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PHYSICAL PROPERTIES OF THE VARIATIONS OF THE ELECTRIC FIELD OF THE EARTH PRECEDING EARTHQUAKES, I

P. VAROTSOS and K. ALEXOPOULOS

Department of Physics, University of Athens, Solonos Str. 104, Athens 144 (Greece) *

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ABSTRACT

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The electric field variations of the earth that occur before earthquakes have been studied in a network of eighteen stations in Greece. These precursor seismic electric signals (SES) occur 6–115 h before the earthquake (EQ) and have a duration of 1 min to $1\frac{1}{2}$ h. The duration and the lead-time in contrast to other precursors, do not depend on EQ-magnitude (M). These signals appear as a transient change of the potential difference measured between two electrodes (up to a few millivolts for electrodes at a distance of about $L = 50$ m) depending on M, the epicentral distance r and the local inhomogeneities. The components of electric field are measured in two perpendicular directions (E–W and N–S). The totality of experiments showed that the interesting quantity of each SES is the maximum value ΔV of the potential change. The SES of an impending EQ appears simultaneously at a number of stations without being accompanied by any significant change in the magnetic field.

The following rules have been established:

(1) Seismic electric signals recorded on a single line (e.g. E–W) of a given station and emitted from various seismic regions have ΔV-values that decrease with the epicentral distance according to a $1/r$-law (for $r \geqslant 50$ km).

(2) For a given line of a given station the SES emitted from a given seismic region ($r \simeq$ const.) have ΔV-values that increase with the magnitude; to a good approximation log ΔV versus M gives a straight line with a slope between 0.3 and 0.4. If for the same station and line another seismic region is considered, the straight line is parallel to the previous one but shifted by a constant amount that depends purely on the ratio of the epicentral distances. Therefore, if the quantity $\log(\Delta V \cdot r)$ for earthquakes emitted from various seismic regions is plotted versus M, a unique linear relation for each station appears with the same slope.

(3) The simultaneous ΔV-values of a given EQ recorded at various stations do not follow a $1/r$ dependence. The value $\Delta V/L$ of the electric field in each direction, divided by a suitable factor—an empirically determined effective resistivity—gives a quantity characteristic of the variation of the component of the current density in the earth which can be designated as the intensity of the signal in this direction. By combining the values of the two directions the total intensity J of the SES results. This quantity is found to attenuate with the distances of the stations according to a $1/r$-law so that $\log(J \cdot r)$ is an unique linear function of M for all stations and seismic regions.

* Mailing address: Knossou Str. 36, Ano Glyfada, 16561 Athens, Greece.

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INTRODUCTION

Since March 1981 a record of the telluric field has been made at various sites in Greece. Beyond the usual disturbances due to atmospheric or magnetic storms transient changes of the electric field have been observed which appear many hours before an earthquake. The latter—henceforth called *seismic electric signals* (SES)—have been observed in very many cases (Varotsos et al., 1981a, b, 1982a,b) of earthquakes (EQ) with a magnitude * (M) between 3.2 and 6.8 and epicenters almost all over Greece. Since October 1982, eighteen stations have been in operation (Fig. 1); they are telemetrically connected to Athens (through telephone lines) and the results are depicted on multipen recorders in the central station placed at Glyfada (GLY) about 15 km from Athens. This telemetric system enables the on-time observation of the possibly simultaneous variations of the electric field at various sites.

The electric field is determined by measuring the potential difference between two electrodes; they constitute what shall be called a line. Two such lines (E–W and

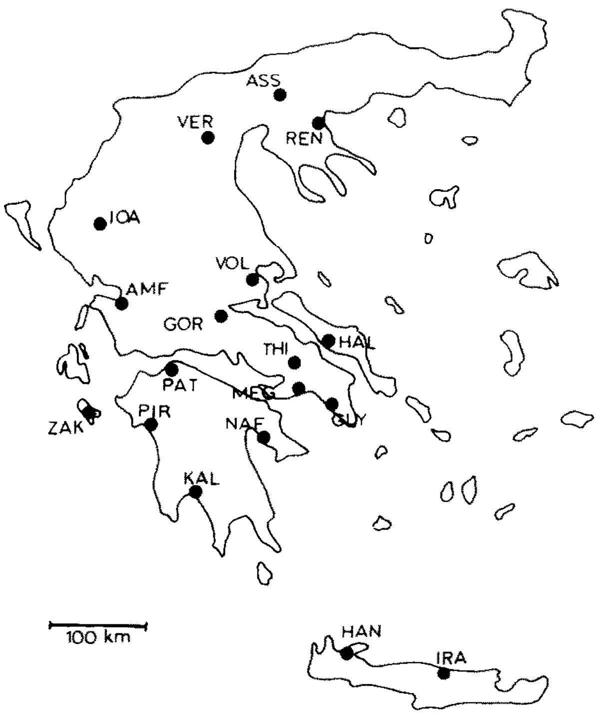

Fig. 1. Map of Greece showing the sites of the stations. MEG and IRA are not telemetrically corrected.

* Magnitudes throughout this paper equal to M_S taken from the officially certified edition of the preliminary seismological bulletin of the National Observatory of Athens; if M_S is not given, we estimate it from $M_S = M_L + 0.4$.

N–S) allow the determination of the total electric field at the site of the station. The SES are collected in the experimental technique commonly applied to geophysical measurements: a pair of brass (or lead or non-polarized) electrodes are buried in the earth at a depth of 2 m and at a distance L between 30 and 200 m. The potential difference V is measured after amplifying and filtering out frequencies higher than 0.3 Hz and the result is displayed on a strip chart recorder (with a speed 1–20 cm/h) for each line separately. The cultural noise at the site should not exceed 0.1–0.2 mV for a line of 100 m; the station therefore has to be installed as far as possible from electric power sources and should be operated with batteries.

It is the aim of the present paper to describe the properties of the SES and to compare them with the parameters of the impending earthquakes.

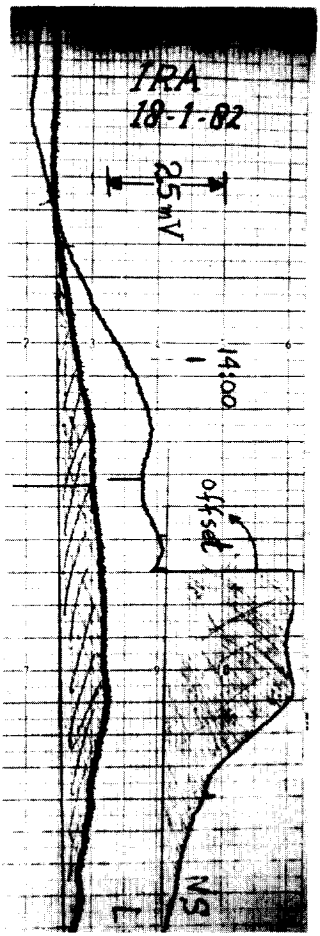

Fig. 2. An SES recorded on both lines ($L = 100$ m) of IRA almost $6\frac{1}{2}$ h earlier than the $M = 6.8$ event which occurred at 19:27, Jan. 18, 1982, at a distance 500 km north of the station. Chart speed: 3 min per line. The SES starts at 13:06.

Many hours before an EQ the potential difference V of one line (or simultaneously of both lines) shows a variation which lasts for a time τ and then recovers its initial value. The value ΔV of this variation for a line of $L = 50$ m ranges from a few tenths of a millivolt up to 10 mV depending on the station, on the magnitude of the impending EQ and on the epicentral distance. The minimum value of τ observed until now is 1 min and the maximum is about $1\frac{1}{2}$ h.

The strongest EQ so far for which SES have been observed had a magnitude $M = 6.8$; in this case the SES were recorded at distances up to 500 km. Figure 2 shows the SES recorded at 13:06 GMT of Jan. 18, 1982 at IRA, a station on the island of Crete. Almost $6\frac{1}{2}$ h later (i.e. at 19:27 GMT) an event of $M = 6.8$ occurred

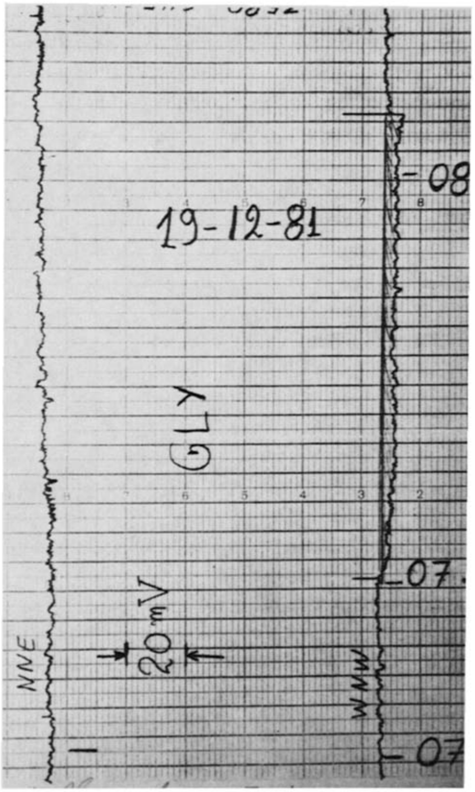

Fig. 3. The SES which preceded the $M = 6.8$-event that occurred at 14:11, Dec. 19, 1981, in the North Aegean Sea. The signal was recorded at GLY at a distance 160 km from the epicenter. The ΔV-value of this signal is more than twice that of the SES depicted in Fig. 2 (note the difference in scale) but has been recorded on a line of only 20 m. Chart speed: 3 min per line. The three time marks refer to: 07:30, 07:48 and 08:30 respectively.

close to Limnos Island, i.e. at 39.9° N, 24.5° E. No significant variation of any component of the magnetic field was observed at the time. Another, almost equally strong event ($M = 6.7$) occurred at 14:11 GMT on Dec. 19, 1981 with an epicenter close to Agios Efstratios Island, about 160 km northeast of Athens; at that time the IRA-station had not yet been installed but the precursor SES was clearly recorded—almost 7 h prior to the shock—at GLY station which usually exhibits much noise. The corresponding SES is given in Fig. 3. The magnetic field did not show again any noticeable variation.

The smallest epicentral distance at which SES have been recorded is around 10 ± 5 km. In Fig. 4 we give the SES recorded at PIR-station at 20:50 GMT on Feb. 17, 1983; 56 h later an $M = 4.8$ event occurred, the epicenter of which was only 10 ± 5 km away from the station.

The SES usually starts gradually although in some cases instantaneous onsets (rise-time 1 min or smaller) have been noticed. In the majority of cases the end of

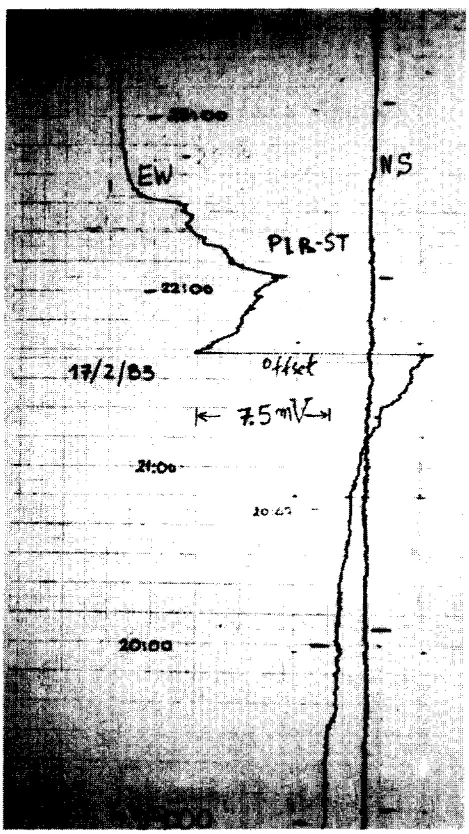

Fig. 4. A strong SES, that belongs to group II, recorded on an E–W line ($L = 50$ m) at PIR at 20:50, Feb. 17, 1983; it was followed by an $M = 4.8$-EQ that occurred at 05:45 on Feb. 20, 1983, with an epicenter close to the station ($r \simeq 10 \pm 5$ km).

the signal decays gradually. The end is abrupt only in cases when an SES shows an instantaneous onset and a duration of only a few minutes (Figs. 5 and 6). The opposite case i.e. a gradual onset and instantaneous end of the signal has never been observed. A smooth signal, i.e. gradual onset and end has also never been observed for duration less than 5 min. In some rare cases the recording shows a remarkable overshoot at the start and the end (Fig. 3).

When comparing SES for earthquakes from the same seismic region or for aftershocks of a given strong event we find that they usually do not have the same form; in order to visualize this we give in Fig. 7 the SES of the main shock 6.5 of Jan. 17, 1983, that occurred close to Kefallinia Island in the Ionian Sea (38.1° N, 20.23° E) whilst in Figs. 8 and 9 the SES of the two largest aftershocks $M = 6$ (Jan.

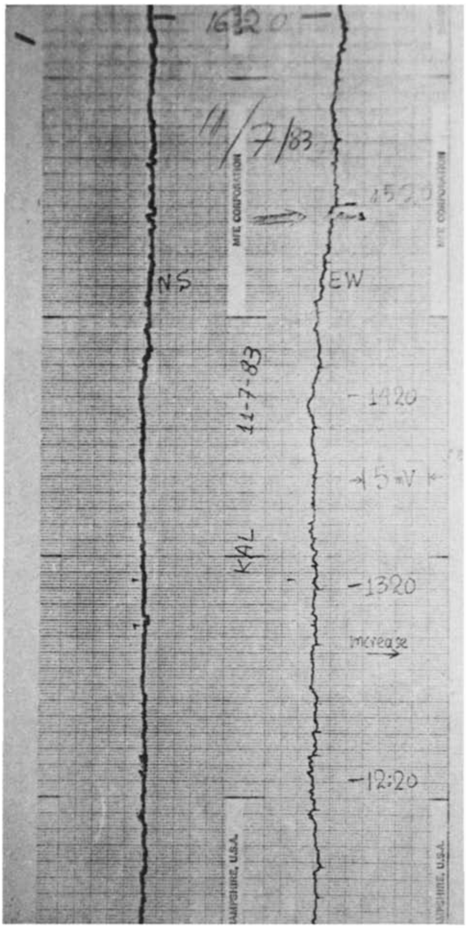

Fig. 5. An SES with abrupt edges and of small duration recorded on an E–W line (L = 200 m) at KAL on 15:15, July 11, 1983; it was followed by an M = 5.8-event that occurred at 02:55 on July 14, 1983 with an epicenter 150 km south of the station. The arrow shows the SES.

19) and $M = 5.7$ (Jan. 31) are depicted. No obvious similarity emerges from the comparison of these last three figures; however in some other cases earthquakes from the same seismic region (i.e. from Kalavrita region, compare Figs. 10, 11 and 12) give strikingly similar SES.

An SES, as mentioned, is characterized by a change of the potential difference that recovers its normal value long before the earthquake occurs. As well as the SES, *gradual changes of the whole background* with a duration of a few days may appear before an EQ, as has been reported by Myachkin et al. (1972) and by Sobolev (1975). *This effect* is evident in Figs. 7 and 8 and may, when strong, cause difficulties with the measurement of the true value ΔV of the SES since the latter signal is superimposed on a continuously varying background.

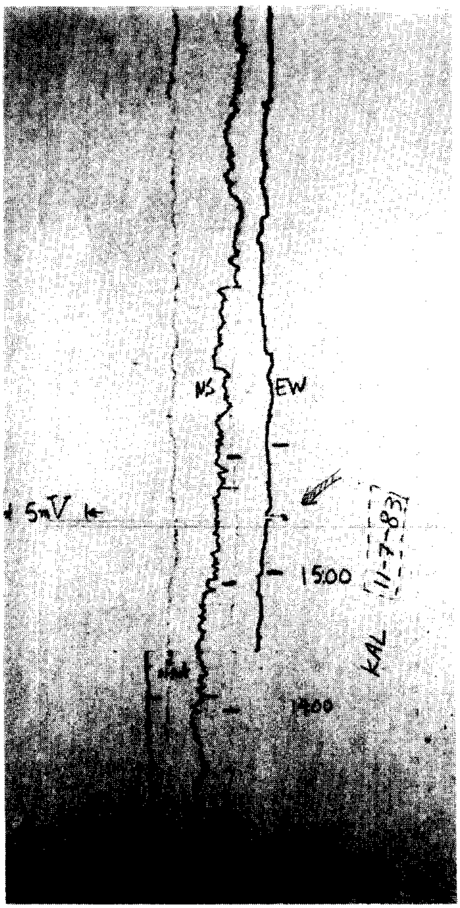

Fig. 6. The same SES as in Fig. 5 simultaneously recorded at the same station but on a parallel line (i.e. E–W) of half the length ($L = 100$ m). By comparing this to Fig. 5 one sees a strict verification of the "($\Delta V/L$)-test".

DISCRIMINATION OF THE SEISMIC SIGNALS FROM ELECTRIC DISTURBANCES OF OTHER SOURCES

Magnetic disturbances

As already mentioned disturbances of the telluric electric field can be induced by usual magnetic variations. They can be excluded if the magnetic field is continuously monitored. It should be noticed, however, that during magnetic storms the corresponding electrical variations are so strong that the SES cannot be recognized when the impending EQ is weak or the epicentral distance large. The SES can usually be simultaneously collected at a *small* number of stations depending on the distance from the epicenter and the magnitude of the impending EQ. This is in contrast to

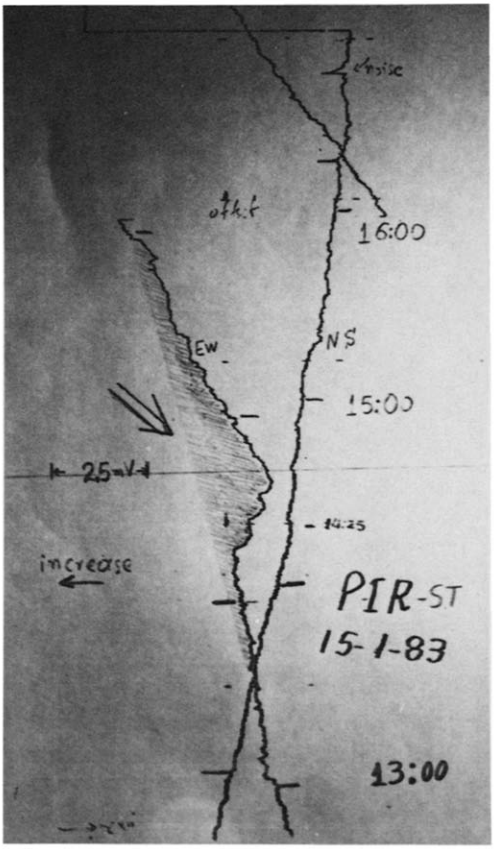

Fig. 7. An SES collected on the E–W line ($L = 50$ m) of PIR at 14:22, Jan. 15, 1983; it was followed by a $M = 6.5$ event which occurred at 12:41, Jan. 17, 1983 in the Kefallinia region, i.e. at a distance 120 km WNW of the station. Note the gradual strong variation of the background (Sobolev's effect) that usually starts 3–16 days before the strong events (see the text).

magnetotelluric disturbances which are usually recorded at *all* sites practically simultaneously but with varying strength depending on the line (E–W, N–S) and the station.

Magnetotelluric disturbances are nevertheless a very serious shortcoming but we are planning to solve the problem by determining the transfer functions between the electric and magnetic field for each station. The magnetotelluric disturbances can then be separated from our data by measuring the three components of the magnetic field. An automatic on-line substraction has not been done yet but is planned in cooperation with the University of Uppsala.

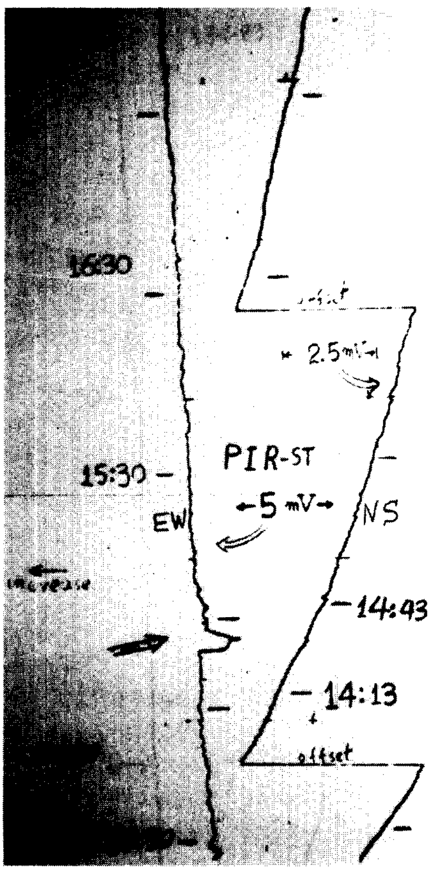

Fig. 8. An SES recorded on the E–W line ($L = 50$ m) of PIR at 14:30, Jan. 18, 1983 which preceded the $M = 6.0$ event that occurred in the Kefallinia region $10\frac{1}{2}$ h later, i.e. 00:02, Jan. 19, 1983. Note the Sobolev effect, i.e. the strong gradual variation of the background in the N–S line, an effect that also appeared before the main shock (see the previous figure) but then on both lines. One should notice that this SES has an appreciably smaller duration and a different form than the SES of Fig. 8 that preceded the main shock.

The total magnetic field is continuously monitored at IOA-station with a proton magnetometer and, when indispensable, with the three-component magnetic recording of the Penteli station (10 km from Athens). Furthermore, by combining the

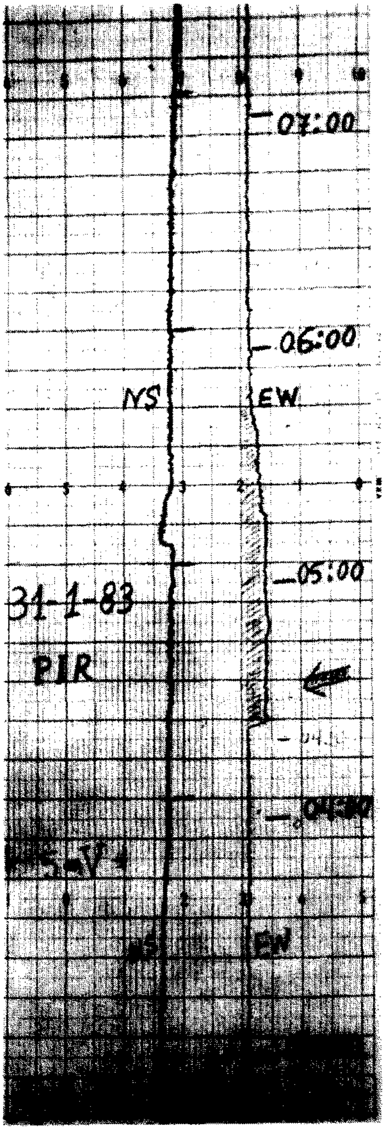

Fig. 9. The precursor SES of a $M = 5.7$ event that occurred in the Kefallinia region at 15:27, Jan. 31, 1983; it was recorded on the same line ($L = 50$ m) of PIR with the same polarity as the SES of the main shock depicted in Fig. 7. Note that although this SES and that depicted in Fig. 8 refer to roughly equal aftershocks of the same main event, they do not have the same form, nor the same duration but they do have the same polarity (see the text).

electrical recordings of all stations we can distinguish the SES from the electric variations due to magnetic causes. We would like to emphasize again that *no* significant variation of the magnetic field is produced by the signal.

As expected, in most of our stations the two lines are not equally sensitive to the

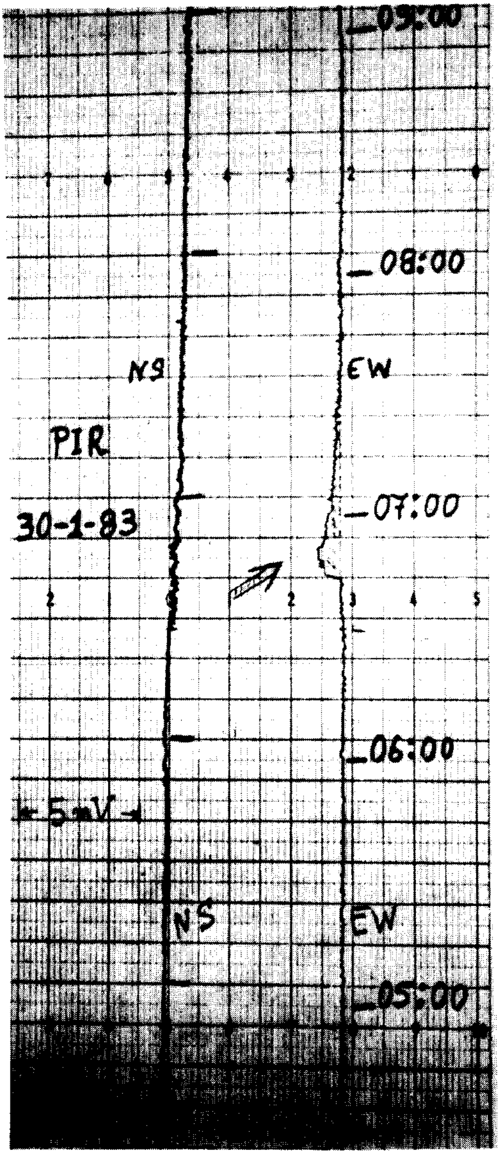

Fig. 10. An SES collected on the E–W line of PIR ($L = 50$ m) at 06:45, Jan. 30, 1983; it is a precursor of a $M = 4.3$ event that occurred at 17:06 on Jan. 30 at a distance of 60 km from the station (i.e. at the Kalavrita region 37.9 ° N, 21.8 ° E). Note that the lead-time is around 10 h, i.e. the SES belongs to group I.

magnetotelluric disturbances; this anisotropy creates difficulties when—during a magnetic disturbance—an SES is recorded on the line of the station which is strongly sensitive to magnetic variations. On the other hand this anisotropy helps towards the recognition of the SES when it is recorded on a magnetically "insensitive line". A striking example of the latter case can be seen in Fig. 13: a clear SES has been recorded at 20:30, June 13, 1983, on the "insensitive" N–S-line of the REN station (it corresponds to an $M = 4.3$ event that occurred almost 7 h later at a distance 40 km south of the station); the other line, E–W, of this station is appreciably more sensitive to magnetic variations as can be roughly seen in the same figure by considering the "magnetic signals" from 18:30 until 20:00 or in Fig. 14 from 21:46 to 22:46 which are also clearly recorded on the E–W line of VOL and the N–S line of GOR.

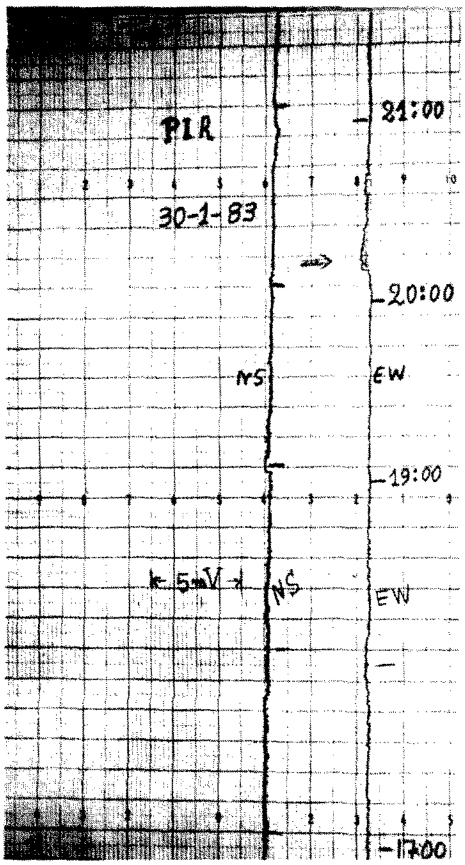

Fig. 11. An SES collected on the E–W line ($L = 50$ m) of PIR at 20:10, Jan. 30, 1983 which was followed 10 h later (i.e. at 05:33, Jan. 31, 1983) by an $M = 3.4$ event with the same epicenter as the EQ in Fig. 10. Note the similarity of this signal with that depicted in the previous figure.

Electrochemical disturbances

The metal electrodes cause, especially after rain, anomalous discontinuities which are due to electrochemical effects and may be sometimes confused with SES. In order to avoid this source of error we have installed two lines in each direction, the length of which have a ratio between 2 and 4. The electrical disturbance due to the contact effects of the metal with moisture does not start and end simultaneously on both parallel lines; but even in the rare case that they do occur simultaneously, they do not have ΔV-values with a ratio equal to the ratio of the corresponding lengths of the lines because they are chemical effects on the surface of the electrodes. On the contrary the electrical disturbance of a true SES gives the same field-strength $\Delta V/L$

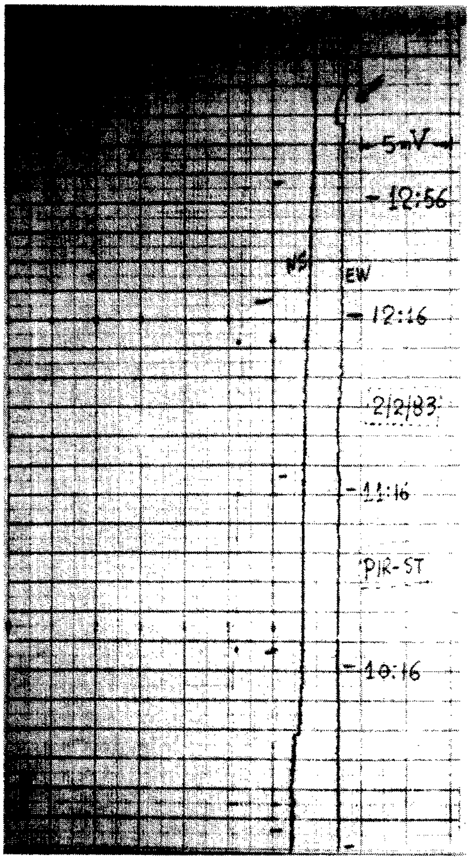

Fig. 12. At 13:20, Feb. 2, 1983 an SES was collected on the E–W line ($L = 50$ m) of PIR; an $M = 4.2$ event occurred at 05:51, Feb. 4, 1983 with an epicenter at (38° N, 22.0 E), i.e. roughly only 20 km away from the EQ of Figs. 10 and 11. Attention is drawn to the fact that the SES of Figs. 10, 11 and 12 were emitted from the same seismic region and were all recorded on the same line (i.e. E–W) of PIR with the same polarity (i.e. increase of the E–W component).

for both parallel lines. In practice if one considers a relatively strong event (M about 6) and an epicentral distance of 100 km, the ΔV-value is roughly 1 mV for $L = 50$ m (the exact ΔV-value depends on the relative "resistivity" of the line as will be discussed below) whereas the corresponding cultural noise is about 0.1–0.2 mV. It is obvious that a parallel line of $L = 200$ m will give a ΔV-value of 4 mV and hence will allow the verification of the constancy of the ratio $\Delta V/L$. This so called "$\Delta V/L$-test" gives a check beyond experimental error when the ratio of the lengths is larger than 2; furthermore the proportionality between ΔV and L is a proof that the origin of an SES is a change of the field strength. An example for the verification of the $\Delta V/L$-test for an $M = 5.8$ event is given in Figs. 5 and 6.

Strong cultural electrical disturbances

Local disturbances due to strong currents introduced into the earth, for example by a factory, cannot be recognized by the $\Delta V/L$-test because they produce devia-

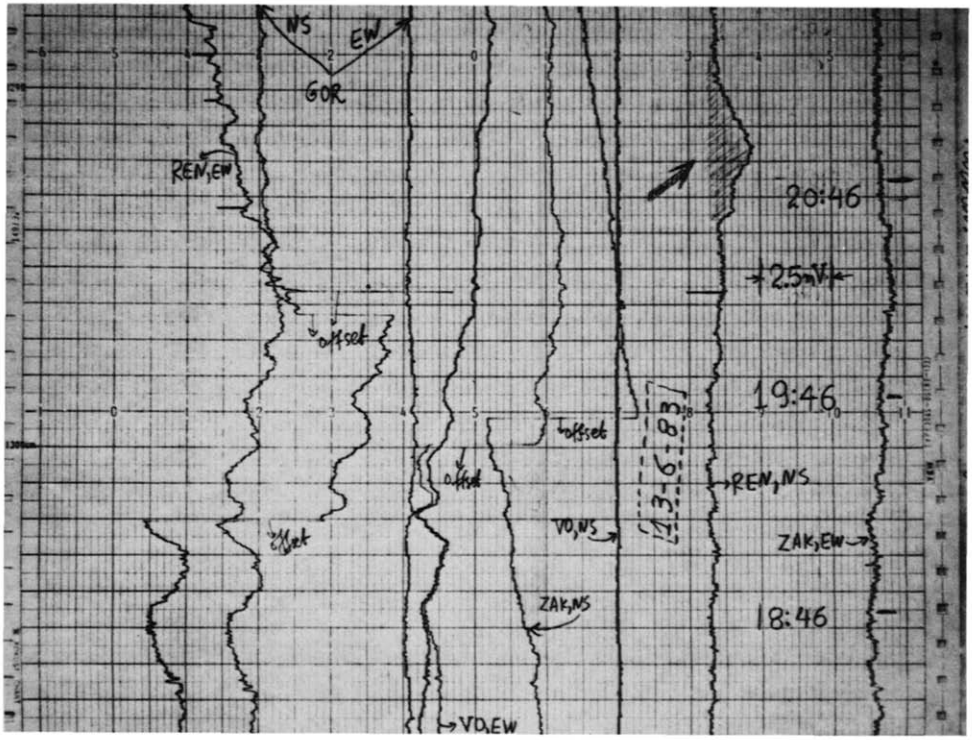

Fig. 13. An SES recorded on the N–S line of REN ($L = 30$ m) during a period of magnetic disturbances. Eight hours later (i.e. 04:40, June 14, 1983) an $M = 4.3$ event occurred ~ 40 km south of the station. Note that magnetic variations have induced electric disturbances from 18:26 until 19:56 which are simultaneously recorded on REN (E–W), GOR (N–S), GOR (E–W), VOL (E–W) but not on the N–S lines of REN and VOL that are "magnetically insensitive" lines.

tions that are proportional to L. To recognize an SES in such cases one has to revert to the simultaneous appearance of a signal (see next paper, i.e. Varotsos and Alexopoulos, 1984) at two stations separated by a distance of tens or hundreds of kilometers. At these distances simultaneous electrical variations of an artificial nature are precluded.

THE LEAD TIME OF SEISMIC ELECTRIC SIGNALS

Extensive time charts of SES and EQ have already been published for observations from a small number of stations (Varotsos et al., 1981a, b, 1982a). The corresponding correlation curves, i.e. plots in which the ordinate gives the percentage of correlated events that occur within a time span of 1 h as a function of the time difference $\Delta t = t_{EQ} - t_{SES}$ (where t_{EQ} is the time of the EQ and t_{SES} is the starting time of the signal), were also published for time differences Δt between -24 and $+24$ h. A maximum emerged for Δt between $+6$ and $13\frac{1}{2}$ h; it exceeds the statistical background noise by almost one order of magnitude. However, after the seismic

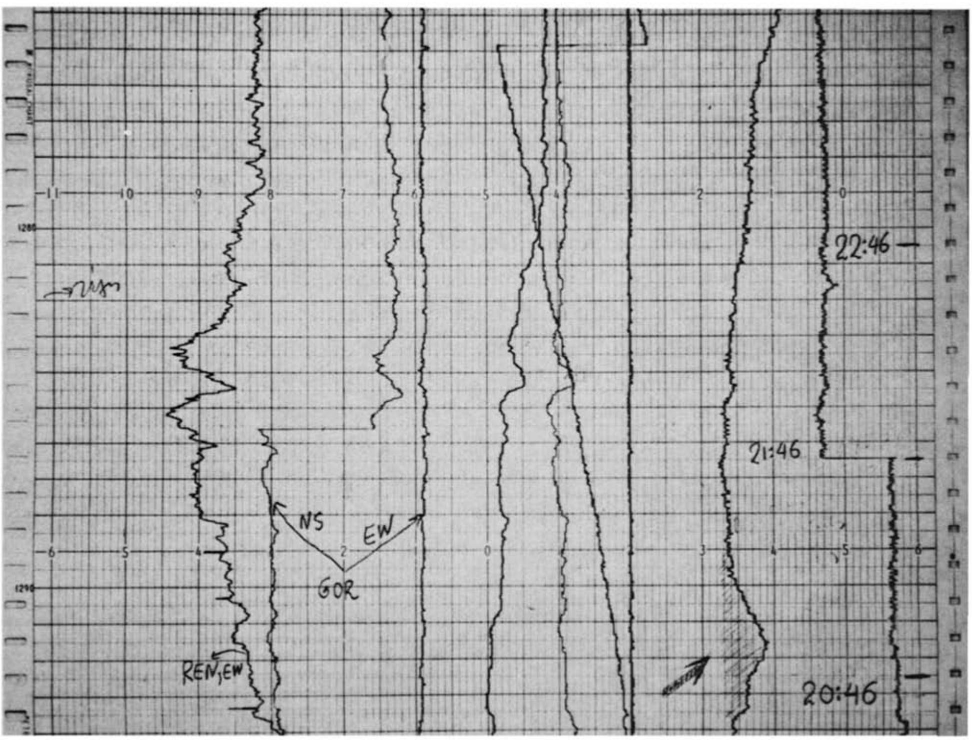

Fig. 14. Continuation of Fig. 13 in which we see that a magnetic variation at 21:56 has induced clear electric disturbances in various lines but only creates a scarcely visible disturbance on N–S of REN; in this line we see the SES from ~ 20:35 until 21:26 while a gradual deviation of the background starts almost one hour later. The latter is the Chinese effect discussed in the text.

activity that started on Jan. 17, 1983, in the Ionian Sea we were able to verify a prior suspicion that the lead time in some cases can reach values up to around 115 h (Varotsos et al., 1983). We shall now describe our present knowledge concerning the values of Δt.

The lead times can be classified mainly into two groups: *group I* with values from 6 to $13\frac{1}{2}$ h with a strong maximum around 7 h. The lead times of about 60% of all events fall into this group. *group II* with lead times between 43 and 60 h with a flat maximum between 45 and 54 h. Around 25% of all Δt-observations fall into this second group. Lastly there are two intermediate groups with lead times of between 24–36 h and 60–115 h which are however rarely observed. * It should be stressed that the form and the duration of SES are independent of Δt.

We have not observed a dependence between a seismic region and the lead time; as an example we refer to the Kefallinia region which emits signals with different Δt; their lead times belong either to group I or to group II: the $M = 6.5$ event of Jan. 17, 1983 (Fig. 7) belonged to group II in contrast to the events of Jan. 19, Jan. 31 (Figs. 8 and 9) and March 23, 1983 ($M = 6$, 5.7 and 6.4 respectively) which belonged to group I.

In spite of the fact that the Δt-values vary roughly by one order of magnitude (6–115 h) there is also no correlation between Δt and the corresponding earthquake magnitude.

EMPIRICAL RULES CONCERNING THE SIGNALS

In this section we describe the empirical relations that connect the measured ΔV-values to the site of the station, the seismic region, the epicentral distance and the magnitude. As already mentioned the precursor signals appear either on one of the two lines or on both of them. In order to answer the question whether a rule is independently valid or is valid only under certain restrictions we have proceeded to a systematical study in which the above parameters are changed one at a time while considering one given direction (e.g. E–W). The investigation is therefore separated into the following cases: (a) signal strength measured on a given line of a given station in function of magnitude of earthquakes from a given seismic region; (b) signal strength of earthquakes of a given magnitude measured on a given line of a given station but for different seismic regions, i.e. for different epicentral vectors; (c) signal strength that corresponds to earthquakes of a given magnitude measured on the same line (i.e. E–W) of various stations. In this way it is possible to find in what way the strength is connected to the station, the site, the seismic region or the intervening route of the current. This detailed, systematic approach might also give some insight into the mechanism of the current emission. At this point we stress that

* In a few percent of (isolated) cases there is some evidence that Δt can reach values up to 1 week.

the present study is possible because of the following empirical property: the ΔV-values of signals emitted from a given seismic region and registered on a given line of a given station are always the same for earthquakes of equal magnitude (in contrast to Δt and τ which vary with M in a non-systematic way). Furthermore, under the same conditions, i.e. for a given seismic region and at a given station, if signals from one EQ only appear on one line they will do so for all EQ. The explicit role of the epicentral distance r on the connection between ΔV and M is found by comparing for each magnitude the ΔV-values registered at a given station for various epicentral distances. Having thus clarified the influence of M and r on the ΔV-values of a given line of a given station we proceed finally to a "true" comparison of values recorded on lines of the same direction at two different stations. A direct comparison is not possible as a ΔV-value recorded at a station is not solely influenced by the epicentral distance but also by the resistivity and the inhomogeneities under each station.

Dependence of ΔV on the magnitude

In Fig. 15 we plot log ΔV vs M for SES measured on the E–W line ($L = 50$ m) of PIR-station (see Table 1). Curve A refers to earthquakes with their epicenters in the

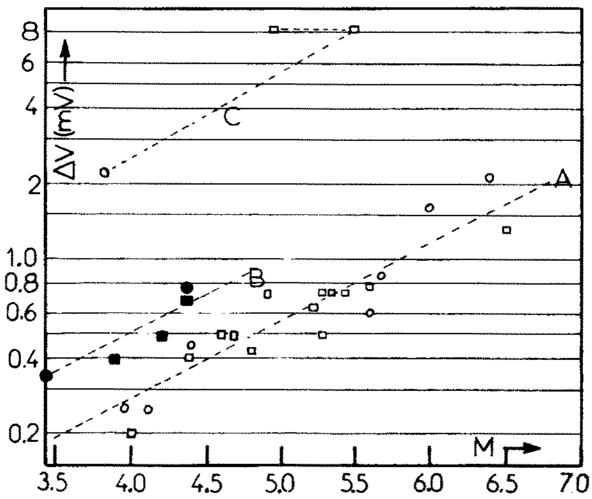

Fig. 15. The log ΔV vs M plot for SES recorded solely on the E–W line of PIR. Curve A—earthquakes from the Kefallinia region, i.e. 120 km from the station; ○ = group I, □ = group II; slope 0.31. Curve B—SES from the Kalavrita region, i.e. 50–65 km NE of the station; ● = group I, ■ = group II; slope 0.33. Curve C—two SES that correspond to earthquakes that occurred close to the station (10 ± 5 km); the magnitude of the stronger EQ is uncertain and hence they were not further considered in the calculation made in the text. It may be that curve C has a higher slope than A and B; if this is so the deviation might be due to the fact that the distance at which the SES were recorded is not orders of magnitude larger than the dimensions of the seismic volume.

TABLE 1

Earthquakes collected on the E–W line of PIR (L = 50 m)

Date (1983)	Time (GMT)	M (M_S)	ΔV (mV)	Group
A. From Kefallinia region (r = 120 ± 20 km)				
17-1	15:54	5.3	0.5	II
17-1	16:54	5.3	0.75	II
17-1	12:41	6.5	1.3	II
19-1	05:42	5.2	0.65	II
19-1	00:02	6	1.6	I
22-1	16:02	4.6	0.5	II
22-1	12:54	4.9	0.75	II
28-1	17:43	4.4	0.45	I
31-1	15:27	5.7	0.85	I
15-2	03:38	4	0.3	I
16-2	16:50	4.8	0.43	II
2-2	04:25	4.1	0.25	I
13-3	13:53	4	0.2	II
15-3	21:20	4.4	0.4	II
15-3	23:31	4.2	0.4	II
23-3	23:51	6.4	2.1	I
24-3	04:17	5.6	0.6	I
13-5	23:50	5.4	0.73	II
14-5	23:14	5.6	0.77	II
14-5	23:26	5.3	0.75	II
B. From the Kalavrita region (r = 50 ~ 65 km)				
30-1	17:06	4.4	0.8	I
31-1	05:33	3.4	0.35	I
4-2	05:51	4.2	0.5	II
9-2	12:41	3.9	0.4	II
2-1-84	07:07	4.4	0.7	II

TABLE 2

Earthquakes * from a given seismic region and collected at the same station

Date	Time (GMT)	M (M_S)	ΔV (mV)
A. Earthquake from Limnos Island collected on the E–W line of VOL (L = 100 m. r = 140 ~ 160 km)			
6-8-83	15:43	6.6	3.5
6-8-83	16:46	5	1.5
8-8-83	08:10	5.2	1.6
8-8-83	14:43	4.4 **	0.7
12-8-83	07:29	4.4 **	0.5
23-8-83	05:42	4.6	0.8
B. Earthquake from Agios Efstratios collected on the E–W line of GLY (L = 50 m; r = 140 ~ 160 km)			
19-12-81	14:11	6.8	15.6
27-12-81	17:39	6.4	11.2
29-12-81	08:01	5.9	7.5

Kefallinia region (latitude 38° N, longitude 20° E), i.e. at an epicentral distance of around 120 ± 20 km. An inspection of the plot indicates that there is no significant systematic difference between group I and group II and that a least squares solution for a straight line reveals a slope 0.31 and a correlation factor 0.92. Considering the same station PIR and component E–W, we plot in curve B the (absolute) values of ΔV for SES obtained from earthquakes from a different seismic region, i.e. from the Kalavrita region (lat. 38° N, long. 22° E). Note that the epicentral distance r of the latter seismic region is roughly half (55 km) of that of the previous one and has a different azimuthal angle. The best fit to a straight line has a slope of 0.33 with a correlation factor 0.91. Similar graphs have been made for SES collected at other stations and for other seismic regions (see Table 2 and Fig. 16); they have slopes

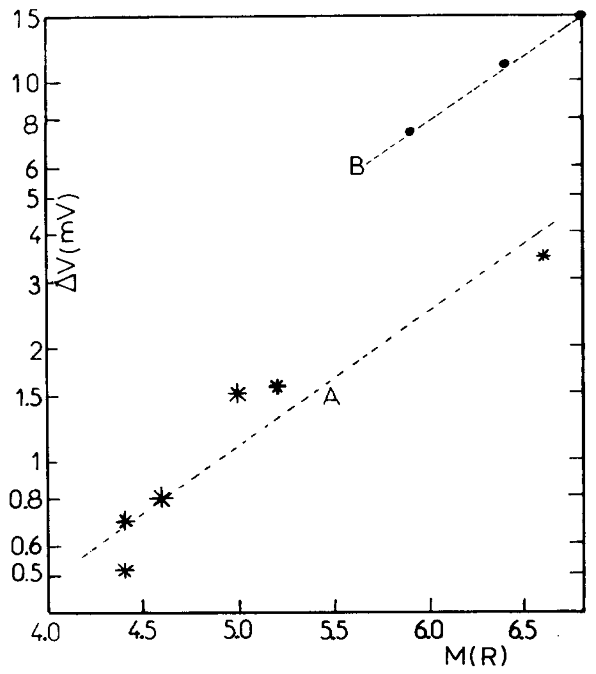

Fig. 16. ΔV-values for earthquakes of a given seismic region and recorded on a single line of a given station. Curve A: E–W line ($L = 100$ m) of VOL corresponding to EQ from Limnos Island (235 km NE of Athens). Curve B = E–W line ($L = 50$ m) of GLY corresponding to EQ with epicenters close to Agios Efstratios (160 km NE of Athens). This line is almost ENE–WSW; here we shall treat it as an E–W line. $M(R) \equiv M_S$.

Notes to Table 2:

* Seismic data from the preliminary seismological bulletin, National Observatory of Athens, officially certified; in the monthly bulletin the epicenters and magnitudes are slightly revised. When M_S is not given in the bulletin we estimate it according to: $M_S = M_L + 0.4$.

** These magnitudes have been later revised by −0.1 and −0.2 units respectively.

practically equal to the previous ones. We suggest that this slope-value does not reflect a physical property of a station nor of a seismic region but has to do with the current producing mechanism.

An increase of the magnitude by 1 R increases the log ΔV by 0.3 (for $r = \text{const.}$) which means that the amplitude of the signal increases by a factor 2 to 2.5. At first glance this is unexpected: considering the formula $E = \frac{1}{2}\delta^2 V$ where E is the energy of the stressed *spherical* volume V one finds:

$\log V = 1.5\ M + \text{const.}$ (because $\log E = 1.5\ M + \text{const.}$).

Then the effective surface $S \sim (V^{2/3})$ is:

$\log S = 1\ M + \text{const.}$

and hence one would expect (for the points for a given station and a given seismic region):

$\log \Delta V = 1\ M + \text{const.}$

This comes from the consideration that the current and hence ΔV (because the resistivity is the same as we consider a given station) is proportional to the emitting surface. In other words one would expect that when the magnitude increases by 1 1-unit, the ΔV value would increase by a factor of 10 in contrast to the experimental data. On these grounds one could reach the speculative assumption that the volume is non-spherical with axes l, w, h and that the current-emitting surface (e.g. $w \times h$) increases appreciably slower than the other surfaces ($l \times h$ or $l \times w$) when magnitudes increase. Our experimental data are compatible with:

$\log(w \times h) \approx (0.3 \sim 0.4) M + \text{const.}$

and hence:

$\log l \approx (1.1 \sim 1.2) M + \text{const.}$

This last conclusion should be compared with various empirical studies that connect the logarithm of the length of the fault with the magnitude (Kasahara, 1981; Purkaru and Berckhemer, 1982); in some of these studies the coefficients of M reach values up to 1.2.

The above very speculative result is based inherently on the assumption that the transient SES is not a result of the variation of any physical property of the ground under the station but is caused by precursor changes in the region of the focus. It is appropriate to indicate here that Fuye et al. (1983) have recently reported variations of the resistivity ρ prior to earthquakes according to:

$M = 3.84 \log(\Delta\rho/\rho) + \text{const.}$

which gives $\log(\Delta\rho/\rho) \simeq 0.3\ M + \text{const.}$

It is curious that this expression has the same slope as $\log \Delta V$ vs. M although the $\Delta\rho/\rho$-technique measures variations of a physical property of the ground under the station at a depth of a few hundred meters (depending on the method it is measured with) while the SES is related, in our opinion, to the current production at the focus.

Dependence of ΔV on the epicentral distance

We now examine the ΔV-values registered on the same line of the same station for earthquakes of given magnitude from different seismic areas; for such measurements the resistivity (and the inhomogeneities) of the station is a constant and only the epicentral azimuth and distance change (Table 3). In Fig. 17 we plot the results for the E–W line of PIR-station for the magnitude range 3.8–4.2 as function of the epicentral distance (curve A); note that the values of ΔV and r vary by one order of magnitude. The least squares fit shows that the data can be smoothed by a function that is proportional to $1/r$ and that a r^{-2} law must be precluded. Curve B in the same figure exhibits the same feature for earthquakes with a magnitude of around 4.8. Note that for small distances, i.e. 10–20 km, we have some reservations concerning the validity of the $1/r$-behaviour because the few existing data indicate that there is a faster decrease of ΔV with distance. The same behaviour has been confirmed for other stations as well. Another curious effect will be discussed in Part II (Varotsos and Alexopoulos, 1984).

The above conclusion must not be interpreted as a general $1/r$-law for the attenuation of the ΔV-signals for one EQ recorded at various stations and hence with various epicentral distances: these ΔV-values do not follow an $1/r$-dependence; in this case, as will be seen below, the quantity which should be considered as attenuating with distance is the current density that results from the measured

TABLE 3

Seismic electrical signals collected on the E–W line of PIR (L = 50 m) and corresponding to earthquakes with various epicentral distances but with constant magnitude

Date (1983)	Time (GMT)	M	ΔV (mV)	r (km)	Group
A. $M = 4 \pm 0.2$ R					
6-1	04:24	4.2	0.6	50	I
2-2	04:25	4.1	0.25	100	I
4-2	05:51	4.2	0.5	60	II
9-2	12:41	3.9	0.4	65	II
14-2	10:37	3.9	0.7	43	I
14-2	23:14	4	0.4	78	I
15-2	03:38	4	0.3	96	I
19-2	23:46	4.1	0.3	86	I
17-3	20:18	3.8	1.3	20	I
B. $M \simeq 4.7$–5 R					
8-5	22:44	4.7	0.8	60	I
16-2	16:50	4.8	0.43	130	II
19-2	15:56	5	0.6	84	II
20-2	05:45	4.8	8	10	II

$\Delta V/L$-value by dividing by an "effective" resistivity which also deals with the local inhomogeneities.

It is worthwhile further clarifying the results of this paragraph by taking two examples:

Consider two seismic regions A and B that are at equal distances, e.g. $r = 50$ km, from station C and an earthquake from A of magnitude, e.g., 4 R which gives a signal ΔV_A on the E–W line of the station C. If we now examine another EQ with the same magnitude from region B it could in principle give a signal ΔV_B on the same line of station C that is appreciably different than ΔV_A (due to possibly different geological conditions in the two regions A and B). The fact that for a given r (e.g., 50 km) Fig. 17 gives a single value, e.g. 0.5 mV, shows that $\Delta V_A = \Delta V_B$, i.e. that ΔV does not depend on the seismic region (for M, r = const.). This implies that not only the physical mechanism that emits the current from A or from B is the same but also that some physical quantities associated with the current emission have the same values at A and B.

In the second example we assume that the epicentral distances of A and B from station C are different and have a ratio, e.g. $r_A/r_B = 3$. By considering two EQ with

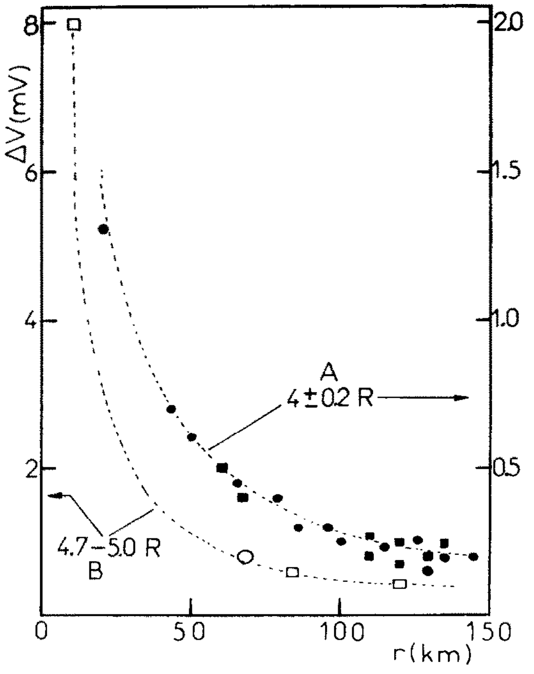

Fig. 17. ΔV-values for earthquakes with constant magnitude recorded on the E–W line ($L = 50$ m) of PIR versus the epicentral distance. Curve A: for $M = 4$ (± 0.2); ● = group I, ■ = group II. The points not included in the tables refer to the Kefallinia region. Curve B: for $M \approx 4.8$; ○ = group I, □ = group II. Note the difference in scale. The lines have been drawn only as a visual aid.

equal magnitudes—one from A and one from B—one finds from Fig. 17 that $\Delta V_A/\Delta V_B \simeq \frac{1}{3}$. This means not only that the mechanism (and the properties) producing the current at A and B were the same but further that the two signals, until reaching station C obeyed an attenuation according to r^{-1}-law.

By combining the conclusions drawn from the study of Figs. 15, 16 and 17 we find that for a given line of a given station, $\log(\Delta V \cdot r)$ should be a linear function irrespective of the seismic region. In effect if we combine that data of curves A and B of Fig. 15, we find a single linear connection for the E–W line of PIR-station (plot A of Fig. 18). A least squares fitting to a straight line gives a slope 0.35, comparable to those of curves A and B in Figs. 15 and 16, with a correlation factor 0.95.

Due to the fact that the curves $\log \Delta V$ vs M (for r = const.) have the same slope irrespective of the station and that ΔV is proportional to $1/r$ (for M = const.) the plots $\log(\Delta V \cdot r)$ vs M—for various stations—have to have the same slope. They do not have the same ordinates.

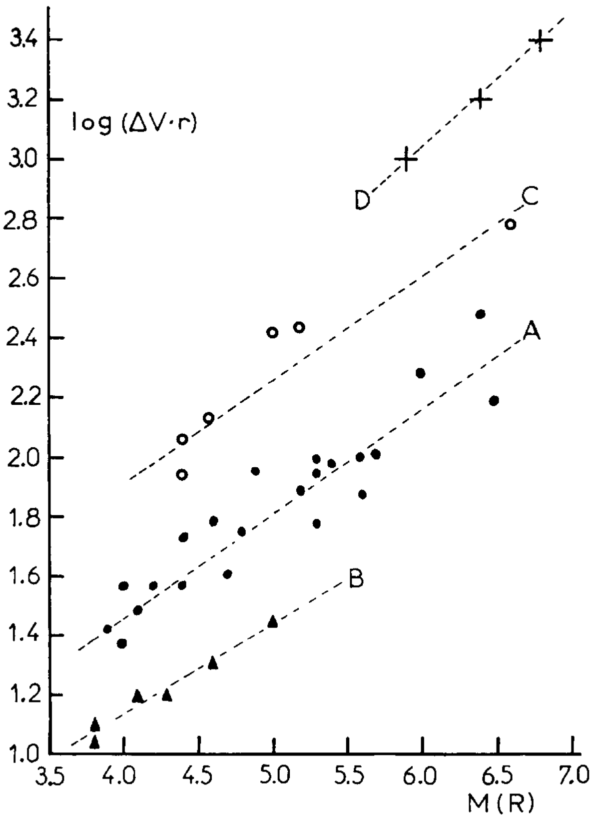

Fig. 18. $\text{Log}(\Delta V \cdot r)$ versus the magnitude for SES emitted from various seismic regions but recorded at the same station. Curve A: E–W line of PIR (L = 50 m); curve B: E–W line of ASS (L = 50 m); curve C: E–W line of VOL (L = 100 m); curve D: E–W line of GLY (L = 50 m). $M(R) \equiv M_S$.

Definition of the relative effective resistivity and the relative signal strength

As we have seen the plots of $\log(\Delta V \cdot r)$ for a line of the same direction (e.g. E–W, $L = \text{const.}$) do not coincide in general for two different stations. However, as mentioned, they all have the same slope and therefore their $(\Delta V \cdot r)$-value must have a constant ratio. The value of this ratio ($L = \text{const.}$), which depends on the stations (i) and (k) shall be labelled with:

$$\left.\frac{(\Delta V \cdot r)_i}{(\Delta V \cdot r)_k}\right|_{E-W} \equiv \left.\frac{\rho_i}{\rho_k}\right|_{E-W}$$

Once the values of ΔV and r on the left side of the equation are known (for a certain EQ) the quantity ρ_i/ρ_k can be determined for each pair of stations.

In Fig. 18 we have plotted the $\log(\Delta V \cdot r)$-values versus M for E–W lines of stations PIR, VOL, ASS and GLY (we intentionally present them for different lengths). By reducing the values to the same length the comparison of the ordinates gives:

$$\rho_{E-W,VOL}/\rho_{E-W,PIR} \simeq 1.4 \qquad \rho_{E-W,GLY}/\rho_{E-W,PIR} \simeq 8 \qquad \rho_{E-W,ASS}/\rho_{E-W,PIR} \simeq 0.4$$

By following the same procedure we have determined such ratios for each line of the stations of the network in comparison to the corresponding line of PIR, which we consider as a base station. The ratio $\rho_{E-W,i}/\rho_{E-W,PIR}$ can be called the *relative effective resistivity of the E–W line of station (i).*

One can write an expression of the form:

$$\Delta V/L = j\rho$$

The absolute value of the current density j, however cannot be determined from this equation because the actual resistivity is not known. A measurement with the usual resistance methods might not be representative of the true situation because of layers of varying resistivity under the station. However in practice when comparing signals from various stations one only needs the value of the relative resistivities because j can be expressed by:

$$j = \frac{\Delta V/L}{\rho} = \frac{\Delta V/L}{\rho/\rho_{bas}}\,\frac{1}{\rho_{bas}} = \frac{\Delta V/L}{\rho_{rel}}\,\frac{1}{\rho_{bas}}$$

where ρ_{bas} is a constant for all stations.

From the two components one gets for the total current density:

$$J = \left(\left|\frac{\Delta V/L}{\rho_{rel}}\,\frac{1}{\rho_{bas}}\right|^2_{E-W} + \left|\frac{\Delta V/L}{\rho_{rel}}\,\frac{1}{\rho_{bas}}\right|^2_{N-S} \right)^{1/2}$$

We have introduced different relative effective resistivities for E–W and N–S because $\rho_{i,E-W}/\rho_{i,N-S}$ may be far from unity (e.g. for the VER station it is around $\frac{1}{3}$)

TABLE 4

Complementary list of earthquakes inserted in Fig. 21

Date	Time (GMT)	*M* (R)	Epicenter
18-1-82	19:27	6.8	230 km NE of ATH
23-3-83	19:04	5.3	90 km S of IOA
15-4-83 *	06:05	3.8	35 km ENE of REN
15-4-83 *	06:12	3.6	35 km ENE of REN
1-6-83	14:44	5	300 km E of GLY
9-6-83	02:39	4.5	330 km E of GLY
13-6-83	17:14	3.7	140 km WNW of GLY
14-6-83	04:40	4.3	40 km S of REN
5-7-83	12:01	6.5	Dardanelles
14-7-83	02:55	5.8	150 km S of KAL

* EQ announced from THES-seismic network.

because of local inhomogeneities. Setting for the base station $\rho_{\text{bas,E-W}}$ and $\rho_{\text{bas,N-S}}$ equal to 1 we obtain a working formula for a quantity that can be called the *relative signal strength* defined by:

$$J_{\text{rel}} = \left(\left.\frac{\Delta V/L}{\rho_{\text{rel}}}\right|^2_{\text{E-W}} + \left.\frac{\Delta V/L}{\rho_{\text{rel}}}\right|^2_{\text{N-S}} \right)^{1/2}$$

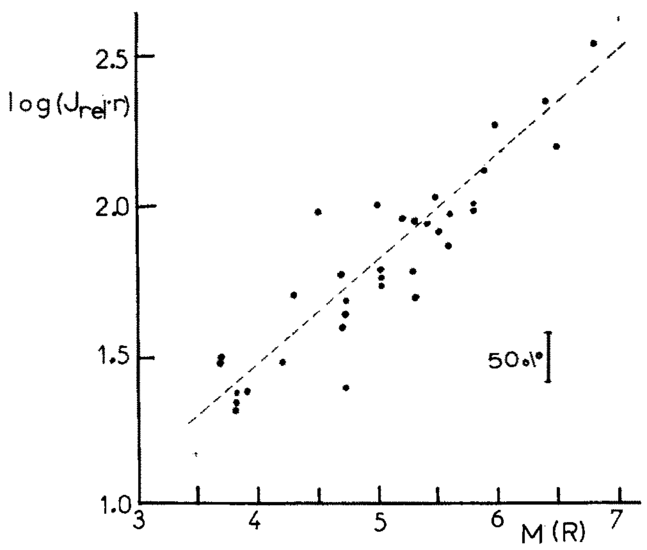

Fig. 19. Log($J_{\text{rel}} \cdot r$) versus the magnitude for SES recorded at various stations and corresponding to earthquakes from all over Greece. The error bars refer to an error of 50% in $J_{\text{rel}} \cdot r$. $M(R) \equiv M_S$.

Dependence of the relative signal strength on the magnitude

By using the values $\rho_{rel,E-W}(ASS) = 0.4$, $\rho_{rel,E-W}(VOL) = 1.4$ and $\rho_{rel,E-W}(GLY) \simeq 8$ we obtain the plot of Fig. 19 for $\log(J_{rel} \cdot r)$ versus M for the data mentioned in Tables 1–3. We emphasize that the events in these tables were chosen so as to have no N–S component. In this figure we also insert points corresponding to EQ mentioned in Table 4. A least squares fitting to a straight line gives a slope of 0.37 with a correlation factor 0.92. Experimental points for events recorded only in the N–S direction or in both directions fall on to the same line. It is therefore clear that $J_{rel} \cdot r$ is an unique function of M, valid for all stations, directions and seismic areas.

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