of electrical anisotropy. All rocks in situ exist under some stress field which is likely to impose non-uniformities, and the overall relationship between rheology, stress field, timescale, connectivity and electrical anisotropy in the deep crust is likely to be a complex and most likely region dependent one, which simplified interpretations involving brittle-ductile rheology in the 350–450 °C temperature range controlling enhanced conductivity due to fluids fail to take proper account of.

Explanation of deep crustal earthquakes may require a strong, mafic lower crust in some regions. Other geophysical observations – e.g., seismic layering and magnetic permeability also support the idea of a mafic lower crust. Eclogisation of mafic lower crust could have significance for recycling of lower continental crust into the mantle by delamination (Austrheim, 1987) and could explain the unexpected chemical signatures of some oceanic and flood basalts (Arndt and Goldstein, 1989).

Interpretation of deep crustal seismicity is ambiguous but shows that a ubiquitous brittle-ductile transition, occurring at temperatures of 300 to 450 °C can not be taken for granted. Furthermore, the transition from brittle to ductile rheology in fact occurs over a much broader and less well constrained range of temperatures than generally credited in electromagnetic literature. Under realistic geological strain rates, the temperatures at which different crustal rock types undergo transition from brittle to ductile rheology may vary from approximately 275 °C for wet quartz to more than 600 °C for wet pyroxene, with a broad intermediate range of temperatures applying to feldspars (e.g., Rutter and Brodie, 1992 and references therein) – a rather broad temperature span in which to fit some kind of deep crustal correlation. The brittle-ductile transition zone as a trap to lower crustal fluids is a concept which may not 'hold water' in many regimes, whilst other electrical conductivity mechanisms, notably graphitisation along cleavage planes aided by ductile shearing (which significantly lowers the thermal activation energy (Ross and Bustin, 1990)) may also become effective above 300–450 °C isotherms. Shear stresses, which may have a more dominant role to play than dihedral angle at lower crustal temperatures, may reduce the connectivity of fluids.

In interpreting electromagnetic models, it is important to remember that it is not only these models which are non-unique, but that other methods and associated assumptions have their inherent uncertainties and ambiguities too. Prime examples are temperature gradients leading to temperature-depth profiles, and brittle-ductile transition temperatures. Deep crustal drilling at both Kola and KTB have revealed just how wrong predicted temperature-depth profiles can be. 'Statistical' compilations relating electrical conductivity to other parameters are likely to be prejudiced by subjective views and may be counter-productive to our science. Science generally advances by disproving hypotheses and we should design experiments to illuminate the exceptions to statistical norms. Whilst integrated studies are a worthwhile objective, joint explanation of seismic, thermal and electrical properties should not necessarily be seen as a universal goal. The likely causes of enhanced conductivity should be considered in an environment specific and not global sense.

The lower crust is generally considered to be an aseismic, weak zone where (based on data which is likely to be inappropriate to lower crustal conditions) fluids might be governed by textural equilibrium geometries. Counter-evidence has been presented that the lower crust in some regions may be both dry and strong, that fluids if at all present at such depths may not be connected, and that fluid mobility in the lower crust may be more limited and heterogeneous than commonly attributed.

It is emphasised that the counter-idea that sections of the lower crust may be dry, mafic and brittle is in no way advanced as a global model. The lower continental crust is likely to be petrologically, physically and mechanically heterogeneous on a variety of scales.

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Appendix 1: References to Figure 1

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