

Electrical conductivity of the continental lower crust

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1. Introduction

The continental middle to lower crust (CLC) remains one of the most enigmatic parts of the Earth about which comparatively little is known. Remote-sensing geophysical and geochemical data illustrate our gaps in complete knowledge of the state and composition of the CLC. Generally, it is thought that the CLC, when compared to the upper crust, is somewhat more uniform in its properties. However, studies undertaken in the last decade have revealed that the CLC can be as heterogeneous as the upper crust, but that some regions of surprising homogeneity exist. In this chapter, I will describe attempts to image one particular physical parameter of the CLC, namely its electrical conductivity σ .

Electromagnetic (EM) sounding methods, as with most geophysical methods, are sensitive to the structure that exists *today*, and as such are dependent on the *current* state and composition of the CLC, rather than its state at formation or as a consequence of tectonic or metamorphic activity as are petrological studies on exhumed samples from the CLC. All EM methods give responses that are volumetric averages of the Earth's conductivities sensed by the diffusive EM fields, and as such are in the same class of geophysical techniques as seismic surface wave studies, potential field methods, and geothermal investigations. These are distinct from seismic reflection and refraction methods which deal with non-diffusive waves. However, with EM methods the governing field equations (§2) and wide range of the physical parameter being sensed (electrical conductivity, §1.1) ensure a far greater resolving power to anomalies than with the other diffusive geophysical techniques. EM and seismic methods are the only geophysical techniques for which probing of the deep crust is assured; with all others there are inherent screening effects and depth ambiguities.

Imaging the electrical conductivity structure beneath the surface has a number of uses, classified mainly by the depth of investigation:

- (1) Identifying zones of mineralization, which are of economic importance.
- (2) Detecting fluids in the deep crust, such as those above downgoing slabs, e.g., the Juan de Fuca plate (Kurtz et al., 1986a, 1990; EMSLAB, 1989).
- (3) Resolving one physical property of crustal zones. As expressed by Dohr et al. (1989): "... magnetotellurics works like a broad paint brush, colouring the layers bounded by the seismic reflections".

(4) Determining the structure of the mantle, in particular the depth to the “electrical asthenosphere”, which is a zone of enhanced conductivity in the upper mantle below the lithosphere (e.g., Jones, 1982). This depth often correlates with the depth to a seismic low velocity zone.

Of these uses, (2) and (3) are relevant to the electrical conductivity of the CLC. Haak and Hutton (1986) have reviewed this topic, and I intend this chapter to complement their work by adding new results and interpretations, only repeating certain of their points for completeness. After a discussion of the physical parameter that EM methods are sensing, I will review EM methods appropriate for determining the conductivity of the CLC and their problems and limitations. A summary of recent (1980s) EM results pertinent to the topic follows with specific details from two geographic areas: the Kapuskasing structure in northern Ontario and the Valhalla complex in southeastern British Columbia. Finally, I examine the proposed causes of the observed enhanced electrical conductivity of the CLC focussing on the two currently most popular: saline fluids and grain-boundary films of carbon.

1.1. The parameter being sensed: electrical conductivity (σ)

The physical parameter that is being imaged by EM methods is electrical conductivity, σ , measured in Siemens per metre (S/m). The effects of magnetic permeability variations (e.g., Kao and Orr, 1982) are not considered as they are of little consequence for the CLC. For historical reasons it is more common to discuss the reciprocal of conductivity, which is electrical resistivity, ρ . Usually, as will be described in §2.4.3, it is not the conductivity or thickness of a zone of enhanced conductivity that is resolved from the observations, but the product of these two. This conductivity-thickness product, $S = \sigma h$ where h is the thickness of the layer, is called the *conductance* and is measured in units of Siemens (S). Its complementary product, which is usually poorly resolved, is the resistivity-thickness of the zone, $T = \rho h$, or *resistance* in units of ohms (Ω).

Electrical conductivity is sensitive to very small changes in minor constituents of the rock, and hence is a complementary parameter to other geophysical techniques, such as potential field methods and most seismic methods, which are sensitive to the bulk properties of the medium. As the conductivity of most rock matrices is very low, the conductivity of a rock unit is generally a function of the interconnection of a minor constituent, such as fluids, partial melts, or highly conducting minerals like graphite, that provide ionic or electronic pathways. Rarely is the rock so competent that the resistivity of the rock matrix is measured, although such highly resistive blocks have been found in the upper crust (e.g., Bailey et al., 1989; Kurtz et al., 1989; Beamish, 1990).

The well known Archie's Law (Archie, 1942), which was initially proposed to describe the conductivity of saturated sediments, is often an appropriate first-order model for the total conductivity of a medium:

$$\sigma_m = \sigma_f \eta^m \quad (1)$$

where σ_m and σ_f are the conductivities of the bulk medium and of the fluid respectively, η is the porosity, and the exponent m has a value between 1 and 2, with 2 being shown empirically to be valid for a wide range of rocks to mid-crustal depths (Brace et al., 1965; Brace and Orange, 1968). Note that the rock matrix conductivity, σ_r , is assumed to be sufficiently high to be of little effect (see §4.6 for a more complete discussion of pore geometry considerations). The electrical resistivity of zones within the Earth's crust can vary by over eight orders of

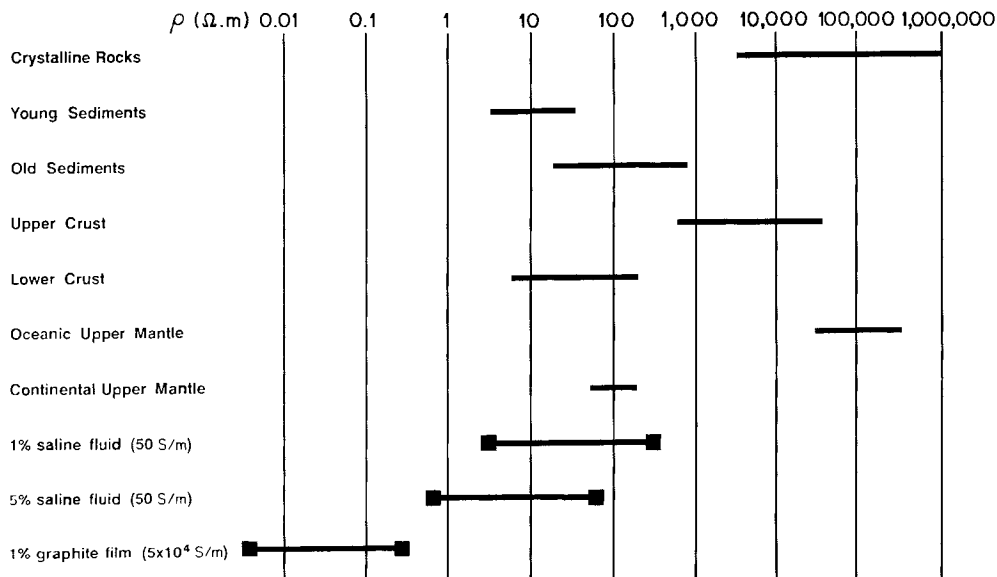


Fig. 3-1. Range of electrical resistivity of earth materials: dry crystalline rocks; “young” porous brine-saturated sediments; “old” less porous sediments; typical upper crust; typical lower crusts; oceanic upper mantle resistivity; continental upper mantle resistivity; fluid-saturated resistive rock of 1% porosity (fluid of 50 S/m conductivity): the upper and lower bounds are for Archie’s Law exponents of 2 and 1 respectively; fluid-saturated resistive rock of 5% porosity; thin graphite film (graphite of 5×10^4 S/m conductivity) (modified from Haak and Hutton, 1986).

magnitude (Fig. 3-1, redrawn from Haak and Hutton, 1986), which is the widest range of any of the physical parameters that can be remotely sensed from the surface of the Earth.

2. Review of methods

Imaging the electrical conductivity structure at CLC depths requires either natural-source or deep-probing controlled-source EM techniques. I will review these briefly; for more complete discussions see Berdichevsky and Zhdanov (1984), Vozoff (1986), Nabighian (1987, 1991), and Keller (1989a).

Natural-source EM methods have the advantages that:

- (1) penetration to all depths is assured by the skin depth phenomenon;
- (2) no transmitter is required; and
- (3) the mathematics of the assumed uniform source-field are relatively tractable, with the consequence that interpretation techniques are more advanced than controlled-source methods.

In comparison, controlled-source EM methods have the advantages that:

- (1) the source is controllable in space, so it can be configured to excite optimally the structure of interest;
- (2) the source is controllable in time (or frequency) and is thus repeatable thereby enabling signal enhancement techniques; and
- (3) in the transition zone between the near field (where the source geometry dominates the responses) and far field (where the fields approach those of a uniform source) there is greater resolving power.

Obviously, given the advantages and disadvantages of both techniques one should not be exclusive, but instead choose the method, or combination of methods, most appropriate and with the highest chance of success for solving the particular problem at hand. However, for probing to great depths within the continental crust, logistical considerations usually exclude controlled-source methods.

The governing field equations can be written in terms of potential functions. However, in comparison to gravity, magnetics and geothermal formulations, the potential functions are *vector* not scalar potentials; this enables the construction of model uniqueness theorems (§2.4.3), whereas scalar potential methods are inherently non-unique. The skin depth property of EM fields (Eq. 5 below) ensures penetration to all depths, although the resolution is strongly affected by the conductance of the material overlying the depth of interest.

2.1. Magnetotelluric method

The most commonly used technique for determining the conductivity distribution within the CLC is the magnetotelluric (MT) method. MT is a natural-source technique utilizing the time-varying electromagnetic fields due to electric storms (for frequencies above 8 Hz) and solar activity (for periods longer than 0.125 s). A good collation of fifty-five papers on the MT method, detailing its historical development and the state-of-the-art up to the early 1980s, is given in the Society of Exploration Geophysicists book edited by Vozoff (1986).

The external magnetic fields penetrate into the ground and induce electric (also called telluric) fields and secondary magnetic fields, and components of the total electric and magnetic fields are measured on the Earth's surface. In the case of the magnetic field, H , all three components are measured. These are h_x , h_y , and h_z , where x usually denotes either north (geographic or geomagnetic) or along strike (for a 2D body), y denotes either east or perpendicular to strike, and z denotes vertically downwards. For the telluric field, E , only the two horizontal components, E_x and E_y , are measured. The fields are measured in the time domain and are transformed into the frequency domain where cross-spectra are computed, and from these spectra the MT response function estimates are derived. The amplitude and phase relationship at a particular frequency, ω , between $E(\omega)$ and $H(\omega)$ are indicative of the conductivity distribution below, and the horizontal field components are related by the complex MT impedance tensor:

$$\begin{bmatrix} E_x \\ E_y \end{bmatrix} = \begin{bmatrix} Z_{xx} & Z_{xy} \\ Z_{yx} & Z_{yy} \end{bmatrix} \begin{bmatrix} H_x \\ H_y \end{bmatrix} \quad (2)$$

(dependence on ω assumed). The principal impedances Z_{xy} and Z_{yx} are converted to apparent resistivities (ρ_a) and phases (ϕ) using:

$$\rho_{a,xy}(\omega) = (1/\omega\mu)|Z_{xy}(\omega)|^2 \quad (3)$$

and

$$\phi_{xy}(\omega) = \tan^{-1}[\text{Imag } Z_{xy}(\omega)/\text{Real } Z_{xy}(\omega)] \quad (4)$$

[similarly for $\rho_{a,yx}(\omega)$ and $\phi_{yx}(\omega)$], where μ is the permeability (usually taken as the free space value, $\mu_0 = 4\pi \times 10^{-7}$ H/m). The MT apparent resistivity represents the scaled total impedance of the layers and zones of varying conductivity in the Earth and may be modelled by equivalent series and parallel electrical circuits. Apparent resistivity and phase are not

independent quantities, but are related to each other: it can be shown that for responses from all one-dimensional (1D) Earth models, and almost all two-dimensional (2D) Earth models, the two form a Hilbert transform pair (Weidelt, 1972; Fischer and Schnegg, 1980). However, because of the unknown constant in the Hilbert transform integral, although the phase curve may be predicted from the apparent resistivity curve, only the *shape* of the apparent resistivity curve can be predicted from the phase curve; there is an unknown multiplicative scaling constant on the apparent resistivities. For responses from three-dimensional (3D) Earth models the relationship has yet to be proven analytically, but empirically the two appear to obey Hilbert transformation.

For a uniform Earth, the apparent resistivities are the true resistivity of the half-space, and the phases are 45° (actually the ϕ_{yx} phases would be -135° , i.e., $Z_{yx} = -Z_{xy}$, because of the sign convention adopted). For regions of high electrical conductivity there is little induced E -field, and so the apparent resistivities are reduced and the phases are high (close to 90° for ϕ_{xy} , -90° for ϕ_{yx}). Conversely, for regions of low conductivity the apparent resistivities are high and the phases are low (close to 0° or -180°). In contrast to DC resistivity, penetration to all depths is assured because of the skin depth phenomenon of EM fields. The skin depth, $\delta(T)$, in metres, given by:

$$\delta(T) = [2\rho_a(T)/\omega\mu]^{1/2} \approx 503 [\rho_a(T)T]^{1/2} \quad (5)$$

where T is the period in seconds ($T = 2\pi/\omega$), can be a reasonable measure of the depth of investigation of the MT method. Spies (1989) considers that the *maximum* depth of investigation is given by $1.5\delta(T)$, whereas for Niblett-Bostick transform the depth is given by $(1/2^{1/2})\delta(T)$ (Niblett and Sayn-Wittgenstein, 1960; Bostick, 1977; Jones, 1983a). For Schmucker's ρ^*-z^* transform (Schmucker, 1970), z^* is $(1/2)\delta(T)$ and is thought to be the "centre of gravity of the in-phase induced current system" by Weidelt (1972).

In a one-dimensional (1D) environment when the conductivity varies with depth alone, $\sigma(z)$, the MT impedance tensor reduces to the simple form:

$$\mathbf{Z}_{1D} = \begin{bmatrix} 0 & Z_{xy} \\ -Z_{xy} & 0 \end{bmatrix} \quad (6)$$

The negative sign for the lower off-diagonal element indicates that the sign convention is such that the phases of this element are in the third rather than the first quadrant. Such a 1D description of the conductivity distribution may be valid at sufficiently short periods for sites over a sedimentary basin, or at long periods when the EM fields are sampling the continental upper mantle. The MT method has advanced to the stage at which 1D MT data are well understood and conductivity-depth profiles can be determined using a variety of inversion methods (e.g., Parker, 1980; Parker and Whaler, 1981; Constable et al., 1987; Oldenburg, 1992), each of which has a fundamentally different philosophy. MT apparent resistivity and phase curves for two three-layered 1D earth models are illustrated in Figure 3-2. Note that the phase curve for the model with a monotonically decreasing resistivity-depth function (solid line) is always above 45° , and conversely for the monotonically increasing resistivity-depth model (hatched line).

In a two-dimensional (2D) model of the Earth, when the axes of the co-ordinate frame are aligned parallel (x) and perpendicular (y) to strike, the impedance tensor becomes:

$$\mathbf{Z}_{2D} = \begin{bmatrix} 0 & Z_{xy} \\ Z_{yx} & 0 \end{bmatrix} \quad (7)$$

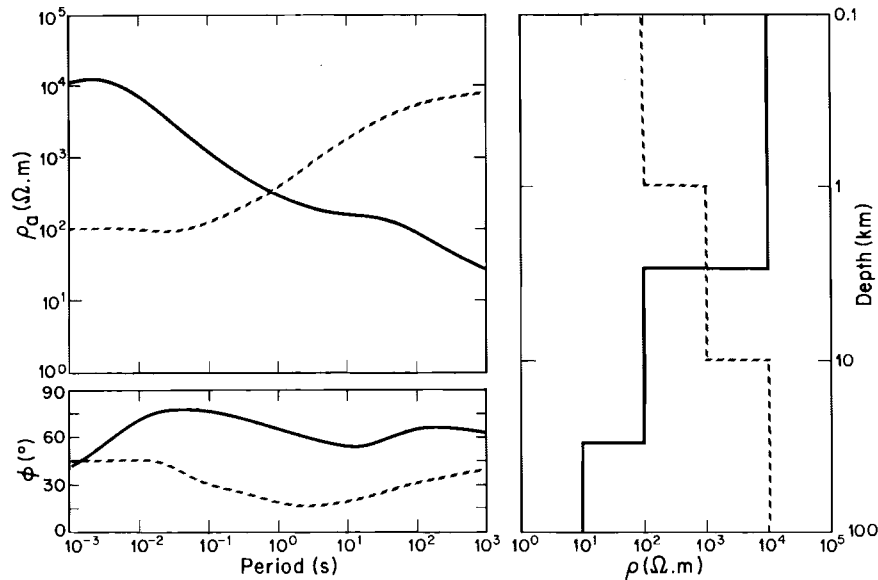


Fig. 3-2. One-dimensional apparent resistivity (top left) and phase (bottom left) curves for the two 1D three-layer models illustrated (right).

One approach for handling 2D data is to find a suitable approximate 1D response that may be inverted to illustrate, perhaps in only a qualitative sense, the structure beneath the recording site (e.g., Jones and Hutton, 1979a). However, it is becoming common to undertake more thorough trial-and-error forward 2D modelling of the data (e.g., Kurtz et al., 1986a, 1990; Wannamaker et al., 1989b; Gupta and Jones, 1990; Jones and Craven, 1990), and recent advances in 2D inversion methods are very promising (deGroot-Hedlin and Constable, 1990; Smith and Booker, 1991; Oldenburg, 1992).

In 2D, Maxwell's equations separate into two modes: the *TE-mode* (transverse electric) describes the field components (E_x , H_y and H_z) observed when the currents are flowing *along* (parallel to) the structure, and the *TM-mode* (transverse magnetic) relates the field components (H_x , E_y and E_z) when the currents are *crossing* (perpendicular to) the structure. The TE-mode is also often called *E-polarization* or *E-parallel*, and the TM-mode *B-polarization* or *E-perpendicular*. An instructive 2D body to consider is the fault model with two infinite quarter spaces of differing resistivity juxtaposed (Fig. 3-3). The TE-mode responses (solid lines) are continuous with lateral distance with the apparent resistivities at a given frequency going from ρ_1 to ρ_2 smoothly. In contrast, because physics requires continuity of electric current J , the electric field is discontinuous across the boundary (J is given by $J = \sigma E$, so if σ is discontinuous then so must be E to ensure continuity of J). Accordingly, the TM-mode apparent resistivity curve (dashed line) is discontinuous; this leads to a fundamentally higher resolution for the TM-mode responses to lateral variation in conductivity than for the TE ones.

In a fully three-dimensional (3D) Earth model, where Z takes on the general form as expressed in Eq. 2 above, MT data are not as well understood. Due to prohibitively high computational costs, full 3D modelling is usually undertaken only for representative structures, not to model actual field data. Approximate 3D solutions, such as "thin sheets", are computationally much faster and are adequate for modelling continent-ocean boundaries where $0.3 \Omega\cdot\text{m}$ sea-water is juxtaposed against 10^3 – $10^4 \Omega\cdot\text{m}$ land (e.g., Weaver, 1982).

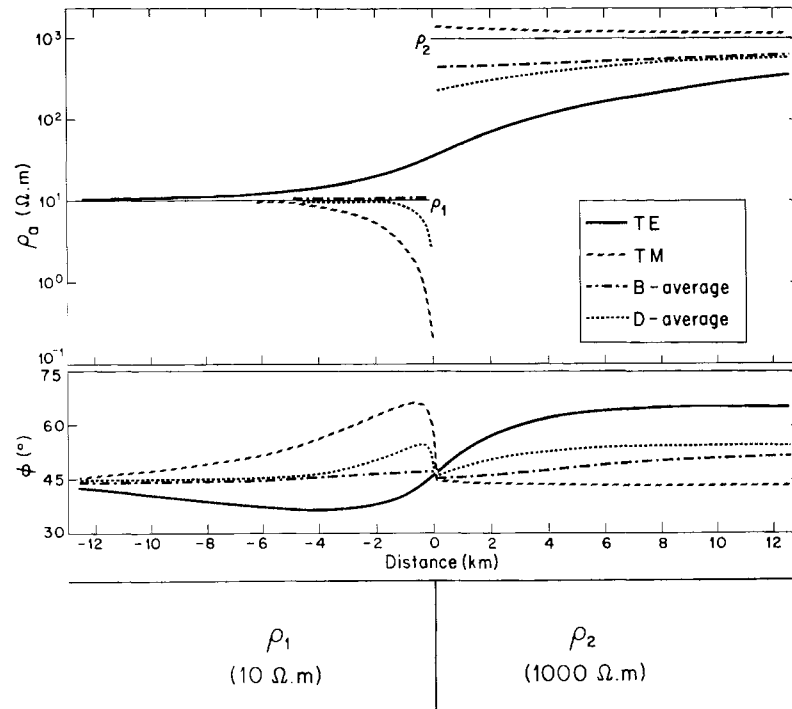


Fig. 3-3. Apparent resistivity (top) and phase (middle) curves for the fault model illustrated (bottom) at 1 s. The TE-mode responses (solid curves) are continuous across the structure, whereas the TM-mode (dashed curves) apparent resistivity curve is discontinuous. Also shown are the Berdichevsky averages (B-average; dashed-dot curves) and the determinant averages (D-average; dotted curves). Note that on the conductive side of the discontinuity the Berdichevsky average gives virtually the correct resistivity (ρ_1) and phase (45°) for valid 1D interpretation right up to the contact. On the resistive side a 1D interpretation of the TM data will yield the most correct model.

Advances have been made to reduce the dimensionality of the data, from 3D to 2D or even 1D, and still yield a valid model. In certain cases, the TM-mode responses from a 3D situation can be interpreted in a 2D manner to give a reasonably correct conductivity distribution (Jones, 1983b; Wannamaker et al., 1984).

The use of two invariant forms of the impedance estimates is becoming more prevalent in MT studies. These two forms, called the “Berdichevsky average”, Z_B , and the “determinant average”, Z_D , are defined by:

$$Z_B = (Z_{xy} - Z_{yx})/2 \quad (8)$$

and

$$Z_D = (Z_{xx}Z_{yy} - Z_{xy}Z_{yx})^{1/2} \quad (9)$$

(Berdichevsky and Dmitriev, 1976). The negative sign in Z_B is again an expression of the coordinate convention used with the phase of Z_{yx} in the third quadrant. The Berdichevsky and determinant averages for the fault model are shown on Figure 3-3. Various authors have illustrated that under certain conditions, even with 3D data, a 1D interpretation of Z_B or Z_D can give a reasonable indication of the conductivity distribution beneath the recording site (Ranganayaki, 1984; Ingham, 1988; Park and Livelybrooks, 1989). Note that for 1D and 2D

Earth models Z_B and Z_D are the arithmetic and geometric means respectively of Z_{xy} and Z_{yx} . The determinant phase, ϕ_D , has the particular appeal of being unaffected by galvanic static shift distortions of the MT impedance tensor (see §2.4.1 for a description of static shift and its implications for MT interpretations).

2.2. Geomagnetic Depth Sounding (GDS) and Horizontal Spatial Gradient (HSG)

Two methods are used for sensing the Earth's conductivity structure in the CLC by recording the magnetic field components alone. In the Geomagnetic Depth Sounding (GDS) method, the two transfer response functions T_x and T_y which relate the vertical magnetic field component to the two horizontal magnetic field components:

$$H_z = [T_x \quad T_y] \begin{bmatrix} H_x \\ H_y \end{bmatrix} \quad (10)$$

(dependence on frequency assumed) are interpreted. They are often displayed as "induction arrows" (Schmucker, 1970), and the real arrows generally point towards zones of enhanced conductivity (Jones, 1986). This method is excellent as a mapping tool for locating anomalous structures, and array studies by Gough and others have mapped continental-scale anomalies in many parts of the world (Gough, 1981, 1989). A full description of the analysis and interpretation techniques used for magnetometer array studies can be found in Gough and Ingham (1983).

The Horizontal Spatial Gradient (HSG) method differs from both MT and GDS in that the external source fields are assumed to be non-uniform. Over large-scale laterally homogeneous regions, the spatial gradients of the horizontal components of the magnetic field induce a vertical magnetic component, and the ratio of the two can be interpreted in terms of conductivity-depth structure (Schmucker, 1970; Kuckes, 1973a,b; Jones, 1980). This ratio, called Schmucker's C -function (Schmucker, 1970), is given by:

$$C(\omega, k) = H_z(\omega) / \left[\frac{\partial}{\partial x} H_x(\omega) + \frac{\partial}{\partial y} H_y(\omega) \right] \quad (11)$$

where k is the wavenumber representative of the non-uniform source field, and $C(\omega, k)$ is related to the MT impedance for a layered earth model $Z_{1D}(\omega, k)$ by:

$$C(\omega, k) = (1/i\omega\mu)Z_{1D}(\omega, k) \quad (12)$$

Thus, $C(\omega, k)$ can be interpreted using the same advanced techniques available for $Z_{1D}(\omega)$. One advantage of the HSG method over MT is that, because only magnetic fields are measured, galvanic static shift problems associated with the telluric fields (§2.4.1) are avoided. The HSG method is not more routinely used because of the requirements for: (a) a large laterally homogeneous region; (b) a sufficient gradient in the source fields; and (c) synoptic observations of the magnetic fields at at least five locations and preferably many more.

2.3. Controlled-source EM methods

Very few controlled-source techniques are capable of resolving structure at CLC depths due to the requirement for large source energy and large source-receiver separations. Ward (1983), Nabighian (1987), Keller (1989a), and recently Boerner (1992), have reviewed the methods and results obtained from controlled-source EM surveys. A number of large-scale

experiments have been carried out using devices developed for other purposes, such as power transmission lines (van Zijl, 1969; Blohm et al., 1977; Lienert, 1979; Lienert and Bennet, 1977; Towle, 1980), decommissioned telephone lines (Constable et al., 1984) and the Kola peninsula magneto-hydrodynamic (MHD) generator (Velikhov et al., 1986). In these experiments, EM fields penetrated and sampled the CLC. Conventional deep-probing controlled-source techniques, e.g., LOTEM (Long Offset Transient ElectroMagnetic system: Keller et al., 1984; Strack, 1984), CSAMT (Controlled Source Audio MagnetoTellurics: Boerner et al., 1990) and UTEM (University of Toronto ElectroMagnetic system: Kurtz et al., 1989; Bailey et al., 1989), usually do not penetrate into the CLC, although the two UTEM studies are important because they sample an upthrust lower crustal block (see §3.5). However, using MEGASOURCE (Keller et al., 1984), a very high-current electric bipole source, Keller, Skokan and colleagues believe that they are able to map the CLC and the Moho (Skokan, 1990). Boerner (1992) discusses problems and limitations of controlled-source methods for deep probing of the crust.

2.4. Problems and limitations of the MT method

The periods at which MT data sample CLC depths are usually in the range 10–100 s (frequencies of 0.1–0.01 Hz). Over highly resistive upper crust, periodicities of 1 s and above may be important, and over thick sedimentary basins periodicities of 30 s or greater may be required. At these periods, the sources are mainly due to solar plasma ejected by the Sun that enters the Earth's magnetosphere and induces magnetic fields in the Earth's ionosphere. Usually, there are excellent signal levels at periods greater than 10 s. Between 0.1 and 10 s there is a minimum in the telluric field spectrum and a maximum in the natural noise (due to microseismic activity, mainly wind noise but also ocean effects) and cultural noise spectra, with resultant low signal-to-noise ratios. This is the so-called MT “dead-band” and until the late 1970s it was difficult to obtain good quality MT response estimates in this band. During the 1980s there have been many significant advances in the instrumentation used to acquire MT data, in the acquisition procedures used (e.g., remote-reference acquisition; Gamble et al., 1979), and in the processing methods employed to estimate the MT impedance elements from the time series (see Jones et al., 1989) to the extent that *modern* MT data have errors of only a few percent over the whole period range of observation. This is compared to the typical one quarter of an order of magnitude errors for data collected and processed during the 1970s. Precise data, which are required if one wants to resolve certain features (e.g., Cavaliere and Jones, 1984), demand highly sophisticated modelling methods for their interpretation.

Equally important is that the data are now wide-band: a modern MT system typically covers six orders of magnitude in frequency (10^3 Hz– 10^3 s) compared to two or three orders of magnitude previously (10 – 10^3 s). Although the data in the range 10–100 s are the most sensitive to the CLC, high frequency data are important for determining the upper crustal structure in order to understand its effects on the longer period data.

Perhaps the most pressing current problem is the effects that local near-surface inhomogeneities have on the MT impedance estimates. These effects have been termed “static shifts” but should more generally be referred to as “static distortions”. It is necessary to correct the MT responses for these distortions prior to any attempt to derive a conductivity model. Intimately coupled with determination of the static distortions is determination of the appropriate strike angle of the coordinate frame into which to rotate the data for 2D modelling. Finally, MT data are limited in their resolving power, and this is discussed with particular reference to the CLC below.

2.4.1. Static distortions

Advances are being made in understanding the effects that near-surface inhomogeneities have on MT responses. A spectacular example of this phenomenon is illustrated by Poll et al. (1987) in which a decrease of more than an order of magnitude in the electric field amplitude perpendicular to a fault was observed over 50 m. This is a problem which expresses our current difficulty in dealing with the very wide range of scales in a typical crustal-sounding MT experiment, from the metres-scale of the locations of the telluric electrodes themselves to the hundreds of kilometres-scale of regional tectonic features. To address this problem approximate techniques have been developed which deal with local features as if the EM fields are at the galvanic limit. The only observable EM effects in the data of the inhomogeneities are due to electric charges bound to the surfaces of them. The effects of these charges are predominantly on the electric field components, but at sufficiently high frequencies the magnetic field components are also affected (Zhang et al., 1987; Groom, 1988; Groom and Bailey, 1991).

In its simplest form for data from either 1D or 2D (in the correct strike direction) Earth models, this problem manifests itself as a shift of the apparent resistivities by a frequency independent multiplicative constant without affecting the phases, and is termed "static shift" (e.g., Jones, 1988a; Sternberg et al., 1988). An example is given in Figure 3-4 for data from two MT sites on a thick sedimentary basin that were separated by only 50 m. Three of the four ρ_a curves are at the same level, and all four phases are similar, but one of the ρ_a curves is shifted downwards by local conditions. A 1D interpretation of the shifted ρ_{yx} curve (dashed line, Fig. 3-4) would lead to an erroneous model with the interfaces closer to the surface by a factor that is the square root of the multiplicative shift constant, and layer resistivities too small by a factor equal to the multiplicative shift constant.

Interpretations of MT data possibly affected by static shifts can be found in the published literature. For example, Figure 3-5 is an interpretation of MT data from a profile of four sites

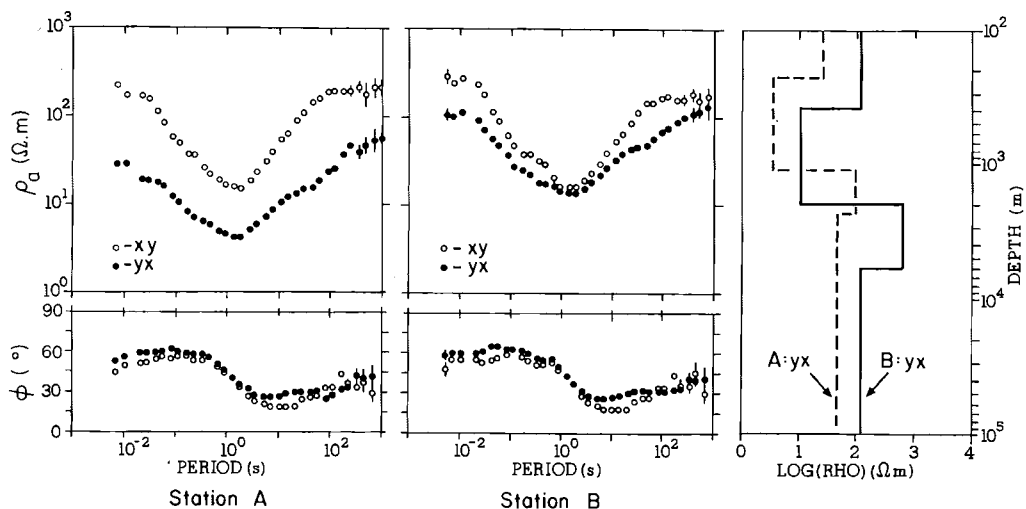


Fig. 3-4. Apparent resistivity and phase curves from two MT sites that shared a common electrode, i.e., were centred 50 m apart. Note that one ρ_a curve (ρ_{yx} at station A) is static shifted downwards by half an order of magnitude compared to the other three, but that all four phase curves are similar. The 1D model of the static shifted data (A: yx ; dashed line) is in error compared to the correct model (B: yx ; full line), with layer interfaces and resistivities underestimated.

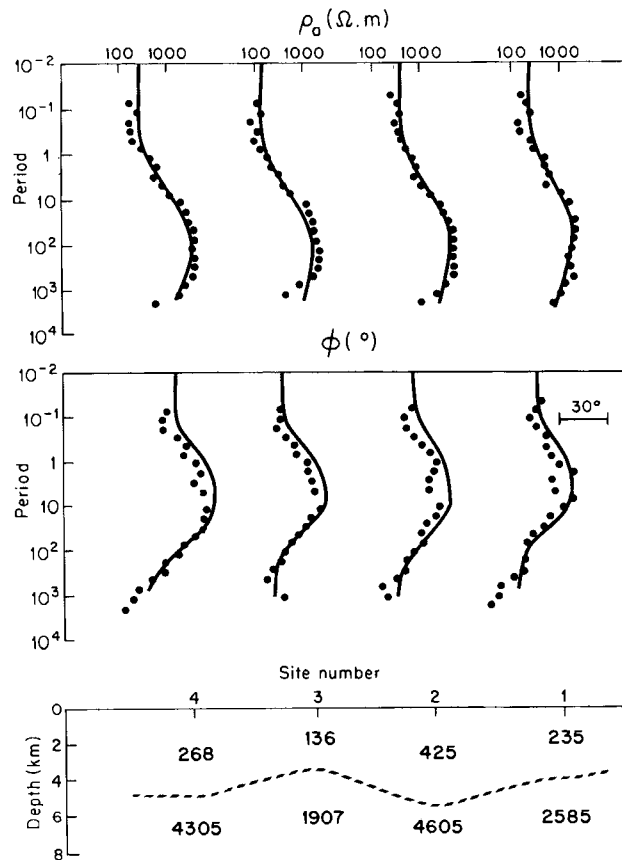


Fig. 3-5. Apparent resistivity and phase curves, and their 1D inversions, for four locations in northern Australia (redrawn from Cull, 1982). Note that the four phase curves are virtually identical, and that the four apparent resistivity curves have the same shape but are at different base levels. Static shift of the resistivity curves is an alternative explanation for the implied lateral variation in basement topography.

in the McArthur Basin of northern Australia (Cull, 1982). Note that the resistivity curves are similar in shape, and the phase curves are virtually equal to one another, but that the levels of the resistivity curves are shifted by a factor of nearly 5. The pseudo 2D model, obtained by stitching together the 1D models from each site, shows purported basement topography beneath site 3. However, the sediment and basement resistivities beneath this site are both smaller by a factor of approximately 2 than at site 4, and the depth to basement is smaller by a factor of approximately $2^{1/2}$. Accordingly, whereas one might expect the three Earth parameters ρ_1 (resistivity of the sedimentary layer), h_1 (thickness of the sedimentary layer) and ρ_2 (resistivity of the basement) to vary with distance, it is suspicious that they vary in a manner consistent with a static shift of 2. Without *a priori* knowledge, it is impossible to be certain of the correct level of the MT apparent resistivity curves, and thus perhaps the data from *all* of the sites are affected by varying amounts. The model may be correct, but there is sufficient doubt to warrant independent assessment of the depths to basement. Many other examples abound in the older MT literature of suspect interpretations due to possible static shift problems, and recently a variety of methods have been advanced to estimate the shift factor (Jones, 1988a; Sternberg et al., 1988; Craven et al., 1990; Pellerin and Hohmann, 1990).

Others have noted that, whereas the phases from a number of sites all lie within a narrow range, the apparent resistivity curves all have the same shape but are displaced by up to two orders of magnitude (e.g., Kurtz et al., 1986a, 1990; Jones, 1988a; Vanyan et al., 1989; Beamish, 1990; Volbers et al., 1992), which is also a manifestation of static shift. Sternberg et al. (1988) illustrated that, for their MT data from 70 sites, the static shifts on the apparent resistivity curves had a standard deviation of about one-quarter to one-third of an order of magnitude. Vanyan et al.'s (1989) compilation for almost 400 MT sites on the eastern Siberian shield covering a very large area (1500×1000 km) exhibits a statistical standard deviation of the static shift factors of half an order of magnitude.

In its more general form these distortions have been addressed by modelling their effects as 3D galvanic bodies over a 2D regional earth:

$$\mathbf{Z} = \mathbf{C}\mathbf{Z}_{2D} = \begin{bmatrix} a & b \\ c & d \end{bmatrix} \mathbf{Z}_{2D} \quad (13)$$

where a , b , c and d are real and frequency-independent. Methods have been developed to determine the 2D strike direction, and to decompose the MT tensor observed, \mathbf{Z} , into component tensors that describe the distortions and the regional 2D conductivity distribution separately (see §2.4.2). No methods currently exist for dealing with the completely general case of a 3D distorting tensor \mathbf{C} over a region with a 3D conductivity distribution. However, because the phase of the determinant of a complex matrix is unaffected by multiplication with a real matrix, the determinant phase of \mathbf{Z} (Eq. 9) is a true estimate of the determinant phase of the regional structure. Accordingly, a pseudo-section plot of this phase with distance along the abscissa and period down the ordinate can give a good image of the first-order conductivity features beneath the profile (e.g., Jones et al., 1988). One note of caution, however, is that the estimation of \mathbf{Z}_D is unstable in the presence of noise and the condition number of \mathbf{Z} should be determined to indicate if the matrix is singular.

2.4.2. Strike determination

If the conductivity structure of the Earth near a site can be reasonably approximated by a 2D model, at least for a range of frequencies, then it is necessary to determine the appropriate strike direction, θ , so as to obtain a tensor that best approximates the form of Eq. 7 in some manner. The observations in co-ordinate frame θ , \mathbf{Z}_θ , are related to the 2D impedance tensor by:

$$\mathbf{Z}_\theta = \mathbf{R}\mathbf{Z}_{2D}\mathbf{R}^t \quad (14)$$

where \mathbf{R} is the Cartesian rotation matrix and the superscript t denotes transpose. This angle can either be obtained from the MT data or can be assumed from *a priori* knowledge of the region.

Quite a number of methods were developed through the 1970s for deriving the strike angle from the MT data. The majority of workers used a form of Swift's (1967) algorithm that leads to an analytical solution for an angle which maximized the power in the off-diagonal components (or equivalently minimized the power in the diagonal components) of \mathbf{Z} at each frequency. Generally, this was applied independently at each frequency and at each site, so that sites very close together (in an inductive scale length sense; Eq. 5) had data rotated to angles that could differ by up to 90° . However, local static 3D distortions and noise contributions can cause meaningless angles to be determined from Swift's formula; this leads

to frequency- and site-dependent angles that result in estimates of Z_{2D} which are not rotated to the correct axes with possible mode-mixing of the TM and TE responses.

Accordingly, new methods have been advanced to determine the correct strike angle of the 2D regional structure in the presence of local distortions (Bahr, 1985, 1988, 1991; Zhang et al., 1986). These methods suffer in the presence of noise as the regional strike angle is the least stable distortion parameter that can be extracted from the data. Groom and Bailey's (1989, 1991) mathematical decomposition gives the parameters required as well as the confidence with which to believe those estimates, and the Groom-Bailey approach is now being used routinely by many workers in the field for deriving the appropriate θ and the distortion parameters.

Alternatively, regional geology/tectonics can indicate the appropriate strike angle to choose; for example, the strike of the coast (Vancouver Island: Kurtz et al., 1986a, 1990; Oregon/Washington (EMSLAB): Wannamaker et al., 1989a,b), the strike of accreted terrains (southeastern B.C.: Jones et al., 1988), or the strike of major faults. No matter how the strike angle θ is determined however, it is important that the data from the profile all be rotated into a *consistent* reference frame to permit valid 2D modelling.

2.4.3. Model resolution and uniqueness

The current state-of-the-art for MT data modelling is that 1D inversions, 2D trial-and-error forward model fitting, and 3D thin-sheet modelling are routine, and 2D inversions and full 3D model studies for generic structures are becoming more used. Full 3D inversions for arbitrarily-shaped realistic bodies of widely differing scales is our ultimate goal and, given advances at the current rate, should be attainable by the turn of the century.

One question that must be asked is just how well resolved and how unique are our conductivity models of the CLC? In the majority of studies, especially the older ones, 1D inversions were used to estimate the conductivity of the CLC. For perfect 1D MT data at all frequencies *only one* $\sigma(z)$ model exists that will satisfy the observations (Bailey, 1970; Weidelt, 1972). In theory therefore, MT data are distinct from other geophysical responses of diffusive phenomena (gravity, magnetism, geothermal) in that there is no inherent non-uniqueness. The realization of perfect data is obviously unrealistic, but the existence of the uniqueness theorem gives us a rationale for striving to obtain the highest precision data possible.

Consider a typical crust of 40 km thickness with a conducting CLC beginning at a depth of 20 km. Assuming that there are no sedimentary sequences, we may adopt a resistivity of $10^4 \Omega\cdot\text{m}$ for the upper crust, and a value of $10^3 \Omega\cdot\text{m}$ for the continental upper mantle. The response to a single-layer CLC of $100 \Omega\cdot\text{m}$ material is illustrated in Figure 3-6 (solid curves). A singular value decomposition (SVD) analysis of this model (e.g., Edwards et al., 1981; Jones, 1982), for an observation range of $10^3 \text{ Hz} - 10^3 \text{ s}$, indicates that the parameters are resolved in the following order: the thickness of the upper crust (h_1), the resistivity of the upper crust (ρ_1), the conductance of the lower crust ($S_2 = \sigma_2 h_2$), and the resistivity of the upper mantle (ρ_3). Least well resolved is the resistance of the lower crust ($T_2 = \rho_2 h_2$). This indicates that the correct (orthogonal) parameterization of the CLC of this model is not one in terms of layer thickness and resistivity, but one with a Dar Zarrouk parameterization (Maillet, 1947; Koefoed, 1979) in terms of S_2 and T_2 . This phenomenon, of sensitivity to $S(z)$ rather than $\sigma(z)$, has been known for a decade. In the COPROD project (COMparison of PROfiles from One-dimensional Data) a number of workers derived a wide variety of 1D models that fit the MT data from a site in southern Scotland (the NEW (Newcastleton) data of

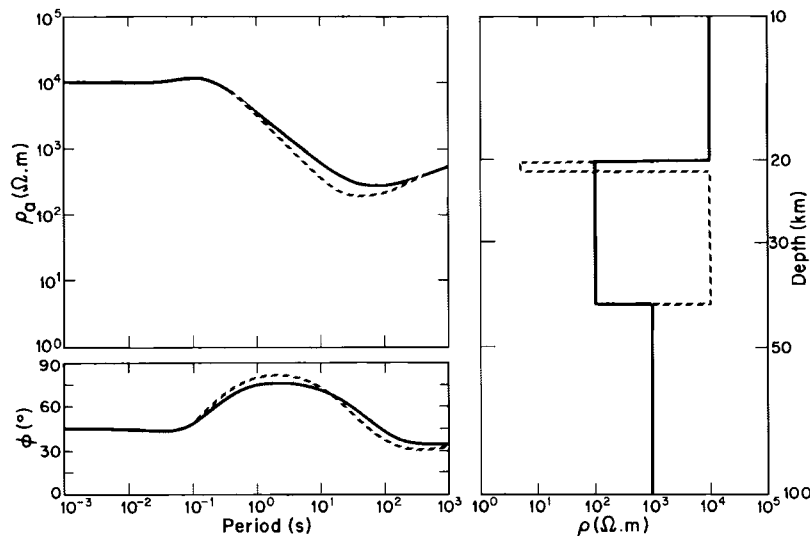


Fig. 3-6. Apparent resistivity and phase curves for two models of the continental lower crust of total conductance 200 S: in one model the conductance is distributed in a single 100 $\Omega\cdot\text{m}$ layer (solid curves) whereas in the other model the conductance is distributed in a thin top layer of 1 km thick and 5 $\Omega\cdot\text{m}$ underlain by a resistive lowermost lower crust (dashed curves).

Jones and Hutton, 1979a,b), but it can be shown that the conductances determined from all of the models fell within a narrow range (Weidelt, 1985). Some inversion algorithms favour thin zones — without constraints the natural tendency to minimize misfit is for infinitely-thin zones (D^+ : Parker, 1980) — whereas others tend towards thicker zones, e.g., the smooth inversions of Constable et al. (1987) and Smith and Booker (1988).

The conductance of the lower crust in this model is 200 S (a reasonable average for the CLC, see §3.4), which could also be distributed as a thin layer (layer 2) of 1 km thickness and 5 $\Omega\cdot\text{m}$ overlying a 19-km-thick layer (layer 3) of much higher resistivity (taken as $10^4 \Omega\cdot\text{m}$). For such a model the parameters that are well-resolved are the thickness of the upper crust (h_1), the conductance of the conducting zone (S_2), the resistivity of the upper crust (ρ_1) and the resistivity of the upper mantle (ρ_4). For MT data with 5% errors the thickness of the resistive part of the lower crust (h_3) is marginally resolved, whereas the resistance of the conducting zone (T_2) and the resistivity of the resistive part of the lower crust (ρ_3) are unresolved. The responses of this two-layer CLC are also illustrated on Figure 3-6 (dashed curves), and the difference between the two responses is a few percent. Discriminating between these two end-member models for the distribution of the conductance in the CLC — either a thick zone of moderate resistivity (100s $\Omega\cdot\text{m}$), or a thin zone of low resistivity (10s $\Omega\cdot\text{m}$) overlying a resistive lowermost lower crust — depends on resolving the resistive part of the CLC. This requires MT data with 2% errors or better and a region which is 1D electrically to within these errors. Thus, many of the zones classified by Jones (1981b) as “intermediate” and by Haak and Hutton (1986) as “normal”, of resistivity of 100–300 $\Omega\cdot\text{m}$ and thickness some 20 km could, in fact, be 10–30 $\Omega\cdot\text{m}$ and only 2 km thick for example.

As a general rule-of-thumb it is not possible to resolve conducting zones of conductance less than the conductance of all the zones above it. Figure 3-7 shows the responses of two models of the lower crust, one with a single 1-km-thick zone of conductance 200 S at 20 km depth (solid curves) and the other with an additional zone of enhanced conductivity at 30

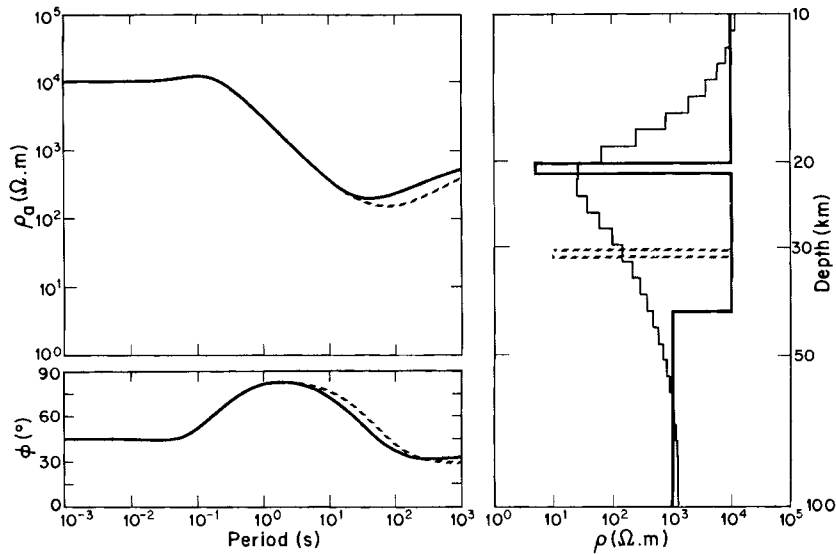


Fig. 3-7. Apparent resistivity and phase curves for two models of the continental lower crust: one model (solid curves) is of a two-layer CLC with the top layer being 1 km thick of 5 $\Omega\cdot\text{m}$ (thin layer model of Fig. 3-4), and the other model is a four-layer CLC with a second 5 $\Omega\cdot\text{m}$ conducting zone of 100 S at a depth of 30 km (dashed curves). A smooth inversion of the theoretical data with 5% errors of the four-layer model (thin line) illustrates that the lower conducting zone cannot be detected.

km with half the conductance (100 S, or 10 $\Omega\cdot\text{m}$) of the upper one (dashed curves). Note that there is little difference in the responses between the two models. A smooth inversion (Constable et al., 1987) of the MT data with 5% errors from the two-conducting zone model of the CLC is also illustrated in Figure 3-7 (thin line), and the lower conducting zone cannot be detected. This shielding effect could be important for certain geological symmetries, such as anticlines and synclines.

This aspect, of screening by overlying conductive structures, has an important consequence for resolving the conductivity structure of the CLC beneath sedimentary basins. Figure 3-8 compares the MT responses that would be observed on a 2-km-thick basin with fill of 5 $\Omega\cdot\text{m}$ sediments (dashed curves), i.e., total conductance of 400 S, to the responses that would be observed on resistive upper crust with no sedimentary cover (solid curves) for the thin-layer CLC example of Figure 3-6. For no sedimentary cover, the existence of the conducting layer within the CLC is indicated by the drop in the ρ_a curve at periods greater than 0.1 s, and its conductance is indicated by the level of the minimum at 30 s. For the ρ_a curve on the basin neither of these features is readily apparent in the curves. Smooth inversion (Constable et al., 1987) of the MT data, with 5% errors, on the basin illustrates that the conducting zone within the CLC cannot be detected (thin line). For MT data with 2% errors the existence of the zone can be inferred, and with 1%, or better, data its conductance and the depth to its centre can be marginally resolved.

In 2D studies, because of the two polarizations of the EM fields, it is possible with high-quality data to discriminate between thin zones and thicker zones even beneath sedimentary basins, as shown, for example, by Jones and Craven (1990) in their modelling study of the North American Central Plains conductivity anomaly (NACP, §3.3.4). However, few MT studies of the lower crust are of sufficient quality or are interpreted two-dimensionally. The

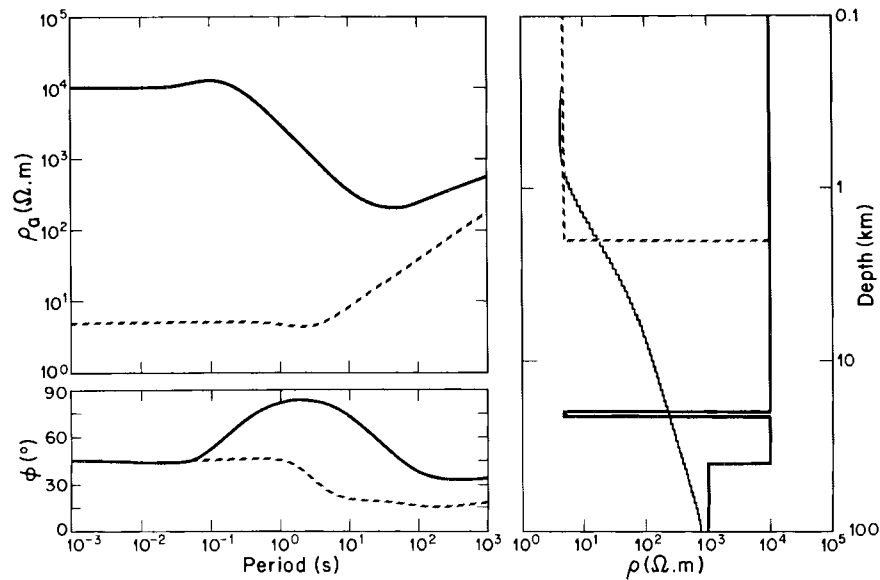


Fig. 3-8. Apparent resistivity and phase curves for a model of the continental lower crust of 1 km thickness, 5 $\Omega\cdot\text{m}$ beneath an exposed upper crust (solid curves) and beneath a 3-km-thick sedimentary basin with fill of 10 $\Omega\cdot\text{m}$ (dashed curves). A smooth inversion of the theoretical data with 5% errors of the basin model (thin line) illustrates that the conducting zone in the lower crust cannot be detected.

shielding effect is still important and conductors beneath other conductors may be masked (see fig. 1 of Jones, 1987). The inherent sensitivity to lateral variations of conductivity is greater for TM data than for TE data.

The MT data that can be acquired today are of sufficient precision (1% is possible) to resolve theoretically the details of the conductivity structure of the CLC beneath regions of little sedimentary cover. The important limiting factor is now the geological noise, rather than instrumentation noise or cultural noise, which is an expression of our modelling limitations.

3. Summary of results

Interpretations of the conductivity distribution within the CLC abound in the published literature. However, many models were derived from responses that were of insufficient precision for the quoted resolution due to older technologies used for data acquisition and processing (see Jones et al., 1989). Accordingly, I will concentrate on the more recent results and refer to older results when they are significant or historic.

I have grouped the specific results into three broad categories; (1) shields, (2) rifts, and (3) continental margins, though there is obviously much overlap between these. Another feature that warrants consideration in a general treatise on crustal EM studies is sedimentary basins. However, in this paper we are concerned with the CLC beneath the basins, not the structure of the basins themselves. It should be noted that although thick sedimentary sequences can mask the features of the CLC that we are attempting to elucidate, strongly anomalous structures within the CLC can create secondary EM fields that can be observed on the surface of the Earth, e.g., the NACP anomaly (see Trans-Hudson Orogen, §3.3.4) and the anomaly in the CLC beneath the Flathead basin in southeastern British Columbia (Gupta and Jones, 1990).

The generic results for the resistivity of the CLC from statistical compilations are also considered (§3.4), as are the results from two special areas: the Kapuskasing Uplift in Ontario (§3.5) and the Valhalla complex in British Columbia (§3.6). The former is important as it is an exposure of rocks from deep levels within the crust, whereas the latter is unusual in exhibiting high conductivity but no seismic reflectivity in the lower crust.

In the following I will not discuss an electrical analogue to the seismic Moho. In my opinion there is not a single EM study which shows unequivocally a change in electrical conductivity at the base of the crust as defined seismically by the depth of the Moho discontinuity. Indeed, studies on resistive windows, which would enable the greatest chances of resolving an electrical Moho, suggest that there is no dramatic change in conductivity across it (Jones, 1982; Beamish, 1990).

3.1. Shields

Electromagnetic investigations have been carried out on all of the Earth's shields and platforms, and a review of the work up to the mid-1970s was given by Kovtun (1976). Three shield regions are chosen for discussion; namely the Baltic, Canadian and Siberian shields.

3.1.1. Baltic shield

By far the most extensively studied shield region using modern equipment, analysis and interpretation techniques is the Baltic shield of northern Europe. Results and models for predominantly the Finnish and Soviet portions of the shield were recently collated by Hjelt and Vanyan (1989).

The first results from the Baltic — or Fennoscandian — shield (Jones, 1982, 1984a; Jones et al., 1983) from long-period (>100 s) MT and HSG measurements indicated that the CLC beneath the northern parts of the shield was moderately resistive — of the order of some hundreds of $\Omega\cdot\text{m}$ — whereas the CLC beneath southern Finland appeared to be more conductive — some tens of $\Omega\cdot\text{m}$. The dividing line between these two was conjectured to be the Ladoga–Bothnian Bay zone (LBBZ), which is discussed in greater detail below (§3.3.5). The LBBZ, described as a 1.9–1.87 Ga Svecofennidic schist, separates Archean basement (2.8–2.6 Ga) to the northeast from a Svecokarelian (1.9–1.8 Ga) complex to the southwest. Although the actual resistivity values of the CLCs on either side of the LBBZ may be in question, there is little doubt that there must be an increase in the conductance of the CLC of approximately an order of magnitude to the south of the LBBZ compared to the north (about 1000 S to the south compared to 100 S to the north). This difference is also evident in the vertical magnetic fields observed by the Scandinavian International Magnetospheric Study (IMS) array (Küppers et al., 1979; Jones et al., 1983).

The shield has been further studied using MT methods by Swedish and Finnish groups with Soviet and Hungarian participation. Rasmussen et al. (1987) conducted an extensive regional MT survey of forty sites along a 1250 km profile following the N–S Fennolora (Fennoscandian Long Range profile) seismic refraction profile in Sweden, but the stations are generally too widely spaced (separation typically 40 km) to be certain of the results. The three most northerly sites show a zone of increased conductivity in the CLC, with estimates of conductance in the range 100–1000 S. This wide variation may be explained by a downward static shift of the data at one anomalous site giving a regional average value for the

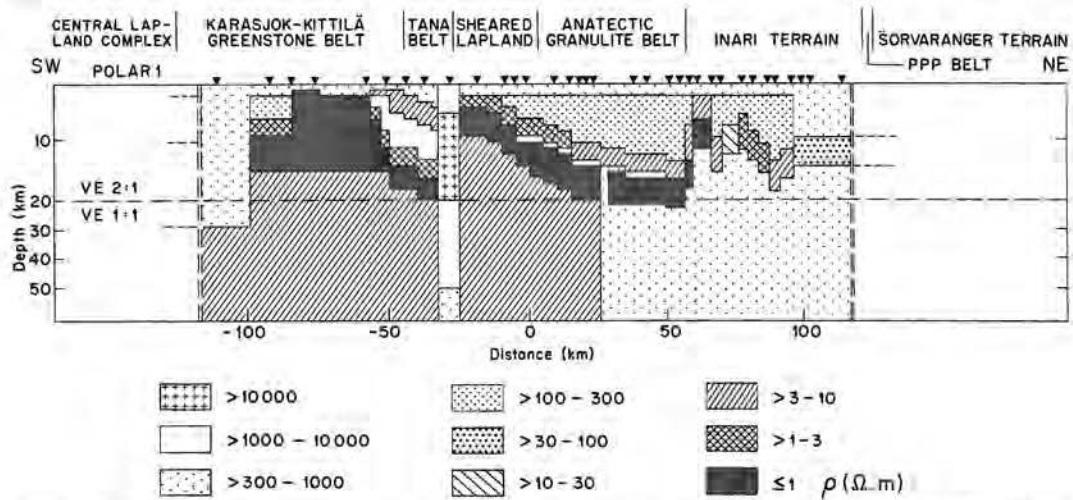


Fig. 3-9. A 2D model of the POLAR profile data from northeastern Finland (redrawn from Korja et al., 1989). The locations of the MT sites are indicated by the inverted triangles.

conductance of the CLC of 100 S. In the south there does not appear to be a zone of markedly high conductance in the CLC.

An extensive MT study was performed along the POLAR profile, the northernmost segment of the European Geotraverse (EGT), which strikes northeast-southwest in northeastern Finland (Korja et al., 1989). MT measurements were made at 40 sites along 300 km of the transect, with a shorter parallel transect of ten sites to the southeast. After attempting to correct the data for static distortion effects, the authors derived a 2D model which fitted the MT data (Fig. 3-9) with greatest weight being given to the phase information. (The resistivity shading scheme used is consistent on all 2D models illustrated in this chapter to facilitate comparison.) Note that the northeastward-dipping conducting zones within the granulite belt appear to stop at mid-crustal levels, ≈ 20 km. Within the CLC the conductance decreases from SW to NE; the actual conductances are not well resolved and the boundary between the high- (resistivity about $5 \Omega \cdot m$) and low- (resistivity about $100 \Omega \cdot m$) conductivity CLC regions is not well defined. Features in the upper and middle crust correlate well with other geophysical information (seismic reflection, refraction and gravity). In the lower crust, however, there is less coincidence.

An MT study in southern Sweden across the Mylonite Zone by Rasmussen (1988) led to two possible families of models to explain the observations: either (1) a 2D model with inductive scale lengths that are large compared to the profile length so that lateral variations in resistivity are modelled to occur off the ends of the profile (more than one profile length away), or (2) a 1D model incorporating a transversely anisotropic layer between 12 km and 33 km of resistivities of $400 \Omega \cdot m$ and $17000 \Omega \cdot m$ in the two directions (Fig. 3-10). Moho is thought to be around 40 km in this region. Rasmussen (1988) expresses a preference for the latter interpretation because it explains better the observations over the whole frequency range of observation. This is discussed further in §4.5.

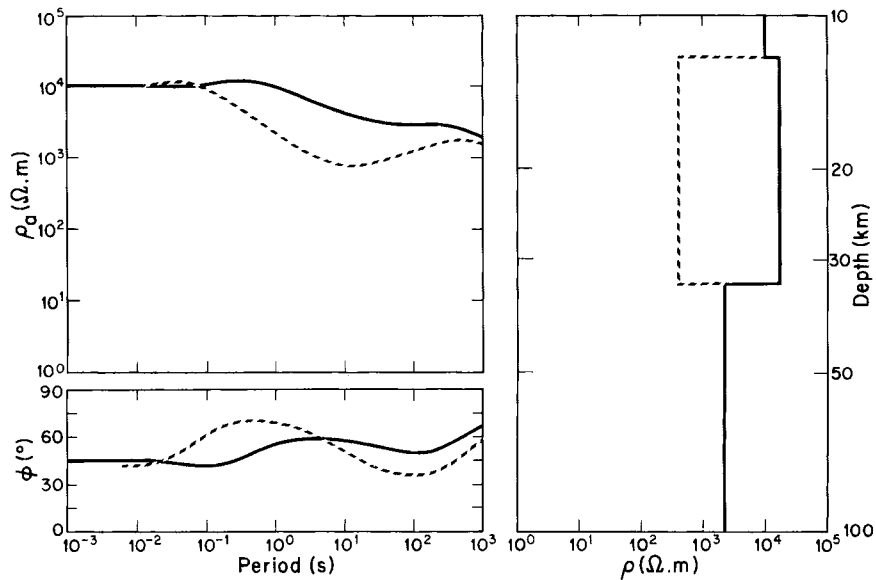


Fig. 3-10. A 1D model with an anisotropic middle and lower crust of Rasmussen (1988) and the theoretical responses.

3.1.2. Canadian shield

Studies on the Canadian shield up to the early 1980s were reviewed by Jones (1981b, 1984b) who proposed that the shield's CLC could be zoned into three general types (1981b, fig. 4) with the central part (beneath Alberta and Saskatchewan) highly resistive (10^3 – 10^4 $\Omega\cdot\text{m}$), an edge of intermediate resistivity (100–300 $\Omega\cdot\text{m}$), and some zones of anomalously low resistivity (10–50 $\Omega\cdot\text{m}$). Given the screening effects of the sediments (§2.4.3) and our knowledge of the sensitivity to the conductance of the CLC rather than its intrinsic conductivity (§2.4.3), this zonation should be treated with caution. However, Keller's (1989b) recent map of the zonation of the United States shows some similarities to Jones' map and, while some details may be in error, there may be a general validity to the zonation.

The early works of greatest reliability were the deep-probing controlled-source measurements of Connerney et al. (1980) in the Adirondacks and those of Duncan et al. (1980) in northern Ontario. In both areas a marked decrease in electrical resistivity was observed at mid-crustal depths; 23 ± 2 km to the top of a zone of 20 ± 10 $\Omega\cdot\text{m}$ for the Adirondacks and 17–29 km to a zone of 270 ± 90 $\Omega\cdot\text{m}$ for northern Ontario. The differences in resistivity are not considered significant: for both studies the resistivity of the lower half-space (the conducting zone) was the least-resolved parameter (see Table 4 of Duncan et al., 1980). HSG measurements in the Adirondacks by Connerney and Kuckes (1980) indicated that the conducting zone has a conductance of ≈ 400 S.

In contrast to Scandinavian work, deep-probing EM studies in Canada during the 1980s were more target-oriented, such as sedimentary basins (Prince Edward Island: Jones and Garland, 1986; Williston Basin: Jones, 1988a; Fredericton basin: Jones, unpubl.), earthquake zones (Buchbinder et al., 1983; Kurtz and Gupta, 1992), tectonic features (Trans-Hudson Orogen: see §3.3.4; Wopmay Orogen: Camfield et al., 1989; Kapuskasing Uplift Structure: see §3.5), the Abitibi greenstone belt (Chouteau et al., 1991), and plutons as possible sites for the disposal of radioactive waste (Kurtz et al., 1986b).

The MT study by Kurtz et al. (1986b) on the East Bull Lake pluton in northern Ontario showed a highly resistive CLC ($2,500 \Omega \cdot \text{m}$) underlain by an $800\text{-}\Omega \cdot \text{m}$ mantle at a depth of 200 km. Given that this mantle resistivity is high compared to global averages of about $100 \Omega \cdot \text{m}$ (Schmucker, 1985), the apparent resistivity data are possibly static shifted upwards at the longer periods by an order of magnitude. Correcting for such a shift would imply that the resistivity of the CLC here is about $250 \Omega \cdot \text{m}$, which is consistent with Duncan et al.'s (1980) estimate, although it would be much shallower (6 km compared to 17–29 km for Duncan et al., 1980). The longest-period apparent resistivity values of Kurtz et al.'s (1986b) data are $3,000 \Omega \cdot \text{m}$ at 10^3 s, which is an order of magnitude greater than Vanyan's global average curve (Vanyan et al., 1980; Vanyan and Cox, 1983), and this would also be consistent with a static shift of that order. Obviously, the distorting effects of the resistive pluton, as illustrated by the Nelson batholith in British Columbia (Jones et al., 1988), need to be taken fully into account.

MT studies of the earthquake-prone Miramichi region of Canada's maritimes (New Brunswick) by Kurtz and Gupta (1991) showed a highly resistive crust ($> 10^5 \Omega \cdot \text{m}$) down to 20 km then a transition to $1,000\text{s} \Omega \cdot \text{m}$ underlain by a zone of $100\text{s} \Omega \cdot \text{m}$ at a depth of ≈ 30 km.

3.1.3. *Siberian shield*

The Siberian shield has been much studied by U.S.S.R. groups, and recently Vanyan et al. (1989) gave a review of the interpretations from more than 2500 MT soundings made in its eastern part since the late 1960s. They showed that the responses could be grouped into four type areas, each with its own conductivity-profile. Types 1 and 2 curves, which are representative of most of the area studied, had a conducting zone in the CLC of about 350–650 S with their centres between 30 and 40 km in a crust of approximately 45–50 km thickness. Type 3 curves are insensitive to a CLC conducting layer because of the screening by sediments of 500 S total conductance on the surface (see §2.4.3). Type 4 curves are indicative of a highly conducting zone of resistivity $< 0.5 \Omega \cdot \text{m}$, at a depth of 3–5 km which also screens the CLC. Although the MT data were not acquired by modern instrumentation and were not corrected for static shifts, the existence of a zone of enhanced conductivity in the CLC beneath most of the eastern Siberian shield cannot be doubted due to the vast quantity of data available.

3.1.4. *Shields: Conclusions*

It is apparent that, in detail, the shields are almost as heterogeneous in the conductivity of their lower crusts as younger or active regions. Although statistically (§3.4) the conductance of the CLC beneath shield regions is low — of the order of 40 S — it is still far higher than would be expected for dry rock (< 2 S for an CLC of 20 km and resistivity of $10^4 \Omega \cdot \text{m}$, §4). Also, there is evidence that the conductance of the CLC can be an order of magnitude greater, such as beneath the eastern part of the Siberian shield and the southern part of the Baltic shield in Finland. What is causing the enhanced conductivity of these old regions? This question is discussed further in section §4.

3.2. *Rifts*

Because of their intriguing nature quite a few studies of continental rifts have been undertaken. One particular problem with imaging the conductivity-distribution of the CLC beneath rifts is that the sediments filling the rift, or the rift structure itself, are often highly

conducting and mask the deeper structure. In a comparison of interpretations from four rifts, Jiracek et al. (1979) concluded that a ubiquitous feature of rifts is a zone of enhanced conductivity with $\rho < 50 \Omega \cdot \text{m}$ at 10–30 km depth beneath them.

3.2.1. East African Rift

Two-dimensional model studies of MT and GDS data from the East African Rift (Banks and Ottey, 1974; Rooney and Hutton, 1977; Banks and Beamish, 1979) suggest that much of the crust within the Gregory Rift is highly conducting ($10 \Omega \cdot \text{m}$) from the surface down to >35 km depth, with crustal thickness in the region being about 36 km (Henry et al., 1990). Although these MT data are not of sufficient bandwidth (30–3000 s period range) or quality for resolution of fine structure within the rift, the conclusion that much of the crust beneath the rift is conducting is not easily amenable to other interpretation. A recent interpretation of the KRISP85 (Kenya Rift International Seismic Project 1985) seismic refraction study in the same region (Henry et al., 1990) presented a cartoon (their fig. 4) to explain the observations. One obvious interpretation for the enhanced conductivity is magma that permeates the whole of the crust beneath the Rift Valley. A more likely explanation is that the uppermost part of the crust is conducting because of the valley sediments, of thickness ≈ 3 km, below which there are zones of interconnected partial melt, as suggested by Rooney and Hutton (1977) and Banks and Beamish (1979). Any resistive layer between these two conducting zones would not be detectable in the data (Rooney and Hutton, 1977).

3.2.2. Rio Grande Rift

The Rio Grande Rift has been the subject of EM studies for three decades (Schmucker, 1964, 1970; Swift, 1967; see historical review in Keshet and Hermance, 1986), and recently MT measurements have been made across it by Jiracek and his colleagues (Jiracek et al., 1979, 1983, 1987). Jiracek's initial studies followed COCORP's* seismic reflection profiling of the rift in 1975 and 1976 (Oliver and Kaufman, 1976) that imaged a seismic reflection bright spot at 18–22 km depth beneath the rift. Jiracek's data are reasonably broad-band (0.1–1000 s), and are of good quality. Considerable along-strike variation in the conductivity distribution is evident from the MT data along two east–west profiles separated by ≈ 30 km.

On the northern line, through Bernardo, a thick zone of low resistivity, $10 \Omega \cdot \text{m}$, which is thicker and closer to the surface to the west, was modelled in the middle to lower crust (Fig. 3-11a). This was interpreted as due to fluid trapped beneath an impermeable cap (Jiracek et al., 1983). In contrast, on the southern line through Socorro no such zone was present (Fig. 3-11b). Given the enhanced earthquake activity in the region of Socorro compared to further north, Jiracek et al. (1987) interpreted the lack of a zone of enhanced conductivity as due to magmatic fracturing of the impermeable cap which led to loss of the trapped fluid. One puzzling aspect is that where there is active magma injection to upper crustal levels on the southern profile, the crust is relatively resistive ($400 \Omega \cdot \text{m}$). Given that the geothermal gradient and heat flow are higher in the Socorro area (Reiter et al., 1978), this observation appears to contradict laboratory results that partially molten rocks are more conducting (Waff, 1974; Sato and Ida, 1984). Could this be due to lack of interconnectivity of the magma? **.

* COCORP: COnsortium for COntinental Reflection Profiling based at Cornell University.

** *Note added in proof:* A recent study across the Socorro magma body by Hermance and Neumann (1991) challenges Jiracek et al.'s (1987) model (Fig. 3-11b). Hermance and Neumann (1991) suggest that there is indeed a conducting zone, of 10–30 $\Omega \cdot \text{m}$, beneath the rift at intermediate depths (15–20 km).

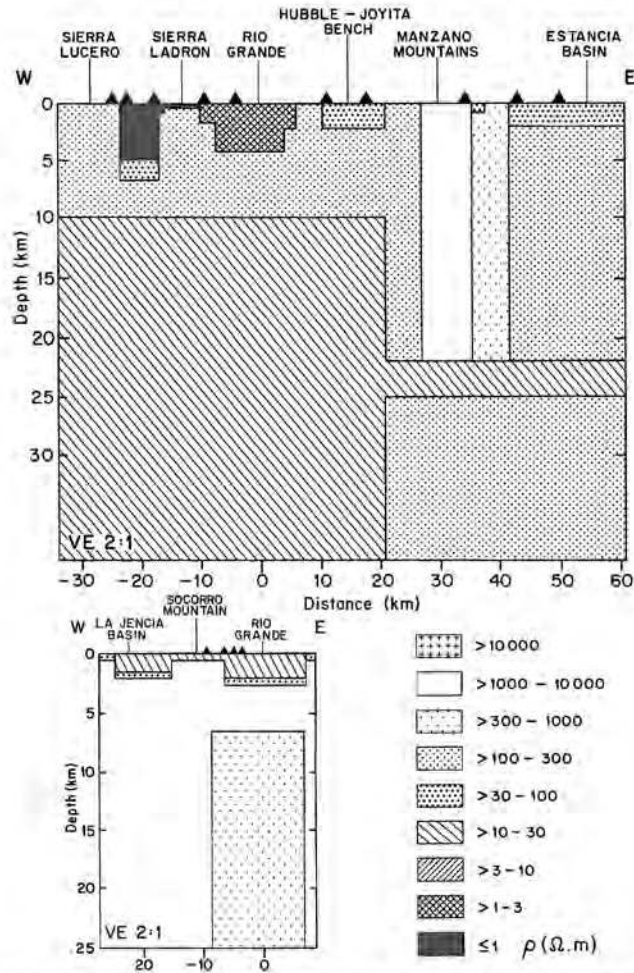


Fig. 3-11. Two 2D models of the Rio Grande Rift from profiles 30 km apart (redrawn from Jiracek et al., 1983). Note that whereas the lower crust beneath the northern profile is interpreted to be of very low resistivity (3–10 $\Omega\cdot\text{m}$), the lower crust beneath the southern profile is thought to be of much higher resistivity (>100 $\Omega\cdot\text{m}$).

Approximately 250 km to the south (32°N latitude), Keshet and Hermance (1986) interpret their observations in terms of a conducting zone in the CLC, with $\rho = 1\text{--}10 \Omega\cdot\text{m}$ beginning at ≈ 20 km depth. Recent results by Jiracek and his colleagues from an E–W profile 160 km north of the Bernardo line are indicative of a strongly conducting zone, with $\rho = 1 \Omega\cdot\text{m}$, at depths below ≈ 15 km (G.R. Jiracek, pers. commun., 1991).

3.2.3. Rhinegraben

The Rhinegraben has been the focus of much EM work since the early 1970s (see the review by Schmucker and Tezkan, 1988), and initial interpretations were of a zone of enhanced conductivity in the lower crust and upper mantle (Reitmayr, 1975). However, the highly attenuating effects that the conductive sediments can have on the observations were not initially appreciated. Interpretations of the GDS and MT responses at periods shorter than 1000 s in terms of bodies of enhanced conductivity in the deep crust or mantle have

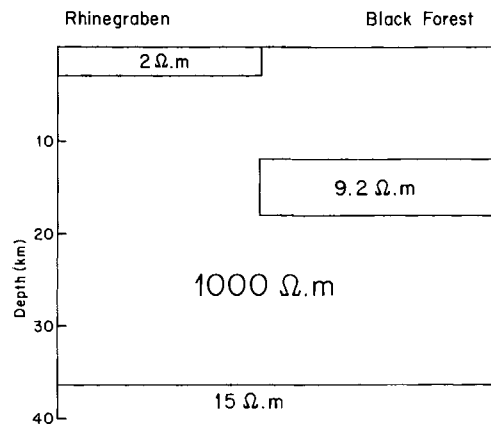


Fig. 3-12. A 2D model of the Rhinegraben with a zone of enhanced conductivity at a depth of 12 km beneath the Black Forest which is absent beneath the Rhinegraben (redrawn from Wilhelm et al., 1989).

been shown to be in error, and conductive sedimentary fill can explain these shorter period responses (Dupis and Thera, 1982; Jones, 1983b). At longer periods, however, there are features in the MT and GDS data that require explanation, and Richards et al. (1981; reported in Fuchs et al., 1987) interpret their MT data as indicating a zone of slightly enhanced conductivity between 20 and 40 km depth. Limited 3D thin-sheet modelling of the Rhinegraben area by Kaikkonen et al. (1985) was interpreted in terms of current channelling, but the modelling algorithm did not allow for poloidal current flow so that currents were confined to the surface thin-sheet and thus were insensitive to lateral variation in conductivity beneath the surficial layer.

Recently, MT and LOTEM measurements were made in the Black Forest as a component of the multidisciplinary studies for the KTB * drill site (Wilhelm et al., 1989; Strack et al., 1990). The MT data were interpreted, using a 2D model (Fig. 3-12), as indicative of a conductive zone about 12–18 km beneath the eastern flank of the Rhinegraben, with no conductive zone, apart from the sedimentary fill, beneath the Rhinegraben itself (Berkold et al., 1985; Tezkan and Schmucker, 1985; both reported in Fuchs et al., 1987; Schmucker and Tezkan, 1988; Wilhelm et al., 1989). The LOTEM data, which were inverted for 1D structure, revealed a zone of enhanced conductivity at depths of 7–9 km, at least 500 m thick (Strack et al., 1990). As the shortest period of the MT data was 10 s, it is possible that the zone imaged by the LOTEM results was not resolvable from the MT data, as suggested by Wilhelm et al. (1989). However, an alternative explanation is that the MT data are static shifted (§2.4.1): Tezkan's model comprises of a layer of 1000 Ω·m material overlying 9.2 Ω·m material at a depth of 12 km, whereas the LOTEM results indicate that the uppermost part of the crust is about 400 Ω·m with a conductive zone of about 4 Ω·m material (Wilhelm et al., 1989; Strack et al., 1990). Applying a static shift factor of 2.5 to Tezkan's model, so that the MT upper crustal resistivity agrees with LOTEM, gives a depth of ≈ 7.5 km (i.e., $12/2.5^{1/2}$) to a zone of 3.7 Ω·m (i.e., $9.2/2.5$), which agrees well with the LOTEM results. The LOTEM results are insensitive to the thickness of this conducting zone, but impose a lower bound of at least 500 m (Strack et al., 1990). However, if the MT model (Tezkan and Schmucker, 1985) is scaled with a shift of 2.5,

* KTB (Kontinentales Tief Bohrung): the German Continental Deep Drilling Program.

then the base of the zone is at 11.4 km ($18/2.5^{1/2}$). Thus, Tezkan's model, when corrected for static shift with a factor of 2.5, has a zone of enhanced conductivity between 7.5 and 11.4 km, which coincides with a pronounced low-velocity layer between depths of 7–14 km, above a laminated lower crust (Gajewski and Prodehl, 1987; Wilhelm et al., 1989). Alternatively, the LOTEM data could be affected by static shift (K.-M. Strack, pers. commun., 1991), or both the MT and the LOTEM could, but the correlation of the depth to the LOTEM anomaly and the depth to the top of the seismic low-velocity zone suggests that the LOTEM data are reliable. Also, the magnitude of any static shifts of the LOTEM data would be in terms of percent rather than factors of 2.5 as for the MT. The conductive zone has been interpreted in terms of either fluids or graphitic metasediments (Wilhelm et al., 1989), with the fluid explanation being supported by the geothermal results.

3.2.4. Baikal rift

Generally, the lower crust in the Baikal region exhibits a zone of increased conductivity in the middle to lower crust (Popov, 1987, and references therein), and this zone broadly correlates with a seismic low-velocity zone. Popov (1987) illustrates that the conductance of this zone increases from 200–300 S to 800 S as the rift is approached from either flank, which implies a four-fold decrease in resistivity beneath the rift for a layer of constant thickness. The depth to this zone is considered to rise from 25 km on the flanks to 14 km beneath the rift itself. Generalized 1D electrical and seismic structures for the platform region and for the rift zone are illustrated in Figure 3-13. Note that for the lower crust the models indicate a $V_p \approx 7.0$ –7.5 km/s and $\rho \approx 200 \Omega \cdot \text{m}$, whereas beneath the rift these parameters are 6.5–7.0 km/s and $30 \Omega \cdot \text{m}$, respectively.

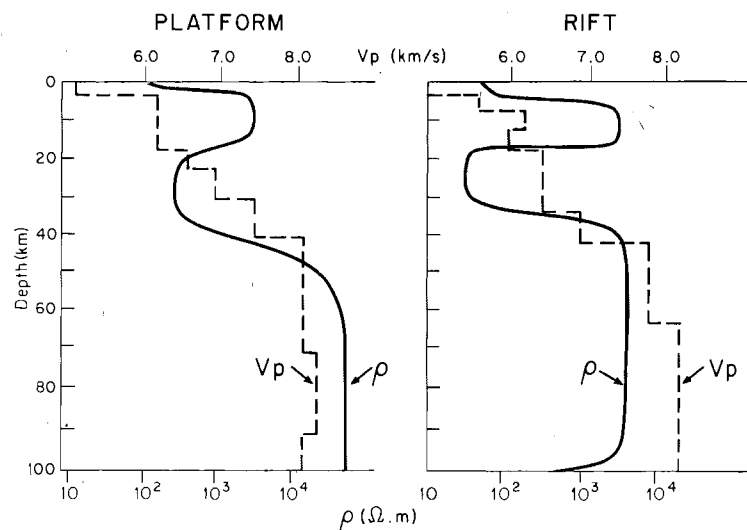


Fig. 3-13. Generalized electrical and seismic 1D models for the flank and the rift in the Baikal (redrawn from Popov, 1987).

3.2.5. Midcontinent Rift

Although the thrust of this section of the chapter is tectonically active rift zones, for comparison I have included studies over an ancient rift. The 1.1-Ga Keweenaw Midcontinent Rift (MCR) in North America has not been extensively studied by EM methods, and the only recently published results are those of Young and Wunderman (Wunderman et al., 1985; Wunderman, 1986; Wunderman and Young, 1987; Young et al., 1989; Adams and Young, 1989). One significant result in the MT data over the MCR is the asymmetry of conductivity structure beneath each of the flanks. Beneath the western flank, MT surveys imaged an easterly-dipping conducting zone, of $\rho < 10 \Omega\cdot\text{m}$, to a depth of ≈ 5 km at the rift. This zone, when traced updip, is thought to correspond to rocks of the Animikie basin, with the upper contact representing a decollement zone (Wunderman and Young, 1987). The zone was also seen on proprietary commercial MT data on the western flank of the MCR in Iowa (C.T. Young, pers. commun., 1991). COCORP data from Kansas image a prominent east dipping feature in the basement to the west of the MCR which has been interpreted as “a low angle detachment similar to those seen in the Basin and Range” (Serpa et al., 1984). In contrast, the eastern flank does not exhibit such a conductive feature. Beneath the MCR itself, the MT data illustrated in Wunderman et al. (1985, site MR) indicate that the crust is resistive, with resistivity greater than $10,000 \Omega\cdot\text{m}$ down to a depth of some 30 km.

3.2.6. Rifts: Conclusions

It is apparent that Jiracek et al.'s (1979) general conclusions regarding modern rifts have to be modified in the light of more recent investigations. No single conductivity model can be proposed for rift zones, and certainly variations of structure along strike should be kept in mind. In most locations over modern rifts the lower crust generally has a high conductivity (low resistivity), which is most probably associated with an interconnected fluid phase (either partial melt or saline fluids). Intriguingly, in at least the central portion of the Rio Grande Rift there is the suggestion that a partially molten zone is resistive — an interpretation that is at odds with laboratory studies (§4.4).

3.3. Modern and ancient continental margins

Imaging the conductivity structure in the vicinity of modern continental margins is difficult because of the dominant coast-effect. The large contrast in electrical conductivity between land (typically $\rho = 1,000\text{--}10,000 \Omega\cdot\text{m}$) and sea water ($\rho = 0.3 \Omega\cdot\text{m}$) results in electric current redistributions at coastlines. In the TE-mode (current flow parallel to the coast) there is virtually a line of current flowing offshore, whereas in the TM-mode (current flow perpendicular to the coast) the currents flowing in the sea water deepen significantly on entering the land. The observed MT responses are therefore dominated by the geometry of the ocean–continent boundary, and conductivity structures beneath margins are usually of second-order on the responses. Thus, in studies close to margins it is very important to obtain the highest possible accuracy and precision in the responses in order to resolve structure.

As suggested initially by Law and Riddihough (1971), EM methods are one possible geophysical technique for delineating the locations of ancient margins interior to the continents. One cause for conductive anomalies at ancient plate margins, proposed by Drury and Niblett (1980), is the presence of subducted and obducted sediments saturated with saline brines.

Whilst this explanation may not be correct ubiquitously (e.g., Trans-Hudson Orogen, §3.3.4), the existence of conductive zones as identifiable markers in the CLC of ancient tectonic boundaries is an important observation. An alternative explanation is that the zones are associated with conductive minerals — possibly graphite — in metamorphosed and fractured rocks in the basement (Camfield and Gough, 1977).

3.3.1. West coast of North America

The most recent, and most precise, EM investigations on a continental margin have been carried out along the west coast of North America. These studies were aimed at imaging the Juan de Fuca (JdF) plate as it dips beneath the mainland of Oregon, Washington and British Columbia. The first of these studies, by Kurtz et al. (1986a, 1990; Green et al., 1987), was undertaken as part of LITHOPROBE's* Vancouver Island transect investigations. Kurtz et al. (1986a, 1990) made high quality measurements at twenty-seven locations coincident with a SE–NW seismic reflection profile. Their interpretation yielded a model (Fig. 3-14) that has a dipping conductive zone, of $\rho = 30 \Omega\cdot\text{m}$, which correlates spatially with a strong reflection zone. The conductivity of the zone is consistent with a porosity of $\approx 1.6\%$ infilled with saline fluid (Kurtz et al., 1986a; Hyndman, 1988). Initial interpretation of the coincident reflecting

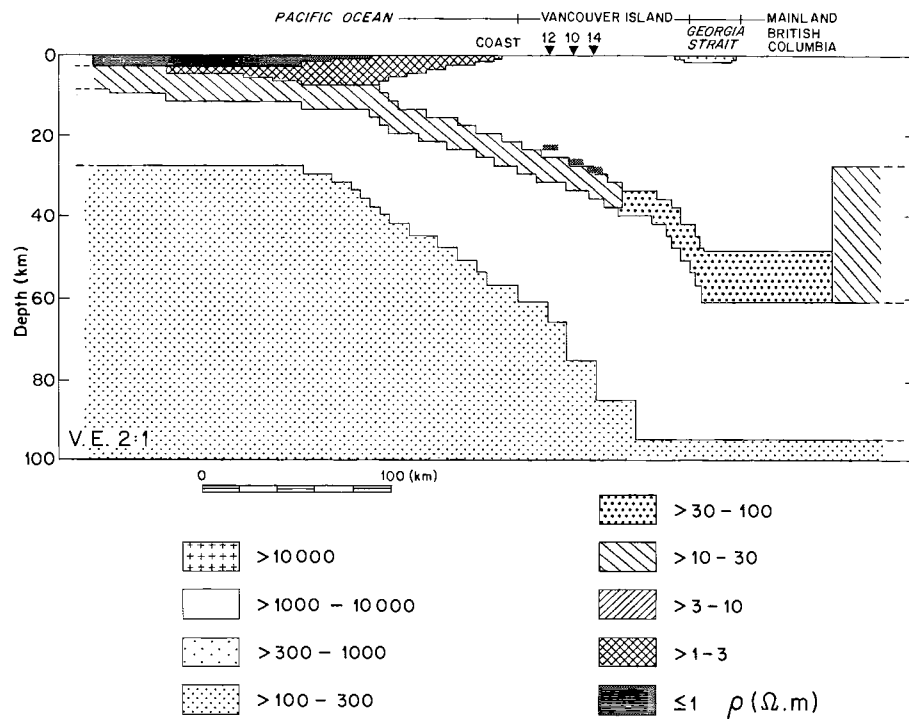


Fig. 3-14. A 2D resistivity model of the crust and upper mantle beneath Vancouver Island and the nearby mainland to account for the observed responses (redrawn from Kurtz et al., 1986).

* LITHOPROBE is Canada's national, collaborative, multidisciplinary earth science research program designed to answer fundamental questions on the nature and evolution of the lithosphere beneath Canada and its surrounding oceans.

and conductive zone suggested that it represents the top of the downgoing Juan de Fuca plate, but more recent studies, including the locating of earthquakes and further offshore seismic experiments, infer that the top of the plate is below this zone. It has been suggested that the anomalous zone may be due to sediments derived from the accretionary wedge, with fluids causing the high reflectivities and conductivities. Alternatively, a recent interpretation of the strength of seismic reflections concludes that the reflection coefficients of up to 0.2 are too high to be caused by fluids alone and that the reflections are explained by shear zones (Calvert and Clowes, 1990a,b).

Just to the south of Vancouver Island, onshore and offshore of Oregon and Washington, the largest EM experiment to date was carried out in 1985. This experiment, named EMSLAB*, involved 24 collaborating institutes supplying over 100 instruments (EMSLAB, 1989). The main results of the EMSLAB experiment are discussed in the special issue devoted to the study (Booker and Chave, 1989). Essentially, the same electrical structure was found beneath Oregon as beneath Vancouver Island. However, whereas Vancouver Island's upper crust is resistive (some thousands of $\Omega\cdot\text{m}$), the Coast Range of Oregon is relatively conductive (some hundreds of $\Omega\cdot\text{m}$) with total conductance of the same order as that of the zone associated with the downgoing plate. Accordingly, whereas for Vancouver Island the presence of the conducting zone was a first-order feature of the data (phase differences of 15°), essentially the same zone beneath Oregon was a second-order feature (phase differences of 2°) that demanded sophisticated processing, modelling and interpretation (see Booker and Chave, 1989, for a review).

3.3.2. East coast of North America

The east coast of North America is distinct from the west coast in that it is a passive margin rather than an active one. Along four widely-spaced transects, highly conducting lower crust appears to extend from the continental edge inland over 600 km beneath much of the Appalachians (Fig. 3-15; see review by Greenhouse and Bailey, 1981). Some specific details of this interpretation may be suspect because the models were derived from GDS data alone, but the general result of a highly conducting lower crust is probably correct. MT measurements in South Carolina (Young et al., 1986) show a conducting zone beneath the survey area, with the indication of a sharp rise in the depth to the layer from 20 km to some 5 km closer to the coast. The conducting layer is modelled to be of around 100 $\Omega\cdot\text{m}$. These results have to be followed up by more precise EM studies of the CLC beneath this and other passive margins.

3.3.3. Iapetus suture

A prominent and well known conductivity anomaly is the "Eskdalemuir anomaly" in the Southern Uplands of Scotland. This anomaly has been studied in Scotland using MT and GDS techniques since the mid-1960s (Jain, 1964; Edwards et al., 1971; Jones and Hutton, 1979a,b; Ingham and Hutton, 1982; Beamish and Smythe, 1986; Sule and Hutton, 1986), and most recently in Ireland by Whelan et al. (1990). The conductive anomaly is related to the Iapetus suture formed by the closure of the Iapetus — or Proto-Atlantic — ocean originally proposed to have existed by Wilson (1966) mainly from consideration of faunal realms. The Iapetus

*EMSLAB: ElectroMagnetic Study of the Lithosphere and Asthenosphere Beneath the Juan de Fuca plate.

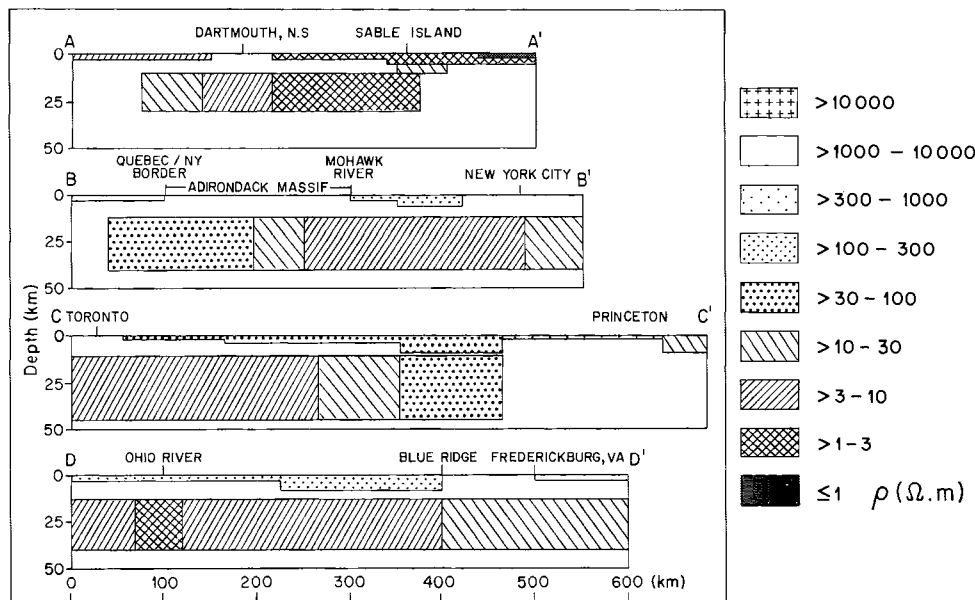


Fig. 3-15. 2D resistivity models for four profiles across the Appalachians (redrawn from Greenhouse and Bailey, 1981).

suture (Fig. 3-16) is believed to be represented in Britain by the Solway Line (Phillips et al., 1976; McKerrow and Soper, 1989a) and in Ireland by the Navan–Silvermines Fault (Phillips et al., 1976; McKerrow and Soper, 1989a), although the latter interpretation is debated (Murphy and Hutton, 1986; Hutton and Murphy, 1987; Harper and Murphy, 1989; McKerrow and Soper, 1989b). A variety of tectonic models have been proposed for the closure of the Iapetus Ocean, including:

- (1) northward-dipping subduction beneath the Midland Valley of Scotland along a line now marked by the Southern Uplands Fault (Garson and Plant, 1973);
- (2) southward-dipping subduction beneath northern England (Fitton and Hughes, 1970);
- (3) two subduction zones (Dewey, 1969), with a 14–16° oblique angle of convergence between them (Phillips et al., 1976); and
- (4) northward subduction initially located close to the present Southern Uplands Fault, but which migrated southwards with time, with subduction of continental crust from eastern Avalonia during the final stages (Legget et al., 1983; McKerrow and Soper, 1989a, and references therein). Northward subduction appears to be supported by BIRPS reflection data in the Irish Sea (Brewer et al., 1983; Beamish and Smythe, 1986) and in the North Sea (Klemperer and Matthews, 1987). These data imaged a reflective horizon dipping at 25–40° from 10 km below northern England to a depth of 25 km beneath the Southern Uplands. There appears to be a correlation between this seismic image and a northward-dipping conductive structure determined from three MT sites (Beamish and Smythe, 1986). However, updip projection of this reflecting horizon would intersect the surface in the Lake District, and thus McKerrow and Soper (1989a) consider it not to be the plate boundary but probably an intra-crustal shear zone.

Hutton and her colleagues have recorded MT data at over thirty locations in the Southern Uplands and have imaged several conducting zones in the upper and lower crust. The locations of these conducting zones are depicted in Figure 3-16 together with zones recently

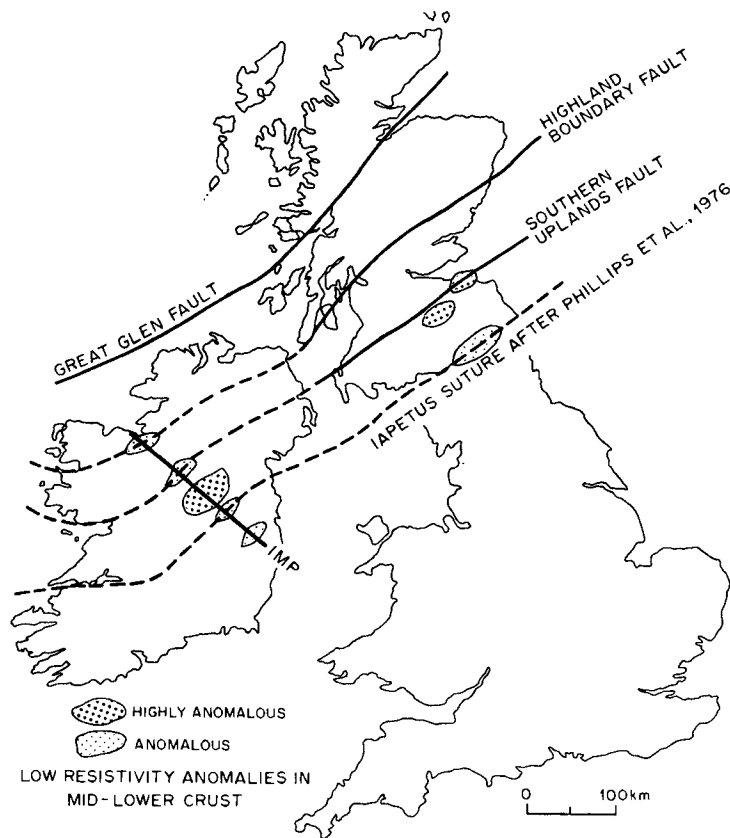


Fig. 3-16. Ireland and Britain showing the location of the two major tectonic features, the Southern Uplands fault and the Iapetus Suture, and the location of lower crustal conductivity zones (redrawn from Whelan et al., 1990).

identified in Ireland (Whelan et al., 1990 and references therein). The anomalous zone in the Northumberland Trough is the same as that identified by Beamish and Smythe (1986). The more major conductive features, labelled as “highly anomalous” zones on Figure 3-16, are all close to the surface (4–6 km in Scotland and 4 km in Ireland) and all exhibit the same geometry. The locations of these major zones also are in qualitative agreement with the induction arrows for both Scotland and Ireland (Edwards et al., 1971; Hutton and Jones, 1980). Interestingly, the trace of these zones is not parallel to either the proposed Iapetus Suture (Solway Line — Navan-Silvermines Fault) or the Southern Uplands Fault, but is oblique to them both at an angle of $\approx 5^\circ$. The geometry of these zones appears to support Phillips et al.’s (1976) hypothesis of oblique subduction along two zones.

3.3.4. *Trans-Hudson Orogen*

The enigmatic North American Central Plains conductivity anomaly (NACP, Fig. 3-17) has been mapped since the late 1960s using GDS array studies (cross-hatched shading) by Gough and others (see Alabi et al., 1975 and references therein; Handa and Camfield, 1984; Gupta et al., 1985) and was postulated to be an EM expression of a Proterozoic plate boundary by Camfield and Gough (1977). In regional terms their hypothesis linked a postulated

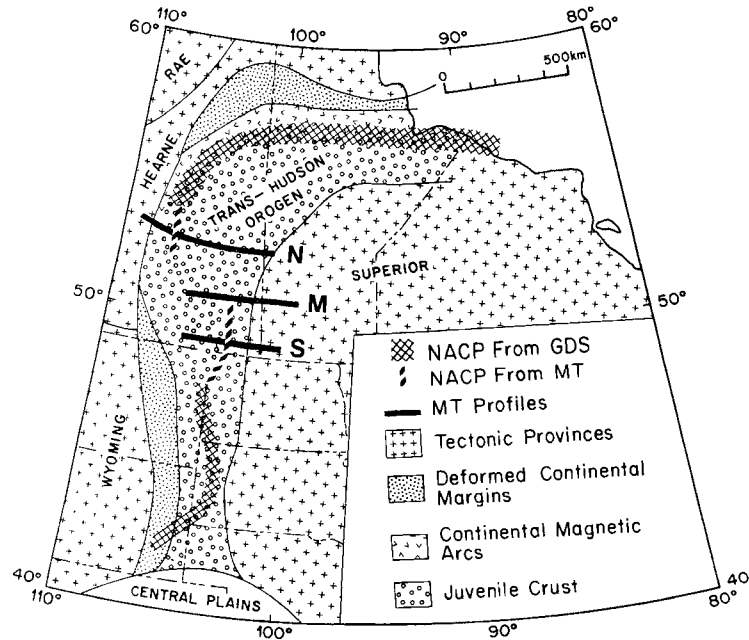


Fig. 3-17. Trace of the NACP from GDS and MT studies and Hoffman's (1988) tectonic map of North America (redrawn from Jones and Craven, 1990).

Proterozoic subduction zone in southeastern Wyoming (Hills et al., 1975) with a proposed suture in the Canadian shield (Gibb and Walcott, 1971), although the spatial correlations were not coincident. The NACP (Fig. 3-17) lies within an Early Proterozoic assemblage of juvenile crust between the Superior and Churchill (now Hearne and Rae) cratons termed the Trans-Hudson Orogen (Hoffman, 1981, 1988).

Recent MT studies in Canada over the NACP (Jones and Savage, 1986; Jones, 1988b; Jones and Craven, 1990; Rankin and Pascal, 1990) have imaged the structure with greater resolution than possible with the GDS data. They have shown that between 49° and 50° latitude the NACP is some 75 km east of the location indicated by the GDS studies, and that the NACP is not a continuous feature but exhibits offsets (Fig. 3-17; dashed shading of NACP); a feature also identified by Thomas et al. (1987) from the horizontal gravity gradient map of North America. Two-dimensional modelling of the MT responses (Jones and Craven, 1990) shows that the top of the structure is at a depth of about 9–10 km and that it is a zone of high conductivity, of $\rho < 1 \Omega \cdot \text{m}$, with a truncated asymmetric anticlinal geometry and predominantly westward dip (Jones and Craven, 1990). A single-body model proposed by Jones and Craven (1990) to explain the observations is illustrated in Figure 3-18. Two-dimensional smooth inversions of the data from line S in Figure 3-17 essentially substantiate the main features of this model (deGroot-Hedlin and Constable, 1990; J.R. Booker, pers. commun., 1991). A body of approximately equal shape with a positive density contrast of 270 kg/m^3 was shown to explain the locally observed gravity high coincident with the NACP (Jones and Craven, 1990). Furthermore, the trace of the NACP is coincident with a zone of anomalously high heat flow and with a magnetic quiet zone.

The high conductivities of the NACP cannot be explained in terms of fluids of 50 S/m conductivity (see §4.1.4) in a porous medium, as this would require implausibly high porosities

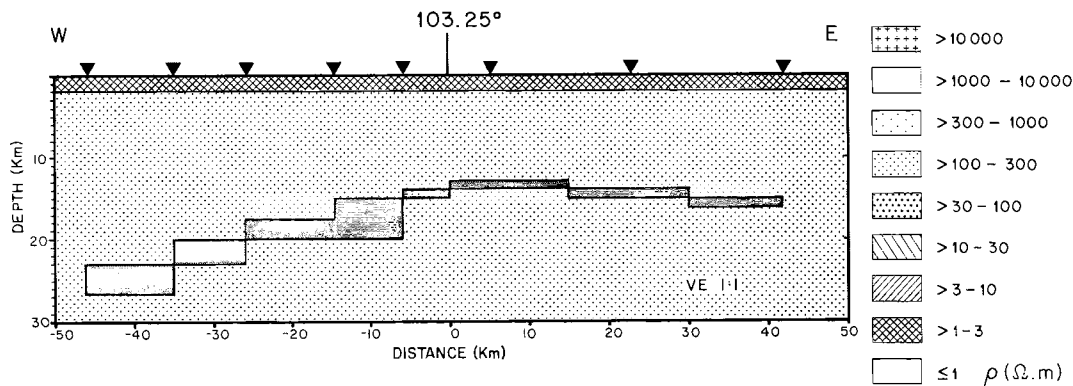


Fig. 3-18. A single body 2D model derived to explain the MT observations along profile S of Fig. 3-17 (redrawn from Jones and Craven, 1990).

of $>10\%$. Shortly after its discovery, Camfield et al. (1971) suggested that the NACP structure might be associated with conductive minerals such as graphite in schistose rocks. Camfield and Gough (1977) discussed the spatial correlation of the ends of the NACP anomaly with exposures of graphite in southeastern Wyoming and in northern Saskatchewan, although these correlations must be revised in the light of more recent studies in northern Canada. Resolution of other more subtle variations in the conductivity of the lower crust beneath the Trans-Hudson Orogen beneath the Phanerozoic cover is not possible due to the attenuating effect of the thick conducting sediments (3 km of $3 \Omega \cdot \text{m}$ material, i.e., $S = 1000 \text{ S}$, see §2.4.3).

3.3.5. Ladoga–Bothnian Bay–Skellefteå zone

By far the most interesting results from the Fennolara MT profile (§3.1.1) of Rasmussen et al. (1987) are from the Skellefteå region in northern Sweden, which is a mining district believed to be a relic of a Proterozoic island arc that separates Archean basement to the north from Svecokarelian rocks to the south. A significant EM anomaly (Storavan anomaly) in this region was first mapped by Jones (1981a) using the coarsely-spaced IMS magnetometer array. Various tectonic models have been presented for the region, including a northeast-dipping subduction zone. A high station density (average of 20 km separation compared to 40 km for the rest of the Fennolara profile, §3.1.1) ensured greater resolution of the Skellefteå structure, and the proposed 2D model to explain the observations is illustrated in Figure 3-19. Note that the conducting zone, of conductance 2000–4000 S, dips to the northeast with its base at the base of the crust. This model is similar to Jones and Craven's (1990) model for the NACP (Fig. 3-18), which is also associated with a Proterozoic collision zone of similar age (1.9–1.8 Ga). Very recent reflection profiling in the Bothnian Bay (the BABEL* experiment, BABEL Working Group 1990) shows enhanced reflectivity in the zone of increased conductivity mapped by Rasmussen et al. (1987).

Certain workers have postulated that the Skellefteå zone is an extension of the Ladoga–Bothnian Bay Zone (LBBZ, §3.1.1) into Sweden. The Finnish group has studied the LBBZ with magnetometer arrays, MT profiles around Oulu, and a 550 km profile (called SVEKA) of 62 MT sites in central Finland. The array studies (Pajunpää, 1984, 1986, 1987) confirmed

* BALTic and Bothnian Echoes from the Lithosphere.

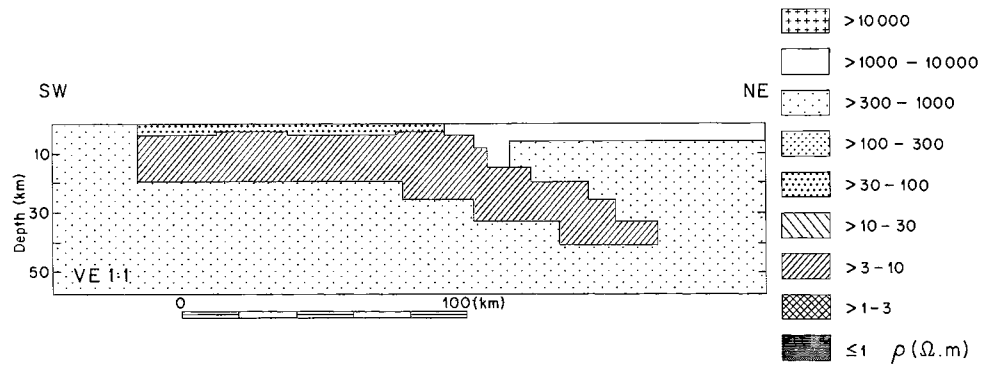


Fig. 3-19. A 2D model of the Skellefteå area in northern Sweden (redrawn from Rasmussen et al., 1987).

the existence of an anomaly around Oulu previously identified by Küppers et al. (1979), and revealed an important anomaly in southeastern Finland (the “Outokumpu” anomaly), which is along strike from the Ladoga conductivity anomaly to the northeast of Leningrad in the southeastern part of the shield (Rokityansky et al., 1981). The Ladoga anomaly was modelled as a thin sheet of anomalously high conductivity with a depth to top less than 10 km and with a dip to the north of total horizontal extent of the order of 40 km. This model was obtained from GDS measurements, which are sensitive to strong anomalies in Earth structure rather than the actual host earth structure itself. The body was modelled with a total longitudinal conductance (conductivity times thickness times length extent) of the order of 2×10^8 S·m, which for the proposed length extent of 45 km implies a conductance of the order of 4,500 S. MT measurements (Rokityansky, 1983) essentially confirmed the interpretation, although there were differences in some minor details. Interpretation of densely-spaced GDS and MT studies in the region of Oulu (Pajunpää et al., 1983; Korja et al., 1986) required a highly conducting feature (some $0.5 \Omega\cdot\text{m}$) very close to the surface (4–7 km), with a slightly less conductive deeper block ($5 \Omega\cdot\text{m}$, 12 km) to the southwest. This anomaly was thought to be due to Archean brines. An alternative model of the TM-mode only data was presented by Golubev and Varentsov (1989), who used an unconstrained inversion program, in which the conducting zone was 4 km thick, of resistivity greater than $10 \Omega\cdot\text{m}$, and total conductance less than 500 S. This exemplifies the need for both objective 2D inversion and for interpretation of all the data available, not just a limited subset.

Although interpretation of the SVEKA MT data has yet to be completed, initial results (Korja, 1990) indicate the presence of a conducting zone beneath the LBBZ which may be linked with the Ladoga and Storavan anomalies (Skellefteå zone). Such an elongated feature from northeast of Leningrad to northern Sweden was originally suggested by Rokityansky (1983), which he termed the “Transscandinavian anomaly”. Comparing Rokityansky et al.’s (1981) model with that of Rasmussen et al. (1987) (Fig. 3-19), there is excellent agreement with the approximate shape of the top of the dipping portion and with the vertical conductance (of the order of 4,000 S).

3.3.6. Continental margins: Conclusions

It is apparent that at various geological times and locations the process of subduction has emplaced bodies of enhanced conductivity within the middle and lower crust which can survive

for half the age of the earth. These bodies can be imaged by high-quality MT studies, and possibly by deep-probing controlled-source methods. Also, the seismic impedance contrast between these bodies and the material above them is such that there is a reflecting horizon associated with them. What is the nature of these bodies, and what causes their enhanced conductivity? Is this a ubiquitous feature of subduction? If so then the locations of all ancient active margins can be mapped using EM methods. If not, then what are the differences between the processes?

Beneath the passive margin of eastern North America there is a zone of enhanced conductivity also. What is the nature of this anomaly? Is it a ubiquitous feature of all passive margins? Certainly both active and passive margins, in particular the east coast of North America, warrant further studies to elucidate these points.

3.4 "Generic" results

A number of compilations have been made of the resistivity of the CLC (e.g., Jones, 1981b; Shankland and Ander, 1983; Haak and Hutton, 1986; Ádám, 1987; Hjelt, 1988; Keller, 1989b; Schwarz, 1990) without regard for the quality of data or interpretations and, as such, these compilations could be misleading. However, as discussed in §2.4.3, one parameter that is reasonably well resolved in MT studies and that is reasonably robust (i.e., will not change greatly after re-measuring or re-interpretation) is the total conductance S of the CLC. If an MT apparent resistivity curve is static shifted by a factor of c , the estimate of the conductance of the CLC is reduced by $c^{1/2}$. Given that static shifts display a standard deviation of one-quarter to one-half an order of magnitude (Sternberg et al., 1985), statistically 95% of the estimates of S_{CLC} will be correct to within factors of 2–3, and will be both underestimates and overestimates.

Hyndman and Shearer (1989) normalized the results of the resistivities and thicknesses of conductive zones in the CLC presented in compilations of Shankland and Ander (1983) and Haak and Hutton (1986) for a 10-km-thick zone, and also classified them into Phanerozoic and Precambrian areas. Representing their histogram in terms of conductances (Fig. 3-20) it is apparent that there is a statistical difference between the average conductance for Phanerozoic and for Precambrian areas, with the mean for the former of 400 S and the mean for the latter of 20 S.

Jones (1981b) had previously compared seismic and electromagnetic CLC values, and had surmised that the Canadian Shield may be zoned with a central resistive "core" and a more

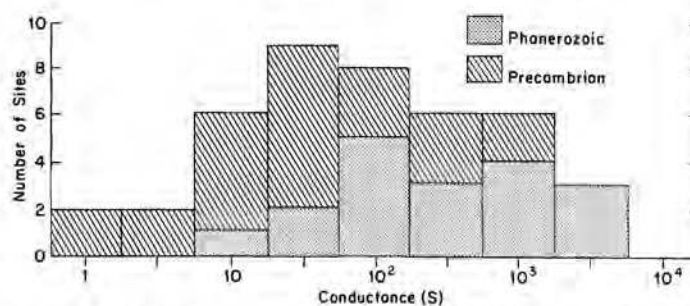


Fig. 3-20. The conductance of the continental lower crust compiled from a variety of studies and classified into Phanerozoic and Precambrian terrains (modified from Hyndman and Shearer, 1989).

conducting outer edge. This model must be modified in the light of more recent investigations, but nevertheless the lateral variation in resistivity of the CLC, even within an old shield region, is apparent.

3.5. Kapuskasing structure

The Kapuskasing uplift in northern Ontario, Canada, is of particular importance because it represents deep crustal material that has been exposed on the surface by tectonic thrusting along a fault with ramp-and-flat geometry.

Originally discovered in the late-1940s by its very large gravity anomaly (Garland, 1950), it has recently been the subject of intense geoscientific study as one of the LITHOPROBE targets. The 500-km-long structure strikes NE, lies between Lake Superior and James Bay, and consists of high-grade (granulite facies) metamorphic rocks. The latest tectonic models of the region, based initially on geobarometry and geothermometry studies and gravity modelling, included a lower crustal block upthrust from 30 km depth in the Proterozoic along a major thrust fault with attendant crustal thickening (Percival and Card, 1983, 1985). A short pilot reflection study appeared to confirm this model (Cook, 1985).

Accordingly, it was considered an excellent opportunity to address the question as to whether the conductivity of the lower crust is *positionally-dependent* or *compositionally-dependent*, i.e., what happens to the conductivity when CLC rocks are brought to the surface. The initial EM experiment was a GDS array study by Woods and Allard (1986), and the results indicated that the uplifted lower crustal section was even more resistive than the surrounding upper crust. This demonstrated that the location of the rocks in the crust is important, and was interpreted by Woods and Allard (1986) as indicating that intrinsic conditions at depth rather than mineralogical composition controlled lower crustal conductivity.

An MT study by Kurtz et al. (1988) over one part of the Kapuskasing structure showed remarkable lateral homogeneity. A virtually 1D model consisting of a 40,000- $\Omega\cdot\text{m}$ upper crust to 15 km depth overlays a 4,000- $\Omega\cdot\text{m}$ mid-crust to 25 km. This is underlain by a 100- $\Omega\cdot\text{m}$ lower crustal zone to 35 km and a 900- $\Omega\cdot\text{m}$ upper mantle (to 250 km). The model was not completely 1D; the conducting zone in the lower crust was laterally bounded with a width of 30–60 km and a strike of N65°E (Fig. 3-21). However, this model generates vertical magnetic field components that should have been observed by the GDS array study of Woods and Allard (1986). As these components were not observed, Kurtz et al. (1988) proposed a model that included electrical anisotropy (see §4.5). Perhaps of significance is that the electrical strike of N65°E is similar to a number of trends on the magnetic anomaly map of the region and to diabase dyke swarms of various ages (Kurtz et al., 1988).

A similar virtual-1D conductivity structure was interpreted by Mareschal et al. (1988) and Mareschal (1990) for MT data from a profile further north over a different part of the Kapuskasing structure. Controlled-source studies across the master thrust fault (Bailey et al., 1989; Kurtz et al., 1989) confirmed the highly resistive nature and lateral homogeneity of the upthrust lower crustal material, and located some conductive features — where conductive here refers to zones of resistivity of 5,000 $\Omega\cdot\text{m}$ — which were interpreted as due to brines in rocks with 0.5% porosity.

Seismic refraction and reflection studies led to modifications of Percival and Card's original tectonic model with the conclusion that, in fact, the lowermost crust is not exposed but that deep crustal levels (to 30 km) are brought to the surface along ramp-and-flat structures (Boland et al., 1988; Boland and Ellis, 1989; Percival et al., 1989; Geis et al., 1990), and that

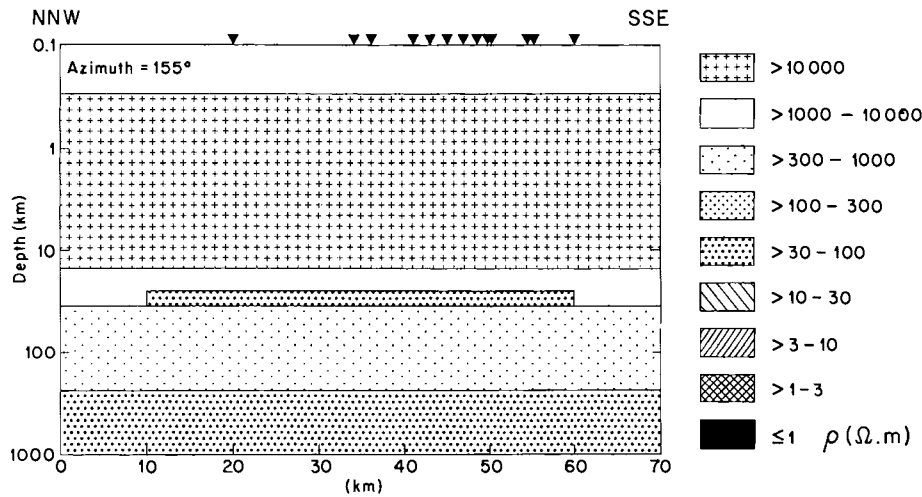


Fig. 3-21. A two-dimensional model of MT data observed across the Kapuskasing uplift structure in northern Ontario (redrawn from Kurtz et al., 1988).

there exists a decollement in the mid-crust (≈ 20 km in Percival et al., 1989 and ≈ 12 km in Geis et al., 1990). The lowermost crust is not actually uplifted and exposed, but upper lower crustal levels are. Thus, the conducting regions in the lowermost lower crust (Fig. 3-21) detected by Kurtz et al. (1988) and Mareschal et al. (1988; Mareschal, 1990) remain hidden from inspection.

The proposed crustal thickening was verified by the seismic studies, and it is of interest to note that the $100 \Omega\cdot\text{m}$ lower crustal block in Kurtz et al.'s model (Fig. 3-21) spatially correlates with the depression in the Moho, although the orientations of the two features are not equivalent ($N45^\circ\text{E}$ for the Moho depression, $N65^\circ\text{E}$ for the conductivity anomaly).

3.6. Valhalla complex

The Valhalla gneiss complex is a metamorphic core complex located in southeastern British Columbia which has been the subject of multidisciplinary geoscientific studies as part of the LITHOPROBE Southern Cordillera transect. Seismic reflection studies (Cook et al., 1987, 1988; Eaton and Cook, 1990) over the complex imaged a shear zone in the uppermost crust that was interpreted as bounding the complex on its western edge and being structurally related to the Slocan Lake Fault, which bounds the complex to the east. Beneath this shear zone is a major reflector, the Valhalla reflector, which is associated with a shear zone exposed to the north.

Other than these features, the seismic data show moderate reflectivity for the upper 8-s two-way travel time (TWT), after which there are few reflections to about 12 s where a weak band of events is probably associated with the Moho. This latter TWT, equivalent to ≈ 36 km, corresponds with a crustal thickness estimate of 35 km from a seismic refraction experiment (Cumming et al., 1979). There can be no question that the seismic energy was sufficient for sampling the lower crust; Moho reflections, although weak, were recorded.

Data from two MT stations on the Valhalla complex were reasonably 1D for periods that sampled the crust, and the conductivity model that matches the data has a highly resistive uppermost layer of $\approx 1,000 \Omega\cdot\text{m}$ to a depth of ≈ 9 km below which there is a less resistive zone

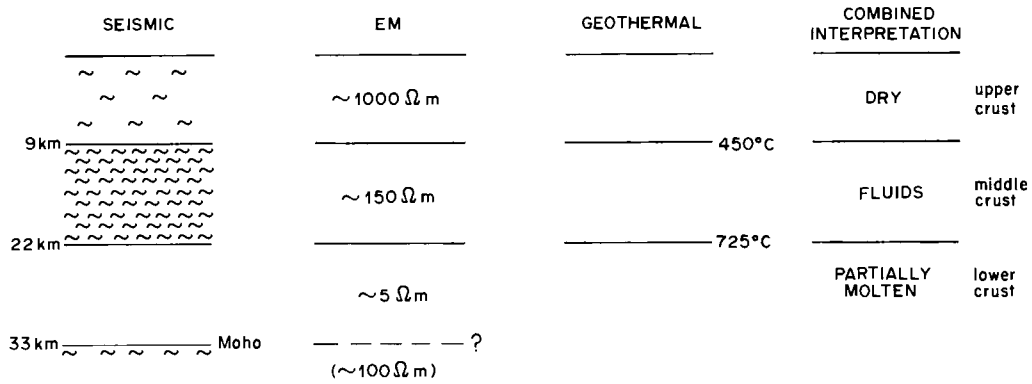


Fig. 3-22. A combined interpretation for the Valhalla gneiss complex showing a cold, highly resistive, virtually unreflective uppermost crust to ≈ 10 km, a warm, somewhat conductive, reflective mid-crust to ≈ 22 km, underlain by a hot, highly conducting, unreflective lower crust to the Moho at ≈ 33 km.

of $150 \Omega \cdot \text{m}$ down to 19 km, underlain by a highly conducting lower crust of $5 \Omega \cdot \text{m}$ (Jones et al., 1987). That all four ρ_a curves from the two stations lie on or close to one another is an indication that there are no severe static shift problems associated with these data. The resolution of this model is good with standard errors on the layer parameters of the order of 5–10%. Lewis (1990) and Lewis et al. (1992) recently presented a geothermal model for the complex in which he concluded, from analyses of the heat generation, that temperatures of 450°C are reached at a depth of 12 km and 730°C at 21 km. In summary, the upper crust (depths < 10 km) is moderately reflective, highly resistive (thousands of $\Omega \cdot \text{m}$), and cold ($< 450^\circ\text{C}$). The mid-crust (10–22 km) is reflective, moderately conductive (hundreds of $\Omega \cdot \text{m}$), and warm (450 – 730°C). The lower crust (> 22 km) is non-reflective, highly conductive ($\approx 5 \Omega \cdot \text{m}$), and hot ($> 730^\circ\text{C}$). This crustal stratification into three zones is shown in Figure 3-22. Lewis's (Lewis, 1990; Lewis et al., 1992) interpretation of this stratification involves a "dry" upper crust, a "wet" mid-crust, and a partially molten lower crust.

3.7. Results: Conclusions

It is apparent from the above information that it is not possible to construct a generic model of the conductivity of the CLC. Detailed investigations illustrate that there are many anomalous conductivity features within the CLC, and given that EM data are sensitive to high-conductivity anomalies rather than featureless zones, any global compilations are prone to bias and should be treated cautiously — every locale must be considered unique unless otherwise shown differently.

Notwithstanding these remarks, there is the strong indication that at many localities the electrical conductivity increases sharply at mid-crustal depths (15–20 km). Whether these zones of enhanced conductance are typically thin and highly conducting, or thick and moderately conducting, is unknown at this time. Also, there is the suggestion that the CLC becomes more resistive with age (§3.4), and that perhaps shields that are sufficiently large may have a central, cold resistive core surrounded by an edge displaying more transitional geophysical parameters (Jones, 1981; Keller, 1989b). Whether there is an age dependency or a function of proximity to the edge, where processes such as fluid infiltration may be controlling the electrical conductivity, should be determined by EM studies on shields of varying sizes.

4. Causes of enhanced electrical conductivity in the continental lower crust

The high electrical conductivity of the lower crust observed in many parts of the world is an enigma that needs to be addressed. During the 1960s and 1970s, any structure within the CLC of resistivity less than thousands of $\Omega\cdot\text{m}$ was considered “*anomalous*” because it was known from laboratory measurements that dry rocks at the appropriate temperatures and pressures have resistivities in excess of $10^4 \Omega\cdot\text{m}$ (Brace, 1971). More recent work on the resistivities of dry rocks that are possible lower crustal material yield values in excess of $10^4 \Omega\cdot\text{m}$ at mid-crustal temperatures (400°C) decreasing to $10^2 \Omega\cdot\text{m}$ for silicic rocks and to $10^3 \Omega\cdot\text{m}$ for mafic rocks at 800°C temperature (Kariya and Shankland, 1983).

Recent compilations of lower crustal resistivities from various parts of the globe (§3.4) illustrate that *resistive* CLCs (resistivities in excess of some thousands of $\Omega\cdot\text{m}$) are the exception rather than the rule, and that resistivities of hundreds of $\Omega\cdot\text{m}$, with the top of the zones in the mid-crust, are more common. Thus, according to Haak and Hutton (1986) we should now consider that the CLC is *normally* within the range 10^2 – $10^3 \Omega\cdot\text{m}$, and that values above and below this range can be classified as anomalously high or anomalously low respectively. Such a classification was also adopted by Keller (1989b) in his zonal subdivision of the continental United States. As stated in §2.4.3, it is virtually impossible with 1D models to differentiate between a thin zone of high conductivity and a thicker zone of lower conductivity. Thus, many of the zones interpreted to be of 300 $\Omega\cdot\text{m}$ and 20 km thickness *could*, in fact, be 30 $\Omega\cdot\text{m}$ and only 2 km thick, for example. Accordingly, for the present the CLCs should be classified by their conductance values *not* their interpreted resistivities. This has the additional attraction that static shifts of the apparent resistivity curves only bias the estimates of the conductances of the CLC by the square root of their multiplicative factors, whereas they bias the estimates of the resistivities of the zones by the factors themselves. Thus, whereas the definitions of the CLCs into “*normal*” and “*anomalously high*” based on the resistivities is suspect, a categorization in terms of S_{CLC} (Fig. 3-20) may have some meaning, as shown (indirectly) by Hyndman and Shearer (1989). Obviously, one first-order question that needs to be addressed by the EM community is the thickness of the conductive zone within the CLC. MT methods are not likely to provide the answer and specifically-designed controlled-source EM experiments must be undertaken.

What are the causes of the observed enhanced conductivities? Many have been proposed during the last two decades but consideration will be given here to the four most likely candidates: (1) saline fluids; (2) carbon grain-boundary films; (3) conducting minerals; and (4) partial melts.

During the mid- to late-1970s, it was common to interpret enhanced conductivity as due to hydrated minerals, especially serpentine or serpentinized rocks (e.g., van Zijl, 1977; Connerney et al., 1980; Jones, 1981b; de Beer et al., 1982) as laboratory studies had shown that serpentinized rocks exhibited high electrical conductivity (Stesky and Brace 1978). However, in an experiment in which the outgasses were carefully controlled, Olhoeft (1981) showed that at lower crustal conditions a hydrated hornblende schist with structural water had an electrical resistivity very similar to that for dry granite (Fig. 3-23). This illustrated that structural water appears to have no significant effect on electrical properties, contrary to the comments by Lee et al. (1983). For this reason, compositional contrasts, as advocated most recently by Baker (1990) to explain the conductivity of the CLC, are not considered here.

Duba (1976) questioned whether interpretations of laboratory measurements on rock samples were meaningful given that it is impossible to recreate the P - T - t (pressure-temperature-

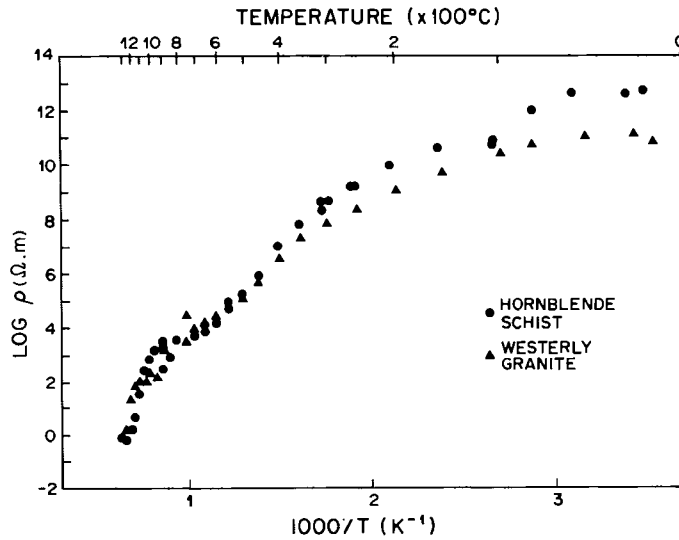


Fig. 3-23. Log electrical resistivity versus reciprocal temperature for Westerly Granite and hornblende schist. The two curves are essentially identical at lower crustal temperatures even though the hornblende has considerably more structural water than the granite (redrawn from Olhoeft, 1981).

time) conditions and environment of the CLC. In particular, simply wetting rocks is inappropriate because of the chemical reactions that take place, and, most importantly, the time element can never be reproduced. Given our understanding of the “ductile” nature of the CLC on geological time scales, this latter point is critical. It has been shown that holding a sample at certain P - T conditions for even different short lengths of time (1500 h instead of 500 h) can lead to an increase in the conductivity of albite by three orders of magnitude at critical temperatures just below the onset of partial melting (Piwinski and Duba 1974; Duba 1976).

4.1. Saline fluids

Saline fluids have been proposed as an explanation of lower crustal conducting anomalies since such features were first discovered (Hyndman and Hyndman, 1968; Brace, 1971; Parkomenko et al., 1972). The dramatic effect that even small amounts of saline fluids have on the electrical conductivity is well illustrated in Figure 3-24, and many workers have appealed to the presence of aqueous fluids with a high ionic content to explain the conducting CLC (Shankland and Ander, 1983; Lee et al., 1983; Haak and Hutton, 1986; Jones, 1987; Vanyan and Shilovski, 1989). A model proposed to explain both the observed enhanced electrical conductivity and the increased seismic reflectivity presented by Jones (1987) consists of a thin zone of interconnected fluids trapped beneath an impermeable zone, which is at a temperature of the order of 400°C (Fig. 3-25). At greater depths, where the temperatures are in excess of about 600°C, the resistivity increases because the fluids are no longer interconnected. This model is based upon the fluid circulation model of Etheridge et al. (1983, 1984) shown in Figure 3-26. Field evidence for the growth of an impermeable zone sealed by mineral deposition at a temperature of 370°C is given in Fournier (1991). Jones’ model (1987) was examined by Hyndman (1988), who considered the petrological consequences in terms of metamorphic facies. Hyndman (1988) showed that the temperature range 400–700°C covers greenschist to amphibolite facies conditions, for which petrological studies indicate

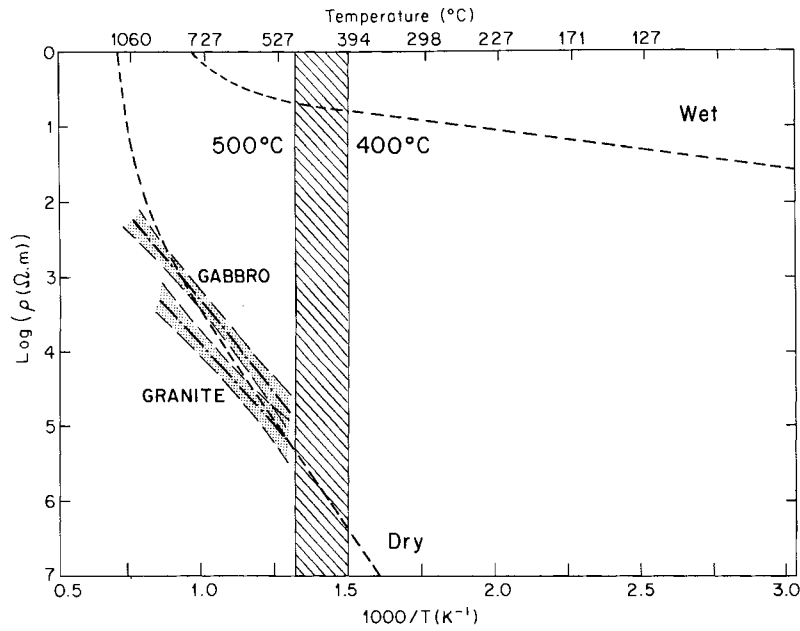


Fig. 3-24. Log electrical resistivity versus reciprocal temperature for dry granite and gabbro compared to their wet counterparts. (redrawn from Shankland and Ander, 1983). Note that in the temperature range believed to be appropriate for the upper part of the lower crust (400–500°C; shaded region) there is five orders of magnitude difference in resistivity.

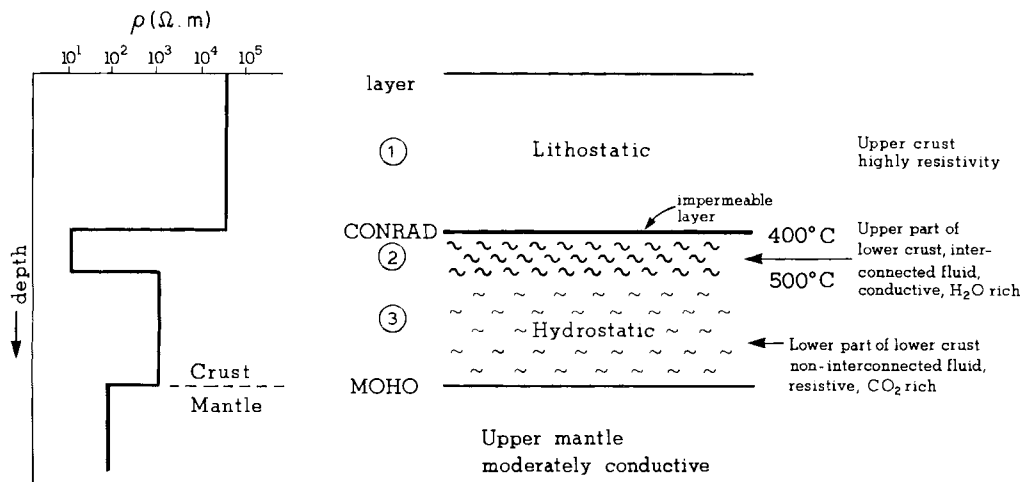


Fig. 3-25. A model for the continental crust with a zone of interconnected H₂O-rich fluids trapped beneath an impermeable layer (the seismic “Conrad” discontinuity?) at mid-crustal depths at a temperature of 400°C. At deeper depths within the lower crust the temperature is greater, the fluids are CO₂-rich, and the resistivity is greater (modified from Jones, 1987). The difference in pressure regime, with the upper crust at hydrostatic pressure and the lower crust at lithostatic, is possibly an explanation for the highly resistive nature of the upper crust.

chemical equilibrium with free saline water at close to lithostatic pressure.

A number of questions arise when a “wet” lower crust is postulated: (1) Do the fluids exist? (2) What are the fluids made of? (2) Do they stay there? (3) Do they move? and (4) What is the likely conductivity of the saline fluids?

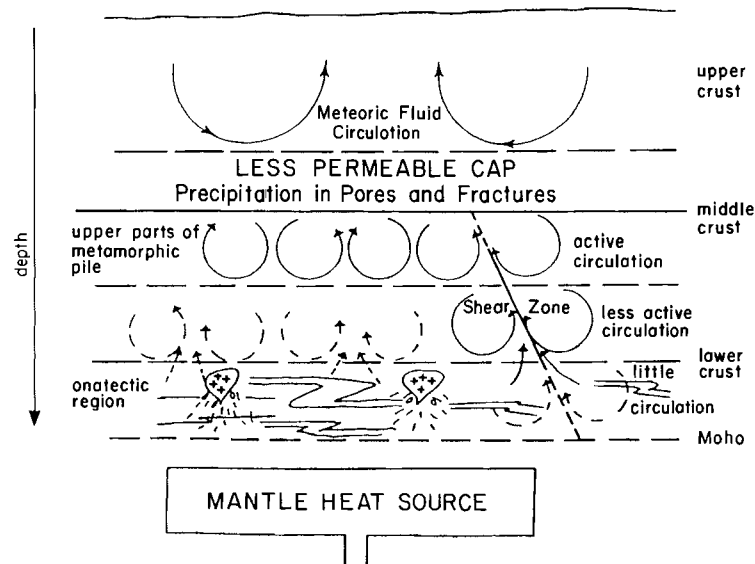


Fig. 3-26. A model of the fluid circulation within the continental crust with the development of an impermeable cap at mid-crustal depths (redrawn from Etheridge et al., 1983).

4.1.1. Existence and types of fluids

The existence of fluids within the deep continental crust is a topic that provokes intense debate within much of the geoscientific community, and recent workshops have focussed on the general themes of crustal fluids and fluid movements (see Bridgwater, 1989; Fountain and Boriani, 1990; Langseth and Moore, 1991; Torgersen, 1991).

In certain geographical areas there is general acceptance that fluids are present at least locally. This is particularly so in active subduction zones where it is well-known that at depths less than 40 km there is a large amount of free water available from expulsion of pore waters and from $\text{CH}_4\text{-H}_2\text{O}$ fluids produced by diagenetic and low-grade metamorphic reactions (Peacock, 1990). At greater depths, H_2O and CO_2 are released by metamorphic reactions in the subducting oceanic crust. It is estimated that the transport rate of H_2O down subduction zones is of the order of 10^{15} g/yr (Fyfe et al., 1987). These fluids may have been detected using the MT method in two studies of the subducting Juan de Fuca plate (§3.3.1).

Is there evidence for free fluids in stable regimes?

There appears to be acceptance by some geologists and geochemists since the mid-1970s that fluids can exist in the CLC and that they account for a variety of metamorphic and petrological processes. It has been proposed that in certain cases magma generation in the deep crust by crustal anatexis requires the ingress of external fluid (Wickham and Taylor, 1987). In his review of the evidence for fluid flow during regional metamorphism, Rumble (1989) emphasized the advances that are currently taking place in metamorphic petrology due to an understanding of the dynamic physical processes responsible for metamorphism such as heat flow, fluid flow and tectonic advection. As noted by Rumble (1989), compared to grain-boundary permeability for which fluid-rock ratios of 1:1 to 4:1 are required, fracture permeability requires fluid-rock ratios of 330:1 for quartz to 600:1 for graphite. The presence of hydrothermal graphite in a sillimanite zone in New Hampshire is evidence

for pervasive fluid flow throughout the 2000 km² of the studied zone (Rumble et al., 1986), and Rumble (1989) tentatively concluded that there is significant intergranular permeability at depths of 20–25 km.

One enigma is the apparent paucity of data for inferring the existence of fluids in the CLC (Yardley, 1986a). This is discussed further below (§4.1.5).

What types of fluids are there at various crustal depths?

Studies of fluids in the C-O-H system on metamorphic rocks from the Grenville Orogen and the Alps are interpreted as suggesting a three-layer stratification of the crust: a methane-rich upper crust, a H₂O-dominant mid-crust, and a CO₂-dominant lower crust (Frey et al., 1980; Turek, 1987; Skippen, 1988). This is consistent with Frost et al.'s (1989a) hypothesis that magmas may form conduits by which mantle CO₂ is transported into the lower crust while H₂O is moved to higher levels from the lower crust. The temperatures inferred from microprobe analyses of the Alpine rocks are 450–550°C for the mid-crust, and 600–800°C for the lower crust (Frey et al., 1980), which is consistent with CO₂ being less soluble in silicate melts than H₂O leading to CO₂-saturated melts crystallizing at higher temperatures than those that are H₂O-saturated (Frost et al., 1989a). H₂O-rich fluids cannot exist in the deep crust at temperatures in excess of 650°C because silicate melts would replace them (Newton, 1990), and beyond even 500°C wollastonite crystals begin to grow by which process CO₂, released by the metamorphic reactions, begins to dilute the H₂O (Rumble, 1989). That crustal fluids at high temperatures are H₂O-poor and CO₂-rich was also interpreted from fluid inclusion studies of the Laramie Anorthosite Complex (Wyoming) which crystallized at around 950°C (Frost and Turek, 1989). These mid-crustal temperatures of 450°C are close to those proposed by Jones (1987) for the thin crustal conductor, and the fluid consequences are included in the model of Figure 3-25.

4.1.2. Retention/replacement of fluids

One objection to the proposal that saline fluids are the cause of high conductivity in the CLC is, in the absence of a recharge mechanism, the apparently unreasonably long retention times required. Etheridge et al. (1984) suggested that there exists an impermeable cap in the mid-crust beneath which fluids are trapped. Such an impermeable cap is considered to be untenable because the pressure at the top of a sealed lithostatic reservoir will soon exceed the local lithostatic pressure and hydraulic fracturing will break the cap (Yardley, 1986b). Subsequently, the trapped fluids will escape and migrate upwards (McCaig, 1988). This then raises the question of the rate of fluid migration. A recent theoretical study (Bailey, 1990) showed that the retention time is critically dependent on crustal temperature. For 1% initial porosity, fluids at 30 km depth would take 100 years and 3×10^8 years to rise to the surface for crustal temperatures of 800°C and 400°C, respectively. Assuming an average crustal temperature of 400°C yields retention times of 0.1–1 Ga. Degassing rates of ⁴He from depths of 40 km are also consistent with a 1-Ga-time scale (Torgersen, 1989). Also, once the cap is broken will all the fluid escape or will the cap re-seal after loss of sufficient fluid to reduce the pressure to lithostatic as discussed by Fournier (1991)?

Alternatively, the fluids could be continuously recharged from the mantle, as proposed by Haak and Hutton (1986). This poses the question of why the upper crust is so resistive if fluid is passing up through it. Gough (1986) proposed that the electrical state of the crust depends on the different stress regimes. In his (Gough, 1986) generalized model of the continental

crust, the upper crust is electrically resistive, seismically transparent and brittle. In contrast the lower crust is conductive, reflective and ductile. Gough suggested that saline fluid permeating the whole crust is under compressive stress in the upper crust, and thus is in separated cavities, whereas in the lower crust the fluid forms interconnected films on the crystal surfaces.

4.1.3. Large-scale fluid circulation

Fluid-driven mass transport in the crust is a necessary prerequisite for emplacement of ores in faults and fractures, and fluids from the surface are often interpreted to convect to great depths (10–20 km) within the continental crust (see McCaig, 1989, and review by Torgersen, 1990). Seismic pumping is thought to be a mechanism by which the fluids are brought to the surface along wrench faults (Sibson et al., 1975).

Crustal fluid circulation models have been presented by Etheridge et al. (1983, 1984) in which there is depth-dependent circulation rates beneath an impermeable cap that is sealed by precipitation in pores and fractures (Fig. 3-26). This model has not been generally accepted (Wood and Walter, 1986; Valley, 1986; Yardley, 1986b), but other evidence exists to support large-scale fluid circulation. For example, ^4He degassing fluxes suggests that the whole crust degasses on a 1-Ga time scale. This is possibly the first observational support for a grain-boundary diffusion process that operates on crustal/continental scales (Torgersen, 1989, 1990). Furthermore, studies of Lewisian gneisses in NW Scotland were interpreted to indicate that fluid circulation along cracks had been active for over 2 Ga (Hay et al., 1988). These studies support the concept of rapid, ubiquitous and large-scale mass transport by fluids in the Earth's crust (Torgersen 1989, 1990).

Obviously, fluid movement will be at two rates. It will be very slow, of the time scales given by Bailey's (1990) calculations, in a relatively unfractured rock matrix. Once fluids enter a fracture/fault system, however, they will move rapidly. Such systems can exist at the frictional-quasiplastic ("brittle-ductile") transition, above which fluid pressures are hydrostatic but below which are lithostatic (McCaig, 1988) (Fig. 3-25). This is an alternative explanation to Gough's (1986) for the high resistivity of the upper crust: fluids cannot exist in the upper crust for long geological times and are quickly expelled.

4.1.4. Fluid conductivity

Pore geometry (§4.6) is obviously an important factor when considering "effective porosity". As important, however, for interpretation of EM data is an accurate assessment of the fluid conductivity, σ_f . Hyndman and Shearer (1989), using data from Quist and Marshall (1968) and Uçok (1979), illustrated the effect of increasing salinity on fluid resistivity at 500°C and 4 kbar (Fig. 3-27), and adopted a fluid conductivity of 50 S/m (resistivity of 0.02 $\Omega\cdot\text{m}$), equivalent to an NaCl concentration of 5%, for their calculations. As stated by Hyndman and Shearer (1989), the values from Figure 3-27 should be considered *maximum* values of resistivity because of the necessary extrapolation from 4 kbar to 6–10 kbar pressure appropriate for the CLC.

However, fluid inclusion studies from Alpine thrust faults show typically 20–25 wt.% dissolved salts (Grant et al., 1991), from deep-penetrating shear zones up to 40 wt.% NaCl (S. Tempest, reported in McCaig, 1989), and from the Laramie Anorthosite Complex indicate 50–60 wt.% NaCl (Frost and Toubert, 1989). The occurrence of highly saline waters at depth beneath the Canadian Shield has long been known from mines (Lane, 1914), and borehole

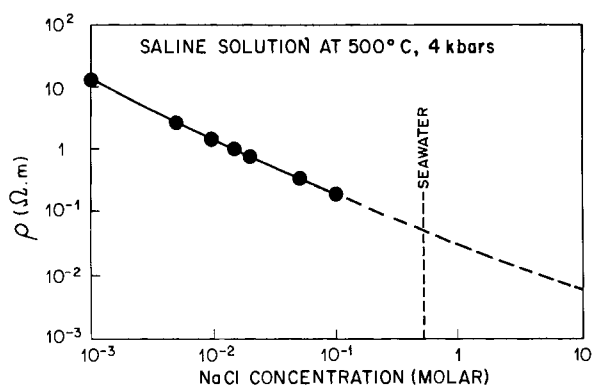


Fig. 3-27. Resistivity as a function of NaCl concentration at 500°C and 4 kbar pressure (redrawn from Hyndman and Shearer, 1989).

studies show increasing salinity of the groundwater with depth, with salt concentrations in excess of 30% by weight from depths of 1.5 km (Frape and Fritz, 1987). These borehole data are interpreted by Frape et al. (1984) to indicate the existence of a common source brine at great depth beneath the whole Canadian shield, with total salt concentrations of ≈ 40 wt.% in the parent brines (Gascoyne et al., 1987). The origin of the salinity is still debated, but theories are either for an allochthonous origin, such as sedimentary basin brines or modified Paleozoic seawater, or an autochthonous origin, such as leaching of fluid inclusions, grain boundary salts or weathered mineral phases (see Frape and Fritz, 1987).

From Figure 3-27 it is apparent that an order of magnitude increase in salinity concentration produces approximately an order of magnitude decrease in the electrical resistivity of the solution. This, in turn, will reduce the thickness required to produce observed conductances by an order of magnitude.*

4.1.5. Petrological arguments against free fluids

The interpretation of the observed enhanced electrical conductivity being due to aqueous fluids is not accepted by some petrologists who argue for a dry lower crust. Their thesis, summarized succinctly by Yardley (1986a), is that the rocks which predominate within the CLC are granulite facies assemblages. Such rocks are only stable in the presence of fluid at the extreme metamorphic formation temperatures of 700–800°C, whereas the temperature at the top of the zone of enhanced conductivity in the mid-crust is believed to be around 400°C. Water penetrating granulites at these lower temperatures are consumed by hydration reactions to form lower-grade hydrous minerals until the granulite facies assemblage is entirely retrograded. These retrograde reactions should be detected in the CLC rocks exhumed. Accordingly, areas undergoing active metamorphism or plutonism will be full of fluids, but large-scale fluid flow should not occur in stable regions (B.R. Frost, pers. commun., 1990). Thus, the fluid explanation is acceptable for areas of active subduction or rifting, but from these petrological arguments it is thought to be invalid for the CLC beneath stable regions.

It is interesting to note the results of Hulsebosch and Frost (1989), which suggest that large quantities of saline brines penetrating into the deep crust down zones of ductile deformation

*Note added in proof: Recent work by Nesbitt (1992) shows that the resistivity decrease asymptotically approaches $\approx 10^{-2}$ with increasing salt concentration, rather than the simple increase shown in Fig. 3-27.

do not produce major chemical modifications. If this is so, then perhaps it is not surprising that the petrology of rocks from the CLC brought to the surface does not indicate the existence of saline brines.

Sanders (1991) proposed that the petrological evidence for a dry predominantly granulitic lower crust, and the proposal that fluids are the cause for the enhanced electrical conductivity, can be resolved by considering the lower crust to consist of “kilometre-scale granulite-facies lenses (mega-augen) separated from each other by a system of anastomosing, gently-dipping brine-soaked ductile shear zones”. The brine in the shear zones is postulated to be stable, and that retrogression into the mega-augen fails because the hydration reactions on the margins form an impermeable seal preventing further ingress of water. Sanders (1991) details the observations that are consistent with this model, but concludes that the model raises many questions which can only be satisfactorily addressed when the interaction of brine and water-deficient rock is better understood.

4.2. Grain-boundary films

The most recent explanation for observed high conductivities concerns films of carbon on grain boundaries (Duba et al., 1988; Frost et al., 1989b). Duba and Shankland (1982) proposed that grain-boundary carbon was a possible cause of a high-conductivity layer in the upper mantle and Jödicke (1985) had suggested that carbon on the grain boundaries was the cause of the low resistivity of black shales. Fluid-inclusion and Raman studies on rocks from the 1.4 Ga Laramie Anorthosite Complex indicate that the rocks crystallized at pressures of 3 kbar and temperatures of 950–1030°C in the presence of graphite and a nearly-pure CO₂-CO vapour phase (Frost et al., 1989b). Graphite films of $\approx 1,000$ Å thickness were observed using Auger spectrometry on the grain boundaries, and Frost et al., (1989b) suggest that an interconnected carbon film may explain the high conductivities of the CLC.

This phenomenon appears to be depth-dependent: at depths where the temperatures exceed rock crystallization temperatures, CO₂ is excluded as a candidate because it is an insulating liquid and a non-polar gas, and thus has little or no effect on electrical properties (Olhoeft, 1981; Kay and Kay, 1981). Equally, rocks crystallizing at shallow depths may not reach graphite saturation (Frost et al., 1989b). Also, as shown by Frost et al. (1989b) there is a grain-size dependence of the expected conductivity that is given approximately by $\log[\sigma] = -4 - \log[x]$, where σ is in S/m, and x is the grain size in metres. This converts to $\rho = 10^4 x$, where ρ is the resistivity in $\Omega \cdot m$. Small-grain rocks (<1 cm) are conductive ($\rho < 100 \Omega \cdot m$), whereas large-grain rocks (>10 cm) are resistive ($\rho > 1000 \Omega \cdot m$). Such a grain-size dependence was previously noted by Kariya and Shankland (1983) in their statistical study of the conductivity of dry lower crustal rocks. In the ductile conditions of the lower crust one would expect that large-sized grains would predominate.

Given the petrological arguments advanced against fluids as a cause of enhanced conductivity in stable regions, EM practitioners are beginning to appeal to the explanation that grain-boundary films of graphite are responsible, e.g., Mareschal (1990), Haak et al. (1991) and Jödicke (1991). For some zones of enhanced conductivity for which there exists careful corroborating evidence this explanation may be acceptable. However, certain questions must be addressed if we are not to repeat the same mistake that we made with serpentine in the 1970s (see §4) by interpreting ubiquitously such regionally-widespread zones in the CLC as due to interconnected grain-boundary thin films over many hundreds of kilometres.

4.2.1. Wetting angle paradox and carbon mobility

Studies of the dihedral wetting angle θ in the presence of C-O-H fluids indicate that the angle will be too great for wetting by pure CO₂ or pure H₂O (Watson and Brenan, 1987; Yund and Farver, 1990; Lee et al., 1991). This has the consequence that, in the absence of existing fractures, fluid transport in such systems is limited to transient events during which the fluid pressure exceeds the confining pressure and crack propagation occurs. However, introduction of only small amounts of solutes (e.g., NaCl) to H₂O causes a substantial reduction in θ with consequent wetting along all grain boundaries and possible pervasive fluid transport. If this is the case, why is graphite observed by Frost et al. (1989b) along the grain boundaries and not just at the grain junctions? Perhaps the fluids present are not either pure CO₂ or pure H₂O, but a mixture of the two. However, as discussed previously (§4.1.1) fluid stratification into a H₂O-rich middle crust and a CO₂-rich lower crust is indicated, with little mixing likely, and Frost et al. (1989b) remarked that their rocks crystallized in the presence of a nearly-pure CO₂-CO vapour phase.

Laboratory studies of the mobility of carbon in olivine (Tsong et al., 1985) and in dunite (Watson, 1986) grain boundaries illustrated that carbon is sufficiently immobile as to preclude development of a continuous carbon film by diffusion from an initially localized source. With diffusivities of the order of 10^{-10} cm²/s, carbon would move only a few metres in 200 million years (Watson, 1986), from which one must conclude that regional continuity over hundreds of kilometres of a carbon film derived initially from localized sources of carbon cannot be achieved. Thus, the existence of a CO- and/or CO₂-rich fluid that had previously infiltrated the region in question must be postulated to explain the existence of a hypothetical carbon film.

4.2.2. Untestable hypothesis?

One unfortunate aspect of Frost et al.'s (1989b) proposal is that the hypothesis is possibly untestable. It is argued that when lower crustal rocks are tectonically exhumed, decompression destroys the continuity of the fine graphite film along the grain boundaries, leading to an increase in the resistivity of the rock. In addition, laboratory studies are thought to be of little consequence because the discontinuous film cannot be reconnected under laboratory conditions, thus making it impossible to observe the high conductivity either in the field or in the laboratory.

As a possible example of this conundrum, measurements on graphitic pelites from northern Canada, which were of such high graphite content that they were black in colour, gave values in excess of 10^5 Ω·m (Camfield et al., 1989). This is explained by Frost et al. (1989b) as destruction by fracturing of the grain-boundary films, and that the relatively coarse individual grains seen are unimportant for conductivity-enhancement as they are not interconnected.

4.2.3. Black shales

Jödicke (1985) suggested that carbon films on grain boundaries were the source of the low resistivity (2–3 Ω·m) of black shales. Duba et al. (1988) undertook laboratory studies on Carboniferous rock samples abundant in black shales, the colour of which is due to 5–8% organic carbon, extracted from a 5425-m-deep borehole in northern Germany. Rock analyses showed that there was no graphite present in the samples, and that the organic component

consisted of $\approx 95\%$ carbon. At high frequencies strong dependence of the electrical impedance with bedding direction and between dry and wet samples was observed. At frequencies of interest for deep studies (frequencies lower than 17 Hz in their experiments) the data were nearly frequency-independent. Their experiments showed that the conduction mechanism was controlled by an interconnected phase, and they concluded that they had confirmed Jödicke's earlier suggestion that the low resistivity is caused by a thin film of carbon on the grain boundaries.

When the samples were heated to 417°C in the absence of distilled water, their resistivities were observed to be frequency dependent and were very high ($> 10^6 \Omega\cdot\text{m}$ at 17 Hz). When saturated with distilled water they were observed to be frequency-independent and much lower ($10 \Omega\cdot\text{m}$) (Duba et al., 1988). After cooling back to room temperature it was observed that the carbon had oxidized and the interconnectivity was obviously broken. Pyrite, also an excellent conductor with a resistivity of $5 \times 10^{-5} \Omega\cdot\text{m}$, is stable to 743°C but it did not contribute to the low resistivities because there was no continuous contacts between the pyrite spheres.

Stanley (1989) suggested that metamorphosed black shales subducted to depths in excess of 20 km may be responsible for conductivity anomalies beneath the Carpathians and beneath the Cascade mountains of northwestern U.S.A. Given the observation by Duba et al. (1988) regarding the oxidation of the carbon film at temperatures in excess of 400°C, it appears unlikely that this explanation is valid for the whole depth extent (to 20 km) postulated for the Cascades conductor which lies coincident with the Cascades volcanic belt.

Jödicke et al. (1983) interpreted their thin conducting zones in the mid-crust beneath the Rhenish Massif as due to graphite-bearing metamorphic black shales of Precambrian age. Fluids were excluded because of the lack of evidence for large amounts of free water in the middle and lower crust from xenolith studies.

4.3. Sulphides/metallic ores/graphite

EM methods have been used by the exploration industry to locate highly conducting mineralized zones for over half a century. Many long linear zones of enhanced conductivity in the CLC (see Haak and Hutton, 1986, their Table 2) may be associated with mineralized zones. For the longest of these, the 3000-km NACP anomaly (§3.3.4), Camfield and Gough (1977) discussed a spatial correlation of its ends with known exposures of graphite in both southeastern Wyoming and northern Saskatchewan, and concluded that the structure is probably comprised of graphite sheets in highly metamorphosed and folded basement rocks. Although this correlation must now be revised, for at least the northern segment, the high conductivity of the NACP ($> 1 \text{ S/m}$) appears to preclude fluids as an explanation (Jones and Craven, 1990).

Olhoeft (1981) suggested that sulphides might be the cause of enhanced conductivity of certain zones. Conducting zones located by commercial MT studies in northern Montana and southeastern British Columbia were interpreted as due to sulphides in belt rocks; an interpretation that was confirmed by drilling.

4.4. Partial melt

Partial melt has been shown in laboratory studies to increase the conductivity of rock by orders of magnitude (Waff, 1974; Sato and Ida, 1984). Tyburczy and Waff (1983) showed that

conductivities are far less dependent on pressure than on temperature, and even at pressures as high as 25 kbar the conductivity approaches 70% of the zero pressure values.

4.5. Anisotropy?

Seismic anisotropy of the continental upper crust and upper mantle has been known for over a decade (Crampin et al., 1984), and evidence for electrical anisotropy in orthogonal horizontal directions of the oceanic upper mantle was recently presented by Chave et al. (1990).

Previous studies on land have also proposed transverse anisotropy to explain some observations (e.g., Jones and Garland, 1984; Kurtz et al., 1986b), but the anisotropy was confined to the upper crust and was “macro anisotropy” rather than intrinsic anisotropy. Two-dimensional models were found that satisfied the data but they included zones that had resistivities that, when averaged over a lateral distance of the order of the appropriate inductive scale length, differed in the two horizontal directions (see discussion of Cull, 1985, below).

Rasmussen (1988) concluded that the CLC of southern Sweden is electrically anisotropic (Fig. 3-10), with an anisotropy factor, given by the ratio of the two horizontal resistivities of the layer, of 42.5, or 4,250% (§3.1.1). Such an anisotropic model of the CLC has also recently been proposed by Kurtz et al. (1988) for the Kapuskasing structure in the Canadian shield (§3.5) and by Kellet et al. (1992) for the Abitibi belt. Given our generalized rheological model of the continental crust with a “ductile” lower crust that cannot withstand long-term stress, it is difficult to understand how a layer deep within the crust could maintain intrinsic anisotropy over long geological times. Is this proposed electrical anisotropy evidence for 1D intrinsic micro-anisotropy within the CLC with a preferential fluid-filled crack orientations? Or is it evidence for 2D macro-anisotropy with a conductive CLC (of ρ presumably $\approx 400 \Omega\cdot\text{m}$) intersected by numerous resistive dykes (of ρ presumably far greater than $17,000 \Omega\cdot\text{m}$)? Such a model was given as an alternative interpretation by Cull (1985) for MT data recorded on Precambrian terrain in Australia (Fig. 3-28), although Cull (1985) preferred the 2D contact model (model A in Fig. 3-28) to the dyke model (model B).

In Sweden there are no surface geological observations to support the “dyke” model, and Rasmussen (1988) preferred intrinsic micro-anisotropy as the explanation. This concept, of possible anisotropic CLCs beneath at least three Precambrian shield regions — the Baltic, Canadian and Australian shields — needs to be tested further.

In this regard, it is of import to note that the rheology of the lower crust beneath shield regions is such that deep earthquakes do occur (Chen, 1989), albeit rarely; the most recent on the Canadian shield was the Saguenay earthquake just north of Quebec City on November

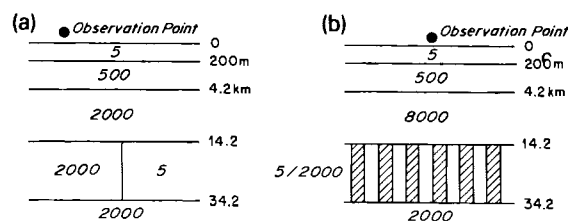


Fig. 3-28. Two models of the lower crust which fit MT observations on Precambrian terrain in Australia. Model A is of a contact in the lower crust, whereas model B is of numerous repetitive resistive dykes ($2000 \Omega\cdot\text{m}$) cutting a conducting ($5 \Omega\cdot\text{m}$) lower crust (redrawn from Cull, 1985).

25th, 1988, measuring 5.9 on the m_b scale located at a depth of 29 km (North et al., 1989) in a region of 39 km crustal thickness (Berry and Fuchs, 1973). This unusually large depth for eastern North America was confirmed by an aftershock study which located many events deeper than 25 km (North et al., 1989). Most of these deeper intracratonic earthquakes appear to be close to the crust–mantle boundary (Chen, 1989). Thus, our rheological model of the continental lower crust beneath shield regions may have to be modified somewhat to permit sufficient stress build up for earthquakes to occur and possibly for the existence of intrinsic electrical anisotropy in two horizontal directions.

4.6. Pore geometry considerations

Whatever is responsible for the enhanced electrical conductivity of the CLC — fluids, carbon films, partial melt — a mathematical model of the effect can be described by considering possible pore geometries. As stated in the Introduction (§1.1), Archie's Law (Eq. 1) has been used frequently for estimating the porosity from electrical conductivity estimates and assumed pore fluid conductivity. An exponent of 2 has been shown to be appropriate where the porosity is due largely to high aspect-ratio microcracks, whereas 1 is appropriate for interconnected low aspect-ratio cracks (Evans et al., 1982). Departures from Archie's Law have been noted (Lee et al., 1983, and references therein), particularly at very low or high porosities, and Hermance (1979) proposed a modified form for low porosities:

$$\sigma_A = \sigma_r + (\sigma_f - \sigma_r)\eta^2 \quad (15)$$

where σ_r and σ_f are the rock matrix and fluid conductivities, respectively, and η is the porosity. Note that in Eq. (15), in contrast to Eq. (1), the rock matrix conductivity is considered. Evans et al. (1982) suggest using Hoening's (1979) formulation modified for cigar-shaped cracks, and showed that, for pores with aspect ratio greater than 0.01, there is a critical porosity above which the effective conductivity increases rapidly to approach Archie's Law with unit exponent, and below which the effective conductivity is negligible.

More detailed consideration of possible pore geometries leads to a wide range of conductivities for a given porosity (Schmeling, 1986), with the highest conductivity values being obtained for interconnected thin films that wet the grain boundaries. These conditions are described by the Hashin-Shtrikman upper bound (Hashin and Shtrikman, 1963):

$$\sigma_{HS+} = \sigma_f + (1 - \eta)[1/(\sigma_r - \sigma_f) + \eta/3\sigma_f]^{-1} \quad (16)$$

which approximates to:

$$\sigma_{HS+} \approx (2/3)\sigma_f\eta \quad (17)$$

at low porosities and $\sigma_r \ll \sigma_f$. Of historical note is that the Hashin-Shtrikman upper bound is identical to one developed by Maxwell (1892) for widely dispersed spheres embedded in a fluid. Waff (1974) considered the problem of arbitrary-sized spheres and generally confirmed the Hashin-Shtrikman formula with the modification that:

$$\sigma_w \approx (2/3)\sigma_f\eta + (1 - \eta)\sigma_r \quad (18)$$

at very low porosities. This formula describes a parallel electric circuit with ionic or electronic conduction along fluid pathways and electronic conduction through the rock matrix itself.

For fluid-filled tubes along grain edges, Grant and West (1965) showed that an effective conductivity is given by:

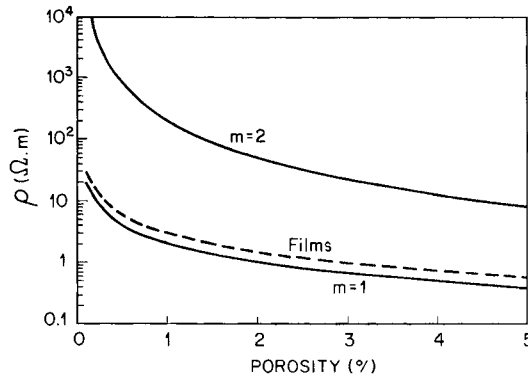


Fig. 3-29. Resistivity vs. porosity curves for fluid of 50 S/m conductivity in a resistive host matrix with Archie's Law exponents of 2 and 1 and for thin film geometry (calculated from the equations in §4.6).

$$\sigma_{GW} \approx (1/3)\sigma_f\eta + (1 - \eta)\sigma_r \quad (19)$$

Hyndman and Shearer (1989) have recently undertaken a comparison of the resistivity of a medium for various pore geometries for a host-rock resistivity of $10^4 \Omega \cdot m$ and fluid resistivity of $0.02 \Omega \cdot m$ (equivalent to an NaCl concentration of ≈ 1 M (molar), or 5%, at $500^\circ C$; Quist and Marshall 1968). Using these parameters in the above equations (Fig. 3-29) it is clear that pore geometry is important for conclusions regarding possible porosities. For example, Kurtz et al. (1986a) assumed an Archie's Law equation for their $30\text{-}\Omega \cdot m$ layer with a fluid of 20 wt.% NaCl of $0.008 \Omega \cdot m$ resistivity for a calculated porosity of 1.6%. However, a thin connected film of the same resistivity would only require a porosity of 0.04% according to Eq. 17.

Assuming a resistivity for the saline fluid of $0.002 \Omega \cdot m$ and 100 S for the CLC (§3.4) leads to a relationship between films of porosity η and thickness of the conducting zone h (in m) of:

$$\eta = 3/10h \quad (20)$$

that for a zone of 6-km thickness (i.e., a resistivity of $60 \Omega \cdot m$), requires a porosity of only 0.005%. For more conducting zones of 1000 S, this value rises to 0.05%.

4.7. Causes: Conclusions

One should not fall into the trap of searching for a single cause for the enhanced conductivity of the CLC — it is clear that there may be many explanations, and that different effects are operating in different tectonic environments and at different depths. For currently-active subduction zones there are large quantities of free water available from expulsion of pore fluids and from $CH_4\text{-}H_2O$ fluids produced by diagenetic and low-grade metamorphic reactions at depths less than 40 km (Peacock, 1990). It is expected that after the end of metamorphic activity the fluids can be maintained for more than 70 Ma (Thompson and Connolly, 1990) or even much longer (Bailey, 1990). For the Early Proterozoic subduction zone within the Trans-Hudson Orogen, the values of conductivity are too high (>3 S/m) to be explained by fluids alone; for fluid of 50 S/m conductivity the porosity would have to be implausibly high ($>10\%$). Explanations involving conducting minerals, such as graphite sheets, must be appealed to for this type of structure. For zones where the temperature exceeds $730^\circ C$, such as the lower crust beneath the Valhalla Gneiss Complex in British

Columbia (§3.6, Fig. 3-22), partial melting may be the cause of the very high conductivities observed.

Given our observations that the conductance beneath old shield regions is two or three orders of magnitude larger than would be expected for dry rocks, we still do not have a satisfactory explanation free of considerable objections. A small percentage (less than 1%) of saline-filled porosity is currently favoured by the electromagnetic community. Is it reasonable to expect free fluids? Many petrologists see no evidence for free water within the lower crust in their data. Accordingly, other explanations — such as carbon films on grain-boundaries — are being advanced. The required porosities could be very low if the fluids are highly saline with conductivities of 500 S/m rather than 50 S/m. Would the effects of saline fluids in thin zones, of such very low porosities, in the mid-crust be observed in exhumed rocks from the CLC, especially given Hulsebosch and Frost's (1989) remarks concerning the lack of chemical modification of rocks by deeply-penetrating saline brines? Is it reasonable to expect interconnectivity at such low porosities? Is the model proposed by Sanders (1991) an explanation of this apparent dichotomy between petrologists and EM geophysicists?

Alternatively, is it reasonable to expect interconnected thin carbon films over many thousands of kilometres, such as the Siberian shield (§3.1.3), especially given the observed immobility of carbon (Tsong et al., 1985; Watson, 1986). Also, can we further expect these films to remain stable over very long geological times?

It would narrow the field of possible causes considerably if we could determine precisely the thicknesses of the various conducting zones observed in the CLC. As discussed in §2.4.3, the deep-probing EM methods used are virtually all detecting the conductance, or conductivity-thickness product, of the zone of enhanced conductivity and not either parameter independently. Stacked multiple thin zones, rather than a single thick zone, may be more consistent with the seismic reflectivity patterns observed for the CLC, but we must try to obtain evidence of this from the EM data themselves.

5. Conclusions

Mapping the lower continental crust is only in its infancy, and this activity will engage us for many years to come. As envisioned by Oliver (1991), by the year 2100 all the continents will be criss-crossed by deep seismic reflection profiles with complimentary information from other geophysics that will give us a far more comprehensive understanding of the continental crust than we have today. As a component of such a multidisciplinary geoscientific program, imaging the electrical conductivity can add much to understanding the state and composition of the crust. The EM results from Kapuskasing (§3.5) and beneath the Valhalla complex (§3.6) are giving key information, and EM methods may be the only method for detecting crustal-scale fluids and fluid transport systems of the type proposed by Torgersen (1990).

One enigma that must be addressed is "*what is the cause, or causes, of the zones of enhanced electrical conductivity beneath stable regions?*" Interpretations of such zones beneath active or young (<100 Ma) regions in terms of fluids or partial melt are generally acceptable. Linear anomalies of enhanced conductivity associated with ancient subduction zones can be interpreted as due to conducting minerals. But what of the CLCs beneath stable regions such as the Canadian, Siberian and Baltic shields which have conductances one to two orders of magnitude greater than would be expected for dry granulite facies assemblages? Compositional contrasts appear to be excluded. Less than 1% of an interconnected saline fluid

is sufficient to account for the observations — but there have been petrological arguments advanced against such a hypothesis which must be addressed. Alternatively, is it reasonable to expect an interconnected grain boundary film of carbon beneath extensive stable regions? Are there other conduction mechanisms operating that we have yet to discover?

Such questions can only be answered by the reference to other geophysical, geological and geochemical data. Statistical correlations between EM results and heat flow (Ádám, 1976, 1980), seismic refraction (Jones, 1981b; Hyndman and Klemperer, 1989) and seismic reflection (Jones, 1987), and between heat flow and seismic reflection results (Klemperer, 1987), must be treated with caution, but are the necessary first steps.

Perhaps it should be re-emphasized that the whole of the CLC may not be conducting, but a thin zone of high conductivity at mid-crustal depths may be the cause for our EM observations. Obviously, the prime task of the EM community in the near-term is to resolve the actual depth-distribution of conductivity within the CLC rather than just its conductance; this will involve specifically-designed experiments at well-chosen sites. To aid our interpretation of the conductivity structure determined, the geological and geochemical community must come to a consensus on the role of saline fluids in the deep crust of stable regions and either permit or exclude them. I predict that by the beginning of the 21st century we will have a far clearer picture of the electrical conductivity of the lower continental crust and its interpretation.

Acknowledgements

I wish to acknowledge with deep gratitude the invitation of the editors, in particular David Fountain, to contribute this chapter to their book, and their fortitude and patience whilst waiting for its completion. My colleagues responded abundantly to my requests for reprints and preprints of their work, and Steve Constable kindly provided the code for his OCCAM inversion scheme. Finally, I thank Alan Green, David Fountain, Ron Frost, David Boerner and Ron Kurtz for comments on an earlier version. Their contributions, especially Alan Green's, were substantial; the errors remain mine alone, however.

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