On a Type Classification of Lower Crustal Layers Under Precambrian Regions

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Abstract. Various parameters pertinent to the lower crustal layer under Precambrian regions are listed for locations where seismic, and geomagnetic or geoelectric, studies have been undertaken. The parameters define three distinct types of lower crustal layer with certain dominant characteristics: Type I -- "Normal" -- typical continental seismic parameters and a high electrical resistivity ($10^3$-$10^4$ $\Omega$m); Type II -- "Intermediate" -- high compressional wave velocity (either fixed $V_p=7.0$ km s$^{-1}$ or transitional $V_p=6.7$-$7.3$ km s$^{-1}$) and a moderate resistivity ($100$-$300$ $\Omega$m); Type III -- "Low" -- a low shear wave velocity layer ($V_s < 7.3$ km s$^{-1}$) and a low electrical resistivity ($10$-$50$ $\Omega$m). Possible conditions and rock types, existing at the $P$-$T$ environment of the lower crust and which could account for the observations, are suggested. The zoning of Canada into types implies that Type II layers are shield "edge" effects, and that inability to observe what is regarded as the final stage of development of a shield region under certain shields may be due to their being too small.

Key words: Precambrian regions -- Crustal structure -- Electromagnetic induction studies -- Seismic studies

Introduction

Various seismic and electrical studies over Precambrian regions of the world have been able in the past to detail some of the parameters of the lower crustal layers, e.g., compressional wave velocity ($V_p$), shear wave velocity ($V_s$), density ($\rho$), Poisson’s ratio ($\sigma$) from multi-method seismic studies, and electrical resistivity ($R$ -- the normal symbol for electrical resistivity $\rho$ is not used to avoid confusion with density) from geoelectric and geomagnetic methods. However, confusion is often apparent in the literature as to why certain features are found in some parts of the world, but not in others. This confusion is perhaps highlighted by Kovtun (1976) who states, with reference to geomagnetic studies, that "at present, we have almost no results on which a study of the ‘normal’ distribution of conductivity vs depth in Precambrian shields could be based.”

It is with these points in mind, that the author has attempted a broad classification of lower crustal Precambrian layers into three types. In the following section, each of these types will be described in turn and examples of their locations will be specified. The locations and their corresponding data sets, both seismic and geoelectric, were chosen for the following reasons:

(i) all come from stable regions of low seismicity and heat flow,
(ii) the data sets have all been thoroughly scrutinized by the groups involved, and
(iii) the information is readily available.

Finally, possible candidates for the rocks at lower crustal depths causing the various observed responses are suggested.

Classification

Type I -- “Normal”

The prime example of a Type I lower crustal layer is that observed in the centre of the Canadian shield. Seismic models presented by Wickens and Buchbinder (1980), based on surface waves (Wickens 1971) and refraction information (Mereu and Hunter 1969), Hall and Hajnal (1973), and Green et al. (1980), representative of middle and northern Manitoba, are illustrated in Fig. 1a, and their estimated parameters for the lower crustal layer are detailed in Table 1. The model of Wickens and Buchbinder (1980) was shown by them to be consistent with the $S$-wave residuals observed in the central shield region.

A transient electromagnetic sounding investigation undertaken by Jacobson (1969; reported in Keller 1971) in Manitoba inferred that there must exist a layer of uniform resistivity, of $R \approx 2000$ $\Omega$m, to depths greater than 20 km, whilst magnetotelluric (MT) investigations by three groups in Alberta (Srivastava and Jacobs 1964; Vozoff and Ellis 1966; Reddy and Rankin 1971) all infer resistivities for this layer of $10^3$-$10^4$ $\Omega$m.

It appears that the Ukrainian shield also falls into this Type I category. Models P2 and P3 of Jentsch (1979) display a lower crustal layer with $V_p=6.2$-$6.4$ km s$^{-1}$ (the very thin transitional layer between 32.5-33.4 km depth, of velocity increase from 6.8 to 7.6 km s$^{-1}$ can be ignored for this comparison), and the data from the majority of magnetotelluric stations recorded by Tkachev (1973; reported in Kovtun 1976) display apparent resistivities of greater than $10^3$ $\Omega$m over the period range 1–$10^3$ s. Such values require a highly resistive crust, i.e., resistivity greater than about 4000 $\Omega$m.

The heat flow observed on the Canadian shield is, on average, 39 mW m$^{-2}$ (Rao and Jessop 1975), i.e., 0.93 HFU, whilst on the Ukrainian shield values of 25–35 mW m$^{-2}$ (Kutas et al. 1979), i.e., 0.60–0.84 HFU, are reported.

Type II -- “Intermediate”

The three main examples of a Type II lower crust are to be found in northern Scotland, eastern Canada, and northern Sweden. No heat flow estimates are available for northern Scotland (Bloomer...
Table 1. Lower crustal parameters as inferred by the investigations reported in the text

<table>
<thead>
<tr>
<th>Type</th>
<th>Code</th>
<th>Area</th>
<th>$V_p$ (km s$^{-1}$)</th>
<th>$V_s$ (km s$^{-1}$)</th>
<th>$\rho$ (g cm$^{-3}$)</th>
<th>$\sigma$</th>
<th>$R$ (Ωm)</th>
<th>Reference</th>
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<td>W</td>
<td></td>
<td>Central Canadian shield</td>
<td>6.6</td>
<td>3.8</td>
<td>0.252$^a$</td>
<td>Wickens and Buchbinder (1980)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>G</td>
<td></td>
<td></td>
<td>6.55</td>
<td></td>
<td></td>
<td>Green et al. (1980)</td>
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<td>6.85</td>
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<td>0.241$^a$</td>
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<tr>
<td>JA</td>
<td></td>
<td>Manitoba</td>
<td></td>
<td></td>
<td></td>
<td>Jacobson (1969)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>SR</td>
<td></td>
<td>Alberta</td>
<td></td>
<td></td>
<td>2,000</td>
<td>Srivastava and Jacobs (1964)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>R</td>
<td></td>
<td>Alberta</td>
<td></td>
<td></td>
<td>3,500</td>
<td>Reddy and Rankin (1971)</td>
<td></td>
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</tr>
<tr>
<td>VO</td>
<td></td>
<td>Alberta</td>
<td></td>
<td></td>
<td>10,000</td>
<td>Vozoff and Ellis (1966)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>JE</td>
<td>T</td>
<td>Ukrainian shield</td>
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<td></td>
<td>&gt;4,000</td>
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<tr>
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<td></td>
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<td></td>
<td>4.04$^a$</td>
<td>Assumpção and Bamford (1978)</td>
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<td></td>
</tr>
<tr>
<td>A</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.249</td>
<td>Hutton et al. (1980, 1981)</td>
<td></td>
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</tr>
<tr>
<td>O</td>
<td></td>
<td>Southern Superior Province</td>
<td>7.0</td>
<td>7.2</td>
<td>0.21</td>
<td>O'Brien (1968)</td>
<td></td>
<td></td>
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<tr>
<td>ST</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>Sternberg (1979)</td>
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<td></td>
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<tr>
<td>BI</td>
<td></td>
<td>Eastern Superior Province</td>
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<td>7.1</td>
<td>0.22</td>
<td>Berry and Fuchs (1973)</td>
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<td>HA</td>
<td></td>
<td>Southwestern Superior Province</td>
<td>7.1</td>
<td>±0.04</td>
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<td>Hall and Hajnal (1973)</td>
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<td>II</td>
<td>G</td>
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<td>7.2</td>
<td>0.21</td>
<td>Green et al. (1980)</td>
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<td>KU</td>
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<td></td>
<td>200</td>
<td>Kurtz and Garland (1976)</td>
<td></td>
<td></td>
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<tr>
<td>D</td>
<td></td>
<td>Superior Province</td>
<td></td>
<td></td>
<td>270</td>
<td>Duncan et al. (1980)</td>
<td></td>
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<td></td>
<td>Southeastern Grenville Province</td>
<td>7.14</td>
<td>4.11</td>
<td>3.11</td>
<td>Jordan and Frazier (1975)</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>Northern Sweden</td>
<td>6.75</td>
<td>7.15</td>
<td>0.21</td>
<td>Hirschleber et al. (1975)</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>J</td>
<td>Southern Norway</td>
<td></td>
<td>7.17</td>
<td>0.21</td>
<td>Lund (1970)</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>140–450</td>
<td>Jones (1981)</td>
<td></td>
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<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td>Massé and Alexander (1974)</td>
<td></td>
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<tr>
<td>BL</td>
<td>V</td>
<td>Southeastern Africa</td>
<td>6.8</td>
<td>3.75–3.85</td>
<td>2.95</td>
<td>0.28$^a$</td>
<td>Block et al. (1969)</td>
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<td></td>
<td></td>
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<td>van Zijl (1977)</td>
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<td></td>
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<td></td>
<td></td>
<td></td>
<td>Blohm et al. (1977)</td>
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<tr>
<td>III</td>
<td>JO</td>
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<td>&gt;6.8</td>
<td>3.4</td>
<td>3.11</td>
<td>0.30</td>
<td>10–30</td>
<td>Jordan and Frazier (1975)</td>
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$^a$ denotes a calculated value assuming perfect elasticity on the rock.

Fig. 1. a Seismic profiles as determined by several investigations over Type I lower crustal layers (for key see Table 1). b Resistivity profiles for the same locations as in a, dashed lines denote minimum values.
et al. 1979), but in eastern Canada values of 40–50 mW m\(^{-2}\) (1.0–1.25 HFU) are most commonly observed (Lachenbruch and Sass 1977), and in northern Sweden 38–47 mW m\(^{-2}\) (0.91–1.12 HFU) are reported (Eriksson and Malmquist 1979).

In northern Scotland there exists between the Highland Boundary Fault and the Great Glen Fault a lower crustal layer with \(V_p=7.0\) km \(s^{-1}\) (Bamford et al. 1978) and \(\sigma=0.249\) (Assumpção and Bamford 1978), giving a calculated \(V_r=4.04\) km \(s^{-1}\), which corresponds with a layer of electrical resistivity \(R=100–300\ \Omega\text{m}\) (Hutton et al. 1980, 1981). The original seismic interpretation of this layer was of a velocity gradient layer from 6.7 to 7.3 km \(s^{-1}\) (Bamford et al. 1976), but such a layer was later found not to be consistent with ray-tracing and a uniform layer of 7.0 km \(s^{-1}\) was found to be most effective (Bamford et al. 1978).

In eastern Canada, various seismic and geomagnetic studies have been carried out. In southern Superior Province, O’Brien (1968) reported a zone with \(V_p=7.0–7.2\) km \(s^{-1}\) at a depth of 19.5 km for stations on the northern edge of Lake Superior, whilst Berry and Fuchs (1973; reproduced in Berry and Mair 1977) interpret their data as inferring a velocity gradient layer with \(V_p=6.7–7.5\) km \(s^{-1}\) between 27–40 km depth. This layer correlates with a geoelectric layer of 270 \(\Omega\text{m}\), the top of which deepens from 19 km in the south-west to 29 km in the north-east, defined by Duncan et al. (1980) for the region around Timmins, and with a layer of \(R=50–1,500\ \Omega\text{m}\) for Wisconsin found by Sternberg (1979).

In Grenville Province, Berry and Fuchs (1973) find another velocity gradient layer with \(V_p=6.7–7.1\) km \(s^{-1}\) between 22–40 km, whilst Jordan and Frazer (1975) report a layer between 15–25 km with \(V_p=7.14\) km \(s^{-1}\) and \(V_s=4.11\) km \(s^{-1}\). The 2-dimensional geoelectric model for eastern Canada presented by Kurtz and Garland (1976) has a uniform crust under Grenville Province of \(R=200\ \Omega\text{m}\).

In northern Sweden, the best fitting gradient model to the “Blue Road” seismic data includes a layer with \(V_p\) increasing from 6.5 km \(s^{-1}\) at 20 km depth to \(V_p=7.15\) km \(s^{-1}\) at 42 km depth (Hirschleber et al. 1975). Values of around 7 km \(s^{-1}\) for the bottom crustal layer were confirmed in a more sophisticated analysis of the data by Lund (1979). Monte-Carlo inversion (Jones and Hutton 1979) of the geomagnetic response function observed at Kiruna (Jones 1980), with the constraints that i) the top layer have a resistance of \(10^4\ \Omega\text{m}\) (Westerlund 1972) and ii) that the second interface be at 46 km, to correspond with the latest values of Moho depth (Bungum et al. 1980), yielded a lower crustal layer of \(R=140–450\ \Omega\text{m}\) (Jones 1981).

Type III — “Low”

The two examples for a Type III lower crustal layer are from the Adirondacks, New York State, and the southeastern African shield region.

In the Adirondacks, a joint controlled-source and horizontal spatial gradient (HSG) investigation by Connerney et al. (1980) and Connerney and Kuckes (1980) determined a zone of low electrical resistivity (LRL — low resistivity layer) of \(R=10–30\ \Omega\text{m}\) at a depth of 25–34 km. This is exactly the depth at which Jordan and Frazer (1975) interpret a shear wave low velocity layer (LVsL) from \(S_p\) phases for stations mostly situated close to Ottawa. The studies of Jordan and Frazer inferred that the layer must have a minimum Poisson’s ratio of \(\sigma=0.30\), hence the value of the compressional wave velocity for the layer \(V_p=6.80\) km \(s^{-1}\) is to be regarded as a minimum. Thus, the existence of a compressional wave low velocity layer (LVL) is not proven, but the existence of an LVPL has been confirmed by Wickens and Buchbinder (1980).

Two geoelectric studies on southern African cratons inferred LRLs. These were on the Kaapvaal craton (van Zijl 1977) and the Zimbabwe (formerly Rhodesian) craton (Blohm et al. 1977). Data from the Limpopo mobile belt also were explained in terms of an LRL, but heat flow estimates on mobile belts have been observed to be anomalously high when compared with cratonic terrains (Carte and van Rooyen 1969) hence the model is not included here. Bloch et al. (1969), in an analysis of surface waves observed by an array of seismometers centered on Johannesburg, infer from their data the existence of two LVsLs, one at 12 km and the other at 24 km (Fig. 3a). The depth of the deeper LVPL corresponds exactly to the depth of the LRL as interpreted by Blohm et al. (1977), and the LRL under the Limpopo mobile belt (van Zijl 1977). It may be significant that the model of Bloch et al. (1969) does not display an LVPL.

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**Fig. 2.** a, b as Fig. 1 but for investigations over Type II lower crustal layers
Discussion

The notable characteristics of the three types classified are:

a) Type I – Normal
   - typical continental $V_p$, $V_s$ and $\sigma$ values
   - highly resistive, i.e., greater than $10^3 \Omega$m.

b) Type II – Intermediate
   - $V_p$ of around 7.0 km s$^{-1}$, either fixed (northern Scotland) or transitional from 6.7 to 7.2 km s$^{-1}$ (eastern Canada, northern Sweden)
   - typical $\sigma$ values of around 0.25
   - moderately resistive, $R=100-300 \Omega$m.

c) Type III – Low
   - low shear wave velocity layer ($L_V$, $L_s$), possibly without a corresponding low compressional wave velocity (no $L_V$, $L_s$)
   - high Poisson’s ratio, of $\sigma > 0.30$
   - highly conductive, with $R=10-50 \Omega$m.

The association of $L_V$, $L_s$ layers with LRL layers in more tectonic regions has been known since the early 1970’s (see, for example, Landisman et al. 1971), even though the existence of both was initially in question (Healy 1971; Porath 1971). Many theories have been proposed to explain these layers, the two most often mentioned being the role of water of dehydration escaping into pore spaces (Hyndman and Hyndman 1968), or of hydrated rocks under typical $P$–$T$ conditions in the lower crust (Landisman et al. 1971; Jordan and Frazer 1975). For Precambrian regions, the formers of these is however considered untenable (see later).

At Moho depths under the centre of the Canadian shield (∼35 km, Wickens and Buchbinder 1980), the temperature is believed to be in the range 400°–500° C (Rao and Jessop 1975), preferably towards the upper end of this limit (475° C as given by Eq. (5) of Hall 1977), with a pressure of around 10 kbar (9.5 kbar, Eq. (7), Hall 1977). At such $P$–$T$ and anhydrous conditions, eclogite is the stable mineral assemblage for basalt of quartz tholeiite composition, rather than garnet granulite (Ringwood 1975, p. 25, Figs. 1–6). However, the inferred compressional wave velocities for eclogite under these $P$–$T$ conditions, of $V_p=7.8-8.6$ km s$^{-1}$ (Press 1966, Table 9–2; Manghnani et al. 1974; Christensen 1979), require the presence of large amounts of minerals of relatively low $V_p$, e.g., quartz and alkali feldspars (Ringwood 1975, p. 39). This implies a dioritic rock in the granulite-eclogite facies as a suitable candidate, e.g., quartz diorite, $V_p=6.71$ km s$^{-1}$, $V_s=3.8$ km s$^{-1}$ (Press 1966, Tables 9–2 and 9–3), $\rho=2.68-2.96$ g cm$^{-3}$ (Daly et al. 1966, Table 4.1). Measurements of electrical conductivity on alkali basalts at 500° C indicate values in the range $R=10^3-10^4 \Omega$m (Bondarenko and Galdin 1972; reported in Haak 1980) whilst Volarovich and Parkhomenko (1976) specify a range of $R=10^3-10^5 \Omega$m for granodiorites, quartz diorites and diorites at this temperature.

The compressional wave velocity for the Type II layer, $V_p=6.8-7.2$ km s$^{-1}$, can be explained by either a garnet granulite in an anhydrous crust, as suggested by Manghnani et al. (1974) and Ringwood (1975, p. 39), or by amphibolites in a hydrous crust (Ringwood 1975, p. 41). Measurements of Poisson’s ratio at 10 kbar by Manghnani et al. (1974) on their garnet granulite samples give a mean of $\sigma=0.282 \pm 0.017$ (one standard deviation), which is too large compared with field observations (northern Scotland, $\sigma=0.249 \pm 0.020$, Assumpção and Bamford 1978; eastern Canada, $\sigma=0.25$, Jordan and Frazer 1975). Also, the electrical resistivity of the majority of dry rocks expected to be found at lower-crustal depths under Precambrian regions is never below $10^3 \Omega$m for temperatures less than 500° C (Brace 1971, Fig. 3; Haak 1980), with the notable exception of a Wisconsin gabbro with $R=650 \Omega$m at $T=500°$ C (Housley and Oliver 1977). The effects of increasing the pressure to 10 kbar at this temperature were shown by Volarovich and Parkhomenko (1976) to have little effect on the electrical conductivity, i.e., less than one third an order of magnitude. However, Ringwood (1975, pp. 35–47) believes that gabbro is not stable at the $P$–$T$ conditions of the lower crust. Seismic evidence appears to confirm this suggestion because although gabbros display compressional wave velocities of the correct order when at 10 kbar ($V_p \sim 7.2$ km s$^{-1}$, Press 1966, Table 9–2), the effects of temperature will reduce this to $V_p \sim 6.8$ km s$^{-1}$ (Press 1966, Fig. 9–7; Christensen 1979). Also, gabbro do not show sufficiently large shear wave velocities, e.g., $V_s=3.84$ km s$^{-1}$ (Press 1966, Table 9–4). Thus, having apparently disqualified all suitable rocks that might exist in an anhydrous lower crust, it is necessary to conclude that hydrous conditions prevail.
in Type II lower crustal layers. Amphibolites, as suggested by Ringwood (1975, p. 41) as the probable rock at the $P$-$T$ conditions of a hydrous lower crust, display the correct order of seismic velocities ($V_p = 7.18$–$7.22$ km s$^{-1}$, Christensen 1965; $V_p = 4.03$ km s$^{-1}$, Christensen 1966) but it is probably necessary to infer a slightly seismically faster epidote amphibolite ($V_p = 7.67$–$7.75$ km s$^{-1}$, Christensen 1965) due to the reduction in velocities at temperatures of around 400–500°C (Christensen 1979). Unfortunately, not many measurements of the electrical resistivity of amphibolites at these $P$-$T$ conditions have been made, but the data of Volarovich and Parkhomenko (1976) illustrate that certain amphibolites display the required resistivity of a few hundreds of $\Omega$m at 500°C.

For the Type III lower crustal layer, Jordan and Frazer (1975), van Zijl (1977), and Connerney et al. (1980) all independently suggested the effects of serpenitisation to explain their data. Jordan and Frazer (1975) observe that a serpenitised peridotite has the high Poisson's ratio required, and van Zijl (1977) and Connerney et al. (1980) both remark that serpenitised rocks display to the appropriate low resistivities, of 10–50 $\Omega$m, provided that the water of hydration does not escape (Zablocki 1964; see van Zijl 1977).

An interpretation of the high conductivities observed in Type III crusts in terms of electrolytic conduction is considered untenable by Connerney et al. (1980) because investigations by Richter and Simmons (1977) on Precambrian igneous rocks showed that the majority of the original cracks were healed or sealed by secondary minerals. However, the somewhat speculative models presented by Dewey (1969a) and Ringwood (1975, p. 298–304) for the development of a Benioff zone require, as a passive initial state, a thick accumulation of sediments, the prime example of which is the non-volcanic margin of the spreading Atlantic Ocean (Drake et al. 1959). Assuming that the east coast of North America represents such an initial state, and considering that there is much support in the Appalachians for a progressive migration of deformation and metamorphism towards the stable continental foreland (Dewey 1969b), it is not inconceivable that the lower foreland crust be fractured and fissured. If so, then electrolytic conduction effects would become significant, which would explain the results of Connerney et al. (1980) for the Adirondacks, and also those of Edwards and Greenhouse (1975), of a highly conducting lower crustal layer, $R = 2$ $\Omega$m, underlying the southern Appalachians.

An eigenparameter investigation by Edwards et al. (1980) of the model presented by Blohm et al. (1977) revealed that the resistivity and depth of the lower crustal conducting layer are not independently estimated when geolectrical sounding data are inverted. The resolvable eigenparameter pertinent to this layer is the layer's depth-integrated conductivity, i.e., thickness-conductivity product. Hence it may be possible that the data of van Zijl (1977) and Blohm et al. (1977) are also consistent with a slightly less conducting but thicker zone of Type II class. A similar possibility cannot be suggested for the conducting zone delineated by Connerney and Kuckes (1980) and Connerney et al. (1980) as their controlled-source electromagnetic induction experiment inferred a resistivity in the range $R = 12.5$–25 $\Omega$m for this layer. The conductance of this layer was defined by Horizontal Spatial Gradient (HSG) data as being approximately 400 S (Connerney and Kuckes 1980), hence inferring a thickness of 5–10 km. This estimate is fully consistent with the 10 km thick LV$L$ delineated by Jordan and Frazer (1975).

There is evidence from Jordan and Frazer (1975) that a Type III crustal layer underlies a Type II layer. The data from Berry and Fuchs (1972) and Kurtz and Garland (1976) imply that the Type III layer pinches out to the north, leaving solely a Type II layer, and the work of Connerney et al. (1980) suggests the opposite is true on going further southeast, i.e., that the Type II layer pinches out and a Type III layer exists alone. For the Canadian shield, a sketch map is shown in Fig. 4 illustrating the lower crustal layers as classified in this paper. The illustration includes the recent results of Green et al. (1980) with a layer of $V_p = 7.1$–7.2 km s$^{-1}$ underlying southeastern Churchill Province at a depth of 36–45 km and those of Hall and Hajnal (1973) of a layer with $V_p = 7.1 \pm 0.04$ km s$^{-1}$, observed south of Lake Winnipeg.

It is apparent from Fig. 4 that a Type II lower crustal layer is an “edge” phenomenon of the Canadian shield, where “edge” means not the actual physical boundary of the shield, but a wide zone encompassing the central core of the shield. This may be evidence of a general feature of shield regions in that their centres...
Table 2. Classification system with values of the parameters inferred for the lower crustal layer and conditions and possible candidates for the rock responsible

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<th>Classification</th>
<th>Parameters</th>
<th>Conditions and possible rock types</th>
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<td>Type I &quot;Normal&quot;</td>
<td>$V_p$ (km s$^{-1}$) 6.6  $V_s$ (km s$^{-1}$) 3.8  $\rho$ (g cm$^{-3}$) 0.25  $\sigma$ 10$^{3}$--10$^{4}$</td>
<td>Anhydrous: Quartz Diorite</td>
</tr>
<tr>
<td>Type II &quot;Intermediate&quot;</td>
<td>$V_p$ (km s$^{-1}$) 6.8--7.3  $V_s$ (km s$^{-1}$) 4.1  $\rho$ (g cm$^{-3}$) 3.11  $\sigma$ 0.25  $R$ (Ωm) 100--300</td>
<td>Hydrous: Amphibolite</td>
</tr>
<tr>
<td>Type III &quot;Low&quot;</td>
<td>$V_p$ (km s$^{-1}$) 6.8  $V_s$ (km s$^{-1}$) 3.4--3.7  $\rho$ (g cm$^{-3}$) 3.11  $\sigma$ &gt;0.30  $R$ (Ωm) 10--50</td>
<td>Hydrous: Serpentinite</td>
</tr>
</tbody>
</table>

It must be remarked that it is not expected that all Precambrian lower crustal layers fall into one of the three general categories detailed in Table 2. One seemingly notable exception to this classification is the East European Platform. Seismic models of this platform include a lower crustal layer with $V_p$=6.8--7.2 km s$^{-1}$ (Bozholo and Starovoit 1969; Kosminskaya and Pavlenkova 1979; Pavlenkova 1979; Patton 1980), hence defining it as Type II. However, magnetotelluric data (Kovtun and Chicherina 1969), particularly the "generalised curve" for the region (Vanyan et al. 1977), imply a highly resistive crust, i.e., a Type I lower crustal layer. This latter result is, in the opinion of the author, not totally conclusive as it is based solely on apparent resistivity information, whereas the magnetotelluric phase is far more discriminating between Types I, II, and III.

Finally, this study has shown the decided advantages of undertaking cooperative seismic and geomagnetic or geoelectric measurements in the same area. Independent interpretations of the data recorded may differ wildly, hence either, or both, may be wildly wrong. This may be particularly true for seismic data from shield regions because, as noted by Pavlenkova (1979), compressional wave velocities of between 6.6--7.7 km s$^{-1}$ are not usually observed in the first arrivals, hence implying that seismically Type II layers are difficult to delineate. A good example of this is Massé and Alexander’s (1974) reinterpretation of Sellevoll and Warrick’s (1971) data from southern Norway. The initial model had a lower crustal layer of $V_p$=6.51 km s$^{-1}$ (Sellevoll and Warrick 1971), whereas a more sophisticated analysis gave a value of $V_p$=7.17 km s$^{-1}$ (Massé and Alexander 1974). Joint interpretation of dual data sets, in the manner described in this work, is likely to reduce drastically the number of possible likely candidates for the lower crustal layer.

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