ESTIMATION OF THE PARAMETERS CONTROLLING STRONG GROUNG MOTION FROM SHALLOW EARTHQUAKES IN GREECE

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A B S T R A C T

A simple stochastic model for the generation of strong ground motions is partially validated for applications in Greece. The partial validation is restricted to the far field of medium size earthquakes and consists of independent estimates of the underlying physical parameters from geophysical data, comparisons with empirical strong motion spectra and comparisons with selected accelerograms; the validation encourages the use of values determined in California. The use of the stochastic model adds to the capabilities of the strong motion data bank in Greece for engineering applications, helps to define the underlying physics and paves the way for the incorporation of low amplitude geophysical data.

ΕΚΤΙΜΗΣΗ ΤΩΝ ΠΑΡΑΜΕΤΡΩΝ ΠΟΥ ΚΑΘΟΡΙΖΟΥΝ ΤΗΝ ΙΣΧΥΡΗ ΕΔΑΦΙΚΗ ΚΙΝΗΣΗ ΑΠΟ ΕΠΙΦΑΝΕΙΑΚΟΥΣ ΣΕΙΣΜΟΥΣ ΣΤΗΝ ΕΛΛΑΔΑ.

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Π Ε Ρ Ι Λ Η Ψ Η

Ενα απλό στοχαστικό μοντέλο σύνθεσης επιταχυνσιογραμμάτων επαληθεύεται μερικώς για εφαρμογές στον ελληνικό χώρο. Η μερική επαλήθευση περιορίζεται στο μακρυνό πεδίο σεισμών μέσου μεγέθους και συνίσταται σε ανεξάρτητους γεωφυσικούς προσδιορισμούς των αιτιοκρατικών φυσικών παραμέτρων, συγκρίσεις με εμπειρικά φάσματα και επιλεγμένα επιταχυνσιογράμματα. Η επαλήθευση ενθαρύνει την υιοθέτηση τιμών που έχουν προσδιορισθεί στην Καλιφόρνια. Η χρήση του στοχαστικού μοντέλου αυξάνει τις δυνατότητες που προσφέρει η ελληνική τράπεζα δεδομένων ισχυρής κίνησης στην τεχνική σεισμολογία και στο αντισεισμικό σχεδιασμό, βοηθάει στον προσδιορισμό των φυσικών παραμέτρων που διαρφώνουν ισχυρούς σεισμικούς κραδασμούς και διευκολύνει την ορθολογική χρήση σεισμολογικών δεδομένων.

INTRODUCTION

Strong ground motions can be specified at different levels, depending on the data availability and the scope of a particular investigation. Many investigations, either in theoretical geophysics or in earthquake engineering, are conducted at the level of strong motion time histories. Realistic time histories can be selected from strong motion data banks and scaled to a particular application (Heaton et al., 1986) or synthesized from basic geophysical data. The synthesis can rely on kinematical models of fault rupture (Trifunac, 1971) or on basic parameters describing the physics of the earthquake fault. Because of the heterogeneities in the fault rupture mechanism and the dependency of strong ground motion on such heterogeneities (Joyner and Boore,1988), stochastic source models are required for the prediction of such motions. Still, stochastic models may differ vastly on the complexity of the underlying physics.

The data bank of Greek strong motion records is acquiring a respectable size amenable to statistical analysis and selection of representative accelerograms. Statistical analysis has lead to the definition of empirical frequency spectra (Theodulidis and Papazachos, 1992) and the time is wripe for parallel studies of stochastic simulations. The latter have the distinct advantage of utilizing other available geophysical data, promoting the understanding of the underlying physical parameters and thus rationalising comparisons with data from other regions.

Here we summarise the results of a more general study on stochastic strong ground motion generation in Greece. We present the stochastic model, our best estimate of the underlying parameters for Greece and also comparisons with empirical spectra and accelerograms.

THE METHOD

Ground motion prediction with Stochastic Source Models (Joyner and Boore, 1988, DiBona and Rovelli 1990) proceeds in two steps: in the first step a target amplitude spectrum is set on the basis of seismological parameters. In the second step a time history is generated from band limited white noise shaped to the target spectrum. Variations of the method (Silva, 1987) implement the amplitude target spectrum with an appropriate phase difference spectrum and then transpose to the time domain.

The target amplitude spectrum R(f) represents in the frequency domain the chain-action from the source to the site with the product:

$$R(f)=C\,S(f)\;A(f)\;D(f)\;I(f) \qquad (1)$$

where the factors C,S,A,D and I, are, respectively, a scaling factor, the source spectrum, an amplification factor, a diminution factor and an instrument-response factor. The S(f) factor depends on the seismic moment, M_0, and a corner frequency, f_0, representing in the frequency domain the geometric characteristics of the fault rupture. For sources that do not break the seismogenic layer the fracture geometry is represented with a single corner frequency, f_0, whereas larger earthquakes or barrier models require more. The corner frequency, f_0, is related to the seismic moment through the stress drop, $\Delta\sigma$, across the fault. The scaling, C, factor averages the radiation pattern around the source and introduces the density and shear wave velocity in the source region, the geometric spreading of seismic waves and the free surface effects. The amplification factor, A, is the frequency-dependent transfer function that results from wave propagation in a stack of surface layers with a strong impedance contrast at the base of the stack (Boore and Joyner, 1984). The diminution factor, D, represents the anelastic attenuation in the crust and includes a high frequency energy cut-off, f_m. This frequency cut-off is assigned to the diminution factor as a near source attenuation effect (Hanks and McGuire, 1981). The same frequency is explained by the barrier model (Papageorgiou and Aki, 1983) as a source effect. Finally the instrument-response factor, I, is used to adjust the target spectrum to the particular ground motion parameter of interest (Boore, 1983). Clearly, the above definition of the target spectrum conforms to the expected far field characteristics of ground motions from point seismic sources i.e. at distances large compared to the dimensions of the source. At closer distances the generating functions must operate on sections of the rupture surface (Joyner and Boore, 1988, Papageorgiou and Aki, 1983).

In the second step Gaussian white noise is generated with a random-number generator and windowed by a shaping function or a box car with duration T_w:

$$T_w=\frac{1}{f_o}+0,05r \qquad (2)$$

where the first term represents the duration of rupture and the second accounts for scattering and wave propagation effects (Herrmann, 1985). The resulting noise sample has a mean spectral level of unity and is shaped with R(f). In order to go back to the time domain, a Fast Fourier Transformation is carried out. The process is repeated by varying the seed number for the random-number generator. Between 20 and 100 simulations are generally adequate to give a good estimate of the peak ground motion.

The alternative method employs an appropriate phase shift spectrum to implement the amplitude target spectrum and relies on the selection of a representative recording (phase generator). In this case, additional simulations require more phase generators (time-histories). Taking into consideration all the parameters involved in the generation and propagation of seismic waves makes it clear that a meaningful selection of phase generators should be guided by site specific criteria.

ESTIMATION OF SEISMIC MOTIONS IN THE AREA OF GREECE.

TARGET SPECTRA. Accelerograms from shallow earthquakes in Greece have been processed and analysed in terms of peak ground motions, duration and response spectra (Theodulidis and Papazachos, 1992). The statistical analysis was performed in terms of a broad classification of site conditions as rock or alluvium, the surface wave magnitude, M_s, and the distance, R, from the source. The generation of target spectra utilizes the additional geophysical parameters involved in the definition of R(f), eq. (1). For some of these parameters, such as mechanical properties of the crust, universal values are usually assumed. Other parameters, such as the quality factor Q, the high frequency cut-off, f_m, the stress drop, $\Delta\sigma$, and the layering amplification have become the focus of serious regional investigations. Such investigations involve, apart from strong motion records, registrations of low amplitude seismological networks.

In Greece, limited investigations on the quality factor Q have indicated values in the range 100-300 (Papazachos, 1992) and suggest a highly attenuating crust similar to California. Either definition of the high frequency cut-off, f_m, (source or near field crustal property) requires near field recordings that are very sparse in Greece. The static stress drop, $\Delta\sigma$, controlling long period motions has been attributed low values in the range 4.5-12 bars (Kiratzi et al., 1985) in Greece. For the shorter periods of interest to engineering design we repeated the calculations of Hanks and McGuire,(1981) using Greek accelerograms. The calculations give a dynamic stress drop, $\Delta\sigma$, that is correlated, through the theory of random vibrations, to the rms peak acceleration. From the Greek strong motion data bank we selected the records on rock shown on Table(1). On this table the corner frequency, f_0, was calculated from Fourier amplitude (or velocity response) spectra and rms peak ground accelerations from integration of the strong motion section of the accelerograms. Stress drops, $\Delta\sigma$, from the last column of Table(1) were taken as the average of the two horizontal components, when available and plotted against M_s in Figure (1). They indicate a mean stress drop of 50 bars.

Target spectra for two magnitudes (M_s=5 and 6) and two epicentral distances (R_{ep}=10 and 50km) are compared on Figure(2) with the 10-90% band of empirical predictions from Greek accelerograms. To obtain a good fit of the target spectra it was necessary to apply the amplification factors developed in California (Boore,1986). Similar amplification factors are not available in Greece. Also, the comparison was improved by a slight correction of the surface wave magnitude, M_s, to obtain, the moment magnitude, M_w, as suggested from the analysis of Greek data (Papazachos and Papazachou, 1989).

TIME HISTORIES. For a visual test of the predictability of stochastically generated time histories in Greece we performed an experiment. We looked into the earthquake sequence of 1983 in W Cephalonia, an Ionian Island (Skordilis et al., 1985). Figure (3) shows the sequence and the fault mechanism of the main shock and the larger aftershock. Out of the sequence we selected as phase generators two aftershocks in a rather narrow magnitude, distance range, as indicated in Figure(3). We then declared a later aftershock (ARG83-8) record as target for the stochastic ground motions. Stochastic ground motions were generated with two different methods: from white noise and from the phase shift spectra of the preceding recordings(phase generators).

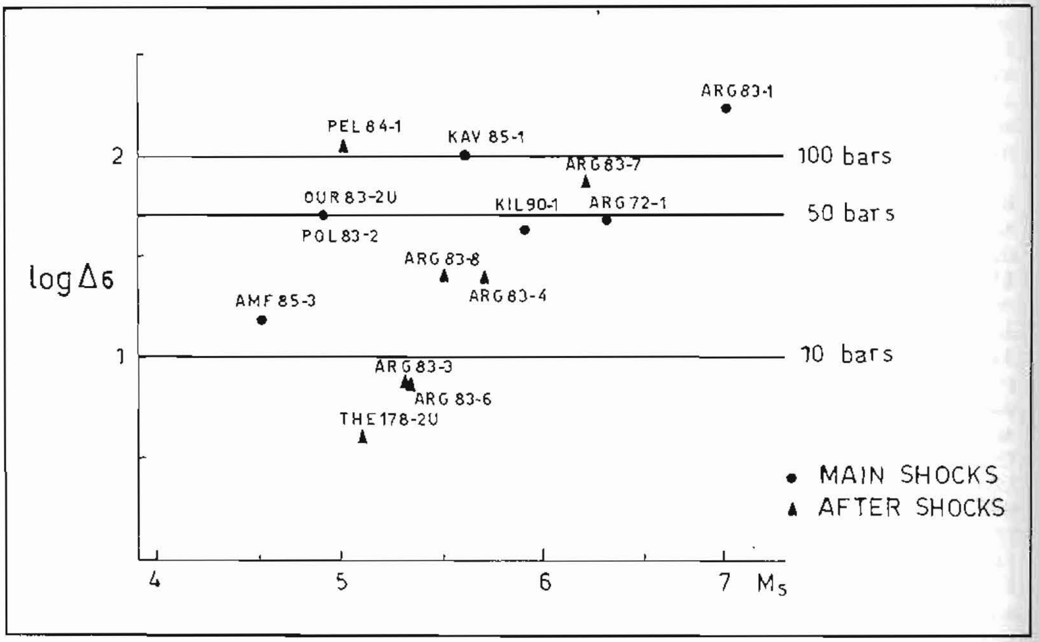

Fig.1. Stress drop, $\Delta\sigma$ (in bars), versus the surface wave magnitude, M_s calculated from Greek accelerograms on rock.

Figure(4) shows the results of the test. The top row presents the two actual records analysed for the stochastic prediction (phase generators). The second row shows predicted motions for the target aftershock based on the phase generators on the first row. For the comparison we reproduce on the third row, the two recorded horizontal components of the target aftershock. The bottom row shows two white noise randomly generated motions. The prediction of the target time history appears better on the basis of ARG83-3, from the nearby aftershock of similar magnitude (the same crustal path and, probably, the same source mechanism), although the simulation reflects the smaller magnitude of the phase generator. The white noise simulations are, in general, very satisfactory. On the contrary, the simulation based on the largest aftershock gives lower acceleration spread over a much longer strong motion duration. In this case, responsible for the deviation from the target time history is the magnitude difference and probably, to a lesser extent, the different azimuthal disposition.

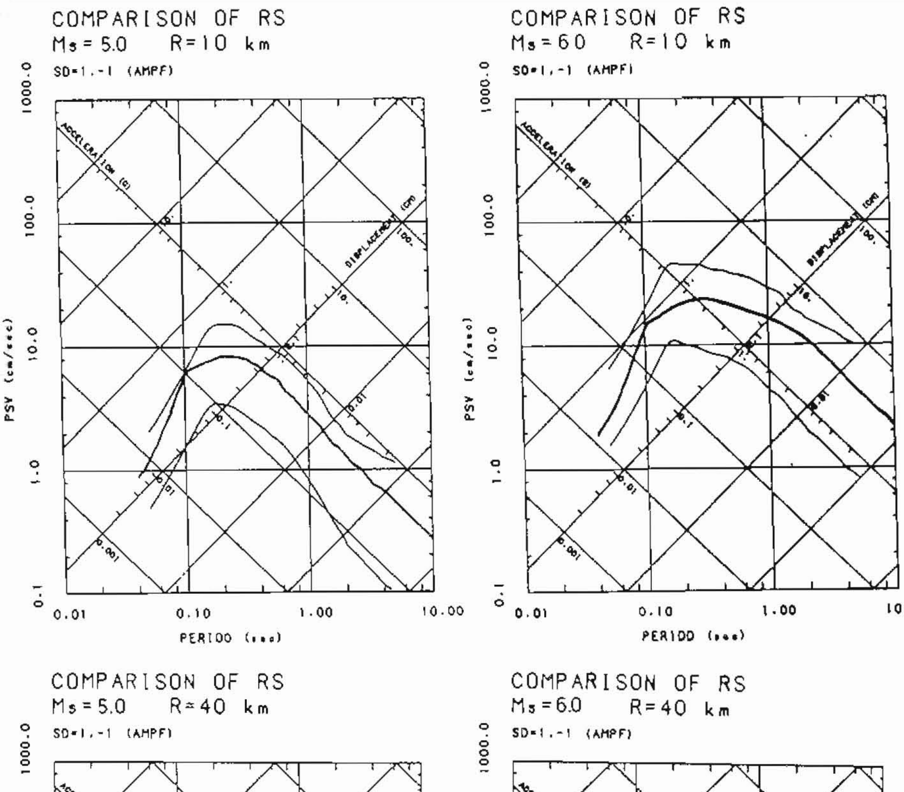

Fig.2. Target spectra (thick lines) for two surface wave magnitude, (M_s=5 and 6) and two epicentral distances, (R_{ep}=10 and 40 km) compared to empirical prediction from Greek accelerograms. The empirical predictions are plotted with thin lines corresponding to ±1 standard deviation.

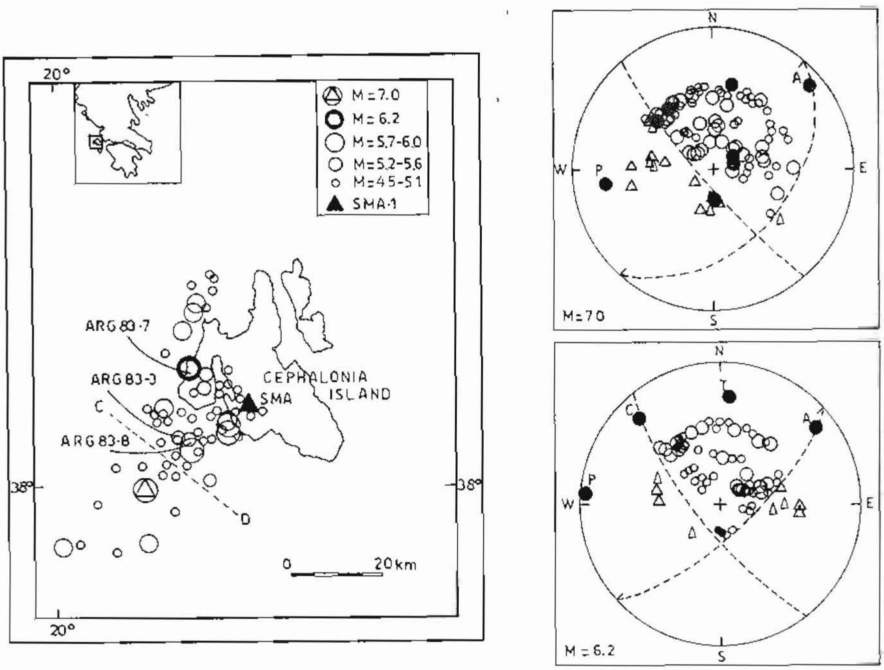

Fig.3. Epicentral map of the best located earthquakes of the 1983 Ionian island seismic sequense and the fault plane solution of the main shock (Jan. 1983, M_s=7) and the largest aftershock (Mar. 1983, M_s=6.2) (Scordilis et al., 1985). The map shows the location of the accelerograph and the aftershocks selected for the analysis:

ARG83-3	M_s=5.3	R_{ep}=13 km
ARG83-7	M_s=6.2	R_{ep}=13 km
ARG83-8	M_s=5.5	R_{ep}=10 km

CONCLUSIONS

We presented partial validations of a simple stochastic model generating strong ground motions with data form Greece. The stochastic model is defined with a spectral shape band limited between f_m and f_o, corrected for surface layering and scaled to the moment magnitude, M_w. The validation was satisfactory for f_m=10Hz, a stress drop $\Delta\sigma$=50bars (for the definition of the corner frequency,f_o) and amplification from surface layering. In general, we found good fit of the stochastic predictions for the values of the parameters determined for California. This validation was based on empirical spectra derived from Greek records of shallow earthquakes but without taking into account the source mechanism.

More site specific motions require relevant data within a narrow magnitude and distance range. Relevance of the data renders the method inpractical in most cases where a representative record of source and propagation effects is

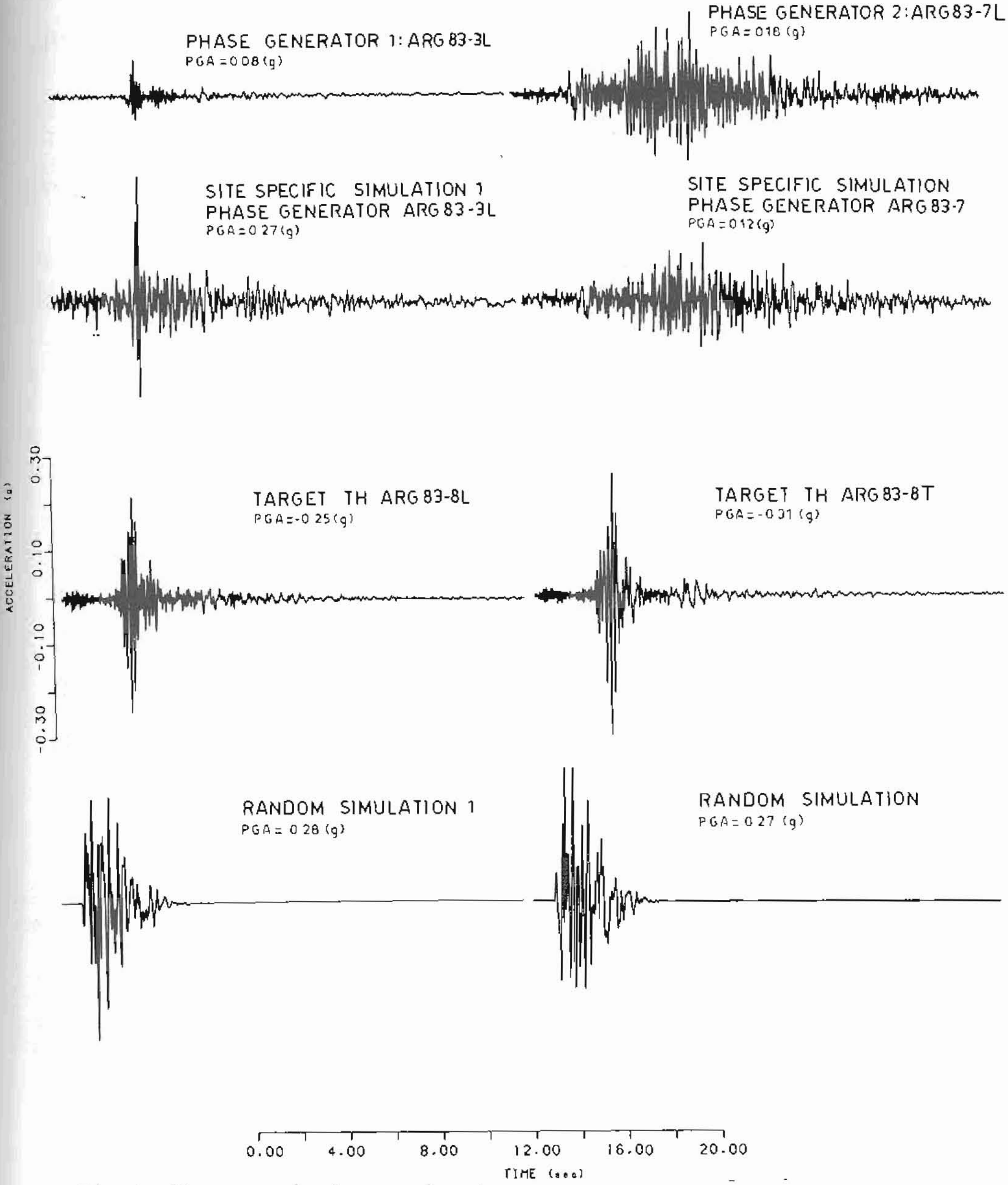

Fig.4. The numerical experiment:
Top row : The phase generator recordings.
Second row : Site specific simulations from the target spectrum and the two phase generators in the first row.
Third row : Recorded target motions.
Botton row : Random simulations from the target spectrum and white noise for two different seed numbers.

Table 1. The calculation of source parameters from Greek accelerograms on rock

DATE	GMT ORIG. TIME	EPICENTRAL COORDINATES φ^{o}N	λ^{o}E	R_{ep} (km)	M_s	log M_o dyn-cm	a_{rms} (%g)	f_o Hz	$\Delta\sigma$ (bars)	Recording Code
17 SEP 72	14:07	38.18	20.23	22	6.3	25.34	12.3	0.83	47.69	ARG72-1T
							13.7	0.90	55.32	L
4 JUL 78*	22:23	40.73	23.12	17	5.1	23.54	1.2	2.5	4.32	THE178-2Y-L
							1.2	2.0	3.87	T
17 JAN 83	12:41	37.96	20.23	33	7.0	26.39	32.7	0.85	189.03	ARG83-1L
							30.8		178.09	T
17 JAN 83*	16:53	38.11	20.37	13	5.3	23.84	2.8	1.7	9.02	ARG83-3L
							2.0	1.7	6.4	T
19 JAN 83*	00:02	38.08	20.38	15	5.7	24.44	7.7	-	-	ARG83-4L
							6.8	1.7	25.27	T
31 JAN 83*	15:27	38.14	20.45	6	5.3	23.84	4.5	-	-	ARG83-6L
							5.2	1.7	7.73	T
23 MAR 83*	23:51	38.22	20.35	13	6.2	25.19	31.2	1.0	77.06	ARG83-7L
							29.8	1.0	73.61	T
24 MAR 83*	04:17	38.11	20.42	10	5.5	24.14	9.7	2.0	26.06	ARG83-8L
							10.5	-	-	T
26 AUG 83	12:52	40.47	23.96	16	4.9	23.24	4.8	3.0	25.27	OUR83-UL
							6.8	-	35.81	T
							5.4	5.0	105.53	POL83-L
							2.4	-	46.90	T
25 OKT 84*	09:49	36.83	21.72	27	5.0	23.39	17.8	3.0	111.84	PEL84-L
							19.7	-	123.77	T
22 MAR 85	20:37	38.98	21.11	14	4.5	22.64	2.1	5.0	12.49	AMF85-L
							3.9	-	23.20	T
9 NOV 85	23:30	41.26	24.02	49	5.5	24.14	7.4	2.5	108.93	KAV85-L
							7.4	-	108.93	T
21 DEC 90	06:57	40.93	22.37	44	5.9	-	5.6	1.0	46.82	KIL90-L
							5.0	-	41.80	T

* Aftershocks.

unavailable. In other words, the derivation of site specific bedrock motions is, in general, beyond the resolution of the available information.

Clearly, our validation is restricted to medium size events, up to about M_s=6, that are not associated with elongated fault ruptures. Moreover, the simple validated model does not cover near field configurations for which very little strong motion data exist in Greece. For example, the 1986 Kalamata (Papastamatiou et al., 1987) earthquake had a magnitude M_s=6, just broke the entire seismogenic layer and produced near field effects within a radius of about 5km.

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