Monitoring temporal change in conductivity in the central Vancouver Island region, an area with past large earthquakes[1]

D. R. AULD
Pacific Geoscience Centre, Geological Survey of Canada, 9860 West Saanich Road, Sidney, B.C., Canada V8L 4B2

S. E. DOSSO[2] AND D. W. OLDENBURG
Department of Geophysics and Astronomy, The University of British Columbia, Vancouver, B.C., Canada V6T 1Z4

AND

L. K. LAW
Pacific Geoscience Centre, Geological Survey of Canada, 9860 West Saanich Road, Sidney, B.C., Canada V8L 4B2

Received July 30, 1991
Revision accepted November 22, 1991

Two major earthquakes, magnitude 7.0 in 1918 and magnitude 7.3 in 1946, have occurred this century in the central region of Vancouver Island, British Columbia, Canada. Levelling data in the region indicate relative uplift of 4 mm/year from 1977 to 1984, followed by subsidence at approximately the same rate over the next 2 years. In response to the observed elevation changes, a program was initiated to investigate if temporal changes in the geoelectrical conductivity might be associated with earthquake occurrence. Beginning in 1986, magnetotelluric (MT) data have been measured annually at a number of sites on central Vancouver Island to monitor the long-term variability of the conductivity of the crust and upper mantle in the region. Robust processing techniques now used in the analysis of MT data enhance the possibility of detecting changes in the conductivity.

Past studies involving the monitoring of MT stations have considered temporal change only in terms of the measured responses. However, formulating the inverse problem of constructing conductivity–depth models that vary minimally from year to year allows quantitative investigation of the changes required in the models to accommodate the yearly variations in the data. This provides a method of evaluating the processes and depths involved in observed changes in the data. Our modelling study indicates a small but systematic yearly decrease in conductivity from 1987 to 1990 localized in a conductive zone overlying the subducting Juan de Fuca Plate.

Deux grands tremblements de terre, de magnitude 7,0 en 1918 et de magnitude 7,3 en 1946, se sont produits durant ce siècle dans la région centrale de l'île Vancouver, Colombie-Britannique, Canada. Les données de nivellement topographique dans cette région indiquent un soulèvement relatif de 4 mm/an entre 1977 et 1984, suivi d'un affaissement au même taux, approximativement, sur les deux années suivantes. Ces variations de l'élévation observées ont incité le développement d'un programme de recherche pour vérifier si des changements dans le temps de la conductivité géoélectrique pouvaient être associés aux évènements de tremblement de terre. Depuis 1986, les données magnétotelluriques (MT) sont enregistrées annuellement sur des sites au centre de l'île Vancouver pour tenter de mesurer les variations à long terme de la conductivité de la croûte et du manteau supérieur dans la région. Les techniques très efficaces utilisées actuellement pour traiter les données magnétotelluriques renforcissent la possibilité de détecter des changements dans la conductivité.

Les études dans le passé qui effectuaient des enregistrements aux stations MT ne considéraient ces changements dans le temps qu'en termes de réponses mesurées. Cependant, la formulation du problème inverse de l'élaboration de modèles de conductivité–profondeur qui varient au minimum d'une année à l'autre permet une mesure quantitative des changements, indispensable dans les modèles qui accomodent ces variations annuelles dans les données. Ce qui fournit une méthode pour évaluer les processus et les profondeurs impliqués dans les changements observés dans les données. Notre étude de modélisation indique un faible accroissement annuel, cependant systématique, dans la conductivité depuis 1987 jusqu'à 1990, localisé dans une zone conductrice sus-jacente à la plaque Juan de Fuca subductante.

[Traduit par la rédaction]

Can. J. Earth Sci. **29**, 601–608 (1992)

Introduction

Earthquake prediction is a challenging and important goal. Predictions are generally based on the monitoring of geophysical phenomena such as tilt, strain, seismic velocity, magnetic field, or geoelectrical conductivity to detect anomalous changes that may occur prior to an earthquake. Changes in these quantities may precede the onset of an earthquake by a matter of hours, days, or years. China, Japan, the United States, and the former U.S.S.R. have extensive programs to monitor possible earthquake precursors. Perhaps the most successful prediction to date involved the 1975 earthquake of magnitude 7.3 near Haicheng, China, where precursors were observed in most of the phenomena noted above (Raleigh *et al.* 1977).

Changes in the conductivity of the Earth can be caused by accumulating tectonic stress. Laboratory results of Brace and Orange (1968) show that the conductivity of saturated rock generally decreases slightly during the early stages of applied stress due to the closure of existing microcracks. However, as the stress continues to increase, new microcracks are formed and the rock dilates. Water diffusing into the new microcracks can cause an increase in conductivity. Measurements of crustal conductivity using artificial electrical current sources have detected precursor changes as large as 30% (Sumitomo and Noritomi 1986). The lead time of the precursor depends on the magnitude of the earthquake and the distance between the epicentre and the recording site (Rikitake 1987).

The magnetotelluric (MT) method, which uses measurements

[1]Geological Survey of Canada Contribution 22391.
[2]Present address: Defence Research Establishment Pacific, Fleet Mail Office, Victoria, B.C., Canada V0S 1B0.

Printed in Canada / Imprimé au Canada

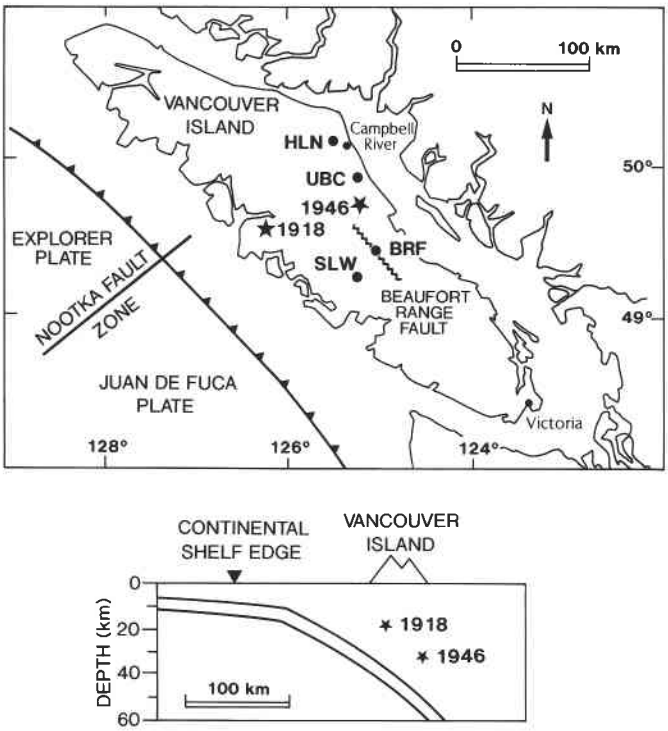

FIG. 1. The tectonic setting of the study area in plan view and cross section. ★, epicentres and focal depths of the 1918 and 1946 earthquakes; ●, MT sites. The dipping structure in the cross section indicates the subducting oceanic crust (after Cassidy *et al.* 1988).

of naturally occurring electromagnetic fields to investigate the conductivity of the Earth, has also been used to monitor earthquake precursors. Canadian studies using MT for earthquake prediction began in 1974 in Charlevoix County, Quebec, near the centre of a zone of seismicity on the north shore of the St. Lawrence River. Kurtz and Niblett (1978) monitored a number of MT sites, and their results show annual changes in the impedance of approximately 14%. They did not, however, find any clear association between changes in impedance and seismic activity.

The central region of Vancouver Island, where two major earthquakes have occurred in this century, is another area in Canada of interest for earthquake prediction. Figure 1 shows the location of these earthquakes: the earthquake on the western side of the island occurred in 1918 and had a magnitude of 7.0 and an estimated average focal depth of 15 km (Cassidy *et al.* 1988); the earthquake on the eastern side occurred in 1946 with a magnitude of 7.3 and an estimated hypocentral depth of 30 km (Rogers and Hasegawa 1978). Relevelling surveys in the Campbell River region adjacent to the hypocentre of the 1946 earthquake showed relative uplift of 4 mm/year from 1977 to 1984, followed by subsidence of approximately the same rate over the next 2 years (Dragert and Lisowski 1990). In response to this change in the direction of vertical deformation, a program was initiated to monitor temporal variation in electromagnetic parameters. Since 1986, MT data have been measured annually at a number of sites on central Vancouver Island. Figure 1 shows the present study region, with MT measurement sites denoted as BRF, UBC, SLW, and HLN.

This paper presents some of the results of 4 years of MT monitoring at these sites. Past studies using the magnetotelluric method have indicated few, if any, unambiguous changes in the measured responses associated with seismic events. However, new robust processing techniques constrain the MT responses to a greater degree than previous methods (Jones *et al.* 1989); therefore, the possibility of detecting significant changes through these measurements is enhanced. In addition, previous studies have only considered temporal changes in the MT responses, usually represented as apparent conductivity–resistivity or impedance. Although examining the responses in this manner may be sufficient to detect a precursor signal, it is difficult to quantitatively interpret changes in the conductivity at depth simply by inspecting temporal changes in the data. The required conductivity changes can be found only by solving an explicit inverse problem.

Geologic and tectonic setting

Vancouver Island is composed of a number of terranes which are part of a series of accreted terranes that make up the western segment of the Canadian Cordillera (Gabrielse and Yorath 1989). The Vancouver Island study region is part of Wrangellia, a large composite terrane consisting of volcanic, plutonic, sedimentary, and metamorphic rock of Paleozoic to Jurassic age. Overlying Wrangellia and underlying part of the study area is the Nanaimo Group, composed of conglomerates, sandstones, mudstones, and shales of Late Cretaceous age. This generally conductive complex occurs beneath site BRF and along the eastern side of Vancouver Island, extending from approximately 100 km north of Victoria to the Campbell River area. The Nanaimo Group extends to depths of 200 m near BRF and to depths of at least 500 m in the coastal region near UBC (England 1990) and causes some attenuation of the telluric signals at sites BRF, UBC, and HLN.

The tectonic setting of central Vancouver Island is complex. The plate boundaries of the northeast Pacific region are shown in Fig. 1. Riddihough (1977) concluded that the subducting Juan de Fuca and Explorer oceanic subplates interact independently with the lithosphere beneath Vancouver Island. These subplates are bounded by the Nootka fault zone which extends from the northern end of the Juan de Fuca Ridge to the continental margin off central Vancouver Island (Hyndman *et al.* 1979). There are a large number of old crustal faults on Vancouver Island, most of which strike northwest–southeast. Of these, the Beaufort Range fault is within the present study area. It is possible that the tectonic forces which cause large earthquakes on Vancouver Island are a result of some form of stress coupling between the subduction zone and these crustal faults.

Hyndman (1988) and Hyndman *et al.* (1990) traced a number of seismic reflective zones beneath Vancouver Island. These are shown in cross section in Fig. 2. The zone designated the F reflector delineates the top of the subducting oceanic plate. The two shallower reflective bands, designated the E and C reflectors, may be either structural features, such as underplated oceanic material (Green *et al.* 1986), or areas of contrasting physical properties, such as zones of trapped fluids. Hyndman (1988) suggested that such fluids, generated by dehydration in the subducting oceanic plate, may be the source of many reflective bands and zones of high conductivity in the lower crust.

Kurtz *et al.* (1990) studied the electrical structure of the central Vancouver Island region as part of the Lithoprobe multidisciplinary geoscience research program. Their results

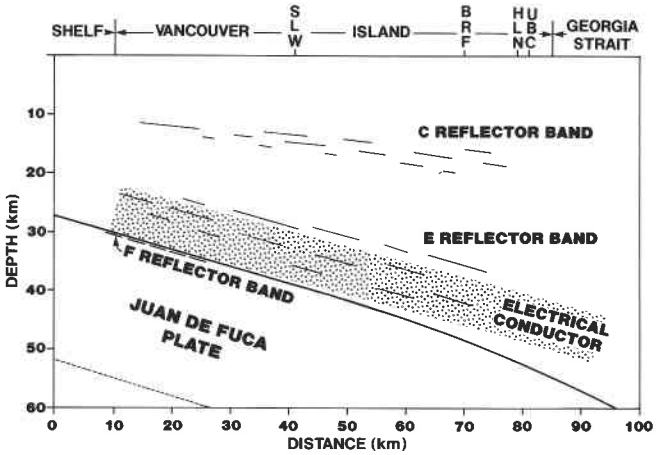

FIG. 2. Cross section of the Cascadia subduction zone, showing the high-conductivity zone of Kurtz *et al.* (1990) and the C, E, and F seismic reflectors of Hyndman (1988). The locations of the MT sites (SLW, BRF, HLN, and UBC) are indicated at the top of the figure.

indicate that a highly conducting zone is present beneath Vancouver Island, dipping inland with its top at depths from approximately 20 to 40 km. These depths correlate well with the seismic E reflector, as shown in Fig. 2. Their MT recording sites were approximately 50 km southeast of the present study area.

MT data collection and processing

Natural electromagnetic fields at the Earth's surface consist of two components: a primary component of external origin; and a secondary or internal component that arises due to telluric currents induced in conductive regions of the Earth by the primary field. In the magnetotelluric method, electric and magnetic fields are measured as a function of time and are Fourier transformed to obtain values in the frequency domain. The ratio of (orthogonal) horizontal components of the electric and magnetic fields, known as the impedance, depends primarily on the subsurface conductivity distribution and is relatively insensitive to the properties of the source. The depth of penetration of the electromagnetic fields is proportional to the period, and hence impedances measured at increasing periods provide information about the conductivity to progressively greater depths. Rather than displaying the complex-valued impedance, MT responses may also be expressed as apparent conductivity σ_a and phase ϕ. The apparent conductivity represents an average of the conductivity distribution over the depth of penetration of the fields, and the phase provides information on variations in conductivity with depth. Plots of σ_a and ϕ as a function of period T provide some clues about the behaviour of the conductivity distribution; however, quantitative information about the conductivity as a function of depth can only be obtained by solving the MT inverse problem.

The locations of the MT recording sites are shown in Fig. 1. Sites BRF, located on the Beaufort Range fault, and UBC, located to the northeast of the fault, were established in 1986. Unfortunately, it was required to change the instrumentation at BRF in 1988. Although the instruments were calibrated, transporting and installing the equipment can affect the calibration, and the effect of this on the measurements is not known. In 1989, sites HLN and SLW were added. Data were recorded at each site for 1–2 months at the same time each year (in the spring) to minimize variations in groundwater level and temperature. Three components of the geomagnetic field were recorded using EDA fluxgate magnetometers (Trigg *et al.* 1971), and potential differences between pairs of electrodes oriented in magnetic north–south and east–west directions were measured using the telluric system of Trigg (1972). The five components were recorded digitally on a cassette recorder.

The data were processed using robust techniques developed by Egbert and Booker (1986). Robust processing methods provide more accurate estimates of the MT responses and significantly reduced error estimates. The advantages of such methods are clearly illustrated by Jones *et al.* (1989). The data were then Fourier transformed using an approach similar to cascade decimation, and Fourier harmonics with less than a certain minimum power in the horizontal magnetic field were rejected. Responses were computed for nonoverlapping frequency bands of width equal to 25% of the centre frequency. The data extend over a bandwidth of 100–5000 s period.

Temporal change in MT responses

The MT data recorded at the sites established in 1986 (BRF and UBC) are generally of good quality. Typical processed data from BRF are shown in Fig. 3. These results have been rotated into a principal axes framework in which the x and y directions maximize and minimize the apparent resistivity (averaged over period), respectively. These axes approximately coincide with true north–south (x direction) and east–west (y direction). Data from the other sites were also rotated into these coordinates. The upper panels display apparent conductivities, σ_{xy} and σ_{yx}, for 1988 and 1989; the corresponding phases, ϕ_{xy} and ϕ_{yx}, are shown in the lower panels. The curves are well constrained, with two standard deviation error bars varying from less than 1% to approximately 3%.

As an example of the annual variation in the data, Fig. 4 shows the percentage change in σ_{xy} and ϕ_{xy} at sites BRF and UBC between 1989 and 1990. The corresponding results for σ_{yx} and ϕ_{yx} are shown in Fig. 5. The percentage change for each period is calculated by subtracting the datum for 1989 from that for 1990, dividing by the 1989 value, and converting to percent. The error bars are computed as twice the standard deviation of a sum of random variables.

The amount of temporal variation in the measurements at the two sites appears to be quite low. At BRF the difference in σ_{xy} between 1989 and 1990, shown in Fig. 4, is a decrease of about 2% averaged over all periods. The corresponding results for UBC are similar. The percentage change in ϕ_{xy} between 1989 and 1990 also indicates that little change has taken place. Figure 5 shows that the error bars associated with σ_{yx} and ϕ_{yx} are somewhat smaller; however, the percentage change is still small. At BRF there is an indication of an increase of a few percent for periods less than about 1000 s and as much as 10% for the longest periods. The average increase over the entire period range is about 3%. The corresponding results for UBC are similar, with an average increase of about 2%. The percentage change in phase between 1989 and 1990 at BRF shows essentially no change, whereas there is an increase of about 2% at UBC.

The variations in MT responses between 1989 and 1990 at BRF and UBC shown in Figs. 4 and 5 are fairly representative of the small yearly variations recorded throughout the study. No clear trends are evident in the data over the duration of the study. Given the estimated error bars and annual variations of

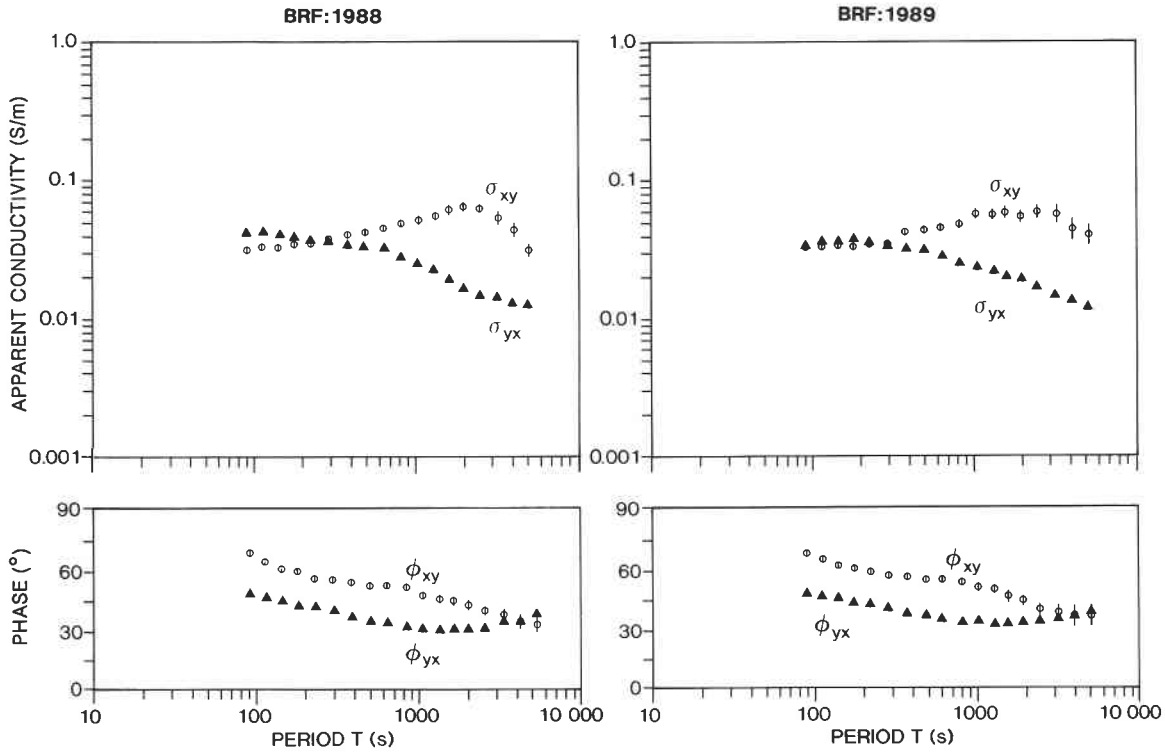

FIG. 3. Apparent conductivities and phases measured in 1988 and 1989 at BRF. The subscript *xy* indicates that the values are derived from the electric field in the *x* direction and magnetic field in the *y* direction, and vice versa for subscript *yx*. Error bars are two standard deviations.

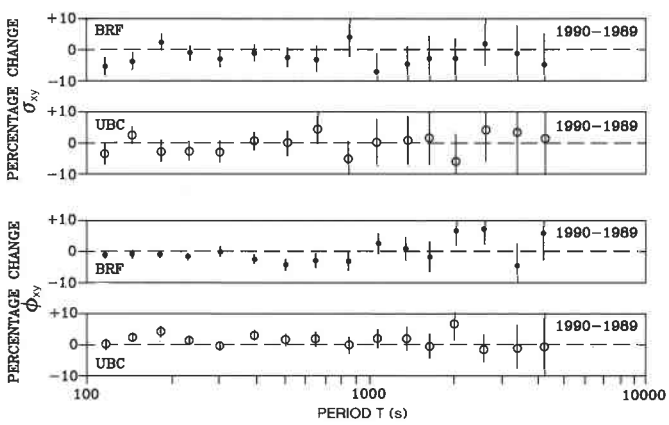

FIG. 4. Percentage change in apparent conductivity σ_{xy} and phase ϕ_{xy} at BRF and UBC between 1989 and 1990.

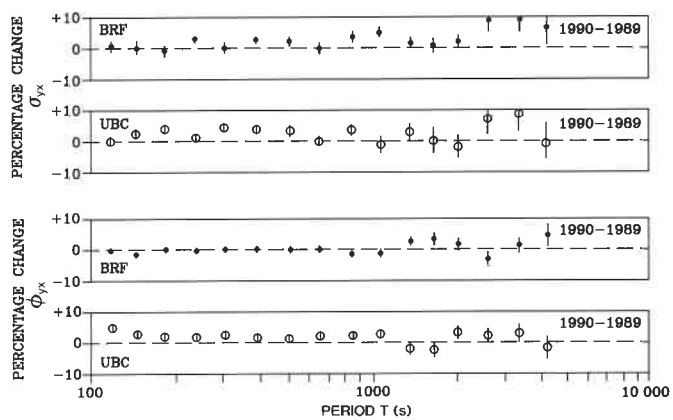

FIG. 5. Percentage change in apparent conductivity σ_{yx} and phase ϕ_{yx} at BRF and UBC between 1989 and 1990.

at least a few percent due to changes in groundwater level and ground temperature which affect electrode coupling (Xu 1986), it is difficult to assign any clear significance to the observed variations in the responses. In order for a change in the apparent conductivity to be interpreted as an earthquake precursor signal, it would have to exceed the observed annual variations of 1–3%.

Temporal change in conductivity models

Figures 4 and 5 illustrate the yearly changes in the measured responses expressed in terms of apparent conductivity and phase. However, to quantitatively interpret changes in the conductivity structure at depth, the inverse problem must be solved. The one-dimensional (1D) MT inverse problem involves determining a conductivity–depth model $\sigma(z)$ that adequately reproduces the measured data. Solving the inverse problem allows investigation of the changes required in conductivity models of the Earth to accommodate the temporal variations in the data. This approach can provide information regarding the depths and processes involved in the observed changes.

A fundamental difficulty with the geophysical inverse problem is that of nonuniqueness: if one model exists which adequately reproduces the measured data, then infinitely many such models exist. However, linearized model construction methods can be applied to find the particular solution that minimizes a given functional of the model (e.g., Oldenburg 1983; Dosso 1990). Models of different character may be constructed by minimizing different functionals. Two particularly

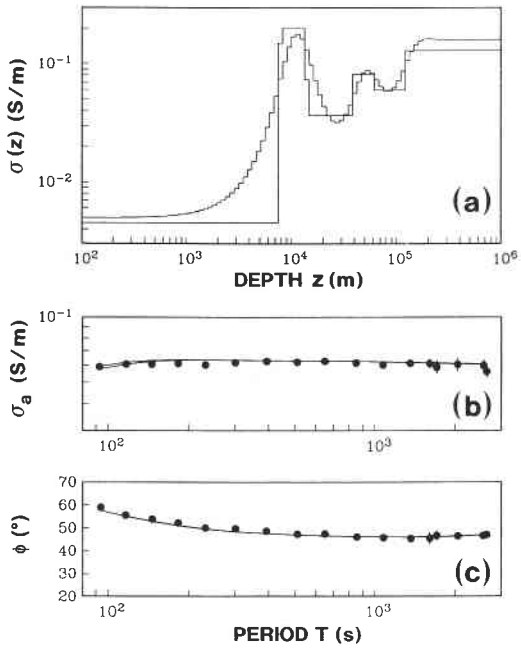

FIG. 6. Minimum-structure conductivity models constructed by inverting the MT data recorded at BRF in 1988. (*a*) The l_2 solution (smooth model) and l_1 solution (blocky model). (*b*, *c*) Comparison of the measured apparent conductivities and phases (circles with error bars) with those predicted for the constructed models (solid lines).

useful solutions are the minimum-structure and smallest-deviatoric models. The minimum-structure solution explicitly minimizes a functional measure of the model structure, subject to an acceptable fit to the data, to produce the simplest model. The smallest-deviatoric solution minimizes the deviation of the model from an arbitrary base or reference model.

One method of using model construction to investigate changes required in the conductivity between any two years uses the minimum-structure solution for data measured in one year as the base model in a smallest-deviatoric inversion of responses measured in a subsequent year. Any differences between the constructed smallest-deviatoric model and the base model are definitely required by the data; therefore, this procedure determines the smallest change from the minimum-structure base model that is consistent with the measured responses.

Figure 6*a* shows two minimum-structure models constructed for the MT responses measured at site BRF in 1988 (this data set was chosen for inversion for the base model, as it appears to be the best in quality). The smooth model minimizes the l_2 norm of the model gradient, and the blocky model minimizes the l_1 norm of the model variation, as described by Dosso and Oldenburg (1989). These norms represent different measures of model structure and can be used to construct models that represent conductivity variations as either gradients or layers. The responses used in the inversion consist of determinant averages of the impedance tensor (e.g., Ranganayaki 1984) and are shown as apparent conductivities and phases in Figs. 6*b* and 6*c*. These responses combine both the *xy* and *yx* orientations in a rotationally invariant form. The l_2 and l_1 models fit the data correctly according to the χ^2 and χ^1 misfit criteria, respectively, which indicates that 1D model solutions are justified. In fact, according to the criterion of Parker (1980), 1D models are justified for each of the 4 years of data recorded at BRF at a misfit considerably less than the expected values. Figures 6*b* and 6*c* compare the measured apparent conductivities and phases (circles with error bars) with those predicted for the constructed model (solid lines). The minor differences in the number and periods of the data as compared with Figs. 3–5 are due to a slightly different scheme of averaging the responses in the frequency domain.

The l_2 and l_1 minimum-structure models shown in Fig. 6*a* are very similar and indicate a high-conductivity zone at about 10–15 km depth. Unfortunately, the depth of this zone could not be accurately determined in this study. Near-surface inhomogeneities in conductivity can introduce a static shift into the measured apparent conductivities which has the effect of displacing the conductivity as a function of depth (e.g., Jiracek 1990). In order to use model construction techniques to reliably determine depths, an independent measurement of the static shift is required (Sternberg *et al.* 1988).

The validity of the constructed models shown in Fig. 6 also depends on the applicability of the 1D assumption. One method of investigating this for the central Vancouver Island region is to apply the 1D inversion methods to responses predicted for the two-dimensional (2D) conductivity model of the region derived by Kurtz *et al.* (1990). Figure 7*a* shows the 1D minimum-structure model (solid line) constructed by inverting determinant-average responses computed for the 2D model of Kurtz *et al.* (1990). The broken line indicates the 1D conductivity structure directly beneath the field site obtained from the Kurtz *et al.* (1990) 2D model. Figure 7*b* shows the 1D minimum-structure model (solid line) constructed by inverting responses computed for the 1D conductivity structure directly beneath the field site of the 2D model (broken line). Since the data computed for the 2D and 1D models have no experimental uncertainty, they were fit as closely as possible; the fit to the data is represented in the panels to the right. Figures 7*a* and 7*b* indicate that the 1D inversion provides a good estimate of the vertical conductivity structure. The 2D effects are not strong, although they do affect the conductivity at great depths. In addition, since the purpose of the modelling study is to detect temporal changes in the conductivity, as long as the inversion methodology is consistent, the computed changes are likely to be meaningful even if the 1D conductivity structure is not precisely correct.

To obtain a representation of the yearly changes required in the conductivity, smallest-deviatoric models may be constructed. As an example of this procedure, the 1988 l_2 minimum-structure model shown in Fig. 6*a* is taken to be the base model. A model may then be constructed which fits the 1987 data but deviates least (in an l_2 sense) from the 1988 base model. This smallest-deviatoric solution is shown in Fig. 8*a* by the solid line; the base model is indicated by the broken line. At most depths the two models are indistinguishable, indicating that the structure of the 1988 model is consistent with the 1987 data. However, the models differ slightly in the high-conductivity zone at about 10–15 km depth; this region is shown in detail in Fig. 8*b* by an enlarged view of the five partition elements at the peak of the conductive zone. The 1987 data require a slightly higher conductivity in this region. The panels to the right of Fig. 8*b* show the measured responses for 1987 (circles with error bars) and those predicted for the smallest-deviatoric model (solid line).

Figure 8*c* shows the smallest-deviatoric model constructed for the 1989 responses (solid line), as well as the 1988 base model (broken line). At most depths the two models are

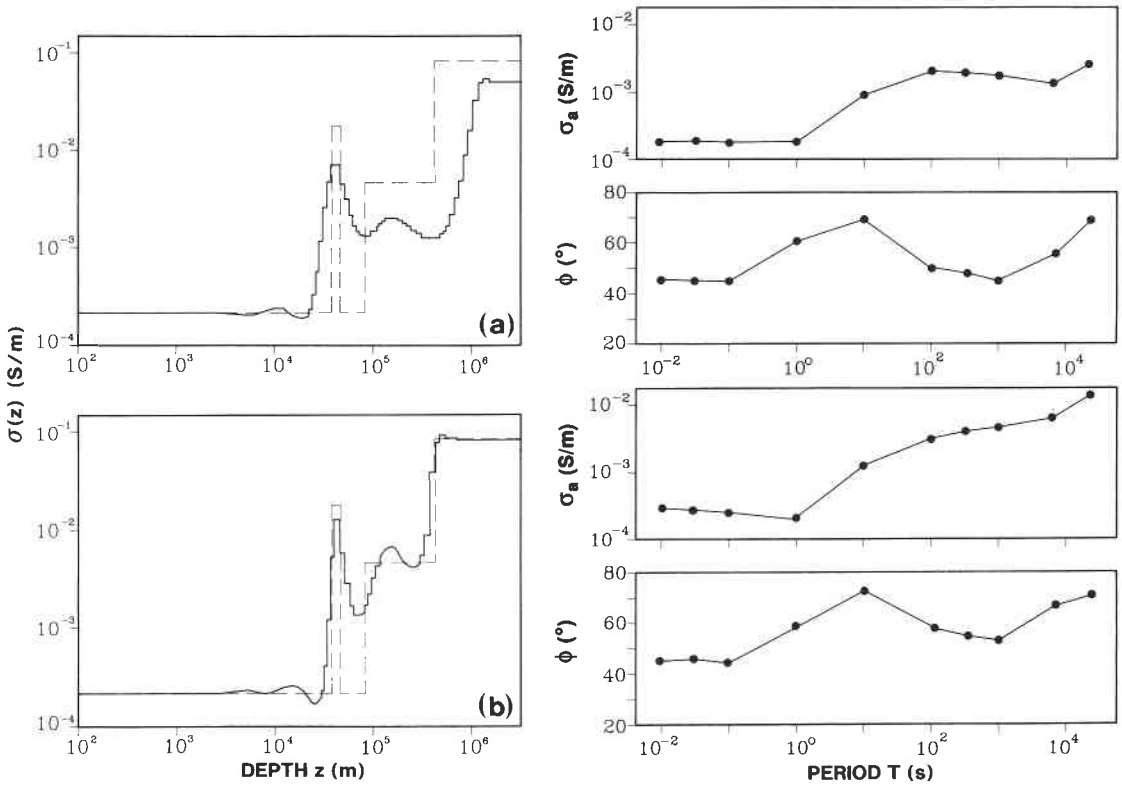

FIG. 7. Investigation of 2D effects. (*a*) The 1D minimum-structure model (solid line) constructed by inverting responses computed for the 2D conductivity model of Kurtz *et al.* (1990). The broken line indicates the 1D conductivity directly beneath the field site obtained from the 2D model. The panels to the right show the apparent conductivities and phases computed for the 2D model (●) and those predicted for the constructed 1D model (solid line). (*b*) The 1D minimum-structure model (solid line) constructed by inverting responses computed from the 1D conductivity structure directly beneath the field site of the 2D model (broken line). The panels to the right show the responses computed for the 1D conductivity structure (●) and those predicted for the constructed model.

indistinguishable, but there are slight differences in the high-conductivity zone. This zone is shown in detail in Fig. 8*d*. Even though the difference is small, it indicates that the 1989 data require a slightly lower conductivity in this region. The 1990 smallest-deviatoric model, shown in Figs. 8*e* and 8*f*, indicates a similar difference. Although not shown, the 1986 smallest-deviatoric model was also constructed; it indicates that no change was required for the 1986 and 1988 responses. However, this result may be due to the relatively large uncertainties associated with the 1986 data set.

The changes in the constructed models in Fig. 8 indicate that the data require a decrease in the conductivity of the conductive zone from 1987 to 1988, from 1988 to 1989, and from 1988 to 1990. The percentage decrease in the conductivity averaged over the three partition elements at the peak of the conductive zone is 14% between 1987 and 1988 (however, this may be influenced by the instrumentation change), 4% between 1988 and 1989, and 7% between 1988 and 1990.

Summary

We have established baselines for MT measurements of apparent conductivity and phase based on yearly sampling at four sites on central Vancouver Island. Annual average variations in measured apparent conductivity and phase are about 1–3% and do not indicate any clear trend. Constructing conductivity models that vary minimally from year to year provides a method of quantitatively investigating changes required in the conductivity at depth to accommodate the yearly variations in the data. Our modelling study at site BRF indicates a small but systematic decrease with time in the conductivity of a crustal conductive zone overlying the subducting Juan de Fuca Plate. Within the context of dilation theory, this decrease in conductivity may possibly be related to the closing of microcracks and expulsion of water due to tectonic stress. This may also be compatible with the subsidence observed through levelling measurements in the region.

At the time this analysis was carried out, only 1 year of data was available at sites SLW and HLN, and therefore temporal change could not be investigated at these sites. Five years of data were available at UBC; unfortunately, the quality of this data was not as high as at BRF. This is believed to be due to the greater thickness of the Nanaimo Group of sediments at UBC. Modelling studies have not yet resolved temporal change at UBC. This does not mean that change is not occurring, but rather that due to the larger uncertainties associated with the responses, the data can accommodate the same model solution over a period of 4 years.

Continued observations and improved measurements are required to validate long-term trends. To date, we have not observed conductivity changes that might be a precursor to a large seismic event, and no earthquakes of magnitude greater than 3.0 have occurred in the study region over the duration of the MT experiment. Future plans include employing newly developed instrumentation to improve data quality, obtaining measurements at higher frequencies to resolve the shallow structure, and increasing the regional coverage by installing

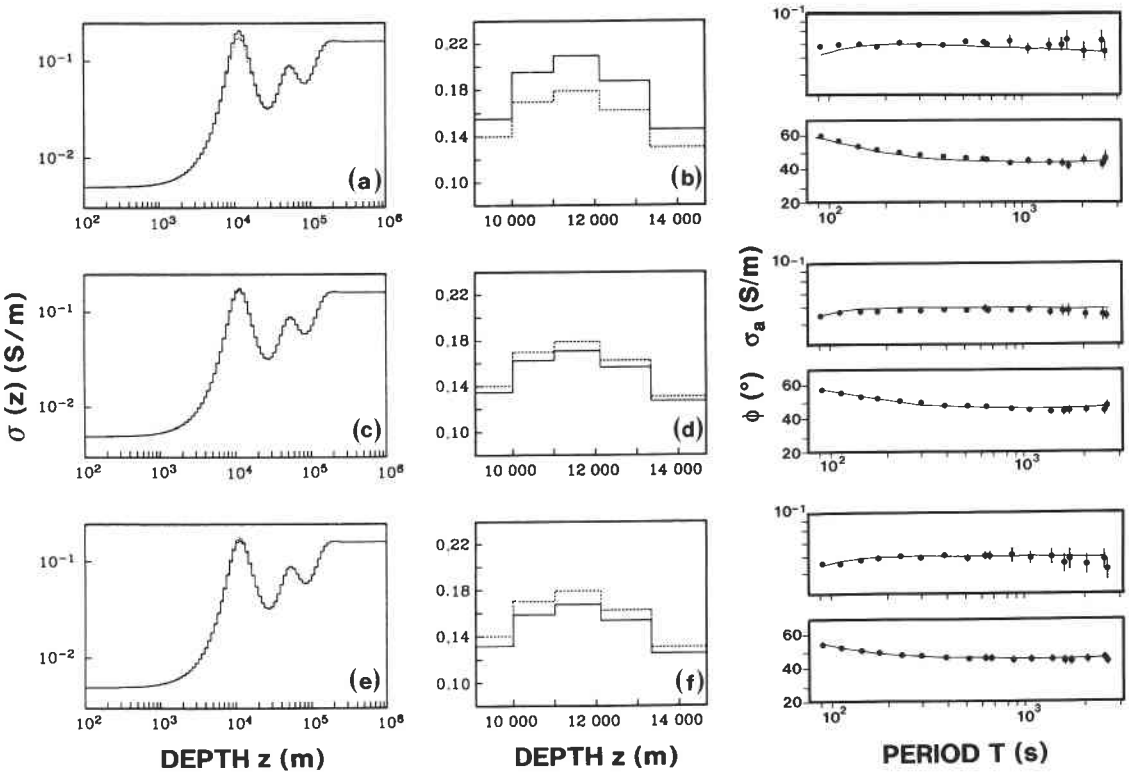

FIG. 8. Smallest-deviatoric models for BRF and the 1988 l_2 minimum-structure solution as the base model. (*a*, *c*, and *e*) The smallest-deviatoric models (solid line) constructed by inverting the 1987, 1989, and 1990 responses, respectively. The 1988 base model is indicated by the broken line. Detail of the high-conductivity region at 10–15 km depth is shown in (*b*, *d*, and *f*). The panels to the right compare the measured apparent conductivities and phases (circles with error bars) with those predicted for the constructed models (solid line).

new sites on the west coast of Vancouver Island near the Nootka fault.

Acknowledgments

We are grateful to M. H. J. Wilde for his assistance with the fieldwork over the 5 years of the experiment and to G. D. Egbert for providing the robust processing software. The research was funded in part by Natural Sciences and Engineering Research Council of Canada (NSERC) grant 5-84270; S. E. D. was supported by an NSERC postgraduate scholarship.

Brace, W. F., and Orange, A. S. 1968. Electrical resistivity changes in saturated rocks during fracture and frictional sliding. Journal of Geophysical Research, **73**: 1433–1445.

Cassidy, J. F., Ellis, R. M., and Rogers, G. C. 1988. The 1918 and 1957 Vancouver Island earthquakes. Bulletin of the Seismological Society of America, **78**: 617–635.

Dosso, S. E. 1990. Inversion and appraisal for the one-dimensional magnetotellurics problem. Ph.D. thesis, University of British Columbia, Vancouver.

Dosso, S. E., and Oldenburg, D. W. 1989. Linear and non-linear appraisal using extremal models of bounded variation. Geophysical Journal International, **99**: 483–495.

Dragert, H., and Lisowski, M. 1990. Crustal deformation measurements on Vancouver Island, British Columbia: 1976 to 1988. Proceedings, International Association of Geodesy Symposium, Edinburgh, Scotland, pp. 241–250.

Egbert, G. D., and Booker, J. R. 1986. Robust estimation of geomagnetic transfer functions. Geophysical Journal of the Royal Astronomical Society, **87**: 173–194.

England, T. D. J. 1990. Late Cretaceous to Paleogene evolution of the Georgia Basin, southwestern British Columbia. Ph.D. thesis, University of British Columbia, Vancouver, B.C.

Gabrielse, H., and Yorath, C. J. 1989. The Cordilleran orogen in Canada. Geoscience Canada, **16**: 67–83.

Green, A. G., Clowes, R. M., Yorath, C. J., *et al.* 1986. Seismic reflection imaging of the subducting Juan de Fuca plate. Nature (London), **319**: 210–213.

Hyndman, R. D. 1988. Dipping seismic reflectors, electrically conductive zones, trapped water in the crust over a subducting plate. Journal of Geophysical Research, **93**: 13 391 – 13 405.

Hyndman, R. D., Riddihough, R. P., and Herzer, R. 1979. The Nootka Fault zone — a new plate boundary off Western Canada. Geophysical Journal of the Royal Astronomical Society, **58**: 667–683.

Hyndman, R. D., Yorath, C. J., Clowes, R. M., and Davis, E. E. 1990. The northern Cascadia subduction zone at Vancouver Island: seismic structure and tectonic history. Canadian Journal of Earth Sciences, **27**: 313–329.

Jiracek, G. R. 1990. Near-surface and topographic distortions in electromagnetic induction. Surveys in Geophysics, **11**: 163–203.

Jones, A. G., Chave, A. D., Egbert, G., *et al.* 1989. A comparison of techniques for magnetotelluric response function estimation. Journal of Geophysical Research, **94**: 14 201 – 14 213.

Kurtz, R. D., and Niblett, E. R. 1978. Time dependence of magnetotelluric fields in a tectonically active region in Eastern Canada. Journal of Geomagnetism and Geolectricity, **30**: 561–577.

Kurtz, R. D., Delaurier, J. M., and Gupta, J. C. 1990. The electrical conductivity distribution beneath Vancouver Island: a region of active plate subduction. Journal of Geophysical Research, **95**: 10 929 – 10 946.

Oldenburg, D. W. 1983. Funnel functions in linear and nonlinear appraisal. Journal of Geophysical Research, **88**: 7387–7398.

Parker, R. L. 1980. The inverse problem of electromagnetic induction: existence and construction of solutions based on incomplete

data. Journal of Geophysical Research, **85**: 4421–4425.

Raleigh, B., Bennett, G., Craig, H., *et al.* 1977. Prediction of the Haicheng earthquake. Transactions of the American Geophysical Union, **58**: 236–272.

Ranganayaki, R. P. 1984. An interpretative analysis of magnetotelluric data. Geophysics, **49**: 1730–1748.

Riddihough, R. P. 1977. A model for recent plate interactions off Canada's west coast. Canadian Journal of Earth Sciences, **14**: 384–396.

Rikitake, T. 1987. Magnetic and electric signals precursory to earthquakes: an analysis of Japanese data. Journal of Geomagnetism and Geolectricity, **39**: 47–61.

Rogers, G. C., and Hasegawa, H. S. 1978. A second look at the British Columbia earthquake of June 23, 1946. Bulletin of the Seismological Society of America, **68**: 653–675.

Sternberg, B. K., Washburne, J. C., and Pellerin, L. 1988. Correction for the static shift in magnetotellurics using transient electromagnetic soundings. Geophysics, **53**: 1459–1468.

Sumitomo, N., and Noritomi, K. 1986. Synchronous precursors in the electrical earth resistivity and the geomagnetic field in relation to an earthquake near the Yamasaki fault, southwest Japan. Journal of Geomagnetism and Geoelectricity, **38**: 971–989.

Trigg, D. 1972. An amplifier and filter system for telluric signals. Publications of the Earth Physics Branch (Canada), **44**: 1–5.

Trigg, D. F., Serson, P. H., and Camfield, P. A. 1971. A solid state electrical recording magnetometer. Publications of the Earth Physics Branch (Canada), **41**: 67–80.

Xu, S. 1986. Quantative estimation of an annual variation of apparent resistivity. Journal of Geomagnetism and Geoelectricity, **38**: 991–999.