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Latest aspects of earthquake prediction in Greece based on seismic electric signals

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On the occasion of the 80th birthday
of Professor Kessar Alexopoulos

ABSTRACT

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Since 1983, continuous monitoring of the electrotelluric field has been carried out using an array of measuring stations located at various sites in Greece. The basic physical properties of the transient changes—seismic electric signals (SES)—in the electrotelluric field that are forerunners of earthquakes were first described six years ago. Since then a large body of data has been collected resulting in new insight into various aspects of the method. The present paper reviews the latest developments in SES-based earthquake prediction and describes the current procedures used to predict the epicenter and magnitude of an impending earthquake.

A detailed list of the predictions officially issued in Greece during the past 3 years (January 1, 1987–November 30, 1989) is also given. Public warnings were issued well before the most destructive seismic activity.

Introduction

Since January 1983, continuous measurements of the earth's electric field have been made at various sites in Greece. The data (8 differential channels per station, sampling rate: 3 readings/sec/channel) are continuously transmitted via telephone lines to a central station (Glyfada, GLY) located in an Athens suburb. The present configuration of the telemetric network, shown in Fig. 1, is slightly different from that given in Varotsos and Alexopoulos (1984a). Except for station SER, all stations are located in continental Greece. Varotsos and Alexopoulos (1984a) reported that transient variations in the earth's electric field — hereafter called seismic electric signals (SES)—are detectable before the occurrence of earthquakes (EQ).

In 1984, several examples, together with a short description of the basic properties of SES, were presented by Varotsos and Alexopoulos (1984a, b). A large body of data has since been collected resulting in new insights into various aspects of the method. The present paper details the most up-to-date knowledge of the physical properties of SES, as well as the current procedure used in predicting earthquake parameters. Some examples of recent SES are given in Figs. 2–7.

Latest insights into the basic physical features of SES

Data collected since 1983 necessitate a reconsideration of certain aspects formerly presented by Varotsos and Alexopoulos (1984a, b).

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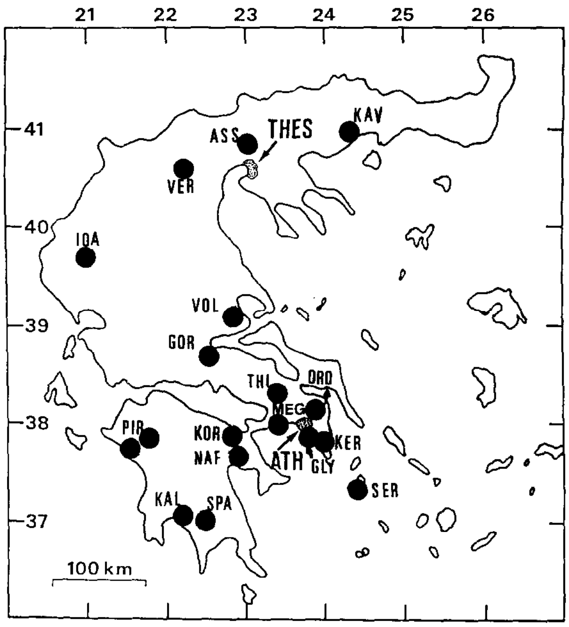

Fig. 1. Map showing the distribution of the Greek stations at which variations in the electric field of the earth are measured. The two largest cities Athens (ATH) and Thessaloniki (THES) are also shown. ASS = Assiros; GLY = Glyfada; GOR = Gorgopotamos; IOA = Ioannina; KAL = Kalamata; KAV = Kavala; KER = Keratea; KOR = Korinthos; MEG = Megara; NAF = Nauplio; ORO = Oropos; PIR = Pirgos; SER = Serifos; SPA = Sparta; VER = Veroia; VOL = Volos; THI-Thiva.

SES duration τ

The duration of SES lies between 1/2 min and several hours (Varotsos and Alexopoulos, 1987) and does *not* depend on the earthquake magnitude, *M*.

Time lag between SES and earthquake

For *isolated* events (i.e. when a single SES and a single earthquake allow a one-to-one correlation), the time lag Δt lies between 7 hours and 11 days. No correlation between Δt and M has been observed.

For cases of prolonged *electrical activity* (i.e. when a number of SES, detected within a time period comparable with the time lag Δt, is followed by a number of earthquakes) it has occasionally been observed that, although the time lag between the onsets of the electrical and seismic activity does not usually exceed 11 days, the time lag between the largest SES and the strongest earthquake may, however, be much longer, e.g., around 22 days. An example is provided by the destructive earthquakes in the Killini area, on Sept. 22, 1988 and Oct. 16, 1988. This is due to the fact that the time lag, Δt, is independent of the magnitude of the earthquake (see Table 1, Figs. 7 and 8). Consequently a sequence of SES with different amplitudes does not necessarily correspond to a series of earthquakes with magnitudes in the same order. This is illustrated in Fig. 9 which depicts a sequence of four SES with the first being the largest while in the corresponding sequence of four earthquakes the last one is the strongest (see also Appendix 1). In this example, the time lag between the *largest* (in amplitude) SES and the *strongest* earthquake obviously exceeds the aforementioned usual maximum of 11 days.

Furthermore, small values of Δt, of the order of 10 h, are usually related to aftershocks. On the other hand, seismic areas that have been quiescent

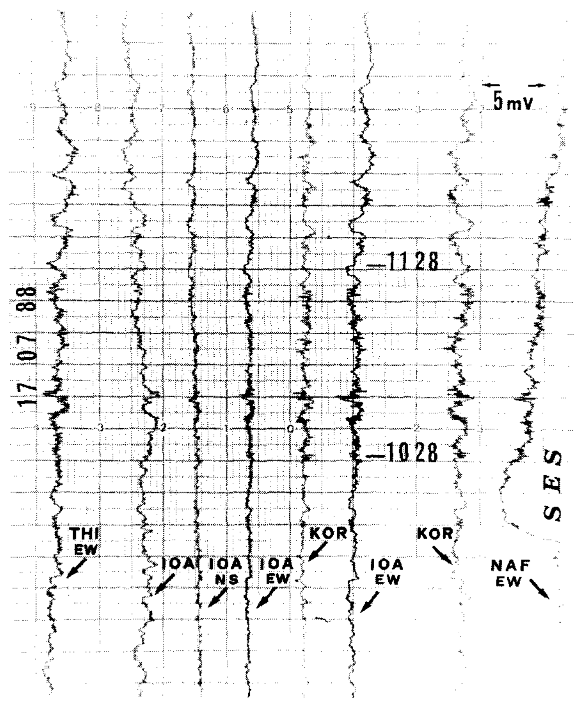

Fig. 2. SES detected by the EW dipole (L = 300 m) of NAF on July 17, 1988. The signal starts at 10:02 GMT. It was also recorded on the NS short dipoles (not shown) and provoked the issue of telegram No. 9 (see Table 1). The corresponding earthquake occurred on July 23 with $M_s \approx 4.4$, 200 km southwest of Athens. The scale applies only to the NAF dipole.

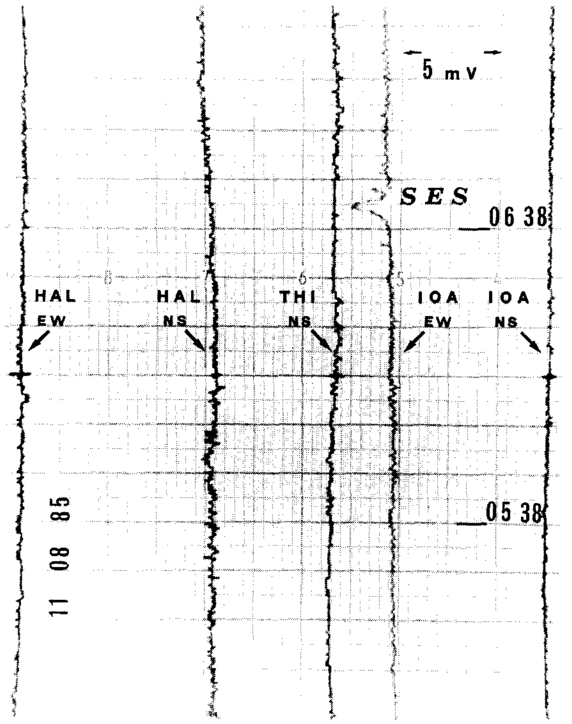

Fig. 3. SES recorded on August 11, 1985 on the short EW IOA dipole ($L = 47.5$ m). The corresponding earthquake ($M_s = 5.1$) occurred on August 13, 1985 with epicenter at 38.0° N, 21.4° E, i.e. close to PIR. For HAL-station see fig. 1 of Varotsos and Alexopoulos (1984a).

for a long time produce SES with Δt values of the order of a few days or more.

Form of the SES

In most cases, SES are recorded at a station by both component dipole arrays, i.e. EW and NS, usually with different values of $\Delta V/L$, where L is the length of the dipole (e.g. Figs. 4, 8). However, cases have been observed (e.g. Fig. 3) where the SES was recorded on all the parallel dipoles only in one direction.

The SES may have a gradual or abrupt (i.e. within 20 s) onset and a gradual or abrupt cessation. The combination "gradual onset/abrupt cessation" has *never* been observed. It should also be mentioned that SES originating from the same seismic area and recorded at the same station occasionally have strikingly similar shapes, though of course different amplitudes, since the latter depend on the magnitude (see below) of the corresponding earthquakes.

Relation between SES amplitude and earthquake magnitude

Consider the case of a given station S_A located in the area A and a seismic area B which emits SES that can be recorded at this station (see section on selectivity effect). The following two cases can be distinguished, depending on the number of components that appear:

(a) When only one SES component is recorded, the relationship between the SES amplitude, expressed as $\Delta V/L$, and the earthquake magnitude, M, is such that a plot of $\log(\Delta V/L)$ versus M gives a straight line with a slope between 0.32 and 0.37. If another seismic area C produces SES that are recorded at the same station but solely on the other component dipole array, the plot of $\log(\Delta V/L)$ versus M is again a straight line, with the *same* slope but with a different intercept. This

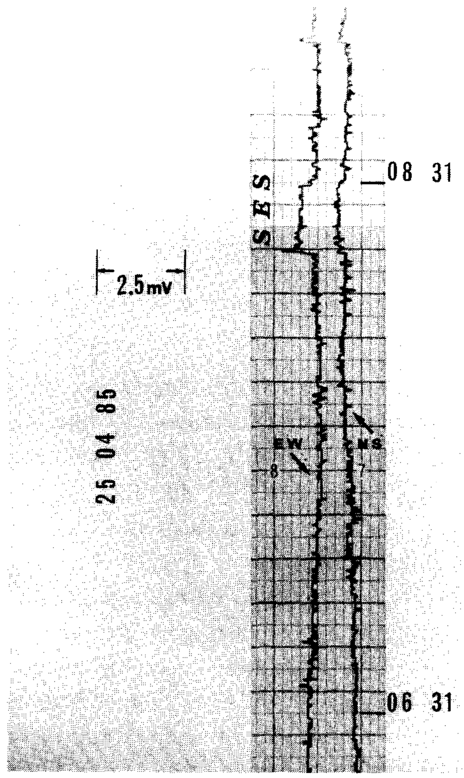

Fig. 4. Example of an SES recorded by two perpendicular dipoles. Signals detected by the short dipoles (L = 70 m) at station ASS on April 25,1985. The corresponding earthquake ($M_s = 5.8$) occurred on April 30, 1985 with an epicenter at 39.3° N, 22.9° E, i.e. close to VOL. Note that this SES was *not* recorded at station VOL due to selectivity.

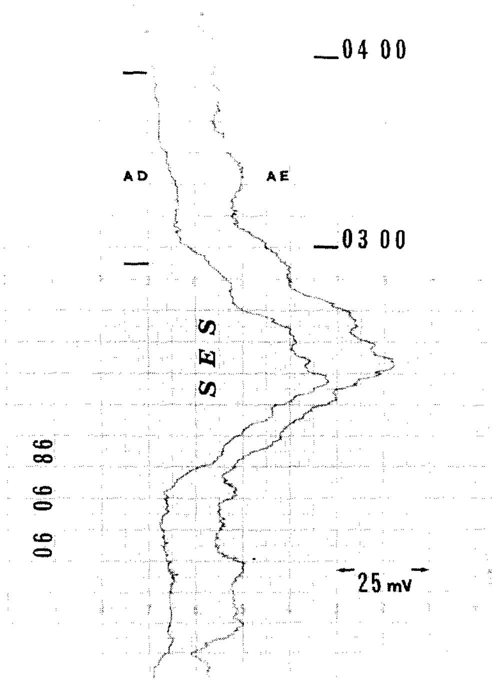

Fig. 5. Example of an SES lasting more than 1 hour, recorded at station KER on June 6, 1986 by (almost) parallel dipoles AD (L = 1.3 km) and AE (L = 1.5 km) oriented approximately NNW. Note that electrode A is common to both dipoles and lies within a dyke (granodiorite) with a surface width of the order of some meters, while points D and E are located in different rocks (granite and marble, respectively). From other dipoles it was verified that the variation shown in the figure is not due to an electrochemical variation at the common point A. The corresponding earthquake ($M_s = 4.8$) occurred at 15:35 GMT on June 6, 1986 with the epicenter at 38.9° N, 22.9° E.

variation in the intercept is not solely due to a difference in the epicentral distances r_{AB} and r_{AC} since it is also observed when $r_{AB} \approx r_{AC}$. The effect appears to be, at least partially, due to a significant difference in the resistivities of the station's two component dipole arrays.

(b) When SES are recorded by both perpendicular dipole arrays at S_A, for earthquakes from area B, the *ratio* of the amplitudes of the two components *remains the same* for all earthquakes and, as said before, the amplitude of each component increases with earthquake magnitude so that the resulting plots of $\log(\Delta V/L)$ versus M, for each component are straight lines with the same slope * (0.32–0.37) but with different intercepts (Figs. 10 and 11). This difference in the intercepts of the two components (EW and NS) is partially attributed to a difference in the so-called "relative effective resistivities" ρ_{rel} as defined by Varotsos and Alexopoulos (1984a, b).

Comments on the ρ_{rel} values of a given station

Several points about the value of ρ_{rel} at a given station are worthy of note:

(a) The ρ_{rel} values of two parallel, short dipoles (e.g. 50 m and 200 m) are the same only for a station located in an area with homogeneous geology beneath the station (which, however, may be anisotropic). If there are strong inhomogeneities close to the station, the ρ_{rel} values of two parallel dipoles may differ even by one order of magnitude. An example of this is given later. In such cases a vertical SES-component is usually expected.

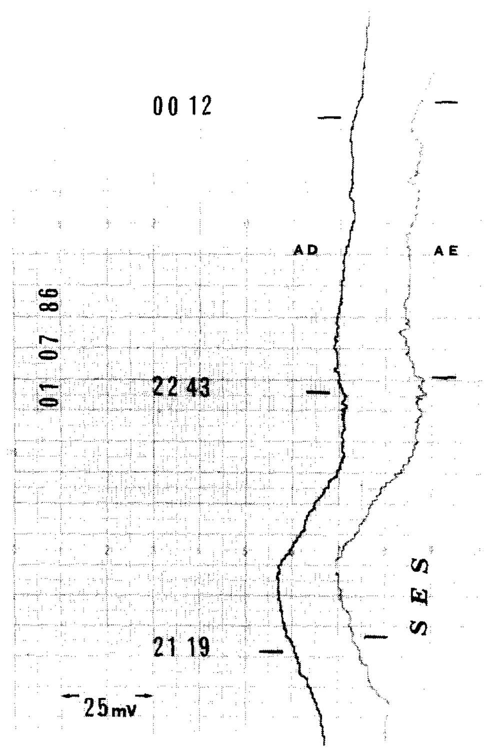

Fig. 6. SES recorded at KER on July 1, 1986 by the same dipoles as in Fig. 5. The corresponding earthquake ($M_s = 4.6$) occurred on July 5, 1986 with epicenter at 37.8° N, 22.4° E and depth of around 80 km.

* ΔV is measured in mV and L in meters.

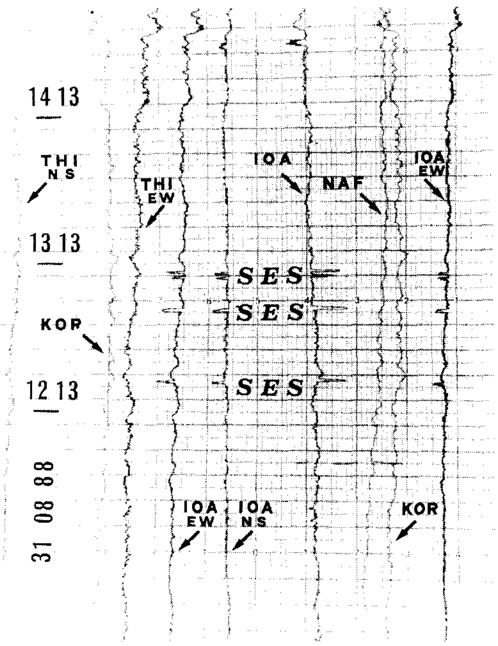

Fig. 7. An example of electrical activity recorded at IOA with 4 SES of the same polarity within 50 min on August 31, 1988 on both dipole arrays (for scales and lengths see Fig. 8). This electrical activity resulted in the telegram shown in Fig. 21. It preceded the catastrophic earthquakes of the Killini area, for which a public warning was issued (see text). The apparently reversed polarity of the SES on the IOA trace is due to the connections of the dipole having been intentionally reversed in order to better distinguish between the recordings of the 8 dipoles at station IOA (see Appendix 2). All SES correspond to an *increase* of both components.

(b) The ρ_{rel} value of a short dipole is not necessarily the same as the ρ_{rel} value of a parallel long dipole.

(c) Two earthquakes of equal magnitude with epicenters at equal distances from the station but from *different* seismic areas do not necessarily give SES with comparable $\Delta V/L$ values when registered by the same long dipole. They are not comparable when the SES of these two earthquakes are recorded by the short dipoles of the station with *different* ratios $(\Delta V/L|_{EW})/(\Delta V/L|_{NS})$; the latter difference will generally reflect different $\Delta V/L$ values on a given long dipole provided that the ρ_{rel} values of the two perpendicular short dipoles have a ratio other than unity (and the long dipole is *not* parallel to any of these perpendicular short dipoles).

Recognition and elimination of noise

Noise obstructing the clarity of SES can generally be classified into three categories, depending on the nature of the cause: electrochemical, magnetotelluric and cultural.

Noise of electrochemical origin

Noise of electrochemical origin, which is generally ascribed to a change in the contact potential between the electrodes and the ground, e.g. due to rain, can easily be recognised when a number of parallel dipoles at every station for each measuring direction are installed (for details see Varotsos and Alexopoulos, 1984a). Furthermore, the effect caused by the daily temperature variation (especially when the electrodes are buried at shallow depth, e.g. 0.5 m) can easily be discerned. The influence of electrochemical disturbances does not seriously affect the measurements when the dipoles have lengths of the order of 10 km as in the case of *grounded* telephone lines (Varotsos and Alexopoulos, 1986).

Magnetotelluric disturbances

These are induced by changes in the earth's magnetic field. As they are (almost) simultaneously recorded at all the stations, they cannot be misinterpreted as SES.

This type of noise can be eliminated by determining for each station the impedance tensor Z which interrelates the variations in the earth's magnetic (H) and electric (E) fields (Varotsos and Alexopoulos, 1984a). Once this tensor has been determined the elimination of the magnetotelluric noise is easily achieved by calculating the difference: $E - Z \times H$ (which should be equal to zero for both measuring directions when SES are absent).

The determination of Z for each station leads to the calculation of the so-called "apparent resistivities" ρ^a_{NS} and ρ^a_{EW} as a function of frequency (and hence as a function of depth). The following comments concerning these apparent resistivities may be useful:

(a) The ratio of the two apparent resistivities is

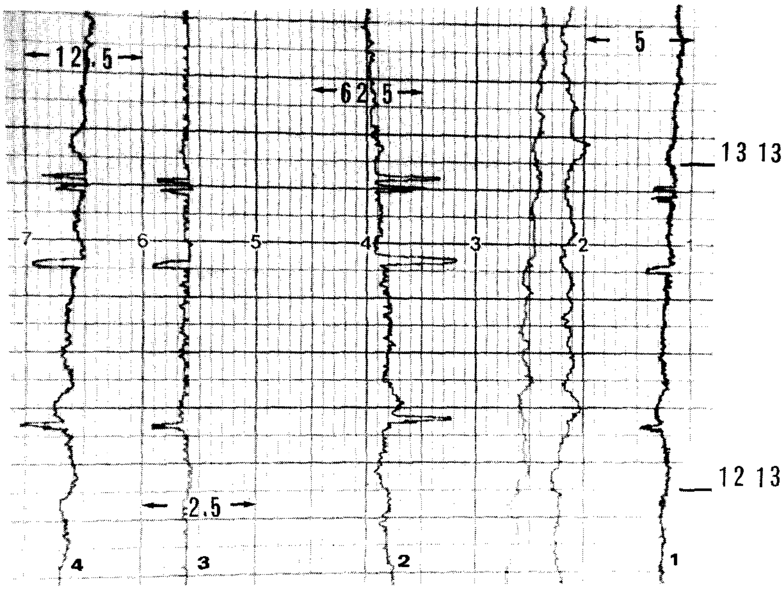

Fig. 8. Magnified portion of Fig. 7. The corresponding directions and lengths of dipoles 1, 2, 3 and 4 are as follows (the scales in the figure are in mV). 1: EW, $L = 47.5$ m; 2: NNE, $L = 2.5$ km (see Fig. 25); 3: NS, $L = 48$ M; 4: EW, $L = 181$ m. These SES were also recorded by four other dipoles with the following directions and lengths: NS, $L = 100$ m, NS, $L = 184$ m, EW, $L = 48$ m and EW, $L = 100$ m (see also Appendix 2).

not equal to the ratio of the two SES- "relative effective resistivities" $\rho_{rel,EW}$ and $\rho_{rel,NS}$ which are empirically determined for each station (Varotsos and Alexopoulos, 1984a, b).

(b) For most of our stations, the ratio of the two apparent resistivities is appreciably different from unity. If periods of the order of 10 s to 1 min are studied (Lazaridou-Varotsou and Papanikolaou, 1987), it seems that the geographical distribution of the stations falls into three zones distinguished by different ρ^a_{EW}/ρ^a_{NS} ratios. For example, the stations lying along the west coast of

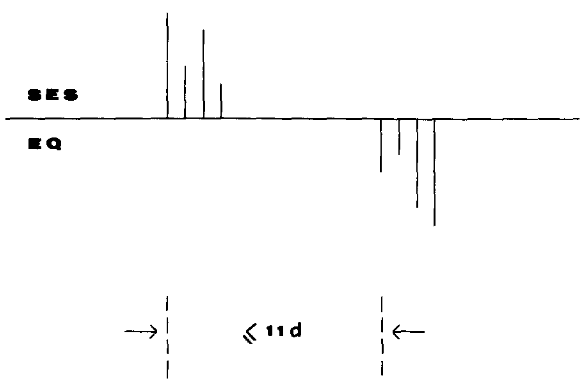

Fig. 9. Diagrammatic representation of a sequence of SES followed by a sequence of earthquakes. Note that the time lag between the initiation of the electrical and the initiation of the seismic activity is around 11 days, while the time lag between the largest SES and the strongest earthquake may exceed this. (For an explanation see Appendix 1.)

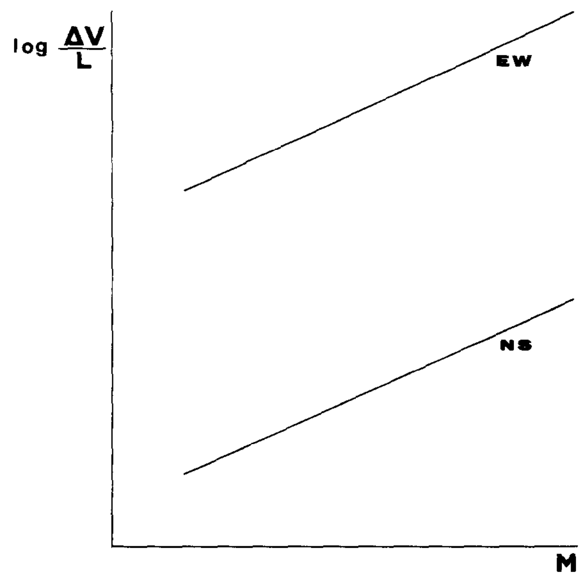

Fig. 10. Relation between the amplitudes of the two components of a SES versus the earthquake magnitude (M_s). The data correspond to SES recorded at the same station and emitted from the same seismic area.

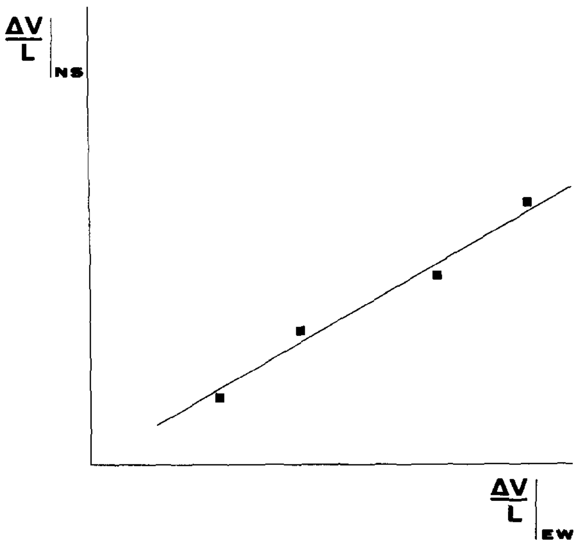

Fig. 11. Typical example of the relation between the two components of SES collected at the same station and emitted from the same seismic area for various earthquake magnitudes (M_s). Note their constant ratio.

Greece have $\rho^a_{EW} > \rho^a_{NS}$ in contrast to stations in the central part of the country where $\rho^a_{NS} > \rho^a_{EW}$. A more thorough examination of the magnetotelluric data indicates that although $\rho^a_{EW} > \rho^a_{NS}$ along the west coast, the ratio ρ^a_{EW}/ρ^a_{NS} is not constant. It seems that the direction of maximum resistivity is perpendicular to the Afro-Asian trough (after having excluded the so-called coast effect). This observation, however, needs further confirmation by studying a larger number of stations along the west coast.

Discrimination between cultural noise and true electrotelluric disturbances

Cultural noise arises from electrical installations, e.g. industrial plants, high-voltage lines, electric trains, etc. Noise from a d.c. electric train can be detected from a distance of around 15 km. Although there is no general technique to avoid this type of noise (except, of course, by installing stations far enough from noise sources) some suggestions are made that might, at least partially, eliminate such interference.

Assume that the noise source N is close to a short dipole E_1W_1. In general such proximal noise can be easily recognized because (in a homogeneous area) the cultural signal voltage ΔV does not generate the same field strength $\Delta V/L$ in neighbouring dipoles with the same orientation but of different lengths. After the signal has been identified as being due to a proximal source it can be distinguished from SES by the following procedure: If the noise source N lies in the vicinity, but not between the electrodes, of a dipole E_1W_1 (Fig. 12a) a long dipole $E'W'$ several km long is installed so that $E'N \gg NW'$ (e.g. $E'N = 5$ km and $NW' = 100$ m). It is evident that a disturbance emitted from the noise source N will be recorded

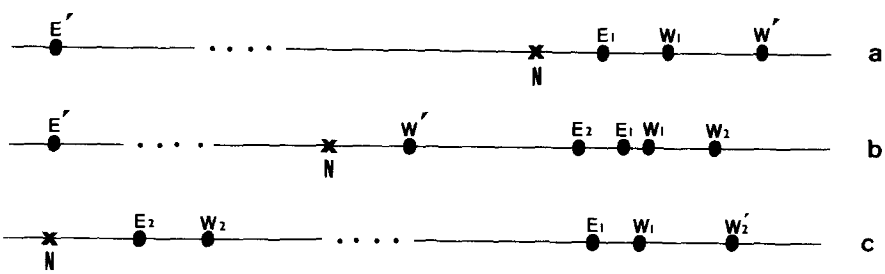

Fig. 12. Disposition of electrodes of short dipoles and long dipoles that allows the recognition of signals originating from a noise source N, for different positions of N. (a) N lies very close to a short dipole E_1W_1. The noise signals detected by the long dipole $E'W'$ have a polarity opposite to that of the short dipole E_1W_1. (b) N lies at distances of the order of a few km from the short dipoles E_1W_1 and E_2W_2. The short dipoles detect noise signals that obey the rule $\Delta V/L \approx$ constant but exhibit polarities opposite to those recorded by the long dipole $E'W'$ or $E'W_2$. (c) N lies at distances of the order of 10 km from the short dipole E_1W_1. The installation of a short dipole E_2W_2 very close to the noise source N (or of a long dipole E_2W_2' with the electrode E_2 close to N) may allow permanent elimination of the noise signal (see text).

by the dipoles E_1W_1 and $E'W'$ with appreciably different amplitudes and with *opposite* polarities. On the other hand, SES (or magnetotelluric disturbances) will be recorded by both of these dipoles with the *same* polarity since their source is assumed to be more distant and on the same side for all the measuring points.

The same method of recognizing cultural noise can in principle also be applied to cases where the noise source lies at distances of the order of some km from the short dipoles (Fig. 12b). Then, when installing a long dipole, the electrodes $E'W'$ are placed so that a disturbance emitted from N will be recorded by the long dipole and the parallel short dipoles (e.g. E_1W_1 and E_2W_2) with opposite polarities. In such cases, the noise source generates signals at short dipoles that obey the rule $\Delta V/L$ independent of the dipole length, L. Therefore, when short dipoles (for a homogeneous area) are found to obey the rule $\Delta V/L =$ constant, two possibilities exist for the interpretation of a voltage variation: either it is due to a noise source located at a distance appreciably larger than the length of the short dipoles, or it is an SES. (As already mentioned magnetotelluric disturbances are identified in a different way.) We conclude therefore that a combination of long *and* short dipoles, as indicated in Fig. 12, is *absolutely necessary* for the discrimination of SES from cultural noise. Otherwise a correlation between SES and EQ cannot be achieved.

Elimination of cultural noise

When signals from a noise source N appear only for short periods of time (e.g. once or twice a day with a duration of a couple of minutes) the techniques described above can be of immediate use in identifying them as noise and distinguishing them from SES. But when a noise source operates continuously it becomes necessary to eliminate it permanently from the records. In some cases, this can be achieved by assuming that a short dipole E_1W_1 (Fig. 12c) has been installed some kilometers away from a known noise source N. Another dipole is installed with one extremity, say E_2, close to N; the other end is installed either very close to E_2 (at W_2), or close to the dipole E_1W_1 (at W_2'). With such an arrangement the disturbances ΔV_2 recorded by the dipole E_2W_2 (or E_2W_2') will exceed the corresponding simultaneous disturbances ΔV_1 at the dipole E_1W_1 by one or more orders of magnitude. In other words, the dipole E_2W_2 (or E_2W_2') is used to record the same type of noise recorded simultaneously with the dipole E_1W_1, but greatly magnified. Therefore, for disturbances from the noise source N:

$$\frac{\Delta V_2}{L_2} \gg \frac{\Delta V_1}{L_1}$$

On the other hand, variations due to SES, from a presumably distant epicenter, give:

$$\frac{\Delta V_2}{L_2} \approx \frac{\Delta V_1}{L_1}$$

Thus, when $\lambda = (\Delta V_2/L_2)/(\Delta V_1/L_1)$ is much larger than unity, the signal is due to noise (as-

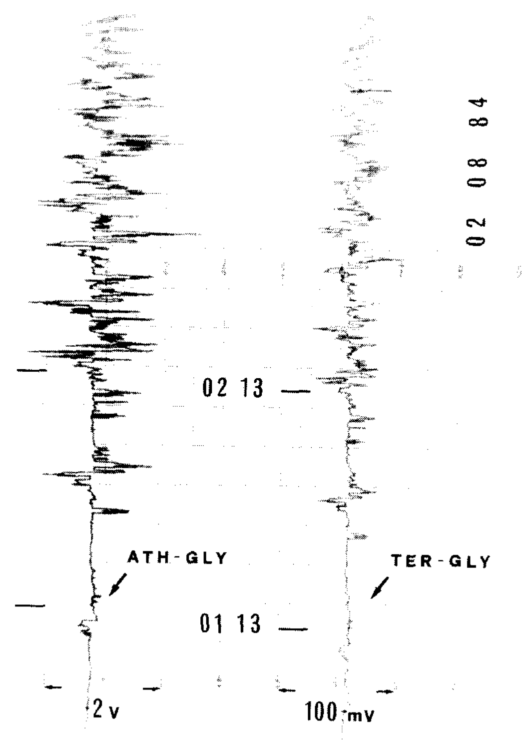

Fig. 13. Example of the decrease of cultural noise with increasing distance from the noise source. The electrode ATH is in the center of Athens while electrodes TER and GLY lie on a straight line with ATH so that ATH–TER = 9 km and TER–GLY = 1 km. Note that the noise ΔV variations at dipoles ATH–GLY and TER–GLY have the same form with a ratio of around 40, although their lengths differ by a factor of 10. It is evident that at night the noise, as expected, is quite low.

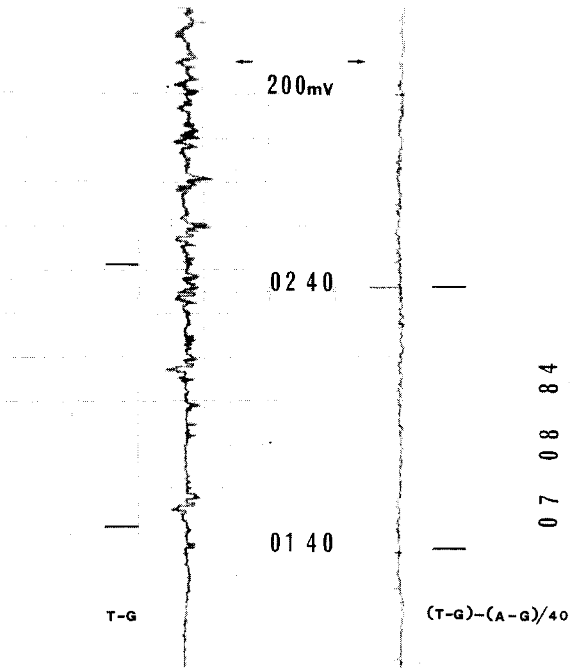

Fig. 14. Elimination of noise of the dipole T–G (i.e., TER–GLY) by subtracting from the records the quantity (A–G)/40, i.e. (ATH–GLY)/40 (compare with Fig. 13).

suming that the two dipoles have comparable resistivities). In practice, λ is not found to be exactly constant, but to depend slightly on frequency. The "transfer function" relating the noise signals of E_2W_2 and E_1W_1 can be readily determined through Fourier analysis and by evaluating the corresponding coefficients for various frequency bands.

Since the duration of SES is usually appreciably larger than 2 s, we can filter out signals of shorter duration and assume that for the remaining frequencies the factor λ is approximately constant. The value of λ can be determined by comparing the amplitudes of ΔV_2 and ΔV_1 measured over a time interval much larger than the period T up to which the noise signal is significant. Once this factor is known, any single voltage variation (SES or noise) that is recorded as ΔV_1 and ΔV_2 can be used to calculate the value:

$$\Delta V_1 - \frac{1}{\lambda} \times \frac{L_1}{L_2} \times \Delta V_2$$

By recording the above difference (instead of ΔV_1) the noise coming from N is eliminated. For an example see Figs. 13 and 14. Note that the information in the SES has not been lost since for the SES we have $\Delta V_1/L_1 \approx \Delta V_2/L_2$, provided that the corresponding resistivities are equal and that the distance of the impending earthquake is appreciably larger than the lengths (L) of the dipoles. The true amplitude of the SES will now be given by $(\Delta V_1/L_1) \times (1 - 1/\lambda)$. If $\lambda \gg 1$, this expression gives the approximate amplitude of the SES at the dipole E_1W_1 *free from noise*.

Selectivity

Regional characteristics of selectivity

Selectivity is defined as the sensitivity of a station to signals from a *restricted number* of seismic areas while remaining insensitive to SES from other areas which may be closer by. For instance, station S_A, located at A, can record SES from a given seismic area B (of course, for earthquakes with the *same* source characteristics and above a certain magnitude) but cannot detect those (for earthquakes of comparable magnitude) from another seismic area C, even when the epicentral distance r_{AC} is appreciably smaller than the distance r_{AB}. As a result, SES emitted from certain seismic areas cannot be "felt" by some stations in the network, irrespective of the earthquake magnitude and epicentral distances.

That the above phenomenon cannot be attributed solely to one of the following three factors: "travel path", "source characteristics" or "station characteristics", is demonstrated by the following empirical observations:

(a) *Selectivity is not reversible.* If selectivity were only governed by the conditions along the main travel path between A and B, this would imply that when a station S_A is sensitive to an area B, a station at B should also be sensitive to the seismic area A. However, this is not always observed.

(b) *Two stations and one seismic area in straight line.* If two stations, S_A and S_B, lie along a straight line from seismic area C, they do not necessarily have the same selectivity as for C. An example is provided by the stations KER and NAF which lie almost on the same straight line with the epicenter of the destructive earthquake that occurred close to Kalamata (KAL) city on September 13, 1986. In this case, although the epicentral distance KAL–KER is roughly twice that of NAF–KAL, the SES was clearly recorded at KER but not at all at NAF (for details of the SES of the Kalamata earthquake see Varotsos and Alexopoulos, 1987).

The above example indicates that the selectivity phenomenon is not purely a directional phenomenon in the sense that for each seismic area the SES can be detected at some specific azimuths only.

(c) *Two seismic areas and one station in straight line.* If two seismic areas C and D lie along a straight line with a recording station S_A, it may happen that this station is sensitive to SES from area D but not to those from area C, even if the epicentral distance r_{AC} is significantly smaller than the distance r_{AD}. Even in the case where a station S_A is sensitive to both seismic areas D and C, the corresponding SES may not have the same $\Delta V_{\mathrm{EW}}/\Delta V_{\mathrm{NS}}$ ratio (for the same dipole length). These ratios have been found to have quite different values, even when the distance between the two seismic areas is appreciably smaller than the distance of their epicenters from the station S_A (for an example see the case described in Appendix 2).

The above observation indicates that in addition to the properties of the main path between a station and seismic areas (and geological inhomogeneities in the vicinity of the station) some other characteristics of the emitting sources must be important. For instance, the directional properties of the current emitted (which might be related to the earthquake mechanism) could also play a significant role in governing the occurrence of SES at remote points.

Observations (a), (b) and (c) suggest that *selectivity depends simultaneously* on: (i) physical properties(e.g. conductivity) of the main path between the station and the seismic area; (ii) source properties (e.g. the mechanism at the focus of the earthquake from which the directional properties of the emitted signal might result) and (iii) the geological structure in the vicinity of the recording station (e.g. inhomogeneities that produce different resistivities along different azimuths from the station).

In spite of its complexity the selectivity phenomenon is *reproducible in time and space* in the following sense: once a station S_A has been identified as being selective to one earthquake from a seismic area B, then this station will also be selective to all future earthquakes (with the same source characteristics) from B (above a certain magnitude of course). This reproducibility in time suggests that the physical properties of the path which influence the selectivity are permanent. As all earthquakes (with the same source characteristics) of *equal magnitude* originating at B give SES of *equal amplitude* at station S_A, it appears that the reproducibility also exists for some aspects of the physical process which leads to "similar electricity changes for similar preparatory stages of earthquakes". This "similarity" refers not only to earthquakes of equal magnitude from B giving equal maximum SES amplitudes at A but also to the *polarity*. However, it does not always apply to the form and duration of the SES. Thus signals coming from a *broad* seismic area and recorded at the same station may in one case have a duration of 2 min starting and ending abruptly and, in another case, a duration of 20 min with gradual onset and finish.

Local characteristics of selectivity

Selectivity seems to be a phenomenon in which both *large-scale* (or "regional") and *small-scale* (or "local") properties of the earth's crust play a significant role. It has been observed that in a strongly inhomogeneous area some sub-areas, with surface dimensions of the order of a few meters only, may provide localities which play quite an important role in the detection of SES. When one or both electrodes of a dipole are located within such a "sub-area", SES are greatly enhanced. In such cases, the length of the dipole is irrelevant; the SES collected with a dipole of a few meters in length have ΔV-values comparable to those collected with a dipole a few kilometers long. The effect can be better described with the aid of Fig. 15 which represents a real case observed at station

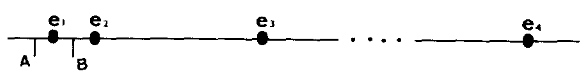

Fig. 15. Arrangement of electrodes near a dyke AB, which acts as an amplifier of SES and magnetotelluric disturbances.

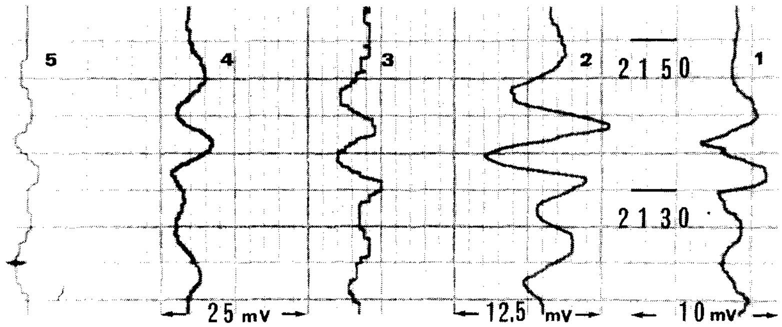

Fig. 16. Magnetotelluric variations registered simultaneously by dipoles 1, 2 and 4 at station KER: *1* = dipole A_3A_1, Scale = 1.00 mV/div; *2* = dipole AA_1, 1.25 mV/div; *4* = dipole AE, 2.50 mV/div. The points A and A_3 lie on a dyke whereas A_1 is just outside it; the lengths of the perpendicular dipoles A_3A_1 and AA_1 are of the order of some meters. The electrode E of the dipole AE ($L = 1.5$ km) is located in marble and the vectors AE and AA_1 have opposite directions. One electrode of dipoles *3* ($L = 100$ m) and *5* ($L = 200$ m) lies on a different dyke from dipoles *1*, *2* and *4* and the scale is 1.0 mV/div.

KER, which is located close to ancient mines in the Lavrion area (see also Figs. 16–18). In Fig. 15 "AB" represents a dyke with an electrode (e_1) located within the dyke; a second electrode (e_2) is installed so that the distance e_1e_2 is of the order of a few meters. Let us now consider three more dipoles e_1e_3, e_1e_4 and e_3e_4 with the same direction and with lengths of the order of a hundred meters. Cases have been observed where an SES was detected by the dipoles e_1e_2, e_1e_3 and e_1e_4 but not by the dipole e_3e_4. Note that in such a case, the SES have practically the same ΔV value at the dipoles e_1e_2, e_1e_3 and e_1e_4. Hence the corresponding $\Delta V/L$ values are completely different, i.e. the $\Delta V/L$ value of the dipole e_1e_2 is found to be several orders of magnitude larger than the corresponding values of e_1e_3 and e_1e_4 (see Fig. 18). In other words, the inhomogeneity "AB" acts as an "amplifier"; an

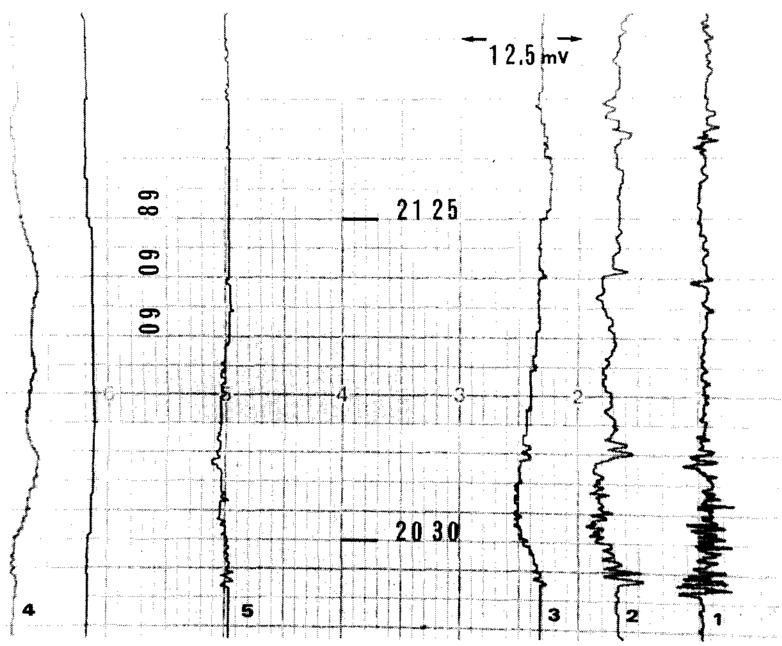

Fig. 17. Magnetotelluric disturbances with a different frequency content recorded at the same dipoles as Fig. 16. (They were also recorded at a vertical dipole installed close to a dyke.)

SES-*vertical* component is also observed in this case.

A question arises concerning the origin of the above effect. From a physical point of view, the electric field intensity perpendicular to the contact between two media should show a discontinuity which depends on the ratio of the respective conductivities of the two media. (This was in fact the expectation that motivated the experiment depicted in Fig. 15). It is noticeable that as shown in Figs. 16 and 17, the ΔV-values of magnetotelluric variations exhibit similar behaviour to that mentioned above for SES. However, the data indicate that the "amplification" of the magnetotelluric variations due to a dyke is not identical to the "amplification" of an SES (Fig. 18) and is frequency dependent.

To summarize the above results, we can state that, in a strongly inhomogeneous area, there are "sensitive" localities", i.e. small sub-areas, sometimes with linear dimensions of the order of meters, that amplify the SES signal. This *local* characteristic of the selectivity effect is superimposed onto the *regional* characteristics. Hence, "sensitive" localities can only amplify those signals for which the region (in which they are located) is selective. As an example, the regional characteristics of selectivity mean that the "sensitive" localities found in the KER area preclude the reception of SES from the Killini seismic area (which lies 240 km west of Athens) but permit the reception of SES from the Kalamata seismic area. Two earthquakes of comparable magnitude occurred in the Kalamata area (on September 13, 1986) and in the Killini area (on October 16, 1988); in spite of the fact that the distances "KER–Kalamata" and "KER–Killini" are comparable, the "sensitive" localities of the KER station did not detect any

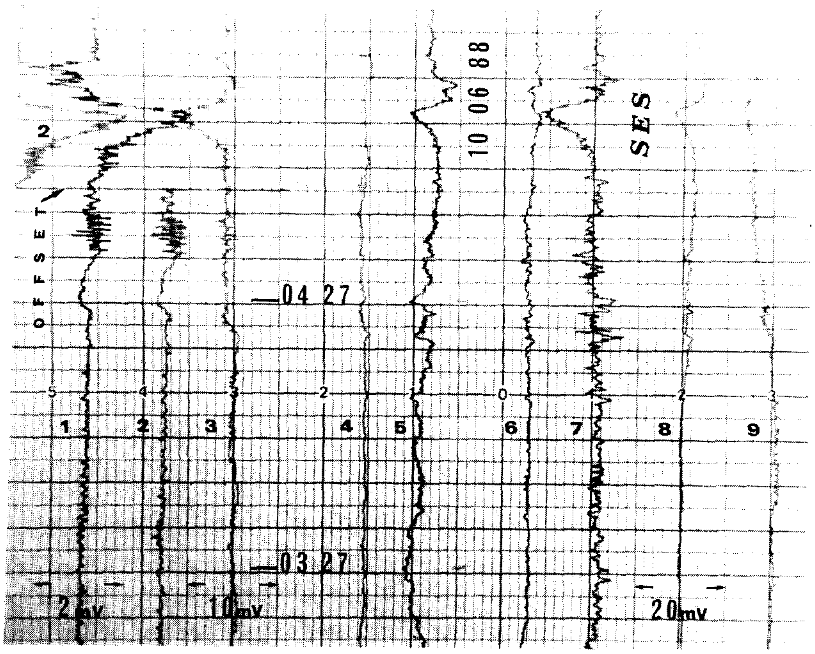

Fig. 18. Detection of an SES by dipoles *1–9* at KER close to and far away from a dyke (with surface dimensions of the order of some meters). Scale: for the traces of dipoles *1* and *2* = 0.2 mV/div., *3* and *7* = 1.0 mV/div; *4*, *5*, *6* and *8* = 2.0 mV/div. Dipoles *1*, *2* and *7* have lengths of some meters and one of their electrodes lies within the dyke, whereas the other is placed just outside the dyke in various directions. Dipoles *5* and *8* have lengths of 1.5 and 1.1 km respectively, with one of their electrodes in the same dyke as *1*, *2* and *7*. Dipole *3* ($L = 200$ m) has one of its electrodes in a different dyke. Both the electrodes of dipoles *9* ($L = 100$ m) and *4* ($L = 1.3$ km) lie outside the two dykes mentioned above. The dipoles *4*, *5*, *7*, *8* and *9* are parallel (almost in N–S) whereas dipole *3* is almost E–W. This SES resulted in telegram number 5 of Table 1; a *vertical* SES-component was also observed.

TABLE 1

Complete list of telegrams issued from May 15, 1988 to August 10, 1989 [1]

	Date of telegram (dd-mm-yy)	Prediction (epicenter–magnitude)	Date of earthquake (dd-mm-yy)	Time of earthquake	Epicenter of earthquake	Earthquake magnitude (M_s)
1	15-05-88	NW330—5.0 (or W300—5.3)	18-05-88	05:17	W310	5.8
2	21-05-88	W300—5.3 (or NW350—5.0)	22-05-88	03:44	W290	5.5
3	30-05-88	W300—5.4 (or NW350—5.0)	02-06-88	10:35	W300	5.0
4	04-06-88	W300—5.0	06-06-88	05:57	W300	5.0
5	10-06-88	SW200—5.1 (or 60 km from ATH—4.7)	13-06-88	20:32	SW240	4.3
6	21-06-88	W300—5.0 (or NW350—4.8)	26.06.88	06:05	W300	4.5–4.7
7	10-07-88	W170—4.7 (or WSW240—5.2)	12-07-88	02:27	NNW95	5.0
8	13-07-88	W70—5.0	16-07-88	01:54	SW100	4.9
9	18-07-88	NNW80—uncertain (or SW100)	23-07-88	09:20	SW200	4.4
10	01-09-88 [2]	W240—5.8 (or NW300—5.3)	22-09-88	12:05	W250	5.1–5.5
11 12	30-09-88 [2,3] 03-10-88 [2]	W240—5.3 (or NW330—5.0) Number of SES from area W 235; attention, activity has not finished.	30-09-88 15-10-88 16-10-88	13:03 07:00 12:34	W215 W235 W240	4.9 4.9 6.0
13	21-10-88	Several tens of km away from W 240—6.3—6.5 (or NW400—5.5)	22-10-88 31-10-88 08-11-88 11-11-88	09:34 03:00 08:18 17:52	W250 W230 SW170 W270	4.9 4.9 5.3 5.0
14	02-03-89	W300—5.4 (or NW330—5.0)	05-03-89 08-03-89	16:44 05:57	NW440 NW470	4.7 4.9
15	03-06-89	W300—5.5 (or NW350—5.0)	07-06-89	19:45	W185	5.2–5.4
16	13-06-89	W200—5.2 (or NW350—4.8)	17-06-89	20:56	W140	4.5
17	23-07-89	NE40—5.0	01-08-89	02:24	N130	5.0

[1] The Table does not contain the M_s 5.8 earthquake (of March 19, 1989 with an epicenter at 39.3° N; 23.6° E) that was missed (see text). It contains all the other earthquakes with $M > 5.2$ within the area 36–41° N, 19–25° E (except two that occurred in Albania).

[2] Attention is drawn to the fact that these telegrams mentioned *sequences of SES* (i.e. case of electrical activity, see Appendix 1 and Figs. 8, 9 and 21). The seismic activity in the Killini area started at 03:39 on September 5, with a 4 mag units event, i.e. the Δt-value between the initiation of the electrical and seismic activities is actually smaller than 11 days (see Fig. 9 and Appendix 2).

[3] For the exact text of this telegram see Fig. 26 and Appendix 2; a displacement of the epicenter by a few tens of kilometers was mentioned.

SES related to the Killini earthquake, although they recorded very intense SES due to the Kalamata earthquake.

Epicenter and magnitude determination

Epicenter determination

In cases when SES have been recorded simultaneously at a number of remote stations, the epicenter can be found by applying the empirical observation that j is proportional to $1/r$ (Varotsos and Alexopoulos, 1984a, b) where j denotes the relative current density of the SES at various epicentral distances r. However, in most of the cases (and especially for the present network in which the average distance between the stations is 150 km) the SES of an impending earthquake is usually observed at only one station. In such cases, (after application of the criteria described in Appendix 2) the epicentral area of the impending earthquake is predicted using the restricted data from this single station. Examples are given in Table 1 (for the period May 15, 1988 to August 10, 1989) where the epicentral coordinates and magnitude are given for the 17 events that were forecast from the SES recorded at a single station.

The epicenter can be determined from the data of a single station by the systematic elimination of possible seismic areas. Let us assume that the SES was recorded at a single station S_1 and no simultaneous SES traces were recorded at any other station $S_2, S_3, \ldots, S_n$. The following effects are then considered in order to exclude certain regions and determine the epicentral area by a process of elimination:

(a) *Selectivity effect.* Using earlier experience, we can exclude as possible epicentral regions all seismic areas which had previously emitted SES which had been recorded at the $S_2, S_3, \ldots, S_n$ stations but had never been observed at S_1. The lack of any SES traces at a number of stations thus plays a prominent, empirical role in the determination of the epicenter. This leaves us with a restricted set of seismic areas which contains only those that had either already produced SES, recorded only at the station S_1 or "new" seismic areas (new to the station S_1). A "new" area is defined as one which has not been active since the installation of the network of SES recording stations.

(b) *Polarity effect.* From the above restricted set of seismic areas it is possible to further exclude those known seismic areas to which station S_1 is sensitive but which emit SES with polarity components opposite to those of the signal recorded.

(c) *Ratio of the two components of the SES.* When the ratio $(\Delta V/L|_{EW})/(\Delta V/L|_{NS})$ of the signal in question is determined it can be compared to the corresponding ratios of SES formerly collected from the seismic areas that remain after exclusions based on (a) and (b). It is clear that the identification of the epicenter is straightforward for seismic regions which have been active in the past so that the above mentioned ratio is known. A difficulty arises when the ratio for the signal under consideration does not coincide with any value that is known. In such a case, a less reliable prediction of the expected epicentral area may be achieved by interpolating values from neighbouring areas.

Magnitude determination

Here, we will restrict ourselves to the case where the SES data are recorded at a single station. After the epicentral area has been predicted the magnitude can be estimated as follows:

Let us assume that data from the recording station S_1 leads to the conclusion that the expected epicenter lies within the seismic area B. Earlier data from the station S_1 for earthquakes in this particular seismic area provide linear plots of $\log(\Delta V/L)$ versus M (see Fig. 10) for the short EW and NS dipoles. These two plots and the corresponding amplitudes $(\Delta V/L)$ of the two components of the new SES provide an estimation of the magnitude of the impending event. Therefore, for an accurate estimate of M for a future earthquake from data collected at a single station, earlier data from the same seismic area is necessary. In other words, a "*calibration*" of a station must be made for *each* seismic area to which it is sensitive. The collection of such data requires long periods of recording. However, since the plots of $\log(\Delta V/L)$ versus M always have a slope between

0.32 and 0.37 a single, clear SES is sometimes enough for the calibration.

As far as the *time* of the expected earthquake is concerned, it should be noted that once the arrival time t_{SES} of a *single* SES has been recorded, the time t_{EQ} of the expected (single) earthquake is already restricted since, as mentioned earlier, the time lag Δt ($= t_{EQ} - t_{SES}$) does not usually exceed 11 days. For the case of an *electrical activity* see Fig. 9.

Recent predictions

In a series of earlier papers (Varotsos and Alexopoulos, 1984a, b; Varotsos, Alexopoulos, Nomicos and Lazaridou, 1986; Varotsos and Alexopoulos, 1987; Varotsos, Alexopoulos, Nomicos and Lazaridou, 1988) several earthquake predictions, based on the interpretation of seismic electric signals (SES), were described. Here a complete list of the predictions issued during the period January 1, 1987 to August 10, 1989 is presented.

Prior to the occurrence of the corresponding earthquakes, the predictions were announced by telegrams sent to the Greek Government. Since May 15, 1988 our forecasts have *also* been sent to scientific institutions in other countries. Each telegram usually contains the following information:

(a) arrival time of the SES;

(b) station(s) at which the SES were recorded;

(c) location of the predicted epicenter of the impending earthquake given by the epicentral distance(s) (in km) and the direction with regard to Athens;

(d) surface wave magnitude, M_S, of the impending earthquake.

Amplitudes of the SES recorded at our stations have been calibrated (Varotsos and Alexopoulos, 1984a, b) with respect to the M_S values reported by the Seismological Institute of the National Observatory of Athens (SI-NOA). In cases when the M_S values were not announced by the SI-NOA they are estimated by means of the approximate formula:

$$M_S \approx M_L + 0.5$$

where M_L denotes the local magnitude given in the Preliminary Seismological Bulletin (PSB) of SI-NOA. When the magnitude M_D (i.e. that obtained from the "duration" of the earthquake) differs significantly from M_L we also make an estimation based on $M_S = M_D + 0.5$. In such cases, two M_S values are given in the tables accompanying the present paper.

Predictions for the period May 15, 1988 to August 10, 1989

Table 1 lists, in chronological order, all telegrams issued during the period May 15, 1988 to August 10, 1989. The values given in parenthesis refer to alternative solutions mentioned in the telegrams. In these cases the data were not sufficient for a unique solution. As already mentioned, the telegrams were sent (prior to the occurrence of the earthquake of course) not only to the Greek Government but also to: (a) Earthquake Research Institute, University of Tokyo (Professor S. Uyeda); (b) Commissariat à l'Energie Atomique, Laboratoire de Détection et de Géophysique (Dr. B. Massinon); (c) Professor Haroun Tazieff (former Minister for Major Disasters in France); and (d) Dr. J. Labeyrie (Paris, France).

Telegrams listed in Table 1, can be classified into four categories according to the actual epicentral area (Fig. 19):

(a) Kefallonia area; almost 300 km west of Athens. Five telegrams of May 15, May 21, May 30, June 4 and June 21, 1988 predicted this activity.

(b) Killini–Vartholomio area; 230–250 km west of Athens. Three telegrams issued on September 1, September 30 and October 3, 1988 relate to a predicted epicentral area between the northwestern coast of the Peloponnesus and Zakynthos.

(c) Patras area, 160–200 km west of Athens. Two telegrams, on June 3 (Fig. 20) and June 12, 1989 were issued.

(d) Other seismic areas. Seven telegrams (June 10, 1988; July 10, 1988; July 13, 1988; July 18, 1988; October 21, 1988; March 2, 1989 and July 23, 1989) were issued.

The above four categories are discussed separately below. The first three (Kefallonia, Killini–Vartholomio, Patras) refer to events that resulted in damage. For the most destructive events

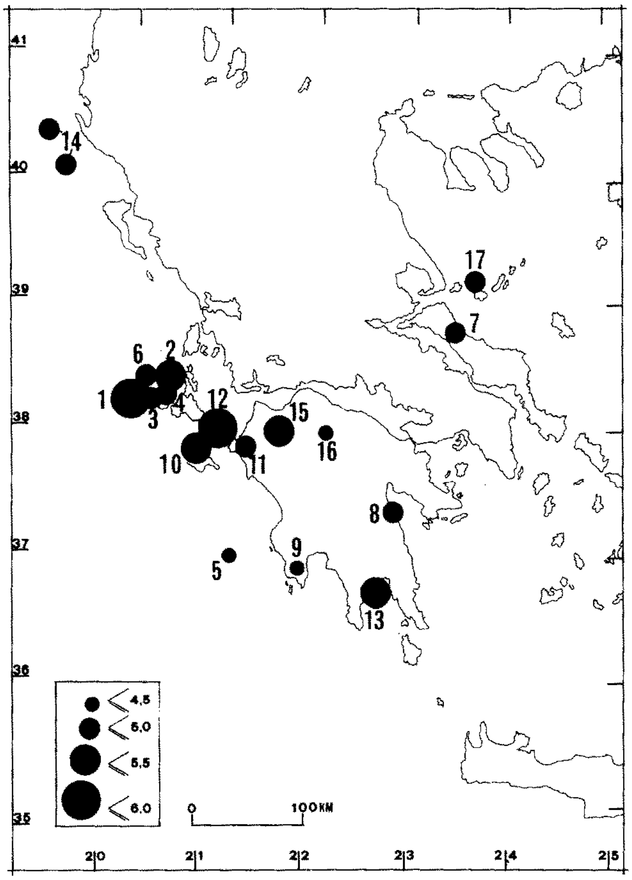

Fig. 19. Epicenters of the earthquakes that correspond to the 17 predictions issued during the period May 15, 1988–August 10, 1989 (Table 1).

(Killini–Vartholomio), *public warnings* were issued.

(a) Kefallonia area

An earthquake, $M_S = 5.8$, occurred on May 18, 1988 and caused minor damage in Akarnania and Aitolia provinces. It was felt as far away as Lecce, Italy. This event was isolated in time and space if one considers that no other earthquake with $M_S > 5.6$ had occurred within the area 35–42° N and 19–28° E (i.e. roughly 700 × 700 km) during the previous 14 months.

Confirmation of the prediction of this event was made by scientists participating in the Conference on Nuclear Test Ban Verification (Linköping, Sweden, May 17–19, 1988). The authors, also participating in this meeting, presented a copy of the telegram addressed to the Greek Government and issued at 18:38 GMT on May 15, 1988 (see Table 1). On May 18, 1988, the Conference participants witnessed, in the conference room, the occurrence of the predicted earthquake. An on-line connection between the conference room and the Swedish seismological array station (Hagfors) and also the German array station (Grafenberg) had been made so that the participants were able to witness the recording of the earthquake.

The $M_S = 5.8$ event of May 18, 1988 was followed by a number of weaker shocks, three of which reached a M_S between 5.0 and 5.5. These events occurred on May 22, June 2 and June 6. These had been preceded by telegrams sent on May 21, May 30 and June 4, respectively.

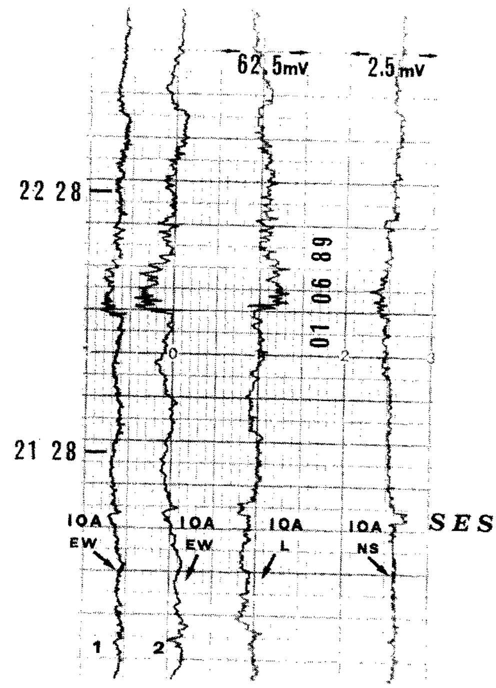

Fig. 20. The SES that gave rise to telegram number 15 in Table 1. The SES is clearly recorded by the short NS dipole ($L = 48$ m) and by the long dipole IOA,L (with $L = 2.5$ km) with a NNE direction; for the reversal of its polarity see Fig. 8 and Appendix 2. Traces of the SES could be also recognized on the two short EW dipoles; IOA EW[1] (with $L = 47.5$ m, scale 0.50 mV/div) and IOA EW[2] (with $L = 181$ m, scale 1.25 mV/div. The date and time indicated correspond to GMT. In view of the difference of 3 h between local time and GMT the "local date" was June 2, 1989; the telegram was issued on June 3, 1989.

(b) Killini and Vartholomio areas

On September 22, 1988 an earthquake with $M_S = 5.1$ (to 5.5) occurred between Zakynthos island and the northwestern coast of the Peloponnesus, in the Killini area, i.e. 250 km west of Athens (around 37.9° N; 20.9° E), and caused significant destruction. Smaller shocks occurred during the following weeks and on October 16, a $M_S = 6.0$ earthquake destroyed many hundreds of houses, mainly at Vartholomio village. This strong event occurred at 37.9° N; 21.0° E, i.e. 10 km away from the previous site (240 km west of Athens).

A series of strong SES, beginning on August 31, 1988 were recorded by both the perpendicular dipole assemblies of station IOA (Figs. 7 and 8). The interpretation of these signals was forwarded to the Greek Government as well as to our Japanese and French colleagues by a telegram on September 1 (Fig. 21) announcing that a $M_S = 5.8$ earthquake was anticipated with an epicenter 240 km west of Athens. In the same telegram an alternative possibility, i.e. a $M_S = 5.3$ shock 300 km NW of Athens was given. After the occurrence of the earthquake on Sept. 22 a new series of SES was recorded which obliged us to issue two more

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ΥΠΕΧΩΔΕ ΑΘΗΝΑ

SIGNIFICANT ELECTRICAL ACTIVITY WAS RECORDED AT IOA - STATION
ON AUGUST 31 1988 EPICENTER AT N.W 300 OR W 240 WITH MAGNITUTES
5,3 AND 5,8
PROFESSOR P.VAROTSOS

1-9-1988

Fig. 21. Copy of the telegram addressed to the Greek Government at 11:30 local time on September 1, 1988. The phrase "significant electrical activity..." was occasioned by the four SES depicted in Figs. 7 and 8. The same telegram was also sent to scientists in France and Japan.

telegrams on September 30 and October 3 drawing the attention of the authorities to the prediction that the seismic activity should continue with magnitudes around 5.3 in an area lying within a few tens of kilometers away from the epicenter of the previous earthquake.

As already mentioned, the above telegrams were also sent to scientists in Japan and France. After the receipt of our first telegram and based on his personal experience, Professor H. Tazieff *publicly announced* on September 3 through A.F.P. and Antenne 2 of the French television that a destructive earthquake was expected in the western part of Greece. After the first destructive earthquake of September 22 a public declaration made by Greek seismologists claimed that *no further destructive events were expected in the area*. Professor Tazieff, on the basis of our telegrams of September 30 and October 3 issued a second public warning on October 5 insisting that the destructive activity would continue; the latter announcement was fulfilled on October 16, 1988 (Tazieff, 1989; pers. commun.) [1]

(c) Patras area

The earthquake that occurred on June 7, 1989 extensively damaged 173 houses, mainly in villages between Patras and Pirgos. The telegram sent to the Government 4 days before the earthquake (June 3, 1989) was based on an SES that was recorded at station IOA (Fig. 20). A second SES, recorded at the same station on June 12 (but announced on June 13) was followed by a $M_S = 4.5$ event on June 17 in the same area.

(d) Other seismic areas

Among the telegrams that refer to other seismic areas, the one of October 21, 1988 showed an unusual degree of error from the actual earthquake parameters (at least 1.0 M_s unit and 120 km for the epicenter location). For this case we list in Table 1 all events with $M_S \geqslant 4.9$ that followed this telegram within 20 days. The prediction was based on a series of SES that had been recorded on the two (NS and EW) dipole assemblies at IOA station; as explicitly stressed in this telegram, the ratio of the two components was *different* from that observed for the SES that motivated the telegrams issued on September 1 and 30, and October 3, 1988. This difference in the ratio was interpreted by our group as an indication of a *displacement* of the epicentral area from the previously active zone (i.e. 215–250 km west of Athens) by several tens of kilometers. If this interpretation is correct this telegram should not be correlated with the events that occurred on October 22 and 31 and November 11, 1988 with epicenters in the Killini area. These SES might, instead, be correlated with the $M_s = 5.3$ event that occurred on November 8 and had an epicenter 150 km away from the previously active area. In this case, however, the discrepancy between the predicted magnitude and that announced by SI-NOA remains unusually large.

Evaluation of the results of the period May 15, 1988 to August 10, 1989

The above results lead to the following conclusions (see also Fig. 19):

(a) Two predictions (out of seventeen, see Table 1) can be considered as erroneous (telegrams of July 10, 1988; October 21, 1988). Although these two predictions were followed by earthquakes, the deviation between the predicted and the true parameters was too large.

(b) Fifteen (or fourteen if one disregards telegram number 9) predictions out of seventeen showed differences between predicted and true magnitudes of between zero and 0.7 M_s units; discrepancies in the distance between the predicted and the true epicenters were of the order of 100 km or less.

(c) If one restricts the inspection to earthquakes with $M_S > 5.6$, three events occurred within the area 36–41° N, 19–25° E (about 500 × 600 km)

[1] Professor Tazieff decided to make these public announcements, firstly because earthquakes with $M > 5.5$ are potentially destructive and lethal; secondly, because some Greek (as well as Italian) seismologists had too frequently stated that forecasts made by the present method had been delivered after the earthquake and, thirdly, because the Greek organization in charge (i.e. OASP) had never told the Greek population that some destructive earthquakes that had happened in the country had been officially forecast by the method described in this paper.

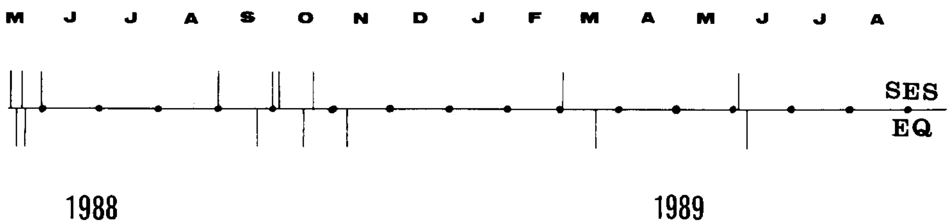

Fig. 22. Time chart for all the predictions issued for expected seismic events of $M_s \geqslant 5.3$ and all earthquakes (EQ) with $M_s \geqslant 5.3$ during the period May 15, 1988–August 10, 1989. The data are given in Table 1.

during the period under consideration. For one of them, with $M_S = 5.8$ (March 19, 1989 with an epicenter at 39.3° N, 23.6° E, lying outside our network) no telegram was issued since the authors were out of the country at that time.

For the other two earthquakes, with relatively high magnitudes of $M_S = 5.8$ and 6.0 (May 18, 1988 and October 16, 1988 respectively), accurately specified telegrams had been issued. As already mentioned in one of these cases (May 18, 1988) our prediction was confirmed during an international conference (Linköping, Sweden), while advance *public warnings* were issued *internationally* in the case of the most destructive seismic activity (Killini and Vartholomio area on September 22 and October 16, 1988).

(d) In evaluating the results of earthquake prediction * two ratios are usually employed: the "success rate" and the "alarm rate":

$$\text{success rate} = \frac{\text{total number of "successful" predictions}}{\text{total number of issued predictions}}$$

$$\text{alarm rate} = \frac{\text{total number of "successful" predictions}}{\text{total number of earthquakes}}$$

By considering the numbers given in the second conclusion (and remembering that predictions are issued only for earthquakes with predicted values of $M \approx 5.0$ or larger) we find:

$$\text{success rate} = \frac{15 \text{ (or 14)}}{17}$$

* Two types of probabilities p_1 and p_2 should be considered: p_1 is the probability that a prediction will be successful and p_2 is the probability that an earthquake will be predicted (Hamada, 1991).

It is evident that the value of the "alarm rate" depends on: (1) the threshold value chosen for M_s and (2) the area considered. For example, if we consider the total number of earthquakes with $M_s \geqslant 5.3$ that occurred within the area 36–41° N, 19–25° E we find:

alarm rate

$$[M \geqslant 5.3,\ \Delta r \leqslant 120 \text{ km},\ \Delta M \leqslant 0.7\text{-units}] = \frac{5}{7}$$

In this last calculation the following seven events have been considered: May 18, 1988; May 22, 1988; September 22, 1988; October 16, 1988; November 8, 1988; March 19, 1989 and June 7, 1989. From these, one (March 19, 1989) was not predicted, one (Nov. 8, 1988) was predicted with considerable inaccuracy, while five were predicted with an accuracy of $\Delta M \leqslant 0.7$ units and $\Delta r \leqslant 120$ km, (for four of these five events an accuracy of $\Delta r \leqslant 30$ km was achieved: for one $\Delta r = 120$ km).

(e) The degree of correlation between SES and earthquakes can be best visualized when one selects events with large magnitude e.g. $M_s \geqslant 5.3$. In Fig. 22 a time chart is given that depicts *all* the predictions of $M_s \geqslant 5.3$ issued, together with all the $M_s \geqslant 5.3$ earthquakes that actually occurred within the area 36–41° N, 19–25° E. The time chart shows a *non-uniform* time distribution of the SES and earthquakes. Two periods might be distinguished: during the first, 6 months long (May 15, 1988 to November 10, 1988), seven predictions ($M \geqslant 5.3$) were issued, whereas only five earthquakes of $M_s \geqslant 5.3$ occurred; on the other hand during the last 9 months (November 10, 1988 to August 10, 1989) seismic activity had significantly decreased, so that only two predictions with $M_s \geqslant 5.3$ were issued while two earthquakes of $M_s \geqslant 5.3$ occurred.

Predictions for the period April 1, 1987 to May 15, 1988

Table 2 lists in chronological order all the predictions issued during the period April 1, 1987 until May 15, 1988. Some comments are appropriate:

(a) The prediction of April 27, 1987 relates to a "gradual variation of the electric field" (GVEF), which appeared at station PIR. This unusual type of variation sometimes precedes the detection of an SES at the same station. It always has the *same polarity* and is detected by the *same dipole array(s)* as the subsequent SES (Varotsos and Alexopoulos, 1986). As mentioned in the same publication, a GVEF usually appears a few weeks before the occurrence of a strong earthquake ($M_s \geqslant 5.5$) and it has an amplitude one order of magnitude larger than that of the corresponding SES. On April 27, 1987, immediately after the detection of the GVEF, an urgent letter was consequently sent officially to the Government (Ministry of Public Works) explaining that this type of signal is usually followed by a strong seismic activity. In fact, 4 weeks after the initiation of this GVEF, an event of $M_s = 5.5$ occurred (on May 29, 1987) in the Pirgos area. Two weeks later, on June 10, 1987, another, $M_s = 5.6$, occurred some tens of kilometers away from the epicenter of the previous shock. The May 29 event had been preceded by an SES registered simultaneously by both the short and the long dipoles of station PIR. This was in contrast with the SES of June 10, 1987 which had been detected only by the long dipoles of the same station.

The second telegram in Table 2 refers to a SES that, again, had been registered only by the long

TABLE 2

Complete list of predictions issued from April 1, 1987 to May 15, 1988 *

Date of prediction (dd-mm-yy)	Prediction	Date of earthquake (dd-mm-yy)	Time of earthquake	Epicenter	Magnitude (M_s)
27-04-87 [1]	50 km from the PIR-station with M_s 5.5	29-05-87	18:40	30 km from PIR	5.5
		10-06-87	14:50	70 km from	5.6
13-06-87	W200—5.2	21-06-87	06:13	WSW240	5.0
01-02-88	NE200—5.0	10-02-88	10:08	ENE287	4.3
		18-02-88	11:11	N130	5.1
10-03-88	NW350—5.0	16-03-88	20:02	NW396	4.6
	(or WNW260—5.0)	26-03-88	20:35	NW438	5.5
02-04-88	W250—5.0	05.04-88	06:24	W200	4.3
	(or SW300—5.5)	08-04-88	05:57	WSW290	4.4
03-04-88	N100—5.0	05-04-88	09:17	ENE144	3.8 False
07-04-88	WNW250—5.0	12-04-88	19:48	W300	4.5
	(or NW360—5.0)	24-04-88	10:10	WSW320	5.0
21-04-88 [2]	40 km from ATH—4.3	23-04-88	10:28	NE95	3.4
28-04-88	W300—5.0 (or WNW300—5.0)	09-05-88	16:52	W350	5.0

[1] Detection of GVEF at station PIR; this type of variation is usually detected a few weeks before the occurrence of strong (i.e. $M \geqslant 5.5$) events.

[2] This is the only telegram that predicts $M < 5.0$ and was sent at the request of the Authorities.

* During the previous period of January 1, 1987 to April, 1987 *only one* prediction was issued on February 26; it announced a $M_s \approx 6.5$ earthquake with an epicenter at W300 and was *actually* followed by a $M_s \approx 5.9$ earthquake on February 27 at W295.

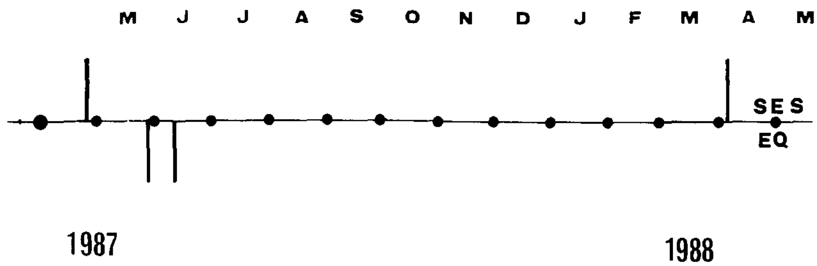

Fig. 23. Time chart for all the predictions issued for expected seismic events of $M_s \geqslant 5.3$ and all earthquakes (EQ) with $M_s \geqslant 5.3$ during the period April 1, 1987–May 15, 1988. The data are given in Table 2. Note the long period of quiescense.

dipoles of station PIR. This telegram was followed, on June 21, 1987 by an earthquake of $M_s = 5.0$, the epicenter of which was close to that of the June 10, 1987 event (according to USGS the epicenter of the event of June 21 was located at 21.32° E; 37.23° N).

(b) Although our telegrams are generally sent *only* for earthquakes predicted as having $M_s \approx 5.0$ or more, one of these (April 21, 1988) concerns the prediction of an event of relatively small magnitude, $M_s = 4.3$. This telegram was sent because of a special request made by the Authorities, motivated by a weak event, with $M_s = 3.7$, that had been felt at Athens on April 17, 1988.

Comments on the results of the period April 1, 1987 to May 15, 1988

In order to compare the results of this period with those of the subsequent period until August 10, 1989, Fig. 23 shows a time chart of all the predictions announcing seismic events with $M_s \geqslant 5.3$ and all the earthquakes with magnitudes above this threshold that occurred. An inspection of this figure shows that:

(a) only two predictions were issued for $M_s \geqslant 5.3$. The first (April 27, 1987), as mentioned above, reported a GVEF at station PIR and was actually followed by the strong seismic activity of May–June 1987 with its epicentral area located close to this station. The other (April 2, 1988) was also followed by an event of smaller magnitude.

(b) A comparison of Figs. 22 and 23 shows a significant difference in the seismic activity during the two periods. It is also evident that predictions and earthquakes are not only grouped together, but also that a long period of seismic quiescence coincides with a long period during which no significant SES were recorded and, consequently, no predictions were issued.

TABLE 3

Summary of the results: Alarm rate for various magnitude thresholds

Period	Magnitude threshold	Total number of earthquakes	Number of missed events	Number of successful predictions		Alarm rate	
				$\Delta M \leqq 0.7$, $\Delta r \leqq 120$	$\Delta M \leqq 0.7$, $\Delta r \leqq 30$	$\Delta M \leqq 0.7$, $\Delta r \leqq 120$	$\Delta M \leqq 0.7$, $\Delta r \leqq 30$
May 15, 1988	$M_s \geqq 5.3$	7	1	5	4	5/7	4/7
to August 10,	$M_s \geqq 5.5$	5 [1]	1	4	4	4/5	4/5
1989	$M_s \geqq 5.8$	3	1	2	2	2/3	2/3
	$M_s \geqq 6.0$	1	0	1	1	1/1	1/1
April 1, 1987	$M_s \geqq 5.5$	7 [1]	1	6	6	6/7	6/7
to August 10,	$M_s \geqq 5.8$	3	1	2	2	2/3	2/3
1989	$M_s \geqq 6.0$	1	0	1	1	1/1	1/1

[1] By including the EQ on Sept. 22 (i.e. the first destructive EQ of the Killini-activity) and excluding, as mentioned, two EQ that occurred in Albania.

The results of the present paper are summarized in Table 3 where the alarm rate is given for various magnitude thresholds. A detailed statistical analysis of the totality of the predictions described in this paper has just been completed by Prof. K. Hamada (1991).

Acknowledgments

The authors would like to express their gratitude to Professor H. Tazieff who has made many useful corrections to the first draft of this paper. Numerous useful comments by Profs S. Uyeda, K. Hamada and O. Kulhanek are gratefully acknowledged.

Appendix 1: Physical aspects of electrical activity

The object of this section is to present some considerations that may explain the empirical observation that a sequence of SES is not necessarily followed by a sequence of earthquakes with magnitudes altering in the same order as the amplitude of the SES. These considerations are based on a model that was presented by Varotsos and Alexopoulos (1986) and found to give plausible explanations for various empirical observations. The model assumes that, during the final stage of the preparation of an earthquake, the stresses, σ, gradually change with a rate b $[=(\mathrm{d}\sigma/\mathrm{d}t)]$. A physical mechanism (i.e. orientation of electric dipoles formed by point defects) has been suggested which gives rise to the emittance of a transient current when σ reaches a certain value (σ_{cr}). The model also assumes that σ_{cr} is smaller than the stress σ_{fr} at which fracturing (i.e. the earthquake) occurs so that a time lag $\Delta t = (\sigma_{fr} - \sigma_{cr})/b$ exists. For the case of a stressed ellipsoidal volume of rock the maximum amplitude of the SES detected at a given site at the surface of the earth is proportional to the cross section S perpendicular to the long axis of the largest ellipsoid. Note that for a stressed spherical volume the plots $\log(\Delta V/L)$ versus M should have a slope equal to unity: this is in contrast to the observed value of 0.32 to 0.37 (Varotsos and Alexopoulos, 1986).

Time lag variations: Consider now a large volume, underlying a seismic area A, throughout which the stresses change gradually. It is to be expected that, in general, the σ value at a given moment will not be the same in the various sub-volumes A_1, A_2, A_3 (see Fig. 24) and hence in each sub-volume the value σ_{cr} is reached at different times t_1, t_2, t_3, etc; it is also assumed that volume A_1 is the largest and the first to emit SES, the amplitude of which is proportional to the surface S_1 (see Fig. 24), i.e. $t_1 < t_2 < t_3$. Therefore, a sequence of SES will be observed with amplitudes similar to those depicted in Fig. 9. It is now evident that σ will not have to reach the value σ_{fr} in all the sub-volumes A_1, A_2, A_3, etc., *simultaneously*. If σ reaches the value σ_{fr} first in volume A_2 and subsequently in A_1, an earthquake of magnitude M_2 will occur first followed later by another of larger magnitude M_1. Of course, due to the redistribution of stresses, the rate of change, b, for volume A_1 will have changed somewhat after the occurrence of the earthquake originating in volume A_2.

Appendix 2: Additional details of the prediction of the Killini–Vartholomio destructive earthquakes, September–October, 1988

The station IOA at which the SES activities corresponding to the above events were recorded comprises eight dipoles as follows:

Four short dipoles in the E–W direction with $L = 47.5$, 48, 100 and 181 m, respectively.

Two dipoles in the N–S direction with $L = 100$ and 184 m, respectively.

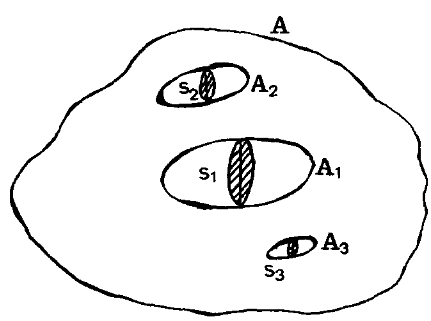

Fig. 24. During the last preparatory stage before a seismic event in area A, the stresses σ_i gradually change. However the critical value σ_{cr} (and the value σ_{fr}) is not reached simultaneously in the various sub-volumes A_1, A_2, A_3 etc., so that the corresponding SES are not emitted at the same time (see text).

One long dipole, $L = 2.5$ km, one electrode of which is very close to the short dipoles of the station, while the other electrode lies close to the village Pérama (see Fig. 25); for reasons of brevity this long dipole is labelled IOA.

One short dipole ($L = 48$ m) which is (almost) parallel to the long one; it is labelled IOA NS.

In measuring the potential difference between the electrodes of the short dipoles the following convention is used: E^+W^-, N^+S^-; and for the long dipole: (Pérama village)$^+$-(station)$^-$. This convention implies that for a true SES signal the *polarity* of ΔV for the long and parallel short dipole *should be opposite*.

In Figs. 2, 7, 8 and 20 the records of the following four dipoles are depicted: (a) Two EW dipoles with $L = 47.5$ and 181 m IOA EW^1, IOA EW^2; (b) the long dipole (IOA); and (c) the short (almost) parallel dipole (IOA NS).

Three periods of SES activity were recorded before the Killini–Vartholomio earthquake: August 31 (Figs. 7 and 8), September 29 and October 3. These show the following features:

(a) The SES activities were recorded at the following times: from 12:25 to 13:10 on August 31; almost from 17:00 to 20:50 on September 29 and from 01:50 to 03:00 on October 3 (all times GMT). These results together with those of many other SES studies clearly show that the "mean value" of the number of SES recorded at night is equal to that recorded during the day.

(b) The SES activities of September 29 and October 3 are strikingly similar; however, when comparing them to the activity recorded on August

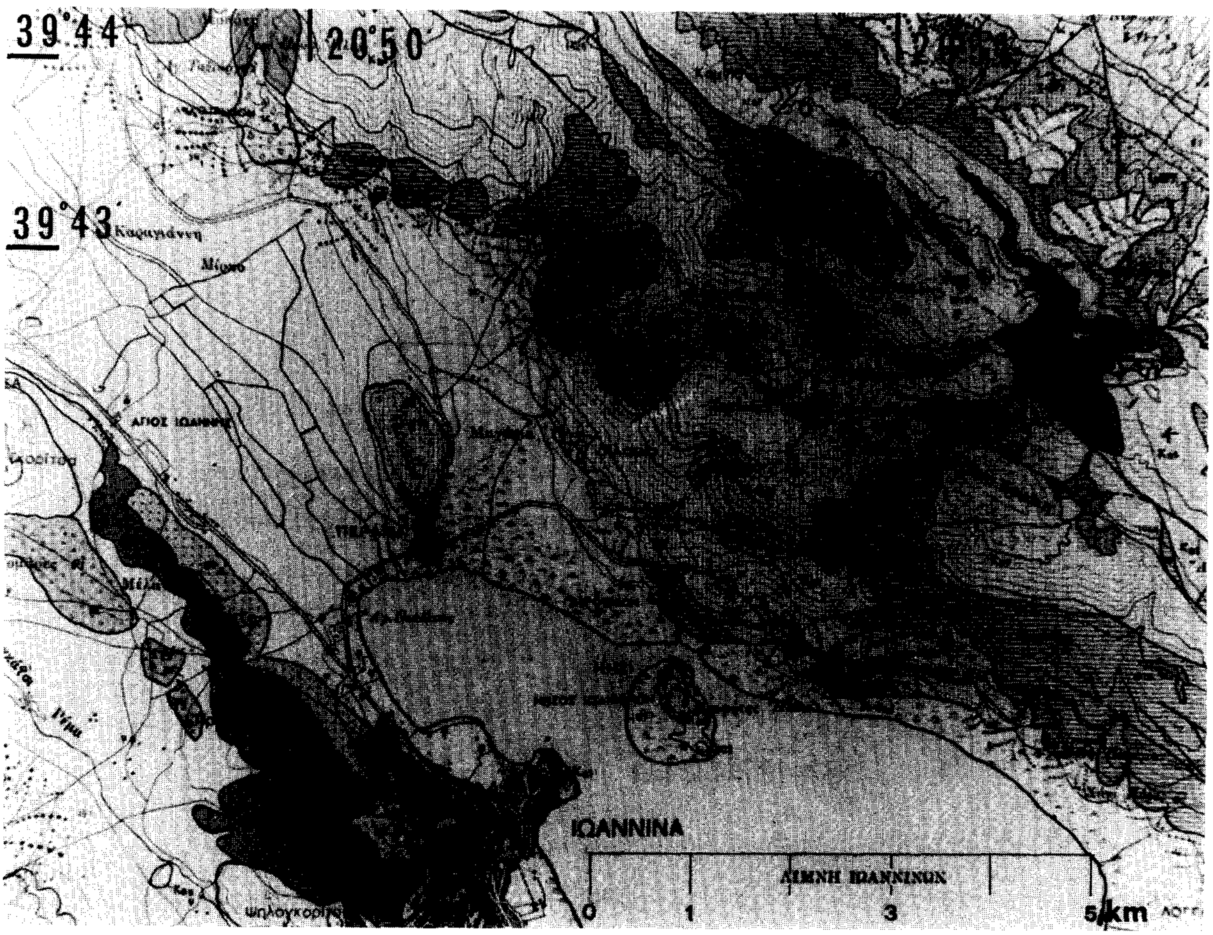

Fig. 25. Map showing the location of station IOA. The two dots indicate the site of the short dipoles of the station and Pérama (ΠΕΡΑΜΑ) village respectively. The two electrodes of the long dipole are placed at the two dots.

31 (Fig. 8), a difference in the ratio of the two components emerges. *This difference* was the reason why in the prediction issued on September 30 (see Fig. 26) it was stated that the expected epicenter would *not* coincide with that of the earthquake on September 22, but that it would be a few tens of kilometers away from the previous one. The two epicenters were in fact different: the September 22 event destroyed the mole of Killini harbour whereas the earthquake of October 16 seriously damaged the village of Vartholomio. It is therefore of interest to study the difference in the mechanisms of these earthquakes.

(c) If we study the time lag between the SES and the earthquakes it should be noted that, though the first destructive earthquake (September 22) occurred 22 days after the first SES activity (August 31, 1988), the seismic activity could be considered to have started *earlier*, since two smaller shocks (with $M_s \approx 4.0$) occurred on September 5 (at 00:39 GTM; epicenter at 37.96°N, 20.9°E) and September 11, 1988 (at 14:57 GMT,

OTE ΟΡΓΑΝΙΣΜΟΣ ΤΗΛΕΠΙΚΟΙΝΩΝΙΩΝ ΤΗΣ ΕΛΛΑΔΑΣ ΑΕ

ΑΝΤΙΓΡΑΦΟ ΤΗΛΕΓΡΑΦΗΜΑΤΟΣ

ΑΘΗΝΩΝ 155 862479 50/47 30/9 13.14
TF 1-64463787
J.LABEYRIE PROFESSOR
H.TAZIEFF
CHEMIN DE LA FEVERIE
91190 GIF/YVETTE (ΓΑΛΛΙΑ)
ELECTRICAL ACTIVITY RECORDED AT IOA. ON SEPTEMBER 29, 1988 STOP
EPICENTER A FEW TENS OF KILOMETERS FAR FROM THE EPICENTER OF
SEPTEMBER 22 1988 OR NW 330 STOP MAGNITUDES 5,3 AND 5,0
RESPECTIVELY
PROFESSOR P.VAROTSOS

5-10-1988

8/0766/11-10.88

Fig. 26. Copy of the telegram issued at 13:14 local time (i.e. 11:14 GMT) on September 30, 1988. It refers to the electrical activity recorded in September 29, 1988.

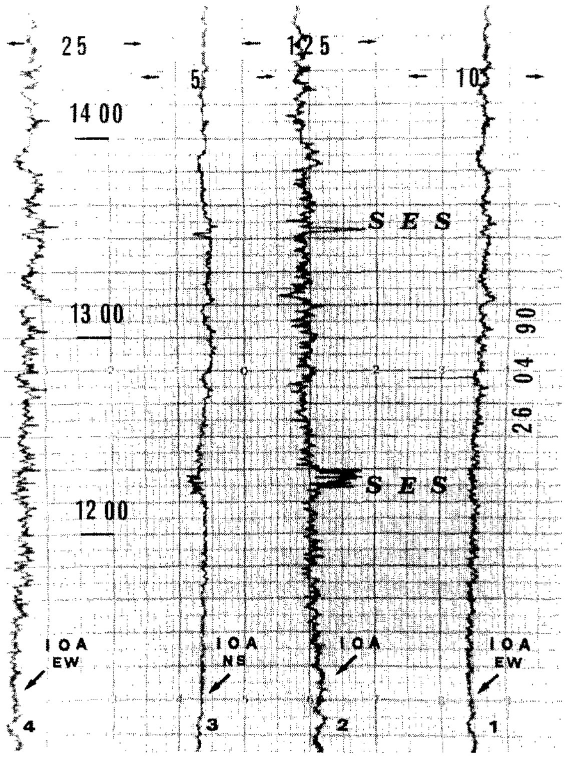

Fig. 27. Electrical activity recorded at station IOA on April 26, 1990. The directions, scales and lengths are the same as in Figs. 7 and 8.

epicenter at 38.1° N, 20.7° E). Therefore, the time lag between the initiation of the SES and seismic activities is actually *less* than 11 days, as indicated in Fig. 9.

Criteria that have to be followed in order to recognise SES at station IOA

An electric disturbance is classified as a SES *after* it has met the following *four* criteria:

(a) it is not recorded at all the stations and, therefore, is not a magnetotelluric effect;

(b) the "$\Delta V/L =$ constant" test should be obeyed for the short dipoles;

(c) the SES appears simultaneously at the short and long dipoles;

(d) the polarity of ΔV recorded by the long dipole has to be opposite to that of its parallel short dipole, as mentioned above.

Attention is drawn to the last criterion and its *decisive* importance.

TABLE 4

Telegrams issued between August 10, 1989 and November 30, 1989

	Date of telegram (dd-mm-yy)	Prediction (epicenter–magnitude)	Date of earthquake (dd-mm-yy)	Time	Epicenter	Magnitude (M_s)
18	16-08-89	WNW200—5.0	20-08-89 24-08-89	18:32 02:13	WSW245 W310	5.9 5.7
19	24-08-89	zone: WNW190—WSW240 $M = 5.2$–5.8	31-08-89	21:29	W170	4.8–5.0
20	11-09-89	activity from the same area as in telegram 19 $M =$ the same as in 19	25-09-89 25-09-89	07:35 07:38	WSW220 WSW235	4.7–5.0 4.6–5.0
21	15-09-89	SES from ASS-THES $M =$ uncertain	19-09-89	07:57	NNW265	5.0
22	18-10-89	NW300—4.8 (or W240—5.5)	29-10-89	19:36	NW280	4.5

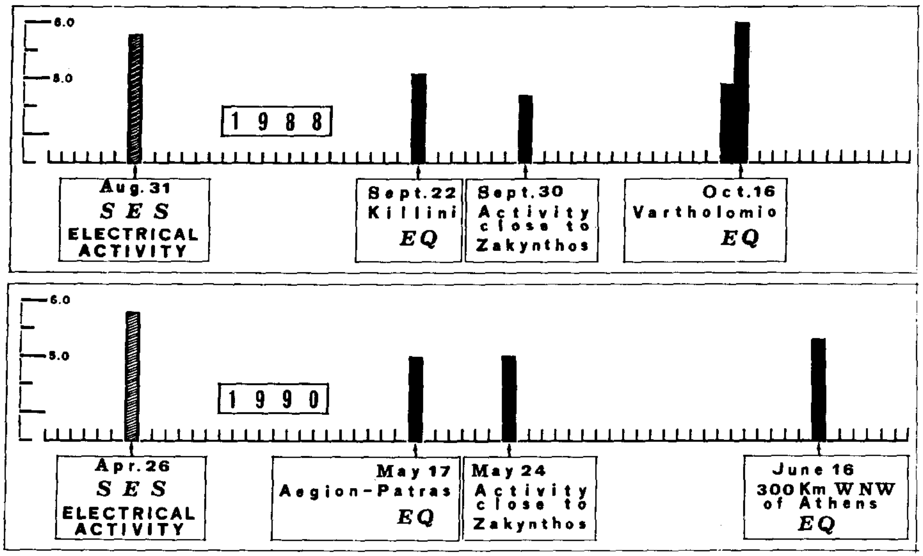

Fig. 28. Comparison of the Killini–Vartholomio destructive seismic activity to the recent one that occurred in 1990 at western Greece. Note that only the corresponding initial electrical activities are depicted.

Appendix 3: Predictions made after completion of this paper; August 10, 1989 to November 30, 1989

Table 4 lists five telegrams which were issued between August 10, 1989 and November 30, 1989. In fact, it is a continuation of Table 1.

Appendix 4: Recent electrical activity

An *electrical* activity was recorded at station IOA on April 26, 1990 (Fig. 27). Comparison with Figs. 7 and 8, shows that the SES recorded at 13:32 GMT on April 26, 1990 is *strikingly similar* to those recorded on August 31, 1988, as far as the form, amplitude and polarity are concerned.

As the ratios $(\Delta V/L|_{EW})/(\Delta V/L|_{NS})$ are slightly *different*, a *displacement* of the epicentral zone by a few tens of kilometers might be expected.

Appendix 5

Twenty-two days after the electrical activity described in Appendix 4 (i.e. on May 17, 1990), a $M_s = 5.0$ earthquake actually occurred with an epicenter 140 km west of Athens causing damage in the area of Patras and Aegion. This represents one of the events of the predicted seismic activity and its time lag equals that of the first destructive Killini event in 1988. The seismic activity could be considered to have started *earlier* since two smaller shocks (with $M_s \approx 4.5$ and 4.0) occurred on May 4 and May 7 with epicenters 280 and 170 km west of Athens. Therefore the time lag between the *initiation* of the SES activity and the *initiation* of the electrical activity is actually *less* than 11 days as in the case of the Killini–Vartholomio destructive sequence in 1988. The latter activity is compared to the recent one in Fig. 28 where all earthquakes with $M_s \geqq 4.9$ have been included. The following points should be emphasized:

(1) The electrical activity of April 26, 1990 resulted in a telegram on April 27, 1990 the content of which is just that described in Appendix 4 (which was added to the present paper *before* the initiation of the seismic activity).

(2) The aforementioned telegram of April 27, 1990 was publicly announced by Prof. H. Tazieff as in the case of the Killini–Vartholomio activity in 1988. It is the *only* telegram issued after the last one of Table 4.

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