Monitoring of Time Dependent Electrical Resistivity by Magnetotellurics

I.K. REDDY,*1 R.J. PHILLIPS,* J.H. WHITCOMB,**
D.M. COLE,** and R.A. TAYLOR**

*Jet Propulsion Laboratory, California Institute of Technology,
Pasadena, California, U.S.A.
**Seismological Laboratory, California Institute of Technology,
Pasadena, California, U.S.A.

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A feasibility study is being made to monitor possible electrical resistivity changes preceding the earthquakes. Numerical calculations, based on dilatancy hypothesis, indicate that the resistivity changes as much as 60 percent can be possible at some of the Southern California sites. Preliminary magnetotelluric measurements in the eight month period indicate that changes in the resistivities greater than 10 percent can be monitored by magnetotelluric measurements. The observed scatter in the magnetotelluric measurements are within the random errors of the estimates, and no major earthquake activity is reported in the vicinity of these sites, during this period.

1. Introduction

Crustal resistivity, as judged from recent advances in laboratory and field observations and in physical models of earthquake premonitory effects, holds promise as a parameter for monitoring pre-earthquake tectonics. Measured resistivities of crustal rocks using deep electrical sounding techniques, such as magnetotellurics, show that such resistivities are 4 to 5 orders of magnitude less than the values obtained by laboratory measurements on completely dry rocks of similar composition. It is very unlikely that the temperature, except in geothermal areas, is the contributing factor for this reduction in resistivity at crustal depths. The most likely explanation seems to be related to the presence of fluid-filled pores and cracks in the crustal rocks. Laboratory measurements on saturated crystalline rocks produced resistivities at least within two orders of magnitude of those obtained by field measurements (Brace and Orange, 1968). This favors the idea of a network of solution-filled cavities in typical crustal rocks (Brace, 1972).

1 National Research Council, Resident Research Associate.
Laboratory measurements of the effect of stress on water-saturated crystalline rocks show that the resistivity will drop by a factor of ten before the fracture of rock occurs (Brace and Orange, 1968). Brace and Orange (1968) have also shown that the effect is much smaller if the load is reapplied to cause slip again along an existing fracture surface. They attributed the change in resistivity with stress in unfractured crystalline rocks as due to the increase in volume by opening of new cracks and subsequent filling with fluids. This phenomenon, which is also known as dilatancy-diffusion process (Brace and Orange, 1968), is believed responsible for the observed seismic velocity anomalies preceding the earthquakes (Nur, 1972; Whitcomb et al., 1973; Scholz et al., 1973).

Actual drops in resistivity have been measured preceding earthquakes Barsukov (1972) in the Garm region of the U.S.S.R. and by Mazzella and Morrison (1974) in central California. Both these groups used artificial source dc soundings and the effective sensing depth for these experiments was few km. High surface resistivity in certain of our field areas makes it difficult to transfer large amounts of dc power into the ground. Thus, the artificial source dc methods are unattractive for continuous measurements of crustal resistivities. Fedotov et al. (1970) detected anomalous changes in the telluric field preceding the earthquakes in Kamchatka. However, measurements of telluric field alone can produce ambiguous results unless knowledge on the source fields is firmly established. Hence, we have opted for the magnetotelluric method. There are two main disadvantages of the magnetotelluric method as applied to detect small percentage changes in the crustal resistivity.

(a) The knowledge on the effect of sources and their long-time variations on the magnetotelluric fields are limited. However, it has been shown (Price, 1962; Srivastava, 1965) for moderate depths of investigation and periods less than a few hundred seconds, source effects can be neglected.

(b) The accuracy to which one can estimate the resistivities depend on several parameters, since magnetotelluric signals are assumed to represent gaussian random process of finite bandwidth, and the resistivities are estimated using statistical procedures, as discussed in the later sections.

A feasibility study is being made in Southern California to see if temporal variations in the electrical resistivity preceding the earthquakes can be measured by the magnetotelluric method. Based on observed seismic velocity anomalies (Whitcomb et al., 1974; Kanamori and Hadley, 1975) we have chosen three sites in and around the Los Angeles basin. We are very limited in usable sites because of high cultural noise. The location of these three sites are shown in Fig. 1. The solid black dot in Fig. 1 indicates the last major earthquake (San Fernando earthquake, $M=6.4$, 9 Feb. 1971) in this area. Site 1, China Flat, is located in Paleocene sediments. Site 2, in the West Antelope Valley of the
Mojave Desert, is underlain by Quaternary alluvium. Site 3, Lytle Creek, in the San Gabriel Mountains, is on the Pre-cretaceous metamorphic rocks.

2. Expected Changes in Magnetotelluric Apparent Resistivities Based on Dilatancy-Diffusion Model

To give some idea on the magnitude of observable changes in the apparent resistivity due to the dilatancy-diffusion process, one- and two-dimensional model studies have been performed for the field sites and the results are shown in Figs. 2 and 3. The zone of dilatancy is assumed to exist in the strongest and most brittle part of the crust, that is, where earthquakes occur. This is between 0 and 20 km deep in Southern California. The thickness of this dilatant zone is unknown but must be at least 5–10 km to explain the seismic velocity data prior to the San Fernando earthquake. We have assumed the resistivity will drop by a factor of 5 due to the dilatancy-diffusion process. This is half of what Brace and Orange (1968) measured for some saturated crystalline rocks in the laboratory. The rest of the parameters, viz., thickness of individual layers and their resistivities for these hypothetical models are assigned on the basis of local geology and magnetotelluric sounding data.

The one-dimensional model results in Fig. 2 show that maximum change
Fig. 2. Expected changes in magnetotelluric apparent resistivity (positive change indicates decrease in resistivity) at the three magnetotelluric sites, based on one-dimensional hypothetical models.

Fig. 3. Expected changes in the magnetotelluric apparent resistivities (positive change indicates decrease in resistivity) due to dilatancy-diffusion process at China Flat site (2), based on a hypothetical two-dimensional model.
 (>60%) will occur for the periods in between 1 and 10 sec for the China Flat and Lytle Creek sites. Unfortunately, geomagnetic pulsations in this band have very low power and are less common in mid-latitudes. The small percentage (<20%) in the apparent resistivity at West Antelope Valley site is expected because of thick conductive layer at the surface corresponding to the sedimentary section.

To study the effect of contact between the zone of dilatancy and the non-dilatant zone on the magnetotelluric apparent resistivities, we have constructed a two-dimensional hypothetical model for China Flat site. Model results are computed using the Galerkin Finite Element technique (Reddy and Rankin, 1975). The areal extent of zone of dilatancy is rather unknown. Anderson and Whitcomb (1975) developed a relation indicating an anomalous zone of 135 km in diameter for the magnitude 6.4, San Fernando earthquake. Here, we assume the longitudinal (along-strike) extent of the anomalous zone is at least a few skin depths so that a two-dimensional assumption is valid. The lateral (cross-strike) extent of the dilatant zone is assumed to be 160 km. We note that it is the "edge effect" but not the lateral extent that effects the behavior of the apparent resistivities outside the dilatant zone. The results depicted in Fig. 3 for the two-dimensional model show one can theoretically detect changes in the resistivity due to dilatancy-diffusion at the stations outside the dilatant zone by the magnetotelluric method. For the stations outside the dilatant zone, depending on the distance of the magnetotelluric station from the contact, the apparent resistivity perpendicular to the contact shows an increase whereas the apparent resistivity parallel to the contact decreases. If the zone of contact is gradual rather than sharp, the difference between parallel and perpendicular resistivities disappear and they approach the one-dimensional model calculations at any given site (Rankin et al., 1976).

3. Instrumentation

The magnetotelluric data acquisition system being used is similar to the one described by Allsopp et al. (1974). It has a frequency range $10^{-3}$ Hz to 20 Hz. The magnetic field variations are monitored using induction-type sensors with sensitivity to 70 $\mu$V/Hz at 1 Hz. Telluric field variations are detected by measuring the voltage differences between copper electrodes 70 to 200 m apart. The total gain of the magnetic system is 125 db and that of the telluric system is 85 db. The data are recorded in digital form on 7-track tape with two sampling intervals, viz., 0.8 sec for low mode and 0.025 sec for high mode.

The recording is made at each site for 12–24 hours depending on the geomagnetic activity. Each site is being occupied at monthly intervals.
4. Data Analyses

Several data sets, each 27 min long from low mode of recording and 51 sec long from high mode of recording, free from visual noise are chosen from each recording period. The signals are conditioned (removed mean and trend, tapered both ends) and Fourier coefficients are computed. The Fourier coefficients are smoothed using a Hamming window. Power and cross power spectral estimates are obtained by simple multiplication of complex Fourier coefficients. Spectral estimates thus obtained are smoothed using a constant Q filter which gives ten estimates per decade. Smoothed spectral estimates for several data sets for a given recording period are averaged and these averaged spectral estimates are used to compute the magnetotelluric tensor impedances, polarization parameters, etc.

A linear relationship between the horizontal magnetic and telluric field components, at each frequency band, is written as

\[
\begin{align*}
E_x &= Z_1 H_x + Z_2 H_y \\
E_y &= Z_3 H_x + Z_4 H_y
\end{align*}
\]

(1)

where \(Z_1, Z_2, Z_3\) and \(Z_4\) are known as tensor impedances representing the electrical properties of the earth in the vicinity of the measuring site. \(E_x, E_y\) and \(H_x, H_y\) are the telluric and magnetic field components, and \(x\) and \(y\) any two orthogonal measuring axes.

Assuming that magnetotelluric process represent a stationary random process, the tensor impedances are expressed in terms of the power and cross power spectral density estimates at each frequency band, as (BENDAT and PIERSOL, 1971; REDDY and RANKIN, 1974).

\[
\begin{align*}
Z_1 &= S_{13} \left[ 1 - \frac{S_{23} S_{33}}{S_{22} S_{33}} \right] / S_{11} (1 - \gamma_{12}^2) \\
Z_2 &= S_{23} \left[ 1 - \frac{S_{21} S_{13}}{S_{11} S_{33}} \right] / S_{22} (1 - \gamma_{12}^2) \\
Z_3 &= S_{14} \left[ 1 - \frac{S_{24} S_{43}}{S_{22} S_{44}} \right] / S_{11} (1 - \gamma_{12}^2) \\
Z_4 &= S_{24} \left[ 1 - \frac{S_{21} S_{14}}{S_{11} S_{44}} \right] / S_{22} (1 - \gamma_{12}^2)
\end{align*}
\]

(2)

For writing convenience, we have abbreviated magnetic and telluric components \(H_x, H_y, E_x\) and \(E_y\) to 1, 2, 3 and 4 respectively, where \(x\) and \(y\) represent measuring directions. \(S_{ij}\) (\(i, j = 1, 2, 3, 4\)) are the elements of spectral matrix and \(\gamma_{12}\) is the sample coherency between the magnetic field components \(H_x\) and \(H_y\). The required accuracy of estimates to determine small percentage changes in appar-
Two kinds of errors exist in estimating the tensor impedances by the above procedure: (1) bias errors and (2) random errors.

Bias errors in impedances result from a number of sources: (a) bias in the power and cross power spectral density estimates. This error can be quite significant at frequencies where spectral peaks occur. These errors can be suppressed by obtaining properly resolved estimates of power and cross spectra, thus making bandwidth sufficiently narrow to accurately define peaks in the spectra. On the other hand, a narrow bandwidth decreases the number of degrees of freedom and thus contributes to the random error. A compromise is required. (b) Measurement noise in the magnetic field components is another source of bias error. Uncorrelated noise in the electric field components does not cause the bias error in the impedances, because of the way the linear relationships are written in (1). For example, assuming other bias errors are negligible and a zero coherence between the magnetic field components ($\gamma_{32} = 0$, $S_{32} = 0$), if the spectral density of the noise in one of the magnetic field components is 10 percent of the signal, the resulting impedance estimate would be biased downward by about 10 percent. Other bias errors are caused by nonlinearity and time dependency of the impedances, and these can be assumed small compared to the other bias and random errors.

Following GOODMAN (1965) and BENDAT and PIERSOL (1971) the expression for random error in the impedance estimate can be written as

\[
 r^2 = \frac{4}{n-4} F_{4, n_2-4} \frac{[1 - \gamma_{32}^2]}{S_{33}} \frac{S_{33}}{S_{jj}}
\]

for $Z_1$ and $Z_2$, and

\[
 r^2 = \frac{4}{n-4} F_{4, n_2-4} \frac{[1 - \gamma_{41}^2]}{S_{44}} \frac{S_{44}}{S_{jj}}
\]

for $Z_3$ and $Z_4$

where

- $n = 2BT =$ number of degrees of freedom.
- $B$ is the bandwidth and $T$ is the length of the record.
- $F_{4, n_2-4} = 100\alpha$ percentage point of an $F$ distribution with 4 and $n_2 = n - 4$ degrees of freedom.
- $S_{jj} =$ power spectrum estimate of the magnetic field component ($j = 1, 2$).
- $\gamma_{32}^2, \gamma_{41}^2 =$ sample estimates of the multiple coherency (REDDY and RANKIN, 1974).
- $\gamma_{12} =$ ordinary coherence function between $H_x$ and $H_y$. 
The (1—α) confidence interval for the magnitude of the impedances, Z, and phase factors Φ are given for each frequency band.

\[ |\hat{Z}_i| - t_i \leq Z_i \leq |\hat{Z}_i| + t_i \]
\[ \hat{\Phi}_i - \Delta \hat{\Phi}_i \leq \Phi_i \leq \hat{\Phi}_i + \Delta \hat{\Phi}_i \]

(4)

where \( \Delta \hat{\Phi}_i = \sin^{-1}\left[ t_i/\hat{Z}_i \right] \) and "-" indicates the estimated values.

From (3) the accuracy in estimating impedances depends on four parameters:

(a) The number of degrees of freedom, \( n \), as \( n \) increases \( r \) decreases. For given bandwidth this means increasing \( T \), the length of the record.

(b) The multiple coherence estimates \( \gamma_{313}^2 \) and \( \gamma_{413}^2 \). As the multiple coherence estimates approach unity, the random error approaches zero. For the special case, \( \gamma_{313}^2 \) and \( \gamma_{413}^2 = 1 \), \( r \) is independent of the number of degrees of freedom. However, for fewer degrees of freedom coherence functions are dependent on the degrees of freedom. When multiple coherence functions are zero, then no linear relationship exists between the electric and magnetic field at the surface of the earth.

(c) The coherency between \( H_z \) and \( H_y(12) \). As \( \gamma_{12}^2 \to 0 \) the accuracy of the impedance estimates improves.

(d) The spectral ratios, \( S_{33}/S_{11} \), \( S_{33}/S_{22} \), \( S_{44}/S_{11} \), and \( S_{44}/S_{22} \). As these decrease, the accuracy of the impedance estimate improves. However, these spectral ratios are dependent on the electrical properties of the earth.

5. Results

Directions of principal axes are obtained from the tensor impedances in the measured directions. Spectral and cross spectral estimates are rotated into principal directions and coherence functions and tensor impedance are computed in principal directions. As a general rule, data points having multiple coherencies \( (\gamma_{313}^2 \text{ and } \gamma_{413}^2) \) greater than or equal to 0.9 are documented and less acceptable data points are used with discretion where better data points are sparse. Multiple coherencies for periods less than 10 sec and greater than 300 sec are usually low (<0.8). Pulsation activity in the range 10 and 300 sec depend on the time and day of the recording, hence several gaps exist even in this period range. However, pulsation with periods 20–30 sec are present almost every day and the results for this period band are presented in Figs. 4a–4f and Tables 1 to 3. Results for other period bands also exhibited similar features. There are several missing data points 20–30 sec band which are mostly due to instrumental failure and human errors.

Figures 4a to 4f contain tensor apparent resistivities and phases (\( \rho_{12} \) and \( \phi_{12} \))
along the major resistivity axis, $\rho_{21}$ and $\phi_{31}$ along the minor resistivity axis) along the principal directions. Tensor apparent resistivities from all the three sites exhibit anisotropies of varying degree. The China Flat (1) and Lytle Creek (3) sites exhibit larger anisotropy compared to West Antelope Valley (2) site. The dimensionality coefficients for all the three sites are small indicating that the geoelectric structure at all the three sites can be represented by a two-dimensional geometry. Relatively low and uncorrelated vertical magnetic field variations are measured at these sites which suggest that the anisotropy exhibited by tensor apparent resistivities may be due to the anisotropy in the rock formations and dipping strata rather than lateral inhomogeneity.

<table>
<thead>
<tr>
<th>Table 1. China Flat (1), Period: 23–30 sec.</th>
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<tbody>
<tr>
<td>Date</td>
</tr>
<tr>
<td>14 Jan.</td>
</tr>
<tr>
<td>30 May</td>
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<tr>
<td>10 July</td>
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<table>
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<tr>
<th>Table 2. West Antelope Valley (2), Period: 23–30 sec.</th>
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<tbody>
<tr>
<td>Date</td>
</tr>
<tr>
<td>20 March</td>
</tr>
<tr>
<td>29 May</td>
</tr>
<tr>
<td>17 July</td>
</tr>
<tr>
<td>6 Aug.</td>
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<table>
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<tr>
<th>Table 3. Lytle Creek (3), Period: 23–30 sec.</th>
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</thead>
<tbody>
<tr>
<td>Date</td>
</tr>
<tr>
<td>13 Feb.</td>
</tr>
<tr>
<td>28 May</td>
</tr>
<tr>
<td>8 July</td>
</tr>
<tr>
<td>5 Aug.</td>
</tr>
</tbody>
</table>

List of symbols used in tables:

- $n$: Number of data sets,
- $\tau_{12}^1$: Sample coherency between magnetic field components in principal directions,
- $\tau_{312}^1$, $\tau_{412}^1$: Multiple coherencies in principal directions,
- $\phi_H$, $\phi_E$: Azimuths of magnetic and electric field ellipses measured clockwise from geographical north (+ve),
- $\varepsilon_H$, $\varepsilon_E$: Ellipticities of magnetic and electric fields respectively,
- $\theta$: Direction of maximum resistivity principal axis, measured clockwise from geographical north (+ve),
- $\alpha$, $\beta$: Dimensionality coefficients, $\alpha=(Z_1+Z_4)/(Z_2-Z_3)$, skew; $\beta=(Z_1'-Z_4')/(Z_2'+Z_3')$, ellipticity index in the principal directions.
1. CHINA FLAT
PERIOD: 23-30 sec

2. W. ANTELOPE VALLEY
PERIOD: 23-30 sec

3. LYTLE CREEK
PERIOD: 23-30 sec
The thickness of the sedimentary strata at the China Flat (1) site, which is located in the Ventura basin of the transverse range province, is greater than 18 km (BAILEY and JAHNS, 1954). The central part of the Ventura basin has been subjected to direct north-south compression, resulting in overturning of beds and the development of thrust faults (north of site 1 in Fig. 1). The major axis of tensor apparent resistivity which is roughly north-south, is perpendicular to the strike of the basin. The anisotropy in the tensor apparent resistivity may be caused by the currents flowing parallel and perpendicular to the strike of the dipping strata. Although several thrust- and strike-slip faults are present in the vicinity of this site, the resistivity contrasts across the fault zones may be small because of the same lithological units on either side of the faults.

The anisotropy observed at the West Antelope Valley (2) site is surprisingly small, given the fact that this site, which is in wedge-shaped portion of the western Mojave Desert is in between two major faults in this area, that is, the San Andreas and Garlock. Low tensor apparent resistivities measured at this site may be reflecting the resistivities of sedimentary formations. Some of the formations may be highly saline which is not uncommon for the desert areas of this part of the country. The major axis of the tensor apparent resistivity which is approximately east-west is parallel to the San Andreas fault in this area. Contacts between the highly conductive sedimentary rocks in the desert region and highly resistive crystalline rocks in both the Garlock and the San Andreas fault zone may be contributing to the observed small anisotropy at this site.

The Lytle Creek site which is located in the eastern edge of the San Gabriel Mountains and just south of the San Andreas and San Jacinto faults is underlain by Pre-cretaceous rocks. Moderately high tensor apparent resistivities (Fig. 4e) observed at this site may be attributed to the water-saturated rocks. The major axis of the tensor apparent resistivity is perpendicular to the strike of the San Andreas and San Jacinto faults. Lateral inhomogeneity resulting from the contact of conductive sedimentary rocks in the Mojave Desert, north of the San Andreas fault, and the resistive rocks in the San Gabriel

Fig. 4. a; Tensor apparent resistivities in principal directions vs. time in months for China Flat site (1). Error bars represent 95 percent confidence limits. b; Phases of the tensor impedances in principal directions vs. time in months for China Flat site (1). Error bars represent 95 percent confidence limits. c; Tensor apparent resistivities in principal directions vs. time in months for West Antelope Valley site (2). Error bars represent 95 percent confidence limits. d; Phases of the tensor impedances in principal directions vs. time in months for West Antelope Valley site (2). Error bars represent 95 percent confidence limits. e; Tensor apparent resistivities in principal directions vs. time in months for Lytle Creek site (3). Error bars represent 95 percent confidence limits. f; Phases of the tensor impedances in principal directions vs. time in months for Lytle Creek site (3). Error bars represent 95 percent confidence limits.
Mountains, may be contributing to the observed anisotropy at this site. This will also explain the orientation of telluric field ellipse, which is perpendicular to the San Andreas fault.

6. Changes in Estimated Resistivity with Time

General expressions for multiple coherence functions, in terms of the elements of the spectral matrix are written as (Reddy and Rankin, 1974)

\[ \gamma_{213}^2 = \frac{S_{22}|S_{13}|^3 + S_{11}|S_{23}|^3 - 2 \text{Re} (S_{12}S_{23}S_{31})}{S_{22}(S_{11}S_{22} - |S_{12}|^2)} \]

\[ \gamma_{415}^2 = \frac{S_{22}|S_{14}|^3 + S_{11}|S_{24}|^3 - 2 \text{Re} (S_{12}S_{24}S_{41})}{S_{22}(S_{11}S_{22} - |S_{12}|^2)} \]

where \( S_{ij} (i, j = 1, 2, 3, 4) \) are the elements of magnetotelluric spectral matrix; subscripts 1, 2, 3, 4 represent the magnetotelluric field components \( H'_x, H'_y, H'_z, E'_y \) respectively. Prime indicates the components are along principal directions.

Assuming that the magnetic field components \( H'_x \) and \( H'_y \) in the principal direction are incoherent \( (S_{12} = 0) \) and by the definition of principal axes \( (S_{13} = 0; S_{44} = 0) \) in the magnetotelluric case, the above expressions for multiple coherence functions in principal directions reduce to

\[ \gamma_{213}^2 = \frac{|S_{23}|^2}{S_{22}S_{33}} \]

\[ \gamma_{415}^2 = \frac{|S_{14}|^2}{S_{44}S_{11}} \]

These expressions are identical to ordinary coherence functions in the principal directions. Assuming the noise and signal are uncorrelated, multiple coherences less than unity indicate noise in either the electric or magnetic field components or both. However, the way the impedances are expressed in (1) suggest that the impedances are not influenced by the noise in the electric field. Noise in the magnetic field components biases the appropriate impedances downward. For example, noise in \( H'_y \) biases impedances \( Z'_1 \) and \( Z'_2 \) downward. The amount of downward bias in impedances can be estimated, for incoherent magnetic field components, using multiple coherence estimates. For example, a multiple coherence of 0.9 (assuming the electric field components are noise-free) can be explained by an approximately 10 percent added noise power in the appropriate magnetic field component. The corresponding impedances will be biased downward by the same percentage, whereas the corresponding tensor apparent resistivities will be biased downward by approximately 17 percent.

Phases of the impedances are unaffected by the uncorrelated noise in the electric and magnetic field components.
Results presented in Figs. 4a to 4f and in Tables 1 to 3 are consistent with the above discussion. For example, the tensor apparent resistivity estimates ($\rho_{21}$) in minor principal axis at the Lytle Creek site for the months of May and July show a drop of 20% which can be easily explained by low multiple coherence estimates ($\gamma_{12}^2=0.88, 0.89$). Similarly a drop of approximately 15% in $\rho_{12}$ for July at the same site can be explained by the low multiple coherency of 0.93 (Table 3). Similar observations can be made from the data at the other two sites. As expected, the phase values are fairly consistent. The overlapping of 95 percent confidence bands in almost all cases indicate that the bias errors are within the limits of random errors.

No significant changes, after accounting for bias and random errors, are noticeable either in the tensor apparent resistivities or phases during the first eight months of 1975 at these sites. No earthquakes of magnitude greater than 4 occurred in the vicinity (<50 km) of these sites during this period. There were some earthquakes with magnitudes less than or equal to 3, but it is doubtful that earthquakes of this low magnitude produce premonitory effects in the magnetotelluric resistivities. Even if they did they may be within the errors of measurement and analyses and thus not noticeable.

Insignificant changes in tensor apparent resistivities, phases, direction of principal axes during this eight-month period suggest that the linear relationship (1) is valid and the effect of source parameters are relatively unimportant. Further, the major axis of the telluric field in most cases is independent of the incident magnetic field and is oriented along the major axis of resistivity.

Correlation between low multiple coherences and downward bias in the tensor apparent resistivities indicate the magnetic field channels are contaminated by uncorrelated noise, which is consistent with low signal-to-noise ratio observed for those data sets. The low signal-to-noise ratio in the magnetic channels is due to the magnetic amplifier because of low signal strength during the recording period. The pulsation activity has been generally low because of latitude of the sites and the year of minimum solar activity. Efforts are being made to improve the signal-to-noise ratio. Once this has been achieved one should be able to estimate the tensor apparent resistivities within ±10 percent accuracy.

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