Calibration of input parameters in volcanic areas and an enlarged dataset by stochastic finite-fault simulations

D. Galluzzo, F. Bianco, P. Cusano, H. Langer, L.Scarfi,
G. Tusa & G. Zonno
Istituto Nazionale di Geofisica e Vulcanologia, Italy

A. Carvalho
Laboratorio Nacional de Engenharia Civili, Portugal

S. Olafsson, R. Rupakhety & R. Sigbjornsson
Earthquake Engineering Research Centre, Iceland

SUMMARY:
The calibration of input parameters is an important task for stochastic finite-fault simulation in volcanic areas, and we manage this in the framework of the European project UPStrat-MaFa. The stochastic simulation method requires the knowledge of fault geometry, source, crust properties of the region, and local site effects. At first, we focused the present study in the pilot test areas: Mt Vesuvius, Campi Flegrei and Mt Etna. Later, we performed two applications for a large magnitude event in the Azores Islands and the South Iceland regions. A general preliminary database of ground-motion records was collected in the test areas, to set up the empirical laws of the ground-motion parameters. The results of the simulations have been compared with observed waveforms and response spectra, to determine the suitability of the parameters used. The results show good agreement between the observed and simulated time histories and response spectra, thus encouraging further efforts towards quantitative high resolution studies on input parameters.

Keywords: stochastic simulation, calibration, volcanic areas, UPStrat-MAFA

1. INTRODUCTION

Estimation of seismic hazard and risk is based both on empirical relations that exploit already existing data, as well as on synthetic approaches that make certain assumptions on models and parameters. The advantage of using empirical relations is because they represent the "ground truth"; i. e., they are not influenced by the ideas and preferences of the modeller. On the other hand, the available databases are typically incomplete, and often they do not provide samples for all of the relevant scenarios. It is known that strong events with a large magnitude are rare, and so related instrumental data might not be available. For the aim of the UPStrat-MaFa European project, a hybrid strategy is followed that exploits the existing instrumental datasets to establish the empirical relationships between earthquake source parameters and seismic signal features, as basically the peak amplitudes of ground motion. These empirical relationships can be used for calibrating synthetic modelling, and in particular, can account for wave propagation effects. Once the parameter set of synthetic simulation of ground motion is verified for the reference data, scenarios for strong and rare earthquakes (for which real data may not be available) can be extrapolated.

It is well known that waveform simulation in volcanic areas is a challenging task due to the complexity of seismic sources, propagation medium heterogeneity, and surface geomorphological features. Hereafter, we briefly describe the tectonic setting and general characteristics of the areas investigated.

Mt Vesuvius and Campi Flegrei are densely urbanised volcanic areas southern Italy, for which the peculiar knowledge of peak ground-motion parameters are crucial to define the safety plans for these volcanic zones. The ground-motion parameters were recently investigated in both of these areas by following stochastic approaches (Galluzzo et al., 2004, 2008, hereafter G2008) and through a probabilistic seismic hazard analysis (Convertito and Zollo, 2011, hereafter CZ2011).

The volcanic edifice of Mt Etna covers an area of ca. 1200 km^2, with a population of ca. 1 million people living in this region of Catania, as well as a number of smaller cities and villages located on the flanks of the volcano. Seismic risk originates both from large earthquakes along major fault zones, as well as modest but superficial events that occur frequently on the eastern and southern flanks of the volcano. Seismic risk analyses was carried out recently by Azzaro et al. (2008). Gresta and Langer (2002) prepared maps of peak ground-motion parameters and response spectra, defining various earthquake scenarios.

The Azores islands are located in the North Atlantic Ocean, and they are positioned along a narrow area that extends for about 600 km. The seismic activity observed in the Azores plateau is the result of its location on an active plate boundary. Although most are low to moderate magnitude seismic single events or sequences, the islands are occasionally struck by large-magnitude (M_W >7) earthquakes (Bezzeghoud et al., 2008).

Iceland is situated on the Mid-Atlantic Ridge, which is the boundary of two continental plates that are drifting apart at a rate of approximately 2 cm/yr. The largest earthquakes in Iceland occur in two transform zones: the South Iceland Seismic Zone (SISZ), which is located in a populated area of the South Iceland Lowland, and the Tjörnes Fracture Zone (TFZ), which is positioned off the northern shore of Iceland.

Our main goal is the preliminary calibration of the input parameters linked to the source, propagation and site effects for recent data hat were collected through high dynamic acquisition systems equipped with a short period, broad-band sensors and accelerometers. The calibrated model is used for the calculation of response spectra and the synthesis of strong ground-motion acceleration; the validation of the method is carried out by comparison with strong ground-motion records and parameters, such as peak ground motion, and 5% damped pseudo-absolute response spectra.

Mt Vesuvius, Campi Flegrei and Mt Etna have been used as pilot test grounds for design of the methodology, which has been applied also in other areas (Azores Island and South Iceland). The novelty of the present study consists of the peculiar calibration of the input parameters based on a more recent dataset, as well as of the consideration of more recent results relating to source, propagation and site characteristics.

2. THE DATASET

At first, we describe briefly the characteristics of the preliminary dataset used for the first calibration analyses. The datasets of the volcano-tectonic (VT) events are composed of 30 selected seismic events for both Mt. Vesuvius ($M_D \in [1.5,2.7]$, focal depth $\in$ [0.0,4.0] km b.s.l.) and the Campi Flegrei area ($M_D \in [0.0,2.2]$, focal depth $\in$[0.0,3.0] km b.s.l.). These low-to-moderate magnitude earthquakes occurred in the period 2000-2010 and were collected by the Mobile Seismic Network of the *Istituto Nazionale di Geofisica e Vulcanologia* (INGV)-*Osservatorio Vesuviano*. In both areas, local earthquakes were recorded by highly dynamic, 3-component, digital seismic stations equipped with broadband, short period sensors, and in a few cases, with Episensor Kinemetrics accelerometers. The two maps in Figures 1 and 2 show the positions of the epicentres and of the selected station/ sites for both of the investigated areas.

For Mt Etna area, the dataset covers events from 2006 to the present. The data were recorded by the stations of Rete Sismica Permanente della Sicilia Orientale (RSPSO), operated by INGV–*Osservatorio Etneo* (Fig. 3). The seismic network consists of ca. 80 digital stations that are equipped with broadband seismometers, and in some cases, also with accelerometers. These stations are located in the area between the volcanic archipelago of the Aeolian Islands and the Hyblean Plateau. The dataset includes some 120 seismic events of magnitude $M \in [3.0,4.8]$. Overall, a total of ca. 4800 three-component registrations are available.

For the Azores, there are 828 records that correspond to 176 events of magnitude 3 to 6, for the period 1973-2006, with focal depths $\in$[0.0,30.0] km, for a total of 26 stations. The most installed of the digital instruments are GeoSIG SSA-320 force balance, and tri-axial accelerometers with 12, 16 or 18 bit GSR digitisers.

Three earthquakes with $M_w \in [6.3,6.5]$ are considered for Iceland; these occurred in an area that might potentially cause damage in Reykjavik and the surrounding areas. These data were recorded on the Icelandic Strong Motion Network (IceSMN) of the Earthquake Engineering Research Centre of the University of Iceland (Sigbjörnsson et al., 2004).

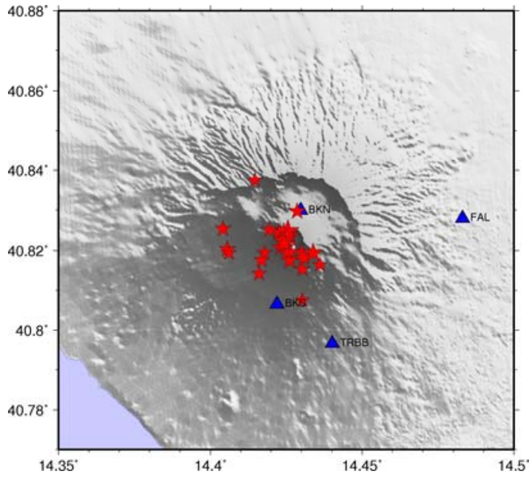

Figure 1. Epicentres of selected earthquakes (red stars) and seismic stations (blue triangles) for the Mt Vesuvius area. The size of the red stars symbols is proportional to the magnitude of the seismic event.

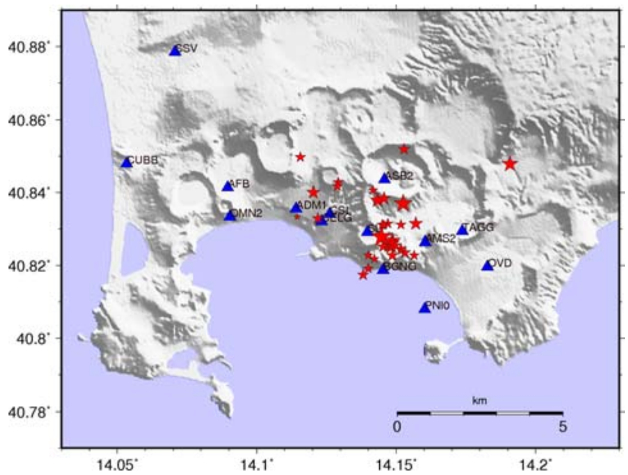

Figure 2. Epicentres of selected earthquakes (red stars) and seismic stations (blue triangles) for the Campi Flegrei area. The sizes of the red stars symbols are proportional to the magnitude of seismic event.

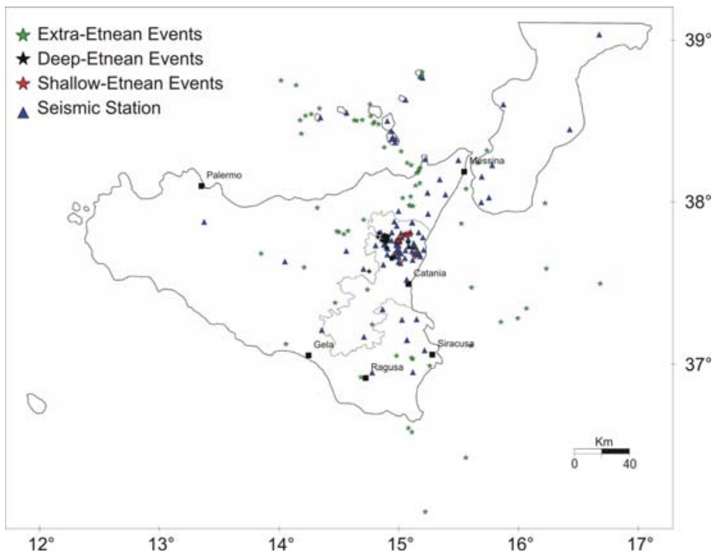

Figure 3. Epicentres of selected earthquakes (stars) and seismic stations of the RSPSO (blue triangles) for the Etnean zone.

3. INPUT PARAMETER SETTING FOR STOCHASTIC PROCEDURES

In the following we separately consider the source, propagation and site parameters for the investigated areas, as summarized in Table 1.

Table 1. Set of parameters for each area. D and R are the epicentral and hypocentral distance respectively.

Parameter	**Areas**				
	Campi Flegrei	**Mt Vesuvius**	**Mt Etna**	**Azores Island**	**South Iceland**
Fault geometrical parameters	Strike = 270° Dip = 85°	Strike = 170° Dip = 85°	Strike slip	Strike = 165° Dip = 85°	Strike=0° Dip= 90°
Magnitude	$M_D = 2.2$	$M_D = 3.6$	$M_w = 3.8$	$M_w = 6.2$	$M_w = 6.5$
Fault Dimension (km^2)	0.4 × 0.4	0.5 × 0.5	1.2 × 1.0	16.5 × 9.4	18.0 × 10.0
Depth of the top (km)	2.0	4.0	1.0	1.1	7
Stress Drop (bar)	20	50	5	200	100
Average S-wave velocity ()	2.0 (from CZ2011)	2.0 (from G2008)	3.0	3.5	3.5
Attenuation parameter Q	Q = 110	Q=150	$Q(f) = 90\ f^{0..5}$	$Q(f) = 239\ f^{1.06}$	$Q(f) = 47\ f^{0.91}$
Geometrical spreading	D < 15 km: 1/R	D < 15 km: 1/R	D=0-35 km: 1/R D=35–70 km: R^0 D > 70 km: $1/R^{0..5}$	R < 30 km: 1/R R >=30 km: $R^{-0.5}$	D <= 30 km: $30^{-1}D^2$ D > 30 km: D
Density (g/cm^3)	2.5	2.2	-	2.8	2.8
Site effects	Amp vs frequency	Amp vs frequency	Not applied	Not applied	Not applied
f_{max} / k	f_{max} = 50 Hz (antialias filter)	f_{max} = 20 Hz (antialias filter)	Not applied	k = 0.03 s	k = 0.027 s
Ground-motion duration	T = 0.9R +1.5	T = 0.9R+1.5	T = 10+0.15D	$T = T_0 + 0.1R$	T = 1.5Rb+121.7D

The Campi Flegrei caldera is an active volcanic area (400 km^2) that is located to the western sector, with respect to the city of Naples. Over the last 12 years, Campi Flegrei has experienced two main seismic crises: the first in 2000 (ground uplift equal to 4 cm), and the second in 2006 (maximum $M_D = 1.4$). Faults and fractures mainly trend NE-SW and NW-SE, with only few around N-S and E-W (Selva et al., 2011). The focal mechanisms evaluated for 2006 VT events show a dominance of normal faults, with stress-drop values from 1 bar to 10 bar. Attenuation analyses (Del Pezzo et al, 1987) have shown an average Q = 110; we take into account the site effects from the study of Tramelli et al. (2010).

Mt Vesuvius is a composite central volcano that is located in the Campanian plain near the suburbs of Naples. The most recent largest event occurred in 1999 ($M_D = 3.6$), with a recorded maximum horizontal acceleration of 0.03 g (Galluzzo et al., 2008). Focal mechanisms show northwest-southeast, northeast-southwest-oriented oblique normal slip faults, and east-west-oriented normal faults (Bianco et al., 1999). The recorded seismicity since 1999 has been characterized by an average stress drop of 10 bar (Galluzzo et al., 2009). For the Vesuvius seismicity, we obtain an average Q for S-waves of 150 (Bianco et al., 1999). To complete the set of parameters adopted for the stochastic simulations, a detailed empirical evaluation of local site effect (amplification vs. frequency) has to be taken into account (Galluzzo et al., 2009), as well as the empirical relationship between the time duration of seismograms and the source–site distance (Galluzzo et al., 2008).

We used values from Giampiccolo et al. (2006) to calibrate the input parameters for Mt. Etna. These authors found a scaling law with a global stress-drop of ca. 5-10 bars for earthquakes with M >3. High variability of the Q-parameters was found for the different seismic stations, recording the earthquake swarms that occurred before the 2001 and 2002-03 eruptions. We extracted the following relationship: $Q = 30\ f^{0.5}$, which should be valid for the whole network and for the dataset that covers 2006-2011.

At a first stage, simulations for the Azores were performed with parameters hat were defined by making assumptions and using information from Zonno et al., 2010. Here we present the simulation of the 1998 Faial earthquake, $M_w = 6.2$, located 8 km NE of Faial island.

Source and path parameters for the South Iceland zone were studied by Sigbjornsson and Olafsson (2004), Halldorsson et al. (2007) and Snaebjornsson et al. (2008) (Table 1). Our analyses were performed using the dataset of the large earthquakes ($M_w = 6.5$) of June 2000, for which an extensive work for near-fault and far-fault simulations was previously performed by Halldorsson et al. (2007), who considered a hybrid approach.

4. RESULTS

At the first stage of the UPStrat-MaFa project, we checked the stochastic procedure by comparing simulated and observed waveforms for each volcanic area at a wide range of magnitude and hypocentral distances.

Below, we show the results starting from the low-magnitude events of the Campi Flegrei and Mt Vesuvius volcanic areas, and moving to the moderate-to-large magnitude ones for Mt. Etna, the Azores Islands and South Iceland. We show the results obtained through the FINSIM code (Beresnev and Atkinson, 1998) for $M_D = 2.2$ earthquakes (hypocentral distance = 2.5 km) that occurred in the Campi Flegrei area in the year 2000. By setting a fault dimension equal to (400 × 400) m^2, a stress drop value of 20 bar, and a $\Delta l = 100$ m, we obtain a value of Amax of 39 cm/s^2 which is close to the 40 cm/s^2 observed in accelerograms at the SLF station (Fig. 4).

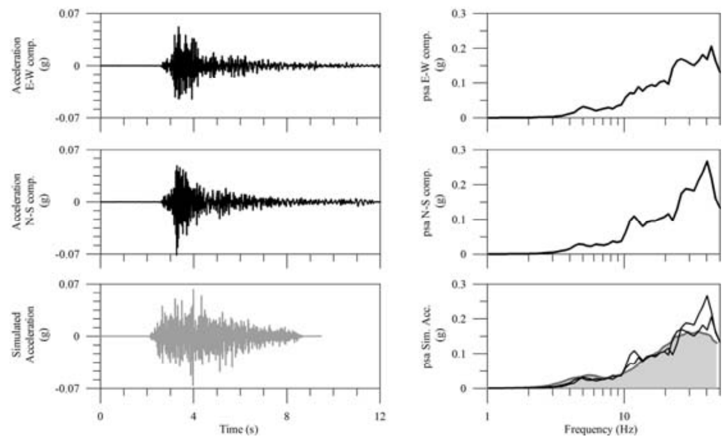

Figure 4. Comparison between observed (black lines) and simulated (grey line) waveforms and response spectra (5% damping) for the $M_D = 2.2$ earthquake that occurred at Campi Flegrei on August 22, 2000, using FINSIM code. Observed waveform were recorded at the SLF station.

The observed and simulated response spectra show similar trends. Moreover, a frequency peak around 4.5 Hz, which is peculiar of the chosen site, is well reproduced, and this is a clear indication of the necessity to use local-site-effect transfer functions to correctly reproduce the ground-motion characteristics.

The same good results were obtained for Mt Vesuvius for an M_D=3.6 event at a hypocentral distance of 4.5 km. The simulated and recorded waveforms at station BKE show good agreement in terms of the maximum acceleration and response spectra (Fig. 5). The differences between the observed and the simulated maximum acceleration are within one standard deviation.

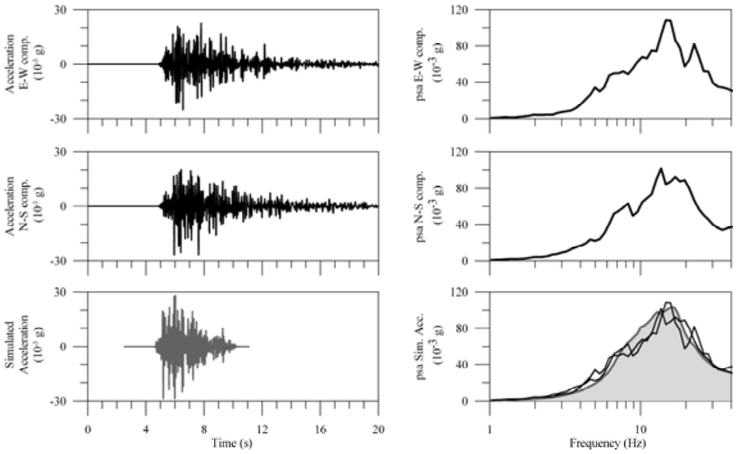

Figure 5. Comparison between observed (black lines) and simulated waveforms and response spectra (5% damping) for an $M_D = 3.6$ earthquake that occurred at Mt. Vesuvius on October 9, 1999, using FINSIM code. The observed seismograms were recorded at BKE station. (Redrawn from Galluzzo et al., 2008)

FINSIM code allows for the choice of the subfault size Δl, which is related to: (1) the total number of subfaults; (2) the amplitude of the summed radiation from each subfault; and (3) the corner frequency of the subfault spectrum (Beresnev and Atkinson, 1998). In both cases shown above (Campi Flegrei and Mt Vesuvius), by applying a trial and error procedure, subfault dimension Δl for the selected seismic events were allowed to vary in the range [100–200 m].

For these previous cases, we checked for the peak ground motion parameters to evaluate the reliability of the stochastic approach. A more complete way to compare observed and synthetic peak ground motion parameters consists of evaluating the magnitude/ distance dependent decay laws for peak ground acceleration (PGA), for peak ground velocity (PGV), and for peak ground displacement (PGD). For the Mt Etna case, we considered the following:

$$LogY = a + bM + cLogR \pm \sigma_{LogY} \tag{1}$$

where Y represents either PGA, PGV or PGD (measured in cm s^{-2}, cm s^{-1}, cm, respectively), M_L is the local magnitude, R is the hypocentral distance (in km), and σ_{LogY} is the standard deviation associated with the logarithm of Y. Our dataset covers both Mt Etna and its adjacent zones, which makes the use of a unified relation questionable. We therefore decided to distinguish between sources falling outside and inside the area of the volcanic edifice. From the inspection of the waveforms of Mt Etna events, we can infer the necessity of a further distinction among Mt Etna sources; i.e., separating superficial events with a depth of 5 km or less from deeper events. We separated the dataset into three different groups on the base of their epicentral location: (i) "*S*uperficial *E*tna *E*vents, SEEs", (ii) "*D*eeper *E*tna *E*vents, DEEs", and (iii) "*E*xtra *E*tna *E*vents, EEEs". The regression coefficients obtained and their standard deviations are provided in Table 2. The PGA, PGV and PGD models for the three groups of events are characterized by a geometrical attenuation coefficient *c* higher than unity, which reflects the low magnitude values of the events included in the available dataset. This should suggest that small events characterized by high frequency values and low energy contents are more affected by the attenuation phenomena than larger events, where lower frequencies prevail. Indeed, the values of the magnitude coefficient *b* are higher than those by Sabetta and Pugliese (1997), and most of the worldwide GMPEs obtained using events with magnitudes higher than about 5, for which *b* is generally lower than 0.5. Vice versa, our *b* values reflect a stronger dependence of PGA, PGV and PGD on magnitude, which is similar to the relations derived considering M_L up to 6.0 (e.g., Frisenda *et al.*, 2005; Massa *et al.*, 2007; Emolo *et al.*, 2011).

For a given magnitude, the predicted PGAs for the SEE group are lower than those predicted for the DEE and EEE groups (see Fig. 6), which suggests a lower high-frequency energy content of the SEEs with respect to the DEEs and EEEs. The differences slightly decrease if we compare the predicted PGVs and PGDs for the three groups of events, even if the higher magnitude coefficient *b* of the SEEs again causes increasing differences in the PGDs with magnitudes, above all, at short distances.

Table 2. Regression coefficients and associated standard errors for PGA, PGV and PGD for the three groups of events considered in this study. Uncertainties (□) of coefficients refer to a confidence interval of 95%. SEEs, shallow Etnean events; DEEs, deep Etnean events; and EEEs extra Etnean events.

Group	Y	a	b	c	σ_{LogY}	Determination coefficient (r^2)
SEE	PGA (cm s^{-2})	-0.805±0.213	0.817±0.059	-1.989±0.039	0.372	0.91
	PGV (cm s^{-1})	-2.898±0.212	1.060±0.059	-1.829±0.038	0.365	0.92
	PGD (cm)	-4.580±0.223	1.279±0.063	-1.668±0.040	0.382	0.90
DEE	PGA (cm s^{-2})	-0.303±0.144	0.846±0.035	-1.999±0.055	0.372	0.80
	PGV (cm s^{-1})	-2.072±0.127	0.883±0.031	-1.812±0.049	0.328	0.85
	PGD (cm)	-3.472±0.132	0.862±0.032	-1.524±0.051	0.345	0.80
EEE	PGA (cm s^{-2})	-0.293±0.177	0.809±0.039	-1.835±0.060	0.410	0.72
	PGV (cm s^{-1})	-2.322±0.150	0.946±0.030	-1.704±0.050	0.347	0.80
	PGD (cm)	-4.061±0.159	1.003±0.035	-1.402±0.052	0.369	0.75

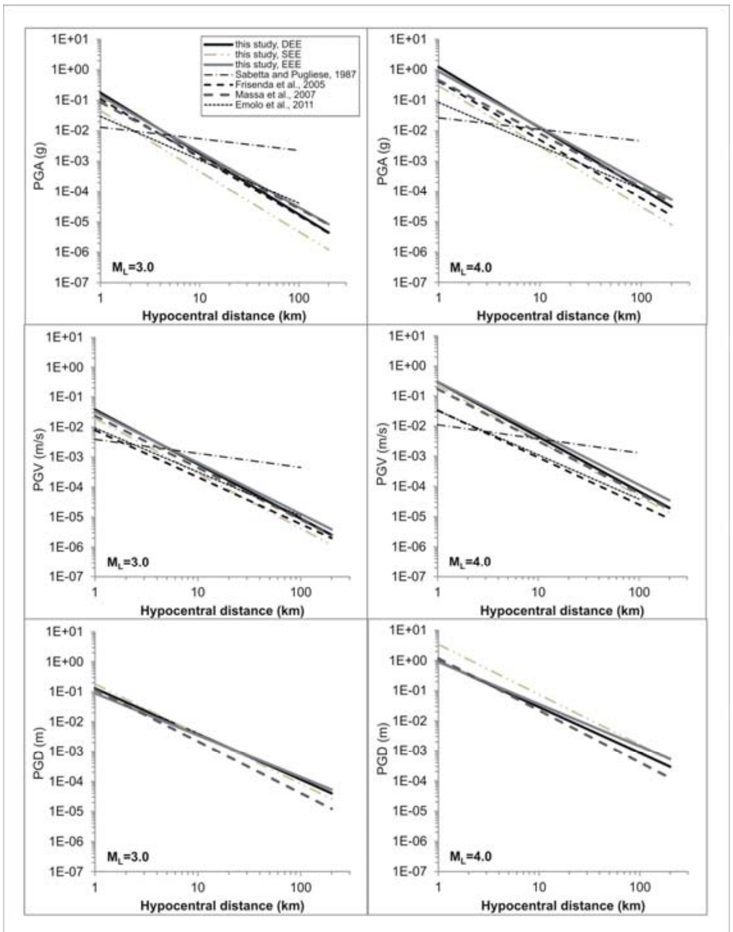

Figure 6. Comparison between the GMPEs obtained in our study with the SP87, FRI05, MAS07 and EMO11

The comparison of our GMPEs with those proposed by Sabetta and Pugliese (1987) (the earliest GMPEs in Italy; hereinafter referred to as SP87), Frisenda *et al*. (2005) (hereinafter, FR05), Massa *et al*. (2007) (hereinafter, MAS07) and Emolo *et al*. (2011) (hereinafter, EMO11) suggests that the SP87 model is not adequate to estimate ground shake parameters in the investigated area.

This underestimates/ overestimates the values of both PGA and PGV at short/ long distances, respectively, and predicts a weaker attenuation with distance. Conversely, for the propagation term, the attenuation models obtained appear to be in better agreement with other relations provided by using peak data from small earthquakes. However, there are still significant differences between our trends and those of MAS07, FR05 and EMO11, and these could be attributed to a difference in the tectonics of the region where the data have been collected.

Moreover, we show synthetic simulations based on the stochastic approach proposed by Motazedian and Atkinson (2005) using the EXSIM code. The model parameters used are given in Table 1. Here we show the attempt to calibrate the input parameters for the SEEs, to match the empirical relations obtained for PGA. The essential parameters (stress drop, Q, and shear wave velocities) were deduced using the results of Giampiccolo et al. (2006); and we adjusted the Q values that were overestimated. We also adjusted the durations, exploiting the real data that showed considerable durations even for small events recorded at stations close to the source.

In Figure 7, we show synthetic PGAs ha match the empirical relation with reasonable accuracy. Empirical relations for the DEEs, as well as the EEEs, predict higher PGA for a given magnitude than expected for SEEs. A different scaling law according to a higher stress drop might be a cause. An in-depth study of this problem will be carried out within the UPSTRAT-MAFA project.

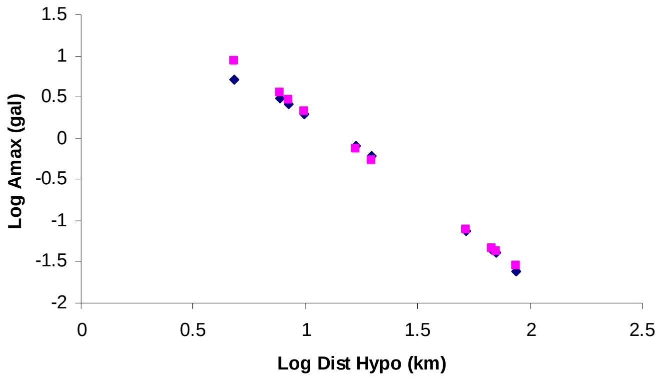

Figure 7. Empirical (pink) and simulated (dark) PGAs. Simulations

Growing with magnitude (M_W=6.2), an extensive use of stochastic simulation was performed through the EXSIM code for the 1998 Faial earthquake (Azores Islands), by Zonno et al. (2010). The results are shown in Figure 8, where recorded and simulated ground motion for the Prince of Monaco Observatory site (Faial Island) are compared. At this station, relatively large ground shaking was recorded (PGA = 390 mg) due to large topographic effects

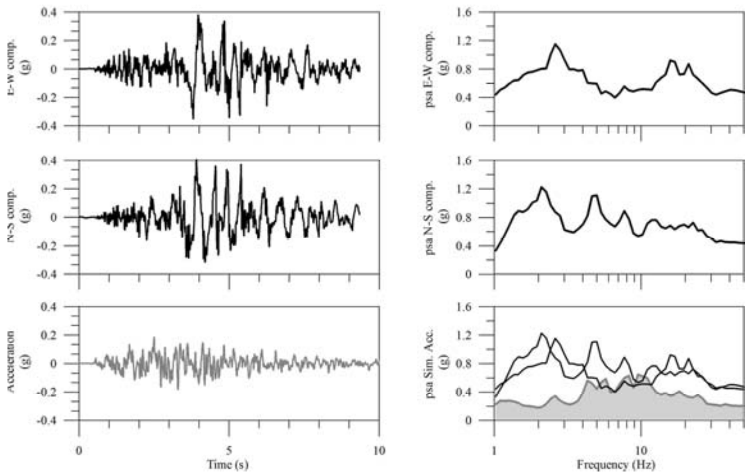

Figure 8. Faial Island M_W 6.2 earthquake. Top two panels (black lines): accelerograms recorded at the Horta site. Bottom panel (grey line): simulated horizontal accelerogram. (Redrawn from Zonno et al., 2010)

(the station is located on the top of a scoria cone).

In the nearby downtown area of Horta, the PGAs estimated from the behaviour of simple structures (Oliveira et al., 2002) was 200 to 250 mg. These values are more consistent with the light damage suffered by the stock of buildings in Horta City, and with simulated ground motion.

A similar application was performed for the M_w 6.5 South Iceland earthquake of June 21, 2000. In this case, we follow a finite-fault simulation approach using FINSIM code, considering the waveforms recorded at Thjorsarbru (free-field station in a structure classified as a rock site [Halldorsson et al., 2007]). It appears clear from Figure 9 that the low frequency near-fault pulse is not well reproduced by the stochastic simulation, and the same occurs for the low frequency part of the response spectra (<10 Hz). As shown by Halldorsson et al. (2007), complete waveform modelling was obtained using a specific barrier model combined with a mathematical model of near-fault velocity pulses.

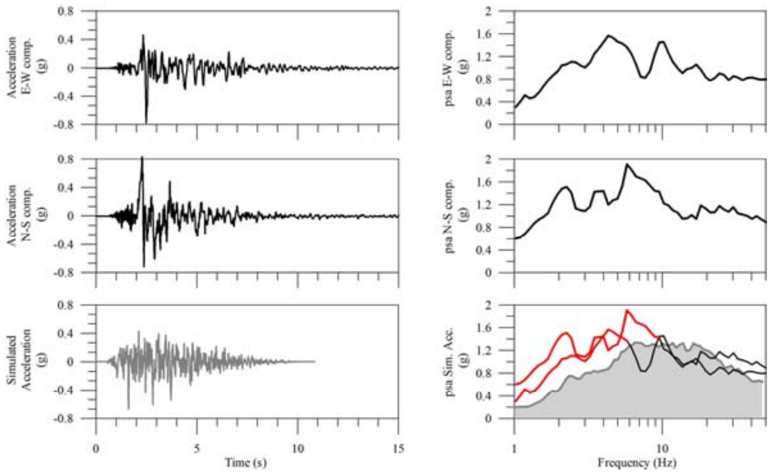

Figure 9. June **21,** 2000, South Iceland M_W 6.5 earthquake. Top two panels (black lines): accelerograms recorded at the Thjorsarbru site. Third panel: simulated accelerogram and response spectra (grey lines) overlapped with observed spectra (black/red lines). The red part of the response spectra is not well reproduced.

5. CONCLUSIONS

The calibration of the input parameters in volcanic areas using instrumental records is an important part of the chain for the procedure proposed by the project. The stochastic finite-fault simulations will be used to: (i) generate the ground-motion shaking through stochastic high-frequency simulation; and (ii) create a synthetic intensity field database that complements the existing data for different intensity scales and considering site-effect corrections. Indeed, in the UPStrat-MaFa project, one of the objectives is to evaluate seismic hazard using macroseismic fields and fault sources.

In this specific case, the preliminary results that calibrate the stochastic finite-fault simulation have shown that:

a) taking into account the peculiar parameterization of the source in terms of the path parameters, local site effects and duration parameters, the FINSIM stochastic approach reproduces the observed waveforms well for low seismic events of Mt. Vesuvius (M_D = 3.6) and Campi Flegrei (M_D = 2.2);

b) the stochastic approach can be considered a useful tool to investigate empirical relationships that link the PGA, PGV and/or PGD to hypocentral/ epicentral distance and magnitude. The results obtained for the Mt Etna area (M_W = 3.8) are promising, in the sense that empirical relations can be reproduced with the EXSIM synthetic calculations. We are aware that the empirical dataset has to be inspected carefully, distinguishing groups of events belonging to different seismotectonic regimes.

In the future, we will test the application of stochastic finite-fault modelling based on a dynamic corner frequency (EXSIM code) to realistically reproduce the low frequency amplitude and the impulsive long-period velocity pulses.

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