J. Geomag. Geoelectr., **45**, 933–955, 1993

The COPROD2 Dataset: Tectonic Setting, Recorded MT Data, and Comparison of Models

Alan G. JONES

Geological Survey of Canada, 1 Observatory Crescent, Ottawa, Ontario, Canada, K1A 0Y3

(Received September 28, 1993; Revised September 30, 1993; Accepted October 5, 1993)

To facilitate a comparative study of approaches and algorithms for two-dimensional interpretation of magnetotelluric (MT) data, a dataset of thirty-five sites, called COPROD2, was distributed to the electromagnetic induction community. The data are from stations along a 400 km east-west profile in southern Saskatchewan and Manitoba, Canada, crossing the thick Paleozoic sediments of the Williston basin. Within the basement beneath the sediments lies one of the world's longest and most enigmatic crustal conductivity features—the North American Central Plains (NACP) conductivity anomaly. Also, at the eastern extremity of the profile is a second basement anomaly (TOBE) interpreted to be associated with the Thompson Nickel Belt at the Superior-Churchill boundary. The MT data were corrected for static shifts, and only the off-diagonal impedance data together with the transfer function data, and their errors, were made available. These MT data are of wide bandwidth (384 Hz–1820 s) and high quality (impedance errors typically <2%), and require sophisticated modelling and inversion in order to extract as much meaningful information as possible from them. A challenge for those interpreting these data is that there is a very small, but critical, response in the B-polarization data to the presence of the basement anomalies. In this introductory paper, the previous electromagnetic, and other geophysical, studies of the NACP anomaly, and its tectonic setting within the Trans-Hudson orogen, are described. Representative data from the COPROD2 data are illustrated, and models derived by various groups are shown. Comparisons of these models reveals that a single measure of misfit is an inadequate description of how well a model fits a dataset; one must compare the model fit to the data at virtually each datum to avoid trends in the misfit residuals. Additionally, reliable and consistent error and static shift estimates are essential in order to obtain high resolution images of the Earth's conductivity structure. Finally, the COPROD2 data highlight the sensitivity in B-polarization mode data to breaks in the resistive layer, whereas the E-polarization data sense regions of enhanced conductivity.

1. Introduction

During the late-1970s and early-1980s, apparent resistivity and phase responses from a site in southern Scotland (site NEW, JONES and HUTTON, 1979) were distributed to the EM induction community to serve as a baseline dataset for comparison of one-dimensional (1D) inversion algorithms. This comparative effort, called COPROD for "Comparison of One-dimensional PROfiles from MT Data", was viewed by the community as a worthwhile exercise which contributed both towards the leap in our understanding of 1D inversion and towards the number of algorithms and approaches available to the data interpreter (see WEIDELT, 1985). Without question, the COPROD data are poor by today's standards; they are relatively narrow-band with impedance errors typically >10%.

For the last five years there has been a virtual explosion in methods for two-dimensional (2D) inversion of MT data after a relative hiatus of fifteen years since the mid-1970s. Many of the programs are being made available to the community, and publications have been appearing using them. It was thought useful to repeat the COPROD experience, but with an MT dataset from a profile crossing a strongly 2D geoelectric structure. Hence, COPROD2, where the "2" refers to two-dimensional data, was born.

It must be emphasized from the outset that the intent of this exercise is certainly *not* to hold a "competition" and declare a winner! Given the vast range of complexity of structure existing within the Earth's crust, that we can never hope to model completely, we must always remember that our "image" of that structure is biased by the limitations and underlying assumptions of our modelling and inversion algorithms. Accordingly, it behoves us to use as many inverse codes as possible on our data, as each will give a different search through model space. There is no code which can be called "best"; some have computational efficiency advantages at the cost of approximate solutions, whereas others are ideally suited for hypothesis testing. What we wish to accomplish is a better understanding of our modelling and inverse codes, and thereby appreciate fully the information that they are revealing about the actual conductivity structure of the Earth.

Chosen for this first COPROD2 comparison was a dataset of thirty-five MT sites in southern Saskatchewan and Manitoba, Canada. These data were acquired in 1984 and 1985 by Phoenix Geophysics (Toronto) Ltd., using their MT-16 systems, for PanCanadian Petroleum Ltd. who were interested in investigating basement features as part of their studies of possible basement control of sedimentary structures. The 400 km east-west profile crosses the hydrocarbon-rich, intracratonic, Paleozoic Williston basin and, within the basement, the North American Central Plains (NACP) conductivity anomaly. In addition, within the basement at the eastern end of the profile was discovered a second zone of anomalously high electrical conductivity, named the TOBE anomaly by JONES and SAVAGE (1986) because of its spatial correlation with the subsurface extrapolation of the Thompson belt based on aeromagnetic data (GREEN *et al.*, 1985).

JONES and SAVAGE (1986) presented a brief, principally qualitative, interpretation of these data illustrating that the location of maximum response to the NACP structure lay some 75 km east of the position mapped by the coarse GDS array studies (see below). JONES (1988a) described a method for removing static shifts from MT data, relying on there existing within the section a layer which could be described parametrically, and showed that for these data a reasonable assumption to make is that the conducting sediments above the resistive basement have a laterally-invariant resistivity of 3 Ω·m. JONES and CRAVEN (1990) modelled these static shift-corrected data, and qualitatively interpreted data from two MT profiles further north.

The data that were distributed to the COPROD2 participants prior to the First Magnetotelluric Data Interpretation Workshop (MT-DIW1) were the shift-corrected MT apparent resistivities and phases together with vertical field transfer functions, and their errors. Only the off-diagonal terms were distributed as the intent of this COPROD2 exercise was to focus effort on 2D modelling and inversion and not concern ourselves with galvanic distortion effects. The data are in a geographic coordinate system, with the Z_{xy} impedances representing northward-directed electric field and are defined as the E-polarization data. The Z_{yx} impedances are the B-polarization data.

One intriguing aspect about these data is that the B-polarization responses appear to be virtually insensitive to the presence of either the NACP or TOBE anomalies. However, on closer examination the B-polarization phases do display local features which dictate the shape of acceptable models.

Herein I describe the previous EM studies of the region, which are mainly geomagnetic array studies and two geomagnetic profiles on the exposed Precambrian rocks of the Trans-Hudson Orogen (THO). The tectonic setting of the THO is also outlined, together with other geophysical results. Representative data from the COPROD2 dataset are illustrated to show data quality and anomaly responses. Models derived by the contributors to this special issue of *J. Geomag. Geoelectr.* of the COPROD2 dataset are shown, and finally conclusions drawn about the interpretation of these data.

2. Previous EM Studies in the Region

Fortunately for THO studies in general, and the MT-DIW1 which occurred twenty-five years later, one station in the corner of a geomagnetic depth sounding (GDS) magnetometer array operated in September, 1967, in the western United States proved the well-known "Gough's rule" of being highly anomalous (REITZEL *et al.*, 1970). This station, located close to the town of Crawford in western Nebraska, displayed a very large intensification in both the H_z and H_y component data at all periods, and the Fourier spectra maps of REITZEL *et al.* (1970) are visually dominated by its data. The data were so anomalous that they caused problems for PORATH *et al.* (1970) who attempted to separate the external and internal parts of the array data using two-dimensional Fourier integrals.

Again fortunately for THO and MT-DIW1, the response at this station was not ignored (one may wonder what "robust interpretation" would have done!), much to the credit of those involved, but was taken into consideration in the design of an array operated during July of 1969 in northwestern United States and southwestern Canada by CAMFIELD *et al.* (1970). The main purpose of that array was to study the relationship between the North American craton and the exotic terranes of the Cordillera, and to link together GDS profiles in southwestern U.S. with those in southern British Columbia (see JONES, 1993, for a review of the B.C. studies). However, CAMFIELD *et al.* (1970) chose to extend their array eastwards to study the anomaly at Crawford. What they discovered was a large and very narrow crustal structure running from the Black Hills of South Dakota north to the Canadian border along the boundary between Montana and the Dakotas. This feature they termed the "North American Central Plains" (NACP) conductivity anomaly, and it was the most prominent zone of enhanced conductivity found by their array study. The anomalous H_y field ($H_{y,a}$) was shown to be more than double the normal H_y field ($H_{y,n}$), which, CAMFIELD *et al.* (1970) argued, *"cannot be accounted for by local induction in an elongated structure"*. They suggested that the NACP anomaly could not be explained by the sedimentary structures in the Williston basin, nor by currents in the upper mantle, and proposed that the high conductivity and linear form could be due to a graphitic schist body in the basement, noting that high conductivities had been observed in graphite bodies in western South Dakota (MATHISRUD and SUMNER, 1967). Analysis of daily variation data showed that the NACP anomaly persisted in H_z, but was weak in the H_y component (CAMFIELD and GOUGH, 1975). It was concluded that the NACP conductor channels current induced in regions of unknown geometry by $H_{y,n}$ at substorm periods, and by $H_{z,n}$ at daily variation periods.

Data from a profile of the GDS array across the NACP at the latitude of the Black Hills were modelled by PORATH *et al.* (1971) using the 2D forward code of WRIGHT (1969) based on the transmission surface analogue of MADDEN and SWIFT (1969). They found that the normalized H_z data ($H_z/H_{y,n}$) at 50 min period (the period of maximum $H_z/H_{y,n}$ response) could be explained by a body of 0.1 Ω·m, 30 km wide, 3 km thick and 2 km deep. Although PORATH *et al.* (1971) believed the H_z results from Wright's code, they thought that the H_y results were highly erroneous as $H_{y,a}$ was greater than $H_{y,n}$. Accordingly, they derived the anomalous horizontal fields from a Hilbert transformation of the vertical fields. They were unable to explain the observed period-dependency of the normalized H_z amplitudes, and concluded that the NACP conductor must connect large regions in which the currents were actually induced and subsequently "channelled" through the NACP.

GOUGH and CAMFIELD (1972) considered that the graphitic schist interpretation for the source body of the NACP anomaly was strengthened by LIDIAK's (1971) mapping of a metamorphic belt striking just west of north through the Black Hills.

RANKIN and REDDY (1973) undertook a 15 station MT survey of the Black Hills, but only displayed the results of 3 of them in their publication. The two stations on the flanks of the Black Hills gave anisotropic apparent resistivity curves, with their minor curves (E-polarization) indicating conducting layers beneath the sediments at depths of 12.3 km and 3 km to the east and west respectively. In contrast, the data, albeit poor, from the station in the centre of the Black Hills gave isotropic apparent resistivities at around 1 Ω·m (0.4–35 Ω·m) in the period range 50–600 s. This information was not taken into consideration in their interpretation.

During August and September of 1972 a large 41 station GDS array was operated from Wyoming/Nebraska to northern Saskatchewan/Manitoba, over 13° in latitude and 10° in longitude, by ALABI *et al.* (1975) specifically to study the NACP anomaly. This array defined a north-south concentration of current from the Southern Rockies to just south of the exposed shield. ALABI (1974) and ALABI *et al.* (1975) suggested that the NACP anomaly marked a major structure in the Precambrian basement spatially linking mapped faults in the Wollaston belt of northern Saskatchewan with the metamorphic belt mapped by LIDIAK (1971) in the Black Hills.

This latter suggestion was elaborated on by CAMFIELD and GOUGH (1977) who courageously and perceptively proposed that the NACP anomaly traces a Proterozoic plate boundary for some 1800 km beneath the Paleozoic sediments of the Central Plains. In support of this interpretation, they cited HILLS *et al.* (1975) who had suggested that there was a Proterozoic subduction zone in southeastern Wyoming. CAMFIELD and GOUGH (1977) reiterated their earlier interpretation that the NACP structure is due to graphitic sheets in highly metamorphosed and folded basement rocks.

To identify definitively the spatial correlation of the NACP anomaly with one of the domains of the Churchill Province of the exposed Precambrian structures in northern Saskatchewan, HANDA and CAMFIELD (1984) undertook a profile of 7 GDS sites in July, 1981, across the Wollaston belt into the THO internides. The stations were located in the belief that the NACP anomaly was associated with the Wollaston domain, and only two sites were positioned within the THO. As is always the case ("Gough's rule", otherwise known as "Murphy's Law"), the anomaly was found to lie between the two sites at the eastern end of the line; one on the Rottenstone domain and the other at the La Ronge/Glennie domains boundary (see JONES *et al.*, 1993, for a domain map and further geological review of the exposed bedrock). The NACP anomaly was modelled two-dimensionally as a 10 Ω·m body at a depth of between 5–10 km with a thickness extent of some 20 km, i.e., a vertically-integrated conductance of 2,000 S.

Data from an International Magnetospheric Study (IMS) chain of magnetometers running north-south alongside Hudson's Bay in Manitoba were shown to exhibit a strong reversal in H_z between two stations; BCK (Back: 57.7°N) and GIM (Gillam: 56.4°N) (GUPTA *et al.*, 1985). These data were modelled two-dimensionally as a 25 Ω·m body at a depth of 5 km, of weakly-resolved thickness extent but with a definite northward dip. On the basis of spatial correlation with exposed tectonic units of the THO internides, this anomaly was associated with the NACP structure.

As mentioned in the Introduction, the PanCanadian MT surveys took place in 1984 and 1985. The 1984 survey was based on the NACP location from the GDS studies, but the strongest responses were observed on the easternmost sites (*PCSE01–PCSE04*), not on the central sites (*PCS003–PCS005*) as expected. Accordingly, a second survey was commissioned during 1985 to continue the MT profile to the east (sites *PC5014–PC5000*).

Subsequent to these surveys, MT measurements were made just north of the PanCanadian profile by MAIDENS and PAULSON (1988; see also JONES, 1988b), and RANKIN and PASCAL (1990). The former interpreted static shift effects in terms of structure, whereas the latter, notwithstanding their imaging of the NACP anomaly, concluded, somewhat perplexingly, that there was a gap in the NACP structure so that it did not cross their profile.

The TOBE feature was initially discovered by RANKIN and KAO (1978) in their data from 9 MT stations across the Superior-Churchill boundary in southern Manitoba. However, RANKIN and KAO (1978) did not fully appreciate the magnitude of their discovery, because they concluded that the E-polarization responses from the site which exhibited the largest anisotropy in apparent

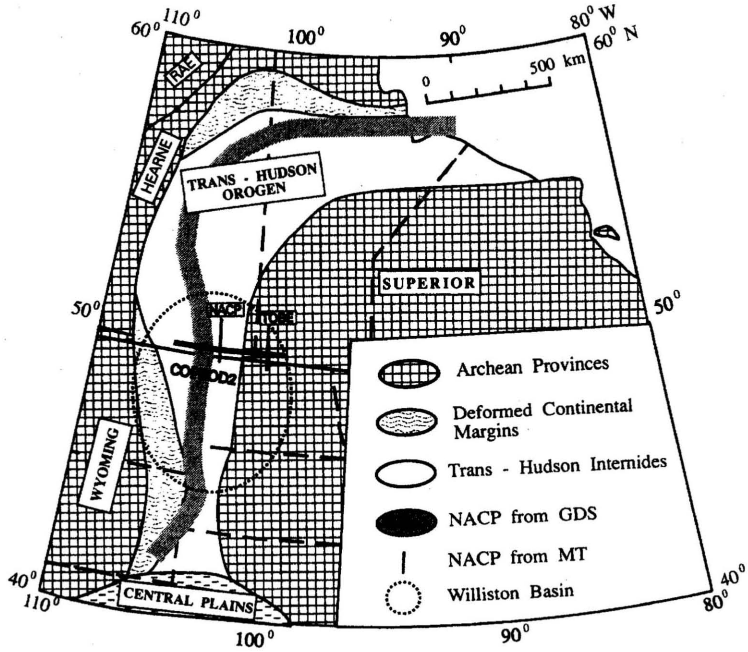

Fig. 1. The Trans-Hudson orogen and bounding Archean provinces of North America (based on HOFFMAN, 1988). The trace of the NACP anomaly from GDS studies is shown (based on ALABI *et al.*, 1975), and its position along the COPROD2 profile, which is 75 km east of the GDS location.

resistivity curves (site 9) could not be explained by any "reasonable" model. Accordingly, their pseudo-2D interpretation, constructed by stitching together 1D models from their E-polarization data, with the exception of site 9 where they used the B-polarization data, completely missed the vertical zone of enhanced conductivity causing the TOBE anomaly.

3. Tectonic Setting and Other Geophysical Results

The Trans-Hudson Orogen (THO) was named, as an afterthought and without precise definition, by HOFFMAN (1981). Its trace is shown in Fig. 1 (taken from HOFFMAN, 1988), together with the bounding Archean provinces and the 1.80–1.63 Ga Central Plains orogen (SIMS and PETERMAN, 1986) which truncates the THO to the south. The THO lies in the eastern segment of the previously named Churchill Province, which was a motley assemblage of ill-defined geological entities left over after all others on the Canadian Shield had been defined. Geological mapping during the 1970s (see HOFFMAN, 1990, and LEWRY and COLLERSON, 1990, for brief historical accounts) revealed the duality of the province, with an eastern 400 km wide Paleoproterozoic juvenile crustal segment and the rest underlain by Archean crust ("Rae" and "Hearne" provinces, HOFFMAN, 1988).

In northern Saskatchewan and Manitoba the orogen preserves, albeit cryptically but better than any other coeval orogen, virtually a complete record of Paleoproterozoic continent-arc-continent subduction and collision processes associated with the closure of the Manikewan ocean (STAUFFER, 1984), estimated to have been as wide as the Atlantic Ocean (5000 km) at about 1865 Ma (SYMONS, 1991). A recent special volume of the Geological Association of Canada (LEWRY and STAUFFER, 1990) details the extent of knowledge of the THO up to 1987. Recently, geophysical surveys have been undertaken under the auspices of LITHOPROBE (LUCAS *et al.*, 1993; JONES *et al.*, 1993) across the THO in northern Saskatchewan and Manitoba. Various tectonic models have been proposed for the northern segment (for example GREEN *et al.*, 1985, see LEWRY *et al.*, 1986, and GREEN *et al.*, 1986), but the southern segment beneath the Paleozoic sediments is more problematic (KLASNER and KING, 1990). What is now known with some certainty is that the Wyoming and Superior provinces did not share a common Archean history (FROST, 1993), contrary to earlier speculation (PETERMAN, 1979; and references in DUTCH, 1983).

Confirmation of CAMFIELD and GOUGH's (1977) proposal that a Proterozoic tectonic collision zone lies beneath the Central Plains came from interpolation to the south of known tectonic domains in the exposed THO internides beneath the sediments using principally aeromagnetic and gravity maps (GREEN *et al.*, 1979, 1985; DUTCH, 1983; KLASNER and KING, 1986; THOMAS *et al.*, 1987) combined with geological and geochronological analyses of chips and cores from basement-reaching boreholes (PETERMAN, 1981; VAN SCHMUS and BICKFORD, 1981; FROST and BURWASH, 1986; KLASNER and KING, 1986; SIMS and PETERMAN, 1986).

During the late-1970s and early 1980s the COCRUST[1] group undertook a series of seismic refraction profiles in southern Saskatchewan (MOREL-Á-L'HUISSIER *et al.*, 1986, 1990). The targets of study were the eastern edge of the Superior craton, the THO, the NACP anomaly, and the Williston basin. Unfortunately, the survey design was based on the location of the NACP anomaly from the GDS array measurements, and so the strike profiles intended to sample the structure were located too far to the west. The only profile crossing the NACP structure was a dip profile with little inherent sensitivity to the body. There is a lower crustal low velocity zone within the interpreted east-west cross-section, but it appears just west of 104° longitude at a depth of about 22 km.

COCORP[2] recently recorded a profile of deep seismic reflection data across the THO just south of the U.S./Canadian border in northern Montana and northern North Dakota (NELSON *et al.*, 1993). Their data showed arcuate reflections centred about the depth and lateral position of the NACP anomaly. The interpretation presented by NELSON *et al.* (1993) is of an arcuate structure within the basement cored by an unreflective Archean microcontinent, of unknown affinity, over which are draped reflective and conductive continental margin arc and oceanic arc rocks. This view contrasts sharply with the previously prevailing interpretation of the THO being underlain by the Superior Province.

Gravity gradient maps of the Central Plains (THOMAS *et al.*, 1987) showed linear features correlative with the NACP anomaly. In South Dakota, the NACP anomaly is thought to lie along the eastern boundary of the Wyoming province (THOMAS *et al.*, 1987; KLASNER and KING, 1986, 1990), although the Black Hills are interpreted to have been a marginal fold belt (HOFFMAN, 1988, and references therein). Further north, in North Dakota and southern Saskatchewan, THOMAS *et al.* (1987) suggest that the NACP anomaly lies well to the east of the suture between the THO's internides and the Archean province boundaries. In northern Saskatchewan and Manitoba, the NACP anomaly and the suture are again interpreted to be coincident.

At the location of the COPROD2 profile, there is a magnetic quiet zone spatially coincident with the NACP anomaly (JONES and CRAVEN, 1990; Fig. 7). Its sharp onset to the east at 102°33' is striking. The high amplitude magnetic anomaly just to the east of this onset is interpreted to be the expression of the extension of the Tabbernor fault beneath the Paleozoic sediments (GREEN *et al.*, 1985), and JONES and SAVAGE (1986) suggested that the Tabbernor fault bounds the NACP structure to the east. There is a local gravity high, of some 20 mgal above background, which appears to be spatially coincident with the NACP structure (JONES and CRAVEN, 1990; Figs. 6 and 15).

[1]Canadian Consortium for Crustal Reconnaissance Using Seismic Techniques.
[2]Consortium for Continental Reflection Profiling.

Also spatially coincident with the NACP anomaly is a north-south trending heat flow anomaly, of greater than 100 mW/m^2, below the Paleozoic surface (MAJOROWICZ *et al.*, 1986, 1988, 1989). MAJOROWICZ *et al.* (1988) discuss this correlation, and show that the trace of the NACP anomaly is also that of the Nesson anticline and of major oil occurrences in North Dakota. The Nesson anticline is a surface structure that was active throughout the Phanerozoic history of the basin (GERHARD *et al.*, 1982), and was originally a positive feature during initial basin subsidence (AHERN and MRKVICKA, 1984). The earliest recorded Paleozoic deposition in the Williston basin, the Deadwood Formation, was at 495 Ma in an elongated north-south trough, up to 270 m in depth, at about the location of the Nesson anticline (LEFEVER, 1992). LEFEVER (1992) suggested that there was a basement movement contemporaneous with onset of deposition. This age (495 Ma) is unusual in that there are no known coeval tectonic events elsewhere in North America.

The cause of the subsidence of the intracratonic Williston basin is still an open question. There are currently three hypotheses that warrant serious consideration: lithospheric contraction from cooling of a local thermal anomaly in the lithosphere (e.g., TURCOTTE and AHERN, 1977; SLEEP *et al.*, 1980); thermally-driven gabbro/eclogite phase change in the lower crust and/or upper mantle (FOWLER and NISBET, 1985); and localized rifting (e.g., DEV KLEIN and HSUI, 1987). Understanding the nature of the NACP anomaly can possibly aid in discriminating between these.

Finally, the lithosphere beneath the THO is now confirmed to be anomalous compared to the bounding Wyoming and Superior continental lithospheres (SILVER and KANESHIMA, 1993). Beneath the Archean cratons the lithospheres are strongly anisotropic in their seismic propagation. In contrast, the THO lithosphere beneath the Dakotas appears to be seismically isotropic.

It is certainly perplexing that right on top of the 2000 km long linear Paleoproterozoic THO is the circular Williston basin. However, initial subsidence was in an elongated north-south trench, so perhaps the Proterozoic lithosphere beneath the THO did control formation of the basin. The time scale of onset of subsidence, about 1 By after the THO collision, is of the right order for the phase transformation hypothesis (FOWLER and NISBET, 1985; HAID *et al.*, 1992).

4. The COPROD2 Dataset

The apparent resistivity and phase curves from the COPROD2 dataset have been published in a number of papers (JONES and SAVAGE, 1986; JONES, 1988a, 1988b; JONES and CRAVEN, 1990). Accordingly, all that will be displayed here are the data from representative stations, to show the data quality and the magnitude of the responses to the NACP and TOBE anomalies.

The site locations are listed in Table 1 in both geographic co-ordinates and UTMs. Also given is the location of the town of Macoun, which was chosen as the centre of the profile. All locations in the resulting data and models refer to positions east of Macoun.

Prior to distribution, the data were corrected for static shifts (JONES, 1988a). The technique applied was to try to find a parametric description of the conducting sedimentary layers which would provide the correct determination of the depths to resistive basement, known from well-log information from nine boreholes along the profile. At periods less than about 10 s, the responses are consistent with a 1D structure, and can be inverted as such for the layered conductivity structure of the Williston basin. It is important to note that the resistive basement is not the geological basement. For much of the profile, resistive basement, defined as the depth to an increase in resistivity from 3–10 Ω·m to 30–100 Ω·m, is the top of the Lower Paleozoic sequences, given by the Ashern dolomite marker bed. However, to the east the resistive basement interface is within the Lower Paleozoic sequences due to high fluid salinity from leeching of the Prairie evaporite, and to the west the resistive basement interface is higher in the section within the Mississippian (Carboniferous). With these *a priori* controls, the best parametric description of the conductive sedimentary layer was shown to be when it was assumed that the resistivity

Table 1. Co-ordinates of the COPROD2 dataset sites.

Station	Latitude	Longitude	UTM Easting	UTM Northing	Elevation (m)
Macoun	49.31310	103.2596	626500	5463500	
PC5000	49.33391	100.0633	422750	5464900	495
PC5001	49.31874	100.3140	404500	5463500	495
PC5002	49.33227	100.5959	384050	5465400	457
PC5003	49.31349	100.7735	371100	5463600	442
PC5004	49.34145	100.9369	359300	5467000	442
PC5005	49.33937	101.1316	345150	5467150	465
PC5006	49.33893	101.4063	325200	5467700	488
PC5007	49.32947	101.6535	307200	5467250	533
PC5008	49.33726	101.9445	286100	5468900	556
PC5009	49.31078	102.1658	706000	5465650	586
PC5010	49.32113	102.3751	690750	5466250	586
PC5011	49.33128	102.5088	681000	5467050	586
PC5012	49.26210	102.6861	668350	5458950	586
PC5013	49.32367	102.9447	649350	5465250	586
PC5014	49.31789	103.1913	631450	5464150	579
PCSE04	49.29901	103.3419	620550	5461800	579
PCSE03	49.32224	103.4602	611900	5464200	568
PCSE02	49.30353	103.6169	600550	5461900	586
PCSE01	49.28967	103.7417	591500	5460200	602
PCS001	49.31187	103.8898	580700	5462500	609
PCS002	49.30411	104.0261	570800	5461500	628
PCS003	49.29881	104.1527	561600	5460800	732
PCS305	49.28843	104.2114	557350	5459600	719
PCS004	49.27137	104.2845	552050	5457650	701
PCS405	49.28746	104.3441	547700	5459400	693
PCS005	49.30497	104.4236	541900	5461300	713
PCS006	49.31404	104.5391	533500	5462250	746
PCS007	49.30183	104.6479	525600	5460850	740
PCS008	49.29233	104.8212	513000	5459750	747
PCS009	49.31496	104.9684	502300	5462250	770
PCS010	49.31222	105.0970	492950	5461950	701
PCS011	49.32057	105.2319	483150	5462900	732
PCS012	49.29408	105.3851	472000	5460000	746
PCS013	49.30180	105.4855	464700	5460900	709
PCS014	49.30544	105.6582	452150	5461400	876

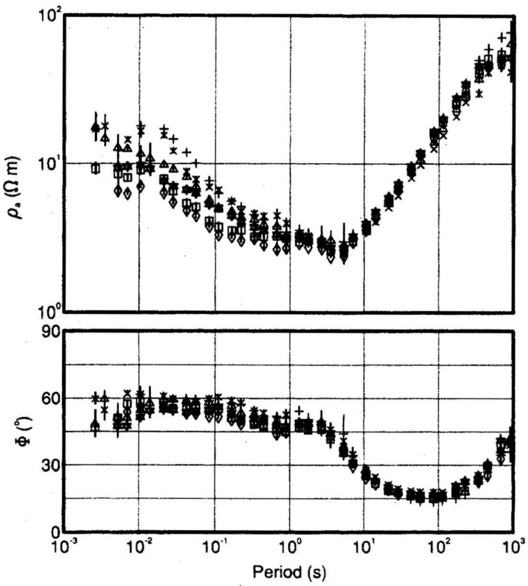

Fig. 2. The MT responses for both modes from the three stations at the western end of the profile (*PCS014, PCS013 & PCS012*).

decreased linearly from east to west, with 3.2 Ω·m at the eastern end, 2.6 Ω·m in the middle, and 2.2 Ω·m at the western end (JONES, 1988a, Fig. 13). When the appropriate static shift corrections were applied to the data from all sites, the depths to the base of the conductive sedimentary layer from 1D MT models then correlated well with the known depth to resistive basement. However, in order to propose a method for use where there is no well-log control, the modal value of all the estimates of the conducting layer's resistivity was chosen. This value was 3 Ω·m (JONES, 1988a, Fig. 14), and biasing was observed with the resistive basement depths being overestimated, by 1.16 (i.e., $\sqrt{3.0/2.2}$), beneath those sites at the western end of the profile. Accordingly, all apparent resistivity curves for both modes were shifted so that they gave a conducting layer of 3 Ω·m in the sedimentary section. This procedure also assures that the apparent resistivity curves from both modes lie on top of each other to be consistent with the assumption of a 1D shallow Earth.

Figure 2 displays the MT responses for both modes from the three stations (*PCS014, PCS013 & PCS012*) at the western end of the profile. There are differences at periods less than 1 s due to the lateral variation in the topmost resistive sedimentary layer. At periods greater than 1 s however all curves for both modes lie virtually on top of one another. There are subtle differences between the two modes, of up to 4° in phase at 100 s for site *PCS014*, but these are second order compared to the dominant 1D lateral homogeneity expressed by these data.

Figure 3 shows the responses from the three stations over the top of the NACP anomaly (*PCSE03, PCSE04 & PC5014*). Whereas the anomaly has a very strong effect on the E-polarization responses at periods beyond 20 s (the three minor resistivity curves and higher phase curves), it has no noticeable effect on the B-polarization curves up to 100 s period (compare these curves with Fig. 2). However, there is a weak response, of less than 2°, in the B-polarization phases at periods in the range 40 to 125 s which is discussed in detail below.

The responses from three sites closest to 102°W longitude (*PC5009, PC5008 & PC5007*) are shown in Fig. 4. These sites lie between the two anomalies, NACP and TOBE. Their responses

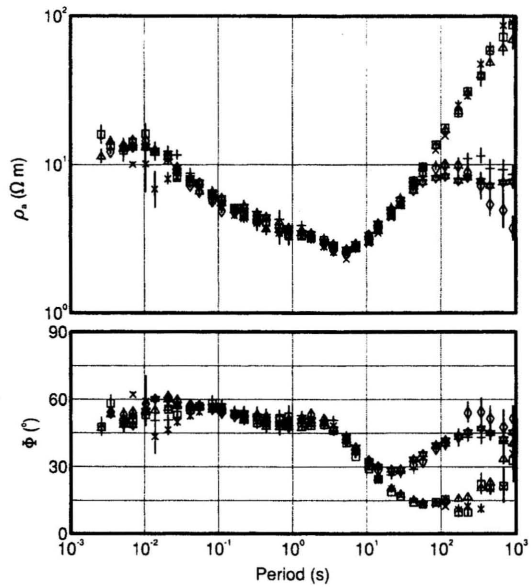

Fig. 3. The MT responses for both modes from the three stations exhibiting the greatest E-polarization mode sensitivity to the NACP anomaly (*PCSE03 (pluses)*, *PCSE04 (diamonds) & PC5014 (stars)*). The B-polarization responses exhibit no visible effects due to the presence of the NACP on this scale (compare with Fig. 2).

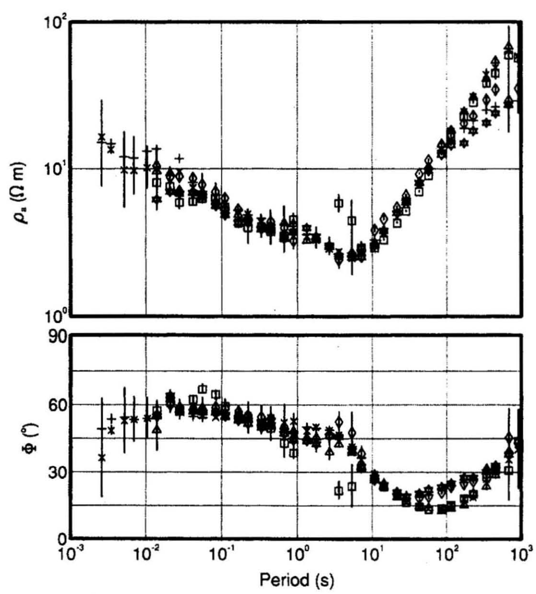

Fig. 4. The MT responses for both modes from the three stations between the NACP and TOBE anomalies (*PC5009, PC5008 & PC5007*).

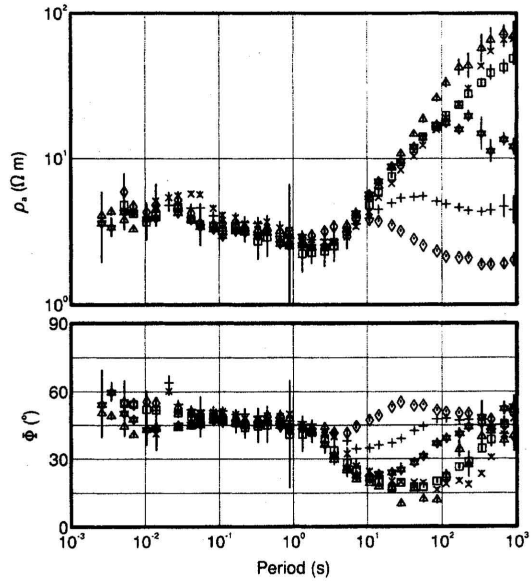

Fig. 5. The MT responses for both modes from the three stations exhibiting the greatest E-polarization mode sensitivity to the TOBE anomaly (*PC5004* (*pluses*), *PC5003* (*diamonds*) & *PC5002* (*stars*)). Note that the response drops off faster to the east (*PC5002*) than it does to the west (*PC5004*).

are close to 1D, which means that the two anomalies can be modelled independently of each other.

Above the TOBE anomaly, the data from site *PC5003* shows a very strong E-polarization response (Fig. 5, *diamonds*). This response drops off more rapidly to the east (*PC5002*, Fig. 5, *stars*) than to west (*PC5004*, Fig. 5, *pluses*). There is also a noticeable response in the B-polarization apparent resistivity data to the TOBE anomaly.

Finally, Fig. 6 shows the MT responses from the three sites at the eastern end of the profile (*PC5002, PC5001* & *PC5000*). Again, the dominant response is from a 1D homogeneous Earth with some variation in the sedimentary layers.

5. Comparison of Models

At the outset, the intention of COPROD2 was to compare inversion *algorithms*. However, as the reader will discover on studying the contributions in this special issue, this did not occur. Due to the differing approaches and computational requirements, virtually no two individuals or groups interpreted exactly the same data. Most used subsets of both sites and frequencies to make the problem tractable. Many contributors did not use the error estimates as given, but either used an error *floor*, or set a constant value for all data. In addition, DEGROOT-HEDLIN and CONSTABLE (1993) and WU *et al.* (1993) computed their own static shift factors from the original data to test approaches for their evaluation (see DEGROOT-HEDLIN, 1991a, for a description of the former). Accordingly, this COPROD2 exercise actually compares modelling and inversion *methodologies*, rather than just the algorithms alone.

Figure 7 shows the thirteen models described in this special issue, together with the two models in JONES and CRAVEN (1990), of the NACP anomaly. For this comparison only the 100 km on either side of Macoun are shown. Many contributors also modelled the TOBE structure, but

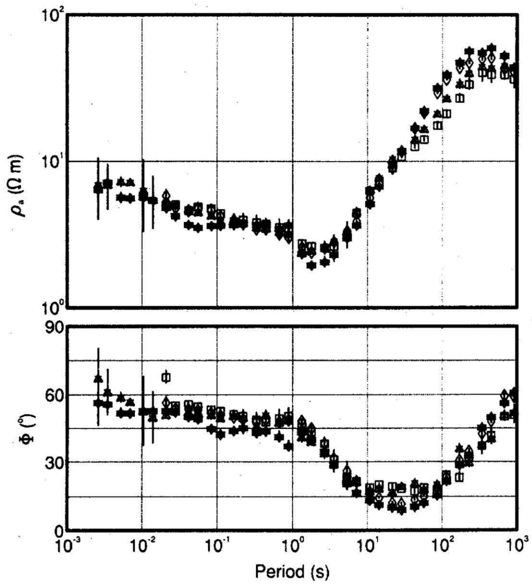

Fig. 6. The MT responses for both modes from the three stations at the eastern end of the profile (*PC5002, PC5001 & PC5000*).

the site separation was too large for adequate resolution, and no comparisons will be made of it. The data used, and the misfit statistics, are listed in Table 2. The final column gives the estimated average phase misfit assuming that the misfits were equi-partitioned between the two modes and between the apparent resistivities and phases. These phase misfit estimates were derived by multiplying the RMS misfit with the adopted phase error floor, or as described in the notes accompanying the table. (Due to RASMUSSEN's (1993) error renormalization scheme, it is not known what equivalent phase misfit his RMS of 1.0 represents.) Accordingly, these phase misfits should be used as an approximate qualitative guide only.

The first four of these (labelled *jones-1* and *jones-2*, from JONES and CRAVEN, 1990; *zhao*, from ZHAO *et al.*, 1993; *takasugi*, from TAKASUGI *et al.*, 1993) were derived by forward 2D model fitting. For *jones* and *zhao*, start models were derived from 1D inversions of the E-polarization data. TAKASUGI *et al.* (1993) took great care to fit the sedimentary section of the profile before turning their attention to the deeper resistivity structure. For this they used the well-log data published in JONES and CRAVEN (1990). Trial studies indicated to TAKASUGI *et al.* (1993) that the locations of their conducting mid-crustal bodies are well constrained by the data. The quantitative misfit to the data are not given for any of these four models, but qualitative fits can be seen in the relevant publications.

The next seven models form a class in which the model space is overparameterized, and a lowest misfit residual model with the smoothest variation in parameterization (either *conductivity* or log(*conductivity*)) is sought. Model *ellis-1* (ELLIS *et al.*, 1993) sought to minimize the *L1*-norm of this misfit statistic, whereas all the other models sought to minimize the *L2*-norm. The three models *degroot-1*, *degroot-2* (DEGROOT-HEDLIN and CONSTABLE, 1993) and *rasmussen* (RASMUSSEN, 1993) used the Occam2 code of DEGROOT-HEDLIN and CONSTABLE (1990). Model *degroot-1* is the smoothest model without any constraints, whereas for model *degroot-2* the upper layers (sedimentary section) are penalized against a surface structure derived by inverting the high frequency data alone. RASMUSSEN (1993) employed a error-renormalization scheme, based

Table 2. COPROD2 model misfits.

Model	No. sites	Period. range	No. periods	RMS Misfit	Av. Phase Misfit
jones-1	22	10 - 1000	7		?
jones-2	22	10 - 1000	7		?
zhao	25	0.03 - 1000	14		?
takasugi	35	0.001 - 1000	7		?
degroot-1	23	14 - 910	7	1.0$^{(1)}$	2.9
degroot-2	23	14 - 910	7	1.0$^{(2)}$	4.35
rasmussen	35	0.05 - 910	11	1.0$^{(3)}$	?
wu	28	10 - 450	12	1.0$^{(4)}$	4.35
ellis-1	35	5 - 680	8	1.15$^{(5)}$	5.75
ellis-2	35	5 - 680	8	2.5$^{(5)}$	12.5
uchida	35	0.67 - 680	6	2.6$^{(6)}$	0.8
agarwal	20	10 - 680	7	0.00123$^{(7)}$	2.5/0.15$^{(7)}$
schmucker	20	85 - 680	4	0.0456$^{(8)}$	5.3
schnegg	10	100 - 1000	3	0.085$^{(9)}$	4.8
everett	10	85 - 680	4	1.48$^{(10)}$	4.3

1: Error floor set to 10% in ρ_a and 2.9° in phase.
2: Error floor set to 15% in ρ_a and 4.35° in phase.
3: Errors scaled to fit D^+ inversions. RMS is 1.0 for 11 periods used in inversion, but is 1.53 for 29 periods.
4: Error floor set to 10% in ρ_a and 4.35° in phase.
5: Error floor set to 10% in ρ_a and 5° in phase.
6: Error floor set to 1% in ρ_a and 0.3° in phase.
7: This misfit value is not an RMS, but a normalized weighted misfit statistic. The average phase errors are those for the E-polarization mode (2.5°) and B-polarization mode (0.15°) separately.
8: Not an RMS, but a mean squared residual.
9: This misfit value is not an RMS, but a normalized misfit based on a 2D version of the ϵ of FISCHER *et al.* (1981). If the misfit between the apparent resistivity and phase data for both modes are equi-partitioned, 0.085 corresponds to an average phase misfit of 4.8° (ϵ_ϕ=0.1 corresponds to 5.7° in phase, FISCHER *et al.*, 1981).
10: Errors set to 10% in ρ_a and 2.9° in phase.

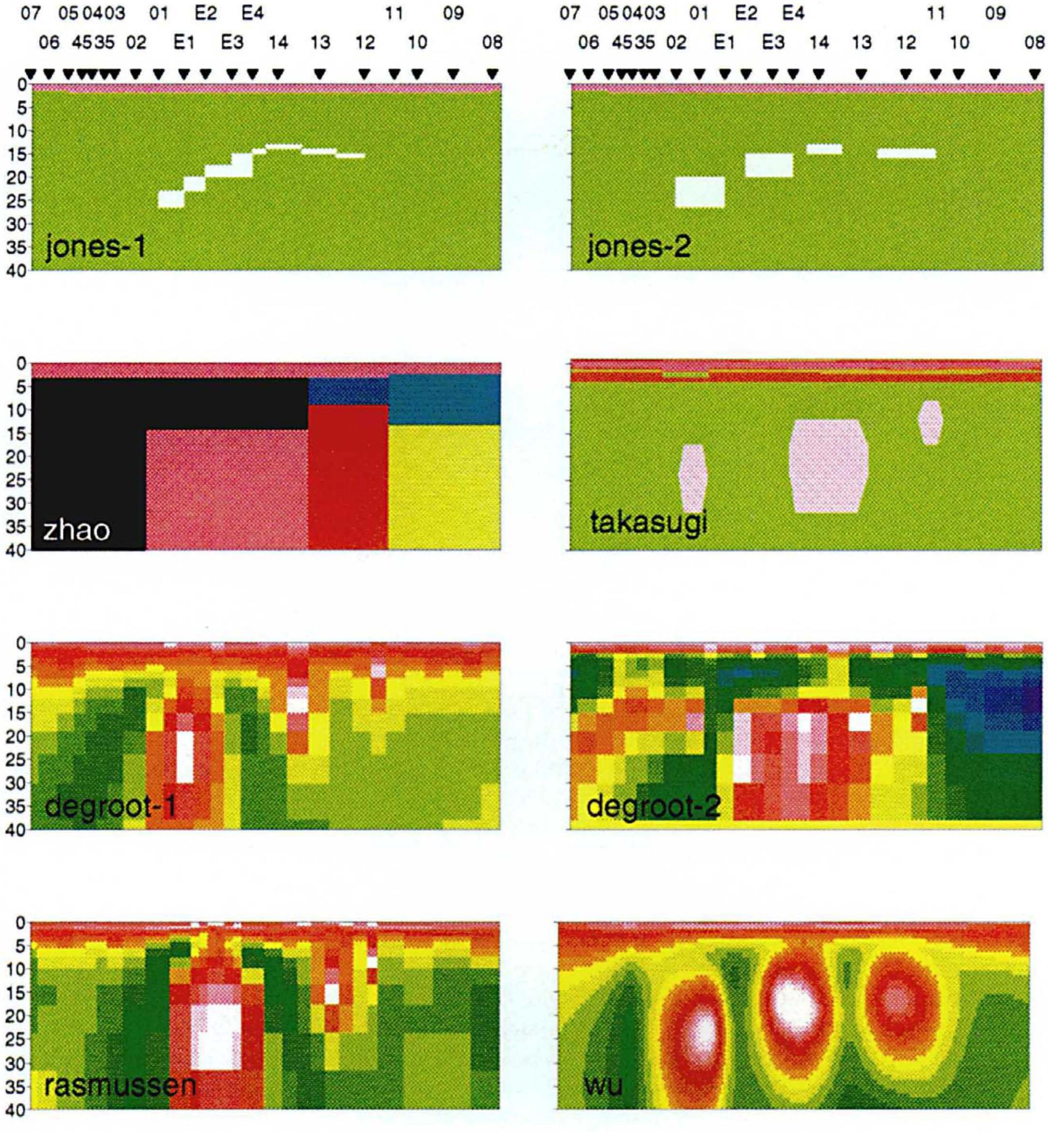

Fig. 7. Final models of the COPROD2 data from the contributions in this issue, plus the two models of JONES and CRAVEN (1990). The models are plotted for a distance of 200 km on either side of the central point (Macoun, 103.25°W) with a vertical exaggeration of 2:1. The colour scheme used is consistent for all models, ranging from white (<1 Ω·m) through red (10 Ω·m), yellow (100 Ω·m), green (1,000 Ω·m) and blue (10,000 Ω·m) to black (>100,000 Ω·m).

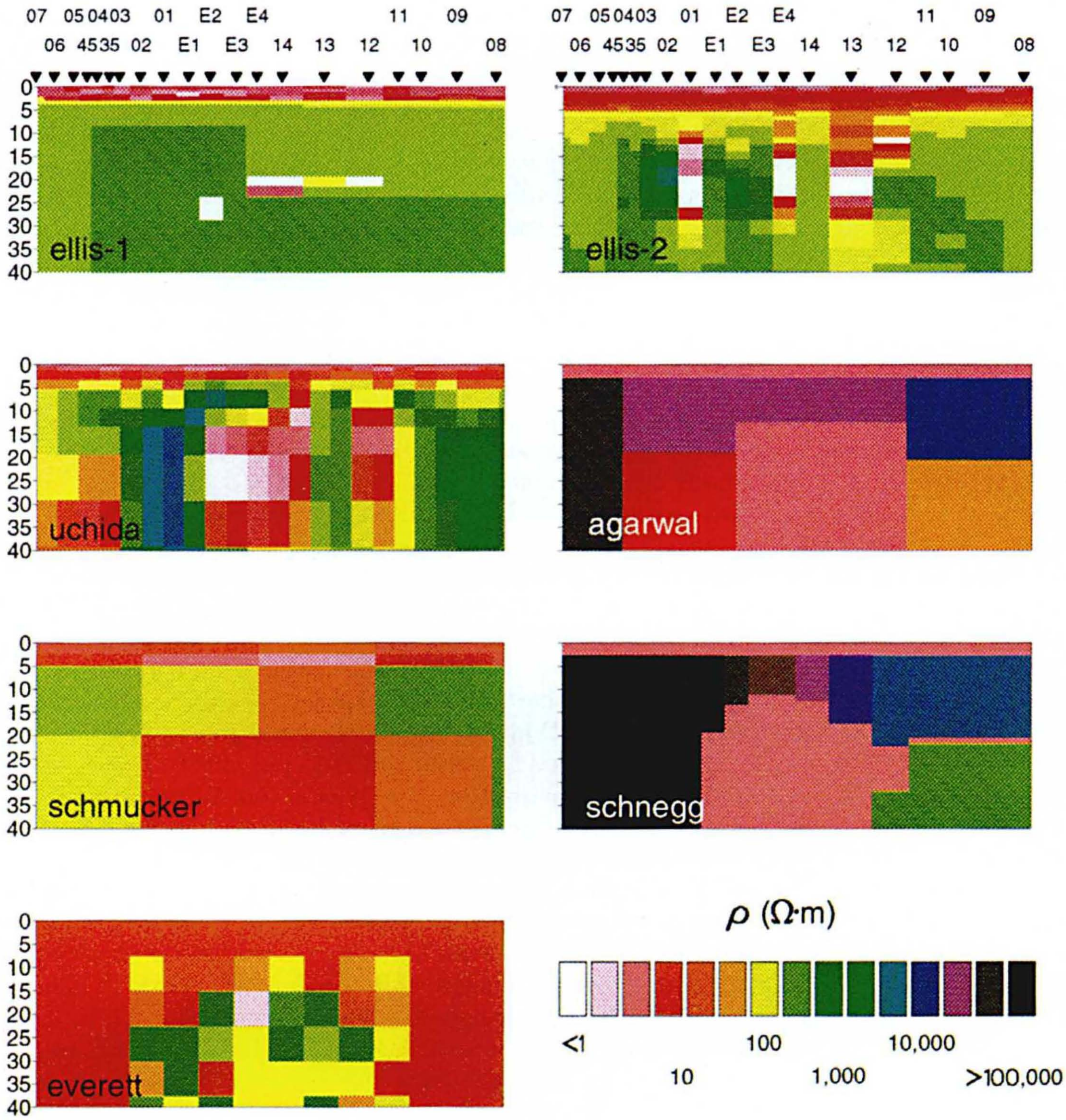

Fig. 7. (continued).

on application of PARKER's (1980) D^+ algorithm, to ensure that model-structure was as free as possible from error-structure. Model *wu* (WU *et al.*, 1993), derived using a modified version of SMITH and BOOKER's (1991) RRI algorithm, bears a striking resemblance to models *degroot-1* and *rasmussen.* The two models of ELLIS *et al.* (1993) were obtained using their AIM scheme (OLDENBURG and ELLIS, 1993). The approach of UCHIDA (1993) is similar to Occam with the exception that a statistical Bayesian criterion (ABIC) is used to define the optimum level of smoothness. Note the similarity between the *uchida* model and *degroot-2.*

The other four inversions are all different from those above. AGARWAL and WEAVER (1993) seek a model with *least structure* in the sense of having fewest blocks of differing resistivity. The optimum number of blocks is found when the misfit does not decrease significantly when the number of blocks is increased. SCHMUCKER (1993) inverted the E-polarization data alone with a philosophy akin to that of AGARWAL and WEAVER (1993); the fewest number of blocks, and their resistivities, are sought. The approach of SCHNEGG (1993) is radically different from the others using low-order polynomials to parameterize the variations in the conductivity boundaries in the Earth. Finally, EVERETT and SCHULTZ (1993) used a genetic algorithm (GA) search, basically a refined brute-force approach, on a high-speed computer to show what is possible in terms of hypothesis testing. Their approach is useful for checking the results of linearized inversions.

In the fifteen models presented, two fundamentally different models of the Earth to describe the COPROD2 data have emerged. The common feature in all of them is that there must exist an anomalous region of low resistivity, below 10 Ω·m, within the mid-crust. However, some models suggest that the anomaly must consist of several zones of very low resistivity (<1 Ω·m) separated by higher resistivity ("multi-body"; *jones-2, takasugi, degroot-1, degroot-2, rasmussen, wu, ellis-1, ellis-2, uchida, everett*), whereas others consider that there is no sensitivity in the data to such a feature and model the anomaly as a large block ("single-body"; *zhao, agarwal, schmucker, schnegg*). (It should be noted that model *jones-1*, although termed a "single-body" model by JONES and CRAVEN (1990), is actually a "multi-body" model as the conducting regions are only touching at their corners, hence there is no continuous conductivity path for B-polarization current.) This apparent dichotomy is significant, and warrants further consideration.

As noted by many, if the E-polarization data alone are inverted, then there is little sensitivity to a multi-body anomaly and the resulting 2D inversion will differ little from a 1D inversion of the E-polarization data (see, e.g., JONES and CRAVEN, 1990, and WU *et al.*, 1993). Hence, sensitivity to any breaks in a body comes from the B-polarization data (see Fig. 7 of WU *et al.*, 1993). These COPROD2 data are challenging because there is only a very weak variation in the B-polarization data due to the existence of the NACP anomaly.

Inverting the B-polarization data alone leads, perhaps perplexingly, to anomalous regions of high *resistivity*, not conductivity, in the basement (DEGROOT-HEDLIN, 1991b; WU *et al.*, 1993). The explanation for this apparent contradiction comes from close examination of the long period B-polarization data (Fig. 8). There is a local minimum, of the order of 1.5°, in the B-polarization phases at the location of the NACP anomaly centred around 70 s period. Whilst 1.5° may be only twice or three times the statistical error of a single phase datum, the site-to-site and frequency consistency of the phase minimum require that this feature be fit. It is important to note the lateral uniformity of B-polarization phases at periods shorter than 30 s. Local minima in B-polarization phases can be generated by varying the sedimentary section, but these minima will be at *all* periods. The phase minimum in Fig. 8 is a long period phenomenon, which is unequivocal evidence that it is a *basement*-related feature, and nothing to do with variation in the sedimentary sequences.

Two models were taken as representative of the two classes of models; model *uchida* (UCHIDA, 1993) for the multi-body anomaly, and model *agarwal* (AGARWAL and WEAVER, 1993) for the single-body anomaly. These two were chosen because they both have the smallest equivalent average phase misfit for their respective class. The B-polarization data, and the theoretical

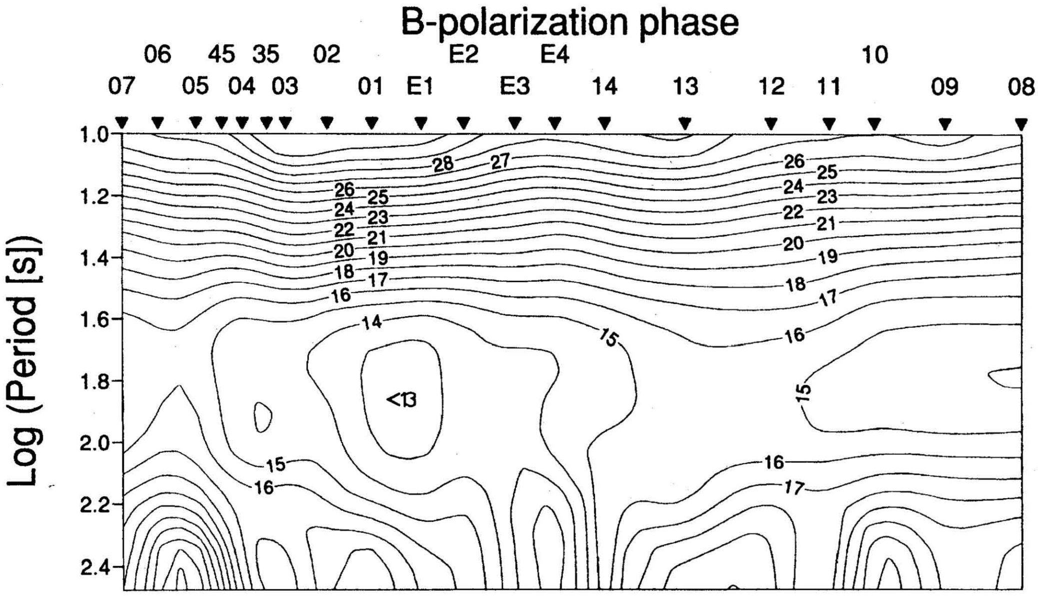

Fig. 8. B-polarization phase pseudosection for data in the period range 10 to 300 s and for stations within 100 km of the centre of the profile (Macoun, 103.25°W). Note the minimum with phases <14° in the period range 40 to 125 s for stations above the NACP anomaly.

responses for these two models, at periods of 56.9 s and 85.3 s (the periods in the centre of the local phase minimum of Fig. 8) are plotted in Fig. 9. It is apparent that, although visually one may estimate that both have approximately the same RMS misfit, the *uchida* model responses (*dashed lines*, Fig. 9) follow the undulations of the phase data closer that do the *agarwal* model responses (*solid curves*, Fig. 9). The misfit of the *agarwal* model is "coloured", i.e., the *shape* of the model response curve does not match the *shape* of the field data. In particular, taking into consideration the trend in the model data, there is a local *maximum* in the two *agarwal* model phase curves, in the region of 0 km to −50 km, where the field data show the minimum.

The fit of the *uchida* model is perhaps more impressive given that UCHIDA (1993) did not use these periods in his inversion. The closest period that Uchida did use was 42.7 s, and the fit of the *uchida* model at that period is very good (to within 0.5° on average).

Accordingly, in order to match the data well, and not to have misfit residuals that are "coloured", i.e., do not scatter randomly about the data but have a trend, the anomaly must be split into electrically-distinct segments—a "multi-body" anomaly.

A model which appears not to fit the data as well, but which also has multi-bodies, is that of WU *et al.* (1993). The responses of the *wu* model are also shown on the figure (*dashed curves*, Fig. 9). It is clear that although the model responses are not close to the field data in a misfit sense, they do display the correct curve *shape* with a local minimum. This minimum is centred about −75 km though, whereas the minimum in the data is centred around −30 km to −50 km. Why is it that this model appears to have, on the face of it, a higher misfit but to display higher resolution? The question becomes more perplexing given that Wu *et al.* assumed a phase error floor which is more that twice the magnitude of the observed phase minimum (4.5° cf. 1.5°)? Perhaps the crucial aspect about the *wu* model is that, due to the computational efficiencies of RRI, it was derived from the data at *all* twelve periods in the range 10 s to 450 s. Thus, the *trends* of subtle features may be fit, even though at individual periods there may be a substantial

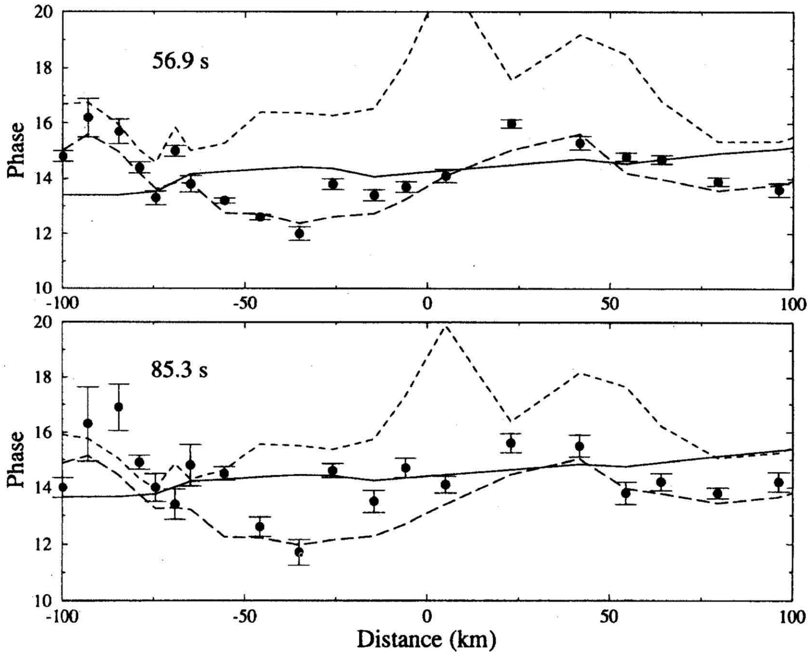

Fig. 9. B-polarization phases, and their errors, at periods of 56.9 s (upper graph) and 85.3 s (lower graph) together with the model responses to the *agarwal* (*solid lines*), *uchida* (*dashed lines*), and *wu* (*short dashed lines*) models.

misfit. A thorough analysis needs to be undertaken to determine why the models with large phase misfits, such as *wu*, display multi-bodies rather than single-bodies, especially since they are supposed to find the smoothest models that fit the data.

One other interesting aspect about the B-polarization data is that they permit a superior estimate of the resistivity of the bulk of the crust beneath the sediments to be obtained. For a 1D Earth, representative of sites at the western end of the profile, there is no resolution of the resistivity of the crust as it is sandwiched between two conducting layers—the sediments of the Williston basin and the conducting asthenosphere. All we can obtain is a minimum resistivity value for this region of the order of 200 Ω·m, the value adopted by JONES and CRAVEN (1990). Note, however, that in the multi-body inverse models the resistivity of the crust on the boundaries of the model is higher—at least 1,000 Ω·m and up to as much as 10,000 Ω·m (*uchida* model). The B-polarization data sense not the conductive region, but variations in conductivity, i.e., the breaks in the resistive layer, and accordingly give estimates of the resistivity of this layer in the vicinity of the conducting anomalies.

6. Conclusions

The COPROD2 dataset is a good start for comparison of the relative merits of various 2D modelling and inversion algorithms and methodologies. The underlying geological structures are dominantly 2D, and the data are, in the main, of high quality with small errors and small scatter. However, the dataset is challenging because to fit it properly means balancing the gross effects seen in the E-polarization data with the weak, but highly significant, effects in the B-polarization data. This is akin to the EMSLAB results—the successful imaging of a conductive region above the Juan de Fuca plate beneath the Coast Range mountains came from a 2° phase anomaly in the B-polarization data (WANNAMAKER *et al.*, 1989a, 1989b). The magnitude of this anomaly was close to the data errors, but it was consistent from site-to-site and so could not be discounted.

For various reasons, it is difficult herein to compare the resulting models in a quantitative manner. Errors were changed, static shifts were recalculated, and disparate subsets of both sites and frequencies were used. For future comparisons, it is essential that the model responses at all frequencies are reported, thus permitting measures of misfit to the actual data to be determined Nevertheless, much was learned both prior to and at the MT-DIW1 as participants analyzed the COPROD2 data.

Analyses and inversions to date of these data have highlighted the necessity for obtaining estimates of the errors that are as high quality as the estimates of the response functions themselves. Robust schemes which make no assumptions about the error distribution and use jackknife estimation procedures lead to such high quality error estimates. However, for many of our existing datasets, the time series are not available, only cross-spectral estimates, and accordingly effort should be expended in deriving error estimation algorithms which use cross-spectra. The fine structure of the responses, for example the B-polarization phases at a period of 56.9 s, must be fit to within reasonable tolerances. Enlarging the errors has been used by some in this comparison as a mechanism to ensure that their programs resulted in a model that fit to an RMS of one. In the author's opinion, there is little justification for making the error tolerances much larger than one degree for most of these data. The error renormalization approach of RASMUSSEN (1993) has potential merit, but it must be extended to 2D. For a variety of reasons, we need a tool in 2D that is as powerful and useful as Parker's D^+ is in 1D.

One note of caution, emphasized in some respects by Fig. 9 and the above discussion regarding error estimates, is that we must be careful about putting too much confidence into a single all-encompassing measure of misfit, such as RMS or equivalents. As a graphic example, the RMS misfit for the *agarwal* model at 85.3 s period is 2.80, which is only marginally lower than that for the *wu* model (2.83), but is much lower than that for the *uchida* model at 3.62. However, visually one would say that the *uchida* model response fits the field data better than does the *agarwal* model response, and certainly much better than does the *wu* model response. The problem is that the *uchida* model response misses two points with nominally small error, the 8th (−55.7 km) and the 11th (−25.9 km) from the west, whereas the *agarwal* model response goes directly through them. Without these two points, the RMS of the *uchida* model reduces to 2.48. Also, the *agarwal* model fits the B-polarization data to within 0.15° on average (A. Agarwal, personal communication, 1993), but it is apparent that it does not fit the data to such a tolerance at periods of 56.9 s and 85.3 s (Fig. 8). Obviously, the smoothness constraint ensures that the data are fit very well where there is little structure in the data, i.e., for B-polarization phases at periods less than 30 s or at sites outwith the range of the NACP anomaly, but are not fit well at exactly those periods/sites where there is significant structure in the data. A number of publications now discuss the use of a measure of trend misfit, such as the Spearman's statistic to test for systematic bias in the misfit residuals (SMITH and BOOKER, 1988; AGARWAL and WEAVER, 1993). However, this statistic has not been used as part of the objective function, only as model appraisal. We must try to develop objective functions that include not only misfit measures and model constraints, but also misfit trend measures. (Such complex objective functions are possible with EVERETT and SCHULTZ's (1993) genetic algorithm approach.) At the very least, once a model has been found the fit to the data should be plotted on a scale commensurate with the data errors; comparisons using plots of the size of Figs. 2 to 5 are inadequate for modern, high quality, MT data. Also, modelling the response at every period over the whole bandwidth should become the norm, rather than the exception.

Another key conclusion one can draw is that static shifts are very important, and their estimation should be treated with as much care as the model finding and appraisal exercises (see contributions by DEGROOT-HEDLIN and CONSTABLE (1993) and WU *et al.* (1993)). The approach suggested by JONES and CRAVEN (1990) worked reasonably well for the COPROD2 data, but should be thought of as a first step. Refining the shift factors should be part of the inversion methodology. In connection with static shifts, inversion of the topmost layers should be undertaken initially, and these layers fixed, or only changed if absolutely necessary, during the inversion of the longer period data, as undertaken by TAKASUGI *et al.* (1993) and DEGROOT-HEDLIN and CONSTABLE (1993). Models *degroot-1*, *rasmussen* and *wu* show sedimentary structure, particularly the deepening to 5 km to the east and west, which does not occur.

These COPROD2 data have highlighted for us the sensitivity in the B-polarization data to multiple breaks in the resistive crustal layer, whereas the E-polarization data sense the regions of enhanced conductivity. Note that if the anomaly is modelled as a single body, e.g., *agarwal* model, then the B-polarization phase response is positive, i.e., an increase in phase is observed, and so this mode becomes sensitive to the enhancement in conductivity.

Finally, the models of the COPROD2 data fall into two dichotomous classes. All models are in agreement that there is a region of low resistivity beneath the Central Plains, and that the data can be validly modelled in 2D, notwithstanding earlier comments to the contrary (see Introduction). However, some models suggest that there is little intrinsic resolution in the data for the fine structure of the anomalous region, and prefer to characterize it as a single conducting body. In stark contrast, other models suggest that there is sufficient information within the data to be able to state definitively that the anomalous region must consist of multiple bodies of very high conductivity (>1 S/m) separated by resistive regions. These latter models do appear to explain better some of the fine details of the B-polarization responses at periods in the range 40 to 200 s (Fig. 9). Also, inversion of recently-acquired MT data from the edge of the Phanerozoic sediments in northern Saskatchewan show unequivocally that the NACP anomaly at that latitude exists as at least two zones of enhanced conductivity (JONES *et al.*, 1993). (Spatial correlation with exposed bedrock just to the north implies that the NACP anomaly is caused by sulphides rather than graphitic metasediments.)

It may be possible that comparisons should be made of integrated conductances between different depths (see RASMUSSEN, 1993), as was shown to be useful for the COPROD data by WEIDELT (1985). However, given that the B-polarization data are sensitive to charges on conductivity gradients, this is less likely to be as useful.

The original data were supplied to the Geological Survey of Canada by Pete Savage of PanCanadian Petroleum Ltd. (Calgary). The author wishes to thank all participants of this COPROD2 exercise; most of the comments herein come from discussions at the MT-DIW1 and from reading the submitted manuscripts. In particular, extensive discussions with John Booker eventually led to a better understanding by the author of data modelling and inversion. Reviews and discussion of an earlier version of this manuscript by John Booker, John Weaver, Ron Kurtz and David Boerner are appreciated. Ashok Agarwal, John Booker and Toshi Uchida are thanked for providing their model responses shown in Fig. 9.

The COPROD2 dataset is available, either in SEG-EDI format or in impedance format, on application to the author by email to jones@cg.emr.ca.

Geological Survey of Canada contribution no. 28893.

REFERENCES

AGARWAL, A. K. and J. T. WEAVER, Inversion of the COPROD2 Data by a method of modelling, *J. Geomag. Geoelectr.*, this issue, 969–983, 1993.

AHERN, J. L. and S. R. MRKVICKA, A mechanical and thermal model for the evolution of the Williston Basin, *Tectonics*, **3**, 79–102, 1984.

ALABI, A. O., A study of the North American Central Plains conductivity anomaly, Unpublished Ph.D. Thesis, Univ. Alberta, 1974.

ALABI, A. O., P. A. CAMFIELD, and D. I. GOUGH, The North American Central Plains anomaly, *Geophys. J. R.*

Astr. Soc., **43**, 815–834, 1975.
CAMFIELD, P. A. and D. I. GOUGH, Anomalies in daily variation magnetic fields and structure under north-western United States and south-western Canada, *Geophys. J. R. Astr. Soc.*, **41**, 193–218, 1975.
CAMFIELD, P. A. and D. I. GOUGH, A possible Proterozoic plate boundary in North America, *Can. J. Earth Sci.*, **14**, 1229–1238, 1977.
CAMFIELD, P. A., D. I. GOUGH, and H. PORATH, Magnetometer array studies in the northwestern United States and southwestern Canada, *Geophys. J. R. Astr. Soc.*, **22**, 201–222, 1970.
DEGROOT-HEDLIN, C., Removal of static shift in two dimensions by regularized inversion, *Geophysics*, **56**, 2102–2106, 1991a.
DEGROOT-HEDLIN, C., Occam2: Regularized inversion to generate smooth, two-dimensional resistivity models from magnetotelluric data, Ph.D. Thesis, Univ. Calif., San Diego, 1991b.
DEGROOT-HEDLIN, C. and S. CONSTABLE, Occam's inversion to generate smooth two-dimensional models from magnetotelluric data, *Geophysics*, **55**, 1613–1624, 1990.
DEGROOT-HEDLIN, C. and S. CONSTABLE, Occam's inversion and the North American Central Plains electrical anomaly, *J. Geomag. Geoelectr.*, this issue, 985–999, 1993.
DEV KLEIN, G. and A. T. HSUI, Origin of cratonic basins, *Geology*, **15**, 1094–1098, 1987.
DUTCH, S. I., Proterozoic structural provinces in the north-central United States, *Geology*, **11**, 478–481, 1983.
ELLIS, R. G., C. G. FARQUHARSON, and D. W. OLDENBURG, Approximate inverse mapping inversion of the COPROD2 data, *J. Geomag. Geoelectr.*, this issue, 1001–1012, 1993.
EVERETT, M. E. and A. SCHULTZ, Two-dimensional nonlinear magnetotelluric inversion using a genetic algorithm, *J. Geomag. Geoelectr.*, this issue, 1013–1026, 1993.
FISCHER, G., P.-A. SCHNEGG, M. PEGUIRON, and B. V. LE QUANG, An analytic one-dimensional magnetotelluric inversion scheme, *Geophys. J. R. Astr. Soc.*, **67**, 257–278, 1981.
FOWLER, C. M. R. and E. G. NISBET, The subsidence of the Williston Basin, *Can. J. Earth Sci.*, **22**, 408–415, 1985.
FROST, C. D., Nd isotopic evidence for the antiquity of the Wyoming province, *Geology*, **21**, 351–354, 1993.
FROST, C. D. and R. A. BURWASH, Nd evidence for extensive Archean basement in the western Churchill Province, Canada, *Can. J. Earth Sci.*, **23**, 1433–1437, 1986.
GERHARD, L. C., S. B. ANDERSON, J. A. LEFEVER, and C. G. CARLSON, Geological development, origin, and energy mineral resources of Williston Basin, North Dakota, *Amer. Assoc. Petrol. Geol.*, **66**, 989–1020, 1982.
GOUGH, D. I. and P. A. CAMFIELD, Convergent geophysical evidence of a metamorphic belt through the Black Hills of South Dakota, *J. Geophys. Res.*, **77**, 3168–3170, 1972.
GREEN, A. G., G. L. CUMMING, and D. CEDARWELL, Extension of the Superior-Churchill boundary zone into southern Canada, *Can. J. Earth Sci.*, **16**, 1691–1701, 1979.
GREEN, A. G., Z. HAJNAL, and W. WEBER, An evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north-central United States, *Tectonophys.*, **116**, 281–322, 1985.
GREEN, A. G., Z. HAJNAL, and W. WEBER, Reply to Discussion by J. F. Lewry, K. D. Collerson, M. E. Bickford and W. R. Van Schmus on "An evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north-central United States", *Tectonophys.*, **131**, 188–197, 1986.
GUPTA, J. C., R. D. KURTZ, P. A. CAMFIELD, and E. R. NIBLETT, A geomagnetic induction anomaly from IMS data near Hudson Bay, and its relation to crustal electrical conductivity in central North America, *Geophys. J. R. Astr. Soc.*, **81**, 33–46, 1985.
HAID, J. H., E. G. NISBET, and C. M. R. FOWLER, Tectonic subsidence of the Williston basin, *EOS*, **73**, 321, 1992.
HANDA, S. and P. A. CAMFIELD, Crustal electrical conductivity in north-central Saskatchewan: the North American Central Plains anomaly and its relation to a Proterozoic plate margin, *Can. J. Earth Sci.*, **21**, 533–543, 1984.
HILLS, F. A., R. S. HOUSTON, and G. V. SUBBARAYUDU, Possible Proterozoic plate boundary in southern Wyoming, *Geol. Soc. Am. (Abstracts)*, **7**, 614, 1975.
HOFFMAN, P., Autopsy of Athapuscow aulacogen: a failed arm affected by three collisions, in *Proterozoic Basins of Canada*, edited by F. H. A. Campbell, Geol. Surv. Canada, Paper 81-10, pp. 97–102, 1981.
HOFFMAN, P., United plates of America, the birth of a craton: early Proterozoic assembly and growth of Laurentia, *Ann. Rev. Earth Planet. Sci.*, **16**, 543–603, 1988.
HOFFMAN, P. F., Subdivision of the Churchill Province and the extent of the Trans-Hudson Orogen, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, Geol. Assoc. Can. Spec. Paper 37, Geol. Assoc. Canada, pp. 15–39, 1990.
JONES, A. G., Static shift of magnetotelluric data and its removal in a sedimentary basin environment, *Geophysics*, **53**, 967–978, 1988a.
JONES, A. G., Discussion of "A magnetotelluric investigation under the Williston Basin" by J. M. Maidens and K. V. Paulson, *Can. J. Earth Sci.*, **25**, 1132–1139, 1988b.
JONES, A. G., The BC87 dataset: Tectonic setting, previous EM results, and recorded MT data, *J. Geomag. Geoelectr.*, this issue, 1089–1105, 1993.

JONES, A. G. and J. A. CRAVEN, The North American Central Plains conductivity anomaly and its correlation with gravity, magnetics, seismic, and heat flow data in the Province of Saskatchewan, *Phys. Earth Planet. Inter.*, **60**, 169–194, 1990.

JONES, A. G. and R. HUTTON, A multi-station magnetotelluric study in southern Scotland—I. Fieldwork, data analysis and results, *Geophys. J. R. Astr. Soc.*, **56**, 329–349, 1979.

JONES, A. G. and P. J. SAVAGE, North American Central Plains conductivity anomaly goes east, *Geophys. Res. Lett.*, **13**, 685–688, 1986.

JONES, A. G., J. A. CRAVEN, G. A. MCNEICE, I. J. FERGUSON, T. BOYCE, C. FARQUARSON, and R. G. ELLIS, The North American Central Plains conductivity anomaly within the Trans-Hudson orogen in northern Saskatchewan, *Geology*, 1993 (in press).

KLASNER, J. S. and E. R. KING, Precambrian basement geology of North and South Dakota, relationship to exposed Precambrian terranes, tectonic implications, *Can. J. Earth Sci.*, **23**, 1083–1102, 1986.

KLASNER, J. S. and E. R. KING, A model for tectonic evolution of the Trans-Hudson orogen in North and South Dakota, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, Geol. Assoc. Can. Spec. Paper 37, Geol. Assoc. Canada, pp. 271–285, 1990.

LEFEVER, R. D., Earliest stratigraphic record in the North Dakota Williston Basin: constraints on the age of origin and early subsidence history of the basin, *EOS*, **73**, 321, 1992.

LEWRY, J. F. and K. D. COLLERSON, The Trans-Hudson Orogen: extent, subdivision, and problems, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, Geol. Assoc. Can. Spec. Paper 37, Geol. Assoc. Canada, pp. 1–14, 1990.

LEWRY, J. F. and M. R. STAUFFER (editors), *The Early Proterozoic Trans-Hudson Orogen of North America*, Geol. Assoc. Canada Spec. Paper 37, 1990.

LEWRY, J. F., K. D. COLLERSON, M. E. BICKFORD, and W. R. VAN SCHMUS, Discussion on"An evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north-central United States" by A. G. Green, Z. Hajnal, and W. Weber, *Tectonophys.*, **131**, 183–197, 1986.

LIDIAK, E. G., Buried Precambrian rocks of South Dakota, *Geol. Soc. Am. Bull.*, **82**, 1411–1420, 1971.

LUCAS, S. B., A. G. GREEN, Z. HAJNAL, D. WHITE, J. LEWRY, K. ASHTON, W. WEBER, and R. CLOWES, Deep seismic profile across a Proterozoic collision zone: surprises at depth, *Nature*, **363**, 339–342, 1993.

MADDEN, T. R. and C. M. SWIFT, Magnetotelluric studies of the electrical conductivity structure of the crust and upper mantle, in *The Earth's Crust and Upper Mantle*, edited by P. J. Hart, AGU Monograph, **13**, pp. 469–479, 1969.

MAIDENS, J. M. and K. V. PAULSON, A magnetotelluric investigation under the Williston Basin of south-eastern Saskatchewan, *Can. J. Earth Sci.*, **25**, 60–67, 1988.

MAJOROWICZ, J. A., F. W. JONES, and A. M. JESSOP, Geothermics of the Williston Basin in Canada in relation to hydrodynamics and hydrocarbon occurrences, *Geophysics*, **51**, 767–779, 1986.

MAJOROWICZ, J. A., F. W. JONES, and K. G. OSADETZ, Heat flow environment of the electrical conductivity anomalies in the Williston Basin, and occurrence of hydrocarbons, *Bull. Can. Petrol. Geol.*, **36**, 86–90, 1988.

MAJOROWICZ, J. A., F. W. JONES, and M. E. ERTMAN, Model of terrestrial thermal effects in the area of crustal electrical conductors of the Williston Basin, *Can. J. Explor. Geophys.*, **25**, 45–49, 1989.

MATHISRUD, G. C. and J. S. SUMNER, Underground induced polarization surveying at the Homestake Mine, *Mining Congress J.*, **53**, 66–69, 1967.

MOREL-Á-L'HUISSIER, P., A. GREEN, and C. PIKE, Results from crustal refraction surveys in south-central Canada, *J. Geophys. Res.*, **92**, 6403–6420, 1986.

MOREL-Á-L'HUISSIER, P., A. G. GREEN, A. G. JONES, T. LATHAM, J. A. MAJOROWICZ, M. J. DRURY, and M. D. THOMAS, The crust beneath the intra-cratonic Williston Basin from geophysical data, in *The Potential of Deep Seismic Reflection for Hydrocarbon Exploration*, edited by B. Pinet and C. Bois, Les Editions Technip, pp. 141–160, 1990.

NELSON, K. D., D. J. BAIRD, J. J. WALTERS, M. HAUCK, L. D. BROWN, J. E. OLIVER, J. L. AHERN, Z. HAJNAL, A. G. JONES, and L. L. SLOSS, Trans-Hudson orogen and Williston basin in Montana and North Dakota: New COCORP deep profiling results, *Geology*, **21**, 447–450, 1993.

OLDENBURG, D. W. and R. G. ELLIS, Efficient inversion of magnetotelluric data in two dimensions, *Phys. Earth Planet. Inter.*, 1993 (in press).

PARKER, R. L., The inverse problem of electromagnetic induction: existence and construction of solutions based on incomplete data, *J. Geophys. Res.*, **85**, 4421–4425, 1980.

PETERMAN, Z. E., Geochronology and the Archean of the United States, *Economic Geol.*, **74**, 1544–1562, 1979.

PETERMAN, Z. E., Dating of Archean basement in northeastern Wyoming and southern Montana, *Geol. Soc. Am. Bull.*, **92**, 139–146, 1981.

PORATH, H., D. W. OLDENBURG, and D. I. GOUGH, Separation of magnetic variation fields and conductive structures in the western United States, *Geophys. J. R. Astr. Soc.*, **19**, 237–260, 1970.

PORATH, H., D. I. GOUGH, and P. A. CAMFIELD, Conductive structures in the north-western United States and south-western Canada, *Geophys. J. R. Astr. Soc.*, **23**, 387–398, 1971.

RANKIN, D. and D. KAO, The delineation of the Superior-Churchill transition zone in the Canadian Shield, *Can.*

Soc. Explor. Geophys., **14**, 50–54, 1978.
RANKIN, D. and F. PASCAL, A gap in the North American Central Plains conductivity anomaly, *Phys. Earth Planet. Inter.*, **60**, 132–137, 1990.
RANKIN, D. and I. K. REDDY, Crustal conductivity anomaly under the Black Hills: a magnetotelluric study, *Earth Planet. Sci. Lett.*, **20**, 275–279, 1973.
RASMUSSEN, T., Two-dimensional Occam model of COPROD2 data—first order description of resolution and variance, *J. Geomag. Geoelectr.*, this issue, 1027–1037, 1993.
REITZEL, J. S., D. I. GOUGH, D. I. PORATH, and C. W. ANDERSON, III, Geomagnetic deep sounding and upper mantle structure in the western United States, *Geophys. J. R. Astr. Soc.*, **19**, 213–235, 1970.
SCHMUCKER, U., 2D modeling with linearized integral equations, *J. Geomag. Geoelectr.*, this issue, 1045–1062, 1993.
SCHNEGG, P.-A., An automatic scheme for 2-D magnetotelluric modelling, based on low-order polynomial fitting, *J. Geomag. Geoelectr.*, this issue, 1039–1043, 1993.
SILVER, P. G. and S. KANESHIMA, Constraints on mantle anisotropy beneath Precambrian North America from a transportable teleseismic experiment, *Geophys. Res. Lett.*, **20**, 1127–1130, 1993.
SIMS, P. K. and Z. E. PETERMAN, Early Proterozoic Central Plains orogen: a major buried structure in north-central United States, *Geology*, **14**, 488–491, 1986.
SLEEP, N. H., J. A. NUNN, and L. CHOU, Platform basins, *Ann. Rev. Earth Planet. Sci.*, **8**, 17–34, 1980.
SMITH, J. T. and J. R. BOOKER, Magnetotelluric inversion for minimum structure, *Geophysics*, **53**, 1565–1576, 1988.
SMITH, J. T. and J. R. BOOKER, Rapid inversion of two and three-dimensional magnetotelluric data, *J. Geophys. Res.*, **96**, 3905–3922, 1991.
STAUFFER, M. R., Manikewan: an Early Proterozoic ocean in central Canada, its igneous history and orogenic closure, *Precam. Res.*, **25**, 257–281, 1984.
SYMONS, D. T. A., Paleomagnetism of the Proterozoic Wathaman batholith and the suturing of the Trans-Hudson orogen in Saskatchewan, *Can. J. Earth Sci.*, **28**, 1931–1938, 1991.
TAKASUGI, S., T. SATO, and Y. HONKURA, Two-dimensional forward modelling for the COPROD2 MT data, *J. Geomag. Geoelectr.*, this issue, 957–962, 1993.
THOMAS, M. D., V. L. SHARPTON, and R. A. F. GRIEVE, Gravity patterns and Precambrian structure in the North American Central Plains, *Geology*, **15**, 489–492, 1987.
TURCOTTE, D. L. and J. L. AHERN, On the thermal and subsidence history of subsidence basins, *J. Geophys. Res.*, **82**, 3762–3766, 1977.
UCHIDA, T., Inversion of COPROD2 magnetotelluric data by use of ABIC minimization method, *J. Geomag. Geoelectr.*, this issue, 1063–1071, 1993.
VAN SCHMUS, W. R. and M. E. BICKFORD, Proterozoic chronology and evolution of the midcontinent region, North America, in *Precambrian Plate Tectonics*, edited by A. Kroner, pp. 261–296, Elsevier, Amsterdam, 1981.
WANNAMAKER, P. E., J. R. BOOKER, J. H. FILLOUX, A. G. JONES, G. R. JIRACEK, A. D. CHAVE, H. S. WAFF, C. T. YOUNG, J. A. STODT, M. MARTINEZ, L. K. LAW, T. YUKITAKE, J. S. SEGAWA, A. WHITE, and A. W. GREEN, Magnetotelluric observations across the Juan de Fuca subduction system in the EMSLAB project, *J. Geophys. Res.*, **94**, 14111–14125, 1989a.
WANNAMAKER, P. E., J. R. BOOKER, A. G. JONES, A. D. CHAVE, J. H. FILLOUX, H. S. WAFF, and L. K. LAW, Resistivity cross-section through the Juan de Fuca subduction system and its tectonic implications, *J. Geophys. Res.*, **94**, 14127–14144, 1989b.
WEIDELT, P., Construction of conductance bounds from magnetotelluric impedances, *J. Geophys.*, **57**, 191–206, 1985.
WRIGHT, J. A., The magnetotelluric and geomagnetic response of two-dimensional structures, GAMMA 7, Inst. Geophys. Met., Techn. Univ. Braunschweig, pp. 102, 1969.
WU, N., J. R. BOOKER, and J. T. SMITH, Rapid two-dimensional inversion of COPROD2 data, *J. Geomag. Geoelectr.*, this issue, 1073–1087, 1993.
ZHAO, G., J. TANG, and Y. ZHAN, Two-dimensional modelling by using COPROD2 magnetotelluric data, *J. Geomag. Geoelectr.*, this issue, 963–968, 1993.