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The resolution of the Graefenberg array for earthquake locations in the eastern Mediterranean

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Abstract. One hope of making short-term predictions of upcoming mainshocks lies in the rapid location and identification of foreshocks. An advantage of the Graefenberg digital seismic array (GRF) for this task is that seismic signals are available for analysis at the array in real time. In this paper we ask the question: Is the location accuracy of the Graefenberg array in the Hellenic Arc high enough to allow foreshock recognition? We found that the delay patterns for events located throughout the Hellenic Arc were not sufficiently different to allow locating events using these patterns. We also examined waveform cross-correlation techniques and found the location errors to be too large (200–300 km). These errors are due to complexity in the upper mantle. The best technique for locating the events combines the azimuth estimate from the GRF array with the arrival time difference between *P* and *P*4 (the *P* wave reflected off the 400-km discontinuity). The distance estimate from this travel-time difference has errors of about 60 km. The azimuth estimate has errors of about 3° which yields location errors of about 100 km. These errors are of the order of rupture zone sizes in the Hellenic Arc and, therefore, the Graefenberg array may be useful for approximate monitoring of seismic activity in real time. The techniques involved in determining the locations are, however, difficult for an automated location system.

Key words: Arrays – Earthquake location – Epicentre resolution

Introduction

Several regions of the Hellenic Arc south of Greece are presently experiencing seismic quiescence. Such quiescence has been observed before several large earthquakes in other subduction zones and, therefore, these regions may be in the late preparation stage for such events. It is important to monitor these regions for short-term precursors which may occur before these events. Foreshocks are one of the most commonly observed short-term precursors. The real-time analysis capability of the Graefenberg seismic array in West Germany provides a tool for monitoring a region for foreshocks. In this paper we investigate the location capability of this array in the Hellenic Arc to determine if it is good enough to use the array for foreshock monitoring.

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Foreshocks generally occur only during several days to several weeks prior to the mainshock (e.g. Jones and Molnar 1978). This means that one must recognize the foreshocks as soon as possible to allow time for difficult decisions and meaningful warnings. The first step in this recognition is the detection and location of the events.

Wyss and Baer (1981) presented a detailed study of long- and intermediate-term seismicity patterns in the Hellenic Arc south of Greece. They concluded that most of the arc could be considered a seismic gap of the first or second type (highest seismic potential) and that a small section of the arc had no clear history of large earthquakes. They also examined temporal seismicity patterns throughout the arc and found two regions which were presently experiencing seismic quiescence. In the long-, intermediate- and short-term prediction framework, therefore, two segments of the Hellenic Arc have reached the intermediate stage and should be monitored for possible short-term precursors.

In the circum-Aegean area about 27% of mainshocks with $m_b \geq 5.5$ have foreshocks (swarmlike sequences excluded). Wong and Wyss (1984) showed that these foreshock sequences have a high level of spatial and temporal clustering. This observation suggests that foreshock sequences in the study area may be detected and identified in real time. Therefore, we investigated the possibility of using the Graefenberg digital seismograph array for foreshock monitoring in the circum-Aegean area. The primary advantage that the GRF data have over local data is that they can be analysed in real time which is not yet done with local data. The GRF array may, therefore, provide an important source of rapidly available information to local researchers. The source dimensions of the largest ruptures in the Hellenic Arc are expected to be about 100 km (Wyss and Baer 1981). Therefore, the GRF location capability must achieve an accuracy better than 100 km to monitor potential source volumes.

The array and the data

The Graefenberg array consists of 13 vertical and 6 horizontal Wielandt broadband seismometers located near Erlangen in southeast West Germany (Harjes and Seidl 1978; Seidl and Kind 1982; Fig. 1). The distance from the stations to the Hellenic Arc varies between 1,500 and 1,700 km. Digital seismic data are transmitted continuously from the seismographs to the central site in Erlangen where they

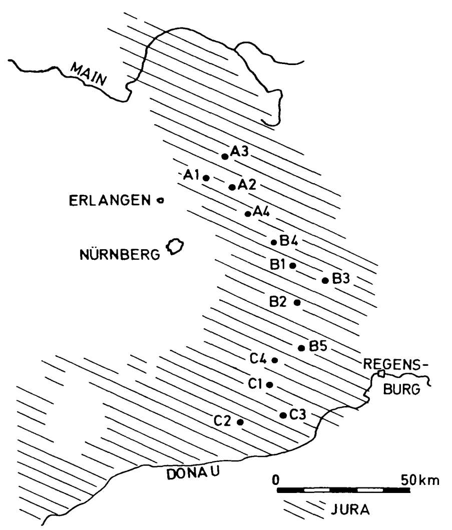

Fig. 1. Map of the GRF array stations. All stations have vertical Weilandt seismometers. The stations used in this study included A1–4 and B1–3

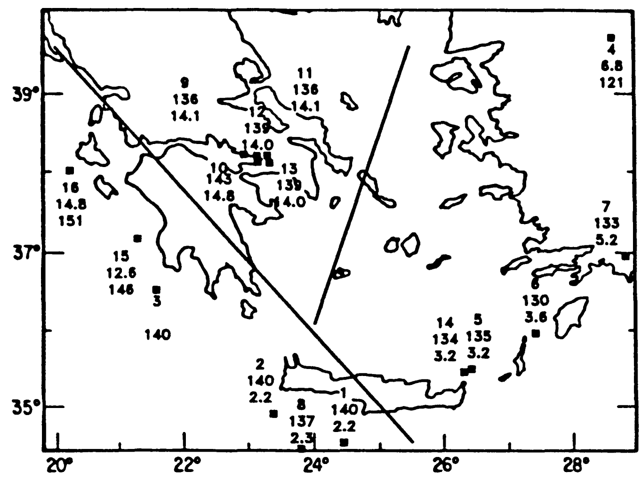

Fig. 2. Epicentre map showing the locations of the events used in this study. Each epicentre is labelled by event number, azimuth to the events from the centre of the array and the time difference between the arrival of the direct *P* wave and *P*4 at station B3. This time is a function of distance, decreasing toward the south. The *P* arrival for event no. 3 could not be measured at GRF; therefore, this event was excluded from further study

are recorded on magnetic tape. Generally the signals are examined within one day. This speed of analysis is crucial for the foreshock recognition problem. Faster analysis is possible, if necessary.

An advantage of using the Graefenberg array for this study is that one can use array techniques for phase recognition and event location. The most common array technique for event location is beam forming. In beam forming the array is focused into a particular region by applying time delays appropriate to a plane wave arriving from that region to the seismic traces. After these delays have been applied, the traces are summed and coherent plane waves from that region are amplified. By finding the best beam for a given event, one has located that event in terms of its azimuth and slowness.

Several features of Greek seismograms recorded at Graefenberg noted by Rademacher et al. (1983) are important for the work reported here. Greek events clearly show two major phases in the *P*-wave group at Graefenberg. The direct *P* wave has small amplitudes and is difficult to recognize in many cases. The second phase has mostly high amplitudes in the northern part of the array with smaller amplitudes to the south, causing large differences in the waveform across the array. This second phase has been interpreted by Rademacher et al. (1983) as a reflection from the 400-km discontinuity and is termed *P*4.

It is difficult to use *P* waves from Greek earthquakes recorded at GRF for accurate epicentre calculations because of their small amplitude. In many cases the phase would be missed and picking the generally emergent, and rather long-period, first arrival could be subject to large reading errors. We used the *P*4 phase for array processing. To avoid difficulties in waveform correlation, we considered only data from seven of the instruments in the northern part of the array (Fig. 1) where the amplitude of this phase was largest and the waveforms most consistent. We considered only events which occurred when all of these stations were operating to avoid problems related to changing numbers of stations. We excluded station B4 which is in the northern part of the array because this station had been down during unusually many events. The *P*4 phase is complicated by interference from depth phases as well as other reflections and converted phases. These interfering phases often travel with different slownesses so they merge with and split from one another as they travel across the array. These complications make it difficult to pick the same phase consistently at all stations. For this reason we used only events which were large enough to be well recorded on broadband data.

Events from various locations in the Hellenic Arc were examined. The locations of these events are shown in Fig. 2 and listed in Table 1 (the distances and azimuths are calculated from array station A1). The pattern of arrival times of the *P*4 phase across the array was determined by comparing waveforms of each event at different array stations and by comparing waveforms for different events at the same stations. The arrival time differences between all stations and a reference station (A1) were computed to form an empirical delay set which corresponds to the PDE location of the event being considered. Our hypothesis was that delays from new events with unknown locations could be automatically compared to the delays for the calibration events to provide a quick estimate of their location. The first step in determining if this would be possible was to check for regional consistency of the observed delay patterns.

Regional consistency of delays

The calibration events we chose are distributed fairly uniformly in space along most of the Hellenic Arc from the Ionian Sea to Turkey (Fig. 2). By examining the delay sets we determined that three regions showed fairly consistent delay patterns. These are divided by straight lines in Fig. 2 into West, Corinth and East, and the delay patterns for each of the three regions are shown in Fig. 3. A linear trend corresponding to an average velocity of *P*4 across the array of 9.6 km/s was removed from the delays. Thus, the delays

Table 1. Greek earthquakes

No.	Date	PDE			Azimuth		Distance		m_b
		Lat. N	Long. E	P_4-P	GRF	PDE	GRF	PDE	
1	15. 5.79	34.530	24.437	2.2	140	143	19.65	19.55	5.6
2	18. 5.79	34.909	23.351	2.2	140	145	19.65	18.71	4.9
3	27. 5.79	36.544	21.563		140	147		16.37	4.9
4	18. 7.79	39.672	28.660	6.8	121	122	17.98	17.23	5.2
5	23. 7.79	35.483	26.322	3.2	135	137	19.28	19.53	5.2
6	22. 8.79	35.946	27.417	3.6	130	134	19.14	19.66	5.3
7	4.10.80	36.937	28.847	5.2	133	129	18.56	19.58	4.9
8	10. 2.81	34.379	23.779	2.3	137	145	19.61	19.41	4.2
9	24. 2.81	38.222	22.934	14.1	136	140	15.32	15.34	5.9
10	25. 2.81	38.125	23.141	14.8	143	139	15.07	15.52	5.6
11	4. 3.81	38.209	23.288	14.1	136	139	15.32	15.52	6.0
12	5. 3.81	38.207	23.129	14.0	139	139	15.36	15.44	5.1
13	7. 3.81	38.186	23.320	14.0	139	139	15.36	15.55	5.5
14	1. 6.81	35.445	26.307	3.2	134	137	19.28	19.55	5.1
15	22. 6.82	37.160	21.273	12.6	146	147	15.85	15.64	5.1
16	17. 1.83	38.026	20.228	14.8	151	148	15.07	14.29	6.1

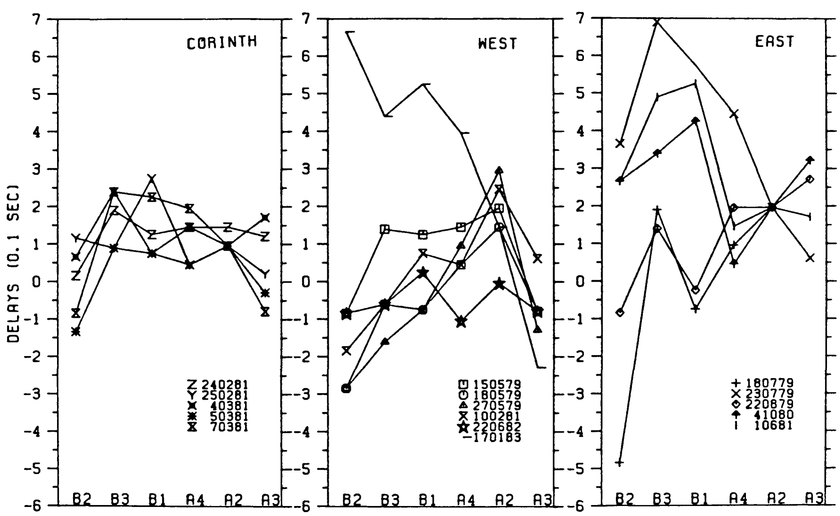

Fig. 3. Delays at the Graefenberg array stations relative to station A1 for events in three subregions of the Hellenic Arc. A constant depending on distance to the events from the array (distance/9.6 km/s) was subtracted from all delays to remove a linear trend. Note how the delays are fairly consistent in the Corinth and West regions

for an event which moved out across the array at this velocity would plot as a straight line across the plot at delay = 0.0.

The Corinth group is made up of five events which occurred during February and March, 1981 (Table 1). The average distance between these events is 18 km. The standard deviations for the delays at a given station in this group are all less than 0.1 s (see Fig. 3).

The West region includes six events from the Ionian Sea and Crete. This group shows surprisingly consistent delays considering that the average distance between the events is 264 km. The exception is the event of January 17, 1983 in the Ionian Sea. The delays for this event are clearly higher than for the others in this group in the B sub-array. The explanation for this difference is unknown. This event is large and well recorded, but the *P*4 phase is long-period and very complex. This resulted in a questionable correlation of the waveforms across the array. Perhaps this complexity reflects complexity in the larger source of this event. If this earthquake is excluded, the standard deviations for delays in this group varies from 0.07 to 0.12 s. This variation is essentially the same as that observed for the Corinth group. This observation suggests that the variance of delays is only weakly dependent on the distance between events in this distance range.

The East group includes five events from east of Crete and Turkey. These events cover a large region, like the West group, and their delays are not consistent. The standard deviations in this region vary from 0.0 to 0.4 s.

Two of the three regions show consistent delay patterns. However, the patterns are not sufficiently different to allow one to place an event with an unknown location unambiguously in any one region on the basis of its delay pattern alone. Therefore, we conclude that the delay patterns do not provide the location capability needed.

Techniques using additional information

We clearly need to use some technique for locating the events which considers more than the delays. The goal is to find the delay pattern from the set of delays for the calibration events (Fig. 3 and Table 1) which most closely fits an event with an unknown location. In order to test these techniques we treated the calibration events as if their locations were unknown. We picked one of the calibration delay sets and tested all of the calibration events against it to see if we could find the event whose delays we had picked. If this event could be recognized, we considered the test a success. If it could not, we considered the distance between the event picked by the test and the actual event as a measure of the location error of the technique.

First we tried a delay and sum technique based on the assumption that the correct delay pattern should be the one which maximizes the energy in a trace formed by summing the signals from all stations after they have been delayed according to that pattern. In order to test this technique, the signals for each event from all stations were delayed by the amounts determined for one of the calibration events (Fig. 3) and summed to form a sum trace. The integral of the squared sum trace was then computed over a time window including the phase used in identifying the delays to determine the total energy for each event. The same calculation was done for the calibration event whose delay set was being used to delay the other events and, because that calibration event determined the delays, we would expect it to show the highest energy of all of the events. In 5 of the 16 cases the calibration event did have the highest energy using this technique (as expected). In the other cases the distance between the event with the highest energy and the correct event varied from 17 to 510 km (average = 286 ± 158 km). This technique, therefore, does not provide satisfactory results.

We next calculated single lag correlation coefficients for all seven traces by multiplying together the delayed traces and summing the product trace over a time window which included the phase used to determine the delays. This technique was more successful than the summing technique at recognizing the correct event (9 of the 16 cases). Yet, in the other cases the distance between the correct event and the event found was again very high (263 ± 213 km).

In summary, the distances between the correct events and other events with high sums or correlation coefficients averaged between 250 and 300 km. This number provides an estimate of the location errors expected if these techniques are used for quick locations. These errors are too large to be acceptable.

A possible solution

In the previous section we discussed three techniques for quick event locations using delays observed across the Graefenberg array. These techniques do not provide the resolution that we need in this work. In this section we describe a distance and azimuth measurement which may allow reasonably quick locations of events in the Hellenic Arc.

Rademacher et al. (1983) made the observation that the *P*-wave group observed at Graefenberg for events from Greece characteristically contained two major phases. The first is the direct *P* wave and the second is a *P* wave reflected from the 400-km discontinuity (*P*4). These phases travel across the array with different average slowness, 13.6 and 11.5 s/deg (corresponding to 8.1 and 9.6 km/s) and the time difference between the two phases decreases as the sources move farther away (see Rademacher et al. for examples). This time difference, therefore, provides a measure of the distance to the earthquakes. We used the time difference between the first arrival of the *P* wave and the first maximum of *P*4, rather than the *P*4 arrival time, because the

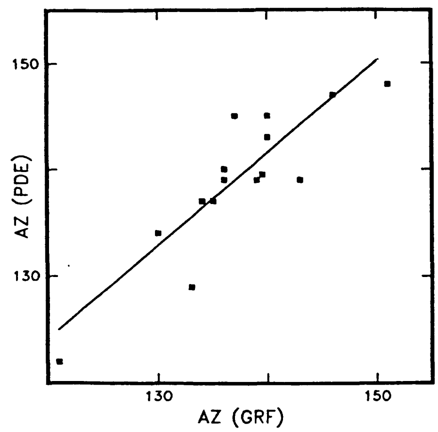

Fig. 4. Azimuth from the centre of the Graefenberg array to the epicentres listed in the PDE [Az(PDE)] versus the azimuth estimated by the GRF data alone [Az(GRF)]. The least-squares fit through the data was obtained minimizing the dependent variable only. The standard deviation of the latter is ±3.25°

first arrival of *P*4 is sometimes hard to recognize. The errors in these measurements are probably a few tenths of a second (provided that the correct phase is picked), which is small relative to the magnitude of the variations. The time differences at station B3 for the calibration events we examined are given in Fig. 2. They range from 2 s near Crete to 14s near Corinth. This suggests that if the time differences can be measured even with only 1 – s accuracy, distances accurate to less than 100 km can be measured. This is clearly better than the other techniques we described.

The second part of the earthquake location we need is the azimuth. The stations we used at Graefenberg were oriented along a line with an azimuth of N 140° E, pointing directly at the region we studied. This configuration is the worst possible for azimuth determinations. The azimuths determined from a least-squares fit to a plane wave crossing the array have errors of between 2° and 6°. The azimuths we determined are shown in Fig. 2 with the epicentres and difference times. They show the expected pattern of smaller azimuths (near 120°) for Turkey, the low 130s for the area between Rhodes and Crete, the upper 130s near Corinth and the 140s near Crete and in the Ionian Sea.

The correlation between the Graefenberg azimuths and those calculated on the basis of the PDE locations is shown in Fig. 4. Although only the northern part of the array was used, the standard deviation from the least-squares fit is not large (±3.25°). However, at the epicentral distances in question, this translates to a mislocation of approximately 100 km. The error introduced through uncertainties in the transformation of (*P*4–*P*) into distance is smaller. Figure 5 shows the epicentral distance (based on the PDE location) versus (*P*4–*P*). The straight-line fit in Fig. 5 was obtained by minimizing the dependent variable only and is described by the equation

Distance (in degrees) = 20.45 – 0.36 (*P*4–*P*) (in seconds).

Based on this relationship one could calculate distance for future events based on the observed (*P*4–*P*) value. The standard deviation for the distance estimate is ±0.53°, corresponding to a location error of about 60 km.

Originally it was our intention to use only *P*4 and not *P* for the location because *P* is usually weak at GRF. Now, with the method just described, we need the arrival time of *P* at one station. This is certainly a disadvantage for

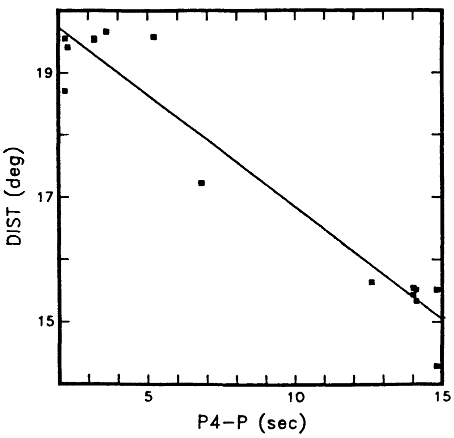

Fig. 5. Distance between the PDE location and Graefenberg, versus the time difference between the reflection off the 400-km discontinuity, $P4$, and the direct P phase. The straight line is a least-squares fit assuming a linear correlation between the two parameters for the relatively short distance range. The standard deviation in the dependent variable is $\pm 0.54° = \pm 60$ km

weak events, but it is still much better than using all the P delays, which standard techniques do.

Discussions and conclusions

In this paper we examined the question of whether the location capability of the Graefenberg array was high enough to allow the use of the array for foreshock recognition in the eastern Mediterranean. The similarity of delays for events from all around the Hellenic Arc, the lack of an increase in the delay scatter in the West region relative to the Corinth region and the difficulty of picking out the calibration events using waveform correlations based on observed delays for those events all indicate that the locations determined using conventional array location techniques are not accurate enough for the foreshock problem. We found that the errors were between 200 and 300 km using such techniques.

Two parameters are used to describe locations by the Graefenberg array: slowness and azimuth. The configuration of stations we used is the worst possible for azimuth determinations and the best possible for slowness determinations for Greece. In spite of this, we found the azimuth resolution to be better than the slowness resolution. This is because the slowness in this distance range is controlled by complex vertical structure in the upper mantle and varies only slightly with distance. Azimuth, on the other hand, does not depend on vertical earth structure and can be measured well in any distance range.

Ironically, the same earth structure which makes it difficult to use slowness to determine distance provides the reflection ($P4$) which we use for our distance estimate. If this reflection did not occur it would be very difficult to determine the distance to these events. Using the difference between the $P4$ and the P arrival, epicentral distances can be estimated rapidly with an expected error of ± 60 km.

In conclusion we offer a technique for determining quick locations of earthquakes in the Hellenic Arc, using the time difference between the direct P wave and the reflection off the 400-km discontinuity and an azimuth estimate from the Graefenberg array. The error in the distance measurement may be as small as 60 km. The error in azimuth is about $\pm 3.3°$ (100 km). This resolution is comparable to the size of rupture zones in the Hellenic Arc and may be sufficient for preliminary determinations of foreshock locations.

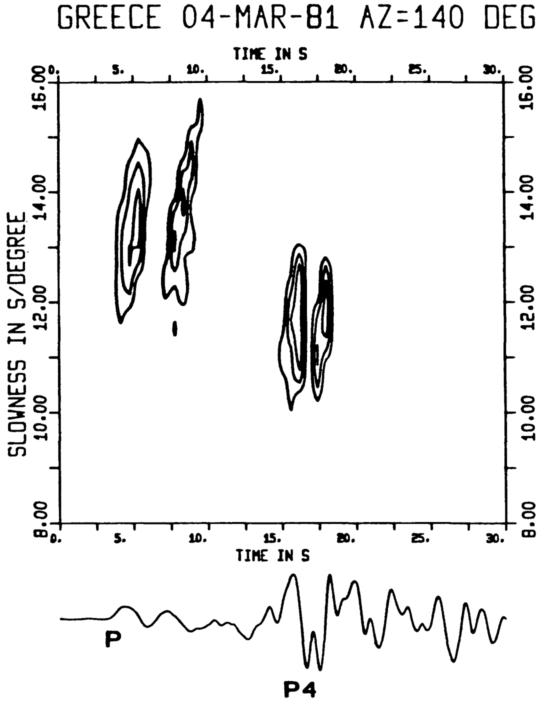

Fig. 6. Slowness analysis of one Greek earthquake. The slowness is measured as a function of time using all array data. The azimuth is fixed. Only the contour lines from the peaks of the phases P and $P4$ are plotted. The different slowness of P and $P4$ can easily be seen. This kind of data could be used to measure the time difference between $P4$ and P and the azimuth on-line

A number of problems limit the usefulness of this technique. First, it depends on the correct identification of the P and $P4$ phases. As mentioned above, the P phase from events in the eastern Mediterranean is weak at GRF, but it can usually be identified. Late P phases, like $P4$, can have a number of causes including surface reflections, source complexity and reflections from upper-mantle discontinuities. Large location errors could occur if some other phase is mistakenly identified as $P4$. The waveforms of Greek events at GRF are very distinctive, primarily because of the different slownesses for the two main phases, so it is unlikely that such a misidentification would occur. These problems can, therefore, be overcome by an experienced analyst at GRF. However, our original goal of developing a foreshock recognition tool which could run automatically could not be achieved. A second problem is that we could only study events which were large enough to be well recorded at GRF. This includes most events with $m_b \geq 4.7$–4.9 (see Table 1). We must hope that foreshock sequences which occur will contain enough events of this size to be clearly recognized as anomalous.

It is possible that a more sophisticated array technique may provide a satisfactory solution to the problem. Figure 6 shows a slowness analysis of a Greek event using all array data. The slowness is plotted as a function of time with fixed azimuth. Only two dB contour lines from the peaks of the two phases P and $P4$ are plotted. The slowness difference of the two phases is clearly visible. It seems possible to measure the time difference between *P*4 and *P* and the azimuth to the event from data like that shown in Fig. 6. This might provide an on-line loction of Greek earthquakes accurate enough to identify events from the quiescent regions of the arc. Extensive computations are required for such a method, which works in the frequency and wave-number domain. Unfortunately, the computing facilities of the Graefenberg array are not sufficient for doing this analysis in real time at the present.

The possibility of using the Graefenberg array to monitor other regions of the world remains good. Our present results indicate that monitoring regions in Europe may be difficult or impossible unless the local structure provides some aid, like the *P*4 phase used in this work. The resolution may improve with increasing distance when the slowness-distance relationship becomes more reliable and the signal shape is clearer.

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